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상부 체사파크 만에서의 Estuarine Turbidity Maximum에 대한 수치모델 연구

A Model Study of the Estuarine Turbidity Maximum in the Upper Chesapeake Bay (US)

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ABSTRACT

A suspended sediment (SS) transport model is developed and internally linked to the CH3D-WES model. The model is based on the mass-balance equations for three size classes representative of the Upper Chesapeake Bay. The first class (3 μm) consists of fine particles that are in more or less continual suspension in water column and thus constitute the background concentration. The second class (18 μm) consists of medium particles that are alternately eroded and deposited by tidal current and the last class (65 μm) of coarse particles that stay at the bed most of the time. No interaction among size classes and no effect of the SS on the hydrodynamic field are assumed. The present model employs an empirical formulation to account for depth-limited erosion by varying the critical shear stress for erosion, $\tau_{e}$, as a function of eroded mass. A point model is employed to compare continuous deposition and mutually exclusive erosion and deposition. Based on the point model results, the present model employs for the fine class mutually exclusive erosion and deposition with small constant values for $\tau_{e}$ (0.02 Pa = $\tau_{cd}$) to assure quick erosion of deposited fine particles, without allowing erosion of consolidated, hard bed sediments, and continuous deposition with depth-limited erosion for the medium and coarse classes.

The model is applied to the Upper Chesapeake Bay to simulate annual conditions in 1996. The model is calibrated against a coefficient in the formulation of depth-limited erosion with the intensive data in Sanford et al. (2001) and verified with the Chesapeake Bay Program monitoring data. The three size classes appear to behave as they are intended to and the model gives a reasonable reproduction of the data, not only for intertidal but for intratidal variations in salinity and SS concentration. The model results for 1996 are analyzed to investigate the variations of the estuarine turbidity maximum (ETM) in intertidal and intratidal time scales.
Tidal average model results reveal that both the limit of salt intrusion and the ETM migrated up and down largely in response to the Susquehanna River discharges, with the movement of the ETM lagging behind that of the salt limit. During the high flow conditions except for the enormous flood in January, the downstream limit of the ETM was trapped between km 40-42, marked by steep longitudinal SS gradient. During the low flow conditions in June-September, a relatively weak ETM was developed and lasted a few days between km 20-28 with a period of about 30 days, which coincided with the occurrence of strong spring tides.

Analysis of the 15-day average model results for each of the low-to-moderate flow (LMF) and high flow (HF) conditions reveals that the relative positions and strengths of bottom convergence of SS, net vertical flux of SS, erosion and deposition combine to affect the intertidal variations in location and strength of the ETM. The relatively large bed shear stress during the HF condition results in net erosion in the upstream reach of the Upper Chesapeake Bay. The resuspended sediments and those from the Susquehanna River are transported downstream and much of them are settled down to form the ETM very close to the bottom convergence of SS flux at the null zone. The sediments, not trapped in the ETM zone and transported to the downstream of the ETM zone, are settled down to the lower layer due to the suppressed vertical mixing by stratification in saline water. As the freshwater discharges diminish, everything is rebounded back to upstream. Resuspension of the bed sediments deposited during the HF condition and the subsequent upstream transport results in the formation of the ETM at a farther upstream location. The sedimentation rates in the ETM zone are estimated 1.1-4.0 g cm$^{-2}$ yr$^{-1}$, depending on the freshwater discharges, which are on the same order of magnitude with, though slightly larger than, the previous estimates.
The model results exhibit large intratidal variations in the SS distribution, with their characteristics depending on the relative importance of various forcing functions, including tidal currents, gravitational circulation and stratification. Neither tidal resuspension lag nor tidal asymmetry, or any given combination of both, is not likely to explain the intratidal variations at all times and at all locations. With large intratidal variations, caution should be taken in interpreting the SS data for spatial distributions if the data had not been collected at the same tidal phases for all stations.
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1. INTRODUCTION

In the upper or middle reaches of many estuaries, the concentrations of suspended sediment (SS) are considerably higher than those either upstream in the source river or farther seaward in the estuary in spite of settling of fluvial sediments and dilution with less turbid seawater. Such feature is referred to as the estuarine turbidity maximum (ETM), one of the most distinctive features of sediment transport in estuaries (Postma, 1967; Schubel, 1968a; Nichols and Biggs, 1985; Dyer, 1986). The existence of the ETM has been observed in many estuaries. Examples include the Upper Chesapeake Bay (Schubel, 1968b, 1969a, and 1972; Nichols, 1986; Sanford et al., 2001), York River (Lin, 2001; Lin and Kuo, 2001), Rappahannock River (Nichols and Poor, 1967; Nichols, 1977; Officer and Nichols, 1980), James River (Nichols, 1972; Officer and Nichols, 1980), Patuxent River (Roberts and Pierce, 1976), Hudson River (Geyer et al., 2001), Saint Lawrence Estuary, Canada (Silverberg and Sundby, 1979; Hamblin, 1989), Ems Estuary, the Netherlands (van Leussen, 1988), Weser Estuary, Germany (Grabemann and Krause, 1989; Lang et al., 1989; Grabemann et al., 1997), Tamar Estuary, U.K. (Uncles et al., 1985; Uncles and Stephens, 1989 and 1993; Grabemann et al., 1997), and Aulne and Gironde estuaries in France (Allen et al., 1980).

The ETM encompasses a zone within which the physicochemical and compositional properties of water change rapidly from those of freshwater to those of seawater and the area is a major site for chemical and biological reactions. In particular, organics, inorganics, and heavy metals react with the cohesive silts and clays in the ETM zone (Owen, 1977). The size range of particles is narrow in the ETM zone because of a very effective sorting mechanism (Postma, 1967;
Schubel, 1969a and 1971; Officer, 1981; Jay and Musiak, 1994). If the material supplied is too fine, it passes out the estuary. If it is too coarse, it deposits further upstream of the ETM zone. The ETM zone is often the site of net deposition and rapid shoaling (Schubel, 1968b; van Leussen, 1988; Sanford et al., 2001). The high SS concentrations in the ETM zone reduce the light intensity in the water column (Champ et al., 1980) and may inhibit the growth of phytoplankton and aquatic plants (Moore et al., 1997).

1-1. Previous Studies

The formation and variation of the ETM are influenced by many processes including two-layer gravitational circulation, stratification, intratidal variations in tidal currents, tidal asymmetry, and topography. A brief description of the previous studies of the ETM is given below.

Two-layer gravitational circulation is the most commonly attributed formation mechanism of the ETM (Dyer, 1986). As a result of gravitational circulation in partially mixed or highly stratified estuaries, there is a horizontal convergence in the bottom flow at a null zone near the upper limit of salt intrusion (Dyer, 1988) and the ETM is maintained by convergence (Postma, 1967). Sediment introduced either from the river, the sea, or recycled by gravitational circulation within an estuary is effectively trapped in the null zone (Schubel, 1968a and 1968b; Schubel and Carter, 1984; Uncles and Stephens, 1993). Schubel (1968a, 1968b, and 1974) explained the ETM at the Chesapeake Bay is produced and maintained by the sediment trap produced by gravitational circulation. Festa and Hansen (1978) developed a steady-state, two-dimensional (x-z) model, with the assumption of no net deposition and erosion and of a
constant eddy viscosity and diffusivity. They supported the hypothesis in Postma (1967) that gravitational estuarine circulation is the major reason for the occurrence of the ETM. Officer (1980a and 1980b) was able to reproduce the model results of Festa and Hansen (1978) using a steady-state box model with linear algebraic equations, which assumed dominant longitudinal gravitational effects over longitudinal tidal exchange effects. Officer and Nichols (1980) applied a box model including net erosion and deposition flux with slack water data from the James and Rappahannock rivers, tributary estuaries of Chesapeake Bay, and showed that turbidity maximum is an effect related to two-layer gravitational circulation.

Stratification affects the formation of the ETM by suppressing turbulence in partially mixed or highly stratified estuaries. The sediment is kept in suspension in water column by the balance between gravitational sinking and turbulent mixing. Stratification in an estuary generally is determined by salinity and varies directly with river discharge and inversely with tidal range (Dyer, 1997). In partially mixed or highly stratified estuaries, after the homogeneous freshwater reaches the limit of salt intrusion, the water column becomes stratified. Suppression of turbulent mixing by stratification, therefore, may result in sinking of SS (Simpson et al., 1990; Lewis, 1996). Hamblin (1989) and Lang et al. (1989) showed the importance of stratification-induced inhibition of turbulence in determining the vertical distribution of SS near the ETM zone through a comparison between data and model results. Geyer (1993) developed an intertidal two-dimensional (x-z) model similar to that of Festa and Hansen (1976) with the exception that the turbulent mixing was affected by density stratification through the Munk-Anderson approximation. He showed that the suppression of turbulent mixing by stratification has much influence on the intensity of the ETM. Uncles and Stephens (1989 and 1993) demonstrated that intratidal variations in salinity
stratification contribute to the stability and turbulent mixing, and also have a strong influence on the location of the ETM in the Tamar Estuary, U.K.

Intratidal variation in magnitude and direction of tidal currents may affect the formation and variation of the ETM. The magnitude and position of the ETM have been observed to vary with respect to the phase of tidal currents. As tidal currents increase, sediment is eroded and resuspended from the bed resulting in an increase in SS concentration in water column. On the other hand, as tidal currents decrease, SS is deposited to the bed resulting in a decrease in water column concentration. The change of bottom SS concentration with tidal currents has been observed in many estuaries, including the Upper Chesapeake Bay (Schubel, 1971), Aulne Estuary (Allen et al., 1980), Weser Estuary (Lang et al., 1989), and Tamar estuary (Uncles and Stephens, 1993). The position of the ETM shifts landward during flood tide and toward down estuary during ebb tide (Dyer, 1986).

Tidal asymmetry is another factor that may affect the ETM. As the tidal wave propagates upstream in an estuary, it becomes asymmetrical by the effect of topographic convergence and friction, such that the flood current becomes stronger than the ebb current but lasts a shorter time, and the duration time of slack-before-ebb is longer than that of slack-before-flood (Nichols and Biggs, 1985; Dyer, 1988). Thus, the combination of the greater velocity on the flood tide and the more settling at slack before ebb creates a tidal sediment trap by causing a net upstream transport of SS until net transport is directed downstream by the seaward flowing river water. This tidal trapping zone, or tidal node, is usually located near the upper limit of tides, generally farther upstream than the null zone of gravitational circulation occurring near the upper limit of salt intrusion (Allen et al., 1980). Nichols and Poor (1967) argued that tidal asymmetry superimposed on the net density flow results in a net upstream transport of sediment in the Rappahannock Estuary, a microtidal estuary. Allen et al. (1980) showed that
bottom flood currents exceed ebb currents, even above the limit of salt intrusion, and the ETM is more pronounced during flood tide than ebb tide in both the Gironde and Aulne estuaries, macrotidal estuaries in France. Hamblin (1989) demonstrated that landward sediment flux calculated by observed data in the lower layer is maintained by the slack time asymmetry near the ETM of the upper Saint Lawrence Estuary. The synoptic surveys and earlier measurements of SS transport demonstrated that tidal pumping of SS by tidal asymmetry maintains the ETM in the upper reaches of the Tamar Estuary (Uncles et al., 1985; Uncles and Stephens, 1993). Dyer (1988) and Dyer and Evans (1989) showed that a phase lag of sediment resuspension relative to near-bottom currents can produce an ETM in the presence of asymmetrical tidal currents without convergence of the two-layer gravitational circulation. Sanford et al. (2001) showed in the Upper Chesapeake Bay that convergence in gravitational circulation and its associated salinity structure contribute to strong tidal asymmetries in sediment resuspension and transport that collect and maintain a resuspendible pool of rapidly settling particles near the limit of salt intrusion.

Topography is another factor that may affect the ETM in some estuaries. A high current velocity due to topographic constriction may result in strong local erosion or resuspension, increasing SS concentration in the water column (Lin, 2001). The tidal waves, as they propagate in an estuary, are modified by convergence of the estuary cross-section and the energy loss by friction. Depending on the relative importance of these two processes, the amplitude of tidal waves can either increase or decrease upstream. Allen et al. (1980) proposed that a hypersynchronous topography, in which the convergence effect is more pronounced than the energy loss by friction, of the Gironde Estuary can lead to the zone of maximum current velocity in the upper reaches of the estuary at which an ETM may develop. More than one ETM has been observed in some estuaries
and one of the multiple ETMs has been demonstrated to be associated with
topographic convergence in the Patuxent Estuary (Roberts and Pierce, 1976) and
Columbia River estuary (Jay and Musiak, 1994).

The processes responsible for the formation of the ETM may vary for
different estuaries and/or different times in an estuary. The ETM in an estuary
may be caused by any, or a combination, of several processes. Because of the
very complicated nature of the ETM, numerical models have been employed to
study the formation mechanisms and variations of the ETM. Models ranging
from relatively simple tidal average box models (e.g., Officer, 1980a) to highly
sophisticated intratidal, three-dimensional models (e.g., Lin and Kuo, 2001) have
been employed. One of the most unique features in modeling of SS transport is
the erosion/resuspension and deposition at the water-bed interface. Some of the
earliest models assumed no erosion and deposition (Festa and Hansen, 1978;
Officer, 1980a; Geyer, 1993) or required input specification of erosion and
deposition (Officer and Nichols, 1980). Some models have attempted to simulate
erosion and deposition processes, by assuming that the rate of erosion/deposition
is a linear function of the difference of the bed shear stress and critical shear stress
for erosion/deposition. Examples include Kuo et al. (1978), Uncles and Stephens
(1989), Burchard and Baumert (1998), Brenon and Le Hir (1999), and Lin and
Kuo (2001). These models, however, assumed unlimited supply of erodible
sediments at the bed without considering the consolidation processes of the bed
sediments.

The erosional flux from the bed in reality is a function of the availability
of unconsolidated, easily erodible sediments at the bed, which may vary
depending on the previous history of erosion and deposition. Some of the recent
models have incorporated bed sediment models to simulate the depth of erodible
sediments at the bed to account for the previous history of erosion and deposition
(Spasojevic and Holly, 1994; McDonald and Cheng, 1997; Brenon and Le Hir, 1999; Reed et al., 1999; Harris and Wiberg, 2002). While they are closer to the reality, these models have suffered from the lack of data to verify the model results, especially for the consolidation processes of the bed. These models could also be quite time consuming, making the incorporation of bed sediment models impractical for long-term simulations of a three-dimensional hydrodynamic-sediment transport model (Lin et al., 2003). Using the recent experimental results of Sanford and Maa (2001), Lin et al. (2003) proposed a simple method that accounts for the previous history of erosion and deposition in modeling of erosion process, without explicitly incorporating bed sediment models. The method employs an empirical relationship in which the critical shear stress for erosion varies as a function of the previously eroded mass.

1-2. Objectives

A number of studies have been conducted to understand the characteristics of the ETM in the Upper Chesapeake Bay. Many of them were based on the data, showing that the ETM are mainly affected by the Susquehanna River discharge (Schueler, 1972; Schueler, 1974; Schueler and Pritchard, 1986), tidal resuspension (Schueler, 1971; Schueler et al., 1978), convergence in two-layer gravitational circulation near the limit of salt intrusion (Schueler, 1968a), and a combination of these processes (Sanford et al., 2001). The modeling studies of the ETM in the Upper Chesapeake Bay include the box model in Biggs (1970) and pseudo-hydrodynamic model in Schueler (1968a), which assumed a steady-state and thus did not simulate tidal resuspension. One-dimensional (vertical) model has been
employed to study tidal resuspension in Sanford et al. (1991), Sanford and Halka (1993), and Sanford and Chang (1997).

The objective of this research is to develop a three-dimensional SS transport model for the Upper Chesapeake Bay, to calibrate and verify the model with the data for 1996. One year-long model simulation, with real geometry and detailed hydrodynamics, is used to study the processes affecting the formation and variation of the ETM in the Upper Chesapeake Bay. The characteristics of the study area, the Upper Chesapeake Bay, are described in Chapter 2. The numerical model is described in Chapter 3 and the model calibration/verification for the year 1996 in Chapter 4. A study of the characteristics of the ETM using the model is described in Chapter 5. A summary is provided in Chapter 6.
2. STUDY AREA: UPPER CHESAPEAKE BAY

Chesapeake Bay, the largest estuaries in the United States, is a drowned river valley formed by the most recent rise in sea level and is less than 10,000 years old (Schubel and Pritchard, 1986). Chesapeake Bay is approximately 314 km long, varies in width from 5.5-56 km. Its long axis runs approximately north-south, and its mouth faces east connecting to the Atlantic Ocean. Chesapeake Bay including its tributary estuaries has a surface area of $11.5 \times 10^3$ km$^2$, a mean low water volume of 74 km$^3$, and a mean depth of 6.46 m (Schubel and Pritchard, 1986). Chesapeake Bay is a microtidal estuary with a tidal range of 0.2-0.9 m (Nichols and Biggs, 1985). Tides propagate up to the fall line and the upper limit of salt intrusion oscillates near about 290 km from the sea depending on the Susquehanna River discharge. Sand is characteristic near the Bay mouth, while the Bay channel, large areas west of it, and most of the tributary channels are floored with black and gray clayey silt. The major mineral constituent of the silt is quartz and the major clay minerals are chlorite, illite, and kaolinite (Schubel, 1968b).

The Upper Chesapeake Bay, the study area of this research, includes the region extending from the southern boundary at Cove Point, Maryland northward to the mouth of the Susquehanna River (Fig. 2-1). The Upper Chesapeake Bay is clearly “the estuary of the Susquehanna” (Nichols and Biggs, 1985). The drainage basin of the Susquehanna River, which accounts for 42% of the Bay’s total drainage basin, has an area of about 71,250 km$^2$ in New York, Pennsylvania, and Maryland. Nearly 87% of the total river discharge into the Upper Chesapeake Bay is supplied by the Susquehanna River with long-term average of approximately 985-1099 m$^3$ s$^{-1}$ (Schubel and Pritchard, 1986). The Susquehanna
River has a discharge pattern typical of mid-latitude rivers: high discharge in spring and other brief periods produced by snow melt and spring rains followed by low-to-moderate flow throughout most of the remainder of the year. The seasonal variation of the Susquehanna River discharge is large, producing significant variations in the circulation patterns, distributions of suspended sediment (SS), and the biological responses in the milieu of the Upper Chesapeake Bay (Schubel and Pritchard, 1986). The average tidal range in the Upper Chesapeake Bay increases northward from 0.36 to 0.5 m, with typical maximum tidal current speeds of 0.5 m s\(^{-1}\) in the channel and 0.3 m s\(^{-1}\) over the broad shoals (Sanford et al., 2001).

The sources of SS in Chesapeake Bay come from the rivers, shore erosion, primary productivity, and the sea (Biggs, 1970; Schubel and Carter, 1977; Eaton et al., 1980; Dyer, 1986). The Susquehanna River is the dominant source of SS, accounting for more than 80%, for the Upper Chesapeake Bay (Biggs, 1970). The discharged sediment is nearly all fine-grained silt and clay because the coarser particles are trapped upstream in the reservoirs along the lower reaches of the Susquehanna River (Schubel, 1968b, 1969a, and 1974). In most years, the Upper Chesapeake Bay receives about 0.5-1.0×10\(^6\) tons yr\(^{-1}\) (Schubel, 1968b, 1968c, and 1969b; Biggs, 1970), which accounts for more than 90% of the total input of fine-grained sediment (Hirschberg and Schubel, 1979). It has been estimated that denudation provides 3.0×10\(^6\) tons yr\(^{-1}\) of sediments to the river headwaters (Williams and Reed, 1972) and shore erosion supplies about 0.1×10\(^6\) tons yr\(^{-1}\) of silt and clay (Hirschberg and Schubel, 1979).

Fluvial sediments entering the Upper Chesapeake Bay from the Susquehanna River are transported seaward through freshwater reaches near the river mouth and the upper estuarine layer and landward through the lower estuarine layer to the inner salt limit where it is retained for long periods in the
ETM zone. The ETM zone, an area of net deposition, is contained within the uppermost part of the estuary and is almost always found between latitudes 39°10′N and 39°28′N, a range of about 40 km (Sanford et al., 2001). The mean position of the ETM was located approximately 20-30 km downstream from the Havre De Grace (Schubel, 1974). The location, however, varies seasonally and at a shorter time scale: the location during the freshet may migrate up to about 45 km seaward from its normal position. The location also is affected by strong tidal asymmetry associated with salinity structure (Sanford et al., 2001).

Approximately 50-70% of the sediments discharged from the Susquehanna River during the freshet are deposited within the upper 45 km and the remainder is carried farther seaward (Schubel, 1968b; Biggs, 1970; Hirschberg and Schubel, 1979; Schubel and Prichard, 1986). Schubel (1968b) using simple box models estimated the sedimentation rate in the upper 45 km reach of about 0.6 cm yr⁻¹. Radiometric methods estimated the mean sedimentation rate of about 0.45 cm yr⁻¹ in the same segment of the upper Bay excluding flood events (Schubel and Hirschberg, 1977; Hirschberg and Schubel, 1979). Officer et al. (1984) reported the sedimentation rates in the same segment of about 0.3-1.2 g cm⁻² yr⁻¹ (0.11-0.45 cm yr⁻¹ with sediment density of 2.65 g cm⁻³), higher than that of downstream.
Fig. 2-1. A map of the Upper Chesapeake Bay showing CBP monitoring stations (● for those used for model verification and ✗ for other stations), one station (▲) for intratidal data (Sanford et al., 2001) in Fig. 4–9, and three stations for intratidal variations in Fig. 5–14 to 5–16 (□).
3. MODEL DESCRIPTION

The CH3D-WES (Curvilinear Hydrodynamics in Three Dimensions - Waterways Experiment Station) model has been successfully applied to simulate the hydrodynamic conditions of the entire Chesapeake Bay and its major tributaries (Johnson et al., 1993; Wang and Johnson, 2000). The model had a good representation of the observed surface elevation, flow velocity, salinity, and temperature. The CH3D-WES model was utilized to provide the flow fields for the simulation of eutrophication model for the entire Chesapeake Bay (Cerco and Cole, 1993; Cerco, 1995). Wang et al. (in press) applied the CH3D-WES model to the Upper Chesapeake Bay with a very fine grid to resolve the main channel and major tributaries. The model was applied for the period from 1992 to 1997 and was calibrated and verified for water elevation, salinity, and temperature. The simulated information of physical mass transport was used for water quality modeling (Liu, 2002).

In this study, a suspended sediment (SS) transport model is developed and internally linked to the CH3D-WES model. The SS transport model is based on the mass-balance equations for three size classes, each representing a fine, medium and coarse size class. No interaction among different size classes and no effect of the SS concentration on the hydrodynamic field are assumed. A very brief description of the CH3D-WES model is given here, followed by detailed description of the SS transport model, particularly for erosion and deposition. Detailed description of the CH3D-WES model can be found in Johnson et al. (1991 and 1993) and Wang and Johnson (2000).
3-1. Hydrodynamic Model (CH3D-WES)

The CH3D-WES model is based on the continuity, momentum, salt-balance, and heat-balance equations, with the assumptions of incompressible flow, hydrostatic approximation, Boussinesq approximation, and eddy viscosity approach. For turbulence closure, the k-ε turbulence model (Rodi, 1980; ASCE Task Committee, 1988) is employed with the assumption of local equilibrium of turbulent kinetic energy. The model uses a boundary-fitted or generalized curvilinear grid in the horizontal direction to resolve irregular shoreline and deep navigation channels. The governing equations hence are transformed from the Cartesian coordinates (x, y) to the generalized curvilinear coordinates (ξ, η). In the vertical direction, Cartesian coordinate is used. The model solves non-dimensional forms of the governing equations to facilitate relative magnitude comparison of the various terms in the governing equations (Johnson et al., 1991).

The governing equations are solved numerically with finite difference methods. For the time derivative terms, a two-time level scheme is used. For the spatial derivative terms, an upwind difference scheme is used for the advective terms in the momentum equations, whereas a third-order accurate QUICKEST scheme (Leonard et al., 1978; Leonard, 1979) is used for the advective terms in the mass-balance equations for salinity and temperature. A staggered grid is used both in the horizontal and vertical directions of the computational domain. The CH3D-WES model employs a mode splitting that separates external and internal modes in the horizontal momentum equations. The divergence terms in the vertically-averaged continuity equation and the barotropic pressure gradient terms (surface slope) in the vertically-averaged external mode equations are treated implicitly to relieve the strict stability criteria caused by fast-propagating free-
surface gravity waves. The vertical mixing terms in all governing equations and the bottom friction terms are treated implicitly. The finite difference equations are solved by Alternating Direction Implicit (ADI) method and the resulting linear, tri-diagonal algebraic equations are solved by Gaussian elimination method.

3-2. Suspended Sediment Transport Model

3-2-1. Governing Equation

The SS transport model is based on the mass-balance equations for three size classes. The equations are identical to the salt-balance equation in the CH3D-WES model except for the terms representing gravitational settling in the water column and erosion/deposition at the water-bed interface. With the assumption of no interaction among different size classes and no effect of the SS concentration on the hydrodynamic field, the mass-balance equation, expressed in terms of dimensionless variables (those denoted with *), for any size class of SS over a vertical layer in generalized curvilinear coordinates may be expressed as (Spasojevic and Holly, 1994).

\[
\frac{\partial h^* \rho^* C^*}{\partial t^*} = \frac{E_v^*}{S_{ev}^*} (VD_{top}^* - VD_{bottom}^*) - \frac{R_o^*}{J^*} \left( \frac{\partial (HA^* \cdot u^*)}{\partial \xi^*} + \frac{\partial (HA^* \cdot v^*)}{\partial \eta^*} \right) + R_o^* (VA^*_{top} - VA^*_{bottom}) + R_{of}^* (VS^*_{top} - VS^*_{bottom}) + \frac{E_v^*}{S_{eh}^*} \cdot HD^*
\]

(3-1)

\[
VD^* = D_v^* \frac{\partial \rho^* C^*}{\partial z^*} \quad HA^* = h^* J^* \rho^* C^* \quad VA^* = w^* \rho^* C^* \quad VS^* = w_s^* \rho^* C^*
\]

t^* = time
\( \xi^* \) and \( \eta^* \) = along and cross channel directions, respectively
\( z^* \) = vertical direction (positive upward)
\( h^* \) = layer thickness
\((u^*, v^*, w^*) = velocities in (\xi^*, \eta^*, z^*) \) directions
\( \rho^* \) = water density
\( C^* \) = suspended sediment concentration
\( R_{o}^* \) = Rossby number
\( R_{of}^* \) = Rossby number for settling velocity
\( E_h^* \) and \( E_v^* \) = Ekman numbers, respectively, in horizontal and vertical directions
\( S_{ch}^* \) and \( S_{cv}^* \) = Schmidt number, respectively, in horizontal and vertical directions
\( J^* \) = square root of the Jacobian-matric determinant = \( x_\xi y_\eta - x_\eta y_\xi \) where the subscript indicates difference
\( w_s^* \) = settling velocity of suspended sediment particle
\( D_v^* \) = vertical eddy diffusivity
\( HD^* \) = horizontal diffusion terms as described in Johnson et al. (1991).
The nondimensional numbers are defined as

\[
R_{o}^* = \frac{U_r}{f \cdot X_r} \quad R_{of}^* = \frac{w_{s,r}}{f \cdot Z_r} \quad (3-2a)
\]

\[
E_h^* = \frac{A_{h,r}}{f \cdot X_r^2} \quad E_v^* = \frac{A_{v,r}}{f \cdot Z_r^2} \quad (3-2b)
\]

\[
S_{ch}^* = \frac{A_{h,r}}{D_{h,r}} \quad S_{cv}^* = \frac{A_{v,r}}{D_{v,r}} \quad (3-2c)
\]

\( f \) = Coriolis parameter (s\(^{-1}\))
\( U_r \) = reference value of velocity (m s\(^{-1}\))
$X_r$ and $Z_r$ = reference values, respectively, for horizontal and vertical dimensions (m)

$w_{s,r}$ = reference value for settling velocity ($\text{m s}^{-1}$)

$A_{h,r}$ and $A_{v,r}$ = reference values, respectively, for horizontal and vertical eddy viscosities ($\text{m}^2 \text{s}^{-1}$)

$D_{h,r}$ and $D_{v,r}$ = reference values, respectively, for horizontal and vertical eddy diffusivities ($\text{m}^2 \text{s}^{-1}$).

The terms on the right-hand side of Eq. 3-1 represent, respectively, the transport by vertical diffusion, horizontal advection, vertical advection, settling, and horizontal diffusion.

### 3-2-2. Settling Velocity

In a clear, still fluid the settling velocity of a solitary sand particle smaller than about 100 $\mu$m may be estimated by (van Rijn, 1984; Spasojevic and Holly, 1994)

$$w_s = \frac{1}{18} \frac{(s - 1)gD_s^2}{\nu}$$

(3-3)

$w_s$ = settling velocity ($\text{m s}^{-1}$)

$s$ = dimensionless sediment density = $\rho_s/\rho_r$

$\rho_s$ = density of sediment particle ($2.65 \times 10^3 \text{ kg m}^{-3}$)

$\rho_r$ = reference density ($1 \times 10^3 \text{ kg m}^{-3}$)

$g$ = gravitational acceleration ($9.8 \text{ m s}^{-2}$)

$D_s$ = particle size (m)

$\nu$ = kinematic viscosity ($1.14 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$).
3-2-3. Surface Boundary Condition

The surface boundary condition of no sediment flux across the surface may be expressed in a dimensional form as

\[
\left( D_v \frac{\partial C^*}{\partial z} \right)_{\text{surface}} + \left( w_s \rho C^* \right)_{\text{surface}} = 0
\]  

(3-4)

3-2-4. Bottom Boundary Condition

At the bottom, the flux from the bed to the water column is represented with the sum of erosional and depositional fluxes (Sheng and Lick, 1979; Lick, 1982). The bottom boundary condition in a dimensional form then may be expressed as

\[
- \left( D_v \frac{\partial C^*}{\partial z} \right)_{\text{bottom}} - \left( w_s \rho C^* \right)_{\text{bottom}} = E - D = S_o
\]  

(3-5)

\( E = \) erosional, or resuspension, flux (kg m\(^{-2}\) s\(^{-1}\))
\( D = \) depositional flux (kg m\(^{-2}\) s\(^{-1}\))
\( S_o = \) net erosional flux (net erosion if \( S_o > 0 \) and net deposition if \( S_o \leq 0 \)).

3-2-4a. Deposition

The depositional flux (D) may be expressed as
D = P \cdot w \rho C^* \quad (3-6)

where \( P \) is the probability that a sediment particle will stick to the bed. The probability of deposition is usually expressed as (Einstein and Krone, 1962; Dyer, 1986; Mehta, 1989; Self et al., 1989)

\[
P = 1 - \frac{\tau_b}{\tau_{cd}} \quad \text{for } \tau_b < \tau_{cd} \quad (3-7)
\]

\[
P = 0 \quad \text{for } \tau_b \geq \tau_{cd}
\]

\( \tau_b = \) bottom shear stress (Pa)

\( \tau_{cd} = \) critical shear stress for deposition (Pa).

In cohesive sediment transport models, the assumption of mutually exclusive erosion and deposition is usually employed with \( \tau_{cd} \) is less than, or equal to, the critical shear stress for erosion \( \tau_{ce} \) (Ariathurai and Krone, 1976; Dyer, 1986; Mehta, 1986 and 1988; Partheniades, 1986 and 1993; O'Connor and Nicholson, 1988; Odd, 1988; Mehta et al., 1989; Uncles and Stephens, 1989). In non-cohesive sediment models, on the other hand, simultaneous erosion and deposition, i.e., continuous deposition, is usually assumed (Smith, 1977; Dyer, 1986; Glenn and Grant, 1987). Several researchers, however, employed the simultaneous erosion and deposition in transport modeling of cohesive sediments (Lick, 1982; Lavelle et al., 1984; Bedford et al., 1987; Lang et al., 1989). Setting \( P = 1 \) in Eq. 3-7, for example, will ensure continuous deposition. Sanford and Halka (1993) suggested that allowing continuous deposition of a single suspended particle class may often give quite satisfactory results for practical modeling purpose. They showed that the model using continuous deposition for a mud size simulates the dynamics of the SS in the Upper Chesapeake Bay well.
3-2-4b. Erosion

Various formulations have been employed to model erosional flux \((E)\), including a simple linear relationship (Odd, 1988; Lang et al., 1989; Uncles and Stephens, 1989; Sanford et al., 1991; Sanford and Halka, 1993; Mei et al., 1997), a power law relationship (Lick, 1982; Lavelle et al., 1984; Maa et al., 1998), and an exponential form (Mehta, 1988; Amos et al., 1992). In all these formulations, erosional flux is expressed as a function of excess bed shear stress, \((\tau_b - \tau_{ce})\). The larger the \((\tau_b - \tau_{ce})\), the more erosion. For example, a simple linear relationship for erosional flux may be expressed as

\[
E = M(\tau_b - \tau_{ce}) \quad \text{for } \tau_b > \tau_{ce} \\
= 0 \quad \text{for } \tau_b \leq \tau_{ce}
\]  

(3-8)

where \(M\) is the erosion rate \((\text{kg m}^{-2} \text{s}^{-1} \text{Pa}^{-1})\). Although many previous modeling studies had considered constant \(\tau_{ce}\) in Eq. 3-8, it in reality is not constant but depends not only on the characteristics of the bed sediments but on the consolidation processes of the bed. Depending on the previous history of erosion and deposition, the degree of consolidation of the bed may vary, which will affect the availability of unconsolidated, easily erodible sediments at the bed.

3-2-4c. Depth-Limited Erosion

To account for the consolidation processes of the bed, some recent modeling efforts have incorporated bed sediment models to simulate the depth of
erodible bed sediments (Spasojevic and Holly, 1994; McDonald and Cheng, 1997; Brenon and Le Hir, 1999; Reed et al., 1999; Harris and Wiberg, 2002). These models, while they are closer to the reality, have suffered from the lack of data to verify the model results, particularly for the consolidation processes of the bed. These models could also be quite time consuming, making the incorporation of bed sediment models impractical for long-term simulation of a three-dimensional hydrodynamic-sediment transport model (Lin et al., 2003).

Sanford and Maa (2001) examined the in situ data from a submersible seabed flume at the Baltimore Harbor and developed a simple extension of the standard linear erosion formulation, in which $\tau_{ce}$ is expressed as a function of previously eroded mass. The set of erosion formulations proposed by Sanford and Maa (2001) is

$$E(t, m_E) = M(m_E)[\tau_b(t) - \tau_{ce}(m_E)]$$
$$= 0$$

for $\tau_b > \tau_{ce}$

for $\tau_b \leq \tau_{ce}$

(3-9)

$$M(m_E) = a_E (m_E - c_E)^p$$

(3-10a)

$$\tau_{ce}(m_E) = b_E (m_E - d_E)^q$$

(3-10b)

$m_E$ = eroded mass (kg m$^{-2}$)

$a_E$, $b_E$, $c_E$, $d_E$, $p$, and $q$ = empirical constants.

Sanford and Maa (2001) estimated with the data from the Baltimore Harbor that $a_E = 0.027$, $b_E = 0.86$, $c_E = d_E = 0.017$, $p = 0.54$, and $q = 0.5$. In Eqs. 3-9 and 3-10, the two parameters, $\tau_{ce}$ and $M$, are no longer constants but vary as a function of the eroded mass $m_E$. As erosion proceeds, $m_E$ increases, resulting in increase in $\tau_{ce}$ (Sanford and Maa, 2001: Fig. 2), which reflects less available unconsolidated bed sediments for erosion. $M$ also increases with increasing $m_E$ but the rate of
increase in $M$ levels out as $m_E$ keeps increasing (Sanford and Maa, 2001: Fig. 4), which again reflects less available unconsolidated sediments at the bed.

To account for the effect of bed consolidation without explicitly incorporating a bed sediment model, Lin et al. (2003) presented an algorithm that implements the relationship in Sanford and Maa (2001) to a SS transport model, in which $m_E$ is defined as

$$m_E = \int_{t^*} E(t)dt$$  \hspace{1cm} (3-11)

$m_E$ = accumulated eroded mass since the beginning of an erosion cycle (kg m$^{-2}$)
$t^*$ = total time since the beginning of an erosion cycle (s).

Lin et al. (2003) employed a point model, a model with one box on top of the bed and in which the change in SS concentration is solely determined by erosion and deposition, to examine the behavior of the algorithm. The calculated SS concentrations with the depth-limited erosion (Eqs. 3-9 to 3-11) give more realistic results in terms of the magnitude of concentration and the timing of erosion compared to those with constant $\tau_{ce}$. Little change in the point model results with varying $a_E$, thus $M$, in Eq. 3-10a was noted, indicating the insensitivity of the choices of $M$ that ranges from 0.002-0.006 kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$ (Sanford and Maa, 2001). Appreciable differences were noted with varying $b_E$, thus $\tau_{ce}$, in Eq. 3-10b.

3-2-4d. Exclusive versus Simultaneous Erosion and Deposition

Lin et al. (2003) assumed mutually exclusive erosion (Eqs. 3-9 to 3-11) and deposition (Eqs. 3-6 and 3-7) with $\tau_{cd} \leq \tau_{ce}$. An alternative approach is to assume simultaneous erosion and deposition, i.e., continuous deposition, as
suggested by Sanford and Halka (1993). To adopt continuous deposition in the algorithm of Lin et al. (2003), depositional flux is calculated every time step using Eq. 3-6 with $P = 1$ and erosional flux using Eqs. 3-9 and 3-10, with the accumulated eroded mass $m_E$ defined as

$$m_E = \int_s \max\{(E - D), 0.0\}\,dt$$  \hspace{1cm} (3-12)

The point model in Lin et al. (2003) is employed to compare exclusive erosion and deposition (EED) and continuous deposition (CD). Sanford and Maa (2001) showed that their relationship in Eqs. 3-9 and 3-10 can be used for both depth-limited (Type I) erosion and unlimited (Type II) erosion, with a seamless transition between the two types. Type I erosion refers to the situations where erosion can be limited by the availability of unconsolidated, easily erodible bed sediments. Type II erosion refers to the situations with unlimited amount of unconsolidated sediments, so that erosion is never limited by the availability of erodible bed sediments. Hence, the comparison between EED and CD is made for both types of erosion. The comparison also is made for three different particle sizes, 3, 18, and 65 μm: they are chosen to represent fine, medium and coarse particle size classes characteristic to the Upper Chesapeake Bay (Section 4-1-3). The respective corresponding settling velocities, estimated from Eq. 3-3, are 0.007, 0.26, and 3.3 mm s$^{-1}$.

For the fine particle size with Type I erosion, EED results in a virtually constant water column concentration of SS (Fig. 3-1b), indicating that the limited amount of erodible bed sediments is readily resuspended and kept in suspension all the time due to small settling velocity and short time duration of deposition. CD gives the same pattern (Fig. 3-1c), with slightly larger, but still very small, tidal fluctuations. For Type I erosion, EED with constant $\tau_{ce} = \tau_{cd} = 0.02$ Pa
(results not shown) also gives the SS concentration almost identical to those in Figs. 3-1b and 3-1c. Therefore, for the fine particles that are mostly in suspension in the water column due to very slow settling (Schubel, 1968a and 1971; Sanford et al., 2001), the modeling method of erosion is not critical as long as the method assures quick resuspension of deposited fine particles. For Type II erosion, the SS concentration keeps increasing in both cases of EED (Fig. 3-1d) and CD (Fig. 3-1e) due to the unlimited erodible bed sediments and relatively small settling velocity. The condition of Type II erosion may not be realistic for the fine particles with the size of about 3 μm (0.007 mm s⁻¹), since they may not stay in the bed long enough to be consolidated (Schubel, 1968a and 1971; Sanford et al., 2001).

The point model results for the medium particle size are shown in Fig. 3-2. EDD for Type I erosion (Fig. 3-2b) shows little tidal fluctuations in the SS concentration. The maximum SS concentration remains constant over the time periods of the relatively large τ_b, during which too large a τ_b prevents deposition and lack of erodible bed sediments results in no erosion. The behavior of the constant maximum concentration maintained into the decelerating tide in Fig. 3-2b does not agree with the observation in Sanford and Halka (1993) that the SS concentration decreases as τ_b starts to decrease. EED for Type II erosion (Fig. 3-2d) shows that the SS concentration keeps increasing due to the unlimited erodible bed sediments and relatively short time duration of deposition. CD, on the other hand, gives the SS concentration with tidal fluctuations for both types of erosion. For Type I erosion (Fig. 3-2c), the SS concentration increases as τ_b starts to increase, and all of the erodible bed sediments are resuspended before reaching the maximum τ_b, beyond which CD decreases the concentration, resulting in the maximum concentration occurring earlier than the maximum τ_b. For Type II erosion (Fig. 3-2e), termination of erosion is not controlled by the availability of
the erodible bed sediments but by the increase in $\tau_{ce}$ due to increase in $m_F$ (Eq. 3-10b), resulting in the maximum concentration occurring closer to, still earlier than the maximum $\tau_b$. The behavior that the maximum erosion rate occurs at the midpoint in the accelerating tide and that erosion has ceased by the time $\tau_b$ reaches its maximum in Figs. 3-2c and 3-2e agrees with the observation in Sanford and Maa (2001).

The point model results for the coarse particle size are shown in Fig. 3-3. EDD for both types of erosion (Figs. 3-3b and 3-3d), same as in Fig. 3-2b, shows the prolonged time duration of the constant maximum SS concentration, contrary to the observation in Sanford and Halka (1993). CD for both types of erosion (Figs. 3-3c and 3-3e) exhibits similar concentration variations with distinctive tidal fluctuations as in Figs. 3-2c and 3-2e. The maximum concentration occurs earlier than the maximum $\tau_b$ for Type I erosion but close to the time of the maximum $\tau_b$ for Type II erosion. The coarse particles with the size of about 65 $\mu$m (3.3 mm s$^{-1}$) will stay at the bed most of the time and spend less time in suspension in the water column, which is well represented by CD for both types of erosion (Figs. 3-3c and 3-3e).

3-2-4e. Erosion and Deposition in the Present Model

The present SS transport model simulates three size classes of SS, including fine, medium, and coarse classes (Section 4-1-3). The fine class is to represent the fine particles that are mostly in suspension in water column. They hardly settle to the bed and will be readily resuspended even when they reach the bed. Since the fine particles are not likely to stay in the bed long enough to be consolidated, it is assumed that only the fine particles deposited to the bed will be eroded, without allowing erosion of consolidated, hard bed sediments. The
present model employs mutually exclusive erosion and deposition with constant values for $\tau_{ee}$ (Eq. 3-8) and $\tau_{ed}$ (Eq. 3-7) for the fine class. A small constant value needs to be assigned to $\tau_{ee}$, and to $\tau_{ed}$, to assure quick erosion of deposited fine particles.

The medium class is to represent the particles that are alternately suspended and deposited by tidal current. The coarse class is to represent the particles that stay at the bed most of the time and are resuspended into the water column only when bed shear stress is relatively large. For the particles in these two classes, both unconsolidated and consolidated particles may be eroded, with the erosion of the consolidated particles being more difficult. For the medium and coarse classes, the depth-limited erosion with continuous deposition (Eq. 3-12) is employed by varying $\tau_{ee}$ as a function of accumulated eroded mass (Eqs. 3-9 and 3-10). The coefficient $b_E$ in Eq. 3-10b is considered as a calibration parameter for the medium and coarse classes.

3-2-5. Numerical Method

The mass-balance equations for the three size classes are solved with finite difference methods similar to those for the salt-balance equation. A two-time level scheme is used for the time derivative terms. An upwind difference scheme is used for the horizontal advection terms, although the QUICKEST scheme is used for the salt-balance equation. It is because of the negative SS concentrations, although small in magnitude, inherent to the QUICKEST scheme. The vertical mixing terms are treated implicitly. The settling terms are solved implicitly in time and with an upwind scheme in space. A staggered grid is used both in the horizontal and vertical directions, with the SS concentrations defined at the center of the grid and the settling velocity at the top end of the grid (Fig. 3-4).
Fig. 3-1. Point model results for a fine size particle (3 um): bed shear stress (a) and SS concentrations for depth-limited (Type I) erosion with exclusive erosion and deposition (b) and with continuous deposition (c), and unlimited (Type II) erosion with exclusive erosion and deposition (d) and with continuous deposition (e).
Fig. 3-2. Point model results for a fine size particle (18 um): bed shear stress (a) and SS concentrations for depth-limited (Type I) erosion with exclusive erosion and deposition (b) and with continuous deposition (c), and unlimited (Type II) erosion with exclusive erosion and deposition (d) and with continuous deposition (e).
Fig. 3-3. Point model results for a fine size particle (65 μm): bed shear stress (a) and SS concentrations for depth-limited (Type I) erosion with exclusive erosion and deposition (b) and with continuous deposition (c), and unlimited (Type II) erosion with exclusive erosion and deposition (d) and with continuous deposition (e).
Fig. 3-4. A diagram showing the location of the variables
4. MODEL APPLICATION

The suspended sediment (SS) transport model described in Chapter 3 is applied to the Upper Chesapeake Bay. The modeling domain covers from the southern boundary at Cove Point northward to the lower portion of the Susquehanna River (Fig. 4-1). The modeling domain is gridded with a fine resolution to resolve main channel and its tributaries (Fig. 4-1), with approximately 1.0 km resolution longitudinally and 0.25 km laterally. Vertical resolution is 1.52 m with the exception of the surface layer with 2.44 m thickness to cover the tidal fluctuation. There are 3,777 active horizontal cells and a maximum of 19 vertical layers, resulting in 16,307 computational cells. A time step of 180 seconds is used that satisfies the stability criteria.

Application of the SS transport model requires extensive data for model input (e.g., initial condition, boundary conditions, and settling velocity) and for model calibration and verification. The data from various sources have been used for the present model application. Sanford et al. (2001) has collected an intensive data set for the Upper Chesapeake Bay in 1996. Eight hydrographic surveys along the axis of the main channel were carried out between February 1 and October 27, 1996, with each survey employing 8-11 stations from just north of the Bay Bridge (about 30°00'N) to Turkey Point (30°26'N). They presented the vertical distributions of salinity and SS concentration along the main channel for five surveys on February 1, July 21, July 25-26, October 22, and October 27. They also presented the time-depth contours of salinity, SS, and along-channel velocity at one ETM channel station for 25 hours on October 24-25. U.S. Environmental Protection Agency (EPA) Chesapeake Bay Program (CBP) Office has been collecting the monitoring data since 1984 from the stations throughout the Chesapeake Bay. Surveys have been conducted semimonthly in March
through October and monthly in the remaining months. Of Bay-wide stations, forty stations fall inside the present modeling domain (Fig. 2-1).

The model is run to simulate the conditions in 1996: the simulation period is selected to make use of the intensive data set collected by Sanford et al. (2001). The model is calibrated with the data in Sanford et al. (2001) and verified with the CBP monitoring data.

4-1. Model Set-Up

4-1-1. Initial and Open Boundary Conditions

The initial condition for the SS concentration is estimated using the CBP monitoring data. Linear interpolation is employed between forty monitoring stations to construct a matrix table with contributing fractions of forty stations for each model cell. Linear interpolation is also employed in vertical direction and the same matrix table is applied for each vertical layer. For a given model cell, the sum of contributing fractions always is one. The matrix table of contributing fractions is applied to the data collected in December of 1995 to estimate initial condition for each model cell. Since no measurements were made at the shallow stations across cross-sections, e.g., at the stations CB3.3W, CB3.3E, CB4.1W, CB4.1E, CB4.2W, CB4.2E, CB4.3W, and CB4.3E (Fig. 2-1), laterally uniform initial condition is assumed.

Open boundary condition is estimated using the CBP monitoring data at the station CB4.4, located about 3 km north of the open boundary (Fig. 2-1). Linear interpolation using the data at CB4.4 is employed in the vertical direction and lateral uniformity is assumed. Open boundary condition is specified at the
intervals of measurements (bi-weekly or monthly) and the model employs linear interpolation in time.

4-1-2. External Loading from the Susquehanna River

EPA CBP Office has maintained a watershed model, Hydrologic Simulation Program - Fortran, for the entire Chesapeake Bay, the domain of which includes Maryland, Virginia, Delaware, West Virginia, Pennsylvania, and portions of New York. When overlapped with the present modeling domain, the watershed model has four segments contributing to the fall-line loads (Susquehanna, Gunpower, Patapsco, and Choptank rivers) and nineteen segments contributing to the below-fall-line nonpoint source loads. The watershed model provides daily outputs for 1996 including the discharge rates and SS loads. The total loading in 1996 through the fall-line and below-fall-line is about 12,496 tons d\(^{-1}\). About 91% of the total loading is the fall-line loading (11,367 tons d\(^{-1}\)), about 96% of which is from the Susquehanna River (10,920 tons d\(^{-1}\)). The present model application has four river boundaries including the Susquehanna River, Chesapeake and Delaware Canal, Chester River and Choptank River. Although the freshwater discharges are considered for all four river boundaries in the CH3D-WES model, the SS loading is considered only for the Susquehanna River, which is the major source of sediments for the Upper Chesapeake Bay (Hirschberg and Schubel, 1979; Officer et al., 1984; Schubel and Pritchard, 1986).

Daily mean freshwater discharge rates are available for the Susquehanna River from U.S. Geological Survey (USGS), which are used in the CH3d-WES model. The daily SS concentrations for the Susquehanna River are estimated from the USGS daily freshwater discharge data and the daily loads calculated from the CBP watershed model. Figures 4-2 and 4-3 show the USGS daily
freshwater discharge rates and the watershed model output of daily SS loads, respectively, for the Susquehanna River.

4-1-3. Particle Size Classes

Various size distributions have been observed in the Upper Chesapeake Bay (Schubel, 1968a and 1971; Schubel et al., 1978; Sanford et al., 2001). Schubel et al. (1978) showed that the vertical flux density is relatively uniformly apportioned among particles with Stokes’ diameters ranging from about 2-64 μm, but smaller and larger particles than this range contribute relatively little to the vertical flux density at 0.5 m above the bottom. Previous studies have shown that the size distributions in the Upper Chesapeake Bay may be characterized by three classes (Schubel et al., 1978). The first class consists of fine particles that are in more or less continual suspension throughout the water column and thus constitute the background concentration. They settle slower than 0.007 (Schubel, 1968a and 1971) to 0.02 (Sanford et al., 2001) mm s⁻¹, with the corresponding Stokes’ diameters of about 3 to 5 μm (Eq. 3-3). With little erosion and deposition, the fine particles are expected to have relatively little vertical gradient in SS concentration. The second class consists of medium particles that are alternately suspended and deposited by tidal current (Schubel et al., 1978). The last class consists of coarse particles, mostly fine sands composed of sandy delta known as Susquehanna Flats, though coarse particles are trapped upstream in the reservoirs along the lower reaches of the river (Schubel, 1968b, 1969a, and 1974). The coarse particles are expected to stay at the bed most of the time, except the time periods of relatively large bed shear stress.

Based on the previous studies (e.g., Schubel et al., 1978) and the size distributions in Sanford et al. (2001: Fig. 6), three particle size classes, 3, 18, and
65 μm, are considered in the present model application to the Upper Chesapeake Bay. They are referred to as the fine, medium, and coarse classes, respectively. The corresponding settling velocities calculated from Eq. 3-3, respectively, are 0.007, 0.26, and 3.3 mm s\(^{-1}\), which are in the same order of magnitude in Schubel et al. (1978) and Sanford et al. (2001). The total SS concentration is the sum of the three classes.

The external loading from the Susquehanna River (Fig. 4-3) needs to be split into the three classes. The reservoirs of the Susquehanna almost eliminate the introduction of any sand into the Upper Chesapeake Bay (Officer et al., 1984). The USGS data in 1979-2002 at the Conowingo Dam, about 15 km upstream of the river boundary in the present model domain, show that a mean of 94.7%, with a standard deviation of 7.9% (n = 343), of the SS is finer than 0.62 mm (http://va.water.usgs.gov/cgi-bin/falline/qwmd.p?01578310=on&P80154=on&P70331=on). Of the river loading, 5% is assigned to the coarse class. The fine class, which is mostly in continual suspension throughout the water column, is responsible for the background concentration (Schubel, 1968a and 1971; Sanford et al., 2001). From the examination of the model results (Section 4-2-3), it is determined to assign 20% of the river loading to the fine class. The remaining 75% is assigned to the medium class. The same partitioning fractions are employed for the initial and open boundary conditions.

4-1-4. Calibration Parameters

The present model employs mutually exclusive erosion and deposition with constant values for \( \tau_{cc} \) (Eq. 3-8) and \( \tau_{cd} \) (Eq. 3-7) for the fine class (Section 3-2-4e). A small constant value of 0.02 Pa is assigned to \( \tau_{cc} \) to assure quick erosion of deposited fine particles and the same value is assigned to \( \tau_{cd} \).
The present model employs the depth-limited erosion with continuous deposition (Section 3-2-4e) for the medium and coarse classes. The depth-limited erosion is accounted for by varying the $\tau_{ee}$, to make erosion of consolidated sediments more difficult, as a function of $m_E$ (Eqs. 3-9, 3-10, and 3-12). The model results are calibrated against the coefficient $b_E$ in Eq. 3-10b. The values of 0.56 and 0.96 are determined for the coefficient $b_E$, respectively, for the medium and coarse classes. A large value of the coefficient $b_E$ enhances the rate of increase in $\tau_{ee}$ as a function of $m_E$, resulting in relatively more difficult erosion of the coarse class particles. Following Sanford and Maa (2001), the range of 0.002-0.006 kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$ is imposed on the parameter $M$ calculated from Eq. 3-10a.

4-2. Model-Data Comparison

The SS transport model is run to simulate the one year condition in 1996. The model run is started from December 27, 1995 with arbitrary (zero) initial conditions for surface elevation and velocity. At 00:00 on January 1 of 1996, the initial conditions for salinity, temperature, and SS concentration are specified. The model results are calibrated with the intensive data in Sanford et al. (2001) and verified with the CBP monitoring data.

4-2-1. Axial Distributions along Main Channel

The model calculated axial (x-z) distributions of salinity and SS concentration along the main channel are compared with the data in Sanford et al. (2001). Since the times of each survey are given in terms of tidal current phase in Sanford et al. (2001), the hourly average model results in the corresponding tidal phases are compared with the data. There however are surveys, in which the tidal
phase was not the same throughout the stations. For such surveys, the hourly average model results corresponding to the survey tidal phases in the ETM zone are compared with the data.

The 1996 average flow of 1,800 m$^3$ s$^{-1}$ from the Susquehanna River exceeded the nine year (1991-1999) average by 60% and the associated sediment load exceeded the nine year average by 176% (Sanford et al., 2001). The high annual average flow and sediment load were due to several large flood events. The most important was an enormous flood event during the last week of January (Fig. 4-2). The highest daily flow rate, 17,613 m$^3$ s$^{-1}$ on January 21, was about 10 times the January average flow and the total sediment load was about 17 times the total for an average January. The last 100 days of 1996 were much wetter than usual with 140% higher freshwater flow and 370% higher sediment loads than the corresponding averages. The freshwater discharge on October 21 was 5,267 m$^3$ s$^{-1}$, about one third of the enormous January flood (Fig. 4-2). The axial distribution data from five surveys on February 1, July 21, July 25-26, October 22, and October 27 are presented in Sanford et al. (2001). The data on February 1 was collected at the tail end of the enormous January flood. The surveys on October 22 and 27 followed the high flow event on October 21. The remaining two surveys were conducted during low discharge conditions in July (Fig. 4-2).

The data on February 1 (Sanford et al., 2001: Fig. 4) shows the most extreme SS distributions observed in 1996, with the limit of salt intrusion (defined as the intersection of the 1 psu isohaline with the bottom) intruded strongly up to 16 km downstream from Harve De Grace (2 psu isohaline at km 30) after the January flood might have pushed salt out of the upper portion of the Bay. A weak ETM-like near-bottom SS maximum was observed between km 14-30. The model result for February 1 (Fig. 4-4), however, shows the limit of salt intrusion around km 40 and the ETM around km 22. The model particularly fails to
reproduce the highest SS concentrations distributed nearly homogeneous in the vertical upstream of km 20 in the inflowing freshwater (Sanford et al., 2001: Fig. 4). The model result for February 1 indicates that the present model seems not doing well in reproducing the observations immediately after an enormous flooding event in January.

The data on October 22 and 27 (Sanford et al., 2001: Fig. 5) shows the response of salinity and SS concentration after the high freshwater inflow on October 21 (Fig. 4-2). The data on October 22 show the high freshwater and its associated high sediment load as they entered the upper part of the ETM zone. The limit of salt intrusion was around km 35 and a weak ETM was formed between km 20-40 with its center at km 30. Although the model reproduces the locations of the limit of salt intrusion and the ETM (Fig. 4-5), it fails to reproduce the observed highest SS concentrations upstream of km 14 in the inflowing freshwater (Sanford et al., 2001: Fig. 5). The ETM in the model result also is more distinctive than the data. Immediately after very large freshwater discharge events, the model fails to reproduce the high SS concentrations in the incoming freshwater in the very upstream reach (e.g., February 1 in Fig. 4-4 and October 22 in Fig. 4-5). The discrepancy seems to be due, at least in part, to the misalignment in time between the model and data (Section 5-1-2).

The model result for October 27 (Fig. 4-6) agrees very well with the data (Sanford et al., 2001: Fig. 5). Both the model and data show that salinity upstream of km 70 was quite low with the limit of salt intrusion around km 35 and the distinctive ETM was formed with its center around km 35-37. The observed ETM, however, was spatially more widespread along the main channel, compared to the model result. The high sediment load accompanied by the high freshwater discharge several days ago had been effectively trapped in the ETM zone near the limit of salt intrusion, which is well reproduced by the model (Fig. 4-6).
Figures 4-7 and 4-8 show the model results, respectively, for July 21 and 26, the period of which was characterized by low freshwater discharges (Fig. 4-2). Compared to the data on October 27, salinity increased considerably in the region upstream of km 70 with strong stratification and the ETM existed farther upstream and the near-bottom SS concentrations were much lower in two July surveys (Sanford et al., 2001: Fig. 8). The model results agree well with the data for both surveys in terms of the location and magnitude of the ETM. The model also shows the presence of strong stratification but the limit of salt intrusion exists in the model somewhat farther upstream than the data. The model appears to reproduce the observed ETM well in dry periods.

4-2-2. Intratidal Variations

Sanford et al. (2001: Fig. 10) presented the 25-h (14:00 on October 24 till 16:00 on October 25) time-depth contours of salinity, SS and along channel currents from one channel station in the ETM zone (Fig. 2-1). The station is located almost exactly, on the average, at the limit of salt intrusion. The corresponding model result (Fig. 4-9) exhibits almost identical patterns of the observed variations in salinity, SS concentration, and currents.

As ebb current proceeded, the salt water was advected downstream, making the whole water column fresh towards slack-before-flood. As the tidal current turned to flood, the salt water was advected upstream, increasing the bottom salinity up to its maximum near slack-before-ebb. The velocity structure at this location near the limit of salt intrusion was affected by stratification and gravitational circulation, such that vertical velocity shear decreased and the height of maximum current was lowered on flood, while vertical velocity shear increased and the height of maximum current was raised on ebb. Gravitational circulation
also caused the near-bottom currents to turn to flood sooner and to ebb later than
the near-surface currents, with the latter much more distinct than the former.
Bottom SS concentrations increased or decreased due to local resuspension or
deposition in phase with the near-bottom velocity. Resuspension was more
pronounced at maximum flood than ebb, which Sanford et al. (2001) attributed to
the enhanced flood currents below the limit of salt intrusion by gravitational
circulation. The flood-ebb difference in SS concentrations is not as distinct in the
model as in the data, which may be due to the difference in depth. The model
depth (~8 m in Fig. 4-9), shallower than the depth of 11 m at the intratidal station
in Sanford et al. (2001: Fig. 10), may underestimate baroclinic pressure gradient,
resulting in relatively weak landward flowing gravitational circulation in the
bottom layer. The model depths in the region upstream of km 50 (Figs. 4-4 to 4-
8) generally are shallower than the depths observed in surveys (Sanford et al.,
2001: Figs. 4, 5 and 8). Both the model and data show the high bottom
concentrations limited to the lower layer on flood but elevated surface
concentrations, although not as large as bottom concentrations, on ebb.

A discrepancy does exist between the model and data in terms of time
shifting. The model results (Fig. 4-9) lag the data (Sanford et al., 2001: Fig. 10)
by about 3 hours, although the model reproduces almost all of the relative
variations in the data. The applicability of the present model will be limited by
this discrepancy in time shifting and those noted above for February 1 (Fig. 4-4)
and October 22 (Fig. 4-5), which is discussed in Section 4-2-4.

4-2-3. Time-Series Comparison

The SS transport model is verified with the CBP monitoring data from the
ten stations along the main channel, including CB1.1, CB2.1, CB2.2, CB3.1,
CB3.2, CB3.3C, CB4.1C, CB4.2C, CB4.3C, and CB4.4 (Fig. 2-1). The model calculated hourly averages are compared with the CBP monitoring data for surface and bottom. Examination of the CBP monitoring data suggests that the ten stations may be grouped into four. The “boundary stations” include the stations CB1.1 and CB4.4 near the river and open boundaries, respectively. The “ETM stations” are those in the ETM zone and include the stations CB2.1, CB2.2, and CB3.1. The “transition stations” are those in the transition region between the ETM zone and the downstream stations and include the stations CB3.2 and CB3.3C. The “downstream stations” are those located farther downstream and include the stations CB4.1C, CB4.2C, and CB4.3C. The model-data comparison in Fig. 4-10 for the stations CB1.1 and CB4.4 shows the reliability of the river and open boundary conditions used in the present model application.

Figure 4-11 shows the model-data comparison for the ETM stations, CB2.1, CB2.2, and CB3.1. The SS concentrations, especially the bottom concentrations, in the ETM zone are much higher than those either upstream or downstream. The data indicate that the bottom concentrations increase from CB 1.1 (Fig. 4-10a) to CB2.1 (Fig. 4-11a) and keeps increasing downstream to CB2.2 (Fig. 4-11b), then decreasing slightly to CB3.1 (Fig. 4-11c). The model generally gives a good reproduction of the observed spatial variation in the bottom concentrations. The model results show that the concentrations, particularly the bottom concentrations, are strongly influenced by freshwater discharge as well as by tidal current. The model however underestimates the surface concentrations in summer from June till early September. Lin et al. (2003) in the SS transport modeling of Baltimore Harbor reported the same problem of underestimation of the observed surface concentrations. They attributed the discrepancy to the fact that the model did not include the phytoplankton biomass that may considerably contribute to the surface SS.
Figure 4-12 shows the SS concentrations for the fine, medium, and coarse classes at the station CB2.2. The fine particles (Fig. 4-12a) are mostly affected by freshwater discharges with little tidal variation and little vertical gradient between the surface and bottom. The medium particles (Fig. 4-12b) exhibit large tidal variations as well as large variations due to freshwater discharges. The water-bed interaction of erosion and deposition plays an important role for the medium particles that are alternately suspended and deposited by current velocity. The bottom concentration is much higher than the surface concentration, as have been observed by many researchers (Schubel, 1968a and 1968b; Schubel, 1971b; Schubel et al., 1978; Sanford et al., 2001). The coarse particles are resuspended into the bottom water only during the times of high freshwater discharges and virtually no coarse particles exist in the surface water (Fig. 4-12c). Figure 4-12 indicates that the medium particles are the dominant SS particles in water column in the ETM zone.

Figure 4-13 shows the model-data comparison for the transition stations, CB3.2 and CB3.3C. The SS concentrations are lower than those in the ETM stations (Fig. 4-11). The model gives a good reproduction of the data, although slight overestimation of the bottom concentration at the station CB3.3C is noted (Fig. 4-13b). The model results show large variations due to freshwater discharges and tidal currents in the bottom concentration. The bottom concentration is affected by local erosion and deposition, although the effect is not as pronounced as in the bottom concentrations of the ETM stations (Fig. 4-11). The surface concentration exhibits small variations due to freshwater discharges and virtually no tidal variations.

Figure 4-14 shows the SS concentrations for the fine, medium, and coarse classes at the station CB3.2. The general trend for all three size classes in Fig. 14 is more or less the same as that for the ETM stations (Fig. 4-12), with the effect of
freshwater discharge getting smaller and tidal variations getting larger. The fine particles constitute the background concentrations (Figs. 4-13a and 4-14a). The medium particles exhibit large tidal variations associated with erosion and deposition (Fig. 4-14b). Appreciable contribution of the coarse particles to the bottom SS concentration exists only during the times of very high freshwater discharges (Fig. 4-14c). Both the fine and medium particles are, on the average, equally important in contributing to the water column concentrations in this transition zone.

Figure 4-15 shows the model-data comparison for the downstream stations, CB4.1C, CB4.2C, and CB4.3C. The data show that the SS concentrations are quite low, much lower than the transition (Fig. 4-13) and ETM (Fig. 4-11) stations. The concentrations remain relatively constant throughout the year, with little vertical difference between the surface and bottom, particularly in dry summer. The model gives a good reproduction of the observed characteristics. Almost all particles in this region are composed of the fine particle sizes that constitute the background concentrations (Figs. 4-15b and 4-16a), demonstrating the reliability of allocation of 20% of the river loading to the fine class (Section 4-1-3). The solitary concentration peaks in Fig. 4-15 are the result of erosion and deposition of mostly the medium class particles, which however are much smaller compared to the transition and the ETM stations (Figs. 4-12b and 4-14b).

4-2-4. Comments on Model Calibration and Verification

The present model application seems to give a good description of the observed features of the ETM in the Upper Chesapeake Bay. The model results in general are in a good agreement with the observed axial (x-z) distributions of SS concentration (Figs. 4-5 to 4-9) and with the observed intratidal variations (Fig. 4-
10). Favorable comparison exists between the model results and the time-series data of the CBP Monitoring Program (Figs. 4-11, 4-13, and 4-15). The model results in Figs. 4-11 to 4-15 also demonstrate that the three size classes of the fine, medium, and coarse particles, with the corresponding settling velocities, appear to behave as they are intended to (Section 4-1-3).

There are some discrepancies between the model and data, however. The model, immediately after very large freshwater discharge events, fails to reproduce the high SS concentrations in the incoming freshwater in the very upstream reach (Fig. 4-4 and Fig. 4-5). The discrepancy seems to be due, at least in part, to the misalignment in time between the model and data (Section 5-1-2). Another discrepancy in time shifting is noted in Fig. 4-9, where the model results lag the data in Sanford et al. (2001: Fig. 10) by about 3 hours, although the model reproduces almost all of the relative variations in the data. Finally, the model underestimates the surface SS concentrations in summer from June till early September in the ETM zone (Fig. 4-11).

When using the present model to study the characteristics of the ETM in the Upper Chesapeake Bay, the above-noted shortcomings of the model are kept in mind. No attempt has been made with the present model to apply it beyond its applicability. The model has been used in Chapter 5 to compare the variations in the ETM between high and low freshwater discharges and within a tidal cycle, so that a few days or a few hours of time shifting would not affect the analysis. Since the primary analysis has been on the ETM, which tends to be confined to the lower layer during dry periods (Fig. 5-3), the low surface concentrations in dry summer periods have not been considered in Chapter 5.
Fig. 4-1. Model grid with bottom bathymetry
Fig. 4-2. Daily freshwater discharge rates for the Susquehanna River: USGS data
Fig. 4-3. Daily SS loads from the Susquehanna River: the CBP watershed model results
Fig. 4-4. Axial distributions of the model results of hourly average salinity (a) and SS concentration (b) along the main channel at slack-before-flood on February 1, 1996: compared to Fig. 4 in Sanford et al. (2001)
Fig. 4-5. Axial distributions of the model results of hourly average salinity (a) and SS concentration (b) along the main channel at flood current on October 22, 1996: compared to Fig. 5 in Sanford et al. (2001)
Fig. 4-6. Axial distributions of the model results of hourly average salinity (a) and SS concentration (b) along the main channel at ebb current on October 27, 1996: compared to Fig. 5 in Sanford et al. (2001)
Fig. 4-7. Axial distributions of the model results of hourly average salinity (a) and SS concentration (b) along the main channel at ebb current on July 21, 1996: compared to Fig. 8 in Sanford et al. (2001)
Fig. 4-8. Axial distributions of the model results of hourly average salinity (a) and SS concentration (b) along the main channel at ebb current on July 26, 1996: compared to Fig. 8 in Sanford et al. (2001)
Fig. 4-9. Isopleths in a depth-time plane of the model results of hourly average salinity (a), SS concentration (b), and along channel currents with positive velocity indicating flood (c) for October 24-25, 1996 at the intratidal station (ū in Fig. 2-1): compared to Fig. 10 in Sanford et al. (2001)
Fig. 4-10. Time-series comparison of the model results of hourly average SS concentration at surface (blue solid line) and bottom (red solid line) with the CBP monitoring data at surface (●) and bottom (△) at the boundary stations, CB1.1 (a) and CB4.4 (b)
Fig. 4-11. Time-series comparison of the model results of hourly average SS concentration at surface (blue solid line) and bottom (red solid line) with the CBP monitoring data at surface (●) and bottom (△) at the ETM stations, CB2.1 (a), CB2.2 (b), and CB3.1 (c)
Fig. 4-12. Model results of hourly average SS concentration for the fine (a), medium (b), and coarse (c) classes at surface (blue solid line) and bottom (red solid line) at the station CB2.2, with the CBP monitoring data at surface (●) and bottom (△)
Fig. 4-13. Time-series comparison of the model results of hourly average SS concentration at surface (blue solid line) and bottom (red solid line) with the CBP monitoring data at surface (●) and bottom (△) at the transition stations, CB3.2 (a) and CB3.3C (b)
Fig. 4-14. Model results of hourly average SS concentration for the fine (a), medium (b), and coarse (c) classes at surface (blue solid line) and bottom (red solid line) at the station CB3.2, with the CBP monitoring data at surface (●) and bottom (△)
Fig. 4-15. Time-series comparison of the model results of hourly average SS concentration at surface (blue solid line) and bottom (red solid line) with the CBP monitoring data at surface (●) and bottom (△) at the downstream stations, CB4.1C (a), CB4.2C (b), and CB4.3C (c)
Fig. 4-16. Model results of hourly average SS concentration for the fine (a), medium (b), and coarse (c) classes at surface (blue solid line) and bottom (red solid line) at the station CB4.2C, with the CBP monitoring data at surface (●) and bottom (△).
5. MODEL STUDY OF ESTURINE TURBIDITY MAXIMUM

The results of the model application for 1996 are used to study the characteristics of the ETM in the Upper Chesapeake Bay. Both the model and data exhibit large intertidal variations mostly correlated with the freshwater discharges from the Susquehanna River (Figs. 4-4 to 4-8 and 4-10 to 4-16) and large intratidal variations mostly correlated with tidal phase (Fig. 4-9). Variations in the characteristics and its controlling processes of the ETM are examined in both intertidal (Section 5-1) and intratidal (Section 5-2) time scales.

5-1. Variations in Intertidal Time Scales

5-1-1. Annual Variations in 1996

The isopleths of tidal average model results in a distance-time plane for the bottom water in 1996 (Fig. 5-1) show the variations in salinity and SS concentration largely in response to the variations in freshwater discharges. The enormous January flood (Fig. 4-2) made the upper 50 km reach, relative distance downstream from Havre De Grace, virtually fresh and also resulted in high SS concentration in the upper 60 km reach. After the January flood, salinity strongly rebounded upstream, pushing the limit of salt intrusion close to km 20 in early February and the ETM followed behind with its center occurring between km 21-35. Sanford et al. (2001) suggested that the non-tidal movement of the salt front lead to the separation of the limit of salt intrusion and the ETM, with the ETM lagging behind the limit of salt intrusion. Several high freshwater discharge events between late February and mid May moved the limit of salt
intrusion up and down the channel between km 25-50. SS concentration also shows large variations, but the downstream limit of the ETM appeared to be trapped in the region between km 40-42. The longitudinal distributions of SS in Schubel (1974) showed the markedly steep longitudinal gradient between km 40-45. He suggested the region as the seaward boundary of the ETM during high flow conditions and attributed it to the blocking of the SS loading by the limit of salt intrusion.

The high freshwater inflow in late October (Fig. 4-2) resulted in the downstream shift of the limit of salt intrusion and the ETM, again with the limit of salt intrusion leading the ETM (Sanford et al., 2001). As the direct effect of the high SS loading diminished with decreasing freshwater inflow, a distinct, strong ETM was formed in the region between km 30-40 toward the end of October. The variations in the limit of salt intrusion and the ETM with decreasing freshwater discharge after the large flood event on October 21 are examined in more detail in the next section (Section 5-1-2).

During the prolonged low discharges between June and September, the salt water intruded upstream, even close to km 10 in early September. Low SS loading resulted in a relatively weak ETM developed and lasted a few days in the region between km 20-26, with a period of about 30 days. The Tide Tables show that the tides in Chesapeake Bay have such a characteristic that the successive spring tides often alternate in strength between strong and weak springs (e.g., Kuo et al., 1991). With tidal resuspension becoming an important mechanism in developing an ETM in dry periods, strong spring tides (▼ in Fig. 5-1) generated bed shear stress strong enough to develop the ETM.

5-1-2. Effects of Susquehanna River Discharge in Late October, 1996
The model results in late October of 1996 are examined to investigate the correlation between the limit of salt intrusion and the ETM. Figure 5-2 presents the daytime tidal average salinity and SS concentration on October 21, 23, 25, 27, 29 and 30. Heavy rains resulted in a large pulse of freshwater flow that peaked at 5,267 m$^3$ s$^{-1}$ on October 21 (Fig. 4-2). The survey on October 22 in Sanford et al. (2001: Fig. 5) caught high freshwater discharge and its associated SS loads entering the upper part of the ETM zone. It was noted that the model result on October 22 (Fig. 4-5) does not reproduce the observed pattern in SS concentration well (Section 4-2-4). The model result on October 21 (Figs. 5-2b), however, reproduces the observed pattern of the highest SS concentration on October 22 distributed nearly homogeneous in the vertical upstream of km 14 in the inflowing freshwater (Sanford et al., 2001: Fig. 5).

The high freshwater discharges since October 21 kept pushing the limit of salt intrusion and high SS loads downstream till October 25 (Figs. 5-2a to 5-2f), at which both the limit of salt intrusion and the center of the ETM existed around km 36. During this five day period, the downstream shift of the limit of salt intrusion led the center of the ETM (Sanford et al., 2001) and the downstream limit of the ETM zone could not extend seaward beyond km 40-42 with a very sharp gradient in the downstream boundary (Schubel, 1974). After October 25, the limit of salt intrusion and the ETM had rebounded toward upstream, again with upward movement of the limit of salt intrusion leading the center of the ETM With diminishing freshwater discharges, the magnitude of the ETM kept decreasing, with its spatial extent and strength decreasing (Figs. 5-2e to 5-2l). On October 30, the limit of salt intrusion existed around km 29 and the center of the ETM around km 31-32 (Fig. 5-2k and 5-2l). The model results in Fig. 5-2 demonstrate that the high SS loads entering with high freshwater discharges of the
Susquehanna River during heavy rains were effectively trapped in the upper reaches, thus resulting in the distinctive formation of the ETM.

5-1-3. High Flow versus Low-to-Moderate Flow Conditions

The statistics based on the USGS daily average discharge in 34 years (1967-2001) for the Susquehanna River give 1139 m$^3$ s$^{-1}$ for the mean daily flow with the range of 8-31,715 m$^3$ s$^{-1}$ (http://nwis.waterdata.usgs.gov/md/nwis/discharge?site_no=01578310&agency_cd=USGS&format=rdb&begin_date=&end_date=&period=). The 25, 50 and 75 percentile flows, respectively, are 374, 762 and 1,450 m$^3$ s$^{-1}$. The model calibration and verification show that the distributions of SS concentration in the Upper Chesapeake Bay have distinctive characteristics in response to the Susquehanna River discharges. To study the controlling processes of the ETM, the model results in two different freshwater flow regimes are examined. The high flow (HF) condition represents the spring freshet and other brief periods of high flow and the low-to-moderate flow (LMF) condition represents most of the reminder of the year. The 15-day period from July 23 to August 6, with average daily discharge of 719 m$^3$ s$^{-1}$ (between 25 percentile and mean, is considered as the LMF condition and the 15-day period from October 19 to November 2, with average of 2,418 m$^3$ s$^{-1}$ (higher than 75 percentile) as the HF condition. The model results averaged over each of the 15-day periods are considered to represent the LMF and HF conditions.

The axial distributions of salinity and SS concentration along the main channel for the LMF (Fig. 5-3) and HF (Fig. 5-4) conditions illustrate much of the observed characteristics of the limit of salt intrusion and the ETM in the Upper Chesapeake Bay. For the discussion given below, the ETM zone, characterized by relatively high SS concentration compared to either upstream or downstream,
is defined with the contour line of 40 mg l\(^{-1}\) for its spatial extent. For the LMF condition (Fig. 5-3), the limit of salt intrusion is located around km 20 and the ETM zone between km 17-32 with the center of the ETM occurring at km 22. For the HF condition (Fig. 5-4), the limit of salt intrusion and the ETM are located farther downstream, respectively, around km 30 and km 21-41, with the center of the ETM around km 36-38. The maximum SS concentration at the center of the ETM is higher for the HF condition than for the LMF condition. The ETM is confined to the bottom layer only under the LMF condition, while the ETM extends to the surface under the HF condition, despite the stronger vertical stratification for the HF condition than for the LMF condition (Schubel, 1972 and 1974). The SS concentration decreases away from the center of the ETM in both the upstream and downstream directions, but more rapidly in the downstream than in the upstream direction for the HF condition (Schubel, 1972).

5-1-3a. Two-Layer Gravitational Circulation

The axial distributions of 15-day average along-channel velocity (principal axis component) show the presence of two-layer gravitational circulation for both the LMF and HF conditions (Fig. 5-5). Both flow conditions show seaward flowing residual current at the surface layer. At the lower layer, the residual current flows seaward at the upstream of the null zone but landward at the downstream of the null zone, resulting in a horizontal convergence at the bottom of the null zone (Dyer, 1988). Difference, however, exists in the location of the null zone, which meets the bottom around km 20 and km 36-38, respectively, for the LMF and HF conditions. During the LMF condition, the null zone coincides with the limit of salt intrusion, resulting in gravitational circulation existing in the entire saline portion. During the HF condition, however, the null zone occurs
farther downstream than the limit of salt intrusion, leaving the upstream end of the saline portion between km 30-36 under the direct influence of seaward flowing high freshwater discharges from the Susquehanna River.

The axial distributions of 15-day average along-channel SS flux for the LMF and HF conditions (Fig. 5-6) exhibit generally the same pattern as residual current (Fig. 5-5). In both flow conditions, however, the strongest upstream residual SS flux in the lower layer occurs just downstream of the null zone, where the residual currents are not their maxima. Furthermore, the magnitude of the upstream residual SS flux just downstream of the null zone is more or less the same for both flow conditions, but the downstream SS flux just upstream of the null zone is much stronger for the HF condition, resulting in the stronger convergence of SS and thus contributing to the higher SS concentration for the HF condition than for the LMF condition (Figs. 5-3b and 5-4b). Strong convergence of SS exists at the null zone, which coincides with the center of the ETM, regardless of the flow conditions. Two-layer gravitational circulation, therefore, is an important process affecting the ETM, particularly with respect to the intertidal variations in location and strength of the ETM zone.

5-1-3b. Stratification and Vertical Fluxes

Potential energy anomaly (PEA), an index of the vertical stratification, is the work required to bring about complete mixing and may be calculated as (Simpson et al, 1990)

\[
\text{PEA} = \frac{g}{H} \int_{-h}^{\eta} (\rho_m - \rho)z \cdot dz
\]  

(5-1)

\[H = \text{total depth} = \eta + h\]
\( \eta \) and \( h \) = surface and bottom elevation, respectively  
\( \rho \) = density at a depth \( z \) from the mean sea surface  
\( \rho_m \) = mean density over the total depth.

The larger the PEA is, the stronger the stratification of the water column.

For both flow conditions, PEA is virtually zero upstream of the limit of salt intrusion and increases downstream limiting vertical mixing (Figs. 5-7a and 5-8a). For the LMF condition (Fig. 5-7), diffusive flux, which is always upward, is essentially zero in saline water, except close to the bottom, probably due to suppression by stratification, lack of mixing energy, and small vertical gradient in SS concentration. Settling flux, which is always downward, is larger than diffusive flux, particularly in the lower layer of the ETM zone between km 17-32, which probably results in the confinement of the ETM to the lower layer only under the LMF condition (Fig. 5-3b). For the HF conditions (Fig. 5-8), both diffusive and settling fluxes are much stronger, particularly below about 3 m depth. Strong upward diffusive flux occurs upstream of about km 36, downstream of which the relatively large PEA suppresses vertical mixing. Strong downward settling flux, however, extends farther downstream to about km 41. The difference in downward settling and upward diffusive fluxes between km 36-41 may explain the asymmetrical spatial extent of the ETM for the HF condition (Fig. 5-4b): the SS concentration decreases away from the center of the ETM at km 36-38 more rapidly in the downstream than in the upstream direction.

The settling in the downstream region of the ETM zone, where vertical mixing is suppressed due to stratification, makes the sediments subject to landward flowing gravitational circulation in the lower layer. Otherwise, the sediments may have escaped the system by seaward flowing surface water. For both flow conditions, settling flux is larger than diffusive flux downstream of the center of the ETM. Stratification, therefore, is an important process affecting the
ETM, particularly with respect to the intertidal variations in the spatial extent, both longitudinal and vertical, and the trapping of the sediments within the ETM zone. Sanford et al. (2001) discussed that seasonal changes in settling velocity may be an important factor affecting the trapping efficiency of the SS loads within the ETM zone. The present model application does not account for such changes in settling velocity and employs constant partitioning fractions for the river loads (20:75:5 for fine, medium and coarse classes in Section 4-1-3), which might be, at least in part, responsible for the model’s poor behavior in reproducing the observed conditions immediately after large discharge events.

5-1-3c. Erosion and Deposition

The along-channel distributions of 15-day average bed shear stress, erosional flux, and depositional flux are presented in Figs. 5-9 and 5-10, respectively, for the LMF and HF conditions. For the LMF condition, bed shear stress is relatively high in the upper 43 km reach. For the HF condition, bed shear stress increases rather substantially between km 20-35, the region where gravitational circulation used to exist during the LMF condition. Erosional flux is higher between km 18-36 for the HF condition than for the LMF condition and the same trend exists for depositional flux between km 18-42. Fain et al. (2001) showed the increasing of bed shear stress with increasing river discharges. Increased bed shear stress during the HF condition will enhance erosional flux, resulting in increase in the SS concentration in the water column and then in depositional flux, especially with continuous deposition employed in the present model (Section 3-2-4). With more newly deposited, unconsolidated, easily erodible sediments at the bed, more erosion will occur subsequently, especially with depth-limited erosion employed in the present model (Section 3-2-4). More
active water-bed interaction via erosion and deposition maintains higher water column concentration in the ETM for the HF condition than for the LMF condition.

Maximum bed shear stress exists around km 30-32 for both flow conditions, which is downstream of the center of the ETM for the LMF condition (km 22 in Fig. 5-3b) but upstream of the center of the ETM for the HF condition (km 36-38 in Fig. 5-4b). For the LMF condition (Fig. 5-9), maximum fluxes of both erosion and deposition occurring near km 22, result in the formation of the center of the ETM at the same location, just downstream of the null zone at km 20 (Fig. 5-5a). For the HF condition (Fig. 5-10), maximum resuspension of bed sediments by maximum erosion near km 28 and subsequent downstream transport by strong seaward currents (the null zone exists at km 36-38 for the HF condition in Fig. 5-5b) and maximum deposition near km 34 combine to result in the formation of the center of the ETM near km 36-38. The separation of the locations of maximum bed shear stress and maximum erosion, indicating that erosion is not entirely determined by bed shear stress alone, has been observed (Nicholson and O’Connor, 1986; Mehta et al., 1989; Sanford and Maa, 2001). With depth-limited erosion in the present model, erosion depends not only on bed shear stress but availability of unconsolidated, erodible bed sediments (Figs. 3-2 and 3-3).

The bottom velocity distributions in the Upper Chesapeake Bay are such that relatively strong bed shear stress, and thus strong erosion and deposition, exists upstream of about km 43. Their magnitudes generally increase with increasing freshwater discharges, but the degree of increase is not uniform among parameters and locations. The relative positions of the locations of strong erosion, strong bottom convergence of SS, and large deposition, which may vary with freshwater discharges, combine to affect the intertidal variations in location
and strength of the ETM zone. The water-bed interaction via erosion and deposition, therefore, is another important process affecting the ETM.

5-1-3d. Budget Analysis

A budget analysis for SS is conducted for the LMF and HF conditions. Three reaches along the main channel are considered and each reach has two layers vertically, resulting in six boxes (Fig. 5-11). The boxes U1, U2, and U3 are the boxes in the upper layer, and L1, L2, and L3 in the lower layer. The most upstream reach (U1 and L1) represents the ETM zone for the LMF condition and the middle reach (U2 and L2) represents the ETM zone for the HF condition, with the most downstream reach (U3 and L3) representing the downstream portion of the ETM zone.

The results in Fig. 5-11 show the presence of two-layer gravitational circulation and its effect on the location of the ETM (Section 5-1-3a). The longitudinal SS transport is to the downstream at all cross-sections of the upper layer for both flow conditions. For the LMF condition (Fig. 5-11a), the bottom convergence occurs at the box L1 where the ETM exists, with the largest contribution of SS from the longitudinal transport from downstream. For the HF condition (Fig. 5-11b), the bottom convergence occurs at the box L2, indicating the downstream shift of the null and ETM zone with increasing freshwater discharges. The ETM extends to the upper layer, the box U2, for the HF condition (Fig. 5-4b). For the box L2 (Fig. 5-11b), the largest contribution of SS is from the longitudinal transport from upstream, which is significantly augmented by the eroded materials in the box L1 (Section 5-1-3c). For the box U2 (Fig. 5-11b), the largest contribution again is from the horizontal transport
from upstream, maintaining the high SS concentration in the upper layer during the HF condition.

The net vertical transport between the upper and lower boxes for both flow conditions is upward in the upstream region down to the ETM zone, but downward downstream of the ETM zone due to suppression of vertical mixing by stratification (Fig. 5-11). The sediments transported downward to the lower layer are moved upstream by landward flowing gravitational circulation, thus trapping the sediments from escaping from the system (Section 5-1-3b).

The net transport at the water-bed interface indicates that sedimentation occurs in the ETM zone for both flow conditions (Fig. 5-11), consistent with previous studies (Officer and Nichols, 1980; Officer et al., 1984). The sedimentation rate in the ETM zone is 1.1 g cm\(^{-2}\) yr\(^{-1}\) for the LMF condition (box L1) and about 4.0 g cm\(^{-2}\) yr\(^{-1}\) for the HF condition (box L2), which is on the same order of magnitudes with the previous estimates. Estimates from previous studies in the ETM zone of Chesapeake Bay include 0.6 cm yr\(^{-1}\) in Schubel (1968b) and 0.45 cm yr\(^{-1}\) excluding flood events in Hirschberg and Schubel (1979); they respectively are equivalent to 1.6 and 1.2 g cm\(^{-2}\) yr\(^{-1}\) with sediment density of 2.65 g cm\(^{-3}\). Officer et al. (1984) reported the sedimentation rates of 0.3-1.2 g cm\(^{-2}\) yr\(^{-1}\) in the ETM zone. Based on annual maintenance dredging records from the upper 48 km of the shipping channel, Sanford et al. (2001) argued that the annual sedimentation rate in the main channel may be much higher than that on the adjacent shoals.

5-2. Variations in Intratidal Time Scales

5-2-1. Intratidal Variation in x-z Plane
Intratidal variations in salinity and SS concentration are examined with the hourly average model results on July 21, 1996 (Fig. 5-12) when the freshwater discharge was quite low (Fig. 4-2). At maximum ebb current (Figs. 5-12a and 5-12b), the limit of salt intrusion was located around km 22, with the center of the ETM somewhat lagging behind around km 20-22. Toward the end of ebb current at slack-before-flood (SBF), both the salt water and the ETM were pushed downstream, with the strength of the ETM much weakened (Figs. 5-12c and 5-12d). Flood current pushed the salt water upstream and at maximum flood current, the ETM, higher in concentration and over wider spatial extent than that at maximum ebb, was developed below the limit of salt intrusion (Figs. 5-12e and 5-12f). At slack-before-ebb (SBE), the limit of salt intrusion and a much weakened ETM were formed in a little upstream location (Figs. 5-12g and 5-12h). At both maximum ebb and flood, the ETM is confined to the lower layer only, indicating the absence of large SS loads from upstream and weak mixing. The tidal resuspension lag of the limit of salt intrusion leading the center of the ETM in the direction of tidal current, discussed in Dyer and Evans (1989), is well illustrated in Fig. 5-12 for July 21 when the tidal effect was dominant with sustained low freshwater discharge.

Figure 5-13 presents the model results analogous to those in Fig. 5-12, except for October 29, 1996 when relatively strong gravitational circulation existed after the high discharge event on October 21 (Fig. 5-2). Unlike Fig. 5-12 the tidal resuspension lag was not evident in Fig. 5-13. The model results in Fig. 5-13 rather are, at least qualitatively, identical to the conceptual diagram in Sanford et al. (2001: Fig. 11), in which they discussed tidal asymmetry caused by the interaction between tidal currents, gravitational circulation, and stratification. At maximum ebb, gravitational circulation enhances ebb current upstream of the limit of salt intrusion but opposes it below the limit of salt intrusion, resulting in
the ETM developed in the upstream of the limit of salt intrusion where the suspended sediments are mixed upward into the unstratified water column. At maximum flood, however, gravitational circulation enhances flood current below the limit of salt intrusion but opposes it upstream of the limit of salt intrusion, resulting in the ETM developed in the downstream of the limit of salt intrusion where the suspended sediments are confined to the lower layer due to stratification. The model results indicate that either tidal resuspension lag (Fig. 5-12) or tidal asymmetry (Fig. 5-13), or a combination of both, may occur depending on the relative importance of tidal currents, gravitational circulation, stratification, and other forcing functions.

5-2-2. Intratidal Variations in t-z Plane

Figure 5-14 to 5-16 present the hourly average model results for salinity, SS concentration, and along-channel currents at three locations (○ in Fig. 2-1) for July 21 when the freshwater discharge was very low (Fig. 4-2). The limit of salt intrusion existed around km 20-22 on July 21 (Fig. 5-12). The most upstream location at km 18 was in freshwater reach with zero salinity (Fig. 5-14). The SS was nearly homogeneous, with slight increase in bottom concentration at the times of maximum ebb and flood currents, but the concentration was quite low, always less than 25 mg l⁻¹. There also was no vertical gradient in current velocity, with the maximum velocity less than 30 cm s⁻¹ for both ebb and flood currents. The next downstream location at km 24 (Fig. 5-15) was just downstream of the limit of salt intrusion, at which the patterns of variations are almost identical to those at the most downstream location at km 31 (Fig. 5-16) and also to those in Fig. 4-9.

Since the two tidal cycles in Fig. 5-14 to 5-16 repeat similar sequence, the following discussion starts from about 06:00 on July 21 when ebb current started.
As ebb current proceeded, the salt water is advected downstream, making the water column least stratified towards SBF. As the tidal current turned to flood, the salt water was advected upstream, increasing the bottom salinity up to its maximum near SBE. Due to the effects of gravitational circulation and stratification, the velocity structures show that the height of maximum current occurred lower in the water column with decreased vertical velocity shear on flood than on ebb and that the near-bottom currents turned to flood sooner and to ebb later than the near-surface currents. The differences in velocity structures were more evident at km 31 (Fig. 5-16c) than at km 24 (Fig. 5-15c), since the maximum landward residual velocity along the bottom occurs not at but farther downstream of the limit of salt intrusion (Fig. 5-5).

Bottom SS concentrations increased or decreased in phase with the near-bottom velocity, indicating the effects of local resuspension and deposition (Figs. 5-15b and 5-16b). The tidal asymmetry in bottom currents resulted in the larger bed shear stress and thus resuspension on flood than on ebb. Maximum bottom concentrations occurred at the mid-point in the accelerating current, mostly between SBF and maximum flood (Sanford and Maa, 2001). At km 24 and km 31, the ETM was not always developed on ebb currents because of too weak a bottom current: the lack of easily erodible bed sediments on ebb currents could not be responsible since the high bottom concentrations in the preceding flood currents had deposited to the bed during SBE. Figures 5-12, 5-15b and 5-16b show large intratidal variations in SS distribution, with the ETM stronger in strength and more widespread along the channel on maximum flood than on maximum ebb. Hence, caution should be taken in interpreting the SS data for spatial distributions if the data had not been collected at the same tidal phases for all stations. Such large intratidal variations in the SS distribution are more likely to occur during dry periods when tidal forcing tends to be dominant.
Fig. 5-1. Isopleths in a distance-time plane of the tidal average salinity and SS concentration for the bottom water in 1996: ▼ indicates times of strong spring tides
Fig. 5-2. Isopleths in a distance-depth plane of the daytime tidal average salinity and SS concentration for October 21, 23, 25, 27, 29 and 30, 1996
(e) October 25, 1996: Salinity (psu)

(f) October 25, 1996: SS (mg/l)

(g) October 27, 1996: Salinity (psu)

(h) October 27, 1996: SS (mg/l)

Fig. 5-2. Continued
Fig. 5-2 Continued
Fig. 5-3. Isopleths in a distance-depth plane of the 15-day average salinity (a) and SS concentration (b) for the low-to-moderate flow condition.
Fig. 5-4. Isopleths in a distance-depth plane of the 15-day average salinity (a) and SS concentration (b) for the high flow condition.
Fig. 5-5. Isopleths in a distance-depth plane of the 15-day average along-channel velocity for the low-to-moderate flow (a) and high flow (b) conditions: positive velocities indicate flood
Fig. 5-6. Isopleths in a distance-depth plane of the 15-day average along-channel SS flux for the low-to-moderate flow (a) and high flow (b) conditions: positive values indicate landward flux.
Fig. 5-7. Along-channel distribution of 15-day average and range of PEA (a) and isopleths of 15-day average vertical diffusive flux (b) and settling flux (c) in the water column for the low-to-moderate flow condition: positive values indicate upward vertical flux.
Fig. 5-8. Along-channel distribution of 15-day average and range of PEA (a) and isopleths of 15-day average vertical diffusive flux (b) and settling flux (c) in the water column for the high flow condition: positive values indicate upward vertical flux.
Fig. 5-9. Along-channel distributions of the 15-day average bed shear stress (a), erosional flux (b), and depositional flux (c) for the low-to-moderate flow condition.
Fig. 5-10. Along-channel distributions of the 15-day average bed shear stress (a) erosional flux (b), and depositional flux (c) for the high flow condition.
Fig. 5-11. A budget analysis for the low-to-moderate flow (a) and high flow (b) conditions: insert shows the location of three reaches. Values are in tons d-1, except those in the parentheses in g cm-2 yr-1 for net flux of erosion and deposition.
Fig. 5-12. Isopleths in a distance-depth plane of the hourly average salinity and SS concentration on July 21, 1996 at full ebb (a), slack-before-flood (b), full flood (c), and slack-before-ebb (d)
Fig. 5-12 Continued
Fig. 5-13. Isopleths in a distance-depth plane of the hourly average salinity and SS concentration on October 29, 1996 at full ebb (a), slack-before-flood (b), full flood (c), and slack-before-ebb (d)
Fig. 5-13. Continued
Fig. 5-14. Isopleths in a depth-time plane of the hourly average salinity (a), SS concentration (b), and along channel currents with positive velocity indicating flood (c) for July 21, 1996 at km 18 (☐ in Fig. 2-1)
Fig. 5-15. Isopleths in a depth-time plane of the hourly average salinity (a), SS concentration (b), and along channel currents with positive velocity indicating flood (c) for July 21, 1996 at km 24 (□ in Fig. 2-1).
Fig. 5-16. Isopleths in a depth-time plane of the hourly average salinity (a), SS concentration (b), and along channel currents with positive velocity indicating flood (c) for July 21, 1996 at km 31 (□ in Fig. 2-1)
6. CONCLUSIONS

6-1. Conclusions and Summary

A suspended sediment (SS) transport model is developed to study the characteristics of the estuarine turbidity maximum (ETM) in the Upper Chesapeake Bay. The model, internally linked to the CH3D-WES model, is based on the mass-balance equations for three particle size classes, 3, 18, and 65 μm, representative of the characteristic size distributions of the Upper Chesapeake Bay. The first class consists of fine particles that are in more or less continual suspension in water column settling with a very small velocity (~0.007 mm s⁻¹) and thus constitute the background concentration. The second class consists of medium particles that are alternately eroded and deposited by tidal current. The last class consists of coarse particles that stay at the bed most of the time, except the time periods of relatively large bed shear stress. No interaction among different size classes and no effect of the SS concentration on the hydrodynamic field are assumed.

The present model employs an empirical formulation to account for depth-limited erosion by varying τce as a function of eroded mass, a modified version of the algorithm in Lin et al. (2003) based on the empirical relationship in Sanford and Maa (2001). A point model, in which the change in SS concentration is solely determined by erosion and deposition, is employed to compare continuous deposition and mutually exclusive erosion and deposition for three size classes and different types of erosion environment, limited and unlimited unconsolidated, erodible sediments (Figs. 3-1 to 3-3). The point model results show virtually identical results for the fine class regardless of the modeling method of erosion. Since the consolidation effect would be minimal for the fine class particles
because of their relatively short residence time in the bed, the present model employs for the fine class mutually exclusive erosion and deposition with small constant values for $\tau_{ce} (0.02 \text{ Pa} = \tau_{cd})$ to assure quick erosion of deposited fine particles, without allowing further erosion of consolidated, hard bed sediments. The point model results for the medium and coarse size classes show that continuous deposition gives more reasonable behavior in SS concentration than exclusive erosion and deposition. For the medium and coarse classes, hence, the depth-limited erosion with continuous deposition (Eq. 3-12) is employed by varying $\tau_{ce}$ as a function of accumulated eroded mass (Eqs. 3-9 and 3-10). The coefficient $b_E$ in Eq. 3-10b is considered as a calibration parameter for the medium and coarse classes (Lin et al., 2003).

The model is applied to the Upper Chesapeake Bay to simulate one-year conditions in 1996. The SS loading from the Susquehanna River is considered, for which the CBP watershed model output is used (Fig. 4-3). The model is calibrated with the intensive data in Sanford et al. (2001) and verified with the CBP monitoring data. The three size classes appear to behave as they are intended to and the model gives a reasonable reproduction of the data, not only for intertidal variations but for intratidal variations in salinity and SS concentration (Figs. 4-4 to 4-16). Model-data discrepancy did exist, however, particularly in time shifting between the model and data, and was noted (Section 4-2-4). The model results for 1996 are analyzed, within the applicability of the model, to investigate the characteristics of the ETM in intertidal and intratidal time scales.

Tidal average model results (Fig. 5-1) exhibit that both the limit of salt intrusion and the ETM migrated up and down largely in response to the Susquehanna River discharges. The high discharges pushed both downstream but the downstream limit of the ETM was trapped between km 40-42, relative distance downstream from Havre De Grace, marked by steep longitudinal SS
gradient (Schubel, 1974). The enormous flood in January was an exception, which resulted in high SS concentration throughout the upper 60 km reach. After the high freshwater discharges, salinity and SS rebounded upstream and the associated high SS loads entering were effectively trapped in the upper reaches, thus resulting in the distinctive formation of the ETM (Figs. 5-1 and 5-2). In migrating up and down the channel in response to discharges, the ETM lagged behind the limit of salt intrusion (Sanford et al., 2001). During the low flow conditions in June-September, a relatively weak ETM was developed and lasted a few days between km 20-26, with a period of about 30 days (Fig. 5-1), which coincided with the occurrence of strong spring tides. With tidal resuspension becoming an important mechanism in developing an ETM in dry periods, strong spring tides generated bed shear stress strong enough to develop the ETM.

The 15-day average model results are examined to investigate the controlling processes for each of the low-to-moderate flow (LMF) and high flow (HF) conditions. The limit of salt intrusion and the center of the ETM occur close to each other for the LMF condition (Fig. 5-3), while the HF condition pushes both downstream with the center of the ETM occurring farther downstream than the limit of salt intrusion (Fig. 5-4). Two-layer gravitational circulation (Fig. 5-5) results in strong bottom convergence of SS flux at the null zone (Figs. 5-6), which coincides with the center of the ETM for both flow conditions, and thus is an important controlling process, particularly with respect to the location and strength of the ETM zone. Vertical flux of SS determined by relative importance of downward settling flux and upward diffusive flux affects the longitudinal and vertical extent of the ETM zone that varies with freshwater discharges (Figs. 5-7 and 5-8). Settling flux in the downstream of the ETM where vertical mixing is suppressed by salinity stratification, particularly during high discharges, is important in trapping the sediments within the ETM zone. Relatively strong bed
shear stress and thus large erosion and deposition exist at the region upstream of about km 43 (Figs. 5-9 and 5-10). Because of non-uniform increase in their magnitudes among parameters and locations with increasing freshwater discharges, the relative positions of the locations of strong erosion, strong bottom convergence of SS, and large deposition combine to affect the intertidal variations in location and strength of the ETM zone.

A budget analysis based on 15-day averages reveals that the relative positions and strengths of bottom convergence of SS, net vertical flux of SS, erosion and deposition combine to affect the intertidal variations in location and strength of the ETM (Fig. 5-11). The relatively large bed shear stress during the HF condition results in net erosion in the upstream reach of the Upper Chesapeake Bay. The resuspended sediments and those from the Susquehanna River are transported downstream and much of them are settled down to form the ETM very close to the bottom convergence of SS flux that occurs at the null zone of gravitational circulation. The downstream transport, significantly augmented by the eroded materials upstream of the ETM, provides the largest amount of SS to the ETM zone and also is responsible for the extension of the ETM to the surface layer under the HF condition (Fig. 5-11b). The sediments, not trapped in the ETM zone and transported to the downstream of the ETM zone, are settled down to the lower layer because vertical mixing is suppressed by strong stratification, particularly during HF condition, in the saline water. As the freshwater discharges diminish, everything is rebounded back to upstream. Resuspension of the bed sediments deposited during the HF condition and the subsequent upstream transport results in the formation of the ETM at a farther upstream location. For the LMF condition, the largest contribution of SS to the ETM zone is the upstream transport of SS along the lower layer (Fig. 5-11a). A sequence of the above-mentioned processes results in the effective trapping of the sediments.
introduced either from the river, the sea, or recycled by the estuarine circulation near the null zone of the Upper Chesapeake Bay. Regardless of the flow conditions, the ETM zone is characterized with large deposition, with the sedimentation rate of about 1.1 and 4.0 g cm$^{-2}$ yr$^{-1}$, respectively, for the LMF and HF conditions, which are on the same order of magnitudes with the previous estimates in Schubel (1968b), Hirschberg and Schubel (1979) and Officer et al. (1984).

Hourly average model results at different tidal current phases on July 21, when the tidal effect was dominant with sustained low freshwater discharge (Fig. 5-12), show the tidal resuspension lag, in which the limit of salt intrusion lead the center of the ETM in the direction of tidal current (Dyer and Evans, 1989). On October 29, however, when relatively strong gravitational circulation existed after the high discharge event on October 21, the tidal asymmetry due to interaction between tidal currents, gravitational circulation, and stratification results in the formation of the ETM (Fig. 5-13) similar to the conceptual diagram discussed in Sanford et al. (2001: Fig. 5-11). Hence, either tidal resuspension lag or tidal asymmetry, or a combination of both, may occur depending on the relative importance of tidal currents, gravitational circulation, stratification, and other forcing functions. The intratidal variations at three locations, upstream of the ETM and near and farther downstream of the limit of salt intrusion, on July 21 show that the effects of gravitational circulation and stratification resulted in the asymmetrical current pattern, which gave rise to the ETM with distinctively different characteristics between flood and ebb currents (Figs. 5-14 to 5-16). With large intratidal variations in the SS distribution, which are more likely to occur during dry periods when tidal forcing tends to be dominant, caution should be taken in interpreting the SS data for spatial distributions if the data had not been collected at the same tidal phases for all stations.
6-2. Recommendations

During the present study, the following limitations have been noted, which need to be further investigated and to be included in future models to improve the model capability.

1) In SS transport modeling, the region of utmost interest is near the limit of salt intrusion. Largely due to the available salinity data, most hydrodynamic modeling studies have focused on the region farther downstream of the limit of salt intrusion in validating the model capability for salinity simulation. Moreover, the model-data discrepancy of 1 psu in the region with average salinity of, let say, 15 psu is totally different from the same 1 psu discrepancy near the limit of salt intrusion. More efforts are required in collecting salinity data near the limit of salt intrusion with varying freshwater discharge conditions and in modeling the observed data.

2) The model calculated SS concentrations, particularly in the bottom layer, are sensitive to bed shear stress, which is calculated based on the velocity in the model bottom layer. In models employing Cartesian coordinate in the vertical, the bottom layer velocity may not be a good indicator of bed shear stress, because of the thickness of the bottom layer. In the present model application with the bottom layer thickness of 1.52 m, for example, bed shear stress is based on the velocity 0.76 m above the bottom, which itself is not precisely defined in the model. In models employing stretched (e.g., $\sigma$) coordinate in the vertical, the bottom layer thickness changes with total depth, resulting in bed shear stress based on the velocities at different heights from the bottom. A new approach that
may account for the dynamics of the benthic boundary layer faithfully is desirable for accurate estimation of bed shear stress.

3) The present model assumes no interaction among size classes and no effect of the SS on the hydrodynamic field. It has been suggested that particle agglomeration, particularly due to biogenic changes, may be important in affecting the SS transport (Schubel and Kana, 1972; Jahmlich et al., 1999; Sanford et al., 2001). A model may be able to take into account of the seasonal changes in settling velocity by incorporating interactions among particles. It also is desirable to incorporate the effect of the SS on the hydrodynamic field, which may be especially important within the benthic boundary layer, to have a better representation of the boundary layer in the model.
LITERATURE CITED


Schubel, J.R. (1971) Tidal variation of the size distribution of suspended sediment at a station in the Chesapeake Bay turbidity maximum. *Netherlands Journal of Sea Research, 5*, 252-266.


감사의 글

모든 것을 하나님께 감사드립니다.

부족하고 부끄럽지만 오랜 연구를 일단락 지을 수 있도록 도와 주신 분들께 감사를 드리고자 합니다.

먼저, 세심한 지도와 따뜻한 격려로 학문의 길을 인도해 주신 승영호 교수님께 깊은 감사를 드립니다. 또한 많은 가르침과 예리한 지적으로 부족하고 어리석은 저를 지도해주시고 인도해 주신 박경 교수님께 고개 숙여 깊은 감사를 드립니다. 바쁘신 중에도 더 나은 논문이 되도록 심사해주신 세밀한 부분까지 지도해 주신 오재경 교수님, 김승찬 박사님, Harry V. Wang 교수님께 깊은 감사를 드립니다. 이 연구를 할 수 있도록 기회를 주시고 많은 도움을 주신 Harry V. Wang 교수님과 많은 가르침과 부모님 같은 사랑으로 도와주신 김승찬 박사님, 석봉출 박사님, 신용식 교수님께 깊은 감사를 드립니다. 대학과 대학원 생활동안 격려와 미소로 따뜻이 대해주신 최중기 교수님, 박용철 교수님, 홍재상 교수님, 한명우 교수님, 한경남 교수님께도 깊은 감사를 드립니다.

미국에서 열심히 공부하고 힘과 용기를 준 김충기, 정훈신, 권재일 친구들에게도 고마움을 전하며 좋은 결실이 있길 바랍니다. 대학원 생활 동안 많은 도움과 사랑을 주신 김국진 선배님과 힘과 용기를 주신 김홍성, 임효혁, 김도연, 김영중, 송용식 그리고 다른 많은 선배님들과 후배들에게 감사를 드립니다.

끝으로 한없는 사랑과 오늘까지 보살펴 주신 부모님, 장인, 장모님께 깊이 감사를 드리며, 항상 옆에서 도와주고 사랑과 용기를 준 아내와 딸, 경미, 수민, 찬우, 진환, 세환에게도 감사드리며 이 논문이 조그마한 보답이 되었으면 합니다.