Cohesive Sediment Transport. I: Process Description

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Abstract: Physical processes constituting fine, cohesive sediment transport in estuarial waters are described. These processes, which include settling and deposition, consolidation, erosion and transport in suspension, are typically interlinked by the cyclic nature of the tide dominated environment. Complexities in process characterization arise as a consequence of the dual dependence of sediment aggregate properties on the physico-chemical properties of the sediment-water mixture as well as the turbulent flow field. Present day knowledge of the processes enables reliable predictions of rates of sedimentation and erosion in navigable channels, waterways and harbors through numerical modeling. Further research is required for improving procedures for measuring settling velocities, identification of depth at which a definable bed is encountered, and the behavior of near-bed high density suspensions.

Introduction

Quantification of fine, cohesive sediment transport is required in hydraulic engineering applications including the estimation of erosion and sedimentation in estuarial navigation channels, waterways, harbors, docks, and marinas. In addition there is a need to evaluate the strength of turbidity currents and associated rates of sediment transport over the estuarial shelf and along coasts underlain by mud banks. The high sorptive affinity of fine suspended matter for chemical constituents in water causes fine material to act as a carrier for contaminants with consequent implications for related water quality problems.

The interrelationship among basic cohesive sediment transport processes in estuaries is shown schematically in Fig. 1 (Mehta et al. 1982; Parker and Kirby 1977). A sediment-water mixture can be considered to exist in any one of the four states shown in Fig. 1: a mobile suspension, i.e., suspension in horizontal transport; a horizontally stationary high density suspension which may however possess a vertical velocity component; a consolidating (soft) deposit; or a settled (firm) consolidated bed. A stationary suspension with practically no mechanical strength results from settling of cohesive sediments in transport, particularly at times close to slack water. Under suitable con-

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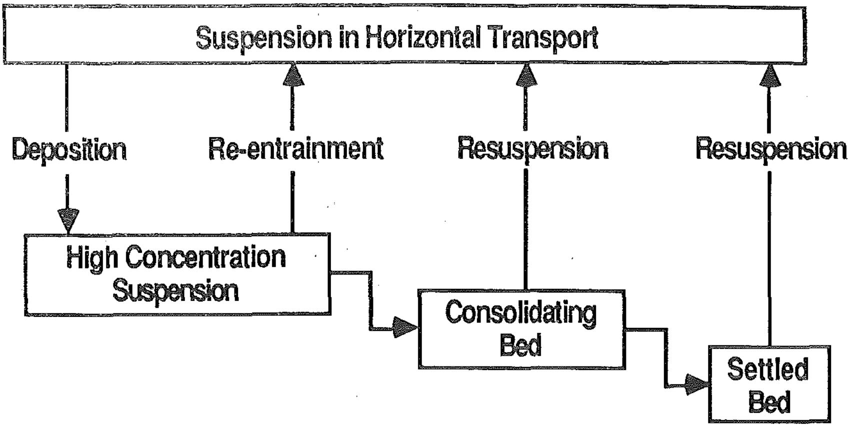

FIG. 1. Physical States and Processes Governing Estuarial Cohesive Sediment Transport (after Mehta et al. 1982)

ditions a bed deposit possessing a small but measurable shear strength begins to develop. Gelling and consolidation of this deposit and associated physicochemical changes eventually result in a settled bed with a lower water content, higher shear strength, and more stable structural configuration.

Entrainment of a stationary suspension, which typically could occur shortly after current reversal following slack water, has been referred to as redispersion or re-entrainment (Parker and Kirby 1977). Entrainment from a consolidating or a settled bed is referred to as erosion or resuspension.

In this paper, basic process-related aspects are briefly reviewed, with emphasis on process descriptions associated with Fig. 1, and process characterizations relevant to hydraulic engineering practice. At the outset, reference is made to parameters which characterize the sediment. Basic processes are elaborated upon next. Definitions and measuring techniques for suspension concentration, and for the bottom boundary or bed important both to modelers of sediment transport as well as to ship navigation, are then considered. The paper concludes with a brief consideration of data requirements for estimation of erosion/sedimentation, with reference to available information and research needs.

Sediment Characterization

When sufficient salt is added to a suspension of dispersed clay particles, the suspended particles become cohesive. Flocs or aggregates with orders of magnitude larger settling velocities are formed when cohesive particles collide repeatedly. Collision mechanisms include Brownian motion, differential settling, and current shear. Current shear, expressed in terms of the rate of fluid shear, is typically the most important factor contributing to aggregation in turbulent flows. The transformation of dispersed particles into cohesive ones is due to ions in solution, which suppress interparticle electrochemical repulsive forces, thereby allowing the attractive London-van der Waals forces to dominate. All suspended clay particles become cohesive when the salinity exceeds 2 to 3 ppt.

Sediment of size greater than 60 μm is considered to be coarse, and less than this size fine-grained. There appears to be some confusion between this mode of classification, and one that considers sediment as either cohesionless or cohesive. The boundary between cohesive and cohesionless sediment is, unfortunately, not clearly defined and generally varies with the type of material. However, dominance of interparticle cohesion over gravitational force increases with decreasing particle size. Thus the effect of cohesion on the behavior of clays (particle size < 2 μm) is much more pronounced than on silts (2 to 60 μm), and, in fact, cohesion in clayey muds is primarily due to the presence of clay-sized sediment.

Muds in aquatic environments are typically composed of clay and nonclay minerals in the clay- and silt-size ranges, organic matter and, sometimes, small quantities of very fine sand. When large amounts of coarse detritus including sand, gravel and shell occurs with mud, the interactive behavior between different-sized sediments becomes quite complex and is not presently well understood. The present practice is to treat the coarse material separately from mud.

The problem of property characterization for cohesive sediment is more complex than that for coarse grained material because aggregate properties depend upon the type of sediment, type and concentration of ions in the water, and on the flow condition. Furthermore, cohesion is influenced by colloidal organic matter, microbes, polysaccharides, etc. (Montague 1986). For characterizing the sediment it is recommended that the following properties be specified through laboratory measurement: (1) Grain size distribution of dispersed, nondried sediment; (2) the relationship between the settling velocity and the suspension concentration of the flocculated sediment in native water (Owen 1976); (3) clay and nonclay mineralogical composition (Grim 1968); (4) total organic matter; and (5) the cation exchange capacity, CEC, as a measure of clay cohesion (Grim 1968).

Inasmuch as consolidation increases bed density, it is important to obtain in situ cores for determining the bed density profile, which is always site-specific. This information enables a conversion between deposited or eroded sediment mass per unit time and the corresponding changes in the suspension concentration (mass per unit volume). In addition, density provides an approximate indication of the shear strength of the bed with respect to erosion.

In studies in which dissipation of fluid energy within the bed plays an important role, e.g. wave-mud interaction, it is essential to evaluate the rheological properties. The most important rheological property is the viscosity, which has been found to be related to sediment density in an approximate manner (Krone 1963; Migniot 1968). Viscometer data indicate that muds typically exhibit a non-Newtonian rheology. Thus it becomes necessary to specify additional parameters. Most commonly this includes the yield stress for a comparatively simplified rheological description. The dynamic behavior of muds under wave-induced loading suggest a visco-elastic response, which is characterized by the viscosity and shear modulus of elasticity (Mehta and Maa 1986).

For characterizing the fluid, it is recommended that the following quantities be specified using standard chemical analyses: (1) Concentration of important cations (e.g., Na^{+}, Ca^{++}, Mg^{++}) and anions (e.g., Cl^{-}, SO_4^{--});

(2) total salt concentration; (3) pH; and (4) fluid temperature during field and laboratory experiments.

Settling and Deposition

Settling characteristics of cohesive sediment affect the rates of deposition and vertical distribution of suspended material. The rate of deposition per unit bed area, D, or flux of material to the bed per unit time, is calculated as the sum over a number of classes of settling flux: $D = \Sigma_i p_i W_{si} C_{bi}$, where W_{si} is the settling velocity of settling class i, p_i is the probability of deposition, i.e., the probability that a particle of class i reaching the bed will remain there, and C_{bi} is the concentration of class i near the bed. The purpose here is to provide information on how W_s, p and C_b can be estimated or measured.

Settling Velocity

Settling is that component of particle motion effected by gravitational forces, viscous drag on the particles and interparticle interactions. Settling velocity is usually defined as the settling rate in quiescent water. In the following, the settling velocity is treated on a time-invariant macroscopic level, but it should be understood that time-dependent aggregation processes, the suspension concentration and water salinity affect the settling velocity. Thus, settling velocities of cohesive materials are properties of a suspension, not unique properties of the sediment.

Several direct and indirect methods have been used to measure settling velocity. Direct methods include visual or photographic observation. Indirect methods include sedimentation balances and settling tubes. A commonly encountered problem with the sedimentation balances approach is preventing the formation of a high density suspension upon entering the area below the balance pan. Accumulation or pipette analyses have been used to reduce settling tube data, which are obtained under quiescent conditions wherein aggregation is typically well-advanced, and the rates of ongoing aggregation extremely low (Federal Inter-Agency 1953; McLaughlin 1959; Owen 1976; Whitehouse et al. 1960).

Settling velocities of sediments under continued aggregation can be estimated from flume tests. By measuring the removal of material from suspension with time by deposition, and using the relationship:

$$\frac{d\bar{C}}{dt} = -\frac{pW_s\bar{C}}{h} \quad (1)$$

where h is the depth and $\bar{C}$ is the depth-mean concentration, the effective settling velocity, pW_s, can be estimated. Samples should be taken in the vertical to check the assumption implicit in Eq. 1 that the suspension has uniform concentration. In Eq. 1 the effects of aggregation and salinity are implicit in W_s.

To determine W_s from the effective settling velocity, p must be estimated. The functional form of p depends on the time-mean value of the bed shear stress, τ_b, and a critical shear stress for deposition, τ_{cd}, which depends on the sediment-water composition. Krone (1962) found: $p = 1 - (\tau_b/\tau_{cd})$ when

TABLE 1. Primary Particle and Aggregate Diameters and Settling Velocities

Primary particle diameter, (μm) (1)	Stokes settling velocity, (mm/s) (2)	Aggregate setting velocity (mm/s) (3)	Aggregate diameter, (μm) (4)	Aggregate velocity divided by Stokes velocity (5)
2×10^1	2.4×10^{-1}	2.7×10^{-1}	8.8×10^1	1.1×10^0
2×10^0	2.4×10^{-3}	1.7×10^{-1}	5.6×10^1	7.1×10^1
2×10^{-1}	2.4×10^{-5}	1.1×10^{-1}	3.4×10^1	4.6×10^3

$\tau_b < \tau_{cd}$, and $p = 0$ when $\tau_b \geq \tau_{cd}$. The critical shear stress for deposition is thus the bed shear stress above which no deposition occurs.

Any physical or chemical factor which influences aggregate size, density and shear strength affects the settling velocity. Marine and estuarial sediments thus exhibit a wide range of settling velocities. Reported values range from 10^{-4} to 10^0 mm/s (Burt 1986; Chase 1979; Krone 1962; Migniot 1968; Owen 1970; Owen 1971; Teeter 1986; Whitehouse and Jeffrey 1952).

An indication of the degree of enhancement of the settling velocity due to flocculation is obtained from the illustrative results of Table 1, which are derived from the studies of Migniot (1968) and Chase (1979) in settling tubes. For primary (dispersed) particle diameters of 20, 2, and 0.2 μm, the corresponding Stokes settling velocity of primary particles, aggregate settling velocity, and aggregate diameter are given. The ratio of aggregate to Stokes velocity given in the last column ranges from 1.1 at a primary particle diameter of 20 μm to 4.6×10^3 at 0.2 μm. Furthermore, it is noteworthy that while Stokes velocity decreases rapidly with particle size, aggregate settling velocity as well as diameter retain the same orders of magnitude due to increasing aggregation with decreasing particle size.

For a given sediment, the effects of suspension concentration, salinity and the flow field on aggregate settling velocity are found to be the most important ones deserving consideration. In general, the settling velocity increases with concentration up to about 5,000 to 10,000 mg/L, above which it begins to decrease with increasing concentration as a consequence of hindered settling. Hindered settling occurs when the sediment forms a nearly continuous network through which pore water must escape slowly upwards for settling to continue. A high density suspension characterized by hindered settling is commonly referred to as fluid mud (Krone 1962).

The settling velocity of flocculated cohesive sediments typically increases with increasing salinity up to about a salinity of 10 ppt (Krone 1962). At higher salinities, the effect is found to be important mainly for predominantly montmorillonitic materials (Whitehouse et al. 1960).

Krone (1962) measured the settling velocity of mud from the San Francisco Bay in both a quiescent tube and in a recirculating flume. Settling velocities estimated from flume tests (0.007 mm/s) were only about 20 percent as great as those obtained from quiescent settling tests. On the other hand, Owen (1971) reported results from field settling tests in the Thames River estuary, England, using a specially designed settling tube sampler which was used to collect the suspension in situ and measure the settling velocity immediately thereafter. This method resulted in settling velocities which were an order higher than those obtained for the same sediment tested in a lab-

TABLE 2. Ratio of Near-Bed to Mean Suspension Concentration, β

Peclet number, P_e (1)	β $p = 0$ (2)	$p = 1$ (3)
0.06	1.1	1.0
0.3	1.2	1.1
1.5	2.2	1.3
3	3.4	1.5

oratory settling tube. This difference is clearly due to increased aggregation under estuarial turbulent flows as compared to quiescent conditions in laboratory tests, and emphasizes the need to obtain in situ settling velocity data as much as possible.

Probability of Deposition

The critical shear stress for deposition, τ_{cd}, required for specifying the probability of deposition is principally characterized by the type of sediment-water mixture, and can be determined from flume tests in which sediment, initially suspended at a high flow velocity, is allowed to deposit at a lower velocity (Krone 1962; Mehta and Partheniades 1975). If this lower velocity corresponds to a τ_b which exceeds τ_{cd}, no deposition will occur provided the sediment has uniform properties. A series of flume tests in which the lower velocity is slowly increased until no deposition occurs enables determination of τ_{cd} corresponding to this critical lower velocity. For the San Francisco Bay sediment, Krone (1962) found $\tau_{cd} = 0.06\ \text{N/m}^2$ when the initial suspension concentration, C_0, was less than ~300 mg/L. At higher values of C_0 (ranging from 300 to 10,000 mg/L), $\tau_{cd} = 0.078\ \text{N/m}^2$ was obtained, indicating the influence of continued aggregation under turbulent flows on the settling process at higher concentrations.

When the sediment has a broad size distribution, e.g., contains particles ranging in size from coarse silt to fine clay, τ_{cd} does not possess a unique value, but a range of values occur. In tests using kaolinite, Mehta and Partheniades (1975) found τ_{cd} ranging from 0.18 N/m^2 to about 1.1 N/m^2. In such a case, the total deposition flux can be calculated by summing the rate of deposition of each sediment class. Data interpretation, however, requires careful scrutiny.

Near-Bed Concentration

Near-bed sediment concentration is related to the depth-mean concentration and to the vertical variation in concentration. Vertical variation can be described using direct observations or by an analytic expression relating the principal quantities involved.

The following analytic expression for the ratio β of near-bed concentration, C_b, to depth-mean concentration, $\bar{C}$, was developed by Teeter (1986) assuming a parabolic distribution of diffusivity with variable boundary flux conditions:

$$\beta = 1 + \frac{P_e}{1.25 + 4.75p^{2.5}} \qquad (2)$$

where $P_e = W_s h/k_z$ is the Peclet number for the suspension and k_z is the depth-mean eddy diffusivity. In Table 2, illustrative values of β are given for P_e ranging from 0.06 to 3 and for probability of deposition, $p = 0$ and 1. In general, increasing magnitude of P_e implies an increasingly deposition-dominated environment. Large vertical concentration gradients ($\beta >> 1$) can occur in estuaries, both under eroding flows and also when the material begins to settle out of the water column.

CONSOLIDATION

A cohesive sediment bed is formed by the combined action of hindered settling and consolidation. Sediment aggregates comprising a stress-free stationary suspension undergo hindered settling, during which the aggregates begin to interact and form a sediment bed. During this transition the weight of sediment mass near the suspension/bed-water interface is balanced by the seepage force induced by the upward flow of pore water from the underlying sediment. As the sediment continues to be brought closer together and the upward flux of pore water lessens, the weight of this near-surface sediment gradually turns into an effective stress (i.e., the difference between total hydrostatic pressure and pore water pressure), which is transmitted by virtue of particle-to-particle contact. These surface stresses are very small and in general are not measurable (Been and Sills 1981).

Primary consolidation, which is caused by the self-weight of sediment in the overlying deposit, begins when effective stresses are first developed. The strains involved in primary consolidation are relatively large, typically greater than $\pi/4$ radians (Parker and Lee 1979). As consolidation continues, a sediment bed is defined to be formed when the water content of the sediment-water suspension decreases to the fluid limit (Parker and Lee 1979). For cohesive sediment beds, the fluid limit is a function of the initial water content of the suspension. Primary consolidation ends when the excess pore water pressure, which is equal to the total stress minus the sum of the effective stress and the static pore water pressure, has completely dissipated. Secondary consolidation, which is the result of plastic deformation of the bed under a constant overburden, begins during primary consolidation and may typically continue for many weeks or months after primary consolidation ends.

Consideration of the consolidation of cohesive sediment beds is essential in modeling bed erosion because: (1) The susceptibility to erosion of a consolidating bed decreases with time due to the continual increase in bed shear strength; and (2) the accompanying density increase changes the mass of sediment eroded per unit bed thickness.

There are several methods for evaluating consolidation of saturated cohesive beds. These include numerical models (Gibson et al. 1981), analytical solutions (Lee and Sills 1981), solution charts developed using the results from a numerical model (Cargill 1984), and empirically based modeling (Hayter 1983). The numerical and analytical methods are based on solution of the governing equation for finite strain consolidation theory for the time- and depth-varying void ratio. Two constitutive relationships—that between void ratio and permeability, and between void ratio and effective stress—are required for solving the consolidation equation. For any sediment-water mix-

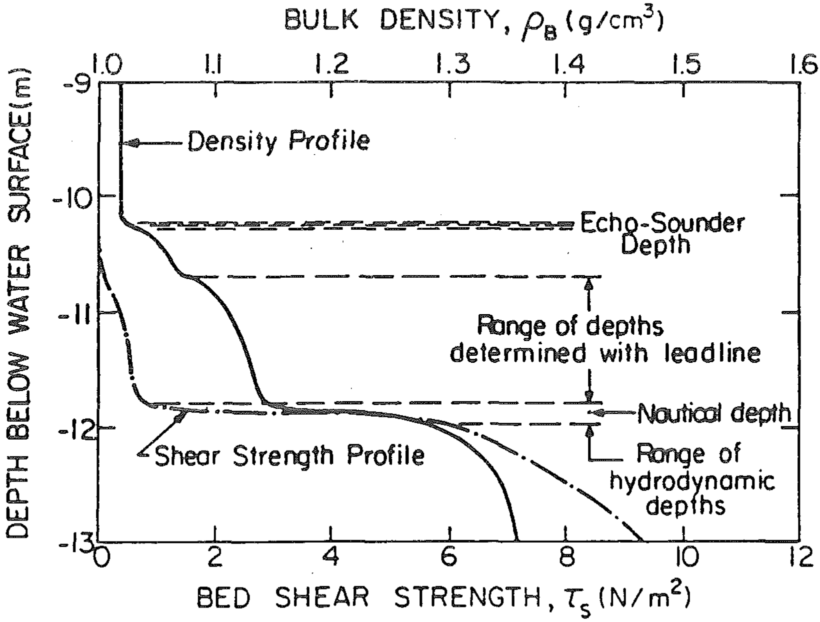

FIG. 2. Bed Bulk Density and Erosion Shear Strength Profiles and Bed Level Definitions

ture, these relationships may be determined using a geotechnical centrifuge, stress-controlled slurry consolidometer, pore pressure probe, and nuclear densimeter.

In order to assess the time-varying erosion potential of consolidating beds, a relationship of the form, $\tau_s = \zeta\rho_D^{\delta}$, where τ_s = cohesive bed shear strength with respect to erosion, ρ_D = time-varying (with consolidation) dry bed density, and ζ, δ are empirical coefficients, has been obtained from laboratory experiments (Migniot 1968). Given τ_s in N/m^2 and ρ in kg/m^3, Owen (1970) obtained $\zeta = 6.85 \times 10^{-6}$ and $\delta = 2.44$. Using different muds, Thorn and Parsons (1980) found similar values of these coefficients ($\zeta = 8.42 \times 10^{-6}$, $\delta = 2.28$). Parchure (1984) however noted that inasmuch as τ_s and ρ_D have different physical meanings, *there may be no unique* relationship between τ_s and ρ_D. In fact this relationship is very approximate, but it is quite useful for estimating τ_s in the absence of a better correlation between properties characterizing bed structure and τ_s (Hayter 1983).

In Fig. 2, a typical profile of the bulk (wet) density, ρ_B, obtained by Watts (1954) is shown together with the corresponding bed shear strength profile calculated using coefficients ζ, δ of Thorn and Parsons (1980). The dry density, ρ_D, is equal to $(\rho_B - \rho_w)\rho_s/(\rho_s - \rho_w)$, where ρ_w is water density and ρ_s is the sediment density. The density profile extends down into the bed, and shows a characteristic step-like "differentiated" structure (Parker and Kirby 1977). At about 12-m depth both ρ_B and τ_s have significant gradients. The problem of identifying the depth at which the bed is encountered is critically important, as noted later.

EROSION

While the bed shear stress is the primary flow-induced parameter characterizing the erosive force, the corresponding resistive force depends on a number of factors including sediment composition, pore and eroding fluid compositions and the manner in which the deposit is formed. The deposit itself may be in the form of a stationary suspension or a bed. The latter may be soft, partially consolidated, with a very high water content, or a more dense, settled bed. The mode of erosion varies both with the magnitude of the bed shear stress and the nature of the deposit.

Three modes of erosion have been identified. These include: (1) Aggregate-by-aggregate erosion of a bed, also referred to as surface erosion; (2) mass erosion of a bed; and (3) re-entrainment of a stationary suspension (Fig. 1). The rate of erosion, ϵ (mass of sediment eroded per unit bed area per unit time), can be expressed functionally as, $\epsilon = \epsilon(\tau_b - \tau_s, \nu_1, \nu_2 \ldots \nu_i)$, where $\tau_b - \tau_s$ is the bed shear stress in excess of the cohesive bed shear strength with respect to erosion, τ_s, and ν_1, $\nu_2 \ldots \nu_i$ are erosion resistance specifying parameters.

Surface erosion typically occurs at low to moderate values of the excess shear stress, $\tau_b - \tau_s$, and is prevalent in estuaries subject to currents of low to moderate strength. Quantification of the range of $\tau_b - \tau_s$ is highly dependent on the degree of resistance provided by the bed, which is site-specific. The specific form of surface erosion rate expression depends on bed properties. For relatively dense, consolidated beds (water or moisture content well below 100%) with uniform properties (τ_s is constant over bed depth), the erosion rate expression is

$$\epsilon = \alpha_1\left(\frac{\tau_b - \tau_s}{\tau_s}\right) \quad \ldots \quad (3)$$

where α_1 is equal to ϵ when $\tau_b = 2\tau_s$. τ_s is often referred to as the critical shear stress for erosion, in analogy with cohesionless sediment transport (Ariathurai and Arulanandan 1978). It has been found that α_1 and τ_s vary with the type of sediment, water content, total salt concentration, ionic species in the water, pH and temperature. However, α_1 and τ_s poorly correlate with bulk soil indices such as the Atterberg limits, which are inadequate indicators of cohesive inter-particle bond strength (Partheniades 1965).

Recognizing that Na^+, Ca^{++}, and Mg^{++} are three commonly found cations in soils, the sodium adsorption ratio, SAR, is found to be a useful parameter representing the influence of ionic species on α_1 and τ_s. SAR is defined as

$$\mathrm{SAR} = \frac{[\mathrm{Na}^+]}{\frac{1}{2}\{[\mathrm{Ca}^{++}] + [\mathrm{Mg}^{++}]\}^{1/2}} \quad \ldots \quad (4)$$

where the square brackets indicate concentration in milliequivalents per liter (Arulanandan et al. 1975).

For soft (water content well above 100%), partially consolidated beds, Parchure and Mehta (1985) found the rate expression for surface erosion to be

$$\epsilon = \epsilon_f e^{\alpha_2(\tau_b - \tau_s)^{1/2}} \quad \ldots \quad (5)$$

where α_2 is an empirical rate constant, ϵ_f is defined as the floc erosion rate, and τ_s generally increases with depth below the bed surface, at least over the top few centimeters, and also increases with the degree of consolidation. For thin beds, e.g., of a few centimeters thickness, the time dependence of τ_s becomes practically negligible after one or two weeks. In general, α_2, ϵ_f and τ_s in Eq. 5 depend on the same physico-chemical factors as α_1 and τ_s in Eq. 3.

When $\tau_b - \tau_s$ becomes large, or when rapidly accelerating flows occur, the bed may fail at some plane below the surface and clumps of material are mass eroded. Erosion by this process can be described approximately by an expression of the form of Eq. 3, although the rate is typically much greater than for surface erosion. Mass erosion is dominant in areas of strong tidal currents and also under storm-generated flows.

Re-entrainment of a stationary suspension is not a well-understood phenomenon. During this process, which, for instance, occurs at times following slack water and also when wind-generated waves superimposed on tidal currents act on recently formed fluid mud, the suspension density decreases with the progress of erosion, and wave-like forms tend to develop at the suspension/clear water interface with associated sediment entrainment. As the flow velocity increases the rate of entrainment can become quite rapid (Wells 1983).

Shallow and intermediate depth waves can substantially enhance the rate of bed erosion or resuspension. This rate enhancement results from bed softening under wave-induced oscillatory loading. Furthermore, if waves occur in the presence of currents, the combined wave-current bed shear stress can be quite large, and the resuspended material is easily transported by the currents. Alishahi and Krone (1964) studied erosion by wind-generated waves, and Thimakorn (1984) used mechanically generated waves. In both studies suspended sediment concentration variation with time was measured. Although no erosion rate expression was proposed in either study, the data suggest that the rate expression would be analogous to Eqs. 3 or 5. In a later study (Mehta and Maa 1986) it was found that highly stratified suspensions develop during erosion with a fluid mud layer near the bed, and a much lower concentration layer above.

In conjunction with wave-induced erosion of relatively soft beds, the associated phenomenon of bed softening, mud motion and attenuation of surface wave amplitude due to energy dissipation in the mud must be taken into consideration. In addition, the shear stress at the mud-water interface differs from the shear stress over a rigid bed as a result of the oscillatory response of the bed itself. Early observations of this type of response were made by Migniot (1968) in a laboratory flume in which he showed that the amplitude of mud oscillation is influenced by the type of mud-water mixture, bed density, mud viscosity and the degree of consolidation. These observations have since been confirmed (Mehta and Maa 1986).

SUSPENSIONS

Mobile and Stationary Suspensions

The structure and horizontal movement of mobile suspensions has been examined by workers using optical turbidity sensors, sampling and other conventional oceanographic techniques. Relatively low density suspensions, less than ~1,000 mg/L, often exhibit vertical profiles which can be approximated by classical exponential functions. Parker and Lee (1979) observed in the Severn, England, a well-mixed high energy estuary, the following characteristics of high density mobile and stationary suspensions:

1. Continuous, vertical turbidity profiles revealed that concentration was not typically a smooth exponential function of height above the bed. The most common profiles were either vertically homogeneous or, more commonly, were stratified and temporally variable.
2. As the spring to neap cycle progressed, the suspension settled through the water column so that concentrations in the upper part of the column decreased while concentrations near the bed increased. Thus, stratification in the suspension became more pronounced.
3. At times when the currents were weak the near-bed layers became stationary on the bed, but were redispersed on the succeeding tide.
4. As neap tides approached, stationary suspensions persisted progressively longer into the succeeding tidal cycle, both as a result of typically lower bed shear stresses near neap tides and higher bed shear strength due to settling and consolidation.

These observations, as well as similar ones in the Maas estuary in the Netherlands (Parker and Kirby 1977), the Pao Kye estuary in Thailand (Allersma 1980) and elsewhere along open coasts (Wells 1983) emphasize the critical need to detect and quantify near-bed sediment transport governed by the motion of high density fluid mud layers. In the majority of estuarial and coastal environments in which measurable rates of cohesive sediment transport occur, it is not unusual to find that a significant portion of the transport is in fact confined to the near-bed region, particularly in areas of low current speeds or when the flow regime is wave dominated (Mehta and Maa 1986). The ratio, β, of the near-bed concentration to depth-mean concentration (Table 2) is well above unity in these cases.

Dispersive Transport

Cohesive sediments in natural flows have four components to their motion: Brownian motion, gravitational settling, the motion of the suspending fluid, and rebounds from inter-particle collision. Brownian motion produces no net flow but can be important to aggregation and diffusion, especially for clay particles at relatively high suspension concentrations. Cohesive sediments move with a velocity component equal to that of the surrounding water. This applies to both mean flows and turbulent fluctuations in homogeneous fluids.

Appropriate diffusion coefficients must be selected to account for dispersive transport of the suspended material. Jobson and Sayre (1970) verified the Reynolds analogy for sediment particles in the Stokes range (less than about 100 μm in size). It was found that the turbulent Schmidt number is approximately equal to one, and decreases with increasing particle size. As a result, it is acceptable to use analytic formulations relating the effective sediment mass dispersion coefficients for fine sediment to the mean flow parameters as are used for momentum transfer coefficients. For instance Hayter (1983) in his numerical procedure used the dispersivity tensor based on the work of Fischer (1978) for solving the depth-integrated advection-dispersion equation for tidal flows. For computing the rates of sedimentation in small basins, Askren (1979) proposed other formulations for the longitudinal and transverse dispersion coefficients.

Under waves, both longitudinal and vertical diffusion is typically much more significant in the near-bed layer than in the upper portion of the water column. Near the bed, particularly if it is soft, the high frequency oscillatory wave boundary layer, wave-induced bed interface undulations and associated vortex shedding contribute to high diffusion rates (Maa and Mehta 1986).

Measurement Techniques

Suspended Sediment Concentration

Suspended cohesive sediment concentrations in estuaries range from 10^0 to 10^5 mg/L. There are few absolute measurement techniques adequate to the requirements of spatial resolution. As a result, it is essential to choose the most appropriate measurement technique. The choice is governed by: (1) The need to have rapid and unambiguous response over a wide range of concentrations; (2) the need to prevent the sampling method used from affecting the parameter being measured; and (3) the need to have an output which is readily amenable to automated data handling (Parker 1986). Available measurement techniques are mentioned below.

Gravimetric Analysis

In this well-known and accurate but slow technique water bottles and shipboard pumps are the two most common water sampling devices (Parker 1986).

Optical Methods

Instruments include the transmissometer, nephelometer and Secchi disk. Nephelometers are not very practical for use in estuaries since they are sensitive to only very low concentrations. Secchi disks can be used to estimate surface values only. Transmissometers, or electro-optical turbidity meters, have been successfully used to measure vertical turbidity profiles in, among others, the Severn (England), Maas (The Netherlands), James and Rappahannock estuaries (Kirby and Parker 1977; Nichols 1986).

Acoustical Methods

Several investigators (Jansen 1978; Thorne et al. 1983) have studied the use of ultrasound for sediment transport studies as well as the acoustic properties of particle dispersions (Ahuja 1974). It has been found that both attenuation and scattering are influenced as a function of frequency, by concentration and particle size (Parker 1986). Ambiguities occur at differing concentrations as attenuation, for example, increases and then decreases across a particular concentration range; thus, calibration by gravimetric analysis is essential. Ultrasonic measurements are particularly sensitive to compressible components (e.g., gas microbubbles or cellular plant tissue) in the sediment. Nevertheless, acoustic developments offer distinct advantages of non-intrusive measurement (Orr and Hess 1978; Orr and Grant 1982).

Nuclear Methods

Both the gamma-ray transmission densimeter and the backscatter densimeter measure in situ sediment density profiles over the mass concentration range of 0 to 1,600 g/L (Parker et al. 1975). The accuracy of these probes may be reduced by sorbed contaminants such as heavy metals and organic matter. If the contaminant concentrations are significant, corrections are required (Brolsma 1983). Nuclear densimeters provide a method for near absolute determination of mass concentration, although it is limited by spatial and temporal integration. However, new detector technology offers promise for practical studies over a wide range of concentrations (Parker 1986).

Bed Level

In many cohesive sediment areas the position of the bed is occupied by a transition from muddy water through watery mud to firm mud (Fig. 1). The transition may be discontinuous and may span a few centimeters or several meters. Thus, in contrast with cohesionless sediment areas, the plane of the bed is usually ill defined. In such circumstances the method used to measure the depth of water to the sediment/water interface will have a considerable influence on the result obtained.

In most practical circumstances the bed level which is of interest relates either to:

1. The level within the sediment bed which is resistant to ambient flow-induced stresses and over which water flows—this may be designated the *hydrodynamic bed* and the depth to it the *hydrodynamic depth.*
2. The level at which sediment properties deleteriously affect the safe navigation of vessels and which is of main interest to dredging—this has been referred to as the *nautical bed* and the depth to it is the *nautical depth* (Kirby et al. 1980; Nederhof and van Bochove 1981).

Thus hydrodynamic depth is the depth of interest to modelers for predicting sedimentation or erosion, loss of storage capacity in reservoirs, or fluxes of sorbed contaminants. Nautical depth is the depth of interest to design of navigation channels, formulation of dredging strategies, or management of dredging works. As observed by Parker (1986), there is no physical relationship between the hydrodynamic depth or the nautical depth and the most common techniques used in surveys, e.g., leadlines (sounding weights) or acoustic sounders (echo sounders, fathometers, sonar). For example, the ambiguity that arises from the use of fathometers for defining bed levels in areas with high density suspensions (Kirby et al. 1980; Nederhof and van Bochove 1981; Parker and Kirby 1977; Parker and Kirby 1982) lies in the parameters which affect the detection of muddy substrates by echo sounders.

The principal parameters of interest in hydrographic surveys of muddy areas are sound speed and target strength (Akal 1972; Gupta 1966). The latter can be calculated from the reflection coefficient, which in muddy areas is strongly controlled by the density (Parker and Kirby 1977). It has, however, been demonstrated that reflectors on echo sounder records depend principally on the density gradient rather than the density magnitude such that echo sounder reflectors may appear due to consolidation rather than deposition, and that the reflectors do not follow specific values of density (Kirby et al. 1980; Parker and Kirby 1982). Furthermore, the critical density gradient for detection is frequency-dependent but not in a consistent manner, as illustrated in Fig. 2. Here, the spatial variability between reflectors from commonly used (200 kHz frequency) survey echo sounders, as related to the density profile, is exemplified. Echo sounders define a target range, which has no unique correspondence with either hydrodynamic or nautical depth.

Leadlines are used as a standard method of checking echo sounder accuracy; however, leadline data are no more dependable. Watts (1954) reported a spread of more than 0.6 m in leadline depths measured under laboratory conditions and up to 2.0 m in field trials in muddy channels. Referring to Fig. 2 it can be stated that technique-defined depth (echo sounder, leadline) is in general shallower, and shows more spread than parameter-defined depth (hydrodynamic, nautical).

The parameter defined bed levels are essentially related to density. The hydrodynamic bed is contingent upon bed resistance, which is defined, albeit in a very approximate way, by bed density. Based on acceptable levels of ship hull resistance through muddy waters in port areas, the nautical bed has been specified at a bulk density between 1.1 and 1.2 g/cm^3. A range of density measuring devices have been used, including nuclear gage systems (Caldwell 1960; Parker and Kirby 1982), advanced acoustic transmissometers (Chaumet-Lagrange 1984), and electrical systems (Ariathurai and Arulanandan 1986).

For calculating the rates of suspended cohesive sediment transport from field measurements, density measuring devices are significantly more reliable than echo sounders. However, in practical terms these devices are not as efficient in covering large areas which are readily examined by acoustic techniques, and are thus more suited to problems involving soft muds or "fluff" in relatively confined areas of critical importance.

Concluding Remarks

As a consequence of complexities due to the oscillatory nature of tidal flows and associated cycles of erosion and deposition, modeling of cohesive sediment transport in many cases is carried out via solution of the advection-dispersion equation of sediment mass conservation using numerical techniques. Data collection efforts in field investigations for prediction of deposition or erosion are therefore guided by the need to provide input to numerical models.

It is recognized that cohesive sediments, by virtue of their low settling velocities, are typically advected over relatively large horizontal distances during each tidal cycle, and thus numerical model results have been found to be highly sensitive to the hydrodynamic (flow field) description. Therefore, the flow field must be characterized and simulated with a good degree of accuracy to avoid gross errors in prediction, e.g., of zones of deposition and erosion.

Particular attention must be paid to the need to obtain settling velocities in situ, as well as density (suspension) profiles from the water surface to depths at which a stable bed is encountered, i.e., bed which is not periodically eroded either under spring tides or under seasonably dependent episodic conditions. Inasmuch as scaling laws for settling velocity of cohesive aggregates in turbulent flows are not presently known, laboratory measurement of settling velocities should be used only for aiding in the interpretation of field determined values. Along these lines, the correspondence (if any) between field and laboratory determined settling velocities deserves special consideration in future research efforts. Density profiles are highly site-specific and must likewise be measured in situ, as they cannot be deduced using data from other sites or from laboratory tests.

Most other process-characterizing coefficients (e.g., τ_{cd}) for deposition, consolidation and erosion can, and in many cases need to be obtained from laboratory tests. Numerical model results indicate that presently known process descriptions in many instances yield fairly reliable predictions for estuarial erosion and deposition rates. At the same time significant gaps in understanding certain processes have been revealed. These gaps are principally associated with the identification of hydrodynamic depth and the behavior of stationary and mobile near-bed suspensions under currents and waves. Several sub-areas of research are essential for this purpose. These include the development of suitable instrumentation for measuring current velocity profiles in high density suspensions, understanding the interaction between high density suspension and near-bed hydrodynamics, and measuring the effective stress response of the bed during its formation and erosion.

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Appendix I. References

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Appendix II. Notation

The following symbols are used in this paper:

$\bar{C}$ = depth-mean suspension concentration;
C_b = near-bed sediment concentration;
C_{bi} = near-bed sediment concentration of settling class i;
C_0 = initial, depth-mean suspension concentration;
D = rate of deposition;
h = suspension depth;
k_z = depth-averaged eddy diffusivity;
P_e = Peclet number;
p = probability of deposition;
p_i = probability of deposition of settling class i;
SAR = sodium adsorption ratio;
t = time;
W_{si} = settling velocity of settling class i;
α_1, α_2 = erosion rate constants;
β = ratio of C_b to $\bar{C}$;
δ = empirical exponent in τ_s-ρ_D relationship;
ϵ = rate of erosion (mass eroded per unit bed area per unit time);
ζ = empirical coefficient in τ_s-ρ_D relationship;
ν_i = erosion resistance specifying parameter(s); $i = 1, 2 \ldots$;
ρ_B = bulk density of sediment;
ρ_D = dry density of sediment;
ρ_s = sediment density;
ρ_w = fluid density;
τ_b = time-mean value of the bed shear stress;
τ_{cd} = critical shear stress for deposition; and
τ_s = cohesive bed shear strength with respect to erosion.

Subscript

i = settling class of sediment particles.