Deltas' Responses to Fluvial and Marine Forces

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DELTAS’ RESPONSES TO FLUVIAL AND MARINE FORCES

by

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B.S., Shandong Normal University, 2007

M.S., Nanjing University, 2010

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Department of Geological Sciences

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This thesis entitled: Deltas’ Responses to Fluvial and Marine Forces written by Fei Xing has been approved for the Department of Geological Sciences.

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James P. M. Syvitski

________________________________________
Albert J. Kettner

Date____________________

The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.
Deltas’ Responses to Fluvial and Marine Forces

Thesis directed by Professor James P. M. Syvitski and Dr. Albert J. Kettner

This thesis addresses the responses of river deltas to both fluvial and marine environmental changes. Fluvial sediment flux, the predominant factor providing sediment to deltaic systems, can vary significantly over time, affecting delta formation and evolution. Hurricanes and cold fronts, which cause significant changes in coastal hydrodynamics, also greatly influence delta morphology. This thesis incorporates two areas of interest: a) the Ebro River in Spain, for which numerical simulations of long-term fluvial sediment fluxes were conducted using the hydrological transport model HydroTrend, and b) the Wax Lake Delta (WLD) in Louisiana, for which the morphological changes were investigated under different magnitudes of coastal storms by applying the numerical model Delft3D.

The 4,000-years history of fluvial sediment flux for the Ebro delta was reconstructed based on climate changes and anthropogenic factors. The results demonstrate that long-term fluvial water discharge to the river mouth is controlled by changes in precipitation, which have a high annual variability but no long-term trend. Modeled suspended sediment load, however, has an increasing trend over time, which is closely related to anthropogenic land cover variations. The recent significant decrease in sediment flux (more than 99%) is also attributed to human influences (dam emplacements).

Hurricanes and cold fronts are the two major weather systems influencing the morphology of coastal wetlands along the Gulf of Mexico. Here, simulations show that Hurricane Rita, which made landfall 120 km to the west of WLD as a Category 3 hurricane in 2005, had a significant impact on the delta, with more than 500,000 m³ of bottom sediment eroded, while the simulated largest cold front in the 2008-2009 season (mean wind speed of 11.4 m/s) caused net erosion of 100,000 m³. However, the cumulative impact of cold fronts in 2009 caused 1,900,000 m³ of erosion on the WLD during 29 events, much higher than the erosion caused by the two hurricanes in 2008 (Gustav and Ike caused erosion of 500,000 m³). Winds and waves, hurricane tracks, and aboveground vegetation are the major driving
forces of the morphological changes, while fluvial input and roots have a minor influence on delta morphology during hurricane events.
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Figure A.24 Significant wave height in the WLD domain when cold front event 11 caused water level to rise (data available at wavm_WLD_CF.dat). Wave height was larger in deep areas, so high waves were distributed in the surrounding areas of the WLD and in channels. The highest wave height was 0.6 m, much smaller than the wave height during the Hurricane Rita simulation (1.6 m, Fig. A.20).

Figure A.25 Water level rise in the WLD domain when cold front event 11 caused water level to rise (data available at trim_WLD_CF.dat). The highest water level was 1.5m, much smaller than that during Hurricane Rita (2.5 m).

Figure A.26 Total sediment transport in the WLD domain when cold front event 11 caused water level to rise (data available at trim_WLD_CF.dat). The total sediment transport was higher in channels than shallow areas, as the hurricane case. The magnitude was more than ten times smaller than that during the Hurricane Rita simulation (300 kg s⁻¹ m⁻¹, Fig. A.19).
1 INTRODUCTION

1.1 OVERVIEW

The word “delta” was initially used around 490 BC by the Greek philosopher Herodotus to describe the triangular-shaped geometry of the fluvial and coastal plain at the mouth of Nile River. Charles Lyell was the first to define the modern geological meaning of “delta” in his Principles of Geology (Lyell, 1830): “an alluvial land, formed by a river at its mouth, without reference to its precise shape.” Deltas are among the most economically important coastal environments and are ideal residences for humans due to their rich oil, gas and groundwater resources, high agricultural productivity, and easy means of transportation via abundant waterways (Doust and Omatsola, 1989). Over 300 million people live on 40 major world deltas, and the number is predicted to continue increasing (Ericson et al., 2006; Timmerman and White, 1997). For instance, the Yangtze Delta supports 42.1 million people, and the Nile Delta has a population number of 47.8 million (Ericson et al., 2006; Stanley and Wame, 1998). Deltas also have significant ecological value, with delta wetlands supporting diverse species that influence global biochemical circulation, local temperature, local precipitation patterns, water flow, soil retention and formation, nutrient cycling, and pollution filtering (Costanza et al., 1997; McKee et al., 2004).

A delta is a deposit of sediment formed under the interactions of terrestrial and marine forces. Nowadays, deltas are “at risk” due to decreased sediment flux from the hinterland, intensified erosion, accelerated delta subsidence, and more frequent occurrence of coastal storms under the background of global sea level rise (Syvitski, 2008). A study of 40 deltas covering all continents except Antarctica shows that 70% of deltas are experiencing effective sea level rise as a result of decreased fluvial sediment flux due to dam trapping and agricultural irrigation activities, 20% are undergoing accelerated subsidence caused by mining activities, and 12% are in peril, threatened by eustatic sea level rise (Ericson et al.,

1
Syvitski et al. (2009) classified 33 of the world’s major deltas into 5 groups: not at risk, at risk, at greater risk, in peril, and in greater peril. According to this study, only 6 of the 33 deltas are not at risk, while a number of highly populated deltas, such as the Yangtze, Yellow, and Nile deltas, are in greater peril.

1.2 DELTA SURFACE ELEVATION

Delta surface elevation increases when sediment input is larger than output and subsidence. The delta cycles of the Mississippi River Delta System serve as a great example of this process. The Mississippi River Delta system is composed of 6 sub-delta complexes, each formed as a delta-building event. Sub-deltas are formed when significant amount of fluvial sediment starts to go through the new sub-delta branches, leading to lake filling and bayhead delta aggradation. During this process, delta surface elevation increases as the aggradation rate caused by high fluvial sediment supply outweighs the degradation rate induced by erosional forces (e.g., waves, winds, tides). An example that shows this progradation phase is the Atchafalaya Delta. Here, sub-deltas become abandoned when the increased delta elevation and continuously rising flow routes become unfavorable for runoff, leading to a decrease in fluvial sediment supply and a delta evolution controlled by erosional forces.

Subsidence occurs on deltas when the decreased sediment supply cannot balance the increased degradation rate from sediment compaction and erosional forces. The bird’s-foot Mississippi River Delta system is a good example of this stage (Roberts, 1997). During the Anthropocene, human activities intensify variations in delta surface elevation by, for example, decreasing sediment loads to deltaic systems, accelerating sediment compaction, and indirectly inducing global climate change, which affects sea level and the intensities of marine forces (Syvitski, 2008; Ericson et al., 2006). In tectonically active areas, land motion is another source of variations in delta surface elevation (Soter, 1999).

Syvitski et al. (2009) proposed a quantitative way to calculate delta surface elevation relative to mean sea level \( \Delta_{RSL} \) [L], Equation 1):
\[ \Delta_{RSL} = A - \Delta E - C_N - C_A \pm M, \]  

where \( \Delta_{RSL} \) [L] is positive when the delta is aggrading relative to mean sea level. \( A \) [L] is delta aggradation, which depends on fluvial sediment supply (Goodbred and Kuehl, 2000), marine processes (winds, waves, currents) and delta properties (such as grain size and vegetation cover), in which the latter two factors influence how much sediment can be retained on deltas (Dallas and Barnard, 2009; McLaren and Bowles, 1985; Scruton, 1960; Yang et al., 2008). \( \Delta E \) [L] is the eustatic sea-level variation, which follows changes in global ocean volume determined by terrestrial water storage and ocean water temperature (Meier et al., 2007; Peltier, 2002). \( C_N \) [L] is the compaction of sediment due to natural processes, such as pore water expulsion, grain-packing realignment and organic matter oxidation (Törnqvist et al., 2008; van Asselen et al., 2009). \( C_A \) [L] is the accelerated compaction of sediment on deltas due to anthropogenic activities, such as subsurface mining (e.g. oil, gas or groundwater), soil drainage and human-induced oxidation by lowering the groundwater table. \( C_A \) is becoming critical in modern times and can exceed natural compaction by an order of magnitude or more (Syvitski et al., 2009; Shi et al., 2012; Higgins et al, 2013; Erban et al., 2014). \( M \) [L] is the local vertical movement of land due to changes in earth mass distribution, which might be caused by tectonics, variations in delta sediment volume, local sea level, and local glacier volume (Belknap, 1987; Goodbred and Kuehl, 2000; Ivins et al., 2007).

Most of the world’s deltas began to form when global sea level stabilized within a few meters of its present level, approximately 6,500-8,000 years ago (Amorosi and Milli, 2001; Stanley and Warne, 1994). Eustatic sea level has risen significantly since the 19th century, which is observed in tide gauge records (Bindoff et al., 2007). The global mean sea level was rising at a rate of 1.7 (1.5 – 1.9) mm yr\(^{-1}\) between 1901 and 2001, but accelerated to 3.2 mm yr\(^{-1}\) between 1993 and 2010 (Church et al., 2013). Global mean sea level is predicted to rise an additional 0.21-0.71 m by 2070 and 0.26-0.98 m by the year 2100 (Church et al., 2013). Sea level rise will significantly increase the vulnerability of deltas. Digital elevation models demonstrate that 18.1% of the Nile Delta lies below mean sea level, and a 1 m rise of
local sea level would cause inundation of 30.8% of the Nile Delta, while a 2 m increase of local sea level would flood 43.9% of the delta, causing loss of residence for more than 8 million people (Hereher, 2010). Sea level rise will also cause several coastal regions to be exposed to coastal flooding and storm surges, such as the coastal areas of southeast Asia and Africa, which are characterized by dense populations, low elevations, appreciable rates of subsidence, and/or inadequate adaptive capacity (Nicholls and Cazenave, 2010).

Deltas are composed of loose fluvial sand, silt, mud and peat deposits, which are highly compressible under pressure from new deposits of sediment. The natural compaction rate of deltas is typically reported to be less than 3 mm yr$^{-1}$ (Syvitski, 2008), but the rate varies significantly for different fluvial materials. Sand is the least and peat is the most compressible material among fluvial sediment. Peat compaction was responsible for a 5 mm yr$^{-1}$ subsidence rate of Holocene sediment in the Mississippi River Delta (Tornqvist et al., 2008). Forty percent of the subsidence observed at the Rhine-Meuse delta was attributed to peat compaction during the Holocene (van Asselen, 2011), and the compaction rate more than doubled the basin subsidence rate (0.62 mm yr$^{-1}$ over 4,000 – 6,000 years compared to 0.1 – 0.3 mm yr$^{-1}$). Sediment compaction rates on deltas have significantly accelerated due to modern anthropogenic activities. For instance, parts of the Po Delta subsided 3.7 m during the last century, of which 81% was due to methane mining (Caputo et al., 1970), and the Chao Phraya Delta subsided 50 – 150 mm/yr in the 1980s due to groundwater withdrawal (Saito et al., 2007). Some parts of the Mississippi River Delta experienced subsidence rates between 6 – 7 mm yr$^{-1}$ and 11 – 12 mm yr$^{-1}$ between the 1960s and the 1990s in response to hydrocarbon extraction (Morton and Bernier, 2010).

Tectonic land motion is not universal, but it can be of significant importance to some deltas. For instance, a high subsidence rate of 16.9 mm yr$^{-1}$ recorded for part of the Mississippi River Delta from 1969 to 1971 was associated with motion of the Michoud Fault (Dokka, 2006). The large earthquake that happened in 1762 at the Ganges-Brahmaputra Delta (GBD)-Burma Arc collision zone caused rupture of 250 km of the southern part of the GBD (Steckler et al., 2008). The Helike Delta on the southwestern coast of the Gulf of Corinth subsided by at least 3 m during a major earthquake in 373 BC (Soter, 1999).
Isostatic adjustment caused by changes in overlying material also widely affects deltas, where the significant amount of additional water and sediment added during flood seasons can increase the weight on a delta plain and thereby accelerate subsidence (Ivins et al., 2007; Steckler et al., 2008). Ice sheet melting and ocean mantle movements influence delta elevation as well by modifying the distribution of earth mass (Syvitski et al., 2009).

Delta aggradation is the major process that increases delta surface elevation. Modern deltas have experienced fast progradation in the last several thousand years due to high fluvial sediment discharge under stabilized sea level (for example, Po delta: Trincardi et al., 2004; Nile Delta: Coutellier and Stanley, 1987; Yangtze Delta: Li et al., 2002; Mississippi River Delta: Saucier, 1994). Humans play a significant role in increasing sediment loads by changing land cover from natural vegetation (forest, grasslands) to agricultural lands, or through mining activities, which dramatically increase soil erosion (Saunders and Young (1983); Syvitski and Kettner, 2011). For instance, the sediment load of the Ebro river basin increased by 56% due to human related land use changes in the last 4,000 years before the advent of hinterland dams (Xing et al., 2014). Recent human activities have significantly reduced the amount of sediment that was previously transported to the ocean as well by the emplacement of dams and flood control structures (Syvitski et al., 2005a, 2005b). Syvitski et al. (2005a) reported a 1.4±0.3 GT yr\(^{-1}\) decrease in sediment flux compare to a more pristine period, while Saito et al. (2007) stated that the sediment flux has decreased by 1.5 GT yr\(^{-1}\) for Asian rivers during the last century.

1.3 AGGRADATION AND EROSION

Aggradation is the increase in delta surface elevation due to sediment deposition (+), while erosion is the loss in delta surface elevation due to sediment removal (−). Aggradation and erosion rates vary widely between different delta systems and could have significant temporal and spatial variations in one delta system (Day et al., 2007). Typical aggradation rates for deltas are millimeters to centimeters per year (Syvitski et al., 2003), but major storms and river floods can lead to aggradation rates of 50 mm yr\(^{-1}\)
or more (Yellow River Delta: Syvitski et al., 2009; Ganges-Brahmaputra Delta: Rogers et al., 2013). A significant decrease in aggradation rates for deltas during modern times has caused 27 world major deltas to experience net subsidence (Ericson et al., 2006).

The aggradation rate \( \dot{A} \) in Equation 1 is usually calculated with the Exner Equation (Exner 1920, 1925):

\[
\frac{\partial \eta}{\partial t} = \frac{1}{\varepsilon_0 (1-\phi)} \nabla \cdot q_s
\]  

(1.2)

where \( \eta \) is bed elevation [L], \( t \) is time [T], \( \varepsilon_0 \) is grain density [M L\(^{-3}\)], \( \phi \) is porosity [-], and \( q_s \) is sediment flux [M T\(^{-1}\)]. Sediment flux is the sum of deposition (+) and erosion (-), which is represented as:

\[
q_s = D - E,
\]  

(1.3)

where \( D \) is deposition [L T\(^{-1}\)] and \( E \) is erosion [L T\(^{-1}\)]. Deposited delta sediments come from two sources: contributing fluvial systems and marine sediments transported by winds, waves, tides, and coastal currents (Frihy et al., 1991; Maldonado, 1975; Syvitski et al., 2009; Xing et al., 2012). Fluvial and marine forces also cause erosion in delta systems. For instance, a river flood event may erode sediment from channels and transport it downstream to the delta front (Roberts, 1998; Dumars, 2002). Marine processes such as coastal storms are able to erode sediments from the delta plain and redistribute them to deeper areas (Barras, 2006; Turner and Cahoon, 1987; Wiseman Jr et al., 1986). Beyond these factors, delta properties such as shape, grain size, vegetation coverage, and local morphology also influence deposition and erosion patterns on deltas (Kettner and Syvitski, 2009; Orton and Reading, 1993; Syvitski and Saito, 2007).

1.3.1 Fluvial Sediment Flux

Fluvial sediment flux is the most critical source of sediment to delta systems and has experienced significant changes recently (Fig. 1.1). Syvitski and Milliman (2007) proposed a global model (BQART) to calculate long-term suspended sediment loads to the coast through a dimensional analysis of 294 river basins (Milliman and Syvitski, 1992). The analysis indicates that basin geomorphology (area and relief), hydrology (discharge), climate (temperature), geology (lithology and ice cover), and anthropogenic
factors (reservoir trapping and human induced soil erosion by measure of population density and economic wellbeing) are the main factors that control fluvial sediment flux to the ocean.

Basin area and relief have a primary influence on controlling fluvial sediment loads to the ocean. For instance, the Amazon River and the Yangtze River both belong to the high mountainous river group (>3000 m). The Amazon River with a catchment area of $6.1 \times 10^6$ km$^2$ produces a sediment load of 1200 Mt yr$^{-1}$ while Yangtze River with an area of $1.9 \times 10^6$ km$^2$ generates a sediment load of 480 Mt year$^{-1}$ (Milliman and Syvitski, 1992). Mountainous rivers, which have higher relief, produce more sediment than low-elevation basins due to their higher flow energies (Milliman and Syvitski, 1992). As a result, mountainous river basins that have areas of about 10,000 km$^2$ produced averaged sediment yield of 140 – 1700 t km$^{-2}$ yr$^{-1}$, while lowland basins with the similar basin size have an averaged sediment yield of 20 – 60 t km$^{-2}$ yr$^{-1}$ (Milliman and Syvitski, 1992).

The influence of climate on sediment load is complex, with no simple correlations (Walling and Webb, 1983). There is evidence that both extreme wet and dry climates can intensify soil formation processes, supplying more sediment to rivers (Gaillardet et al., 1999; Molnar, 2001; Syvitski and Milliman, 2007). Basin temperature affects sediment loads through several processes: the sediment-forming frost cycle, which influences the breakdown of rocks and soil formation; and the freeze-thaw cycle, which impacts mechanical soil erosion. Temperature also influences fluvial sediment loads indirectly by controlling the storage or release of glacier water and the related latitudinal variations in lapse rate and weather systems (frontal vs. convective, monsoons, and typhoons) (Hallet et al., 1996; Morehead, 2001; Syvitski and Andrews, 1994; Syvitski and Milliman, 2007). For Arctic rivers, a 2°C increase in temperature causes a 22% increase in fluvial sediment flux (Syvitski, 2002).

Glacier erosion contributes significantly to sediment loads in glacier-covered basins, as turbid water generated by ice melt is able to carry large amounts of sediment downstream. A study of fluvial systems of Greenland demonstrates that Greenland contributes to 12% of the global sediment flux with only a land area of 1.5% due to the high erosion rates of glacier basins. High sediment concentrations are
observed at many river mouths, and several rivers reach hyperpycnal flow conditions at the river mouth (Hudson, 2014).

Basin geology significantly influences sediment erodibility, which further impacts basin sediment loads (Gaillardet et al., 1997; Hinderer, 2001; Milliman and Syvitski, 1992; Schaller et al., 2001). The erodibility of poorly consolidated rocks is seven times higher than massive rocks (Pinet and Souriau, 1988), and that of sedimentary rocks is three times higher than crystalline rocks (Schaller et al., 2001). Hicks et al. (1996) state that, for any given rainfall, sediment load increases by an order of magnitude in New Zealand when rivers flow through weak sedimentary rocks as compared to hard metamorphic rocks. In order to better represent the influence of basin geology on sediment flux, Syvitski and Milliman (2007) classified the global basins into six lithology classes (Tab. 1.1).

<table>
<thead>
<tr>
<th>Basin composition</th>
<th>L</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hard, acid plutonic and/or high-grade metamorphic rocks</td>
<td>0.5</td>
</tr>
<tr>
<td>Mixed, mostly hard rocks, sometimes shield</td>
<td>0.75</td>
</tr>
<tr>
<td>Volcanic, mostly basaltic rocks, or carbonate</td>
<td>1</td>
</tr>
<tr>
<td>Softer rocks, with a significant area of hard rocks</td>
<td>1.5</td>
</tr>
<tr>
<td>Sedimentary rocks, unconsolidated sedimentary cover, or</td>
<td>2</td>
</tr>
<tr>
<td>alluvial</td>
<td></td>
</tr>
<tr>
<td>Weak material</td>
<td>3</td>
</tr>
</tbody>
</table>

There is no doubt that human activities affect the sediment flux towards the ocean (Syvitski et al., 2005a; Walling, 2006). Syvitski et al. (2005a) reported that humans simultaneously increase sediment flux to the ocean through soil erosion activities and decrease the flux by reservoir emplacements, leading to a net decrease of sediment flux by ~1.4 GT yr⁻¹ compare to pre-human loads. The World Commission on Dams (WCD, 2000) reported that there are 48,000 large dams (>15 m) in the world, which have an average height of 31 m and an average reservoir area of 23 km². There are approximately 75,000 dams in the continental United States, most of which were constructed since the 1940s, and the total amount of water stored almost equaling one year’s runoff (Graf, 1999; Syvitski and Milliman, 2007). Both the Mississippi and Yangtze basins have more than 50,000 dams of various sizes (Vörösmarty et al., 2003; Xu et al., 2006; Yang et al., 2006). Globally, reservoirs have trapped ~26% of the total amount of
sediment that would otherwise have been delivered to the coastal ocean (Syvitski et al., 2005a). The Aswan Dam located in the main stream of the Nile River reduced the Nile’s sediment loads to the Mediterranean Sea from 110 Mt yr\(^{-1}\) to almost zero (Walling and Fang, 2003), and the sediment discharge of the Colorado River that formerly entered the Gulf of Mexico decreased from 138 – 165 Mt yr\(^{-1}\) to 0.11 Mt yr\(^{-1}\) since 1930 (Meade and Parker, 1985). The world’s largest dam, the Three Gorges Dam, has decreased the sediment flux of the Yangtze River by 85 Mt yr\(^{-1}\) (31% of the annual sediment discharge; Yang et al., 2006). Before the booming dam construction of the 20\(^{th}\) century, the major impact of human activities on sediment flux was to increase sediment flux by intensifying soil erosion through land clearance for agriculture and other facets of land disturbance, such as mining (Willing, 2006). By doing so, more sediment was eroded from hillsides, which then could be transported through river networks towards the ocean (Alatorre et al., 2010; Syvitski and Kettner, 2011). In a review of 270 publications, Saunders and Young (1983) stated that moderate land use is able to increase sediment yield by a factor of 2 – 3, while intensive land use increases it by an order of magnitude. The global sediment discharge to the ocean has doubled since 500 BC due to agricultural activities (Milliman and Syvitski, 1992; Syvitski, 2008). Other human activities, such as flood control structures and coastal engineering projects, also influence sediment flux to the ocean. For instance, channel stabilization and flood-wave mitigation have decreased sediment dispersion across the Yellow River Delta from 82% of the annual load to almost nothing (Syvitski et al., 2005b; Syvitski and Saito, 2007). The fluvial sediment that is delivered to the Po delta is not allowed to disperse onto the surrounding flood plains, but stays within the distributary channels (Syvitski et al., 2005a). A coastal engineering project on the Mississippi River Delta system created a new outlet, which transported 7.5% of water and sediment from the Mississippi River and formed the Wax Lake Delta (Dumars, 2002). The implementation of soil conservation and sediment control measures and sand mining activities also change sediment loads to deltas (Walling, 2006). Controls on fluvial fluxes in an entire basin could counteract with each other. For instance, the increase in sediment load caused by deforestation could be compensated by a reduction in sediment loads for other tributaries caused by activities such as dam emplacements (Lu and Higgitt, 1998).
Fluvial sediment fluxes towards the ocean are determined by the interactions of all these factors. The functions of these factors in space and time are distinct to each river system. The anthropogenic factor is the most critical factor that has significantly changed the amount of sediment flux to deltas since human civilization for those highly populated river basins (Yellow River and Yangtze River: Saito et al., 2001; Nile River: Milliman and Ren, 1995; Ebro River: Xing et al., 2014), while climate changes might be more critical for rivers when studying fluvial fluxes for larger time scales. It is therefore critical to study temporal variations in these controlling factors, in order to better understand how deltas respond to changes in fluvial fluxes.

Figure 1.1 Comparison between Pre-Anthropocene and modern sediment loads, summarized from 217 global rivers with sediment flux data before and after dam emplacements

Note: The 1:1 line represents no human influences. The two lines on the left show the increase of sediment flux due to human activities (mainly deforestation), and the lines on the right show the decrease of sediment flux due to reservoir trapping (Reproduced from Syvitski et al., 2005a, copyright 2005 American Association for the Advancement of Science).

1.3.2 OCEANOGRAPHIC PROCESSES

The major marine forces that might cause changes in delta elevation include winds, waves, tides and coastal currents. Strong winds and waves during coastal storms can cause significant water
movement, leading to sediment transport (Allison et al., 2005; Cahoon et al., 1995; Dufois et al., 2014; Ulses et al., 2008). The roles of these forces can be considerably different in distinct delta systems. For instance, tides are the predominant force in the Ganges-Brahmaputra Delta, forming sand ridges, sand bars and intertidal channels with medium to high tidal flats. Waves are the dominant force in the Nile and Niger delta systems, resulting in regular shorelines with protrusions and beach ridges due to the sudden increase in sediment deposition at the coast when fluvial energy rapidly decreases and fresh and salt water mix through wave breaking (Allen, 1965). The Orinoco River Delta, which is located on the northeastern coast of South America, is exposed to the Equatorial Current and Northeast Trade Winds, which reshape the delta and form relatively straight shorelines in the area of the main distributaries, compared to adjacent irregular coastlines (Scruton, 1960). Most deltas are influenced by different configurations of these forces (Bhattacharya and Giosan, 2003; Giosan et al., 2006). For instance, for the Danube delta system, waves are the dominant factor at the St. George Lobe, but the influence of the river is more pronounced for the Chilia Lobe (Giosan et al., 2006).

Coastal storms, although occurring infrequently, are able to produce significant sediment transport in a short time and cause significant morphological changes on deltas (Barras, 2006; Fan et al., 2006; Turner et al., 2006; Yang et al., 2003). Coastal storms can be caused by frontal systems or cyclones. The former type, which is also called a cold front, is the interface or transition zone between cold, dry airflows and warmer, lighter air (Hsu, 1988). When there is sufficient moisture, the rising air of a cold front system would thus condense and form storms. Cold fronts occur frequently in subtropical areas during transition seasons in fall and spring. Cyclones typically occur in summer and fall with higher intensities. The strong tropical cyclones, which are named as hurricanes (Atlantic Ocean) or typhoons (Pacific Ocean), are among the greatest hazards in coastal areas, imperiling coastal wetlands, properties and human lives (Huang et al., 2001; Li and Ellingwood, 2006; Pielke et al., 2008).

Cold front passages cause significant changes in temperature, pressure and humidity (Roberts et al., 1987). According to these changes, the passage of a cold front is divided into three stages: prefrontal, frontal passage, and postfrontal stages. The prefrontal stage is characterized by warm temperatures and
steadily decreasing pressures. During frontal passage, temperature drops suddenly, and pressure decreases to its lowest values, usually accompanied by strong rainfall or thunderstorms. The postfrontal stage starts with increasing pressure, falling temperatures, and easterly winds (Fig. 1.2, Moeller et al., 1993; Roberts et al., 1987). The significant changes in weather factors can cause water levels setting up and down, leading to coastal flooding and sediment transport.

![Wind field of a cold front system](image)

*Figure 1.2 Wind field of a cold front system. Reproduced from Roberts et al. (1987). Copyright 1987 American Society of Civil Engineers*

Cold fronts play a significant role in transporting sediment from the coast towards offshore. A study of the Rhone delta showed that a strong winter storm transported 2.1 Mt of sediment from the Rhone prodelta (Dufois et al., 2014), and Ulses et al. (2008) reported that marine storms play a crucial role in sediment dispersal on the shelf. Walker and Hammack (2000) argued that cold fronts cause rapid flushing of sediment from the shallow Atchafalaya Bay that reduces the progradation rates of the Atchafalaya and Wax Lake delta system (Fig. 1.3). The Yellow River subaqueous delta erodes in winter and accumulates in summer due to the powerful effect of cold front events (Yang et al., 2011). Cold fronts also cause deposition in some coastal areas. Kineke et al. (2006) showed that frequent cold front
passages (20 – 30 times per year) along the coast areas of Louisiana are responsible for high progradation rates (up to 29 mm yr\(^{-1}\)) at Chenier plain coast because sediment is transported towards onshore during both the pre- and post-frontal stages. Mossa and Roberts (1990) demonstrated that cold fronts cause either erosion or deposition depending on sediment supply and grain size (Roberts et al. 1987, 1989). As all of the former studies are based on field observations and satellite image analyses, they cannot provide a clear figure of the underlying sediment transport mechanisms and how individual factors, such as winds, waves, and vegetation, influence morphological changes of coastal areas during cold front events.

*Figure 1.3 Suspended sediment expansion during strong cold front events at the Atchafalaya Bay. a)*
Tropical cyclones are developed over tropical or subtropical waters, which have warm core, non-frontal low-pressure systems of synoptic scale and a definite organized surface circulation (Holweg, 2000). Depending on wind intensity, tropical cyclones in the Atlantic Ocean are classified into tropical depressions, tropical storms, and hurricanes (Tab. 1.2). Strong winds, high waves, powerful storm surges, and accompanying intense precipitation during tropical cyclones pose threats to coastal lands, properties, infrastructures, and human lives. During the last two centuries, around 1.9 million people died due to tropical cyclones, and each year around 10,000 people perish because of tropical cyclones (Adler, 2005). Cyclone Bhola (1970) that hit the densely populated Ganges Delta region of Bangladesh killed 500,000 people (Bolonkin, 2007). The western part of the North Pacific is the most active and intensive tropical cyclone region (Fig. 1.4, 1.5), and the North Atlantic and Gulf of Mexico are also greatly influenced by tropical cyclones (Woodruff et al., 2013). In general, the southern hemisphere has fewer and weaker tropical cyclones than the northern hemisphere because of comparatively low ocean temperature and greater wind speeds that prevent cyclones from forming.

**Table 1.2** Winds, pressures and estimated wave heights for different categories of hurricanes. *TD:* Tropical Depression; *TS:* Tropical Storm. From [http://www.nhc.noaa.gov/aboutsshws.php](http://www.nhc.noaa.gov/aboutsshws.php) (2015)

<table>
<thead>
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<th>Type</th>
<th>Category</th>
<th>Pressure (mb)</th>
<th>Winds (knots)</th>
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<td>Depression</td>
<td>TD</td>
<td>---</td>
<td>&lt;34</td>
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<tr>
<td>Tropical Storm</td>
<td>TS</td>
<td>---</td>
<td>34-63</td>
</tr>
<tr>
<td>Hurricane</td>
<td>1</td>
<td>&gt;980</td>
<td>64-82</td>
</tr>
<tr>
<td>Hurricane</td>
<td>2</td>
<td>965-980</td>
<td>83-95</td>
</tr>
<tr>
<td>Hurricane</td>
<td>3</td>
<td>945-965</td>
<td>96-113</td>
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<tr>
<td>Hurricane</td>
<td>4</td>
<td>920-945</td>
<td>114-135</td>
</tr>
<tr>
<td>Hurricane</td>
<td>5</td>
<td>&lt;920</td>
<td>&gt;135</td>
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</tbody>
</table>
Figure 1.4 Tracks and intensities of tropical cyclones between 1851-2006.

Note: Colors show Saffir-Simpson Hurricane Intensity, from NASA: http://www.goesr.gov/education/comet/satmet/remote_sensing/navmenu.php_tab_2_page_1.2.0_type_text.htm (2015)

Figure 1.5 Global distribution of typical tropical cyclone tracks (Reproduced from Abbott (2006) and Copyright 2006 McGraw-Hill Higher Education.)
Tropical cyclones have distinct behaviors for different water depths. In deep water, where water depth is larger than the mixing layer of the hurricane (e.g., 60 m for a storm with a maximum wind speed of 17.2 to 24.4 m s$^{-1}$), the momentum is diffused downward into the deep ocean as the surface wind stress creates a rotating mound, or vortex, of water. As a result, only a small magnitude of water elevation rise occurs at the hurricane eye due to the hydrostatic uplift caused by the low center pressure (Jelesnianski, 1899). The wide and deep ocean is favorable for the formation of significant waves. For instance, Hurricane Katrina (2005) caused wave heights of 16.91 m at buoy station 42040 (National Data Buoy Center http://www.ndbc.noaa.gov/). When cyclones approach shallow coastal areas, water elevation increases significantly and forms high storm surges because the conservation of the potential vorticity of the mound requires development of marked divergence. The water level increase is also strengthened by local bathymetry and coastal refection. Nevertheless, wave heights decrease significantly during this process because waves begin to contact the sea floor when water level is smaller than half the wave height. Waves transform and finally break when the wave steepness ratio reaches 1/7, leading to most of the wave energy being dissipated. The dissipated energy plays an important role in eroding sediment from the seabed. In the northern hemisphere, the areas located on the eastern side of the cyclone center encounter the most severe damages due to the anticlockwise structure of the storm (Coch, 1994). Tropical Storm Isidore (2002) and Hurricane Lili (2002) transported and deposited more than 0.16 Mt muddy sediments onto the inner shelf (Allison et al., 2005). Cahoon et al. (1995) observed that Hurricane Andrew (1992) caused significant deposition on the coastal marshes of Louisiana, and stated that hurricanes are important for the survival of coastal marshes. These studies provide an overview of the general impacts of cyclones on the morphology of coastal areas. Nevertheless, none of them are able to explain the mechanism how hurricanes cause these morphological changes, or how a hurricane would influence the morphology of a specific coastal stretch. The influences of individual factors (winds, waves, vegetation) on delta morphology during hurricanes are still unclear.
1.3.3 DELTA PROPERTIES

Not all of the fluvial sediment that is transported to deltas remains on the deltas. Most of the coarse sediment will be deposited at the river mouth, while the fine sediment will be transported further out by a fluvial surface plume or marine forces (winds, waves, coastal currents). The amount of sediment that can be deposited on a delta is controlled by the receiving basin geometry and the delta properties, such as local bathymetry and vegetation coverage, under the influence of oceanographic processes (Galloway, 1975; Kenyon and Turcotte, 1985; Lap Nguyen et al., 2000; Pulham, 1989).

Delta topography significantly influences the destination of fluvial water and sediment. Deltas with high slopes and short channels are favorable for sediment to escape offshore, while flat deltas with long channels have higher sediment retention rates (Gao, 2007). The geometry of the receiving basin influences delta morphology by modifying the strength of marine forces (Wright et al., 1973). For instance, a flat offshore slope profile is able to attenuate wave energy, while a steep offshore slope profile usually forms strong near-shore waves, which significantly influence deltas (Wright et al., 1973). A slight inclining receiving basin is in favor of accepting sediment, promoting delta aggradation (Lap Nguyen et al., 2000).

Vegetation coverage and properties have a significant impact on sedimentation-erosion patterns of deltas. Vegetation can attenuate flood erosion on coastal wetlands (Fosberg, 1971; Costanza et al., 2008; Gedan et al., 2011) by locally increasing resistance to flow (Cowan, 1956; Abernethy and Rutherfurd, 1998; Hickin, 1984; Chen et al., 2009), reducing the flow velocity between and above plants (Harlin et al., 1982; Hickin, 1984), decreasing turbulence (Temmerman et al., 2005), and influencing sediment distribution and transport. Vegetation can also eliminate bank and floodplain erosion by increasing the soil strength with root networks (Abernethy and Rutherfurd, 1998; Gyssels et al., 2005; Knighton, 1998; Murray and Paola, 2003; Pollen-Bankhead and Simon, 2009; Smith, 1976). As a result, native vegetation has been used as an effective and economical way to protect inlands from coastal storms (Chen et al., 2009).
Other factors, such as sediment grain sizes and the overall amount of fluvial water and sediment, also influence sediment movement on deltas (Kettner and Syvitski, 2009; Orton and Reading, 1993; Syvitski and Saito, 2007). Grain sizes influence the margin property (such as dissipative or reflective), which in turn determine the roles of discharge, waves and tides on delta morphology (Orton and Reading, 1993). For a river-dominated delta system, highly cohesive sediment is favorable to the formation of bird-foot deltas, while less cohesive sediment results in fan-like deltas (Edmonds and Slingerland, 2010). Many deltas show patterns of significant morphological changes during large river floods, which are able to erode sediment from channels and transport it to the shallower subaqueous and subaerial areas of a delta and offshore. For instance, the subaerial part of Wax Lake Delta, which is a new branch of the Mississippi River Delta system, appeared in 1973 during a large river flood (Roberts, et al., 1997).

1.4 OBJECTIVES OF THESIS

1.4.1 HOW DO CHANGES IN CLIMATE AND ANTHROPOGENIC FORCES CONTROL THE FLUVIAL SEDIMENT FLUX TOWARDS DELTAIC SYSTEMS?

This thesis investigates the response of delta surface elevation to fluvial and oceanographic processes. Rivers transport sediment to the coastal zone (an annual global sediment yield of 12.6 BT yr⁻¹, Syvitski et al., 2005a) and form deltas. A delta is an integral part of a large river system and must be considered in that context to be adequately understood. Each component subsystem contributes in varying degrees to the characteristics of a delta. Fluvial water and sediment is the predominant factor supporting delta formation and guaranteeing its sustainability. Deltas adjust their shapes to variations in sediment flux. Ashton et al. (2013) demonstrated that the significant changes in the morphology of the river-wave dominated Ebro Delta system over time might show the delta’s response to the extreme increases in fluvial sediment flux within short time periods that exceed the capacity of longshore sediment transport by waves (Fig. 1.6). Many of world’s deltas are experiencing significant changes in morphology due to a decrease of sediment flux in the modern times (Nile delta: Frihy et al., 1991; Ebro delta: Guillén and Palanques, 1997; and many others). Thus, it is of critical importance to study how deltas respond to
fluvial sediment flux variations. The monitoring of sediment loads to the ocean only occurs for 10-15% of world rivers, most of which are not gauged anymore (Syvitski et al., 2003). So the observed data is not enough to analyze the mechanism that controls delta responses to fluvial sediment flux. Numerical models provide a method to reconstruct fluvial sediment fluxes for long temporal scales. They also serve as a tool to study the impacts of the different forcing factors in controlling sediment flux and how their roles changes over time. For instance, there is no doubt that human activities are significantly controlling sediment flux to the ocean currently (Syvitski et al., 2005b; Saito et al., 2007), whereas humans’ role might have been minor thousands of years ago. Climate changes significantly change sediment flux to the ocean (Langbein and Schumm, 1958; Schumm, 1977; Inman and Jenkins, 1999), but whether these changes are the main factor causing variations in sediment flux to a delta over time is less known. All of these problems can be solved with a precise numerical modeling approach. This study analyzes how individual factors, including climate changes and human activities, temporally control the fluvial sediment flux to delta systems over a long period of time. This in turn provides insight on delta evolution under environmental changes.

*Figure 1.6 Sketch of the Ebro Delta evolution. The white area constrained by the black line shows the current shape of the Ebro Delta. Reproduced from Canicio and Ibanez (1999), copyright 1999 Science Press*
1.4.2 What is the impact of extreme events on the morphology of deltaic systems?

In modern times, deltas are becoming more vulnerable due to a decrease in fluvial sediment and probably more intensified coastal storms (Syvitski, 2008). So, in addition to studying delta responses to fluvial forces, it is also critical to understand how deltas respond to different magnitudes of coastal storms in order to provide valuable information for coastal management and restoration projects. Coastal storms have been well studied, including storm formation, progradation, and resulting hydrodynamic fields in the coastal areas (Cañizares and Irish, 2008; Lintern et al., 2013a; Resio and Westerink, 2008). However, the morphological changes of coastal areas during storms are less well studied. Most related research uses field observations and satellite images to estimate the morphological changes of deltas and coastal wetlands after storms (Fan et al., 2006; Jimenez et al., 1997; Yang et al., 2003b). These studies usually cover general information for a large area. However, coastal storms are argued to bring both erosion and deposition to different parts of coastal areas, which is hard to extract from big maps or few observation spots (Mossa and Roberts, 1990). Numerical models are able to simulate the morphological responses of a specific area for realistic hydrodynamic fields, which provide insight into sediment transport dynamics during coastal storms. Models also enable an evaluation of different factors’ impacts on sediment dynamics during coastal storms. This study analyzes delta’s responses to different magnitudes of coastal storms. The evaluations of individual factors (winds, waves, vegetation and fluvial input) help to shed light on delta dynamics and protecting deltas against storms.

1.5 Model development

Numerical models provide a method to test theories and identify parameters that control physical processes. In this study, hydrological models and fluid dynamic models are applied to study how deltas respond to changes in fluvial and marine environments.

Drainage basin numerical models have been developed to simulate runoff and sediment yields in watersheds, which can be used to better understand fluvial water and sediment transport processes.
Physically based, spatially distributed hydrological models are designed to study basin responses and variations to changing climate and/or human activities by simulating key hydrological and sedimentological processes. These models include the Hydrological Simulation Program-FORTRAN (HSPF, Bicknell et al. 1997), TOPMODEL (Beven and Kirkby, 1979; Beven et al., 1984), Système Hydrologique Européen (SHE, Beven et al., 1980), and the Precipitation-Runoff Modeling System (PRMS, Leavesley et al., 1983). However, these models require a large amount of data for model input and validation and are extremely time-consuming, making them impractical for long-term simulations.

Empirical models use empirical algorithms and state variables to represent the hydrological processes in catchments. Using such models can greatly increase simulation time efficiency, making it possible to reproduce basin hydrology for long time periods with reasonable accuracy, especially in small basins which have comparatively low spatial variability. For example, LOAD ESTimator (LOADEST) develops a regression model for the estimation of constituent load with a time series of stream flow, additional data variables, and constituent concentration (Runkel, et al., 2004). HydroTrend, a climate-driven hydrological transport model, can produce synthetic river runoff and sediment flux for both short (yearly to decadal) and long ($10^5$ years) time periods with high time efficiency using relatively easy to obtain parameters (Syvitski et al., 1998, Kettner and Syvitski, 2008). In this thesis, HydroTrend will be used to study long-term water and sediment discharge for the Ebro River over thousands of years. The model has successfully reproduced long-term fluvial fluxes for several other Mediterranean drainage basins like the Po and Rhone as well as the Têt (Kettner and Syvitski, 2009).

Numerical models have been widely adopted to simulate coastal hydrodynamics during storm events. Tides, storm surges and waves are the major factors causing water and sediment movement during storm events, so models that can successfully predict coastal changes during storms have to take into account all these factors. Tides are very predictable, and all of the ocean tides models are very accurate with errors limits of only 2 – 3 cm in the deep ocean (Shum et al., 1997). However, these models become less accurate when predicting tidal water levels in shallow water regions, where bathymetry, current amplification and nonlinear effects become critical and cannot be omitted (Tierney et al., 2000). In order...
to realistically represent tide propagation in shallow waters, observational techniques and satellite altimeter data have been coupled into these global ocean tide models to achieve better predictions (Savcenko and Bosch, 2007; Tierney et al., 2000).

Storm surges are generated by strong winds moving over shallow water, so surge height is proportional to wind stress divided by water depth and to wind duration (Wolf, 2009). Wind stress is typically determined to be positively correlated to the square of wind-speed, expressed with a drag coefficient, which increases with wind speed (Brown and Wolf, 2009). So high-resolution wind fields and a detailed coastal bathymetry are required to generate realistic surges in the coastal area, especially for tropical storms, which are smaller and more intense than other systems (Emanuel et al., 2008). The appropriate estimation of wind shear stress is also critical for generating accurate surge heights (Williams and Flather, 2000; Brown and Wolf, 2009). For instance, De Vries et al. (1995) reported that five different surge prediction models (IFREMER, MUMM, KNMI, POL, UA/AUT) all underestimate surge height when inappropriately low wind shear stress coefficients were used.

The state of the art wave modeling theory was established by Komen et al. (1994), in which the nonlinear interaction terms were explicitly incorporated in the third-generation Wave Model (WAM) (WAMDI Group, 1988). This theory significantly improved the ability for wave models to simulate accurate waves and storm surges in shallow water regions and led to a wide-application model that can compute both day-to-day average wave conditions and higher temporal resolution hurricane conditions. This theory has become the standard for both global and regional wave model applications, with further development mainly focusing on shallow water physics (WISE Group, 2007).

Tides, waves, and storm surge models need to be coupled together to accurately simulate the hydrodynamics during storm events (Wolf et al., 1988). The nonlinear effects of tide-surge interactions in shallow water can cause a surge peak appearing on the rising tides and advanced tidal phase due to the existence of a positive surge level. This theory has been shown to be important in shallow water areas with large tidal ranges (Prandle and Wolf, 1978; Wolf, 1981; Davies and Lawrence, 1994; Horsburgh and Wilson, 2007; Jones and Davies, 1998). Because wave transformation, propagation, breaking, and energy
dissipation processes are all influenced by water depth, water level variations due to tides and surges will significantly influence wave behavior in shallow water areas.

Current-wave interactions are also critical in shallow water environments (Wolf et al., 1988; Osuna and Wolf, 2005; Ozer et al., 2000; Rosales et al., 2008). Waves influence water levels and currents through wave setup and longshore drift due to wave radiation stress (Longuet-Higgins and Stewart, 1962). Waves also affect surface water roughness and bottom friction, which impact surge generation and current friction in shallow water areas (Brown and Wolf, 2009; Janssen, 1991, 1989; Wolf and Prandle, 1999; Davies and Lawrence, 1994). The coupled tide, surge, and wave models have been shown to produce a better result than computing each factor separately (Ozer et al., 2000). Two-way coupling of wave and surge models has been widely used and shown to be more efficient in applications all over the world (Choi et al., 2003; Osuna and Monbaliu, 2004; Osuna and Wolf, 2005; Xing et al., 2012; Zhang and Li, 1996).

Sediment transport and morphodynamic models are essential in highly dynamic environments like the coastal zone. Morphological models are classified into process-based (deterministic) and behavior-based models. The first type computes all relevant sediment transport processes, while the latter uses simple parameters to represent general behaviors of morphological systems for larger scales or long time spans (Amoudry and Souza, 2011). Deterministic morphological models have been widely used in coastal sediment dynamic studies to investigate the morphological evolution and therefore have been applied for this study.

In order to calculate sediment transport and bed morphological changes, a sediment transport model must be coupled to a hydrodynamic model. Wave and turbulence models are also interlinked in order to achieve reliable simulation results for sediment movement (Fig. 1.6). Both suspended sediment transport and bed evolution are based on the principle of conservation of sediment mass, where the suspended sediment concentration is calculated through the advection-diffusion equation and the change in bed level is calculated through Exner equation (Paola and Voller, 2005). The output of bathymetry change will then be taken into account in the hydrodynamic computations. The sediment dynamic theory
includes boundary layer dynamics (current, current-wave boundary layer calculation for bottom shear stress, bottom roughness), bed load transport, sediment erosion and deposition, sediment diffusivity, and cohesive sediment transport governing equations, all of which have been described in detail by former studies (Delo, 1988; Engelund and Fredsøe, 1976; Garcia and Parker, 1991; Grant and Madsen, 1982; Harris and Wiberg, 2001; Li and Amos, 2001; Neumeier et al., 2008; Ribberink, 1998; Richardson and Zaki, 1954; Sanford, 2008; Smagorinsky et al., 1965; Soulsby, 1995; Soulsby and Clarke, 2005; Soulsby and Damgaard, 2005; van Rijn, 1984; Winterwerp, 2002; Xu and Wright, 1995). Using these theories, numerous sediment models have been applied to describe sediment transport in coastal areas.

Widely used sediment models include: 1) the U.S. Community Sediment Transport Model (CSTM) which is embedded in the Regional Ocean Model System (ROMS, Warner et al., 2008), the Finite Volume Coastal Ocean Model (FVCOM) and Proudman Oceanographic Laboratory Coastal Ocean Modeling System (POLCOMS); 2) the sediment model embedded in Delft3D modeling system (Lesser et al., 2004; Delft Hydraulics, 2007); 3) the Sediment Transport (ECOMSED) model embedded in the Estuarine Coastal Ocean Model (Blumberg, 2002); 4) the sediment transport model (SISPHE) included in TELEMAC (Villaret, 2004); and 5) the sediment model of the MIKE modeling system (DHI, 2007a, 2007b). All these models can be coupled with wave models to calculate wave-induced bed shear stress and wave-current interactions. Although the accuracy of sediment models has improved over the last few decades, some important processes are not accounted for. For instance, the biological effects on sediment dynamics, the impact of mixed beds, and several cohesive sediment physical processes are still poorly understood. Also, the parameterizations of erosion, sedimentation rate, and bed load transport rate are less understood partly due to insufficient validation datasets (Amoudry and Souza, 2011).
There is no doubt that vegetation is of critical importance in determining hydrodynamics and morphodynamics in coastal areas, but the mechanism by which vegetation influences flow structure and sediment movement is a topic of ongoing research. Numerical models have been used as a useful approach to study this process. Vegetation was initially represented in 1D and 2D numerical models by locally increasing the bed roughness coefficient in vegetated areas (Anderson et al., 2006; Beffa and Connell, 2001; Darby and Thorne, 1996; Helmiö, 2002; Leu et al., 2008; Nepf, 1999). This approach accounts for the impact of vegetation on slowing flow velocities but misinterprets its influence on sediment transport. 3D numerical modeling of vegetation has resolved this problem by realistically representing plants in calculations of vertical flow profiles. In a 3D setup, plants are represented as cylinders that add additional drag forces on flow and influence turbulences, which have been verified to produce reliable results in laboratory flumes (Neary, 2003; Su and Lin, 2005). Recently, 3D vegetated models have been applied to simulate sediment dynamics for more realistic landscapes. Wilson et al. (2006) simulated two flood events on a 170 meter-long floodplain with the 3D finite volume program (Sediment Simulation in Intakes with Multiblock Option (SSIIM)) and found that the existence of
willows can significantly change velocity distributions. Temmerman et al. (2005) applied Delft3D to simulate the Paulina Marsh, SW Netherlands, for a single tide cycle, demonstrating that the presence of vegetation cover significantly changes the long-term geomorphology of tidal marshes. Baptist et al. (2005) used Delft3D to simulate vegetation-induced roughness on a floodplain by changing bed roughness and flow resistance, and applied this approach to a flood event of 858 m$^3$ s$^{-1}$, for the Allier River, France, in May 2001. Results indicate that the model performance can be significantly improved by taking into account vegetation. This approach has been used by Facchini et al. (2009) and Arboleda et al. (2010), to study the effects of different types of vegetation on water flow, sediment transport and floodplain morphological changes, producing reliable results. All of these applications verify the importance of vegetation on flow pattern and related sedimentary systems and provide insights into the influence of vegetation on sedimentation and erosion.

In this study, the hydrodynamic model system Delft3D, in which the Flow module takes into account morphology and vegetation, is coupled with the wave model called Simulating Waves Nearshore (SWAN). The Delft3D model system has been used successfully to simulate coastal storms focusing on hydrodynamics and flood inundation (Cañizares and Irish, 2008; Thuy and Tien, 2005; Lintern et al., 2013b; Voukouvalas, 2010). However, morphodynamic studies, specifically in vegetated wetlands, are still rare. In this thesis, we focus on the morphological changes of a river-dominated delta system during different types of coastal storms, which will provide more insights into coastal morphological processes, coastal land management and protection under a scenario of increasing coastal storm frequency.

1.6 THESIS OUTLINE

This thesis aims to highlight environmental changes in delta systems and explore how deltas respond to these changes. The 2nd chapter focuses on reconstructing long-term fluvial sediment flux for a river-wave dominated delta system and exploring the mechanisms that control this process. The roles of different factors (temperature, precipitation, forest cover, dam emplacements) are evaluated through an
application of the hydrological model HydroTrend to the Ebro Delta, Spain, for the last 4,000 years. The 3rd chapter discusses the hydrodynamics and morphological responses of a river delta (the Wax Lake Delta) to hurricanes, and the roles of winds, waves, rivers, and vegetation on the delta morphology during extreme events. In this chapter, the widely used model Delft3D is applied for the simulations. The 4th chapter numerically compares the impacts of different magnitudes of coastal storms on delta morphology. The cumulative impacts of frequent winter cold fronts (20 – 30 times per year) on delta morphology are compared to those from one-year hurricane events (Hurricanes Ike and Gustav), which are more intense events but occur less frequently. Salinity variations during the two types of events are also examined. The 5th chapter concludes this thesis and outlines the major findings and results from this study, as well as highlighting the importance of this study in a broader scientific context.
2 FLUVIAL RESPONSE TO CLIMATE VARIATIONS AND ANTHROPOGENIC PERTURBATIONS FOR THE EBro RIVER, SPAIN IN THE LAST 4000 YEARS


2.1 ABSTRACT

Fluvial sediment discharge can vary in response to climate changes and human activities, which in return influences human settlements and ecosystems through coastline progradation and retreat. To understand the mechanisms controlling the variations of fluvial water and sediment discharge for the Ebro drainage basin, Spain, we apply a hydrological model HydroTrend. Comparison of model results with a 47-year observational record (1953–1999) suggests that the model adequately captures annual average water discharge (simulated 408 m$^3$ s$^{-1}$ versus observed 425 m$^3$ s$^{-1}$) and sediment load (simulated 0.3 Mt yr$^{-1}$ versus observed 0.28 ± 0.04 Mt yr$^{-1}$) for the Ebro basin. A long-term (4000-year) simulation, driven by paleoclimate and anthropogenic land cover change scenarios, indicates that water discharge is controlled by the changes in precipitation, which has a high annual variability but no long-term trend. Modeled suspended sediment load, however, has an increasing trend over time, which is closely related to anthropogenic land cover variations with no significant correlation to climatic changes. The simulation suggests that 4000 years ago the annual sediment load to the ocean was 30.5 Mt yr$^{-1}$, which increased over time to 47.2 Mt yr$^{-1}$ (AD 1860–1960). In the second half of the 20th century, the emplacement of large dams resulted in a dramatic decrease in suspended sediment discharge, eventually reducing the flux to the ocean by more than 99% (mean value changes from 38.1 Mt yr$^{-1}$ to 0.3 Mt yr$^{-1}$).

2.2 INTRODUCTION
Rivers transport significant amounts of sediment to the ocean, affecting the evolution of river deltas, and producing valuable land for human settlements (Ashton et al., 2013; Giosan et al., 2012; Kirwan et al., 2011). Currently nearly half a billion people live on or near deltas, among which some changed significantly during the last century, mostly due to human-induced global changes (Sánchez-Arcilla et al., 2008; Syvitski and Saito, 2007; Syvitski et al., 2009; Vörösmarty et al., 2009). Delta coastlines evolve in response to both terrestrial and oceanic forcing. Terrestrial factors determine the amount of riverine sediment flux to the ocean, which provides the most significant source of coastal sediments, while oceanic forces determine the deposition of these sediments (Reimnitz et al., 1988; Ashton et al., 2013; Giosan et al., 2012; Goodbred and Kuehl, 2000).

For many rivers, the fluvial fluxes towards the ocean have decreased over the last few decades (Syvitski and Kettner, 2011). River sediment flux to the ocean varies over time, corresponding to both climate changes and human activities. Wet and warm climate typically favors high sediment loads when the river has high transport capacity and the basin has a high erosion rate (Xu, 2003; Zhu et al., 2007). Anthropogenic perturbations in the hinterland can either increase or decrease the sediment flux at the river mouth. Transforming a landscape from its natural condition to an anthropogenic land cover, e.g. farmland, typically increases sediment load (Guillén and Palanques, 1997; Kettner et al., 2007; Milliman et al., 1986; Saunders and Young, 1983). For example, Kettner et al. (2007) found that intensive land cultivation could increase sediment flux by more than an order of magnitude. However, modern human activities, such as irrigation and emplacement of dams, have significantly reduced the sediment flux to the ocean. At certain locations the reduced fluvial sediment flux is so severe that coastline retreat becomes noticeable (Frihy et al., 1998; Vörösmarty et al., 2003). Modern human activities are estimated to control more than half the flow of all accessible fresh water (Vörösmarty et al., 1997; Crutzen, 2002) and reduce the global terrestrial sediment to the ocean by ~10% (Syvitski et al., 2005). Given this changing environment, it is critical to explore the mechanisms that control fluvial sediment fluxes towards the ocean over time in order to better understand deltaic evolution. Such knowledge can subsequently contribute to more sustainable management practices of the river system as a whole. Because rivers are
often not monitored or data is inaccessible, the mechanisms controlling fluvial sediment fluxes have not been well understood. Currently, less than 10% of Earth’s rivers have observational time series of sediment delivery to the ocean (Syvitski et al., 2005). Among the available records, some span 100 years at best, which is too short to unravel critical processes that influence the fluvial fluxes (Wilby et al., 1997).

Numerical models have been developed to simulate discharge of watersheds. Empirical models can significantly increase the simulation time efficiency and are able to reproduce basin hydrology across long time periods with high accuracy, especially for smaller basins, which typically have relatively low spatial variability. HydroTrend is a climate-driven hydrological transport model, developed to produce synthetic water discharge and sediment flux for both short (yearly) to long (10^5 years) time periods with relatively easy to obtain input parameters (Kettner and Syvitski, 2008; Syvitski et al., 1998). The model has successfully reproduced long-term fluvial fluxes for several Mediterranean drainage basins like the Po and Rhone as well as the Têt (Kettner and Syvitski, 2009).

The Ebro Delta, located along the Spanish Mediterranean coast, has experienced significant changes in riverine output due to both climate and human activities. The delta began prograding rapidly since around 4000 years ago (Canicio and Ibáñez, 1999). However, modern human activities have altered this condition since the middle of the 20th century, transforming the Ebro Delta towards a more “sediment-starved” system (Jiménez and Sánchez-Arcilla, 1993). This is mainly caused by intensified dam emplacements, which lead to a reduction of more than 95% in sediment load to the river mouth (Palanques et al., 1990). As a result, the Ebro Delta is evolving from a river-wave dominated system to a more exclusively wave-dominated system (Jiménez et al., 1999). This shift will significantly affect social and ecological systems, such as agriculture, fishery industry, tourism and biodiversity in the near future (Fatoric and Chelleri, 2012). Therefore, the Ebro River serves as an ideal case for studying mechanisms controlling fluvial sediment flux, and offers insights for future river and delta management.

This paper provides an overview of the evolution of potential fluvial sediment fluxes of the Ebro River over the last 4000 years by applying HydroTrend. We test the hypothesis that the increase of
the fluvial sediment fluxes is primarily controlled by anthropogenic land cover changes. Currently, modern human activities such as dam construction significantly constrain the fluvial sediment flux. Through numerical simulations, we attempt to 1) obtain the general trend of water and sediment discharge variations over time, 2) analyze the potential impacts of different factors on fluvial water discharge and sediment flux, and 3) compare the effects of natural and human activities on the Ebro river system.

2.3 MATERIALS AND METHODS

2.3.1 REGIONAL SETTING

The Ebro basin, located in Northeastern Spain, has a drainage area of 85530 km². Over 928 km long, the Ebro River is longest and has the largest water discharge among all Spanish rivers (Ibáñez et al., 2008). Confined by several mountain ranges (the Cantabrian Range and the Pyrenees in the north and the Iberian Massif in the south), the river flows through a triangular-shaped drainage basin (Fig. 2.1) before reaching a 320 km² large subaerial delta. The catchment is heterogeneous in terms of geology, topography, land use, and climatology. Four major geological rock types can be found (Guillen and Palanques, 1992): 1) sedimentary and metamorphic Paleozoic rocks in the Iberian and Pyrenees mountains; 2) sedimentary Mesozoic rocks in the Iberian Chain; 3) sedimentary Tertiary rocks in the Pre-Pyrenees and the Ebro Valley; and 4) unconsolidated sediments in the lower part of the Ebro River. Topography varies from the surrounding mountains towards the fluvial system with a total basin relief exceeding 3 km. Land cover changed over time according to climate changes and human activities. The Ebro basin was dominated by xerophytic Mediterranean woodlands that locally graded into both tall forests and semi-desert steppe vegetation depending on climate and soils during the mid Holocene (6000-4000 BP) when human impact was still minor (Davis and Stevenson, 2007). At present, the Ebro basin is mainly covered by irrigated cropland, pastures and fragmented patches of natural vegetation (Water Resources eAtlas).
Figure 2.1 Map of the Ebro basin and its fluvial system

Note: The map shows four major reservoirs including the Ebro, the Canelles, the Mequinenza and the Riba-roja reservoirs (the Ebro, the Canelles, and the Mequinenza dams are incorporated in the model because of their volume (>0.5 km$^3$) and therefore significance impact. The Riba-roja dam is included as it is close to the river mouth). Mora d’Ebre Monitoring Section (MEMS) is the gauging station used to obtain water discharge and suspended sediment load observations from Tena et al., 2011.

Precipitation varies spatially across the basin due to topographical and climatological diversity. The mean annual precipitation varies from 2000 mm in the Pyrenees to 300 mm in the lower centrally located valley that includes the main stream (Vericat and Batalla, 2006). Precipitation experiences high seasonal variability. Observations (AD 1950–2008) of the Spain Meteorological Agency indicate that November is the wettest month (monthly averaged precipitation of 66.12 mm) and July is driest (monthly averaged precipitation of 26.41 mm). Water discharge varies accordingly. Observed discharges from the Tortosa gauging station (2002–2008; Fig. 2.1) ranges from less than 50 m$^3$ s$^{-1}$ in the dry season to more
than 10000 m$^3$ s$^{-1}$ during flood conditions (Guillén and Palanques, 1997). The mean annual discharge between 1914 and 1935 was 592 m$^3$ s$^{-1}$, which decreased by 28% to 426 m$^3$ s$^{-1}$ in the last few decades (1960-1990) (Ibàñez et al., 1996). This decrease was mainly caused by irrigation practices, domestic water consumption and intensified evaporation from reservoirs that had recently been constructed. At present, there are 187 dams in the drainage basin, regulating 57% of the river flow and influencing more than 97% of the basin area (Négrel et al., 2007).

2.3.2 MODEL THEORY, INPUT PARAMETERS AND VALIDATION

2.3.2.1 Model description

HydroTrend is able to simulate daily water and sediment discharge at a river mouth with high accuracy over a long period of time (centuries to millennia) (Syvitski et al., 1998). The model incorporates basin properties and biophysical processes to compute the hydrological balance (Kettner and Syvitski, 2008). For long-term simulations, HydroTrend is capable of reproducing reliable fluvial sediment flux if appropriate assumptions of past climate and land use are made (Syvitski and Morehead, 1999).

The structure and modules of HydroTrend have been described in detail by Syvitski et al. (1998) and Kettner and Syvitski (2008), and will not be iterated here. However, key equations for this study, to compute river discharge, sediment load, and reservoir trapping efficiency, will be described below.

Fluvial water discharge is determined by basin area ($A \text{ [m}^2\text{]}$), precipitation ($P \text{ [m} \text{ yr}^{-1}\text{]}$), evapotranspiration ($Ev \text{ [m}^3\text{ s}^{-1}\text{]}$) and water storage and release ($Sr \text{ [m}^3\text{ s}^{-1}\text{]}$), based on the classic water balance equation (Eq. 2.1). Here $ne$ is the number of simulated epochs and $i$ is the daily time step. Following equation 2.1, five hydrological processes are taken into consideration by the model: rain ($Q_r$), snowmelt ($Q_n$), glacial melt ($Q_{ice}$), evaporation ($Q_{Ev}$) and groundwater discharge ($Q_g$) (all in [m$^3$ s$^{-1}$]):

$$Q = A \sum_{i=1}^{ne} (P_i - Ev_i \pm Sr_i)$$  \hspace{1cm} (2.1)

$$Q = Q_r + Q_n + Q_{ice} - Q_{Ev} \pm Q_g$$  \hspace{1cm} (2.2)
In the model, sediment load is classified into two categories: suspended sediment load and bedload. This study focuses on suspended sediment load \( (Q_s \text{[kg s}^{-1}\text{]}) \), which is determined through a 2-step approach. In the first step, long-term suspended sediment load (~30 years: \( \bar{Q}_s \text{[kg s}^{-1}\text{]} \)) is computed by applying a semi-empirical relationship described by Syvitski and Milliman (2007):

\[
\bar{Q}_s = \omega B \bar{Q}^{0.31} A^{0.5} \bar{R} T, \tag{2.3}
\]

Here the term \( B \) (non-dimensional) is estimated as:

\[
B = IL(1 - T_E)E_R, \tag{2.4}
\]

where \( \omega \) is the proportionality coefficient defined to be 0.02 kg s\(^{-1}\) km\(^{-2}\) °C\(^{-1}\), \( \bar{Q} \) and \( \bar{R} \) are respectively non-dimensional water discharge at the river mouth and maximum basin relief, following the procedure as \( \bar{Q} = \left( \frac{Q}{Q_0} \right) \) and \( \bar{R} = \left( \frac{R}{R_0} \right) \), where \( Q_0 \) is equal to 1 m\(^3\)/s, and \( R_0 \) equals 1 m. \( T \) is the temperature at sea level for each simulated epoch (°C). \( I, L, T_E, \) and \( E_h \) are non-dimensional parameters, where \( I \) is a glacial erosion factor to represent the impact of glacial erosion processes, \( L \) is the basin-averaged lithology factor to express the hardness of rock, and \( T_E \) is the trapping efficiency of man-made reservoirs to incorporate sediment trapping by reservoirs. \( E_h \) is the soil erosion factor related to human activities, which is in earlier studies (Kettner and Syvitski, 2008; Syvitski and Milliman, 2007) expressed through population density and Gross National Product. For this study we derive \( E_h \) through variations in agricultural/usable land (croplands and pasture) and natural/unusable land (forest and other natural land types). These two types of land cover are represented in the model by applying different non-dimensional erosion parameters (respectively 0.5 versus 1.4 for natural/unusable land and agricultural/usable land) based on a study of Garcia-Ruiz et al. (1996). Equation 2.3 explains 96% of suspended sediment variance when compared to 294 globally distributed rivers (Syvitski and Milliman, 2007b), and shows good consistency in previous studies (Kettner et al., 2010, 2007). Therefore, we are confident that this relationship is applicable to the Ebro basin as well.

Once the long-term suspended sediment load is determined, the stochastic model \( (\Psi) \) is applied to generate daily suspended sediment loads (Eq. 2.5; Morehead et al., 2003).
\[
\left( \frac{Q_{S[i]}}{Q_S} \right) = \psi_{[i]} \left( \frac{Q_{[i]}}{Q} \right)^{C_{(a)}}_{i=1:m},
\]

where \( m \) is the total number of days \((i)\) being modeled of a climate interval, also called an epoch. The \( Psi \) model can capture the inter- and intra-annual variability of suspended sediment load at the river mouth following Eq. 2.6-2.9:

\[
E(\psi) = 1 \tag{2.6}
\]

\[
\sigma(\psi) = 0.763 (0.99995)^Q \tag{2.7}
\]

\[
E(C) = 1.4 - 0.025T + 0.00013R + 0.145 \ln (Q_S) \tag{2.8}
\]

\[
\sigma(C) = 0.17 + 0.000183Q \tag{2.9}
\]

Parameters \( E \) and \( \sigma \) represent the mean and standard deviation of a random variable, respectively. Parameter \( C \) is a normal distributed rating coefficient that varies over a time step of one year. The standard deviation of \( C \) depends on the average water discharge. Parameter \( \psi \) is a random variable which changes on a daily time step and has a log-normal distribution—its standard deviation is also calculated from the mean discharge but by applying a power relationship (Eq. 2.7). As a result, small rivers (large variances in \( \psi \)) will generate a more variable hydrograph and vice versa. For short-term simulations (years to decades), the daily mean suspended sediment load \( (Q_{S[i]}) \) might not precisely match the mean suspended sediment load \( (Q_S) \), however, these two parameters converge in long-term simulations (hundreds to thousands of years).

Reservoirs formed by dams can trap a significant amount of sediment which otherwise would have reached the river mouth. In the model, the influence of reservoirs is represented by imposing a trapping efficiency factor on the calculating of suspended load in Eq. 2.4, which is calculated as (Brune, 1953; applied for dams with a reservoir volume larger than 0.5 km\(^3\)):

\[
Te = \sum_{j=1}^{m} \left( 1 - \frac{0.05}{\sqrt{\Delta \tau_j}} \right), \tag{2.10}
\]

where \( \Delta \tau_i \) is the approximated residence time per sub-basin \( j \) and is estimated by

\[
\Delta \tau_j = \frac{\sum_i^{n_i} V_i}{Q_j}, \tag{2.11}
\]
here: $V_i$ is the operational volume of a reservoir $i$ (m$^3$), and $Q_j$ is the water discharge at the mouth of a sub-basin $j$.

**2.3.2.2 Input parameters for short-term simulation (1953~1999)**

The Shuttle Radar Topography Mission (SRTM) version 2.1 with a spatial resolution of three arc seconds (Farr, et al., 2007) is used to obtain basin relief ($\bar{R}$) for this study. The lithology factor ($L$) is assigned using the classification scheme of Syvitski and Milliman (2007), based on the Ebro basin geological conditions (http://oph.chebro.es/ContenidoCartoGeologia.htm 09/03/2013). According to this scheme, the basin is divided into alluvial sediment, sedimentary rock, and igneous rock, with lithology factors of 2, 1.5, and 0.75, respectively. Basin climate is derived from daily measurements by the Spain Meteorological Agency with a spatial resolution of 0.5° for both temperature and precipitation. The gridded precipitation is classified by elevation (<1000 m, 1000~2000 m, and >2000 m), and then averaged by area. Temperature data is derived from one station close to the river mouth (elevation 3 m) to eliminate the impact of elevation on temperature. The daily data is averaged to obtain the monthly means and standard deviations for model input.

The impact of dams on the sediment flux became more significant since AD 1950–1975. During this time, most of the dams were emplaced in the catchment, trapping more than 95% of the total suspended sediment load (Palanques et al., 1990). Data from the Global Reservoir and Dam database (Lehner, et al., 2011) is used to determine where and when dams with a volume larger than 0.5 km$^3$ were constructed. Located close to the river mouth, the Mequineza and Riba-roja dams (Fig. 2.1), which were constructed during the 1960s, have the most significant impact on reducing the sediment flux towards the river mouth.

In the last several decades, irrigation diversion has redirected a large amount of water that otherwise would have reached the river mouth. To take this into account in the model, water loss due to irrigation is represented by manually decreasing the mass balance coefficient.

**2.3.2.3 Model validation**
The 47-year simulation results demonstrate that the model is capable of effectively simulating water discharge. The modeled monthly mean water discharge is 408 m$^3$ s$^{-1}$, versus observed 425 m$^3$ s$^{-1}$ over a 47-year period (AD 1953–1999) at the Tortosa gauging station. Because the model uses a stochastic routine to calculate daily output, the modeled results cannot be compared directly with the observed data. In order to evaluate the model performance, simulated monthly water discharges are ranked by volume and compared with observed data for the computational period, illustrating that the model is capable of reproducing present day discharge characteristics (Fig. 2.2). The simulation also produces a reliable mean annual suspended sediment load (0.3 Mt yr$^{-1}$), closely matching observations ($0.28 \pm 0.04$ Mt yr$^{-1}$, Vericat and Batalla, 2006). The relationship between modeled daily water discharge and suspended sediment load has been compared with 600 observations over a seven-year period (data described by Tena et al., 2011), illustrating that HydroTrend captures suspended sediment load variability for low flows as well as for medium to peak flows (Fig. 2.3). In addition, the model is shown to accurately simulate sediment transport during flood events. According to the model results, 45% of suspended sediment is transported during 2% of the time during 2002-2003, compared to the observed 50%, and 82% of the simulated suspended sediment is transported in 10% of the time during 2003–2004, compared to the observed 99% (Vericat and Batalla, 2006). The minor discrepancy likely arises from the different timespans of the measurements and the model simulation. The observations were taken from October 2002 to September 2003 (1 year), but the simulation period is from January 1st, 2002 to December 31st, 2003 (two years). Accordingly, the model describes a more general status of the river’s behavior.
Figure 2.2 Ranked monthly water discharge comparison of a 47-year record (1953-1999) of the Ebro River

Figure 2.3 Visual comparison of water discharge versus suspended sediment load from 600 observations taken at the Mora d’Ebre monitoring section (MEMS) (Fig. 2.1) over 2002–2008 (black dots, Tena et al., 2011) with the daily records from the 47-year simulation (gray dots)

To statistically test the model performance for water discharge and sediment load, we applied a Chi-square analysis, incorporating 15 equally spaced subclasses (Davis, 1986). The comparison indicates
that the model is able to effectively reproduce the observed variability in river water discharge \( (P<0.05, \chi^2 = 15 \text{ with 14 degrees of freedom}) \). A two-sample Chi-square test is then used to compare the simulated suspended sediment loads with 313 observations taken during 2002~2008. The observed sediment loads are classified into 25 subclasses based on both water discharge and suspended sediment load (Fig. 2.4). In order to compensate for the nonrandom observed data, the same amount of simulated sediment loads are randomly selected for 1000 subsets within each of the 25 subclasses. After that the observed dataset is compared with the 1000 subsets of the simulated suspended sediment load, using:

\[
\chi^2 = \sum_{y=1}^{m} \sum_{x=1}^{n} \frac{(N_1xy-N_2xy)^2}{(N_1xy+N_2xy)} \text{ where } N1, N2 > 0
\]  

(2.12)

Here \( m \) and \( n \) are the number of discharge and suspended sediment load subclasses, respectively. The resulted \( \chi^2 \) are ranked and shown in figure 2.5. The intrinsic variability of simulated sediment load is evaluated through comparison between the simulated subsets applying the same method (Fig. 2.5). Comparison indicates that the mean representative \( \chi^2 \) value of the data-model comparison (the 50\textsuperscript{th} percentile of 1578 ± 3) is less than the critical \( \chi^2 \) value from the model intrinsic comparison with a confidence of 99\% (1584 ± 5). This demonstrates that HydroTrend is able to catch the main characteristics of water discharge and sediment load for the Ebro River and therefore we feel confident applying the model over the past 4000 years.


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Figure 2.4 Distribution of observed water discharge and suspended sediment load for the Ebro River over the period 2002–2008

Figure 2.5 Ranked $\chi^2$ values for model-model comparison (comparison between 10000 simulated subsets) and model-observed comparison (observed dataset compared with 10000 simulated subsets)
2.3.2.4 Long-term simulation

2.3.2.4.1 Historical climate reconstructions

Historical precipitation and temperature have been reconstructed using multiple datasets to generate climate input for a 4000-year simulation. Basin-wide averaged monthly mean precipitation and its standard deviations are obtained from daily observations made over the last 58 years (AD 1950–2008) by the Spain Meteorological Agency. Half a millennia of daily data is then reconstructed with correlation analysis, combining the 500-year European seasonal precipitation data from Pauling et al. (2006) with the daily data from the Spain Meteorological Agency. Pauling et al. (2006) reconstructed this gridded seasonal dataset with a horizontal resolution of 0.5° by using principal component regression to combine long quality-checked instrumental precipitation series and precipitation indices based on documentary evidence (non-instrumental data on past weather and climate that was reported as well as instrumental observations that were taken before the set-up of continuous meteorological networks) and natural predictors (tree-ring chronologies, ice cores, corals and a speleothem) that are sensitive to precipitation signals (Fig. 2.6B). The data is then extended by merging the 500-year daily precipitation with an 8000-year rainfall trend to obtain the last 4000-year yearly precipitation (Fig. 2.6A). The 8000-year trend is reconstructed from the lipid distributions and total organic carbon (TOC) of a peat bog recovered in Northern Spain, reflecting changes of plant distribution over time, which was related to palaeohydrological and palaeoenvironmental conditions. From their study, ten dry/humid episodes were distinguished according to the n-alkane carbon proxies during the last 8000 cal. years (Ortiz et al., 2010). The standard deviations of precipitation for the long-term simulation are inferred from the means based on the assumption that a time period with more precipitation tends to have a higher standard deviation.
Figure 2.6 Data used to generate paleoclimate input as well as forest coverage for a 4000-year period

Note: A) 4000-year precipitation index from Ortiz et al. (2010); B) 1500–2000 AD yearly precipitation for seasonal analyses from Pauling et al. (2006); C) observed 1950–2008 AD monthly precipitation for daily analyses (Spain Metrological Agency); D) 4000-year temperature index from Martínez-Cortizas et al. (1999); E) observed 1950–2008 AD monthly temperature for daily analyses (Spain Metrological Agency); F) 4000-year yearly basin-averaged forest coverage data for the Ebro basin from Kaplan et al. (2009) based on gridded data.

Long-term temperature time series at the river mouth are reconstructed following the similar method described above. The monthly means and standard deviations are obtained from the same source as precipitation. The 58-year data is then correlated with a long-term trend (4000 years) from Martínez-Cortizas et al. (1999) to acquire the last 4000-year yearly mean temperature (Fig. 2.6D, E). The long-term temperature trend is inferred from the analysis of a 4000-year record of the net accumulation of atmospheric mercury, based on the phenomena that mercury has lower thermal stability and therefore tends to be preserved and accumulated during colder episodes. For warmer episodes, mercury will become more thermally active, leading to less preservation and accumulation (Martínez-Cortizas et al., 1999).
2.3.2.4.2 Anthropogenic land cover changes for the Ebro basin

Humans have altered the Earth’s surface by, for example, cultivating land that was once forested. By doing so, more sediment erodes from hillsides, eventually reaching rivers. The eroded sediment is then transported through river networks towards the ocean (Alatorre et al., 2012; Mitchell and Thornes, 1990; Syvitski and Kettner, 2011). In this study, we used anthropogenic land cover change scenarios (KK10; Kaplan et al., 2012, 2009) to estimate the effects of land use change on sediment delivery to the Ebro River. The KK10 scenarios have 5 arc-minute spatial and annual temporal resolutions. Anthropogenic land cover change is determined by a combination of country-scale estimates of past population and maps of land suitability for agriculture and pasture. Suitability for land use in turn depends on climate and soil properties. Since the first half of the 20th century, irrigation became important in the Ebro basin, which was the first part of Spain where large-scale irrigation projects were established (Pinilla, 2006). Irrigation substantially changed the pattern of land use, concentrating agricultural land into the hottest and driest parts of the central Ebro basin that were formerly less intensively used because they have low suitability for (rained) agriculture and pastures. Thus, the significant shift in the spatial pattern of land use from the preindustrial to the modern era may have an additional impact on sediment flux to the Ebro and its delta.

For the Ebro basin, anthropogenic land cover change was important over the entire 4000 years of our study period. In the earliest part of our time period of interest, during the Bronze Age, the basin was mainly used for pasture (Fig. 2.6F), which, given low densities of animals and preference for naturally semi-open landscapes, may not have had a substantial impact on natural land cover. A period of deforestation began at around 2700 BP due to a considerable increase of settlers in the Ebro basin (van Zuidam, 1975). Anthropogenic impact continues to intensify until about AD 150 (1800 BP) when land use reaches a maximum (39%). The population in the Ebro basin diminished afterwards because of a Roman political act that required redistribution of indigenous settlements (Harrison, 1988). Severe aridity in this time period may have also intensified the process, leading to recovery of the natural vegetation cover. The natural vegetation cover increased until AD 550 (1400 BP), approaching a peak of 69%, which
was followed by increased land use due to the establishment of nomadic pastoralism in the region (Davis, 1994). Another maximum in anthropogenic land cover (48%) occurred during the late Muslim times in Iberia (AD 1150-1250; Chocarro et al., 1990; Davis, 1994). The European Black Death epidemic of AD 1348-1353 led to a second shorter period of depopulation and recovery of the natural vegetation (58%; Chocarro et al., 1990). The 17th century saw a marked agricultural expansion in the Ebro basin (Davis, 1994), which combined with the cold-dry climates during the Little Ice Age (LIA 600–200 BP), led to a minimum natural vegetation cover at around AD 1750 (48%) (Chocarro et al., 1990; Valero-Garcés et al., 2004), Most recently, intensification of agriculture in irrigated areas, industrialization, urbanization and afforestation schemes have led to land abandonment and, in some areas, recovery of natural vegetation following the 18th century minimum (Gallart and Llorens, 2004).

As climate and anthropogenic land cover changes are primary drivers of our river discharge and sediment flux calculations, uncertainties in these scenarios are discussed in more detail in section 2.1.

2.3.3 Sensitivity Analysis

Three numerical experiments have been designed in order to determine the most dominant factor (climatic versus human impact) that controls changes in water and suspended sediment discharge. Climate changes are expressed as variations in temperature and precipitation, and human impacts are represented by changes in land use and emplacement of dams. For experiment 1 (EX1), climate varies according to the derived input data discussed in section 2.2.4.1, and human influence is kept constant, using the land use parameters of 4000 BP and without dam emplacement. For experiment 2 (EX2), climate is kept constant at the current climate characteristics derived from the period 1950–2008. Human impacts on riverine output over time are taken into account by changing the long-term land use parameter ($E_h$) according to section 2.2.4.2 and by adjusting the dam trapping efficiency ($T_e$) since the second half of 19th century (1960) following section 2.2.1. For the last experiment (EX3) both climate and human impacts vary over time. Similarly, we highlight the impact of reservoirs by designing two numerical experiments: one that does not include the impact of dam emplacement and one that does. For both runs
the settings are kept identical to the model validation case (1953–1999) discussed in section 2.2.3. The weighting method is adopted in order to explore the impacts of human activities and climate on the riverine output. Four additional simulations are preformed to analyze the model output sensitivity by incorporate a 20% increase / decrease of the main climate input parameters (temperature, precipitation) for EX1, and a 20% increase / decrease of the main anthropogenic factor (land use) for EX2, respectively.

2.4 RESULTS

2.4.1 LONG-TERM SIMULATION RESULTS

A 4000-year HydroTrend simulation of the Ebro basin has been conducted. Nelson (1990) estimated that the basin transported a minimum of 22.0 Mt yr\(^{-1}\) of suspended sediment load to the ocean during 1912–1935 based on stratigraphic analysis, and Ibáñez et al. (1996) suggest a value of 28.1 Mt yr\(^{-1}\) for the same period. The model simulates that, on average, suspended sediment load and its standard deviation to the ocean was 47.6 ± 11 Mt yr\(^{-1}\) during that period. Nelson (1990) also reported that the annual suspended sediment load was ~0.3 Mt yr\(^{-1}\) after the emplacement of dams in the 1980s, close to our model findings of 0.3 ± 0.3 Mt yr\(^{-1}\), although a more recent study indicated that this might be lower (0.1 Mt yr\(^{-1}\)) for the same period (Rovira and Ibáñez, 2007). Dams are represented in the model since 1960, influencing 36.9% of the Ebro basin area, which increased to 99.7% in 1966 when the Mequineza and Riba-roja dam were emplaced. After 1966 we assume all dams are operational and no additional dams are placed, while in practice, dams were constructed more gradually. This difference leads to some discrepancies over the short term, but does not have an impact when comparing suspended sediment load to the ocean over two time periods: before and after emplacement of all dams. Overall, the comparisons demonstrate that modeled results match observed data.

Long-term variations in water discharge reflect changes in precipitation. The simulated annual mean water discharge has a moderate temporal variability (665 ± 44 m\(^3\) s\(^{-1}\)) without an apparent trend (Fig. 2.7). The water discharge shows no significant changes in the first thousand years of simulation, due
to the comparatively stable precipitation input (Fig. 2.4A). Discharge then decreases to a minimum of 543 m$^3$ s$^{-1}$ at around 2600 BP (Low1, Fig. 2.7) before the Roman Humid Period (RHP: 2600–1600 BP). Increased precipitation leads to a rise in discharge during the RHP. The humid condition continues and the discharge keeps increasing till 1800 BP, reaching an annual maximum of 690 m$^3$ s$^{-1}$ (High, Fig. 2.7). The high discharge lasts until 1600 BP when precipitation approaches the RHP maximum. After that, discharge decreases to another comparatively low of 604 m$^3$ s$^{-1}$ at around 1150 BP (Low2, Fig. 2.7) when precipitation reduces widely across Europe during the Medieval Warm Period (MWP: 1200–600 BP) (Jalut et al., 2000; Julià et al., 1998; Ortiz et al., 2010). Water discharge rebounds until the LIA (600–200 BP), afterward it decreases and reaches a minimum of 590 m$^3$ s$^{-1}$ (Low3, Fig. 2.7) when the LIA is at its maximum at around 200 BP. A rapid ephemeral rise in water discharge appears after the LIA, followed by a sudden decline (40%) after AD 1956 when the adjusted mass balance coefficient is applied to account for fresh water uses for irrigation purposes.

Compared to water discharge, long-term suspended sediment load does consistently increase over time (Fig. 2.7). Starting with a hundred-year average of 30.5 ± 8.8 Mt yr$^{-1}$, the suspended sediment load slowly increases to 47.2 ± 12.4 Mt yr$^{-1}$ for the period 1860-1960, before emplacement of dams. There are no severe climatic or human disturbances before 2700 BP, and therefore the variations of the simulated sediment load follow closely the water discharge (Fig. 2.7). Discrepancy between simulated water discharge and suspended sediment load arises around 2700 BP, when a considerable population increase in the Ebro basin (van Zuidam, 1975) leads to deforestation and a subsequent increase in suspended sediment load (Increase 1, Fig. 2.7). A relative maximum sediment load of 43.5 Mt yr$^{-1}$ appears at around 1800 BP (Max1, Fig. 2.7), related to a maximum of land use during Roman times (Fig. 2.6F). Following this population maximum, population decreases with the collapse of classical empires (Kaplan et al., 2009) and increased aridity. As a result, a minimum load (39.9 Mt yr$^{-1}$) occurs at around 1400 BP (Min1, Fig. 2.7). Another remarkable maximum of suspended sediment load (44.0 Mt yr$^{-1}$) (Max2, Fig. 2.7) happens at around 700 BP, coinciding with the Muslim times in Iberia. During that
period, population increases abruptly and, as a consequence, land use increases to a maximum (42%). Later, the population declines during the Black Death Period, contributing to recovery of natural vegetation and decreases in suspended sediment load to 40.0 Mt yr\(^{-1}\) (Min2, Fig. 2.7), followed by a rise of sediment load to 47.4 Mt yr\(^{-1}\) (Max3, Fig. 2.7) during the LIA (Increase 2). Modern agricultural development over the last century increases sediment load. However, this tendency is altered with the emplacement of dams. Suspended sediment load to the Ebro River mouth decreases sharply to less than 1% of its original output (from 38.1 to 0.3 Mt yr\(^{-1}\)), illustrating how modern human activities completely control the sediment load of the Ebro River to the ocean (Fig. 2.7).

![Figure 2.7 Simulated annual water discharge and suspended sediment load and their 50-year averages for the last 4000 years](image)

Note: LIA: the Little Ice Age; MWP: the Medieval Warm Period; RHP: the Roman Humid Period. Highs and Lows indicate maximum and minimum values of water discharge, and Maxs and Mins indicate maxima and minimal of suspended sediment load. Increase1 and Increase2 are the two periods when suspended sediment load has a steady increasing trend.

### 2.4.2 Results from Sensitivity Analyses

Three experiments are set up to study in isolation the impact of humans versus climate changes (see section 2.3) on river water discharge and suspended sediment flux. The results are analyzed for two periods: before and after dam emplacement. Before dam emplacement, humans show a marginal influence on water discharge but do significantly alter the suspended sediment load to the ocean (Fig. 2.8). EX1 (climate-driven) and EX3 (both climate- and human-driven) show high consistency in simulated water discharge, producing an almost identical signal containing comparatively high temporal variations (Fig. 2.8A). Contrastingly, EX2 (human-driven) produces a more stable water discharge signal
with less variation over time (Fig. 2.8A). When dams are included in the model, the simulated water discharge of EX3 deviates from that of EX1. The mean water discharge decreases from $665 \pm 37.6 \text{ m}^3\text{ s}^{-1}$ to $461 \pm 13.7 \text{ m}^3\text{ s}^{-1}$ for EX3 and EX2 after dam emplacement, but stays at $640 \pm 37.2 \text{ m}^3\text{ s}^{-1}$ until the end of the simulation for EX1 (Fig. 2.8A). The weighting experiments show that when climate parameters (temperature and precipitation) increase/decrease by 20%, the 4000-year mean river water discharge changes from $665 \pm 37.6 \text{ m}^3\text{ s}^{-1}$ to $885 \pm 44.7 \text{ m}^3\text{ s}^{-1}$, and $486 \pm 26.1 \text{ m}^3\text{ s}^{-1}$, respectively. There is no significant impact on water discharge when the human-related factor ($E_h$) varies by 20% (Fig. 2.8A).

Sensitivity analyses indicate that, for the Ebro River, suspended sediment load is more susceptible to anthropogenic land cover changes than climate changes (Fig. 2.8B). There are no significant differences in simulated suspended sediment load between the three experiments during the first 1300 years (4000–2700 BP). After that, the annual suspended sediment load increases by 41% from $34.3 \pm 11.2 \text{ Mt yr}^{-1}$ at ~2700 BP to $47.2 \pm 12.5 \text{ Mt yr}^{-1}$ before 1960s for EX2 and EX3 (Fig. 2.8B). EX1 generates a less variable sediment signal, with only small temporal variations around the mean value of $31.5 \pm 9.2 \text{ Mt yr}^{-1}$. The weighting experiments suggest that river suspended sediment flux changes in response to both climate (temperature and precipitation) and anthropogenic (land use) impacts. Twenty percent increase of climate parameters changes the long-term mean sediment load by 5.9 Mt yr\(^{-1}\), from 38.1 Mt yr\(^{-1}\) to $44.0 \pm 11.2 \text{ Mt yr}^{-1}$, while a 20% decrease reduces the flux to $20.5 \pm 6.1 \text{ Mt yr}^{-1}$. When anthropogenic land use is set to increase by 20%, the 4000-year mean sediment load changes from 38.1 Mt yr\(^{-1}\) to $45.9 \pm 16.2 \text{ Mt yr}^{-1}$, while a 20% decrease reduces the flux to $30.6 \pm 10.5 \text{ Mt yr}^{-1}$ (Fig. 2.8B). However, it is highly unlikely that the climate parameters could change by 20% (see also Fig. 2.6A, D).

For the long-term simulation, EX1 produces a stable suspended sediment load, compared to the rising trend for EX2, in which natural vegetation coverage has the general trend of decreasing over time. To identify the isolated role of climate changes (EX1) and human impacts (EX2), the simulated suspended sediment load of EX1 and EX2 are subtracted from that of EX3 (human impacts & climate changes) and normalized to the average suspended sediment load of EX3 (Fig. 2.9). The two tests show different patterns. According to the model, before dam emplacement, anthropogenic land cover changes are the
main factor accounting for the increasing trend of suspended sediment load over time (41%). Once dams are placed, the sediment load diminishes to less than 1% of its natural signal for the period 4000 BP−1960 AD (from 38.1 ± 12.5 Mt yr⁻¹ to 0.3 Mt yr⁻¹; Fig. 2.9). Although changes in climate are clearly reflected in the trend of water discharge (Fig. 2.7 and 2.8A), they have less impact on the long-term suspended sediment load (Fig. 2.9). However, climate changes do affect short-term variations of the annual suspended sediment flux to the ocean, which can be as high as 10%.

Figure 2.8 Comparison of climate changes and human activities on: (A) simulated water discharge and (B) suspended sediment load at the Ebro river mouth over a 4000 year period using a 50-year running average

Figure 2.9 Normalized suspended sediment load: EX1 (climate) and EX2 (human) normalized to EX3 (both climate and human)
Note: Grey line shows the normalized value by subtracting the results of EX3 from that of EX2; black line is the normalized value by subtracting the results of EX1 from that of EX3).

In the 20th century, modern agricultural activities and dam emplacement have significantly affected the Ebro’s flux to the ocean. Results of the 47-year sensitivity analysis indicate that dams alone can reduce the mean suspended sediment load from 21.5 Mt yr\(^{-1}\) to 0.3 Mt yr\(^{-1}\). This suggests that nowadays only ~1.3% of the historical suspended sediment load is able to reach the ocean. When ranking the simulated sediment discharges for an environment without dams, 80% of the sediment would reach the ocean during flows that carry suspended sediment load of less than 1000 kg s\(^{-1}\), and 2% of the sediment would reach the ocean in flows that carry suspended sediment load of more than 9700 kg s\(^{-1}\) (Fig. 2.10). Because of the way that dams are currently operated in the Ebro Basin (water flow is mostly controlled by dam operators, and this process is not included in model), these peaks are not likely to occur anymore. If we assume that all dams reach their maximum storage capacity at all time during the simulation, then 80% of the sediment loads would reach the ocean during flows that carry sediment load of less than 50 kg s\(^{-1}\), and 2% of the sediment would get to the ocean during flows that carry sediment load exceeding 91.7 kg s\(^{-1}\) (Fig. 2.10). During most of the time, these dams are not at their maximum storage capacity and therefore these peak events might be even less.

Figure 2.10 Ranked suspended sediment load with and without dam emplacement (using the 47-year simulation settings)
2.5 DISCUSSION

2.5.1 INPUT PARAMETER UNCERTAINTIES

The reliability of the long-term simulation depends for a large extent on the accuracy of input parameters, which for this study mainly includes paleoclimate and anthropogenic land cover change scenarios. The long-term temperature trend, which is derived from mercury preservation (Martínez-Cortizas, et al., 1999), has been used in many other studies (Diz et al., 2002; Martin-Puertas et al., 2008; Oliva and Gomez-Ortiz, 2012; Oliva and Ortiz, 2011; Parcero-Oubiña, 2003) and shown to be consistent with former findings. For example, former studies estimate that the temperature during the LIA was about 1–2 °C lower than the present day in Spain, and that the MWP and the RWP was 1.5 °C and 2 °C warmer than present, respectively (Tullot, 1996; Martín-Vide and Barriendos, 1995; Ramil et al., 1996). The reconstructed temperature series produce similar results, which are −1.7 °C, +1.9 °C, and +2.2 °C for these three periods, respectively. Given the mean annual temperature of 13.3 °C (4000-year average), the reconstructed data should be effective in producing reliable results.

The long-term precipitation data, derived from the analysis of peat total organic carbon content, is in agreement with other studies as well (Ortiz et al., 2010). For instance, the humid condition during 4000 BP – 2900 BP was consistent with the high lake level episodes in Europe (Magny, 1998, 2004; Magny et al., 2007, 2003) and the observed humidity rise in the Southeastern Mediterranean basin (Schilman et al., 2001). The wetter climate during the RHP was also reported for Southern Spain by other studies (Gil-Garcia et al., 2007; Martin-Puertas et al., 2008). The more recent frequent dry-wet variations, such as the humid condition during 830–230 BP, and the dry condition during 230–80 BP, coincide with the transition from the MWP to the LIA and the maximum onset of the LIA, characterized by increased rainfall in Mediterranean areas (Issar, 2003), and widely spread drought events (Gil-Garcia et al., 2007; Martin-Puertas et al., 2008).

Another important input parameter of the model is the anthropogenic land cover change,
inferred from historical population and technological progress (Kaplan et al., 2009). The KK10 scenarios were evaluated by Kaplan et al. (2009) on available literature for Europe and suggests that significant events, such as the Black Death, the Renaissance, and the industrial Revolution may all have had a distinct expression in the history of deforestation and afforestation across the continent (Darby, 1956; Williams, 2000). The land-use scenario is transformed into an erosion parameter in Hydrotrend through adjusting factor $E_h$ (Eq. 2.4). The comparison of two 4000-year simulations, one with fully covered usable land (crops and pastures) (Run1) and one with fully covered unusable land (natural land use) (Run2), indicates that Run1 produces suspended sediment loads 2.8 times higher than Run2. This result is congruent with findings from a former study in the Aisa Valley of the central Spanish Pyrenees, in which river sediment concentrations were 2.3 times higher when land use was farmland compared to land under forest (Garcia-Ruiz, et al., 1996).

2.5.2 FACTORS CONTROLLING WATER DISCHARGE AND SUSPENDED SEDIMENT LOAD (CLIMATIC VS. HUMAN FACTORS)

Before the last century, alterations in simulated water discharge are highly correlated with precipitation (see Fig. 2.6A and 2.8A). The sensitivity analysis shows there is no difference between the trend of water discharge in EX3 (both climate and human) and that of EX1 (climate)—both follow the main trend of precipitation over time (Fig. 2.6A and 2.8 A). Experiment EX2 (human) provides a different trend, showing stable water discharge over time, illustrating that human activities do not account for the disturbance in river water discharge. The weighting experiments also verify that climate changes determine river discharge over the long-term simulation. A 20% increase of climate parameters leads to river discharge to rise by 33%, and a 20% decrease of climate parameters leads to river discharge to reduce by 37%. Twenty percent changes in land use show no impact on river water discharge. The comparison of climate changes and human influences on water discharge is also expressed by the fluctuation of water discharge. Variations of precipitation causes water discharge to fluctuate around its mean value by ~11% (between 619–691 m³ s⁻¹, mean discharge is 665 m³ s⁻¹), while EX2 (with varied
human factors) produces a mean water discharge of 640 m$^3$ s$^{-1}$, and range 3.4% (from 629 m$^3$ s$^{-1}$ to 650 m$^3$ s$^{-1}$). Water discharge decreased significantly in recent times. Measurements from the Ministerio de Medio Ambiente (2000) demonstrate that the river flow has been reduced by 40% at Tortosa (40 km upstream of the river mouth) from 1960s to 2000s. As a result, the relationship between water discharge and precipitation is highly disturbed. The significant decrease of water discharge was mainly caused by human activities such as irrigation, which is not included in the model simulation. To follow the observed water discharge (Ministerio de Medio Ambiente 2000), we reduced the water discharge by 40% through manually changing model parameters. In general, climate changes contribute to temporal variations of water discharge, which are quasi-periodic and do not change the long-term millennium-scale trend. Conversely, modern human activities can fundamentally change both the overall quantity and peak magnitudes of water discharge towards the river mouth.

Apart from water discharge, model simulations demonstrate that suspended sediment load is mostly affected by human activities. Before dam emplacement, the suspended sediment load primarily follows variations of natural vegetation coverage (Fig 2.6F, 2.7 and 2.8B). Sensitivity analysis shows that EX2 (human) produces similar sediment load as EX3 (both climate and human), in which increased land use accounts for a 41% increase in the sediment load (34.3 Mt yr$^{-1}$ to 47.2 Mt yr$^{-1}$), based on the sediment load at 2700 BP. Before this, the sediment load is more closely determined by river discharge and basin characteristics (basin area, basin relief, temperature, lithology, glacier impact, and initial land use). Experiment EX1 (climate) generates a more stable suspended sediment load of 31.5 Mt yr$^{-1}$, with a variation range of ~10% over the last 4000 years (Fig. 2.9). Over the last century when dams are emplaced, fluvial suspended sediment load to the ocean decreased to less than 1% of its original value (EX2 and EX3). The weighting experiments also show that a 20% change in anthropogenic parameter lead to suspended sediment load variations by more than 20% (20% increase and 25% decrease). Although a 20% change in climate parameters can also lead to more than 15% change in sediment load, such a significant change of climate parameters is highly unlikely, (Fig. 2.8A and 2.6). Overall, long-term simulations demonstrate that climate changes have less of an impact on the sediment load, mostly
contributing to small temporal variations. Human impact is the more dominant factor, controlling suspended sediment load variations over time, first by increasing the available sediment that can be eroded to the river system through deforestation whereafter a dramatic decrease in fluvial sediment load appears due to the emplacement of dams. These findings would therefore support the hypothesis that humans have contributed to the progradation of the Ebro Delta through increasing river sediment flux to the ocean, and that humans could held at least partly responsible for the current vulnerability of the delta as a result of fluvial sediment starvation.

The dominant role for humans on the increase of Ebro’s sediment flux is consistent with previous studies (Albert and Jorge, 1998; Guillén and Palanques, 1997; Canicio and Ibáñez, 1999). However, our model results suggest a different magnitude of historical sediment flux and variations caused by anthropogenic factors. Guillén and Palanques (1997) and Albert and Jorge (1998) suggest that the sediment transport by the Ebro River increased from 6 Mt yr$^{-1}$ two thousand years ago to 25 Mt yr$^{-1}$ for the 19$^{th}$ century (417% increase). Accordingly, the Ebro delta progradation rate increased from 10 m yr$^{-1}$ two thousand years ago to 30 m yr$^{-1}$ between 13$^{th}$ and the 15$^{th}$ centuries, and 50 m yr$^{-1}$ during 1500 and 1650 AC (Albert and Jorge, 1998). They attributed this significant increase of sediment flux to intense deforestation of the river drainage basin. However, this study suggests that the sediment flux may have been as high as 30 Mt yr$^{-1}$ 4 000 years ago, and flux increased over time with a much smaller magnitude compared to the former studies (41% from 2700 BC to the nineteenth century). Varying the anthropogenic input parameter by 20% (the weighting experiments; Fig. 2.8) changes the sediment load by about 20% to 25%. The differences may be caused by uncertainties in input data, both for climate and land uses, as there are no direct observations to validate the reconstructed data. The basin land uses are simply classified into two categories: usable land and unusable land for the simulation, which did not take into account the impact of different usable/unusable land uses on the sediment flux. For instance, grassland and forest are treated similarly, which could increase the uncertainty in the output. However, even taking into account all of these uncertainties, the model is unable to produce a large increase in sediment flux as the former studies.
2.6 CONCLUSIONS

For this study, a numerical model HydroTrend is applied to study the long-term Ebro fluvial sediment flux alterations to the ocean due to climate changes and human influences. First, a 47-year validation test (1953–1999) is implemented to determine how well the model could simulate the modern river hydrological characteristics. These results suggest that the model is capable of producing water discharge and sediment fluxes close to observations (simulated average water discharge of 408 m³ s⁻¹ versus the observed 425 m³ s⁻¹; simulated mean suspended load of ~0.3 Mt yr⁻¹ versus observed 0.28 ± 0.04 Mt yr⁻¹).

Excluding modern human activities, the hydrograph of the Ebro River is determined by climate changes, primarily following the trend of precipitation. High discharges at 1800 BP (690 m³ s⁻¹) are well correlated to high precipitation in the Roman Humid Period, and the relatively low discharge period (604 m³ s⁻¹) at 1150 BP coincides with the dry condition during the MWP. Low discharges at about 200 BP (590 m³ s⁻¹) occur during the LIA maximum. Sensitivity analysis demonstrates that precipitation is the dominant factor accounting for temporal variations (±11%) in river water discharge. Human activities, represented as land use changes, show minor impact on water discharge. However, water discharge has been significantly decreased by modern human activities that consume a large amount of water through irrigation practices and other domestic water consumptions. Weighting experiments, incorporating changes in the main climate parameters (temperature and precipitation) and human factors (anthropogenic land use), verify the dominant role of climate on water discharge (20% changes in climate parameters produces up to 37% changes in water discharge), and the marginal role of human impact (negligible changes of water discharge for 20% changes in human factors). Anthropogenic factors significantly influence suspended sediment flux (20% changes in human factors cause sediment flux to change by 25%). Although sediment load also responds to the changes of climate parameters, the realistic variation range of climate change constrains their impact to a lower level than human influence.
Model simulations also suggest that transport of suspended sediment load to the ocean is more significantly affected by human activities than climate changes. Suspended sediment load at the river mouth increases by more than 41% over the last 4000 years as anthropogenic land cover of the Ebro Basin progresses over time, especially after 2700 BP, as the suspended sediment load increases from 34.3 Mt yr\(^{-1}\) to 47.2 Mt yr\(^{-1}\) in AD 1860-1960. The rising trend of fluvial sediment load over time is interrupted by several maximums and minimums. Three high sediment load periods can be distinguished around 1800 BP, 700 BP, and 200 BP and are correlated with maximum land use due to intensified settlement in the basin, Muslim times, and modern agriculture, respectively. Two low sediment load periods occur when population decreases due to the collapse of classic empires and the Black Death event. Ignoring recent dam emplacement, Anthropogenic land cover is the dominant cause of changes in suspended sediment over the last 4000 years, while climatic changes contribute to small temporal variations. When modern human activities (emplacement of dams and diverting water for irrigation practices) are incorporated in the model, the suspended sediment load declines to less than 1% of its original value (from 38.1 Mt yr\(^{-1}\) to 0.3 Mt yr\(^{-1}\)), indicating that human impact on sediment load is becoming more significant when compared to historical times.

The simulation results show a different pattern of sediment flux increases over time with a much smaller magnitude than the one proposed in former studies (modeled 41% increase of sediment flux from 2700 BC to the 19\(^{th}\) century, compared to 417% increase according to former studies).

2.7 ACKNOWLEDGEMENTS

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3 MORPHODYNAMICS DUE TO HURRICANES: WAX LAKE DELTA, LOUISIANA

3.1 ABSTRACT

This study examined the influence of hurricanes on the morphodynamics of wetlands on the Wax Lake Delta (WLD) in Louisiana, USA, using the Deflt3D numerical model. The simulation showed that Hurricane Rita, which made landfall 120 km to the west of WLD as a Category 3 hurricane in 2005, had a significant impact on the delta where more than 500,000 m$^3$ of bottom sediment was eroded. The simulated Hurricane Rita’s storm surge reaches 2.5 m, with maximum currents of 2.0 m s$^{-1}$, and wave heights of 1.4 m on the WLD. The wind generated flow and waves are the major driving forces of the morphological changes. The northwestern-directed flow pattern causes erosion on the eastern banks of the deltaic islands and deposition in channels that located on the west of these islands. Waves intensify erosion and the total amount of erosion doubled when waves are included in the simulation. Hurricane tracks are critical in controlling sediment transport as it determines the wind fields. A simulated hurricane making landfall 120 km to the west of the WLD produces twice as much erosion and deposition at the delta compared to a hurricane of a similar intensity that makes landfall on the delta. Further, densely distributed aboveground vegetation is numerically shown to slow floodwater propagation and decrease flow velocity on islands, leading to an approximately 47% reduction in the total amount of erosion compared to the simulation excluding vegetation.

3.2 INTRODUCTION

Hurricanes are among the greatest hazards in coastal areas, imperiling coastal wetlands, properties and humans lives (Huang et al., 2001; Jonkman et al., 2009; Li and Ellingwood, 2006; Michener et al., 1997; Pielke et al., 2008). The normalized damage in the continental United States caused by hurricanes, which takes into account inflation, wealth, population and house unit, is estimated to be
$10 \text{ billion yr}^{-1}$ over the period 1900 – 2005 (Pielke et al., 2008), which is predicted to rise with increasing cyclone intensities (Emanuel, 2005; Landsea, 2005) and higher coastal vulnerability caused by subsidence, sea level rise and population growth (Irish et al., 2014; Li and Ellingwood, 2006; Pielke et al., 2008; Syvitski and Kettner, 2011; Syvitski, 2008). Both empirical and numerical studies have shown that the existence of wetlands can effectively reduce the damage from hurricanes by decreasing storm surges and wave heights (Badola and Hussain, 2005; Barbier, 2007; Barbier et al., 2008; Bayas et al., 2011; Chen et al., 2009; Corps of Engineers, 1963; Das and Vincent, 2009; Farber, 1987; Moeller et al., 1993; Resio and Westerink, 2008; Wamsley et al., 2010). Yet few studies have explored the influence of hurricanes on wetland systems. Hurricanes are reported to render great damage to wetlands (Barras, 2006; Miner et al., 2009), but they also transport large amount of nutrients into wetland systems, aiding wetland maintenance (Conner et al., 1989; Goodbred and Hine, 1995; Turner et al., 2006). Barras (2006) demonstrated that 15% of the total wetlands within the Louisiana coastal plain were converted to open water due to Hurricanes Katrina and Rita (2005), while Turner et al. (2006) stated that more than 131Mt of inorganic sediments were transported to the coastal wetlands of Louisiana during the same events.

Hurricane and wetland properties govern the hydrodynamic and morphodynamic patterns of coastal wetlands during hurricane events (Resio and Westerink, 2008; Wamsley et al., 2010). Hurricane intensity, track, and forward speed determine storm surge distribution and wave fields, while wetlands topography alters the floodwater propagation routes. So current and wave shear stress on wetlands is controlled by both hurricane and wetland properties (Resio and Westerink, 2008). Native vegetation reduces flow velocities, attenuates waves on vegetated island tops, changes turbulence structures and modifies soil strength through the belowground root system (Chen et al., 2009; Shepard et al., 2011; Smith, 1976; Temmerman et al., 2005). However, the mechanisms by which different factors quantitatively impact wetland morphology during hurricanes are still unknown. Numerical models provide an effective approach to predict potential flood damage on coastal wetlands. Storm surge and circulation models such as Sea, Lake and Overland Surges from Hurricanes (SLOSH) (Jelesnianski et al., 1992) and ADvanced CIRCulation model (ADCIRC) (Hench et al., 1994) can be used to predict storm
paths, water levels and wave fields in the deep ocean. When hurricane systems approach the shallow coastal area, fluid dynamic models such as Delft3D (Lesser et al., 2004) or the Finite Volume Coastal Ocean Model (FVCOM) (Chen et al., 2007) are able to simulate storm surges and calculate sediment transport within coastal wetlands. Wave models, such as the Steady-State Spectral Wave Model (STWAVE) (Davis, 1992) and Simulating WAves Nearshore (SWAN) (Booij et al., 1996), are usually coupled with flow models to predict the combined impact of flow and waves on coastal hydrodynamics and morphology during hurricanes (Bunya et al., 2010; Dietrich et al., 2010; Rego et al., 2010).

Vegetation is represented in models by changing the model’s hydrodynamic scheme. The most widely used approach is to increase bed roughness (Augustin et al., 2009; Zhang et al., 2012), which accounts for the momentum transfer at the bottom of the water column caused by plants, and works properly for hydrodynamic simulations (Baptist, 2005; Loder et al., 2009). But the method generates erroneous bed shear stress, leading to unrealistic bed level changes in the morphodynamic calculation (Baptist, 2005). Baptist (2005) develops a different approach to calculate hydraulic resistance of vegetation by adding a vegetation drag force to bed roughness, which improves the morphological computation results. A more physically realistic method that has been tested in laboratory is to represent plants as rigid cylinders with vertical structures, which influence the vertical flow and turbulence profiles within and above the vegetation canopy (Nepf and Vivoni, 2000; Nepf, 1999; Temmerman et al., 2005; Uittenbogaard, 2003). This approach has been validated with field observations to be effective for cohesive sediment transport in tidal marshes (Temmerman et al., 2005).

In this study, we applied the Delft3D model to explore the impacts of hurricanes on the hydrodynamics and morphological changes of coastal wetlands through application to the Wax Lake Delta (WLD) during Hurricane Rita (2005). WLD, a prograding lobe of the Mississippi River Delta system, is characterized by low elevation and abundant freshwater wetlands (Howes et al., 2010), which are vulnerable to frequent hurricane events in this area, e.g., 16 major hurricanes (Category 3, 4 or 5 on the Saffir-Simpson scale) influenced Louisiana between 1941 and 2008 (Bunya et al., 2010).
A huge amount of valuable coastal wetlands on the Mississippi River Delta are missing due to decreased fluvial sediment loads, and accelerated compaction caused by mining activities and natural sediment compaction (Blum and Roberts, 2009; Syvitski et al., 2009). A lot of studies have tried to save the delta from disappearance, including open new outlets to induce water and sediment to shallow areas, forming deltas similar as the WLD (Kim et al., 2009). This study investigates how the new-formed deltas respond to hurricanes, which shares light on the protection and management of the coastal wetlands along the Louisiana coast. The roles of storm surges, hurricane tracks, waves, and vegetation on influencing morphological changes of wetlands during hurricane events under natural status are also evaluated through numerical experiments.

3.3 REGIONAL SETTING

3.3.1 WAX LAKE DELTA (WLD)

WLD is a new fluvial depositional lobe at the Atchafalaya Bay, along the Gulf of the Mexico in Louisiana. It was formed by rapid deposition of sediment following the construction of a canal (the Wax Lake Delta Outlet, or WLO) in 1941, which connected the upstream Six-Mile Lake and the Atchafalaya Bay (Figure 3.1B). The accumulation of fluvial sediment near the mouth of the outfall canal led to the formation and progradation of the subaqueous delta. The delta became subaerial in 1973 after a large river flood (peak discharge of 20,000 m$^3$ s$^{-1}$, compared to the mean discharge of 5,781 m$^3$ s$^{-1}$ from 1935 to 2007) that transported a significant amount of sediment to the river mouth, both from fluvial sediment input and the erosion of sediment from the canal (Roberts et al., 1997).

Fluvial water and sediment discharge have strong seasonal variations for the WLD system, with annual river floods occurring in spring (Mossa and Roberts, 1990). The mean flood velocity is approximately 2 – 2.5 times higher than the non-flood velocity for an average-discharge year, and suspended sediment concentration during floods can be up to 20 times higher than that in non-flood conditions (DuMars, 2002). The WLO transports approximately 30 MT of sediment per year to the
shallow (~2m deep) Atchafalaya Bay (Kim et al., 2009), leading to a yearly area increase of 0.81 km² (FitzGerald, 1998; Roberts et al., 2003). The mean fluvial sediment layer (topset) thickness of the delta is 2.4 m, of which 70% is sand (Roberts et al., 1997).

Tides and waves are relatively mild in the Atchafalaya Bay under typical conditions. The tidal range is around 0.3 m (DuMars, 2002), and the significant wave height is less than 0.5 m (Wright, 1977). This area is exposed to two major weather systems: winter cold fronts and tropical storms. Cold fronts occur every 4 – 7 days from October through March, or 20 – 30 times per year in total (Chuang and Wiseman, 1983). During cold fronts, wind speed rises to >10 m/s (Walker and Hammack, 2000). The high wind and wind-generated waves can effectively re-suspend and transport the bottom sediment, contributing to ~15% of the fluvial sediment transporting outside the bay system every year (Roberts et al., 1997). Tropical storms affect this area primarily in summer and fall, producing significant damage to coastal wetlands (Barras, 2006; Howes et al., 2010).

The WLD has been colonized by freshwater species on the high-lying areas of the islands due to the freshwater sequestration in this river-dominated system. The plant community is in a dynamic early successional phase, and its composition is mainly determined by fluvial discharge, elevation, disturbance, salinity, organic content of sediment, and nutrient availability (Holm and Sasser, 2001; Rejmánek et al., 1987; Shaffer et al., 1992). The dominant species in the freshwater wetlands of the WLD include Salix nigra, colocasia esculenta, Polygonum punctatum, Vigna luteola, Schoenoplectus americanus, Alternanthera philoxeroides, Leersia oryzoides, Sagittaria platyphyla, Nelumbo lutea, Potamogeton nodosus (Viparelli et al., 2011).

3.3.2 Hurricane Rita

Hurricane Rita (2005) is the fourth most intense Atlantic hurricane ever recorded in the Gulf of Mexico (Fig. 3.1A, Beven et al., 2008). Rita initially formed as a tropical depression on September 17th, 2005 near the Turks and Caicos Islands and moved westward through the Florida Straits. The system intensified with time but remained a Category 2 hurricane until it entered the Gulf of Mexico. Rita
strengthened rapidly when it passed over the warm Loop Current during midday September 21st, reaching Category 5 at 1800 Coordinated Universal Time (UTC) with a minimum barometric pressure of 897 mbar and a maximum wind speed of 175 mph. The system weakened on September 23rd as it approached the north-central coast of the Gulf of Mexico. Simultaneously, the track changed from westerly to northwesterly. Rita made landfall in western Cameron Parish, Louisiana, 120 km west of the WLD, on September 24th as a Category 3 hurricane, causing a storm surge of 4 – 5 m close to the landfall area (Williams, 2009). Significant inundation and wetlands damage was observed along the Louisiana coast (Bunya et al., 2010; Howes et al., 2010; Rego and Li, 2010). The high water mark showed storm surge was more than 3 m in Vermilion Bay (Fig. 3.1B, Rego and Li, 2010). After entering the inland area, Rita weakened, moving northward through Texas and western Louisiana, eventually turning northeastward and disappearing when it merged with a cold front on September 26th, 2005.

![Figure 3.1 Bathymetry and topography of model domains](image)

*Figure 3.1 Bathymetry and topography of model domains*
Note: Fig. 3.1A shows the largest model domain: Gulf of Mexico (GoM). The dotted line displays the track of Hurricane Rita, and the colors demonstrate the hurricane intensities. Round red dots show six National Data Buoy Center (NDBC) stations for wave observations (significant wave heights and peak wave periods). Black dots show four NOAA tidal stations along the Louisiana and Texas coasts for storm surge observations. Figs. 3.1B and 3.1C show the detailed bathymetry and topography of the Atchafalaya domain and the Wax Lake domain. 1, 2, 3, 4 in Fig 3.1B are four river flows that are included in simulations: the Wax Lake Outlet, Atchafalaya River, Jaws River, and Vermilion River, respectively. I is Atchafalaya Bay, II is West Cote Blanche Bay, and III is Vermilion Bay. The black line in Fig. 3.1C shows the deltaic area used in calculations of sediment balance.

3.4 Method

3.4.1 Model description

The Delft3D software package, a widely used computational fluid dynamics model (Lesser et al., 2004), is applied for this study. The Delft3D FLOW module uses a finite difference solution of the three-dimensional shallow water equation and the k-ε turbulence closure model (Rodi, 1980) to compute flow characteristics under the hydrostatic pressure assumption. The model adopts the flooding and drying algorithm to determine the computational grid cells in the way that a cell is included in the calculation when its water depth is higher than 0.01 m. The Morphology module (MOR) allows sediment transport (both cohesive and non-cohesive) to be calculated simultaneously with flow computation using the advection-diffusion equation (for suspended sediment load, not for bedload). Bedload is calculated with Van Rijn (2000) and Van Rijn et al. (2003) equations. Bed level is updated every time step based on hydrodynamic results using the Exner equation (WL|Delft3D Hydraulics, 2011). The third-generation fully spectral wave module SWAN is coupled with Delft3D FLOW to calculate wave parameters with the discrete spectral action balance equation (WAVE|Delft3D Hydraulics, 2011). A 3D vegetation routine is also incorporated in the model framework to explore the impact of vegetation on flow hydrodynamics and sediment transport. More detailed explanations of the model structures are given in Lesser et al. (2004) and will not be reiterated here. However, the 3D vegetation routine will be described below as it is an important component of this study which has not been widely applied.
The FLOW model accounts for the influence of vertical plant structures on drag and turbulence, expressed as a flow drag term in the momentum equation (Eq. 1) and extra source terms for turbulent kinetic and dissipation energy equations (Eq. 2 and Eq. 5). The flow drag force \( F(z) \) in N m\(^{-3} \) caused by plant vertical structures at height \( z \) (m) is given by:

\[
F(z) = \frac{1}{2} \rho n(z) C_D \phi(z) |u(z)| u(z),
\]

where \( \rho \) is water density \([\text{kg m}^{-3}]\), \( n \) is plant density \([\text{m}^{-2}]\), \( C_D \) is the drag coefficient [-], \( \phi(z) \) is the plant diameter at height \( z \) [m], and \( u \) is the time-averaged horizontal flow velocity \([\text{m s}^{-1}]\).

Plant vertical structures generate an extra source term of turbulent kinetic energy \( k \) \([\text{m}^2 \text{s}^{-2}]\) in the turbulence calculating \( k-\varepsilon \) equation:

\[
\left( \frac{\partial k}{\partial t} \right)_{\text{plants}} = \frac{1}{1 - A_p(z)} \frac{\partial}{\partial z} \left\{ \left( 1 - A_p(z) \right) \left( \nu + \nu_T / \sigma_k \right) \frac{\partial k}{\partial z} \right\} + T(z),
\]

where \( A_p(z) \) [unitless] is the horizontal cross-sectional plant area per unit area at height \( z \):

\[
A_p(z) = \left( \frac{\pi}{4} \right) D^2(z)n(z),
\]

and \( T(z) \) \([\text{m}^2 \text{s}^{-3}]\) is the work spent by the fluid at height \( z \):

\[
T(z) = F(z)u(z)/\rho,
\]

the parameter \( \nu \) \([\text{m}^2 \text{s}^{-1}]\) is the molecular fluid viscosity, \( \nu_T \) \([\text{m}^2 \text{s}^{-1}]\) is the eddy viscosity, \( \sigma_k \) [unitless] is the turbulent Prandtl-Schmidt number for turbulence self-mixing process (\( \sigma_k=1 \)), \( D \) is plant diameter [m], and \( t \) is time [s].

The extra source term in the turbulent energy dissipation equation generated by plant vertical structures is expressed as \( \varepsilon \) \([\text{m}^2 \text{s}^{-3}]\):

\[
\left( \frac{\partial \varepsilon}{\partial t} \right)_{\text{plants}} = \frac{1}{1 - A_p(z)} \frac{\partial}{\partial z} \left\{ \left( 1 - A_p(z) \right) \left( \nu + \nu_T / \sigma_\varepsilon \right) \frac{\partial \varepsilon}{\partial z} \right\} + T(z)\tau_\varepsilon^{-1},
\]

where \( \sigma_\varepsilon \) [unitless] is the turbulent Prandtl-Schmidt number for mixing of small-scale vorticity (\( \sigma_\varepsilon=1.3 \)), and \( \tau_\varepsilon \) [unitless] is the minimal dissipation time scale of free turbulence \( \tau_{free} \):

\[
\tau_{free} = \frac{1}{c_2 \varepsilon \left( \frac{k}{\varepsilon} \right)}
\]
The coefficient $c_{2E}$ is determined by calibration (suggested value of 1.96). The eddy dissipation timescale ($\tau_{\text{veg}}$) between plants is calculated as:

$$\tau_{\text{veg}} = \frac{1}{c_{2E}c_{\mu}} \left(\frac{L}{T}\right)^{1/3},$$

where the coefficient $c_{\mu}$ equals 0.09. The sizes of these eddies are limited by the smallest distance between plant stems $L(z)$ [m]:

$$L(z) = C_l \left\{ \frac{1-A_P(z)}{n(z)} \right\}^{1/2},$$

where the coefficient $C_l$ is used to reduce the geometrical length scale to the typical volume averaged turbulence length scale.

### 3.4.2 MODEL INPUT PARAMETER SETTINGS

The Delft3D FLOW, WAVE and MOR modules are coupled and applied to three nested domains in order to acquire a detailed hydrodynamic field in the WLD area. The largest domain (GoM) covers a large part of Gulf of Mexico with a resolution of 0.020 degrees; the second domain includes the coastal shallow areas of the WLD (Atchafalaya, Vermilion, and Cote Blanche Bay) and surrounding low-elevated wetlands with an average resolution of 200 m; the third domain (WLD) covers the WLD and surrounding wetlands, with an average resolution of 50 m (Fig. 3.1). The GoM model is triggered by tides at the ocean boundaries which is extracted from the TPXO 7.2 Global Inverse Tide Model (http://volkov.oce.orst.edu/tides/TPXO7.2.html), and an equal-distance wind field with a spatial resolution of 0.05° and a temporal resolution of 15 minutes, achieved from the combination of NOAA Hurricane Research Division Wind Analysis System (H*WIND, Powell et al., 1998), and the Interactive Objective Kinematic Analysis (IOKA) kinematic wind analysis Cox et al., 1995). Bathymetry is derived from the Louisiana Virtual Coast Data Archive (http://virtual-coast.c4g.lsu.edu/), in which NOAA’s bathymetry sounding database, the Digital Nautical Charts database, and the 5-minute gridded elevations/bathymetry for the world (ETOPO5) database are combined (Mukai et al., 2002; National Geophysical Data Center, 1988; National Ocean Service, 1997; U.S. Department of Defense, 1999).
Three types of sediment: sand, silt and mud, are included in the model and seabed properties in the GoM domains are extracted from seabed sediment distribution map (dbSEABED) reported by Jenkins (2002). In deep area the bed roughness (Manning’s n) is calculated through an empirical equation:

\[ n = 0.015 + \frac{0.01}{|\text{depth}|}, \quad (3.9) \]

and this equation has been validated in coastal ocean simulations (Xing et al., 2012). For the WLD domain, a more detailed bathymetry and island topography obtained from 1998 hydrological survey with an averaged resolution of 80 m are used, combining with Light Detection and Ranging (LIDAR) survey data for overbank areas that were not covered in the hydrographic survey (USACE, 2010). The river discharges at fluvial boundaries are acquired from observed daily data from USGS (Baumann et al., 2005), and sediment concentrations are set up with annual averaged values from (DuMars, 2002). The major morphological parameters used for the Atchafalaya and WLD domains are shown in Tab. 3.1.

Table 3.1 Morphological parameter settings for Atchafalaya and WLD domains.

<table>
<thead>
<tr>
<th>D50 (m)</th>
<th>Mud (N/m²)</th>
<th>Bed layer thickness (m)</th>
<th>Bed roughness (Chezy: m¹/₂ s⁻¹)</th>
<th>Bed sediment proportion (Sand:Silt:Mud)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>Silt</td>
<td>CrERO</td>
<td>CrDEP</td>
<td>Channels, Ocean</td>
</tr>
<tr>
<td>1.0E-04</td>
<td>3.0E-05</td>
<td>1.0E+00</td>
<td>1.0E-02</td>
<td>20</td>
</tr>
</tbody>
</table>

Note: \( CrERO \) is critical shear stress for erosion of cohesive sediment (mud), and \( CrDEP \) is critical shear stress for deposition of cohesive sediment (mud)

3.4.3 DATA ANALYSES

We apply a widely-used, low-pass digital filter program to eliminate the influence of tides, through which waves with periods longer than 12 hours are filtered from the simulated results (Thompson, 1983; Walters and Heston, 1982). In this way, we obtain the water level and flow velocity caused by interactions of hurricanes and rivers. Residual currents (\( R_j \) in m s⁻¹) are calculated for each grid cell \( j \) on the WLD using the following equation for both \( x- \) and \( y- \) direction:

\[ R_j = \frac{1}{dep_j} \sum_{i=1}^{i=n} v_{i,j} \cdot d_{i,j}, \quad (3.10) \]
where \( \text{dep}_j [\text{m}] \) is mean water depth at cell \( j \), \( v_{i,j} [\text{m} \text{s}^{-1}] \) is velocity at time step \( i \), \( d_{i,j} [\text{m}] \) is water depth at time step \( i \), and \( n \) [unitless] is the total number of calculating time steps.

The numerical settings for Hurricane Rita are assigned as the base case, and experiments EX1, EX2, EX3, EX4, EX5 are designed to study the influence of waves, fluvial input, aboveground vegetation, roots, and hurricane tracks on delta morphology, respectively. Because there was no available data for vegetation distribution and properties on the WLD, and studies show that the distribution of vegetation is significantly dependent on water depth (Shaffer et al., 1992), the above ground vegetation is set to be uniform on the delta where water depth is lower than 0.2 m. Stem height is set to 1.0 m (plants height varies from 0.3 m to maximum 3.4 m, mostly ~1m (Carle et al., 2013) and density to 150 stems per \( \text{m}^2 \) (stem density of the dominant 3 Sagittaria species was estimated to be ~150 stems per \( \text{m}^2 \) at the coast of Louisiana during September (Martin and Shaffer, 2005)). Roots can significantly increase soil strength, especially for cohesive sediment, but because the WLD is dominated by sand (average 70%), the influence of roots is limited. The influence of roots within the model domain is explored by slightly increasing the critical shear stress for erosion of cohesive sediment from 1 N \( \text{m}^{-2} \) to 1.5 N \( \text{m}^{-2} \) (Tab. 3.2).

### Table 3.2 Numerical model settings of different experiments

<table>
<thead>
<tr>
<th>Hurricane Track</th>
<th>Fluvial process (river)</th>
<th>Waves</th>
<th>Aboveground Vegetation</th>
<th>Roots</th>
</tr>
</thead>
<tbody>
<tr>
<td>Base</td>
<td>Side</td>
<td>Y</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>EX1</td>
<td>Side</td>
<td>Y</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>EX2</td>
<td>Side</td>
<td>N</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>EX3</td>
<td>Side</td>
<td>Y</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>EX4</td>
<td>Side</td>
<td>Y</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>EX5</td>
<td>Direct</td>
<td>Y</td>
<td>Y</td>
<td>Y</td>
</tr>
</tbody>
</table>

Note: Side track indicates that hurricane makes landfall on one side of the delta (here we use the same settings as Hurricane Rita), and direct track means that the hurricane makes landfall on the delta.

#### 3.4.4 MODEL VALIDATION

The observed waves (significant wave height and peak period) and water levels derived from the National Data Buoy Center (NDBC [http://www.ndbc.noaa.gov/]) and NOAA coastal tides stations (NOAA/NOS/C0-OPS:

http://tidesandcurrents.noaa.gov/noaatidepredictions/NOAATidesFacade.jsp?Stationid=8762075),

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Figure 3.2 Comparisons of significant wave heights and peak wave periods at six NOAA buoy stations (See Fig. 3.1A), showing that the modeled results catch both the trends and the magnitudes of the observed wave parameters respectively, (Fig. 3.1A) are used to test the model’s effectiveness. The comparisons between simulations and observations (Fig. 3.2, 3.3) indicate that the modeled results match well the realistic wave dynamics in the deep ocean and the storm surges in the coastal areas during Hurricane Rita. The slight difference between observation and simulation might be caused by uncertainty in input parameters, e.g., winds and bathymetry. The simulated wave height is within a bias (RMSE) of 1 m (mean value of 3.2 m), and the
simulated peak wave periods have a RMSE of 2s (mean value of 6 s). The filtered water levels from simulation are overall smaller than that from observations but match the peak surges, with a RMSE bias of 0.16 m (mean value of 0.24 m) (Fig. 3.4). The modeled maximum water level in Vermilion Bay also shows a great consistency with the high water marks measured by Federal Emergency Management Agency (FEMA) during Hurricane Rita (measured 3.6 m, compared to observed 3 – 4 m, FEMA, 2006).

Because of data scarcity, the Delft3D MOR module is not validated with field observations during Hurricane Rita. Instead, the parameter sets are derived from the study of Meselhe et al. (2015, in preparation), in which these parameters have been validated against successive mapping of the WLD growth from 1998 to 2014 where the morphological evolution produced by the model compared favorably with the mapping data.

Figure 3.3 Comparisons of water levels during storm surge caused by Hurricane Rita at four coastal stations (See Fig. 3.1A), indicating that the model successfully simulates the water level rise during the hurricane event.
Figure 3.4 Evaluation of model performances in simulating significant wave heights, peak wave periods, and storm surges, demonstrating that the model is efficient in modeling significant wave heights, but the simulated peak wave periods and storm surges are slightly smaller than observations.

3.5 SIMULATED HYDRODYNAMICS AND MORPHOLOGICAL CHANGES DURING HURRICANE EVENTS

Hurricanes cause significant water level changes and high waves, which disturb coastal hydrodynamics and cause notable changes in coastal morphology (Allison et al., 2005; Yang et al., 2003). Storm surges are determined by the interactions between storms and coastal topography (Chen et al., 2008). Hurricane Lily (2002) that made landfall 80 km to the west of the WLD as a Category 1 hurricane produced a significant wave height of more than 2 m and a storm surge of 2 m in the Atchafalaya inner shelf (Allison et al., 2005). Typhoon Paibaian (2000), which made landfall at southern Shanghai as a Category 1 Typhoon, caused tidal flat and lower marsh to be eroded by maximum 0.21 m, and the middle-to-upper marsh and the subtidal channels were dominated with significant accretion in the Yangtze Delta (Yang et al., 2003). Our numerical results indicate that Hurricane Rita that made landfall 120 km to the west of WLD as a Category 3 hurricane produced 5 times higher water levels and flow velocities, and 3 times higher wave heights than those under normal conditions, leading to a pattern of erosion on the islands and deposition in the channels. In general, the eastern banks of islands and channels that located on the west of islands have higher morphological changes, due to the northwestern currents.
Rita formed an atmospheric pressure head at the hurricane center, leading to a water level rise of 1.0 m in the deep ocean. As the hurricane system approached the shallow area, the reduction in water depth limited the strong vertical momentum diffusion, leading to significant increase in storm surge because of the conservation of the potential vorticity of the mound (Jelesnianski et al., 1992). Combined with local bathymetric reflections and onshore-directed winds, the surge height raised to 4 – 5 m when the hurricane made landfall (Williams, 2009). Our simulation shows that Hurricane Rita drastically disturbs the hydrodynamic field in the WLD area. Before the hurricane center approaches the coast, the anti-clockwise wind structure produces offshore-directed winds in the north of hurricane system, leading to offshore-directed water flow and significant drop of water level in the WLD area (-2.0 m, Fig. 3.5A). Water then flows back toward the coastal area when the winds changes direction to southeast during the storm surge, resulting in water level rise of 2.5 m (Fig. 3.5B). The simulated water level variations during Hurricane Rita event are 4 – 5 times higher than the typical tidal range (+/- 0.5 m, Fig. 3.5C), and flow velocity reaches 2.4 m s^{-1}, 5 times higher than velocities under non-storm conditions (0.5 m s^{-1}). Hurricane Rita produces a maximum significant wave height of 1.4 m (Fig. 3.5D), ~3 times higher than the wave heights under normal conditions (0.5 m from Wright (1977), Fig. 3.5E). Waves are higher in river channels than on the shallow islands because wave energy dissipates in shallow areas (Fig. 3.5D).

Floodwaters spread beyond the channel banks and overtop the islands and coastal wetland areas during the hurricane simulation. The flow velocities are higher on the island tops than within the channels during floodwater propagation because the floodwater propagating northward counteracts with the fluvial discharge flowing southward within the channels. The WLD is then dominated by a uniform flow toward northwest, following the wind field (Fig. 3.5B, 3.6A). Residual currents flow towards northwest in most of the WLD area following the wind direction (Fig. 3.6A, 3.7A), which are higher on the islands and lower in the channels, and the highest values occur on the eastern banks of islands due to the significant decrease in water depth from channels to islands during the northwestern directed flow (Fig. 3.7A). The residual current field significantly influences sediment transport, leading to the pattern of erosion on islands and deposition in channels, which is intensified by waves as waves increase shear stress in
shallow areas (Fig. 3.8A, B). Erosion is the most severe on the eastern banks of islands (windward), while accretion is highest in the channels that are located on the west of islands (leeward) due to the significant decrease in shear stress from islands to channels during the northwestern directed flow (Fig. 3.8A). Hurricane Rita leads to erosion in the calculated WLD area (Fig. 3.1C), and 500,000 m$^3$ of seabed sediment is removed in 100 hours during the storm surge (1600 UTC, September 21$^{st}$ to 1900 UTC, September 25$^{th}$).

Figure 3.5 Simulated water levels and significant wave heights on the WLD at different time steps. A): water level drops before storm surge; B): water level rises at peak storm surge; C): water level rises during high tides; D): significant wave height distribution at peak storm surge; E): significant wave height distribution during high tides. Water level drops significantly before hurricane system approaches the deltaic area due to the dominated offshore winds during that time period. Water level then rises when the hurricane pushes water onto the delta. The water level rise caused by Hurricane Rita is ~5 times higher than that under normal condition (dominated by tides). The significant wave height caused by Hurricane Rita is ~3 times higher than that under normal condition.
Figure 3.6 Time series of wind vectors in the center of the WLD from 0000 UTC September 21st to 0100 UTC, September 25th, 2005 for A): base case; B): route 1 (EX5). Line lengths show wind speeds, and line directions show wind directions. The base case is dominated by onshore-directed wind, and EX5 is dominated by offshore-dominated winds.

Figure 3.7 Simulated residual current distributions on the WLD within 100 hours for A): base case; B): EX1, which excludes waves; C): EX2, which excludes fluvial input; D): EX3, which includes aboveground
vegetation; E): EX4, which includes aboveground vegetation and roots; F): EX5, which applies a directly-landfalling hurricane track. Waves and fluvial input have slight impacts on residual currents, but aboveground vegetation significantly decreases the high residual currents on deltaic islands, and the high currents concentrate on the southwest of the delta. Roots have a minor impact on residual currents. Hurricane tracks significantly change the distribution pattern of residual currents on the WLD.

![Figure 3.8 Simulated morphological changes of the WLD after hurricanes for: A): base case; B): EX1, which excludes waves; C): EX2, which excludes fluvial input; D): EX3, which includes aboveground vegetation; E): EX4, which includes aboveground vegetation and roots; F): EX5, which applies a directly-landfalling hurricane track. The deltaic islands are dominated by erosion, and channels are dominated by deposition. Waves are the most critical factor contributing to total erosion on the WLD. Fluvial input has a minor influence on deltaic morphological changes during Hurricane Rita. Aboveground vegetation dramatically decreases the erosion on vegetated areas. Roots have a minor impact on sediment transport pattern. The shifts in wind patterns significantly change the pattern in morphological changes when a different hurricane track is applied.](image)

3.6 WAVE AND FLUVIAL INFLUENCE ON MORPHOLOGICAL CHANGES DURING HURRICANE EVENTS

Hurricane driven flow and waves are the major causes of morphological changes during extreme events. For instance, Typhoon Paibaian caused the suspended sediment concentration on tidal flats to rise 10-20 times higher than under normal conditions (Yang et al., 2003), and typhoons are stated to be responsible for shaping tidal flat profiles on the Yangtze Delta (Fan et al., 2006). Wave breaking and energy dissipation on rough beds produce significant erosion in shallow areas, while the high current
shear stress caused by strong flow also contributes to sediment re-suspension. The interactions between flows and local topography determine the pattern of sedimentation-erosion at the coastal area, while waves intensify sediment erosion, particularly in shallow areas. The influence of fluvial processes on delta morphology during extreme conditions such as hurricanes is minor.

We quantitatively evaluate the impacts of waves and riverine inflows on delta morphological changes during hurricane events by numerically deactivating the wave module in EX1 and excluding fluvial input in EX2. Results indicate that waves have a minor influence on the residual current pattern (Fig. 3.7A, 3.7B), but significantly increase the amount of erosion (Fig. 3.9A). When the wave module is deactivated, the area that has the erosion of bed is larger than 0.2 m on islands decreases significantly and the maximum erosion reduces from 0.3 m to 0.2 m compared to the base case (Fig. 3.8A, 3.8B). The reduced sediment erosion leads to smaller suspended sediment loads in the water column, and consequently decreases the amount of accretion in the channels when water flows northwesterly from islands to channels. The total erosion in the calculated area decreases by 48%, and the total deposition decreases by 8%. As a result, the sediment balance in the WLD area changes from net erosion of 500,000 m$^3$ to net deposition of 100,000 m$^3$ (Tab. 3.3).

Although the WLD is a river-dominated system, the fluvial input has only a minor influence on the hydrodynamics and sediment transport pattern during the Hurricane Rita simulation. The only significant difference between EX2 and the base case is the direction of residual currents in the channels, which change from downstream to upstream when fluvial input is not included (Fig. 3.7A, 3.7C). EX2 produces minor increases in both accretion in the channels and erosion on the islands compared to the base case (Fig. 3.9B) and the total amount of erosion and deposition increase by 2% and 1%, respectively. Runoff therefore mainly influences the residual currents and sediment transport in channels, and its influence on morphological changes of the WLD is minor under hurricane conditions.
Table 3.3 Numerically simulated total sediment balance on the Wax Lake Delta after hurricanes

<table>
<thead>
<tr>
<th>Hurricane Rita Base Case</th>
<th>Total change (m$^3$)</th>
<th>Total erosion (m$^3$)</th>
<th>Total accretion (m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EX1</td>
<td>-500,000</td>
<td>-1,400,000</td>
<td>900,000</td>
</tr>
<tr>
<td>EX2</td>
<td>100,000</td>
<td>-700,000</td>
<td>800,000</td>
</tr>
<tr>
<td>EX3</td>
<td>-300,000</td>
<td>-700,000</td>
<td>400,000</td>
</tr>
<tr>
<td>EX4</td>
<td>-200,000</td>
<td>-700,000</td>
<td>400,000</td>
</tr>
<tr>
<td>EX5</td>
<td>-200,000</td>
<td>-700,000</td>
<td>500,000</td>
</tr>
</tbody>
</table>

Figure 3.9 Differences in simulated sedimentation and erosion patterns between A): (EX1 - base case), which shows winds and waves are the most critical factor contributing to erosion on deltaic islands and erosion in channels; B): (EX2 - base case), which shows fluvial input slightly decreases deposition in channels and erosion on islands; C): (EX4-EX3), which show roots have small impact on morphological changes of the deltaic area, but decrease erosion on the upstream wetlands

3.7 VEGETATION WEAKENS THE INFLUENCE OF HURRICANES ON COASTAL WETLANDS

Vegetation increases the flow resistance of wetland areas, leading to flow amplification and erosion over unvegetated areas and flow reduction and deposition over vegetated areas on tidal flat (Temmerman et al., 2005). Vegetation also effectively attenuates waves, reduces surge height and decreases area that is inundated, depending on the coastal topography and storm properties (Augustin et al., 2009; Wamsley et al., 2010). Yang et al. (2003) observed deposition in vegetated marshes and erosion between the edges of unvegetated tidal flat during Typhoon Jelawat (2000) and Fan et al. (2006) found that mudflat erosion during typhoons is highly correlated to local vegetation. Not surprisingly, vegetation
decreases residual currents and the amount of sediment transport on wetlands during hurricane events. The underground roots increase soil strength and protect bed layers from erosion (Smith, 1976). However, Howes et al. (2010) observed significant erosion due to the existence of roots (when roots are pulled out because of strong shear stress, they carry along a block of soil and form rootballs) in coastal wetlands after Hurricanes Katrina and Rita, indicating that roots might also cause wetland erosion in extreme events.

In this study, we design an experiment (EX3) with uniformly distributed plants on the high-elevated islands of the WLD and surrounding wetlands (water depth < 0.2 m) to study the impacts of vegetation on delta hydrodynamics and morphology during hurricane events. EX3 produces a similar pattern of high surges in the northwestern part of the WLD as the base case, caused by the wind field and elevated coastal wetlands in the north of the WLD (Fig. 3.6A, 3.10A, 3.10B). However, the propagation of floodwater is slowed by vegetation, leading to wide distributed high surges on the deltaic area in the EX3, compared to the base case at the same time step (Fig. 3.10A, 3.10B). High velocities are restricted in deep water to the southwest of the islands (Fig. 3.10B). Vegetation does not change the distribution pattern of wave heights, but decreases the magnitude of significant wave height by 0.1 m on islands and increases its magnitude by a maximum of 0.23 m in the channels (Fig. 3.10C) during the flood wave propagation. Vegetation also significantly decreases the magnitude of residual current on islands from 0.3 – 0.6 m s\(^{-1}\) to less than 0.1 m s\(^{-1}\) on the eastern banks of islands, but increases the magnitude of residual current in channels by a maximum of 0.16 m s\(^{-1}\) (Fig. 3.7A, 3.7E). The changes in both waves and residual current pattern are favor of decreasing sediment erosion on islands and deposition in channels. When aboveground vegetation is included in the simulation, the maximum erosion on the eastern side of islands (windward) decreases from 0.3 m to 0.1 m, and the accretion in the channels (leeward) decreases from 0.1 – 0.2 m to less than 0.05 m (Fig. 3.8A, 3.8E). The total amount of sediment that is eroded and deposited in the WLD area decreased by 50% and 51%, respectively (Tab. 3.3). Thus, the existence of vegetation can significantly reduce morphological changes on wetlands during hurricane events.
Winter cold fronts and tropical storms are the major events causing erosion on the islands of the WLD (Roberts et al., 1997) and the existence of roots might decrease the erosion during these events. In this study, we explore the influence of roots by increasing critical shear stress for mud erosion on the vegetated wetlands in EX4, because no root ball erosion is observed on the WLD (Twilley group, LSU, personal contact). The simulated residual currents show identical patterns between EX3 and EX4 (Fig. 3.7D, 3.7E), indicating that roots have a minor influence on flow field during hurricane simulations. Surprisingly, the simulated morphological patterns are also identical for EX3 and EX4, except for a slight decrease in both erosion on islands and deposition in the channels (Fig. 3.9C). The reason is that the current-wave shear stress has been significantly decreased on islands when vegetation is included, and the resulting shear stress is mostly smaller than the critical shear stress for erosion in EX3. As a result, the increase in erosional critical shear stress in EX4 would not create a big difference in sedimentation-erosion pattern. However, the rootball erosion might be critical in determining sedimentation-erosion patterns on other parts of wetlands along the Louisiana coast where sediment-roots interactions are more important, e.g. on the saltmarsh near New Orleans where wetland is mainly composed of mud and organic matters (Howes et al., 2010).

Figure 3.10 Tide-filtered water level and velocity distribution on the WLD for A): base case; and B): EX3, which shows vegetation slows down the propagation of flooding water, leading to high water level on the high-elevated wetlands at the northwest of the WLD. Fig. 3.10C shows the differences in significant wave heights between the base case and EX3, demonstrating that vegetation decreases wave heights on islands.
3.8 A SIDETRACK HURRICANE CAUSE LARGER MORPHOLOGICAL CHANGES THAN A DIRECT TRACK

Hurricane tracks have been reported to be critical in determining the magnitude of coastal damages (Chen et al., 2008; Coch, 1994; Dietrich et al., 2010; Weisberg and Zheng, 2006). A coast-parallel track that keeps the weaker left side of a storm against the coastline has a small impact on the coast, while a coast-normal track that strikes with the powerful right side of storm can cause significant damage to coastal areas, such as Hurricane Hugo (1989) and the Long Island-New England Hurricane (1938) (Coch, 1994). Weisberg and Zheng (2006) reported that hurricane storm surges in Tampa Bay are significantly influenced by landfall position where a sidetrack hurricane that makes landfall in the north of the bay mouth causes the bay mouth to experience the maximum winds and the highest storm surge, while a storm that made landfall at the bay mouth yields the smallest surge. Our study also shows that the variations in hurricane tracks cause great changes in wind and flow fields, consequently a sidetrack hurricane that makes landfall 120 km away on the western side of the study area produces higher storm surges and more sediment erosion at the coastal area than a direct track.

We investigate the influence of hurricane tracks on coastal wetland morphology by moving the wind field to the east so that the hurricane would make landfall directly on the WLD in EX5, based on the setting of Hurricane Rita. The magnitude of winds is similar for the base case and EX5, with a maximum speed of 30 m s\(^{-1}\). However, the main wind direction during the storm surge changes significantly from southeast to northwest (Fig. 3.6A, 3.6B), leading to dramatic differences in flow pattern and consequently morphological changes. Compare to the base case, EX5 produces smaller storm surges both on islands (1.1 m vs. 2.0 m) and in the channels (0.3 m vs. 1.6 m) because the study area is dominated by strong offshore-directed winds, leading to less water and sediment transport towards inland (Fig. 3.6B, 3.7F, 3.8F). The changes in the wind and flow fields result in weaker residual currents in EX5 (0.1 – 0.4 m s\(^{-1}\) vs. 0.3 – 0.6 m s\(^{-1}\) in the base case), which direct towards the southeast with high values occurring on the
western banks of islands, apart from the base case. Sediment transport is significantly weakened, too, both for erosion and deposition. Erosion occurs on the western side of the delta with a maximum value of 0.05m (vs. 0.3 m for the base case) and deposition occurs occasionally on the eastern side of islands (maximum 0.1 m vs. more than 0.2 m for the base case) due to the wind field (Fig. 3.6B, 3.8A, 3.8F). The total amounts of sediment erosion and deposition reduce to 50% and 47% of the base case, respectively. This analysis demonstrates the importance of hurricane tracks on morphological changes of coastal wetlands.

3.9 Model Uncertainty

Uncertainty analysis is critical in evaluating a model’s reliability, which determines whether a model could be used in decision-making applications (Li et al., 2013; Scheel et al., 2014; Radwan et al., 2004; Katz, 2002). However, uncertainty analysis is usually omitted in numerical studies due to the complexity of uncertainty analysis and lack of data, especially for complicated numerical models. Model uncertainty can be introduced by uncertainty in input and boundary data, model structure, and uncertainty propagation (Li, et al., 2013; Katz, 2002; Beck, 1987). Although Delft3D has been widely used in coastal studies, the studies on quantification of model uncertainty are rare. Giardino et al. (2011) evaluated the uncertainty of the Delft3D model structure and listed the most uncertain terms for the model due to simplification of model equations. Plüß and Kösters (2013) stated that surface sediment porosity influences the modeled sediment dynamics within Delft3D by changing sediment mobility. Scheel et al (2014) used the Probabilities Morphodynamic framework to evaluate the sensitivity of the model morphological results to input parameters such as d50. For this study, we use Dakota software to qualify model uncertainty caused by input parameters, including observational uncertainty (sand and silt grain size) and uncertainty caused by parameter estimation (critical shear stress for mud erosion and sedimentation). The analysis shows that the model is very sensitive to the uncertainty caused by
estimation of critical shear stress for erosion for hurricane case study, and that the sensitivity of the model to a specific parameter varies within different environmental settings.

Critical shear stress for mud erosion and deposition, determined by grain size, consolidation, and biological processes, are important in calculating morphodynamics (WL|Deltares Hydraulics, 2011; Sanford and Maa, 2001; Grabowski et al., 2011). However, due to lack in observations, most of studies use calibration or parameter estimation to determine these values, which vary significantly for different environments and models (Sahin et al., 2012; Xu et al., 2014; Wright, 1997). For the WLD area, we apply the range of critical shear stresses for erosion and deposition according to the studies of Hanegan (2011) and Eesehle et al (2015, in prep.), which have been validated in studying the long-term evolution of the WLD using Delft3D (Tab. 3.4). The largest possible ranges sand and silt grain sizes are used in this model (Tab. 3.4).

Given the complexity of the Delft3D model and the high computational intensity, we apply 3 samples for each of these parameters with Latin hypercube sampling method (Stein, 1987), leading to 81 simulations in total. Our simulation results show that critical shear stress for mud erosion is the predominant factor determining the total sediment balance in the study area, and the changes from 0.50 N m\(^{-1}\) to 1 N m\(^{-1}\) leads to model uncertainty of 332,000 m\(^3\) (standard deviation), with the mean value of 889,000 m\(^3\). The cells having the highest morphological changes also have a larger value of uncertainty (Fig. 3.8A, 3.11A) for the reason that variations in critical shear stress greatly influences cells that encounter strong flow and wave shear stress. The temporal uncertainty of the model is also explored through computing model uncertainty over time for the hurricane simulation, demonstrating that the model uncertainty varies significantly with changes in hydrodynamics, so that the uncertainty of the model is much higher under intensive coastal hydrodynamics than in comparatively peaceful periods (Fig. 3.6A (wind speed) and Fig. 3.11B). The significant increase in model uncertainty appears when water starts to flood the WLD area. The parameters silt grain size and critical shear stress for deposition become more important during peaceful periods. Our study demonstrates that uncertainty in input parameters
significantly influences model results, and their roles need to be evaluated based on specific environmental settings.

*Table 3.4 Uncertainties of major input parameters*

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Morphology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Contents and Sources</td>
<td>Parameters</td>
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<td>Silt Grain size (m)</td>
</tr>
<tr>
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</tr>
<tr>
<td>(62.5 – 200)×10⁻⁶</td>
<td>(8 – 62.5)×10⁻⁶</td>
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<tr>
<td>References</td>
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<td>0.05 – 0.1</td>
</tr>
<tr>
<td>[Pankow et al., 1990]</td>
<td>0.5 – 1</td>
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<tr>
<td>[Hanegan, 2011; Meselhe et al., 2015, in pre.]</td>
<td></td>
</tr>
<tr>
<td>[Hanegan, 2011; Meselhe et al., 2015, in pre.]</td>
<td></td>
</tr>
</tbody>
</table>

*Figure 3.11 Simulated spatial uncertainty (A) and temporal uncertainty (B) on the WLD due to Hurricane Rita, demonstrating that the areas with large morphological changes also have large uncertainty (Fig. 3.8), and that model uncertainty is highly correlated to the magnitudes of events so uncertainty is very low under normal condition, which increases significantly during Hurricane Rita*

### 3.10 CONCLUSIONS

The numerical simulation of Hurricane Rita (2005) with Delft3D demonstrates that hurricanes significantly influence the hydrodynamics and cause tremendous morphological changes on coastal deltas in a short time period. Hurricane Rita that made landfall 120 km to the west of the WLD as a Category 3 hurricane causes the maximum storm surges of 2.5 m and the maximum significant wave heights of 1.4 m, 3 and 5 times higher than normal conditions, respectively. Following the dominant southeastern winds,
the simulated residual currents direct towards northwest, and the maximum occur on the eastern banks of islands \((0.3 – 0.6 \text{ m s}^{-1})\). Combined with the higher wave shear stress in shallow areas, high erosion occur on the eastern banks of islands \((0.3 \text{ m})\), accompanied by high deposition in the channels located on the west of these islands \((0.1 – 0.2 \text{ m})\), caused by the significant decrease in shear stress from islands to channels during the northwestern flow. In total, \(500,000 \text{ m}^3\) of sediment is removed from the WLD area during Hurricane Rita based on our simulation.

Local topography and flow field determine the main patterns of residual current and sediment transport, while waves significantly intensify erosion in shallow areas, causing the amount of erosion to rise by 48%. Dense vegetation slows flow propagation and decreases flow velocities on the island tops, leading to flow amplification in the channels. Accordingly, amounts of both erosion on islands and deposition in channels are reduced by half when vegetation is included in the simulation. The increase of soil strength by roots has minor influence on the hydrodynamic and morphological patterns as vegetation has already significantly decreased the shear stress on islands. The major southeast (offshore)-directed winds during a direct-landfall hurricane induce smaller storm surge and less sediment erosion compared to the settings of Hurricane Rita. The magnitude of residual currents decreases from \(0.3 – 0.6 \text{ m s}^{-1}\) to \(0.1 – 0.3 \text{ m s}^{-1}\) on islands, and the amount of sediment that is eroded from the WLD area dropped to 56% of the original value.

The uncertainty analysis with Latin hypercube sampling method shows that uncertainty from parameter estimations, specifically critical shear stress for mud erosion, is the major factor causing model uncertainty during Hurricane Rita simulation, and that the model uncertainty under dramatic changes in hydrodynamic condition is larger than that under comparatively peaceful condition. Model uncertainty varies spatially i.e. cells on the island boundaries that encounter the strongest currents have the largest uncertainty.
3.11 ACKNOWLEDGMENTS

We respectfully acknowledge Stephanie Higgins for the kind assistance in editing the manuscript.

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4 THE IMPACTS OF HURRICANES AND WINTER COLD FRONTS ON THE MORPHOLOGICAL EVOLUTION OF THE WAX LAKE DELTA, LA

4.1 ABSTRACT

Hurricanes and winter cold fronts are the two major weather systems influencing the morphology of coastal wetlands along the Gulf of Mexico. Delft3D is applied to the Wax Lake Delta (WLD), Louisiana, to study the impact of coastal storms with different magnitudes and frequencies on wetland morphology. Our simulations of a strong cold front (mean wind speed of 11.4 m/s) and a single hurricane event (Hurricane Ike, made landfall 337 km to the west of the WLD as a Category 2 hurricane) demonstrate that although the hurricane event causes more sediment transport, many similarities exist between the two events: winds and waves significantly increase the amount of sediment transport; erosion occurs on islands; and there is a negative sediment balance for the WLD system (erosion). The simulations of 11 cold fronts from the 2008-2009 season show that cold fronts that cause significant water level variations produce stronger residual currents and move more sediment than events that cause minor water level variations, and that mean wind speeds are positively correlated with hourly-averaged erosion and deposition caused by winds and waves ($R^2$ of 0.94 for erosion and 0.81 for deposition). The relationships are applied to 29 cold front events with available wind data (in total 41 events were recognized for the season), leading to cumulative erosion of 1,900,000 m$^3$ on the WLD, higher than a single hurricane event such as Ike (412,000 m$^3$). The results illustrate that cold fronts are more critical in determining deltaic morphology than a single hurricane event. Vegetation that grows in the hurricane season significantly decreases the amount of erosion (from 412,000 m$^3$ to 308,000 m$^3$ for Hurricane Ike). Saline water intrusion threatens the survival of freshwater species on the WLD during hurricane events but would not influence the species during most of cold front events.
4.2 INTRODUCTION

Most wetlands of the United States are located along the coast of the Gulf of Mexico. These wetlands began forming when sea level stabilized 5,000 - 7,000 years ago (Bahr et al., 1983). Wetlands have critical ecological roles and serve as natural barriers to coastal storms (Bell et al., 2005; Chen et al., 2009; Shepard et al., 2011; Smith, 1976). Tropical cyclones and winter cold fronts have been classified as normal parts of the climatic regime of the Gulf of Mexico and are the major episodic marine disturbances for the development and maintenance of coastal wetlands (Bahr et al., 1983; Conner et al., 1989; Georgiou et al., 2014). Fifty-five tropical storms have made landfall along the Louisiana coast in the past 100 years (Stone et al., 1997), while winter cold fronts occur 20 – 30 times per year (Chuang and Wiseman, 1983). Hurricanes are observed to bring sediment into coastal wetlands (Turner et al., 2006; Yuan et al., 2004), but also cause damages in low-elevated wetlands and lead to shoreline retreat (Barras et al., 2005; Stone et al., 1997). The saline water intrusion into coastal wetlands during hurricanes influences plant community evolution and significantly imperils the survival of freshwater wetlands communities (Liu et al., 2008). Most studies generate a big-picture view of the impacts of hurricanes on coastal areas, and studies of how hurricanes influence the morphology of a specific coastal area are rare. In chapter 3 we quantitatively computed the morphological changes of a specific delta (Wax Lake Delta) during a significant hurricane event and illustrated that hurricanes cause erosion to the deltaic system. However, the importance of hurricanes on wetland morphology, which has to be evaluated by taking into account all the major events that impact delta morphology (such as cold fronts, hurricanes, and river flood events for some fluvial influenced areas), is still unclear.

Winter cold fronts are argued to cause more cumulative morphological changes to coastal wetlands because of their higher frequency compared to the more powerful and impressive but less frequent hurricanes (Moeller et al., 1993; Roberts et al., 1987). Single strong cold fronts have been observed and simulated to produce high flow velocities and sediment movement compared with non-storm periods (Cobb et al., 2008; Kineke et al., 2006; Walker and Hammack, 2000). However, each cold
front is distinctive with its own wind structure (both wind speed and direction) and influence on coastal morphodynamics. Studies that account for the cumulative impacts of both strong, typical events and smaller, atypical events during a cold front season are based on field observations, which could be of significantly different for specific areas. For instance, Cahoon and Turner (1987) observed accretion of wetlands after cold fronts and stated that cold fronts are important factors contributing to the accretion of saline marshes in southern Louisiana, while other studies observed erosion on the subaerially exposed Atchafalaya River delta during winter months and attributed it to winter cold front passages (Roberts et al., 1980; Rouse et al., 1978; Van Heerden, 1983). Mossa and Roberts (1990) demonstrated that winter cold fronts cause both erosion and deposition on the Atchafalaya River delta depending on fluvial sediment supply and grain size (Roberts et al., 1987, 1989). More studies are needed to evaluate the overall impacts of cold fronts on wetlands morphology, which should be studied within specific environments.

WLD is a new prograding lobe of the Mississippi River Delta system located at the center of Louisiana. Previous studies have focused on the impacts of fluvial processes on delta morphology and evolution (Kim et al., 2009; Olariu and Bhattacharya, 2006; Roberts et al., 1997; Shaw and Mohrig, 2014), and few explain the roles of coastal storms (Muller and Stone, 2001; Roberts et al., 1980). Sixteen major hurricanes (Category 3, 4 or 5 on the Saffir-Simpson scale) affected Louisiana between 1941 and 2008 (Bunya et al., 2010), while cold fronts transport 15% of Atchafalaya fluvial sediment to the deep Gulf of Mexico per year (Walker and Hammack, 2000). However, the influences of hurricanes and cold fronts on the deltaic morphology have not been well evaluated. In this study, we apply the fluid hydrodynamic model Delft3D (Lesser et al., 2004) to investigate the hydrodynamics and morphological changes of the Wax Lake Delta (WLD) during hurricane events and the cold front season between 2008 and 2009.
Figure 4.1 Bathymetry and topography of model domains (depths that are larger than zero are below sea level)

Note: A) the largest model domain: Gulf of Mexico (GoM), the sector black line includes the Atchafalaya domain. The dotted line displays the track of Hurricane Ike, and the colors indicate the hurricane intensities. Round dots show eight NDBC stations for wave observations (significant wave heights and peak wave periods). Black dots are four NOAA tidal stations along the Louisiana and Texas coasts. B) and C) show the detailed bathymetry and topography of the Atchafalaya and the Wax Lake Delta domains, respectively. 1, 2, 3, 4 in Fig. 4.1B are four rivers that are included in simulations: the Wax Lake Outlet, Atchafalaya River, Jaws River, and Vermilion River, respectively. I is Atchafalaya Bay, II is West Cote Blanche Bay, and III is Vermilion Bay. The gray line in Fig. 4.1C shows the deltaic area used in calculating the sediment balance.

4.3 REGIONAL SETTING

4.3.1 WAX LAKE DELTA (WLD)

WLD is a new fluvial depositional lobe belonging to the Atchafalaya Delta complex of the Mississippi River Delta system (Roberts, 1997). It was formed by rapid deposition of sediment following the construction of a canal (the Wax Lake Delta Outlet, or WLO) in 1941, which connected Six-Mile
Lake and the Atchafalaya Bay (Fig. 4.1A). Since then, the subaqueous delta began to form and prograde at the mouth of the outfall canal due to accumulation of fluvial sediment. The delta became subaerial in 1973 after a large river flood (peak discharge of 20,000 m$^3$ s$^{-1}$, compared to the mean discharge of 5,781 m$^3$ s$^{-1}$ from 1935 to 2007) that transported a significant amount of sediment to the river mouth, both from fluvial sediment input and erosion of the channel bed layer (Roberts et al., 1997).

Fluvial water and sediment discharge have strong seasonal variations in the WLD system (Mossa and Roberts, 1989). Annual river floods occur in spring when the mean flow velocity and suspended sediment concentration become 2 – 2.5 times and up to 20 times higher than that in non-flood conditions for an average-discharge year, respectively (DuMars, 2002). The WLO transports approximately 30 MT of sediment per year to the shallow Atchafalaya Bay (~2m deep, Kim et al., 2009), leading to a yearly area increase of 0.81 km$^2$ (FitzGerald, 1998; Roberts et al., 2003). The mean fluvial sediment layer (topset) thickness of the delta is 2.4 m, of which 70% is sand (Roberts et al., 1997). The WLD has been colonized by freshwater species on the high-lying areas of islands due to the freshwater sequestration in this river-dominated system. The plant community is in a dynamic early successional phase, and its composition is mainly determined by fluvial discharge, elevation, disturbance, salinity, organic content of sediment, and nutrient availability (Holm and Sasser, 2001; Rejmánek et al., 1987; Shaffer et al., 1992).

Tides and waves are relatively mild in the Atchafalaya Bay under typical conditions. The tidal range is around 0.3 m (DuMars, 2002), and the significant wave height is less than 0.5 m (Wright, 1977). This area is exposed to two major marine weather systems: winter cold fronts and tropical storms. Tropical storms affect this area primarily in summer and fall, and cold fronts occur every 4 – 7 days from October through March, or 20 – 30 times per year in total (Chuang and Wiseman, 1983). The high winds and wind-generated waves during coastal storms can effectively re-suspend and transport the bottom sediment, leading to significant morphological changes on coastal wetlands (Barras et al., 2005; Howes et al., 2010). More detailed information about hurricanes and cold fronts will be explained in the next section.
4.3.2 HURRICANE IKE

The WLD was influenced by two hurricanes systems in 2008: Hurricane Gustav from August 25th to September 4th (Category 4, made landfall 82.4 km to the east of the WLD as a Category 2 hurricane), followed by Hurricane Ike from September 1st to September 14th (Category 4, made landfall 337 km to the west of the WLD as a Category 2 hurricane). Although the track of Hurricane Gustav was closer to the WLD, Hurricane Ike caused higher flow velocities and sediment transport in this area as the WLD is located on the east side of hurricane systems, which experienced the strongest energy (Coch, 1994). Here we focus on Hurricane Ike and its impacts on delta morphology.

The development and movement of Ike was very complicated. It first formed as a tropical wave over the western coast of Africa on August 28th, 2008, then moved west and became a tropical depression at 0600 Coordinated Universal Time (UTC) on September 1st. The Ike system strengthened quickly at the beginning of September 3rd and reached its maximum sustained winds of 230 km h$^{-1}$ and lowest pressure of 935 mbar at 0600 UTC on September 4th (Berg, 2009). Before entering the Gulf of Mexico on September 10th, Ike made two landfalls in the Bahamas and Cuba as Category 3 and 1 hurricane, respectively, and the highest intensity reached Category 4. Hurricane Ike slightly weakened when it moved into the Gulf of Mexico, then strengthened and reached a secondary minimum barometric pressure of 944 mbar at 0000 UTC on September 12th when it passed over the warm waters of the Loop Current. The system made its third landfall on the northern end of Galveston Island in Texas at 0700 UTC on September 13th, as a Category 2 hurricane with wind speed of 175 km h$^{-1}$ (Berg, 2009, Fig. 4.1A).

4.3.3 COLD FRONTS

A cold front is the interface or transition zone between cold, dry airflows and warmer and lighter air (Hsu, 1988). Passages of cold fronts cause changes in temperature, wind speeds and directions, barometric pressures, and humidity (Mossa and Roberts, 1990; Roberts et al., 1989). Based on these changes, a cold front event is divided into three stages: prefrontal, frontal passage, and postfrontal. The prefrontal stage is characterized by significant drop in pressure and strong southerly winds, causing water
level setup and overbank flooding into wetlands during extreme cases (Mossa and Roberts, 1990). During the frontal passages, pressure drops to its lowest value, which is usually accompanied by intense rainfall. Winds become stronger with a sharp shift in wind direction from southerly to westerly. The postfrontal phase starts with rise in barometric pressure and drop in air temperature and humidity (Hsu, 1988; Roberts et al., 1989). Winds shift towards northerly, leading to offshore directed flow and decreased water level.

4.4 METHODS

4.4.1 MODEL DESCRIPTION

Delft3D, a widely used and validated computational fluid dynamics model, is applied for this study (Lesser et al., 2004). Comprised of several integrated modules, Delft3D is designed to simulate hydrodynamic flow, transport of water-borne constitutes, short wave generation and propagations, sediment transport and morphological changes, ecological processes and water quality parameters in coastal, riverine and estuarine areas (Lesser et al., 2004; Roelvink and Van Banning, 1994). Delft3D FLOW is the hydrodynamic module that uses a finite difference solution of the three-dimensional shallow water equation and the k-ε turbulence closure model (Rodi, 1980) to compute flow characteristics under the hydrostatic pressure assumption (WL|Deltres Hydraulics, 2011). When the Morphology module (MOR, inside of the flow module) is activated, sediment transport is calculated simultaneously with flow computation using the advection-diffusion equation (for suspended sediment load, not for bedload). Waves are calculated with the third-generation fully spectral wave module SWAN, which computes wave parameters with the discrete spectral action balance equation (WAVE|Deltres Hydraulics, 2011). A three-dimensional vegetation routine is incorporated in the FLOW module to explore the impact of vegetation on flow hydrodynamics and sediment transport during coastal storms (Temmerman et al., 2005). More detailed explanations of the model structures are given in Lesser et al. (2004).
4.4.2 Model Input Parameter Settings and Data Analyses

In this study, we analyzed the hurricanes that happened in 2008 and cold fronts between October 2008 and April 2009 (two hurricanes and a cold front season). In order to acquire detailed hydrodynamics and morphological fields on the WLD, the Delft model is applied to three nested domains: the GoM domain that covers a large part of the Gulf of Mexico with a resolution of 0.020 degrees; the Atchafalaya domain that includes the coastal shallow areas of the WLD (Atchafalaya, Vermilion, and Cote Blanche Bay) and surrounding low-elevated wetlands with an average resolution of 200 m; and the WLD domain that covers the WLD and surrounding wetlands, with an average resolution of 50 m (Fig. 4.1A, B, C). The GoM model is triggered by tides at the ocean boundaries extracted from the TPXO 7.2 Global Inverse Tide Model (http://volkov.oce.orst.edu/tides/TPXO7.2.html) and winds derived from Global Forecast System (GFS, Environmental Modeling Center, 2003) with spatial resolution of 0.1667 degrees and temporal resolution of 3 hours. Bathymetry is obtained from the Louisiana Virtual Coast Data Archive (http://virtual-coast.c4g.lsu.edu/), in which NOAA’s bathymetry sounding database, the Digital Nautical Charts database, and the 5-minute gridded elevations/bathymetry for the world (ETOPO5) database are combined (Mukai et al., 2002; National Geophysical Data Center, 1988; National Ocean Service, 1997; U.S. Department of Defense, 1999). Sediment sand (mean grain size of 1.0E-04), silt (mean grain size of 3.0E-05), and mud are included in the model and their distributions on seabed are extracted from the sediment distribution map (dbSEABED) estimated by Jenkins (2002). In deep area the bed roughness (Manning’s n) is calculated through an empirical equation:

\[ n = 0.015 + \frac{0.01}{\text{depth}}, \]

and this equation has been validated in coastal ocean simulations (Xing et al., 2012; Xing et al., 2015 in prep.). A more detailed bathymetry and island topography obtained from 1998 hydrological survey with an averaged resolution of 80 m is used for the WLD domain, combining with Light Detection and Ranging (LIDAR) survey data for overbank areas that were not covered in the hydrographic survey (USACE, 2010). The daily water discharges and annual average sediment concentrations at fluvial...
boundaries are acquired from observations at USGS (Baumann et al., 2005), and DuMars (2002), respectively. The morphological settings are derived from a former study (Meselhe et al., 2015, in prep.), in which correlated parameters have been validated against successive mapping of the WLD growth from 1998 to 2014 where the morphological evolution produced by the model compared favorably with the mapping data.

According to the Hydrometeorological Prediction Center's surface analysis archive (http://www.hpc.ncep.noaa.gov/archives/web_pages/sfc/sfc_archive.php), 41 cold fronts were identified between October 2008 and April 2009, and cold front characteristics (duration, wind structures) are analyzed with hourly winds, water levels, and pressure data from NOAA station Amerada Pass (8764227 http://tidesandcurrents.noaa.gov/met.html?id=8764227, Fig. 4.1C). Due to lack of wind data between January 7th and February 17th, 2009, 12 events are removed. The remaining 29 events show significantly different properties, with a maximum and minimum mean wind speed of 11.4 m s\(^{-1}\) and 4.3 m s\(^{-1}\), respectively. 20 events have typical pre-frontal, frontal, post-frontal structures, and 5 are dominated by onshore-directed winds and the other 5 are dominated by offshore-directed winds, among which 11 events with distinct properties are simulated for this study (Tab. 4.1). As Hurricane Ike occurred when vegetation existed, the impact of aboveground vegetation is explored in the model by applying uniform vegetation with a height of 1m and density of 150 stems m\(^{-2}\) on the delta where water depth is shallower than 0.2 m (Carle et al., 2013; Martin and Shaffer, 2005; Shaffer et al., 1992). Vegetation is not included for the cold front simulations as the aboveground parts of vegetation died in winter. In order to explore the impact of winds and waves on delta morphology, simulations that use the same parameter settings as shown above but exclude winds and waves are also implemented for both cold front and hurricane cases.

Table 4.1 Properties of 2008 cold front events with wind data from the Amerada Pass station (8764227, Fig. 4.1A), and hourly averaged sediment deposition and erosion caused by winds and waves

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<th>Event</th>
<th>Front passage time</th>
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<th>Duration (h)</th>
<th>Deposition (m(^3))</th>
<th>Erosion (m(^3))</th>
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</table>

Note: The duration of cold front events is obtained by comparing winds, water levels, and pressure at the Amerada Pass station, and front passage time is derived by analyses of spatial pressure fields from the Hydrometeorological Prediction Center's surface analysis archive. The events that are marked with * are simulated for this study, and the number that are marked with ‘₉’ are typical cold front events with prefrontal, frontal, and postfrontal phases. All the events without mark of ‘₉’ are atypical cold fronts.

A widely-used, low-pass digital filter program is applied to simulated and observed water levels and velocities to eliminate the influence of tides, by which waves with periods longer than 12 hours are filtered from the simulated results (Thompson, 1983; Walters and Heston, 1982). In this way, we obtain the hydrodynamic results generated by interactions of winds, waves and rivers. The filtered flow velocities are then calculated with the following equation to obtain the residual currents for each cell during hurricanes and cold fronts:
\[ R_j = \frac{1}{\text{dep}_j} \sum_{i=1}^{n} v_{i,j} \cdot d_{i,j}, \]  

where \( \text{dep}_j \) [m] is mean water depth at cell \( j \), \( v_{i,j} \) [m s\(^{-1}\)] is velocity at cell \( j \) for time step \( i \), \( d_{i,j} \) [m] is water depth at cell \( j \) for time step \( i \), and \( n \) (unitless) is the total number of calculating time steps.

### 4.4.3 Model validation and uncertainty analysis

Delft3D’s performance is tested with observed wave parameters from National Data Buoy Center (NDBC: http://www.ndbc.noaa.gov/) and water levels from tide gauge stations (http://tidesandcurrents.noaa.gov/tide_predictions.html) for simulation periods of August 25\(^{th}\) to September 19\(^{th}\), 2008 and November 22\(^{nd}\) to December 13\(^{th}\), 2008 (Fig. 4.2, 4.3, 4.4, and 4.5). Comparisons show that the simulated wave fields closely match observed wave dynamics (Fig. 4.2, 4.3): the simulated mean wave height is 1.1 m, compared to observed value of 1.4 m, and the correlation coefficient is 0.83 (\( n = 15,247 \), Fig. 4.5A); the simulated mean peak wave period is 4.6 s, compared to observed value of 6.4 s, and the correlation coefficient is 0.66 (\( n = 15,082 \), Fig. 4.5B). Simulated tide-filtered water levels catch the observations both during the Hurricane Ike and cold front simulations (Fig. 4.4). The average observed water level is 0.05 m and the simulated value is 0.04 m, and the correlation coefficient between them is 0.66 with \( n = 10,993 \) (Fig. 4.5C). The consistency between simulations and observations demonstrates that the model is reliable for producing realistic hydrodynamics. As the morphological parameters used in this study have been validated in Meselhe et al. (2015, in prep.), we assume that the MOR module of Delft3D system is capable of providing reasonable morphological results for this study.
Figure 4.2 Comparisons of simulated significant wave heights and peak periods at 8 NDBC buoy stations between August 25th and September 19th, 2008 (See Fig. 4.1C), showing that the model catches the high waves caused by the two hurricanes (Gustav and Ike)

Note: $H_{\text{sig}_O}$ represents observed significant wave height, $H_{\text{sig}_S}$ represents simulated significant wave height, $T_{\text{p}_O}$ shows observed peak wave period, and $T_{\text{p}_S}$ shows simulated peak wave period. A): 42001; B): 42002; C): 42019; D): 42020; E): 42035; F): 42036; G): 2039; H): 42040.
Figure 4.3 Comparisons of simulated significant wave heights and peak periods at 8 NDBC buoy stations between November 22nd and December 12th, 2008 (See Fig. 4.1C), demonstrating that the model successfully simulates the variations in wave parameters caused by cold fronts.
Note: $H_{sig\_O}$ represents observed significant wave height, $H_{sig\_S}$ represents simulated significant wave height, $TP\_O$ shows observed peak wave period, and $TP\_S$ shows simulated peak wave period. A): 42001; B): 42002; C): 42019; D): 42020; E): 42035; F): 42036; G): 2039; H): 42040.

Figure 4.4 Comparisons of simulated tide-filtered water levels at 4 NOAA stations for time period: (i) August 29th to September 18th, 2008 and period: (ii) November 25th to December 12th, 2008 (See Fig. 4.1C), showing that the model successfully simulates the water level rise caused by hurricanes (i) and cold fronts (ii)
Figure 4.5 Evaluation of model performance in simulating wave heights, wave peak periods and storm surges, demonstrating that the model catches the observed significant wave height well, but has more uncertainty in simulating peak wave periods and water levels.

The uncertainty of the model is explored by setting up 36 simulations within Dakota software with the Latin hypercube sampling method (Stein, 1987), taking into account the uncertainties of input observed parameters (sand grain size, silt grain size) and input estimated parameters (critical shear stress for mud erosion ($\text{Cr}_{\text{ERO}}$) and critical shear stress for mud deposition ($\text{Cr}_{\text{SED}}$))(Tab. 4.2). In order to save computational time, the uncertainty analysis is implemented only for the Hurricane Ike event and cold front event 11, demonstrating that the uncertainty for total sediment balance is 271,000 m$^3$ (mean value of 700,000 m$^3$) and 147,000 m$^3$ (mean value of 278,000 m$^3$) for Ike and event 11, respectively. Although model uncertainty is larger for Hurricane Ike than for cold front event 11, the critical shear stress for erosion is the predominant factors for both studies, which accounts for more than 95% of the uncertainty. Similar to the result from Chapter 3 (section 3.9), the cells with higher morphological changes shows larger uncertainty (Fig. 4.6).

Table 4.2 Uncertainties of major input parameters

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Morphology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Contents and Sources</td>
<td>Morphology</td>
</tr>
<tr>
<td>Uncertainty</td>
<td></td>
</tr>
<tr>
<td>References</td>
<td></td>
</tr>
<tr>
<td>Wellner et al.</td>
<td>Pankow et</td>
</tr>
</tbody>
</table>
Figure 4.6 Spatial distribution of model uncertainty during (A): Hurricane Ike; and (B): cold front event 11, showing that the model uncertainty increases significantly with the magnitudes of events.

4.5 DISTINCT CHARACTERISTICS OF COLD FRONTS AND DELTA RESPONSES

Cold fronts are critical in transporting sediment along the coasts of the Gulf of Mexico (Moeller et al., 1993; Roberts et al., 1987). Former studies demonstrate that a strong cold front can significantly disturb coastal hydrodynamics and sediment transport, and a cold front season causes either deposition or erosion on specific coastal wetlands (Kineke et al., 2006; Roberts et al., 1980; Rouse et al., 1978; Van Heerden, 1983; Walker and Hammack, 2000). However, few studies have focused on the inter-variations of cold fronts and their impacts on coastal hydrodynamics and morphological changes. For this study, we simulate 4 cold fronts (events 9, 11, 20, and 21 in Tab. 4.1) within the cold front season of 2008-2009, including both typical and atypical events. Results illustrate that each cold front has a distinct wind structure, producing distinct hydrodynamics and sediment transport patterns. Mean wind speeds are the most critical factor in controlling the magnitude of residual currents and the amount of sediment transport,
and the events that cause more significant water level variations cause more sediment transport than the events that cause minor water level variations.

Figure 4.7 Times series of wind vectors at the Amerada Pass station (8764227, Fig. 4.1C) during cold front event 9, 11, 20, 21 and Hurricane Ike. Fig. 4.7A and B show wind fields for typical cold fronts which are dominated with onshore-directed winds, followed by offshore-directed winds. Fig. 4.7B is the strongest cold front event during the 2008-2009 cold front season. Fig. 4.7C shows a minor atypical cold front with majorly offshore winds and Fig. 4.7D shows an atypical event with onshore winds. Fig. 4.7E shows the wind structure for Hurricane Ike which is dominated by onshore-dominated winds.

A typical cold front is characterized by onshore-directed winds followed by offshore-directed winds, correlated with typical frontal structure (Fig. 4.7 A, B, section 4.3.3). Variations in frontal structures (e.g., two parallel fronts in the same system) will generate distinct wind fields, leading to atypical cold fronts. For instance, events 3, 4, 16, and 20 are dominated by offshore directed winds while event 11, 14, 21, 29 are dominated by onshore directed winds. Four simulated events represent cold fronts with different wind structures and wind speeds: event 9 represents a typical cold front with medium mean
wind speed (7.7 m s\(^{-1}\)); event 11 is the largest cold front event with the typical wind structure (11.4 m s\(^{-1}\)); event 20 is an offshore wind dominated cold front event (6.7 m s\(^{-1}\)), and event 21 is an onshore wind dominated cold front event (8.8 m s\(^{-1}\), Fig. 4.7).

Wind structures significantly influence water level variations (Fig. 4.7, 4.8). Events 9 and 11 (typical) produce water levels setup followed by setdown, while events 20 (atypical) generate minor negative water levels (Fig. 4.7), and event 21 (atypical) produces positive water levels throughout the event, respectively (Fig. 4.8). Fluvial input plays a great role in controlling hydrodynamics and sediment transport in channels during cold fronts, leading to high residual currents and erosion in channels for all of the 4 events (Fig. 4.9 A, C, E, G). Wind properties determine the distributions of residual current and sediment transport on islands: higher onshore wind speeds are able to blow water onto high-elevated islands and re-suspend sediment, producing higher residual currents and erosion on islands (Fig. 4.7, 4.9), while offshore winds, combined with fluvial downstream forces, lead to a large amount of sediment transportation downstream. When winds and waves are excluded, all of the 4 events produce similar patterns of erosion in channels and deposition on islands, demonstrating that winds and waves significantly increase erosion on islands and decrease erosion in channels (Fig. 4.10 A, C, E, G). The comparisons of the sediment balance for the 4 events show that every cold front brings about erosion to the WLD and that the amount of erosion is highly correlated to mean wind speed – so event 11 produces the largest amount of sediment transport, followed by event 9, and then events 21 and 20 (Tab. 4.3). The magnitude of water level change also controls the amount of erosion, so event 21 with a larger mean wind speed (8.8 m/s) produced less sediment erosion than event 9 (7.7 m/s), as water level dropped suddenly by ~0.5 m for event 9 but only gradually increased by ~0.4 m during event 21 (Tab. 4.1, 4.3). Wind directions influence the deposition and erosion patterns on the islands so the high erosion on the eastern sides of islands occurs during events 11 and 21 due to the strong easterly onshore-directed winds, which did not occur during event 20 (Fig. 4.7, 4.8).
Figure 4.8 Time series of tide-filtered water level at the Amerada Pass station (8764227, Fig. 4.1C) during cold front events 9, 11, 20, and 21. Event 9 and 11 has abrupt water level variations, leading to high morphological changes. Event 20 has minor water level variations, leading to the lowest morphological changes. Event 21 is dominated by slowly increasing water level, leading to less erosion than event 9 and 11. Most erosion occurs on the deltaic islands during event 21, apart from event 9 (Fig. 4.10).

Figure 4.9 Simulated residual current distributions on the WLD A): event 9 including winds and waves; B): event 9 excluding winds and waves; C): event 11 including winds and waves; D): event 11 excluding
winds and waves; E): event 20 including winds and waves; F): event 20 excluding winds and waves; G): event 21 including winds and waves; H): event 21 excluding winds and waves. The simulations show that fluvial input majorly determines the high residual currents in channels, while winds and waves increase the residual current on deltaic islands.

Figure 4.10 Simulated sedimentation-erosion distribution on the WLD for A): event 9 including winds and waves; B): event 9 excluding winds and waves; C): event 11 including winds and waves; D): event 11 excluding winds and waves; E): event 20 including winds and waves; F): event 20 excluding winds and waves; G): event 21 including winds and waves; H): event 21 excluding winds and waves. The simulations demonstrate that winds and waves are the most critical factors contributing to large morphological changes, and wind speed is the most critical factor that determines the amount of morphological changes (Fig. 4.7)

Table 4.3 Total amount sediment transport for the calculated WLD area (Fig. 4.1C) during the cold front event 9, 11, 20, 21 and Hurricane Ike, respectively

<table>
<thead>
<tr>
<th>Cold Front Event 9 (138 hours)</th>
<th>Scenarios</th>
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<th>No winds, waves</th>
</tr>
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<tr>
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<tr>
<td>Erosion</td>
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<td>-11,000</td>
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<tr>
<td>Accretion</td>
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<table>
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<th>Base</th>
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<td>Net</td>
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<td>-3,000</td>
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</tr>
<tr>
<td>Erosion</td>
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<td>-11,000</td>
<td></td>
</tr>
<tr>
<td>Accretion</td>
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<td>9,000</td>
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<table>
<thead>
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<th>No winds, waves</th>
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<tbody>
<tr>
<td>Net</td>
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<td>7,000</td>
<td></td>
</tr>
<tr>
<td>Erosion</td>
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<td>-3,000</td>
<td></td>
</tr>
<tr>
<td>Accretion</td>
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Cold Front Event 21 (94 hours)

<table>
<thead>
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<tr>
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<tr>
<td>Net</td>
<td>-13,000  -11,000   21,000</td>
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<tr>
<td>Erosion</td>
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Hurricane Ike (100 hours)

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</thead>
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<tr>
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<td>Accretion</td>
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</tr>
</tbody>
</table>

Note: Base case indicates that winds, waves, tides, and morphology are calculated. The units are $m^3$, and the numbers that are marked with * are calculated for model uncertainties.

4.6 COMPARISONS OF DELTA RESPONSES TO A HURRICANE AND A STRONG COLD FRONT

In addition to frequent cold fronts, the rare but more intense hurricanes cause significant water and sediment movement in coastal wetlands that should not be ignored in studies of wetland evolution (Dietrich et al., 2010; Rego and Li, 2010). For this study, we numerically simulate the hydrodynamics and morphological changes on the WLD during Hurricane Ike and a strong cold front event (event 11), which generate similar patterns of residual currents and morphological changes, although the magnitude is much larger during hurricane simulation. Both of the simulations are characterized by downstream currents in channels due to strong fluvial forces and erosion on islands due to the shear stress caused by winds and waves. The channels are dominated by erosion during the cold front simulation, which change to be depositional for the hurricane case, demonstrating that the winds and waves outweigh the fluvial forces in channels during Hurricane Ike but not occur during the strong cold front event. The strong downstream flow caused by fluvial forces and offshore winds causes significant erosion in channels and deposition in front of the channels at the delta front for both cold front event 11 and Hurricane Ike simulations. Aboveground vegetation significantly decreases the amount of sediment transport during the Hurricane Ike simulation but has no impact on cold front events as vegetation dies in winter.
The residual current field from the hurricane simulation shows similar patterns to that of the cold front, although the magnitude is higher, characterized by high currents in channels and low currents on islands. This demonstrates that fluvial forces play the predominant role over winds and waves in the channels (Fig. 4.9C, 4.11B). However, winds and waves outweigh fluvial forces in determining sediment transport in the channels during Hurricane Ike, illustrated by the major depositional pattern (Fig. 4.10C, 4.11E). The high residual currents during the hurricane simulation lead to more widely distributed erosion on the islands compared to the cold front event (Fig. 4.10C, 4.11E). Both events show high erosion at the eastern sides of islands (0.015 m for cold front, Fig. 4.10C, vs. 0.025 m for hurricane, Fig. 4.11E), attributed to the easterly-dominated winds. Significant erosion in channels and deposition in front of channels occur at the delta front, caused by the strong downstream flow due to offshore winds and fluvial forces, which has also been observed by Shaw and Mohrig (2013). This pattern does not show up for other small cold front events as the low offshore winds are not able to cause the morphological changes (Fig. 4.10). Vegetation dramatically decreases residual currents on islands from as high as 0.3 m s\(^{-1}\) to less than 0.1 m s\(^{-1}\) and eliminates the areas with high currents (0.4 m s\(^{-1}\)) at land boundaries (Fig. 4.11B, C). As a result, the sedimentation-erosion pattern on the WLD changes significantly: the maximum erosion on the eastern sides of islands decreases from 0.2 m to 0.025 m, and deposition in the channels decreases from as high as 0.035 m to 0.025 m, accompanied with increased erosional area (Fig. 4.11E, F). Hurricane Ike causes ~4 times more sediment transport than cold front event 11 (412,000 m\(^3\) vs. 103,000 m\(^3\)), which decreases to less than 3 times when aboveground vegetation is taken into account (308,000 m\(^3\)) (Tab. 4.3).
Figure 4.11 Simulated residual current and morphological changes on the WLD during Hurricane Ike. 
A): residual current field excluding winds and waves; B): residual current field including winds and waves; C): residual current field including winds and waves and aboveground vegetation; D): sedimentation-erosion pattern excluding winds and waves; E): sedimentation-erosion pattern including winds and waves; F): sedimentation-erosion pattern including winds and waves and aboveground vegetation. Winds and waves significantly increase the residual currents on the deltaic islands and the upstream wetlands. Aboveground vegetation decreases residual currents on islands and causes residual currents to focus in channels. Winds and waves determine the pattern and the amount of morphological changes on the delta during Hurricane Ike, and aboveground vegetation significantly decreases the amount of erosion on islands.

4.7 STATISTICAL ANALYSES OF THE IMPORTANCE OF HURRICANES AND COLD FRONTS ON THE WLD MORPHOLOGY

Cold fronts are able to transport 15% of the fluvial sediment from the Atchafalaya Bay to offshore areas (Roberts et al., 1997), while hurricanes are responsible for 90% of Louisiana shoreline retreat in the past century (Stone et al., 1997). However, the roles of the two events on the WLD morphology have not been evaluated. We estimate the sediment balance from October 2008 and April
2009 on the WLD to show annual cumulative impacts of cold fronts and use the total sediment balance for Hurricane Gustav and Ike to demonstrate the influences of hurricanes on delta morphology for 2008. Our research demonstrates that winter cold fronts lead to 1,900,000 m$^3$ sediment leaving the delta area in 2008, while hurricanes, although they cause intense sediment transport in a short time, move far less sediment than cold fronts (erosion of 500,000 m$^3$).

We analyze the statistical features of cold fronts over the 2008 – 2009 season (29 events with available winds out of 41 events) and simulate 11 cold fronts in October, November, December, February and March. The analysis of wind vector during October to April between 2008-2014 demonstrates that the 2008 cold front season has similar wind vector distribution compared to the 6-year statistics (Fig. 4.12). However, high winds occur more frequently (15.6% of the time, wind speed is larger than 8.8 m/s) in the 2008 – 2009 season than in the next 6 years (13.7% of the time, wind speed is larger than 8.8 m/s), and the maximum winds blow from south to north, favoring to water level setup. So the study of cold fronts in 2008 – 2009 represents a strong cold front season for the WLD area. The simulations show that winds and waves significantly increase the amounts of sediment erosion and deposition in the calculated WLD area (Tab. 4.3), and both the amounts of erosion and deposition per hour are positively correlated to mean wind speeds, with $R^2$ equal to 0.94 and 0.81 for erosion and deposition, respectively (Fig. 4.13). Using these relationships, we reconstruct the sediment balance due to winds and waves during the 29 events, which show that 1,900,000 m$^3$ of sediment would be removed from the calculated WLD area.

The year 2008 was an active hurricane year for the WLD during which two hurricanes influenced the area. Hurricane Gustav made landfall to the east of the study area, so it caused a small water level rise (<1 m) in the study area, and the mean wind speed was about 5.6 m s$^{-1}$, leading to net sediment erosion of 8,000 m$^3$ on the WLD. The total sediment erosion of the WLD caused by hurricanes in 2008 was ~500,000 m$^3$ (Hurricane Ike caused erosion of 412,000 m$^3$), much less than the amount of sediment that was removed during cold front events. Chapter 3 also demonstrated that the strong Hurricane Rita (2005) caused erosion of 500,000 m in the study area. It is therefore reasonable to state that the frequent cold fronts have a larger impact than hurricanes on the morphology of the WLD system. However, hurricanes
are shown to be the only events (other than cold fronts and river floods) that are able to change the erosional pattern in the channels to depositional, which might offset the influences of river floods and slow the delta progradation (Xing et al., 2015).

Figure 4.12 Wind rose and frequency distribution compiled from station Amerada Pass (8764227, Fig. 4.1C) for the period: October 2008 - April 2009 (A, B) and the period: October – April 2008 to 2014 (C, D). The cold front season of 2008 – 2009 is dominated by more frequent southwest winds and a larger proportion of high winds (larger than 8.8 m s⁻¹).

Note: Percentage frequencies are shown for each 22.5° wind delineations. The map is made with Lake Environment Software for wind rose plots software, Wind Rose Plots for Meteorological Data (WRPLOT View™ Version 7.0.0).
4.8 SALINE WATER INTRUSION TO THE WLD DURING COLD FRONTS AND HURRICANES

Saline water exposure influences the live biomass and stem density of wetlands species, among which freshwater species are more vulnerable (McKee et al., 2004). Temporary saltwater intrusion (1 week) can cause plants to stop growing and even begin dying (Grace and Ford, 1996). Vegetation death has been widely observed in coastal wetlands along the Gulf of Mexico after Hurricanes Audrey (1957), Andrew (1992), and Rita (2005) due to saline water intrusion (Ensminger and Nichols, 1957; Guntenspergen et al., 1995; Neyland, 2007; Shiflet, 1963), and Walker (2001) observed a significant salinity increase of 20 ppt in Atchafalaya-Vermilion Bay during Hurricane Georges (1998). Significant salinity rise is also observed after strong cold front events in the Atchafalaya Bay (Walker and Hammack, 2000; Teeter and Johnson, 2005). The subaerial parts of WLD are occupied with freshwater plant communities due to the dominant role of fluvial freshwater input, which is sensitive to salinity variations (McKee et al., 2004). In this study, we explore saline water intrusion during a strong cold front (event 11) that represents the most intense of the strong cold fronts (Fig. 4.12) and Hurricane Ike that represents a
strong hurricane event. Salinity data derived from Regional Ocean Modeling System (ROMS) simulations by Rutgers University (http://tds.marine.rutgers.edu/thredds/roms/gom/catalog.html) is used to set up the model initial and boundary conditions. The results indicate that the Hurricane Ike is able to increase the salinity at the WLD area to more than 20 ppt, while saline water (maximum 14 ppt) is limited to the western edge of WLD during the cold front (Fig. 4.14). This would not influence the survival of freshwater species on the islands.

Under non-storm conditions, saline water is limited to 5 m isobath lines, which hardly touches the edge of the delta, so the freshwater species continue to grow on the wetlands (Fig. 4.14A). During the biggest winter cold front events in this study (event 11), saline water with maximum salinity of 14 ppt occurs to the west of the WLD, and the western part of the delta is dominated by saline water where salinity is higher than 10 ppt (Fig. 4.14B). The saline water tongue stays in the delta edge for 15 hours from December 11\textsuperscript{th} 2000 UTC to December 12\textsuperscript{th} 1100 UTC, then retreats outside of the WLD, having minor influence on the survival of freshwater species. It is reasonable to state that winter cold fronts would not cause saline water intrusion to the extent that it would influence the survival of freshwater species on the WLD for the following reasons: 1) this event (11) is the largest winter cold front event of the year, which happened when fluvial water is in its minimum status; and 2) the statistical analyses (August 26\textsuperscript{th}, 2008 to November 1\textsuperscript{st}, 2014) at station Amerada Pass show that, in the past 6 years, wind speed was larger than 11.1 m s\textsuperscript{-1} (mean wind speed for event 11) only 3.7% of times including hurricane conditions (Fig. 4.1C, 4.12B).

The simulation shows that Hurricane Ike causes significant saline water expansion to the deltaic area, and the WLD is covered with saline water (higher than 20 ppt) for 35 hours (September 12\textsuperscript{th}, 0700 UTC to September 13\textsuperscript{th}, 1700 UTC, Fig. 4.14C). The water then retreats due to high fluvial freshwater discharge. The overbank flow carries the saline water into the freshwater wetlands, leading to significant damage to plant species. The impact of saline water on plant survival may be more severe in the low-elevated area where saline water is less likely to flow back through the river system. More study focusing
on saline water intrusion into wetlands is needed to protect coastal wetlands from extreme coastal storms with increased intensity (Emanuel, 2005).

Figure 4.14 Simulated vertical-averaged water salinity distributions on the WLD during (A) non-storm tidal condition; (B) Hurricane Ike; (C) a strong cold front event 11. All of them represent the time step with the most upstream saline water intrusion during the events. The simulations show that saline water (>20 ppt) is able to expand all over the WLD during Hurricane Ike, but is limited to the western edge of the delta during cold fronts (maximum 14 ppt). The delta and its surrounding area is dominated by freshwater (<5 ppt) under normal condition.

4.9 CONCLUSIONS

The numerical model Delft3D is applied to study hydrodynamics and morphological changes on the Wax Lake Delta during cold fronts and hurricanes. Simulated wave heights, peak periods and water levels match well with observations. The uncertainty analysis shows that the critical shear stress for erosion is the predominant factor causing uncertainty in the model results, and that the uncertainty is larger for Hurricane Ike (271,000 m³) than a cold front (147,000 m³). Cells with higher morphological changes also have larger uncertainty than those with smaller changes. Mean wind speeds are the predominant factor controlling residual currents and sediment transport during cold front events, and erosion occurs mainly when the water level has significant variations, so the cold front events with higher wind speeds that cause larger water level variations over a short time produce more sediment erosion than other events. Although hurricanes cause stronger residual currents and more sediment transport than a strong cold front, both events produce high residual currents in the channels and low residual currents on islands,
demonstrating the dominant role of fluvial forces in channels. However, winds and waves become more important than fluvial forces in determining morphological changes in channels as the erosional pattern caused by the river changes to be mostly depositional during the Hurricane Ike simulation. Winds and waves cause widespread erosion on islands and severe erosion on the eastern banks of islands during both cold fronts and hurricanes, due to the high current-wave shear stress in shallow areas and the dominant easterly winds. The significant erosion in the channels at the delta front and accompanying deposition in front of the channels are caused by the strong downstream flow due to offshore winds and fluvial forces, which occurs only for strong events such as cold front event 11 and Hurricane Ike. Aboveground vegetation significantly decreases the amount of erosion on islands during the Hurricane Ike simulation (from maximum 0.2 m to 0.025 m) but has no impact during cold fronts as the aboveground parts die in winter. The statistical analyses of cold fronts between 2008 and 2009 show a positive relationship between mean wind speed and hourly-averaged erosion and deposition caused by winds and waves in the WLD area. Using these relationships, the total amount of erosion caused by winds and waves during the season is estimated to be as high as 1,900,000 m³ for 29 events with available winds, which is around 4 times as high as erosion for Hurricane Gustav and Ike (500,000 m³). We therefore estimate that cold fronts cause more sediment transport and morphological changes on the WLD than hurricanes. Saline water intrusion significantly influences the survival of local freshwater wetlands during the Hurricane Ike simulation, while having minor impacts during most of the cold fronts.

4.10 ACKNOWLEDGMENTS

This paper is supported by the National Science Foundation (NSF), award # 1135457.
5 CONCLUSION

This thesis explores fluvial and marine environments of delta systems and quantifies how two deltas respond to environmental changes. A long-term fluvial sediment flux for the Ebro River Delta in Spain is reconstructed to provide insights into how its morphology has changed over the past four thousand years. This study demonstrates the critical role of human activities in controlling long-term fluvial sediment flux. The study of the Wax Lake Delta in Louisiana investigates the response of deltas to marine extreme events, providing insight about delta sustainability in modern times. The latter study also quantifies the cumulative impacts on delta morphology of a cold-front season and yearly hurricane events.

All deltas experience morphological changes over time. For instance, the Mississippi River Delta system is evolving with the abandonment of its “Bird’s foot” subdelta and with the progradation of the Atchafalaya subdelta. The Yellow River Delta has changed its location from the Yellow Sea to the Bohai Sea in 1855, forming a completely new delta system (Roberts, 1997; Zhu, 1993). The river-wave dominated Ebro Delta in Spain has significantly changed its shape over time, resulting in the modern ‘ping wing shaped’ delta (Canicio and Ibanez, 1999). Significant changes in deltas’ morphology are attributed to variations in environmental forces, such as fluvial sediment flux, delta topography and flow path, and wave activities (Canicio and Ibanez, 1999; Roberts, 1997; Saito et al., 2001). The interactions between fluvial and marine forces are a possible cause of morphological changes for the Ebro Delta system, so the study of delta evolution requires reconstructing marine and fluvial environments (Ashton et al., 2013). Chapter 2 of this thesis reconstructs the long-term fluvial sediment flux for the Ebro River, which is a critical step towards better understanding the evolution of the Ebro Delta. In this chapter, the numerical model HydroTrend is applied to the basin, which takes into account climate changes (temperature and precipitation), basin geology, basin relief, and anthropogenic activities (land use and dam emplacements) in order to estimate long-term fluvial sediment flux to the ocean. Both numerical simulations and sensitivity analyses indicate that, before the 18th century, precipitation was the dominant
factor controlling fluvial water discharge to the Ebro river mouth, while fluvial sediment flux was most
influenced by anthropogenic activities. Significant alterations in fluvial sediment flux correspond to
changes in intensities of human activities. For instance, the increased sediment flux during AD 1150 –
1250 is correlated to the Muslim times when agricultural activities were more intensive (43.5 Mt yr⁻¹),
while the decreased sediment flux (40 Mt yr⁻¹) during AD 1348 – 1353 is attributed to the decrease of
population during the Black Death Period. The Ebro basin has been observed to have immense decreases
in sediment flux since the last century (99%) (Vericat and Batalla, 2006), which is also captured by the
model simulations. The decrease is attributed to anthropogenic activities such as dam emplacements and
irrigation activities (Pinilla, 2006; Vericat and Batalla, 2006). However, our simulations show that
emplacement of dams is the most significant factor contributing to the decrease of fluvial sediment flux to
the Ebro Delta in modern times. Chapter 2 demonstrates that anthropogenic forces should be taken into
account when studying fluvial sediment flux for both the 20th century and long-term in basins that have
strong anthropogenic disturbance history.

Most of the world’s major deltas are currently subsiding relative to sea level due to decreased
fluvial sediment flux and intensified marine forces (Ericson et al., 2006; Woodroffe et al 2006; Syvitski et
al., 2009). Coastal storms are one of the greatest hazards to low-elevation and high-population deltas
(Nicholls, 2004). Understanding a delta’s response to coastal storms will be critical in determining how
humans can best protect deltas from marine hazards. Louisiana River Delta is one of the most vulnerable
deltas due to decreased fluvial sediment loads, and natural and anthropogenic factor caused compactions
(Blum and Roberts, 2009; Syvitski et al., 2009). One way to save the delta would be open new channels
along the old channels, which will introduce sediment to shallow areas and form new delta systems (Kim
et al., 2009). It is critical to understand whether these new delta systems will be sustainable under the
frequent coastal storms. Chapter 3 and 4 focus on the morphological changes of a new-formed delta in
response to hurricane events and cumulative cold front events. The study applies the numerical model
Delft3D to a river-dominated delta, the Wax Lake Delta (WLD), in Louisiana, USA. As a newly-formed
and fast-prograding deltaic system (area increases 200 acres per year, FitzGerald, 1998; Roberts et al.,
the major controlling factor of the evolution of the WLD is thought to be fluvial forces (Hanegan, 2011; Roberts et al., 2003, 1997). However, weather events, which in this deltaic system include cold fronts and hurricanes, also play a significant role in determining the evolution of the delta. Stone et al. (1997) stated that 90% of coastline retreat in Louisiana might be caused by hurricanes, and Walker and Hammack (2000) estimated cold fronts transport ~15% of fluvial sediment outside of the bay. The study presented in this thesis is the first numerical simulation that captures morphological changes of the WLD during a hurricane event. The impacts of waves, fluvial forces, hurricane tracks, and vegetation in determining the overall morphological changes on the deltaic system are evaluated. The study illustrates that hurricanes cause erosion on the delta. In total 500,000 m$^3$ of sediment was removed within 100 hours when Hurricane Rita (2005) made landfall 120 km to the west of the WLD as a Category 3 hurricane. The simulation results reveal that erosion mostly occurs on islands and deposition occurs in the major channels, attributed to the influence of winds and waves, which is opposite to the sedimentation/erosion pattern during river flood events (deposition on islands and erosion in channels, Shaw and Mohrig, 2014). The sedimentation/erosion pattern demonstrates that marine forces outweighed fluvial forces in both islands and channels during Hurricane Rita (2005). Aboveground vegetation, which in this study is simulated as 3D cylinders in the water column, decreases flow velocity and sediment transport, leading to the stabilization of half of the sediment that would otherwise be removed when vegetation is not taken into account. This stabilization occurs mainly on the islands. When roots are considered through adding soil strength on vegetated islands in the simulation, the sedimentation/erosion pattern shows minor changes, because the aboveground vegetation has already significantly decreased the flow-wave shear stress on islands. The study also explores the influence of different hurricane tracks on the delta morphology, demonstrating that a side track with the hurricane center located on the west side of the delta causes higher flow velocities and transports more sediment on the delta than a direct track of which the hurricane center passes through the delta. The differences are caused by the distinct wind fields: onshore winds occurring during the sidetrack case cause a higher surge than offshore winds during a direct track.
Chapter 4 compares the morphological responses of the WLD to cold fronts and a single hurricane event in 2008. As each cold front has distinct characteristics, the morphological changes of the WLD to different events (classified based on wind structures) are simulated and compared, demonstrating that wind fields determine the hydrodynamics and that mean wind speeds are the predominant factor controlling the amount of sediment in motion. Events that cause larger water level changes in a short period of time produce a larger amount of sediment transport than events with minor and gradual water level changes. The morphological changes of the WLD during a hurricane event and a strong cold front are compared, showing a similar spatial pattern of erosion on islands although the magnitude is much more pronounced for the hurricane event. The major differences occur in the channels, where the erosional pattern becomes depositional during hurricanes, indicating that marine forces outweigh fluvial forces during hurricanes, in contrast to cold front events. The computations of sediment transport for a cold front season in 2008-2009 and yearly hurricanes in 2008 demonstrate that, although a single hurricane event causes more sediment movement than a single cold front event, the cumulative impacts of cold fronts largely outweigh the influences of hurricanes. Four times more sediment was removed from the study area during the season’s 29 cold front events than during the Hurricane Ike and Gustav. Simulated erosion estimates are 1,900,000 m$^3$ for the 29 cold fronts and 500,000 m$^3$ for the hurricanes (412,000 m$^3$ for Ike and 8,000 m$^3$ for Gustav). The aboveground vegetation, which is active during hurricane season but die in winter, decreased the amount of erosion to 308,000 m$^3$ during the simulation of Hurricane Ike. This study indicates that, although the WLD shows larger morphological changes during a hurricane event, the frequent cold fronts play a more critical role in determining the delta morphology at the coastal areas of the Gulf of Mexico. The influences of these weather systems should be taken into account when studying the evolution of coastal deltas, even for river-dominated systems. Saline water intrusion significantly threatens freshwater wetlands on the WLD system during hurricanes, but its influences during cold fronts are minor.

Uncertainty analysis is implemented for all of the three studies. The extreme uncertainty of modeled fluvial sediment flux for the Ebro river is $47.2 \pm 12.4$ Mt yr$^{-1}$ for the period 1860 – 1960 before
emplacement of dams, which is achieved by estimating uncertainties in climate change factors (precipitation and temperature) and anthropogenic factors (land use) and applying the extreme combinations to model simulations. The uncertainty qualification of Delft3D for Chapter 3 and 4 is implemented through Dakota software, which analyzes the model uncertainty automatically based on the uncertainty in input parameters. Our study shows that the uncertainty in simulated sediment balance on the WLD varies with different environmental settings – the strongest hurricane, Hurricane Rita, has the largest uncertainty (332,000 m³) and the weakest cold front event has the smallest uncertainty (147,000 m³). The contributions of different factors to model uncertainty vary over time so that critical shear stress for mud erosion is the major contributor to the model uncertainty during all three simulated events (Hurricane Rita, Hurricane Ike, and cold front event 11) when the study area is dominated by erosion. The influence of critical shear stress for erosion becomes less important during normal condition when critical shear stress for deposition and silt grain size becomes more important. The areas with the highest residual currents during these events have the largest uncertainty in sediment balance for the reason that variations in critical shear stress greatly influence calculations of sediment dynamics in cells that encounter strong flow and wave shear stress. So model uncertainty has to be quantified within specific environmental settings.
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APPENDIX A: MODEL INPUT AND OUTPUT DATA

All of the following data is available online at http://csdms.colorado.edu/pub/users/fexi8823/ or ftp://csdms.colorado.edu/pub/users/fexi8823/

Figure A.1 Bathymetry of the GoM model domain (from Louisiana Virtual Coast Data Archive)
Figure A.2 Model roughness (Manning) calculated from water depth with equation 3.9
Figure A.3 Wind speed distribution in the GoM domain when Hurricane Rita was over the deep ocean (data available at Rita.amu, Rita.amv, Rita.amp). The largest wind speed appeared in the eyewall of the hurricane, and wind speed was low at the hurricane eye. Wind speed decreased from the hurricane center to the surrounding area.
Figure A.4 Significant wave height distribution in the GoM domain when Hurricane Rita was over the deep ocean (data available in wave-gom_rita.dat). Wave height was highest in the eyewall and low at the hurricane eye, in correspondence with the wind field. The high waves on the west of the hurricane system may have been caused by storm swell.
Figure A.5 Mean wave direction distribution when the Hurricane Rita system was over the deep ocean (data available at wave-gom_rita.dat). Wave direction distribution had a circular structure with the center located at the hurricane eye, and wave direction followed wind direction (anti-clockwise in the northern hemisphere). The sudden changes in wave direction at the coastal area were caused by wave refraction.
Figure A.6 Water level distribution in the GoM domain when the Hurricane Rita system was over the deep ocean (data available at trim-GoM_rita.dat). Water level had a smaller rise in the deep ocean, with a magnitude of around 1 m, because hurricane systems have huge vertical structures that transfer momentum to deep water. There was only a small water elevation rise occurring at the hurricane eye due to the hydrostatic uplift caused by the low center pressure.
Figure A.7 Wind speed distribution in the GoM domain when the Hurricane Rita system approached the coastal area (data available at Rita.amu, Rita.amv, Rita.amp). Hurricane Rita made landfall at Sabine Pass, Texas, with the strongest winds appearing on the east of the system.
Figure A.8 Significant wave height distribution in the GoM domain when the Hurricane Rita system approached the coastal area (data available in wave-gom_rita.dat). Wave height decreased when the hurricane approached the coastal area due to wave dissipation, and the largest wave height appeared on the east of the system where the wind speed was highest.
Figure A.9 Peak wave direction distribution when the Hurricane Rita system approached the coastal area (data available at wave-gom_rita.dat). Wave direction had an anti-clockwise circular structure with the center located at the hurricane eye. The significant changes at the model open boundaries were caused by boundary errors.
Figure A.10 Water level distribution in the GoM domain when Hurricane Rita approached the coastal area (data available at trim-GoM_rita.dat). The storm surge caused by Hurricane Rita was more than 4 m, with the highest value appearing on the east of the system. The areas located on the eastern side of the hurricane experienced a higher storm surge than the western areas.
Figure A.11 Wind speed distribution in the GoM domain after Hurricane Rita made landfall (data available at Rita.amu, Rita.amv, Rita.amp). The color points show the points where wind parameters were given (spatial resolution of 0.05°), which had significantly different values compared to surrounding points.
Figure A.12 Significant wave height distribution in the GoM domain after Hurricane Rita made landfall (data available in wave-gom_rita.dat). Wave height distribution closely followed the wind field (Fig. A.11).
Figure A.13 Mean wave direction distribution in the GoM domain after Hurricane Rita made landfall (data available at wave-gom_rita.dat). Wave direction was mainly determined by wind direction (mostly western).
Figure A.14 Water level distribution in the GoM domain after Hurricane Rita made landfall (data available at trim-GoM_rita.dat). Although at this time step wind speed was larger in the ocean than in the coastal area, the onshore winds caused significant water level variations at the coastal area. Coastal bathymetry magnified the water level rise.
Figure A.15 Bathymetry for the WLD domain (from USACE)
Figure A.16 Model roughness (Chezy) for the WLD domain (from Meselhe et al., 2015)
Figure A.17 Significant wave height in the WLD domain during flooding caused by Hurricane Rita (data available at wavm_WLD_rita.dat). Wave height was larger in deeper areas and influenced by wind direction (southeast directed winds), so here wave height was larger to the west of the islands and in channels.
Figure A.18 Storm surge at the WLD domain during flooding caused by Hurricane Rita (data available at trim_WLD_rita.dat). The storm surge expanded from downstream to upstream, and the highest surge appeared at the land and ocean boundary on the west of the delta due to the southeastern wind field.
Figure A.19 Total amount of sediment transport in the WLD domain during flooding caused by Hurricane Rita (data available at trim_WLD_rita.dat). Sediment was eroded from islands, and most water and sediment was transported through channels, leading to high upstream sediment transport in channels during water flooding.
Figure A.20 Wind speed distribution in the GoM domain when cold front event 11 caused water level to rise (data available at wavm_GoM_CF.dat). The highest wind speed for the largest cold front event in 2008-2009 cold front season was around 14 m s\(^{-1}\), and the mean wind speed at the WLD domain was 11.1 m s\(^{-1}\).
Figure A.21 Significant wave height distribution in the GoM domain when cold front event 11 caused water level to rise (data available at wavm_GoM_CF.dat). Wave height closely follows the wind field.
Figure A.22 Mean wave direction in the GoM domain when cold front event 11 caused water level to rise (data available at wavm_GoM_CF.dat). Wave direction followed wind direction (mainly north directed winds).
Figure A.23 Water level distribution in the GoM domain when cold front event 1 caused water level to rise (data available at trim_GoM_CF.dat). Although wind speed was larger in the ocean than in the coastal area, the onshore winds and coastal bathymetry caused water level to rise significantly (>0.5 m) in the coastal area.
Figure A.24 Significant wave height in the WLD domain when cold front event 11 caused water level to rise (data available at wavm_WLD_CF.dat). Wave height was larger in deep areas, so high waves were distributed in the surrounding areas of the WLD and in channels. The highest wave height was 0.6 m, much smaller than the wave height during the Hurricane Rita simulation (1.6 m, Fig. A.20).
Figure A.25 Water level rise in the WLD domain when cold front event 11 caused water level to rise (data available at trim_WLD_CF.dat). The highest water level was 1.5m, much smaller than that during Hurricane Rita (2.5 m).
Figure A.26 Total sediment transport in the WLD domain when cold front event 11 caused water level to rise (data available at trim_WLD_CF.dat). The total sediment transport was higher in channels than shallow areas, as the hurricane case. The magnitude was more than ten times smaller than that during the Hurricane Rita simulation (300 kg s\(^{-1}\) m\(^{-1}\), Fig. A.19)