1996

Modeling storm-induced sediment transport on the inner shelf: Effects of bed microstratigraphy

Baeck Oon Kim
College of William and Mary - Virginia Institute of Marine Science

Follow this and additional works at: https://scholarworks.wm.edu/etd
Part of the Geology Commons, and the Oceanography Commons

Recommended Citation
https://dx.doi.org/doi:10.25773/v5-gzcd-n430

This Dissertation is brought to you for free and open access by the Theses, Dissertations, & Master Projects at W&M ScholarWorks. It has been accepted for inclusion in Dissertations, Theses, and Masters Projects by an authorized administrator of W&M ScholarWorks. For more information, please contact scholarworks@wm.edu.
INFORMATION TO USERS

This manuscript has been reproduced from the microfilm master. UMI films the text directly from the original or copy submitted. Thus, some thesis and dissertation copies are in typewriter face, while others may be from any type of computer printer.

The quality of this reproduction is dependent upon the quality of the copy submitted. Broken or indistinct print, colored or poor quality illustrations and photographs, print bleedthrough, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send UMI a complete manuscript and there are missing pages, these will be noted. Also, if unauthorized copyright material had to be removed, a note will indicate the deletion.

Oversize materials (e.g., maps, drawings, charts) are reproduced by sectioning the original, beginning at the upper left-hand corner and continuing from left to right in equal sections with small overlaps. Each original is also photographed in one exposure and is included in reduced form at the back of the book.

Photographs included in the original manuscript have been reproduced xerographically in this copy. Higher quality 6" x 9" black and white photographic prints are available for any photographs or illustrations appearing in this copy for an additional charge. Contact UMI directly to order.

UMI
A Bell & Howell Information Company
300 North Zeeb Road, Ann Arbor MI 48106-1346 USA
313/761-4700 800/521-0600

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
MODELING STORM-INDUCED SEDIMENT TRANSPORT
ON THE INNER SHELF: EFFECTS OF BED MICROSTRATIGRAPHY

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

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

by
Baeck Oon Kim
1996
APPROVAL SHEET

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

Doctor of Philosophy

Baeck Oon Kim

Approved, June 1996

L. Donelson Wright, Ph.D.
Committee Chairman/Advisor

John D. Boon, III, Ph.D.

Jerome P.-Y. Maa, Ph.D.

Linda C. Schaffner, Ph.D.

Ole S. Madsen, Sc.D.
Massachusetts Institute of Technology
Cambridge, Massachusetts

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
### TABLE OF CONTENTS

ACKNOWLEDGMENTS .................................................................................................................. v

LIST OF TABLES ................................................................................................................................vi

LIST OF FIGURES ............................................................................................................................vii

LIST OF SYMBOLS ........................................................................................................................ x

ABSTRACT ........................................................................................................................................xvii

1. INTRODUCTION ....................................................................................................................... 2

2. METHODS AND DATA ANALYSIS ...................................................................................... 6
   2-1. Introduction .................................................................................................................. 6
   2-2. Field Experiments ......................................................................................................... 6
   2-3. Background of Field Site .............................................................................................. 8
   2-4. Data Analysis ..................................................................................................................12
      2-4-1. Wind ............................................................................................................. 12
      2-4-2. Wave ............................................................................................................. 15
      2-4-3. Current ......................................................................................................... 18
      2-4-4. Suspended Sediment .......................................................................................21
      2-4-5. Grain Size .......................................................................................................23
   2-5. Correction of Measurement Heights ...............................................................................26

3. BOUNDARY LAYER PROCESSES COUPLED WITH BED STRATIGRAPHY ............. 33
   3-1. Introduction ................................................................................................................ 33
   3-2. Wave-current Boundary Layer Model .......................................................................... 35
   3-3. Bottom Roughness ....................................................................................................... 41

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
ACKNOWLEDGMENTS

I would like to make a grateful acknowledgment for Dr. L. Donelson Wright who guided this work and provided support, insight and understanding with great patience. I would like to thank my committee members, John Boon, Jerome Maa, Linda Schaffner, and Ole Madsen. They provided valuable comments and helped me to make this thesis more readable. I am also grateful to Dan Hepworth, Frank Farmer, Todd Nelson, and Bob Gammish for helping with data acquisition. Thanks to Cynthia Harris and Beth Marshall for their hospitality in helping paperwork. Thanks to Sungchan Kim and Jingping Xu for many valuable conversations. Special thanks to Tom Chisholm for English corrections. I am thankful that many people of the Virginia Institute of Marine Science including my fellow students helped my education and daily life.

I would like to thank Pat Wiberg of the University of Virginia for providing valuable discussion. The Field Research Facility of the U.S. Army Corps of Engineer provided wind data. Permission to use these data is appreciated. This research was supported by the National Science Foundation, Grant No. OCE-9123513.

I would like to thank my parents and brothers for their love, support, and encouragement. Finally, I would like to thank my wife, Jihyun, for her constant love and great patience through five years.
LIST OF TABLES

2-1. Summary of instrumentation ......................................................... 9
2-2. Summary of measurement height .................................................. 29
2-3. Initial measurement heights measured above bed ................................ 30
3-1. $f_1(\delta_m)$ versus $f_1(\eta_1)$ ......................................................... 62
4-1. Comparison of the JMGM and GM model predictions with field measurements ........... 124
LIST OF FIGURES

2-1. Map showing study area ........................................................................................................7
2-2. Three-dimensional representation of bathymetry .................................................................11
2-3. Grain size distribution and cumulative curve of bottom sediment ................................. 13
2-4. Vertical distribution of sand composition in a sediment core ............................................. 14
2-5. Time series of the FRF wind data .......................................................................................... 16
2-6. Time series of wave characteristics .................................................................................... 19
2-7. Time series of current velocity ............................................................................................ 20
2-8. Time series of OBS data .................................................................................................... 22
2-9. Input of bed stratigraphy .................................................................................................... 25
2-10. Time series of DSA data .................................................................................................. 28
2-11. Time series of bed elevation changes .................................................................................. 32

3-1. Schematic representation of vertical one-dimensional model .............................................. 36
3-2. Flow chart of vertical one-dimensional model .................................................................... 67
3-3. Bed stratigraphy formation in vertical one-dimensional model ...................................... 68
3-4. Comparison of predicted concentrations \( (\gamma_0 = 0.001) \) with measurements, at 20 m ......72
3-5. Comparison of predicted concentrations \( (\gamma_0 = 0.001) \) with measurements, at 12 m ......73
3-6. Predicted ripple heights ..................................................................................................... 76
3-7. Comparison of predicted \( u_c \) and \( z_{ca} \) with measurements at 20 m depth .................... 79
3-8. Comparison of predicted \( u_c \) and \( z_{ca} \) with measurements at 12 m depth ....................... 80
3-9. Predicted concentrations using coupled and decoupled models ($\gamma_0 = 0.001$) .......... 83
3-10. Predicted concentrations using coupled and decoupled models ($\gamma_0 = 0.0004$) ........ 84
3-11. Predicted concentrations using coupled and decoupled models ($\gamma_0 = 0.0002$) .......... 85
3-12. Predicted concentrations using coupled and decoupled models ($\gamma_0 = 0.0003$) .......... 86
3-13. Predicted concentration profiles using coupled and decoupled models ....................... 87
3-14. Predicted $u_c$ and $z_s$ using coupled and decoupled models ........................................ 89
3-15. Predicted erosion depth, total depth and sediment size at 20 m depth ......................... 91
3-16. Predicted erosion depth, total depth and sediment size at 12 m depth ......................... 92
3-17. Predicted erosion depth using coupled and decoupled models ($\gamma_0 = 0.001$) .......... 93
3-18. Predicted erosion depth using coupled and decoupled models ($\gamma_0 = 0.0003$) .......... 94
3-19. Predicted bed stratigraphy at 20 m depth ............................................................... 96
3-20. Predicted grain size distributions of individual layers at 20 m depth ......................... 97
3-21. Predicted bed stratigraphy at 12 m depth ............................................................... 98
3-22. Predicted grain size distributions of individual layers at 12 m depth ......................... 99

4-1. Flow chart of horizontal, one-dimensional model .................................................. 105
4-2. The FRF wave height measured at 8 m and 17 m depths .......................................... 110
4-3. Predicted wave height with no energy dissipation ...................................................... 111
4-4. Wave transformation model errors in comparison with wave period and wind speed .... 112
4-5. Predicted wave height with bottom friction ............................................................... 113
4-6. Predicted $u_c$ and $u_c^{*}$ using GM and JMGM models ............................................. 122
4-7. Predicted $z_s$ using GM and JMGM models .............................................................. 123
4-8. Comparison of predicted current velocity using JMGM model with measurements .... 125
4-9. Bed stratigraphy formation in horizontal one-dimensional model ............................ 130
4-10. Comparison of predicted concentrations ($\gamma_0 = 0.002$) with measurements, at 20 m . 133
LIST OF SYMBOLS

$A^b, A^s$: Complex coefficient
$A_b$: Semi-excursion amplitude
$B^b, B^s$: Complex coefficient
$b$: Distance between wave rays
$C$: Wave celerity
$C_s$: Wind-drag coefficient
$C_b$: Volumetric sediment concentration in the bed
$C_g$: Group velocity
$C_i$: Suspended sediment concentration of the $i$th size class
$C_r$: Reference concentration
$C_b$: Mixing depth coefficient
$c$: Local bed wave celerity
$D, D_s$: Grain diameter
$D_s$: Grain diameter related to the total depth
$d_{50}$: Median grain diameter
$d_o$: Near-bed wave orbital diameter
$E$: Wave energy
$f$: Frequency or Coriolis parameter
$f_i$: Fraction of the $i$th size class
\( f_i (\Delta Z_i) \) Fraction of the \( i \)th size class in net bed elevation change

\( f_i (\delta_k) \) Fraction of the \( i \)th size class in the \( k \)th layer

\( f_i (\delta_m) \) Fraction of the \( i \)th size class in the mixing depth

\( f_i (\eta_t) \) Fraction of the \( i \)th size class in the total depth

\( f_{w_e}, f_w \) Wave friction factor

\( f_{w_c} \) Skin friction factor

\( g \) Gravitational acceleration

\( H \) Wave height

\( h \) Water depth

\( i \sqrt{-1} \)

\( K \) Number of bed layers involved in the total depth

\( K_d \) Damping coefficient

\( K_s \) Shoaling coefficient

\( K_r \) Refraction coefficient

\( k \) Wave number

\( k_o \) Total bottom roughness

\( k_{tm} \) Movable bed roughness due to sediment transport

\( k_r \) Ripple roughness

\( k_g \) Grain roughness

\( L \) Wavelength

\( l \) Wave-boundary-layer length scale \( (\equiv u_{*w} / \omega) \)

\( N \) Total number of sediment size classes

\( <p> \) Reynolds average of pressure

\( Q_x \) Across-shelf sediment flux
Q_y  Along-shelf sediment flux
q_b  Mean bedload transport rate
Re  Reynolds number
Re_p  Grain Reynolds number
S  Normalized excess skin friction shear stress
S_w(f)  Spectrum of wave orbital velocity
s  Relative sediment density
T  Wave period
t  Time
U_a  Wind speed
\bar{U}  Mean depth-averaged velocity in across-shelf direction
\bar{U}_{BL}  Depth-averaged, mean wave velocity
u  Velocity in x direction
<u>  Reynolds average of horizontal fluid velocity
u'  Turbulent velocity
u_b  Near-bottom wave orbital velocity amplitude
u_c  Current velocity
u_{nb}  Shear velocity on bottom boundary
u_c  Current shear velocity
u_{cr}  Critical shear velocity
u_{cw}  Wave-current shear velocity
u_s  Shear velocity on surface boundary
u_{wmax}  Maximum wave shear velocity
u'_{wmax}  Maximum skin friction shear velocity
\[ u_w \quad \text{Wave velocity} \]
\[ u_{wm} \quad \text{Maximum wave velocity} \]
\[ u_\infty \quad \text{Free stream velocity} \]
\[ V_t \quad \text{Volume of sediment in transport} \]
\[ V_a \quad \text{Volume of sediment available for transport} \]
\[ V_b \quad \text{Volume of sediment in bedload} \]
\[ V_s \quad \text{Volume of sediment in suspended load} \]
\[ v \quad \text{Velocity in y direction} \]
\[ w' \quad \text{Turbulent velocity in z direction} \]
\[ w_f \quad \text{Fall velocity} \]
\[ x \quad \text{Horizontal axis, East-West direction} \]
\[ y \quad \text{Horizontal axis, North-South direction} \]
\[ z \quad \text{Vertical axis} \]
\[ z_{b,i} \quad \text{Bed elevation contributed by sediment of the } i \text{th size class} \]
\[ z_0 \quad \text{Roughness length } (= k_r / 30) \]
\[ z_m \quad \text{Apparent bottom roughness} \]
\[ z_r \quad \text{Reference height} \]
\[ \alpha \quad \text{Empirical constant} \]
\[ \beta \quad \text{Constant} \]
\[ F_a \quad \text{Fractional availability index} \]
\[ Y_0 \quad \text{Resuspension coefficient} \]
\[ \Delta t \quad \text{Increment of time step} \]
\[ \Delta x \quad \text{Increment of spatial step} \]
$\Delta Z_b$  Net bed elevation change
$
\delta^k$
Bed thickness of the $k$th layer
$
\delta_b$
Background mixing depth
$
\delta_m$
Mixing depth
$
\delta_{we}$
Wave-current boundary layer thickness
$
\varepsilon_D$
Energy dissipation rate
$
\zeta$
Sea surface elevation
$
\zeta_o$
Nondimensional roughness length ($= z_o / l$)
$
\eta$
Bed elevation
$
\eta_e$
Erosion depth
$
\eta_r$
Ripple height
$
\theta$
Incident angle of wave ray
$
\kappa$
von Karman constant
$
\lambda_r$
Ripple spacing
$
v$
Kinematic viscosity
$
v_s$
Eddy diffusivity
$
v_t$
Eddy viscosity
$
\rho$
Water density
$
\rho_a$
Air density
$
\rho_s$
Sediment density
$
\sigma_w$
Standard deviation of wave orbital velocity
$
\tau_b$
Bottom shear stress
$
\tau'_b$
Instantaneous skin friction shear stress
$
\tau_c$
Current shear stress

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_{cr}$</td>
<td>Critical shear stress</td>
</tr>
<tr>
<td>$\tau_{cw}$</td>
<td>Wave-current shear stress</td>
</tr>
<tr>
<td>$\tau_s$</td>
<td>Wind stress</td>
</tr>
<tr>
<td>$\tau_{sx}$</td>
<td>Wind stress, $x$ component</td>
</tr>
<tr>
<td>$\tau_{sy}$</td>
<td>Wind stress, $y$ component</td>
</tr>
<tr>
<td>$\tau_{wm}$</td>
<td>Maximum wave shear stress</td>
</tr>
<tr>
<td>$\tau'_{wm}$</td>
<td>Maximum skin friction shear stress</td>
</tr>
<tr>
<td>$\tau_x$</td>
<td>Turbulent Reynolds' stress, $x$ component</td>
</tr>
<tr>
<td>$\tau_y$</td>
<td>Turbulent Reynolds' stress, $y$ component</td>
</tr>
<tr>
<td>$\phi_s$</td>
<td>Wind direction</td>
</tr>
<tr>
<td>$\phi_c$</td>
<td>Current direction</td>
</tr>
<tr>
<td>$\phi_{cw}$</td>
<td>Acute angle between wave and current</td>
</tr>
<tr>
<td>$\phi_{ub}$</td>
<td>Upper limit of the $i$th size class in the $\phi$ unit</td>
</tr>
<tr>
<td>$\phi_{lb}$</td>
<td>Lower limit of the $i$th size class in the $\phi$ unit</td>
</tr>
<tr>
<td>$\psi_{cr}$</td>
<td>Critical Shields parameter</td>
</tr>
<tr>
<td>$\psi'_{wm}$</td>
<td>Maximum skin friction Shields parameter</td>
</tr>
<tr>
<td>$\omega$</td>
<td>Radian frequency</td>
</tr>
</tbody>
</table>

Subscripts and superscripts

- 16: of 16th percentile of bed grain size distribution
- 50: of 50th percentile of bed grain size distribution
- 84: of 84th percentile of bed grain size distribution
- ano: Anorbital
- orb: Orbital
sub Suborbital

i Grain size class

j spatial step

n time step

x x component of the coordinate system

y y component of the coordinate system
ABSTRACT

Sediment transport during a storm event on the inner continental shelf was detailed through the development of models based on field experiments conducted at Duck, North Carolina in October 1994. A vertical one-dimensional model (1DV model) was developed by coupling the Grant and Madsen (1986) model with bed stratigraphy to consider real seabeds. Sediment was divided into seven size classes and fractional transport was estimated. Mixing depth and total depth from a simplified sediment conservation equation provided the basis for changing bottom sediment, sediment availability for transport, and armoring processes. These processes involve a feedback between hydrodynamics and bed stratigraphy. A horizontal one-dimensional, depth-resolved model (1DH model) was developed to predict inner-shelf morphological changes. Flow and shear stress fields were calculated using a simple wave transformation model combined with the Jenter and Madsen (1989) model. Sediment flux was computed in relation to fractional transport and armoring processes. The sediment conservation equation was numerically solved to yield bed elevation changes associated with individual size classes. Predictions of suspended sediment concentrations from both models were adjusted by the resuspension coefficient $\gamma_0$, resulting in $\gamma_0 = 0.001$ for the 1DV model and $\gamma_0 = 0.002$ for the 1DH model, respectively.

The coupling in the 1DV model was critical to predicting suspended sediment concentrations. Hydrodynamic variables, however, were not significantly affected by changing bottom sediment. Predicted suspended sediment concentrations were higher during the waning phase of the storm than during the erosional phase. Modeled bed stratigraphy showed fining upward sequences. Wind-driven processes on the inner shelf were interpreted using the 1DH model. The magnitude and the direction of horizontal sediment flux were explained in terms of wind-driven currents. Waves produced a sigmoidal-shaped vertical concentration distribution, explaining horizontal gradients of suspended sediment concentrations. The steepness of the sediment flux gradient due to the waves was correlated with wave height. Synchronization of currents and waves was necessary for large flux divergence and morphological changes. During downwelling currents, deposition occurred on the shoreface whereas upwelling currents were accompanied by shoreface/inner shelf erosion. The inner shelf thus responded as either the sink of sediment or the source of sediment.

Baeck Oon Kim
Department of Physical Sciences
Virginia Institute of Marine Science

Thesis supervisor : L. D. Wright, Ph.D.
Title : Professor, Virginia Institute of Marine Science

xvii
MODELING STORM-INDUCED SEDIMENT TRANSPORT
ON THE INNER SHELF : EFFECTS OF BED MICROSTRATIGRAPHY
1. INTRODUCTION

The inner continental shelf exists in a morphological continuum from the beach-surf zone complex to the deeper mid to outer continental shelf (in general 20 to 30 m depth). It can be defined, from a dynamical point of view, as the region in which nonbreaking waves normally agitate the bed (Wright, 1995). Breaking and dissipation of shoaling waves drives most surf zone processes. Seaward of the surf zone, however, various forcing factors such as waves, winds, tides and their nonlinear interactions govern the complicated inner-shelf dynamics (Nittrouer and Wright, 1994; Wright, 1995). On the inner shelf, wind stress directly affects water flow and generates advective currents that affect the bed (Mitchum and Clark, 1986). The inner shelf boundary layer is characterized by wave-current interactions and involves overlapping surface and bottom boundary layers. Thus, it is important to understand boundary layer processes in the study of sediment transport on the inner shelf. For this reason, wave-current interaction models describing comprehensive boundary layer processes such as the Grant and Madsen (1986) model become important components of more complex sediment transport models.

Application of benthic boundary layer process models to natural environments demands considerations of differences between simplified assumptions of theoretical and empirical models and complicated realities existing in nature. Some important factors include flow acceleration, wave irregularities, nonhomogeneous sediment and biological effects (Dyer, 1984; Wright, 1989). However, quantifying each factor is still very difficult because of the lack of field experiments. Nielsen (1993) showed that formulae for wave-generated ripple geometry obtained from laboratory
experiments required modification in order to give agreement with data obtained from measuring natural bedforms under irregular waves. It has been noted that simple models result in unrealistic predictions for the direction of sediment transport and values of roughness parameters because of micro-scale bedforms (Wright et al., 1991; Wright, 1993). In relation to this, Wright (1993) emphasized the consideration of natural sediment to account for the time-related composition of bottom sediment (bed microstratigraphy).

A primary goal of the present study is to relate time-varying hydraulic roughness to bed stratigraphy. Bottom sediment in natural environments shows poor sorting and vertical stratification. Therefore, representative parameters for size distribution of nonhomogeneous sediment, that is, median grain size and sorting, cannot be constant as is the case for results from controlled laboratory experiments. By realizing that the size distribution of bottom sediment changes with time-varying flows through enhanced suspension of sediment, it can be assumed that the size parameter representing both bottom and suspended sediment is a function of both time-varying shear stress and elapsed time since the onset of an event. Since the geometry of bedforms is governed by the size of bed material (Middleton and Southard, 1984), hydraulic roughness, which is likely to be influenced by sediment size and bedforms, is also a function of time.

Since boundary layer processes are mutually dependent on bottom sediment, we need to consider interactions between flow and the bed. When the mixed and stratified sediment is incorporated into a sediment transport model, there exists a complicated feedback that involves changing bottom sediment and armoring processes (Holly and Karim, 1986). Sediment transport models developed for river channel beds, composed of sand and gravel mixtures, have evolved to account for the armoring processes (Borah, et al., 1982; Park and Jain, 1987; van Niekerk et al., 1992). The armoring processes play important roles in controlling sediment transport because these processes influence flows as well as availability of sediment for transport. Several studies on the
inner shelf included armoring processes (Kachel and Smith, 1986; Wiberg et al., 1994) but lack of field measurements constrained fully developed models. In this study, a model accounting for interactions between the flow and the bed is developed to deal with real seabeds. For this purpose, an approach to simplify the sediment conservation equation is employed to calculate changing bottom sediment. Also, this provides insights as to the coupling between boundary layer processes and bed stratigraphy.

Process-based models predicting beach profile changes and coastal morphology have been used to understand coastal morphodynamic systems (Roelvink and Broker, 1993; De Vriend et al., 1993). Although various mathematical models were developed by combining different constituent models of waves, currents, and sediment transport, the models were similar in that the constituent models were closely coupled to each other based on morphodynamic principles. These principles state, in brief, that boundary layer processes are the link between hydrodynamics and morphology in terms of the friction forces. The exerted forces trigger sediment transport and erosion or accretion of the morphology result from sediment flux gradients. Again, the boundary layer processes depend on the changed morphology (Wright, 1995). In this study, a process-based model associated with wind-driven processes is developed to understand the relationship between forcing and response of the inner-shelf bed. The continental shelf of the Middle Atlantic Bight is classified as storm dominated (Swift et al., 1985). It is during high energy events that the most significant sediment transport occurs. Gradients in transport (divergence of sediment flux) cause morphological changes. Across-shelf sediment transport on the inner shelf exerts important influences on morphological changes (Swift et al., 1985; Wright et al., 1991; Wright, 1995). Therefore, a model accounting for across-shelf transport during high energy events has great potential to describe inner-shelf morphodynamic processes.

One objective of this study is to understand interactions between flows and mixed, stratified
bottom sediment in order to improve our understandings of boundary layer processes on the inner continental shelf. Another objective is to integrate these boundary layer processes with a process-based model to understand relationships between sediment transport triggered by wind-driven processes and morphological changes of the inner shelf. These objectives are achieved through the development of two models: a boundary layer process model and a sediment transport model.

Chapter 2 provides description and analysis of data used in the models. In Chapter 3, a vertical one-dimensional model is developed to link boundary layer processes to bed stratigraphy based on a simplified sediment conservation equation and armoring processes. Using this model, the effects of bed stratigraphy on hydrodynamics and sediment transport will be examined. However, there are no laboratory and field measurements to prove these interactions. In Chapter 4, a horizontal one-dimensional, depth-resolved model is developed assuming a simplified geometry for the inner shelf. This model gives insights as to the roles of storm-driven waves and currents in causing sediment flux divergence. Also, the effects of sediment transport during a storm event on morphological changes of the inner shelf is examined using this model. In Chapter 5, the results of these two models are summarized and conclusions of this study are presented.
2. METHODS AND DATA ANALYSIS

2-1. Introduction

The development of models (presented in Chapter 3 and 4) is based on field experiments. Field data are used as inputs to the models. The model inputs are the measured quantities such as winds, waves, currents and bottom sediment. In addition, measured suspended sediment concentrations, shear velocity, and roughness are used to compare with the model predictions. Since this study relates the models to storm events, the field experiments require to catch those events. Therefore, the portion of field data from high-energy events is most useful for the models.

This chapter describes equipments for field experiment, the field site, the methodology for data analyses, and the results of data analyses. Winds, waves, currents and bottom sediment data corresponding to a storm event are selected for the model inputs. Because instrumented tetrapod experienced vertical displacement during the deployment, a methodology to correct measurement heights due to instrument displacements is also introduced.

2-2. Field Experiments

As a part of the multi-disciplinary CoOP'94 experiments (Coastal Ocean Processes), the Virginia Institute of Marine Science deployed two instrumented tetrapods on the inner shelf off the U.S. Army Corps of Engineers’ Field Research Facility (FRF) at Duck, North Carolina (Fig. 2-1) to measure the benthic boundary layer flows and resultant sediment resuspension. The two tetrapods were deployed at depths of 12 m and 20 m on the same transect. Four data sets were retrieved from
Figure 2-1. Map showing study area.
the two missions held in August and October 1994. The October data sets were used in this study due to stronger and prolonged storm activities.

The instrumentation on each tetrapod during the October experiment is summarized in Table 2-1. The instruments used in this study were similar to those installed on the tripod described in Wright et al. (1991). They consisted of a vertical array of four Marsh-McBirney electromagnetic current meters (EMCM), a Seadata Model 635 directional wave gage incorporating a pressure transducer and a current meter, a digital sonar altimeter (DSA) and a vertical array of five Downing optical backscatterance sensors (OBS). Tetrapods were deployed from the R/V Cape Hatteras on 4 October 1994. The measurements were recorded every four hours in bursts consisting of 1024 samples. The sampling rate was 1 Hz.

The individual current meters were calibrated before the deployments with steady flows in a recirculating flume. The OBS sensors were calibrated after the deployments using a calibration cylinder (Kim, 1991) with suspended sediment collected in a sediment trap mounted on the tetrapod.

2-3. Background of field site

The study area has several physical characteristics: sand grain size typical of US east coasts, wave climate and storm exposure representative of US east coasts, regular bottom topography in the inner shelf, a microtidal range, and a straight coastline (Birkemeier et al., 1985). The dominant wind direction during the summer season is from the southwest. The wind direction during the autumn and winter, however, is frequently from the northeast as a result of the extra-tropical storms (northeasters); westerlies often prevail during intervening periods of high pressure. The high wave activities occur during October through March and waves come predominantly from the northerly quadrants. The wave height, for example, reached 3.5 m during an October 1980 storm (Birkemeier et al., 1985).
Table 2-1. Summary of instrumentation for the October experiment.

<table>
<thead>
<tr>
<th>Description</th>
<th>Inshore tetrapod</th>
<th>Offshore tetrapod</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>36° 11.994 N 75° 42.304 W</td>
<td>36° 11.960 N 75° 42.480 W</td>
</tr>
<tr>
<td>Depth</td>
<td>12 m</td>
<td>20 m</td>
</tr>
<tr>
<td>Period</td>
<td>October 4 - November 1</td>
<td>October 4 - November 1</td>
</tr>
<tr>
<td>Initial time</td>
<td>1600 EST</td>
<td>1600 EST</td>
</tr>
<tr>
<td>Instrument</td>
<td>No. of burst</td>
<td>No. of burst</td>
</tr>
<tr>
<td>EMCM</td>
<td>137</td>
<td>168</td>
</tr>
<tr>
<td>Press. Transducer</td>
<td>169</td>
<td>168</td>
</tr>
<tr>
<td>DSA</td>
<td>129</td>
<td>129</td>
</tr>
<tr>
<td>OBS</td>
<td>169</td>
<td>173</td>
</tr>
<tr>
<td>No. of samples in one burst</td>
<td>1024</td>
<td>1024</td>
</tr>
<tr>
<td>Sampling rate</td>
<td>1 Hz</td>
<td>1 Hz</td>
</tr>
<tr>
<td>Burst time interval</td>
<td>4 hour</td>
<td>4 hour</td>
</tr>
</tbody>
</table>
Using the bathymetry data from several sources (Data Announcement 88-MGG-02; Data Announcement 95-MGG-02), a 3-D representation of bathymetry was constructed (Fig. 2-2). It shows that the inner shelf (approximately 10 to 20 m depth range) is relatively narrow, i.e., the across-shelf dimension is less than 10 km compared with the along-shelf dimension which is order of 100 km. Its shore-normal gradient is steep compared with the mid-shelf where water depth is deeper than 20 m. The shore-parallel variation of the inner shelf is not significant, while the deeper area shows irregular topography. For this reason many studies of the inner shelf assume the simplified geometry in which the alongshore variation of the morphological features is negligible.

Bathymetric profiles and seismograms collected in this area show the concave upward inner shelf profile (Tiedeman, 1995). The side-scan sonograms indicated the existence of orbital ripples and a change in sediment type near the 20 m contour. Micromorphodynamic study of the sea bed in this area has shown variation of bed characteristics depending on the hydrodynamic conditions (Wright, 1993). During the summer fairweather period, sea beds are composed of large ripples and biogenic roughnesses. The range of the ripple length and the ripple height was found to be 8 to 15 cm and 2 to 4 cm, respectively. In both post-storm and winter swell-dominated conditions, small ripples were observed. The irregular micromorphology was replaced with the highly mobile plane bed during storm periods.

The grain size analysis of the surface sediment along the shore-normal transect shows a fining seaward trend, encountering a zone of sandy silt between the 13 m isobath and the 15 m isobath (Birkemeier et al., 1985). The median grain size along the transect ranges from 0.028 to 0.012 cm. The medium to fine sand at the 12 m contour consists of an approximately 1.5 m thick sand sheet (Birkemeier et al., 1985). A core collected at 20 m depth was composed of more than 80% sand from the surface to 20 cm depth and had a fining downward trend (Tiedeman, 1995). A core collected at 9 m depth was composed of more than 90% sand in all depths and had a nearly
Figure 2-2. Three-dimensional representation of bathymetry shows that the inner shelf of study area is very narrow and steep.
uniform grain size.

Figure 2-3 shows the grain size distribution and the cumulative curve of the sediment collected at the surface of 13 m depth during 1992 field experiments. The bottom sediment and suspended sediment of a sediment trap collected during 1994 field experiment were not confidently analyzed due to unknown systematic errors. Although there is slight variation of sediment composition along the transect, the bottom sediment composition shown in Figure 2-3 was used as input for the models described in Chapter 3 and 4. It is also assumed that the horizontal variation of bottom sediment composition is insignificant. This surface sediment consists of 84% sand fraction, 11% silt fraction and 5% clay fraction with a median diameter $d_{50} = 0.0113$ cm. The variation of sand composition from the surface to 15 cm depth is presented in Figure 2-4. The sand composition from 1 cm to 2 cm depth shows the highest value, 85%. Below this layer, sand composition decreases to the minimum at 5 cm depth. It increases to about 80% at 9 cm depth and below this depth it remains constant with small fluctuations.

2-4. Data Analysis

For the right-handed coordinate system adopted in this study, the East-West component was taken to be the x axis, the offshore direction being positive. The North-South component was taken to be the y axis. The direction of currents, waves, and winds was referenced to true North which is the positive y axis. These directions were measured clockwise from the true North. For the wave and wind data, the directions are shown where they are going toward. The data were rotated such that the new coordinate system was aligned to be parallel with the local coast line (-20° with respect to true North).

2-4-1. Wind
Figure 2-3. Grain size distribution (a) and cumulative curve (b) of bottom sediment. The median diameter is 0.0113 cm and geometric sorting is 1.77. Note that silt and clay fraction (> 4ϕ) is less than 20%.
Figure 2-4. Sediment core showing vertical distribution of sand composition in 1 cm intervals.
The local wind data used in this study were provided by the FRF. Since they were measured at 34 minute intervals, it was necessary to select samples corresponding to the sampling interval of the VIMS data. The magnitude and direction of the local wind in October 1994 showed two events in which the wind speeds exceed 10 m/s with a highest speed of 18 m/sec (Fig. 2-5). The dominant direction was from the north during the high wind activities. The two peaks occurred between 10 and 18 in October were separated by a short gap in which the wind speed was reduced to about 5 m/sec and the direction was from the east.

2-4-2. Wave

Following the method described in Chisholm (1993) and Madsen et al. (1993), the wave characteristics were calculated from the near-bottom velocity measurements. The time series of wave orbital velocity was obtained by removing current velocity. The rms near-bottom wave orbital velocity amplitude, \( u_b \), is defined by

\[
    u_b = \sqrt{2} \sigma_w
\]

(2-1)

where \( \sigma_w^2 \) is the total variance of the time series of the wave orbital velocity. To estimate wave direction, \( u^2 + v^2 \) and direction were calculated using each \( u \) and \( v \) pair of wave orbital velocities. These were converted to a directional distribution with individual 1° directional bins containing the sum of variances in that direction. A running average using an 11° window smoothed directional distribution of the wave orbital velocity. It results in two peaks of the directional distribution and they represent the forward and backward wave orbital velocity differing by 180°. These wave orbital velocity directions were calculated as the mean of the centroid of each peak. Then the wave direction was given as the mean of the two directions.

Each wave orbital velocity was projected to the wave direction. Giving a time series of
Figure 2-5. Time series of the FRF wind data showing wind direction (a) and magnitude (b) in October 1994. The wind direction is defined as blowing toward.
projected wave orbital velocities, a frequency spectrum, $S_w(f)$, was then calculated. From this spectrum the rms wave period was taken as

$$T = \frac{\int_0^\infty S_w(f) \, df}{\int_0^\infty f S_w(f) \, df} \quad (2-2)$$

where $f$ is the frequency.

Using the values computed from (2-1) and (2-2) and the wave dispersion relationship for monochromatic waves

$$\omega^2 = g k \tanh kh \quad (2-3)$$

where $\omega$ is the radian frequency, $k$ is the wave number and $h$ is the water depth, the significant wave height $H_s$ can be estimated by

$$H_s = \sqrt{2} u_b T \frac{\sinh(kh)}{\pi} \quad (2-4)$$

with linear wave theory (Dean and Dalrymple, 1992). To get $kh$ in (2-4), the equation (2-3) needs to be solved by an iterative method. Alternatively an accurate approximate solution for $kh$ (Hunt, 1979) can be used, which gives

$$(kh)^2 = y^2 + \frac{y}{1 + \sum_{n=1}^{6} d_n y^n} \quad (2-5)$$

where
\[ y = \frac{\omega^2 h}{g}, \tag{2-6} \]

\[ d_1 = 0.666..., \quad d_2 = 0.355..., \quad d_3 = 0.1608465608, \quad d_4 = 0.0632098765, \quad d_5 = 0.0217540484, \quad \text{and} \quad d_6 = 0.0065407983. \]

The time series of the wave characteristics observed in October 1994 is presented in Figure 2-6 and shows that the maximum near-bottom wave orbital velocities and the maximum wave periods reached 60 cm/sec and 15 sec, respectively. The only one prominent peak of wave orbital velocities occurred at the time of the strongest wind (see Figure 2-5) but the trend of wave orbital velocities was not similar to that of wind speeds. This indicates that the waves were not locally generated waves. At the onset of the storm, on 10 October, the shortest wave period was recorded and then it increased monotonously to its maximum value on 20 October when wind speeds had already died out. This also proves that the wave generation area was further away from the measurement site, so the swells arrived after local wind ceased. The dominant direction of the waves was from the southeast except during the period of the onset of the storm when waves arrived from the northeast. Afterwards wave direction changed, so waves came from the southeast. The relationships between the local winds and the waves are presented in detail in Kim et al. (in preparation).

2-4-3. Current

The measured currents of the top current meter (93 cm above bed) at the 20 m site are presented in Figure 2-7. The largest velocity was about 50 cm/sec occurred at the time of the strongest local wind, and it was directed predominantly to the south in the along-shelf direction. The major features of both current speed and directions were closely related to wind speed and directions,
Figure 2-6. Time series of wave characteristics showing near-bottom wave orbital velocities (a), wave periods (b), and wave directions (c) at 20 m depth.

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 2-7. Time series of current velocity showing current direction (a) and magnitude (b) of the highest sensor (93 cm above bed) at 20 m depth.
indicating a wind-driven current. For example, two prominent peaks in the velocity time series from 10 to 17 in October were clearly caused with the peaks in the wind time series. On the other hand, the minor peak in the late phase of the storm from 17 to 20 in October was not related to the local wind occurred in that period. The onshore/offshore velocity components appeared to be minimal due to the much larger along-shelf velocity components. It was argued that the across-shelf velocity components are small enough to be in the range of the measurement errors, so that this small velocity can cause an erratic estimation of the sediment flux (Kim et al., in preparation).

2-4-4. Suspended Sediment

Measured values of winds, waves, and currents corresponding to the days from 10 to 20 in October were selected for the model inputs. This time segment was characterized by a broad peak of high energy flows which were responsible for most of the resuspension events during the field experiment. As a result, suspended sediment concentrations increased continuously during the period from 10 to 16 in October (Fig. 2-8). Both time series of concentrations at 12 m and 20 m depths appeared to follow those of the wave orbital velocities shown in Figure 2-6. The concentrations at the 12 m depth were roughly twice of those at the 20 m depth.

The OBS sensors of the tetrapods at both sites showed a steady increase of suspended sediment concentrations (Fig. 2-8). Thus, the OBS data from all elevations at the 12 m depth and the lower three elevations at the 20 m depth increased beyond the preset maximum concentration of approximately 5 g/l during the waning storm phase. Since a maximum concentration is expected to be measured at the storm peak, the OBS data greater than this value has no physical meaning. This unrealistic behavior may result from either an electronic drift or a fouling problem. Therefore, the OBS data were cut off on 16 October. Except this truncation, no data correction was made further. These truncated OBS data were used for comparisons with predicted suspended sediment
Most OBS sensors give values greater than the preset maximum concentration, 5 g/l, after 16 October.
concentrations.

The truncated OBS data indicate some problems in performance of the OBS sensors. As shown in Figure 2-8 b, the sensor 3 showed greater concentrations than the sensor 2 at the 20 m depth during the period from 13 to 16 in October. This contradicts a physically reasonable trend in which we expect a decreasing concentration with elevation above the bed during high energy flows. If we exclude the sensor 3, the truncated OBS data from the rest of sensors appear to be reasonable because measured concentrations decrease with elevation. More damage was encountered in the OBS data at the 12 m depth (Fig. 2-8 a). Measured concentrations from the sensor 1 (not shown in Figure 2-8 a) were greater than the preset maximum capacity of the sensor, 5 g/l, during low energy flows. The higher three sensors (sensor 3, 4, and 5) showed an unrealistic trend of an increasing concentration with elevation, so these were not acceptable. The quality of the OBS measurements from the inshore tetrapod was possibly bad.

2-4-5. Grain Size

For the grain size analysis, the general operation for the wet sieving followed Folk’s method (Folk, 1968). After dividing sediment samples into coarse and fine grain size fractions, the fine grain size fractions were analyzed by pipette methods to yield the ratio of silt to clay. Then weighted sand:silt:clay ratio was found to give the preliminary result of the grain size analysis. In order to estimate grain size parameters using a graphic method (Folk, 1968), grain size distributions from coarse and fine grain size fractions, each grain size fraction being analyzed separately, were combined to a grain size distribution. A Rapid Sand Analyzer (automated settling tube) was used for the sand fraction and a Micrometrics SediGraph was used for the silt and clay fraction. Since the fine grain size fractions in a typical surface sediment from the study area were not enough to be analyzed using the SediGraph, subsurface samples from a sediment core collected near the 20 m isobath
(during 1995 field experiment) were used instead. This substitution for the fine fractions was acceptable because the ratio of the silt to clay fractions in the subsurface samples was similar to that of the surface sediment. They were combined to form the grain size distribution shown in Figure 2-3 which was used to estimate grain size parameters. The median diameter $D$ in the $\phi$ unit was obtained by observing the 50 percentile of the cumulative curve. This median diameter in the $\phi$ unit was converted to $D$ in the mm unit using the formulation

$$D [\phi \text{ unit}] = -\log_2 D [mm \text{ unit}]$$

In the same way, the 84 percentile and the 16 percentile (percent coarser than) were calculated and converted to $D_{84}$ and $D_{16}$, respectively. The geometric sorting was found by $\sqrt{D_{16}/D_{84}}$ which is about 1.77. This value satisfies the criterion for the armoring processes, in which the sorting value must be greater than 1.5 (Chin et al., 1994).

A stratified bed generally consists of various layers and each layer can be identified with its thickness and its grain size composition. When we describe microscale features of bed stratigraphy (mm to cm scale) as is the case in this study, it is practically difficult to get a unique bed stratigraphy for certain conditions from field measurements. This makes it difficult to obtain an input of bed stratigraphy for the model because the surface layers of sea beds are susceptible to continuous change due to either physical and biological processes. Therefore, it might be reasonable to choose a simple input, for example, a massive bed (vertically uniform bed showing no structure inside) shown in Figure 2-9 instead of expending great effort specifying complicated bed stratigraphy. As shown in Figure 2-9, a simple input can be represented as a matrix form where identical layers with regard to layer thickness and grain size composition are repeated downward from the surface of a vertical sediment column. The modeling of the formation of bed stratigraphy is, however, a different problem. The proposed model itself can deal with even complicated input of bed stratigraphy,
### Input of bed stratigraphy

<table>
<thead>
<tr>
<th>bed thickness cm</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>6</th>
<th>8</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
<tr>
<td>1</td>
<td>0.0220</td>
<td>0.0151</td>
<td>0.4026</td>
<td>0.4203</td>
<td>0.1058</td>
<td>0.0170</td>
<td>0.0172</td>
</tr>
</tbody>
</table>

Figure 2-9. Input of bed stratigraphy is a matrix consisting of bed thickness and grain size composition of each layer. A massive bed is chosen for simplicity. It is also noted that the number of size classes in silt and clay fraction is reduced to reduce computational time.
although this study uses a simple massive bed as input.

The size interval of the fine grain size fractions (> 4 φ) in Figure 2-9 is modified to reduce the number of the input size fractions as well as the number of iterative processes. This reduces computational time in a complicated model. Since the fine grain size fractions are not abundant at the study area, this modification will not cause a significant effect. However, this is not always the case when a different type of the bottom sediment is encountered.

2-5. Correction of Measurement Heights

The rate of the change of hydrodynamic quantities becomes greater with increasing proximity to the bed. This boundary effect can cause significant errors in evaluating velocity and concentration profiles if there are some appreciable discrepancies in measurement heights. Therefore, it is important to know the variation of measurement heights with time. A theoretical consideration for this problem is presented in Appendix A. In the following a method to calculate measurement heights using DSA measurements will be presented.

The time series of the displacement heights can be obtained using the time series of the relative distance from DSA to seabed if we know the initial value of this distance. Temporal changes in this distance can be produced by a combined effect of the sink of the instrumented tetrapod and the bed elevation changes due to sediment erosion and deposition. At the current stage of our technique, it is impossible to differentiate between these effects. Since displacement heights are relative distance changes, the DSA measurements are appropriate for correction of measurement height changes. However, an extended use of these data, such as describing the absolute change of bed elevation, requires careful attention. The reason is that these data involve the combined effects and there is uncertainty as to the absolute changes of the displacement heights. So the absolute bed elevation changes are not separable from the DSA measurements.
The time series from the DSA of both the inshore and offshore tetrapods are presented in Figure 2-10. Both showed overall variations of about 25 cm. The DSA ranges varied with time, forming approximately two maxima and two minima. During the storm period, from 10 to 20 October, the DSA time series from the instrument located at the 12 m contour increased to form a local maximum and then decreased sharply until it had a second local maximum at the time of strongest flow condition. After the second local maximum, the time series decreased to a minimum and then finally increased. The DSA time series from the 20 m depth generally followed the time series from the 12 m depth but with some time lag and minor differences.

In situ instrument heights above the seabed are routinely observed by divers as the tetrapods are deployed and retrieved. Unfortunately it was not possible during the October 1994 deployment due to the high wave activities at the time of deployment. However, measurement heights of instruments were recorded on the deck of the deployment vessel. In addition to these measurement heights above the deck, diver observations during the deployment are summarized in Table 2-2. By using DSA time series presented in Figure 2-10, the initial measurement heights were determined and are summarized in Table 2-3. In case of the inshore tetrapod, the most reliable record came from a diver observation. The lowest OBS sensor was 5 cm above bed (hereafter, ab) in early October, although the exact time of observation was missing. This indicated that the instrument sank approximately 6 cm (the difference between the lowest OBS height above the deck and the diver record) early in the deployment. The DSA time series possibly ranges between 118 and 120 cm during the time of diver observation. The initial displacement height, 6 cm, was added to these values, getting a range from 124 to 126 cm, which turns out to be close to the initial measurement height of DSA above the deck, 127 cm. The error 1 to 3 cm, which is within 3% error, is acceptable considering that the deck surface was not flat. Therefore, the lowest EMCM and OBS sensors were estimated to be 5 cm ab at the time of the deployment.
Figure 2-10. Time series of DSA data measured at 12 m (a) and 20 m (b) depths.
Table 2-2. Summary of measurement heights for the October experiment.

<table>
<thead>
<tr>
<th>Instruments</th>
<th>Inshore tetrapod 12 m</th>
<th>Offshore tetrapod 20 m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ab†</td>
<td>ad ‡</td>
</tr>
<tr>
<td>DSA</td>
<td>N/A</td>
<td>127</td>
</tr>
<tr>
<td>EMCM</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>N/A</td>
<td>11×</td>
</tr>
<tr>
<td>2</td>
<td>41</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>71</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>105</td>
<td></td>
</tr>
<tr>
<td>OBS</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>N/A</td>
<td>11×</td>
</tr>
<tr>
<td>2</td>
<td>42</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>71</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>107</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>125</td>
<td></td>
</tr>
<tr>
<td>Diver measurement</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(OBS 1, in early October)</td>
<td></td>
</tr>
</tbody>
</table>

† above the bed (cm)
‡ above the deck (cm)
× buried during deployment of tetrapod
Table 2-3. Initial measurement heights measured above the bed (cm).

<table>
<thead>
<tr>
<th>Instruments</th>
<th>Inshore tetrapod</th>
<th>Offshore tetrapod</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>12 m</td>
<td>20 m</td>
</tr>
<tr>
<td>EMCM</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>5(^\times)</td>
<td>4</td>
</tr>
<tr>
<td>2</td>
<td>35</td>
<td>33</td>
</tr>
<tr>
<td>3</td>
<td>65</td>
<td>64</td>
</tr>
<tr>
<td>4</td>
<td>99</td>
<td>93</td>
</tr>
<tr>
<td>OBS</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>5(^\times)</td>
<td>7</td>
</tr>
<tr>
<td>2</td>
<td>36</td>
<td>33</td>
</tr>
<tr>
<td>3</td>
<td>65</td>
<td>64</td>
</tr>
<tr>
<td>4</td>
<td>101</td>
<td>93</td>
</tr>
<tr>
<td>5</td>
<td>119</td>
<td>120</td>
</tr>
</tbody>
</table>

\(\times\) buried during deployment of tetrapod
Diver observations of measurement heights of instruments on the offshore tetrapod were collected at the end of the experiment. In spite of the truncation of the DSA time series, it was assumed that the later DSA values (about 140 cm) can be extrapolated to the end of the experiment and related to the observed 3 cm ab of the lowest EMCM. Using the fact that the initial DSA ranges were similar to the values at the end, it was proposed that the tetrapod initially sank approximately 11 cm. So the initial value of the DSA time series was related to a position of 4 cm ab for the lowest EMCM sensor. The smallest value of the DSA time series over the deployment was not much different from the initial value. This small variation doesn’t allow burial of the lowest EMCM sensor during the deployment.

Based on the time series of the displacement heights, the time-varying measurement heights were used for the calculation of shear velocity and roughness parameters. The fact that the inshore tetrapod experienced much more vertical displacement than the offshore tetrapod and the lowest sensors of the inshore tetrapod were buried during the early phase of the storm reduces confidence in the inshore data set.

The time series of the DSA could be converted to a time series of bed elevations (Fig. 2-11 c), if sinking of the instrument after the deployment is not considered. The initial value of the DSA time series is simply translated to the zero elevation. Time series of two hydrodynamic variables, wave orbital velocity and current from both inshore and offshore sites, are also presented in Figure 2-11. The major features of the bed elevation changes in this study are somewhat different from those described in Wright et al. (1986), where bed elevation changes were closely related to the wave orbital velocities and so consisted of an erosional phase at the storm peak and a depositional phase after the event. In contrast, the two depositional phases of bed elevation changes were apparently associated with the waning phases of two peaks in the current velocity time series, interrupted by an erosional phase during the storm peak coinciding with the strongest wave orbital velocity.
Figure 2-11. Time series of near-bottom wave orbital velocities (a) and current velocities (b) at 12 m and 20 m depths are compared with the time series of bed elevation changes (c) at the two locations.
3. BOUNDARY LAYER PROCESSES COUPLED WITH BED STRATIGRAPHY

3-1. Introduction

It has been proposed that time-varying roughness should be considered in the bottom boundary layer processes when we deal with beds of mixed and stratified bottom sediment. This is because the roughness varies with the changes in grain-size distribution in the uppermost part of the sediment column due to sorting during sediment transport (Parker and Klingeman, 1982; Hill et al., 1988; Kuhnle, 1989; Wilcock and Southard, 1988; 1989; Wilcock and McArdell, 1993). In armored beds of river channels with unidirectional flows, variations of grain-size distribution with time have been simulated using various numerical models (Borah et al., 1982; Lee and Odgaard, 1986; Park and Jain, 1987; van Niekerk et al., 1992). In a sensitivity analysis conducted by Bedford and Lee (1994), the grain-size distribution parameter was found to be most sensitive using a bottom boundary layer model which followed the Glenn and Grant model (Glenn and Grant, 1987). In addition, a range of bottom sediment conditions will occur during the course of erosion/deposition for a given stratified sediment. Therefore, assumptions of constant sediment parameters which have been widely used in the sediment transport literature become inadequate from the theoretical point of view.

Since flow conditions and responses of bottom sediment are mutual (Holly and Karim, 1986; Wright, 1993), there are significant interactions between the bed and the flow. It is natural to model these interactions with a bed of mixed and stratified sediment to completely describe the boundary layer processes. Numerical models dealing with sediment transport for alluvial channels have focused on the behavior of the mixed or active layer in attempts to quantify changes in the sediment size of
the armor or surface layer (Borah et al., 1982; Lee and Odgaard, 1986; Park and Jain, 1987; van Niekerk et al., 1992). These models linked successfully the concept of mixing depth in the bottom sediment to the controls of sediment transport. As Lu and Shen (1986) pointed out, most models were similar in the respect of the hydrodynamic routine but treated mixing depth and armoring processes in different ways.

A few researchers have tried to include beds of mixed sediment in the models for continental shelf bottom boundary layers (Shi et al., 1985; Kachel and Smith, 1986; 1989; Lyne et al., 1990; Wiberg et al., 1994). The lack of field measurements over the inner shelf with regards to time-varying roughness parameters has severely limited the development of a fully described model such as those used for river beds. These simple equilibrium models have neglected the fact that changes in the grain size composition affect sediment transport and the resulting sediment transport alters bottom roughness. In this chapter, the feedback relation between flow and sediment is a major consideration in the development of a new model in which the procedure of sediment stratigraphy formation is incorporated with this feedback relation. Although this model is based on a steady state assumption, it can also be used as a time-dependent model in limited conditions, assuming that equilibrium state is reached in a time scale of several hours. This time scale is similar to the time interval of the field measurements in this study. Although there is no way to verify changes in bed sediment, there are some indirect ways to determine whether the results of model calculations are reasonable. For example, one possible way comes from monitoring suspended sediment concentration, which is relatively easier to measure than monitoring bottom sediment, under the assumption that changes in the bed are directly linked to resuspension activity.

In this chapter, a vertical one-dimensional boundary layer model (1DV model) that is coupled with bed stratigraphy is developed to examine the effects of mixed and stratified sediment on hydrodynamics and sediment transport. Figure 3-1 shows all components which are integrated to
construct the model. These include a wave-current interaction model, a bed roughness model, sediment transport equations associated with the sediment conservation equation, and bed stratigraphy formation. The combined boundary layer model based on the two-layer eddy viscosity profile by Grant and Madsen (1979; 1986) is used because its simplicity makes it easy to incorporate this component into the model. Fractional sediment transport of each sediment size class is used to estimate for both bedload and suspended load. Procedures to calculate the changing bottom sediment and the sediment availability for suspended sediment transport are proposed by simplifying the sediment conservation equation. These procedures are associated with an armoring process. A mutually dependent feedback between hydrodynamics and bed stratigraphy based on these processes is proposed but it is emphasized that the approach utilized is not unique. Solution procedures are described, followed by results of the model. Finally the limitations of the model are discussed.

3-2. Wave-current Boundary Layer Model

Grant and Madsen (1979; 1986) and Madsen (1994) developed the wave-current boundary layer model used in this study. Their model calculates the velocity profiles and the shear velocities for the combined wave-current boundary layer. The shear velocities are necessary for the calculation of sediment transport. Neglecting the Coriolis term and surface wind stress, the simplified, linearized, Reynolds-averaged equation of motion within the bottom boundary layer (Grant and Madsen, 1979; 1986) becomes

\[
\frac{\partial \langle u \rangle}{\partial t} = -\frac{1}{\rho} \frac{\partial \langle p \rangle}{\partial x} - \frac{\partial \langle u'w' \rangle}{\partial z}
\]

(3-1)

in which \(t\) is time, \(z\) is the vertical coordinate measured positive upward from the bottom, \(\rho\) is the water density, \(u\) is the horizontal fluid velocity in the \(x\)-direction, \(w\) is the vertical fluid velocity, \(p\) is
Figure 3-1. Schematic representation showing boundary layer processes coupled with bed stratigraphy. It consists of wave-current interaction model, sediment transport associated with sediment conservation equation, and bed stratigraphy formation.
the pressure, the prime denotes the turbulent fluctuation and the angle brackets represent the Reynolds-averaging process. Using a turbulent eddy viscosity $v_e$, the turbulent flux of momentum $-\langle u'w' \rangle$ in equation (3-1) can be related to the vertical gradient of the mean fluid velocity by

$$-\langle u'w' \rangle = v_e \frac{\partial \langle u \rangle}{\partial z} \quad (3-2)$$

and the eddy viscosity is modeled under the assumption that the eddy viscosity varies linearly with the distance from the bottom (Grant and Madsen, 1979; 1986). Incorporating the shear velocity $u_{cw} = \sqrt{\tau_{cw}/\rho}$, based on the combined bottom shear stress inside the wave boundary layer, and the shear velocity $u_{wc} = \sqrt{\tau/\rho}$, based on the current bottom shear stress, the eddy viscosity is given as a function of the elevation by

$$v_e = \kappa u_{cw}^2 \quad \text{for } z < \delta_{wc}$$
$$v_e = \kappa u_{cw}^2 \quad \text{for } z > \delta_{wc} \quad (3-3)$$

where $\delta_{wc}$ is the wave boundary layer thickness that is defined as $\delta_{wc} = \kappa u_{cw}/\omega$ in which $\kappa = 0.4$ is the von Karman's constant and $\omega$ is the wave radian frequency (Grant and Madsen, 1979).

With the assumption of the time-invariant eddy viscosity, the pressure and velocity terms can be resolved into the current and wave components. Dropping the angle brackets for clarity and following the manipulation shown in Grant and Madsen (1986), we have the equation for the wave

$$\frac{\partial (u_w - u_\infty)}{\partial t} = \frac{\partial}{\partial z} \left[ v_e \frac{\partial (u_w - u_\infty)}{\partial z} \right] \quad (3-4)$$

where $u_w$ is the wave velocity and $u_\infty$ is the free stream velocity. And the equation for the current is given by
where $u_e$ is the current velocity. The approximate wave solution for (3-4) is found in Grant and Madsen (1986). This assumes a periodic wave motion specified by the near-bottom wave orbital velocity $u_w = u_o \cos(\omega t)$ and a no-slip boundary condition at the bottom roughness $z_0$.

$$u_e = \frac{u_b \ln(z/z_0)}{[\ln(\zeta_0 - 1.15)^2 + \left(\frac{\pi}{2}\right)^2]} \cos \omega t$$  \hspace{1cm} (3-6)

where $u_b$ is the maximum near-bottom wave orbital velocity and $\zeta_0$ is the nondimensional roughness length given by $\zeta_0 = z_o/\kappa = z_0/(\kappa u_w/\omega)$. The current solutions for (3-5) are given as follows. With a no-slip boundary condition at the top of the bottom roughness $z_0$, this equation is valid for $z < \delta_w$.

$$u_e = \frac{u_w}{\kappa u_w} \ln\frac{z}{z_0}$$  \hspace{1cm} (3-7)

For $z > \delta_w$, the current velocity profile can be written as follows with the apparent bottom roughness $z_{aw}$.

$$u_e = \frac{u_w}{\kappa} \ln\frac{z}{z_{aw}}$$  \hspace{1cm} (3-8)

Since $u_w$ is an unknown, we need a condition for the maximum wave shear stress $\tau_{wm}$ to close the problem, that is
and the combined shear stress \( \tau_{cw} \) is given as the vector addition of the shear stress \( \tau_{wm} \) due to the wave and the shear stress \( \tau_c \) due to current, which leads to

\[
u^2_{cw} = \frac{|\tau_{cw}|}{\rho} = \frac{|\tau_{wm}| + |\tau_c|}{\rho}
\]

(3-10)

Inserting (3-6) into (3-9), Grant and Madsen (1986) gave

\[
u_{wm} = \frac{\kappa u_{cw} H_b}{\left\{ \ln \left( \frac{\kappa u_{cw}}{z_0} \right) - 1.15 \right\}^2 + \left( \frac{\pi}{2} \right)^2}^{1/2}
\]

(3-11)

They further express equation (3-10) in terms of \( u_{wm} \) as

\[
u^2_{cw} = C_\mu \nu^2_{wm}
\]

(3-12)

where

\[
C_\mu = (1 + 2\mu \cos \phi_{cw} + \mu^2)^1/2
\]

(3-13)

in which \( \phi_{cw} \) is the angle between wave and current and \( \mu \) is

\[
\mu = \left( \frac{u_{cw}}{u_{wm}} \right)^2
\]

(3-14)
In the case of weak currents with strong waves, which is a typical case on the inner shelf, $\mu << 1$ and alternately $C_\mu = 1$. Under this condition, $f_{wc} = f_w$. We may have the following relationship for the wave friction factor

$$u_{rms}^2 = \frac{1}{2} f_{wc} u_b^2$$

(3-15)

However, in the case of storms, this relationship may be not well justified because the currents are more strongly forced. Based on equations (3-11), (3-12), and (3-15), Grant and Madsen (1986) gave an approximate expression for the wave friction factor related to bottom roughness in the presence of waves and currents.

$$\frac{1}{4 \sqrt{f_{wc}/C_\mu}} + \log \frac{1}{4 \sqrt{f_{wc}/C_\mu}} = \log \left( \frac{C_\mu u_b}{\omega z_0} \right) - 1.65 + 0.24 \sqrt{f_{wc}}$$

(3-16)

When current, periodic wave characteristics, and bottom roughness are given, equations from (3-4) to (3-16) can be solved for the velocity profile and shear velocities. Depending on the specification of the current, the method of solution has different procedures (Madsen et al., 1994). In this section, solutions are sought for the case in which the current is specified at a reference height. When flows interact with the bottom sediment, if there are wave-generated ripples and sediment transport, both bedforms and sediment transport increase the total bottom roughness. This enhanced bottom roughness affects the flows. Since the ripple generation and sediment transport are estimated by the flow parameters which are solutions of the above equations, there is mutually dependent relationship between the bottom roughness and flow parameters. Therefore, these equations are solved iteratively until an equilibrium state is established. In the following section, formulations of the bottom roughness are described.
3-3. Bottom Roughness

The solution for the velocity profile requires knowledge of the bottom roughness, $k_b$ which is given by

$$k_b = k_g + k_{br} + k_{bm}$$  (3-17)

The grain roughness $k_g$ is equal to the grain diameter $D$, and the ripple roughness $k_{br}$ is given by Nielsen (1981) in terms of the ripple geometry, the ripple height $\eta_r$ and the ripple spacing $\lambda_r$,

$$k_{br} = 8 \eta_r \left( \frac{\eta_r}{\lambda_r} \right)$$  (3-18)

By way of comparisons of several models, Xu and Wright (1995) proposed a movable bed roughness $k_{bm}$ due to sediment transport as

$$k_{bm} = 5 D_s (\psi'_m - \psi_\alpha)$$  (3-19)

In equation (3-19), $\psi_\alpha$ is the critical Shield parameter and $\psi'_m$ is the maximum skin friction Shield parameter which can be calculated as follows

$$\psi'_m = \frac{\tau'_{wm}}{\rho(s-1)gD_s}$$  (3-20)

where $s = \rho_s/\rho = 2.65$ is the relative sediment density. The maximum skin friction shear stress $\tau'_{wm}$ is related to $u_g$ through the skin friction factor $f'_{uc}$,

$$\tau'_{wm} = \frac{1}{2} \rho f'_{uc} u_g^2$$  (3-21)
Predictive models for the ripple geometry in a wave-dominated environment have been proposed by several studies (Nielsen, 1981; Grant and Madsen, 1982; Wiberg and Harris, 1994). Wiberg and Harris (1994) developed equations for predicting ripple type and geometry based on sediment size, near-bed wave orbital diameter \( d_0 \) and estimated anorbital ripple height. The wave orbital diameter is defined as the wave-induced water particle excursion near the bed and this is easily estimated using \( d_0 = H/\sinh(kh) \), where the wave height \( H \) should be based on the significant wave height. Nielsen (1981) and Grant and Madsen (1982) used wave orbital amplitude (half of the wave orbital diameter) and a nondimensional skin friction factor through a friction factor. Wiberg and Harris (1994) compared the proposed models using a large data set of laboratory and field measurements and argued that their model is simple and accurate. Since Wiberg and Harris’ model unified laboratory and field measurements, it doesn’t differentiate the case between laboratory and field as does the Nielsen’s method. Their model also provided much better estimates of ripple geometry than the Grant and Madsen’s method. For this reason, this study followed the Wiberg and Harris’ model for ripple geometry.

Wiberg and Harris (1994) described ripple types as follows. There are three categories depending on the relationship between ripple spacing \( \lambda \) and wave orbital diameter. When ripple spacing is proportional to \( d_0 \), these ripples are defined as orbital ripples. When ripple spacing is almost independent of \( d_0 \) and is a constant multiple of sediment size, these ripples are defined as anorbital ripples. Anorbital ripples occur at large values of the wave orbital diameter compared to sediment size. The transitional group between orbital and anorbital ripples falls into suborbital ripples. In the continental shelf, most ripples are found to be suborbital and anorbital ripples. These ripple types can be predicted in terms of the ratio of the wave orbital diameter to the anorbital ripple height \( \eta_{\text{an}} \). They provided the classification criteria for the predicted ripple types as follows.
Following Wiberg and Harris (1994), ripple geometry is calculated in terms of ripple spacing and steepness, defined as $\eta_r/\lambda_r$. The ripple height $\eta_r$ is simply calculated from the product of ripple spacing and steepness. The ripple spacing of the anorbital ripple is given by

$$\lambda_{ano} = 535 \ D_s$$

(3-23)

and the steepness of the anorbital ripples is expressed as

$$\frac{\eta_r}{\lambda_r} = \exp \left[ -0.095 \left( \ln \frac{\eta_r}{d_o} \right)^2 + 0.442 \ln \frac{\eta_r}{d_o} - 2.28 \right]$$

(3-24)

Equation (3-24) is solved for the ripple height by an iterative method using (3-23). Then the estimated ripple height is used to determine ripple types according to (3-22) and if this procedure indicates anorbital ripples, the evaluation of the ripple geometry parameters is satisfied. In the case of orbital or suborbital ripples, the parameters of the ripple geometry should be reevaluated as follows (Wiberg and Harris, 1994). For the orbital ripples,

$$\lambda_{orb} = 0.62 \ d_o$$

$$\left( \frac{\eta_r}{\lambda} \right)_{orb} = 0.17$$

(3-25)

For the suborbital ripples, the ripple spacing is obtained from

\[ d_o/\eta_{ano} < 20 \quad \text{for orbital ripples} \]
\[ d_o/\eta_{ano} > 100 \quad \text{for anorbital ripples} \]
\[ 20 < d_o/\eta_{ano} < 100 \quad \text{for suborbital ripples} \]
\[
\lambda_{sub} = \exp \left\{ \ln\left(\frac{\ln(d_/\eta_{med}) - \ln 100}{\ln 20 - \ln 100}\right) \left[ \ln \lambda_{orb} - \ln \lambda_{med} + \ln \lambda_{med} \right] \right\} \tag{3-26}
\]

and the suborbital ripple height \( \eta_{sub} \) is computed by \( \eta_{sub} = (\eta_r / \lambda_r) \lambda_{sub} \) with (3-24).

It is noted that the Wiberg and Harris' model for the ripple geometry doesn't involve any explicit consideration of the nondimensional Shields parameter \( \psi'_m \). When the flow condition is such that \( \psi'_m \) is less than the critical Shields parameter \( \psi_w \), there is no motion of sediment and consequently no ripples are generated. In some cases, there exist ripples which are inherited from earlier events but these are not predictable. There is also a critical value of \( \psi'_m (> 0.8) \) where ripples are washed out and this results in a flat bed (Wilson, 1989). Thus, the range of \( \psi'_m \), 0.05-0.8, can be defined as the conditions under which ripples may be active. Wiberg and Harris (1994) defined these active ripple conditions as the range of \( d_/D_s \) which was roughly \( 3 \times 10^3 - 10^4 \) when the wave period of \( T = 10 \) sec was considered. So it is necessary to examine whether the range of \( d_/D_s \) is similar to that of \( \psi'_m \) for the active ripple conditions when \( d_/D_s \) is converted to \( \psi'_m \). The orbital diameter \( d_o = 2u_o/\omega \) can be related to the shear velocity \( u'_w = (\psi'_w / \rho)^{1/2} \) through the wave friction factor \( f'_w = 2(u'_w / u'_b)^2 \). On hydraulically rough flows \( (Re = u'_w k_b / \nu > 70) \), most of the cases in field environments, the wave friction factor depends on a nondimensional parameter \( A_o/k_b \) where \( A_o = d_/2 \). Since \( A_o/k_b \) is unknown, the computation of wave friction factor or shear velocity from an orbital velocity is not straightforward. Alternately, as a rough estimation, the wave friction factor can be assumed to be of the order 0.01 for the range of \( A_o/k_b \), 10-10^4. For the active ripple conditions of \( d_/D_s \) the range of \( u_b \) becomes approximately 10-30 cm/sec when \( D_s \) and \( T \) are taken as 0.01 cm and 10 sec, respectively. Using \( f'_w = 0.01 \) and \( g (s-1) D_s = 15 \) in equation (3-20), the range of \( \psi'_m \) corresponding to the range of \( u_b \) is 0.03-0.3 which is on the order of \( \psi'_m \) (= 0.05-0.8). This indicates that the range of \( d_/D \) is similar to that of \( \psi'_m \) for the active ripple conditions.
3-4. Sediment Transport

The sediment transport equations can be separated into bedload and suspended load. In coastal areas, a considerable amount of sediment moves in suspension (Wright et al., 1991; Nielsen, 1993). Although bedload contributions are negligible, it is interesting to treat bedload as analogous to suspended load in an attempt to make both bedload and suspended load have an equivalent dimension of volume. Thus, these can be easily incorporated in a sediment conservation equation which is discussed later.

Individual sediment transport equations corresponding to the size classes of the bottom sediment, which are defined as fractional transport equations, can be expressed for each size class. Fractional transport equations, for both bedload and suspended load, are appropriate to deal with mixed sediment. The number of equations depends on the number of the size classes, which are determined based on the sediment size distribution. The transport equation of each size class behaves as if it is based on the single grain size representative of the size class. There is no interaction between each size class.

The representative sediment size of each size class, denoted as $D_i$, can be given as the geometric mean of the size class. So the size parameter of the $i$th sediment size class in the mm unit is expressed as

$$D_i = 2^{-\frac{\phi_{u,i} + \phi_{l,i}}{2}} \quad (3-27)$$

in which $\phi_{u,i}$ and $\phi_{l,i}$ represent the upper and lower limit of the $i$th size class in the $\phi$ unit. The fraction of a size class is assigned as $f_i$ and the sum of all $f_i$ must be unity, that is

$$\sum_{i=1}^{N} f_i = 1 \quad (3-28)$$
in which $N$ is the total number of the size classes. In this study sediment are divided into seven size classes (Fig. 2-9).

3-4-1. Mean Suspended Sediment Concentration Model

The following suspended sediment concentration model which is used for a sediment transport module is described in Madsen et al. (1994). Suspended sediment concentration profiles are calculated by a Rouse type equation which uses an eddy viscosity profile from the wave-current interaction model. In designating the reference concentration, it is necessary to consider the problem of sediment availability which is discussed later.

The time-averaged suspended sediment concentrations of the individual size classes are computed from the diffusion equation (Madsen et al., 1994). Assuming steady state, the governing equations are

$$\frac{\partial}{\partial z}(-w_{f,i}C_i) = \frac{\partial}{\partial z} \left( v_x \frac{\partial C_i}{\partial z} \right)$$

(3-29)

in which $C_i$ is the local suspended sediment concentration of the $i$th size class, $w_{f,i}$ is the fall velocity of the $i$th size class and $v_x$ is the eddy diffusivity. The eddy diffusivity is assumed to be related to the eddy viscosity by $v_x = \beta v_t$. The value of $\beta$ may be controlled by vorticity due to bedforms and particle inertial effects as summarized in Hill et al. (1988). Glenn and Grant (1987) proposed a correction for sediment-induced stratification of the near-bed flow through an eddy diffusivity scheme. This correction may be important for coarse sediment, fine to medium sands, during high energy flows. There is, however, no conclusion for the value of $\beta$ (Hill et al., 1988). When we assume a neutral flow, this constant can be taken as unity (Vincent and Green, 1990). We assume further that eddy diffusivity is identical in all size classes for simplicity. Then, as the similar form of eddy
viscosity (3-3), eddy diffusivity becomes

\[ v_s = \kappa u_\infty c \quad \text{for} \quad z < \delta_{wc} \]

\[ = \kappa u_\infty c \quad \text{for} \quad z > \delta_{wc} \]

(3-30)

Assuming that \( w_{yi} \) is a constant and integrating (3-29) with the condition of zero vertical flux through upper boundary results in

\[ \frac{dC_{i}}{dz} = -\frac{w_{yi}}{v_s} C_i \]

(3-31)

The boundary condition of (3-31) is taken as the reference concentration of the \( j \)th size class, \( C_{r,i} \), at a certain height above bed. The mean reference concentration is determined following the Smith and McLean (1977), which reads

\[ C_{r,i} = \int_0^1 C_b \frac{Y_0 S_i}{1 + Y_0 S_i} \]

(3-32)

where \( C_b \) is the volume concentration of the sediment in the bed, \( Y_0 \) is the resuspension coefficient, and \( S_i \) is the mean normalized excess skin friction shear stress of the \( j \)th size class. It may be appropriate to take the reference height \( z_r \) as the bedload layer thickness, so that from (3-17) and (3-19) we have

\[ z_r = k_{tm} \]

(3-33)

Following Madsen et al. (1994), \( S_i \) can be obtained from the instantaneous normalized skin friction, \( \frac{\left| \tau' \right|}{\tau_{cr,i}} \), where \( \tau' \) is the instantaneous skin friction shear stress and \( \tau_{cr,i} \) is the critical shear stress of the \( j \)th size class. For wave-dominated storm conditions, the skin friction shear stress is assumed
to be dominated by wave motion in the suspension of bottom sediment. For this reason, 
\[ |\tau^r_{gr}| = \tau_{\text{wem}}' \cos \omega t \]. After averaging the instantaneous normalized skin friction over a wave cycle, one obtains

\[ S_i = \frac{2}{\pi} \left( \frac{u_{\text{wem}}'}{u_{\text{wcr},i}} \right)^2 - 1 \quad (3-34) \]

in which \( u_{\text{wem}}' \) is the shear velocity based on the maximum skin friction shear stress and \( u_{\text{wcr},i} \) is the critical shear velocity of the \( i \)th size class.

With two-layer linear eddy diffusivity given by (3-30) and the boundary condition given by (3-32), Madsen et al. (1994) gave the solution for the elevation-dependent time-averaged concentration as follows

\[ C_i = C_{r,i} \left( \frac{z}{z_r} \right)^{-\frac{\psi_{fi}}{\nu_{\text{wcr}}} - \frac{\psi_{fl}}{\nu_{\text{wcr}}}} \quad \text{for } z < \delta_{\text{wcr}} \quad (3-35) \]

and

\[ C_i = C_{r,i} \left( \frac{\delta_{\text{wcr}}}{z_r} \right)^{-\frac{\psi_{fl}}{\nu_{\text{wcr}}} \left( \frac{z}{\delta_{\text{wcr}}} \right)^{-\frac{\psi_{fi}}{\nu_{\text{wcr}}}}} \quad \text{for } z > \delta_{\text{wcr}} \quad (3-36) \]

### 3-4-2. Bedload Transport

For the bedload sediment transport rates of the individual size classes, Sleath's (1978) formula was adopted because it expresses explicitly the grain size parameter. This allows fractional sediment transport to be easily obtained. The mean transport rate \( q_{bl,i} \), averaged over half a wave cycle, can be written by
\[ \frac{q_{k,i}}{\omega D_i^2} = f_i \cdot 47 \left( \psi'_{m,i} - \psi_{cr,i} \right)^{1.5} \]  \tag{3-37}

where \( \psi_{m,i} \) is the maximum skin friction Shield parameter and \( \psi_{cr,i} \) is the critical Shield parameter for the \( i \)th size class \( D_i \), respectively. It is noted that equation (3-37) is not a net transport rate which will be zero due to wave symmetry. Here, the mean transport rate during half a wave cycle is of theoretical interest in the following way.

We must convert the dimensions of the bedload transport rate (cm\(^2\)/sec) to volume of the sediment per unit area (cm) in order to add it to suspended sediment concentration which has a unit of volume per unit area. Then, we can calculate the total amount of eroded sediment. To do this, it is necessary to define a mean velocity for the bedload material velocity. Dividing a mean transport rate by this velocity, the bedload volume can be estimated. Some studies (Madsen, 1991; Nielsen, 1993) showed that mean velocity is on the order of several times the friction velocity based on theoretical analysis. Shi et al. (1985) suggested that the mean velocity of the bedload material for the combined wave-current flow is \( 10 u_{cr}/\pi \).

In a different way, the velocity of the bedload material can be estimated using the wave-current interaction model described in the above. Since a sediment particle responds and reaches its terminal velocity in an extremely short time compared to the time scale of the mean turbulent flow (Madsen, 1991), the velocity of the near-bed flow can be used for calculating that of the bedload material. A depth-averaged, mean wave velocity during half a wave cycle can be obtained as follows

\[ \bar{U}_{BL} = \left[ \frac{z_r}{z_r - z_0} \ln \frac{z_r}{z_0} - 1 \right] \left( \frac{2}{\pi} \right) \frac{u_b}{\left[ (\ln \zeta_0 - 1.15)^2 + (\pi/2)^2 \right]^{1/2}} \]  \tag{3-38}

Here, the range of integration of the near-bed flow velocity with respect to elevation starts from the
bed surface to the height of bedload layer $z_r$.

3-4-3. Critical Shear Stress

There has been debate about critical shear stresses for bedload transport when heterogeneous sediments are involved (Parker and Klingeman, 1982; Wilcock and Southard, 1988). If mobility of all sizes is equal, the grain size distribution of a poorly sorted sediment is not important and complicated fractional transport can be simplified using a representative value of the grain size distribution (Parker et al., 1982). But some geological evidences suggest that entrainment is selective. In this study, the relationship suggested in Wilcock and Southard (1988) is adopted, which is given by

$$u_{cr,i} = u_{cr,m} \left( \frac{D_i}{D_{50}} \right)^{\alpha}$$

(3-39)

where $u_{cr,i}$ and $u_{cr,m}$ are the critical shear velocity of the $i$th size class and the median diameter of the sediment, respectively, $D_i$ is the geometric mean of the $i$th size class, $D_{50}$ is the median diameter, and $\alpha$ is 0.08. The coefficient $\alpha$ was determined from wave flume experiments using sandy sediments in which size classes range from 0.01 to 1 cm (Tanaka, 1988).

To calculate the critical shear velocity of the median diameter, the formulae presented by van Niekerk et al. (1992) are used. Approximate expressions of the Shields curve are given by the following three equations

$$\psi_{cr} = 0.1(Re_D)^{-0.3} \quad \text{for} \quad Re_D < 1$$

(3-40)
\[
\ln(\psi_{cr}) = -2.26 - 0.905 \ln(Re_D) + 0.168 \ln^2(Re_D) \quad \text{for } 1 < Re_D < 60
\] (3-41)

\[
\psi_{cr} = 0.045 \quad \text{for } Re_D > 60
\] (3-42)

In these equations, \( \psi_{cr} \) is the critical Shields number and the grain Reynolds number \( Re_D \) is given by

\[
Re_D = \frac{u_{cr} D_{so}}{v}
\] (3-43)

in which \( u_{cr} \) is the critical shear velocity that is unknown and \( v \) is the kinematic viscosity. Both \( \psi_{cr} \) and \( Re_D \) are functions of the shear velocity and sediment size. Thus equations (3-40) to (3-42) can be solved for the critical shear velocity if \( D_{so} \) is given.

3-5. Sediment Conservation Equation: One-dimensional Model

The exchange of sediment between suspension in the water, moved as the bedload and stay as the bed can be solved using the continuity equation of sediment which is a function of time and space

\[
C_b \frac{\partial \eta}{\partial t} = \frac{\partial V}{\partial t} + \frac{\partial Q_x}{\partial x}
\] (3-44)

where \( C_b \) is the bed concentration, \( \eta \) is the bed elevation that is positive downward from a reference level, \( t \) is time, \( V \) is the volume of the sediment in transport per unit area \( (\text{cm}^3/\text{cm}^2) \), and \( Q_x \) represents the sediment flux in the \( x \) direction. The added physical processes, i.e., the processes of...
bed armoring and grain size changes of bottom sediment, make it impossible to get a formal
algebraic equation of (3-44) (Holly and Karim, 1986). However, a simplification of the sediment
conservation equation is made in this section to get a physical meaning of mixed, stratified sediment
processes. The schematic representation for this purpose is shown in Figure 3-1.

The conservation equation (3-44) says that the change in the bottom elevation with time
equals the change in the volume of sediment in transport with time plus the change of horizontal flux
with space. By neglecting the spatial gradient of horizontal flux and assuming that flows reach an
equilibrium with bed stratigraphy at each time step during the sequence of time-varying flows,
integration of (3-44) over time elapsed from an initial state gives

$$C_b \eta_e = (V_s + V_b) \text{ at time } t$$

(3-45)

where \(\eta_e\) is defined as an erosion depth at time \(t\) below the initial bed surface and \(V_s\) and \(V_b\) are the
volume of sediment at time \(t\) due to time-integrated suspended load and bedload, respectively. It is
noted that the calculation of \(\eta_e\) and \(V_s + V_b\) at each time step always refers to the initial bed surface
(Fig. 3-1). The initial bed surface is set at the time of no transport and the initial erosion depth equals
to zero. Whenever there exists sediment transport, the erosion depth has a positive value. Since the
erosion depth is always positive, it is not possible to distinguish between an erosional and
depositional phase using one equation given by (3-45). This requires comparison of the erosion
depth at time \(t\) with that at the previous time \(t-1\). The amount of \(V_s\) and \(V_b\) are obtained by vertically
integrating the concentration profile. The concentration profile of the bedload sediment has not been
well established compared to that of the suspended load sediment. Equation (3-45) indicates that
erosion depth at time \(t\) is directly related to the amount of sediment in transport. The water-sediment
boundary moves up or down depending on the budget of sediment between the water column and
bed. This simplified relation seems to be sufficient to address the primary objectives of this study.
Under these assumptions we can extend the sediment conservation relationship to more realistic bottom sediments. Real sea beds exhibit nonhomogeneous sediment as well as micro structure (sedimentary structure).

When mixed sediment is considered at first, the input of bottom sediment is not a single grain diameter but a spectrum of multiple grain sizes. In this case the terms for sediment transport can be replaced by the summation of $V_{s,i}$ and $V_{b,i}$ of individual size classes, which gives

$$C_b \eta_e = \sum_{i=1}^{N} (V_{s,i} + V_{b,i})$$

(3-46)

The $\eta_e$ values calculated by (3-45) and (3-46) are not necessarily the same because the different sizes respond differently to a given flow. From these constituent equations the following statement can be made that the sorting processes during sediment transport may be reflected in the change of the size distribution in the uppermost bottom sediment. In turn, the change of the grain size distribution means the change of the grain roughness. This argument supports qualitatively the idea that the grain roughness may depend on the flow conditions.

Second, the grain roughness parameter changes as a stratigraphic unit undergoes erosional/depositional processes. As flows exhume the bottom sediment during an erosional phase, different sediment sizes become exposed at the surface. During the depositional phase, the sorting process in the water column makes deposits of graded sediment while flows experience gradually changing roughness. So the new aspect of the model developed in this study is clearly to treat a grain roughness parameter as a variable changing with flow conditions as the input variables of flow, representative current and wave parameters.

Unlike the flow variables, monitoring or measuring continuously the grain roughness is practically difficult even in a well-controlled laboratory experiment. An approach for determining the
grain roughness may be developed by further applications of the conservation equation (3-46). Two conditions should be predetermined to quantify the change in the bottom sediment. One is the initial sediment stratigraphy, which can be taken at the time of the lowest hydraulic energy. The other is to set a boundary below the water-sediment interface. Beneath this boundary, there is no change of sediment composition. To locate this boundary in the sediment column, this model adopts the mixing depth concept in which a length scale from the water-sediment interface is provided in relation to the flow and sediment properties. The layer of mixed layer is equivalent to the frequently disturbed layer caused by both physical processes due to resuspension and bedforms and biological processes. This layer is referred to the surface mixed layer (Nittrouer and Sternberg, 1981, Lu and Shen, 1986). The background and the numerical formulation of the mixing depth are given in the following section.

The conservation equation (3-46) gives sediment stratigraphy one length scale (erosion depth) that is measured from the initial bed surface (Fig. 3-1). This erosion depth accounts for the amount of sediment transferred between the water column and bed. Through the calculation of sediment transport, for example, the volume of suspended load and bedload, the erosion depth is easily determined. However, determination of a size parameter in the newly formed water-sediment interface is arbitrary in the case of stratified sediment because it depends strongly on the thickness involved in the calculation. The mixing depth concept, which provides an additional length scale beneath the surface (Fig. 3-1), is a very convenient way to close this system. Thus, it eliminates the concern regarding the lower limit of bottom sediment.

With the sub-interface imposed by a mixing depth, the system becomes divided into three layers: the water column, the mixed layer, and the bed of no motion (Fig. 3-1). Including the volume of the mixed layer (its thickness is $\delta_m$) to the conservation equation (3-46) and rearranging it as an individual equation of each size class, the sediment conservation equation can be written as
\[
\int_0^{\eta_t} f(\eta) \, C_b \, d\eta = \int_0^{\eta_t} f(\delta_m) \, C_b \, d\eta + (V_{s_d} + V_{b_d})
\]  

(3-47)

It is noted that \( \eta_t = \eta_e + \delta_m \) and is defined as total depth. In (3-47), \( f(\eta_e) \) and \( f(\delta_m) \) represent the fraction of the \( i \)th size class in the total depth and mixing depth, respectively. \( f(\delta_m) \) is what we want to know at time \( t \). The conservation equation (3-47) includes the volume of sediment transferred from bed to the water column as well as the volume of the mixed layer. The total depth, sum of the erosion and mixing depth, defines the boundary in the stratigraphy which separates the active layer from the layer of no motion. The usage of this length scale will be discussed later.

In summary, the simplified conservation equation has the potential to determine the change in grain roughness with time-varying flow conditions when we deal with real sea beds. The fractional transport and mixing depth concept should be included. From the literature dealing with sediment column mixing, armoring process and initiation of motion have been considered important. All these components should be involved in building a model to simulate the sediment conservation equation. The following section will discuss the physical meaning of these components and show how they relate within the model.
3-6. Mixing Depth and Armoring Process

3-6-1. Review of Mixing Depth Concept

The mixed layer (or the layer extending from the bed surface to the mixing depth) is defined as a layer in which bed material can be exchanged within this layer due to the hydraulic and bedform related processes. Biological processes may also alter the mixed layer. However, they are not considered in this study. Within the mixed layer all particles are susceptible to movement and exposure to the surface, so that they can be entrained by near-bed flows. The mixed layer is closely related to the armoring process in that the preferred winnowing of fine grain size fractions leaves behind the coarse grain size fractions in the surface layer, which block eroding the fine sediment below.

This definition is similar to that used in the researches of sand-gravel mixtures which are composed of river bed sediments. Early studies, including Gessler (1970) and Little and Mayer (1976), presented methods of predicting the size distribution of the static armored bed and the conditions resulting in a stable bed, based on laboratory experiments conducted in a range of flows below the critical armoring condition. Many studies regarding channel bed degradation have used a mixed layer or active layer as a significant physical process as well as a numerical tool to calculate the change of bed sediment (Lu and Shen, 1986). As noted in van Niekerk et al. (1992), there has been division of the bed into two or three layers; a top horizon of an active layer overlain a mixed layer, the mixed layer corresponding to the space of bedforms, and a lower stationary bed. Rahuel et al. (1989) discussed the physical-numerical interpretation of the mixed layer and pointed out that the mixed layer should be associated with a time scale. When the instantaneous time scale is considered, the mixed layer may be very thin as in the case of an active layer (thickness of one or two grain diameters). As the time scale increases to the time required for a bedform to traverse its wavelength, the mixed layer may increase to the space occupied by the bedform traversing downstream.
Field experiments on the vertical depth of disturbance of sand in the surf zone were initiated by King (1951). She correlated the depth of disturbance with the wave height as well as the sediment grain size and found that the greatest disturbance depth was located at the breaker point. Williams (1971) showed that the breaker height and the wave period might be the major factors controlling the depth of disturbance of sand. The depth of disturbance of sand was predicted to be 40% of the breaker height, which was one order of magnitude greater than King’s estimation. Gaughan (1978) obtained a smaller value, about 0.5 to 1% of the breaker height, and attributed it to a shorter duration of experiment compared to King’s sampling interval which lasted a semi-diurnal tidal cycle. One theoretical analysis (Madsen, 1974) suggested that the passage of nearly breaking waves could cause momentary bed failure due to horizontally acting pressure gradients in the bed. Based on the field observations using video cameras, Conley and Inman (1992) showed that a pronounced asymmetry in instantaneous sediment transport exists between the wave crest and trough under nearly breaking waves. They suggested that this is caused by wave-induced boundary ventilation which was referred to the flow normal to the bed surface resulting from the spatially varying pressure field on the bed. Flow under the crest, characterized by a more rapid boundary layer development due to normal flow into the bed, induced strong sediment mobilization. This could be a basis for the mixing depth hypothesis in wave-dominant environments.

Based on field experiments on high-energy beaches over the time scale of hours, Kraus (1985) gave an empirical equation in which the experiment-average mixing depth (surf-zone averaged) was represented by $0.027 H_b$ where $H_b$ is the significant breaking wave height. Parallel to this result, Sunamura and Kraus (1985) suggested a relationship of the mixing depth $\delta_m$ with a wave-induced maximum skin friction Shields parameter and sediment grain size. This predictive model is given as
The predicted mixing depth increased in a linear fashion with breaker height and also depended on wave period. Similarly, van Niekerk et al. (1992) related the mixing depth from unidirectional flows to skin friction Shields parameter and sediment grain size. However, the mixing depth coefficient was equal to 2 and this small value compared to (3-48) from other studies may result from different mechanisms for mixing processes between surf zone and river channel. The mixing processes in the surf zone are greatly influenced by resuspension induced by wave action as well as effects of wave-induced boundary ventilation.

For modeling mixing depth on the continental shelf, a few empirical relationships were proposed but they have not been well established. Some researches used one or two grain diameters as the mixing depth (Lyne et al., 1990; Shi et al., 1985). Kachel and Smith (1986; 1989) proposed that the mixing depth is related to bedform geometry and that it should be a function of ripple height. Wiberg et al. (1994) modified Kachel and Smith's (1986) formulation in terms of the ripple length \( \lambda_r \) and the bedload transport rate \( q_b \) and applied to all sediment transport stages including sheet flow.

\[
\delta_m = \frac{q_b}{C_b} \frac{T}{\lambda_r} + \delta_b
\]  

(3-49)

They also added a background mixing depth \( \delta_b \) in the formula based on the fact that the availability of sediment for entrainment always constrains any sediment transport stage. The background mixing depth is useful when flow conditions are so weak that there is no bedload transport. After bedload transport begins, it may be reasonable to drop the background mixing depth. However, the background mixing depth can be generalized as a layer one grain diameter thick that can support the
availability of sediment at any time.

The bedload transport rate adopted in (3-49) was proportional to \((\tau' - \tau_c)^{1.5}\). Thus (3-49) can be converted to the form for mixing depth as a function of nondimensional Shields parameter, \((\psi'_m - \psi_c)^{1.5}\), in analogy to the expressions of Sunamura and Kraus (1985) and van Niekerk et al. (1992). Since the power factor is 1.5, the mixing depth increases in a nonlinear fashion with \(\psi'_m\). In other words, the rate of increases in the mixing depth increases with increasing wave height. This contradicts to Sunamura and Kraus’ experimental results in which the rate of increases in the mixing depth is reduced at high energy flows, i.e., the power factor gets less than unity. However, the relationship between \(\delta_m\) and \(\psi'_m\) on the inner shelf has not been tested due to lack of data. Thus, a formula to relate the mixing depth linearly with \(\psi'_m\) is adopted in this study for simplicity, although the power factor other than unity may be appropriate for the inner-shelf mixing depth.

The mixing depth for the inner shelf was proposed as follows

\[
\delta_m = C_\delta D_s (\psi'_m - \psi_c) + D_s
\]

(3-50)

where \(C_\delta\) is the mixing depth coefficient given as 81.4 and the second term is the background mixing depth equal to one grain diameter. The value of \(C_\delta\) for the inner shelf is not available because of lack of empirical relationships and field measurements. Since mechanisms of mixing processes on the inner shelf are different from those of surf zone, the value of \(C_\delta\) on the inner shelf may be smaller than that in the surf zone. Sherman et al. (1994) argued that the predictability of the empirical equation (3-48) might lead to an underestimation on low-energy reflective beaches where significant wave heights are less than on high-energy beaches such as the field sites of Kraus (1985). The rapidly migrating bedforms observed on low-energy beaches increased the mixing depth significantly. So equation (3-48) may present moderate or underestimated surf-zone mixing depth. Therefore, this mixing depth coefficient from (3-48) may not lead serious errors in predicting the
inner-shelf mixing depth if we apply it to a limited condition of inner shelf such as storm.

The biological mixing is different in nature because it is mainly related to a secondary process for the primary bed. Biological mixing is a post depositional process of reworking. This is important in the study of Nittrouer and Sternberg (1981), where the degree of physical sediment mixing over a long time scale (more than an event scale) was related to the degree of biological activity and the intensity of the sediment recycling process. The time scale of hours associated with the physical mixing process is, in some cases, comparable to that of biological mixing (Myers, 1977; Grant, 1983); however, the biological mixing contributes to the physical process is unknown and difficult to quantify. Since this study is focussed on the objective of determining how a nonhomogeneous and stratified sediment column affects sediment transport on an event time scale, the neglect of a possible biological mixing will not affect the outcomes.

3-6-2. Availability of Sediment

Two length scales, mixing depth and total depth, are defined in Figure 3-1. In this section the physical meaning of these two parameters is discussed in association with the availability of sediment. We can see that the sediment size distribution of each layer, which is denoted as \( f(\delta_{m}) \) and \( f(\eta_{s}) \) respectively, will be different. Here the question appears to be which grain size distribution is appropriate in order to support the grain size distribution in suspension if the model deals with a mixed sediment. When a single grain size is considered, there is no problem using the resuspension relationship which relates the reference concentration of suspended sediment to the bed concentration (Smith and McLean, 1977). However, in the case of mixed suspended sediment which is at equilibrium state with the changed bottom sediment, the grain size distribution of the suspended sediment is different from that of the bottom sediment. In an extreme case, if the surface armored bed has no available sediment in some finer size classes, then these size classes will contribute no
material to the suspended sediment concentration because of their zero bed concentration.

One way to overcome this problem is proposed here, assuming that the gradient of horizontal flux is negligible in the conservation equation (3-45). For bottom sediment we use the size distribution of the mixed layer $f_1(\delta_m)$. However, either the size distribution of the mixed layer $f_1(\delta_m)$ or the size distribution of total depth $f_1(\eta_t)$ can be used in sediment transport calculations (Table 3-1). $f_1(\delta_m)$ is related to a coarsened mixed layer. $f_1(\eta_t)$ includes sediment in a mixed layer as well as sediment eroded from the bed. Since fine-grained sediment is abundant in suspended sediment, the mean sediment size of the total depth is smaller than that of the mixing depth. For suspended sediment concentration calculations, it seems more reasonable to use $f_1(\eta_t)$ because $f_1(\eta_t)$ includes suspended sediment as well as bottom sediment. This means that $f_1(\eta_t)$ should be applied to the size fraction term $f = f_1(\eta_t)$ in equation (3-32) in order to avoid zero reference concentration caused by an armored bed. Since bedload sediment is not permanently separated from the bottom sediment, $f_1(\delta_m)$ should be retained for bedload transport calculations, i.e., $f = f_1(\delta_m)$ in equation (3-37). In the similar way, the above argument can be applied to critical shear stress.

It is speculated that $f_1(\delta_m)$ changes with time-varying flows. According to this time-varying $f_1(\delta_m)$, the relevant parameters shown in Table 3-1 will vary with time. However, $f_1(\eta_t)$ is not changing with time specially during an erosional phase. So this will be the same as the average size distribution of input bed stratigraphy, i.e., the availability of sediment is constant during the erosional phase. This is because the bed stratigraphy is assumed as a simple massive bed. If we use a complicated bed stratigraphy, $f_1(\eta_t)$ will change at any time. Although there is no change in $f_1(\eta_t)$ during the erosional phase, the calculation of suspended sediment concentration is influenced by $f_1(\delta_m)$. The time-varying bottom sediment affects the parameter $S$ in equation (3-32) as well as the shear velocity and roughness parameters in equations (3-35) and (3-36). Therefore, suspended sediment concentrations are calculated based on the changing bottom sediment process.
Table 3-1. The relationship between sediment transport parameters and the fractions of the
ith size class, \( f(\delta_m) \) and \( f(\eta_l) \).

<table>
<thead>
<tr>
<th>parameter</th>
<th>( f(\delta_m) )</th>
<th>( f(\eta_l) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>skin friction</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>grain roughness</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>availability of sediment</td>
<td></td>
<td></td>
</tr>
<tr>
<td>for suspended load</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>for bedload</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>critical shear stress</td>
<td></td>
<td></td>
</tr>
<tr>
<td>for suspended load</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>for bedload</td>
<td>X</td>
<td></td>
</tr>
</tbody>
</table>
When a value of total depth is given in a stratified bed, the fraction of the \(i\)th size class in the total depth can be expressed by

\[
f_i(\eta) = \left( \sum_{k=0}^{K} \delta_k f_i(\delta_k) \right) / \eta_t
\]

(3-51)

in which \(K\) is the number of the layers involved in the total depth, \(f_i(\delta_k)\) is the fraction of the \(i\)th size class in the \(k\)th layer and \(\delta_k\) is the bed thickness of the \(k\)th layer. Here the total depth includes a layer which is contributed from suspended sediment concentration. We get the size distribution of this total depth when \(f_i(\eta_t)\) is calculated for all size classes. The median diameter \(D_a\) related to the total depth can be calculated with the method described in Chapter 2. Using the sediment transport model, the erosion depth \(\eta_e\) can be obtained by solving equation (3-46), which is given by

\[
C_b \eta_e = \sum_{i=1}^{N} \frac{q_{h,i}}{U_{BL}} + \sum_{i=1}^{N} \left( \int_{z_r}^{h} C_i \, dz \right)
\]

(3-52)

where \(h\) is the water depth. The second term of the right-hand side in (3-52) can be replaced by the analytical solution for the depth-integrated suspended sediment concentration giving

\[
\int_{z_r}^{h} C_i \, dz = C_{r,i} \frac{z_r}{p+1} \left( \left( \frac{\delta_{wc}}{z_r} \right)^{p+1} - 1 \right) + C_{r,i} \left( \frac{\delta_{wc}}{z_r} \right)^{p} \frac{\delta_{wc}}{q+1} \left( \left( \frac{h}{\delta_{wc}} \right)^{q+1} - 1 \right)
\]

(3-53)

in which
The size distribution in the mixed layer is simply obtained by

\[ p = -\frac{w_{q,1}}{Ku_{oc}} \]

\[ q = -\frac{w_{q,2}}{Ku_{oc}} \]  

(3-54)

in which \( f(\eta_t) \) and \( f(\eta_e) \) are the fractions of the \( i \)th size class in the total depth and the erosion depth, respectively. Based on the estimated size distribution of the mixed layer, a new sediment size and mixing depth are calculated. This procedure consists of iteratively updating the total depth using the calculated values of the mixing depth and the erosion depth until values converge.

In the calculation of sediment transport, an armoring process can be implemented in the above scheme based on the availability of sediment. The volume of the \( i \)th size class available for transport, \( V_{a,i} \), is given by

\[ V_{a,i} = f(\eta_t) C_b \eta_t \]  

(3-56)

and the estimated volume of sediment in transport becomes \( V_{s,i} + V_{a,i} \) for the \( i \)th size class. The volume of sediment transport must be constrained so as not to exceed the available sediment. An fractional availability index \( \Gamma_{a,i} \) of the \( i \)th size class can be defined by

\[ \Gamma_{a,i} = \frac{V_{a,i}}{V_{s,i} + V_{b,i}} \]  

(3-57)

as one of the simplest forms. As a nested loop, the fractional availability index is applied to
equations (3-32) and (3-37) as follows

$$C_{r,i} = \Gamma_{w,i} f(\eta) C_b \frac{g \sigma S_i}{1 + \sigma S_i}$$  \hspace{1cm} (3-58)

and

$$\frac{q_{b,i}}{\omega D_i^2} = \Gamma_{w,i} f(\delta_m) 47 (\psi_{m,i} - \psi_{cr,i})^{1.5}$$ \hspace{1cm} (3-59)

This loop repeats until the volume of sediment in transport becomes less than that available for transport, in other words, \( \Gamma_{w,i} > 1 \).

Although it is not related to the availability of sediment, the model assumes a 1 cm thick bed below the bed surface to predict ripple geometry. Assuming that the average ripple height is 2 cm during all flow conditions, half of this value (1 cm) can be an average space occupied by ripples. In this way, the top 1 cm thick bed is defined to calculate the sediment size which is used to predict ripple geometry. During low flow conditions the predicted mixing depth using (3-48) is smaller than the ripple height. On the contrary, ripples disappear during high flow conditions when the mixing depth is comparable to the ripple height. The sediment size distribution in the top 1 cm thick bed is computed by (3-51) and its median diameter is calculated by the previously mentioned methods.

3-7. Solution Procedures

Several effects may result from changes in grain roughness. One is a change of the total roughness parameter due to the change of bedform geometry and bedload transport. The total roughness will affect the friction that affects flows. Eventually sediment transport will be influenced by the new flow conditions. The other effect may come from changes in the amount of sediment
available for transport. If the surface layer is covered with coarse material, it will directly influence suspended sediment concentration by preventing the entrainment of fine sediment beneath the armored layer. The change in the composition of sediment will affect the critical shear stress, skin friction, and bedload transport. As a result, these effects control the sediment transport. Again the altered sediment transport will give a new grain-size distribution to the mixed layer and thus make a feedback loop until flow and bed reach an equilibrium state.

The model is solved by an iteration method with two-level processes (i.e., inner loop and outer loop). The flow chart for the model is shown in Figure 3-2. An equilibrium state is assumed to be reached during one time step and the bed stratigraphy created in the previous time step is used in the subsequent time step (Fig. 3-3). When the bed is put to the next time step, the sediment in transport is considered to settle out to recover the initial bed surface following the method used in Harris (1994). This creates a surface layer (identical to the erosion depth) above the mixed layer (Fig.3-3). Any specific settling process is not applied, so that the surface layer above the mixed layer has the same grain size distribution as the sediment in transport.

The inner loop consists of the one-time step process and involves two loops; one updates the friction factor due to changes in grain roughness and the other is the armoring process. The result of the inner loop is a mixing depth (or other parameter such as a median diameter) when flow and bed reach an equilibrium state. The step-by-step procedures of the model are described in the following.

1) Input of the flow parameters (waves and currents) and previous bed stratigraphy (bed thickness and grain size distribution of each layer) at a time step \( t \).

2) Set an initial mixing depth as the background mixing depth and an initial erosion depth equal to zero, noting that the total depth (sum of mixing depth and erosion depth) is the same as the mixing depth.

3) Calculate the fractions of size classes and the median diameter of the mixed layer and the total
Figure 3-2. Flow chart of vertical one-dimensional model (1DV model).
Figure 3-3. Schematic representation showing bed stratigraphy formation.
depth; at the first cycle there is no difference in these parameters.

4) Calculate shear stress, skin friction, roughness parameters, and friction factor using the Grant and Madsen (1986) model and ripple roughness using the Wiberg and Harris (1993) model.

5) Calculate the volume of suspended load and bedload sediment.

6) Compare the potential of sediment entrainment to the availability of the sediment imposed by the total depth and if there is a shortage of sediment availability in any size class, then adjust reference concentration to have the potential of sediment transport equal to the amount of the sediment available for transport (armoring process).

7) Calculate the erosion depth and new mixing depth based on the new size distribution in the mixed layer.

8) Compare new mixing depth with the previous one, and until convergence, repeat procedures from (3) through (8).

After finding a mixing depth, the procedure goes into the outer loop in which the new bed stratigraphy (at time \( t \)) is made using the bed stratigraphy at time \( t-1 \). Figure 3-3 illustrates the method used to construct the new stratigraphy. Depending on the result of the comparison between the new erosion depth and the previous one, the procedure is diverted into an erosional phase or into a depositional phase. The first layer from the initial bed surface (L1) is made by settlement of the volume of sediment in transport. The second layer (L2) includes mixing depth at time \( t \). Both L1 and L2 consist of the total depth, \( \eta_{T,T} \), and there is no difference between the erosional phase and the depositional phase in constructing these two upper layers. At this point, the total depth plays a significant role in the construction of the third layer (L3). During the erosional phase, the sediment from the previous bed above the total depth at time \( t \) is transferred and redistributed into layers L1 and L2. Below the total depth in the previous bed, one layer may be a partial remnant. This remnant is put into L3 in the new stratigraphy. The layers below L3 are identical to those of the previous bed.
During the depositional phase, the total depth will be located somewhere in the mixed layer of the previous bed. In this case, the layer $\eta_{k,T-1} - \eta_{k,T}$ in the mixed layer is transferred into L3. When the total depth is located above the bed surface, not shown in Figure 3-3, the whole mixed layer and the layer $\eta_{k,T-1} - \eta_{k,T}$ are put into L3 and then a mixing process is applied to make a homogeneous layer. After this procedure, the new bed stratigraphy is used as an input for the next time step.

The common operations in the procedure to update bed stratigraphy are; (1) the determination of the sediment size distribution in a given bed thickness, (2) the parallel transformation of a single layer or a group of layers, (3) the division of a layer, and (4) the addition of two layers. The first operation is performed using equation (3-51). The second and third operations are straightforward and there is no change in the sediment size distribution. The number of layers in the bed stratigraphy increases due to the second operation. However, the fourth operation eliminates one layer and involves the mixing process which is similar to the first operation. Frequent usage of the division and addition operations can cause physically unrealistic thin layers. An arbitrary operation, basically operation 4, can be implemented to eliminate the thin layers. In the model developed here, the minimum thickness of a bed layer is 0.01 cm.
3-8. Results and Discussion

Input parameters describing flows include wave characteristics and a current specification including current velocity magnitude and direction at a reference height for use by the Grant and Madsen (1986) model. Input parameters describing the bed are rather complicated because the model deals with multiple grain size classes as well as bed stratigraphy. Since bed stratigraphy is characterized by the thickness and the grain size distribution of the bed layers, we need a two-dimensional data (matrix): one dimension represents the bed thickness and the other represents the grain size distribution. As a simple form, the bed stratigraphy presented in Figure 2-8 was used.

The resuspension coefficient $\gamma_0$ was the only tuning parameter to adjust the calculation of suspended sediment concentration. The resuspension coefficient was iteratively adjusted until the model predictions of suspended sediment concentration agreed with field measurements from 10 to 20 October 1994 at the 20 m depth (Fig. 3-4). It was possible to find a good agreement between model predictions and field measurements using $\gamma_0 = 0.001$. However, it was not possible to have good agreement between the measured and predicted suspended sediment concentration at the 12 m depth (Fig. 3-5). Especially predicted suspended concentrations were as twice as the measured concentrations at higher three elevations during the storm peak. The poor agreement at the 12 m depth may be attributed to the quality of the inshore OBS measurements as mentioned in Chapter 2.

The resuspension coefficient $\gamma_0$ is the essential parameter in predicting suspended sediment concentration (see equation 3-32). This parameter has been estimated from many field experiments and showed variability over several orders of magnitude. Smith and McLean (1977) estimated $\gamma_0 = 2.4 \times 10^{-3}$ using field data from the Columbia river. The estimated resuspension coefficients on the California shelf ranged from $1.5 \times 10^{-5}$ to $3 \times 10^{-4}$ and these values were inversely proportional to excess bed shear stress (Drake and Cacchione, 1989). They suggested two possible mechanisms for this decreasing trend of $\gamma_0$ values with excess bed shear stress. (1) Due to bed armoring, bed
Figure 3-4. Predicted suspended sediment concentrations (solid line) are compared with field measurements (plus signs) at 20 m depth. The resuspension coefficient is $\gamma_0 = 0.001$. 

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 3-5. Predicted suspended sediment concentrations (solid line) are compared with field measurements (plus signs) at 12 m depth. The resuspension coefficient is $\gamma_0 = 0.001$. 

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
concentration $C_b$ may not be a constant. Since fine-grained sediment is winnowed into suspension
during early low-energy to medium-energy flow conditions, the fine fractions are depleted in the
surface layer of bed during high energy flows. This change in the bed concentration can lead to low
$\gamma_0$ values. (2) The grain cohesiveness beneath the loose surface sediment can cause change in critical
shear stress and erosion rate. This may also lower the $\gamma_0$ values. Vincent et al. (1990) showed the $\gamma_0 - S$
relation similar to that suggested by Drake and Cacchione (1989) but the estimated $\gamma_0$ values were
an order of magnitude larger. They suggested that the variation of $\gamma_0$ was related to wave-induced
ripples. Kim (1991) and Li et al. (1996) also showed that the strong vortex activity near the seabeds
due to wave-generated ripples caused the resuspension coefficient to have a higher apparent value
under low-energy fair-weather conditions. Under storm conditions, ripples were washed out with the
increase of bed shear stress and this reduced vortex activity, resulting in lowering $\gamma_0$ values to order
of $10^{-4}$. Under an extreme storm at the Duck site during the 1991 Halloween storm, Madsen et al.
(1993) obtained $\gamma_0 = 4 \times 10^{-4}$ for sheet flow conditions.

Hill et al. (1988) suggested a constant resuspension coefficient $\gamma_0 = 1.3 \times 10^{-4}$ based on
laboratory experiments and argued that the $\gamma_0 - S$ relation shown in most field experiments may
result from improper treatment of the eddy diffusion coefficient or measurement errors. As Xu
(1993) noted, the effects of stratification and multiple sediment composition in suspended sediment
concentration complicated the inverse calculation of the resuspension coefficient using field
measurements even if he assumed a constant resuspension coefficient. In this study, bed
concentration for an individual size class is not a constant as that mentioned in the above field and
laboratory experiments in the above. This does not allow a simple $\gamma_0 - S$ relation because both $f_c C_b$
and $S_j$ in equation (3-32) are variables in association with flow conditions. Thus, the solution
procedure have to use a preset $\gamma_0$ value. Since $f_c C_b$ depends on bed stratigraphy of the previous time
step, it is difficult to calculate inversely the resuspension coefficient at a certain time step using
concentration measurements. Therefore, we assume that $\gamma_0$ is a constant over flow conditions but consider $f_i C_b$ as a variable. This may suggest that deviations between model predictions and field measurements of suspended sediment concentrations are attributable to this constant resuspension coefficient. The iteratively estimated $\gamma_0$ value as the tuning parameter in the model is an order of magnitude larger than those values obtained in storm conditions but it is close to the $\gamma_0$ value suggested by Smith and McLean (1977).

The OBS measurements from 12 to 15 October at the lower two elevations ($z = 7$ cm and 33 cm) show poor agreement with model predictions of suspended sediment concentrations (Fig. 3-4). There are several possible explanations for these deviations. First, Figure 3-6 shows predicted ripple height from 10 to 20 October at the 20 m depth. The predicted ripple height was about 2 cm at the beginning and it decreased with the increasing wave orbital velocities (see Fig. 2-6). Ripples were washed out during strongest flow conditions from 15 to 17 October. After this period, ripples were regenerated for two days as flows were weakened. They disappeared again on 19 October when wave orbital velocities increased moderately. This disappearance may be due to either the increased wave orbital diameter induced by swells or the decreased sediment size. Since the sediment size of the top 1 cm thick bed was not significantly changed (see Fig. 3-15 b) during the waning storm phase at the 20 m depth, ripples were washed out by swells. The model indicated existence of ripples during the period of poor predictions and vortex action due to ripples may have enhanced resuspension of sediment. The diffusion model used in this study probably underestimates suspended sediment concentrations because it doesn't include the vortex action due to ripples. Alternatively, the actual $\gamma_0$ value may be larger than the constant $\gamma_0$ value used in the model. A diffusion-advection model is capable to deal with the effects of vortex action on resuspension of sediment as suggested by Nielsen (1993).

A second source of the poor agreement may be that the model ignored possible contribution
Figure 3-6. Predicted ripple heights show that ripples disappear during sheet flow conditions.
caused by the spatial gradient of sediment flux in the sediment conservation equation. This will be discussed in Chapter 4. Additionally, malfunctions of OBS sensors can cause some damage to the data quality. As shown in Chapter 2, the apparently incorrect data in Figure 2-8 b were removed. There may exist some drift in the truncated OBS data which may start at an early stage of the storm. The lowest OBS sensor, positioned at 7 cm above the bed, shows obviously a sudden jump of OBS data right after 12 October while the OBS data of other four sensors change little (Fig. 2-8 b). This indicates the small drift of the lowest OBS sensor. Considering overall time series of the OBS data, all sensors have some drift problem. The upper two OBS sensors (sensor 4 and 5) appear to have smaller drift than the lower three OBS sensors (sensor 1 to 3). This may result in better agreement between model predictions and field measurements of concentrations in the upper two OBS sensors than the lower sensors (Fig. 3-4). However, these small drifts are difficult to remove because these are not explicitly separated from the correct signal.

An interesting feature of the model predictions in Figures 3-4 and 3-5 is the asymmetric variation of the suspended sediment concentration in response to the more or less symmetrical variation of flow conditions with time (see Fig. 2-6). Suspended sediment concentrations during the erosional phase (October 10-16) were smaller than those during the depositional phase (after October 16). Unfortunately, the truncated OBS measurements could not support the behavior of the model predictions during the waning storm phase. This kind of behavior of the suspended sediment concentration was observed in other field experiments (Shi et al., 1985) and interpreted as hysteresis effects which were related to time-dependent processes. Wiberg et al. (1994) noticed that steady models can underestimate suspended sediment concentrations during waning flows because the sediment suspended at peak flow conditions is neglected between calculation time steps. They suggested that time-dependent effects should be considered in steady models to improve model predictions. The model developed here satisfies two conditions for steady models as described in
Wiberg et al. (1994): (1) suspended sediment concentrations are in equilibrium with the imposed flow and (2) the bed is well-mixed at all times. The difference of this model from most steady boundary layer models is to take care of the contribution of sediment suspended at high flow conditions through the total depth $\eta$, which is related to the sediment availability for resuspension. Also, the armoring process may be responsible for this model behavior rather than time-dependent processes. During an erosional phase, the amount of sediment in suspension is less than the capacity of the flow to suspend sediment because the coarse-grained armored layer prevents the erosion of the fine-grained sediment. This means that the bed controls the suspended sediment concentration during the resuspension period. On the contrary, the armoring process ceases during the depositional phase and thus the capacity of the flow to suspend sediment plays an important role in governing the suspended sediment dynamics. Therefore, flows reaching equilibrium with resuspended sediment can support higher concentrations.

The model also predicted the shear velocity due to the current above the wave boundary layer and the apparent roughness. The comparisons of the predicted values with the field measurements of these parameters are presented in Figure 3-7 and Figure 3-8. The field measurements of both shear velocity and apparent roughness at the 20 m depth agree well with model predictions (Fig. 3-7). The field measurements of these variables at the 12 m depth showed very poor agreements with the predicted values (Fig. 3-7). Again, the poor quality of field measurements at the 12 m depth was responsible for discrepancies mainly because only three measured values were involved in the calculation of the velocity profile.

Comparison of the predicted $u_*$ and $z_m$ values between the two locations showed similar variation with time. Also, the magnitudes of predicted values from the two locations were similar. So it can be said that shear stresses due to currents were quite similar at both sites. But there was significant difference in the model predictions of suspended sediment concentrations (Fig. 3-4 and 3-
Figure 3-7. Predicted values of shear velocity (a) and apparent roughness (b) are compared with field measurements at 20 m depth. Solid lines represent the model predictions and plus signs represent the field measurements.
Figure 3-8. Predicted values of shear velocity (a) and apparent roughness (b) are compared with field measurements at 12 m depth. Solid lines represent the model predictions and plus signs represent the field measurements.
5). The predicted values at the 12 m depth location were two to three times greater than those at the 20 m depth location. Considering the higher wave-orbital velocities at the 12 m depth (Fig. 2-11), waves were the main agents agitating the bottom sediment as observed in Wright et al. (1986; 1991).

To examine the effects of coupling between the boundary layer process model and bed stratigraphy on the prediction of hydrodynamic and sediment transport parameters, a decoupled model was constructed and its predictions were compared with the coupled model predictions. The decoupled model was different from the coupled model in a few respects; it did not account for the armoring processes or changes of the bottom sediment with time. Since the armoring process is not separable from the process of changing bottom sediment from the theoretical point of view, comparison of the two models indicated combined effects of the couplings. Predicted values of suspended sediment concentrations at the 20 m depth from both coupled and decoupled models are compared in Figure 3-9. It shows that the decoupled model overestimated the suspended sediment concentration at the time of the storm peak from 15 to 16 October. Except at the time of the storm peak, both predicted values were not significantly different. The overestimation of the decoupled model is attributable to the resuspension coefficient $\gamma_0 = 0.001$ which is too high for sheet flow conditions. Adjustment of the decoupled model to use other $\gamma_0$ values is required.

To adjust the decoupled model the $\gamma_0$ value was lowered to $4 \times 10^{-4}$ which was suggested by Madsen et al. (1993) with the reference concentration calculated at the height $z = 7d_{so}$ for sheet flow conditions. As shown in Figure 3-10, except at the lower two elevations, the decoupled model overestimates at the upper two elevations during the storm peak. Predicted values of both the coupled and decoupled models are comparable during low energy conditions. In an attempt to improve predictions of the coupled model at higher elevations, the $\gamma_0$ value was lowered to $2 \times 10^{-4}$ (Fig. 3-11). But this caused underestimation at lower elevations during the storm peak. During the low energy conditions, the decoupled model predicted much smaller concentration than the coupled
model. Based on these two analyses, one may choose an intermediate \( y_0 \) value between \( 2 \times 10^{-4} \) and \( 4 \times 10^{-4} \) to compromise the adverse results caused by these two \( y_0 \) values. When the resuspension coefficient is taken as \( y_0 = 3 \times 10^{-4} \), the decoupled model overestimates at higher elevations and underestimates at lower elevations during the sheet flow conditions (Fig. 3-12). This can be seen in a concentration profile (at Hour 12 on 15 October) in which the decoupled model gives a gentle concentration profile compared with the coupled model (Fig. 3-13). As we have attempted to adjust the decoupled model using several values of the resuspension coefficient (Fig. 3-10 to 12), slopes of predicted concentration profiles will be consistent between these different values. This gentle slope may be an intrinsic nature of the concentration profile predicted by the decoupled model. This demonstrates that it is difficult to get good agreement between decoupled model predictions and the OBS data.

The underestimation at lower elevations by the decoupled model probably results from low coarse-grained sediment composition in suspended sediment because of low \( y_0 \) value. In the coupled model, however, coarse sediment is enriched on the bed surface as an armored bed is developed. This coarsened surface layer provides large amount of coarse sediment to be suspended. This contributes high concentration at lower elevations but low concentration at higher elevations because the settling velocity of the coarse sediment is large. This results in a steep concentration profile. Contrary to the coarse sediment situation, fine-grained sediment has small settling velocity and gives a gentle concentration profile. Fine-grained sediment contributes more effectively to suspended sediment at higher elevations than coarse sediment. Since there is no constraint for the erosion of fine-grained sediment in the decoupled model, the bed supplies large amount of fine-grained sediment to the water column. Therefore, the overestimation at higher elevations is likely attributable to fine-grained sediment as the degree of overestimation increased at higher elevations. However, an armored layer limits the supply of fine-grained sediment and results in low concentration at high elevations in the
Figure 3-9. Suspended sediment concentrations predicted by coupled model (solid line) are compared with those predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.001$. Plus signs represent the OBS data at 20 m depth.
Figure 3-10. Suspended sediment concentrations predicted by coupled model (solid line) are compared with those predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.0004$. Plus signs represent the OBS data at 20 m depth.
Figure 3-11. Suspended sediment concentrations predicted by coupled model (solid line) are compared with those predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.0002$. Plus signs represent the OBS data at 20 m depth.
Figure 3-12. Suspended sediment concentrations predicted by coupled model (solid line) are compared with those predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.0003$. Plus signs represent the OBS data at 20 m depth.
Figure 3-13. Suspended sediment concentration profile predicted by coupled model (solid line) is compared with that predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.0003$. Plus signs represent the OBS data at 20 m depth. This shows that the coupled model predicts suspended sediment concentrations better than the decoupled model.
This implies that the armoring process was critical to obtain a correct gradient of concentration profile. In addition, it is noteworthy that the coupled model predicted higher suspended sediment concentrations during the waning phase of the storm than the decoupled model.

In contrast to the difference presented in Figure 3-9, the predicted values for shear velocity and apparent roughness from both models were too close to allow the effects of couplings on the hydrodynamic parameters to be differentiated (Fig. 3-14). Small grain size changes may result from the sediment having a relatively narrow range of grain size distribution (i.e., well sorted). In other environments such as a river channel, bottom sediment is composed of sand-gravel mixtures which show a much wider range of grain size distribution and multiple modes. In that condition, grain size changes in the bed surface can be anticipated to affect significantly the shear stress and roughness parameters (Kuhnle, 1989). Apparently, the hypothesis that the changing bottom sediment affects the hydrodynamics may not be useful for the boundary layer processes in inner-shelf environments. Instead, these effects were most profound in sediment transport. It implied that the model is still useful. Small changes in the sediment size in the armored layer were very effective on sediment transport but the shear stress terms. It can be said that a model showing good agreement with the hydrodynamics doesn’t necessarily provide good predictions of sediment transport if it is not coupled with bed stratigraphy. The fact that hydrodynamic variables can be accurately estimated without couplings is very helpful in designing a complicated hydrodynamic model. A partly decoupled model in the calculation of the hydrodynamics avoids time-consuming procedures.

In the following, model outputs, which have no field data to compare with, are presented. As mentioned earlier, the acquisition of bed characteristics in field experiments is not easy. So there is no way to calibrate and validate the model outputs using field data. Nevertheless, the validity of model predictions of bed stratigraphy can be evaluated indirectly by comparison of the model predictions of suspended sediment transport with field measurements on the basis that the model
Figure 3-14. Shear velocity (a) and apparent roughness (b) predicted by coupled model (solid line) are compared with those of decoupled model (dotted line).
couples the boundary layer processes with bed stratigraphy. As mentioned previously, a model coupled with bed stratigraphy appears to be superior for the prediction of suspended sediment concentration. For this reason, it is proposed that the predicted bed stratigraphy may be valid, at least qualitatively.

The model predictions of the length scales, erosion depth and mixing depth, and the grain size parameters at both the 10 m and 20 m depths are presented in Figure 3-15 and Figure 3-16 respectively. The predicted length scales at both locations showed the same patterns that were closely related to wave orbital velocities. The maximum erosion depth at the time of the storm peak was 0.5 cm at the 20 m depth and 2 cm at the 12 m depth. The increased erosion depth at the 12 m depth reflected increased wave-current interactions. The time series of sediment size at both locations showed the tendency that sediment size reached a maximum value during the storm peak and then decreased to be finer than the initial values (Fig. 3-16). The maximum sediment size in the armored layers at both locations was 2.5 $\phi$ and 2.8 $\phi$ respectively. Some differences between the two sites occurred during the depositional phase. The sediment size in the top 1 cm thick bed at the 20 m depth did not change much after the model run but at the 12 m depth it decreased from 3 $\phi$ to 3.5 $\phi$. Since the sediment became finer after the storm event at the 12 m depth, small ripples were expected. This demonstrates that model predictions of sediment size were consistent with observations described in Wright (1993). It seems that the model predicts qualitatively the roughness changes of seabeds in response to storm events.

In Figure 3-17, the model predictions of the erosion depth using the decoupled model were compared with values obtained using the coupled model when both models were adjusted by same $\gamma_0$ value, $\gamma_0 = 0.001$. At the time of the storm peak, the erosion depth predicted by the decoupled model was 3 cm which is six times greater than that predicted by the coupled model. During low energy conditions, both models predicted comparable results. When the resuspension coefficient of the
Figure 3-15. Model predictions of bottom sediment length scales and sediment size at 20 m depth. (a) shows the erosion depth (cross signs) and the total depth (star signs). (b) shows the median diameter of the mixed layer (circle signs) and that of the top 1 cm thick bed (plus signs).
Figure 3-16. Model predictions of bottom sediment length scales and sediment size at 12 m depth. (a) shows the erosion depth (cross signs) and the total depth (star signs). (b) shows the median diameter of the mixed layer (circle signs) and that of the top 1 cm thick bed (plus signs).
Figure 3-17. The erosion depth predicted by coupled model (solid line) is compared with that predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.001$. 

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 3-18. The erosion depth predicted by coupled model (solid line) is compared with that predicted by decoupled model (dashed line) when the $\gamma_0$ value of the decoupled model is $\gamma_0 = 0.0003$. During high energy flows, the decoupled model predicts the erosion depth as twice as the coupled model. This implies that the decoupled model has the potential to overestimate sediment transport.
decoupled model was lowered to $3 \times 10^{-4}$, the erosion depth predicted by the decoupled model was as twice as the coupled model during the storm peak (Fig. 3-18). Since the erosion depth equals to the depth-integrated suspended sediment concentration, a steep concentration profile gives smaller value of the erosion depth. The concentration profile calculated by the coupled model is steeper than that calculated by the decoupled model (Fig. 3-13). Therefore, the decoupled model has the potential to overestimate sediment transport during storm events.

The predicted bed stratigraphy at the 20 m depth is shown in Figure 3-19. The thickness of storm deposit was predicted to be approximately 1.3 cm. This storm deposit showed fining upward sequences. The base of the storm deposit consisted of the coarsest material. The median diameter of the top layer was smaller than the initial sediment size. The size distributions of the individual layers in the storm deposit are presented in Figure 3-20. It shows clearly a fining upward trend in a graded bed. The layers 4 to 6 must be armored layers in that fine-grained sediment is depleted. As for the predicted bed stratigraphy at the 12 m depth, thicker storm deposits resulted from the high intensity of wave orbital velocities (Fig. 3-21). Compared with the bed stratigraphy at the 20 m depth, the base of the storm deposit was much coarser. To compensate for this coarsening, the top layer of the storm deposit contained fine-grained sediment with the sand size fractions depleted, as seen in Figure 3-22.

The observed bed stratigraphy is shown in Figure 2-4. Since this sediment core was analyzed at 1 cm intervals, precise structures within the 2 cm thick layer which indicates fining upward trend was not provided. From this surface layer to about 5 cm depth of the core, sediment shows coarsening upward trend. The thin surface layer showing a fining upward trend is attributable to a storm event. Variations below the surface layer may be related to long term processes including bioturbation. In order to compare bed stratigraphy predicted by the model developed here with field observations, it is necessary to collect sediment carefully focusing on the upper few centimeters of
Figure 3-19. Predicted bed stratigraphy at 20 m depth. Thickness of the predicted storm deposit is about 1.3 cm. The storm deposit shows a fining upward trend.
Figure 3-20. Predicted grain size distributions of individual layers in the predicted bed stratigraphy at 20 m depth. Each layer in the bed stratigraphy is denoted by the numbers on the vertical axis. Number 1 represents top layer. Size distribution of layer number 7 equals to that of initial bed stratigraphy.
Figure 3-21. Predicted bed stratigraphy at 12 m depth. Thickness of the predicted storm deposit is about 3.5 cm.
Figure 3-22. Predicted grain size distributions of individual layers in the predicted bed stratigraphy at 12 m depth. Each layer in the bed stratigraphy is denoted by the numbers on the vertical axis. Number 1 represents top layer. Size distribution of layer number 8 equals to that of initial bed stratigraphy.
bottom sediment. This sediment should not be disturbed to preserve micro-scale structures. Also, the
time to collect sediment samples after storm events is important because these are temporal
structures. Without these careful controls, comparisons between predicted bed stratigraphy and
observed sediment core will not be robust.

3-9. Limitations of the Model

Several assumptions underlie the model and are sources of weakness when applying the
model to realistic field conditions. In the derivation of the sediment conservation equation, the
horizontal flux term was dropped. This is not always justified in the inner-shelf environment where
across-shelf transport induces morphological changes during high energy events (Wright et al.,
1991). But the lack of data concerning this spatial term makes it difficult to include the horizontal
flux term.

The assumption of a steady state introduces errors regardless of the time scale and interval
used. For example, suspended sand has a relatively short response time to changes of flows because
of its large settling velocity. The response time increases as the sediment size decreases. In spite of
the fact that fine-grained sediment is a small fraction of the total composition of the inner-shelf
sediment, it is a large fraction of suspended sediment and can cause significant effects on the OBS
measurements. OBS sensors are more sensitive to fine-grained than coarse-grained sediment
(Ludwig and Hanes, 1990; Green and Boon, 1993). Predicting the behavior of the fine-grained
sediment accurately requires a time-dependent model.

The model mainly focuses on the physical processes, the relationship between the sediment
transport and bed responses, during a resuspension event. This makes it of limited applicability to the
fair-weather period when the biological mixing process dominates. So the time scale of the model is
limited to the range of hours to days which is long enough to cover a high energy event. Biological
activity may be so efficient during a calm period that the sedimentary structures of the surface layer can be destroyed in a time scale less than one day (Myers, 1977; Grant, 1983). Including fair-weather processes into a long term model such as that developed by Harris (1994) would be a virtual pursuit but beyond the scope of this study. Also, the model in this study results in one unit of bed stratigraphy, a graded bed, whose thickness is in the order of mm to cm (Niederoda et al., 1989). This range of the bed thickness is not comparable to a geological stratigraphic unit.

The wave-current interaction model of Grant and Madsen (1979; 1986) has been successfully applied to the inner-shelf environment. It is based on a time invariant eddy viscosity which increases linearly in the lower wave boundary layer as well as in the upper current boundary layer, but has a discontinuity between two layers. Consistent with this eddy viscosity profile, a Rouse type equation has been used to predict suspended sediment concentrations (Drake and Cacchione, 1989). When wave-generated ripples occupy the seabeds, it was observed that vortex ejection from the bedforms causes variation of concentration profiles (Vincent et al., 1991; Nielsen, 1993). They deviate from the predictions of the simple diffusion model.

Another limitation is that the model assumes noncohesive sediment and cohesive properties are assumed to be minor. In cohesive sediment transport, thresholds of erosion are different from those of deposition. However, this differentiation in noncohesive sediment is not taken into account because erosion and deposition occur at the same time (Sanford and Halka, 1993). Combining these two processes may be necessary in a time-dependent numerical model. So it is reasonable to apply the model developed in this chapter to an environment where cohesive properties in the sediment are not significant.
4. ACROSS-SHELF TRANSPORT: 1DH DEPTH-RESOLVED MODEL

4-1. Introduction

Across-shelf sediment flux on the inner continental shelf involves numerous processes (Swift et al., 1985; Wright et al., 1991). Tide-induced and wind-driven currents are relatively more important on the inner shelf than in the surf zone. The significant role of mean currents for advective transport has been recognized. For engineering applications within surf zones, Dean (1991) applied simple equilibrium beach profile concepts to profile changes. Equilibrium profiles were constructed based solely on considerations of wave dissipation and with the unrealistic assumption of a shallow depth of closure beyond which there is no sediment movement. During disequilibrium, cross-shore sediment transport is assumed to occur by a diffusive process and unidirectional currents are not considered (Pilkey et al., 1993). Wave-induced bed shear stress is considered to entrain sediment particles but to contribute little to advective movement. In nature, mean currents are responsible for cross-shore sediment flux although their magnitude is much smaller than that of wave orbital velocities. Therefore, to extend such simple equilibrium concepts to inner shelf without considering currents would be unrealistic.

It has been noted that the direction of sediment transport depends strongly upon mean currents (Wright et al., 1991). Also, asymmetrical wave orbital velocities promote shoreward mass transport (Swift et al., 1985). With a sloping plane bed, equilibrium is attained when there is a balance between onshore sediment flux due to asymmetrical orbital velocities and down-slope gravitational flux. This profile is attacked by storms which generate intensified wind-driven
circulations. On the middle Atlantic shelf, wind stress is commonly onshore during storms. The rates of offshore sediment transport due to strong downwelling mean currents can be an order of magnitude greater than onshore transport by fair-weather swells (Wright et al., 1991). Niederoda et al. (1985) interpreted a cross-shore deficit of inner-shelf sediment budget on the basis of geological time scales with a sea level rise as an important factor affecting the erosional retreat of barrier islands along the east coast of the United States.

Relative rates of cross-shore sediment transport, with regard to steady currents, incident waves, long-period infragravity waves and the gravitational component caused by a sloping bed, were obtained from a sediment transport model (Bailard, 1981) based on Bagnold's energetics model. The predicted values were not consistent with observed values except in terms of the qualitative direction of each component (Wright et al, 1991). Mean flows were found to dominate observed cross-shore sediment flux in all conditions. While high-frequency waves were predicted to be significant in all conditions, they were observed as a major mechanism for shoreward sediment transport only during swell-dominated conditions. Since energetics models are based on instantaneous velocities, it was impossible to represent the phase relationship between the maximum near bottom velocity and the maximum concentration of suspended sediment under fair-weather and moderate wave energy conditions. However, both quantities were in phase under high energy conditions, i.e., storms. Wright et al (1991) suggested that the discrepancies corresponding to low and moderate energy conditions may be attributed to the presence of wave-generated ripples.

To take the horizontal flux term into consideration, we need to know the flow field. During a storm event, wind may be the dominant force driving circulation on the inner shelf, but wave and Coriolis forces are not negligible. Here, these three terms are included to derive a simple analytical solution of the two-dimensional equation of motion following the Jenter and Madsen model (Jenter and Madsen, 1989; Chisholm, 1993; Madsen et al., 1994). In this chapter, the appropriate
hydrodynamic models are implemented in a horizontal one-dimensional model scheme (1DH model) in order to investigate the wind-driven processes for the erosion/deposition of the inner-shelf morphology. Components of this model are shown in Figure 4-1. Assuming a simple geometry in which along-shelf variation is negligible, a one-dimensional set of grids was constructed in the across-shelf direction. Based on this simple geometry, simple wave transformation is used to calculate wave fields. Using the computed waves and the input of winds, current profiles at each grid point were predicted by the Jenter and Madsen (1989) model which is coupled with a wave-current interaction model. Depth-integrated sediment flux was estimated using the predicted hydrodynamics and bed stratigraphy in association with fractional transport and armoring processes. The horizontal flux divergence was numerically solved to yield bed elevation changes related to individual size classes.

4-2. Wave Transformation Model

4-2-1. Formulation

Assuming a simple geometry consisting of straight and parallel contours, wave transformation along wave rays on the inner shelf can be predicted to the lowest order using simple linear wave theory which is described in Dean and Dalrymple (1992). Assuming that the wave energy is conserved and neglecting reflection, diffraction and frictional dissipation, the wave height \( H \), at intermediate or shallow water (at \( x = 1 \)) can be determined by

\[
H_1 = H_0 \ K_s \ K_r
\]

(4-1)

where \( H_0 \) is the wave height at deep water (at \( x = 0 \)), \( K_s \) is the shoaling coefficient and \( K_r \) is refraction coefficient. The shoaling coefficient \( K_s \) is given by
Figure 4-1. Flow chart of horizontal one-dimensional, across-shelf sediment transport model.
\[ K_r = \sqrt{\frac{C_0}{2C_{gt}}} \]  

(4-2)

where \( C_o \) is the deep water wave celerity \((L/T)\) and \( C_{gt} \) is the group velocity at \( x = 1 \). The wave numbers at both locations are calculated by Hunt’s approximate equations (2-5) and (2-6) and also,

\[ C_1 = C_0 \tanh kh \]  

(4-3)

With the wave speed \( C_j \), the group velocity at \( x = 1 \) is given by

\[ C_{gt} = \frac{C_1}{2} \left( 1 + \frac{2kh}{\sinh kh} \right) \]  

(4-4)

The refraction coefficient \( K_r \) can be determined directly provided the assumption of straight, parallel contours is valid,

\[ K_r = \sqrt{\frac{b_0}{b_1}} = \sqrt{\frac{\cos \theta_0}{\cos \theta_1}} \]  

(4-5)

where \( b_0 \) and \( b_1 \) are the distance between the wave rays at locations 0 and 1 respectively and \( \theta_0 \) and \( \theta_1 \) are the incident angles at \( x = 0 \) and 1, made by the wave ray and the across-shelf \( x \) axis. Except where contours protrude seaward (convex seaward), this refraction coefficient is always smaller than unity. The incident angle at \( x = 1 \) is calculated using Snell’s law which is

\[ \frac{\sin \theta_1}{C_1} = \frac{\sin \theta_0}{C_0} \]  

(4-6)

Considering the effects of real sea beds, wave damping due to bottom friction becomes
significant as waves propagate into shallow water and may reduce the wave height (Liu and Tsay, 1985; Maa and Kim, 1992). The average rate of the energy dissipation related to the rough turbulent boundary layer can be formulated as

\[
\varepsilon_D = \tau_b \mu_b = \frac{2}{3\pi} \rho f t \mu_b^3
\]

(4-7)

where \( \tau_b \) is the maximum bed shear stress, \( u_b \) is the rms bottom orbital velocity, \( \rho \) is water density and \( f_t \) is Jonsson’s (1966) wave friction factor. The energy flux equation leads to

\[
\frac{\partial E_C}{\partial x} = -\varepsilon_D
\]

(4-8)

Inserting \( E = (1/8) \rho g H^2 \) and equation (4-4) into (4-8), we get (4-9) after some manipulation.

\[
\frac{\partial H^2}{\partial x} = -\frac{2f_e}{3\pi g C_x} \left( \frac{\omega}{\sinh kh} \right)^3 H^3
\]

(4-9)

It is noted that (4.9) can be applied only to a flat bed. Solving (4.9) for the wave height with the boundary condition, \( H_0 \) at \( x = 0 \), we have

\[
\frac{H_1}{H_0} = K_f = \frac{1}{1 + \frac{f_e}{3\pi} \left( \frac{k^2 H_0}{2kh + \sinh 2kh} \right) } \frac{1}{\sinh kh}
\]

(4.10)

where \( x \) is the traveling distance in the wave propagation direction. The damping coefficient \( K_f \) can be included in (4-1) leading to

\[
H_1 = H_0 \frac{K_f}{K_f}
\]

(4-11)
To apply (4-10) to a gentle slope bottom, the bottom of each segment of grids is assumed to be a flat bed and then the damping coefficient $K_f$ is calculated locally along the wave ray. In addition, the distance at each segment which a wave propagates depends on the wave incident angle $\theta$ and it can be roughly expressed as $x/\cos \theta$. The effects of wave damping along a refracting wave ray are thus cumulative.

The friction factor in (4-10) is the only unknown and it must be determined using field measurements. Using Swart's formula (Swart, 1974), which is a function of the roughness $k_b (=30z_0)$ and the semi-excursion amplitude $A_b$, the friction factor $f_e$ is given by

$$f_e = \exp \left[5.213 \left( \frac{k_b}{A_b} \right)^{0.194} - 5.977 \right] \quad \text{for } \frac{k_b}{A_b} < 0.63$$

$$f_e = 0.3 \quad \text{for } \frac{k_b}{A_b} > 0.63$$

(4-12)

For simplicity, the roughness $k_b$ is considered to be a constant through the comparison with field measurements although it varies with time and space.

5-2-2. Determination of Wave Friction Factor

The FRF wave data measured at 8 m and 17 m depths were used to determine the friction factor of the wave transformation model. Wave heights during October 1994 reached 4 m and included three events where wave height exceeded 1 m (Fig. 4-2 a). Comparison of model predictions with field measurements mostly used the broad peak of wave heights from 10 to 20 October. The time series of the difference between the wave heights at the two locations (Fig. 4-2 b) appeared to be proportional to the magnitudes of the wave heights and tended to be positive except during the period from 16 to 21 October. The positive difference means that the wave height in shallow water is
smaller. Several factors may have caused the changes of the wave height along the across-shelf transect. The inner shelf is where wave transformation due to shoaling and refraction occurs and also the wave dissipation due to friction or input from a wind source is possible. These processes depend on wave and wind conditions. As discussed in Battjes et al. (1990), we need to investigate the effect of whitecapping which may be responsible for wave dissipation in an actively wind-driven wave field.

To begin with, the wave transformation model was run using (4-1), assuming that there was no dissipation of the wave energy flux. The model predictions of the wave heights were compared with the field measurements (Fig. 4-3). In general, good agreement occurred for high energy waves and long-period waves (not shown here). During the period when the wave height was less than 1 m, for example, from 5 to 10 October, wave heights were not predictable and probably depend on wind conditions. However, the wave transformation during the period of high energy waves, from 10 to 20 October, was strongly governed by the wave period. As seen in Figure 4-4, the errors between the model predictions and the field measurements (Fig. 4-3) appeared to follow the trend of wave periods rather than that of wind speeds. Dissipation of the wave energy flux, objecting the assumption, was responsible for these errors. Wave dissipation due to friction depends on the wave period. Therefore, we could apply a wave damping factor to the model to improve predictions of the wave height variations with changing depth. The simulation produced by the wave transformation model is presented in Figure 4-5, using the adjusted roughness parameter $k_b = 0.1$ cm. There has good agreement between the model predictions and the field measurements during the period of the high-energy and long-period waves.
Figure 4-2. The FRF wave heights measured at 8 m (dashed line) and 17 m (solid line) depths in October 1994. (a) shows time series of the wave heights (in the m unit). (b) represents difference (in the cm unit) between the wave heights at the two locations and this shows a tendency that the wave height at the 8 m depth is smaller.
Figure 4-3. Predicted wave height (heavy line) is compared with the FRF wave height (line with circles) when the wave transformation model runs assuming no dissipation of wave energy flux.
Figure 4-4. Errors (a) between model predictions and field measurements of wave height are compared with wave period (b) and wind speed (c).
Figure 4-5. Predicted wave height (heavy line) is compared with the FRF wave height (line with circles) when the roughness parameter $k_b$ is 0.1 cm.
4-3. Current Model

4-3-1. Formulation

The current model used in this study is described in Jenter and Madsen (1989), Poon and Madsen (1991), Chisholm (1993) and Madsen et al. (1994). The momentum equations of a depth-resolving hydrodynamic model for a steady, linear, inviscid, and homogeneous water column in horizontal $x$-$y$ plane are given by

\[ -f v = -g \frac{\partial \zeta}{\partial x} + \frac{1}{\rho} \frac{\partial \tau_x}{\partial z} \]  

\[ f u = -g \frac{\partial \zeta}{\partial y} + \frac{1}{\rho} \frac{\partial \tau_y}{\partial z} \]  

where $f$ is the Coriolis parameter, $g$ is the acceleration of gravity, $\rho$ is the water density, $\zeta$ is the sea surface elevation, $z$ is the vertical coordinate, $u$ and $v$ are the local velocity components in the $x$ and $y$ direction, and $\tau_x$ and $\tau_y$ are the turbulent Reynolds' stresses. The turbulent stress profile depends on the eddy viscosity $v_t$, and velocity shear

\[ \frac{1}{\rho} \{ \tau_x, \tau_y \} = v_t \left\{ \frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right\} \]  

and the eddy viscosity, following Madsen (1977), is

\[ v_t = \kappa u \eta \frac{\partial z}{\partial z} \quad \text{for} \quad 0 < z < z_m \]

\[ = \kappa u \eta (h-z) \quad \text{for} \quad z_m < z < h \]  

where $\kappa = 0.4$ is von Karman's constant, $h$ is the water depth, and $z_m$ is defined as
\[ z_m = h \frac{u_{sb}}{u_{sr} + u_{sb}} \] (4-17)

in which \( u_{sb} = \sqrt{\tau_b / \rho} \) and \( u_{sr} = \sqrt{\tau_s / \rho} \) are the shear velocity on the bottom and surface boundary respectively.

Inserting (4-15) into (4-13) and (4-14), then multiplying (4-14) by \( i = \sqrt{-1} \), and adding to (4-13), the equations are manipulated to a single equation in a complex form

\[
\frac{\partial}{\partial x} \left( v \frac{\partial W}{\partial z} \right) - i f \ W = 0 \] (4-18)

where

\[
W = (u + iv) - \frac{g}{f} \left( -\frac{\partial \zeta}{\partial y} + i \frac{\partial \zeta}{\partial x} \right) \] (4-19)

Jenter and Madsen (1989) gave solutions of (4-18) with eddy viscosity (4-16) as follows

\[
W = W^s = A^s \left( \text{ber } 2\sqrt{f(h-z)/k u_{sr}} + i \text{ bei } 2\sqrt{f(h-z)/k u_{sr}} \right) + B^s \left( \text{ker } 2\sqrt{f(h-z)/k u_{sr}} + i \text{ kei } 2\sqrt{f(h-z)/k u_{sr}} \right) \] (4-20)

for \( z_m < z < h \)

\[
W = W^b = A^b \left( \text{ber } 2\sqrt{f(z)/k u_{sb}} + i \text{ bei } 2\sqrt{f(z)/k u_{sb}} \right) + B^b \left( \text{ker } 2\sqrt{f(z)/k u_{sb}} + i \text{ kei } 2\sqrt{f(z)/k u_{sb}} \right) \] (4-21)

for \( 0 < z < z_m \)

where \( A^s, B^s, A^b, \) and \( B^b \) are arbitrary complex coefficients to fit the boundary conditions and ber,
bei, ker, and kei are zero-order Kelvin functions (Hildebrand, 1976).

As shown in Signell et al. (1990) and Keen and Glenn (1994), the current model can be coupled with a bottom boundary layer model which account for wave-current interactions. The boundary condition used to couple the current model with the wave-current interaction model is suggested in Madsen et al. (1994). The coupled model only differs from the Jenter and Madsen (1989) model in the respect that the apparent bottom roughness $z_{oa}$, the roughness in the presence of the wave, is used instead of $z_p$. The apparent bottom roughness (Grant and Madsen, 1986) is

$$z_{oa} = \delta_{wc} \left( \frac{z_0}{\delta_{wc}} \right) u_{wb} / u_{wb}$$

(4-22)

in which $z_0$ is the physical bottom roughness and $\delta_{oa}$ and $u_{wb}$ are the wave bottom boundary layer thickness and the shear velocity based on the combined wave-current bottom shear stress respectively.

The boundary conditions are shown in Jenter and Madsen (1989) and Madsen et al. (1994). The no-slip condition at the bottom leads (4-21) to

$$[W^b]_{z=zm} = -\frac{g}{f} \left( -\frac{\partial \zeta}{\partial y} + i \frac{\partial \zeta}{\partial x} \right)$$

(4-23)

And the surface boundary condition is

$$\lim_{z-h} v \frac{dW^t}{dz} = \frac{\tau_{xx} + i \tau_{xy}}{\rho}$$

(4-24)

where $\tau_{xx}$ and $\tau_{xy}$ are the $x$ and $y$ component of the wind stress $\tau$, respectively. The wind stress $\tau$ can be calculated by the measured wind speed $U_a$ and wind direction $\phi_a$. 

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
\[ \tau_{xx} + i\tau_{xy} = \rho_a C_a U_a z e^{i\phi_a} \]  
\( (4-25) \)

where \( \rho_a \) is the air density and \( C_a \) is the wind-drag coefficient. Here the equation to estimate the wind-drag coefficient is given by Wu (1980)

\[ C_a = (0.8 + 0.065 U_a) \times 10^{-3} \quad \text{where } U_a [m \ s^{-1}] \]  
\( (4-26) \)

Additional conditions come from the requirement for continuous velocity and shear stress profiles at the height of the discontinuity in eddy viscosity (Jenter and Madsen, 1989).

\[ [W^b]_{z=z_m} = [W^z]_{z=z_m} \]  
\( (4-27) \)

\[ \left[ v_{t} \frac{dW^b}{dz} \right]_{z=z_m} = \left[ v_{t} \frac{dW^z}{dz} \right]_{z=z_m} \]  
\( (4-28) \)

Hence, the complex constants can be obtained by solving (4-23), (4-24), (4-25), and (4-26) simultaneously using the Kelvin functions and their derivatives shown in Abramowitz and Stegun (1972).

4-3-2. Solution Procedures

Although two boundary conditions and two matching conditions give solutions for the complex constants, the problem is still not closed. The bottom shear stress \( u_{\tau_b} \) is the solution sought and the apparent bottom roughness is related to this unknown. And if the pressure gradients are not given, we have more unknowns. The model is solved with an iterative process.
To determine the convergence of the bottom shear stress, we need the following condition.

\[
\lim_{z \to z_{\infty}} \left| \frac{dW}{dz} \right| = \frac{|\tau_{bx} + i \tau_{by}|}{\rho} = u^* b^2
\]  

(4-29)

This equation (4-29) specifies the magnitude and direction of the bottom shear stress when convergence is achieved. With this reference to the current, in addition to inputs of wave characteristics and physical roughnesses calculated from wave and current specifications, the apparent bottom roughness can be obtained by (4-22).

When determining the unknown pressure gradients, it is assumed that the along-shelf component of the pressure gradient is negligible, that is, we follow the case of the semi-infinite ocean (Jenter and Madsen, 1989). The across-shelf component of the pressure gradient can be computed by (4-19), (4-20) and (4-21) with the condition that the mean depth-integrated across-shelf velocity becomes zero, that is

\[
h \bar{U} = \int_{z_{\infty}}^{h} u \, dz = \int_{z_{\infty}}^{h} Re \left( W - \frac{G}{f} \frac{\partial \varepsilon}{\partial y} \right) \, dz = 0
\]  

(4-30)

where \( \bar{U} \) is the mean depth-averaged velocity in the across-shelf direction, with the \( x \) axis shore-normal. Rearranging (4-30) with (4-18), we get

\[
Re \left[ -i \frac{\nu_t}{f} \frac{dW}{dz} \right]_{z_{\infty}}^{h} + Re \left[ -i \frac{\nu_t}{f} \frac{dW}{dz} \right]_{z_{\infty}}^{h} = 0
\]  

(4-31)

and this analytical solution can be calculated using the same Kelvin function values and their derivatives used for the general solution (4-20) and (4-21).

The iterative process basically consists of an inner loop which updates the bottom shear...
stress and the apparent bottom roughness and an outer loop which updates the pressure gradient. The procedure begins with an initial guess of the bottom shear stress and its direction for the inner loop and an initial guess of the pressure gradient for the outer loop. The initial values of the bottom shear stress and its direction can be obtained from the along-shelf component of the wind stress. With this stress specification, the apparent bottom roughness can be calculated by the wave-current interaction model (Grant and Madsen, 1986; Madsen, 1994). After the inner loop meets the condition of zero depth-integrated across-shelf velocity, it gives the solution for the across-shelf pressure gradient and returns to the outer loop. The updated pressure gradient is compared with the previous one and the inner loop is repeated until convergence of the pressure gradient is achieved.

4-3-3. Test of Current Model

The depth-resolved current model coupled with the wave-current interaction model (hereafter JMGM) was evaluated to examine its performance, together with the Grant and Madsen model runs. Both the JMGM and the GM models include the same roughness model as described in Chapter 3 (see section 3-3). To calculate the ripple roughness $k_r$, using equation (3-18), ripple geometry is predicted using Wiberg and Harris’ (1994) model. For the movable bed roughness $k_{m,b}$, both models use the relationship (3-19) proposed by Xu and Wright (1995). One of the major differences between the JMGM and the GM models is the eddy viscosity profile. When calculating bottom shear velocity and roughness parameters, the different eddy viscosity profiles will not influence model predictions. They may significantly affect model predictions of parameters such as depth-integrated transport. For this reason, evaluation of the JMGM model may be achieved by comparing model predictions of shear velocity and roughness parameters between the two models. There is also difference in current specifications between the two models. The GM model included in the JMGM model uses the current specification as bottom shear velocity and direction which are computed using the Jenter and Madsen
model. However, the current specification for just GM model is current velocity magnitude and
direction measured at a reference height.

For input parameters, the JMGM model uses field measurements of winds, waves, and
bottom sediment while the GM model uses field measurements of currents, waves, and bottom
sediment. Field measurements of waves and bottom sediment are the common inputs for both
models. The FRF wind data presented in Figure 2-5 were inputs for the JMGM model. The current
and wave data were measured at the 20 m depth in October 1994. The current data shown in Figure
2-7 were inputs for the GM model. The wave characteristics shown in Figure 2-6 were inputs for
both models. Since multiple grain sizes do not affect the hydrodynamics model as discussed in
section 3-8, the median diameter $d_{50}=0.012$ cm was used for both models. The Wiberg and Harris
model predicts ripple geometry based on waves and bottom sediment which are the common inputs
for both models, so that predicted ripple roughnesses using both models will be identical.

Model predictions of the shear velocity due to the current above the wave boundary layer and
the shear velocity due to the combined flow are presented in Figure 4-6. There is generally good
agreement between the JMGM and the GM model predictions. In detail, the JMGM model slightly
underestimated $u_c$ values during strong currents. Predicted $u_{cm}$ values showed excellent agreement
between the two models. Based on these facts, the JMGM model was equivalent to the GM model in
predicting shear velocity parameters.

In the same way, predicted apparent roughness parameters are presented in Figure 4-7. Good
agreement between the two models occurred during the storm period from 10 to 20 October. During
the post storm period the JMGM model underestimated. This might result from the application of the
Jenter and Madsen model to low wind conditions. During the storm period, the performance of both
models was nearly identical in predicting the apparent roughness parameter.

Predicted current shear velocity and apparent roughness using both models are compared
with field measurements in Table 4-1. The field data were obtained using a velocity profile fit. The velocity profiles having $R^2 > 0.98$ were chosen except that at Hour 12, 15 October which was measured at the time near the storm peak. The predicted $u_c$ values using both the JMGM and the GM models agreed well with the field measurements. At Hour 16, 13 October, the difference between the JMGM model and the field data was about factor 5. This time corresponded to the gap of wind peaks (see Fig. 2-5) and this low wind probably caused an unstable behavior of the JMGM model. In comparisons of predicted $z_0$ values using both models with field data, there was good agreement except at Hour 12, 17 October where the difference between the JMGM model and the field data was greater than factor 4. Those times of poor agreements in shear velocity and apparent roughness at days of 13 and 17 may be associated with weak wind conditions. During strong winds or strong currents corresponding to days of 10, 11, 12, and 15, both model predictions are very good.

Another attempt for the comparison was made, in which predicted current magnitudes and directions 1 m above the bed were compared with the field measurements (Fig. 4-8). As expected from the results shown in Figure 4-6, the peaks of the predicted velocity magnitudes using the JMGM model were slightly less than the field measurements although there was good agreement in general. The predicted values of velocity direction agreed well with the field measurements within the range of $20^\circ$ during the storm period. But agreement during fair-weather conditions was poor.

In summary, the JMGM model has the same potential to predict shear velocity and apparent roughness as the GM model as long as it is applied to strong wind conditions. Modeling of current fields using the JMGM model can be justified based on the above analysis. Also, the analysis supports the conclusion that surface wind stress is important to inner-shelf processes during storm periods. The JMGM model shows good performance during high energy events, which means that the model has limited applicability. Therefore, the 1DH across-shelf transport model developed in this study has the same limitation in principle.
Figure 4-6. Comparisons between the GM model (dashed line) and the JMGM model (solid line) in terms of predicted shear velocities due to current (a) and due to combined flow (b) at 20 m depth.
Figure 4-7. Comparisons between the GM model (dashed line) and the JMGM model (solid line) in terms of predicted apparent bottom roughness at 20 m depth.
Table 4-1. Comparison of predicted $u_c$ (cm/sec) and $z_o$ (cm) using the JMGM and the GM models with field measurements at 20 m depth in October 1994.

<table>
<thead>
<tr>
<th>Day</th>
<th>Hour</th>
<th>JMGM</th>
<th>GM</th>
<th>Velocity Profile</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>$u_c$</td>
<td>$z_o$</td>
<td>$u_c$</td>
</tr>
<tr>
<td>10</td>
<td>12</td>
<td>1.56</td>
<td>0.14</td>
<td>2.20</td>
</tr>
<tr>
<td>11</td>
<td>12</td>
<td>1.30</td>
<td>0.15</td>
<td>1.36</td>
</tr>
<tr>
<td>12</td>
<td>16</td>
<td>1.38</td>
<td>0.13</td>
<td>1.42</td>
</tr>
<tr>
<td>13</td>
<td>16</td>
<td>0.19</td>
<td>0.81</td>
<td>0.99</td>
</tr>
<tr>
<td>15</td>
<td>12</td>
<td>2.41</td>
<td>0.08</td>
<td>2.52</td>
</tr>
<tr>
<td>17</td>
<td>12</td>
<td>0.89</td>
<td>0.18</td>
<td>0.84</td>
</tr>
</tbody>
</table>

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 4-8. Predicted values of velocity magnitude (a) and directions (b) at 1 m above the bed using the JMGM (solid line) model are compared with field measurements (dashed line) at 20 m depth.
4-4. Sediment Conservation Equation

Assuming that the along-shelf sediment flux gradient is negligible and ignoring the local change of sediment concentration with respect to time, the sediment conservation equation of individual size classes becomes

\[
C_b \frac{\partial z_{b, i}}{\partial t} = -\frac{\partial Q_{x, i}}{\partial x}
\]  

(4-32)

in which \( z_{b, i} \) is the bed elevation contributed by sediment of the \( i \)th size class and \( Q_{x, i} \) is the across-shelf sediment flux of the \( i \)th size class. It is noted that \( z_b \) is positive upward from a reference level.

Equation (4-32) must be solved numerically and the numerical scheme is not trivial. Equation (4-32) can be converted to a difference equation using the backward difference scheme

\[
C_b \frac{z_{b, i}^{n+1} - z_{b, i}^n}{\Delta t} = -\frac{Q_{x, i}^n - Q_{x, i}^{n-1}}{\Delta x}
\]  

(4-33)

where the superscript \( n \) denotes the time step, the subscript \( j \) denotes the spatial step, and \( \Delta t \) and \( \Delta x \) represent the increments of the time step and the spatial step respectively. The spatial scale in the onshore/offshore dimension for this study is the approximately \( 3.5 \) km distance between the 8 m and 22 m depth contours. These were set to be the onshore and the offshore boundaries of the computation field, respectively. The field was equally divided into one-meter depth intervals with one grid being about 250 m. We must choose a time step which gives stable solution. Stability condition of the backward difference scheme is given by \(|c \frac{\Delta t}{\Delta x}| \leq 1\), in which \( c \) is the local wave celerity associated with movable bed (Tan, 1992). However, we don’t know the exact value of \( \nu \) and thus we assume that this value is an order of \( 1 \) m/10 min. Somewhat arbitrarily, the value of \( \Delta t/\Delta x \) is chosen as \( 1 \) min/m resulting in a time interval of 120 minutes. Equation (4-33) requires one known
end condition of sediment flux as a boundary condition. Since there was no measurement of sediment flux at the boundaries, zero sediment transport at the offshore boundary was used.

Suspended sediment concentrations and current velocity profiles at each spatial node need to be determined to calculate sediment flux. The mean suspended sediment concentration corresponding to individual size classes can be obtained as described in section 3-4. Following Madsen et al. (1994), the eddy diffusivity is given by

\[ v_r = \kappa u_{*a} z \quad \text{for } 0 < z < \delta_{wc} \]

\[ = \kappa u_{*b} z \quad \text{for } \delta_{wc} < z < z_m \]

\[ = \kappa u_{*c} (h - z) \quad \text{for } z_m < z < h \] (4-34)

Equations describing the mean suspended sediment concentration of the \( i \)th size class are

\[ C_i = C_{r,i} \left( \frac{z}{z_r} \right) \left( \frac{z}{\delta_{wc}} \right) \left( \frac{z}{\delta_{wc}} \right) \quad \text{for } 0 < z < \delta_{wc} \] (4-35)

\[ C_i = C_{r,i} \left( \frac{\delta_{wc}}{z_r} \right) \left( \frac{\delta_{wc}}{\delta_{wc}} \right) \left( \frac{z}{\delta_{wc}} \right) \left( \frac{z}{\delta_{wc}} \right) \quad \text{for } \delta_{wc} < z < z_m \] (4-36)

and

\[ C_i = C_{r,i} \left( \frac{\delta_{wc}}{z_r} \right) \left( \frac{\delta_{wc}}{\delta_{wc}} \right) \left( \frac{z}{\delta_{wc}} \right) \left( \frac{h - z_m}{h} \right) \left( \frac{h - z}{h} \right) \quad \text{for } z_m < z < h \] (4-37)

The armoring process described in section 3-6-2 can be used in the calculation of suspended sediment concentrations. When applying the armoring process to equations from (4-35) to (4-37),
the sediment flux divergence in the sediment conservation equation is ignored. Equation (3-47) is used instead of equation (4-32).

The current velocity profile within the wave boundary layer, following Madsen (1994), is given by

\[ u + iv = \frac{u_0}{\kappa U_{wave}} \ln \frac{z}{z_0} e^{i \phi_x} \quad \text{for} \ z < \delta_{wave} \] (4-38)

in which \( \phi_x \) is the current direction. Using (4-38) and the complex velocity profile \( u + iv \) from (4-19) subtracting the component of geostrophic current, we can calculate the across-shelf sediment flux of the \( i \)th size class, \( Q_{x,i} \), taking the real part of the depth-integrated complex sediment flux

\[ Q_{x,i} = \text{Re}(Q_{x,i} + iQ_{y,i}) = \text{Re} \left[ \int_{z_i}^{h} (u + iv) \ C_i \ dz \right] \] (4-39)

Similarly, the along-shelf sediment flux of the \( i \)th size class, \( Q_{y,i} \), is

\[ Q_{y,i} = \text{Im}(Q_{x,i} + iQ_{y,i}) = \text{Im} \left[ \int_{z_i}^{h} (u + iv) \ C_i \ dz \right] \] (4-40)

Since there is no analytical solution for (4-39) and (4-40), the sediment flux is calculated by numerical integration. Trapezoidal integration was found to be very accurate when logarithmically spaced elevations were used.

Solving (4-33) for bed elevation changes due to the individual size classes and integrating all changes due to all size classes gives the net bed elevation change, \( \Delta Z_y \), during one time step. The net bed elevation change is a resultant change in which some size classes may show a gain and others a loss in the sediment budget at a grid point. To take care of the gain/loss of individual size classes,
negative fractions in associated with $\Delta Z$, are allowed but they should not violate the rule that the sum of all the fractions is always equal to unity. The net bed elevation change is related to whether an erosional or depositional phase in bed stratigraphy is occurring. If there is net deposition, the lower portion of the mixed layer is moved up by the amount equal to the thickness of the deposited sediment (Fig. 4-9). The sediment size distribution of the mixed layer can be updated by

$$f(\delta_m)^{\text{new}} = \frac{(\delta_m - \Delta Z_0) f(\delta_m) + \Delta Z_0 f(\Delta Z_0)}{\delta_m}$$

(4-41)

in which $f(\delta_m)^{\text{new}}$ is the fraction of the $i$th size class in the mixed layer of the newly formed bed stratigraphy and $f(\Delta Z_0)$ is the fraction of the $i$th size class in the net bed elevation change. In the case of net erosion, the layer below the mixed layer is added to the mixed layer in the amount of the eroded sediment. The sediment size distribution of the mixed layer is given by

$$f(\delta_m)^{\text{new}} = \frac{(\delta_m + \Delta Z_0) f(\delta_m) - \Delta Z_0 f(\Delta Z_0)}{\delta_m}$$

(4-42)

Bed stratigraphy formation (Fig. 4-9) is different from that described in Chapter 3 (1DV model). The net bed elevation change determines how the bed stratigraphy is formed. For the erosional phase, two layers are formed and this process is actually the same as the 1DV model when we don't consider the eroded sediment. L1 and L2 are, therefore, equal to L2 and L3 in the 1DV model respectively (Fig. 3-3). For the depositional phase an interfacing layer, L2, is generated because the deposited sediment raises the base of the mixed layer. The layer L2 having the same thickness of the deposited layer may have a different size distribution compared with that of the mixed layer. The formation of L3 uses the same procedure as L2 in the erosional phase.

The current method solving the governing equations using both wave and current models has
Figure 4-9. Schematic representation showing bed stratigraphy formation in the 1DH model.
the wave model passing the output to the current model (Fredsøe and Deigaard, 1993). The two
models run independently. The overall procedure to calculate one-dimensional sediment transport in
an across-shelf transect follows
1) One-dimensional set of grids is made in the across-shelf direction.
2) The wave field over the transection is predicted by a wave transformation model.
3) Current profiles at each grid point are predicted by hydrodynamic models using wind and wave
   conditions as input. This current model is coupled with a wave-current interaction model.
4) Calculate the suspended sediment flux.
5) The bed elevation profile in the across-shelf direction is calculated using the sediment
   conservation equation. This step also calculates size distribution and stratigraphy.
Procedures (4) and (5) are decoupled for the hydrodynamic model and can have several time
steps for each time step of the hydrodynamic model.

It is noted that waves and wind are two-dimensional forces but sediment flux is considered one-
dimensional.

As seen in Figure 4-1, there is no feedback allowing calculated bed elevation change to alter
the current. In the next time step, the procedure begins again with new hydrodynamic inputs and
updated bed stratigraphy. First, the wave field is calculated and it is used by the current field module.
The JMGM current field model is called at each grid point using inputs of wind, wave, and bottom
sediment. It calculates velocity profiles and shear stresses. Shear stresses are transferred to the
sediment flux module.

During each time step, it is assumed that both changing bottom sediment and armoring
processes (the coupling with the bed stratigraphy) reach an equilibrium state. This is assumed to
occur in the time scale of hours as shown in Chapter 3. These procedures are implemented in
somewhat different ways. The shear stresses determined by the current module are used repeatedly
for the calculation of the sediment flux, which is associated with procedures calculating changing bottom sediment and armoring processes. This reduces the time consumed calculating the flow fields. After the loop calculating sediment flux is finished, the sediment conservation equation is solved for the net bed elevation change. The bed stratigraphy is updated depending on the condition of the net bed elevation change.

4-5. Results and Discussion

Inputs for the hydrodynamic model were the local winds measured at the FRF and the wave characteristics measured at the 20 m water depth. The wind was applied uniformly over all grid points. Wind speeds less than 4 m/s were not used because the current module failed under low wind conditions. The model was coded to handle different inputs of bed stratigraphy at each grid point. However, the same initial bed stratigraphy as that used by the 1DV model was used at all grid points. Therefore, the effects of different initial bed stratigraphy on model predictions were not evaluated.

As with the 1DV model, the resuspension coefficient was the only parameter used to adjust the calculation of suspended sediment concentration. Using a resuspension coefficient $\gamma_0 = 0.002$, which was twice the value of the 1DV model ($\gamma_0 = 0.001$), gave good agreement between model predictions and field measurements (Fig. 4-10). The higher $\gamma_0$ value in the 1DH model can be associated with the low $u_c$ values predicted by the JMGM current model compared to the GM model used in the 1DV model (see Fig. 4-6). The increment of the resuspension coefficient in the 1DH model is, however, trivial compared with the wide range of $\gamma_0$ values in the literature, i.e., 0.01 to 0.0001.

It appeared that the 1DH model predictions are somewhat better than the 1DV model predictions (see Fig. 3-4). There are minor peaks predicted by the 1DH model which give better agreement between predicted suspended sediment concentrations and field measurements than the
Figure 4-10. Predicted suspended sediment concentration (solid line) are compared with field measurements (plus signs) at 20 m depth. The resuspension coefficient is $\gamma_0 = 0.002$. 

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
1DV model. These small variations are not prominent in the 1DV model and may result from the horizontal flux divergence in the sediment conservation equation. Considering the calculated minor peaks are consistent with current velocity peaks (Fig. 2-7), these current peaks may induce some gradient in sediment flux. This sediment flux gradient causes difference between the sediment flux moved into a grid and the sediment flux moved out of that grid. Thus, there is a net amount of sediment that moves horizontally from a grid to the adjacent one. This amount of sediment can be contributed to suspended sediment concentrations in the following way. The amount of sediment moved into or moved out of a grid results in the net bed elevation change $\Delta Z_n$ in the sediment conservation equation. Since the formation of bed stratigraphy depends on $\Delta Z_n$ and the time-stepping procedure allows a feedback of bed stratigraphy, the sediment added or subtracted through the net bed elevation change can affect the calculation of suspended sediment concentrations. If this contribution is ignored, the calculation of suspended sediment concentration is based on the only diffusion process from the bed. Ignoring this term may be attributed to poor agreement of the 1DV model as suggested in Chapter 3. Another difference between 1DV and 1DH models in the calculation of suspended sediment concentration profiles is the eddy viscosity profile; the former uses the linear eddy viscosity and the latter the bilinear eddy viscosity. Since the near-bed eddy viscosity profiles are almost identical in both models, they are not invoked to improve the model predictions.

Predicted current velocity fields along the across-shelf transect at the time near the storm peak are presented in Figure 4-11 and 4-12. As expected, across-shelf current velocity profile was predicted as two layers in which the current directions in the upper layer were onshore due to the surface wind drag and those in the lower layer were offshore as a downwelling flow. Along-shelf current velocity profiles showed southerly flows in the whole water column over the transect. The magnitudes of the across-shelf current velocities were almost uniform along the transect. However, those of the along-shelf current velocities increased toward the offshore boundary, reflecting the
Figure 4-11. Predicted current velocity fields along the transect showing across-shelf current velocity profiles at the time near the storm peak.
Figure 4-12. Predicted current velocity fields along the transect showing along-shelf current velocity profiles at the time near the storm peak.

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 4-13. Predicted concentration fields along the transect at the time near the storm peak.
Figure 4-14. Predicted local sediment flux fields along the transect showing across-shelf local sediment flux profiles at the time near the storm peak.

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
Figure 4-15. Predicted local sediment flux fields along the transect showing along-shelf local sediment flux profiles at the time near the storm peak.
frictional effect of inner-shelf geometry.

Suspended sediment concentration fields along the transect are presented in Figure 4-13. The contour lines of the various concentrations were distributed regularly as a sigmoidal shape. As a result, they showed larger horizontal gradient than the current velocities and also showed more or less increased vertical gradient at the water surface. This characteristic distribution of suspended sediment concentration may be produced by the wave field. The intensity of wave orbital velocity increases with decreasing depth and more sediment is resuspended. It results in larger horizontal gradient of suspended sediment concentration.

Using two fields presented in above, we obtained local suspended sediment flux fields, mean concentration multiplied by mean velocity at a certain elevation, along the transect (Fig. 4-14 and 4-15). The distribution of across-shelf flux, UC, was similar to that of the across-shelf current velocity. The distribution of along-shelf flux, VC, was closer to the concentration field rather than that of the along-shelf current velocity. It is noted that the local suspended sediment flux near the surface in Figure 4-14 is not negligible.

Temporal and spatial variations of predicted shear velocities are presented in Figure 4-16 and 4-17. Bottom shear velocities due to wind-driven currents $u_{bw}$ varied predominantly with respect to time but their spatial variation was negligible. Shear velocities due to the combined flows, $u_{cw}$, varied with time and space. Spatial gradient of $u_{cw}$ was highly correlated to its spatial mean value which was chosen as its representative value at a specified time (Fig. 4-18). This means that higher wave energy is accompanied by larger spatial gradient of $u_{cw}$. High waves increase the magnitude as well as the across-shelf gradient of suspended sediment concentration. So waves are very important in determining flux divergence in the sediment conservation equation and eventually morphological changes. Wind-driven currents must play an important role in determining not only the magnitude but also the direction of sediment flux. However, the currents are not deterministic for the spatial
Figure 4-16. Contour map showing distribution of predicted shear velocities due to current with respect to space and time.
Figure 4-17. Contour map showing distribution of predicted shear velocities due to combined flow with respect to space and time.
Figure 4-18. Correlation between predicted shear velocity due to combined flow, $u_{cw}$, and spatial gradient of $u_{cw}$. 

Corr. coef. = 0.9254
gradient of sediment flux. As shown in Figure 4-19 and 4-20, the across-shelf and along-shelf sediment fluxes were simultaneously related to both waves and currents. The distributional pattern of sediment flux was not exclusively related to either waves or currents. Both magnitude and spatial gradient of the sediment flux were maximum at the storm peak. To examine relationships between the sediment flux and both waves and currents, the time series of sediment flux at the 20 m depth was taken from the contour map (Fig. 4-21) and the time series of $u_b$ and $u_w$ were made (Fig. 4-22). Comparisons between both figures clearly demonstrated that synchronization of waves and currents resulted in a higher sediment flux. Therefore, we can postulate that morphological changes on the inner shelf during a high energy event occur with synchronized wave and current activities. In other words, either wave or current events alone, for example, the wind peak at the onset of the storm, will not give significant effect on sediment flux and then morphological changes.

Predicted direction of across-shelf sediment flux was dominantly offshore. Onshore sediment transport occurred at the end of the storm (Fig. 4-21) when southerly winds caused upwelling currents. But the peaks from 17 to 18 October were related to unstable behaviors of the Jenter and Madsen model which occur when the wind direction is nearly parallel to the coastline (see Fig. 2-5).

The time series of bed elevation along the transect showed accretion of inner-shelf bed during the storm period (Fig. 4-23). A rapid increase in the net bed elevation change occurred during the storm peak and overall accretion of the bed elevation was about 2 mm (Fig. 4-24). The time series representing the net bed elevation change $\Delta z_n$, which is similar to the time series representing the sediment flux, present the erosion/deposition sequences during the storm period (Fig. 4-25). Offshore sediment flux was directly related to deposition and onshore sediment flux to erosion. Another feature was that the bed elevation change was greater in shallow water depths than deeper water depths, implying that source of sediment was in shallow water depths, i.e., the surf zone. These
Figure 4-19. Contour map showing distribution of mean depth-integrated across-shelf sediment flux with respect to space and time.
Figure 4-20. Contour map showing distribution of mean depth-integrated along-shelf sediment flux with respect to space and time.
Figure 4-21. Time series of across-shelf (a) and along-shelf (b) sediment flux at 20 m depth.
Figure 4-22. Time series of shear velocity due to current (solid line) and shear velocity due to combined flow (dashed line) at 20 m depth.

Reproduced with permission of the copyright owner. Further reproduction prohibited without permission.
facts can be explained by the characteristic concentration field produced by the waves and the
direction of wind-driven currents. The horizontal gradient of suspended sediment concentration can
be regarded as supporting two sources of sediment surplus and deficit, which are located at shallow
water and deep water depths, respectively. When the direction of wind-driven currents is offshore
(downwelling), the higher concentration moves toward deep water, so the inner shelf acts as a sink of
sediment. On the contrary, the lower concentration (deficit) moves toward shallow water during
upwelling and consequently the inner shelf acts as a source of sediment. Therefore, both waves and
wind-driven currents are crucial to determining the erosion/deposition of bed elevation. The inner-
shelf bed responds dynamically to these variables during a high energy event.

Compared with field measurements (Fig. 2-11), the time series of near-bottom orbital
velocities appears to qualitatively agree with predicted $u_{ow}$ values. The differences between wave
orbital velocities at the 12 m and 20 m depths were proportional to the intensity of the waves. Also,
current velocities at the two locations were generally close during the storm event except during the
storm peak when some difference existed. Another discrepancy was that current velocity was larger
at the onshore site than at the offshore site. These discrepancies between the model predictions and
the field measurements might come from non-wind-induced processes. Since currents play an
important role in determining sediment flux, we need to predict them with confidence. The predicted
currents were almost uniform with respect to space, so that they could have led to an underestimation
of sediment flux.

Field measurements of bed elevation changes were quantitatively and qualitatively different
from the model predictions. The predicted bed elevation changes were about two orders of magnitude
smaller than the field measurements presented in Figure 2-11. Even if we include along-shelf
sediment flux that is an order of magnitude greater than across-shelf sediment flux, the model still
underestimates by about an order of magnitude. Currently, no resolution of this problem is
Figure 4-23. Time series of bed elevation along the across-shelf transect. Solid line is the bed elevation which varies with time and dashed line is the initial bed elevation. Horizontal axis represents time (hour/day/month).
Figure 4-24. Contour map showing bed elevation changes with respect to space and time.
Figure 4-25. Time series of the net change of bed elevations (a) and the cumulative elevations (b) at 12 m, 16 m, and 20 m depths.
immediately available. Qualitatively, the field measurements at the 12 m depth showed two depositional phases concurrent with the current velocity peaks. However, the model predicted just one depositional phase at the storm peak. Both model predictions and field measurements seemed to agree in the last stage of the storm when the erosional phase occurred. It is worth noting that the bed elevation changes at two locations were not in phase during the first depositional phase when erosion occurred at the 20 m depth. The first depositional phase at the 12 m depth does not agree with the model predictions and might have resulted from sinkage of the tetrapod. If this is the case, approximately 15 cm of the change can be regarded as caused by the sink.

The distribution of bed stratigraphy along the across-shelf transect as expressed by the predicted spatial variation of storm deposits is shown in Figure 4-26. The thickness of the storm deposits as well as the vertical gradient of sediment size was associated with depth. Both thickness of storm deposits and the vertical gradient of sediment size increased toward shallow water. This implied that the surface sediment in shallow water had the smallest grain size after the storm event (Fig. 4-27). This can be explained by the analysis in Chapter 3 where the modeled erosion depth is relevant to the segregation of sediment size in the sediment column, assuming a negligible flux divergence. As expected, the vertical gradient of sediment size was related to the erosion depth shown in Figure 4-28. The erosion depth was generally much larger than the bed elevation changes induced by the sediment flux divergence. Sediment flux weakly influenced the sediment size distribution. However, it is noted that sediment transport will be important to bed stratigraphy on a long-term basis.
Figure 4-26. Predicted bed stratigraphy along the across-shelf transect. Circle represents the median diameter of sediment layer.
Figure 4-27. Contour map showing distribution of the median diameter of the top 1 cm thick bed with respect to space and time.
Figure 4-28. Contour map showing distribution of the erosion depth with respect to space and time.
5. CONCLUSIONS

Two process-based models for sediment transport and strata formation on the inner shelf during a high energy event were developed in this study. One was the vertical one-dimensional model (1DV model) which deals with the coupling of boundary layer processes with bed stratigraphy, neglecting the flux divergence term. The Grant and Madsen (1986) model was implemented for the hydrodynamics module. The armoring processes and the processes that cause changing sediment size were solved explicitly using the simplified sediment conservation equation. One of the basic assumptions was that the system reaches an equilibrium state in a time scale of hours and this was used to close the problem.

The other model was the horizontal one-dimensional, depth-resolved model (1DH model) which treats across-shelf sediment transport by including the horizontal flux term in the sediment conservation equation. To estimate wave and current fields, a simple wave transformation model and the Jenter and Madsen (1989) model were used. In the calculation of sediment flux, the bed stratigraphy was coupled under the same assumptions as used in the 1DV model. Bed elevation changes were solved explicitly by the backward difference scheme with an offshore boundary condition of zero sediment transport.

The unique feature of the 1DV model was to defining the total depth of change as the sum of the erosion depth and the mixing depth. This length scale was introduced to provide sediment availability to the suspended sediment. It supports a more rigorous treatment than the mixing depth alone can. Another feature of this model was coupling of the bed stratigraphy using parameterized
bed inputs. Theoretically, the mixing depth, armoring process, and changing bottom sediment
processes are not independent but mutually dependent. The problem is not closed if any one of these
processes is neglected. The following features must be included to incorporate the armoring process
in a composite model; (1) the variation of the grain size distribution in an armored layer, (2) the
variation of the armored layer thickness, (3) the independence of the armored bed from the suspended
sediment concentration, and (4) the bed stratigraphy as an output of the model. A feedback procedure
between the hydrodynamics and the bed stratigraphy was established.

The field measurements were performed in October 1994 as a part of the CoOP'94
experiment off Duck, North Carolina. Wind, wave, current measurements made during an extra-
tropical storm were used for model inputs. A simplified bed stratigraphy provided sediment
parameters. Both models were adjusted using field measurements of the suspended sediment
concentration to determine the resuspension coefficient $\gamma_0$. The adjusted models resulted in the
resuspension coefficient values $\gamma_0 = 0.001$ for the 1DV model and $\gamma_0 = 0.002$ for the 1DH model,
respectively. Suspended sediment concentrations predicted by the 1DH model were closer to the field
observations than those predicted by the 1DH model. This was attributed to the contribution of
horizontal sediment flux. The $\gamma_0$ values were close to the Smith and McLean's (1977) value. The
resuspension coefficient is assumed to remain constant over the time-varying flow conditions in order
to treat bed concentration as a variable. This makes it difficult to estimate the $\gamma_0$ values inversely
using field observations of suspended sediment concentration. The major features of the model
predictions are described in the following.

Coupling of boundary layer processes with bed stratigraphy was critical to improving
predictions of suspended sediment concentration. The decoupled model resulted in the resuspension
coefficient $\gamma_0 = 3 \times 10^{-4}$ which was close to the value suggested by Madsen et al. (1993). Good
agreement between model predictions and field measurements of suspended sediment concentration
was not achieved because the decoupled model overestimated suspended sediment concentration at higher elevations and underestimated at lower elevations. However, in terms of ability to predict hydrodynamic variables, the decoupled model was equivalent to the coupled model. Small changes in the bottom sediment had little effect on the hydrodynamics but greatly affected sediment transport. The bed was as important in controlling the sediment transport as the flow and both bed and flow controls should be modeled simultaneously to predict sediment transport with confidence.

Predicted bed stratigraphy showed that the major influence of a resuspension event was to increase the vertical gradient of the sediment size in a sediment column. Deeper erosion depths resulted in larger vertical gradients of sediment size, forming the typical fining upward sequence of storm deposits. This may explain the reduced bed micro-morphology immediately after a high energy event. In this qualitative aspect, the behavior of the model was accurate. Thin surface layers of bottom sediment appeared to show fining upward trends which may result from a storm event. However, more rigorous comparisons require a long-term model and careful treatment of sediment size variations.

The 1DH model showed the effects of wind-driven processes on across-shelf sediment transport and inner-shelf morphology. Wind-driven currents, forming a more or less uniform current field, were responsible for the magnitude as well as the direction of the sediment flux. A characteristic sigmoidal shape of the concentration field was produced by the wave field, meaning that waves were important to the horizontal gradient of the suspended sediment concentration. Therefore, flux divergence depended on the waves. The coexistence of waves and currents was responsible for large flux divergence and morphological changes. The results also showed that shoreface deposition occurred during downwelling currents whereas erosion was associated with upwelling currents. Thus, inner-shelf morphology responded dynamically to imposed wind-stresses and waves as either a sink of sediment or a source of sediment.
APPENDIX A

The Effects of Measurement Height Changes on the Velocity Profiles

INTRODUCTION

From mean velocity profiles measured in the benthic boundary layer, bed shear stress and roughness can be calculated based on the assumption of the logarithmic 'law of the wall'. These parameters are crucial to estimating quantities of sediment transport. This technique is typically performed using an instrumented tripod equipped with current meters vertically arranged at several heights, recorders, and control systems [Grant and Madsen, 1986; Wright, 1989]. Numerous uncertainties in field instrumentation may cause the ideal logarithmic profile to be a curved line on a log-linear plot. Sources of uncertainty generally are properties of real flows such as low-frequency internal waves, stratification, and flow accelerations [Soulsby and Dyer, 1981; Grant et al., 1984; Dyer, 1986] or measurement errors including inaccuracies of a velocity sensors and uncertainty in the distance from the velocity sensors to the seabed [Wiberg and Smith, 1983; Grant et al., 1984; Gross et al., 1992].

Wiberg and Smith [1983] related a velocity profile with a concave-upward curvature to instrument settlement. A zero-plane adjustment resulted in better prediction of the roughness parameter [Wiberg and Smith, 1983; Grant et al., 1984]. Their method was to choose the displacement height that maximized the $R^2$ values for all velocity profiles. This ignored the possibility of time-varying displacement. The displacement height might be defined as a collective
term accounting for relative change in the measurement height due to both the vertical movement of
the instrument frame and the deposition/erosion of sediment. Therefore, the time variant
displacement height was natural in this definition. Moreover the positive displacement might be
acceptable although Gross et al. [1992] argued that it is not physically reasonable in light of
instrument settling.

When a real velocity profile was assumed to be logarithmic, a simple equation having a
displacement height term, similar to Jackson’s [1981], was found to represent the displaced velocity
profile. This equation was solved using the least square method and only velocity profile
measurements [Robinson, 1962; Stearns, 1970]. One goal is to test the usefulness and examine the
nonlinear property of the equation.

Since the equation contains one more parameter to be determined compared to the
conventional velocity profile fitting method, it requires an independent technique to find one of the
three parameters [Soulsby and Dyer, 1981]. An instrument such as the Digital Sonar Altimeter can
be deployed to keep tracks for the measurement height. In an alternate way, the shear velocity term
can be determined by the inertial dissipation method which requires the turbulent part of velocity
measurements [Huntley, 1988; Xu et al., 1994]. However, it was not always easy to get highly-
qualified data for these methods. Therefore, it is desirable to seek a method to determine all
parameters without relying on an independent method.

THEORETICAL APPROACH

When the instrument moves an unknown distance vertically, the actual measurement heights
of velocity sensors become unknown. The heights of sensors will be different from the initially
measured ones on the hard surface before deployment. All sensors experience the same amount of
change in the distance from sensors to seabed. In case that the distance is reduced, the sensors will
record lower values of velocity than what should be seen if the instrument is not moved. On the contrary, the higher velocity will be measured when the instrument goes upward. This means that the velocity measurements without accurate information about measurement heights will not give realistic velocity profiles. Therefore, the departure from the logarithmic velocity profile can occur because of the use of misleading measurement heights.

This simple relationship can be shown in the log-linear plot of a velocity profile (Fig. 1). The straight line in Figure 1 represents an assumed logarithmic profile before the instrument is moved. When the instrument is lowered in some distance, for example, 10 cm in Figure 1 the logarithmic profile is changed to the concave-upward displaced profile in which the degree of curvature (i.e., the departure from the logarithmic profile) increases toward the bed. In a hypothetical upward displacement of the instrument, the displaced profile becomes concave-downward. The form of curvature is reversed compared with the profiles induced by accelerating flows [Soulsby and Dyer, 1981]. As going away from the bed, the displaced profile approaches asymptotically to the logarithmic velocity profile.

Derivation of equation

When a velocity profile is logarithmic, we have

\[ U(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right) \]  

(1)

where \( U(z) \) is the mean velocity, \( u_* \) is the shear velocity, \( \kappa \) is the von Karman's constant, \( z \) is the vertical elevation, and \( z_0 \) is the roughness. This logarithmic velocity profile corresponds to the solid straight line in the log-linear plot (Fig. 2). The straight line is changed to the broken curve when measurement heights are lowered as much as a displacement height \( d \). This displaced profile can be expressed as equation (4) through a simple coordinate transformation. However, the following
Figure 1. Departure of logarithmic profile due to instrument displacements when instrument is displaced by ± 10 cm.
Figure 2. Definition sketch showing the relationship between logarithmic profile and displaced profile.

$U(z) = u_r / \kappa \ln(z/z_0)$
analysis provides some physical meaning for this equation. Let the velocities, $U$ and $U_d$, at a certain elevation $z$ from both velocity profiles as seen on the horizontal axis in Figure 2. On the triangle $\Delta ABC$ the slope ($u./\kappa$) can be written as

$$\frac{u_*}{\kappa} = \frac{BC}{AB} = \frac{U-U_d}{\ln(z) - \ln(z+d)}$$

(2)

It is noted that the sign of $d$ is set to be positive for the downward displacement of instrument.

Rearranging (2) for $U_d$ gives

$$U_d = U - \frac{u_*}{\kappa} [\ln(z) - \ln(z+d)]$$

(3)

Substituting (1) into (3), with some manipulation, the displaced profile is given by

$$U_d = \frac{u_*}{\kappa} \ln \left( \frac{z+d}{z_0} \right)$$

(4)

This equation is similar to that derived by Jackson [1981] but its physical basis is quite different.

$u./\kappa$ in (4) appears to represent the slope of a displaced profile. But this is not true because the displaced profile is a curve, so that the slope is not a constant. Therefore, the term, $u./\kappa$, in (4) can be interpreted as the slope of the straight line which shows the actual logarithmic profile, (1).

The slope of (4), designated by $u.^/\kappa$, can be found approximately on $\Delta BCD$.

$$\frac{u.}{\kappa} = \frac{BC}{CD} = \frac{U-U_d}{\ln(z-d) - \ln(z)}$$

(5)

Eliminating the common term, $U-U_d$, with (2) gives

$$u_* = u_* \left[ \ln \left( \frac{z+d}{z+d} \right) / \ln \left( \frac{z-d}{z} \right) \right] = u_0 F(z, d)$$

(6)

Equation (6) means that the slope of the displaced profile depends upon both the elevation and the
displacement height. The vertical gradient of function $F(z,d)$ is greater near the bed and decreases asymptotically with increasing height.

The intercept of (4), $z_0'$, is given by $z_0-d$ when $U(z) = 0$. Hence the term $z_0$ in (4) does not stand for the roughness parameter in the displaced profile. It is the roughness of the assumed logarithmic velocity profile which is not displaced.

In summary, if the shear velocity and roughness parameters are taken from the equation (4) as usual, they will contain properties of not the displaced profile but the actual logarithmic profile. This simple relationship explored in this manner shows that additional terms are not necessary for calculating the real shear stress and roughness parameters.

**Least square method**

The method to solve for three parameters, $u_*$, $z_0$ and $d$ is adopted with those described in Robinson [1962] and Stearns [1970]. The observed variables, $U$ and $z$, are treated as input variables, denoted as a subscript $i$ in the equation given by

$$U_{d,i} = U_* \ln \left( \frac{z + d}{z_0} \right), \quad \text{where} \quad 1 \leq i \leq M$$

(7)

in which $M$ is the number of observations and $u_*\kappa$ is substituted into $U_*$ for simplicity. Here $z_i$ is an initially measured height of the $i$th velocity sensor rather than the actual observed data. A solution for the three profile parameters is sought by the least square method which minimizes the sum of error squares between the observed velocity data at several heights and the values calculated by the model equation (7). Since the derivation of solution is identical to the previous work conducted by Robinson [1962], many of the details will be skipped here.

$M$ elements of an ordered array can be denoted by $M$ dimensional components of a vector...
and then the sum of the product between the elements of two arrays can be represented simply by the scalar product of two vectors (a dot product). If we take a vector notation with regards to the elements of a data array, the derivation to the solution will be simplified. So the error vector between the observed data and calculated quantities with \((7)\) is defined as

\[
\varepsilon = \bar{U}_{obs} - \bar{U}_{cal} = \bar{U} - U_* (\bar{x} - \bar{w})
\]

(8)

where \(X = \ln(z + d)\) and \(W = \ln(z_0)\). The sum of error squares, therefore, is given by

\[
\varepsilon \cdot \varepsilon = [\bar{U} - U_* (\bar{x} - \bar{w})] [\bar{U} - U_* (\bar{x} - \bar{w})]
\]

(9)

The condition to minimize \((9)\) is that the partial derivatives of \((9)\) with respect to \(U_*\), \(d\), and \(z_0\) are zero, i.e.,

\[
\frac{\partial \varepsilon \cdot \varepsilon}{\partial U_*} = 0, \frac{\partial \varepsilon \cdot \varepsilon}{\partial d} = 0, \frac{\partial \varepsilon \cdot \varepsilon}{\partial z_0} = 0
\]

(10)

Hence we have three equations from \((10)\) that are given successively for \(z_0\), \(U_*\), and \(d\),

\[
\bar{W} = \bar{W}_m = \bar{x}_m - U_*^{-1} \bar{U}_m
\]

(11)

\[
U_* = (\bar{u} \cdot \bar{v}) / (\bar{x} \cdot \bar{x})
\]

(12)

\[
f(d) = (\bar{u} \cdot \bar{x}) (\bar{v} \cdot \bar{b}) - (\bar{x} \cdot \bar{x}) (\bar{u} \cdot \bar{b})
\]

(13)

where \(W_m, X_m, U_m,\) and \(B_m\), denote the vector of mean in which all components are equal to its mean, \(B\) is a vector substituted for \(\partial X / \partial d\), and vectors subtracted by its mean, for example, \(U - U_m\) are indicated by lowercase vectors \(u, x,\) and \(b\).

An iteration technique may be applied to solve for \((13)\) which is a function of the only
unknown $d$. The secant method is appropriate in practice because this numerical technique requires
no derivatives and converges at the efficient rate, the convergence order 1.62 [Robinson, 1962].
However, it needs the caution that two successive initial guesses (a bracket) should be properly set in
the small side from the maximum point of (13). The tolerance on $d$ is good enough at 0.1 cm. After $d$
is determined, $U.$ and $z_0$ are solved in (12) and (11) by successive substitutions.
REFERENCES


169


Kim, S. C., L. D. Wright, and B. O. Kim, 1996. The combined effects of synoptic and local-scale meteorological events on bed stress and sediment transport on the inner shelf of the Middle
Atlantic Bight. submitted.


VITAE

Baeck Oon Kim