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Sediment transport in wind-exposed shallow, vegetated aquatic systems

Allen Michael Teeter
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SEDIMENT TRANSPORT IN WIND-EXPOSED SHALLOW, VEGETATED AQUATIC SYSTEMS

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Doctor of Philosophy

in

The Department of Oceanography & Coastal Studies

by

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B.S., U. S. Merchant Marine Academy, 1970
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May 2002
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Abstract

Ecosystems with submersed vegetation are relatively shallow, physically stable and of moderate hydrodynamic energy. Submersed vegetation affect hydrodynamic friction for currents and waves resulting in increased overall frictional loss. Seagrasses shelter the sediment bed and reduce wind-wave resuspension. Bed sheltering factors were estimated from previous flume data on Laguna Madre seagrass species. Data from Laguna Madre indicate that total suspended material levels for bare areas are about an order of magnitude higher than some areas with submersed vegetation.

Waves in Laguna Madre at depths less than 2 m were found to be smaller than those expected for the same non-dimensional depths based on studies in slightly deeper waters. Waves were depth-limited and in the transition wave lengths between deep-water and shallow-water waves. A scaling of wave energy and wave period by atmospheric shear stress, rather than the conventional wind speed, was found to improve prediction of wave characteristics. Atmospheric roughness height was related to wave height and “age” (the ratio of wave celerity to atmospheric friction velocity).

Modeling requires process descriptions to be organized and prioritized resulting in model structures which might be different for different aquatic systems. Model formulations were defined with a single grain-class and simultaneous erosion and deposition (type I), and single (type II) and multiple (type III) grain classes with mutually exclusive erosion and deposition. Model formulations were compared. A type I sediment resuspension model was developed for Florida Bay, validated, and coupled to a water quality/ecological model of the system.
For Laguna Madre, a two-dimensional depth-averaged type III sediment model was developed to make annual simulations for fixed seagrass characteristics, with and without dredged material disposal. Dredged material disposal involves more sediment than the total natural sediment input to this system, and near-field deposit areas expose an appreciable sediment source to possible resuspension. Measurements near a dredge-pipeline discharge indicated that a highly-stratified fluid-mud underflow slowly moved material hundreds of meters downslope as sediment deposited. Underflow layer-averaged concentration did not change much with distance from the discharge.
Chapter 1
Introduction

Accurate observations and predictions of sedimentation are needed for various environmental and engineering assessments of aquatic systems. Often some existing condition and/or contemplated works require that effects be predicted. Assessments rely on numerical hydrodynamic and sediment models to predict flows of sediment and water, and perhaps to provide information to other models of water quality and ecosystem productivity. This dissertation is about enhancing model formulations for a specific type of aquatic system: wind-exposed, shallow water areas with submersed aquatic vegetation. These systems might include estuaries, lagoons, bays, reservoirs, or lakes.

One feature that makes such systems unique is the scale or magnitude of the feedbacks that occur between biological and physical components. One such ecosystem-level feedback loop was described by James and Barko (1994) and is shown schematically in Figure 1.1. This feedback loop includes sediment resuspension and the growth and distribution of submersed vegetation. Resuspension reduces water clarity short-term through the optical properties of suspended sediment particles and dissolved substances, and long-term by resuspending nutrients and organic material which stimulate plankton growth. This dissertation deals with the short-term, physical aspects of sediment resuspension and transport.

RESEARCH NEEDS

This research highlights the ecological and economic importance of these areas. Ecological import comes for their diverse habitat and high productivity. Economic import
Figure 1.1. Feedbacks between resuspension, water clarity, and submersed vegetation
comes for their many commercial, residential, and recreational uses. Ecosystems with submersed aquatic vegetation are particularly fragile and precious to the coastal zone.

The U. S. Army Corps of Engineers spends $500 million annually to maintain the nations 40,000 km of waterways (McAnally 1999), some of which cross or skirt shallow-water vegetated areas. Flood control and/or diversion projects control flows of freshwater water and nutrients to some of these areas. Municipal and industrial waste water, and contaminants may be present and possibly in transport. Suspended sediments are vectors for nutrients and contaminants. Sediments can be boons or curses depending on the magnitude of the local supply relative to a perceived optimum. For all these reasons, accurate sediment transport predictions are needed.

OBJECTIVES

The objective of this study is to enhance model formulations for sediment transport calculations in ultra-shallow waters (< 2 m) by improving physics-based descriptions of the following processes: cohesive interaction among particles, deposition of fluid mud, wind-wave and shear stress distributions, and bed sheltering effects of vegetation.

APPROACH AND SCOPE

Many technical aspects of fine, cohesive sediment transport deserve research and present formidable problems. Good reviews were presented by Dyer (1989) and Teisson (1991). The main issues addressed are presented in the paragraphs that follow.

Hydrodynamics in Shallow, Vegetated Systems

Submersed vegetation alters friction and bed shear stress conditions important to sediment transport. Wave and current friction in the presence of submersed vegetation, and the bed sheltering effects of vegetation are described in Chapter 2. The effects of submersed
vegetation on atmospheric friction and wave characteristics are included in the analyses presented in Chapter 5.

**Sediment Transport Formulation Issues**

A basic issue in the modeling of fine, cohesive sediment transport is whether erosional and depositional fluxes at the bed surface should be represented as simultaneous or mutually exclusive processes. A discussion of evidence and a classification scheme for model formulations are presented in Chapter 4 and 2, respectively. A comparison between model formulations is presented in Chapter 4. Example lagoon model applications are presented in Chapter 2.

**Multi-Grain Class Representation**

Fine, cohesive sediment modeling has hitherto represented sediments as a single grain class, while shallow vegetated aquatic systems normally have wide-ranging sediment particle size. The development of a multiple grain-size numerical sediment transport formulation is presented in Chapter 3 and demonstrated in Chapter 4. The formulation is tested against an analytic function, laboratory, and field data. An example application of the multi-grain formulation to sediment transport in Laguna Madre is described in Chapter 2. That application divided the size distribution into classes representing clay < 8 φ (0.004 mm); fine-silt 8 to 6 φ (0.004 to 0.016 mm); coarse-silt 6 to 4 φ (0.016 to 0.062 mm); and sand > 4 φ (0.062 mm) where φ units are the negative base-2 logarithm of the diameter in mm.

**Atmospheric and Wind-Wave Shear Stresses**

Levels of friction and frictional effects on waves can be extraordinary in vegetated areas. Chapter 2 analyzes the frictional effects of submersed aquatic vegetation on waves. Ultra-shallow waters have decreased atmospheric friction and increased bed friction because
of the interaction between depth and wave characteristics. Chapter 5 presents analyses based on field data, and the development of improved methods for predicting wave characteristics and budgets to insure shear stress is conserved globally.

**Sediment Dispersion from Pipeline Disposal**

One activity under close scrutiny in Laguna Madre, and responsible for moving large quantities of sediment, is dredging and dredged material disposal. In Chapter 6, a unique set of field measurements is presented which document the initial dispersion of the sediment material immediately after disposal. Measurements of fluid mud concentration or density, thickness, and water-column suspended solids concentration were made around an ongoing disposal operation. This information was used to formulate and validate a near-field numerical model of fluid mud underflow spreading and deposition and a previous analytic model of sediment entrainment into the water column.

**PRESENTATION OUTLINE**

This dissertation is arranged in journal-style chapters. Chapters 2 through 4 have been previously published (Teeter et al. 2001; and Teeter 2001a and b). However, the content of Chapter 2 has been changed appreciably to restore information from the original manuscript that was deleted from the published version because of space limitations, to reduce duplication with other chapters, and to add additional material.

Chapter 2 presents an overview of general modeling concerns, and alternate hydrodynamic and sediment model formulations. Physical features and model requirements are described for shallow aquatic systems with submersed vegetation, stressing the frictional and bed sheltering effects of vegetation and wind-wave forcing of resuspension. The hypothesis is that the frictional effects of submersed aquatic vegetation (both total friction
and bed sheltering) are critical to sediment transport. Example sediment model results are presented and compared to observations for two systems: Florida Bay and Laguna Madre.

A model framework for multi-grain class clay-silt sediment modeling is presented in Chapter 3 and tested against some existing data sets in Chapter 4. The hypothesis is based on that sediment cohesion couples grain classes and restricts sorting by grain size, yet grain-size affects both erosion and deposition processes.

Atmospheric shear stress and wind-wave models are developed for ultra-shallow water in Chapter 5 based on analyses of wind and wave data. Hypotheses addressed in Chapter 5 are that depth is important to both atmospheric shear stress and to waves, that the atmospheric shear stress is better for scaling dimensionless wave energy and frequency than is wind speed, and that the turbulent-rough, rather than the laminar, friction formulation is best for estimating wave friction.

Sediment dispersion by gravity-driven underflow and entrainment into the water column is analyzed in Chapter 6 using field data and models. A new underflow model formulation is presented and validated in a general way to observations of fluid mud underflow and deposit layer thicknesses and underflow concentrations from the field. The hypothesis used in the underflow model formulation is that during deposition the underflow layer collapses, losing volume to both the deposit in terms of both solids and liquid, and to the overlying water column in terms of liquid only.
Chapter 2
Hydrodynamic and Sediment Transport Modeling with Emphasis on Shallow-Water, Vegetated Areas (Lakes, Reservoirs, Estuaries, and Lagoons) 1

ABSTRACT

Modeling capabilities for shallow, vegetated systems, are reviewed to assess hydrodynamic, wind and wave, submersed plant friction, and sediment transport aspects. Typically, ecosystems with submersed aquatic vegetation are relatively shallow, physically stable, and of moderate hydrodynamic energy. They are open systems that receive flows of material and energy to various degrees around their boundaries. Wind-waves are often important to sediment resuspension. Feedbacks exist, both forward and backward, between bed shear-stress, erosion, light extinction, and submersed aquatic vegetation. Therefore, it is difficult to uncouple these components in model systems. Spatial changes in temperature, salinity, dissolved and particle material depend on hydrodynamics. Water motions range from wind-wave time scales on the small end, which might be important to erosion, to subtidal or seasonal scales on the large end, which are generally important to flushing. Instantaneous shear stresses and residual flows are both important to sediment transport.

Presently, solution of the well-known hydrodynamic and sediment transport equations over extensive spatial domains is limited by computational power, which is, however, improving. A more fundamental issue is our understanding of the basic bed/suspension

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exchange processes that comprise the model formulation. Other limitations include information on seagrass effects expressed in frictional resistance to currents, bed-sheltering, and wave damping in very shallow water under conditions of normal and high bed roughness. So far, various empirical equations have been used with wind or wave forcing to describe resuspension in shallow water. These equations have been reasonably successful at predicting suspended sediment concentration, but they require site-specific data. More detailed laboratory and field measurements are needed to improve the resuspension equations and model formulation pertaining to seagrass beds.

**INTRODUCTION**

Hydrodynamic and sediment transport models can be used as parts of integrated models, along with water quality and submersed vegetation models, to address possible ecosystem response to some event, change, or condition of interest. Submersed aquatic vegetation (SAV) is sensitive to (a) underwater light conditions; (b) flow and wave action; (c) physical bed-sediment characteristics; and (d) nutrient availability in the water column and sediment (Best et al. 2001). Recent advances in computational fluid dynamics and sediment transport model formulations enable computation of flows and sediment fluxes in three-dimensions (3D). However, spatial and temporal scales and model processes must be carefully selected for a particular study to ensure the model will solve the problem at hand and computations can be carried out in the practical sense.

Understanding the integrated effects resulting from some ecosystem change involves many areas of research, including flow modeling, turbulent mixing, transport processes, sediment transport, and/or nutrient dynamics. This chapter focuses on hydrodynamics, macrophyte interactions with hydrodynamics, and fine-sediment resuspension and transport.
Typical Physical Ecosystem Characteristics

Typically, ecosystems with submersed aquatic vegetation are shallow and of moderate hydrodynamic energy. In this chapter most examples pertain to shallow coastal systems where the SAV is composed of seagrass, such as Laguna Madre, Texas, and Florida Bay, Florida, described here. They are both micro-tidal and can have hyper-saline conditions due to low freshwater inflow, evaporation, and sluggish flushing. Both Florida Bay and the Laguna Madre have surface areas of about 1,500 km², and each is more than 75 percent covered by seagrass. Mean depth for all of Laguna Madre is about 0.8 m to mean tide level. Florida Bay is also shallow with depths of 1 to 4 m. Wind-generated waves are important to sediment resuspension in both systems.

The depth range for SAV in coastal waters can be limited by light conditions (Gallegos 1994; Czerny and Dunton 1995; and Gallegos and Kenworthy 1996). The total suspended sediment concentration (TSS) limiting growth can be estimated in a simplistic way by assuming that SAV requires 20 percent of incident light to survive long-term and assuming a relationship between TSS and water-column extinction coefficient for photosynthetically active radiation (PAR). Based on the TSS-PAR relationship found by Burd and Dunton (2000) developed for Laguna Madre, it is demonstrated in Table 2.1 that even low levels of resuspension and TSS can limit the depth range for seagrass.

Typical median total suspended material (TSM) concentrations in and near seagrass beds in the Laguna Madre are 18 mg/l, over a water column of about 1 m. At a bare area in the Laguna Madre (station LLM1), however, an annual-median TSM of 150 mg/l has been reported for a 2-m water column (Brown and Kraus 1997). The difference in TSM concentration levels between bare and vegetated areas in the Laguna Madre is striking and has been attributed partly to high wind at this area. That contrast can be seen in the remotely
sensed image of the southern part of Laguna Madre shown in Figure 2.1 that also shows station locations for LLM1 and LLM2. Light areas in Figure 2.1 have high turbidity. Over the same time period, mean current magnitudes at the un-vegetated site LLM1 were 7 cm/sec, while they were 4.5 cm/sec at the 1.3-m-deep *Thalassia-testudinum* (turtlegrass) vegetated station LLM2. Mean significant wave heights for these sites were 0.14-m for the bare and 0.04-m for the vegetated site (see Chapter 5). To a large extent, lower levels of currents and waves are caused by the increased hydrodynamic friction in seagrass areas, as will be described later.

<table>
<thead>
<tr>
<th>Depth, m</th>
<th>Growth Limiting TSS Concentration, mg/l</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>18</td>
</tr>
<tr>
<td>2</td>
<td>7</td>
</tr>
<tr>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>4</td>
<td>1</td>
</tr>
</tbody>
</table>

In Florida Bay, tripton (TSM minus algal biomass) levels range from 8 to 30 mg/l (Phlips et al. 1995). The average station median-TSS value was 18.2 mg/l for 16 stations within Florida Bay, with a mean station depth of 2.0-m (standard deviation = 0.6 m), and no significant trend between TSS and depth was observed over a 24 month period.

These coastal systems are not dynamic with respect to sediment transport under typical, non-storm conditions, as indicated by the relatively low TSS and TSM values above. Water column suspended sediment and light levels occurring most frequently are most important to SAV growth. Therefore, an integrated sediment transport modeling study might be tasked
Figure 2.1. LANDSAT 5, band 3 (0.60 to 0.69 nm wavelength) image of LLM from Port Isabel to Arroyo Colorado taken 20 August 1994 with 0.7- and 1.4-m depth contours and station locations superimposed (coordinates are UTM zone 14, NAD83, in meters)
with making long simulations of suspended sediment concentrations, which are usually low, depending on wind and wave forcing, and occasionally can be moderately high.

Concepts of Scales and Scale Interactions

Energy inputs into SAV-dominated systems originate from instantaneous wind speed, diurnal fluctuations in wind speed and irradiance, climatic fluctuations, and variations in season and catchment hydrology. In tidal waters, strong inputs can occur at both tidal and sub-tidal periods. Purging and flushing rates vary on larger time scales, while shear stresses vary over smaller time scales (although riverine shear stresses may be dependent on seasonal fluctuations in river discharge). Geometry is important to instantaneous and residual flows. Small-scale geometric features, such as a channel, can have important large-scale effects on residual flows.

Ideally, all scales of space and time should be included in a model but this is not the case today. In practice, averaging must be done to eliminate some ranges of spatial and temporal scales due to current limits on computer memory and computational speed. Averaging leads to closure assumptions. Closure assumptions require skill, experience, and often a-priori knowledge on the part of the modeler. This a-priori knowledge must increase as the averaging interval is increased, and the averaging interval increases as the intended prediction period is extended for the sake of computational economy. In general, it can be argued that a model formulation must fulfill the following requirements, given the model purpose:

1. The model must contain relevant physical processes and geometrical/topographical detail. This requires accurate computation of flows and fluxes, possibly in 3D. To accomplish this, turbulence or turbulent mixing, the transport of salt and/or heat, and sediments might also require computation.
2. A stable numerical solution must be obtained within a reasonable amount of computational effort. This practical consideration generally entails compromising on the flow and closure equations, as well as on spatial resolution.

3. The solution must be sufficiently accurate. This requires the use of numerical methods that are more accurate than the boundary and geometry data.

To fulfill these modeling requirements the modeler has to focus on the most important questions and associated processes, and make correct judgements with the information at hand. The modeler must often compromise between what is desirable and what is possible. Prediction will only provide additional information beyond statistical description as long as prediction errors do not exceed variations in the quantities being predicted. This limit will depend on the types of processes considered, errors in initial state specification, and the quality of the predictive method.

Integration of Model Processes

Bed shear stress, sediment resuspension or erosion, light extinction, and submersed aquatic vegetation (SAV) influence each other through positive or negative feedbacks (James and Barko 1994). Bed shear stresses are dependent on wind-waves (usually) and/or tides and/or freshwater inflows, water depth, wind fetch lengths, and importantly on the presence of SAV. Momentum transfer from winds to waves depends on wave steepness and is reduced by the presence of SAV. Erosion depends on bed shear stress, which is related to the fraction of the total shear-stress reaching the bed (100 percent in un-vegetated areas), and therefore on the sheltering effects of seagrass. Submersed vegetation reduces local resuspension due to wind-generated waves (Ward et al. 1984; James and Barko 1994; and Hamilton and Mitchell 1996). Sediment resuspension affects seagrass areas mainly through its impact on water clarity and light penetration. Light extinction depends on total suspended sediment
concentrations, particle size and/or flocculated state of the particles, absorption, and water chemistry as discussed by van Duin et al. (2001). SAV depends on light and nutrient availability, temperature, and physical stability.

The interactions between wind, wave, seagrass, and bed shear-stresses, and the influences of wave steepness, seagrass roughness, and resuspension are shown schematically in Figure 2.2. Several shear-stress feedbacks will be described in this chapter - seagrass to wave, wave to wind shear stress, and seagrass to bed shear stress. Resuspension affects seagrass through its influence on water-column light extinction. The feedback from resuspension to seagrass roughness, shown in Figure 2.2, operates over a much longer time scale than the other interactions shown.

Non-vegetated areas are prone to resuspension that can decrease water clarity and may prevent seagrass establishment or cause further seagrass decline (Onuf 1994). On the other hand, seagrass beds slow water movement, damp waves, trap and hold sediments (Fonseca 1996). Seagrass reduces shear stress at the sediment bed to lower levels than would occur on an un-vegetated bottom. Even though seagrass greatly increases total resistance to flow and wave damping, it mechanically transmits shear stress in its stems and thereby shelters sediment beds. Previous studies have documented the effects of submersed aquatic vegetation on total flow friction (see Madsen et al. 2001) and on wind-wave damping. Thus, feedbacks occur between the presence of seagrass, water clarity, and the establishment of new seagrass.

Model components can be linked together or coupled in various ways to reflect ecosystem functioning. Direct coupling of hydrodynamic and sediment-transport models admits the density effects associated with transported water-borne constituents (dissolved and
Figure 2.2. The coupling of shear stresses under the influences of seagrass roughness, wave steepness, and resuspension
particulate) into the momentum equations, and changes in bed elevation into the mass continuity equation. Hydrodynamic and sediment transport models can be operated un-coupled if suspended concentrations are low, and bed elevation changes are insignificant. Sediment transport can be un-coupled from SAV provided change in SAV distribution and density is small or otherwise unimportant to the problem at hand.

**Layout of the Chapter**

In this chapter, the various ecosystem components and couplings presented in the last subsection are described as they pertain to hydrodynamic and sediment transport modeling in shallow systems with SAV. The two coastal marine systems described earlier will be used as examples for seagrass dominated systems to illustrate physical characteristics, modeling requirements and approaches. In the next section, hydrodynamic modeling is reviewed and model features described using examples from three models. SAV frictional characteristics which affect hydrodynamics and wind-generated waves are described next. Sediment transport modeling is described in the last section with emphasis on model formulations and example results.

**HYDRODYNAMIC MODELING METHODS**

Geometric flexibility, high-dimensional representation, and computational efficiency are desirable attributes of a numerical hydrodynamic model. Domains to be modeled usually have irregular bathymetric and boundary features which are most accurately represented using similarly irregular, unstructured grids or meshes. Numerical simulations are worthless unless the model is an accurate representation of the system being modeled. Flow variations are best calculated using a three-dimensional (3D) model, although depth-averaged models may be appropriate for near-homogeneous systems. Hydrodynamic model simulations used
for sediment transport or water quality computations have long simulation times that place significant demands on computer resources.

Because of limits in our ability to perform massive computations, compromises are usually made to insure that a given model can be operated in an economical manner, for the problem for which it was developed. Some tradeoffs are built into models which can be traced to the developer's background in fluid dynamics and/or the flow problems he/she worked on during development. As a result, many approaches for the casting of equations, fitting of geometry, and solving equations are used in numerical models. For any specific model domain and ecosystem, other tradeoffs may be required in the model spatial resolution, state variables, temporal resolution, and simulation duration.

The features of three models are given as examples in the sub-sections that follow. CH3D (Johnson et al. 1993) is a finite-difference model that uses curvilinear grids with boundary-fitted coordinates. This gives more geometric flexibility than afforded by regular, rectilinear grids. The finite element approach offers the greatest geometric flexibility. The RMA10 model (King 1993) uses finite elements and a Galerkin variant of the weighted residual solution method. ADCIRC (Luettich et al. 1992) is a model which uses a generalized wave-continuity equation and finite elements. CH3D and RMA10 have been applied in 3D and 2D modes. RMA10 also has the capability for 1D and 2D-laterally averaged modes.

**Three-Dimensional Modeling**

Over the past decade, many researchers have developed and applied 3D hydrodynamic models. Models can assume hydrostatic or non-hydrostatic vertical pressure distributions. The latter models solve for vertical as well as horizontal momentum but at a stiff computational penalty. Most free-surface models assume that the vertical pressure
distribution is hydrostatic, and reduce the 3D momentum equations to a set of two momentum equations, and an integrated continuity equation.

The governing equations of fluid motion can be cast differently, e.g., the primitive or wave equation forms. For the spatial integration of the governing equations, either the finite difference, finite volume, or the finite element approaches can be traced to the concept of the method of weighted residuals. For model time integration, finite differences are generally employed.

Friction treatment, vertical discretization, and mixing-coefficient specification can be handled differently in models. Some models, such as RMA10, have capabilities to wet and dry areas as water level changes. Some solve additional equations such as wave-form momentum or turbulence transport equations.

The basis for governing flow equations are the 3D Navier-Stokes equations. With Boussinesq and hydrostatic assumptions, and with an eddy-viscosity closure on the Reynolds stress terms, they are expressed for RMA10 in two momentum equations and an integrated continuity equation as follows:

\[
\rho \frac{Du}{Dt} - \nabla \cdot \sigma_x + \frac{\partial P}{\partial x} - \Gamma_x = 0
\]  

(2.1)

\[
\rho \frac{Dv}{Dt} - \nabla \cdot \sigma_y + \frac{\partial P}{\partial y} - \Gamma_y = 0
\]  

(2.2)

\[
\frac{\partial h}{\partial t} + u_x \frac{\partial \zeta}{\partial x} - u_x \frac{\partial a}{\partial x} + v_x \frac{\partial \zeta}{\partial y} - v_y \frac{\partial a}{\partial y} + \int_0^\infty \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dx = 0
\]  

(2.3)

where
\[
\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} = -\nabla \cdot \left( \frac{\rho g (u_a^2 + v_a^2)}{C^2} \right) + \psi \frac{U_a^2}{g} \cos(\Theta) + \psi \frac{U_a^2}{g} \sin(\Theta)
\]

\[
D/Dt = \frac{\partial}{\partial t} + u \left( \frac{\partial}{\partial x} \right) + v \left( \frac{\partial}{\partial y} \right) + w \left( \frac{\partial}{\partial z} \right), \quad u,v,w = x,y,z \text{ velocity components}, \quad t \text{ is time}, \quad P \text{ is pressure}, \quad \rho \text{ is density}, \quad \nabla \text{ is the gradient operator}, \quad h \text{ is the depth}, \quad u_\zeta, v_\zeta = x,y \text{ velocity components at the water surface}, \quad \zeta \text{ is the water surface elevation}, \quad u_a, v_a = x,y \text{ velocity just above the bed}, \text{ and } a \text{ is the bed elevation}. \quad \Gamma_x, \Gamma_y = x,y \text{ combined Coriolis, bed-friction, and wind forces:}
\]

\[
\Gamma_x = \rho \Omega v - \frac{\rho g u_a (u_a^2 + v_a^2)^{1/2}}{C^2} + \psi \frac{U_a^2}{g} \cos(\Theta)
\]

\[
\Gamma_y = -\rho \Omega u - \frac{\rho g v_a (u_a^2 + v_a^2)^{1/2}}{C^2} + \psi \frac{U_a^2}{g} \sin(\Theta)
\]

\[
\Omega = 2\omega \sin(\phi), \quad \omega \text{ is the rate of angular rotation of the earth}, \quad \phi \text{ is the local latitude}, \quad g \text{ is the gravitational acceleration}, \quad C \text{ is a Chezy or Manning friction formulation}, \quad \psi \text{ is a coefficient from Wu (1980)}, \quad U_a \text{ is the standard-height wind speed}, \text{ and } \theta \text{ is the wind direction counterclockwise from easterly.}
\]

CH3D and RMA10 are both 3D hydrostatic models, and each uses the continuity equation

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]

(2.5)

to solve for \( w \). In RMA10, the continuity equation is solved in the differentiated form

\[
\frac{\partial^2 w}{\partial x^2} = -\frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)
\]

(2.6)
after applying appropriate boundary conditions to the surface and bottom. In the $z$-plane version of CH3D used for the environmental modeling of large, partially-stratified bays, $w$ is solved with a vertically-integrated continuity equation. The $z$-plane model has constant vertical node spacing or layer thicknesses over the model domain.

In RMA10, the $z$ coordinate is transformed to allow computational nodes to be located at the time-varying water and bed surfaces and variable vertical node spacing. The transformed vertical coordinate is

$$z^\gamma = \frac{(z - a)}{(b - a)} h + z_b$$

(2.7)

where $b$ is the fixed vertical location of the water surface, $h$ is the time varying water surface, $a$ is the elevation of the fixed bed, and $z_b$ is the time-varying location of the bed. This transformation allows more accurate calculation of horizontal pressure terms for Equations 2.1 and 2.2 than the original RMA10 transformation. For example:

$$\frac{\partial P}{\partial x} = \frac{\partial P}{\partial x^\gamma} + \frac{\partial P}{\partial z^\gamma} \frac{\partial z^\gamma}{\partial x}$$

(2.8)

where $x^\gamma, y^\gamma = x, y$ are the transformed horizontal coordinates. In this transformation $z^\gamma \approx z$ so that the transformed vertical coordinate surfaces are nearly horizontal.

ADCIRC uses a wave-form of the governing equations derived by taking the time differential of the primitive continuity equation, and the $y$- and $x$-derivatives of the $x$- and $y$-momentum equations. By substituting the latter into the time-differentiated continuity equation and adding the primitive continuity multiplied by a user-specified constant $\tau_0$, the wave-continuity equation is cast as
\[
\frac{\partial^2 \zeta}{\partial t^2} + \tau^* \frac{\partial \zeta}{\partial t} + \frac{\partial}{\partial x} \left( \frac{\partial u^2 h}{\partial x} - \frac{\partial \nu v h}{\partial y} + \Omega \nu h - h \frac{\partial}{\partial x} \left( \frac{p}{\rho} + g(\zeta - \alpha \eta) \right) + (MDS)_{x} + \tau_{\nu} u h \right)
+ \frac{\partial}{\partial y} \left( - \frac{\partial \nu v h}{\partial x} + \frac{\partial \nu^2 h}{\partial y} - \Omega \nu h - h \frac{\partial}{\partial y} \left( \frac{p}{\rho} + g(\zeta - \alpha \eta) \right) + (MDS)_{y} + \tau_{\nu} \nu h \right) = 0
\] (2.9)

where \( \alpha \) is the effective earth elasticity factor, \( \eta \) is the Newtonian equilibrium tidal potential, and MDBS is the combined momentum diffusion, momentum dispersion, baroclinic forcing, and stress terms.

The ADCIRC model formulation was developed to eliminate spurious numerical modes from the finite element solution and possesses accurate dispersion relations for long waves (Luettich et al. 1992). A model domain was developed including the U.S. east coast, Gulf of Mexico, and Caribbean Sea, verified to tidal stations, and used to predict the primary diurnal and semi-diurnal tidal constituents over a 30,000-node mesh (Westerink et al. 1993). These results have been used to drive smaller, near-shore models. The ADCIRC model has also been used extensively to predict storm surges, and these results have been entered into storm databases.

**Mixing and Stratification**

Density stratification due to temperature, dissolved substances and suspended material can be critical to vertical mixing, vertical velocity profiles, and transport. If stratification is important, the vertical dimension needs to be modeled.

Vertical eddy-viscosity and mixing coefficients are dynamically computed in CH3D based on two auxiliary variables: turbulent kinetic energy \( (k) \) and the dissipation rate for turbulent
kinetic energy ($\epsilon$). The vertical viscosity terms in Equation 2.4 are computed as $E_x = E_z = c_v k^2/\epsilon$ where $c_v$ is an empirical constant. This two-equation $k-\epsilon$ turbulence closure model includes the effects of wind shear, bottom shear, water-column velocity gradients, stratification, and vertical advection and diffusion. The scheme solves local equations for vertical transport but does not transport turbulence quantities horizontally as do some other $k-\epsilon$ models. A review of these methods is given in ASCE (1988). The vertical diffusivity coefficients used in the CH3D advection-diffusion equations, similar to the 3D Equation 2.14 presented later, are based on the viscosity coefficients adjusted for density-stratification effects.

Since advection generally dominates horizontal transport in estuaries and bays, a simpler algebraic method is used to specify viscosity and diffusivity in the horizontal plane. Such a closure would not be appropriate for problems in which small scale features are to be resolved and accurately computed. A third-order spatial accuracy advection calculation scheme based on Leonard's (1979) QUICKEST is used in CH3D and has proven to work well in partially stratified water bodies such as Chesapeake Bay (Johnson et al. 1993).

An algebraic vertical viscosity and diffusivity scheme (Mellor and Yamada 1982; level 2.5) is used in RMA10. A horizontal turbulence closure scheme (Smagorinsky 1963; see review by Speziale 1998) is used to dynamically set eddy viscosity terms in Equation 2.4 and horizontal diffusion coefficients in transport equations to be described later. The Smagorinsky method of describing horizontal eddy-viscosity and diffusivity coefficients uses mesh-element velocity gradients and areas in a generalization of a mixing-length representation for these terms.
**Layer- and Area-Averaged Models**

Models in one- or two-dimension (1D and 2D) are formulated by integrating the 3D equations of motion over appropriate dimension(s). In doing so, 1D and 2D models lose the capability to predict variations in the missing dimension(s) and some assumptions must be introduced. However, the 2D depth-averaged approach may be justified and well suited for shallow, near-homogeneous systems such as those considered here.

**Wind-Driven Circulation**

Since wind generated shear-stress acts on the water surface, a 3D model is most appropriate for computing wind-driven flow. Standard depth-averaged models use the vector of the mean flow to compute shear-stress and generally poorly reproduce wind-driven circulation in shallow water. Some special 2D depth-averaged model formulations for bed stress have been developed to improve the ability of these models to simulate wind-driven circulation especially in shallow water (Davies 1988; Hearn and Hunter 1988; and Hunter and Hearn 1989).

**Small-Scale Features**

Relatively small man-made structures including channels, depth features, flow features such as gyres, and/or recirculations require appropriate spatial resolution to be resolved in model grids and to be accurately computed. An adequate horizontal mixing formulation, such as that of Smagorinsky (1963) mentioned earlier, may be required. It should be noted that accurate, spatially-resolved bathymetric data are important in general. "Geometry is everything" is a good rule of thumb for hydrodynamic modeling.

Langmuir circulation cells are ubiquitous in shallow aquatic systems, consisting of wind-aligned roll vortices which trap floatable materials in windrows. These long spiraling circulations have alternating rotations. Plant litter often marks the downwelling
convergences between cells. In shallow water these circulations reach the bottom and respond rapidly to wind changes. Downwelling current speeds can be of the same order as the mean flows (cm/sec, Leibovich 1983) and could dominate over- to under-canopy mass exchanges. Langmuir circulations are produced by interactions between wave orbital motions and the background shear-flow (Faller 1969). Cross-cell dimensions are only about three times the water depth so circulation cells are too small to be resolved in hydrodynamic models (and would require special model formulations). Dispersion within Langmuir cells is many times higher than the background turbulent values (Faller and Auer 1988).

Reproducing very small-scale horizontal eddies isn't generally considered necessary when modeling large, deep bays and estuaries for the purpose of providing flow fields to water quality models. However, getting the vertical turbulence right and reproducing the residual circulation are extremely important for water quality modeling.

WINDS AND WIND-GENERATED WAVES IN SHALLOW WATER

Wind-generated waves produce shear stresses important to resuspension in shallow water and SAV areas and are often included in sediment transport modeling. However, some resuspension models have successfully used wind alone without calculation of wave characteristics directly. Aalderink et al. (1985) compared four models, two using maximum near-bed wave orbital velocity and two using wind speed. TSM data for a 1 m deep lake, collected hourly for two weeks, were used in the model evaluation. The two models using wind alone (with and without wind thresholds) better matched the observed TSM than did the two resuspension models using wave-induced flows. All models compared used simultaneous erosion and deposition and a background concentration that was not subject to deposition. These assumptions will be discussed later in the section on sediment transport.
Pejrup (1986) points out that, where wave heights and depths change appreciably, wind speed (being relatively constant over an area) may correlate better to TSM concentrations than wave height measured at a point. Analysis of TSM time-series from a micro-tidal estuary indicated that wind alone, regardless of direction, had the best correlation to TSM levels (Pejrup 1986). Arfi et al. (1993) tested an expression relating wind speed and water column buoyancy to calculate wind-speed thresholds for resuspension and obtained results that were similar in magnitude to the wind speeds estimated to generate waves that reached the bed. Threshold shear stresses for resuspension were about 0.05 Pa. Although wave characteristics are important, wave shear stress and the overall balance of momentum input from the atmosphere are critical to resuspension in large shallow lagoons and estuaries. These shallow water bodies respond to winds at small spatial scales (for example, depth-limited wind-generated waves, Langmuir circulation cells and buoyant eddy overturning).

Resuspension model studies of shallow systems have used wave measurements or results from wave models driven by winds to provide wave parameters for bottom shear stress calculations and have been reasonably successful at simulating TSM levels (Luettich et al. 1990; Hawley and Lesht 1992; Sheng et al. 1992; and Lick et al. 1994). Near-bed wave orbital velocity depends on (a) water depth, (b) wave height and (c) wave period and is a critical parameter for resuspension of bed sediments. Wave friction factor is another critical parameter. Short-period oscillatory currents forced by wind-generated waves are more effective at developing bed shear stress than the same current magnitudes forced by tides due to boundary layer effects (Luettich et al. 1990). Wave models have not been specifically developed for SAV areas, but waves can be predicted using a number of different methods.

In shallow water areas, waves "feel' the bottom when wave length exceeds twice the depth, and the resulting bottom stress can resuspend sediments and dissipate the waves.
Moreover, wave growth in very shallow SAV areas appears to be limited by depth, bottom friction, and fetch as described in Chapter 5.

**Wave Modeling**

Ray-methods can be used to propagate waves from offshore considering refraction and diffraction (CERC 1984). Parametric wave models use parabolic equations to transport wave energy while wave energy moves through a spectrum. Parametric wave models assume a certain wave-spectral shape and predict wave generation by winds. Mild-slope elliptical-equation models can also solve refraction and diffraction problems. However, these types of models have not been specifically developed for very shallow water nor applied to areas of high bed roughness such as occurs in SAV areas.

The most popular wave models for shallow water are the analytical methods presented in the Shore Protection Manual (CERC 1984). Relationships for wave energy loss due to bed friction and percolation were first developed by Bretschneider and Reid (1953). These relationships have been used in successive approximations of shallow-water waves where wind stress was balanced by bed friction (CERC 1984). Luettich and Harleman (1990) compared two analytical methods for estimating wave characteristics, including CERC (1984), for a large, shallow lake with a mean depth of 3.2 m. Wave measurements were collected at unvegetated sites 2.0- and 2.2-m deep. Wind velocity was collected at 2-m height above the water surface. Wave hindcasts were found to give good wave height estimates but wave period estimates were about 20 percent lower than the observations.

**Shear Stress Balance**

Whatever wave modeling approach is employed, a model requirement for an enclosed shallow water system should be that the total bed shear-stress over the model domain be less than the total atmospheric shear-stress. This cannot be assured unless atmospheric shear
stress is calculated and compared to that from waves. Some amount of wave shear-stress is normally expended where waves break at the shoreline. The interest here, however, is the wave shear-stress over the greater area of the sediment bed, and therefore shoreline processes will not be considered.

As stated earlier and shown in Figure 2.2, atmospheric, total-wave, SAV, and bed shear-stresses form a coupled system. Measurements from Laguna Madre are analyzed in Chapter 5 and it is shown there that SAV reduce wave height, wave steepness (wave height divided by wave length), and atmospheric friction factors. SAV also reduce the fraction of total shear stress reaching the bed, as will be shown in the next section.

**FRICTIONAL EFFECTS OF SUBMERSED VEGETATION**

To accurately calculate flows and shear stresses, friction coefficients must be specified in hydrodynamic models. However, our knowledge of frictional effects related to submersed plants is insufficient and incomplete to do this with confidence. Bed shear stress is critical to sediment resuspension modeling, and as mentioned earlier, submersed vegetation shelters the bed from shear stress generated largely by vegetation-induced drag. Model friction coefficients can best be determined by using water level and flow data from a site. When data are lacking, coefficients must come from experience and judgement, taking into consideration reported values from similar areas. Madsen et al. (2001) give some example friction coefficients reported for rivers with macrophytes and the effects of macrophyte cutting on friction coefficients.

**Velocity and Turbulence Effects**

Laboratory studies provide the most detailed measurements of steady-flow velocity profiles and turbulence structure for seagrass-roughened flows. Without submersed vegetation, velocity profiles are logarithmic and shear stress increases with depth down to a
near-bed, constant-stress layer. With vegetation beds, momentum is extracted from the flow mainly near the top of the plant canopy and mechanically transmitted to the bed by the plants.

Gambi et al. (1990) found shear stress and turbulence intensity maxima near the top of a submersed plant canopy. Turbulence intensity is the ratio of stream-wise root-mean-square fluctuations to mean value. Turbulence intensities were greater under seagrass canopies than at similar heights above the bed outside seagrass beds, but the under-canopy mean flow speeds were much reduced. Vortex shedding by plant stems has been found to be the dominate turbulence production mechanism in beds of emergent vegetation (Burke and Stolzenbach 1983; and Nepf et al. 1997), and a similar mechanism may be operating within submersed plant canopies.

The characteristics of vertical velocity profiles are another indication that shear stress is extracted from the flow near the top of seagrass canopies. By assuming the shear stress near the top of a seagrass canopy was equal to the bed shear stress, Christensen (1985) developed an analytic expression for the vertical velocity profile which was similar to observations made over and through a SAV canopy.

Mixing above and below submersed plant canopies can be modeled using approaches proven for other flow systems. Lopez and Garcia (1998) developed a $k - \varepsilon$ turbulence model for a hypothetical submersed vegetation system. Modeling the effects of plants on mixing above and below their canopies might require a 3D or a two-layer modeling approach. As described earlier, Langmuir circulations could have a profound effect on dispersion and vertical transport of materials in shallow systems but as yet have not been incorporated into seagrass models.
Wave Friction

Data on the wave-frictional characteristics of seagrasses are very sparse. Fonseca and Cahalan (1992) performed laboratory wave damping tests on several seagrass species collected from the Laguna Madre, Texas. Seagrass characteristics used in these experiments are presented in Table 2.2.

Percentage reduction in wave energy over a 1-m wave-flume section was measured. Two to four flow-depths were used with each species to vary the water depth/leaf length within each species.

<table>
<thead>
<tr>
<th>Table 2.2</th>
<th>Characteristics of Seagrasses Used in Wave Flume</th>
</tr>
</thead>
<tbody>
<tr>
<td>Species</td>
<td>Leaf length, cm</td>
</tr>
<tr>
<td>Halodule wrightii</td>
<td>17.2</td>
</tr>
<tr>
<td>Syringodium filiforme</td>
<td>41.4</td>
</tr>
<tr>
<td>Thalassia testudinum</td>
<td>19.4</td>
</tr>
<tr>
<td>Zostera marina</td>
<td>23.4</td>
</tr>
</tbody>
</table>

Two wave types were generated with wave periods of 0.7 and 0.4 sec and nominal wave lengths of 0.68 and 0.37 m, respectively. Waves were photographed entering and exiting the flume test section as the fifth generated-wave passed. Percent wave energy reductions were about 40 percent per meter for all species.

Species-averaged test data from Fonseca and Cahalan (1992) were used to estimate wave energy ($E_w$), power ($D_d$) and energy ($W_d$) dissipation, shear stress and friction factors. Reported wave damping per meter was converted to energy dissipation rate per second ($W_d$) using wave celerity. Since $D_d = W_d E_w$ and near bed wave orbital velocity could be estimated
by linear wave theory, shear stresses and friction factors could be estimated. For comparison, values of laminar $f_w$ were also calculated using

$$f_w = 2 \left( \frac{U_{wbm} A_{bm}}{v} \right)^{-0.5}, \quad \frac{U_{wbm} A_{bm}}{v} \leq 10^4$$  

(2.10)

The wave Reynolds numbers for these experiments were low (220 to 850) making the bare-bed resistance hard to estimate. Results for flow depth $h$, significant wave height $H_s$, and total wave friction $\tau_w$ are presented in Table 2.3.

<table>
<thead>
<tr>
<th>Species</th>
<th>$h$, cm</th>
<th>$H_s$, cm</th>
<th>$\tau_w$, Pa</th>
<th>$f_w$ test</th>
<th>$f_w$ laminar</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H.w$</td>
<td>9</td>
<td>2.1</td>
<td>0.24</td>
<td>0.082</td>
<td>0.073</td>
</tr>
<tr>
<td>$S.f.$</td>
<td>19</td>
<td>4.3</td>
<td>0.95</td>
<td>0.885</td>
<td>0.134</td>
</tr>
<tr>
<td>$T.t.$</td>
<td>11</td>
<td>3.1</td>
<td>0.52</td>
<td>0.143</td>
<td>0.069</td>
</tr>
<tr>
<td>$Z.m$</td>
<td>16</td>
<td>3.4</td>
<td>0.48</td>
<td>0.304</td>
<td>0.109</td>
</tr>
</tbody>
</table>

For some species in Table 2.3, $f_w$ values are quite high compared to laminar $f_w$ values or to an expected turbulent $f_w$ value of about 0.005 for the smooth, fine sand bed used in these experiments. Fonseca and Cahalan (1992) state that no wave damping occurred when vegetation was not present in the flume. Plants apparently reached near the water surface in these experiments and how representative they are to field conditions is not known.

Wave friction roughness heights $k_n$ were inferred from wind and wave data from the Laguna Madre (see Chapter 5), instead of from bed grain size. For un-vegetated areas, $k_n$ values were near the expected value of about 0.001 m. For $T.t.$ vegetated areas, $k_n$ values were about 0.2 m.
Current Friction

More is known about current friction than wave friction in submersed vegetation beds (see Madsen et al. 2001). Hydrodynamic friction for seagrass areas is complicated by the fact that it can not be characterized by a single coefficient. The magnitude of the frictional effect depends on the mechanical properties of the plants, how they bend in the current, and the area and distributions of their leaves. Canopy height is a key factor, and since it is dynamic with respect to the flow, the friction coefficient changes with even small changes in flow speed. As the vegetation bends the boundary roughness is much reduced.

The relation between Darcy and Manning's friction factors is

$$\frac{u_*}{u} = \sqrt{\frac{f}{8}} = n \sqrt{\frac{g}{R_h^{1/3}}}$$  \hspace{1cm} (2.11)

where \( f \) and \( n \) are the Darcy-type and Manning's friction factors, \( u_* \) is the current friction velocity, and \( R_h \) is the hydraulic radius (equivalent to depth \( h \) in wide, shallow flow).

Kouwen et al. (1969), Kouwen and Unny (1973), and Kouwen and Li (1980) developed expressions for the friction factor based on the deflected canopy height, \( k_c \), and found

$$\frac{1}{\sqrt{f}} = C_1 + C_2 \log_{10} (h/k_c)$$  \hspace{1cm} (2.12)

Initial estimates for coefficients \( C_1 \) and \( C_2 \) were 0.15 and 1.85 based on flume experiments with flexible plastic strips 0.03-cm thick, 0.5-cm wide, and 10-cm high. Kouwen and Li (1980) compiled mechanical properties for a number of grasses used to line channels, and presented a method for calculating the deflected canopy height from the mechanical properties of the plants and the shear velocity.
where $y_c$ = the un-deflected canopy height, and $MEI$ = the flexural rigidity in Newton-meters squared per unit area roughness. The components of $MEI$ include modulus of elasticity of the plant material ($E$, Pa), second moment of the cross sectional area of the stems ($I$, m$^4$), and a stem density parameter ($M$). The experiments of Kouwen and Unny (1973) used various strip un-deflected heights, densities per unit area, and flexural rigidities. Plastic strip $MEI$ ranged from 0.01 to 0.66 Nm$^2$. From these experiments and from Equations 2.11 and 2.12, it should be expected that seagrass friction is a function of these three parameters (as well as the friction velocity).

Data from 1-m into a seagrass bed, presented by Gambi et al. (1990), were fit to Equations 2.11-13. The overall results for $C_1$ and $C_2$ were 0.18 and 1.07 ($R^2 = 0.86, n = 15$), and $MEI$ was 0.043 Nm$^2$ for seagrass densities ranging from 400 to 1,200 shoots/m$^2$. Apparently, shoot density was not an important factor for friction velocity over Zostera marina in this range of shoot density (Gambi et al. 1990). As indicated earlier, $f$ decreases with increased flow until plants are fully deflected. A plot of the Gambi data, demonstrating the decrease in $f$ with increasing $u$ and the lack of a consistent shoot-density effects, is shown in Figure 2.3. For those data, height $y_c = 0.145$ m, depth $h = 0.235$ m, and the friction parameter values were developed from Equations 2.11-13. Analyses indicated that at flows less than a few centimeters per second the canopy height was unchanged and frictional characteristics were constant.

Seagrass frictional characteristics measured in flume experiments were reported by Fonseca and Fisher (1986) and canopy deflection by Fonseca and Kenworthy (1987). Fits were made to Equations 2.11-13 and results presented in Table 2.4. The hydrodynamic

\[
\frac{C_1}{\sqrt{\frac{MEI}{\rho u^2}}} = 0.14 y_c^{-0.6} \left( \frac{MEI}{\rho u^2} \right)^{0.4} \tag{2.13}
\]
conditions in the flume were not well documented but the data are useful as a comparison of four seagrass species. Fonseca and Fisher (1986) found T. most resistive, H. and Z. species moderately resistive, and S. species the least resistive. The S.f. species did not follow the expected trend of decreasing friction with increasing flow as did the other species, and therefore no fit for $C_1$ and $C_2$ could be made. The values for $C_1$ and $C_2$ derived from the data of Fonseca and Fisher (1986) are probably not as reliable as those developed from the data of Gambi et al. (1990).

<table>
<thead>
<tr>
<th>Species</th>
<th>$h$, cm</th>
<th>$f$ at $U = 20$ cm/sec</th>
<th>$C_1$</th>
<th>$C_2$</th>
<th>MEI, Nm$^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>H.w.</td>
<td>23</td>
<td>0.37</td>
<td>0.33</td>
<td>2.07</td>
<td>0.0044</td>
</tr>
<tr>
<td>S.f.</td>
<td>28</td>
<td>0.23</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>T.t.</td>
<td>23</td>
<td>1.26</td>
<td>0.67</td>
<td>0.96</td>
<td>0.455</td>
</tr>
<tr>
<td>Z.m.</td>
<td>21</td>
<td>0.76</td>
<td>0.39</td>
<td>1.79</td>
<td>0.021</td>
</tr>
</tbody>
</table>

Madsen et al. (2001) report on observed Manning’s $n$ values for rivers with submersed plant beds. Depending on plant biomass, values of 0.03 to 0.3 were found for small rivers to be inversely related to river discharge. In a densely vegetated stream, $n$ was also sensitive to discharge and was reduced from 0.2 to 0.1 when macrophytes were harvested.

**Bed Sheltering by Submersed Vegetation**

Submersed macrophyte vegetation reduces the ability of a flow to resuspend material from the bed (Lopec and Garcia 1998), decreases suspended sediment transport capacity, and reduces momentum transfer to the bed sediment surface. It absorbs momentum and transfers it to the bed directly through its stems. By doing so, plants have been reported to change bed particle size distribution, organic content, and bed morphology. Likewise, changes in the
Figure 2.3. The relation of friction factors $f$ (left) and Manning's $n$ (right) to current speed $u$ for the data of Gambi et al. (1990).
configuration of seagrass beds can appreciably change large-scale bed and near-shore morphology (Fonseca 1996).

Fonseca and Fisher (1986) estimated a Shields’ sediment entrainment function under seagrass canopies ($F_s'$) based on the initiation of sediment movement. A normal-valued Shields entrainment function $F_s$ for an un-vegetated bed with the sediment particle size used in these experiments would be expected to be about 0.06 while Fonseca and Fisher (1986) found much higher values. This indicated that greater total shear stress was required to initiate sediment movement under seagrasses, and much of the total shear stress was not acting on the bed. The increased $F_s'$ compared to $F_s$ can be interpreted as a decrease in the bed shear stress for the same total shear stress. Then the ratio $F_s / F_s'$ expresses the fraction of the total shear stress reaching the bed. Results for the four seagrass species from the data of Fonseca and Fisher (1986) are presented in Table 2.5. The fraction of shear stress reaching the bed appears in general to be inversely related to the frictional effect expected from the seagrass species.

<table>
<thead>
<tr>
<th>Species</th>
<th>$F_s / F_s'$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H.w.$</td>
<td>0.21</td>
</tr>
<tr>
<td>$S.f.$</td>
<td>0.66</td>
</tr>
<tr>
<td>$T.t.$</td>
<td>0.12</td>
</tr>
<tr>
<td>$Z.m.$</td>
<td>0.21</td>
</tr>
</tbody>
</table>

Lopez and Garcia (1998) performed numerical simulations which indicated that the fraction of the total shear stress reaching the bed might be a function of plant density. As indicated in the last sub-section, the experimental data of Gambi et al. (1990) indicated that
over a density range of 400 to 1,200 shoot/m², friction factors were about constant. While Gambi et al. (1990) did not measure bed shear stress, they found under-canopy flow to be affected by shoot density. This might indicate that the sheltering effect is seagrass-density dependent. More detailed laboratory and field measurements are needed to improve the description of sediment resuspension within submersed vegetation (Lopez and Garcia 1998).

**Combined Wave-Current Friction**

Christoffersen and Jonsson (1985) present methods for computing friction factors for waves and currents in combination using a specified roughness height. Apparently, their method is more easily coded into models than some other methods such as those presented by Grant and Madsen (1979). The friction factors are generally greater in a combined wave and current field than in either pure-wave or pure-current conditions. Lick et al. (1994) implemented this scheme to evaluate the effects of extreme meteorological events on sediment resuspension in Lake Erie.

**SEDIMENT TRANSPORT MODELING**

Three components needed to successfully apply a sediment model to a real-world situation are: (a) suspended and bed sediment data adequate to the purpose; (b) a correct model formulation; and (c) accurate shear stresses from a hydrodynamic and/or wave model. Shallow, vegetated aquatic systems have low TSM and currents, and the sediment material in suspension is fine-grained under most conditions. SAV respond to frequently-occurring underwater light conditions and can tolerate short periods of low-light. Therefore, since the interest here is in the normal range of TSM conditions and not extreme events, fine-grained silts and clays are the subject of the modeling described in this section. Fonseca (1996) gives example cases of seagrass effects on sedimentary features and near-shore geomorphology.
For three dimensions, the advection-diffusion equation for sediment transport is

\[ \frac{Dc}{Dt} - \frac{\partial}{\partial x} \left( D_x \frac{\partial c}{\partial x} \right) - \frac{\partial}{\partial y} \left( D_y \frac{\partial c}{\partial y} \right) - \frac{\partial}{\partial z} \left( D_z \frac{\partial c}{\partial z} \right) = 0 \]  

(2.14)

where \( c \) is the concentration of suspended material, \( Dc/Dt \) is the total derivative that includes advection, and \( D_x, D_y, D_z = x, y, z \) eddy diffusivity components for sediment mass.

Diffusivities generally include the effects of small-scale motions such as Langmuir circulations described earlier, as well as turbulence. The effective vertical velocity \( w_e \) is substituted for \( w \) in the total derivative \( Dc/Dt \) where \( w_e = w - W_s \), and \( W_s \) is the settling velocity. The surface boundary is given a no-flux condition and the bottom boundary condition is

\[ \frac{\partial}{\partial z} \left( w_e c - D_z \frac{\partial c}{\partial z} \right) = E - D \]  

(2.15)

where \( E \) is the vertical erosional flux and \( D \) is the vertical depositional flux. Dimensions of the erosion and deposition fluxes are mass per unit area per unit time. Various sediment model formulations specify expressions for \( E \) and \( D \) and conditions over which they operate.

**Model Formulations**

A variety of formulations have been used to model wind-wave resuspension in shallow water. Model formulation depends on study objectives, available data, and the modeler's philosophy of fine-grained sediment transport. Various approaches for the study of resuspension of fine estuarine and lacustrine sediments by wind-waves have been taken by Pejrup (1986), Petticrew and Kalff (1991), Hamilton and Mitchell (1996), Ariathurai et al. (1977), Luettich et al. (1990), Hawley and Lesht (1992), and Lick et al. (1994).
Resuspension models were developed along at least two separate lines of research. One was based on controlled laboratory experiments, and the other based on field measurements of suspended sediments and various forcing functions. Several model paradigms have arisen based on the objectives of the model and availability of data. Some model formulations use single sediment components. Some add a background, non-depositing sediment component. Some models use multiple grain-class components. There are reasons for all these formulations. For discussion, the spectrum of approaches is summarized into three formulation types listed hereafter.

I. **Simultaneous erosion and deposition**

Several modelers have found that, for a single grain and bed component, allowing deposition to occur continuously and erosion to either occur continuously or only when some threshold has been exceeded, improves model verification to field-observed TSM.

II. **Exclusive erosion and deposition**

Based on the preponderance of laboratory results, models with non-overlapping thresholds for erosion and deposition have been used successfully for estuarine sediment shoaling prediction, but have been found not to predict low-level TSM very well.

III. **Multiple grain and bed components**

Multiple grain classes have been used to represent clays and silts in the water column and in a layered bed structure which allows variable depositional, erodibility, and consolidation characteristics to develop in the model domain. Each grain class erodes or deposits, not both. Suspension concentrations can be predicted accurately, and spatial and temporal variations in sediment texture are included in model calculations.

Type-I single-grain sediment models which allow simultaneous erosion and deposition (Aalderink et al. 1985; Luettich et al. 1990; Hawley and Lesht 1992; Lick et al. 1994; and...
Hamilton and Mitchell 1996) have been previously found to predict resuspension more accurately than single-grain models with mutually exclusive erosion and deposition. Sanford and Halka (1993) list references to experimental studies which have addressed simultaneous erosion and deposition. Lau and Krishnappan (1994) is a recent experimental study that found mutually exclusive erosion and deposition. One explanation for the failure of single-grain models with mutually exclusive erosion and deposition is that natural sediments with silt and clay have multiple thresholds (Sanford and Halka 1993). One reason for the success of the models that allow simultaneous erosion and deposition is that concentration-dependent depositional flux acts as a penalty to limit erosion, similar to decreases in sediment bed erodibility with erosion depth into the bed observed in laboratory experiments (Dixit 1982).

One formulation or another may be most appropriate for a certain setting, study objective, or data set. If only TSM is to be predicted, and not sediment transport per se, then the first formulation may be appropriate. If changes in bed texture, appreciable bed erosion, or other transport issues are involved, the type II or III formulations may be best. For accurate TSM prediction, type I and III appear to be the best alternatives.

Cohesive Sediment Characteristics

Important differences between coarse and fine-grained sediment transport characteristics can be attributed to cohesive effects. Cohesive forces act at very small distances and are affected by clay mineralogy, ion content and composition, pH, and temperature. Cohesive bonding under field conditions also includes organic coatings and steric bonds of organic origin. Cohesion acts to form several structural levels of progressively-weaker aggregation for clay minerals. Three general differences between cohesive and coarse-grained (greater than 62 \( \mu m \)) sediment transport under moderate shear stresses are listed as follows:
1. Cohesive sediments are only transported in suspended state
   
   Coarse-grained sediments are also transported in quasi-contact with the bed as bed load.

2. Cohesive sediments are not transported as dispersed, individual particles
   
   Flocculation increases settling velocities by many orders of magnitude and is responsible for deposition.

3. Cohesive sediment beds undergo appreciable volume and erodibility changes with time
   
   When rapid deposition occurs, deposits are light and have little hydraulic shear strength.
   
   Cohesive beds can be uniform but more often are vertically stratified by density and hydraulic shear strength.

   As discussed earlier, short-period oscillatory currents forced by wind-generated waves are more effective at developing bed shear-stress than the same current magnitudes forced by tides due to boundary layer effects. Another important factor is that wave-pressure oscillations weaken cohesive beds and make them more susceptible to erosion (see the review of Mehta et al. 1994).

   Cohesive sediment transport is difficult to study and significant gaps in our understanding remain. The issue of simultaneous erosion and deposition is one example. The transport properties of fine-grained, cohesive sediments cover a bewilderingly wide range. The reviews of Dyer (1989) and Teisson (1991) identify knowledge gaps and areas where further research and model development are needed.

A Type-I Model Formulation and Application to Florida Bay

   The focus of the fine-grained sediment model is the sediment/water interface where deposition and erosion flux boundary conditions are calculated for the suspension and the bed. Numerous type I models have been developed and applied to shallow water systems (Aalderink et al. 1985; Luettich et al. 1990; Hamilton and Mitchell 1996; Hawley and Lesht
As Hawley and Lesht note, model parameters vary spatially, so data from many locations are necessary.

**Model formulation.** A water quality model (WQM) of Florida Bay was developed to address issues relating to freshwater flow, nutrient inputs, water column productivity, and seagrass (Cerco et al. 2000). Water column suspended sediment and phytoplankton concentrations modify the availability of photosynthetically active radiation to seagrasses and benthic algae. Conversely, seagrass, and to a lesser extent benthic algae, modify wave climates and sediment resuspension. To address water quality and seagrass issues, simulations of sediment resuspension are required to be coupled to the WQM.

Simplifying assumptions such as a single grain class and independent erosion and deposition processes could be made (similar to the previous studies cited) to reduce resuspension model complexity and computational burden, as long as changes to sediments and depths are not varied for plan tests.

A type I model formulation with simultaneous erosion and deposition was developed and implemented in the Florida Bay WQM

\[ E - D = c_3 \tau_b^n - W_z c \]  

(2.16)

where \( c_3 \) is a constant and \( n \) is an exponent, both varied over the model domain. The first term on the right hand side of Equation 2.16 is the erosion function from Lavelle et al. (1984). In the model, \( \tau_b \) was calculated spatially using specified bare-bed atmospheric friction factor coefficients, which were modified dynamically depending on local SAV biomass.

Above ground seagrass biomass and patchiness were used to parameterize seagrass frictional effects. The model was formulated such that an order of magnitude increase in this parameter decreased atmospheric friction factor by 20 percent and reduced \( \tau_b \) from 97 to 40
percent of the total shear stress. The latter corresponds roughly to the magnitude of SAV
bed-sheltering described in Table 2.5. Linear ramps were used between end points.

**Florida Bay site description.** Florida Bay is a 1,500-km² lagoon system consisting of
largely interconnecting banks and associated islands semi-enclosing shallow 'lakes' located
between the southern tip of the Florida mainland and the Florida Keys. Banks are very
shallow, restrict flows and flushing, and affect waves. The 237 islands greater than 100 m²
have a mean area of 0.11 km², median of 0.02 km², and maximum of 1.68 km² (Enos 1989).
Islands constitute 1.73 percent of the total bay area. The entire bay is underlain by
Pleistocene limestone at roughly 2 to 5 m depth deepening east to west. Windward (north
and east) edges of islands and banks are mostly erosional and are composed primarily of
sand- and gravel-sized material. Leeward (south and west) edges of banks are depositional
and have finer sediment texture.

Florida Bay sediments are predominantly biogenic carbonates which vary considerably in
texture from place to place. Lake bottoms are primarily carbonate mud with sand- and
gravel- sized components. Sediments migrate from lake bottoms, where most are produced,
onto relatively stable banks, or out of the system (Bosence 1989). The northeastern or
interior portion of the bay is sediment starved, with thin banks and some bare rock bottom
(Bosence 1989). Thicker sediment beds occur in the central bay. The western portion of the
bay contains a relatively deep (18 m) area called the 'sluiceway' which has a rock and shell
bottom, and is reported to be frequently scoured (Schomer and Drew 1982).

Previous, rather sparse, sampling indicated that, overall, bay bed sediments are 52 percent
finer than 62 µm, and on average a bi-modal mix of sand, silt, and clay sized material. A
recent study analyzed 600 samples from the bay (Prager et al. 1996), but results were not
available at the time of this study.
Shallow water carbonate sediments are relatively rare in the U.S. coastal zone, and few studies have reported on the erodibility of sediments such as those which occur in Florida Bay. Calcareous silt from the deep ocean, which might be similar to some fine sediments in the bay, has been previously found to be eroded by low near-bed current speeds (Southard et al. 1971). The erodibility of Florida Bay sediments depends on sediment characteristics as well as the nature and quantity of algae and other organic materials which generally reduce the erodibility of sediments.

Resuspension is driven by bed shear stress generated by tidal and wind-driven mean currents and by wind waves. Tidal currents are generally weak except through bank cuts or passes. Wind-forced currents with subtidal periods are very important to circulation and flushing. Winds commonly move larger volumes of water than do tides in the interior of Florida Bay (Enos 1989), and 85 percent of the alongshore velocity variance on the nearby West Florida Shelf is in the sub-tidal frequency band (Mitchum and Sturges 1982).

Philips et al. (1995) made water column observations and sampled at 17 stations in Florida Bay monthly over a one year period. Tripton (TSM minus algal biomass) levels ranged from 8 to 30 mg/l, with higher values in the western bay. They reported that tripton was responsible for 54 to 92 percent of the water column light attenuation, with chlorophyll-containing particles the next most important contributor.

Model application to Florida Bay. The resuspension model was solved over a 1,060-cell finite-volume grid and coupled to water quality constituent transport and SAV growth (Cerco et al. 2000). Flows for the Florida Bay water quality and resuspension model were developed from a 12,473-element RMA10 2D mesh of the system (McAdory and Kim 1998). Water level boundary conditions for the hydrodynamic model were imported from an
ADCIROC model of the western North Atlantic Ocean. Winds for the resuspension model were interpolated to the center of the study area and input every one-hour time step.

Two wind data sets were used. The verification period for the WQM was 1996-1997, and winds for this period were compiled from National Oceanographic and Atmospheric Administration, National Center for Environmental Prediction (NCEP), Mid-Range Forcast (MRF) winds. MRF winds are produced on a 1° global grid. MRF winds were only available for 1994 through 1997. For the anticipated study prediction period of 1988 through 1997 NCEP’s Reanalyzed Global (RG) winds would be used. This data set covers 1976 through 1998 with a global grid of about 2°. There is tendency for the RG to be lower than the MRF winds, and a comparison of the MRF and RG winds indicated that, indeed, the RG winds were lower for this period.

Initial simulations were performed for the WQM verification period 1996 through 1997 using 13 months of data from Philips et al. (1995) which actually covered portions of 1993 and 1994. Those samples were collected about monthly from a boat moving station to station. A rough verification was performed using means and standard deviations from 17 stations, and the MRF winds. Station locations are shown in Figure 2.4. The coordinates shown in Figure 2.4 are in units of feet east and north of NAD83 geographical coordinates 24° 30'N and 81° 30'W. The model was adjusted by regions or areas within which water column conditions were found to be similar, as previously suggested by Philips et al. (1995) and shown in Figure 2.5.

The Florida Bay model was verified to the two-year period 1996-1997. For the final verification, a longer TSS data set was obtained for 16 stations and covered 1994 through mid-1996 - about 937 days. The winds for the final verification were extracted from the RG wind data set. Settling velocity was set as a constant for the system at 0.03 m/day \((3 \times 10^{-7})\).
m/sec) and \( n \) was assigned a constant value of 4. Coefficient \( c_3 \) in Equation 2.16 was varied spatially, within the areas shown in Figure 2.5, such that 0.123, 0.21, and 0.59 were the 25, 50, and 75 percentile values.

Since TSS and wind data were from different periods, the resuspension model was verified in a statistical sense assuming that the data were representative of “normal” conditions. An example station TSS time-series and statistical distribution are shown in Figure 2.6. As can be seen, the timing of concentration peaks was not always predicted accurately by the model, as expected since wind and TSS data were from different periods. Sampling data did not allow assessing the magnitude of diurnal or other short-period fluctuation in TSS.

The most important requirement of the model was to predict the central part of the frequency distributions critical to light availability and seagrass productivity. The model was reasonably successful in this sense. In Figure 2.6b, inorganic solids are plotted on a log scale against quantiles of the standard deviation about the median. Straight lines indicate log-normal distributions. Central portions of the model and field statistical distributions in Figure 2.6b are similar. Comparisons of 25, 50, and 75 percentile field and model TSS for all 16 stations are plotted in Figure 2.7.

A Type-III Model Formulation and Application to Laguna Madre

Multiple grain class fine-grained sediment model formulations have been presented by Lavelle (1993), Le Hir et al. (1993), Chester and Ockenden (1997), and Teeter (2001a and b). Cohesion causes individual flocs or aggregates to be a mix of particle sizes (Kranck 1980; and Lau and Krishnappan 1994), yet spatial variation of dispersed grain-size distributions correlate to field transport processes (Gibbs 1977; McLaren and Bowles 1985; and Teeter 1993b). By admitting multiple grain classes, this formulation included a higher degree of
Figure 2.4. TSM monitoring station locations and depths in Florida Bay (coordinates are in feet as described in text)
Figure 2.5. Model adjustment areas of similar water column conditions for the area shown in figure 2.4
Figure 2.6. Example Florida Bay results for station 11 for the 1996-1997 verification period (a) model and 1994-1996 field TSS time series, and (b) cumulative statistical distribution
realism than in a single grain class formulation for the case of a system with relatively wide-ranging particle sizes.

Model formulation. A type III sediment model formulation, similar to the one described in Chapters 3 and 4, was implemented. Other model features were developed to account for (a) spatially-varying atmospheric friction coefficients to improve computation in very shallow and vegetated areas; (b) partitioning of atmospheric shear stress to account for that going to waves and currents; and (c) bed-sheltering effects of seagrass. The shear stress sheltering factors for seagrasses were described in an earlier section. Atmospheric shear stress and its partitioning are described in Chapter 5.

Coupling between grain classes, an important consideration for a cohesive sediment model, was suggested by Kranck and Milligan (1992) and is described in Chapter 3. The coupling scheme greatly reduces the sorting that would occur if grain classes were independent or un-coupled, but produces more sorting that would occur if classes were totally coupled (equivalent to a single class).

Because fine-grained cohesive transport processes are not dependent on individual particle size intrinsically, grain-class size designations are nominal in that they do not govern transport properties directly. In the model, the finest class represents the most cohesive clay and fine-silt sized particles, and is referred to as the cohesive fraction. Since cohesion varies with mineralogy and other factors as described earlier, site-specific information was used to characterize grain-class transport properties for Laguna Madre.

For the type III model formulation, a layered bed algorithm was developed with variable silt concentrations by layers, depending on initial conditions and on erosion and deposition history. Erodibility is also linked to the structure of the bed (Dixit 1982) and sediment models often use a layered bed structure (Ariathurai et al. 1977; Teisson 1991; and Hamm et
Figure 2.7. Comparison of field TSS to model results for a 24-month simulation of Florida Bay
The effect of suspended sediment concentration on floc settling rate due to flocculation was included in the model. Descriptions of these features are given in Chapters 3 and 4.

The sediment model formulation was implemented in the U.S. Army Corps' TABS-MDS model. TABS-MDS is an enhanced version of the RMA10-WES and RMA10 models (King 1993) described in an earlier section. TABS-MDS was given new capabilities for this study.

**Laguna Madre site description and sediment issues.** The Laguna Madre is 183 km long, consisting of two shallow tidal bays transected and connected by the Gulf Intracoastal Waterway (GIWW), and extending from Corpus Christi Bay to Port Isabel, Texas, near the Mexican border. The average depth of these bays is about 1 m and the total water surface area is about 1,500 km². The GIWW is 4.3-m-deep and 38.1-m-wide (14-ft-deep by 125-ft-wide) and was completed in 1949. Maintenance dredging to keep the waterway navigable is typically performed on 18 to 24 month cycles, depending on the channel reach, and the average channel sedimentation rate is about 1.6 x 10⁶ m³ (2.1 x 10⁶ yds³) per year (PBS&J 2000). Solids content of clayey-silt channel deposits averaged about 500 kg/m³ (n = 30, range 330 to 630 kg/m³).

The system has very low freshwater inputs and is normally slightly hyper-saline. The tide range is about 0.3-m at tidal inlets, and decreases appreciably away from the inlets. Tidal currents and circulation are weak. Wind-waves and wind stress play important roles in transport and in sediment resuspension (James et al. 1977; Onuf 1994; and Brown and Kraus 1997). The system experiences strong southeast winds most of the year. From October through April, fronts bring strong north winds at intervals of 3 to 7 days. Winds generally have a stronger diurnal component in summer months, while winter winds have strong low-frequency components (Brown and Kraus 1997).
Barrier islands (Brazos and Padre Islands) separate Laguna Madre from the coastal ocean and are mostly sand-sized sediments (125 to 2,000 μm) with little silt and clay. The Rio Grande River previously discharged through southern Lower Laguna Madre (LLM) and formed fine-grained flood-plain deposits. Remnant Rio Grande deltaic deposits affect sediment texture in the southern part of LLM. The Arroyo Colorado was previously a Rio Grande channel. From Port Mansfield southward, the western shore of LLM includes areas of less-than 4,500 year old deltaic deposits and reworked fluvial sediments (White et al. 1986). The bay bottom along that same shoreline ranges from coarse to fine silts (8 to 62 μm), with a few areas of very fine sand (62 to 125 μm).

North of Port Mansfield, the western shoreline and lagoon bottom sediments are coarse grained sediments greater than 62 μm consisting of eolian deposits and a barrier system. The north shore of Baffin Bay consists of older fine-grained deltaic deposits covered with a veneer of sand (White et al. 1989). The bottom sediments along the central axis of Baffin Bay are fine-silts and clays less than 16 μm.

The lagoon system receives sediment loads from streams, the coastal ocean, biogenic sources, the shoreline and dredged-material bank erosion. GIWW construction formed mounds of lagoon sub-bottom sediments along the waterway which have been subject to erosion. In addition, sediment dredging occurs as a matter of routine GIWW maintenance. Much of the maintenance material disposed along the GIWW eventually is resuspended and dispersed, although disposed sediments do not represent a new source of sediment to the lagoon but a recycling of material that deposited out of suspension. Channel maintenance materials are mainly fine-grained silts and clays less than 62 μm.

Placement Areas (PA) 233 and 234 located about 12 km north of Port Isabel receive the highest dredged material volumes in Laguna Madre (PBS&J 2000) and the adjacent channel
area has long been identified as one of the major deposition and channel maintenance problems along the Texas GIWW (James et al. 1977). The high deposition and dredging rates have been attributed to the combined effects of cross-channel flow and high TSM concentrations in this area (James et al. 1977; and Brown and Kraus 1997). A satellite image is shown in Figure 2.8 which is similar to the image presented in Figure 2.1 but with higher resuspension. Station LLM1 is near PA 233, which has received the greatest disposed dredged material volume, and is in the high turbidity zone shown in Figure 2.8 crossing the GIWW. Both PAs have had peaks in disposal volumes related to hurricane and other severe storms as noted by Brown and Kraus (1997). On a per dredging cycle basis, however, disposal volumes have been relatively constant and have not significantly increased or decreased (PA 233 p-value = 0.58 and PA 234 p-value = 0.53).

The annualized disposal data show a pattern of high disposal in the late-1960s and 1970s. For PA 233, PA 234, and the total of PA 233 and 234; years 1952 to 1965 and 1981 to 1994 had statistically lower disposal rates than years 1966 to 1980 (p-value < 0.05) and were not significantly different from each other. Therefore, as pointed out by Onuf (1994), dredging was greater immediately after 1965 than before this date, but by 1981 decreased to pre-1966 levels and has remained at those levels for many years. The lower bound of the 1965 to 1980 period was set to correspond to the date of the seagrass survey. The actual peak period of increased dredging appears to be 1966 to 1978 (total 4,576,860 m$^3$ and 381,477 m$^3$ annual). It appears that this peak dredging period was an anomaly that lasted 12 to 15 years of the 45 year record. We have better information on TSM levels during the last period due to monitoring. However, it is difficult to reconcile the present TSM and the dredging rates of the 1952 to 1965 and 1981 to 1994 periods with the assumption that the Lower Laguna was vegetated in the earlier period. This information does not seem consistent with the
Figure 2.8. LANDSAT 5, band 3 image of LLM near Port Isabel taken 24 November 1994, with 0.7- and 1.4-m depth contours and station locations superimposed (coordinates are UTM zone 14, NAD83, in meters)
seagrass/TSM feedback observed in other systems, and observed spatially over much of Laguna Madre, and the assumption that TSM is responsible for channel deposition.

The basis of much environmental concern is that an appreciable area of LLM apparently became bare sometime between a seagrass survey in 1965 and the subsequent survey in 1974 (Quammen and Onuf 1993; and Onuf 1994). A survey in 1988 (Quammen and Onuf 1993) showed no appreciable increase in bare area. The estimated area that converted from vegetated to bare was 140 km² for the LLM, while in the Upper Laguna Madre 130 km² converted from bare to vegetated. Losses were mainly in the deep areas of the LLM, the turbid area shown in Figure 2.1. Increased turbidity and decreased light penetrations resulting from dredged material disposal and subsequent dispersion were speculated to be the cause of the seagrass decline (Quammen and Onuf 1993; and Onuf 1994). The date of the establishment of the bare area is not known exactly. LANDSAT satellite images presented by James et al. (1977) show that the bare, high turbidity area had been established by 1972. Unfortunately, studies were not performed during, nor immediately after, the period when the seagrass loss was noted, and the cause or causes of the loss remain unknown. The loss of seagrass area observed in the deep parts of LLM between 1965 and 1975 was later attributed to suspended sediment inputs resulting from dredging and in-bay disposal of dredged material (Onuf 1994), or possibly from Hurricane Buelah in 1967 (Brown and Kraus 1997).

Two studies have reported long-term impacts of dredged material disposal and underwater light conditions in Laguna Madre. Onuf (1994) monitored underwater light conditions three months before and for up to 15 months after dredging and disposal operations in 1988-1989. He compared before and after disposal light conditions for seven subdivisions near Port Mansfield intended to represent homogeneous geographic areas. Light attenuation deviations between observed and expected values from a pooled, multiple regression formula using
wind, proximity to other sediment sources and seagrass beds, and depth were estimated. Deviations from before and after dredging 3-month time blocks were compared, and found to be significantly different for the subdivisions north of Port Mansfield where disposal was greatest. Brown and Kraus (1997) reported daily average attenuation coefficients monitored before, during, and one-year after October 1994 dredging at instrumented platforms near PA 233. Significant differences were found between the groups. Pre-dredge attenuation values were the lowest, then post-dredge, and during-dredging values were the highest.

Field monitoring data can be difficult to interpret because of environmental variability and the lack of controls. Light attenuation and TSM levels are sensitive to wind, to antecedent conditions, and to nearby upstream conditions. Environmental conditions are variable and do not repeat on regular cycles. Even when statistically significant differences are found, assigning a cause can be difficult or impossible. One example can be taken from Brown and Kraus (1997). As cited in the last paragraph, post-dredging light levels did not recover to pre-dredging levels. Contrary to this finding, post-dredging TSM levels did recover, but there were differences in sampling methods. Environmental modeling, on the other hand, strives to isolate the effect of disposal activity using with- and without-dredging scenarios.

The dredging data seem to conflict with reported seagrass extent. Channel deposition and dredging are related to TSM and to storm events that produce high sediment transport rates. The previously described feedback between seagrass (and SAV in general) and resuspension would indicate that a significant period of vegetated conditions in LLM should correspond to low TSM levels and low dredging rates. Yet disposal during the 1952 to 1965 period totaled $2.8 \times 10^6$ m$^3$. Since deposition and dredging are related to the suspended load, GIWW dredging should have been lower under vegetated conditions. Also, the disposal of the $2.8 \times 10^6$ m$^3$ dredged material during this period could be assumed to be at least locally detrimental
to seagrass. Two main questions raised by the dredging and seagrass data are: How could open water disposal of this magnitude be accomplished without creating a non-vegetated lagoon area? How could TSM levels have been high enough to cause the channel deposition if the entire area was vegetated?

The key issue from a physical perspective is what contribution the dispersed dredged material has on the natural cycling of sediments in Laguna Madre. The deposition rates in the Laguna Madre over the last 50 years have not been estimated, and the source magnitude of sediment inputs are poorly known. A recent sediment-budget study (Morton et al. 2001) concluded that natural sediment inputs (equivalent to about $1 \times 10^6$ m$^3$ of channel sediment) were substantially less than the quantities dredged from the GIWW and placed in the bay on an annual basis.

**Model application.** Two-dimensional depth-averaged numerical hydrodynamic and sediment transport models were developed for Laguna Madre, Texas, using the U.S. Army Corps' Surface Modeling System (SMS©) and the enhanced TABS-MDS computational model. The purpose was to predict the effects of dredged material resuspension on suspended sediment concentrations along the GIWW and in environmentally sensitive seagrass beds. Wave measurements were analyzed and used to estimate atmospheric drag coefficients and depth-limited wave properties (see Chapter 5). Measurement information from this study, from associated studies, and from previous studies were used to specify initial, boundary, sediment, and seagrass conditions in the model, and to validate model performance.

Field measurements and laboratory experiments on sediment were carried out to develop information on settling, erosion, and depositional properties for the model. Sediment bed erodibility was the key factor under investigation in this study. Erodibility of dredged
material depends on the dredging and disposal procedures as well as sediment properties. Erosion parameters described in Chapter 4 were determined by erosion experiments and characterization tests on material from the system. A description of the erosion test device and procedures is given by Teeter et al. (1997). Erosion experiments were performed on undisturbed box cores obtained in the field. Channel sediments were also used in the laboratory to create simulated dredged material slurries. Slurry samples of various densities, with and without the addition of a sand fraction, were allowed to settle and consolidate for 1 to 27 days before erosion testing. Standing time or age was found to be more important to erosion rates than density and erosion decreased by a factor of about 20 over the standing times observed. Sand additions of 40 percent by weight of solids reduced erosion rates by a factor of about 4.

Information on settling was obtained from the field and laboratory settling experiments. Field settling rates ranged from 0.12 to 0.33 mm/sec (n = 4) and laboratory column rates ranged from 0.05 to 1.0 mm/sec over a concentration range of 80 to 1,080 mg/l (n = 9). Laboratory settling experiments were performed varying both concentration and fluid shear conditions (see Chapter 3 for procedures and results). Field experiments were also carried out to measure floc size on undisturbed suspensions using photographs and image analysis (Knowles 1998). Mean floc sizes were 133 to 266 μm, most means were in the range of 200 to 250 μm.

A limited number of underwater light and turbidity measurements were made in the field which were similar to the results presented by Burd and Dunton (2000). They showed that

\[
TSS = 0.15937 K_d^2 + 13.9 K_d - 4.569
\]  
(2.17)

where TSS values are in mg/l, and \( K_d \) is the diffuse attenuation coefficient for PAR, m\(^{-1}\).
Several hundred water samples were collected for TSM determination, and dozens of these were analyzed for organic and calcium carbonate contents.

Detailed shore-to-shore bathymetric data from 1995 were compiled and used to develop a model mesh of about 20,000 nodes. The mesh had highest resolution near the navigation channel, disposal areas, and in Lower Laguna Madre. Model roughness coefficients were assigned based on the sediment type, bed roughness features, and the species of submerged aquatic vegetation by assigning elemental material types based on the union of these parameters. The effect of aquatic vegetation on hydraulic roughness was obtained from the literature as described in an earlier section.

Sediments were discretized into one clay, two silt and one sand fraction for model simulations. For a given erosion rate (mass per unit area), the model computed the change in bed elevation based on the dry density of the bed layers. The model used a layered bed structure to characterize the density and erodibility horizontally and with depth in the bed, and included bed processes such as consolidation and erodibility relationships for grain size composition and bed density. A more detailed description is given in Chapter 4.

During model adjustment, sediment parameters were perturbed about their estimated values and the response of the model observed. In this way, the combination of parameters that were physically reasonable and which minimizes the model differences with prototype data were determined. Model adjustment runs for a 1-year verification period starting 1 September 1994 were long enough to wash out the effects of initial conditions, and to reach equilibrium concentrations with respect to water mass residence times.

Example results for the first 2,800 hrs of the verification period at station LLM1 are shown in Figure 2.9 as time-series and in Figure 2.10 as statistical distributions. Field TSM data collection procedures were described by Brown and Kraus (1997). The verification
period had a disposal operation less than 1 km north of LLM1 which can be seen in the model at hours 720 to 840. Since in the model LLM1 is on the edge of the disposal area, model concentrations increased during this event but did not appear in the field data. Apparently, currents carried any plume that was generated by the disposal toward the north. Other data platforms located 0.5 km east and west from LLM1 also collected TSM samples at 0600 and 1800 hrs daily. A combined and more complete data set, referred to as “Mean field”, was constructed by averaging all available TSM values from these three platforms at 0600 and 1800 hrs daily. As can be seen in Figure 2.9, the model tracked the field variations in TSM but some peaks were over-predicted and some peaks under-predicted. Part of the cause for this deviation was that the model was driven with a composite wind field constructed from three widely-spaced wind station records rather than with local winds. The distributions of field and model TSM were both approximately log-normal. The model objective was to produce similar TSM values along the central portions of statistical distributions, as in Figure 2.10 where cumulative distributions (of standard deviations from the median) are plotted against TSM values.

In contrast to the statistical distributions at several other stations, field TSM data collected at LLM2 in 1996, shown in Figure 2.11, did not follow a log-normal trend. The model results calculated for the 1994-1995 verification period diverges from the field data at the +1 standard deviation above the median. The spatial distribution of model TSM captured the high turbidity zone in LLM, as shown in Figure 2.12 for the time of the November 1994 LANDSAT image seen in Figure 2.8.

After model verification, initial model scenarios were annual simulations with and without dredged material disposal. Alternate locations for several PAs have also been simulated.
Figure 2.9. Verification time-series TSM results for station LLM1 starting 1 September 1994.
Quantiles of Standard Normal

Figure 2.10. Verification cumulative TSM distribution for station LLM1
One purpose of model simulations was to provide suspended sediment time-series with and without dredged material disposal, at certain points within the system, to a seagrass productivity modeling team (Burd and Dunton 2000) for seagrass growth assessment. Another purpose was to provide spatial distributions of water column impacts on suspension concentrations, and light availability to seagrasses.

**Sediment Effects on Underwater Light Conditions**

Factors affecting underwater light conditions are described by van Duin et al. (2001). Flocculation affects cohesive sediments and light conditions, and is a factor that can be calculated within a sediment transport model. Particle size affects the ratio of mass to surface area, and light scattering. Most fine sediments do not exist in the environment as individual particles but as flocs or particle aggregates. Gibbs and Wolanski (1992) measured the optical effects of varying floc size using two fine-grained sediments. Floc size was manipulated by changing carousel-flume flow. When large flocs were disrupted, backscatter intensity about doubled for the same TSM. Gibbs and Wolanski (1992) concluded that floc-state is important to the optical properties of suspensions.

Flocculation can be modeled by calculating particle motions and collisions which result in floc formation or breakup. However, this approach has not been practical for an ecosystem model. Floc spectra might be used along with site information to dynamically calculate floc surface area per unit volume in a sediment transport model to provide information to a light model.

**CONCLUSIONS AND RECOMMENDATIONS**

Hydrodynamic and sediment transport models have features including appropriate physically-based equations, geometric flexibility, adequate accuracy, and economical computing. Often, tradeoffs are required to successfully complete a particular study with a
Figure 2.11. Comparison of model verification period (1994-1995) with 1996 field cumulative TSM distributions for station LLM2
Figure 2.12. Model TSM results for the time of the 24 November 1994 LANDSAT image (coordinates are state plane, NAD27 Texas South, in meters)
particular numerical model. Alternate spatial discretization schemes are employed. The finite difference method applied with boundary-fitted coordinates, such as CH3D, provides a degree of geometric flexibility while retaining economical computing capabilities. For example, the application of CH3D to Chesapeake Bay produced long-term solutions of acceptable accuracy. The finite element provides a greater amount of geometric flexibility but at some computational cost. The finite element discretization using the Galerkin method of weighted residuals can use standard Navier-Stokes equations, or equations cast into a wave-form. RMA10 is an example of the former and uses mixed linear and quadratic interpolation to avoid spurious modes. The RMA10 model is fully-implicit in the spatial domain and gains efficiency back by taking long time steps. ADCIRC uses the wave form the governing equations and uses linear elemental interpolation. ADCIRC first solves the wave-form of the governing equations, then solves the original momentum equations which contain certain boundary conditions which are solved iteratively. These models have been successfully applied.

Hydrodynamic models applied to seagrass vegetated areas require enhancement to accurately model flows and shear stresses, taking submersed plant deflection by currents into account. This is because hydrodynamic friction in seagrass beds can not be characterized by a single parameter as in standard friction formulations (such as those of Chezy or Manning). To improve the description of sediment resuspension within submersed vegetation, more detailed laboratory and field measurements are needed to quantify the bed sheltering effect of SAV.

Resuspension models have generally used the type I formulation that includes simultaneous erosion and deposition, and have successfully modeled TSS concentrations. A drawback of these models is that parameters vary spatially and require data at multiple
locations. Model comparisons indicate that the type-III multiple grain-class formulation describes erosion and deposition processes better than the single-grain type II formulation. The additional information required by the type III formulation shows less variation than the properties of the cohesive fraction, used in the type II formulation. Type I and III formulations are recommended for modeling resuspension where predicting TSS concentration is the model objective. Where significant erosion and deposition occur, or where bed material properties such as grain size and bulk density vary, the type III formulation is recommended.

So far, application of sediment transport models to seagrass systems have focused largely on calculating suspended solids concentration, which were subsequently related to light penetration in separate SAV models. New avenues for model application may include a wider range of ecological research topics. For example, models might be used to explore the potential equilibrium between bed-sheltering, bed sediment stabilization, and local accretion or mounding, and biogeochemical balancing of organic material accumulation, nutrient uptake and plant substrate requirements.
Chapter 3  
Cohesive Clay-Silt Sediment Modeling Using Multiple Grain Classes: Settling and Deposition \(^2\)

**ABSTRACT**

Settling and deposition process descriptors are developed for a multiple grain-size class numerical sediment transport model. Grain class settling rates are calculated to span floc settling rate distributions. Depositional fluxes are coupled from the coarsest to the finest grain size class in proportion to class concentrations, consistent with the analytic model of Kranck and Milligan (1992) and other previously observed grain-size distributions. Numerical deposition experiments display characteristic features of observed grain spectra, as well as trends in overall distribution moments, and stress-dependent steady-state concentrations. Floc settling experiments are performed to examine the combined effects of suspension concentration and fluid-shear. A new settling function is proposed and compared to experimental results.

**INTRODUCTION**

A challenge in understanding and modeling fine-grained sediment settling and deposition is to describe the relationships between the sediment grains and flocs. Since individual flocs contain appreciable sub-populations of grain sizes (Gibbs 1977; and Mehta and Lott 1987), coupling or interaction between grain classes occurs during settling (Kranck 1980). Flocculation is an important consideration for any fine-grained sediment model. Flocs greatly affect cohesive sediment transport properties, and form in fresh waters and to a

---

greater extent in salt water. Electrostatic cohesion and biochemical adhesion bind together sediment grains, small flocs, and organic matter. Flocculation has been found to produce aggregates of 0.1 to 0.5 mm modal diameter $D$ (Kranck and Milligan 1992) that settle and deposit at orders of magnitude greater rates than their constituent grains.

For modeling multiple sediment classes, sediment grains are better model state variables than flocs since they are conservative constituents in both suspension and bed, affect both erosion and deposition, and are more easily measured in the environment. Sediment grains strongly affect the flocculation process (Kranck and Milligan 1992) along with a host of other conditions such as temperature, salinity, ionic content, pH, clay mineralogy, organic constituents, etc. Grain classes are coupled by cohesion, and this coupling must be accounted for in multiple-grain class models. In this chapter, a previously-proposed grain-size distribution model is reviewed and extended to a numerical model framework for multiple fine-grained clay-silt classes.

Settling rates for cohesive suspensions in natural waters depend on concentration and fluid shear-rate conditions rather than dispersed particle size. A previously-proposed settling function (Malcherek and Zielke 1996; and Teisson 1997) assumes concentration and shear effects to be multiplicative. Settling experiments were performed using flocculators and columns, and a new function relating to settling was developed and compared to laboratory results.

The issues addressed here involve important details of the multiple grain-class implementation, specifically how various particle classes interact in a flocculent suspension and during deposition. In Chapter 4, single- and multiple-fine-grained numerical sediment transport models are compared. The multiple grain class formulation improved the deposition and, to a lesser extent, the erosion process description. Both algorithms used in
that study assumed mutually-exclusive erosion and deposition at a given instant in time for constant hydrodynamic conditions, consistent with previous laboratory investigations. Erosion and deposition can occur nearly simultaneously under varying conditions. The multiple grain class formulation reproduced the steady-state suspensions observed in laboratory tests, and resuspension in a large, shallow lake.

**SIZE-SPECTRA RESPONSE TO DEPOSITION**

Kranck and Milligan (1992) measured floc and particle grain-size spectra at an anchor station over flood and ebb tidal phases in San Francisco Bay. This work extended and generalized previous work performed in the laboratory (Kranck 1980). Floc and grain size particle diameter \((D)\) spectra were parameterized in terms of three variables \(Q, m,\) and \(K\). \(Q\) depends on the total particle concentration and the shape of the distribution, and is defined by a concentration \(C_o(D)\) at 1 \(\mu\)m diameter. The variable \(m\) defines the slope of the fine end (small-size) of the distribution when plotted log-log. The variable \(K\) is related to the fall off at the coarse end of the distribution. Kranck and Milligan (1992) found \(m\) to be constant for both floc and grain distributions. Together with \(Q\),

\[
C_o = QD^m
\]  

defines the fine ends of floc or grain-size spectra. The analytic time-dependent solution for the concentration \(C\) of a well-mixed suspension with settling is

\[
C = C_o \exp \left( -W_s t / H \right)
\]  

where \(W_s\) is settling velocity, \(H\) is the suspension depth, and \(t\) is time (Krone 1962). Kranck
and Milligan (1992) took $W_s$ to be proportional to $D^2$ as in Stokes Law, and $t/H$ to represent a settling decay term $K$. By combining Equations 3.1 and 3.2, and substituting for $W_s/t/H$ the equation describing the distributions is

$$C(D) = Q D^m \exp(-KAD^2)$$

(3.3)

where $AD^2$ defines a Stokes' settling rate, $A = g (\rho_f - \rho) / 18 \nu \rho_f$, $g$ is the acceleration of gravity, $\nu$ is the kinematic viscosity of the fluid, $\rho_f$ is the floc density, and $\rho$ is the fluid density. Their distribution model Equation 3.3 was fit to observed spectra taken during both decreasing and increasing suspension concentrations, and reflects how changing suspension concentrations affect grain spectra. Grain and floc spectra covering the range of distribution parameters found for San Francisco Bay suspended sediments are shown in Figure 3.1. Flocs formed well-sorted size spectra with $m = 2$, $Q_f = 0.0011$, and $K_f = 230C^{-0.92}$ where the subscript $f$ refers to floc components. Dispersed particles, on the other hand, were poorly sorted with $m = 0$, $Q_g = 7.8 \times 10^{-3} K_g^{-0.54}$, and $K_g = 3.3 \times 10^4 C^{-1.86}$ where the subscript $g$ refers to grains.

The values of the $Q_g$, $K_g$, and $K_f$ distribution parameters changed with total suspension concentrations while $m$ values remained about constant. Coupling between grain classes caused fine sediment to deposit when a well-sorted sediment of the same size would have otherwise remained in suspension. Links between floc and grain settling were also demonstrated by Kranck and Milligan (1992) as both maximum grain and floc size varied with total concentration and with bottom shear stress.
Figure 3.1. Floc (right) and grain (left) spectra based on tidal measurements of Kranck and Milligan (1992) for varying suspended sediment concentrations.
Though their distribution model Equation 3.3 was developed based on particle removal by settling, spectra observed during resuspension followed the same patterns in reverse. As material was resuspended, increments of material were added first to the fine and then to progressively coarser regions of the spectra. This is consistent with erosion of all grain sizes up to a limit of erodible grain size (shear stress limited erosion). Features which produce this effect were incorporated into the multiple grain size erosion algorithm described in Chapter 4.

**NUMERICAL METHODS**

The distribution model Equation 3.3 is most useful in describing measured size spectra, and explaining the effects of various settling modes on spectral shapes. Numerical fine-grained sediment models, however, normally use source and sink fluxes at the bed/suspension interface to affect concentration changes. Numerical algorithms have been developed which produce results similar to observed and distribution model grain spectra, and will be presented next. Two numerical algorithms will be described: a settling velocity algorithm to relate floc and grain settling, and a depositional algorithm that couples grain classes during deposition.

**Effects of Concentration on Settling Velocity**

Floc settling velocity is defined as the sinking rate in quiescent fluid. It affects vertical transport and distribution in the water column and maximum rate of deposition. Settling velocity of cohesive sediments varies with concentration and with fluid shear rate (Camp 1946; Krone 1962; Van Leussen 1989; and Kranck and Milligan 1992). The effect of differential settling is ignored here as the suspensions under consideration are assumed to be relatively deep and turbulent. The effect of salinity is also ignored here, though if a model were to be applied through the fresh-to-brackish zone, salinity effects should be considered.
Suspension concentration affects cohesive sediment aggregate collision frequency, floc size, and settling rate. For example, an empirical relation between median settling velocity and concentration were developed from the results of Kranck and Milligan (1992):

\[ W_s = 30.9 C^{0.99} \text{ (mm/sec)} \]

where \( C \) is the total concentration in kg/m\(^3\). Previous laboratory quiescent settling tests indicated \( W_s \propto C^{4/3} \), and the difference in the exponent was attributed to differences in turbulence conditions.

Enhanced settling occurs over a concentration range from a lower concentration limit \( C_{ll} \) to an upper concentration limit \( C_{ul} \). Below \( C_{ll} \), particle collisions are too infrequent to promote aggregation. Concentration limits for enhanced settling are shown schematically in Figure 3.2. \( C_{ll} \) is typically 50 to 300 mg/l depending on sediment characteristics. At \( C_{ul} \), collisions are so numerous that particles interact completely, causing all floc settling rates to converge to one value. At concentrations greater than \( C_{ul} \), particle interactions begin to hinder settling and dense suspensions settle as masses. Camp (1946) found the onset of concentration-hindered settling to be 1 to 5 kg/m\(^3\) for turbid river water. The author has found \( C_{ul} \) to be 1 to 10 kg/m\(^3\) for estuarine sediments.

For the multiple grain size model the general form for grain class settling velocity \( W_{s(gs)} \) is

\[ W_{s(gs)} = a_1 \left( \frac{C}{C_{ul}} \right)^{n(gs)}, \quad C_{ll} \leq C \leq C_{ul} \]  

(3.4)

where \( a_1 \) is a grain-size class average maximum floc settling velocity, \( C \) is the total concentration for all grain size classes, \( gs \) is the grain class index ranging from 1 to the number of grain size classes, \( n(gs) \) is an exponent, and \( C_{ll} \) and \( C_{ul} \) are mass-weighted average lower and upper reference concentrations, respectively, over which concentration-enhanced settling occurs. The mass-weighted averages are taken over grain size classes.
Figure 3.2. Definition of upper and lower limiting concentrations for enhanced settling, and example grain class exponents
For example:

\[ a_i = \frac{1}{C} \sum_{gs=1}^{NS} a_i(gs) C(gs) \]  

(3.5)

where \( NS \) is the number of grain classes. Unlike the power law expressions given earlier, the normalized concentration \( C/C_{\text{ul}} \) is used in Equation 3.4 so that the dimensions of \( a_i(gs) \) and \( a_i \) are mutually consistent with \( W_f \), and not affected by the magnitude of \( n(gs) \). At \( C \geq C_{\text{ul}} \), \( W_f(gs) \) equals \( a_i \) for all \( gs \) classes. At \( C \leq C_{\text{ul}} \), settling rates are independent of concentration, and equal to \( W_f(gs) \) evaluated at \( C_{\text{ul}} \).

The range of \( n(gs) \) defines the span of the floc settling distribution as in the hypothetical case shown in Figure 3.2. Though the smallest grains are not always associated with the smallest flocs and the largest grains with the largest flocs, the effect of Equation 3.4 is to produce distributions of \( W_f(gs) \) that reflect the effect of grain composition on floc settling spectra. Additional coupling between grain size classes is imposed during deposition, as described later, such that deposition of a given grain size class is not necessarily related to its settling velocity.

The exponent \( n(gs) \) can be determined empirically using information from settling tests conducted over a range of concentrations. Then, by selecting appropriate percentiles to represent the grain-size classes, fits to settling data are made. The exponent \( n \) has been determined to range below a value of about 1.33 using this method. Teeter and Pankow (1989a) found that \( n \)'s for the 50 and 75 percentile values were progressively less than for the 25 percent slowest settling fraction. Alternately, settling experiments can be performed with particle size analyses performed concurrently, and settling velocities estimated for specific size classes. For example, settling velocities were determined by size class using sediments
from New Bedford Harbor estuary by Teeter (1993b). The median settling velocity

\[ W_{s0} = 1.13 \ C^{4/3} \ \text{ (mm/sec) where} \ C \ \text{is the total concentration in kg/m}^3. \]

**Effects of Fluid Shear and Concentrations on Settling Velocity**

Fluid-shear promotes particle collisions, and can, up to a point, promote floc growth. Since aggregate collisions and turbulence at the microscale can also break flocs apart (Camp 1946; and Van Leussen 1989), only relatively low shear rates are effective at optimizing floc size. Specific data on the effect of shear rate on floc growth and settling is relatively rare for natural sediments. The measurements of Kranck and Milligan (1992) captured fluid shear effects on floc characteristics in the field. Much work has been done on shear coagulation in waste water treatment (Camp 1946). Fluid shear rate \( G \) has been defined as the root-mean-square velocity gradient, related to work input per unit volume per unit time and viscosity, and can be related to overall mean velocity and to overall turbulence intensity (McConnachie 1991). Fluid shear rate \( G \) has units of per time or Hz.

A settling velocity function including the effects of both concentration and fluid shear rate has been proposed by Malcherek and Zielke (1996), and Teisson (1997). In their models

\[ W_s = \frac{W_{s0} \ (1 + a_2 G)}{(1 + a_3 G^2)} \]

(3.6)

where \( a_2 \) and \( a_3 \) are constants, and \( W_{s0} \) is a concentration-dependent settling velocity function like those given earlier. According to Equation 3.6, increasing concentration will increase \( W_s \) for any given \( G \). However, data from Lick et al. (1993) shown in Figure 3.3 suggest that, as concentration increased from 5 to 200 mg/l, maximum floc size and floc settling rate steadily decreased. Tests used natural river sediments in salt water, and rotating disk flocculators. They found similar trends in fresh water, and for chemically treated sediments. Winterwerp
Figure 3.3. Effect of concentration on floc settling rate (data from Lick et al. 1993)
(1998) presented data from a series of settling tests on Elms estuary sediment in a flocculator column. Three tests at a constant \( G = 0.9 \) Hz indicated that as concentration increased (from 150, to 790, and 970 mg/l), floc settling velocity first decreased then increased, contrary to Equation 3.6. In both data sets, reductions in settling velocity occurred at concentrations well below those associated with concentration-hindered settling. Since little information is available to check the applicability of Equation 3.6, laboratory tests were performed, and will be described in the remainder of this section.

Experiments on the effects of both concentration and shear on settling velocity were performed on a model clay and on resuspended estuarine channel sediments. A model clay mix was prepared from pure, dry clays. Dispersed particles greater than 10 \( \mu \)m Stokes-settling equivalents were removed, and a stock mix with 40 percent of both illite and kaolinite, and 20 percent montmorillinite by dry weight was prepared. All three minerals were found by Whitehouse et al. (1960) and Edzwald et al. (1974) to flocculate at ionic concentration lower than those used here. The stock mix was allowed to stand a week after 15 ppt Instant Ocean salts were added. Clay stock mix was admixed with 15 ppt artificial seawater to make test suspensions. A natural estuarine sediment from the Gulf Intracoastal Waterway channel about 11.2 km north of Port Isabel Texas was also tested. This sediment had mean and median dispersed grain diameters of 26 and 7 \( \mu \)m, 7 percent sand (> 74 \( \mu \)m) and 64 percent less than 16 \( \mu \)m. The material had an \textit{in situ} bulk wet density of 1,408 kg/m\(^3\), and 8 percent organic content by weight.

Rotating flocculators were used to generate large flocs. Cylindrical flocculators 12.5-cm diameter by 21-cm long/deep were used to allow for settling velocity determinations after floc generation. Cylinders of 11.75-cm-diameter by 28-cm-deep were used for two higher shear-rate and a quiescent settling experiments. For higher shear-rate model clay
experiments, these cylinders were filled to 18-cm depth in a Particle Entrainment Simulator (PES). The PES is a standard device developed by Tsai and Lick (1986) that has a 2.5-cm stroke oscillating grid.

The flocculators were filled, rotated such that the outside edge or wall moved at about 1 cm/sec for 3.5 to 7 hours until floc sizes stabilized, then removed from rotator, up-ended slowly to bring the axis to the vertical, and sampled over time with a pipette to determine floc settling velocities. In the first series of model-clay experiments, laminar shear rates were varied by partially filling the cylinders, and spanned those which produced maximum or optimum floc sizes. Shear rates were estimated visually by observing small particle movement at 5 to 10 locations in the disk flocculator. Photo-micrographs and grids were used to visually estimate floc dimensions which were found to vary between < 0.2 mm and 0.8 mm diameter depending on the test. In the second series of model-clay experiments, only the optimum shear rate was used, and compared to quiescent settling tests.

The quiescent cylinder experiments were conducted by mixing and introducing sediments, and withdrawing samples from 20-cm depth for 6 hours. Care was taken to keep cylinders under constant temperature conditions. PES settling experiments were performed while grid oscillation was underway and producing mild turbulence. Oscillation rates were 70 and 140 per second producing turbulent shear rates of about 20 and 40 per second estimated using procedures presented by McConnachie (1991). The PES was operated at a rate below previous calibration range to produce mild turbulence and a shear stress of less than 0.05 Pa on the bed. Samples were withdrawn from 12.5 cm depth for 6 hours.

Two series of tests were performed on the natural sediment: one using rotating cylindrical flocculators and the other in a 10-cm-diameter by 1.9-m-high column. The disk flocculator was operated at the shear rate found to be optimum for the model clay and produced large
flocs. Samples were analyzed for total suspended material, and median settling velocities estimated.

The effects of shear rate $G$ alone on settling are shown in Figure 3.4 for model clay suspensions with initial concentrations of 46 mg/l. Also shown is a fit to Equation 3.6 with $W_s$ set to a constant equal to the median $W_s$ value observed in the corresponding quiescent test. The median $W_s$ for the quiescent test was about the same with a $G$ value of about 30 Hz. The maximum floc settling rates were more than 100 times greater than with $G = 0$, occurred at a very low shear rate (about 0.5 Hz), and dropped off sharply with increasing $G$.

Results on the effects of concentration and shear rate on $W_s$ for model clay and natural sediments are shown in Figures 3.5 and 3.6. Test data are $W_s$ values for disrupted floc (quiescent tests which began with floc disruption), and for optimum flocs ($G = 0.5$ Hz). The model clay floc settling velocity decreased as initial concentration increased from 57 to 196 mg/l (Figure 3.5). Eleven flocculator tests were performed at $G = 0.5$ Hz with the natural sediment over a concentration range of 60 to 660 mg/l (Figure 3.6). There was a slight upward trend in the linear regression line between initial concentration and median $W_s$, but the slope was not statistically different from zero. Nine quiescent column tests were performed between initial concentrations of 72 and 1,012 mg/l (Figure 3.6). Above 122 mg/l median $W_s$ values increased steadily, indicating that suspension concentrations were not hindering settling, and were generally in the range of $C_h \leq C \leq C_w$. Optimum-floc and disrupted floc $W_s$ values matched at about 500 mg/l (Figure 3.6). Therefore, contrary to Equation 3.6, floc settling rates did not increase with concentration in either model clay or natural sediment tests.

Based on experimental results, the effects of concentration and shear rate on settling velocity are not taken to be multiplicative as in Equation 3.6, and the following functional
Figure 3.4. Experimental results for the effect of shear rate ($G$) on settling velocity ($W/W_{so}$), and a fit of Equation 3.6 to the data.
INITIAL TEST CONCENTRATION, mg/l

MEDIAN SETTLING VELOCITY, mm/sec

50 100 500 1000

0.005 0.050 0.500 5.000

Flocculator Tests
Column Tests
Equation

G=0.2 Hz
G=2 Hz
G=20 Hz
G=200 Hz

Figure 3.5. Model clay settling velocity results for varying shear and concentration, and a fit of Equation 3.7 to the data
Figure 3.6. Natural sediment settling velocity results for varying shear and concentration, and a fit of Equation 3.7 to the data.
relationship is proposed:

\[ W_{s(gs)} = a_1 \left( \frac{C}{C_w} \right)^{r(gs)} \left( \frac{1 + a_2 G}{1 + a_3 G^2} \right) \exp \left( -\alpha_4 C / C_w \right) + 1 \]  

(3.7)

for \( C_w \leq C \leq C_{sd} \). Equation 3.7 is shown plotted in Figures 3.5 and 3.6 as approximate fits to the data.

**Deposition Rate**

Deposition removes sediment from the water column at a rate equal to the product of effective settling and concentration. To deposit, sediment must transit the zone just above the bed which can have very high shear rates. Previous laboratory experiments (Krone 1962; and Teeter and Pankow 1989b) have observed that effective \( W_s \) based on deposition are lower than those measured in the water column. The calculation procedures presented in this section first assess the deposition process for individual grain classes, then couple grain-size classes such that the final result depends on deposition of the coarsest active class and the grain-size spectra.

Potential deposition of each grain-size class is first assessed. Deposition is assessed differently for the cohesive fraction than for silts. The cohesive fraction is taken to follow Krone's deposition law (Krone 1962) using the concept of a critical shear stress for deposition and the depositional probability. The effective settling velocity is the settling velocity times the depositional probability \( P \) defined by Krone (1962) and for the finest cohesive fraction is

\[ P = \left( 1 - \frac{\tau}{\tau_{cd}} \right), \quad \tau < \tau_{cd} \]  

(3.8)

where \( \tau \) is the bed shear stress and \( \tau_{cd} \) is the critical threshold shear stress for deposition.
According to Equation 3.8, all sediment eventually deposits at shear stresses less than the critical value.

For each silt class, upper and lower shear stress threshold values are defined slightly differently than those used for the cohesive fraction. The lower or critical depositional shear stress is defined as that value below which all material is free to deposit. Below this threshold, silt deposition depends only on concentration and settling velocity \((P = 1.0)\). At shear stresses between the upper and lower critical values, silt sediment fractions erode or deposit at rates linearly related to the bed and threshold shear stresses.

The upper or critical erosional shear stress is defined as that value above which all material of this class will remain in suspension \((P = 0)\), but is not linked explicitly to erosion. Erosion of a silt fraction is first dependent on the erodibility of the cohesive fraction. That is, silt is held in a cohesive matrix and is not free to erode unless the bed shear-stress exceeds the erosional threshold for the cohesive fraction, as described in Chapter 4. Shear stress ranges are specified to be contiguous for contiguous grain classes. That is, the upper shear-stress threshold \(\tau_{ce}\) for one class is the same value as the lower threshold \(\tau_{cd}\) for the next coarser class.

After the potential deposition from each grain class has been assessed, grain classes are coupled such that spectral shapes follow the distribution model of Equation 3.3. As noted earlier, coupling between grain classes causes some sediment to deposit when a well-sorted sediment would remain in suspension. For the deposition of grain classes to be proportional as in Equation 3.3, settling velocities over those grain classes depositing must be equal. Other controls are also introduced to ensure that the algorithm is capable of forming steady-state suspensions at a given bed shear stress level. Thus, if \(F\) is the depositional flux, \(gs\) is a grain class with \(P = 0\), and \(gs+1\) is a larger-sized depositional grain class
where $d_1$ controls the exact proportion between the fluxes, and $d_2$ limits flux of smaller grains as $C(g_s+1)$ tends toward zero. The deposition algorithm introduces two parameters to control the proportional deposition. Data indicate that $d_1 = 1$ or slightly below.

**RESULTS OF NUMERICAL DEPOSITION EXPERIMENTS**

The algorithms Equations 3.4, 3.8, and 3.9 were used in a numerical scheme that included mass conservation equations for each grain-size class in a layered sediment bed and a suspension at a point (1-dimension vertical). Additional details of the numerical scheme and descriptions of other algorithms for erosion and sediment bed processes are presented in Chapter 4. The intention was to demonstrate the algorithm, realizing that application to most real problems would require multi-dimensional numerical modeling.

Model simulations were performed with seven grain-size classes arbitrarily assigned at nominal sizes of 4, 18, 22, 28, 37, 44, and 56 μm. Normally, grain-size classes would be spaced logarithmically by size, but in this case the finest fraction was spaced to accentuate the fine tail of the spectrum. A shear stress was initially imposed (0.16 Pa) greater than $\tau_{cd}$ for the clay fraction and less than the $\tau_{ce}$ for the next larger class. Numerical deposition experiments were performed by initializing suspensions with grain spectral slopes of $m = 1$, 0, and -1.

Time-series grain-size spectra from the three experiments shown in Figure 3.7. The model spectra maintained slopes on their fine ends about the same as initial $m$’s while the coarse ends deposited. This pattern is consistent with field observations described earlier and with Equation 3.3. Portions of the clay fraction of the distribution can be seen in Figure 3.7 to deposit, even though the imposed shear stress exceeds this class’s threshold for deposition.
Figure 3.7. Numerical experiment results for $m = 1, 0, -1$ for deposition under a shear stress of 0.16 Pa
Distribution means correlated with their deposition rates. The distribution with $m = -1$ was the finest and settled the slowest. The distribution with $m = 1$ settled fastest, especially at its coarse end.

Statistical measures of the mean size, sorting (standard deviation), and skewness were calculated in phi units with time for the three experiments. The mean grain sizes became progressively finer and skewness became more negative in all three cases. The sorting for the case of $m = 1$ became larger, for the case of $m = 0$ remained about constant, and for the case of $m = -1$ became appreciably smaller. Trends similar to the latter were reported by Teeter (1993b) who found that sorting decreased slightly during laboratory settling tests on a well-graded sediment ($m$ about equal to 0). Trends similar to the former were reported by Stow and Bowen (1980) who found that clayey-silts with positive $m$ became more poorly sorted in the direction of down-slope transport on the Scotian continental margin.

CONCLUSIONS

Cohesive suspensions have grain spectra that vary according to the composition of particle aggregates with depositional conditions and settling. A limited number of coupled grain-size classes can be used to mimic size distribution changes during quiescent or stress-limited deposition. Coupling of numerical grain classes was based on a previous analytic description of grain-spectra change during settling. Settling experiments using flocculators and columns indicated that concentration and shear rate are not independent multiplicative factors. The concentration effect on settling was found to depend on fluid shear-rate, and a new functional relationship was proposed.
Chapter 4
Cohesive Clay-Silt Sediment Modeling Using Multiple Grain Classes: Application to Shallow-Water Resuspension and Deposition

ABSTRACT

Single and multiple grain-size class model formulations for the erosion and deposition of clay and silt sized particles are compared to each other for a series of laboratory experiments and field data from a shallow lake where resuspension occurs from wind-waves. The approach for coupling between grain-size classes during settling and deposition is presented in Chapter 3. Threshold shear-stresses for mutually-exclusive erosion and deposition are used in the model formulation to be consistent with previous laboratory investigations. Model deposition and erosion laws treat silt and clay fractions differently, yet couple them during certain modes of vertical transport. Models with up to seven grain classes are compared to laboratory flume tests which formed steady state suspension concentrations during deposition. Models with one and four grain classes are compared to field data. An automated, objective method is used to adjust coefficients for both model simulations. Comparisons indicate that the multiple grain-size class formulation improves erosion and deposition process descriptions.

INTRODUCTION

Resuspension of fine estuarine and lacustrine sediments by wind-waves has been studied using empirical methods (Pejrup 1986; Petticrew and Kalff 1991; Arfi et al. 1993; and Hamilton and Mitchell 1996), and numerical models (Luettich et al. 1990; Hawley and

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Lesht 1992; and Lick et al. 1994). Fine-grained sediment transport models are used in hydro-environmental studies of water quality and ecological conditions such as turbidity, dissolved oxygen, and contaminants, often involving relatively low concentrations of fine suspended material. Model process descriptions are not derived from first principles, and much empirical information is needed to successfully describe fine sediment transport at a given site. To qualify as a “good” model, sufficient physical processes must be represented to capture important observed behavior and non-observable system feedbacks that might become important after a physical change. The process descriptions must be conceptually justified. The purpose here is to evaluate two model formulations: single and multiple grain class representations.

Single grain-size sediment models which allow simultaneous erosion and deposition (Luettich et al. 1990; Hawley and Lesht 1992; and Lick et al. 1994) have been found previously to predict resuspension more accurately than single-grain models with mutually exclusive erosion and deposition, despite considerable laboratory experimental evidence supporting the latter assumption (Sanford and Halka 1993). Numerous laboratory studies have determined that the steady state suspensions, which can occur during either deposition or erosion, do not reflect a balance between simultaneous operation of these processes. Sanford and Halka (1993) list many references to experimental studies, and Lau and Krishnappan (1994) is an important recent study. One explanation for the failure of single grain models with mutually exclusive erosion and deposition is that natural sediments with silt and clay have multiple thresholds (Sanford and Halka 1993), while one reason for the success of the models that allow simultaneous erosion and deposition is that concentration-dependent depositional flux acts as a penalty to limit erosion, similar to decreases in sediment bed erodibility with erosion depth observed in laboratory experiments (Dixit
Grain-size coupling or amalgamation occurs due to cohesion, causing individual flocs or aggregates to be a mix of particle sizes (Kranck 1980; and Lau and Krishnappan 1994) and bed sediments to be poorly sorted. The paradox is that, even though these sediments occur as mixed-grain aggregates, they exhibit particle-size dependent transport behaviors. Spatial variation of dispersed grain-size distributions correlate to field transport processes (McLaren and Bowles 1985; and Teeter 1993b). During deposition, suspension particle-size spectra first decrease at their coarser ends, and modal sizes become progressively finer as deposition proceeds (Kranck 1980). Dispersed particle size of fine-grained suspensions increases with shear stress during erosion (Teeter et al. 1997). At a given bed shear stress level, the particles remaining in suspension are generally finer than those which deposit (Mehta and Lott 1987). Both settling velocities and critical shear stresses for deposition vary between clay and silt fractions (Teeter and Pankow 1989a,b). While a well-sorted cohesive suspension will steadily deposit in a flow below a critical shear stress (Krone 1962), a suspension of silts and clays will partially deposit to a steady state constant suspension concentration level (Partheniades et al. 1968). A similar paradox is that clay minerals segregate during transport in a manner that is similar to their settling rates based on dispersed particle size (Gibbs 1977). Thus, even though grain classes are coupled by cohesion, dispersed particle size affects transport properties, and size distribution imprints form clearly detectable patterns in estuarine and lake sediments.
METHODS

This section describes multiple grain-size class transport modeling at and near the sediment/water interface, the data sets used to compare model results, and the method used to adjust model coefficients. Emphasis will be on model features related to multiple grain-size class erosion laws. Deposition laws were described in Chapter 3. A numerical fine-grained sediment transport model was formulated with optionally one or more conservative grain-size classes to test the effectiveness or applicability of the multiple grain class model for resuspension and deposition modeling. For modeling purposes, the fine-grained sediment is assumed to be cohesive in nature, not to liquify under waves, and to follow separate and exclusive process descriptors for particle erosion and deposition.

Since fine-grained cohesive transport processes are not dependent on individual particle size intrinsically, size designations of grain-size classes are nominal. The finest class consists of the most cohesive clay and fine-silt-sized particles, and is referred to here as the cohesive fraction. Evidence suggests that this class behaves as a unit, and dominates several transport processes and the character of the sediments as a whole (Stevens 1991a and b; and Teeter and Pankow 1989b). Silts are divided into a variable number of classes, though evidence suggests that they act as a continuous distribution.

Model Description

The basic model is configured as a vertical suspension and a unit area of sediment bed at a point, similar to the models developed by Luettich et al. (1990), and Hawley and Lesht (1992). Horizontal gradients are ignored in favor of vertical ones. The focus of the model is the sediment/water interface where deposition and erosion algorithms compute flux boundary conditions for the suspension and the bed. Mass conservation equations are solved for each grain class in the suspension and in the bed. A bed model component
describes layers of sediment properties, performs bookkeeping with respect to sediment mass, and adjusts layer properties for consolidation and silt content.

For a well-mixed suspension at a point, model sediment mass conservation equations for each grain-size class are

\[
H \frac{dC(gs)}{dt} = E(gs) - F(gs)
\]  

(4.1)

where \( H \) is the water depth, \( C \) is depth-mean sediment concentration (mass per unit volume), \( t \) is time, \( E \) is the erosion flux, \( F \) is the depositional flux, and the index \( gs \) refers to the grain class. Dimensions of the erosion and deposition fluxes are mass per unit area per unit time. Deposition flux for multiple grain-classes was described in Chapter 3. For the point model, an analytical near-bed concentration \( C_b \) (Teeter 1986) was used in the calculation of \( F \) such that

\[
C_b = \left[ 1 + \frac{Pe}{1.25 + 4.75 P^{2.5}} \right] C
\]  

(4.2)

where a particle Peclet number \( Pe = HW_s/K_z \), \( W_s \) is a concentration-dependent settling velocity described in Chapter 3, vertical diffusivity \( K_z = 0.067U^3H \) (Fischer 1973), \( U^* \) is the friction velocity, and \( P \) is the depositional probability. \( Pe \) was constrained in the model such that \( Pe < 5 \).

The erosion flux depends first on the erosion threshold of the cohesive fraction, and then the erosion thresholds for silt fractions. The form of the cohesive-fraction erosion model is similar to the single-class erosion equation of Alishahi and Krone (1964), and is
where the cohesive fraction is designated $gs = 1$, the layer $bl$ exposed at the bed surface is designated $a$, $M$ is the erosion rate constant for the cohesive grain class and bed layer, $\tau$ is the bed shear stress, and $\tau_{ce}$ is the erosion threshold. With the exponent $n = 1$, Equation 4.3 is similar to the single-grain erosion equations of Kandia (1974) and Ariathurai et al. (1977). If $\tau < \tau_{ce}$ for the cohesive fraction, no sediments are eroded even if the bed shear-stress exceeds the critical threshold for some silts. The critical shear stress for erosion of the cohesive fraction is estimated by a power law depending on the concentration of the cohesive fraction in the bed layer exposed to the flow (Teeter 1987), and generally increases vertically downward in the bed. The erosion rate parameter $M$ is functionally related to the $\tau_{ce}$-value based on Lee and Mehta (1994). The forms of these auxiliary equations are not important to the results to be presented here.

For each silt-size class, upper and lower shear-stress threshold values are defined slightly differently than those used for the cohesive fraction. The lower or critical depositional shear-stress is defined as that value below which all material is free to deposit ($P = 1$). At shear stresses between the upper and lower critical values, a silt sediment fraction will erode or deposit ($0 < P < 1$) at rates linearly related to the bed and threshold shear-stresses, and depending on whether or not the cohesive fraction is eroding. Erosion is given precedence in the model such that, for a particular grain class, bed erosion precludes deposition.

The upper or critical erosion threshold shear-stress is defined as the value above which a silt class will remain in suspension. This definition recognizes that erosion of a silt fraction depends first on the condition that the critical shear stress for erosion of the cohesive
fraction has been exceeded. Silt-fractions erode in proportion to clay-fraction masses to maintain similarity in the shape of bed and suspended grain-size distributions as discussed in Chapter 3, and

\[
E_{(gs>1)} = E_{(gs=1)} \frac{S_{(gs>1,bl=a)}}{S_{(gs=1,bl=a)}}_{gs>1}, \quad E_{(gs=1)} > 0 \text{ and } \tau > \tau_{es_{(gs>1)}} \tag{4.4}
\]

where \(S_{(gs,bl)}\) is the grain-class sediment mass per unit area in a bed layer. Erosion thresholds for silt fractions are taken to be independent of their bed layer location.

Erodibility is also linked to the structure of the bed. A layered bed algorithm was developed with variable silt concentrations by layers, depending on initial conditions, and on erosional and depositional history. A fully-settled near-surface concentration distribution with respect to the cohesive fraction is assumed. After deposition occurs, hindered-settling rate is calculated by bed layer, and material is transported vertically downward in the bed using class-aggregated transport parameters, until the specified density distribution is achieved. The mass conservation equations for bed layer consolidation are

\[
\frac{dS_{(gs,bl)}}{dt} = - \frac{W_h(bl)S_{(gs,bl)}}{H_s(bl)} + \frac{W_h(bl-1)S_{(gs,bl-1)}}{H_s(bl-1)}, \quad H_s(bl) > H_{so}(bl) \tag{4.5}
\]

where \(H_s(bl)\) is the bed layer thickness, \(H_{so}(bl)\) is the specified fully-settled thickness, and the bed-layer hindered settling rate is

\[
W_h(bl) = W_{ho} \left[ 1 - b_1 \sum_{gs=1}^{ns} \frac{S_{(gs,bl)}}{H_s(bl)} \right]^{b_2}, \quad \sum_{gs=1}^{ns} \frac{S_{(gs,bl)}}{H_s(bl)} < \frac{1}{b_1} \tag{4.6}
\]

where \(ns\) is the number of grain-size classes, \(W_{ho}\) is a reference settling rate, and \(b_1\) and \(b_2\) are grain-class-mass averaged coefficients. Hindered settling is inhibited by deposition or
erosion greater than 0.01 g/m²/sec. In the bed, volumes of grain-size classes are taken into account when converting between mass and concentration. The sediment mixture is composed of sediment and water. Specifically, it is assumed that

\[
H_s(bl) = \sum_{gi=1}^{ni} \frac{S(gs,bl)}{\rho_s} + \frac{O_c S(gs=1,bl)}{\rho_l} + \sum_{gi=2}^{ni} \frac{O_s S(gs,bl)}{\rho_l} \tag{4.7}
\]

where \(O_c\) and \(O_s\) are the ratios of clay and silt masses to water masses associated with these fractions, and \(\rho_s\) and \(\rho_l\) are the particle and fluid densities. As mass is transported vertically downward as a result of consolidation, the layer concentration of the cohesive fraction is maintained constant over time, and the condition

\[
H_s(bl) = \frac{S(gs=1,bl)}{\rho_s} + \frac{O_c S(gs=1,bl)}{\rho_l} \tag{4.8}
\]

is imposed. Bed concentration (mass per unit volume) is \(S(gs,bl)/H_s(bl)\).

An example of the bed-layer model operation subsequent to a sudden deposition event is shown in Figure 4.1. The model is based on the sedimentation theory of Kynch (1952) and is a simplification of a number of complex processes. It is intended for calculating the settling and consolidation of thin layers of newly-deposited sediment over times of days. For thicker deposits especially, permeability becomes important as the upward velocity of water must equal downward sedimentation (Tan et al. 1990; and Pane and Schiffman 1997). At longer times and greater deposit thicknesses, inter-particle stresses develop, and self-weight consolidation occurs. Over an important range of times and concentrations, both sedimentation and self-weight consolidation probably occur (see Toorman and Berlamont 1991).
Figure 4.1. Settling of a 100-kg/m$^3$ bed layer deposited instantaneously just after hour 0.0 showing layer boundaries (also lines of constant concentration) and resulting bed profile.
Bed layers are numbered vertically downward. If a layer is withered away by erosion, it disappears at least temporarily. The erosion surface thus descends through the bed, as the surface layer thins, then step-wise through progressively deeper layers. The effects of erosion on bed mass are evaluated as

\[
\left. \frac{dS(gs, bl=a)}{dt} \right|_e = -E(gs)
\]  

(4.9)

where \( a \) is the exposed bed layer index. Deposition, on the other hand, always occurs into the first layer \((bl = 1)\), and the effect of deposition on bed mass is evaluated as

\[
\left. \frac{dS(gs, bl=1)}{dt} \right|_{id} = F(gs)
\]  

(4.10)

In this way, the bed structure is formed by consolidation from the top layer down. After appreciable deposition has occurred, the bed (in the absence of erosion or further deposition) will return to the specified fully-settled structure.

Coupling of grain classes during erosion and deposition is apparent in grain size measurements taken under these conditions, as previously described. The grain-coupling scheme used in this model assumes that fine-grained, cohesive sediments are deposited along with silts even though shear stresses are too high for them to deposit on their own. Serial coupling between grain-size classes is used to promote log-normal trends in the resulting size distributions, as discussed earlier. As a barrier to excessive winnowing of the cohesive fraction and sorting of grain-classes under moderate shear stress conditions, the model shields a small portion of the cohesive fraction, about equal in magnitude to that
proportion deposited with silts, from erosion until coarser grain classes are involved in erosion.

Bed-surface shear stresses generated by currents were calculated using a Manning's equation

\[ U_* = \frac{g^{1/2} n}{H^{1/6}} U \] \quad \text{and} \quad \tau = \rho_f U_*^2 \quad (4.11a \text{ and } 4.11b) \]

where \( g \) is the acceleration of gravity, \( n \) is Manning’s coefficient, and \( U \) is the depth-average velocity. Wave shear stresses were calculated using linear wave theory, and smooth, laminar wave friction formulations as described by Luettich et al. (1990). Wave and current shear stress components were added to estimate the total bed shear stress.

The point model used a fourth-order Runge-Kutta scheme to integrate conservation equations in time using time steps of 90 to 120 seconds for the applications described next.

**Model Comparison Data Sets**

Mehta and Partheniades (1975) performed annular-flume deposition experiments starting at high shear stresses. Initially-suspended fine-grained cohesive sediments deposited when shear stresses were reduced, forming constant, steady-state concentrations that depended on the initial suspension concentrations and the bed shear-stresses. Typical results for one series of experiments are shown in Figure 4.2. Each experiment had 1 g/l initial concentration of kaolinite. The kaolinite sediment material contained about 35 percent coarser than 2 \( \mu \)m, and a maximum particle size of about 45 \( \mu \)m. Similar results were obtained for coarser fine-grained sediments from San Francisco Bay, and Maracaibo Bay, Venezuela. The fractional amount remaining in suspension \( C_f/C_o \) for the kaolinite experiments (Mehta and Partheniades 1975) are summarized in Figure 4.3a. The degree of
Figure 4.2. Relative suspension concentration during kaolinite deposition experiments (Mehta and Partheniades 1975). Initial shear stress was 1.05 Pa for all tests.
Figure 4.3. (a) Observed $C_f/C_o$ corresponding to Figure 4.1, (b) example model result with 1 grain class, (c) example model result for 4 grain classes
deposition \((1 - \frac{C_f}{C_o})\) was found not to depend on initial concentration. This result, plus other experiments on kaolinite suspensions by Partheniades et al. (1968) and Lau and Krishnappan (1994), confirms that these steady-state concentrations were not caused by a balance between erosion and deposition.

The cohesive clay-silt deposition results shown in Figure 4.2 follow Krone's deposition law for bed shear-stresses less than 0.16 Pa, when all sediment eventually deposited. At higher bed shear stresses, however, they do not follow Krone's deposition law as only a certain fraction of material, depending on shear-stress, deposited. The times required for deposition to occur and for suspensions to reach steady-state were not greatly affected by the bed shear stress, as can be seen in Figure 4.2. Material either deposited or remained in suspension, with the transition time consistent with typical settling velocities.

Numerical deposition experiments were performed much the same way as the original experiments. Total suspended material concentrations were initialized at 1 g/l, and shear stresses at 1.05 Pa. A series of 25 model simulations were performed in which shear stresses were reduced to allow deposition.

Wind-wave resuspension in a large shallow lake was studied by Luettich et al. (1990). Data were collected for 15 days in August 1985 in the 600-sq-km, 3.2-m-deep Lake Balaton, Hungary. At the time of the field experiment, the lake had Secchi depths of less than 0.2 m, was eutrophic, and had a fine-grained sediment bed. A tripod station was established in 2 m water depth to carry BASS velocity meters at two depths and a wind instrument 2 m above the water surface. Water samples were collected at mid-depth from an anchored boat. Wave information was extracted from the 2-Hz velocity data.

Two storms occurred during the study but wave information was only collected during the second storm. Wind speeds reached 7 to 9 m/sec. For the present study, current speed,
wave height, and wave period data were digitized and interpolated to 0.1 hours for use as boundary conditions for the model. The model simulations covered 25.5 hours with time steps of 2 minutes. Total suspended material (TSM) data were digitized and interpolated for model comparison at 0.5 hour intervals to make the data coverage uniform in time.

**Coefficient Adjustment for Model Comparison**

The comparison of models for simulating lake resuspension was complicated by the requirement to adjust model coefficients, since optimum model coefficient values could vary between model formulations. An automated model coefficient adjustment method was developed to expedite model adjustment and to systematize this adjustment so that alternate model formulations could be tested objectively.

The adjustment process began with model coefficients and parameters manually set to rough model-to-prototype TSM agreement using threshold values interpreted from field data, or typical values found for other systems. Important coefficients were identified, and adjustment for modeling lake resuspension involved 18 select coefficients for the single grain class model and 27 corresponding coefficients for the multiple grain class model. Each model coefficient in this set was varied by a range of factors (typically 0.6 to 1.4) in 10 model simulations, holding all other parameters constant. The variance in the difference between the model results and the field TSM data was used as a criterion of the goodness of the model simulation. This criterion allowed a constant model-to-prototype offset to occur in case that a washload or background concentration existed. A background concentration was used in the previous modeling of this system (Luettich et al. 1990). If variation of a given coefficient produced a range of variances such that the ratio of the maximum variance to the minimum exceeded a threshold (typically 1.1 to 1.2), then the optimum value of the coefficient was determined for this particular set of other coefficients. After all unknown
coefficients were tested in this manner, a new set of model coefficients was developed by applying a weighting to the difference between the original and optimum values and adding those weighted differences to the old coefficient values. Weighting values were 0.33 to 0.5. This procedure was repeated 20 times for each model.

RESULTS AND DISCUSSION

Plots of example numerical $C_f/C_o$ curves are shown in Figure 4.3b-c for a single grain model ($ns = 1$) and a multiple grain model ($ns = 4$), respectively. Multiple grain classes allow representation of the $C_f/C_o$ curves in a step-wise fashion. The more classes, the better the representation. At times shorter than that required to reach steady-state, multiple grain class model results have smoother transitions between different shear stress levels. Model results with $ns = 7$ after 0.1 and 1 hours for various shear-stress levels are shown in Figures 4.4a and 4.4b. These results indicate that, as suspensions approach steady-state values, $C/C_o$ curves take on the stair-step nature as shown in Figure 4.3, with the finer, slower-settling grain-size classes requiring more time to come to steady-state. The $C/C_o$ curve in Figure 4.4b can be compared to the observed $C_f/C_o$ curve shown in Figure 4.3a. Similarly good results were obtained by a multiple grain-size class analytic deposition model used to simulate the same experimental data set (Mehta and Lott 1987). In the present model, grain class coupling is expected to improve the ability to simulate sediment sorting.

Resuspension in Lake Balaton was modeled with one and four grain classes. Wave and current boundary data are shown in Figure 4.5. Table 4.1 summarized the model to interpolated-prototype TSM differences after 10 and 20 iterations of coefficient adjustment.

Coefficients converged quickly during the first few iterations of the adjustment process to give results similar to those above at 10 iterations, and improved only slowly thereafter. During coefficient adjustment, the multiple-grain model converged more uniformly, while
Figure 4.4. Model relative TSM concentration curves for $ns = 7$ for (a) 0.1 hour, and (b) 1 hour of deposition.
Figure 4.5. Model boundary conditions for wave height, wave period, and current speed. Data interpolated from Luettich et al. (1990)
the single grain-size model oscillated to a greater extent about the minimum variance. The coefficient adjustment process resulted after 20 iterations in the model coefficients presented in Table 4.2. Model coefficients $a_{l}$, $n(gs)$, $C_{ll}$, and $C_{ul}$ were defined in Chapter 3. As indicated in Table 4.2, model coefficients were similar for the two models. While eight bed layers were available in each model, only the top three layers were active in the simulations. Bed layer thicknesses were only 2 to 3 mm for both models indicating that erodibility decreased rapidly with depth in the bed, as has been observed in the laboratory (Dixit 1982).

<table>
<thead>
<tr>
<th>Table 4.1</th>
<th>Summary of Prototype to Model TSM Differences</th>
</tr>
</thead>
<tbody>
<tr>
<td>Iterations</td>
<td>$ns$</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>10</td>
<td>4</td>
</tr>
<tr>
<td>20</td>
<td>1</td>
</tr>
<tr>
<td>20</td>
<td>4</td>
</tr>
</tbody>
</table>

Plots of model predictions using final coefficients for $ns = 1$ and $ns = 4$ are shown in Figure 4.6 along with observed TSM. The single grain model had difficulty with the depositional phases of the time series, especially after hour 54 when shear stresses were higher than the critical threshold for deposition. If the prototype data set had extended through the end of the storm, the single grain model comparisons probably would have been poorer than those presented. The resuspension phase of the data was also better represented in the model with $ns = 4$.

The results of the multiple grain model with $ns = 4$ appear to be better than the single grain model results reported by Luettich et al. (1990). His model adjustment resulted in a root-mean-square error of 8.7 mg/l over the last 10.2 hours of the data set used here.
Table 4.2
Final Model Coefficients

<table>
<thead>
<tr>
<th>Coefficient</th>
<th>Model with $ns = 1$</th>
<th>Model with $ns = 4$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Initial $C(gs)$, mg/l</td>
<td>29.8</td>
<td>28.2, 3.5, 1.0, 0.5</td>
</tr>
<tr>
<td>Manning’s $n$</td>
<td>0.024</td>
<td>0.022</td>
</tr>
<tr>
<td>$a_1$, m/sec</td>
<td>0.00189</td>
<td>0.00175</td>
</tr>
<tr>
<td>$n(gs)$</td>
<td>1.0896</td>
<td>1.138, 0.885, 0.7, 0.5</td>
</tr>
<tr>
<td>$C_{ll}$, mg/l</td>
<td>24</td>
<td>68</td>
</tr>
<tr>
<td>$C_{ul}$, mg/l</td>
<td>923</td>
<td>875</td>
</tr>
<tr>
<td>$\tau_{cd}$, Pa</td>
<td>0.025</td>
<td>0.0375</td>
</tr>
<tr>
<td>$\tau_{ce}(gs=1, bl=1)$, Pa</td>
<td>0.0381</td>
<td>0.0375</td>
</tr>
<tr>
<td>$\tau_{ce}(gs=1, bl=2)$, Pa</td>
<td>0.0915</td>
<td>0.0873</td>
</tr>
<tr>
<td>$\tau_{ce}(gs=1, bl=3)$, Pa</td>
<td>0.104</td>
<td>0.1564</td>
</tr>
<tr>
<td>$M(gs=1, bl=1)$, g/m²/sec</td>
<td>0.1421</td>
<td>0.1278</td>
</tr>
<tr>
<td>$M(gs=1, bl=2)$, g/m²/sec</td>
<td>0.1055</td>
<td>0.0939</td>
</tr>
<tr>
<td>$M(gs=1, bl=3)$, g/m²/sec</td>
<td>0.0984</td>
<td>0.0631</td>
</tr>
<tr>
<td>$\tau_{ce}(gs&gt;1)$, Pa</td>
<td>0.121, 0.226, 0.268</td>
<td></td>
</tr>
<tr>
<td>$S(1 \leq gs \leq ns, bl=1)$, kg/m²</td>
<td>0.0697</td>
<td>0.1444</td>
</tr>
<tr>
<td>$S(1 \leq gs \leq ns, bl=2)$, kg/m²</td>
<td>0.0871</td>
<td>0.1818</td>
</tr>
<tr>
<td>$S(1 \leq gs \leq ns, bl=3)$, kg/m²</td>
<td>0.0879</td>
<td>0.1891</td>
</tr>
</tbody>
</table>
Figure 4.6. Model results compared to field TSM data for a single grain class (top), and for 4 grain classes. Data from Luettich et al. (1990)
CONCLUSIONS

The multiple grain-size class formulation was demonstrated to improve model prediction of resuspension and deposition over a single grain-size formulation. Some of the deficiencies of the single-grain formulation can be overcome by assuming simultaneous erosion and deposition. However, as Lau and Krishnappan (1994) conclude, fine-grained sediment models that assume simultaneous erosion and deposition may not have the correct physical representation. The conclusion of this study is that the multiple grain class formulation can accurately simulate suspension concentrations during resuspension and deposition, although nothing new can be added to the debate concerning simultaneous erosion and deposition. The multiple grain class formulation appears to be a good alternative to a single-grain formulation that assumes simultaneous erosion and deposition. The added physical realism of the multiple grain class formulation allows sediment texture to adjust to hydraulic conditions, but comes at the price of increased computational burden for operating the model and increased numbers of sediment parameters requiring estimation. The suitability of such a formulation is probably site or case specific, depending on the study objectives, availability of laboratory and field data. For sites with high shear stresses and/or significant quantities of silt and for which system changes in sedimentation patterns are to be evaluated, a multiple-grain model formulation may be warranted.
Chapter 5
Wind-Waves and Atmospheric Shear Stress in Ultra-Shallow (< 2 m)
Laguna Madre, Texas

ABSTRACT

Wave and current data from shallow, open-water areas in Laguna Madre, Texas, were
analyzed as a first step in a study of shear stresses. Measured pressures and velocity
components were used to estimate significant wave height, spectral-peak wave period, depth,
mean current speed, and wave orbital flow components at 30 minute intervals. Using
previous scalings, dimensionless wave energy and depth were found to be related differently
in water depths less than 2 m (including areas with submersed aquatic vegetation) than in
deeper, bare areas in this and previous studies. Waves in ultra-shallow water were more
consistently depth limited, and dimensionless energies and periods were less than expected
for the same dimensionless depth. Furthermore, wave characteristics scaled by wind speed
were not consistent among stations of different depths.

A rescaling of wave energy and period by atmospheric friction velocity was performed.
Direct measurements of atmospheric shear stress were not made as part of this study, but
previous relationships for the atmospheric roughness height and friction factor $C_d$ were
evaluated according to the wind and wave data. Dimensionless wave parameters were better
related, and resulting wave hindcasts had smaller errors, using the new wave scaling by
atmospheric friction velocity. Rescaled dimensionless energies and periods allowed the
development of new expressions for wave height and period that were based on water depth
and $C_d$. 
A new expression for $C_d$ in terms of water depth and wind speed is proposed. In vegetated areas, an effective water depth from the water surface to the seagrass canopy height was used. The fraction of atmospheric shear stress going into waves, as opposed to that going into currents, was found to decrease as the inverse of the square-root of wind speed for wind speeds greater than 5 m/sec.

**INTRODUCTION**

Wind and wave relations for the shallow waters of Laguna Madre were investigated for the purpose of evaluating modeling approaches for wind-wave shear stresses. Wind-waves are primarily responsible for sediment resuspension in Laguna Madre and in many other shallow, wind-exposed estuaries and lagoons. Ecosystem evaluations are being performed for many such systems, and suspended sediment particles have an important role in light penetration. As part of the development of a sediment resuspension model for the lagoon by the U. S. Army Engineer Research and Development Center (ERDC, Teeter et al. 2002), wind and wave measurements were used to characterize the major forcing for resuspension: locally-generated wind waves. Resuspension in open-water areas, rather than shoreline sediment transport, was the primary concern.

Laguna Madre is 183-km long, consists of two shallow tidal bays transected and connected by the Gulf Intracoastal Waterway, and extends from Port Isabel, near the U.S.-Mexican border, to Corpus Christi Bay, Texas. Average depth of Laguna Madre is about 1 m and total surface area is about 1,500 km². A schematic of the system is shown in Figure 5.1. Laguna Madre contains over 85 percent of Texas’ seagrass habitat (Brown and Kraus 1997), which supports diverse ecosystems. There are concerns that resuspension of dredged material placed in open-water might be reducing light penetration and limiting seagrass abundance.
Figure 5.1. The Laguna Madre study area, station locations, and depth contours (coordinates are state plane, NAD27 Texas South, in meters)
(Onuf 1994). To address these questions and to model sediment resuspension, scientists need accurate predictions of wind-wave shear stresses.

The presence of submersed vegetation greatly increased the difficulty involved in the prediction of wind-wave resuspension in Laguna Madre since seagrasses are known to increase total flow resistance, damp waves, and at the same time shelter the sediment bed from shear stress and reduce resuspension (Ward et al. 1984; Fonseca and Fisher 1986; Fonseca and Kenworthy 1987; Fonseca and Calahan 1992; and Madsen et al. 2001). However, data on these effects are sparse.

In this study, measurements were used to estimate wave characteristics and check analytic wave models. Predicted wave characteristics based on previous analytic models were found to differ from observed waves in the case of ultra-shallow water (less than 2 m). Therefore, a rescaling of wave energies and periods based on atmospheric friction velocity was performed. Atmospheric shear stress $\tau_a$ was calculated with previous formulations that included the effects of wave characteristics on roughness height. Wave shear stresses were calculated with the use of various friction formulations and then compared to estimated atmospheric shear stresses. A selected wave-friction formulation was used along with $\tau_a$ to estimate the fraction of $\tau_a$ going to waves as opposed to going to currents.

**BACKGROUND**

**Atmospheric Shear Stress**

Atmospheric shear stress ($\tau_a$, Pa) was calculated on the basis of wind speed $U_a$, in meters per second, at 10-m height:

$$\tau_a = \rho_a C_d U_a^2$$  \hspace{1cm} (5.1)

where $\rho_a$ is the air density (about 1.225 kg/m$^3$), and $C_d$ is the atmospheric friction factor appropriate for wind referenced to 10-m height (CERC 1984).
atmospheric shear stress comes from wave roughness at various scales, and \( C_d \) generally increases with increased wind speed, at least up to some high wind speed. The main transfer of momentum from the atmosphere to waves occurs at relatively short wave lengths of about 0.3-m (range 0.06 to 1 m) wavelength (Gemmrich et al. 1994) but transfer to slightly longer wavelengths is also appreciable (Donelan 1990; and Lionello et al. 1998). Short waves are advected by the long-wave orbitals, reducing wind speed relative to short waves at long-wave crests and diminishing the importance of short-wave roughness to atmospheric drag. The wind field is modified by dominant wavelengths (Lionello et al. 1998). Significant wave height \( H_s \) is the most often used, physically important, length scale used to estimate \( C_d \).

When waves are fetch- and/or duration-limited, however, the stage of wave development affects \( C_d \). For a constant wind speed, \( C_d \) decreases as waves become higher, longer, and less steep.

With an assumed logarithmic velocity profile and neutral atmospheric stability, the atmospheric friction factor is dependent on surface roughness

\[
C_d = \left( \frac{\kappa}{\ln(10/z_o)} \right)^2
\]

(5.2)

where \( \kappa \) is the von Karman constant (0.4), and \( z_o \) is the surface roughness coefficient in meters. The latter is much smaller than \( H_s \). At wind speeds greater than about 2.5 m/sec, those important in this study, air flow becomes aerodynamically rough and \( z_o \) is approximately a quadratic function of wind speed (Donelan 1990). For the turbulent-rough regime, Hsu (1974) related \( z_o \) to both wave steepness (significant wave height \( H_s \) over wave length \( L_w \) ) and wave age (wave celerity \( C \) over atmospheric friction velocity \( U_* \) ) starting with
and then substituting a deep-water relationship for $L_w$ to obtain

$$z_o = \frac{H_s}{2 \pi C^2} U_{*s}^2$$  \hspace{1cm} (5.4)

Hsu originally compared this latter formulation to a number of data sets, and recent comparisons have also found it to be reliable (Donelan 1990).

Various expressions have been developed for $C_d$. For fully-developed oceanic wave conditions, Hsu (1988) developed the following expression for $C_d$ from Equation 5.2 by setting the ratio of $C$ to $U_*^2$ equal to 29 and substituting an analytical expression for $H_s$ into Equation 5.4:

$$C_d = \left( \frac{0.4}{14.56 - 2 \ln U_s} \right)^2$$  \hspace{1cm} (5.5)

Various linear expressions have been proposed that relate $C_d$ to $U_s$. For example, for oceanic conditions and neutral atmospheric stability, Wu (1980) proposed

$$C_d = (8.0 + 0.65 U_s) \times 10^{-4}$$ , while Atakturk and Katsaros (1999) found

$$C_d = (8.7 + 0.78 U_s) \times 10^{-4}$$ for Lake Washington, Washington.

The roughness height has also been related to Charnock's parameter $\alpha_c$ to include the effect of wave development

$$z_o = \alpha_c U_{*o}^2 / g$$  \hspace{1cm} (5.6)

Reported field values for $\alpha_c$ generally range from 0.012 to 0.035 for "old" and "young" waves respectively (Wu 1980; Hsu 1988; and Lionello et al. 1998). During the initial stage of wave
development, roughness heights are much greater. Wu recommended using $\alpha_c = 0.0185$ in
Equation 5.6 and proposed an additional term based on dimensional arguments

$$z_o = \frac{\alpha_c U_{*a}^2}{g} \left( \frac{\mu U_{*a}}{\gamma} \right)^{\beta-2}$$  \hspace{1cm} (5.7)$$

where $\mu$ is the dynamic viscosity of water, and $\gamma$ is the surface tension. Wu suggests that the value of the exponent $2 < \beta < 2.5$ correctly defines the dependence of $z_o$ on $U_{*a}$.

Janssen (1989) developed the following relationship for wave roughness:

$$z_o = \frac{\alpha_{cr} U_{*a}^2}{g(1 - \tau_{aw}/\tau_{aw})^{1/2}}$$  \hspace{1cm} (5.8)$$

where $\tau_{aw}$ is the atmospheric shear-stress going into the waves, and $\alpha_{cr}$ is a reference or reduced Charnock’s parameter ($=0.01$). Lionello et al. (1998) used Equation 5.8 to test two-way coupling for atmospheric and ocean-wave models.

**Shear Stress Budget**

Reported values for the fraction of momentum transferred from the atmosphere to waves vary widely. Lionello et al. (1998) indicate that $\tau_a > \tau_{aw} > 0.15 \tau_a$. As with surface roughness, the stage of wave development affects the fraction of momentum transferred from the atmosphere to waves. “Young”, steep waves absorb a greater fraction of atmospheric shear stress as waves develop. Equations 5.6 and 5.8 suggest that

$$\frac{\tau_{aw}}{\tau_a} = 1 - (\alpha_{cr}/\alpha_c)^2$$  \hspace{1cm} (5.9)$$
which implies that the shear-stress fraction transferred to waves is related to wave age, with about 95 percent of $\tau_a$ transferred to $\tau_{aw}$ during initial wave development and about 40 percent for old waves. Apparently, wave dissipation mechanisms more effectively shunt momentum into currents for old waves.

In a fully-developed wave field, when temporal and spatial variations of wave spectra are minimal, shear-stress input from the atmosphere is about equal to wave dissipation. Wave dissipation comes from various losses: friction, wave-wave interactions, white capping, and wave breaking. Wave breaking occurs in deep water when the wave steepness reaches or exceeds 0.14, and in shoaling water when the wave height exceeds about 80 percent of the depth (CERC 1984). White-capping occurs when wind separates at and de-stabilizes wave crests and especially when $C/U* < 1$ (Wu 1980). The resulting loss of wave energy is converted into mean-flow momentum and to turbulent mixing. Wave dissipation is not well understood in general, but it is recognized as important to momentum transfer from the atmosphere to the water column (Lionello et al. 1998). No consensus exists among researchers about the relative magnitudes of dissipation mechanisms, and more research is probably needed before a consensus can be reached.

For open-ocean, deep-water conditions, most of the atmospheric input eventually goes to the upper part of the water column (Richman and Garrett 1977). In shallow-water, however, wave shear-stresses transmitted to the bed can be of the same order as the atmospheric shear stresses. For example, Sanford (1994) measured wave conditions during a January-1990 wind-wave resuspension event at a 3.4-m deep tripod station near Poole's Island in Upper Chesapeake Bay and estimated the wave shear stress to be 0.6 Pa at the bed. Winds were offshore at 11 m/sec and $\tau_a$ was apparently about 0.25 Pa. If in fact wave shear stress is of the
same order as \( \tau_s \), less of the total \( \tau_s \) input is transferred to currents in shallow water than in deeper water.

**Depth-Limited Waves for Unvegetated Bottoms**

Analytical models for waves in shallow water are based on dimensionless parameters used to collapse data to power-law relations. For depth-limited, but otherwise fully-developed, waves, models are of the form

\[
E^* = a_1 h^{a_2} \quad (5.10)
\]
\[
f^* = a_3 h^{a_4} \quad (5.11)
\]

where \( E^* = g^2 E/U_a^4 \) is dimensionless wave energy, \( h^* = gh/U_a^2 \) is dimensionless depth, \( f^* = U_a/gT_p \) is dimensionless wave frequency, \( E = \sigma^2 \) is the variance of the wave height field, and \( E = H_s^2/16 \), \( U_a \) is the wind speed adjusted to 10 m height in meters per second, and \( T_p \) is the spectral-peak wave period in seconds. CERC (1984) found the coefficients to be \( a_1 = 1.4 \times 10^{-3} \), \( a_2 = 1.5 \), \( a_3 = 0.16 \), and \( a_4 = -0.375 \). Young and Verhagen (1996) found coefficients for Equations 5.10 and 5.11 to be \( a_1 = 1.06 \times 10^{-3} \), \( a_2 = 1.3 \), \( a_3 = 0.20 \), and \( a_4 = -0.375 \).

Formulations that include the effects of fetch length are slightly more complicated. The depth-limited and deep-water cases form asymptotic limits which include the dimensionless fetch length \( X^* = gx/U_a^2 \) where \( x \) is the fetch length. For example, Young and Verhagen (1996) found

\[
E^* = 3.64 \times 10^{-3} \left( \tanh A_1 \tanh \left( \frac{B_1}{\tanh A_1} \right) \right)^{1.74} \quad (5.12)
\]

where

\[
A_1 = 0.493 h^{0.75}, \quad B_1 = 3.13 \times 10^{-3} X^{0.57} \quad (5.13)
\]
Wave Friction Formulations

The wave shear stress at the bed $\tau_{wb}$ is calculated as follows:

$$\tau_{wb} = \frac{1}{2} f_w \rho U_{wbm}^2$$

(5.14)

Some friction formulation must be assumed for $f_w$. For example, Luettich et al. (1990), Bailey and Hamilton (1997), Hamilton and Mitchell (1996), and Hawley and Lesht (1992) used the laminar wave friction formulation where

$$f_w = 2 \left( \frac{U_{wbm} A_{bm}}{v} \right)^{-0.5} \quad \frac{U_{wbm} A_{bm}}{v} \leq 10^4$$

(5.15)

and $\tau_{wb}$ is the instantaneous maximum wave shear stress at the bed, $f_w$ is the wave friction factor, $U_{wbm}$ is the maximum wave orbital speed just above the bed,

$$U_{wbm} = \frac{\pi H_s}{T_p \sinh (2 \pi h/L_p)}$$

(5.16)

$A_{bm}$ is the maximum wave excursion amplitude at the bed,

$$A_{bm} = 2 \pi U_{wbm}/T_p$$

(5.17)

and $v$ is the kinematic viscosity of water. The term in parentheses on the right side of Equation 5.15 is the wave Reynolds number just above the bed. For unvegetated beds, $U_{wbm}$ and $A_{bm}$ are computed at the bed and become appreciable when the wave length is about twice the water depth. In seagrass beds, however, the wave begins to "feel" friction when the wave length is twice the distance from the water surface to the top of the plant canopy since the canopy-top presents appreciable friction to flow. With respect to waves, seagrass beds are shallower (by about the height of the canopy) than the water column depth.
Laminar wave-friction factors are greater than turbulent, smooth, or rough friction factors used in other situations (Kamphuis 1975). Even though wave Reynolds numbers for waves may be low enough to meet viscous-dominated criteria (about $10^4$) developed in laboratories, field conditions most often include a turbulent water column and finite currents. Langmuir circulation cells having appreciable vertical circulations are common in the lagoon. Turbulence intensities are higher than normal in seagrass beds due to the shedding of vorticies by plant components, and a turbulent criterion has not been developed for these flows.

Kamphuis (1975) developed a turbulent-rough formulation

$$f_w = 0.4 \left( \frac{k_n}{A_{bm}} \right)^{0.73}, \quad \frac{A_{bm}}{k_n} < 50$$

(5.18)

where $k_n$ is a roughness height normally about twice the bed-grain diameter at the 90th percentile for plane beds. A similar rough formulation was used by Christoffersen and Jonsson (1985) in their model of combined wave and current friction.

**MEASUREMENTS AND DATA PROCESSING**

**Waves and Currents**

Measurements of waves and currents were specifically carried out for this study. Pressure and velocity data were collected by Texas A&M University, Corpus Christi, Conrad Blucher Institute (CBI) and transmitted to ERDC as part of the interagency study of possible dredged-material dispersal impacts. Pressures were measured with Keller Semi-Conductor ® model PA10 strain gauges. Velocities were measured with Marsh-McBirney ® electromagnetic current meters. Data were logged on Applied Microsystems ® Smart Packs.
Measurements were taken at six sites located in Upper (U) and Lower (L) Laguna Madre. Approximate locations for stations are shown in Figure 5.1 with the exception of station U1w which is actually located in southern Corpus Christi Bay. The fetch lengths to the 0.5-m depth contour for the dominant wind direction were determined for select locations and are presented in Table 5.1.

<table>
<thead>
<tr>
<th>Station</th>
<th>Julian Days of Data</th>
<th>Latitude N</th>
<th>Longitude W</th>
<th>Fetch Length to 130°, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1w</td>
<td>14-50</td>
<td>26° 10.7500'</td>
<td>97° 15.6000'</td>
<td>10.2</td>
</tr>
<tr>
<td>L2w</td>
<td>14-33</td>
<td>26° 08.0833'</td>
<td>97° 12.4666'</td>
<td>5.1</td>
</tr>
<tr>
<td>L3w</td>
<td>15-49</td>
<td>26° 35.4200'</td>
<td>97° 22.9600'</td>
<td>4.6</td>
</tr>
<tr>
<td>U1w</td>
<td>10-43</td>
<td>27° 41.4580'</td>
<td>97° 13.2960'</td>
<td>3</td>
</tr>
<tr>
<td>U2w</td>
<td>23-32</td>
<td>27° 17.2333'</td>
<td>97° 24.7500'</td>
<td>3.7</td>
</tr>
<tr>
<td>U3w</td>
<td>14-38</td>
<td>27° 11.5500'</td>
<td>97° 25.7000'</td>
<td>3</td>
</tr>
</tbody>
</table>

Julian days refer to the days of the year in 1998. Station L1w was located on an azimuth of 310° from L2w, down-wind along the dominant wind direction (130°). Station L2w was located in a *Thalassia testudinum* seagrass area. Station L1w had a bare bottom, and stations L3w and U3w were near the edges of *Syringodium filiforme* and *Halodule wrightii* seagrass beds.

Gauges recorded 5-Hz bursts of data every 30 min. Data were in blocks of about 6 min, each block containing about 3,000 readings of pressure, and $u$- and $v$-velocity components.

Sub-surface measurements were used to estimate significant wave height, spectral-peak wave period, depth, mean-current speed, and wave-orbital flow components at 30-min intervals. The mean pressure was converted to the water depth over the gauge. Pressure fluctuations were adjusted from the measurement depth to the water surface by a calculation
that involved wave length, total depth and depth of submergence of the gauge (Dean and Dalrymple 1991). Significant wave height \( H_s \) was calculated as

\[
H_s = 4\sigma
\]  
(5.19)

where \( \sigma \) is the standard deviation of the adjusted pressure (db) measurements.

Spectra of pressure and velocity components were computed with Splus ® statistical software for each sampling burst. Six-minute bursts of about 3,000 points were detrended, demeaned, tapered (10 percent cosine), and converted to spectral density in the frequency domain. Spectra were smoothed to reduce variability, and the spectrum peaks that occurred between 0.5- to 8-sec period were identified. The magnitudes of the velocity component peaks were determined by first fitting a regression line through the spectral region outside the peak band from 0.66 to 1.5 of the peak frequency. The magnitude was then determined as the difference between the regression line and the spectral line at the wave peak frequency.

Example smoothed spectra for pressure, \( u \)- and \( v \)- velocity components, and an example velocity-component spectrum and locally-weighted regression line are shown in Figure 5.2.

Wave characteristics estimated from bursts were smoothed with a 3-hr low-pass filter and decimated to 1 hr intervals. A comparison of the distributions for raw- and smoothed-wave heights for L1w is shown in Figure 5.3.

Wave length estimation was based on iteration of the dispersion equation:

\[
L_w = \frac{g T_p^2}{2\pi} \tanh \left(2\pi \frac{h}{L_w}\right)
\]  
(5.20)

where an iteration convergence criteria of \( 2 \times 10^{-5} \) m was used, and wave celerity was calculated as \( C = L_w/T_p \).
Figure 5.2. Example measurement spectra (a) for current component and pressure signals, and (b) with regression line fit to velocity component to determine peak height.
Figure 5.3. The statistical distributions of raw and 3-hr low-pass smoothed wave heights.
Figure 5.4. Comparison of measured maximum near-bed wave currents with those calculated from wave characteristics and linear wave theory.
As an indication of the quality and/or consistency of the wave and current data, the root-sum-squares of the maximum wave-induced current components were plotted against the calculated peak currents. Linear wave theory was applied to calculate peak currents from the wave data. Results are shown in Figure 5.4 for L1w. The residual standard error is 1.4 cm/sec and the $R^2 = 0.95$ for this comparison. One source of discrepancy at low values might have come from a zero-flow offset in the electro-magnetic current meter data. Such an offset would be consistent with a comparison made to previous Doppler current meter data taken at the same site to be described later.

**Wind Conditions**

Wind data from South Padre Island, Rincon, and Bird Island stations maintained by CBI were downloaded from their web site (<http://tcoon.cbi.tamcc.edu/data>) and used in the analysis of wave data. These stations span most of the length of Laguna Madre, as shown in Figure 5.1, but were many miles from the wave gauges. For this period, $u$- and $v$- component wind data from the three stations were combined, and a locally-weighted regression was used to perform 3-hour low-pass filtering of the data. To make the comparison of wind and wave data meaningful, wind data were scrutinized to ensure that they were representative of the wave measurement sites. In addition to averaging and filtering, time periods were identified when wind conditions were (a) relatively uniform over the area (identified by plotting and comparing the winds from the three stations) and (b) from the dominant southeasterly direction (identified by averaging wind direction from the three stations and taking wind directions from 110 to 145°). The concern was that data taken during frontal passage events, or when the wind record might be corrupted by some other factor, should not be used. Statistical distributions of raw and 3-hr low-pass filtered wind speeds are shown in Figure 5.5
Figure 5.5. The statistical distributions of raw wind speeds from three stations and 3-hr low-pass smoothed wind speed.
for Julian day 13 through 33. A histogram of wind directions for this period is shown in Figure 5.6.

**Observational Results**

A period from about 1998 Julian day 13 through 50 was found to have the longest continuous data coverage, and all available data were compiled for this period.

**Wave conditions.** The Table 5.2 summarizes wave characteristics, wave steepness \( (H_s/L_w) \), and relative depth \( (h/L_w) \). Histograms of \( H_s \) and \( T_p \) are shown in Figures 5.7 and 5.8 for the six stations. For plotting purposes, wave heights and periods for U1w were cut off at 0.5 m and 3.8 sec, even though a very small number of data reached 0.65 m and 6.7 sec.

**Current conditions.** Current speed \( U \) values were relatively low (median values of 7 cm/sec or less). Current conditions are summarized for \( U \) and for the N-S and E-W components (\( u \) and \( v \)) in Table 5.3. Current magnitude \( U \) statistics were compiled from smoothed-current data, while \( u \)- and \( v \)-component statistics are based on raw data.

Current data were previously collected near L1w in 1994-1995 by CBI, using an acoustic-Doppler velocimeter (Brown and Kraus 1997). The present measurement results indicate a residual flow to the ESE at L1w that was not shown in the previous CBI current data. The axis of the scattered data has a similar direction, but is offset to the ESE by perhaps 5 cm/sec.

**RESULTS AND DISCUSSION**

**Atmospheric Friction factors**

The expressions of Hsu (1974) for atmospheric roughness height, \( z_o \), presented in Equations 5.3 and 5.4 are not equivalent for shallow water when \( h/L_w \) values are less than 0.5 and, therefore, \( \tanh (2\pi h/L_w) \) values are less than 1.0 as in deep water. Based on Hsu's development, it might be assumed that Equation 5.3 is more suitable for shallow water than Equation 5.4 if wave steepness is the primary parameter affecting \( z_o \).
Figure 5.6. Histogram of wind direction for the study period
Figure 5.7. Histograms of significant wave heights for the six measurement stations.
Figure 5.8. Histograms of peak wave periods for the six measurement stations.
Table 5.2
Summary of Wave Gauge Observations

<table>
<thead>
<tr>
<th>Station</th>
<th>L1w (m) Mean (Std. Dev.)</th>
<th>L2w</th>
<th>L3w</th>
<th>U1w (m) Mean (Std. Dev.)</th>
<th>U2w</th>
<th>U3w</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2.07 (0.10)</td>
<td>1.13 (0.10)</td>
<td>1.79 (0.06)</td>
<td>2.39 (0.11)</td>
<td>1.57 (0.06)</td>
<td>1.61 (0.12)</td>
</tr>
<tr>
<td></td>
<td>0.14 (0.055)</td>
<td>0.04 (0.017)</td>
<td>0.09 (0.033)</td>
<td>0.16 (0.109)</td>
<td>0.09 (0.035)</td>
<td>0.10 (0.046)</td>
</tr>
<tr>
<td>Percentiles: 25</td>
<td>0.10</td>
<td>0.03</td>
<td>0.06</td>
<td>0.07</td>
<td>0.06</td>
<td>0.07</td>
</tr>
<tr>
<td>50</td>
<td>0.14</td>
<td>0.04</td>
<td>0.08</td>
<td>0.13</td>
<td>0.08</td>
<td>0.09</td>
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<tr>
<td>75</td>
<td>0.17</td>
<td>0.05</td>
<td>0.11</td>
<td>0.25</td>
<td>0.11</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>1.87 (0.250)</td>
<td>1.22 (0.107)</td>
<td>1.81 (0.357)</td>
<td>2.33 (0.508)</td>
<td>1.54 (0.239)</td>
<td>1.75 (0.250)</td>
</tr>
<tr>
<td>Percentiles: 25</td>
<td>1.72</td>
<td>1.14</td>
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<td>1.93</td>
<td>1.39</td>
<td>1.56</td>
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<tr>
<td>50</td>
<td>1.89</td>
<td>1.20</td>
<td>1.76</td>
<td>2.33</td>
<td>1.53</td>
<td>1.73</td>
</tr>
<tr>
<td>75</td>
<td>2.03</td>
<td>1.27</td>
<td>2.07</td>
<td>2.68</td>
<td>1.70</td>
<td>1.92</td>
</tr>
<tr>
<td></td>
<td>2.46 (0.6)</td>
<td>1.85 (0.4)</td>
<td>1.87 (0.9)</td>
<td>2.03 (1.1)</td>
<td>2.37 (0.3)</td>
<td>2.10 (0.5)</td>
</tr>
<tr>
<td>Percentiles: 25</td>
<td>2.11</td>
<td>1.53</td>
<td>1.25</td>
<td>1.05</td>
<td>1.89</td>
<td>1.79</td>
</tr>
<tr>
<td>50</td>
<td>2.51</td>
<td>1.78</td>
<td>1.69</td>
<td>2.23</td>
<td>2.46</td>
<td>2.08</td>
</tr>
<tr>
<td>75</td>
<td>2.89</td>
<td>2.15</td>
<td>2.34</td>
<td>2.85</td>
<td>2.85</td>
<td>2.46</td>
</tr>
<tr>
<td></td>
<td>0.41 (0.119)</td>
<td>0.50 (0.064)</td>
<td>0.40 (0.137)</td>
<td>0.34 (0.141)</td>
<td>0.46 (0.153)</td>
<td>0.36 (0.088)</td>
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<td>Percentiles: 25</td>
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<td>0.45</td>
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<td>0.24</td>
<td>0.35</td>
<td>0.30</td>
</tr>
<tr>
<td>50</td>
<td>0.38</td>
<td>0.50</td>
<td>0.38</td>
<td>0.29</td>
<td>0.43</td>
<td>0.35</td>
</tr>
<tr>
<td>75</td>
<td>0.45</td>
<td>0.54</td>
<td>0.49</td>
<td>0.41</td>
<td>0.53</td>
<td>0.43</td>
</tr>
</tbody>
</table>
Table 5.3  
Summary of Current Observations

<table>
<thead>
<tr>
<th>Station</th>
<th>L1w</th>
<th>L2w</th>
<th>L3w</th>
<th>U1w</th>
<th>U2w</th>
<th>U3w</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>Mean</td>
<td>Mean</td>
<td>Mean</td>
<td>Mean</td>
<td>Mean</td>
</tr>
<tr>
<td></td>
<td>(cm/sec)</td>
<td>(Std. Dev.)</td>
<td>(cm/sec)</td>
<td>(Std. Dev.)</td>
<td>(cm/sec)</td>
<td>(Std. Dev.)</td>
</tr>
<tr>
<td>U</td>
<td>7.02</td>
<td>4.55</td>
<td>2.19</td>
<td>6.88</td>
<td>4.60</td>
<td>5.07</td>
</tr>
<tr>
<td></td>
<td>(2.44)</td>
<td>(2.97)</td>
<td>(1.17)</td>
<td>(3.68)</td>
<td>(1.01)</td>
<td>(1.91)</td>
</tr>
<tr>
<td>Percentiles: 25</td>
<td>5.30</td>
<td>2.19</td>
<td>1.41</td>
<td>4.54</td>
<td>3.30</td>
<td>3.40</td>
</tr>
<tr>
<td>50</td>
<td>6.55</td>
<td>3.58</td>
<td>1.91</td>
<td>5.78</td>
<td>4.09</td>
<td>4.82</td>
</tr>
<tr>
<td>75</td>
<td>8.35</td>
<td>6.61</td>
<td>2.66</td>
<td>8.72</td>
<td>5.57</td>
<td>6.22</td>
</tr>
<tr>
<td>u</td>
<td>-2.23</td>
<td>-2.30</td>
<td>0.18</td>
<td>-2.89</td>
<td>-0.55</td>
<td>0.25</td>
</tr>
<tr>
<td></td>
<td>(4.04)</td>
<td>(3.51)</td>
<td>(1.59)</td>
<td>(6.89)</td>
<td>(2.24)</td>
<td>(4.20)</td>
</tr>
<tr>
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<td>-4.81</td>
<td>-5.08</td>
<td>-0.77</td>
<td>-7.13</td>
<td>-1.85</td>
<td>-2.33</td>
</tr>
<tr>
<td>50</td>
<td>-2.44</td>
<td>-1.60</td>
<td>0.15</td>
<td>-2.63</td>
<td>-0.54</td>
<td>0.15</td>
</tr>
<tr>
<td>75</td>
<td>0.31</td>
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<td>1.17</td>
<td>2.13</td>
<td>0.90</td>
<td>2.35</td>
</tr>
<tr>
<td>v</td>
<td>3.74</td>
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<td>0.08</td>
<td>-0.26</td>
<td>0.47</td>
<td>3.20</td>
</tr>
<tr>
<td></td>
<td>(4.89)</td>
<td>(2.32)</td>
<td>(2.14)</td>
<td>(4.62)</td>
<td>(1.31)</td>
<td>(1.63)</td>
</tr>
<tr>
<td>Percentiles: 25</td>
<td>0.64</td>
<td>1.03</td>
<td>-1.37</td>
<td>-2.78</td>
<td>-0.23</td>
<td>2.40</td>
</tr>
<tr>
<td>50</td>
<td>3.96</td>
<td>2.50</td>
<td>-0.04</td>
<td>-0.55</td>
<td>0.53</td>
<td>3.15</td>
</tr>
<tr>
<td>75</td>
<td>7.09</td>
<td>4.68</td>
<td>1.26</td>
<td>2.53</td>
<td>1.31</td>
<td>4.09</td>
</tr>
</tbody>
</table>
Equations 5.3 and 5.4 were tested in a comparison to previous observed values and by correlation to wind speed. These equations were substituted into Equation 2, and wind and wave data were used to calculate atmospheric friction factors $C_d$. Iteration was required since $U_a$ depends on $C_d$. New $U_a$ values were computed and $C_d$ re-estimated until the maximum change in $C_d$ over the time series was less than $2 \times 10^{-6}$ per iteration. Those $C_d$ values calculated with Equation 5.4 were consistently better correlated to $U_a$ and had much less scatter than those calculated with Equation 5.3, despite the fact that $h/L_w$ values for the various stations ranged from 0.24 to 0.45 at the 25th-percentile longest waves.

Atmospheric friction factor $C_d$ values calculated with Equation 5.4 were cast into a linear form versus wind speed such that

$$C_d = (6.49 + C_{cd} U_a) \times 10^{-4}, \quad U_a > 2 \text{ m/sec}$$ (5.21)

where the six stations had a common intercept. $C_{cd}$ data were separated into directional bands corresponding to the directional modes seen in Figure 5.6, and regressions were performed on each of three bands. Results for the six stations are presented in Table 5.4.

<table>
<thead>
<tr>
<th>Table 5.4</th>
<th>Coefficients for Equation 5.21 for Wind Direction Bands</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$C_{cd} \times 10$</td>
</tr>
<tr>
<td>Wind Direction</td>
<td></td>
</tr>
<tr>
<td>Station</td>
<td>All</td>
</tr>
<tr>
<td>L1w</td>
<td>9.61</td>
</tr>
<tr>
<td>L2w</td>
<td>7.71</td>
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<tr>
<td>L3w</td>
<td>7.58</td>
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<tr>
<td>U1w</td>
<td>9.24</td>
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<td>U2w</td>
<td>8.12</td>
</tr>
<tr>
<td>U3w</td>
<td>8.77</td>
</tr>
</tbody>
</table>
Station L2w located in a seagrass bed had relatively small waves and lower friction factors (13 percent lower at 10 m/sec wind speed) than L1w. It should be noted that station U1w is located in Corpus Christi Bay, and data obtained here were processed only for comparison. Some stations displayed wind-direction dependence with respect to $C_{cd}$. It appears that the directional $C_{cd}$ values do not correlate well to the corresponding fetches. Differences in $C_{cd}$ appear to be caused by depth and bed condition or seagrass type, as will be discussed below.

The finding that Equation 5.4 is better correlated to wind speed than Equation 5.3 suggests that wave age is more strongly related to $z_o$ than is wave steepness $H_s/L_w$. Wu (1980) also argued that wave steepness based on the dominant waves is not a sound physical scaling for atmospheric roughness. Furthermore, Equation 5.4 suggests there might be a relationship between two dimensionless parameters such that

$$
\frac{z_o}{H_s} \propto \left( \frac{C}{U_{*a}} \right)^n
$$

(5.22)

which is similar to a relationship that Donelan (1990) used. Both these dimensionless parameters are strongly related to the wind speed. Equation 5.22 with $n = -2$ is equivalent to Equation 5.4 and regression results indicated a multiple correlation squared (M-R²) = 0.978 and residual standard error (RSE) = 0.0686 for $n = 2056$. However, it was found through successive approximation of $z_o$ and $U_{*a}$ that $n = -1.57$ gave the best fit for Equation 5.22 (M-R² = 0.984, RSE = 0.0468, $n = 2056$). Thus, a new empirical relationship for the atmospheric roughness height was obtained:

$$
z_o = 0.0493 H_s \left( \frac{U_{*a}}{C} \right)^{1.57}
$$

(5.23)
Equation 5.23 was more consistent with the present data with respect to the residual standard error. However, the exponent of Equation 5.23 is much lower than those compiled by Donelan, whose exponents ranged from about 2 to 4. The addition of the dimensionless term introduced in Equation 5.7 to the right side of Equation 5.22 did not improve the fit or suggest a different value for the exponent.

Values of $C_d$ computed with Equations 5.4 and 5.23 increased more quickly with wind speed than did the linear relationships previously presented, as shown in Figure 5.9 for L1w. In shallow water, $C$ is restricted by water depth, and $C/U_{a}$ depends on depth and decreases linearly with wind speed, thus increasing wind separation and white-capping. Equation 5.23 not only improved the fit for $z_o$ but also improved the correlations between dimensionless wave parameters which use the resulting $C_d$ values, as will be described later.

A depth-dependent function was needed to describe $C_d$. By substitution of a depth-limited wave model for $H_s$, presented later, and expressions relating $C/U_{a}$ to $U_{a}$ and $h$ into Equations 5.2, the following expression was found by trial and is proposed for shallow water:

$$C_d = \left( \frac{0.4}{16.11 - 0.5 \ln (h) - 2.48 \ln (U_{a})} \right)^2, \quad h < 2 \text{ m} \tag{5.24}$$

Equation 5.24 predicts smaller $C_d$ values than Equation 5.5 when $h < \exp (3.25 - 1.2 \ln (U_{a}))$, for example, when $h < 1.6$ m at $U_{a} = 10$ m/sec.

Estimated $C_d$ values from Equation 5.24 are shown in Figure 5.10 along with values determined by use of Equation 5.23 for vegetated and nonvegetated stations in Laguna Madre. However, more comparisons to data are needed to determine the generality of Equation 5.24. It is difficult to discern the relative importance of depth and seagrass to $C_d$. 

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Figure 5.9. Comparison of atmospheric friction factors calculated based on Wu (Wu 1980), A & K (Atakturk and Katsaros, 1999), and Equations 5.4, 23, and 24.
Figure 5.10. Comparisons of atmospheric friction factors calculated Equations 5.23 (symbols) and 5.24 (lines) for four measurement stations.
with available data, but it appears depth, or effective depth as discussed later, is more critical to $C_d$ than seagrass is.

Fetch length is also a factor that affects $C_d$ in enclosed waters and has been included in formulations presented by Hsu (1988). As indicated earlier, wave data from Laguna Madre indicate that previous fetch relationships developed from unvegetated, slightly deeper, areas should be used with caution. The difference here may be that, for seagrass areas with high wave-frictional dissipation, the ratio of wave energy to energy dissipation is smaller than for unvegetated, deeper areas. Where momentum is transferred from the atmosphere to small wave lengths, waves build more quickly to a steady condition in which input and dissipation of momentum balance.

**Shear Stress Budget**

Charnock’s parameter decreases with wind speed, so that according to Equation 5.9, almost all shear stress is transferred to waves at low wind speeds, and greater portions of $\tau_a$ are dissipated by white-capping and/or water-column mixing, and less of the total input is dissipated by wave friction at higher wind speeds. Wave dissipation as a result of mixing was not considered, since the Laguna system is normally wind-mixed and tends to be vertically homogeneous. Wave breaking by excess steepness or relative depth was not observed in the Laguna Madre.

For fully-developed waves (constant wave spectra), wave dissipation through total friction and white capping is assumed to be equal to the atmospheric shear stress at the water surface. Even when only a few percent of $\tau_a$ goes directly into currents as $\tau_{ac}$, some part of $\tau_{aw}$ is subsequently transferred to $\tau_{ac}$ by white-capping, thus, $\tau_a$ is assumed to be partitioned between shear stress imparted to waves ($\tau_{aw}$) and to currents ($\tau_{ac}$):

$$\tau_a = \tau_{aw} + \tau_{ac}$$ (5.25)
Ratios of $\tau_{aw}/\tau_a$ were computed (after the unknown roughness height $k_n$ was removed) with Equations 5.1-2, 12, 14-16, and 23. The peak wave shear stress is 2.38 times the average shear stress. Regressing this ratio against wind speed gives an indication of the amount of atmospheric shear stress going into wave shear stress. By assuming that almost all atmospheric shear stress goes into waves at wind speeds 3 to 5 m/sec where a peak in the ratio occurred (assumed to be 0.97), results indicated that the amount decreases at higher wind speeds as

$$\frac{\tau_{aw}}{\tau_a} = 2.169 U_a^{-0.5}, \quad U_a > 5 \text{ m/sec}$$  \hspace{1cm} (5.26)

Plots of $\tau_{aw}/\tau_a$ calculated by Equations 5.26 and 5.9 versus wind speed are shown in Figure 5.11 for L1w. Charnock’s parameter was calculated with Equations 5.6 and 5.23 and was higher than reported values for open waters. Since $\alpha_c$ did not approach the previously cited value of 0.012 for “old” waves, the fraction of shear stress going into waves remained high. Computed values of $\tau_{aw}/\tau_a$, which lead to Equation 5.26, had a great deal of scatter, as can be seen in Figure 5.11. The regressions for L1w had M-R$^2 = 0.09$, p-value $< 0.01$, and the resulting exponent = -0.435. The regression for L2w had M-R$^2 = 0.15$, p-value $< 0.01$, and the resulting exponent -0.66. Equation 5.26 is, therefore, only approximate. Another possible way to determine $\tau_{aw}/\tau_a$ is to determine $1 - \tau_{sc}/\tau_a$ instead. This calculation might be done by determining water surface slope and velocity along the direction of the wind, and therefore the atmospheric shear stress going into the currents. Depth information indicated that at times of $U_a < 5 \text{ m/sec}$, the water surface slope between L1w and L2w did not vary. At higher wind speeds variation in water levels of 0.1 m occurred. Experience in applying hydrodynamic models to other systems has indicated that an appreciable fraction of shear
Figure 5.11. Ratios $\tau_{aw}/\tau_a$ calculated according to Charnock’s parameter and Equation 5.9, Equation 5.26, and wave characteristics and atmospheric shear stresses.
stress must go into currents in order to reproduce observed water-level fluctuations. Equation 5.26 should be used with caution.

**Depth-Limited Waves with High Bed Roughness Effects**

Wave data from the station U1w, located in southern Corpus Christi Bay, was found to be different from data for those gauges located within Laguna Madre. That station was slightly deeper than those in Laguna Madre proper and provided a valuable contrast of a different shallow-water wave climate. Dimensionless wave energy $E^*$ values are plotted against $h^*$ in Figure 5.12a for station U1w for all wind and for the optimal wind conditions described earlier. The values for the uniform wind condition identified earlier were suspected of being optimum for comparisons and were separated out from the remainder of the data. The data centered at 130° wind direction were largely a super-set of the uniform wind set and were not used separately. Also plotted in Figure 5.12a are CERC (1984) and Young and Verhagen (1996) fits to Equation 5.10, which should correspond to the upper edge of the scattered data. Fetch limitations bring values downward away from this line. It appears that the CERC (1984) coefficients make Equation 5.10 more parallel to the upper edge, although if some data errors were admitted, a slight reduction in $a_1$ might be warranted. Scatter of the data in Figure 5.12a is similar to the data presented by Young and Verhagen (1996).

Dimensionless wave energy $E^*$ values are plotted against $h^*$ in Figure 5.13 for all other stations except U2w. The data set from station U2w had only 219 points total and 71 points in the uniform-wind set, only about one-third as many points as the next smallest data set, and, therefore, will be omitted from subsequent data analyses. The $E^*$ versus $h^*$ data scatter can be seen in Figure 5.13 as more tightly banded than for U1w, especially as individual stations, but the edge of the scattered data is displaced downward from previous equation fits. Station L2w is located in dense *Thallasia* seagrass, and these plotted $h^*$ values were
Figure 5.12. Dimensionless wave parameters for station U1w (diamonds are optimum winds and dots are all other winds) with model fits: (a) wave energy versus depth, and (b) wave frequency versus depth.
Figure 5.13. Dimensionless wave energy $E^*$ versus depth $h^*$ for four Laguna Madre measurement stations with model fits.
computed with $h$ adjusted -0.2 m to allow for the seagrass canopy. *Thallasia* sp. was the most resistive to currents of those tested by Fonseca and Fisher (1986).

Dimensionless frequency $f^*$ values are plotted against $h^*$ in Figure 5.12b for U1w. The depth-limited case falls on the lower edge of the data scatter and can be compared to the previous results from CERC (1984) and Young and Verhagen (1996) in this plot. The slope implied by the present data scatter is different from these previous studies. Results for L1w, L2w, L3w, and U3w are shown in Figure 5.14. As with $E^*$ versus $h^*$, the collective- and individual-station scatter away from the depth-limited case, as described by Equation 5.11, is less for these stations than for U1w and previous studies.

The effect of fetch length on wave conditions can best be examined with use of data from L1w and L2w, which were located inline with each other in the dominant wind direction (110° to 145° with a mean of 128°) and had more uniform depth along this direction. Fetch lengths differed by a factor of two for this wind direction, 10.2 and 5.1 km respectively. Following Young and Verhagen (1996), $E^*$ versus $\chi^*$ for increments of small $h^*$ (high winds) were plotted in accordance with data from L1w and L2w as shown in Figure 5.15. Bounding curves for $h^*$ values were plotted as described by Equations 5.12 and 5.13. Based on those equations, data should plot between the curves presented. Data from Laguna Madre diverge sharply from the results of previous studies with regards to the effect of fetch length, as can be seen from Figure 5.15. The data scatter suggests a constant slope of about two. However, since $E^*$ is a function of $U_a^{-4}$ and $\chi^*$ is a function of $U_a^{-2}$, a slope of 2 represents a spurious correlation between these two parameters (Young and Verhagen 1996). Such a relationship would suggest that wave height is related only to fetch length (that is a constant for this wind direction) and not related to wind speed. It appears that the customary values of dimensionless fetch $\chi^*$ do not describe the effect of fetch in the Laguna Madre data.
Figure 5.14. Dimensionless wave frequency $f^*$ versus depth $h^*$ for four Laguna Madre measurement stations with model fits.
Figure 5.15. Dimensionless wave energy versus fetch length $X^*$ at stations L1w and L2w for four ranges of $h^*$. 

(a) $h^* = 0.1$ to $0.2$
(b) $h^* = 0.2$ to $0.3$
(c) $h^* = 0.3$ to $0.4$
(d) $h^* = 0.4$ to $0.5$
Apparently, fetch effects occur over much smaller $\chi^*$ in Laguna Madre than previously found, possibly due to high bottom frictional effects. The lack of fetch effects is also reflected in the relatively low data scatter about the trends shown in Figure 5.13. Since in this case the depth-limited wave condition is not an asymptote but is assumed to be the main tendency of the data, fits to the data were made through this central tendency rather than at the edges of the scattered data. Scatter about this tendency is considered to be introduced by various errors and not importantly by the effects of fetch.

More appropriate scaling for dimensionless wave energy and wave frequency in ultra-shallow water was found by use of the atmospheric friction velocity, $U_{a*} = C_d^{0.5}U_a$, in place of $U_a$. The wave-model expressions corresponding to Equations 5.10 and 5.11 are

$$E^* = a_5 h^{a_6} \tag{5.27}$$
$$f^* = a_7 h^{a_8} \tag{5.28}$$

where the new dimensionless parameters $E^*$ and $f^*$ equal $g^2E/(C_d^2U_a^4)$ and $C_d^{0.5}U_a/(g T_p)$, respectively. With these scalings, data were brought closer into line (with higher M-R² value) when plotted against $h^*$, as shown in Figures 5.16 and 5.17, than in the comparisons shown in Figures 5.13 and 5.14. Regressions were performed between $E^*$ and $h^*$ with data subsets for uniform wind, winds greater than 3 m/sec, and $H_s$ values greater than the 25th percentile conditions. Results yielded exponents $a_6$ for $h^*$ of between 1.73 and 2.11. The assumption of an exponent of 2.0 implies that $H_s \propto C_d h$, and regression with this form yielded the following empirical expression:

$$H_s = 84.6 h C_d - 0.056 \tag{5.29}$$

(M-R² = 0.765, RSE = 0.0253 m, n = 2056) where the intercept is apparently caused by the range of $C_d$ which does not converge to zero at zero wind speed. Plots of observed versus calculated $H_s$ for the four stations are shown in Figure 5.18.
Figure 5.16. Dimensionless wave energy $E'$ versus depth $h^*$ for four Laguna Madre measurement stations.
Figure 5.17. Dimensionless wave frequency $f'$ versus depth $h^*$ for four Laguna Madre measurement stations.
Figure 5.18. Significant wave heights computed with Equation 5.29 versus observed values for four Laguna Madre stations.
Figure 5.19. Peak wave periods computed with Equation 5.30 versus observed values for four Laguna Madre stations.
Regressions performed between \( f^- \) and \( h^* \) indicated an exponent \( a_s \) of -0.5, thus \( T_p \propto (h C_d/g)^{0.5} \) and further analysis indicated that

\[
T_p = 126.5 (h C_d/g)^{0.5}
\]  

(\( M-R^2 = 0.982, \text{RSE} = 0.241 \text{ sec, } n = 2056 \)). Plots of observed versus calculated \( T_p \) for the four stations are shown in Figure 5.19.

**Wave Friction Factors**

Wave shear stresses \( \tau_w \) were estimated with laminar, turbulent, and transitional laminar/turbulent frictional formulations. Computation of laminar shear stresses was based on observed wave characteristics and Equations 5.14-17, and results were correlated to computed \( \tau_w \) values. The laminar friction formulation produced shear stresses which were often higher than the atmospheric shear stress input. Even though the laminar formulation is often used for shallow-water resuspension calculations, it appears to be physically unrealistic in this case. Currents at these sites are low but almost always above threshold values required to produce turbulent water-column conditions. Apparently a wave-current interaction occurs through the eddy-viscosity profile whereby waves assume the turbulent condition of the water column even at low wave Reynolds numbers.

The turbulent rough formulation (Equation 5.18) had consistently higher correlations than the other formulations for the six stations. A wave model was developed which used iteration to arrive at a fully-developed wave height and period, matching observed values reasonably well, and in balance with atmospheric shear stress and dissipation. Turbulent-rough-bottom friction described by Equation 5.18, and wave breaking dissipation calculated according to the method of Massel and Belberova (1990) were included in the model formulation. Whitecapping and dissipation from spectral wave interactions were lumped together as one dissipation mechanism, and shear stress partitioned according to Equation 5.9.
The wave model was validated by a comparison of observed and model wave height distributions and by a comparison of computed shear stresses to observed wind shear stresses. The main results were the estimates for the roughness height, \( k_n \), at the Lower Laguna Madre stations, which are presented in Table 5.5.

<table>
<thead>
<tr>
<th>Station</th>
<th>( k_n, \text{ m} )</th>
<th>Median peak ( \tau_w, \text{ Pa} )</th>
<th>Median ( \tau_a \times 2.38, \text{ Pa} )</th>
<th>Median ( H_s, \text{ m} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1w</td>
<td>0.0014</td>
<td>0.049</td>
<td>0.058</td>
<td>0.133</td>
</tr>
<tr>
<td>L2w</td>
<td>0.2</td>
<td>0.044</td>
<td>0.065</td>
<td>0.044</td>
</tr>
<tr>
<td>L3w</td>
<td>0.01</td>
<td>0.053</td>
<td>0.051</td>
<td>0.083</td>
</tr>
</tbody>
</table>

The atmospheric shear stress, factored by 2.38, was added to Table 5.5 for comparison to the peak-wave shear stress. The estimated roughness heights are consistent with the bottom type, where L1w is bare bottom, L2w is thick seagrass bed, and L3w is seagrass edge. The peak shear stresses were predicted to be not much different for these stations even though wave heights were very different.

**CONCLUSIONS**

Wave conditions varied markedly over Laguna Madre, depending on water depth and seagrass conditions. Atmospheric friction factors were calculated with the use of wind and wave data, and the methods deemed best used wave age and wave height. Friction factor results were used to fit a simpler, linear drag law for examining possible fetch-length effects, and also to fit a function for atmospheric friction factor in terms of depth and wind speed.

The influences of bed roughness and friction on waves in shallow water have not been well studied and need to be resolved (Young and Verhagen 1996). Previous analytic models of depth-limited waves were not consistent with data from seagrass and unvegetated areas in
Laguna Madre with less than about 2-m water depth. Likewise, previous fetch-limiting wave relationships did not compare to these data. Waves tended to be depth limited, at least for the fetches greater than about 3 km included in this data set. A new wave model based on a rescaling of dimensionless wave energy and period was developed and was reasonably successful at predicting wave characteristics in terms of water depth and atmospheric friction factor. For seagrass areas, it is suggested that canopy height be subtracted from the total depth when an estimation of atmospheric friction and wave characteristics is made.

A number of wave-friction formulations based on combinations of laminar, smooth and rough turbulent theory were tested. Computed wave shear stresses were correlated to estimated wind shear stress. A turbulent-rough formulation performed best. Thus, turbulent water-column conditions apparently interacted with waves. Ratios of wave to wind shear stresses were estimated, and an expression developed to estimate the fraction of wind shear stress converted into wave shear stress. From a partitioning of input shear stress, wave friction roughness heights were back calculated from observed wave characteristics at the three Lower LM stations. The resulting roughness heights appear to be reasonable: relatively low for the bare-sediment station and typical of previously observed trends in frictional characteristics for seagrass species present at two sites.
Chapter 6
Sediment Dispersion Near a Dredge Pipeline Discharge in Laguna Madre, Texas

ABSTRACT

Initial dispersion of dredged material after pipeline discharge is important to deposit area and susceptibility to erosion or resuspension. Dispersion consists of a fluid-mud gravity underflow and, possibly, an overlying water-column plume. Fluid-mud thickness, concentration structure, and overlying water-column suspension concentration were measured in shallow, wind-exposed, micro-tidal Laguna Madre, Texas, within about 500 m of where a dredge pipeline was discharging. Depths were 0.5 to 2 m and currents were weak. The dredged material had a median particle size of 4 to 5 μm. Median fluid-mud thicknesses were 0.45 m of which the top 60 percent was interpreted as underflow and the remainder as deposit. Fluid-mud concentration at the upper surface of the underflow layer was about 3 dry-kg/m³ and increased exponentially with depth to about 48 dry-kg/m³. The deposit was 48 to 110 dry-kg/m³ solids.

A numerical model that would simulate underflow fluid-mud spreading resulting from a pipeline discharge was developed as an aid in the diagnosis and interpretation of field measurements. The model was based on one-dimensional equations for momentum and mass conservation. Model features that limited entrainment and concentration change caused by deposition were incorporated as indicated by field observations. A plume of suspended sediment 200 to 500 mg/l above ambient concentration occurred over the underflow footprint, with resuspension driven by wind-waves. The development of a point model of the water column overlying a fluid mud layer was based on a balance between
entrainment and settling. Settling was prescribed on the basis of a laboratory-developed functional dependence on concentration. Data were used in the model to estimate coefficients for this entrainment process.

**INTRODUCTION**

Maintenance dredging of the Gulf Intracoastal Waterway (GIWW) annually involves more sediment than the total of the natural sediment inputs to Laguna Madre (Morton et al. 2001). Hydraulic dredging followed by open-water disposal through a pipeline is the most common dredging method in shallow, vegetated Laguna Madre, Texas. There are concerns that redispersed dredged material in Laguna Madre is contributing to turbidity and is limiting light penetration to seagrasses over the long term (Onuf 1994). The extent of initial dredged-material spreading is important information for assessing total resuspension and predicting possible sediment impacts outside designated disposal areas.

Measurements of fluid-mud thickness, concentration structure, and overlying suspension concentration were made north of Port Mansfield, Texas, in Laguna Madre, within about 500 m of an ongoing dredged-material pipeline discharge. Such information is scarce but is needed to improve understanding of the behavior of such discharges and subsequent sediment dispersion. Field measurements were compared here to results taken from a near-field, numerical underflow-spreading model for the purpose of improving model formulation. Information was also used to characterize pipeline discharges in a large-scale numerical sediment-transport model of this system (Teeter et al. 2002). A simple water-column point model was also used to estimate coefficients for an entrainment relationship that would describe the flux of underflow sediment into the water column. As will be shown, the extent of the underflow affects the extent of any surface plume of
suspended sediment which might form during or shortly after discharge as a result of entrainment of the underflow into the overlying water column.

The 183-km section of Gulf Intracoastal Waterway (GIWW) in shallow Laguna Madre, Texas, is located between Port Isabel and Corpus Christi Bay, or geographic coordinates 26.1° to 27.7° N latitude and about 97.4° W longitude. This section requires about 1.6 x 10⁶ m³ of maintenance dredging annually. About 75 percent of the 1,500-km² Laguna Madre is vegetated, and seagrasses are sensitive to underwater light conditions (Dunton 1994). A series of studies has recently been undertaken to evaluate the impact of dispersion of sediments from the placement areas on seagrass areas in Laguna Madre (Brown and Kraus 1997; Miletello et al. 1997; Burd and Dunton 2000; Mortin et al. 2001; and Teeter et al. 2002).

BACKGROUND ON PIPELINE DISCHARGE UNDERFLOWS

In shallow water, dredged sediment particles reach the bottom soon after pipeline discharge, settling within a short distance from the discharge point. As the bottom layer thickens at the point of discharge, it behaves as a density flow and spreads under the influence of gravity (Neal et al. 1978). Sediments form layers of fluid mud at the bed, which flow away from the point of discharge, the extent of the flow depending on bottom slope, ambient currents, and their initial discharge trajectory. It has been previously estimated that 95 to 99 percent of discharged sediment mass descends to the bottom layers within about 30 m from the point of a pipeline discharge (Schubel et al. 1978; and Neal et al. 1978). In Mobile Bay, for example, 99 percent of the discharged sediment was found to be dispersed along the bottom as fluid mud (Nichols et al. 1978).

The range of concentrations for the fluid-mud definition used here is roughly 5 to 400 dry-kg/m³ (corresponding roughly to 1,003 to 1,250 wet-kg/m³ density). Concentrations of
pipeline-discharge solids are within this range, and solids are generally about 15 percent by weight or 150 to 200 dry-kg/m³ (Schubel et al. 1978).

The approach channel to the Chesapeake & Delaware Canal in Upper Chesapeake Bay was hydraulically dredged in 1988. About $5.2 \times 10^5$ m³ of clayey-silt sediment were pumped and deposited in Areas D, E, and F near Pooles Island. Depths were 2.5 to 3.5 m before disposal at Area D where most of the material was placed. The movement of sediment was down-slope after discharge. A broad continuous layer formed about 3 km long and 1.5 km wide. The maximum deposit thickness was 1.5 m. Sediment consolidated to a density of 1,130 kg/m³ or greater within several weeks. Dewatering and compaction accounted for 5 percent deposit-volume reduction in 5 months. Another 5 percent reduction occurred during the discharge period. The remaining 22 percent of the 32 percent total reduction was from redistribution by resuspension and transport (Panageotou and Halka 1990).

Near Pooles Island, $5.2 \times 10^5$ m³ of sediment were hydraulically dredged from the nearby channel in 1991 and placed in Areas D and E. Sites were 4.5 to 8 m deep. Sediments deposited in a natural trough and constructed trenches. The sediment remained in the deep, trough area. The volume of the deposit was $1.04 \times 10^6$ m³ with maximum thickness of 3 m. Sediment were clayey silt with minor sand. Bulking factor between in-place and deposited volumes was about 1.75. One year later the deposit was $4.4 \times 10^5$ m³ (58 percent reduction). Four-fifths of the reduction was attributed to dewatering, one-fifth to erosion (Panageotou and Halka 1994).

Underflow spreading controls the configuration of the final deposit. Limited observations indicate that the final deposit is a series of strata laid down as the underflow shifts and grows larger in response to bottom topography. Maximum deposit thickness was
about 0.3 m for a typical 2-day disposal operation in Mobile Bay and about 1.8 m for a 10-day disposal in the James River (Nichols et al. 1978).

FIELD OBSERVATIONS

Field experiments were carried out to take advantage of dredging conducted in February 2000 to remove a 1.8-m layer of material deposited in the GIWW as a result of a hurricane the previous year. *Dredge J.N. Fisher* discharged into open-water disposal sites through a 50.8-cm diameter pipeline, using a 1,500-kw (2,000-horsepower) pump. The dredging rate was about 1,100 m³/hr, and, based on the solids content of the channel material, the sediment discharge rate was about 50 dry-kg/sec.

Fluid-mud thicknesses, or heights, and densities were measured on two days while pipeline discharge was occurring. Locations for the discharges are shown in Figure 6.1. A special push-tube sampler allowed for fluid-mud density determination within only a few minutes of sampling. Samples were collected for analysis of the fluid-mud concentration to supplement the field-density measurements. Fluid-mud particle-size distribution and ambient water column suspended-sediment concentrations were also measured. A composite sample was used in the laboratory to determine velocities in the hindered settling range.

Field Methods

A 5.8-m-long flat bottomed boat with a propeller tunnel to minimize draft was used for sampling. A Starlink ® Differential Global Positioning System was used to locate stations to within ±2 m, and an HP PalmPC ® was used to log positions in the field. Water-column samples were collected with a submersible Rule ® electric pump and 1.5-cm diameter hose. Water samples were collected at mid-depth and 0.3-m depth and stored in 225-ml plastic bottles.
Figure 6.1. Vicinity sketch of Lower Laguna Madre north of Port Manfield, Texas, with depth contours and discharge locations (coordinates are state plane NAD27, Texas South, in meters)
Fluid mud was sampled with a push corer with a clear 3.6-cm diameter core tube and a total length of about 3 m. During the first sampling day, it was found that the in-line check-valve developed too much back pressure, resulting in significant errors in underflow sampling. For the second sampling day, fluid-mud samples were collected with a low back-pressure push-core sampler specially fabricated from parts of a WILDCO ® corer. That sampler can be seen in Figure 6.2. Only the fluid mud measurements from the second day are reported.

The boat was brought to a new location, and the anchor was set. A couple of minutes were allowed for the boat to swing to and for the position to be logged. The corer was pushed vertically downward by hand until it encountered firm bottom. A trip line was then pulled to seal the top of the sampling tube. The corer captured ambient water column, fluid mud (if present), and a short plug of the underlying bottom material. (The bottom material contained an appreciable sand fraction not present in the dredged material and had a bulk density of roughly 1,500 kg/m³). The vertical alignment of the core tube was maintained as it was lifted to the deck and a piston push-rod was inserted into the lower end of the core tube (below the sediment plug). After the core tube was unscrewed from the remainder of the sampler, the piston rod was pushed upward to expell sample from the end of the tube. By incrementally extruding sample from the end of the core tube, scientists could take measurement and sub-samples over the vertical dimension of the fluid mud. Density measurements were made in the field with a PARR ® DMA35 vibrating-tube densitometer (precision of 1 kg/m³). A short length of 2-mm diameter tubing was inserted 2.5 cm into the end of the core tube, and a 5 to 10 cm³ sample was drawn through the densitometer. Field density measurements were made in duplicate and averaged.
Figure 6.2. Fluid mud sampler on deck in the open position
Laboratory Methods

Laboratory bulk wet density determinations were made with the use of 25-cm³ wide-mouth pycnometers. Pycnometers were weighed after being mostly filled with sample and then carefully topped with distilled water. Bulk wet density was calculated from this information and known characteristics of the pycnometers. Pore-fluid density was estimated on the basis of the salinity determined on suspended samples, allowing the calculation of sample solids content from bulk wet density (assuming a solids density of 2,650 kg/m³).

Total suspended material (TSM) was determined by a gravimetric method for non-filterable solids with preweighted Nuclepore ® 0.45 µm pore diam, polycarbonate filters. After being used to filter a known volume, filters were rinsed with distilled water, and dried one hour at 90 °C and then reweighed. Particle-size distribution was measured with a Coulter LS100Q ® laser scattering instrument. Samples were first oxidized with Clorox ® to remove organics and then were dispersed with sodium carbonate/bicarbonate. Three oxidation steps and three dispersion steps were performed before samples were processed through the Coulter instrument to determine particle size. The Coulter has 128 geometrically spaced channels, or bins, for sizing.

Settling velocities in the hindered-settling concentration range were measured on left-over sample that had been compositied to make a slurry. The slurry had a bulk density of 1,109.5 kg/m³, pore-fluid density of 1,025.7 kg/m³ (37.3 ppt salinity), and solids content of 136.7 dry-kg/m³. Sample was incrementally added to a 2 liter glass, graduated cylinder which was 7.74 cm in diam and 42.5-cm high at the 2-liter level. Six tests with concentrations of 6.8 to 66.3 dry-kg/m³ were made at 23 °C. After the sample was mixed in the cylinder, height of the interface between the suspension and the clear layer that formed was observed over time. The duration of the lowest initial-concentration test was about an
hour. During other tests, frequent measurements were collected over 100 to 240 min; these tests lasted a total of 1,100 to 1,450 min. Final data points allowed for estimation of average density after about one day of settling time.

Linear regressions were fit to the data for the period when the interface descended linearly (n = 3 to 20, R² = 0.944 to 0.999, standard error on slope = 1.82 to 0.025 mm/min) to determine the hindered settling velocity ($W_s$) at initial test concentration $C$. Tests with the lowest two concentrations were repeated, and data sets were combined in the regression analysis. Finally, $W_s$ and initial concentration from the six tests were combined and fit to an empirical equation for hindered settling dependence on concentration

$$W_s = W_{h_0} (1 - kC)^n, \quad C > \text{hindered-settling threshold}$$  \hspace{1cm} (6.1)

where the hindered settling threshold is usually in the range of 1 to 10 dry-kg/m³.

**Field and Laboratory Results**

All settling tests were in the hindered settling concentration range. Settling rates decreased about two orders of magnitude over the concentration range tested. Data greater than 6.8 dry-kg/m³ fit Equation 6.1 well with the reference hindered settling velocity $W_{h_0} = 0.5$ mm/sec, coefficient $k = 0.005$ m³/kg, and the exponent $n = 11$. Settling test results are plotted in Figure 6.3 along with results from the low-concentration settling tests performed by Teeter et al. (2001a), using Laguna Madre GIWW sediments collected about 3 km north of Port Isabel. The mean depth-average concentration at the end of the settling tests (about 20 hrs) was 115.5 dry-kg/m³ (with one high outlier of 148 dry-kg/m³ removed, n = 4, 95 percent confidence interval 112.0 to 118.9 dry-kg/m³).

**February 10.** When sampling began at 1000 Central Standard Time (CST) on 10 February, south winds were 9 m/sec, making sampling conditions very difficult. Currents
<table>
<thead>
<tr>
<th>INITIAL TEST CONCENTRATION, kg/cu m</th>
<th>SETTLING VELOCITY, mm/sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>0.001</td>
</tr>
<tr>
<td>0.5</td>
<td>0.005</td>
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<td>1.0</td>
<td>0.050</td>
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<td>50.0</td>
<td></td>
</tr>
<tr>
<td>100.0</td>
<td></td>
</tr>
</tbody>
</table>

Figure 6.3. Hindered settling test results (right) with the (dashed) fit to the data described in the text, and low-concentration settling results (left) from Teeter et al. (2001a)
were weak and toward the south. The pipeline discharge was located at coordinates 26°
37.4358' N and 97° 24.8643' W, about 3.7 km north of the entrance to Port Mansfield at the
disposal area designated PA 218. See Figure 6.1. A plot of the station locations, CST and
depths in a local horizontal coordinate system are shown in Figure 6.4. The origin for the
local coordinate system is at state plane 106,000 m N and 717,000 m E (NAD27, Texas
South). The pipeline was located 20 m west of the 1016 CST station and changed only
slightly as the dredge moved. Water-column samples taken at 0.3 m depth had TSM levels
(mean = 211 mg/l, 95 percent confidence interval 51 to 370 mg/l, n = 12) equivalent to those
from mid-depth (mean = 199 mg/l, 95 percent confidence interval 55 to 344 mg/l, n = 11).
Both sampling depths showed highly variable TSM. Depth averaged TSM are shown in
Figure 6.5. High suspension concentrations were measured both north and south of the
pipeline discharge. Depth-average TSM values of 549, 557 and 572 mg/l were obtained
near and down-drift within about 660 m of the discharge. Stations taken at 1107, 1126, and
1255 CST had what was apparently background level TSM ranging from about 100 to 120
mg/l. One station (1200 CST) taken upstream of the discharge had 262 mg/l TSM, possibly
as the result of local resuspension.

A photograph taken from 1,900-m altitude above the dredging operation at 1019 CST is
shown in Figure 6.6. The pipeline length was about 450 m, and the discharge flowed to the
east of the channel. Ambient and dredged-material plumes at both the dredging and disposal
sites are shown in the photograph. The aerial photograph shows the area directly south of
the discharge to be the heaviest visible plume. A plume emanating directly from the
discharge had a blue coloration, while other plume areas were milky.

**February 16.** Winds were from the south. Waves were 0.30 cm or less. Currents were
weak and moved toward the north (≈2 cm/sec). The pipeline discharge was located at
Figure 6.4. Times/depths (CST/m) for stations taken 10 February 2000
Figure 6.5. Depth-averaged TSM (mg/l) collected 10 February 2000
Figure 6.6. Overflight photo of the dredging and disposal operation taken 10:19 CST 10 February 2000 (coordinates are state plane NAD27, Texas South, in meters)
coordinates 26° 44.9752' N and 97° 27.3349' W, about 3.3 km south of the entrance to the Land Cut, at PA 213. See Figure 6.1. The discharge was to the east of the channel onto a mound with about a 0.3-m water depth. Dredging records indicated that the previous discharge location was 360 m north in PA 213. Discharge started there at about 2200 CST on 15 February; the discharge that was sampled began at about 0800 CST on 16 February. During the sampling period, 12,000 to 15,000 m³ of dredged material was discharged at these two sites.

Station times and depths are shown in Figure 6.7. The origin for the local coordinate system is at state plane 120,000 m N and 713,000 m E (NAD27, Texas South). A turbulent surface flow formed in the vicinity of the discharge jet and extended into deeper water. A photograph of the surface jet and flow is shown in Figure 6.8. A plunge line could be clearly seen in the field at a water depth of about 1 m, and an underflow moved toward the deeper water to the east-southeast. Samples taken at 0915 CST were within the turbulent surface flow, and two field measurements and two pump samples indicated that the turbulent surface flow averaged 17 dry-kg/m³.

Fluid mud formed a sharp interface with the ambient suspension, and its thickness was easily measured through the clear core tube. Fluid-mud profiles are presented in Figure 6.9. Because of concerns about settling effects and time constraints, few samples near the upper underflow interface were made. The fluid-mud layer was highly stratified in the vertical. Gradients indicated that concentrations at the upper interface were low. Several measurements indicated minimum underflow concentrations of 3 to 5 dry-kg/m³. However, the upper surface of the underflow was a distinct, sharp interface, indicating a concentration jump associated with the maximum flux of suspended material (Teeter 1986). Concentrations in the underflow were therefore above the concentration at which the
Figure 6.7. Times/depths (CST/m) for stations taken 16 February 2000
Figure 6.8. Photo of a pipeline discharge into about 0.5 m water depth and the resulting turbulent surface flow
Figure 6.9. Vertical fluid mud profiles taken 16 February 2000
maximum settling flux occurred. The maximum settling flux apparently occurred between the settling flux at 6.8 dry-kg/m³ (0.0051 kg/m²/sec) determined in the hindered settling tests and the previous 1 dry-kg/m³ low-concentration test (0.0006 kg/m²/sec). Thus, 3 dry-kg/m³ was estimated to be the minimum or underflow interface concentration. Points were added to profiles as shown in Figure 6.9 at the measured interface locations and the assumed 3 dry-kg/m³.

Fluid-mud layers consisted of underflow and deposit, as interpreted by the following information. Near the top of the fluid-mud layers, concentrations increased exponentially with depth and in approximately straight lines when plotted on semi-log axes as in Figure 6.9. This distribution would be expected for a turbulent flow with a particle Peclet number \( Pe = W \cdot h / K_z \) where \( h \) is the underflow thickness and \( K_z \) is the layer-average turbulent diffusivity) greater than about one (Teeter 1986). Another steeper gradient was evident below many of these exponential layers, and, taking all measurements together, the statistical distribution of fluid-mud concentrations had an inflection at about 50 dry-kg/m³. Previous laboratory experiments on sediment from nearby Corpus Christi Bay (76 percent clay and 21 percent silt) indicated that the mean concentration of newly-deposited material was 46 dry-kg/m³ (Teeter 1986). Therefore, the lower portion of these fluid-mud layers was interpreted as deposit from the underflow, while the upper layer, with concentrations of about 3 to 48 dry-kg/m³, was interpreted as the underflow. Normalized underflow concentration profiles were very similar, as can be seen in Figure 6.10, and their exponential shapes suggest some degree of vertical mixing, consistent with a flow with some turbulence.

All measured fluid-mud thicknesses and a rough interpretation of the underflow footprint extent are shown in Figure 6.11. The 0915 CST samples were assigned 0.0 fluid-mud thickness in this figure because they were located within the turbulent surface flow. It
Figure 6.10. Interpreted and interpolated underflow profiles normalized by underflow thicknesses and mean concentrations.
Figure 6.11. Fluid mud thicknesses collected 16 February 2000
appeared that the underflow footprint formed by the discharge ongoing during sampling overlapped that formed at the previous discharge location. Underflow mean concentrations $C$, thickness $h$, and deposit thickness $delbed$ were calculated on the basis of a 50-point interpolation over the fluid-mud profiles as shown in Figure 6.12 for these stations, along with the water column depth above the underflow ($H_o$).

Surface TSM levels (mean = 258 mg/l, 95 percent confidence interval 114 to 402 mg/l, n = 17) and mid-depth levels (mean = 262 mg/l, 95 percent confidence interval -205 to 728 mg/l, n = 8) were equivalent again this day. A plot of depth-mean TSM values is shown in Figure 6.13, along with an interpreted underflow footprint extent. As can be seen, the highest suspended concentrations occurred over the underflow, whether upstream or downstream from the discharge point. The implication is that wind-waves were acting to entrain material from the active underflow and from the previous underflow into the water column. Since water-column advection was minimal, the resulting TSM plume did not appear to extend much beyond the footprint extent (and vice versa).

Comparisons between field measurements and laboratory densities indicated that many field samples had sampling errors and were biased toward lower density. These samples took much longer to obtain than the field density measurements did, and settling may have caused the bias. Results from the field densitometer and pycnometers agreed well on samples tested in the laboratory. Therefore, density results from samples were not included in previous figures or reported below. Table 6.1 summarizes laboratory analyses of particle-size characteristics on fluid-mud point samples. In that table, $H$ is water depth to the original bed, $z_i$ is the distance up from the bottom, and D50 is the median dispersed grain diameter.
Figure 6.12. Interpolated underflow mean concentrations, thicknesses, deposit thickness, and water column depth over the underflow for mud profile stations 16 February 2000.
Figure 6.13. Depth-averaged TSM (mg/l) collected 16 February 2000
Table 6.1
Summary of Fluid-Mud Sediment Characteristics from Near Pipeline Discharge

<table>
<thead>
<tr>
<th>Time, CST</th>
<th>$H_m$</th>
<th>$z_v$</th>
<th>D50, $\mu$m</th>
<th>$% &lt; 4 \mu$m</th>
<th>$% &lt; 16 \mu$m</th>
</tr>
</thead>
<tbody>
<tr>
<td>915</td>
<td>0.9</td>
<td>0.6</td>
<td>4.4</td>
<td>47</td>
<td>84</td>
</tr>
<tr>
<td>915</td>
<td>0.9</td>
<td>0-0.15</td>
<td>5.1</td>
<td>43</td>
<td>79</td>
</tr>
<tr>
<td>933</td>
<td>1.8</td>
<td>0-0.15</td>
<td>5.1</td>
<td>43</td>
<td>78</td>
</tr>
<tr>
<td>1007</td>
<td>1.5</td>
<td>0.15</td>
<td>4.2</td>
<td>49</td>
<td>86</td>
</tr>
<tr>
<td>1007</td>
<td>1.5</td>
<td>0-0.15</td>
<td>4.1</td>
<td>50</td>
<td>85</td>
</tr>
<tr>
<td>1057</td>
<td>1.9</td>
<td>0.3</td>
<td>3.9</td>
<td>50</td>
<td>87</td>
</tr>
<tr>
<td>1057</td>
<td>1.9</td>
<td>0-0.15</td>
<td>4.3</td>
<td>49</td>
<td>84</td>
</tr>
<tr>
<td>1110</td>
<td>1.8</td>
<td>0-0.15</td>
<td>4.4</td>
<td>47</td>
<td>85</td>
</tr>
<tr>
<td>1122</td>
<td>1.1</td>
<td>0-0.15</td>
<td>4.2</td>
<td>49</td>
<td>86</td>
</tr>
<tr>
<td>1136</td>
<td>2</td>
<td>0-0.15</td>
<td>4.1</td>
<td>49</td>
<td>86</td>
</tr>
<tr>
<td>1216</td>
<td>2</td>
<td>0-0.15</td>
<td>4.1</td>
<td>50</td>
<td>86</td>
</tr>
<tr>
<td>1233</td>
<td>1.7</td>
<td>0-0.15</td>
<td>4</td>
<td>50</td>
<td>88</td>
</tr>
<tr>
<td>1308</td>
<td>2.1</td>
<td>0-0.15</td>
<td>4.2</td>
<td>50</td>
<td>86</td>
</tr>
</tbody>
</table>

UNDERFLOW SPREADING PROCESSES

To better understand the spreading of the fluid-mud underflow, a mathematical description of underflow processes was developed. Unfortunately, there are no analytic solutions for the case of a particle-driven gravity flow which is entraining and depositing material, so a numerical solution was developed. Important model features were guided by field observations with special attention to entrainment and settling, either of which can appreciably reduce underflow concentration.

Model Description

A model was constructed to compute total flow or discharge ($Q$), sediment flux ($CQ$), breadth ($B$), and height ($h$) along the length ($x$) of an underflow by numerically integrating a
set of governing equations down-slope in the direction of the underflow. The development of both model equations and assumptions were guided by the field observations. The behavior of mobile fluid-mud layers is an active area of research, and more studies will be needed before such computations can be made with confidence.

The bed was assumed to be planar with an arbitrary slope which was allowed to vary in the down-slope direction. The underflow was considered quasi-steady or steady over a short duration of time. Thus, time derivatives were ignored in the governing equations. However, time of travel to every discrete location along the underflow trajectory was calculated by integrating the underflow velocity. After the first full numerical integration sweep from the transition (beginning) to the end of the underflow, subsequent sweeps were made at discrete time intervals and included updated bed elevations based on the cumulative deposit thickness from preceding sweeps. Sweeps were made at intervals of 1,800 seconds so that per-sweep deposit-thickness changes were small. Thus, the model made many sweeps over the underflow domain and duration of the discharge to update deposit thickness \( \Delta \text{bed} \) and bed slope. Other variables were calculated from the basic state variables and used to solve auxiliary equations for entrainment, bed friction coefficient, depositional flux, and lateral spreading. These variables include underflow concentration \( C \), velocity \( U \), deposit thickness, Reynolds and Richardson numbers.

**Equations**

Governing layer-averaged equations are based on mass and momentum conservation. Governing equations for steady, one-dimensional gravity flow momentum have been developed by Ellison and Turner (1959), Parker et al. (1986), and van Kessel and Kranenburg (1996). Ellison and Turner (1959) ignored \( \Delta \rho / \rho \) terms except when combined with \( g \) where \( g \) = the acceleration of gravity, \( \Delta \rho \) = the density difference between the
underflow and the overlying ambient suspension (i.e. \( \rho - \rho_l \) where \( \rho_l = \text{constant} \)), and \( \rho = \)
the layer density \( (\rho = \rho_l + \Delta \rho) \). This is equivalent to ignoring them in inertia terms of the
equations of motion, and thus the fluid is treated as having uniform mass, variable weight,
and conservative buoyancy. Parker et al. (1986) derived equations explicitly for particle-
driven gravity currents where deposition and erosion can occur and can affect layer
buoyancy. They began with the three-dimensional Navier-Stokes equation, similar to those
presented in Chapter 2. However, they also assume uniform mass density.

Van Kessel and Kranenburg (1996) cite Ellision and Turner (1959) as developers of an
appropriate layer-averaged integral momentum equation but made a key change by keeping
the density term inside the momentum derivative. In this way mass density and mass flux
are not assumed constant in the \( x \)-direction. The following paragraphs derive a form of their
equation for a wide, variable-width flow of rectangular cross section.

The momentum equation was developed considering the forces acting on the boundaries
of an elemental flow volume. Some definitions are given in Figure 6.14. The momentum of
a flow volume with width \( B \) between \( x \) and \( x + dx \) and planes normal to the \( x \)-axis and
parallel to the \( z \)-axis is the mass times its velocity. Let \( \rho \) be density of the flow. Then \( \rho Bu \)
\( dt \) is the mass passing through the upstream plane per unit depth, where \( u \) is the mean flow
across the plane. Since the mass moves at \( u \), the time-averaged momentum is \( \rho Bu^2 \) per unit
depth.

Consider the forces acting on the boundaries of a flow volume. They include the
pressure forces (in excess of the hydrostatic pressure) acting on the down-stream and
upstream planes, reactive shear-stresses acting along the bottom and the top \( (\tau_b + \tau_t = \tau \)
where \( \tau_b = \) bed shear stress, \( \tau_t = \) shear stress at the top of the underflow, and \( \tau \) is the total) ,
and a body-force accelerating the volume down-slope. Then the momentum balance is
Figure 6.14. Definitions for some model variables
\[ \frac{d}{dx} \int \rho Bu^2 \, dz = -B \tau - g \cos \theta \frac{d}{dx} \int \Delta \rho B z \, dz + g \sin \theta \int \Delta \rho B \, dz \]  

where \( \theta \) = the angle of the bed from the horizontal. The left-hand side term of Equation 6.2 is the change in momentum along the slope, and the right-hand side terms are (a) the total shear stress resisting the flow; (b) the derivative of the pressure along the flow resulting from changing depth and density where \( g \cos \theta \) is the gravity component normal to the slope; and (c) the gravity force acting to accelerate the layer where \( g \sin \theta \) is the gravity component parallel to the slope.

Assume \( h(x), \Delta \rho(x), B(x), \) and \( U(x) \) vary along the flow. Define variables for mass and momentum: \( \rho Bu h = \rho Bu \, dz \) and \( \rho Bu^2 h = \rho Bu^2 \, dz \). Shape factors \( S_{1,2} \) are defined during integration of pressure and gravity terms taking into account \( z \)-direction variations in density \( \rho(x) \) such that

\[
\begin{align*}
S_1 \Delta \rho B h^2 / 2 &= \int \Delta \rho B z \, dz \\
S_2 \Delta \rho B h &= \int \Delta \rho B \, dz
\end{align*}
\]  

(6.3a,b)

The excess density is integrated and the hydrostatic pressure does not appear. The vertical integration is done under the assumption that shape factors do not vary in the \( x \)-direction.

The momentum equation is then cast as

\[ \frac{d(\rho Bu^2 h)}{dx} = -B \tau - S_1 g \cos \theta \frac{d(\Delta \rho Bh^2)}{dx} + S_2 g \Delta \rho B h \sin \theta \]  

(6.4)

Van Kessel and Kranenburg (1996) assumed that the shape factors were unity for laminar flow and used a form of Equation 6.4 with \( B = 1 \) to analyze turbulent flows. Parker et al. (1986) evaluated shape factors using experimental data and found the top-hat assumption for
all vertical profiles (including momentum) to be a good approximation. Ellison and Turner (1959) found $S_1 \approx 0.2$ to $0.3$ and $S_2 \approx 0.6$ to $0.9$ in their experiments. Density profiles from Laguna Madre were integrated after exponential interpolation at 1 cm increments between measurement points ($n = 19$ to $47$). Profiles are shown in Figure 6.9 and dimensionless profiles are shown in Figure 6.10. The trapezoidal integration of Equation 6.3 yielded $S_1 = 0.58$ ($n = 7$, 95 percent confidence interval 0.53 to 0.64) and $S_2 \approx 1.0$. The shape factors are hereafter assumed equal to unity and omitted from the equations that follow.

The conservation of sediment flux in the flow is

$$\frac{d(CQ)}{dx} = -BS$$  \hspace{1cm} (6.5)

where $S = \text{the depositional flux}$. To adjust the discharge to include entrainment and deposition, such that deposition does not decrease concentration, the conservation of underflow volume becomes

$$\frac{dQ}{dx} = E_wBU - BS/C$$  \hspace{1cm} (6.6)

where $E_w = \text{the entrainment coefficient}$. The first term on the right side of Equation 6.6 increases underflow volume due to entrainment, and the second term was added to decrease flow due to deposition. This term is several times the deposit volume rate of change so the assumption is implicitly made that some volume leaves the underflow at its upper surface. The underflow is not an irreversibly-mixed solution but rather a settling suspension. By the inclusion of the second right-hand term, Equation 6.6 is different from equations used in other models of particle-laden gravity currents and was developed under the assumption that
the fluid mud layer collapses onto the bed during deposition. Thus, deposition decreases \( Q \) (and \( CQ \)) but does not decrease underflow \( C \) in the model.

Equation 6.4 can be manipulated to give the thickness change along \( x \) as follows:

\[
\frac{dh}{dx} = \left[ 1.43 C_f - Ri \tan \theta + \frac{h}{Q} \frac{dQ}{dx} + \frac{s \cdot h}{Q} \frac{d(CQ)}{dx} + (Ri/2 - 1) \frac{h}{B} \frac{dB}{dx} \right] / (1 - Ri) \quad (6.7)
\]

where \( C_f \) is a drag coefficient, the factor 1.43 relates the bed shear stress to the total of the bed and the top shear stresses (Findikakis and Law 1998), and the bulk Richardson number for the flow is

\[
Ri = \frac{g \Delta \rho \, h \cos \theta}{\rho \, U^2} \quad (6.8)
\]

The excess density \( \Delta \rho \) is related to the underflow concentration \( C \) : \( \Delta \rho = s' C \) where \( s' = (\rho_s - \rho) / \rho_s \) and \( \rho_s \) is the sediment particle density.

**Deposition.** The depositional flux \( S = PW_s C_b \), where the probability of deposition \( P = 1 - \tau_b / \tau_{cd} \) (Krone 1962), \( \tau_{cd} \) is the threshold shear stress for deposition, and \( C_b \) is the concentration at the interface with the deposited bed. Deposition does not occur at bed shear stresses greater than \( \tau_{cd} \). Since the underflow was found to be highly stratified with respect to \( C \), an expression from Teeter (1986) was used to estimate \( C_b \) :

\[
C_b = C \left( 1 + \frac{Pe}{1.25 + 4.75 P_{2.5}} \right) \quad (6.9)
\]

where \( Pe = W_s h / K_z \) is a particle Peclet number, and \( K_z = 0.067(\tau_b / \rho)^{0.5} h \) is taken as the layer-average diffusivity. The valid range for Equation 6.9 is limited, and the value of \( C_b / C_o \) is assumed to have a maximum of 4. Since underflow concentrations are in the hindered range,
Equation 6.7 decreased the depositional flux relative to that calculated with use of the mean concentration.

**Flow regime.** A Reynolds number (R) criteria for the turbulent-laminar transition has been proposed for Bingham plastic materials (Liu and Mei 1990), and found to be applicable to mud flows. \( R \) is composed of viscous (\( R_\mu \)) and yield-stress (\( R_\tau \)) components depending on underflow conditions, and

\[
R = \frac{1}{(1/R_\mu + 1/R_\tau)} , \quad R_\mu = \frac{4 \rho Q}{\mu B} \quad \text{and} \quad R_\tau = \frac{8 \rho Q^2}{\tau_y B^2 h^2} \tag{6.10}
\]

where \( \mu = \) the apparent viscosity, and \( \tau_y = \) the yield stress. Experimental evidence indicates that the turbulent-laminar transition occurs at \( R \)'s of about 2,000 (Liu and Mei 1990; and van Kessel and Kranenburg 1996).

Power law relationships with \( C \) were used to specify \( \mu \) and \( \tau_y \). Van Kessel and Kranenburg (1996) and Teeter (1994 and 2000) present some data for laboratory clays such as kaolinite and natural muds. However, data for the range of \( C \) measured in this study are very scarce. Some data for low-concentration fluid muds from Gulfport Harbor, Mississippi, are presented in Figure 6.15. Those data were developed from shear-stress sweeps, starting below the yield stress, with a Carri-med controlled-stress rheometer and specially collected samples (Teeter 1993a). A concentric cylinder geometry was used, and data fit to a Herschel-Bulkley shear stress model (Coussot 1994) and a Sisko viscosity model (Barnes et al. 1989).

The yield stresses and viscosities at 50 sec\(^{-1}\) shear rate, plotted in Figure 6.15, are higher for the overlapping range of \( C \) than values found for commonly-used laboratory clays and other natural muds. Measurements of yield stress were obtained for three channel sediment
Figure 6.15. Viscosity at 50 sec$^{-1}$ shear rate and yield stress for fluid mud from Gulfport Harbor, Mississippi
samples from the GIWW about 6 km north of Port Isabel. The three samples were from the top 7.5-cm of sediment cores and all had densities of about 400 kg/m³. Yield stresses for these samples, and for other natural muds including Gulfport, are shown in Figure 6.16. These data indicate that yield stress values of Laguna Madre mud are not as great as for some other natural muds.

For turbulent conditions, \( C_f = \frac{g n^2}{h^4} \) where \( n \) is Mannings’ friction factor. For laminar conditions, the drag coefficient was estimated as the maximum of the turbulent case or the expression presented by van Kessel and Kranenburg (1996) for laminar flows:

\[
C_f = \frac{12}{R_\mu} + \frac{\tau_y}{\rho U^2} \tag{6.11}
\]

**Entrainment.** In addition to deposition, entrainment of overlying water can reduce \( C \). A theoretical and laboratory investigation of entrainment was presented by Kranenburg and Winterwerp (1997) and the resulting relationship between entrainment coefficient \( E_w \) and \( Ri \) tuned for fluid mud by van Kessel and Kranenburg (1996)

\[
E_w = \frac{5.5 \times 10^{-3}}{3.6 Ri - 1 + \sqrt{(3.6 Ri - 1)^2 + 0.15}} , \quad R > 2,000 \tag{6.12}
\]

where \( Ri \) is defined in Equation 6.9 and the condition implies that turbulent flow is required for entrainment to proceed.

**Lateral spreading.** A buoyant-surface spreading rate developed by Shirazi and Davis (1974) was adopted to the case of negatively-buoyant spreading along the bottom, yielding the expression:
Figure 6.16. Yield stress for Laguna Madre GIWW sediments and other muds
\[ \frac{dB}{dx} = 1.4 \left( \frac{c_r B}{h Ri} - 1 \right)^{-1.2} \]  

(6.13)

where \( c_r \) = a coefficient inversely related to the concentration difference. An upper limit of 2.0 was set on \( dB/dx \) so that lateral spreading rate could not be larger than the down-slope advance rate.

**Initial conditions.** A flow transition formed the initial condition for the underflow observed on 16 February. A horizontal discharge pipe had a diameter equal to an appreciable fraction of the receiving water depth. A turbulent surface flow was created, which, after some initial entrainment and spreading, plunged to form an underflow in slightly deeper water as shown schematically in Figure 6.17 and previously in the Figure 6.8 photograph. The critical Richardson number \( Ri \) for a plunging underflow described earlier is a little greater than 1 (Fang and Stefan 2000).

In addition to the pipeline discharge and discharge concentration \( (Q_i \text{ and } C_i) \), the model requires specification of the initial dilution \( S_o \), initial breadth \( B_o \), and initial \( Ri \) which were estimated from field information. The initial height \( h_o \) of the underflow was then estimated by

\[ h_o = \left( \frac{Q_o}{B_o} \right)^{2/3} \left( \frac{\rho_o \rho_f}{g \Delta \rho_o} \right)^{1/3} \]  

(6.14)

where the subscript \( o \) refers to values at the underflow transition, \( Q_o = S_o Q_i \), and \( C_o = C_i/S_o \).

**Model Results**

Governing and auxiliary equations were used to integrate underflow variables, starting with the initial conditions described in the last section; a fourth-order numerical scheme was used to calculate \( Q, CQ, B, \text{ and } h \) at 1 m intervals along the down-slope trajectory of the
Figure 6.17. Schematic of a shallow horizontal pipeline discharge, turbulent surface flow, and transition to an underflow at a plunge point.
underflow. Depths from the 1 m contour toward the east-southeast were estimated from sample station and chart depths and interpolated out to a distance of 475 m. Coefficients $c_i$ and Manning’s $n$ were taken as 0.033 and 0.02, and model results were insensitive to the exact values used.

Example computed profiles of underflow and deposit heights are shown in Figure 6.18 for 1 and 6 hours after the discharge began. These results are for $B_o = 45$ m, $S_o = 5.5$, initial $R_i = 1.5$, $Q_i = 0.5$ m$^3$/sec, $C_i = 100$ dry-kg/m$^3$, $\tau_{cd} = 0.05$ Pa, and the settling properties described earlier. The rheological properties were set between those of kaolinite and Gulfport Harbor fluid muds. The down-slope extent increased between hours 1 and 6 as the deposit developed and reduced slope between 50 and 150 m $x$ distances. Between hours 1 and 6, breadth-weighted average underflow thickness over the footprint decreased from 0.78 to 0.61 m, the deposit increased from 0.05 to 0.28 m, and the footprint area increased from 19,800 to 24,300 m$^2$.

The underflow extent in plan-view is shown in Figure 6.19 for hour 6. The most rapid spreading occurred as $R_i$ increased. Spreading abruptly stops where the deposit greatly increases bed slope. The variation of time, velocity, and sediment discharge along $x$ are shown in Figure 6.20. The underflow velocity decreased rapidly near the origin, then more slowly over the remainder of the $x$ extent. Deposition starts when $U < 0.10$ m/sec.

Underflow concentrations were 18 dry-kg/m$^3$ at the origin and decreased slightly to about 17 dry-kg/m$^3$ at the maximum down-slope extent of the underflow.

**ENTRAINMENT INTO THE WATER COLUMN**

Entrainment of material from the underflow into the water column, the reverse of that entrainment described earlier, was implied by observed plumes of suspended sediments associated with the location of the underflow footprint. In a deeper estuarine field situation,
Figure 6.18. Computed underflow and deposit profiles at two times.
Figure 6.19. Plan view of computed underflow footprint showing spreading with distance along the x-axis.
Figure 6.20. Computed variations of time to reach $x$ distance, underflow velocity, and sediment discharge along the $x$-axis.
no such plume was observed by Thevenot et al. (1992). Though currents were weak at the Laguna Madre site, winds were relatively strong. The entrainment process depends on the local momentum balance, turbulence at the underflow interface, and the magnitude of density differences. Data collected on 16 February at Laguna Madre were used to make an evaluation of the entrainment process in this section.

Details of the underflow interface with the water column are important to this entrainment process. Either the density step between the water column and the underflow, or, if no such step exists, the gradient at the top of the underflow is used to scale a Richardson number (Turner 1986). As discussed earlier, the underflow was observed to be highly stratified but with a sharp density jump between its upper surface and the overlying water column. Therefore, the magnitude of the jump, along with the depth of the water column above the underflow \( (H_o) \), was used to scale a Richardson number. The magnitude of that jump was much smaller than the overall mean density difference between the two layers.

The major momentum input was from the wind. Wind data from a station at Rincon maintained by Texas A&M University, Conrad Blucher Institute, indicated that mean wind speeds at 10-m height \( (U_a) \) were 12.0 m/sec for 1000 to 1200 CST on 10 February, and were 7.6 m/sec for 0900 to 1300 CST on 16 February (standard deviations for both time periods were about 1 m/sec). An approximate location for Rincon is given in Figure 6.1.

Hydrodynamic forcing was assumed to equal the wind stress, some of which goes into waves and some of which goes into currents. Wind-waves in Laguna Madre tend to be at a fully-developed, depth-limited state such that dissipation is nearly equal to the momentum input (see Chapter 5). Aalderink et al. (1985) compared two models which used wind stress directly with two models which used near-bed wave orbital shear stress, and found that the
models which used wind stress directly were better able to match observed TSM. The in-water friction velocity was estimated from

\[ u_* = \left( \frac{D_a C_d}{D_o} \right)^{1/2} U_a \]  

(6.15)

where \( \rho_a \) and \( \rho_o \) are the atmospheric and water column densities and \( C_d \) is the shallow-water atmospheric drag coefficient taken as a function of depth and wind speed (Teeter et al. 2001). The average value for \( u_* \) on 16 February was about 0.0095 m/sec.

At high interfacial Richardson numbers \( (R_i) \), dimensionless entrainment \( (E) \) is the result of perturbations in the interface between the turbulent water column and the underflow. (Also assuming that the molecular Peclet number \( = u_t l_t/v \) is greater than 200 where \( u_t \) and \( l_t \) are the turbulent velocity and length scales, and \( v \) is molecular diffusivity.) Under conditions of turbulence without mean-flow, the laboratory experiments of Long (1975), and E and Hopfinger (1986) confirmed the -3/2 power law described by Linden (1973) that

\[ E = \frac{u_*}{u_*} = K R_{i_*}^{-3/2} \]  

(6.16)

where \( u_* \) is the entrainment velocity or the downward velocity of the interface, \( K \) is a constant, and the interfacial Richardson number is defined slightly differently from \( R_i \) as

\[ R_{i_*} = \frac{g \Delta \rho H_o}{\bar{\rho} u_*^2} \]  

(6.17)

where the density step across the interface \( \Delta \rho = \rho - \bar{\rho} \), \( \bar{\rho} \) is the average density of the layers, and \( H_o \) is the depth of the water column above the underflow. The scales for \( \Delta \rho \) and
length can be chosen differently in different entrainment systems. Here, although the underflow is stratified, the mechanism causing that stratification involves settling and not diffusion across an interface. Thicknesses of density interfaces are typically about 6 percent of the depth of mixed layers, much thinner than the stratified underflow layers observed here. Values of $Ri_s$ are large, and interfacial perturbations are probably intermittent, consisting of vortex rebounding. Thus, $\Delta \rho$ and $H_o$ were scaled by the overall density step and the depth of the water column.

Entrainment and deposition to the underflow by settling are assumed to be simultaneous processes in this case. Teeter (1994) reviewed laboratory entrainment experiments involving suspensions and found them to be consistent with an assumption of simultaneous entrainment and settling. Thus at a depth-averaged water-column point over an underflow

$$H_o \frac{dC_e}{dt} = F_e - F_s$$

(6.18)

where $C_e$ is the depth-averaged suspension concentration in the water column, $t$ is time, and $F_e$ and $F_s$ are the entrainment and settling flux rates at the interface.

A further simplification can be made by assuming that $C_e$ is constant. This assumption is justified since the water-column depth is small, and the time for settling or turbulent mixing is short compared to the time-scales for underflow spread and/or wind speed changes. Under equilibrium conditions of settling and entrainment, $F_e$ and $F_s$ have the same magnitude. Furthermore, the settling flux for the water suspension can be estimated by use of laboratory settling tests on Laguna Madre channel sediments described by Teeter et al. (2001a) and shown in Figure 6.3. These results were obtained by mixing suspensions with site water and allowing them to settle in a 10-cm diameter by 1.9-m-tall column under quiescent conditions.
Nine initial concentrations were tested, and settling velocity \(W_s\) was found to increase linearly with initial concentration. A function describing concentration-dependent settling rate for Laguna Madre channel sediments is

\[ V_s = a_1 C_o^n, \quad 0.1 < C_o < 1 \text{ kg/m} \]  \hspace{1cm} (6.19)

where \(W_s\) is in m/sec, \(a_1 = 0.806 \times 10^{-3}\) m/sec, and the exponent \(n = 1\).

Settling flux depends on depositional probability, as did the depositional flux \(S\) described in the last section. Here, turbulence at the interface is assumed to be low and intermittent, and the depositional probability is assumed to be unity. The depositional flux depends on the near bed concentration \(C_b\), but in this case it is assumed that \(C_b/C_o = 1\). Therefore, \(F_e = u_e C\) and \(F_s = W_s C_o\); a simple model for the water column suspension concentration is

\[ C_o = \left( \frac{K}{a_1} C_u R_i^{3/2} \right)^{\frac{1}{n+1}} \]  \hspace{1cm} (6.20)

Equation 6.20 was recast to solve for the entrainment coefficient \(K\) with field data. The Laguna Madre measurements and Equation 6.7 indicated that \(C_b/C_o = 1.07\), and column \(C_o\) values were adjusted accordingly. In-water friction-velocities were calculated using Equation 6.19 and a constant wind speed of 7.6 m/sec. Underflow concentration at the interface was assumed to be 3 dry-kg/m³. Results for \(K\) and other select parameters are presented in Table 6.2.

The flux Richardson number \((Ri_f)\) was calculated for Table 6.2 as \((u_e/u_*)Ri_f\) (Turner 1986) and represents the ratio of turbulent kinetic energy dissipation by buoyancy flux and turbulent production. Flux Richardson numbers greater than 0.1 are associated with damping of
Table 6.2
Entrainment Conditions at Underflow Profile Stations

<table>
<thead>
<tr>
<th>Time, CST</th>
<th>TSM, mg/l</th>
<th>$H_o$, m</th>
<th>$Ri_*$</th>
<th>$F_r \times 10^5$, kg/m$^2$/sec</th>
<th>$K$</th>
<th>$Ri_f$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1057</td>
<td>168</td>
<td>0.86</td>
<td>86</td>
<td>2.6</td>
<td>0.7</td>
<td>0.1</td>
</tr>
<tr>
<td>1110</td>
<td>240</td>
<td>0.26</td>
<td>133</td>
<td>5.3</td>
<td>2.8</td>
<td>0.2</td>
</tr>
<tr>
<td>1122</td>
<td>308</td>
<td>0.41</td>
<td>69</td>
<td>8.8</td>
<td>1.8</td>
<td>0.2</td>
</tr>
<tr>
<td>1136</td>
<td>228</td>
<td>0.37</td>
<td>128</td>
<td>4.8</td>
<td>2.4</td>
<td>0.2</td>
</tr>
<tr>
<td>1216</td>
<td>480</td>
<td>0.29</td>
<td>134</td>
<td>21.3</td>
<td>11.4</td>
<td>1</td>
</tr>
<tr>
<td>1233</td>
<td>708</td>
<td>0.2</td>
<td>124</td>
<td>46.3</td>
<td>22.1</td>
<td>2</td>
</tr>
<tr>
<td>1308</td>
<td>260</td>
<td>0.25</td>
<td>151</td>
<td>6.2</td>
<td>4</td>
<td>0.3</td>
</tr>
</tbody>
</table>

Turbulence, and the magnitudes of $Ri_f$ obtained here indicate appreciable suppression of turbulence at the interface.

The $K$ (and TSM) values were log-normally distributed. The median $K$ value was 2.8 in fair agreement with the laboratory result of 3.8 reported by E and Hopfinger (1986). Based on this estimate for $K$, an estimate of $C_o$ was made for 10 February with Equation 6.20. Assuming underflow conditions were the same as before, where $\Delta g/\bar{\rho} = 0.00081$, and that $H_o = 2$ m and $u_*$ = 0.0174 m/sec, then the $Ri_*$ for that day was about 52. Equation 6.20 predicts $C_o = 690$ mg/l or somewhat higher than the high values observed in the field (Figure 6.5).

**DISCUSSION AND CONCLUSIONS**

Some of the important results from the field measurements were as follows: the underflow thickness was relatively uniform and decreased rapidly near the limit of down-slope extent. The underflow had a distinct upper surface or interface with the ambient water column, but this concentration was only about 3 dry-kg/m$^3$. Underflow layers were
sediment and density stratified. Turbulence in the underflow was not sufficient to mix sediment vertically. Thick deposits formed under the underflow. The implication of these observations was that the underflow was slow moving. An apparent absence of appreciable entrainment also indicated slow underflow movement and high Richardson numbers. The concentrations near this upper surface and layer average values were relatively uniform along the length of the underflow. This observation indicated the underflow collapsed vertically while depositing and led to the model feature that tended to maintain underflow concentrations while deposition occurred.

The results from the model were in general agreement with features observed in the field. The model predicted the underflow to be slow moving, starting turbulent and becoming laminar after 85 to 150 m distance down-slope. Calculated underflow concentrations were almost constant. Deposition started near the source, became appreciable after 100 m, and decreased rapidly at the down-slope extent of the underflow. Underflow thicknesses in the model were somewhat greater than those observed in the field. The extent of the underflow could not be accurately measured in the field but appeared to be in rough agreement with the model predictions. Lateral spreading, which was controlled by bed topography, was captured only in a very approximate way in the model. The model did not include path switching that might have occurred in the field nor the complexity of bottom topography that influences underflow spreading.

Laboratory-determined settling velocities were used in the description of underflow spreading and in the analysis of water column entrainment of underflow material. Laboratory measured values are probably not the same as field values, but obtaining measurements in the field is problematic. Wolanski et al. (1992) showed that very low levels of turbulence decrease $W_s$ in the hindered settling concentration range by factors of 2 to 10. Teeter (2001a)
showed that for low-concentration Laguna Madre suspended sediments, quiescent column tests yielded settling rates representative of disrupted flocs. Very mild turbulence produced much larger values, but shear rates greater than 2 sec\(^{-1}\) produced settling rates not much different from quiescent values. The uncertainty in \(W_s\) affects the results of both the underflow and water column analyses, if performed in a predictive mode.

Our understanding of fluid-mud flow properties is incomplete and measurements are difficult. The existence of a yield stress may lead to an unsheared plug flow zone (Coussot 1994) in the underflow and could make \(R\) values based on layer average properties unrepresentative of interface conditions where entrainment occurs. In the present case, the underflow layer was highly stratified, and the velocity profile was difficult to evaluate. The flow-regime evaluation imposed on Equation 6.12 only captures the effect of fluid-mud flow properties on entrainment, if such properties are known or can be estimated.

Rheological data on muds are relatively scarce. While varying rheological parameters over an order of magnitude from those used in the example calculation did not greatly affect results, increases in viscous and yield stress properties eventually caused the model underflow to freeze and become numerically unstable. The range from laboratory clays to cohesive natural muds such as those in Figure 6.15-16 affect computed underflow \(R\), friction, flow, and ultimate extent of spreading. The predictive capability of the model is therefore dependent on rather extensive site-specific field information. Important factors include sediment composition, settling and rheological characteristics, bed topography, ambient currents, winds, and waves.

Elevated water-column suspended sediment concentrations were caused by underflow entrainment into the water column by wind-wave forcing. Entrainment model coefficients
were consistent with previously reported values for high $Ri_s$ situations when used with wind-stress forcing.

The pipeline discharge underflow represents the greatest potential for local turbidity generation, if it is entrained into the overlying flow, since it contains the vast majority of sediment particles discharged. Field observations indicate that at times of high bed shear-stress, entrainment of underflow material can generate a turbid plume extending some distance from the discharge, but not necessarily downstream from the discharge. Thus, the area of concern with respect to water column impacts of a pipeline discharge is not confined to the vicinity of discharge, but also includes the area over the underflow that might extend hundreds of meters.
Chapter 7
Conclusions

In two ultra-shallow systems (< 2-m deep) with submersed vegetation, Florida Bay and Laguna Madre, suspended sediment concentrations varied spatially due to bed sediment conditions, imposed shear stresses (related to depth), and the presence of seagrass (related to depth); and temporally due to winds. In windy Laguna Madre, TSM were about a factor of 10 greater at an unvegetated site than at a seagrass site. Typical TSM concentrations in seagrass beds are 15 mg/l for these systems (Chapter 2). Suspended sediment concentrations responded rapidly to wind-speed changes in and were approximately log-normally distributed in time at a point. Under high-wind conditions in Laguna Madre, monitoring data showed TSM levels approached 500 mg/l at an unvegetated site, and were found to approach 1,000 mg/l in the proximity of a pipeline discharging dredged material (Chapter 6).

The previously-identified feedback between resuspension, water clarity, and seagrass was found to contain several physical feedbacks affecting sediment transport. These feedbacks included seagrass to wave (and current), wave to wind shear stress, and seagrass to bed shear stress. Seagrass friction was found to limit the growth of waves and the transfer of momentum from the atmosphere to the water surface (Chapter 5). Wave growth was limited in seagrass areas by decreased effective-depth/wave-length and by high frictional resistance presented at the top of plant canopies. Wave characteristics affect atmospheric friction factor. Seagrass frictional resistance to flow is also very high. The magnitudes of these feedbacks were investigated and found to be important to sediment transport conditions.

When typical hydrodynamic and sediment models are applied to seagrass areas, they require enhancement to accurately model flows and shear stresses. Hydrodynamic friction
should take plant deflection by currents into account and thus should be characterized using additional parameters: un-deflected canopy height, flexural rigidity of plants, and an additional friction coefficient. Previous current friction formulations for tall, flexible artificial roughness elements fit laboratory flume data for seagrasses well (Chapter 2). Since canopy height varies with flow, hydrodynamic friction can not be characterized by a single parameter friction law, as is customary for plane beds. Using experimental data from a 24-cm deep flume, Manning’s $n$ friction factors were estimated to vary from 0.135 to 0.112 for current speeds of 3 to 20 cm/sec and 1-m water depth. The effect of plant density on friction has not been determined.

Wave friction roughness heights, never before estimated from field data for seagrass areas, were found to be about 0.2 m or roughly equivalent to the canopy height for a dense *Thalassia testudinum* seagrass bed in 1.1-m water depth. The magnitude of the roughness height relative to wave-orbital excursion length indicates that the turbulent-rough is the most appropriate friction formulation.

The shear stress acting on the sediment bed is appreciably reduced by the presence of seagrass and this sheltering effect must be parameterized in sediment models. A seagrass bed-sheltering factor was developed based on previous flume data, which indicated that 12 to 66 percent of the total shear stress was transmitted to the bed surface, depending on seagrass species. The magnitude of the bed sheltering effect was inversely related to the frictional characteristics for four seagrass species, as determined in previous studies (Chapter 2). The remainder of the shear-stress is absorbed by plants and transmitted through stems and roots. When these results were used in a numerical sediment resuspension model of Laguna Madre, realistic suspended-sediment concentration results were obtained.
The atmospheric roughness-height equation that produced the best fits to data from Laguna Madre was similar to previously published equations that included significant wave height and wave age (wave celerity divided by atmospheric friction velocity), but was found to have a smaller exponent on the inverse wave-age term. The difference was attributed to the dependence of wave age on depth and wind speed in ultra-shallow water. A new formulation for the atmospheric friction factor in terms of depth and wind speed was proposed and compared well with field data. This allowed accurate calculation of atmospheric shear stress over space and time for use in sediment transport calculations.

Wave conditions varied markedly over Laguna Madre, depending on water depth and seagrass conditions. Waves tended to be depth-limited for observed fetches (greater than about 3 km). Wave-limiting fetch lengths are apparently shorter in ultra-shallow water than for slightly deeper water, possibly caused by increased wave dissipation through friction. Wave energy and frequency were scaled by the atmospheric friction velocity to better fit data to new analytical models for significant wave height and peak period (Chapter 5). It was proposed that canopy height be subtracted from total depth when estimating atmospheric shear stress and wave characteristics in seagrass areas.

The fractions of atmospheric shear-stress going to waves and currents were estimated from data and a turbulent rough wave friction formulation. A new partitioning formula was proposed, dependent on wind speed. The fraction of atmospheric shear stress going to waves decreased with wind speeds greater than about 5 m/sec. The relationship predicts values that are in general agreement with previous estimates.

The application of numerical sediment models for predicting resuspension and transport requires considerable site-specific information. The availability of light affects plants over long time scales. Since the feedback between water clarity and seagrass acts slowly, the
The objective of sediment transport modeling has been to accurately simulate most conditions that occur. Other types of sediment transport studies, on the other hand, often concentrate on extreme events that rarely occur but which involve most of the transport. Like water quality and ecological models, sediment resuspension models applied to seagrass studies must operate over long simulation periods. In the Florida Bay and Laguna Madre hydrodynamic and sediment transport model applications, results from one- and two-year simulation periods were used to assess seagrass growth (Chapter 2).

Settling velocity of fine, cohesive sediment depends on suspension concentration and fluid shear rate. The combined effects of concentration and fluid shear-rate were found not to be multiplicative factors controlling settling velocity as previously proposed. An alternate formulation was proposed based on laboratory experiments (Chapter 3). Under optimum shear-rate conditions, the floc settling rates for Laguna Madre sediments did not change significantly with concentrations ranging from 60 to 660 mg/l.

Sediment transport model formulations were classified and compared (Chapters 2-4). Model formulations with simultaneous erosion and deposition (type I), previously justified on the grounds that they produced results consistent with field data, predict suspended sediment concentration no better than, and probably not as well as, a multiple grain-size model. When models used mutually-exclusive erosion and deposition, a multiple grain-size class model (type III) was demonstrated to predict suspended sediment fluctuations during wind events more accurately than a single grain-size model (type II). The preponderance of experimental evidence suggests that fine-grained (silt and clays less than 62 μm) erosion and deposition are mutually-exclusive processes, suggesting that type-I formulations should be used with caution. However, since type-I formulations are used by the scientific and
engineering communities, the appropriateness of the simultaneous erosion and deposition assumption remains a debated issue.

A multiple grain-size class numerical sediment model was based on a previous analytic grain/floc model found capable of simulating observed patterns of grain size distribution during deposition. Model statistical trends for distributions with fine-tail slopes ranging from 0 to 1 units followed previously observed trends of decreasing mean size, decreasing sorting, and increasing skewness during deposition. The multi-grain-size model was able to produce steady-state suspensions similar to those observed in laboratory experiments.

The multiple grain-size class model with mutually exclusive erosion and deposition produced reasonable suspended sediment concentrations when applied to Laguna Madre. Variation in the assumed-uniform input wind field was suspected of causing some of the deviation between model and observed time-series suspended sediment concentrations. Statistical distributions of suspended sediment concentrations compared favorably to long-term twice-daily TSM monitoring data. A previously developed single grain-size model formulation with simultaneous erosion and deposition, modified for seagrass application, was coded into a water quality model and used to simulate wind-wave resuspension for Florida Bay. The model results appeared to be reasonable, but detailed suspended sediment data were lacking for the Florida Bay.

Dredged material is a substantial sediment volume relative to natural sediment inputs to Laguna Madre. Assessment of dredged material resuspension required information on sediment erodibility and deposit area for certain discharge sites. Laboratory experiments were performed on slurried channel sediment allowed to settle for varying times to determine dredged material erodibility (Chapter 2). The area of the initial deposit was estimated based on field information. Since field information on such discharges is normally limited, a near-
field model was developed for a particle-driven gravity underflow to establish a predictive capability for dredged material deposit area. The model conserves sediment and fluid mass and momentum, and includes the effects of deposition, entrainment, and lateral spreading. However, the model needs more extensive validation to field data to be useful in the predictive sense.

At a dredged-material pipeline discharge, a turbulent surface flow formed adjacent to the discharge from 0.5 to 1.0 m deep, then plunged to form a slow-moving fluid mud underflow. Underflow thicknesses were about 0.3 m and relatively uniform to the down-slope edge. Underflow deposit thicknesses were about 0.2 m. Underflow concentrations were vertically stratified but did not change much with distance from the discharge point, suggesting that the underflow collapsed as it deposited. A special term was added to the flow continuity equation in an underflow model to reproduce this feature. Average underflow concentrations were 10 to 20 kg/m$^3$.

The underflow concentration profiles were all exponential with depth and similar when normalized. The profile shapes indicate some degree of vertical mixing, consistent with a turbulent flow. Underflow surface concentrations were only about 3 kg/m$^3$. The observed range of viscous and yield-stress characteristics for fine, cohesive fluid-mud material varies greatly, and could also explain the relatively constant underflow concentrations observed in the field. Water column plumes of suspended sediment were observed over the underflow footprint, and attributed to wind-wave forced entrainment of the underflow (Chapter 6).
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Appendix - Letters of Permission

Re: Hydrobiologia (2001), 444/1-3, p. 1-23

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Vita

Allen Michael Teeter was born to Elaine and Tuck Teeter on 1 July 1947. He grew up with two brothers in Minneapolis, Minnesota. He went off to the Merchant Marine Academy, in New York, to satisfy his yearning for sailing and far-away places. After several very long freighter trips to Vietnam, graduation from the academy in 1970, and the collapse of the shipping industry toward the end of the war; he was ready to settle down on land. He married Cynthia L. Duval in Minneapolis.

He began work for the U.S. Environmental Protection Agency in Minneapolis and Duluth, Minnesota, during 1971 and early-1972, servicing instrumentation, sampling, and operating small survey vessels. He was transferred to the Corvallis Environmental Research Laboratory in Oregon, and performed field and modeling work until 1979 as a member of the Marine Division. He did some part-time sailing in southern California with his brother John who had moved to Eugene, Oregon. He attended Oregon State University and received an M.S. degree in 1978.

In 1979, he transferred to the U.S. Army Engineers Waterways Experiment Station (WES), Hydraulics Laboratory, Estuaries Division in Vicksburg, Mississippi, and is now a member of the Tidal Hydraulics Branch, Coastal and Hydraulics Laboratory at WES.

He had two girl children - Natalie and Cameron. Natalie died in 1981. He was divorced in 1994. Cameron married in 1999, and now works for the University of Georgia, Athens, as a landscape architect.

He now lives near Vicksburg with Nancy C. Flowers - reported to be a Gypsy goddess.