Simulation of turbidity maximums in the York River, Virginia

Jae-Il Kwon

College of William and Mary - Virginia Institute of Marine Science

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SIMULATION OF TURBIDITY MAXIMUMS IN THE YORK RIVER, VIRGINIA

A Dissertation
Presented to
The Faculty of the School of Marine Science
The College of William and Mary in Virginia

In Partial Fulfillment
Of the Requirements for the Degree of
Doctor of Philosophy

by
Jae-Il Kwon
2005
APPROVAL SHEET

This dissertation is submitted in partial fulfillment of the requirements for the degree of

Doctor of Philosophy

Jae-II Kwon

Approved by the Committee, June 2005

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US Army Corps of Engineers
Vicksburg, Mississippi
To My Family for all of their unending love
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ACKNOWLEDGEMENTS

I am indebted to my advisor, Dr. Jerome P.-Y. Maa, for his advise, guidance, support, and encouragement throughout my study. Without him, this dissertation would not be possible. I am grateful to Dr. Sung-Chan Kim for his many helpful suggestions and discussions on this study and many various research topics. I am thankful to Dr. Jian Shen for his thoughtful comments and assistance in numerical models, and to Dr. Carl T. Friedrichs for the editorial efforts for the manuscript and sharing of valuable data. I would like to thank Dr. Robert J. Diaz for his comments and encouragement.

I thank Sam Wilson, Bob Gammisch, and Wayne Reisner for their help in collecting the field data used in this study. I am thankful to Mr. Mac Sisson for his many helps in staying at VIMS and in proof reading the manuscript. Many thanks should go to Cynthia Harris, Sue Presson, Beth Marshall, Cindy Hornsby, and many other VIMS faculty and staffs for their support and assistance.

I am grateful to Kyung Ho Cho, Ho Kyung Ha and Drs. Jin Yong Choi, Kyeong Park, Jeong Hwan Oh, and many scientists who were visiting scholars from KORDI for enriching my family’s life with joy in VIMS. Many thanks also go to Drs. Dong Kyu Lee, Seung Hyun Son, and Jun Yong Park for their support and continuous encouragement. I would like to thank my dear friends, Byung Soo Kim and Kyun Soo Ahn for their support. I also thank the EPA program for supporting this study.

My special thanks to my parents, my sister, and my wife, Ja Young, for their precious love and support throughout many years. Finally, a true miracle in my life, my son, Tony.
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ABSTRACT

Two of the most important processes in cohesive sediment transport, erosion rate and settling velocity, were the focus of this study. Settling velocities were estimated by the Owen tube method and the acoustic Doppler velocimeter (ADV) method. A novel erosion model, namely a constant erosion rate model, was implemented in a three-dimensional hydrodynamic eutrophication model (HEM-3D) to simulate the turbidity maximums in the York River system, Virginia.

Two one-month periods of model simulations were conducted to mimic typical dry (November–December, 2001) and wet (March–April, 2002) seasons. In order to have enough data to verify the model, four slack water surveys were carried out during each period to measure salinity and total suspended solid (TSS) profiles. Because of the unexpected extremely low freshwater discharge during both those periods, all survey results showed similar salinity and TSS distributions. The estuarine turbidity maximums were abnormally located about 30 km upstream from West Point, with TSS concentrations on the order of $10^2$ mg/L.

Laboratory Owen tube experiments showed that the settling velocity was related to the TSS concentration, highlighting the importance of sediment availability on settling velocity and the less important salinity effect. The estimated settling velocities from four sets of ADV field measurements were much higher than that from the Owen tube laboratory experiments and better reproduced the turbidity maxima for slackwater simulations. These suggested that turbulence may have a dominant effect on settling velocity, and the ADV method seems to be an effective and suitable way to estimate the settling velocity in turbulence dominated environments.

Based on a newly found erosion behavior, a constant erosion rate model was implemented in a three-dimensional numerical model such that erosion occurs only during accelerating phases of the tide. Specifically, the Four Factor Model was suggested that consists of (i) a reference constant erosion rate, (ii) hydrodynamic effects, (iii) spatial variability of the bed condition, and (iv) temporal variability of the bed condition. The Four Factor Model successfully simulated the turbidity maximums in the York River system.
SIMULATION OF TURBIDITY MAXIMUMS IN THE YORK RIVER, VIRGINIA
CHAPTER I

INTRODUCTION

Estuaries and coastal regions are of great economic and environmental interest. They have been widely and heavily changed by both human activities and Mother Nature. The understanding of estuarine and coastal processes and the capability of predicting their responses are important for an estuary with great economic potential. Among many factors involved in estuarial processes, cohesive sediment transport is one of the difficult but important subjects. For instance, the accumulation of fine cohesive sediments may hinder navigation in channels and bring in contaminants (e.g., heavy metals, insecticides, petroleum by-products, and radio-nuclides). Cohesive sediment can remain in suspension for a long time, and it can damp light penetration and reduce the thickness of the euphotic zone. Consequently, it may limit primary productivity and may prohibit submerged aquatic vegetation (SAV) growth. Therefore, the ability to predict cohesive sediment transport and the formation of turbidity maxima in estuaries is an important step toward mitigation of water quality problems.

The schematic diagram (Fig. 1-1) shows the major processes involved in cohesive sediment transport, and a summary is given below:

- River flow, shoreline erosion, bottom erosion, and/or sea are the major sources of the sediment
- The erosion process is dependent on bottom shear stress and the bed condition.
- Once sediments are eroded, then turbulent diffusion and advection move them into the water column.
- Turbulence is a key factor in controlling the flocs size distribution. It not only can break the flocs, but it also can promote the increased size of flocs. Stratification caused by salinity or suspended sediment may damp out the turbulence.
- Salinity and biological materials also tend to increase the floc size distribution.
- Bioturbation can change the bed condition.
- Settling velocity plays an important role in the redistribution of suspended sediment concentration (SSC) in the water column. It is a function of floc size, shape, density, surface roughness, SSC, and fluid properties such as salinity and turbulence.
- Deposition processes serve as a sink of sediment by moving flocs to the bed.
- Consolidation processes will cause a profile of increasing density and strength with depth within the bed.

Despite decades of studies, prediction of cohesive sediment transport is still hard to achieve. For instance, floc dynamics is so complicated (Tsai et al., 1987; Manning and Dyer, 1999; McAnally and Mehta, 2001; Winterwerp et al., 2002;) and it is difficult to have sufficient and precise data of flocs’ properties for one’s own study.

Biological effects are also important in terms of aggregation (Van Leussen, 1988) and changing the bed condition (Austen et al., 1999; Andersen, 2001; Aller, 2001). The best way to account for these biological effects at present is to modify the parameters such as settling velocity, threshold for erosion, etc. (Winterwerp and Kesteren, 2004). But neither settling velocity nor a critical bed shear stress for erosion is easy to measure.
Settling velocity is one of the most important aspects in assessing the transport and fate of cohesive sediment suspensions in the marine environment (Winterwerp and Kesteren, 2004). The wide ranges of size, density, and fragility characteristic of cohesive sediment make measurement of the settling velocity difficult (Dearnaley, 1996; Jones and Jago, 1996; van Leussen, 1999; Winterwerp, 2002). Decades of studies on the developments of in-situ instrument techniques for the settling velocity are well summarized in a review paper (Eisma et al., 1997), but still there is no perfect method to measure the settling velocity.

The cohesive sediment continues to settle toward the bed owing to its settling velocity (Owen, 1977). The most popular deposition rate formulation is based on the work of Krone (1962). But this formulation is not easy to use because it contains settling velocity, the sediment concentration right above the bed, and a critical shear stress for deposition. Moreover, the existence of a critical bed shear stress for deposition is still debatable because there are conflicting experimental results from various laboratory and field studies (Sanford and Halka, 1993, Winterwerp and Kesteren, 2004).

During the formation of a bed of cohesive sediment flocs are continually being deposited on the surface of the bed while the buried flocs are consolidating, so that there exist a profile of density and strength within the bed (Owen, 1977). The behavior of a fluid mud layer in-situ may be predicted by a mathematical model that uses constitutive relationships obtained from laboratory experiments (Merckelbach et al., 2002; 2001) such as settling column experiments (Migniot and Hamm, 1990; Sills, 1997). Numerical bed models to represent the bed in cohesive sediment transport are still requires computing costs (Gibson et al., 1981; Cargill, 1982; Fox and berles, 1997; Govindaraju et al., 1999;
Experiments in the laboratory and in the field have been aimed at relating a critical erosion velocity, or a critical bed shear stress for erosion, to the properties of the mud (Parchure and Mehta, 1985; Dyer, 1986; Krone, 1993; Maa et al., 1993; Sanford and Halka, 1993; Kranenburg, 1999; Sanford and Maa, 2001; Lin et al., 2003). Commonly, cohesive sediment transport models have used either a dimensional or a non-dimensional excess bed shear stress to define the erosion rate at a particular time and location (Geyer et al., 1998; Teeter, 2001; Liu et al., 2002; Ganaoui et al., 2004). Because there is no way to know the change of the critical bed shear stress for erosion with time, tuning is inevitable in the modeling of cohesive sediment transport.

As briefly summarized above, for better understanding and prediction of the cohesive sediment transport, precise and adequate technology to measure the properties of cohesive sediment, and new approaches to avoid or reduce known shortcomings, are necessary. For instance, acoustic Doppler velocimeter (ADV) can be used to estimate the settling velocity while avoiding existing drawbacks such as blocking the ambient turbulence and sensitivity to sediment characteristics, from other methods (Fugate and Friedrichs, 2002; 2003). Maa and Kim (2002) suggested a novel erosion rate scheme based on a newly found erosion behavior - erosion occurs only during accelerating tidal phases. Unlike the traditional erosion scheme, this simple approach may significantly reduce the difficulty of future modeling efforts by changing two or three unknowns to only one.

Therefore, this study concentrates on the erosion rate and settling velocity needed to simulate estuarine turbidity maxima (ETMs) in York River system, VA. The dynamic
The overall objective of this study is to simulate ETMs in estuaries. The detailed and specific sub-objectives are (1) to implement a newly found simple erosion behavior, namely a constant erosion rate, in a three-dimensional hydrodynamic and eutrophication model (HEM-3D) in the York River, (2) to explore the most appropriate range of settling velocities that can be used in the model, and (3) to predict the formation of the turbidity maxima and the suspended sediment distribution in the York River system.

HEM-3D was used for numerical simulations and for applying new sediment
transport process schemes. It is a fully verified hydrodynamic and eutrophication model, but had limited capabilities in sediment transport modeling (Park et al., 1995). HEM-3D has previously been used in the York River (Shen et al., 1997; Sisson et al., 1997) and the James River (Shen and Kuo, 1999).

The York River System (Fig. 1-2) was selected to implement a sediment transport model because of well documented characteristics of the hydrodynamics and salinity intrusion (Shen et al., 1997), a basic understanding of the suspended sediment distribution (Lin, 2001), and in-situ measurements of erosion rates (Maa and Kim, 2002). Studies of sediment accumulation rates, sediment composition, and bed shear stress measurements have also been conducted (Kim et al., 2000). Indeed, only marginal field work was required to complete the data sets, greatly facilitating the development of a complete 3-D sediment transport model.

Two one-month periods of model simulations were conducted to mimic the typical dry (October) and wet (March) seasons. In order to have enough data for calibrating the model, four slack water surveys were carried out during each period to measure conductivity, temperature and Total Suspended Solid (TSS) profiles. Settling velocities were measured using the Owen tube method. Data from Acoustic Doppler Velocimeters (ADVs) (provided by Malcolm Scully and Carl Friedrichs) were also used to estimate the settling velocity.

Details of field surveys and previous studies in the York River system are given in Chapter 2. Descriptions of HEM-3D as well as the calibration and verification using the field data during two slackwater surveys are discussed in Chapter 3. Various settling velocity measurements and estimated settling velocity formulations from the Owen tube
and ADV methods are presented in Chapter 4. The concept and development of a simple and new erosion rate model, the constant erosion rate, are given in Chapter 5. Results of numerical experiments and the application of the constant erosion rate model, Four Factors Model, are described and discussed in Chapter 6. Discussion and conclusions are given in Chapter 7.
Influx
(River flows, Shorelines, and Sea)

Biological Materials

Flocculation

Salinity

Mixing / Damping

Suspended Cohesive Sediments

Settling

Turbulence

Deposition

Erosion

\( T_b \)

Bioturbation

Consolidation

Bed

Fig. 1-1. Schematic Diagram of Cohesive Sediment Transport Processes.
Fig. 1-2. The York River System, a tributary of the Chesapeake Bay. The lower insert shows a part of the curvilinear model grid that follows the bathymetry. Dark areas in the lower insert is the channel. Only the 1st (YR01) and the last (YR25) survey stations are shown in the diagram.
CHAPTER II
STUDY AREA AND FIELD SURVEYS

2-1. Introduction

The York River system was selected to implement the suspended sediment transport model because of previous work in numerical modeling of hydrodynamics and salinity (Shen et al., 1997), a basic understanding of the suspended sediment distribution (Lin and Kuo, 2001), in-situ measurements of erosion rates (Maa and Kim, 2002), sediment accumulation rates, sediment composition, and bed shear stress measurements available for this river (Kim et al., 2000). Therefore, only a small amount of additional fieldwork was required to complete the data sets, greatly facilitating the development of a 3-D suspended sediment transport model.

Two one-month periods of model simulation were conducted to mimic the expected dry (November 2001) and wet (March 2002) seasons for hydrodynamics and bed conditions, which were anticipated to have different suspended sediment distributions. Four slackwater surveys were carried out during each period to measure conductivity, temperature, and vertical TSS profiles for model calibration and verification.
2-2. Study Area and Previous Studies

The York River system is a tributary of the Chesapeake Bay (Fig. 1-2). It is composed of three rivers, *i.e.*, the York, Pamunkey, and Mattaponi. The upstream two branches, the Pamunkey and Mattaponi Rivers, meet the York River at West Point (about 50 km from the York River mouth). The York River estuary is a drowned river valley formed about 7,000 years ago. The principal bathymetric features consist of an axial channel flanked by broad shoals. These reflect the ancestral river channel and bordering flood constricted zone at Gloucester Point. There are two channels in the York River. The main channel is about 10 m deep and a secondary channel which runs parallel is about 5 m deep near Clay Bank. These channels run downstream to the southeast with the secondary channel on the southwest side of the river. These channels play a critical role in both salt and sediment transport. Therefore, a careful and precise model grid that includes these channels is required. Details of the model grid generation process are given in Chapter 3.

Bottom sediment texture or size distribution was investigated by Nichols *et al.* (1991). In the middle part toward the upstream end of the York River, the mud percentage of the bottom sediment is quite high, bordered with sandier beds in its upstream and downstream ends. Physical mixing to depths from 40 to 120 cm was reported at the secondary channel of the York River (Dellapenna *et al.*, 1998). The water content of the bottom sediment (top 20 cm) in the channel of the middle part of the York River varies from 60 to 80 %, and the porosity ranges between 0.8 and 0.9 (Dellapenna, 1999), which indicates a high rate of bottom sediment resuspension to the water column.
York River is a partially mixed estuary. The York River has over 1 m/s surface tidal current on spring tides, and the mean tidal range is 0.7 m at the mouth and 1 m at the head (Schaffner et al., 2001). From a mooring study (Sharples et al., 1994), lower stratification or complete mixing was associated with the strong spring tides, while significant stratification developed during weaker neap tides. Such a spring-neap signal was caused by the modulation of tidal mixing energy that was competing with the stratifying estuarine circulation. In terms of flood-ebb variation, a tripod study showed that during ebbs, the shear velocities near the bed were consistently greater than those during floods (Friedrichs et al., 2000).

In the York River system, the fine-grained sediment is cycled within the estuary by the estuarine circulation (Nichols et al., 1991). In general, the route is (1) seaward through the freshwater reaches of the Mattaponi and Pamunkey Rivers, (2) seaward through the upper layer from about 10 to 20 km upstream of West Point to the mouth, (3) downward by settling into the lower estuarine layer, and (4) landward through the lower estuarine layer to the salt intrusion limit about 10 to 20 km upstream of West Point. The precise location of the limit of the salt intrusion is highly dependent on river discharge.

The salinity distribution of the York River system is affected by the interaction of freshwater discharge, salinity distribution at the mouth, tidal energy, and wind. Salinity gradients between the surface and bottom are influenced by neap and spring tidal cycles with destratification of the water column occurring at spring tides and stratification developing during the intervening periods (Haas, 1977). During low freshwater flow conditions, salt water may extend up to 30 km upriver from West Point (Lin, 2001). The relationship between river discharge and the locations of the estuarine turbidity maximum
(ETM) and the secondary turbidity maximum (STM) was reported by Lin and Kuo (2003). Their model simulations showed that both the prominent ETM and STM occur when river discharge is relatively low, and, at higher freshwater inflow, the two turbidity maxima move closer to each other. Their model study indicated the location of the ETM is well associated with the null point of bottom residual flow. More details regarding freshwater discharge during the two survey periods associated with the present study will be discussed later.

During 1996 and 1997, a series of slackwater surveys (about once a month over a one-year period) along the York River were conducted and revealed the general salinity and suspended sediment distributions. Two possible ETMs were found. The primary ETM was found near the end of salt intrusion, upstream of the York River. A STM, however, was also observed in some of the surveys in the middle of the York River. Lin and Kuo (2001) suggest that resuspension of bed material may be the major source of the STM and that three water column processes are generally important to the formation and maintenance of the STM: convergence of bottom residual flow, the stratification gradient along the channel, and tidal asymmetry.

The data from the previous study (during 1996 and 1997) may only be sufficient to verify the model results for a relatively long-term simulation. However, the interval between each slackwater survey was about a month, and the TSS concentrations were measured at only three elevations at each station. This sparse data is not suitable to resolve the distinct spring-neap cycle in the York River. Higher sampling frequencies and more vertical resolution of TSS profiles during the simulation period (e.g., four sets of data in one month) are needed so that the dynamics can be understood and simulated.
In this study two one-month periods of model simulation were conducted to mimic the expected dry (November 2001) and wet (March 2002) seasons. In order to have enough data for calibrating and verifying the model, four slackwater surveys were conducted during each period to measure conductivity, temperature, settling velocity, and vertical TSS profiles. The only in-situ measurements of erosion rates used were from Maa and Kim (2002) and details are given in Chapter 5.

2-3. Field Surveys

2-3-1. Sampling Stations

Twenty-five stations along the main channel of the York River and the Pamunkey River were selected (Fig. 2-1). Because of the limited resources and the relatively small dynamic range of freshwater discharge in the Mattaponi River (Fig. 2-2), the slack water stations on the upstream side were selected along the Pamunkey River. Note that the discharge information was obtained from two USGS stations: one is near Hanover on the Pamunkey River (about 170 km from the York River mouth) and the other is near Beulahville on the Mattaponi River (about 135 km from the York River mouth).

Although the original objective was to have one survey period for a dry season and the other for a wet season, the extremely dry year following July 2001 caused the two data sets to be very similar (Fig. 2-2). This was unexpected and beyond our control. The first survey period was extended a little because of bad weather and the seasonal closure of VIMS at the end of the year.

The coordinates of the 25 stations are given in Table 2-1. Notice that the distance between each station was short (between 4 to 5 km) because the objective was to obtain a better axial resolution of salinity and TSS gradient at the places where the TSS gradient
was large. It is obvious that not all stations were needed if the axial gradient is small. For this reason, some upstream stations (i.e., between YR17 and YR24) moved a little in each survey to find the maximum TSS concentrations and gradients. Also, because of this reason, not all of the surveys had measurements at all of the stations listed in Table 2-1. All stations were located in the main channel in order to get the salinity and TSS information where the maximum transport occurs. Also, it should be noted that these surveys were carried out at local slack tide, either after a flood or after an ebb tide.

2-3-2. Sampling Methods and Equipment

Conductivity and temperature profiles were measured using an Apply Micro CTD profiler, model 663. A Seapoint Optical Backscatter Strength (OBS) sensor was mounted with the CTD profiler to get continuous OBS readings. A water pump with its inlet aligned at the same elevation as the OBS was used to take water samples whenever the OBS reading showed a significant change. As a result, water samples were taken at almost all of the surveyed stations to establish an in-situ calibration equation for each survey to convert the OBS readings to TSS readings. All the CTD profiles were reasonably smooth and could be used directly to construct the “snap” shots of salinity distribution. The OBS readings, however, required further processing because of the reasons given next.

2-3-3. Calibration for TSS Concentration

Raw OBS readings showed a large fluctuation in almost all of the profiles because of the possibility that fish, sea grass, or any solid subject moving around the OBS sensor
can cause a spike or abnormal readings. Notice that not all OBS profiles showed a high
gradient near bottom. For some downstream stations (i.e., YR01 and YR02 in general, or
YR06 in Fig. 2-3) and at the upstream end when the water depth was shallow (YR24), the
OBS profiles were nearly uniform. When local convergence occurred or at the ETM
zone, the OBS profiles did have a significant gradient over the lower part of the water
column (i.e., YR09 and YR22 in Fig. 2-3).

An OBS sensor is very sensitive to particle size and the reflection index of
suspended particles (Maa et al., 1992). Thus, it was necessary to conduct in-situ
calibration during the surveys. Also for this reason, an OBS sensor might respond
differently when it is in the top of the water column or in the bottom of the water column.
This is because the size of suspended solids may be different in different parts of the
water column. It was found that the OBS calibration curves were slightly different
depending on the sensor location, e.g., at the top or at the bottom of the water column, at
the upstream or downstream end of the York River (Fig. 2-4). Fortunately, the difference
was not significant, and when considering the data scatter, it was not worth using
different calibration equations for each section. Using all the in-situ calibration data
points to construct an OBS calibration equation was reasonably good ($R = 0.93$). Thus,
eight calibration equations were developed to convert the smoothed OBS vertical profiles
to TSS profiles.

2-4. Survey Results

Survey results were very similar because of the extremely dry year. Nevertheless,
the two data sets provided one extreme case for checking the performance of HEM-3D
for the salinity intrusion and for validating the TSS distribution predicted by the module developed.

2-4-1. Salinity distributions

Eight salinity and TSS distributions along the York-Pamunkey Rivers during survey periods are shown in Figs. 2-5 to 2-12. In general, the salinity at the York River mouth was around 24-25 ppt. For a normal year, the salinity would be about 5 ppt at West Point. During the two periods of slackwater surveys with severe dry conditions, the measured salinities at West Point were around 15 ppt. The maximum salinity intrusion distance was about 90 km from the York River mouth during these slackwater survey periods. The stratification caused by the salinity distribution was not strong; most of the time, it was close to uniform in the vertical direction.

Although the aim of this study is not focused on the salt intrusion, the model must be able to simulate the salt intrusion in order to simulate sediment transport in the York River system. All salinity distribution data were used to verify the model performance (See Chapter 3).

2-4-2. TSS distributions

At the downstream end of the York River, the TSS profiles clearly indicated a gradual increase with water depth. Even at depths that were close to the bottom, the TSS concentrations were still low, and only increased about 10 to 20 mg/L. At stations near the upstream turbidity maximum, the TSS profiles increased quickly and had a significant gradient in the middle of the water column. For stations in the middle section of the York River, vertical profiles varied. Nevertheless, all available vertical profiles for one survey
were smoothed and used to construct a snapshot of the TSS distribution for that particular survey.

In general, the TSS concentrations were low and about the same from cruise to cruise near the York River mouth. The existence of the primary turbidity maximum was obvious and mostly located behind the head of the salt intrusion (salinity from 1 to 4 ppt). On April 2, 2002, the ETM located far behind the limit of the salt intrusion (~11 ppt). A more detailed explanation for ETM location related to the salt intrusion is given in Chapter 7. Among these survey results, Jan 18, 2002 (Fig. 2-8(b)) and Apr. 11 (Fig. 2-12(b)) did not show a clear turbidity maximum. A noticeable STM was observed near Clay Bank on four cruises.
Table 2-1. Water Sampling Stations Coordinates for the Slack Water Surveys.

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<th>Lat. (deg)</th>
<th>Dist.(km)</th>
<th>Depth(m)</th>
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Fig. 2-1. Water Sampling Stations for Slackwater Surveys Along the York River System.
Fig. 2-2. Historiography of Freshwater Discharge and Durations for the Slackwater Surveys in the York River System.
Fig. 2-3. Examples of Measured Vertical Profiles of OBS Readings on March 19, 2002. "+" is raw data with spikes removed and "o" is a half meter averaged. (a) YR01, (b) YR06, (c) YR09, and (d) YR22.
Fig. 2-4. An Example of OBS Calibrations (April 11, 2002). Open circles and black circles represent sampling elevation near the bed and near the surface, respectively. Possible regression lines are plotted (solid line is calculated using all data, dashed line is based on the near surface samples in middle section of the York River, dash-dotted line is from the near surface data in the downstream, line with x mark is for the near bed samples in the middle area, and line with square is from the near bed data in the upstream stations.)
Fig. 2-5. Slackwater Survey on November 29, 2001. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-6. Slackwater Survey on December 5, 2001. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-7. Slackwater Survey on December 10, 2001. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-8. Slackwater Survey on January 18, 2002. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-9. Slackwater Survey on March 19, 2002. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-10. Slackwater Survey on March 25, 2002. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-11. Slackwater Survey on April 2, 2002. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
Fig. 2-12. Slackwater Survey on April 11, 2002. (a) Salinity (ppt) and (b) TSS (mg/L) distributions.
CHAPTER III
NUMERICAL MODEL SETUP

3-1. Introduction

In this study, the three-dimensional Hydrodynamic Eutrophication Model (HEM-3D) was used to simulate the sediment transport in the York River system. In this chapter, the HEM-3D and a fine curvilinear-orthogonal grid are introduced. Tidal calibration and salinity verification are examined using historic data and new measurements. The suspended sediment transport module in the HEM-3D and all necessary boundary conditions except the bottom boundary condition are also introduced.

3-2. HEM-3D

The Environmental Fluid Dynamic computer Code (EFDC, Hamrick, 1992; 1996) developed at the Virginia Institute of Marine Science is a sophisticated hydrodynamic model that is capable of predicting small-scale processes such as salinity front formation (Shen and Kuo, 1999). The EFDC comprises the hydrodynamic portion of the HEM-3D (Park et al., 1995), in which water quality simulation is integrated with the hydrodynamic codes. The EFDC resembles the widely used Princeton Ocean Model (POM) (Blumberg and Mellor, 1987) in both the physics and the computational scheme used. The model uses the finite difference method to solve the full three-dimensional equations of motion.
for tidal flows with free water surface and the continuity equation for water mass, and
conservation of salinity and heat (Hamrick, 1996). It also includes non-linear terms,
bottom friction terms, the Coriolis force, and wind stress terms. Bottom friction is
specified through the bottom roughness height, $r$. This model’s external forcing includes
surface wind stress, heat, salinity fluxes, freshwater discharge and tides.

For turbulence closure, Mellor and Yamada’s level 2.5 model (Mellor and
Yamada, 1982) modified by Galperin et al. (1988) is implemented in the model.
Dynamically coupled transport equations for turbulent kinetic energy and the turbulent
length scale are solved to provide an accurate representation of the turbulent field.

The model uses sigma coordinates in the vertical direction and curvilinear
orthogonal coordinates in the horizontal plane. The finite difference model uses an
internal-external mode splitting procedure to separate the internal shear or baroclinic
mode from the external free surface gravity wave (Hamrick, 1996). The solution for the
external mode uses a semi-implicit scheme to allow large time steps, which are
constrained only by the stability criteria of the explicit central difference or upwind
advection scheme used for the nonlinear accelerations (Hamrick, 1996).

HEM-3D is capable of simulating density and topographically induced circulation
as well as tidal and wind-driven flows, and spatial and temporal distributions of salinity
and temperature. It also has the capability of simulating a moving boundary, which is
especially useful for those areas that have large tidal ranges or large tidal flats. The
wetting and drying process is included in this model to simulate better not only the
hydrodynamics, but also water quality and sediment transport.
The current sediment transport module in the HEM-3D is a preliminary module that supports a single class of sediment sizes, but this module has not been fully verified yet. In this study, a new constant erosion scheme is added (see Chapter 5 for details) to simulate the turbidity maxima in the York River system.

3-3. A High-resolution Curvilinear Orthogonal Grid

In any estuary modeling effort, it is important to have a correct representation of channels for salt and sediment transport. The existing bathymetric grid in the York River system does not have sufficient resolution. For example, this coarse grid caused the naturally continuous channel to be discontinuous (Fig. 3-1). More importantly, a portion of the channel disappeared in the old grid. For example, the secondary channel at Clay Bank and the dredged channel near West Point were missing. These two important deficiencies would definitely affect the salinity intrusion, especially for the extreme dry year for which salinity was about 15-19 ppt at West Point.

A curvilinear orthogonal grid with high resolution was generated with a grid size of about 110 m in the cross-channel and 170 m in the along channel directions (Figs. 3-2 and 3-3). This new grid was fine enough (i.e., having 2 or 3 cells to represent the channel) to represent the channels and bathymetry, and reduced a possible excessive numerical horizontal diffusion problem. Numerical horizontal diffusion caused by large horizontal gradients in water depth may misrepresent the salt intrusion as well as suspended sediment transport.

One of the characteristics of the new grid was that the channels approaching Clay Bank be correctly represented by two or three cells in the transverse direction (see Fig. 3-
3b). Then, it would not be an obstacle to salt intrusion. Another important feature of the new grid was that the grid gradually merged the horizontal two-dimensional (2-D) grid and one-dimensional (1-D) grids of tributaries (Fig. 3-2b). On the upstream side, the Pamunkey and Mattaponi Rivers are still represented as a 1-D system because of the very narrow channel.

A digital bathymetric data set is available for the Chesapeake Bay and its tributaries from National Oceanic and Atmospheric Administration (NOAA). The quality of this data set is excellent. Because of the high density of the raw bathymetry data, it was possible to use a high resolution of grid and consequently to resolve small channels. A further confirmation of bathymetric cross-sections at three selected locations using NOAA data was made with a rough field survey conducted by running a small boat across the channel. This indicated that the modeled bathymetric grid was sufficient to represent the real bathymetry (Fig. 3-3).

3-4. Tidal Calibration

Because the performance of HEM-3D had already been demonstrated by others (Sisson et al., 1997) for the York River, there was no need to test every detail. For example, if the behavior of the $M_2$ tide were correct, then there was no reason that the behavior of the $S_2$ tide would be incorrect. For this reason, only the performance of the $M_2$ tide was checked. Because the energy of $M_2$ tide alone is about 89% of all the major seven constitutive tides ($M_2$, $S_2$, $N_2$, $K_1$, $M_4$, $O_1$, and $M_6$) (Sisson et al., 1997), checking the $M_2$ alone was sufficient for the tidal calibration purposes.

Along the York River System at the time of this study, there was one NOAA-
VIMS cooperative tide station at the Research Pier of VIMS at Gloucester Point with more than 20 years data. There have been 14 other short-term tide stations established by VIMS as part of previous studies (Sisson et al., 1997). The $M_2$ amplitudes at all 14 of these short-term stations were adjusted to be consistent with the long-term $M_2$ amplitudes by using the same period of short-term tidal records obtained from the NOAA-VIMS station. Details were given in Sisson et al. (1997).

The amplitude of the $M_2$ tide, 0.28 m, was used as the downstream boundary condition at the York River mouth. At the two upstream ends, the mean freshwater discharges were used. And they were obtained from two U. S. Geological Survey (USGS) stations: near Hanover (about 170 km from the York River mouth) on the Pamunkey River and near Beulahville (about 135 km from the York River mouth) on the Mattaponi River. For bottom friction, a typical value of $r = 0.001$ m was used.

For calibration, 5 cycles of the $M_2$ tidal period with 10 cycles of spinup were simulated. The comparison of model results with tabulated mean tidal ranges indicates a satisfactory agreement (Fig. 3-4).

3-5. Salinity Verification

To check the modeled salinity distribution, the measured salinity distributions given in Chapter 2 were used. Details regarding measurement of the downstream salinity boundary condition are explained along with the TSS concentration downstream boundary condition later in this chapter.

For the open boundary forcing, the HEM-3D started with a cold start and forced by real-time tidal records from the NOAA-VIMS tidal station (NOAA gage 8637624).
The surface elevation was reduced by 10%. Because of the amplification effect (Fig. 3-4), the tidal range at NOAA-VIMS station is about 10% larger than that at the York River mouth, which is model open boundary. For spinup, the HEM-3D was run for seven days in barotrophic mode without salinity calculation. These seven days were sufficient for HEM-3D to achieve an equilibrium state for its tidal simulation as shown by the tidal calibration.

The baroclinic transport was then activated at the end of this seven-day barotrophic mode spinup period, and the HEM-3D model run was continued for another 10 days without changing the salinity boundary condition (Fig. 3-5) to allow stabilization. At the onset of simulating salinity transport, the initial salinities for all water cells were specified according to the measured salinity distribution obtained from the first slack water survey. Linear interpretation in the along the channel direction was used to obtain salinity information for all the along-channel cells that are the deepest cell in the corresponding cross sections. The initial salinity for all other cells in a given cross-section were then estimated assuming a horizontally uniform distribution of salinity across the channel.

The downstream boundary, the vertical distributions of salinity (Fig. 3-5) and suspended sediment concentration were collected during two slackwater survey periods. There are two possible approaches for obtaining the required downstream boundary conditions of salinity and TSS concentration. It was originally proposed to deploy S4 current meters with an OBS sensor at a location near Station YR01 (Fig. 2-1). This approach would have provided continuous records of salinity and TSS at certain elevations of the mouth. However, this approach could not obtain the important near-
bottom TSS concentration and salinity information in the deep channel because S4
current meters could not be placed in the channel. Considering that the near-bottom TSS
concentration and salinity information in the deep channel are more critical for better
simulations of salinity and TSS concentration distribution and for a better understanding
of the source for TSS concentration, the other approach was used, and details are
provided next.

Station YR01 was chosen to provide the required downstream boundary
conditions. Every two or three days, salinity and TSS profiles in the channel at the river
mouth (Station YR01) were measured either at a high slack tide or a low slack tide,
during the two one-month survey periods. For other places within this cross section,
salinity and TSS profiles were assumed to be the same as those measured at the same
elevation in the channel.

At the downstream boundary, the actual salinity boundary condition may change
with time. During flood tide, higher salinity from further downstream may come to this
boundary. Similarly, lower salinity from the upstream may come to this location during
ebb tide. Fortunately, the observed change of salinity at the boundary did not vary
largely, e.g., 3-5 ppt (Fig. 3-5a), and, thus, the error caused by this inaccuracy in the
boundary condition was limited.

For the first set of slack water surveys, the fourth survey was about one month
after the other three surveys due to the VIMS closure at the end of year holidays. Thus,
there are insufficient open boundary condition data available between December 11, 2001
and January 7, 2002. For this reason, the comparisons between calculated and measured
salinity distribution were only made for the first three surveys (Figs. 3-6 to 3-8).
For the second set of slackwater surveys, a similar spinup scheme with the boundary condition for the first set and a similar process of implementing initial conditions were used. During this second period, the salinity difference between the bottom and surface of the water column at the downstream boundary was also small, e.g., around 2-3 ppt (Fig. 3-5b). The comparisons to the four slack water survey results are given in Figs. 3-9 to 3-12.

Overall, the model reproduced the observed salinity very well for both sets of slackwater surveys.

3-6. Sediment Transport Model and Boundary Conditions

The existing sediment transport module in HEM-3D supports a single size class of sediment, and it is coupled with the hydrodynamic model (i.e., EFDC) with the same time step. The governing equation for sediment mass conservation is:

\[
\frac{\partial C}{\partial t} + \frac{\partial C_u}{\partial x} + \frac{\partial C_v}{\partial y} + \frac{\partial (w - w_s)}{\partial z} =
\]

\[
\frac{\partial}{\partial x} \left( k_h \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left( k_h \frac{\partial C}{\partial y} \right) + \frac{\partial}{\partial z} \left( k_v \frac{\partial C}{\partial z} \right)
\]

where \( C \) is the TSS concentration, \( t \) is time, \( x \) and \( y \) are the horizontal coordinates, \( z \) is the vertical coordinate, and \( u, v, \) and \( w \) are the three flow velocity components in \( x, y, \) and \( z \) directions, respectively. The settling velocity for suspended sediment is \( w_s \) and \( k_h \) and \( k_v \) are the horizontal and vertical turbulent diffusion coefficients, respectively.

Initial conditions and boundary conditions are required to get appropriate results from Eq. 3-1. It was assumed that the amount of water, salt, and sediment that result from precipitation over the York River water surface is negligible. This was a reasonable assumption because the precipitation records at VIMS showed that there was no rainfall.
during the two periods of slackwater surveys. Recent study (Scully et al., 2005) demonstrated that the along-channel wind plays a dominant role in governing the estuarine flow and the corresponding increase or decrease in vertical density stratification. However, the wind shear stresses that acted on the water surface were also assumed negligible because wind records at VIMS between November 2001 and April 2002 indicated that the averaged wind speed was not strong, about 3.8-5.5 m/s including all possible wind directions (Maa, 2003). Limited observations from a monitoring station established at VIMS suggested that the sediment input from side boundaries was relatively small, and thus was ignored in this study (Maa, 2003).

The remaining boundary conditions for suspended sediments are at: (1) upstream, (2) downstream, (3) surface, and (4) bottom boundaries. In this chapter the first three boundary conditions will be described and in Chapter 5 the last one will be described.

Although the freshwater discharge boundary condition is available (see sec. 3.4), unfortunately, at the fall line or landward limit of tidal influence of the Pamunkey and Mattaponi Rivers there was not enough data of TSS concentration collected during the periods of simulation to formulate the boundary condition directly. In this study, therefore, the formulation from Lin (2001) was used to calculate TSS concentrations for the two periods of slackwater surveys. Lin (2001) used 15 years (1979-1994) of data for suspended sediment concentration and freshwater discharge at the two USGS stations to work out the "best fit" coefficients, based on the seven-parameter equation given by Cohn et al. (1992), to simulate TSS influx to the York River. Because both periods of slackwater surveys had low freshwater discharge, these two tributaries did not provide
significant sediment input to the York River, with concentrations less than 10 mg/L (Fig. 3-13).

The TSS downstream boundary condition was obtained similar to that for the salinity downstream boundary condition (see Fig. 3-14). Notice that TSS concentration boundary conditions were specified in a manner similar to salinity (i.e., assuming the initial TSS distribution is the same as the first measurement). The simulation of TSS was also activated at the end of 7 days after the model started and was run another 10 days to reach stability. Similar to that for salinity simulation, amplitudes for TSS variations could be assumed. Fortunately, the amplitude was small, and thus, even assuming a zero amplitude, reasonably good results were obtained (see later in Chapter 6).

In general, the York River mouth was not a significant source of TSS during the simulations because of the small TSS concentration during the two observation periods. The time series of the TSS profiles showed that TSS concentration was low near the mouth (around 10 to 30 mg/L), and the change of TSS concentration was also small.

For the surface boundary condition, no sediment flux is allowed, and thus,

$$w_i C + k_v \frac{\partial C}{\partial z} = 0$$

(3-2)

3-7. Summary

The tidal range calibration with $M_2$ forcing was successful with the new fine resolution curvilinear-orthogonal grid (Fig. 3-4).

In general, the simulated results successfully indicated that salty water intruded into the two upstream branches (i.e., Pamunkey and Mattaponi Rivers), but with a slightly lower than observed salinity in the two upstream rivers. The maximum salinity
intrusion distances for all the 7 cases simulated matched the measurements. The most significant differences are shown in Figs. 3-7 and 3-12, for which the measured gradients of salinity were large near West Point (approximately between 50 and 60 km from the York River mouth). Near West Point, the navigation channel is narrow, and most importantly, there is no high resolution/accurate bathymetry grid available for the two upstream rivers. Although there was a significant improvement in the resolution and accuracy of bathymetry for the main section of the York River, there was only a small improvement for the Pamunkey and Mattaponi Rivers, which are upstream from West Point. This may cause the disagreement between observed and modeled gradients of salinity near West Point. Nevertheless, the above salinity simulation results are sufficiently accurate to warrant the simulation of suspended sediment transport.
Fig. 3-1. The Old Grid With Relatively Coarse Grid Resolution Changes the Continuous Channels to a Discontinued Channel. (a) Near West Point and (b) near Clay Bank.
Fig. 3-2. Parts of a Finer Curvilinear-Orthogonal Grid. (a) Near Clay Bank and Gloucester Point and (b) near West Point. Lines are cross-sectional locations where the comparisons of boat survey and bathymetric grid were made (see Fig. 3-3).
Fig. 3-3. Comparisons of the Model Depth (Solid Line) and Surveyed Depth (Dashed Line). (a) York River mouth, (b) near Clay Bank, and (c) near West Point.
Fig. 3-4. Comparison of Tidal Range From Model Output, VIMS Gauge Data, and NOAA Tide Table Data at Available York River Stations.
Fig. 3-5. Salinity Boundary Conditions Specified at the York River Mouth. (a) Winter season (2001) and (b) spring season (2002). Salinities at the bottom layer and at the surface layer are represented by solid and dashed lines, respectively. Measured at near the bottom and at near the surface salinity are marked with circle and cross, respectively.
Fig. 3-6. Salt Intrusion Simulation on November 29, 2001. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-7. Salt Intrusion Simulation on December 5, 2001. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-8. Salt Intrusion Simulation on December 10, 2001. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-9. Salt Intrusion Simulation on March 19, 2002. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-10. Salt Intrusion Simulation on March 25, 2002. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-11. Salt Intrusion Simulation on April 2, 2002. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-12. Salt Intrusion Simulation on April 11, 2002. (a) Measured and (b) modeled distributions (unit: ppt).
Fig. 3-13. TSS concentrations introduced at the two Upstream Ends of the York River system. Solid line is the Pamunkey River and dash line is the Mattaponi River. (a) Winter season in 2001 and (b) spring season in 2002.
Fig. 3-14. Time Series Contours of the TSS Concentration Boundary Conditions Specified at the York River mouth (unit: mg/L). (a) Winter Season in 2001 and (b) spring season in 2002.
CHAPTER IV
SETTLING VELOCITY

4-1. Introduction

For the simulation of suspended cohesive sediments transport, the settling velocity is one of the most important parameters (Dyer, 1986; Winterwerp and Kesteren, 2004). For primary particles, the Stokes’ formula can be used to estimate the settling velocity. Unfortunately, suspended cohesive sediments rarely exist in primary particle forms. Most likely, they exist as sediment flocs with a big range of floc density and size. Because the ambient environmental variables (salinity, TSS conc. and turbulence) will affect the formation of floc size and density significantly, it is difficult to predict the settling velocity. Even after decades of studies, it still remains as a major obstacle in modeling of sediment transport.

Various methods for the measurement of settling velocity for cohesive sediment will be briefly introduced. Settling velocities estimated from two sets of laboratory measurements using the Owen Tube method (OT) and from field measurements using the Acoustic Doppler Velocimeter (ADV) method follow.

4-2. Settling Velocity in Sediment Transport

The definition of the settling velocity (or fall velocity, or terminal velocity) of a
sediment particle is the rate at which the sediment settles in still fluid. For non-cohesive sediments, the settling velocity can be described as a function of grain size, grain shape, density of the grain, and the viscosity and density of the fluid (Hallermeier, 1981). For cohesive sediments, still more variables such as turbulence, salinity, and the sediment concentration are involved because these extra factors will affect the formation of flocs (Leussen, 1988). The particle size distribution, and its median size, may vary largely (sometimes by orders of magnitude) in time and space as a result of flocculation and sorting processes (Winterwerp and Kestenren, 2004).

To simulate the suspended sediment flux, one needs to calculate profiles of sediment concentration and velocity. The settling velocity, turbulent diffusion and erosion rate control the vertical profile of the sediment concentration. Equation 4-1 is the well-known Rouse Equation that provides a useful distribution of suspended sediment concentration in steady flows.

\[
\frac{C_z}{C_a} = \left( \frac{h-z}{z} \right)^{\frac{ws}{ws} \sqrt{ka}}
\]

where \( C_z \) is the concentration at a height \( z \), \( C_a \) is the reference concentration at the elevation \( a \) that is close to the bottom, \( h \) is water depth, \( ws \) is the settling velocity, \( \kappa \) is the von Karman’s constant, and \( u^* \) is the shear velocity. The parameter, \( \beta \), is a constant of proportionality between the eddy viscosity for momentum \( (K_m) \) to the eddy diffusivity for sediment \( (K_s = \beta K_m) \) and a parabolic shape for the eddy viscosity profile is assumed.

Assuming the eddy diffusivity to be known, the distribution of suspended sediment concentration is determined by \( ws \) and \( C_a \). In general, as the settling velocity decreases, the concentration profile becomes more uniform throughout the water column (Fig. 4-1a). Fig. 4-1b demonstrates that various profiles can be obtained by a different combination of
w_s (0.0005, 0.002, and 0.004 m/s) and C_a (200, 300, and 400 mg/L). Notice that different combinations of w_s and C_a can produce similar profiles (see three boxes in Fig 4-1b) within a certain range of the water column. In box I, the highest settling velocity (0.004 m/s) with the middle value of C_a (300 mg/L) and the middle value of settling velocity (0.002 m/s) with the lowest value of C_a (200 mg/L) make quite similar profiles from the surface to mid-depth. In box II, the middle value of settling velocity (0.002 m/s) with the highest value of C_a (400 mg/L) and the lowest value of settling velocity (0.0005 m/s) with the middle value of C_a (300 mg/L) also make quite similar profiles from the surface to mid-depth. In box III, three profiles with different combinations of settling velocity and the reference concentration are similar.

In numerical simulations, it is seldom possible to have high quality data sets for both the settling velocity and erosion/deposition rates. It is common that most of these properties are used as calibration parameters. Therefore, it is possible that certain combinations of settling velocity and erosion/deposition rates can produce a similar measured suspended sediment concentration profile. For instance, the highest settling velocity (0.004 m/s) with the highest value of C_a (400 mg/L) and the middle value of settling velocity (0.002 m/s) with the middle value of C_a (300 mg/L) make quite similar profiles compared to an arbitrarily measured concentration profile (the thick solid line in Fig. 4-1b) from the surface to 10% of the water depth above the bottom. Unless one of parameters is measured directly or one has solid confidence for its use, there are many possible combinations of these two parameters that may be misused.
4-2-1. Settling Velocity for Non-cohesive Sediments

For non-cohesive sediment, there is little interaction among sediment grains except when the suspended sediment concentration reaches a level of 10 g/L or more. For this reason, each sediment grain can be treated as if there is only one grain in the water.

The terminal settling velocity, $w_s$, of a granular sediment particle is a function of grain size, $D$, the kinematic viscosity of water, $\nu$, and the relative grain density, $\gamma' = (\rho_s - \rho)/\rho$, where $\rho_s$ and $\rho$ are solid mass density and water mass density, respectively. Hallermeier (1981) presented three universal equations to determine $w_s$ based on the Archimedes buoyancy index, $A$, which is defined as $A = \gamma'gD^3/\nu^2$, where $g$ is the gravitational acceleration. Later Ahrens (2000) merged the three equations into one for convenient use, and later Chang and Liou (2001) further improved the formulation for a better use of the one equation for $w_s$ (Eq. 4-2)

$$w_s = \frac{\nu}{D} \frac{aA^n}{18(1 + aA^{n-1})} \quad (4-2)$$

where $a$ and $n$ are two constants and they suggested $a = 30.22$ and $n = 0.463$ for general use. Equation 4-2 reduces to the Stokes' (1851) fall velocity ($i.e., w_s = \gamma'gD^2/18\nu$) for a small sphere with the Reynolds number, $R = w_sD/\nu$, less than one.

For a small granular sediment particle ($i.e., D < 0.2 \text{ mm}$), it only takes a few millimeters of downward motion to approach the terminal settling velocity. Considering that the minimum water depth in most numerical models for estuarine flow is usually on the order of 0.5 m, model depths are usually much larger than that required to approach the terminal settling velocity. For this reason, only the terminal settling velocity is considered.
4-2-2. Settling Velocity for Cohesive Sediment

Because of its crystal structure, a primary particle of cohesive sediments has a large ratio of surface area to grain volume. This means the shape of a primary particle of cohesive sediment looks like a piece of paper or a book (Dyer, 1986). The alignment of the crystal structure also brings negative charges to the large flat surfaces and leaves positive charges on the edges. Thus, these primary particles will attract water to form moving clusters or attract other primary particles to form flocs. It is also possible for primary particles to attract organic matter (if available) and form hybrid components.

Before the fresh sediment-laden water (i.e., no salt) meets seawater, the repulsive electric force on a primary particle surface is much stronger than the attractive force (mainly the Van der Waal’s force). For this reason, cohesive sediments are most likely to form face-to-edge flocs (also called non-salt flocculation). Thus, the void ratio (the ratio of empty space to solid space) is large. After the fresh sediment-laden water meets seawater (i.e., with salt), sodium ions in seawater will replace the attached water layer and depress the repulsive force. Thus, cohesive sediments can now form face-to-face flocs (also called salt-type flocculation) when there are sodium ions around. For this reason, the void ratio of sediment flocs in salt water is relatively small and the floc density is relatively large when compared with those of sediment flocs in freshwater. For example, in laboratory experiments Burban et al. (1989) found that the mean floc size of Lake Erie sediments was larger in freshwater than in sea water (McAnally, 1999).

The wide range of sizes (microns to mms), densities, and fragility characteristic of cohesive sediments makes measurement of the settling velocity of cohesive sediment difficult. Turbulence, salinity, and TSS concentration are the three major factors that
affect the formation of flocs. Turbulence can speed up the formation of flocs (if the
turbulence is weak) or break flocs (if the turbulence is strong). Salinity can affect the
structure of flocs as described above. TSS concentration indicates the abundance or
availability of sediment to form flocs. In other words, sediment in the water is a
necessary condition to form flocs. In summary, the amount of cohesive sediment, the
ambient turbulence, and the existence of salt will determine the floc size and density, and
thus, the settling velocity.

4-3. Measurements of the Settling Velocity for Cohesive Sediment

The best way to obtain a cohesive sediment settling velocity is to carry out *in-situ*
measurement without any disturbance of the floc formation. In reality, however, a
perfect approach for measuring the settling velocity for cohesive sediments does not exist
yet.

Decades of studies on the developments of *in-situ* instrument techniques for the
settling velocity are well summarized in a review paper (Eisma *et al.*, 1997). In their
paper, 17 instruments were classified into 5 categories: (1) Bottom Withdrawal Tubes, (2)
Pipette Withdrawal Tubes, (3) Remote and Automated Instruments, (4) *In-situ* Video
System, and (5) miscellaneous techniques.

Both Bottom Withdrawal Tubes (BWT) and Pipette Withdrawal Tubes (PWT)
have a cylindrical shape with open ends. They are lowered to the sampling depth in a
horizontal position, and after a sufficient time for flushing with ambient water, the valves
at both ends are closed. Once the tube is lifted onto the research vessel, it is set to a
vertical position and the sampling time starts to be counted. The BWT method takes
samples from the bottom at pre-selected times, but the PWT method takes samples at pre-selected levels in the settling tube (Eisma et al., 1997). The median settling velocity is determined from a cumulative weighted curve obtained from the dry weight of the samples. The BWT method was implemented by Owen (1976), and it is still one of the most popular methods. Hereafter this method is called the Owen tube method.

Remote and Automated Instruments also use a similar principle, but use optical sensors to measure sediment concentration continuously after both ends of the tube are closed. Most of them are mounted on a tripod or an instrument frame, and a settling velocity histogram is calculated from a concentration histogram.

In-situ video (or camera) systems (e.g., Knowles and Wells (1998)) observe the flocs settling from a place inside an underwater housing. Through image processing techniques, the flocs size and the settling velocity distribution are determined.

In-situ measurements with SCUBA divers, sediment traps, and vertical profiles of flow velocity and suspended sediment concentration were classified by Eisma et al. (1997) as miscellaneous techniques.

Although various types of instruments for measuring settling velocity have been developed, there is no instrument that is free from shortcomings. The BWT method has disadvantages in that the procedure is relatively time consuming and the sediment particles may stick to the tube and not fall (Van Rijn and Nienhuis, 1985). Both the BWT and PWT methods totally block the ambient turbulence during sampling intervals. Although the thermally induced vertical water circulation inside the tubes may be minimized by using two tubes (inner and outer tubes with the gap filled with ambient water to minimize the temperature difference between the water sample and the air that
causes the thermal circulation), it is impossible not to have some disturbance while handling the tube, closing the valves and lifting the tube to the vessel. For these reasons, it is difficult to maintain consistent data quality. The third and fourth methods are relatively free from the disturbances mentioned above, but they still block the ambient turbulence for as long as the flocs are placed inside of the sampling area, usually a tube. This may cause an unknown effect on flocculation inside the instruments. Optical sensors are nondisruptive devices, but they are relatively sensitive to the particle characteristics (Maa et al., 1992). Video/camera systems have a weakness in their resolution in determining the particle size. If a system does not have enough resolution to measure small particles (a few microns), then the median settling velocity may be overestimated by shifting the settling velocity distribution.

To properly address the settling velocity of cohesive sediment, the following three criteria must be satisfied: (1) The ambient turbulence should not be blocked out, (2) the disturbance should be minimized, and (3) the sensitivity to the particle characteristics, such as shape, density, and concentration, should be minimized.

The recent development of the Acoustic Dopper Velocimeter (ADV) provides an attractive technology for measuring instantaneous velocities in laboratories and in the field because it does not require calibration and is a non-intrusive measurement device (Gratiot et al., 2000). The ADV is relatively insensitive to grain size for fine grained cohesive sediment (Fugate and Friedrichs, 2002) and can be operated at a high TSS concentrations (up to 100 g/L) with negligible scattered echoes (Gratiot et al., 2000). Also, an ADV is an appropriate tool for measuring low Reynolds turbulence. These advantages allow the use of an ADV to estimate settling velocity indirectly with
minimum disturbance of ambient turbulence and flocs formation. Assuming that vertical velocity is always zero, Fugate and Friedrichs (2002) reported that a comparison of the local change, the advection term, the settling term, and the diffusion term in the vertical transport equation of suspended sediment mass suggests that a balance between the settling and diffusion terms is a good first order approximation for their study sites (York River and Cherrystone site in Chesapeake Bay). This finding leads to the following indirect method for estimating settling velocity.

Assuming that the sediment concentration results form a balance between gravitational settling and upward turbulent diffusion, the settling velocity can be estimated form the following equation (Sleath, 1984; Glenn and Grant, 1987; Sherwood et al., 1994; Fugate and Friedrichs, 2002; 2003).

\[-w_{sn} C_n = K d C_n / dz\]  

where \( W_{sn} \) is the settling velocity of particles in size class \( n \), \( C_n \) is the concentration of particles in size class \( n \), and \( K \) is the eddy diffusivity. Turbulent diffusion can be measured from the Reynolds diffusive flux:

\[ K d C_n / dz = -\langle w'C_n' \rangle \]  

where \( w' \) is the vertical fluid velocity fluctuation and \( C' \) is the sediment concentration fluctuation estimated from the ADV backscatter strength. Fugate and Friedrichs (2003) suggested a simple way to obtain settling velocity: dividing both sides of Eq. 4-3 by \( C \), which gives \( w_{sn} = \langle w'C_n' \rangle / C_n \). The slope of a plot of \( C_n \) vs. \( \langle w'C_n' \rangle \) then gives \( w_{sn} \). In practice, this assumes that a linear relation between \( C \) and \( w_s \) exists and that the settling component of the concentration field can reasonably be represented by a single fall velocity. Note that this method can estimate the background concentration of the
non-settling component that is the x-axis intercept of the best-fit regression (Fugate and Friedrichs, 2002). By doing this, actually two classes of floc size were represented first, a small size that never settles down and second, a certain size with a single fall velocity.

### 4-4. Settling Velocities in York River

#### 4-4-1. Owen Tube Method

The Owen tube is not a commercially available product. However, descriptions of the Owen tube and details of the method of data analysis are available (Owen, 1976). For this reason, the details of the data analysis methodology have been omitted and only a brief description of the Owen tube is given here.

The Owen tube used in this study consisted of two 1.2 m long plexi-glass tubes with inside diameters of 5.4 cm and 10 cm, respectively. These two plexi-glass tubes were placed together to have the same center, and the space between the outer and the inner tube was filled with ambient water to form a thermal isolation layer. Because the tube was built after the field work period, settling velocity measurements were carried out in the laboratory using both tap water and salt water using surficial sediments collected from the York River at Clay Bank.

A selected amount of sediment was fully mixed with fresh or salty water. Then, the sediment-water mixture was poured into the Owen tube. To further ensure a homogenous mixture, the Owen tube was shaken before it was placed in a vertical position. Fifteen samples were taken during a 3-hour experiment period with uneven time intervals. These samples were used to measure their accumulated mass for estimating the median settling velocity.
The settling velocities with tap water and salty water (about 14 ppt) are indicated by squares and circles, respectively in Fig. 4-2. The concentration range in the tap water data set was from 20 to 800 mg/L. Because the maximum TSS concentration measured in the York River during two slackwater surveys was about 300 mg/L, all of regression equations (Eq. 4-5) in Fig. 4-4 are fit to data between 10 to 400 mg/L with the following form

$$w_z = aC^b$$ (4-5)

where C is TSS concentration, a and b are constant coefficients. Equation 4-5 shows that the settling velocity increases as the TSS concentration increases.

Note that there is not much difference in the measured settling velocities between tap water \((4.64 \times 10^{-6} C^{0.375} \text{ m/s})\) and salty water \((6.0 \times 10^{-7} C^{0.8} \text{ m/s})\) in Fig. 4-2. Kwon et al. \((in press)\) used line c \((3.5 \times 10^{-5} C^{0.375} \text{ m/s})\) because settling velocity data were only available with tap water at that time. Salinity and turbulence effects on the settling velocity were not available, but it is understood that these two factors should enhance the flocculation and result in a higher settling velocity. But later, results showed that the salinity effect on settling velocity was negligible at relatively low sediment concentrations (10 to 400 mg/L). This implies that the turbulence effect on flocculation may be more dominant at relatively low concentrations.

4-4-2. Enhanced Acoustic Doppler Velocimeter (ADV) Method

Four sets of ADV data from a tripod deployed in the York River near Clay Bank during the spring season (March 15 to April 16, 2002) and the winter season (Dec. 12, 2003 to Jan. 22, 2004) were also used to estimate settling velocity. These data sets were
provided by Malcolm Scully and Carl Friedrichs. Fig. 4-3 shows $<w'c'>$ vs. $C$ plots for the four data sets. TSS concentrations were calibrated in the laboratory for the winter season data, whereas the spring season data were calibrated against an ADCP (Acoustic Doppler Current Profiler) and pump samples obtained from the field.

In this study, the best-fit regression curves for Reynolds diffusive flux, $<w'c'>$, were calculated as a non-linear function of TSS concentration (Fig. 4-3) instead of the linear approach used by Fugate and Friedrichs (2002; 2003). They determined that $w_s$ was $0.6\pm0.1$mm/s and the estimated background concentration was $29\pm12$mg/L for the York River (see a cross mark in Fig. 4-2). All of the best-fit non-linear equations (see Fig. 4-3) have a concave downward form, and the estimated settling velocities using these equations with Eqs. 4-3 and 4-4 indicate that the settling velocities increase with TSS concentration.

4-5. Results and Discussion

Laboratory Owen tube experiments showed that the settling velocity was related to the TSS concentration, highlighting the importance of sediment availability on settling velocity and the less important salinity effect. The initial work (Kwon et al., in press) assumed a higher settling velocity than the measured settling velocity with tap water at that time because of the understanding that both salinity and turbulence should have a positive effect on the settling velocity.

Although the ADV method has minimum influence on the ambient turbulence and is less sensitive to particle characteristics, the effect of ignoring the vertical advective velocity is unknown. Nevertheless, the estimated settling velocities from the ADV
method are much higher than that from the Owen tube laboratory experiments (see line d
to g in Fig. 4-4). This suggested that turbulence may have a dominant effect on settling
velocity, and the ADV method seems to be an effective and suitable way to estimate the
settling velocity in turbulence dominated environments.

The ADV method used in this study was applied to the non-linear best-fit equation
for $<w' C>$. Therefore, settling velocities were estimated as a function of TSS
concentration (i.e., not a single settling velocity). Throughout the settling velocity
sensitivity test, a modified settling velocity was used to simulate the turbidity maximums
in the York River system (details in Chapter 6).
Fig. 4-1. TSS Concentration Profiles Calculated From Rouse Equation (Eq. 4-1), Where $h = 10$ m, $a = 0.5$ m, $\beta = 1$, $\kappa = 0.41$, $\tau = 1$ Pa, 3 settling velocities (0.0005, 0.002, and 0.004 m/s), and 3 $Ca$ values (200, 300, and 400 mg/L). (a) Normalized, (b) non-normalized profiles, and an arbitrarily measured profile (gray thick solid line).
Fig. 4-2. Measured Settling Velocities in the York River. Two laboratory Owen tube experiment results are marked with squares (with tap water) and circles (with salty water), respectively. A cross represents the settling velocities from Fugate and Friedrichs (2002). Regression line a is with tap water using Owen tube method, line b is with salty water using Owen tube method, line c is used by Kwon et al. (in press), and lines from d to g are estimated settling velocities from the ADV method. Line h is a modified settling velocity and it was used in numerical simulations.
Fig. 4-3. Measured Reynolds Flux ($w'C'$) vs. TSS Concentration Plots With Regression lines Near Clay Bank in the York River. (a) and (b) were measured in spring season (Mar.-Apr., 2002) at 77 cm and 111 cm above the bed, respectively. (c) and (d) were measured in winter season (Dec., 2003-Jan., 2004) at 15 cm and 105 cm above the bed, respectively (data from Scully and Friedrichs).
CHAPTER V
BOTTOM BOUNDARY CONDITIONS

5-1. Introduction

In sediment transport modeling, the proper bottom boundary conditions are critical to get reasonable predictions, and the key processes that need to be addressed at the bottom boundary include erosion, deposition, and consolidation (Owen, 1977; Kerssens et al., 1979; Nicholson and O'Connor, 1986; Wang et al., 1990; Cancino and Neves, 1999). Direct observations of these boundary conditions, however, are rarely available. For the study of suspended sediment transport in the York River, fortunately erosion rate measurements exist for four seasons at the Clay Bank site (Maa and Kim, 2002). There are no erosion rate data for other parts of the York River system, which extends more than 100 km, nor any data for both deposition rate and consolidation rate for the sediment in this river. Therefore it is difficult, if not impossible, to accurately simulate the bottom boundary conditions for the entire river. In this study, the consolidation process is not considered. Changes in the river bed morphology is beyond the scope of this study. By simplifying the bottom boundary conditions, only erosion and deposition processes are included.

In this chapter, a simple deposition scheme is selected, and the focus is on the use of a new simple erosion rate scheme (i.e., a constant erosion rate model) based on
measured data. Details on the implementation of those two processes in the numerical modeling are explained.

5-2. Bottom Boundary Conditions

The bottom boundary condition for sediment flux is

\[ w_sC + k_v \frac{\partial C}{\partial z} = D - E \]  

(5-1)

where \( E \) is the erosion rate (the mass of sediment eroded from bottom per unit bed area per unit time) and \( D \) is the deposition rate (the mass of sediment deposited to the bottom per unit bed area per unit time).

The deposition rate \( D \) is usually defined as \( D = Pw_sC_i \), where \( C_i \) is the sediment concentration right above the bed, and \( P \) is the probability of deposition that is commonly defined as \( P = \frac{(\tau_{dc} - \tau_b)}{\tau_{dc}} \) (Krone, 1962). Therefore, deposition depends whether the bed shear stress, \( \tau_b \) is less than a critical shear stress for deposition, \( \tau_{dc} \).

\[
D = \begin{cases} 
  w_sC_i \frac{\tau_{dc} - \tau_b}{\tau_{dc}} & \text{for } \tau_{dc} > \tau_b \\
  0 & \text{for } \tau_{dc} \leq \tau_b 
\end{cases}
\]  

(5-2)

Although Eq. 5-2 is widely used in modeling, all four parameters (i.e., \( \tau_b \), \( \tau_{dc} \), \( C_i \), and \( w_s \)) are not easy to determine accurately for real environments. The difficulties in measuring settling velocity were already mentioned in Chapter 4. The existence of the critical shear stress for deposition is still debatable because there are many conflicting experimental results from various laboratory and field experiments (Sanford and Halka, 1993). \( C_i \) is also difficult to measure. In numerical modeling, the TSS concentration at the lowest cell in the water column (\( C_{low,c} \)) is usually used to represent \( C_i \). It is obvious that \( C_i \) will be
much higher that $C_{\text{low}}$ unless the vertical grid size of the lowest cell is fine enough to represent the concentration very near the bed. In reality, however, it is inevitable that $C_i > C_{\text{low}}$, and thus, an error may result from this replacement.

It is worth noting that downward flux at the center of the lowest cell, $w_i C_{\text{low}}$, will always occur because of gravity. Since the exact deposition condition is not known, an approach of not explicitly calculating the deposition is selected. The downward flux of sediment, if not balanced by the upward turbulent diffusion, will cause a net accumulation of sediment mass at the bed. Whether these net accumulated sediments become a bed or form a fluff layer depends on the $\tau_{\text{cr}}$ and other hydrodynamic conditions. It is believed that not long-before and at slack tides, these net accumulated sediments (if any) will be deposited and effectively become part of the bed.

5-3. Traditional Erosion Rate Model

Either a dimensional or a non-dimensional excess bed shear stress has typically been used to define the erosion rate, $\varepsilon$, at a particular time and location (Geyer et al., 1998; Teeter, 2001; Liu et al., 2002; Ganaoui et al., 2004).

$$E = M \left( \frac{\tau_b}{\tau_{\text{cr}}(z)} - 1 \right)^n$$

(5-3)

But Eq. 5-3 is hard to use because of the difficulty to know the vertical profile of the critical bed shear stress for erosion, $\tau_{\text{cr}}$, at a different bed level, $z$, and a different time. The $\tau_{\text{cr}}(z)$ is a function of the water content (i.e., the degree of consolidation), as well as sediment composition (Fukuda and Lick, 1980; Dyer, 1986). In Eq. 5-3, $M$ and $n$ are two empirical constants. The coefficient $M$ has the units of mass per unit area per unit time and varies from one mud to another, as well as with other factors such as temperature,
salinity, water content, and the presence of organic matter (Owen, 1975; Gularte et al., 1980; Dyer, 1986). Based on an analytical study, Parchure and Mehta (1985) found that \( n \) should be 1/2. For practical applications, however, \( n = 1 \) is often used for its simplicity (Geyer et al., 1998; Teeter, 2001; Liu et al., 2002; Ganaoui et al., 2004). Even with \( n \) specified, Eq. 5-3 still remains impractical because there is no way to know the change of \( \tau_{cr}(z) \) with time, especially in the top millimeters to centimeters of sediment beds because erosion and deposition occur alternatively and frequently. Therefore, an assumption of \( \tau_{cr}(z) \) must be frequently made for modeling purposes. This leads to the inevitable and an impractical tuning of \( M \) and \( \tau_{cr}(z) \) in the modeling of cohesive sediment transport.

5-4. A Constant Erosion Rate Model

Using the VIMS Sea Carousel for in-situ erosion tests (Maa, 1993; Maa et al., 1993; 1998; Maa and Kim, 2002), the observed erosion of cohesive sediment has always exhibited "Type 1 behavior" (see Eq. 5-4, Parchure and Mehta, 1985), which means that, for a given bed shear stress \( (\tau_b) \) that is larger than the \( \tau_{cr} \), the eroded sediment mass decreases with time because of the increase of critical bed shear stress with depth (Fig. 5-1a). This response can be modeled using

\[
E(t) = \varepsilon_0 e^{-\lambda t}
\]  

(5-4)

where \( \varepsilon_0 \) is the erosion rate at \( t = 0 \) for the given \( \tau_b \), \( E(t) \) is the erosion rate at a given elapsed time, \( t \), and \( \lambda \) is the rate constant.

The VIMS Sea Carousel field experiments indicated that the rate constant, \( \lambda \approx 0.005 \text{ s}^{-1} \), appears to be a nearly universal constant if the content of sediment has more than 30% of clay. For example, in the clay-rich Baltimore Harbor, Anacostia River near
Washington D.C., and San Diego Bay, \( \lambda \) in each case was around 0.005 s\(^{-1} \) (Maa et al., 1998; Maa and Chadwick, *in press*). The measured results for \( \lambda \) at the Clay Bank site in the York River were also around 0.005 (Fig. 5-2a). The physical meaning of \( \lambda \approx 0.005 \) s\(^{-1} \) is that \( \varepsilon(t) \approx 0 \) in 900 seconds (15 minutes). This is a condition when \( \tau_b = \tau_{cr} \). Because tidal flows (i.e., with tidally induced \( \tau_b \)) do not change significantly within 15 minutes, tidal erosion is always nearly in equilibrium. In other words, the excess bed shear stress, or the term \( [\tau_b / \tau_{cr}(z) - 1] \), is always small during a tidal accelerating phase. During tidal decelerating phases, however, \( [\tau_b / \tau_{cr}(z) - 1] \) is a negative number because \( \tau_b < \tau_{cr} \).

Therefore, it is reasonable to further simply Eq. 5-5 as follows.

\[
E(t) = \begin{cases} 
\text{constant} & \text{for tidal accelerating phases} \\
0 & \text{for other phases} 
\end{cases} \quad (5-5)
\]

The simplified bottom boundary condition is also justified from *in-situ* tripod observations of TSS time series at the Clay Bank site (Maa and Kim, 2002). In their study, the near bed (10 cm above bed) TSS concentration always increased during tidal accelerating phases and decreased at other phases (Fig. 5-3). The decrease in TSS during the decelerating phases indicated a drop, if not total stop, of upward diffusion. This implies that erosion ceased or at least was significantly reduced. These properties have also been shown in other studies (Sanford and Halka, 1993; Nakagawa, *in press*). The net downward flux increases the TSS concentration right above the bed, which is far below the lowest sensor. When \( \tau_b \) is sufficiently small, less than \( \tau_{dc} \), the accumulated sediment right above the bed will deposit and the consolidation process begins.

A conceptual bed erosion pattern that reflects the above stated process is further illustrated in Fig. 5-1. In this example, it is assumed that (1) tidal forces are exactly the
same for every tidal cycle, (2) the process starts with a slack tide, (3) the bed starts with almost a zero resistance at the bed surface, as shown in Fig. 5-1a, and (4) there is no net horizontal advective transport. At the beginning, the bed resistance ($\tau_{cr}$) profile (Fig. 5-1a) at the bed surface is near 0 because of weak consolidation of the freshly deposited materials during previous slack period. As the bed shear stress increases with time, sediments are eroded, and the bed surface elevation moves downward to a level at which the bed resistance is larger than the bed shear stress (see the lines in zoom-in box of Fig. 5-1a for various times). Note that the bed levels between $t_5$ and $t_9$ are all the same because there is no erosion during those periods because $\tau_b$ is less than $\tau_{re}$, and $\tau_b$ is still larger than $\tau_{dc}$. Only when $\tau_b$ is less than $\tau_{dc}$ does deposition start to build the bed, and the bed level will rise back to its original elevation.

The proposed constant erosion rate model consists of two parts. The first part is for the early stage of the tidal accelerating phases, when most resuspended sediments are contained in fluffy material on the top of the newly deposited sediments that just settled down during the time between $t_9$ and $t_{10}$. These freshly deposited sediments are easily redispersed even with a small bed shear stress. This re-dispersion process gradually changes to a re-suspension process (the second part) as the newly deposited bed has more and more erosion resistance as dispersion/resuspension proceeds. The erosion stops when $\tau_b < \tau_{cr}(z)$. Notice that the dash-dotted line in Fig. 5-1e is identical to the line in Fig. 5-1c. One possible erosion rate in time is based on the traditional erosion rate model with an unknown depth varying $\tau_{cr}$ (Eq. 5-3). Because it is difficult to obtain information for $\tau_{cr}(z)$, and since $[\tau_b/\tau_{cr}(z) - 1]$ is always small during the tidal accelerating phase, a single value for the erosion rate is reasonable approach (solid line in Fig. 5-1e). The difference
between the total amount of sediment eroded by the traditional erosion model with depth varying $\tau_{cr}(z)$ (Fig. 5-1c) and the constant erosion rate model is not significant (Fig. 5-1e), but the constant erosion rate model is much easier to use.

Unlike the above two erosion models, the most common erosion rate model in numerical models has a constant $\tau_{cr}$ in depth (Fig. 5-1d) and starts and stops eroding later. Erosion occurs whenever $\tau_b$ is larger than $\tau_{cr}$. By using $M$ and $\tau_{cr}$ as control parameters in a numerical model, it is possible to make the total amount of eroded mass (shaded area in Fig. 5-1d) close to the real amount of eroded mass (shaded area in Fig. 5-1c). In other words, it is possible that all three erosion models could produce the same total amount of eroded sediments. But an erosion rate model using a constant $\tau_{cr}$ could not explain the decrease of near-bed TSS concentration during the tidal decelerating phases (Fig. 5-3).

During the early stage of erosion for the constant erosion rate model, when the eroded sediment mass is from the fluffy material, it would be better to know the precise amount of fluffy mud relative to the newly deposited mud. But practically, this is beyond the currently available technology. Even if it may be possible to calculate the sediment mass using a numerical model, it may not be possible to verify this because of the small thickness of the active mud layer, which is only a few mm. Therefore, it is highly recommended that techniques be developed for determining the properties of the fluffy layer and very near bed sediment properties in general.

5-5. Implementation of a Constant Erosion Rate Model

Although the constant erosion rate model is relatively easy to apply, “the constant” in this model still has the spatial and temporal variability. This is because the two
controlling factors, bed shear stresses and erosion resistance of the sediments, vary in space and time. Details of how to handle this variability will be explained later.

One of the difficulties faced in this study was how to set a constant erosion rate for each bed cell because there was only one in-situ measurement site (near Clay Bank) in the York River (Fig. 5-4). The VIMS Sea Carousel erosion experiments were carried out in the secondary channel near Clay Bank during 1995 (Maa and Kim, 2002). This site had a water depth of 5 m and was located on the south side of the main channel. In their erosion rate experiment design, the duration of each $\tau_b$ was 25 minutes and a large and unequal $\Delta \tau_b$ was applied. Thus, the difference between any two consecutive bed shear stresses represented the excess bed shear stress (Maa and Kim, 2002) and four different relationships between the erosion rate and the excess bed shear stress were determined.

If an excess bed shear stress is given or estimated, then a possible range of erosion rates is available from Fig. 5-2b. Therefore, the first step is to determine the constant erosion rate at the in-situ measurement site, hereafter referred to as the Reference Constant Erosion Rate (RCER).

5-5-1. RCER

To determine a RCER using Fig. 5-2b, the excess shear stress near Clay Bank was required. In this study the excess bed shear stress near Clay Bank was calculated using the maximum bed shear stress, $\tau_{bmax} = 0.8$ Pa, from the model simulation with only $M_2$ tide. This is because the $M_2$ tide is the dominant tide in the York River. It represents 89% of the total tidal energy (Sisson et al., 1997). Thus, the true maximum bed shear stress and duration of accelerating phases only vary slightly for the conditions simulated.
At this time, the possible variations on $\tau_{cr}$ ($z$) for different beds and the change of $\tau_b$ due to flood-ebb and spring-neap tides are not included.

It was found that the duration of a tidal accelerating phase was about 4 hours at this site (Maa and Kim, 2002), and the maximum bed shear stress must occur over these 4 hours. Assuming (1) $\tau_b$ increases linearly, (2) that excess bed shear stress, $\tau_{ex}$, is distributed uniformly in time, and (3) that time scale for $\tau_{cr}$ to reach $\tau_b$ is about 20 minutes, then, $\tau_{ex}$ can be estimated as $0.8 \text{ Pa} / (240 \text{ min} / 20 \text{ min}) \approx 0.0667 \text{ Pa}$ (dotted line in Fig. 5-2b). Note that the estimated $\tau_{ex}$ could vary because of using the model calculated $\tau_b$ and rough estimation of time to reach steady state (15-20 minutes). But even considering these minor errors, the possible range of RCER was large (see gray bar in Fig. 5-2b).

Using this excess bed shear stress, the erosion rate at the Clay Bank site, i.e., the RCER, was shown to vary approximately 0.02-0.7 g/m$^2$/s (Fig. 5-2b). Of course, this is an indication of a significant change with season and has to be correlated with local bed conditions. However, the bed condition was assumed the same for this study at first because the two simulation periods were both during an extremely low freshwater discharge condition. More discussion will be given regarding this later.

The simulation results given in Chapter 6 were based on a selected RCER of 0.016 g/m$^2$/s (marked as a star in Fig. 5-2b). It is at the lower end of the measured range of erosion rates. However, it is possible because the two simulation periods had such low freshwater discharge from the two tributaries. Fig. 5-5 shows that average freshwater discharge rates for one week prior to three of four erosion rate experiments at Clay Bank in 1995 were much higher than those of the three slackwater surveys in 2001. Moreover,
considering the fact that low freshwater discharge persisted for almost 5 months (July-Nov.) before the slackwater survey in 2001, the erosion rate could have a lower value than the measured range of erosion rates in 1995.

5-5-2. A Constant Erosion Rate Model

Once a RCER was obtained, the next step was to find the constant erosion rate for the entire model domain. To do this, it was necessary to assume that the entire model domain had the same sediment bed condition. This may not be a reasonable assumption, but it was needed at least temporarily. In this study the ratio of the maximum bed shear stress of each cell, $\tau_{b_{\text{max}}}$, to the maximum bed shear stress at the reference site, $\tau_{R_{b_{\text{max}}}}$, was used in two ways. In the first attempt, eight categories of $\tau_{b_{\text{max}}}$ were established with an equal interval of $\tau_{b_{\text{max}}}$, and the constant erosion rate for each cell was prorated according to each cell’s category (Kwon et al., in press). Figure 5-4 shows the maximum bed shear stress distribution in the York River with eight categories. Notice that in the deep channel at Gloucester Point, between Gloucester Point and West Point, and in the Pamunkey River, the $\tau_{b_{\text{max}}}$ was large. On the other hand, $\tau_{b_{\text{max}}}$ was small in shallow areas and downstream from Gloucester Point. The eight different categories of maximum bed shear stress imply eight different constant erosion rates, and each constant erosion rate is proportional to the rate of $(\tau_{b_{\text{max}}}/\tau_{R_{b_{\text{max}}}})^c$, where $c$ stands for a category (Fig. 5-4). When the power index $k$ was selected as 2, model results showed better agreement (Kwon et al., in press). Details about the exponent $k$ will be described later.

The other method was to extend this approach by using the ratio of the maximum bed shear stresses, $\text{RCER}^*[\tau_{b_{\text{max}}}(i,j)/\tau_{R_{b_{\text{max}}}}]^k$, directly at every cell where $i$ and $j$
represent each cell's indices. It was understood that since only one bed condition was assumed for the entire model domain, at those places where \( \tau_{b_{\text{max}}} \) was larger than \( \tau_{R_{b_{\text{max}}}} \) (i.e., 0.8 Pa), the erosion rate was larger than the RCER (0.016 g/m²/s), and similarly, the less the \( \tau_{b_{\text{max}}} \), the less the erosion rate. The difference was dependent on the bed condition of the simulation period. In this attempt, the constant erosion rate model predictions were better when \( k \) was set equal to 2.5. All four measured erosion rate results showed a non-linear response of the erosion rate to the excess bed shear stress (Fig. 5-2b). Therefore, \( k=2 \) and \( k=2.5 \) in the constant erosion rate model reflect a non-linear response of the erosion rate to the excess bed shear stress.

It should be noted that, although all four measured erosion rate experiments showed different erosion rates for a given excess bed shear stress, these four regression lines seem to merge into one point (Fig. 5-2b). This means that the near surface sediment has a large range of erosion resistance because of changing hydro and sediment conditions. However, the bed resistance becomes close to each other after surficial sediments are removed. In other words, the sediment erosion resistance may have the same value below a certain depth (Fig. 5-6). Assuming the bed condition is the same everywhere, and using the pivoting point as a reference, the exponent \( k \) can be estimated \( (k=4.2) \). This is different from the selection of \( k=2 \) or 2.5. At this point, it is difficult to say which selection reflects the real bed condition because of the lack of data to verify. Nevertheless, those attempts were explored to check on how to apply the constant erosion rate with the limited erosion rate data. It is clear that more erosion rate data, especially along the river, would help to better establish this simple erosion scheme. More detailed model results are presented in Chapter 6.
5-5-3. The Effect of Freshwater Discharge on the Bed Condition

In the above discussion, the constant erosion rate approach mainly considered variations associated with hydrodynamics. However, there is another important factor, the bed condition itself, \( \tau_{cr}(z) \), that may also have a significant influence on erosion. This is implied by the large range of observed erosion rate for a given excess bed shear stress (see the gray arrow in Fig. 5-2b).

The bed condition also changes with time (Fig. 5-2b and Fig. 5-6), especially when there are newly introduced sediments from storm-induced freshwater discharge. For example, the high erosion rate for May 10th shown in Fig. 5-2b may have been the results of a significant storm event eight days before the date of the erosion test. This may have caused newly deposited materials to be relatively abundant during that time period at Clay Bank (Maa and Kim, 2002). With time, the consolidation process gradually changes the easily erodible sediment into less erodible sediment, and the erosion rate decreases back to normal conditions. Since the new sediment input from the upstream side at some locations is likely to be proportional to the freshwater discharge, it may be possible to assume in some cases that the change of the erosion rate is simply proportional to the freshwater discharge. Thus, the constant erosion rate model could be tuned as a function of the freshwater discharge to incorporate the changes associated in bed materials (e.g., \( T(x,y,t) \) in Eq. 5-6). Notice that two simulation periods both had extremely dry conditions. Therefore, in this study the change of bed condition in time due to the freshwater discharge was still ignored. However, there should be a time delay
in this process for each section of a long river. In other words, the freshwater discharge may also have an effect on the spatial variability.

With regards to the spatial variability of the erosion rate, another factor needs to be considered. The erosion rate may systematically vary with location. For example, Maa et al. (1998) pointed out a clear spatial variability in the erosion rate in Baltimore Harbor. The rate slowly increased from the outer harbor toward the inner harbor. An analogy can be applied to the York River, because upstream, where the turbidity maximum is located, the sediment bed is likely to be easier to be eroded. Thus, it is assumed that the erosion rate also increases toward the upstream (Fig. 5-6). Thus, an additional function was used to reflect the spatial variability of the bed condition. Additionally, other effects such as bioturbation, sediment composition, and the bottom bed type could be also included in this function. Because there are no data to better constrain this function (see M(x,y,t) in Eq. 5-6), a linear function was used, where \( M(x,y,t) = m_0 \cdot x + m_1 \), \( x \) is the along channel distance (in km) calculated from the York River mouth, \( m_0 \) and \( m_1 \) are constant coefficients, 0.03 and 0.1, respectively. Therefore, near Clay Bank, where \( x = 30 \), the erosion rate was set to RCER and there was a 3% increase of erosion rate given by \( M(x,y,t) \) toward the upstream end. A comparative simulation was performed with and without this modification.

In conclusion, the final form of the constant erosion rate consists of four parts: (i) a reference constant erosion rate based on in-situ measurements, (ii) hydrodynamic effects (the ratio of maximum bed shear stress to the reference maximum bed shear stress), (iii) the spatial variability of the bed condition, and (iv) a temporal variability
associated with freshwater discharge and other time-varying factors. In other words, the following form is recommended.

\[ E(x,y,t) = \varepsilon_{\text{ref}} \times S(x,y,t) \times M(x,y,t) \times T(x,y,t) \]  

(5-6)

where \( \varepsilon_{\text{ref}} \) is the RCER, \( S(x,y,t) \) is a hydrodynamic function equal to \( \left( \tau_{\text{bmax}} / \tau_{\text{Rbmax}} \right)^k \) and \( M(x,y,t) \) is a spatial modification function that represents the spatial variability of the bed condition. Both \( S \) and \( M \) are relatively weak functions in time. \( T(x,y,t) \) is a temporal modification function that reflects the temporal variability of the bed condition and is a relatively weak function in space. \( T(x,y,t) \) has not been applied in this study yet.

One should note that present \( S(x,y,t) \) was a weak function of time because \( \tau_{\text{bmax}} \) in \( S(x,y,t) \) was not considered the spring-neap variation. The further improvement about \( S(x,y,t) \) is given in Chapter 7. With this formulation, the constant erosion rate model can be specified for practical application without embedding a more complicated bed model. The results follow in the next chapter.
Bed resistance ($\tau_{re}$)

Fig. 5-1. A Schematic Diagram for Comparing the Erosion Rate Models. (a) An arbitrary selected bed resistance profiles at different time after erosion ($\tau_{re}$) with enlargement in dash-dotted box, (b) a time series of the bed shear stress ($\tau_b$) with two horizontal lines representing the critical shear stress for erosion ($\tau_{cr}$) and the critical shear stress for deposition ($\tau_{dc}$), (c) a traditional erosion model with the depth varying $\tau_{cr}$ since the $\tau_{cr}(z)$ is not known, the dot-dashed line is a possible redispersion/erosion behavior happened between $t_0$ and $t_5$. (d) a traditional erosion model with a constant $\tau_{cr}$, and (e) a constant erosion rate model (Eq. 5-5). The hatched area within in the rectangular boxes in (e) is approximately equal the area in (c).
Fig. 5-2. VIMS Sea Carousel Measured Erosion Constants for Clay Bank, 1995: (a) $\lambda$ and (b) $\varepsilon_0$ (after Maa and Kim, 2002).
Fig. 5-3. Details of 1995 summer tripod measurements at Clay Bank, York River. (a) Tidal elevation, (b) tidal current, (c) calculated \( u^* \), and (d) suspended matter. (after Maa and Kim, 2002)
<table>
<thead>
<tr>
<th>$\tau_{b,\text{max}}$ (Pa)</th>
<th>$\varepsilon_0$ (g/m²/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt; 0.16</td>
<td>0.0006</td>
</tr>
<tr>
<td>0.16 - 0.32</td>
<td>0.0025</td>
</tr>
<tr>
<td>0.32 - 0.48</td>
<td>0.0056</td>
</tr>
<tr>
<td>0.48 - 0.64</td>
<td>0.01</td>
</tr>
<tr>
<td>0.64 - 0.8</td>
<td>0.0156</td>
</tr>
<tr>
<td>0.8 - 0.96</td>
<td>0.0225</td>
</tr>
<tr>
<td>0.96 - 1.12</td>
<td>0.0306</td>
</tr>
<tr>
<td>&gt; 1.12</td>
<td>0.04</td>
</tr>
</tbody>
</table>

**Fig. 5-4.** Distribution of HEM-3D Calculated Maximum Bed Shear Stress.
Fig. 5-5. Freshwater Discharge Rate at the Pumunkey River in (a) 1995 and (b) 2001. Averaged freshwater discharge rates of one week prior to (a) four erosion rate experiments at Clay Bank and (b) three slackwater surveys are presented.
Fig. 5-6. Variation of Bed Resistance Profiles With Time at the Same Location or at Different Locations at the Same Time. For instance, the solid and the dashed lines could represent the bed resistances during low freshwater discharge and high freshwater discharge conditions at a certain site in the York River (e.g., Clay Bank), respectively. They also could represent the bed resistances at downstream site and ETM (and/or STM and upstream site) in the York River, respectively.
CHAPTER VI
MODEL RESULTS

6-1. Introduction

The main approaches and assumptions used in this study are summarized below:

(1) The settling velocity was treated as a function of TSS concentration and different settling velocity formulae were used to address data obtained from the Owen tube measurements and the ADV method.

(2) The downward flux of suspended sediment in the water column was assumed to always exist.

(3) To incorporate a combination of erosion and deposition at the water-sediment interface, a constant erosion rate occurring only during the tidal accelerating phases was implemented.

(4) A slight change of the reference constant erosion rate (RCER=0.016 g/m²/s) is possible due to a difference between the channel and the erosion experiment site at Clay Bank. Nevertheless, the method to find the RCER was based on the model calculated maximum bed shear stress and in-situ erosion rate measurements at the Clay Bank.

(5) Two approaches for selecting the constant erosion rate for the entire estuary were tested. The first one was a relatively simple approach using 8 categorized constant
erosion rates (Kwon et al., in press), and hereafter is called the “simple categorized model.” The other (hereafter called the “Four Factors Model”) consists of four contributing factors (Eq. 5-6). The first factor is the RCER mentioned in item 4. The second factor represents a hydrodynamic condition at each cell, using the ratio of the maximum bed shear stress at each cell to the maximum bed shear stress at the reference site. The third factor implies a possible spatial variability of the bed condition. The last factor is a possible temporal modification function that has not yet been implemented.

In this chapter, results from both of the “simple categorized model” and the “Four Factor Model” are presented. For a correct comparison between model and survey results, one has to consider the following. Each slackwater survey took about one day to finish and was usually started at the York River mouth at a slack tide. After obtaining data at a station, the survey progressed to the next upstream station. The pace of the survey usually matched the tidal propagation, so measurements were usually done near slack tide at all the survey stations. Sometimes it was impossible to match with the tide, and a time lag was inevitable. The measured data were used to construct salinity and TSS concentration contours given in this study, and one must know that these are not exactly “snap shots.”

The model calculated results (i.e., water level, current velocity, salinity, TSS, etc.), however, were initially saved for the same time steps for the entire York River system. They were true “snap shots.” For this reason, a post-processing of model outputs was done to obtain results with times that matched with the survey times at each station for comparison. This approach was more accurate when compared with other alternatives such as averaging the results over one tidal cycle. The averaging process in
other approaches actually smoothes the output and is difficult to compare with the measurements because they represent two completely different conditions.

In general, the model simulated TSS concentrations in the middle of the water column were slightly higher than the observations, and the locations of ETM were also off a little, on the order of 5 to 10 km (see Fig. 6-1 for an example). For comparing the simulated and the measured surface and bottom TSS concentrations (see Fig. 6-2 for an example), an average of SSC at 2 m below surface and SSC at 2 m above the bed were used, respectively.

6-2. Model Experiments

6-2-1. Simple Categorized Model

When this model was implemented, there was only one measured settling velocity data set from the Owen tube method with tap water in the laboratory (line a in Fig. 4-4). Considering the effect of salinity and turbulence, a higher settling velocity ($3.5 \times 10^{-5} C^{0.375}$ m/s, line c in Fig. 4-4) was applied at that time with RCER=0.0225 g/m²/s.

For Dec. 5, 2001 (Fig. 6-2a), the simulated ETM was predicted quite well. The near bed TSS concentration matched at about 110 mg/L and the location also matched at about 70 km from the river mouth. But the surface TSS concentration at the ETM site was overestimated. The other peak of measured bottom TSS concentration at 85 km from the mouth may represent a newly developed plume that moved downstream. The location of STM also matched at approximately 30 km from the river mouth, but the simulated TSS concentration was overestimated both at the surface and near the bed.
For April 11, 2002 (Fig. 6-2b), the center of modeled ETM was off by about 10 km upstream. However, TSS concentrations both at the surface and near the bed were predicted quite closely compared with measurements. Near Clay Bank, the model predicted a relatively weak STM that also manifested weakly during the survey.

Using the selected approach, a relatively high suspended TSS concentration always appeared at the Clay Bank area, and that may contribute to the existence of the secondary turbidity maximum. Nevertheless, the results presented here indicate that the proposed constant erosion rate is capable of reproducing the turbidity maximum. More refinements are necessary to bring it closer to observations, and that is the purpose of next section.

6-2-2. Sensitivity Test for the Settling Velocity

After completing the modeling effort using the “simple categorized model” to simulate the turbidity maxima, more data on settling velocity became available. The effect of salinity (see Chapter 4) is not critical (at least based on the limited laboratory results), and the effect of turbulence may dominate. Nevertheless, more possible settling velocity results became available, and it was worth testing the effect of different settling velocities before exploring the modified erosion rate model. This is because both the settling velocity and erosion rate control the suspended TSS concentration throughout the water column. Three different settling velocities were tested: (1) case 1, 3.13x10^-6 C^{1.09}, line g in Fig. 4-4, (2) case 2, 3.0x10^-4 C^{0.19}, line e in Fig. 4-4, and (3) case 3, 1.80x10^-6 C^{1.35}, line h in Fig. 4-4. All the remaining parameters were kept constant.
The settling velocity in case 1 was calculated using the ADV method based on the instrument deployed at 111 cm above the bed near Clay Bank during March 2002 (data from Malcolm Scully and Carl Friedrichs). This settling velocity was tried first because it was recorded during the same period as the spring season slackwater survey. In general, this settling velocity successfully predicted both the STM and ETM but sometimes overestimated the STM about 80-300% (Fig. 6-3a and b).

Another settling velocity from the ADV method based on the instrument deployed at 105 cm above the bed near Clay Bank during winter 2004, case 2, was tested. Results using $w_s$ given by case 2 underestimated the surface TSS concentrations, particularly for near surface cases where TSS concentration was low. This is because the settling velocity in case 2 has relatively higher $w_s$ in low concentrations (see Figs 4-4 and 6-3). Therefore, the settling velocity given by case 3, a selection based on judgment, was tried. This selected $w_s$ (case 3) gave a better prediction of the STM (Fig. 6-3a), but still overestimated in some case (Fig. 6-3b). By changing the settling velocity from case 1 to case 3, the model results were improved, especially for the STM site. Thus, the settling velocity from case 3 was applied to the rest of model simulations.

6-2-3. Four Factors Models

The constant erosion rate formula (Eq. 5-6) contains three possible modification functions $S(x,y,t)$, $M(x,y,t)$, and $T(x,y,t)$. $S$ is the hydrodynamic factor. $T$ represents the temporal variability and should be a weak function of $x$ and $y$. $M$ is a function representing spatial variability within the estuary.
The model results in this section include (1) the single bed condition, (2) a variable bed condition (considering a spatial gradient along the estuary, *i.e.*, condition (1) with \( M(x,y,t) \)), and (3) a variable bed condition chosen to be closely similar to the reference site.

6-2-3-1. Single bed condition (case 4)

In this run, the hydrodynamic factor, \( S(x,y,t) = \left( \frac{\tau_{b_{\max}}}{\tau_{R_{b_{\max}}}} \right)^k \) with \( k=2.5 \), was the only factor controlling the local erosion. This means that a single bed condition (*i.e.*, data from a single measurement) with the non-linear response was assumed to be applicable for the entire model domain (case 4). As seen in Fig. 6-4, case 4 results always overestimated TSS concentration near the STM site. This means that an additional modification in the bed condition was required to reduce the errors in the STM area. Due to a lack of data on the spatial gradient of the erosion rate, \( M(x,y,t) \) was set a simple linear function of distance from the York River mouth (see next section).

6-2-3-2. Variable bed condition (case 5)

In this run, the effect of the spatial variability of the bed condition, *i.e.*, the \( M(x,y,t) \) function was added (case 5). In general, the simulation results were improved by the addition of \( M(x,y,t) \). For instance, the errors in bottom TSS concentration of STM site on Dec. 5, 2001 were reduced from about 100% (simple categorized mode, Fig.6-2a) to 45% for case 4 (single bed condition, Fig. 6-4a) and further reduced to 25% for case 5 (variable bed condition, Fig. 6-4a). Fig. 6-4 demonstrates that \( M(x,y,t) \) contributed to reducing the bottom TSS concentration at the STM site.
6-2-3-3. Variable bed condition closely similar to the reference site (case 6)

As mentioned in Chapter 5, it is possible to assume that the entire model domain has the same bed condition as that at the reference site. By assuming the erosion behavior is the same everywhere, the results in Fig. 5-2 can be used everywhere along the York River. Then the proper value of $k=4.2$ should be used if the $RCER=0.016 \, g/m^2/s$ is selected (Fig. 5-2). With the spatial modification function $M(x,y,t)$ given by case 5, the calculated near-bottom TSS concentrations in the ETM were improved, but the near-surface TSS concentrations in STM were generally not as good as case 5 (Fig. 6-5). Nevertheless, this approach is one possible way to apply the constant erosion rate model with limited measured erosion rate data.

6-3. Summary

Based on the sensitivity test for the settling velocity, a slightly different settling velocity formula gave the best results, therefore, it is applied in this study (It gave a slight larger $w_s$ at high SSC, but smaller $w_s$ at low SSC than that estimated from the ADV method).

The constant erosion rate model was successful in predicting the turbidity maxima in the York River. However, because only one erosion site measurement was available for the York River, a method was needed to apply this rate to the entire model domain. The hydrodynamic factor, $S(x,y,t)$ in Eq. 5-6 was introduced to reflect the large $\tau_b$ and the large erosion rate. It was clear that the response should be a non-linear function of the ratio of the maximum bed shear stresses based on the measured erosion rate data. Two
different k values that represent the non-linear responses were tested and showed some difference. The model simulation always showed the STM near Clay Bank because of the relatively high bed shear stresses in that area. The possible gradient of erosion rate along the estuary cannot be addressed by this factor only. A simple linear spatial variability function, $M(x,y,t)$, with distance accounting for the possible effect, was applied and improved the model results.

To enhance the constant erosion rate model, more erosion rate measurements along the river covering nearly the same time period are required. This could help to develop a reliable spatial variability function for the bed condition, $M(x,y,t)$. If this kind of measurement could be conducted at different times, then a possible temporal function $T(x,y,t)$ in Eq. 5-6 could also be estimated.
Fig. 6-1. Comparison of (a) Observed and (b) Modeled TSS Distributions (mg/L) on April 2, 2002 Using the Simple Categorized Constant Erosion Rate Model.
Fig. 6-2. Comparisons of Observed (lines with circles) and Simulated (lines without circles) TSS Concentrations at two Depths Using the Simple Categorized Model. Solid and dashed lines represent near-surface and near-bottom TSS Concentrations, respectively. (a) Second slackwater survey on Dec. 5, 2001 and (b) eighth slackwater survey on April 11, 2002.
Fig. 6-3. Sensitivity Test for the Settling Velocity (case 1: \( w_s = 3.13 \times 10^{-6} \), case 2: \( w_s = 3.0 \times 10^{-4} \), and case 3: \( w_s = 1.8 \times 10^{-6} \)). Circles and thick lines represent observed and simulated near-bottom TSS concentrations, respectively. Crosses and thin lines represent observed and simulated near-surface TSS concentrations, respectively. (a) Second slackwater survey on Dec. 5, 2001 and (b) third slackwater survey on Dec. 10, 2001.
Fig. 6-3. (continued). (c) Fifth slackwater survey on March 19, 2002 and (d) eighth slackwater survey on April 11, 2002.
Fig. 6-4. Sensitivity Test for the Spatial Variability Function, M(x,y,t). Case 5 with M(x,y,t) and case 4 without M(x,y,t). Circles and thick lines represent observed and simulated bottom TSS concentrations, respectively. Crosses and thin lines represent observed and simulated surface TSS concentrations, respectively. (a) First slackwater survey on Nov. 29, 2001 and (b) second slackwater survey on Dec. 5, 2001.
Fig. 6-4. (continued). (c) Fifth slackwater survey on March 19, 2002 and (d) seventh slackwater survey on April 2, 2002.
Fig. 6-5. Sensitivity Test for the non-linear response of the bed, exponent k. Case 5: k=2.5 and case 6: k=4.2. Circles and thick lines represent observed and simulated bottom TSS concentrations, respectively. Crosses and thin lines represent observed and simulated surface TSS concentrations, respectively. (a) Second slackwater survey on Dec. 5, 2001 and (b) third slackwater survey on Dec. 10, 2001.
Fig. 6-5. (continued). (c) Fifth slackwater survey on March 19, 2002 and (d) sixth slackwater survey on March 25, 2002.
CHAPTER VII

DISCUSSION AND CONCLUSIONS

7-1. Discussion

Accurate bathymetry with clearly represented channels is critical for any predictive estuary model, especially for salt and sediment transport simulations. Even the new fine resolution curvilinear-orthogonal grid used in this study which shows good predictions of tidal propagation and the head of salt intrusion, has a grid resolution which is not sufficient for the upstream, and thus, causes an imperfect simulation of the salinity gradient near West Point. As mentioned in Chapter 3, most improvements associated with the new grid are in the main stream, not in the Pamunkey and Mattaponi Rivers that have one cell in the lateral direction. To enhance future simulation, the model may need higher resolution in these upstream regions, especially for simulations during low freshwater discharge periods.

In this study, slackwater surveys were carried out to provide data for comparisons with model simulations. Because all water sampling was done from boats, TSS profiles do not go very close to the bed (usually about 0.5 to 1 m above the bottom). However, the vertical gradient of TSS concentration is high when close to the bed (on the scale of cm) because of continuous settling of particles. As mentioned in Chapter 4, to judge which model TSS profile is the best prediction (in other words, which values of parameters are closest to the true values) it is critical to have observed TSS profiles reach
very close to the bed.

In most estuaries, cohesive sediments seldom exist in the form of primary particle but instead exist as flocs. This means that to predict cohesive sediment transport, the floc dynamics should be accounted for. Through decades of flocs studies, our understanding of floc dynamics has improved enormously. However, it is still hard to find a good numerical model which includes the flocculation process. This is because our current understanding is based on many laboratory experiments that were mostly quite different from real field environments. It is clear that better in-situ measurements of flocs characteristics, including densities, shapes and settling velocities, are urgently needed to improve simulation of cohesive sediment transport.

The ADV method for measuring settling velocity (Chapter 4), which does not affect the ambient turbulence and is less sensitive to the particles' characteristics such as concentration and fragility, is probably the best approach for measuring settling velocity. But this method is only valid when the mean vertical velocity can also be estimated accurately or when the vertical velocity is close to zero (see Eq. 4-3 to 5). The slack tide could be a suitable condition. However, the ADV approach used in this study ignored the vertical velocity, and, thus, the measured \( w_s \) is less precise at this time. Currently estimated settling velocities from this ADV method show some differences with deployed elevation (see lines d to g in Fig. 4-4). The estimated settling velocities based on the deployed distance of 1.1 m above the bed (line g: 111 cmab in 2002 and line e: 105 cmab in 2004) were higher than those from lower elevations (line f: 77 cmab in 2002 and line d: 15 cmab in 2004). Although those data measured during the same period are not much different, the difference in the estimated settling velocities caused by different
elevations could be affected by the different vertical current velocity among other factors. More research is needed to address these shortcomings. Additionally, it may be necessary to improve the formulation of settling velocity as a function of turbulence and TSS concentration. In this study, although the settling velocity was expressed as a function of TSS concentration, the formulation still contained the unknown effects of the turbulence. By better constraining the effects of turbulence, a more realistic settling velocity could be determined.

There are very litter or good data providing in-situ measurements of the deposition rate. Moreover, there is still argument over even the existence of the critical shear stress for deposition, $\tau_{dc}$ (Sanford and Halka, 1993; Winterwerp and Kesteren, 2004). In this study, only the downward flux of suspended sediment in the water column always exist. Since the exact deposition condition is not known, an approach of not calculating the deposition amount is selected. The downward flux of sediment, if not balanced by the upward turbulent diffusion, causes a net accumulation of sediment mass above the bed. Whether these net accumulated sediments become a bed or form a fluff layer may depend on the $\tau_{dc}$ and other hydrodynamic conditions. At least, it is believed that not long before and at slack tides these net accumulated sediments (if any) will be deposited.

In the early stage of the erosion period of the proposed constant erosion rate model, the eroded sediment mass is from the fluffy material. Thus, it would be better to know the precise amount of fluffy mud in the newly deposited mud. This could directly improve the erosion rate model. But practically, this is beyond the currently available technology to verify. Even if it may be possible to calculate the sediment mass using a numerical model, it may not be possible to verify the calculation because of the small
thickness of the fluffy mud layer, which is only a few mm. Therefore, it is highly recommended that techniques be developed for determining the properties of the fluff layer and the near bed sediment.

Another improvement related to the deposition is necessary. In this study the model did not track the deposition amount. This means the model provided unlimited sediment from the bed during the tidal accelerating phase. By tracking the inventory of the bed sediment that was deposited during a previous simulation period, the model could provide more sediment for suspension where the more deposition occurs, such as in the ETM zone. This could be implemented into the spatial modification function $M(x,y,t)$.

Inasmuch as the excess bed shear stress depends on the applied bed shear stress caused by the hydrodynamic forces, and the critical bed shear stress for erosion is controlled by sediment properties, changes in the "constant" erosion rate must reflect possible changes in hydrodynamics and sediment conditions in time and space. Even though the constant erosion rate model gives a simple way to address the erosion process, the suggested "constant" erosion rate model must be able to reflect the change of hydrodynamic and the bed conditions in time and space.

In the current constant erosion rate model, the maximum bed shear stresses were calculated with $M_2$ forcing and normal averaged freshwater discharge. Fig. 7-1b shows that the near bottom TSS concentration predicted by the model did not match well with time-series observations, especially during the spring tide at near Clay Bank. To make the constant erosion rate approach more precise, i.e., to improve the hydrodynamic factor, $S(x,y,t)$, the true maximum bed shear stress and the exact duration of accelerating phase should be found for the simulation period. A simple and easy way would be to determine
an enveloped curve of the maximum bed shear stress with time (Fig. 7-1c) and modify $S(x,y,t)$ accordingly.

The spatial and temporal bed conditions are other important factors in simulations of the suspended sediment transport. It is obvious that if not much erodible bed material exists, then even with higher bed shear stress, there would not be much erosion. This also could be related to the existence or distinct fluffy layers in turbidity maximum zones including the STM. Those areas are expected to have a larger erosion rate than other places. And also, storm events and freshwater discharge could affect the bed conditions (Fig. 6-2b). More erosion rate measurements collected along the river under various conditions will help to fully address the need/justification for these modifications, which could be contained in $M(x,y,t)$ and $T(x,y,t)$ in Eq. 5-6.

The current model always predicts the secondary turbidity maximum (STM) to be near Clay Bank. This could be caused by the relatively high bed shear stresses near Clay Bank as well as other factors. For instance, secondary circulation could produce convergence or divergence at Clay Bank, and that could produce or remove the STM. In other words, detailed simulation and verification of the residual current at the STM site are necessary. Further research is needed to investigate this problem.

During the eight slackwater surveys, the ETM was always established on the downstream side of the head of salt intrusion (see Fig. 2-9, for example) except one case given in Fig. 2-10. Because all surveys were conducted during low river discharge conditions, salt intrusion distances were quite long and the ETMs were formed far upstream direction, but they may still lagged behind the head of the salt intrusion. The relationship between the head of salt intrusion and the ETM has been discussed by other
researchers (Geyer, 1993; Uncles et al., 1993; Lin, 2001). It is a complicated process involving the interactions of several factors (e.g., tide, TSS concentration, estuarine circulation, and stratification). However, in this study another possible factor was observed. Between March 19 and March 25, 2002, there was a set down of the mean water level (Fig. 7a). Due to this set down, the head of salt intrusion (2 ppt) moved about 5 km downstream (see Fig. 2-9a and Fig. 2-10a). This is because the increase in pressure gradient produced more downstream-flux from the Pamunkey River. But the ETM did not adjust immediately to the change of this hydrodynamic condition.

Also, at the upstream side of West Point, a pocket of relatively less saline water with low TSS concentration was observed. Unfortunately there are not sufficient data to verify where this plume originated. It might have come into York River from the Mattaponi River during ebb tide, and when the tide changed to flood, a part of the plume flowed into the Pamunkey River. Although in general, the freshwater discharge from the Mattaponi River is only half of the Pamunkey River, the time to reach the York River may vary with unaccounted for the local rain events. Notice that the freshwater discharge from the Pamunkey River also doubled from 5 to 10 m$^3$/s during this set down event. This increase of freshwater discharge may enhance the stratification somehow, and thus, may affect both the locations of the head of salt intrusion and the ETM. Further study is required to better understand the relationship between the head of salt intrusion and the location of the ETM in the York River system.
7-2. Conclusions

(1) A total of eight field surveys of salinity and total suspended sediment (TSS) profiles along the York River were conducted during two one-month periods (Nov.-Dec., 2001 and March-April, 2002). Because of the unexpected extremely low freshwater discharge during both those periods, the survey results in terms of salinity and TSS distributions were very similar. There were no significant sediment inputs from either the land or the ocean. Therefore, sediment resuspended from the bed within the York River system must be the source of the observed TSS.

(2) The survey results showed that the estuarine turbidity maximum (ETM) was anomalously located about 30 km upstream from West Point. In a normal hydrological year, the ETM is usually located close to West Point. A secondary turbidity maximum (STM) was also identified in the York River with a much weaker signal.

(3) A new curvilinear orthogonal bathymetric grid for the York River system was developed to clearly represent the navigation channels. Model performance for tidal propagation and the salinity intrusion were excellent using this new grid. It was found that a still higher resolution grid is necessary for a better simulation of the upstream rivers.

(4) A series of experiments on the settling velocity of York River sediment indicated that the settling velocity was related to the concentration of TSS and highlighted the importance of sediment availability. The influence of salinity on the settling velocity was negligible based on estimates using the Owen tube in relatively low TSS concentrations (10-400 mg/L). The effects of the turbulence may dominate.
(5) Using the acoustic Doppler velocimeter (ADV) method, which does not affect the ambient turbulence and is less sensitive to the particle properties, the settling velocity in the York River was estimated. The result produced a settling velocity more than 10 times higher than that of the Owen tube method and better reproduced the turbidity maxima for two slackwater simulations. The ADV method seems to be an effective and suitable way to estimate the settling velocity in turbulence dominated environments.

(6) The “constant” erosion rate model for erosion was implemented by assuming that erosion occurs only during the accelerating phases of the tide. Specifically, the Four Factor Model for constant erosion was successfully implemented to simulate suspended sediment transport for the York River system.

(7) Further improvements to the constant erosion rate model are necessary to address the spatial and the temporal variability of this “constant” erosion rate approach to reflect the variations of the bed condition.
Fig. 7-1. Time Series of Observation Data and Model Outputs. (a) Surface elevation at Gloucester Point during the spring season in 2002: The dashed line is the trend of Mean Water Level and the dates of slackwater surveys are marked with triangles. (b) The TSS concentrations of the ADV measurements (thin line with cross marks) and model prediction (thick line) at near Clay Bank. (c) The model calculated bed shear stress with an enveloped maximum bed shear stress (dashed line).


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