Evaluating Sediment Dynamics in Waterways

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ABSTRACT

Cohesive sediments are defined to consist of a mixture of clay- and silt-sized particles, organic matter, and sometimes small quantities of fine sand. Cohesive sediment transport is driven by the hydrodynamic flow field, sediment loading to the water body, sediment loading from marine activities, sediment size gradation, and sediment bed properties. Processes influencing cohesive sediments in the water column and the sediment bed include aggregation, settling, deposition, consolidation, erosion, and transport of suspended sediments by advective and dispersive fluid forces. Presented herein are descriptions of the various processes and constitutive equations for these processes. Understanding the complex nature of the interaction between these processes is crucial in developing numerical models of sediment transport. Predicting the spatial and temporal distribution of sediments in waterways is a precursor to understanding the fate and transport of pollutants adsorbed to cohesive sediments.

INTRODUCTION

Sediments are a ubiquitous component of inland and coastal water bodies (rivers, lakes, bays, estuaries, tidal flats, and the coastal ocean). They are derived from alluvial and marine sources and can generally be classified as cohesive and noncohesive. Cohesive sediments are fine-grained sediments consisting of a mixture of clay- and silt-sized (< 2 μm and < 63 μm, respectively) particles, organic matter, and sometimes small quantities of fine sand. A clay fraction greater than about ten (10) percent is generally sufficient for the sediment to exhibit cohesive properties (van Rijn 1993). Noncohesive sediments are coarser-grained sediments comprised of sand and gravel (> 63 μm).

Because of the affinity of contaminants (e.g., heavy metals, radionuclides, toxic chemical compounds, pesticides, and nutrients) to adhere to cohesive sediments, understanding the dynamics of cohesive sediment transport is crucial in determining contaminant fate and transport. Contaminants in aquatic systems originate from various land use and marine activities (e.g., urban and industrial development, deforestation, mining, agricultural practices, dredging and disposal activities for maintenance of navigation channels, and for remediation of legacy contaminated sediments). These contaminants pose potential threats to the environment, the ecosystem, and human health.

This paper describes a comprehensive review of the current state of understanding of the processes that influence the dynamics of cohesive sediment transport, and includes the key constitutive equations needed to develop numerical models for sediment transport.

COHESIVE SEDIMENT DYNAMICS

Sediments introduced to a body of water are influenced by various processes in the water column, at the sediment-water interface, and in the sediment bed. The principal processes include advection, dispersion, aggregation, and settling in the water column, deposition and erosion at the sediment-water interface, and sediment consolidation that influences the erodibility of bed sediments (Figure 1). These processes depend on the physico-chemical characteristics of the sediments as well as the hydrodynamic flow field. It is the interaction between these processes that makes the issue of cohesive sediment transport challenging. The reader may refer to Figure 1 for terms used in the text.

Advection and Dispersion

Advection is the process by which sediments are transported, while dispersion spreads the sediments depending upon the concentration gradient. The advection-dispersion equation is given by:

$$\frac{\partial C}{\partial t} + \frac{\partial uC}{\partial x} + \frac{\partial vC}{\partial y} + \frac{\partial wC}{\partial x} = \frac{\partial}{\partial x}\left(D_h \frac{\partial C}{\partial x}\right) + \frac{\partial}{\partial y}\left(D_h \frac{\partial C}{\partial y}\right) + \frac{\partial}{\partial z}\left(D_z \frac{\partial C}{\partial z}\right) + Q_s \quad (1)$$

where C is the sediment concentration, $u, v, and\ w$ are velocity components averaged over a specific time in the longitudinal (x), lateral (y) and vertical (z) directions, t is time, D_h and D_z are the horizontal (i.e., in the x- and y-directions) and vertical eddy diffusivities, and Q_s is the sediment mass flux into and out of the water column (i.e., erosion and deposition, respectively).

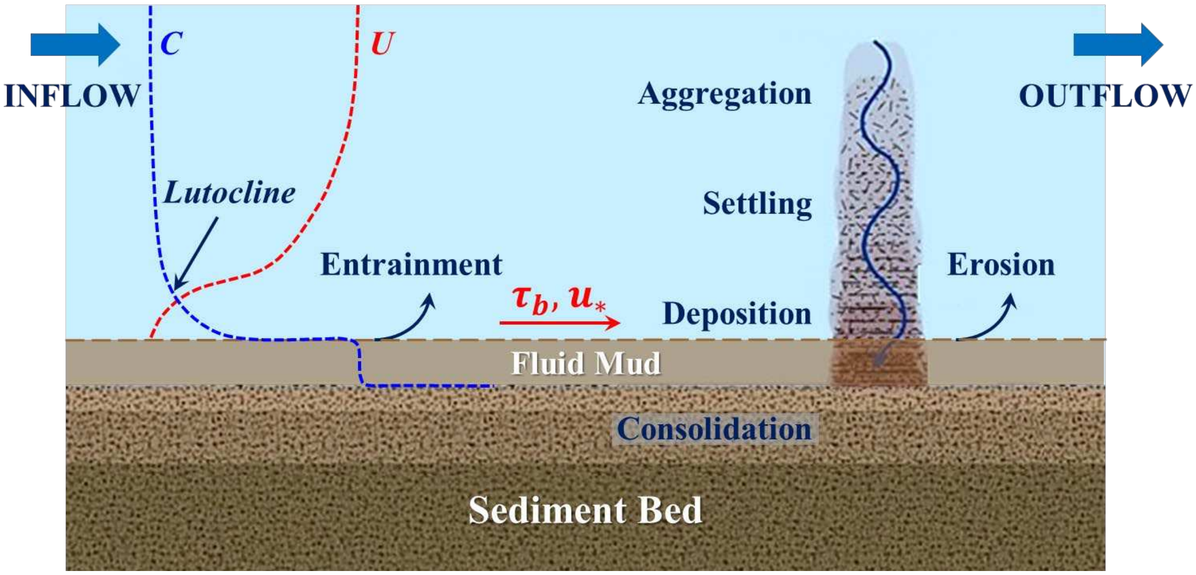

Figure 1. Schematic of cohesive sediment transport processes. $\boldsymbol{C}$ is the suspended sediment concentration profile, $\boldsymbol{U}$ is the velocity profile, $\boldsymbol{\tau_b}$ is the bed shear stress, and $\boldsymbol{u_*}$ is the shear velocity. Adapted from Shrestha et al. (2014) and Mehta and McAnally (2008).

Aggregation (Flocculation)

Cohesive sediments are composed primarily of clay-sized particles that commonly aggregate due to their surface ionic charges. When the double layer around each sediment particle is compressed by high ionic concentration (e.g., divalent ions), the particles are destabilized and van der Waals attractive forces predominate over repulsive forces, creating a condition that is conducive to aggregation (flocculation).

Brownian motion, fluid shear, and differential settling are the three primary collision mechanisms that cause the particles to bind to each other and form aggregates (Krone 1962). The frequency functions for the three collision mechanisms are given by Mehta and McAnally (2008) as:

$$\textit{Brownian Motion:} \quad \beta_c = \frac{2}{3}\frac{\kappa T F_c}{\mu}\frac{\left(d_i + d_j\right)^2}{d_i d_j}$$

$$\textit{Fluid Shear:} \quad \beta_c = \left[\frac{\pi F_c^2}{4}\sqrt{\frac{2}{15\pi}}\right] G_s\left(d_i + d_j\right)^3 \qquad (2)$$

$$\textit{Differential Settling:} \quad \beta_c = \left[\frac{\pi F_c^2}{4}\right]\left(d_i + d_j\right)^2\left|w_{si} = w_{sj}\right|$$

where κ = Boltzmann constant, T = absolute temperature, F_c = collision diameter correction factor (varies between 0 and 1), d_i, d_j = *i* and *j* sizes of the particles, respectively, w_{si}, w_{sj} = settling velocities of *i* and *j* size particles, respectively, G_s = fluid shear $= (\varepsilon/\upsilon)^{1/2} = (\upsilon/\lambda)^2$, where $\varepsilon =$ flow energy dissipation per unit mass of fluid per unit time, $\upsilon =$ kinematic viscosity of fluid, and $\lambda =$ Kolmogorov turbulence micro-scale.

Brownian motion occurs when the sediment particles are agitated by the thermal motion of fluid particles that move the particles in random directions. Brownian motion is predominant in stationary or quasi-stationary waters and the aggregates thus formed are weakly bonded (Mehta and McAnally 2008). Internal shear results from velocity gradients in the suspending medium. Internal shear, the most important of the three collision mechanisms, produces aggregates that are more durable and tightly packed compared to the other two mechanisms (Krone 1986). Differential settling is caused by larger aggregates (with larger settling velocities) colliding with smaller aggregates (with lower settling velocities) and is important during and close to times of slack water (Mehta and McAnally 2008).

Aggregation acts to create large-sized aggregates that can be characterized by their higher porosity, increased irregularity and fragility, and higher settling rate (Krone 1963). Aggregation is influenced by sodium adsorption ratio, pH, salinity, sediment size, shape, gradation, density, turbulence, temperature, and the efficiency of collision between particles. Fluid forces and collisions that exceed the strength of the individual aggregates will break them apart. In numerical models, the aggregation mechanism in generally not considered explicitly but instead included in the constitutive equations for settling velocity.

Settling

The settling velocity depends upon the aggregate properties, i.e., size, density, and shape, which in turn is a function of the frequency and efficiency of interparticle collisions (Mehta and McAnally 2008). The settling velocity is usually described as a function of suspended sediment concentration (SSC) (Krone 1962) because of the convenience of measuring SSC. Mehta et al. (1989) reported settling velocity values in estuarine and coastal waters ranging from 10^{-3} and 10^{-7} m s^{-1}.

Wolanski et al. (1989) divided the settling range into four zones: *free settling, flocculation settling, hindered settling,* and *negligible settling*. As reported in Mehta and McAnally (2008), Hwang (1989) expressed the settling velocity (w) in each zone as:

$$\begin{aligned} &w = w_f && C < C_1 \\ &w = a_w \frac{C^{n_w}}{(C^2 + b_w^2)^{m_w}} && C_1 < C < C_3 \\ &w \sim Negligible && C_3 < C \end{aligned} \tag{3}$$

where w_f = free settling velocity, C = SSC, C_1, C_2 = zone concentration limits, a_w = velocity scaling coefficient, b_w = hindered settling coefficient, n_w = flocculation settling exponent, m_w = hindered settling coefficient. The reader is referred to Mehta and McAnally (2008) for representative values of the various coefficients.

Free settling occurs at low concentrations (between 0.1 and 0.3 kg/m^3). Here, the settling velocity is independent of concentration (Mehta and McAnally 2008; Krone, 1962) and the settling velocity can be calculated using Stokes' law. As the concentration increases, to between 1 and 15 kg/m^3, flocculation settling is triggered (Mehta and McAnally 2008). Krone (1962) reported that *hindered settling* occurs at concentrations greater than 10 kg/m^3, based on experiments on San Francisco Bay sediments. Fluid mud, associated with a lutocline (i.e., strong vertical concentration gradients), forms near the bed when concentrations exceed this value (McAnally et al. 2007).

The following are several settling velocity (w) formulations:

Krone (1962):
$$w = KC^{4/3} \tag{4}$$

where $K =$ empirical constant depending upon sediment type = 0.001, and $C =$ SSC (g/L).

Cole and Miles (1983):
$$w = KC \tag{5}$$

where $K =$ empirical constant = 0.001 - 0.002, and $C =$ SSC.

Nicholson and O'Connor (1986):
$$\begin{aligned} w &= A_1 C^{B_1}, \quad C \le C_H \\ w &= A_1 C_H^{B_1}[1.0 - A_2(C - C_H)]^{B_2}, \quad C > C_H \end{aligned} \tag{6}$$

where $A_1 = 6.0 \times 10^{-4}$ m^4/kg/s, $A_2 = 1.0 \times 10^{-2}$ m^3/kg, $B_1 = 1.0$, $B_2 = 5.0$, and $C_H = 25$ g/L = onset of *hindered settling*.

Burban et al. (1990):
$$\begin{aligned} w &= a d_m^b \\ a &= B_1 (C\tau)^{-0.85} \\ b &= -[0.8 + 0.5 \, log(C\tau - B_2)] \end{aligned} \tag{7}$$

where $B_1, B_2 =$ empirically determined constants = 9.6 x 10^{-4}, 7.5 x 10^{-6}, respectively, $C =$ SSC (g/cm^3), $\tau =$ fluid shear stress (dyne/cm^2), and $d_m =$ median floc diameter (cm).

Van Leussen (1994):
$$w = KC^n \frac{1 + aG}{1 + bG^2} \tag{8}$$

where $a, b =$ empirical constants, $G =$ dissipation parameter $= (\varepsilon/\upsilon)^{1/2}$, $\varepsilon =$ turbulent energy dissipation rate, and $\upsilon =$ molecular viscosity.

Deposition

As aggregates settle toward the sediment bed, near-bed turbulence controls whether they break apart and are re-entrained into the water column or bond with particles on the sediment bed. The stochastic nature of the near-bed turbulence responsible for either floc break-up or floc growth is generally characterized by a "probability of deposition," a value which Krone (1962) defined as the probability that particles reaching the bed will actually stick to the bed. The deposition rate is a function of the settling velocity, SSC, and the probability of deposition expressed in terms of the near-bed shear stress and a critical shear stress for deposition. A widely used equation for the deposition rate is the Krone (1962) formulation:

$$\frac{dC}{dt} = -\frac{wC}{h}\left(1 - \frac{\tau_b}{\tau_d}\right), \qquad \tau_b < \tau_d \tag{9}$$

where $w =$ settling velocity, $C =$ near-bed SSC, $h =$ water depth, $\tau_b =$ bed shear stress, and $\tau_d =$ critical shear stress for deposition.

Shrestha and Orlob (1996) utilized the Krone (1962) experimental data to express the deposition rate over several orders of magnitude of SSC as:

$$\frac{dC}{dt} = -\frac{B_G C^{n_G}}{h}\left(1 - \frac{\tau_b}{\tau_d}\right), \qquad \tau_b < \tau_d \tag{10}$$

where $B_G =$ exp(-4.20706 + 0.1465G), $n_G = 1.11075 + 0.038G$, and $G =$ average shear rate.

If $\tau_b \geq \tau_d$, no sediments deposit. The critical shear stress for deposition is determined via experiments, ranging between 0.06 and 1.1 N/m^2 depending upon sediment type and concentration (Mehta and McAnally 2008). The bed shear stress is computed as:

$$\tau_b = \rho u_*^2 \tag{11}$$

where $\rho =$ density of the suspending medium, and $u_* =$ shear velocity = $\kappa u / ln(z/z_0)$, $\kappa =$ von Karman constant ≈ 0.40, $u =$ near-bed velocity, $z =$ depth at center of near-bed layer, and $z_o =$ bottom roughness length.

Consolidation

Upon reaching the sediment bed, the flocculated structure of the sediment aggregates break down and particle-to-particle contact is made. Interstitial (or pore) water escapes the bed matrix and the sediments consolidate under their own weight. The effective stress, which is the difference between the total hydrostatic pressure and the pore water pressure, increases with the release of pore-water pressure. The void ratio of the bed decreases and the density and shear strength both increase with increasing depth of bed sediments. Empirical relationships are used to quantify the strength of bed sediments as a function of the dry density. It has been observed that shear strength increases with increasing clay content, organic matter, salinity, sodium adsorption ratio, and cation exchange capacity. Conversely, shear strength decreases with increases in temperature, pH-value, and the concentration of noncohesive sediments (e.g., sand) in the bed.

Huang et al. (2006) and Mehta and McAnally (2008) describe the constitutive equations for the consolidation process. Huang el al. (2006) enumerates several relationships by Nicholson and O'Connor (1986), Teisson and Latteux (1986), and Letter et al. (2000) that relate bulk density to consolidation time. In numerical models for sediment transport, consolidation is accounted for by discretizing the bed into a number of layers, each having a specific thickness, density, critical shear stress, and consolidation time (Shrestha et al. 2000). In other treatments, consolidation algorithms are embedded in the numerical model to simulate consolidation.

Erosion

Erosion of bed sediments reflects the balance between bed shear stresses induced by the flow and the resistance of the bed sediments to erosion. In waterways such as coastal waters, the bed shear stress is caused by waves and currents. Resistance to erosion depends upon the sediment type and minerology, the pore and eroding fluid, Huang et al. (2006), the time history of deposition (i.e., history of consolidation), and chemical and biological processes (Mengual et al. 2017). Because of the complex nature of the erosion process, investigators have relied on laboratory and field experiments to derive the constitutive equations for erosion.

Mehta and McAnally (2008) describe four modes of erosion: *surface erosion, mass erosion, fluid mud generation* and *fluid mud entrainment*.

Surface erosion occurs when sediment flocs at the bed surface are dislodged and entrained at low to moderate excess shear stress (i.e., the difference between the bed shear stress and the critical shear stress for erosion), at locations where currents are low to moderate (Mehta et al. 1989). Surface erosion is limited to a finite mass and ceases when the bed shear stress (τ_b) is less than the critical shear stress for erosion (τ_c) (Parchure and Mehta 1985; Tsai and Lick 1987; Amos et al. 1992). A commonly used linear relation for surface erosion presented by Ariathurai (1974) and attributed to Partheniades (1962) is:

$$E = M\left(\frac{\tau_b}{\tau_c} - 1\right); \qquad \tau_b > \tau_c \tag{12}$$

where $E =$ erosion rate (mass eroded per unit bed area per unit time), $M =$ erosion rate constant. τ_c and M are determined from erosion experiments. By plotting erosion rate (E) against the applied shear stress (τ_b), τ_c is the shear stress intercept at zero erosion rate, and the slope of the line is M for the range of shear stresses applied.

Eq. (12) and its variants have been used in various numerical models of sediment transport (Mathew and Winterwerp 2017). τ_c has been expressed as a function of the wet or dry bed density by several investigators (Nicholson and O'Connor 1986; Teisson and Latteux 1986; Hwang and Mehta 1989; van Rijn 1993; Roberts et al. 1998). Huang et al. (2006) present values of τ_c, M, and E based on experiments carried out by various investigators.

Mass erosion occurs when the bed shear stress exceeds the shear strength at some depth and chunks of sediment are eroded *en masse* from the bed. This mode of erosion is considered to occur when the bed is subjected to high shear stresses induced in zones of strong currents. The equation for surface erosion (Equation 12) has been also used for mass erosion because of convenience (Mehta and Lee 1994). Ariathurai et al. (1977) developed the following formulation for instantaneous erosion of a bed layer.

$$E = \rho_b \frac{\Delta z_b}{\Delta t} \tag{13}$$

where E = erosion rate (mass eroded per unit bed area per unit time), ρ_b = dry density, Δz_b = bed layer thickness, and Δt = time. Fluid mud generation represents a phase shift from bed to suspension and hence can be considered as erosion, whereas fluid mud entrainment depends on the turbulent energy at the boundary layer, destabilizing the lutocline interface, and entraining the sediments in the fluid mud into the upper layers (Mehta and McAnally 2008).

Erosion properties have been derived using different devices (Parchure and Mehta 1985; Amos et al. 1992; Maa et al. 1998; Gust and Mueller 1997; McNeil et al. 1996; Jepsen et al. 1997, 2000; Roberts et al. 1998, 2003; Jones and Lick 1999, 2001; Lick and McNeil 2001; Winterwerp et al. 2012). Huang et al. (2006) and Black and Paterson (1997) summarize various laboratory and in-situ methods to determine erosion properties. Erosion properties have also been studied by relying on SSC measurements (van Kessel et al. 2011; van Maren et al. 2015; Mathew and Winterwerp 2017).

NUMERICAL MODELING

The constitutive equations governing the individual processes and methods to incorporate such equations in numerical models are crucial elements in predicting the spatial and temporal distributions in the water column and in the sediment substrate. Numerical models serve as an efficient and practical tool for predicting sediment transport in complex hydrodynamic systems in that they serve as a virtual laboratory for simulating prototype conditions. The advent of sophisticated models and enhanced computing power have provided the means to incorporate complex sediment transport processes into tractable computational techniques.

The computational framework generally consists of coupled modeling of hydrodynamics, waves, and sediment transport, such that output from one model serves as input to another model. Models are calibrated and validated using hind-casting techniques in order to establish a level of confidence in the model's predictive capabilities. Once the model's predictive capabilities have been assessed, model simulations are then performed to forecast future system responses. Numerical modeling studies are generally supplemented by field sampling and monitoring programs to assimilate and analyze data for creating model inputs and for model calibration and validation. A conceptual site model is usually developed in the process of developing a numerical model. The conceptual site model is an evolving document that is initiated at the commencement of the project, periodically updated with site activities, used to inform future sampling and monitoring efforts, and to guide numerical model development (Shrestha et al. 2014).

Hydrodynamic and Wave Models

The hydrodynamics of the water body provides the flow field description to drive the sediment transport model. Martin and McCutcheon (1998) provide a review of hydrodynamic models ranging from one-dimensional to three-dimensional. Choice of the dimensionality of the model is based on the intended use of the model. The governing equations include the continuity, momentum, and constituent transport equations for temperature and salinity, with an equation of state relating density to temperature and salinity. Model inputs include steady-state and/or time-varying freshwater inputs from various point and non-point sources, astronomical inputs such as tides, meteorological inputs such as winds and heat fluxes, and temperature and salinity distributions in the inflows and within the water body. Model outputs include water surface elevations (and depths), velocities, diffusivities, and distributions of temperature and salinity.

Wind waves (i.e., high frequency short waves of periods 3-20 seconds) increase mass and momentum fluxes and induce higher bed shear stresses at the sediment-water interface. The development of wind waves is based on a balance between wind energy input, wave energy, and wave energy dissipation. Wave models used to predict the wave climate include SWAN (Holthuijsen et al. 1993); HISWA (Booij and Holthuijsen 1995); the Great Lakes Environmental Research Laboratory (GLERL) wave model (Schwab et al. 1984); SMB (USACE 1984); WAVD (Resio and Perrie 1989); or ACES (Leenknecht et al. 1992). Input to the wave model requires a time-dependent wind field. Wave models are generally coupled to the hydrodynamic model. Linear wave theory is used to compute the near-bed peak velocity and peak orbital amplitude. A wave-current model is used to predict the bed shear velocity and corresponding bed shear stress, which is the critical parameter required to drive the sediment transport model.

Cohesive Sediment Transport Model

Cohesive sediment transport models include zero-dimensional models (e.g., Krone 1985) that ignore the spatial variability of the sediment properties. A two-layered one-dimensional model was used to simulate cohesive sediment transport in the Thames Estuary (Odd and Owen 1972). Two-dimensional models may be depth-averaged or laterally averaged, in that SSC are well-mixed over the depth or in the lateral direction, respectively. Depth-averaged models include those developed by Ariathurai and Krone (1976), Onishi (1981), Cole and Miles (1983), Lick et al. (1994), Shrestha (1996), and Shrestha and Orlob (1996). These models compute sediment deposition rates from depth-averaged mean SSC. Laterally-averaged models developed by Ariathurai et al. (1977), and Onishi and Wise (1982) describe the longitudinal and vertical distribution of SSC. Lou et al. (2000) developed a quasi-three-dimensional model that accounts for the vertical distribution of sediment concentrations based on the vertical velocity profile. Recent advances in modeling utilize three-dimensional finite difference or finite element models (Hayter and Pakala 1989; Sheng 1991; Onishi et al. 1993; Shrestha et al. 2000; James et al. 2010). These models are suitable for applications to systems where the three-dimensionality of the flow, the salinity and temperature structure, and the associated suspended sediment distribution are important (Shrestha and Blumberg, 2018).

In addition to the hydrodynamic flow field, inputs to the cohesive sediment transport model (depending upon the dimensionality of the model) include sediment bed properties (i.e., bed type, particle-size distribution, sediment density, and critical shear stresses for erosion and deposition), sediment loading and size distribution from various sources, and sediment settling parameters. Model output includes the spatial and temporal distribution of suspended sediment concentrations, the mass of sediment eroded or deposited, and subsequent change in bed elevations.

CONCLUSIONS

Understanding the key processes influencing cohesive sediment dynamics is important for predicting their fate and transport in waterways. The flow field and circulation patterns are important elements that drive sediment transport. Because of the complex mechanisms influencing cohesive sediment transport, it is important to understand the constitutive equations for the various processes and how they are incorporated in numerical models of sediment transport.

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