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Comparison of Land Building by Mississippi River Diversion Using One and Two Dimensional Numerical Models

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COMPARISON OF LAND BUILDING BY MISSISSIPPI RIVER DIVERSION USING ONE AND TWO DIMENSIONAL NUMERICAL MODELS

A Thesis

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Master of Science in Civil Engineering

in

The Department of Civil and Environmental Engineering

by

Gyan Prasad Basyal
B.E., Tribhuvan University, 2009
July 2014
Dedicated to my parents for their unconditional love and support …
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ABSTRACT

River sediment diversions have been identified as one strategy for creating new land and offsetting Mississippi River delta plain land loss. Numerical modeling is one tool for estimating the amount of land, geomorphic features and ecological benefits from diversions. There are a number of models proposed to estimate sediment diversion land building, ranging from simplistic approaches that provide bulk characteristics and use little computational resources to process-based models that require a large amount of input parameters and computing power.

This thesis aims to compare and contrast two approaches to simulating the land building processes in a simplified receiving basin: a 1D spatially averaged model; and a horizontal 2D, process-based Delft3D model. Four scenarios were run: three with varying amounts of non-cohesive sediment; and one with a mixture of non-cohesive and cohesive sediment. A number of simplifying assumptions were made for more direct comparisons of the bulk and detailed delta properties and the computational resources. These included the bulking of cohesive and non-cohesive sediments on deposition are assumed equal; erosion below the pre-delta strata is not allowed; and the river sediment diversion operates continuously at a given flow and sediment concentration. Note that this last assumption was made for easier model comparisons and not how any proposed diversions would be operated. Distributary channel network information, missing in the 1D model but important for ecohydrological processes, is extracted from the 2D model.

The 1D model took less than one minute to simulate the same scenario that required over 20 hours on 32 processors using the 2D model. Results showed the 1D model delta radii and areas were always larger, but relatively close, to those simulated by the 2D model, particularly for non-cohesive sediments. The deltas formed from solely non-cohesive sediments had numerous short, but wide, channels and were roughly fan shaped, thus justifying the radial symmetry assumption.
of the 1D model. The ratios of the 2D to 1D model delta areas were 70% and 55% for non-cohesive and mixed scenarios, respectively. The 2D model results showed that presence of cohesive sediment promoted narrower and weakly sinuous channels that affect delta growth dynamics and result in increased vertical aggradation, thus limiting the area of land built.
Chapter 1. INTRODUCTION

1.1. Mississippi delta

Southern Louisiana is made up of 9,650 sq. miles of delta plain formed by Mississippi River, called the Mississippi River delta plain (Blum & Roberts, 2012). Rich in natural resources and vital to the U.S. economy, the Mississippi River delta plain is in a state of crisis due to changing landscape as a result of natural and human activities (Blum & Roberts, 2009; Blum & Roberts, 2012; CPRA, 2012). Historically, this region was built during Holocene epoch as the sediment laden Mississippi river changed its course (avulsion) repeatedly to form five major deltaic headlands along the active fluvial course (see Figure 1-1) (Blum & Roberts, 2012). Since the beginning of twentieth-century, unprecedented levels of human activities in the Mississippi River basin have unfavorably altered the hydraulics and natural course of the river. Before the human intervention, the total annual suspended load (pre-dam load) in the lower Mississippi River was estimated to be 400-500 MT (Million Tons). The construction of levees along the Mississippi River, since 1920’s, and dams, since 1950s, for navigation and flood protection, and changes in land use within the drainage basin have reduced the sediment supply in the lower Mississippi River by 50 % to 205 MT year\(^{-1}\) (Blum & Roberts, 2009; Blum & Roberts, 2012). This sediment supply is less than the sediment trapping rate of modern deltas (240-300 MT year\(^{-1}\)) formed during last 12,000 years, and certainly not enough to construct new land or sustain existing land (Blum & Roberts, 2009; Blum & Roberts, 2012). Furthermore, the levees in the lower Mississippi River have cut off distribution of sediment (through avulsion and flooding) from a vast majority of the Louisiana’s wetlands. As a result of reduced sediment load in the river and sediment starvation, the wetlands are deprived of sediment minerals that compensate for increased sediment
accommodation volume due to sea-level rise and natural and anthropogenic subsidence. Studies show that approximately 15 sq. miles of wetlands are lost every year (Blum & Roberts, 2012), 1,880 sq. miles has been lost since the 1930s (see Figure 1-2) (CPRA, 2012) and approximately one-third of its original wetland has been lost since the European settlement of North America (Paola et al., 2011). CPRA (2012) has warned that additional 1,750 sq. miles of land may be lost in next 50 years (see Figure 1-2).

Figure 1-1: Holocene deltas (1, 2, 3, 4 and 5) of the lower Mississippi River. Figure obtained from Blum and Roberts (2012).

Figure 1-2 Land surface gains and losses in the Mississippi River delta plain. Figure obtained from CPRA (2012).
The economic and ecological importance of the Mississippi River delta plain to the state and to the nation cannot be underestimated. Barrier islands, marshes, and swamps throughout the Louisiana’s coast not only reduce incoming storm surge during storms and hurricanes but also play an important role in carbon and nitrogen cycles. Nationally, this region provides infrastructure for 25% of hydrocarbon production, 20% of waterborne commerce through the ports of New Orleans and Baton Rouge, and 30% of coastal fisheries (Day et al., 2007; Blum & Roberts, 2012). The cost of losing these lands is high and, if not checked, the cost can be even higher in future as the coastline approaches New Orleans and other coastal cities and towns, exposing important navigation routes, levees and communities to hurricane, storm surge and flooding. CPRA (2012) estimates that the cost of annual damage by flooding will increase ten times within 50 years if bold coastal protection and restoration measures are not taken to prevent rapid coastal decline of Louisiana’s coast.

1.2. Sediment diversions

Submergence of coastal land in Mississippi River delta plain can’t be totally stopped. The lower Mississippi River is sediment starved and the natural processes by which the historic delta was built by the river have mostly been lost due to engineering of the river for navigation and flooding control purposes. By reconnecting the Mississippi River and its sediment load to the coastal wetlands, some of the negative effects can be reversed and the prevailing land loss rates can be counteracted to some extent (Blum & Roberts, 2009; Paola et al., 2011; Blum & Roberts, 2012). The Mississippi River carries most of its annual sediment load during a short (~2-3 month) high flow period that typically occurs in the spring. Diversion of sediment and water during this period to the adjoining drowned wetlands can be effective in building new land and restoring wetland ecosystems.
A certain degree of confidence can be drawn from the examples of delta growth at the mouth of the Atchafalaya River, Wax Lake Delta (WLD) and Cubits Gap regarding the use of a river sediment diversion for Mississippi River delta restoration. To deal with flooding in Atchafalaya River, the Wax Lake diversion was constructed in 1941. After the flood of 1973 and subsequent floods, significant sediment deposition occurred at the outlets of Atchafalaya River and Wax Lake diversion. Since 1980, the Wax Lake outlet started building land seaward in the Atchafalaya bay at an annual rate of ~3 km² (Majersky et al., 1997; Paola et al., 2011). Another relevant example is Cubit’s Gap, a small crevasse dug alongside the main-stem of the Mississippi River between 1862 and 1868 as a shortcut for fisherman. The width of the crevasse kept increasing, diverting enough sediment from the river by 1922 to create a delta the same size as present day WLD (Paola et al., 2011). However, subsequent reduction in sediment supply as a consequence of decreasing crevasse width without human intervention led to the transgressive phase of the Cubit’s gap sub-delta.

Above examples have evoked much interest in building one or more sediment diversions along the Lower Mississippi River. The 2012 Louisiana Coastal Master Plan has placed a high priority on large land building projects, including eight small and large scale sediment diversions (see Figure 1-3). A suitable diversion site is a tradeoff between factors such as availability of sediment in river, diversion location along the river, sustainability of newly formed land, human settlement, etc. Studies have suggested the reach between New Orleans (RK167) and Venice (RK20) is suitable for Mississippi diversion since it is downstream of major infrastructures, has high stream power and sediment load compared to the areas further downstream and the communities are already protected against increasing water level by storm surge levees (Allison & Meselhe, 2010).
1.3. Objective

Sediment diversions have been identified as one of the primary approaches for restoring and/or maintaining some of the Mississippi River delta and coastal wetlands. Similar to natural river deltas, man-made sediment diversions have the potential to create new land and wetlands through the continuous cycle of delta lobe extension, avulsion, and abandonment. Simulations of deltaic evolution, important to delta restoration project planning, design and permitting is necessary to quantify important physical characteristics and to evaluate land building and ecological benefits. Two major hurdles to long-term (e.g., decadal) deltaic geomorphic modeling are: (1) our understanding of complex deltaic processes (i.e., coupled biologic, geochemical and physical processes) is still somewhat limited; and (2) results from reduced complexity numeric models can’t draw high degree of confidence (Paola et al., 2011). Thus, there are still some questions regarding the appropriate level of modeling and application of the results.
Approaches to deltaic geomorphologic modeling range from a relatively simple 1D spatially averaged models (e.g., Kim et al. (2009b) model) that simulate decades of delta growth using a spreadsheet to sophisticated process-based models that require multiprocessor computers and acceleration factors to simulate similar time scales. A one dimensional, mass balance, numerical model has been applied to hypothetical diversions in Barataria Bay and Breton Sound to predict land building capacity of the diversions (Kim et al., 2009a; Kim et al., 2009b; Paola et al., 2011). This land building model has been calibrated and verified against the observed evolution of the Wax Lake Delta. Because of the assumptions and simplifications used in the 1D approach, the models cannot account for the complex receiving basin topography and bathymetry or simulate the geomorphic processes and complexities common to many of these systems and important for healthy ecosystems (Kostic & Parker, 2003b; Kim et al., 2009a; Kim et al., 2009b). On the other hand, a process-based model, such as Delft3D, is able to incorporate many of these important features and processes such as mouth bar formation and deltaic evolution in generalized shallow and deep receiving basin conditions (Edmonds & Slingerland, 2007; Storms et al., 2007; Edmonds et al., 2009; Edmonds & Slingerland, 2009; Geleynse et al., 2010; Geleynse et al., 2011). However, the computational requirements can limit our ability to simulate on decadal time scales. For both kinds of model, unavailability of observed field data or inconsistent quantitative description of landforms can inhibit our ability to have high degree of confidence in the simulated results (Paola et al., 2011; Geleynse, 2013).

The objective of this work is to quantitatively compare sediment diversion simulation results from these two modeling approaches in Louisiana-relevant basin system with simple geometry. Comparisons will be made in two areas: (1) Bulk land building characteristics (e.g., radial extent, subaerial area); and (2) data and computational requirements. In addition, ecosystem-relevant
deltaic properties (e.g., channel network characteristics) will be estimated from the process-based model simulations to get a better understanding of what is missing in the 1D modeling approach.

These objectives will be accomplished by:

1. Developing a simplified horizontal 2D model in Delft3D for proposed Barataria Bay diversion by Kim et al. (2009b).

2. Developing a 1D model with minor modification on the Kim et al. (2009b) delta model, so that the same inflow boundary conditions and diversion operation strategy can be applied to both 1D and 2D models.

3. Operating the diversion at constant flow and sediment inflow values. The sediment inflow values are chosen with reference to sediment rating curve at Belle Chasse.

4. Extracting and quantifying simulation results for both direct and indirect comparisons.

5. Developing of image analysis Matlab codes to process the 2D model results in order to extract delta boundary and area, channel properties, etc.

More details are provided in Chapter 3.
Chapter 2. LITERATURE REVIEW

2.1. River delta

River deltas are coastal features developed from river mouth sediment deposition and dispersal (Wright, 1977; Syvitski, 2008). Syvitski (2008) has defined a river delta in six ways. First off, a delta is the seaward prograding land area that has accumulated over the past 6,000 years, when global sea level stabilized within a few meters of the present level (Amorosi & Milli, 2001). Secondly, it is the seaward area of a river valley after the main stem of a river splits into distributary channels (Syvitski & Saito, 2007). Thirdly, a delta is the area of a river valley underlain by Holocene marine sediments (Kubo et al., 2006). Another definition of a delta is the accumulated river sediment that has variably been subjected to fluvial, wave, and tidal influences. The final definition calls a delta the area drained by river distributary channels that are under the influence of tides. Note that a delta can also be any combination of the above definitions.

Delta plains can be very flat. Steeper mountainous river deltas generally have slopes up to 0.005 while larger deltaic systems have slope as low as 0.00001 (Syvitski, 2008). A delta’s low gradient is both attractive and hazardous to human utilization and occupation. A large flat delta is attractive because it has the potential for easy agricultural development, made further attractive by its rich organic soil. Importantly, population centers are often located in deltas. Fifty-one of the world’s deltas have a combined 2003 population of 325 million (ORNL (Oak Ridge National Laboratory), 2002; Syvitski, 2008). However, low delta gradients can contribute to a dangerous environment for human habitation, allowing river flooding to spread across the flat delta plains through distributary channels that often switch their location and direction. Twentieth-century engineering has partly ameliorated floods with upstream dams, dikes, levees, and other flood control structures. But these structures also encourage people to live in environmentally dangerous
areas, putting themselves at risk when larger flooding event occur. Engineering often provides little protection from ocean-generated storm surges (Syvitski, 2008).

In general, most deltaic features scale with the magnitude of a river’s discharge, sediment load, the number of distributary channels, and the gradient of the deltaic plain. The larger deltas have the lowest gradients (Syvitski et al., 2005; Syvitski, 2008). Delta morphology is primarily dictated by interaction of waves, tides, and river discharge. The coastline of a delta moves as a function of the direction of global ocean volume, more regional earth-surface load change, sediment supply, and compaction of the deposited sediment (Ericson et al., 2006). Reduced complexity numerical models are one of the ways used to study and predict delta’s morphological features. These physics based models consider only the dominant processes and parameters that govern delta building.

2.2. Land building models

Parker et al. (1998) formulated a model for the evolution of alluvial fans, which are fan-shaped zones of sedimentation downstream of an upland sediment source such as river. Fluvial fans may be completely terrestrial or formed in the standing water, the latter also known as fan-deltas (Nemec, 2009). Parker et al. (1998) assumed a conical fan and a single fluvial channel. Their approach did not intended to locate individual channels. The channel distributaries in the fan were modeled with a single effective channel (virtual channel) that ranges over the fan and delivered sediment. As the deposition occurred in the channel, the deposit was spread across the entire width representing the sequence of channel shifts in a long run.

Parker et al. (1998) model was extended by Kostic and Parker (2003a), Parker et al. (2008a), Parker et al. (2008b) and Kim et al. (2009a) to model delta progradation in a standing water. These models composed of delta with four zones: (1) sediment source or vertex (2) delta topset, low slope subaereal surface formed by fluvial (coarse) sediment deposition (3) foreset or delta face formed
by avalanching of sediment to deep water and/or (4) bottomset, deposit of fine sediment on the basin bed.

River diversions, such as planned diversions in Mississippi River for delta restoration, are scaled-down versions of natural delta lobes. Delta building models are also used to study bulk characteristic of land (such as area, extension, etc.) formed by Mississippi River diversions in Barataria Bay and Breton Sound areas (Figure 2-1). This model, called Kim et al. (2009b) model here, was based on earlier works by Parker et al. (1998), Kostic and Parker (2003a), Parker et al. (2008b), Kim et al. (2009a), etc. This model consisted of a diversion channel, delta topset and avalanching foreset (Figure 2-2). This model assumes transport of sand in the active channel using empirical transport formula while transport and deposition of cohesive sediment depends empirically on sand transport. The main goal of Kim et al. (2009b) model was to predict most important bulk properties like overall rate of shoreline advance and growth in delta area, given the water and sediment supply and local relative sea level rise.

Figure 2-1 Sediment diversion scheme for Mississippi delta restoration by Kim et al. (2009b). The wetland areas shown are predictions at 50 years after diversion construction assuming mean subsidence and sea-level rise and other base case parameters as described in Kim et al. (2009b). Figure obtained from Paola et al. (2011).
Dean et al. (2012) have offered much simpler model especially targeted for river diversions. This model utilizes simple analytical method specifically for the purpose of guiding design and preliminary evaluation by examining interrelationship among key variables. Dean et al. (2012) model uses two simple geometries for delta built by diversion: (1) Truncated cone (Figure 2-3) and (2) Uniform width trapezoid (Figure 2-4). The volume of these geometries are equal to the volume of sediment (including bulking after deposition) fed to the system. The key variables include water depth, sea level rise, subsidence, input of sand, operation strategies, and the role of vegetation and changing bulk density and stabilizing the deposits. Model results from both geometries show a clear life cycle of growth and deterioration in a delta that experiences relative sea level rise and subsidence for a constant sediment discharge. A comparison of subaerial deposits in larger versus smaller diversions, assuming the same total sediment discharge in both cases, reveals that the total subaerial land area for the larger diversions is substantially greater than the sum of the two volumes of the smaller diversions.
2.3. Delta formation studies

Deltas evolve from the interaction of numerous complex processes such as multiple flow channelization, sediment transport, vegetation, and biochemical processes, which aren’t explicitly considered by spatially averaged delta models discussed in the previous section. Splitting at the mouth bar deposit plays a major role in the structure of the channel network over the delta top.
The channel network over the delta top has a crucial role in the evolution of the delta surface and its ecosystem. Reduced complexity numerical models that include only the dominant processes and are accurate for the scales and processes of interest can offer a means to examine relationships between delta processes and the resulting morphology (Storms et al., 2007). Such a process-based model provides level of details beyond the capability of a spatially averaged one dimensional delta model (Paola et al., 2011). Studies by Storms et al. (2007), Edmonds and Slingerland (2007), Edmonds and Slingerland (2009), Geleynse et al. (2010) and Geleynse et al. (2011) demonstrate the use of a process-based Delft3D model. The Delft3D modeling package solves the basic equations for fluid flow and sediment transport processes. These studies have demonstrated the ability to simulate delta evolution in a realistic manner as channels bifurcate around mouth bars and avulse producing natural features. A major obstacle for detailed delta modeling, however, is the lack of field or laboratory data for quantitative calibration and validation (Paola et al., 2011). Nevertheless, process-based models have become a useful means to assess the effects of variables in initial stage of delta evolution.

A river mouth is the most fundamental element of deltaic system (Wright, 1977). Fluvial sediments deposit and disperse at the mouth to create a complex deltaic surface with numerous middle bars and distributary channels over time. Using a Delft3D based model, Edmonds and Slingerland (2007) explained that topology of distributary networks in river dominated deltas, esp. the decrease in channel length (the distance between successive bifurcation points) towards the distal ends of distributary networks, is a function of mouth bar dynamics. Their model results suggest that distance of mouth bar from the tip of distributary channel is a positive function of jet momentum flux and negative function of grain size. The jet momentum flux depends on initial channel width, initial channel depth, initial channel velocity, and basin slope (Edmonds &
Slingerland, 2007). The initial channel depth and velocity were important variables for jet momentum flux than the channel width. In each channel resulting from bifurcation, there is decrease in discharge, velocity, depth and turbulent momentum flux. As a result, each bifurcated channel forms mouth bar at shorter distance. Thus, with each bifurcation, the length between successive bifurcations decreases.

Storms et al. (2007) have used a Delft3D based model to simulate the initial delta formation from a river dominant effluent discharging constant flow and sediment loads into shallow (simulation A) and deep (simulation B) receiving basins (Figure 2-5). Though the model was not meant to simulate any real location, the simulation results in the shallow basin resemble the morphology and stratigraphy of the Wax Lake Delta. One of the main differences between Wax Lake Delta and Atchafalaya delta is the width of the river effluent. In order to understand the effects of width of effluent in shallow water basin, two additional simulations C and D, comparable to setup of simulation A but with larger effluent widths, were conducted (Figure 2-6). Simulation C, which had much larger channel width than simulation A, did not allow the formation of middle ground bar. This led to an elongated delta morphology with shoals aligned along the channel levees. For simulation D, which had an intermediate effluent width between simulations A and C, a middle ground bar developed that caused the main channel to bifurcate, similar to simulation A. However, the bifurcation angle of simulation C was much smaller than for simulation A and compared well to that of Wax Lake Delta. Overall, the results suggest that process-based hydrodynamic and transport models are capable of simulating important delta formation processes.
Edmonds and Slingerland (2009) used depth-averaged Delft3D based model to establish sediment bulk cohesion control on overall distributary network and floodplain structure, shoreline smoothness, and bifurcation angle. The bulk cohesion of the sediment was varied by systematically altering the critical shear stress for erosion of the cohesive sediment or the proportion of cohesive to non-cohesive sediment load (Edmonds & Slingerland, 2009). Findings correlated with the delta formation processes observed in field and model studies; the delta distributary network was
generated by the growth of subaqueous levees and mouth bars, mouth bar stagnation and channel bifurcation, breaching of mouth bars and subaqueous levees to form multiple bifurcations, and channel avulsion (Edmonds & Slingerland, 2007). Overall, the results indicated that deltas formed from more cohesive sediment tend to be elongate with long channels, complex floodplains, and rough shorelines; whereas, deltas formed with less sediment cohesion tended to be fan-shaped with approximate radial symmetry and smooth shorelines (Figure 2-7). Cohesion increases the shear stress needed to re-erode deposited sediments, so the depositional structures involved in channel bifurcation were more stable. Subaqueous levees, resistant to erosion due to high cohesion, concentrate the flow into a narrow channel that easily prograded basinward. When a mouth bar was able to form, it generally formed stable bifurcations and was not easily breached by new channels. Similarly, avulsion frequency was low due to the difficulty in breaching high-cohesion levees. Low cohesion formed fewer bifurcations because mouth bars were continually re-eroded and deposited basinward so the stagnation that is required for bifurcation occurred less frequently. Levees were more easily eroded, allowing numerous breaches that distribute flow to the full delta and promote radially-uniform progradation. Deltas with intermediate cohesion formed the most bifurcation-dominant network structures because the balance of mouth bar stability and subaqueous levee breaching formed the most stable bifurcation angles (Edmonds & Slingerland, 2009).

Research described in Geleynse et al. (2010) and Geleynse et al. (2011) have further investigated the effects of fluvial channel length, input sediment composition, initial bed sediment composition and the effects of tide and wave forcings on delta morphology. Geleynse (2010) found that upstream river channel bar migration affects the feeder channel flow distribution to such a degree as to nearly prevent flow towards particular direction, indicating possible upstream controls
on avulsion. Figure 2-8 shows delta morphology results from Geleynse et al. (2011) for fluvial input with varying degree of sand-silt fraction and tide and wave forcing.

Figure 2-7 Variability of delta morphology based on cohesion of the sediment. In the figure total cohesion increases from bottom right to upper left. Figure obtained from Edmonds and Slingerland (2009).

Figure 2-8 Initial delta simulations by Geleynse et al. (2011) showing the influence of sediment fractions and various forcings. Figure obtained from Geleynse et al. (2011).
Chapter 3. METHODOLOGY

Mississippi River sediment diversions have been proposed as a method to restore the natural process of sediment redistribution in the adjoining shallow coastal areas. Such diversions are intended to create new land in a manner similar to natural delta formation. The benefits from these diversions may be estimated from physical characteristics of the formed delta such as land area and radial extent. Components of delta such as channel length and width have been found to govern the productivity of aquatic plants and animals (Twilley et al., 2008). Therefore, ecological benefit of delta may as well be measured upon channel characteristic. This thesis is aimed at comparing two numerical modeling methods for simulating delta growth from a sediment diversion: (1) a process-based 2D numerical model based on Delft3D which simulates the land building capacity as well as other deltaic characteristics such as channel density and mean width; and (2) a 1D delta model that simulates only the bulk characteristics of delta growth. All the employed methodologies are described in this chapter.

3.1. 1D model

The one dimensional model used in this study is a modified version of Kim et al. (2009b) delta model. The governing equations and simulation approach, as well as the modifications made for this thesis, are discussed in the following sections.

3.1.1. Difference between 1D model and Kim et al. (2009b) model

The differences between the two models are follows:

1. Operation of diversion: Kim model assumes operation of diversion during the flood period in Mississippi River. 1D model assumes continuous operation throughout the year.

2. Sediment inflow: Kim model uses average annual sediment load in Mississippi River, representative flood discharge in the river, and intermittency factor to characterize
sediment volume entering the diversion channel. 1D model simply uses user-defined sediment concentration.

Kim et al. (2009b) model used flow Intermittency factor, I, to account for flow variability in the main river. This factor allows the use of a constant flow rate, usually bankfull flow, instead of using individual annual hydrographs or flood events. The Intermittency factor is defined as the faction of time (annual) required for a given bankfull flow to transport the same amount of sediment as the actual annual hydrograph. Since most of the annual sediment load is carried by river like Mississippi River during the flood, it may also be defined as the fraction of time the river is in flood (Wright & Parker, 2005).

Intermittency factor was calculated by Wright and Parker (2005) based on flow duration and sediment (sand) rating curves at Tarbert Landing, LA, a Mississippi River gauging station. The annual sand load was calculated by Wright and Parker (2005) by integrating the flow and sediment rating curves. From flow duration and sediment rating curves, the amount of sediment each flow could carry annually was calculated and plotted. From this plot, the flow that transported the most sediment annually (peak) was selected. This flow was found to be 25,600 m³/s, which closely approximated the bankfull flow condition in the lower Mississippi River. The amount of time (in fraction of a year) this representative flow required to transport the annual sand load was determined to be 0.34, which is the Intermittency factor.

The estimated mean annual suspended sediment load in lower Mississippi River is 124 MT yr⁻¹, 17% of it is sand (Allison et al., 2000). The estimated mean annual sand transport as bed load is 2 MT yr⁻¹ (Nittrouer et al., 2008). This gives an average annual sand discharge of 23.08 MT yr⁻¹ in the lower Mississippi River (Kim et al., 2009b). Sand load or concentration of sand during the operation of diversion or the flood period was obtained by dividing the average annual sand
load by the Intermittency factor. The diverted flow is assumed to have same concentration of sand as in the main river.

As mentioned before, the current model assumes sediment diversion operating continuously all year at a specified flow and sediment discharge. The modified governing equations are discussed in the following section. The sediment inflow scenarios are described in section 3.5.7.

3.1.2. Flow equations

In the diversion and the virtual channel in the delta, normal flow assumption is combined with constant boundary friction coefficient (Kim et al., 2009a). This gives:

$$\tau = \rho C_f U^2 = \rho g H S$$

3-1

where,

- $\tau$ boundary shear stress
- $\rho$ density of water
- $g$ acceleration due to gravity
- $C_f$ boundary friction coefficient (dimensionless)
- $U$ mean flow velocity
- $H$ normal flow depth
- $S$ bottom slope of the channel

The dimensionless Shield’s number, $\tau^*$, describing the sediment mobility in the flow within the channel is assumed equal to channel-forming Shield’s number (1.86) for sand-bed streams (Parker et al., 1998; Kim et al., 2009b).

Non-cohesive sediment (sand) transport, $Q_s$, in the channel is given by Engelund and Hansen (1967) total transport formula. This relation though determined from laboratory data is one of the most simple and accurate transport formula (García, 2008).
Combining all the assumptions, the following primary equation is obtained:

\[
Q_s = B \sqrt{gR D_{50}} \frac{0.1 \tau^*}{C_f} (\tau^*)^{2.5}
\]

where,

\begin{align*}
B & \quad \text{width of the active channel (diversion or virtual channel)} \\
R & \quad \text{submerged specific gravity of sand or quartz (1.65)} \\
D_{50} & \quad \text{median size of sand}
\end{align*}

3-2

Combining all the assumptions, the following primary equation is obtained:

\[
\frac{Q_s}{Q_f} = \frac{0.1 \tau^*}{R \sqrt{C_f}} S
\]

where,

\begin{align*}
Q_f & \quad \text{flow in the channel}
\end{align*}

3-3

Equation 3-3 implies that the sediment carrying capacity is linearly dependent on the channel slope. In this equation, water discharge and the initial profile of the delta is provided. The remaining terms except \(Q_s\) are all constants, therefore, which allows to calculate the value of sediment flux. Once the value of sediment flux is known at a location, the surface profile can be updated using conservation of mass equation, which is discussed in the following section. The boundary conditions for equation 3-3 are inflow sediment load and bed slope at the intake of diversion channel. The bed slope boundary value is such that the inflow sediment load is fully transported.

3.1.3. Morphological equation

Transport of non-cohesive sediment was discussed in the previous section. As for mud, the model assumes that \(\lambda\) fraction of mud is deposited for each unit of non-cohesive sediment deposit. The bed level profile is calculated using the Exner’s equation (Paola & Voller, 2005):
In the diversion channel, the sediment flux is only in the flow direction \((q_y = 0)\). On the topset, the basic Exner’s equation is radially integrated between delta opening angle \((180^\circ)\) due to the radial symmetry condition of the delta. This is simply given by:

\[
(1 - \lambda_p) \frac{\partial \eta}{\partial t} = -(1 + \lambda) \left( \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} \right)
\]

where,

- \(\eta\) bottom elevation
- \(\lambda_p\) porosity of bed
- \(q_x, q_y\) sand flux in flow and perpendicular directions
- \(\lambda\) unit of mud that is deposited for each unit of sand deposit

In the diversion channel, the sediment flux is only in the flow direction \((q_y = 0)\). On the topset, the basic Exner’s equation is radially integrated between delta opening angle \((180^\circ)\) due to the radial symmetry condition of the delta. This is simply given by:

\[
(1 - \lambda_p) B_r \frac{\partial \eta}{\partial t} = -(1 + \lambda) \left( \frac{\partial Q_r}{\partial r} \right)
\]

where,

- \(B_r\) width of delta at radial distance \(r\) \((\pi r)\)
- \(Q_r\) sediment discharge in virtual channel at radial distance \(r\)

### 3.1.4. Prograding delta boundary

The top and bottom of the foreset deposit are moving boundaries of the delta on the receiving basin side (Figure 3-1). It is assumed that the sediment delivered to the shoreline (top of foreset) is wholly deposited in the foreset, resulting in basinward progradation of the delta. Foreset slope, \(S_f\), is a constant parameter. Let \(r_s\) and \(r_u\) be the radial distance of top and bottom of the foreset respectively. By integrating Exner’s equation across the delta foreset \((r = r_s \text{ to } r_u \text{ and } \theta = \pm90^\circ)\), migration speeds of the two boundaries \((r_s, r_u)\) are obtained as:
\[ r_s' = \left( \frac{2 \ 1 + \lambda}{\pi \ 1 - \lambda p} Q_{ss} \right) [(S_f - S_s)(\theta \Psi - r_s^2)]^{-1} \tag{3-6} \]

\[ r_u' = \frac{\dot{Z}}{S_f - S_b} + r_s' \frac{S_f - S_s}{S_f - S_b} \tag{3-7} \]

where,

- \( Z, \dot{Z} \) receiving Basin water level, rate of change of this level (here zero)
- \( S_s \) slope of topset at shoreline
- \( S_b \) Slope of basin bottom
- \( \theta = \frac{1}{\int_{-0.5\pi}^{0.5\pi} \left(1 - \frac{S_b}{S_f} \cos\theta\right)^{-2} \ d\theta} \)
- \( \Psi = \left( r_s + \frac{Z - \eta_{vertex}}{S_f} \right) \)

Figure 3-1 1D model set up: profile view (left); plan view (right)
3.2. Delft3D

In this study Delft3D is chosen as the software to simulate 2D hydrodynamics and 3D morphological changes. In order to understand the model setup, methodologies and results from 2D model runs, it is necessary to understand Delft3D software package. This chapter will first give insight on the general utility of this software and then provide details of the important mathematical equations and numerical methods underlying it and applicable to this study. Most of this information has been extracted from Deltares (2013), Lesser et al. (2004), Elias (2006) and Hillen (2009).

3.2.1. Overview

Delft3D is a modeling software for coastal, river and estuarine areas, developed by Deltares, formerly WL|Delft Hydraulics, in close cooperation with Delft University of Technology (Lesser et al., 2004). It can be used to simulate flow, sediment transport, wave, water quality, morphological development and ecological processes (Deltares, 2013).

The main component of the Delft3D package is the FLOW module, which performs hydrodynamic computations with the capability for simultaneous calculation of salinity, heat, tracer and sediment transport. A large number of processes are included in Delft3D-FLOW: wind shear, waves, tides, density driven flows and stratification due to salinity or temperature gradients, atmospheric pressure changes, drying and flooding events, etc. (Lesser et al., 2004). The most useful feature of the FLOW module is its ability to simultaneously compute and update the flow, sediment transport and morphological change (Figure 3-2).
Delft3D’s general flow chart showing processes that occur in a single time step. Figure obtained from Hillen (2009).

Delft3D-FLOW allows transport of both cohesive and non-cohesive sediments and accounts for non-equilibrium sediment concentration profile as well. The module is flexible so that one can choose a non-cohesive sediment transport formula from the internal library: Van Rijn (1983, 1984), Engelund and Hansen (1967), Meyer-Peter-Muller (1948), Bijker (1971), Soulsby/Van Rijn, Soulsby, Ashida-Michiue (1974), Wilcock-Crowe (2003), Gaeuman et al. (2009) laboratory calibration, Gaeuman et al. (2009) Trinity River calibration, etc. formulae or can even create a custom formula. This capacity allows for more user flexibility in using Delft3D to investigate sedimentation and erosion problems in complex flow situations.

3.2.2. Numerical grid

Delft3D uses orthogonal curvilinear grid in horizontal plane. In this work, the special case of orthogonal curvilinear grid, rectangular grid, is used to formulate Delft3D equations. Delft3D uses Arakawa C-grid for discretization of hydrodynamic variables water level (\(\zeta\)) and velocities (\(u\) and \(v\)). The water level points are defined in the center of a cell. The velocity components are perpendicular to the grid cell faces.
Five set of points can be distinguished as indicated in Figure 3-3: bed level points ($z_b$), water level points ($\zeta$), velocity points in two directions ($u$ and $v$), suspended and bed load transport rates in two directions. Water depth, sediment concentration and sediment transport rates are all determined at the water level points. The staggering becomes important when specifying the boundary conditions, and when interpreting the results in detail.

Figure 3-3 Spatial arrangement of Delft3D’s variables in horizontal grid. Figure from Jagers (2003).

3.2.3. Flow equations

Two and three dimensional flows can be simulated in Delft3D. In this study, two-dimensional flow and transport has been used. The 2D shallow water equations follows from 3D equations by depth integration. Numerically Delft3D-FLOW solves the Navier Strokes equations for an incompressible fluid under shallow water and Boussinesq assumption for unsteady flow
calculation (Deltares, 2013). The shallow water assumption is used in situations where horizontal length scale (i.e. length of river) is significantly larger than the vertical scale (i.e. river depth). The following conditions have to be met in order for the shallow water equations to be applicable:

1. The vertical momentum exchange is negligible and the vertical velocity component is a lot smaller than the horizontal components.

2. The pressure gain is linear with the depth

The shallow water assumptions make it possible to reduce the basic system of equations to only three equations: one continuity equation, and the two (x and y) momentum equations. The 2D continuity equation is given by:

\[
\frac{\partial \zeta}{\partial t} + \frac{\partial HU}{\partial x} + \frac{\partial HV}{\partial y} = 0
\]

where,

\(\zeta\) free surface or water elevation

\(U, V\) depth averaged velocity in x and y directions

\(H\) total water depth

The momentum equations in x and y directions are:

\[
\frac{\partial U}{\partial t} + U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial y} = fV - g \frac{\partial \zeta}{\partial x} + \nu_H \left[ \frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} \right] - \frac{\tau_{\text{bed},x}}{H \rho}
\]

\[
\frac{\partial V}{\partial t} + U \frac{\partial V}{\partial x} + V \frac{\partial V}{\partial y} = fU - g \frac{\partial \zeta}{\partial y} + \nu_H \left[ \frac{\partial^2 V}{\partial x^2} + \frac{\partial^2 V}{\partial y^2} \right] - \frac{\tau_{\text{bed},y}}{H \rho}
\]

where,

\(g\) Acceleration due to gravity

\(f\) Coriolis parameter

\(\nu_H\) horizontal eddy viscosity coefficient
The terms on the left hand side of above equations represent respectively: local flow acceleration, and advection terms in both directions. The terms on the right hand side of above equations represent respectively: coriolis force, horizontal pressure, horizontal Reynolds stress and bed friction. Coriolis force is neglected in this study.

The bed stress is bound to the depth-averaged velocity by a quadratic velocity law:

\[ \tau_{\text{bed},x} = C_f \rho U \sqrt{U^2 + V^2} \]
\[ \tau_{\text{bed},y} = C_f \rho V \sqrt{U^2 + V^2} \]

3-11 (a) 3-11 (b)

\( C_f \) boundary friction coefficient \( g/C_z^2 \)
\( g \) acceleration due to gravity

As mentioned previously, Delft3D-Flow uses a finite difference scheme to discretize the shallow water equations. Alternating direction implicit method is used to solve the discretized continuity and horizontal momentum equations (Stelling & Leendertse, 1992). Accuracy of at least second-order is achieved with ADI method (Stelling & Leendertse, 1992; Deltares, 2013).

**3.2.4. Courant condition**

The Courant number gives an indication of numerical stability and accuracy. Since ADI is semi-implicit method, it doesn’t need to fulfill strict Courant-Friedrichs-Lewy (CFL) condition for stable solution. Courant number for Delft3D is given by:
where,

\[ C = 2\Delta t \sqrt{\frac{gH}{\frac{1}{\Delta x^2} + \frac{1}{\Delta y^2}}} \]  

\[ \Delta t \]  time step in second

\[ g \]  acceleration due to gravity

\[ H \]  water depth

\[ \Delta x, \Delta y \]  grid resolution in x and y directions

Smaller courant number can be obtained by adjusting (decreasing) time step or (increasing) grid size. Though courant number up to 10 is suggested for practical flow problems, an upper limit of \(4\sqrt{2}\) is prescribed for critical cases such as flow around island, along irregular closed boundaries and flow through zig-zag channels, etc.

### 3.2.5. Sediment transport

The sediment transport module supports both bed load and suspended load transport of non-cohesive sediments and suspended load transport of cohesive sediments. For schematization, sediment is distinguished as mud (cohesive suspended load transport mode), sand (non-cohesive bed load and suspended load transport modes) and bed load (non-cohesive bed load only or total load transport mode) fractions (Figure 3-4 and Figure 3-5). The difference between bed load and sand fractions lies in the fact that advection-diffusion is not solved for the former. Silt size fraction represent the transition between non-cohesive and cohesive sediments. It can be modeled either as cohesive or non-cohesive fraction (Deltares, 2014).
Figure 3-4 Classification of sediments in Delft3d. Figure from Deltares (2014).

![Figure 3-4 Classification of sediments in Delft3d.](image)

Figure 3-5 Classification of sediment transport modes in Delft3D. From Deltares (2013).

Suspended sediment transport for non-cohesive sediment is computed using the depth-averaged advection-diffusion equation (Galappatti & Vreugdenhil, 1985; Elias, 2006):

\[
\frac{\partial \bar{c}}{\partial t} + U \frac{\partial \bar{c}}{\partial x} + V \frac{\partial \bar{c}}{\partial y} = +D_H \left[ \frac{\partial^2 \bar{c}}{\partial x^2} + \frac{\partial^2 \bar{c}}{\partial y^2} \right] + E - D \tag{3-14}
\]

where,

- \( \bar{c} \) depth-averaged sediment concentration
- \( H \) flow depth
The net erosion-deposition flux, \( E - D \), on the right hand side of the above equation is given as:

\[
E - D = H \frac{\bar{c}_{eq} - \bar{c}}{T_s}
\]

where,

- \( \bar{c}_{eq} \): depth-averaged equilibrium sediment concentration
- \( T_s \): suspended sediment adaptation time scale (Galappatti & Vreugdenhil, 1985)

Depth-averaged equilibrium sediment concentration requires finding of equilibrium suspended sediment concentration and vertical velocity profiles for assumed steady and uniform flow condition. Vertical velocity profile is calculated according to law of the wall principle. The equilibrium concentration profile, \( c_{eq}(z) \), where \( z \) is vertical distance above the channel bottom, is computed for steady uniform flow by balancing downward settling flux with upward turbulent flux (Schmidt’s equation):

\[
\dot{w}_s c_{eq} + D_v(z) \frac{\partial c_{eq}(z)}{\partial z} = 0
\]

where,

- \( \dot{w}_s \): sediment settling velocity
- \( D_v \): vertical diffusivity or mixing coefficient

“Parabolic constant” vertical diffusivity coefficient \( (D_v) \) is modeled as recommended by Van Rijn (1993):

\[
D_v = \beta \kappa z (1 - \frac{z}{H}) u_* \quad \text{for } z/H < 0.5
\]
where,

- \( u_* \) bed shear velocity
- \( \beta \) Van Rijn Beta factor
- \( \kappa \) von Karman constant (0.4)

\[
D_v = \beta 0.25 \kappa u_* H \quad \text{for } z/H \geq 0.5 \quad 3-17 \text{ (b)}
\]

Finally, depth-averaged equilibrium sediment concentration is calculated by numerical integration as follows:

\[
\overline{c_{eq}} = \frac{\int_{z_b}^{z_b+H} u_{eq}(z)c_{eq}(z) \, dz}{H \sqrt{U^2 + V^2}}
\quad 3-18
\]

Bed load transport is calculated by Van Rijn (1993). The reference concentration \( (c_a) \) required in this formulation excluding waves is given by:
The magnitude of bed load transport rate (kg/m/s) computed at the cell center is given as (Elias, 2006):

\[ c_a = f_{sus} \phi 0.015 \rho_s \frac{d_{50}}{a} T^{1.5} D_{*}^{0.3} \]

\( f_{sus} \)  multiplication factor for suspended sediment reference concentration
\( \phi \)  relative availability of sediment fraction in bed layer
\( c_a \)  mass concentration at reference height
\( d_{50} \)  median sediment size in bed
\( D_* \)  non-dimensional particle diameter

\[ D_* = d_{50} \left( \frac{Rg}{\vartheta^2} \right)^{1/3} \]

\( T \)  non-dimensional bed-shear stress

\[ T = \frac{\tau_b - \tau_{b,cr}}{\tau_{b,cr}} \]

\( R \)  submerged specific gravity of sand or quartz
\( \vartheta \)  kinematic viscosity of fluid
\( \tau_b \)  bed shear stress
\( \tau_{b,cr} \)  critical bed shear stress = \((\rho_s - \rho)g d_{50} \theta_{cr}\)
\( \theta_{cr} \)  Shield’s parameter modelled by Van Rijn (dimensionless)

The magnitude of bed load transport rate (kg/m/s) computed at the cell center is given as (Elias, 2006):

\[ |S_b| = f_{Bed} 0.5 \phi \rho_s d_{50} u'| T \]

\( f_{Bed} \)  multiplication factor for bed-load transport vector magnitude
\( \phi \)  relative availability of the sediment fraction in the top bed layer
The transport rates are corrected for longitudinal and transverse bed slope and sediment availability at the top of the bed. Bagnold-Ikeda (García, 2008) formulation is used for the former.

The transport of cohesive sediment in Delft3D is based on following parameters:

1. Settling velocity
2. Erosion rate
3. Consolidation/dry bed density

The cohesive sediment transport is simulated using the same depth-averaged advection-diffusion equation as 3-14 for non-cohesive sediment. However, the sediment flux (E-D) between bed and suspension for non-cohesive and cohesive sediments differs and is handled using different approach. Unlike non-cohesive transport, there doesn’t exist equilibrium concentration ($\bar{c}_{eq}$) due the supply-limited condition of cohesive sediment. These fluxes are instead controlled by hydrodynamic and sediment-bed properties.

The erosion rate (E) of cohesive sediment is modeled as linear function of excess shear stress, following Partheniades (1965):

$$E = M.S\left(\frac{\tau}{\tau_{cr,e}} - 1\right)$$

where,

- $M$ user-defined erosion parameter
- $S$ step function, becomes 0 if quantity inside parenthesis becomes $< 0$
- $\tau$ mean bed shear stress
- $\tau_{cr,e}$ critical shear stress for erosion
The erosion parameters are controlled by many factors, including bed density, water content, permeability, mineral composition, organic content, salinity, temperature and acidity. Consequently, considerable uncertainty exists in the determination of both $M$ and $\tau_{cr,e}$, even for beds which merely consists of cohesive sediment. Over timescales of days to weeks or months cohesive bed sediment will consolidate into a thinner sediment layer with a higher resistance against erosion. Consolidation is only implemented in special versions of Delft3D, which are very difficult to use (Deltasres, 2014). The common practice is to account for consolidation using appropriate values for the dry bed density, critical shear stress for erosion and static $M$ parameter. In contrast to consolidation, $\tau_{cr,e}$ can decrease due to liquefaction, swell or turbulent stresses. Based on measurements, the erosion coefficient is reported to vary up to three orders of magnitude.

For non-cohesive sediment, the critical shear stress for erosion (Chezy) decreases with grain size. However, within mud the opposite pattern is observed because of consolidation (Figure 3-7). The critical shear stress for erosion may vary between 0.05 and 0.5 Pa in case of freshly-deposited cohesive sediment to values exceeding unity, dependent on bed composition, bed density or clay mineral type.

![Critical shear stress vs. particle size](image)

Figure 3-7 Particle size-critical shear stress for erosion relationship. Figure from Deltasres (2014).
The deposition flux for cohesive sediment is given by Krone (1962):

\[ D = w_s \cdot c_b \cdot S \left( 1 - \frac{\tau}{\tau_{cr,d}} \right) \]

where,

- \( w_s \) settling velocity of sediment
- \( c_b \) sediment concentration in near-bottom layer
- \( \tau_{cr,d} \) critical shear stress for deposition
- \( S \) step function, becomes 0 if quantity inside parenthesis becomes < 0

Critical shear stress for deposition is the bed shear stress above which no deposition occurs. The concept of shear stress for deposition has been questioned, such that the effect of step function \( S \) in the deposition flux in equation (3-22) is ignored by taking a very large value of \( \tau_{cr,d} \) in Delft3D (Deltares, 2014). Also, in contrast to non-cohesive sediment, the settling velocity for cohesive sediment is difficult to determine because of potential for flocculation in saline water. The size of flocs can vary over four orders of magnitudes, as their settling velocities. The settling velocity depends on turbulent shear stress and sediment concentrations. Such a flocculation model is not part of standard Delft3D. Common practice is to define a range of settling velocities, representing well-flocculated, poorly flocculated and non-flocculated mud (Deltases, 2014). When modeling a river stretch or fresh-water system, the settling velocity can be set to depend on salinity. Following this simplification, constant settling velocity is assumed in Delft3D for saline and non-saline flows each, the former being larger in magnitude (Deltases, 2014).
3.2.6. Morphological model

Local variations in the bed load and suspended load transport rates result in bed level changes, which are computed using sediment conservation equation or Exner’s equation (Deltares, 2013):

\[
(1 - \lambda_p) \frac{\partial z_b}{\partial t} = -\frac{\partial S_{b,x}}{\partial x} - \frac{\partial S_{b,y}}{\partial y} - E + D \tag{3-23}
\]

where,

- \( \lambda_p \) bed porosity
- \( z_b \) bed elevation
- \( S_{b,x} \) and \( S_{b,y} \) components of the bed load transport in x and y directions
- \( D \) suspended sediment depositional flux, and
- \( E \) sediment entrainment flux

The elevation of the bed is dynamically updated at each computational time-step by calculating the change in mass of the bottom sediment resulting from the sediment fluxes. This change in mass is then translated into change in bed elevation as shown by equation (3-23).

The control volume for bed level change calculation is centered on the water level point. Since the bed load transport vectors are calculated at the water level points, upwind method is used to calculate the bed load transport vector components at the cell faces (\( S_{b,x} \) and \( S_{b,y} \)). For each active velocity point the upwind direction is determined by summing the bed load transport components at the water level points on either side of the velocity point and taking the upwind direction relative to the resulting net transport direction. The bed load transport component at the velocity point is then set equal to the component computed at the water level point immediately upwind (Figure 3-8).
A morphological acceleration factor (MorFac) is used to reduce computational time by simply multiplying the bed level change by this factor at each time-step. This factor is used since the time scales related to the morphological changes are several orders of magnitude larger than the time scales of the water motion (Elias, 2006). Though choosing large value of morphological factor reduces model run time, the upper limit of MorFac that is applicable to a given simulation is unavailable. Choice of morphological factor is thus based on good judgment through sensitivity testing and past studies and experience (Lesser et al., 2004; Roelvink, 2006). A morphological factor of 60 is used based on past studies of conceptual delta and mouth bar formation using Delft3D (Edmonds & Slingerland, 2007; Storms et al., 2007; Edmonds et al., 2009; Geleynse et al., 2010; Geleynse, 2013). These studies use morphological factor in the range 50 to 175. Edmonds and Slingerland (2007) found that the mouth bar formation is independent of
morphological factor below 200. Nevertheless, sensitivity analysis is done to check the appropriateness of the choice of value used in this study.

The morphology module of Delft3D currently implements two bed composition models: (1) a uniformly well-mixed bed; and (2) multi-layered bed. Uniformly mixed bed model is used in this study. A brief description of both models is given below.

**Uniformly mixed bed:**

The default bed model is the uniformly mixed bed. There is no bookkeeping layers or the order in which sediments are deposited. It simply consists of one layer of sediment, with the sediment fractions uniformly mixed. All sediments, because of the single layer, are directly available for erosion. The basement of the sediment layer is non-erodible. The top of the bed layer coincides with the bed level (Deltares, 2013).

Initial sediment layer thickness at the bed needs to be assigned for each sediment fraction. Spatially varying sediment thickness can be created using Delft3D’s QuickPlot tool. Initial mass of each sediment fraction can be computed by multiplying its thickness with user-defined dry bulk density (CDryB). The dry bulk density is constant in time and space for each sediment fraction. Similarly, the sediment layer thickness is calculated by dividing mass of each sediment fraction per unit area with its dry bulk density (Deltares, 2013).
**Multi-layered bed:**

This model uses a number of bookkeeping layers to keep track of sediment deposits. The bookkeeping layers are of three types: top transport layer, middle under-layers, and bottom base layer. Different initial distributions of sediments can be assigned to each bookkeeping layer. Only sediments in the topmost layer, the transport layer, are available for erosion. The top layer is replenished after erosion with sediment from the layer beneath it. During deposition, sediments are added to the transport layer and homogeneously mixed together. The transport layer has an assigned maximum thickness and therefore pushes the sediments to underneath layer once that threshold thickness is reached (Deltares, 2013). The number of under-layers depend on the thickness of sediment. The maximum thickness of under-layers and the maximum number of these layers are also user-defined. When the number of under-layers reaches the maximum value, the bottommost under-layer merges with the uniformly mixed base layer (Jagers, 2012b).

3.3. Image processing

The purpose of this thesis is also to analyze gross properties of delta and channel networks. At the heart of this lies the problem of delineating delta from the surrounding open water. Delta is combination of dry lands and channel networks, and the shoreline is the land-water interface that separates delta from the open water. Though it is easy to recognize shoreline visually, it is not so easy to quantify it (Shaw et al., 2008). A common definition of shore is the elevation contour of
land above mean water level at land-water interface (Shaw et al., 2008). The problem with this definition occurs when the land-water interface is complex geometrically containing numerous fragmented lands and sheltered lakes that are not directly exposed to open water. Shaw et al. (2008) have used advanced method called Opening Angle Method (OAM) that uses visibility criteria to map locus of shoreline consistently. In this thesis, shore is defined according to the common definition, i.e. elevation contour of land at or above mean basin water level at land-water interface. Much simpler algorithm compared to Shaw et al. (2008) method, using Matlab image processing tool and Adobe Photoshop, is adopted. The following section describes the methodologies to extract delta shoreline, delta surface and channel network properties.

A. Extraction of delta boundary

1. Bed level elevation in receiving basin is converted to a binary image using the threshold elevation of 0 m, which is the mean sea level or the mean water level in the basin. Area with elevation above the threshold value denoted deltaic land that is above the mean water level, white region in Figure 3-11.

![Figure 3-11 Binary image of delta bed elevation obtained using mean water level in basin as the threshold value](image)

2. Following Geleynse (2013) methodology, sequential rows and columns operations are performed on the binary image obtained from step 1 to isolate land-water boundary pixels. The
result of this operation is shown in Figure 3-12. A binary image contains pixels with value 0 (black) and 1 (white) arranged in rows and columns. At first, row operation is conducted. Along each row of the image, the two outermost white pixels are identified. After finishing row operation, along each column the white pixel lying furthest into the receiving basin is identified.

![Figure 3-12 Boundary pixels obtained by row and column operations](image)

3. Matlab’s morphological tool is used to remove isolated pixels lying within the delta. In order to obtain a continuous shoreline enclosing the delta, the boundary pixels are thickened and then thinned. The boundary pixels are thickened using “disk structuring element” matrix in Matlab. The dimension of the matrix varies from 6 by 6 to 10 by 10. Thickening allows connecting prospective shoreline pixels which were otherwise disconnected.

![Figure 3-13 Thickened continuous boundary pixels](image)
4. Thinning algorithm developed by Lam et al. (1992) is used to thin the resulting pixels (Geleynse, 2013). Morphological command using this algorithm is available in Matlab [e.g., bwmorph (binary image matrix, ‘thin’,Inf)]. The thinned shoreline is expected to be a line that is 1 pixel wide. The output from this step is shown in Figure 3-14.

![Figure 3-14 Shoreline pixel resulting from thinning](image)

5. As can be seen in Figure 3-14, the resulting shoreline is not 100% satisfactory. It contains broken shoreline, tentacles resulting from thickening and thinning processes and few isolated pixels. The pixels resulting from step 4 are laid upon pixels representing boundary from step 2. This helps cleaning or closing of pixels in Adobe Photoshop.

![Figure 3-15 Overlay of thinned pixels (blue) over the boundary pixels (white)](image)
**B. Extraction of channel network**

The extracted delta surface, which is a binary image, acts like a mask to extract channel features. Channel networks within a delta can be defined based on minimum thresholds for velocity, water depth or both (Geleynse et al., 2010). If \( d_1 \) and \( v_1 \) are threshold values for depth and velocity, the extracted channel network will have depth \( \geq d_1 \) and velocity \( \geq v_1 \).

The channel network is isolated based on the threshold depth of 0.1 m and a threshold velocity such that the extracted channels captures flow with 10 % error. The captured flow in the channels is measured at three locations. These locations are measured with respect to average delta radius or radius of equivalent semi-circular delta. Figure 3-18 shows these three measurement locations which are at radial distance 25, 50 and 75 % of average delta radius respectively.
Three locations in delta where discharge flux through channel network are measured. These locations are circumference of semicircle with radii equal to 25, 50 and 75 % of delta’s radius.

For each location, a binary image is created with semicircle having radius equal to the radial distance of the location. These images are called 25%, 50% and 75% henceforth. In the image the pixel representing circumference of the semi-circle is assigned value 1 while the rest are assigned 0. Bitwise intersection of the channel (Figure 3-17) with respective (25%, 50% and 75%) binary images were performed. The resulting pixels represented the component of channel’s cross-sectional length at each location (e.g., Figure 3-19).

A pixel in the binary image represents a single grid cell in 2D model’s horizontal grid, which is 50 m by 50 m. Each pixel therefore has (depth-averaged) velocity and depth values, defined at the center of the pixel. The estimated (one-dimensional) width of each channel x-section pixel is
50 m. The total discharge flux at each location is determined by multiplying the three quantities (velocity, depth and width of pixel) and summing up the values. Maximum threshold velocity was selected which gave channel network that captured 90% flow at all the three locations. Since the pixels obtained by bitwise operation aren’t perfectly perpendicular to the channel cross-section or velocity vectors, the calculated discharge flux inherently contains positive or negative errors.

C. Network properties

For this study, the channel network properties are defined as follows:

1. Channel flow area ($A_c$)

   Method for channel network extraction is discussed in the previous section. In the binary image of channel network, a white pixel represented plan area of channel. The plan area of channel network was calculated by multiplying the number of channelized pixels with the area of single pixel (50x50 m$^2$).

2. Distributary length ($L_c$)

   Distributary length is defined as the sum of channel lengths measured along the centerline of distributary channels. The binary image of channel network is thinned to the extent that distributary channel is only 1 pixel wide. The length of each pixel is 50 m. The total length of the channels is obtained by summing up the distributary channel pixels.

3. Average channel width ($W_c$)

   The mean channel width is defined as the average width of distributary channels in the delta. Assuming the distributary channels have rectangular cross-section, the average channel width is the ratio of flow area ($A_c$) to distributary length ($L_c$).
3.4. Parallel computation

One of the objectives of this thesis is to assess the performance of 2D model running in multiprocessors and to compare it against 1D model run time. Louisiana Optical Network Initiative (LONI) and Louisiana State Universities’ High Performance Computing (HPC) computer resources are used for 2D model simulations. The final results are based on SuperMike-II HPC system.

SuperMike-II is named after LSU’s original large Linux Cluster named SuperMike that was launched in 2002. It is a 146 Tflops Peak Performance 440 compute node cluster running Red Hat Enterprise Linux 6 operating system. Each node contains 8-Core Sandy Bridge Xeon 64-bit processors operating at a core frequency of 2.6 GHz. Fifty of the compute nodes also have two NVIDIA M2090 GPUs that provide an additional 66 Tflops total Peak performance. SuperMike-II is open for general use to LSU users (LSU HPC, 2014).

In parallel computation, the Message Passing Interface (MPI) algorithm automatically subdivides the 2D model’s domain into parallel strips along the longest grid direction which is downriver. The number of partition is equal to the number of cores used in the computation. Partitioning is done in such a way that each parallel strip has nearly the same number of active cells. It produces long thin parallel strips for the study domain which is less efficient compared to square block partitioning. Figure 3-20 shows the parallel portioning of model domain while using 32 processors in HPC system. The simulations in this thesis are done using 32 cores. Computation time for different number of cores is also analyzed in this thesis.
3.5. Model setup

3.5.1. Domain and boundary

Model domain of 1D and 2D methods is shown in Figure 3-21. The domain characteristics (length and width of diversion channel, channel and receiving basin bathymetry) are based on Kim et al. (2009b) numerical model for Barataria Bay and Breton Sound diversions. The diversion channel (light blue, West) is 5000 m long and 500 m wide. At the Western edge of the channel, boundary conditions for flow, sediment concentration and (equilibrium) bottom slope are prescribed for the 1D model while only the first two are prescribed for the 2D model. To the East of diversion channel is open water, called receiving basin after here, where land is built by the operation of diversion.
Figure 3-21 Model domain of the 1D and 2D models. The units of axes are meters.

Receiving basin is gray and dark blue region in Figure 3-21. The basin has constant bottom slope of $2 \times 10^{-5}$ or 0.02 m/Km going from West to East in the figure. Mean sea water level is assumed on the receiving basin (Kim et al., 2009b). In the 1D model, the size of the basin is infinite in $n$ and $m$ directions (axes), i.e., this model doesn’t require prescription of basin dimensions. The 2D model setup however requires finite size of the basin for it to be computationally efficient and viable. Offshore open boundary conditions are therefore required to have a basin that is smaller than the actual size of basin. Mean sea level is imposed on the basin by prescribing constant water level of 0 m at the three offshore open boundaries (North, East and South). The open boundaries should be far away from the area of interest so that boundary effects and errors are not introduced in the numerical solution. Considering this requirement, the size of the basin is 15 km by 20 km.
3.5.2. Differences in the receiving basins

There is a small difference between the receiving basins for the two models. For the 1D model, it is necessary to have an initial delta in the basin at the start of the simulation. 1000 m of initial delta length is selected here. The depth at the toe (intersection of delta foreset and basin’s bottom) of the initial delta is 0.8 m. This depth includes the depth or space created by one time settling of the squeezy layer (~ 0.7 m). The prodelta is assumed to have a layer of high porosity mud (or the squeezy layer), which is rapidly compacted on a one time basis as the delta foreset progrades over it (Kim et al., 2009b). It is assumed that the squeezy layer settles by 0.7 m uniformly over the basin whether or not it consists of delta on top of it. For the 2D model, the basin starts at the toe of 1D model’s initial delta which has water depth of 0.8 m. As can be seen in Figure 3-21, the grey region, whose width is equal to initial delta length, is the additional receiving basin the 1D model requires compared to the 2D model. Thus the basin for the 1D model starts at shallower depth (< 0.8 m). In other words, in Figure 3-21, the diversion channel for the 1D model ends at m = 5000 m and for the 2D model at m = 6000 m.

The 1D model uses Engelund and Hansen (1967) sediment transport formula. In the 2D model, there are several sediment transport equations we can choose from. Using the same transport formulation in both models looked to be certain, but due to a bug in the Delft3D software, it wouldn’t allow the use of Engelund and Hansen (1967) formula in parallel processor mode. Personal contact with Deltare’s researchers revealed that the bug wouldn’t be fixed soon. Therefore, Van Rijn (1993) formulation was used for the 2D model simulations.

3.5.3. Definition of delta area and radial distance

Natural deltas contain land as well as channel networks and sheltered water bodies like estuaries and lakes. As the 2D model is able to simulate these natural physical characteristics,
special care is needed to distinguish delta from the open water. The intersection of delta and open water is defined by shoreline. Methods to extract shoreline is provided in section 3.3. The area bounded by shoreline gives the simulated delta area for the 2D model.

Because of the difference in the receiving basin and/or initial delta, it is not possible to directly compare magnitude of surface area and radial extent from the two models. At time “0”, 1D model has finite values for delta area and radius of shoreline but they are zero for 2D model. It is desirable in the 1D model to have zero delta area as well at the beginning since sediment hasn’t yet been introduced into the system at that point in time. Therefore, in the 1D model, initial delta area is subtracted from the simulated delta area in subsequent times to obtain net delta area. This is justified given that initial area of the delta is very small for the given initial delta. Though it appears intuitive to calculate net radius of delta similarly, such reduction is not recommended as area of delta holds square relation with the radial distance and the initial value is quite significant. For this study, the radial distance is untouched and presented as it is. Finally, the vertex of 1D model delta is translated by 1000 m so that both models have common point of reference for radial distance measurement. This is very important while plotting cross-section plot (see Figure 4-11 in Chapter 4). Due to this translation, the toe of the 1D delta doesn’t coincide with the bottom of the basin in Figure 4-11.

Delta simulated by the 2D model apparently isn’t radially symmetrical. For comparison with the 1D model, the average radius of delta simulated by the 2D model is defined as the radius of the semicircle which has same area (A) as the delta. This radial distance is calculated as follows:

$$\sqrt{\frac{2A}{\pi}}$$
3.5.4. Grid and time step

The 1D model uses staggered grid system to discretize elevation (or bed slope) and sediment discharge variables in the diversion and virtual channels. The diversion channel has “N” grid points while the topset has “M” grid points. A grid point comprises a pair of staggered grids. N and M are both 5 in this study. The number of grid points in channel and topset doesn’t have much effect on the result. While the grids on the diversion channel are fixed in horizontal plane, the grids on topset move small distance horizontally at each time step, and thus called “moving staggered grid.” Time step of 0.036525 days is used.

![Moving Staggered Grid](image)

Figure 3-22 Numerical grid for the 1D model

The 2D model grid measures 20 km by 20 km. The model domain is quite simple with straight fluvial channel and rectangular receiving basin. The model domain has grid resolution of 50 m by 50 m. Delft3D-RGFGRID package allows the creation of the grid. This grid resolution is small enough to properly resolve the width of diversion channel. The representation of the diversion channel and distributaries can be inaccurate if their width approaches the grid cell resolution of 50 m (Hillen, 2009). This resolution may not represent the distributaries very accurately, but still good
enough to represent channel switching processes. A time step of 15 second was chosen based on courant’s stability and accuracy condition.

3.5.5. *Bed layer and sediment composition*

Delta is a dominantly depositional system, with erosional process occurring less compared to depositional. The 1D model doesn’t allow incision of bed below the initial topography of the domain. Therefore, the 2D model was set up with initial bed without sediment. The entire spatial domain is characterized by a 0 m thick sediment layer. The initial bottom level of spatial domain represents non-erodible layer. Anytime non-erodible basement is reached, erosional flux is reduced in the 2D model. The diversion channel and basinal beds are modeled following uniformly mixed or single layer bed concept described in section 3.2.6.

3.5.6. *Parameters*

In sections 3.1 and 3.2, transport equations and related numerical parameters in the 1D and 2D models respectively are discussed. The numerical parameter values used in these models are summarized in Table 3-1.

Table 3-1 Numerical parameters for the 1D and 2D models

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Symbol</th>
<th>1D</th>
<th>2D</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chezy’s Constant</td>
<td>$C_z$</td>
<td>62.46</td>
<td>62.46</td>
<td>m$^{1/2}$s$^{-1}$</td>
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<td>Horizontal eddy viscosity coefficient</td>
<td>$\nu_H$</td>
<td>-</td>
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<td>m$^2$s$^{-1}$</td>
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<tr>
<td>Horizontal eddy diffusivity coefficient</td>
<td>$D_H$</td>
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<td>m$^2$s$^{-1}$</td>
</tr>
<tr>
<td>Sand diameter</td>
<td>$d_{50}$</td>
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<td>0.21</td>
<td>mm</td>
</tr>
<tr>
<td>Sediment specific gravity</td>
<td>$G$</td>
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<td>2650</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>Non-cohesive sediment dry bulk density</td>
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<td>1600</td>
<td>1600</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>Cohesive sediment fall velocity</td>
<td>-</td>
<td>-</td>
<td>0.0015</td>
<td>mm s$^{-1}$</td>
</tr>
<tr>
<td><strong>Table 3-1 continued...</strong></td>
<td><strong>Symbol</strong></td>
<td><strong>1D</strong></td>
<td><strong>2D</strong></td>
<td><strong>Unit</strong></td>
</tr>
<tr>
<td>---------------------------</td>
<td>------------</td>
<td>--------</td>
<td>--------</td>
<td>---------</td>
</tr>
<tr>
<td>Cohesive sediment dry bulk density</td>
<td>-</td>
<td>1600</td>
<td>1600</td>
<td>kg m⁻³</td>
</tr>
<tr>
<td>Erodibility coefficient of cohesive sediment</td>
<td>M</td>
<td>-</td>
<td>0.0001</td>
<td>kg m⁻³ s⁻¹</td>
</tr>
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<td>Critical shear stress for erosion</td>
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<td>Pa</td>
</tr>
<tr>
<td>Critical shear stress for deposition</td>
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<td>1000</td>
<td>Pa</td>
</tr>
<tr>
<td>Initial bed thickness</td>
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<td>m</td>
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<td>Morphological factor</td>
<td>MorFac</td>
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<td>Minimum flow depth for sediment transport computation</td>
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<td>-</td>
<td>0.1</td>
<td>m</td>
</tr>
<tr>
<td>Threshold sediment thickness for reduction of transport and erosion</td>
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<td>-</td>
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<td>m</td>
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<td>Maximum fraction of erosion assigned to adjacent dry cell</td>
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<td>Smoothing period</td>
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<td>-</td>
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<tr>
<td>Spin-up period</td>
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<td>-</td>
<td>60</td>
<td>minute</td>
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<tr>
<td>Channel forming Shields number</td>
<td>$\tau^*$</td>
<td>1.86</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Wash load deposited per unit sand</td>
<td>-</td>
<td>0 to 1</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Sediment transport formula</td>
<td>-</td>
<td>Engelund and Hanson (1967)</td>
<td>Van Rijn 1993</td>
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Table 3-2 Summary of physical properties for the two models

<table>
<thead>
<tr>
<th>Model Parameter</th>
<th>Symbol</th>
<th>1D</th>
<th>2D</th>
<th>Unit</th>
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<tbody>
<tr>
<td>One-time settling thickness into prodelta</td>
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<td>0.7</td>
<td>m</td>
</tr>
<tr>
<td>Initial length of delta from vertex</td>
<td>-</td>
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<td>0</td>
<td>m</td>
</tr>
<tr>
<td>Initial slope of diversion channel</td>
<td>-</td>
<td>1.5x10^-5</td>
<td>1.5x10^-5</td>
<td>-</td>
</tr>
<tr>
<td>Subaqueous basement slope</td>
<td>S_b</td>
<td>2x10^-4</td>
<td>2x10^-4</td>
<td>-</td>
</tr>
<tr>
<td>Slope of foreset slope</td>
<td>S_f</td>
<td>0.005</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Delta spreading angle</td>
<td>0</td>
<td>180</td>
<td>-</td>
<td>degree</td>
</tr>
<tr>
<td>Width of diversion channel</td>
<td>B</td>
<td>500</td>
<td>500</td>
<td>m</td>
</tr>
<tr>
<td>Length of diversion channel</td>
<td>-</td>
<td>5000</td>
<td>5000</td>
<td>m</td>
</tr>
</tbody>
</table>

Chezy’s bed roughness parameter ($C_z$), sediment size, one time settling thickness of the basin, slopes of basin and channel, etc., are taken from Kim et al. (2009b) model. Boundary friction coefficient $C_f$ given in section 3.1.2 is related to Chezy’s constant as $C_z = \sqrt{g/C_f^2}$. The chosen value of roughness constant is an estimate for large, low-slope sand-bed streams (Parker et al., 2008a). The sediment size in the lower Mississippi near the diversions is from Nittrouer et al. (2008). The prodelta is assumed to have high-porosity mud (porosity 0.78) which could settle by 0.65 m in one time basis as the delta foreset progrades (Morton et al., 2005; Kim et al., 2009b). This value is rounded to one decimal place (i.e. 0.7 m) while creating the bathymetry for this study.

The parameters used in the 2D model though falls in the range commonly used in morphodynamic studies (Hillen, 2009; Geleynse et al., 2010; Geleynse, 2013) may not necessarily be the most suitable values. Parameters like horizontal eddy viscosity and diffusivity are purely numerical terms necessitated from inability of coarse grid size to resolve eddies smaller than the grid resolution. These parameters need to be calibrated.
3.5.7. Sediment inflow cases

Sediment (sand) rating curve for Mississippi river at Belle Chasse is shown in Figure 3-23. Suspended sediment in this reach occurs for flow above 10,000 m$^3$/s. So diversion operating for lower discharge may not contribute significant sand for land building. For land building purpose, it is best to divert as much sediment as possible to maximize the land building. Studies have shown that the efficiently working diversion should have at least the sediment water (SWR) ratio of 1 to avoid aggradation of sediment in the main channel that could possibly harm navigation. Sediment water ratio is defined as the ratio of sediment concentration in the diversion channel to the concentration in the river upstream of the diversion. SWR greater than 1 means that the diversion can carry more sediment per unit discharge than the main river. It is in the best interest of diversion project to have as high as possible SWR but its viable upper limit is still an ongoing research.

In this thesis, we create medium sized diversions running continuously yearlong with discharge of 2000 m$^3$/s but with four scenario of input sediments as summarized in Table 3-3. The first three have varying concentrations (parts per million or ppm) of non-cohesive (sand) sediment while the fourth has 50-50% mix of cohesive (mud) and non-cohesive sediments. The objective of having these scenarios is to observe what effect sediment concentration and sediment fractions have on land building and channel characteristics. The first two scenario represent the sand concentrations in the river when Mississippi discharge is 19,000 and 24,000 m$^3$/s respectively.

The first three scenario which transport only sand, are assigned bed porosity of 0.39. The representative porosity of purely mud bed or combination of mud and sand bed in the fourth scenario is not so easy to estimate. Porosity can be very high for freshly deposited mud. The mud deposit consolidates thereby shrinking in volume with time. Kim et al. (2009b) assume bed porosity of 0.6 based on bay mud and sand bar deposits (Kuecher, 1994; Meckel et al., 2006, 2007).
Figure 3-23 Mississippi River sediment rating curve at Belle Chasse drawn using best-fit curve given by Nittrouer et al. (2011); dashed lines show the river discharge corresponding to case 40 and case 80 sand concentrations in the diversion channel.

Table 3-3 Input sediment scenarios for the simulations

<table>
<thead>
<tr>
<th>Case</th>
<th>Sand</th>
<th>Mud</th>
<th>2D Bed Porosity</th>
<th>1D Bed Porosity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ppm</td>
<td>ppm</td>
<td>Sand Mud Gross</td>
<td>Gross</td>
</tr>
<tr>
<td>40</td>
<td>40</td>
<td>-</td>
<td>0.39 - 0.39</td>
<td>0.39</td>
</tr>
<tr>
<td>80</td>
<td>80</td>
<td>-</td>
<td>0.39 - 0.39</td>
<td>0.39</td>
</tr>
<tr>
<td>120</td>
<td>120</td>
<td>-</td>
<td>0.39 - 0.39</td>
<td>0.39</td>
</tr>
<tr>
<td>40-40</td>
<td>40</td>
<td>40</td>
<td>0.39 0.39</td>
<td>0.39</td>
</tr>
</tbody>
</table>

Delft3D (the 2D model) treats sand and mud differently; individual dry bulk densities (or porosities) are assigned for the two fractions. Therefore, the net porosity of bed can vary spatially depending on deposited sand and mud fractions. In this study, we like to make comparison among the scenarios and with the 1D model. For that it is necessary to have common combined bed porosity. This condition requires to have same porosity assigned for sand and mud in the fourth
scenario. The net porosity of bed is maintained at 0.39 in all the cases. In the final scenario, sand and mud both are assigned porosity value of 0.39.

3.5.8. Sensitivity analysis

Sensitivity analysis is the study of uncertainty in output of a mathematical model due to different sources of uncertainty in its inputs (Wagner, 1995). This analysis illustrates the robustness and applicability of the model. Case 80 and 40-40 are used as the reference cases as they represent two unique models, one only contains sand, the other contains sand and clay both. For this analysis, only one parameter is varied while other remained constant. The parameters that are used in the sensitivity analysis of the 2D model are shown Table 3-4.

Table 3-4 Parameters for sensitivity analysis

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Symbol</th>
<th>Values</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Morphological factor</td>
<td>MorFac</td>
<td>60, 30</td>
<td>-</td>
</tr>
<tr>
<td>Horizontal eddy viscosity coeff</td>
<td>ν_H</td>
<td>10, 5</td>
<td>m²/s</td>
</tr>
</tbody>
</table>
Chapter 4. RESULTS

Two modeling approaches, a simplified 1D and a process-based 2D, were used to simulate land building in a simplistic receiving basin under four hypothetical sediment diversion scenarios. The objective of this exercise was to directly compare the delta radius and area given by the two models and between the different scenarios. The second objective was to evaluate the morphological features produced by the 2D simulations. Three of the scenarios were based on varying concentrations of non-cohesive sediment: 40 ppm (case 40), 80 ppm (case 80) and 120 ppm (case 120), while the fourth scenario utilized a sediment concentration that was a 50/50 mix of non-cohesive (40 ppm) and cohesive (40 ppm) (case 40-40) sediments. In this chapter, the results from these simulations will be presented and discussed. It is important to note that all of these simulations were based on a continuous inflow of water and sediment into the receiving basin, an operation and management strategy that: (1) may not be possible given the characteristics of the Mississippi River hydraulics and sediment transport; and (2) is not sound from a policy standpoint. Therefore, the times shown in the results are for comparison between simulations and not indicative of what might be expected from future sediment diversions.

For the sake of convenience and clarity, the results are presented in a particular sequence. First, the results for case 40 and case 120, which have the extreme values of non-cohesive sediment, are presented. Those are followed by comparison between case 80 and case 40-40, which have same sediment inflow but vary in cohesive sediment content. Finally, all the cases are compared together.
4.1. Case 40 and 120

Figure 4-1 shows the surface elevation and channelized area of the deltas formed by the 2D model’s case 40 and case 120 during the 40 years of simulation, including shoreline location given by the 1D model’s corresponding cases. It should be noted that land elevations shown in Figure 4-1 are plan area within the shoreline defined as the elevation contour of land-water interface that is at or above mean basin water level at land-water interface (method of shore extraction in section 3.3). The sediment input in case 40 is too small to create any land in 10 years. Only a single mouth bar and two side levees are observable at year 10. First plot in Figure 4-1 (top left) shows the mouth bars and levees and the surrounding bathymetry for case 40 at year 10.

Figure 4-1 Delta surface elevation evolution for case 40 (left) and case 120 (right) at year 10, 20, 30 and 40. The blue dashed line shows the radial extent of the delta from the 1D model for each scenario and time. The network shown in the upper part of each plot shows the plan of the channelized area and its centerline. Each plot is provided with unique color bar for optimum view of elevation.
Figure 4-1 continued.
In both cases, the 1D model produces delta (the blue dashed lines) that, except for a few localized regions, extends farther out into the receiving basin than the 2D model. Both cases in the 2D model show fan-like shape but at larger time, case 120 shows rougher shoreline, with land growth taking place along the major distributary. Though the distributaries induce localized shoreline roughness, at larger time scales the deltas are expected to be relatively uniform radially due to multiple channel switching, thereby supporting 1D model’s assumption of radially symmetrical delta front progradation (Geleynse, 2013).

It is interesting to observe the similarity in the delta size and morphology between case 40 at year 30 and case 120 at year 10 (Figure 4-1). These deltas appear to have geometric but time-lagged similarities. This is not totally unexpected given that same amount of sediment is fed in to the basin for the two deltas over the 30 (case 40) and 10 year (case 120) time periods.

4.1.1. Radial extent and delta area

Figure 4-2 and Figure 4-3 show the plots for radius and area of deltas for the two cases. As a reminder, case 40 was simulated over 100 years while case 120 for only 40 years. As explained in section 3.5.3, the radius of delta for the 2D model is defined as the radius of semicircle which has same area as the given delta. The radii of the 2D model deltas in the two cases are always smaller than the radii of 1D model deltas. Also the rate of increase of delta radial extents are smaller for the 2D model results. Similar trend is observed for delta area as well (Figure 4-3). The area of the 2D model deltas are smaller than the areas of the 1D model deltas. In both the cases, the ratio of the 2D to 1D (Figure 4-3) area is almost constant with time (~70%).
Figure 4-2 Radial extent of delta simulated by the 2D and 1D models during 100 year period for case 40 (left) and during 40 year period for case 120 (right). Smaller inflow of sediment in case 40 allowed longer simulation time without increasing the size of the 2D model domain.

Figure 4-3 Delta area simulated by the 2D ($A_{2d}$) and 1D ($A_{1d}$) models and their ratio during the simulation period. Case 40 (left) and case 120 (right).

4.1.2. Channel characteristics

Figure 4-4 shows the threshold velocities that were used to extract the distributary networks through the delta. By the definition of threshold velocity given in section 3.3, the minimum velocity in the extracted channels is equal to the threshold value and that these channels carry at least 90% of total flow. The velocities in case 120 are slightly higher than those in case 40. While
case 40 has fairly constant time-velocity relationship, case 120 has monotonically increasing relationship.

![Figure 4-4 Threshold velocity. Case 40 (left) and case 120 (right).](image1)

The channel area is defined as the plan area of the distributary channels over the delta surface. Figure 4-5 shows both the channel area and percentage cover on the delta. Case 120 shows steeper decrease in channel area in time compared to case 40.

![Figure 4-5 Channelized area (A_c)-gray line. Case 40 (left) and case 120 (right). Black line is the ratio of A_c to A_{2d} (2D delta area).](image2)

Figure 4-6 shows total length of distributary channels. Case 40 exhibits increase in total channel length. Comparing with Figure 4-1, this increase appears to be because of increase in
number of distributary channels rather than elongation of existing channels. On the other hand, it is hard to draw definite conclusion from plot of case 120 which shows initial increase and then decrease in channel length.

Figure 4-6 Channel length ($L_c$). Case 40 (left) and case 120 (right).

Figure 4-7 shows plots for average width of distributary channels in case 40 and 120. In both cases, the channel width appears to attain a steady value with time. The steady state value is almost same for both cases (~150 m). Case 120 appears to reach this steady width much earlier.

Figure 4-7 Average channel width ($W_c$)-gray line. Black line is the ratio of $W_c$ to $B$ (width of the diversion channel, i.e., 500 m). Case 40 (left) and Case 120 (right)
4.2. Case 80 and 40-40

Figure 4-8 shows the surface elevation and channelized area of the deltas formed by the 2D model’s case 80 (non-cohesive) and case 40-40 (mix-sediment) during the 40 years of simulation, including shoreline location given by the 1D model’s corresponding cases.

Figure 4-8 Delta surface elevation evolution for case 80 (left) and case 40-40 (right) at 10, 20, 30 and 40 years. The blue dashed line shows the radial extent of the delta from the 1D model for each case and time. The network shown in the upper left part of each plot shows the plan of the channelized area and its centerline.
In all cases, the 1D model produces delta that, except for a few localized regions, extends farther out into the receiving basin. The Case 80 delta exhibits a fan-like shape in sharp contrast to case 40-40, which is very asymmetric and has a rough shoreline. The radial symmetry in case 80 can be attributed to the even distribution of channel distributaries in all directions. In non-cohesive deltas, the mouth bar and levees formed by the channels are weak and easily eroded. As a result, water and sediment is fed to the whole delta through numerous breaches in mouth bars (Edmonds & Slingerland, 2009).
From these plots, it can be observed that the individual distributary channels in case 80 are relatively short and straight and that the widths are, on the whole, much larger than those in case 40-40. In addition, the channel widths decrease with each bifurcation. This is in contrast to the case 40-40 distributary channels that are long and weakly sinuous and have widths that remain fairly constant. In case 40-40, channel switching phenomenon is observed from growth of land predominantly to the north at year 10 to growth of land toward the east at year 40. This phenomenon of delta growth by avulsion is consistent with historic growth of Mississippi River delta (Blum & Roberts, 2012) and the observations from cohesive numerical and experimental deltas (Edmonds & Slingerland, 2009; Hoyal & Sheets, 2009; Seybold et al., 2009). Finally, these plots qualitatively highlight the ability of the 2D model to form the distributary channels that are important to delta evolution.

4.2.1. Radial extent and delta area

It was seen from the plots in Figure 4-8 that the radial extent of the 1D delta is larger than that simulated by the 2D model. Figure 4-9 shows that the delta radius, obtained directly from the 1D model, is always greater than the radius calculated from the 2D model results (described in Section 3.5.3). As expected, the rate at which the radii increase slows with time as the delta forms in the deeper basin. What is important to point out is that the radial growth rate of case 40-40, predicted by the 2D model, is different than for the other simulations and results in a slight divergence in the results at later times. This is expected for two reasons: (1) the 1D model only simulates the non-cohesive sediment transport process and adds one unit of cohesive material for every unit of non-cohesive material deposited, thus making the 1D case 80 and 40-40 results nearly identical; and (2) the 2D model simulates the transport behavior of both types of sediment resulting in an uneven spatial distribution of sediment.
Figure 4-9 Radial extent of delta simulated by 1D and 2D models during 40 year period. Case 80 (left). Case 40-40 (right).

Figure 4-10 shows how the case 80 and 40-40 delta areas grow with time along with the ratio of the 2D to 1D delta area. While the ratio of the areas remains relatively constant with time for both cases it can be seen that the ratio for case 80 is higher than for case 40-40 (~70% for case 80 versus ~55% for case 40-40). As will be explained in the following section, part of this difference is due to significant topographic differences between delta composed entirely of non-cohesive sediments and delta composed of cohesive material.

Figure 4-10 Delta area simulated by 2D ($A_{2D}$) and 1D ($A_{1D}$) models and their ratio during the simulation period. Case 80 (left). Case 40-40 (right).
**4.2.2. Delta elevation**

A transect was cut in the basin along the axis of the diversion channel to compare the bed profiles produced by the various simulations at year 40 (Figure 4-11). In general, bed elevations are highest closest to the diversion channel and, as expected due to the simplified approach of the 1D model, the delta topset linearly decreases with distance into the receiving basin. Note, however, the significant variation in the 2D model bed elevations and the extreme variability in the case 40-40. For case 80, the bed elevation rises to nearly 1.5 m just after the diversion channel and then decreases to mean basin water level at a distance nearly 7 km into the receiving basin. The bed elevations in case 40-40 also have a peak just after the diversion channel and a general decrease with distance; however the elevations are significantly higher than in case 80 (and 1D) and show the impact of the significant channeling that occurs and is maintained due to the presence of cohesive material. What these plots highlight is the ability of the 2D model to simulate the spatial variability in the bed elevations and significant impact that cohesive material has on the geomorphology.

![Figure 4-11 Delta bed profile along the diversion channel axis for simulation result of 40 years. Case 80 (left) and Case 40-40 (right). Blue line is 1D delta, red is 2D delta. 1D delta was translated in to the basin by initial length of 1D delta for comparison.](image-url)
10, 50 and 90 percentile elevations were calculated for the 2D model results at year 40 (Table 4-1). A percentile value gives the elevation in meters below which certain percentage of delta elevation lies. The table quantitatively shows that the median elevation (50 percentile) for case 40-40 is higher than for case 80 (by nearly 0.35 m) and that, while the lowest elevations are not significantly different, the highest elevations are significantly higher. The increased aggradation in case 40-40 can be attributed to cohesive nature of the deposits. Mouth bars formed by cohesive sediment are stronger and less easily eroded compared to those formed by less cohesive sediment or by purely sand (Edmonds & Slingerland, 2009). Thus, the smaller delta area for case 40-40 is not unexpected.

Table 4-1 Percentile distribution of bed elevation for the result of 40 year simulation. Percentile value gives the elevation in meters below which certain percentage of delta elevation lies.

<table>
<thead>
<tr>
<th>Case</th>
<th>10 percentile (m)</th>
<th>50 percentile (m)</th>
<th>90 percentile (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>80</td>
<td>-0.67</td>
<td>0.259</td>
<td>0.87</td>
</tr>
<tr>
<td>40-40</td>
<td>-0.75</td>
<td>0.601</td>
<td>2.06</td>
</tr>
</tbody>
</table>

4.2.3. Channel Characteristics

Figure 4-12 shows the threshold velocities that were used to extract the distributary networks through the delta. As expected, the threshold velocities were highest during early times when there were fewer channels available to convey the flow leaving the diversion channel. Comparison of the velocities shows that the velocities in the 40-40 case are always higher than those in the 80 case by at least 50%. Subsequent analysis will quantitatively relate this to the size of the channels.
Figure 4-12 Threshold velocity. Case 80 (left) and case 40-40 (right)

Figure 4-13 shows both the channel area and the percent of the delta area covered by the channels. The case 80 distributary channels have between 50 and 100% larger area than those in case 40-40 and make up a higher percentage of the total delta area.

The total length of distributary channels in both cases show an increase with time (Figure 4-14). The channel lengths are slightly larger for case 80 than for 40-40.

The average channel width is calculated by dividing the total channel area by the total channel length (Figure 4-15). The average channel width of case 80 channels are around 50% larger than case 40 channels. Average channel width is always less than the width of the diversion channel (500 m).
Figure 4-13 Channelized area ($A_c$) - gray line. Case 80 (left) and case 40-40 (right). Black line is ratio of $A_c$ to $A_{2d}$ (2D delta area).

Figure 4-14 Channel Length ($L_c$). Case 80 (left) and case 40-40 (right)
4.3. Comparison of all cases

In the above sections, a comparison was done between cases 40 and 120 and then 80 and 40-40. In this section, comparisons will be across all cases for both 1D and 2D models and the concept of a mass-normalized time will be introduced. The mass-normalized time is defined by multiplying the time by the input sediment concentration and is a way of directly comparing simulation results at times at which the same amount of sediment mass has been introduced into the system.

4.3.1. Radial extent and delta area

Figure 4-16 and Figure 4-17 show the radial extent and area of delta, respectively for all four cases as a function of time and mass-normalized time. Consistent with the results presented previously, for each case, the delta radius and area predicted by the 1D model are always larger than those from the 2D model and, as expected, the delta size is directly related to the sediment concentration, except for the 2D case 40-40. Note that, during the first 40 years of simulation, the radial distance and area for the 2D case 40-40 follows the same path as 1D case 40 but for the latter case only half the sediment mass was supplied. In the mass-normalized plot, it is seen that the data points from the 1D model coincide. This is expected since the 1D model does not simulate the cohesive
sediment transport and simply deposits one unit of cohesive material for every unit of non-cohesive material deposited. The 2D model results for the non-cohesive sediment cases also show this similarity in the mass-normalized plot. Finally, 2D case 40-40 has the least capacity to build delta in terms of plan area and extent.

Figure 4-16 Radial extent of delta for all four cases with horizontal time axis (left) and mass-normalized time axis (right).

Figure 4-17 Area of delta for all four cases with horizontal time axis (left) and mass-normalized time axis (right).
4.3.2. Delta elevation

Difference in the 2D model delta areas can be explained when examining the delta surface bed elevations distributions, particularly in terms of the mass-normalized time (Figure 4-18). The 50th percentile or median elevation is the largest for case 40-40. For the non-cohesive sediment cases, the median elevation is larger for the larger sediment concentration cases but not significant. The 90th percentile elevation was also calculated to find upper extreme value of elevation. Again the elevation in case 40-40 was significantly higher than the non-cohesive sediment cases. Thus, significant vertical aggradation that occurs when including cohesive sediments is one of reasons that case 40-40 has created less deltaic area.

![Figure 4-18 50th percentile delta elevation (left) and 90th percentile delta elevation (right).](image)

4.3.3. Channel characteristics

Figure 4-19 shows that threshold velocities for non-cohesive sediment cases (case 40, 80 and 120) are similar but and are different than the mixed sediment case (case 40-40). The threshold velocities for non-cohesive sediment cases are generally less than corresponding velocities for case 40-40.
Figure 4-19 Threshold velocity plot for case 40, 80, 120 and 40-40 on mass-normalized time axis.

Like the threshold velocity, channelized area in the delta show different pattern for non-cohesive and mix-sediment cases (Figure 4-20). Initially, the proportion of delta surface covered by channels is much higher for non-cohesive sediment cases. With time, the fraction of the delta area composed of channels decreases for non-cohesive sediment cases and the decrease is not as drastic for case 40-40.

Figure 4-20 Left-distributary channel (plan) area; right-distributary channel area to delta area ratio for cases 40, 80, 120 and 40-40 on mass-normalized time axis.
Except case 120, all cases show an overall increase of channel length with time (Figure 4-21). Increase in channel length coincides with the increase in size of delta. For case 120, channel length increases initially and then decreases. It can also be seen from Figure 4-1 that major channels continue to become longer as the delta grows but many minor distributaries disappear. The persisting channels, however, flow with greater velocity (Figure 4-4 and Figure 4-19).

Figure 4-21 Length of distributary channels for all the cases on mass-normalized time axis

Figure 4-22 shows the plot of average channel widths for all cases with mass-normalized time. It is apparent from the figure that the average channel widths of case 40, 80 and 120 are twice as much as case 40-40. With time the channel width decreases and attains a steady value. Given the average width is just two grid size wide in case 40-40, finer grid may be more suitable to reproduce more accurate distributary networks for this case.
Figure 4-22 Average width of distributary channels for all cases on mass-normalized time axis

4.4. Sensitivity analysis

Additional sensitivity simulations are run in order to gain more insight about the response of the 2D model to some of the important model parameters. Sensitivity tests are conducted for case 80 and case 40-40 only. Morphological factor (MorFac) and horizontal eddy viscosity coefficient are altered and their effect on area and median surface elevation of the delta is observed. The base case simulations (simulations used for above results) has MorFac 60 and horizontal eddy viscosity coefficient 10 m$^2$/s. Sensitivity tests are done for half those values.

4.4.1. Morphological factor

Table 4-2 and Table 4-3 summarize the results for change in delta area and median elevation respectively due to change in MorFac from 60 to 30 and Figure 4-23 and Figure 4-24 show the topographic features at year 40. The maximum change in delta area for case 80 and case 40-40 are $\sim$4% and $\sim$7% respectively. The maximum change in median delta elevation are higher in both
cases: ~17% for case 80 and ~25% for case 40-40. Year 40 result in Figure 4-24 shows that the case 40-40 results are quite different in terms of shape for the two MorFac values.

Table 4-2 Change in delta area for case 80 and case 40-40 for using morphological factor 30. Base case uses 60 MorFac.

<table>
<thead>
<tr>
<th>% change in area</th>
<th>10 yr</th>
<th>20 yr</th>
<th>40 yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 80</td>
<td>+0.05</td>
<td>-1.19</td>
<td>-3.82</td>
</tr>
<tr>
<td>Case 40-40</td>
<td>-2.27</td>
<td>+0.22</td>
<td>-6.88</td>
</tr>
</tbody>
</table>

Table 4-3 Change in median elevation for case 80 and case 40-40 for using MorFac 30. Base case uses 60 MorFac.

<table>
<thead>
<tr>
<th>% change in median elevation</th>
<th>10 yr</th>
<th>20 yr</th>
<th>40 yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 80</td>
<td>-17.39</td>
<td>+1.42</td>
<td>+0.77</td>
</tr>
<tr>
<td>Case 40-40</td>
<td>-0.95</td>
<td>-24.69</td>
<td>-18.30</td>
</tr>
</tbody>
</table>

Figure 4-23 Case 80 sensitivity test of MorFac for delta at year 40. Base case (left) and sensitivity case (right).
4.4.2. Horizontal eddy viscosity coefficient

Horizontal eddy viscosity coefficient is a calibration factor in the 2D model whose value depends on the grid resolution. For coarser grid, larger value of horizontal eddy viscosity coefficient is recommended. The simulations presented earlier uses a horizontal eddy viscosity coefficient 10 m$^2$/s and the sensitivity tests uses 5 m$^2$/s. Percent change in area and median elevation of delta for case 80 and 40-40 are summarized in Table 4-4 and Table 4-5 respectively.

There is small change in delta area and large change in median elevation on decreasing the coefficient value. The maximum change that occurred in case 80 is ~8 % and in case 40-40 is ~4 %. The maximum change in median elevation that occurred in case 80 is ~18% and in case 40-40 is ~36 %. In Figure 4-25 and Figure 4-26, plots for base case and sensitivity case deltas for case 80 and 40-40 at year 40 are shown respectively. Again, the two deltas for case 40-40, though not differing in terms of area, differ a lot in shape.

Figure 4-24 Case 40-40 sensitivity test of MorFac for delta at year 40. Base case (left) and sensitivity case (right).
Table 4-4 Change in delta area for case 80 and case 40-40 for horizontal eddy viscosity coefficient 5 m²/s. Base case has horizontal eddy viscosity coefficient 10 m²/s.

<table>
<thead>
<tr>
<th>% Change in Area</th>
<th>10 yr</th>
<th>20 yr</th>
<th>30 yr</th>
<th>40 yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 80</td>
<td>-1.68</td>
<td>-5.40</td>
<td>-5.20</td>
<td>-7.32</td>
</tr>
<tr>
<td>Case 40-40</td>
<td>+0.23</td>
<td>+2.15</td>
<td>-3.35</td>
<td>-3.32</td>
</tr>
</tbody>
</table>

Table 4-5 Change in median elevation for case 80 and case 40-40 due to horizontal eddy viscosity coefficient 5 m²/s. Base case has horizontal eddy viscosity coefficient 10 m²/s.

<table>
<thead>
<tr>
<th>% change in median elevation</th>
<th>10 yr</th>
<th>20 yr</th>
<th>30 yr</th>
<th>40 yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 80</td>
<td>-17.40</td>
<td>-9.22</td>
<td>2.02</td>
<td>3.86</td>
</tr>
<tr>
<td>Case 40-40</td>
<td>-9.52</td>
<td>-35.63</td>
<td>-25.32</td>
<td>-23.29</td>
</tr>
</tbody>
</table>

Figure 4-25 Case 80 sensitivity test of horizontal eddy viscosity coefficient for deltas at year 40. Base case delta (left) and sensitivity case delta (right).
4.5. Computational performance

The 1D model takes less than a minute to complete simulation on a personal computer. On the other hand, Louisiana State University’s SuperMike-II Linux cluster is used to expedite the 2D model simulation run times. For the simulations used in this study, 32 cores are used. The model domain is automatically divided into 32 parallel partitions or strips along the flow direction such that all strips have nearly equal number of active cells. Figure 4-27 shows the wall time required for case 80 and case 40-40 for different number of computational cores in addition to 32 cores used for the current study. The speed up factor for a simulation is calculated by dividing the reference case (i.e., the 8 core simulation) wall time by the wall time of that particular simulation. As can be seen from Figure 4-28, and supported by Jagers (2012a), the computational performance of the 2D model case 80 didn’t increase linearly with the number of cores.
Figure 4-27 Wall time required to simulate 244 days (equivalent to 40 years with MorFac 60 assuming 365.25 days/year) at time step 15 sec in the 2D model.

Figure 4-28 Speed up factor with 8 core simulation time as the reference in the 2D model
Chapter 5. CONCLUSIONS AND RECOMMENDATIONS

5.1. Conclusions

The primary aim of this thesis is to investigate land built by a simplified Mississippi River sediment diversion using two numerical models: a one-dimensional spatially averaged model, and a two-dimensional, process-based Delft3D model. The second objective is to study the channel network properties from the 2D model simulations. In addition, the computation performance of the two models is compared. For comparison purposes, amid uncertainties in many model parameters, the two models are highly simplified. Deposition is assumed to be the dominant process and erosion below initial bottom bathymetry is not permitted. Also, for the purposes of this thesis the sediment diversion operates continuously at a given flow (2000 m$^3$/s) and sediment concentration. Given the simplified assumptions regarding sediment transport in the 1D model and the uncertainty in the sediment concentrations that the sediment diversion could capture, four scenarios are simulated. Three of the scenarios are based on varying concentrations of non-cohesive sediment: 40 ppm sand (case 40), 80 ppm sand (case 80) and 120 ppm sand (case 120); the fourth scenario utilizes a sediment concentration that is a 50/50 mix of non-cohesive (40 ppm) and cohesive (40 ppm) (case 40-40). It is important to note that the quantitative results provided in this thesis are not necessarily representative of the land building capacity of any proposed or future sediment diversions. The scenarios used in this work are developed to facilitate a direct comparison of the two modeling approaches. Also, while the diversion flow and sediment concentrations used here are within the range of potential strategies, the operating strategy, i.e., continuous operation, is not.

For all the simulated cases, the 1D model always produced deltas that were greater in area compared to those from the 2D model. Except for the local variations, the subaerial land in the 2D
model simulations did not extend as far into the receiving basin as in the 1D model. However, the
process-based 2D model produced deltas that were richer in topographic detail and similar in
nature to simulations by Edmonds and Slingerland (2009) and Geleynse et al. (2010). As expected,
given that waves and tides were not included, the deltas produced in the 2D non-cohesive
simulations were somewhat fan-shaped and had smooth shorelines due to the presence of evenly
distributed distributaries on the delta surface. On the other hand, the mixed cohesive and non-
cohesive sediment case delta showed complex delta growth pattern with a very rough shoreline.
From these four simulations, it can be concluded that the radial growth assumption in the 1D model
is more correct for purely non-cohesive sediment delta.

Delta area and average radial extent for the simulations were in the order: 120 (1D) > 80 (1D)
= 40-40 (1D) > 120 (2D) > 80 (2D) > 40 (1D) = 40-40 (2D) > 40 (2D). The ratio of the 2D to 1D
model delta area for the non-cohesive sediment cases was about 70%. This ratio was much smaller,
about 55%, for the mix-sediment case. For a unit volume of input sediment, the land built by the
mix-sediment case was the lowest. Whether the presence of cohesive sediment decreases the
overall land building capacity per unit volume of sediment is not 100% conclusive from this single
simulation. The most important point to be noted here is that it shows cohesive sediment does
affect the dynamics of delta growth (Edmonds & Slingerland, 2009) and, in some cases, even
hinders the growth. Since sediment diversion will be diverting both cohesive and non-cohesive
sediments, these results, to some level, support Edmonds and Slingerland (2009)’s argument that
diverted sediment fractions must be controlled to maximize the amount of land created. The
sensitivity analysis of the 2D model parameters, morphological factor and horizontal eddy
viscosity coefficient, showed little effect on delta area. A 50 % reduction in these parameters
causd less than a 10 % change in simulated delta area.
Elevations of the 1D model delta surface were very low, which is a consequence of spatial averaging of the 1D model over the delta plain. The elevation decreased away from the vertex, which has the maximum elevation, and gently sloped to zero at the shore. On the other hand, the 2D model captured the spatially and temporally varying topography and bathymetry resulting from complex distributary networks that shifted with time. Statistical distribution of the delta surface elevations (50 and 90 percentile) highlighted that the mix-sediment delta had higher elevations than those in the non-cohesive sediment cases. This is in part supported by the fact that cohesive sediment are hard to erode once they are deposited and thus promote vertical aggradation of sediment (Edmonds & Slingerland, 2009). It can also be reasoned from this that more vertical aggradation in the mix-sediment case prevented horizontal expansion of land and resulted in a smaller delta area.

The 1D model represents the distributary network of a deltaic system by a single virtual channel. Qualitative and quantitative results shown in this thesis highlight the significance of the distributary channels and the impact on delta growth and elevations. This difference is particularly significant given the distinct physical and flow characteristics of the distributaries for the 2D non-cohesive and mix-sediment cases. While the addition of cohesive sediment promoted weakly sinuous distributaries, the solely non-cohesive sediment cases showed numerous short channels spread all over the delta surface. Furthermore, cohesive sediment mix contributed to a condition that promoted distributaries that were narrower and had higher flow velocities. The surface coverage of delta by channels was higher for non-cohesive sediment cases.

The 1D model is a robust tool to predict land created by sediment diversions under simplified conditions. All the 1D simulations took less than a minute to run on a personal computer while the
2D model simulations required hours on multi-core computers. Without sufficient computational resources, the long-term simulation of a sediment diversion in a 20 km by 20 km domain wouldn’t have been possible. Despite its simplicity and economy, the 1D model results for the delta radius and area were always larger, but relatively close, to those simulated using the 2D model, particularly for non-cohesive sediment cases. Two critical limitations of the 1D model were highlighted here: (1) the 1D model does not simulate the transport of the cohesive sediments, but simply deposits a user-specified unit of cohesive material for every unit of non-cohesive material; and (2) the 1D model does not provide any details on the distributary network. These distributary channels have a very important influence on biological activity of vegetation and aquatic animals in coastal environments (Paola et al., 2011). Studies have shown that primary productivity is highest along the channel margins (Haas et al., 2004). Across the delta, productivity of aquatic animals is influenced by the availability of shallow channel zones that offer protection from predators. This makes the 2D model a valuable tool in predictive delta ecogeomorphology.

5.2. Recommendations

First of all, the 2D model should be run with Engelund and Hansen (1967) sediment transport formula instead of Van Rijn (1993) once the problem with Delft3D is fixed. Due to a bug in Delft3D at the time of this study, parallel core computation using Engelund and Hansen (1967) formula was not possible. There was a difference in the receiving basin between the 1D and 2D models. Better approach should be sought to eliminate this difference.

The results and conclusions from the 2D model can be solidified in the future by running more scenarios and by better schematization of the model. There was only one mix-sediment case. Additional mix-sediment cases should be run. Though a deltaic environment is by definition net depositional in nature, studies have shown that distributaries channels can erode the pre-delta
substrate and condition the morphological development (Geleynse, 2013). This necessitates better schematization of the initial bed condition than the non-erodible initial bed used in this thesis. In mix-sediment case the delta elevations were as high as 3 m above mean water level near the junction between the diversion channel and basin. This could be due to excessive backing up of flow due to non-erodible bed assumption. Erodible bed could have allowed deepening of the bed in addition to backing up of flow. Mouth bars and deltas are also conditioned by the hydrodynamics in the fluvial channel (Edmonds & Slingerland, 2009). This requires running simulations with different diversion channel dimensions and/or flow conditions. More realistic diversion operation strategy should be implemented. The diversion need to be operated only during the peak flow period in the Mississippi River. Additional simplified (e.g., deeper, varying slope) and complex (e.g., based on bathymetry data of the site) receiving basins should also be implemented.
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