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Quantification of sediment bed – water column exchange processes in the South San Francisco Bay estuary

By

Steven M. Gladding

A dissertation submitted in partial satisfaction of the requirements for the degree of Doctor of Philosophy in Engineering – Civil and Environmental Engineering in the Graduate Division of the University of California, Berkeley

Committee in charge:

Professor James R. Hunt, Chair
Professor Mark Stacey
Professor Céline Pallud

Fall 2011
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Quantification of sediment bed – water column exchange processes in the South San Francisco Bay estuary by Steven M. Gladding

Doctor of Philosophy in Engineering – Civil and Environmental Engineering University of California, Berkeley

Professor James R. Hunt, Chair

The salinity gradient formed by the mixing of saline ocean waters and fresh river waters within estuaries forms diverse habitats and creates the conditions for highly dynamic sediment behavior. Sediments which enter the estuarine system may pass multiple times through the water column and sediment bed before finally being carried out to sea or deposited as a more permanent part of the sediment bed. The exchange of sediments between the water column and sediment bed is affected not only by physical forcing such as tides and winds but also by properties inherent to estuarine sediments including flocculation and consolidation. Like many others, the San Francisco Bay estuary has been highly impacted by human activities over the last 150 years. Faced with an ever changing environment, understanding how the estuary will be impacted in the future depends in part upon understanding how sediments move between the bed and water column.

Two instrument deployments were carried out in the spring and fall of 2009 which measured suspended sediment concentrations, sediment bed properties and water column physical properties in the South San Francisco Bay. Analysis of the in situ measurements from the instrument deployments, sediment cores collected at the study site and results from a coupled sediment bed – water column numerical model of the instrument deployment period were used to estimate several important properties of sediments in the estuary. The depth of the layer of sediments readily resuspended from the bed was found to be less than 1 cm, an order of magnitude smaller than had been previously assumed. Acoustic Doppler velocimeters positioned near the bed were used to make collocated measurements of water velocity and suspended sediment concentrations (SSC). Turbulence data extracted from the velocity time series was used to develop sediment erosion relationships as a function of bed shear stress. Settling velocities were also estimated from the suspended sediment time series data. Upward looking acoustic Doppler current profilers were also deployed to measure water column velocities and collocated SSC. An error analysis showed that, with the proper information, the SSC calibrations produced reliable values. Results from the coupled sediment bed – water column model corroborated many of the results estimated from the experimental data. The results from this study provide several important estuary specific values for sediment exchange processes and demonstrate the need to use in situ measurements to test our mechanistic understanding of these processes.
To my partner and wife, Becca, for supporting me all these years
# Table of Contents

Dedication ......................................................................................................................... i  
Table of Contents ............................................................................................................. ii  
List of Figures ................................................................................................................... v  
List of Tables ...................................................................................................................... vi  
Acknowledgements ......................................................................................................... vii  

1. Introduction ...................................................................................................................... 1  
   1.1. The San Francisco Bay Estuary ................................................................................. 1  
      1.1.1. Physical Setting ................................................................................................. 2  
      1.1.2. Ecological Setting ............................................................................................. 2  
   1.2. A Changing Estuary .................................................................................................. 4  
      1.2.1. Hydraulic Mining .............................................................................................. 4  
      1.2.2. Dredging .......................................................................................................... 4  
      1.2.3. Contaminants .................................................................................................. 5  
      1.2.4. Salt Ponds ........................................................................................................ 6  
   1.3. Modeling Sediments in the Bay ................................................................................. 6  
   1.4. New Methods for Measuring Sediments ................................................................. 6  
   1.5. Objective and Hypotheses ...................................................................................... 7  

2. Current Understanding of Estuarine Sediments ............................................................. 9  
   2.1. Clay Minerals .......................................................................................................... 9  
      2.1.1. Electric Double Layer ..................................................................................... 10  
   2.2. Water Column Processes ....................................................................................... 11  
      2.2.1. Flocculation .................................................................................................... 11  
      2.2.2. Floc Breakup .................................................................................................. 14  
      2.2.3. Modeling Flocculation .................................................................................... 15  
      2.2.4. Settling Velocity .............................................................................................. 17  
   2.3. Cohesive Sediment Beds ......................................................................................... 21  
      2.3.1. Deposition ....................................................................................................... 22  
      2.3.2. Erosion ............................................................................................................ 22  
      2.3.3. Consolidation ................................................................................................. 24  
   2.4. Beryllium-7 ............................................................................................................. 26  
      2.4.1. Beryllium-7 in the Atmosphere ....................................................................... 26  
      2.4.2. Beryllium-7 in Estuarine and Coastal Environments ........................................... 27  
      2.4.3. Sediment Studies Using Beryllium-7 ................................................................. 27  
   2.5. Investigations of the San Francisco Bay Estuary Sediments .................................... 28  
      2.5.1. Sediment Minerals .......................................................................................... 29  
      2.5.2. Sediment Flocs ............................................................................................... 29  
      2.5.3. Sediment Bathymetry ..................................................................................... 30  
      2.5.4. Numerical Models ......................................................................................... 30
3. Experimental Methods .................................................................................................................. 32
  3.1. Field site .................................................................................................................................. 32
  3.2. Water Column Measurements ................................................................................................. 32
    3.2.1. Acoustic Doppler Velocimeter (ADV) ............................................................................. 32
    3.2.1.1. Calibration of ADV for SSC Measurements ................................................................. 34
    3.2.2. Acoustic Doppler Current Profiler (ADCP) ...................................................................... 39
    3.2.2.1. Calibration of ADCP for SSC Measurements ............................................................... 39
    3.2.2.2. The Calibration Process ................................................................................................. 39
    3.2.3. Laser In Situ Scattering and Transmisiometry (LISST) ....................................................... 40
    3.2.4. Optical Backscatter (OBS) Sensor ...................................................................................... 43
    3.2.5. Water Column Samples for Suspended Sediment Concentration Analysis .................. 43
  3.3. Sediment Measurements ......................................................................................................... 45
    3.3.1. Sediment Grab Samples ..................................................................................................... 45
    3.3.2. Sediment Cores .................................................................................................................. 45
      3.3.2.1. Collection and Storage ............................................................................................... 45
      3.3.2.2. Core Sampling and Processing .................................................................................. 45
      3.3.2.3. Radioisotope Analysis ................................................................................................ 47
  3.4. Consolidation Experiment ........................................................................................................ 47
  3.5. Determination of Additional Field Conditions from the Experimental Data ...................... 48
    3.5.1. Turbulence ........................................................................................................................ 48
    3.5.2. Wave Parameters .............................................................................................................. 49
    3.5.3. Bed Shear Stress .............................................................................................................. 49
    3.5.4. Vertical Turbulent Sediment Flux ...................................................................................... 51
  4. Results ........................................................................................................................................ 52
    4.1. Time Series Data ..................................................................................................................... 52
    4.2. Porosity Profiles .................................................................................................................... 61
    4.3. Particle Size Distribution ....................................................................................................... 61
  5. Data Analysis and Discussion .................................................................................................... 65
    5.1. Bed Sediments ....................................................................................................................... 65
      5.1.1. Beryllium-7 ....................................................................................................................... 65
        5.1.1.1. Sediment Bed Activities ............................................................................................ 65
        5.1.1.2. Beryllium-7 Activities in the Near Surface Air ............................................................ 65
        5.1.1.3. Discussion .................................................................................................................. 66
      5.1.2. Erosion Parameters ........................................................................................................... 67
        5.1.2.1. Discussion .................................................................................................................. 67
      5.1.3. Consolidation Parameters ................................................................................................. 70
        5.1.3.1. Discussion .................................................................................................................. 71
    5.2. Water Column Sediments ...................................................................................................... 74
      5.2.1. Second Order Particle Kinetics ......................................................................................... 74
        5.2.1.1. Discussion .................................................................................................................. 76
      5.2.2. Settling Velocity ............................................................................................................... 77
      5.2.3. ADCP Backscatter ............................................................................................................ 80
        5.2.3.1. Error Analysis Equations ........................................................................................... 80
        5.2.3.2. Results ........................................................................................................................ 83
# List of Figures

1.1 Map of the San Francisco Bay estuary ............................................... 2
3.1 Detail of station layout in the South Bay ........................................... 33
3.2 ADV calibrations ........................................................................... 38
3.3 OBS calibrations ........................................................................... 44
3.4 Push corer design ......................................................................... 46
4.1 Conditions during the spring deployment ........................................... 53
4.2 Conditions during the fall deployment .............................................. 56
4.3 Conditions during the fall deployment after DOY 270 ...................... 60
4.4 Sediment core density profiles ....................................................... 62
4.5 Particle size distributions for bed and water column sediments ......... 63
5.1 Determining erosion rate parameters from experimental data ......... 68
5.2 Separation of data by tides ............................................................. 69
5.3 Final density profile from sediment core density profiles ............... 72
5.4 Modeling sediment consolidation experiment results ..................... 72
5.5 Determining 2nd order and 1st order rate constants ....................... 75
5.6 Example fit of 1st and 2nd order models to experimental data .......... 77
5.7 Results from 2nd order flocculation model .................................... 78
5.8 Magnitude of advection – dispersion equation terms ....................... 79
5.9 Estimating settling velocity from ADV data .................................... 81
5.10 ADCP calibration relationship ...................................................... 86
5.11 Estimated SSC and errors derived from ADCP measurements ....... 87
5.12 Comparison between OBS and ADCP measures of SSC ................. 89
5.13 Comparison between grab sample and ADCP measures of SSC ...... 90
5.14 Sources of error for ADCP uncertainties ....................................... 92
5.15 Comparison of ADCP calibrations using different sources of data ... 94
5.16 Comparisons of ADCP calibrations using calibration from other stations 95
5.17 Increase in sediment water column mass as a function of current velocity 98
6.1 Schematic of the sediment bed and water column models .............. 103
6.2 Comparison of modeled and measured sediment values for spring deployment .................................................. 110
6.3 Comparison of modeled and measured sediment values for fall deployment ............................................................... 111
6.4 Modeled values of SSC for the spring deployment ......................... 112
6.5 Modeled values of SSC for the fall deployment ............................... 113
6.6 Comparison of modeled and measured sediment values for fall after DOY 270 ......................................................... 115
6.7 Modeled values of SSC during the fall deployment after DOY 270 .... 117
6.8 Position of sediment bed – water column interface during spring deployment ............................................................. 118
6.9 Position of sediment bed – water column interface during fall deployment ............................................................. 119
6.10 Modeled sediment erosion fluxes ................................................ 121
6.11 Type of erosion behavior ............................................................ 124
List of Tables

2.1 Clay minerals and chemical properties ................................................................. 10
2.2 Clay minerals identified in the San Francisco Bay estuary ................................. 29
3.1 GPS coordinates for experiment stations .............................................................. 34
3.2 Instrument details for the field deployments ....................................................... 35
3.3 Parameter values and standard deviations for ADCP calibration and error analysis ................................................................. 41
4.1 SSC measurements for the LISST PSD shown in Figure 4.5 ........................... 63
5.1 Power law parameters for erosion relationships .................................................... 70
5.2 ADCP S_v to SSC calibration parameters ............................................................ 84
5.3 Average values of SSC and uncertainties for ADCP calibrations ...................... 88
6.1 Input parameters for the sediment and water column models ............................ 107
6.2 Metrics of fit for model predictions of SSC ......................................................... 114
6.3 Power law parameters for model erosion relationships ....................................... 122
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1 Introduction

Estuaries are the transitional areas where fresh water rivers meet and mix with the saline ocean waters forming environments unique from those found in either the rivers or the oceans. Diverse habitats are created along the salinity gradient as the waters transitions from fresh river water to saline ocean water. Estuaries are highly biologically productive owing to the inputs of nutrients from both the ocean and the rivers and are biologically diverse, containing many different habitats supporting innumerable species of flora, fauna and microorganisms. Diverse and abundant natural resources as well as easy access to historic means of transportation via rivers and oceans have led to human settlement of the lands surrounding estuaries. Twenty-two of the 33 largest cities in the world are located on estuaries (Ross, 1988). Industrialization and population growth have impacted estuaries, resulting in contamination, habitat loss and degradation.

The Bay provides habitat for many species of flora and fauna and supports one of the largest shipping ports in the United States. The surrounding urban area was home to 7.15 million people in 2010. The past 160 years of human activity have made the Bay one of the most altered coastal ecosystems in the United States. Recently humans have begun to recognize that estuaries are ecologically important, accumulating sediments and nutrients from river inflows, providing habitat for migrating animals and supporting fish populations caught both commercially and for sport. Estuaries are also essential ecologically, providing a link between freshwater, land and ocean habitats, are responsible for nutrient cycling, nutrient production and the regulation of particle, nutrient and organism fluxes between these habitats and are among some of the most productive natural systems in the world (Levin et al., 2001).

Efforts are underway to restore lost habitat throughout the Bay but these actions must take into account the role the Bay plays in the California economy. Decisions regarding the future of the Bay need to be based upon a scientific understanding of the Bay, its ecosystem and how it will respond to the continually changing environment. Sediments in the Bay are an important piece of the entire system. They are highly dynamic and impact all aspects of the estuarine system including water quality, contaminant transport and commercial shipping. Understanding the behavior of estuarine sediments will allow for better management of the estuary and its resources with less impact upon its ecosystem. Development of the scientific understanding needed requires investigations of sediment behaviour in the field with the intention of quantifying the important sediment processes including erosion, deposition and flocculation. Quantifying these and other sediment processes is important for developing a better understanding of sediment behavior in the Bay with the ultimate goal of improving water quality and ecosystem function.

1.1 The San Francisco Bay Estuary

The San Francisco Bay estuary (the Bay) is one of the largest estuaries in North America with its southern extent near San Jose, California and its northern reach extending to Suisun Bay where the estuary meets the San Joaquin-Sacramento Delta (Figure 1.1) (Conomos et al., 1985).
Figure 1.1: The San Francisco Bay area. Background image is from Google Earth.
The Bay encompasses multiple sub-embayments including the South San Francisco Bay (South Bay), the central San Francisco Bay (Central Bay), San Pablo Bay and Suisun Bay (collectively the North Bay), as well as sloughs, marshes, channels and rivers, covering a total of 1,240 km². Separating the North and South Bays is the Pacific Ocean inlet to the Bay, the Golden Gate. The average depth of the San Francisco Bay is 6 m at mean lower low water (MLLW) but reach 110 m in the Golden Gate channel.

1.1.1 Physical Setting
The Bay receives runoff from 163,000 km² of land, or 40% percent of California. The Delta, an extensive network of channels and islands formed by the confluences of the San Joaquin, Sacramento and other smaller rivers, supplies at least 80% of the mean annual sediment load and 90% of the freshwater which enters the Bay (Conomos et al., 1985). Estimated annual freshwater inflows to the Bay from the Delta are $26 \times 10^9$ m³, with a range of $5 \times 10^9$ m³ to $80 \times 10^9$ m³ for water years 1981 – 2000 (Mckee et al., 2002). The remaining inflow comes from local streams which carry relatively small amounts of sediments (Conomos, 1979; Conomos et al., 1985). In the South Bay during the summer natural inflows are exceeded by sewage water inflows and sediments in the water column are primarily resuspended from the estuary bed (Conomos et al., 1979; Knebel et al., 1977). The average annual sediment flux into the Bay for water years 1994 – 1997 was estimated to be $5.4 \times 10^6$ m³ (Mckee et al., 2002).

The main hydrologic events occur during winter storms and the snowmelt runoff in the late spring. During extreme hydrologic events water originating from the Delta may flow into the South Bay, distinguishable as low salinity high turbidity waters (Carlson and McCulloch, 1974). Winter wind speeds are generally low but remain westerly except as storm fronts pass which typically drive strong winds from the east or southeast. During the summer months prevailing winds are from the west and northwest and are reinforced by the heating of air masses in the Central Valley east of the Bay, resulting in regular afternoon winds. The summer winds can generate waves with a maximum period of 2 – 3 seconds and wave heights of more than a meter. Winter storms can generate waves with a 5 second period.

The tides of the Bay are mixed and semi-diurnal with two high and two low tides each tidal day (24.82 hours). Successive high and low tides are usually different in height although the extent of the difference varies with the lunar cycle. The tidal range at the Golden Gate is 1.3 m while in the southern end of the South Bay the tidal range is 2.6 m. In the North Bay the tidal range is only 1.3 m at Pittsburg, approximately 10 km east of Suisun Bay. The geographic and bathymetric differences which cause the different tidal ranges also cause different tidal waves in the North and South Bays. A standing wave oscillates in the South Bay while a progressive wave propagates through the North Bay (Conomos et al., 1985; Conomos et al., 1979).

1.1.2 Ecological Setting
The benthic and epibenthic bacteria and other microorganisms play a crucial role in the remineralization of organic matter and the cycling of nutrients, including carbon and nitrogen (Levin et al., 2001; Mosier and Francis, 2008; Mosier and Francis, 2010), as well as the transformation or destruction of contaminants, including mercury, selenium, pesticides and polycyclic aromatic hydrocarbons (PAH) (Hollibaugh, 1999; Marvin-DiPasquale et al., 2003). Bacteria are the only significant consumers of dissolved organic carbon within the Bay, converting it into particulate forms accessible to other organisms. Bacteria may also serve as a food source for some benthic organisms (Kimmerer, 2004). Silva (1979) identified 170 different...
The algae and those phytoplankton which settle to the bottom may be a significant food source for the benthic invertebrates including clams, mussels, oysters and polychaetes found in the shallow estuary (Nichols, 1979; Nichols and Thompson, 1985). The benthic invertebrates, along with phytoplankton and zooplankton, are also a major food source for the sport fish found in the Bay, providing a direct path to incorporate sediment bound contaminants like polychlorinated biphenyls (PCB) and mercury into the food web and ultimately to humans (Leatherbarrow et al., 2005). Changes in sediment dynamics can have far reaching affects in the ecosystem. Recent decreases in the suspended sediment concentrations (SSC) are thought to be partially responsible for concurrent changes in the phytoplankton populations as indicated by spring blooms larger than usual, blooms occurring during other seasons and an increase in the baseline concentrations of chlorophyll (SFEI, 2009).

The Bay also supports a number of fish and shellfish populations, some of which have been exploited both commercially and recreationally. More than 100 species of fish and elasmobranchs (sharks and rays) have been identified in the Bay and Delta as have some 70 species of bivalves and species of 30 decapods (shrimp, crabs etc.) (Smith and Kato, 1979). The Bay and Delta also provides an essential connection between the ocean and river habitats of anadromous salmonid populations. Every year the Bay also serves as the wintering and breeding grounds for more than one million water birds (SFEI, 2008). Most of these populations have experience adverse affects due to loss and contamination of habitats and food sources.

1.2 A Changing Estuary

While the areas surrounding the Bay have been inhabited by humans for at least 5,000 years, the last 160 years of human activities have substantially modified the estuarine environment. The effects of water diversions, nonnative species, habitat alteration, destruction and contamination make the Bay one of the most altered coastal ecosystems in the United States.

1.2.1 Hydraulic Mining

The discovery in gold in 1848 in the Sierra Nevada mountain range to the east would have profound and unforeseen effects on the Bay for centuries. Hydraulic mining, introduced in 1853, washed an estimated 9×10^8 m^3 of sediments into the Bay from 1849-1914, or approximately 14×10^6 m^3 yr^-1. The annual supply of sediments prior to hydraulic mining is estimated to be 1.5×10^6 m^3 yr^-1 (Gilbert, 1917). The increased supply of sediments caused rapid siltation and shoaling of the upper arms of the Bay by a few meters. An estimated 9×10^6 kg of mercury, lost during the gold recovery process, was introduced into the Bay along with those sediments (Hornberger et al., 1999). Destruction of agricultural land to the east of the Bay led to a forced end of hydraulic mining in 1884 (Hedgepeth, 1979). Changes in the bathymetry of the Bay as a result of the extra sediment loading are discussed further in Section 2.5.3. Mercury from the gold recovery process now contaminates much of the sediment bed of the Bay (see Section 1.2.3).

1.2.2 Dredging

Large scale dredging operations have occurred in the Bay for more than a century with 17 deep- and shallow-draft channels currently maintained by the US Army Corps of Engineers. Smaller channels, shipping berths and marinas also require periodic dredging. Prior to the introduction of the Long Term Management Strategy (LTMS) in 2001, an average of 4.6×10^6 m^3 of sediment was dredged from the Bay annually, 80% of which was disposed of in the Bay. The
average maintenance dredged volume for 2004 – 2009 was $1.9 \times 10^6$ m$^3$ annually (LTMS, 2001; SFEI, 2009). Maintenance dredging excludes new dredging such as the Oakland Harbor navigation improvement project which deepened the shipping channels to 15 m below MLLW and produced an additional dredged volume of $9.8 \times 10^6$ m$^3$ (Woo et al., 2001). The LTMS goal is to reduce the volume of dredged material disposed of in Bay to $0.8 \times 10^6$ m$^3$, representing 20% of all dredged material, with the other 80% split between ocean disposal and beneficial reuse by 2013 (LTMS, 2001; SFEI, 2009). Dredging has the potential to significantly alter the Bay sediment budget via the dispersal of sediments disposed of in Bay and the removal of those sediments disposed of in the ocean, used for habitat restoration or for other reuse purposes. Additionally, there is the continued concern of disturbing and reintroducing into the ecosystem contaminated sediments that otherwise may have remained sequestered.

1.2.3 Contaminants

Estuarine sediments serve as a source or a sink for many environmental contaminants, particularly those metals and organic compounds that associate with sediment particles once introduced into the Bay. Anthropogenic activities have caused or increased the delivery of a number of environmental contaminants to the Bay, including metals (e.g. mercury, copper and lead), pesticides (e.g. dieldrin, chlordane and dichlorodiphenyltrichloroethane [DDT]) and other organic compounds (e.g. PAH, PCB and polybrominated diphenyl ethers [PBDE]) (Hornberger et al., 1999; Hunt et al., 1998; SFEI, 2007; SFEI, 2009). Some of these contaminants, e.g. PCB, DDT and dieldrin, were banned in the United States for production or most uses in 1979, 1972 and 1974, respectively, but are still found in the Bay and its surrounding watersheds. For these legacy contaminants the greatest hazards is posed by the concentrations already found in the Bay, but even small fluxes into the Bay from upland sources of contaminants may greatly increase the timescale required for concentrations to decrease below recommended levels (Davis, 2004). Others, including the metals and PAH, are still released into the environment today with the annual flux into the Bay contributing a significant source of contamination. Emerging contaminants such as PBDE are found at low concentrations in the Bay but about which little is known in terms of exposure pathways and the environmental risks they pose (SFEI, 2010).

Mercury, PCBs and other contaminants have significant impacts upon the water quality of the Bay and are the cause for the Bay fish consumption advisory in place since 1994 (Russell, 1999). These contaminants are also suspected to be adversely affecting wildlife populations (SFEI, 2007). Sufficient erosion of the sediment bed would reintroduce contaminants back into the Bay ecosystem. Monitoring of these and many other contaminants in the water column and in the sediment bed continue as part of the Regional Monitoring Program for the Bay (SFEI, 2011).

Deposition and burial of contaminated sediments can be an effective means of removing them from the ecosystem; however it is possible that subsequent erosion events will resuspend them, reintroducing them into the ecosystem. Sediment cores indicate the most enriched metal concentrations in the Bay coincide with the rapid industrialization of the San Francisco Bay area following World War II (Hornberger et al., 1999; SFEI, 2007). While the manufacturing, processing and use of PCBs were banned for most applications in 1977, they were still measured in Bay sediments at concentrations as high as 19.9 ppm 10 years later (Hunt et al., 1998). Concentrations were highest in the South Bay and decreased to the north but the Bay wide average was still 9.4 ppb in 2009. Sediment cores indicated that concentrations of PCBs, particularly in the lower South Bay, may be 2-3 times higher below the sediment surface than concentrations found on the surface (SFEI, 2010).
1.2.4 Salt Ponds

The tidal marshes surrounding the Bay have been vastly reduced in size during the last 160 years due to infilling to create new lands and diking to create evaporation ponds for salt production. Over 150 km² of tidal marshes and associated habitat were lost to such ponds in the South Bay (Brew and Williams, 2010). The restoration of 54 km² of salt ponds over the next 50 years is being planned. During salt production the ponds subsided as much as 3 m, caused in part by groundwater overdrafts, requiring an estimated net import of 15 – 50 ×10⁶ m³ of sediments into the ponds to restore them as tidal marshes. As Brew and Williams (2010) point out, two important questions regarding the sediments required for the restoration must be raised: (1) Are there sufficient sediments in the estuarine system to restore the tidal marshes? (2) How will the restoration affect the sediment morphology and intertidal habitats of the entire South Bay? In light of the prior discussion of sediment contamination, if restoration of the salt ponds causes sediment erosion in the Bay the potential for reintroducing contaminated sediments into the ecosystem needs to be evaluated.

1.3 Modeling Sediment in the Bay

A number of numerical models have been developed to predict the transport and fate of sediments and associated contaminants in the San Francisco Bay. Past models focus primarily upon sediment concentrations in the water column, either by design for assessing the impacts of construction projects in the South Bay or because many of the contaminants of concern are associated with sediments and are most readily available to enter the ecosystem when the sediments are actively being mixed into the water column. The result has been simplistic treatment of the sediment bed by the models, caused in part by needs for numerical efficiency but also because of a paucity of data needed to accurately model bed sediments.

With the continually increasing computational power available a 3-dimensional numerical model of the San Francisco Bay estuary is being developed at Stanford University. Currently those efforts are focused upon the hydrodynamics but there is expressed interest in incorporating a sediment module (SFEI, 2009; Wang et al., 2009). The models being developed will be used to predict how the Bay will respond to the changing environment including restoration efforts, sea level rise and water flows from the Delta. However, the parameters needed to model the sediment bed, including bed shear strength, porosity and consolidation relationships are not well known. If accurate models are to be developed, values for these parameters need to be derived from field measurements made in the Bay.

1.4 New Methods for Measuring Suspended Sediments

The development of field deployable instruments capable of measuring suspended sediment concentrations for long periods of time at a high measurement frequency makes it possible to use field data to explore in situ sediment dynamics. This has significant advantages given the difficulty of sampling sediments for later analysis in the laboratory, but in situ measurements pose their own challenges as well.

Current meters for velocity and optical measurements for SSC have long been the standard data collected for studying particle dynamics in estuarine systems (Gross and Nowell, 1983; Kirby and Parker, 1982; Kranck and Milligan, 1992; Noble et al., 2002; Schoellhamer, 1996). These instruments provide the basic information required to calculate sediment fluxes and are capable of producing near continuous data records. Optical instruments, especially in the estuarine environment, are highly susceptible to fouling via the growth of organisms on the
instrument (bio-fouling), rapidly degrading the quality of the data. The use of anti-fouling chemicals or regular maintenance, as frequent as weekly, can reduce these effects. Both of these instruments also make measurements at a point location; the instrumentation of an entire water column requires the deployment of multiple sensors at multiple depths.

While they may never supplant the aforementioned instruments, acoustic instruments offer several advantages and have been widely embraced by the scientific community. Acoustic instruments are less susceptible to bio-fouling allowing them to be deployed unattended in situ for long periods of time, limited primarily by battery life and data storage requirements. Acoustic Doppler velocimeters (ADV) measure water velocities by reflecting an acoustic signal off of particles or bubbles in the water column, providing a 3-dimensional velocity vector and the strength of the returned acoustic signal can be used to measure SSC. ADVs make measurements only at a single point but have the advantage of velocity and SSC measurements being collocated. Acoustic Doppler current profilers (ADCP) operate on similar principles as ADVs but make measurements at multiple locations throughout the water column, providing collocated measurements of velocity and SSC with minimal disturbance to the flow field.

The use of ADCPs to measure SSC has gained considerable traction since the concept was introduced in the late 1990s (Deines, 1999; Holdaway et al., 1999). The process for extracting SSC from the data collected by the ADCP is considerably more involved than for the OBS and ADV instruments mentioned previously and is therefore susceptible to additional sources of error that may affect the accuracy of the SSC measurements. The increasing frequency at which ADCPs are used to measure SSC makes it important to establish the uncertainty associated with those measurements.

1.5 Objectives and Hypotheses

Sediments are a highly dynamic aspect of the San Francisco Bay estuary. In the sediment bed – water column interface sediments may go through numerous cycles of deposition and resuspension prior to becoming part of the consolidating sediment bed or being advected out to sea. These sediments comprise an active layer of sediments which affect water quality and determine the availability of sediment bound and dissolved nutrients and contaminants to the ecosystem.

The primary objective of this research is to quantify the depth of the active sediment layer within the South Bay. Secondary to this is to quantify other sediment processes within the Bay which determine in large part the active sediment layer depth, namely settling velocity, erosion rate parameters, flocculation and consolidation. All of these processes have been well documented in the literature. However, if the best understanding of sediment dynamics in the Bay is to be developed, either experimentally or though numerical models, quantification of these processes in situ will yield the most accurate results. This work seeks to use field measurements of water column suspended sediment concentration and sediment cores to quantify the above processes. The results from these field measurements will also be used to test a simple flocculation model and to determine the efficacy of an acoustic instrument for the measurement of suspended sediment concentrations throughout the water column.

In support of these scientific needs two instrument deployments were completed in the South Bay in February - March and September - October of 2009. Measurements of water velocities, salinity, temperature, and SSC were collected continually for three to four weeks during each deployment. Water samples for SSC measurements and surface sediment samples were collected intermittently and sediment cores were collected at the beginning and conclusion
of each deployment. Together these data comprise a significant source of data for furthering our understanding of the physical processes which affect water column sediment concentrations and water column – sediment bed exchanges. This knowledge allows for more informed decisions about sediment and contaminant management in the Bay to be made to minimize the impacts of historic and current anthropogenic activities.

Chapter 2 presents a review of the current knowledge of estuarine sediments starting with the minerals typically found in the San Francisco Bay estuary. The water column processes of flocculation and settling of estuarine flocs is discussed as are the sediment bed processes of deposition, erosion and consolidation. The use of Beryllium-7 in estuarine sediment research is introduced. Finally, a number of results from previous studies of the sediments of the Bay are introduced, including the clay minerals present, floc sizes and settling velocities, historic changes in the bathymetry of the Bay and previous sediment modeling efforts.

The details for the field observations made in the South San Francisco Bay are provided in Chapter 3. Details of the instrumentation used and samples collected in both the water column and the sediment bed are provided as are details for laboratory analyses made.

Presentation of the data collected appears in Chapter 4. Measurements from the water column include velocity and sediment concentrations using both acoustic and optical instruments. Sediment bed measurements detail particle size distributions, porosity profiles from sediment cores, and the results from laboratory measurements of Beryllium-7 and sediment consolidation.

Analysis of the data is presented in Chapter 5. The measurements of sediment concentration and turbulence were used to test a model of particle flocculation. The data was also used to estimate of floc settling velocities. Measurements from the acoustic Doppler current profilers were calibrated to give sediment concentrations throughout the water column. An error analysis for these measurements is presented along with the concentration measurements.

To better understand the processes of sediment bed – water column exchange a 1-D model of the sediment bed and water column was developed and calibrated, which is presented in Chapter 6. The models will be introduced as will be the values for the parameters required by the models. The model results will be presented followed by the implications of those results with regards to the current understanding of the sediments of the Bay.

Chapter 7 will present the conclusions reached from this work and recommendations for further work. The results from this work indicate the regions of the sediment bed which should be considered when looking at sediment processes in the Bay. Given this information, some of our measurements in the field were at too coarse a resolution. Finer resolution field measurements would be valuable.
2 Current understanding of estuarine sediments

Sediments are a highly dynamic part of the estuarine system. New sediments introduced to the estuary may go through numerous cycles of deposition and erosion before being finally lost to the sediment bed or transported out to sea. In the water column, destabilized sediments are subject to varying levels of turbulence, which at low levels may lead to coagulation but at high levels may lead to breakup. The sediment bed as well is continually consolidating but even this may be disturbed by a strong current or a passing storm front.

Some of these processes have been investigated in the field, others only in the laboratory or numerically. This chapter presents the current state of knowledge of the water column and sediment bed processes to which estuarine sediments are subject. First, clay minerals and clay chemistry are introduced as they are responsible for many of the challenging behaviors of flocculated sediments. Second, the processes of flocculation and floc breakup in the water column are discussed, followed by the modeling of the flocculation process and then floc settling velocity. Third, the processes of deposition, erosion and consolidation in the sediment bed are discussed, followed by a review of Beryllium-7 and how it has been used to investigate sediment processes. Finally, previous studies regarding the San Francisco Bay estuary are reviewed, including measurements of the clay minerals found in the Bay and observations of flocs and settling velocity. Work from the literature detailing the historic bathymetric changes of the Bay is presented followed by numerical models for sediments and contaminants that have been developed for the Bay.

2.1 Clay Minerals

Clay minerals are one of the primary components of estuarine sediments. In freshwater clays form a stable hydrophilic sol. The addition of salts to a stable sol will result in rapid flocculation of the individual particles, destabilizing the suspension. Clay particles in a freshwater river will be stable in suspension resulting in long distance down river transport. In an estuary, the saline ocean waters increases the electrolyte concentration in the bulk solution and destabilizes the clay sol introduced from the river. Destabilization usually occurs at 2,000 – 10,000 ppm salinity or at a few parts per thousand (Gibbs et al., 1983; Krone, 1963). The rapid destabilization and subsequent flocculation of clay minerals is one of the primary causes of the estuarine turbidity maximum observed in many estuaries.

Clays are phyllosilicate minerals comprised of interlayered two-dimensional sheets of silicone-oxygen tetrahedra and two-dimensional sheets of aluminum- or magnesium- oxygen-hydroxyl octahedra. The specific elements contained within the sheets, or the other elements substituted therein as isomorphic substitutions (e.g. the substitution of Al$^{3+}$ for Si$^{4+}$ or Mg$^{2+}$ for Al$^{3+}$), and the proportions in which the sheets occur within the mineral determine the type of clay (van Olphen, 1977). The clays found in the Bay are primarily montmorillonite, illite, kaolinite and chlorite (Knebel et al., 1977; Krone, 1963). The type of clay mineral found in the sediments will affect the size and strength of flocs created in the water column as well as the consolidation
Table 2.1: Clay mineral structure and chemical properties

<table>
<thead>
<tr>
<th>Clay Mineral</th>
<th>Structure</th>
<th>Idealized Structural Formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kaolinite</td>
<td>1:1</td>
<td>$\text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4$</td>
</tr>
<tr>
<td>Montmorillonite</td>
<td>2:1</td>
<td>$\text{Ex}<em>x[\text{Al}</em>{2-x}\text{Mg}_x]\text{&lt;Si}<em>4\text{O}</em>{10}(\text{OH})_2$</td>
</tr>
<tr>
<td>Illite</td>
<td>2:1</td>
<td>$\text{K}<em>{1-x}[\text{Al}<em>2]\text{&lt;Al}</em>{1-x}\text{Si}</em>{3+x}\text{O}_{10}(\text{OH})_2$</td>
</tr>
<tr>
<td>Chlorite</td>
<td>2:1:1</td>
<td>$[\text{Mg},\text{Al}]_3(\text{OH})_6[\text{Mg},\text{Al}]_2&lt;\text{Si},\text{Al}&gt;<em>4\text{O}</em>{10}(\text{OH})_2$</td>
</tr>
</tbody>
</table>

*Tetrahedral layer < >, Octahedral layer [ ], Ex = exchangeable cations.
between the mineral crystal lattices and from desorption of tightly held water molecules at the particle surface (van Olphen, 1977).

The amount of electric double layer compression will determine the stability of clay particle solution. If that repulsive force can be overcome, the attractive forces will bring the particles together resulting in a slow rate of coagulation. With sufficient double layer compression the clay particles will become destabilized and attractive forces will dominate at most separation distances causing rapid flocculation provided the particles are brought within proximity of each other. Mechanisms for bringing particles into close proximity are discussed in Section 2.2.1.

2.2 Water Column Processes

Cohesive sediments within the water column of an estuary generally exist as flocs. In an estuarine setting these flocs may be formed by the destabilization and flocculation of primary clay particles or by biological flocculation in which extracellular polymeric substances from bacteria hold the primary clay particles together to form a stable floc. A single floc contains many primary particles and may attain diameters of 1 cm or more. Turbulence within the water column will cause flocs and particles to collide forming larger flocs during flocculation. Turbulence may also cause weak flocs to breakup into smaller flocs. Finally, some flocs will attain a large enough settling velocity, determined primarily by size, to settle out of the water column altogether.

2.2.1 Flocculation

A stable hydrophobic sol, such as that formed by clay minerals in freshwater, is not truly stable given sufficient time. Settling will remove coarse particles from the sol but the settling of finer particles will be counteracted by Brownian diffusion. Aging of the particles produces larger and more regular particles at the expense of smaller and irregular particles as the system moves towards a lower free energy (van Olphen, 1977). In estuarine systems both of these processes will proceed too slowly to be significant by themselves. Flocculation will produce larger particles over time eventually destroying the stability of the sol as the particles grow large enough to settle out. Even in a salt-free solution, flocculation is known to occur though the rate may be very slow (van Olphen, 1977). In estuarine systems, flocculation can occur rapidly due to the destabilization of the particles by electrolytes in the bulk solution and by increased particle collisions due to hydrodynamics.

Flocculation occurs when two particles collide and stick together forming a new, larger particle. The rate of particle growth was originally modeled by Smoluchowski (1917) but is more commonly presented as:

$$\frac{dn_k}{dt} = \frac{1}{2} \alpha_c \sum_{i+j=k} \alpha_{ij} \beta_{ij} n_i n_j - \alpha_c n_k \sum_{a} \alpha_{ak} \beta_{ak} n_i$$  \hspace{1cm} (2.1)

for the number concentration, $n_k$ [$m^{-3}$], of particles of size $k$ [m]. Particle collisions are modeled by the collision functions $\beta_{ij}$ [$m^3 s^{-1}$] and $\beta_{ik}$ [$m^3 s^{-1}$] and $\alpha_{ij}$ [-] and $\alpha_{ik}$ [-] are size dependent collision efficiency factors included to account for hydrodynamic effects caused by the water in the space between the two particles and interparticle interactions due to van der Waals forces (Han and Lawler, 1992). $\alpha_c$ [-] is an empirical constant used to fit experimental data (Casson and Lawler, 1990). The first term on the right hand side describes the formation of particles of $k$
via the collision of particles of size i and j while the second term on the right hand side describes
the loss of particles of size k by the collision of k sized particles with any other size of particle.
Equations 2.1 is presented in discrete form but can be presented as a continuous function
(Friedlander, 1977).

Three collision functions have been introduced to model different modes of particle
collisions: Brownian motion, fluid shear and differential sedimentation. All three mechanisms
will operate at the same time but for given particle diameters and environmental conditions one
mechanism will usually dominate.

Brownian motion was first described as a method for particle flocculation by
Smoluchowski (1917) in which two particles diffuse toward each other and collide. The particle
diffusion is caused by random collisions between the particles and molecules in the fluid
resulting in random motion modeled by a diffusion coefficient, \( D \) [m² s⁻¹], calculated from the
Stokes-Einstein equation:

\[
D = \frac{k_B T}{3\pi \mu d_p}
\]  

(2.2)

The diffusion coefficient is dependent upon the Boltzmann constant, \( k_B \) [m² kg s⁻² K⁻¹], the
absolute temperature, \( T \) [K], the dynamic viscosity of the fluid, \( \mu \) [kg m⁻¹ s⁻¹], and the particle
diameter, \( d_p \) [m]. The collision frequency function by Brownian motion for particles i and j with
particle diameters \( d_i \) [m] and \( d_j \) [m], respectively, is (Friedlander, 1977):

\[
\beta_{ij}^{br} = \frac{2}{3} \frac{k_B T (d_i + d_j)^2}{\mu d_i d_j}
\]  

(2.3)

In most situations two parcels of water adjacent to one another will travel at different
velocities, whether from turbulence or a shear force applied to the system. Particles travelling
with those parcels of water will also travel at different velocities enabling a faster moving
particle to overtake a slower moving particle. If those particles are close enough they may collide
rather than simply pass by one another, resulting in a collision due to fluid shear. The fluid shear
collision frequency function was first described by Camp and Stein (1943) as:

\[
\beta_{ij}^{sh} = \frac{G (d_i + d_j)^3}{6}
\]  

(2.4)

In the derivation of the above equation \( G \) [s⁻¹] was the absolute velocity gradient at a point.
Camp and Stein (1943) later substitute the root-mean-square velocity gradient, which for a
turbulent fluid may be approximated as (Saffman and Turner, 1956):

\[
G = \sqrt{\frac{\varepsilon}{\nu}}
\]  

(2.5)

where \( \nu \) [m² s⁻¹] is the kinematic viscosity of the fluid and \( \varepsilon \) [m² s⁻³] is the turbulent energy
dissipation rate.
Particle collisions via differential sedimentation occur when a fast settling particle collides with a slower settling particle. The collision frequency function depends upon the difference in settling velocities and the projected area of the two particles when just close enough to collide and is calculated as (Lick et al., 1993):

\[ D_s \beta_{ij} = \frac{\pi}{4} |w_{s,i} - w_{s,j}|(d_i + d_j)^2 \]  

If Stoke’s Law is valid the particle velocities \( w_{s,i} \) [m s\(^{-1}\)] and \( w_{s,j} \) [m s\(^{-1}\)] are proportional to \( d_i^2 \) and \( d_j^2 \), respectively and can be replaced by their Stoke’s Law equivalent to give:

\[ D_s \beta_{ij} = \frac{\pi g}{72 \mu} (\rho_p - \rho_f) |d_i - d_j|(d_i + d_j)^3 \]  

as demonstrated by Ives (1977). In Equation 2.7 the densities of the particles, assumed equal, are \( \rho_p \) [kg m\(^{-3}\)] while \( \rho_f \) [kg m\(^{-3}\)] is the fluid density and \( g \) [m s\(^{-2}\)] is gravitational acceleration. The validity of Stoke’s Law to the settling velocity of estuarine flocs is discussed in Section 2.2.4.

The total particle collision function is the sum of the three mechanisms. However, given two particles of size \( i \) and \( j \) with known settling velocity or densities, and \( T, \mu, G \) for the fluid, one mechanism will typically dominate. In this rectilinear model the collision efficiency factors, \( \alpha_{ij} \) and \( \alpha_{ik} \) in Equation 2.1, are set equal to one. In general, Brownian motion dominates when both particles are smaller than a few \( \mu \)m. Differential sedimentation dominates when one particle is much larger than the other and fluid shear dominates under all other conditions.

Han and Lawler (1992) numerically modeled the approach and collisions of two particles and developed values for the collision efficiency factors to account for the fraction of particle approaches that did not result in collision. In all cases the efficiency factors depended upon the particle diameters. For fluid shear the fluid shear rate, \( G \), is also important while for differential sedimentation the particle densities are important. The result of incorporating these efficiency factors in determining the dominant flocculation mechanism decreased the importance of fluid shear as a collision mechanisms and indicated differential sedimentation was the most significant for a wider range of particle diameters. The model utilizing these collision efficiency factors is referred to as the curvilinear model.

The parameters calculated by Han and Lawler (1992) were based upon numerical models for solid particles. Experiments conducted by Li and Logan (1997a; 1997b) with flocs of latex microspheres showed deviations from the Han and Lawler (1992) curvilinear model as well as from the classical rectilinear model. For flocculation by fluid shear the flocs used by Li and Logan (1997a) had fractal dimensions of 1.89 – 2.47. The fractal dimension, \( n_f \) [-], relates the number of particles in a fractal aggregate, \( N \) [-], to the radius, \( r_a \) [m], as \( N \propto (r_a/r_p)^{n_f} \) where \( r_p \) [m] is the diameter of the primary particle (Kranenburg, 1994). For a solid particle \( n_f = 3 \) and as \( n_f \) decreases to zero the floc becomes infinitely porous. Li and Logan (1997a) found that the fluid shear collision frequency function became less sensitive to \( G \) as \( n_f \) approached 3 but as \( n_f \) decreased towards zero the collision frequency function became proportional to \( G \) as predicted by the rectilinear model.

In laboratory experiments Maggi (2007) found that the fractal dimension was related to the floc size via a power law relationship. During the flocculation process, starting from primary particles and ending with an equilibrium particle size distribution, the power law relationship was found to be constant. It was also anticipated that the changing fractal dimension could
account for changes in floc properties with size, including porosity, excess density, settling velocity and the kinematics of aggregation and breakup. These observations were further explored using numerical modeling in which it was observed that the variable fractal dimension introduced another degree of freedom into fitting laboratory results and that such models were able to recreate the observed PSD satisfactorily (Maggi, 2008; Maggi et al., 2007).

The collision frequency function for differential sedimentation was also found to be dependent upon the fractal dimension (Li and Logan, 1997b). For flocs with \( n_f = 1.81 - 2.33 \) the curvilinear model underpredicted by an order of magnitude while the rectilinear model over predicted by two orders of magnitude. Fluid flowing through the pores of the flocs caused more collisions to occur but the presence of large macropores in the flocs allowed small particles to pass through the floc without contacting the floc. Only a small fraction of the particles with the volume of fluid swept by the floc were captured. Similar results were found by Xiao et al. (2007) using particle image velocimetry.

### 2.2.2 Floc Breakup

Modeling flocculation via the Smoluchowski equation (Equation 2.1), when allowed to progress for some arbitrary long period of time, predicts that flocculation will continue until all of the particles are part of a single, large floc. This contradicts experimental observations that a steady state condition is reached at some point and that some mechanism must keep flocs from growing beyond a certain size. In an early numerical model Fair and Gemmell (1964) imposed a maximum stable floc size in their simulation. They noted fluid shear had been observed to cause floc breakup but were unsure as to whether flocculation simply stopped once the maximum size was reached or if flocs larger than the maximum formed and then broke apart. If they broke apart it was uncertain into how many pieces and into what sizes the flocs would break. Modeling of the breakup process has followed this same trajectory although it is generally accepted that large flocs will form and then break apart into smaller flocs if they are larger than the maximum stable size, \( d_{\text{max}} \), which, as Fair and Gemmell (1964) surmised, is related to the fluid shear rate.

In a turbulent fluid, turbulent eddies will occur along a continuum of scales. The largest eddies are generally limited by a physical characteristic of the system, e.g. fluid depth, on the order of \( 10^1 - 10^2 \) m. The lower limit for the eddies occurs at the scale whereupon viscosity is able to dissipate the energy contained within the eddy, namely at the Kolmogorov microscale, \( \eta \):

\[
\eta = \left( \frac{\nu^3}{\varepsilon} \right)^{\frac{1}{4}}
\]  

(2.8)

in which \( \nu \) is the kinematic viscosity of the fluid and \( \varepsilon \) is the turbulent energy dissipation and \( \eta \) is on the order of \( 10^{-6} - 10^{-3} \) m. Flocs larger than the Kolmogorov microscale will be broken apart by the small scale turbulence (Jarvis et al., 2005). Calculations of \( \eta \) in the Ems Estuary in the Netherlands by van Leussen (1994b) found that during high current velocities the floc sizes were in same range as the Kolmogorov microscale. If the largest stable particle size is proportional to the Kolmogorov microscale then \( d_{\text{max}} \propto \varepsilon^{-1/4} \) and from Equation 2.5 \( G \propto \varepsilon^{1/2} \) it follows that \( d_{\text{max}} \propto G^m \) with \( m = \frac{1}{2} \). Experimental results indicate the exponent \( m \) is between 0 and 2 (Bache, 2004; Bouyer et al., 2005; Jarvis et al., 2005; Manning and Dyer, 1999; van Leussen, 1994b). Variations in the value of \( m \) may be due to differences in the experimental setup including the
use of jar tests, flumes or estuary observations as the experimental systems, the use or lack of flocculating agents like alum or polymers and differences in the material being flocculated, which have included pure clays, estuarine sediments and ferric oxides, among others.

Flocs larger than the maximum stable size are assumed to split into smaller more stable pieces. The number and relative sizes into which the floc splits is not currently understood. Lacking experimental evidence for how this occurs, one of three breakage distribution functions is typically chosen for splitting the floc apart: binary breakage, ternary breakage and normal breakage. Binary breakage splits the floc into two smaller flocs with equal mass. Ternary breakage results in three flocs, one with mass one half that of the original floc and two smaller flocs with masses each one quarter that of the original floc. Normal breakage produces floc fragments that follow a normal distribution with a defined mean and standard deviation (Li et al., 2004; Xu et al., 2008; Zhang and Li, 2003a). The modeling results of Zhang and Li (2003a) found that the binary and normal breakage function produced similar particle size distributions (PSD) while the ternary breakage function shifted the PSD towards smaller particles. Other methods for determining how the floc breaks apart have used random numbers (Kramer and Clark, 1999) and breakage of the unstable particle back into the two original particles (Tambo and Watanabe, 1979).

Floc breakage occurs when the hydrodynamic forces on a floc exceed the interparticle bonding force within the floc. The hydrodynamic force has been parameterized in a number of ways but is usually proportional to the turbulence within the fluid as $G$ or $\varepsilon$, the fluid viscosity, $\mu$ or $\nu$, and the diameter of the floc, $d_p$. (Bouyer et al., 2005; Coufort et al., 2005; Jarvis et al., 2005; Kramer and Clark, 1999; Soos et al., 2008). The interparticle bonding force is thought to be a function of the cohesive forces between primary particles and the number of particle contacts (Feng and Li, 2008; Kobayahsi et al., 1999). A mechanistic understanding of floc strength has yet to be developed as it is not clear how the cohesive and repulsive forces within a floc interact and how, given the complex and fractal nature of flocs, individual particles of the flocs align and interact. Some methods for directly measuring or otherwise inferring floc strength have been developed but this is still not well understood (Jarvis et al., 2005).

2.2.3 Modeling Flocculation

Numerical simulations of particle flocculation generally combine the Smoluchowski equation with a breakage function. Starting with a pulse input of primary particles, flocculation progresses until steady state is reached (Feng and Li, 2008; Kramer and Clark, 1999; Li and Zhang, 2003; Li et al., 2004; Xu et al., 2008; Zhang and Li, 2006). Particle diameter space is divided into a number of bins and masses of particles are moved throughout these bins as they undergo the flocculation and breakup process. With appropriate parameter selection the models are capable of satisfactorily reproducing experimental results. These models can be highly complex; Li et al. (2004) parameterized 9 different models of particles interactions in their model. Parameters for each model of particle interaction require calibration based on experimental data. Such models are difficult to integrate into hydrodynamic models for sediment transport without considerable simplification. Other difficulties arise because most of these models have been developed and calibrated around flocculation as it occurs in water treatment facilities. Experimental data comes primarily from jar flocculation tests often at fluid shear rates higher than those found in estuaries. There is little experimental data for flocculation at fluid shear rates less than 10 s$^{-1}$ e.g. (Hunt, 1982b; Hunt and Pandya, 1984; Zhang and Li, 2003b), or experiments using estuarine sediments e.g. (Manning and Dyer, 1999).
Given the complexity of particle aggregation dynamics a number of simpler models have been developed. Winterwerp (1998) modeled the flocculation process as a diffusive process dependent upon the number and size of particles. Flocculation and breakup were represented as power functions of aggregate diameter, fluid shear rate and floc strength. A final set of simplifications reduced the equation to:

\[
\frac{dd_p}{dt} = k_A CG d_p^2 - k_B G^{3/2} d_p^2 (d_p - d_0)
\]  

(2.9)

where \( k_A [\text{m}^2 \text{kg}^{-2}] \) and \( k_B [\text{s}^{1/2} \text{m}^{-2}] \) are the aggregation and breakup rate constants, \( C [\text{kg} \text{ m}^{-3}] \) is the mass concentration, \( d_0 [\text{m}] \) is the primary particle diameter and \( d_p [\text{m}] \) is a characteristic particle diameter. Winterwerp et al. (2006) found that a similar model to that in Equation 2.9 was able to predict settling velocities over time in three estuaries with overall relative standard deviations of 30-50% once properly calibrated. Trial and error was needed to determine some of the parameters in the model, however.

Hunt (1982a; 1982b) developed a model for flocculation based upon simplifications to Equation 2.1, assuming that for a given particle size one flocculation mechanism would dominate. By dimensional analysis the particle size distribution (PSD) developed by Brownian motion would follow:

\[
\frac{dV}{d(\log_{10} d_p)} = 5.0 A_B \left( \frac{E}{K} \right)^{1/2} d_p^{3/2}
\]  

(2.10)

For fluid shear the PSD would be:

\[
\frac{dV}{d(\log_{10} d_p)} = 6.9 A_{Sh} \left( \frac{E}{G} \right)^{1/2}
\]  

(2.11)

where \( dV [d(\log_{10} d_p)]^{-1} \) is a dimensionless volume distribution function. \( A_B \) and \( A_{Sh} \) are dimensionless constants to be determined experimentally, \( K = k_BT/\mu \) is a dimensional group from the collision frequency function for Brownian motion (Equation 2.3). \( E [\text{s}^{-1}] \) is the volume flux of particles through the PSD. The derivation assumes that for a small particle diameter interval the flux of particles, \( E \), into the interval via coagulation of smaller particles is matched by the flux, also \( E \), of particles out of the interval by coagulation forming larger particles. These assumptions limit the application to long times after the flocculation of an initially mono-disperse suspension begins. No breakup mechanism is included in this model. Particles are removed from the distribution as they flocculate to form particles large enough to settle out of suspension.

With the assumption that the volume flux of particles through the distribution is \( E \) and that the system is at a dynamic steady-state, the flux of particles out of suspension is also \( E \). The change in volume concentration of particles, \( V \) [parts per million by volume (ppm\(_v\))], with time can then be expressed as:

\[
\frac{dV}{dt} = -E(t)
\]  

(2.12)
It has been shown that, for Equations 2.10 and 2.11, the total volume of particles is proportional to volume flux of particles (Hunt, 1980):
\[ E(t) = bV^2(t) \]  
(2.13)
where \( b \) [ppm \( \cdot \) s\(^{-1} \)] is the proportionally constant. From Equation 2.12 and 2.13:
\[ \frac{dV}{dt} = -bV^2 \]  
(2.14)
the solution to which is:
\[ V(t) = \frac{1}{a + bt} \]  
(2.15)
The coefficients \( a \) [ppm \( \cdot \) s\(^{-1} \)] and \( b \) can be determined by plotting \( V(t)^{-1} \) against time. Experimental results for illite and kaolinite agreed favorably with the model predictions in the range of \( G = 0.5 \) s\(^{-1} \) to 8 – 16 s\(^{-1} \). At higher fluid shear rates the particle concentration ceased to decrease with time. From kaolinite data, Hunt (1982a) also found that the rate constant \( b \propto G^{\frac{1}{2}} \). Values of the rate constant for sewage sludge were \( 1 – 10 \times 10^{-3} \) m\(^3\) kg\(^{-1}\) s\(^{-1} \) for fluid shear rates from 0 – 8 s\(^{-1} \) (Hunt and Pandya, 1984).

Farley and Morel (1986) developed a model for solids removal from water by assuming that the flocculation process could be modeled as the sum of three power laws representing the three flocculation mechanisms. Similarity solutions for each flocculation mechanisms acting separately indicated each mechanism could be modeled as:
\[ \frac{dC}{dt} = -BC^n \]  
(2.16)
with the exponent, \( n \) [-], dependent upon the flocculation mechanism, \( B \) [kg\(^{-n}\) m\(^{3n-1}\) s\(^{-1} \)] as the rate constant and \( C \) [kg m\(^{-3} \)] the concentration which would change with time as particles settled out. Calibration of their model against experimental data resulted in:
\[ \frac{dC}{dt} = -B_{Br}C^{1.3} - B_{Sh}C^{1.9} - B_{DS}C^{2.3} \]  
(2.17)
The rate coefficients for Brownian motion, \( B_{Br} \) [kg\(^{-0.3}\) m\(^{0.9}\) s\(^{-1} \)], fluid shear, \( B_{Sh} \) [kg\(^{-0.9}\) m\(^{2.7}\) s\(^{-1} \)], and differential sedimentation, \( B_{DS} \) [kg\(^{-1.3}\) m\(^{3.9}\) s\(^{-1} \)], must be determined experimentally. No mechanism for particle breakup is included in this model. Proper calibration resulted in good predictions of concentration at later times but predicted higher than observed rates of removal early in the flocculation process.

2.2.4 Settling Velocity
Stoke’s Law for calculating settling velocity, \( w_s \) [m s\(^{-1} \)], assumes a balance between gravitational forces and drag forces on a single spherical, solid particle settling with a particle
Reynolds number less than 1. The particle Reynolds number is \( \text{Re}_p = \frac{w_s d_p}{\nu} \) where \( d_p \) [m] is the particle diameter and \( \nu \) [m \(^2\) s\(^{-1}\)] is the fluid kinematic viscosity. Stoke’s Law is a function of \( d_p \) [m], gravitational acceleration, \( g \) [m s\(^{-2}\)], fluid dynamic viscosity, \( \mu \) [kg m\(^{-1}\) s\(^{-1}\)], and the densities of the particle and fluid, \( \rho_p \) [kg m\(^{-3}\)] and \( \rho_f \) [kg m\(^{-3}\)], respectively and is:

\[
w_s = \frac{(\rho_p - \rho_f)gd_p^2}{18\mu}
\]

As estuarine flocs are neither spherical nor solid the validity of using Stoke’s Law, which predicts that \( w_s \propto d_p^2 \), has been questioned. Lick et al. (1993) found that floc settling velocities for Lake Erie sediments flocculated in the laboratory were proportional to \( d_p^{1.56} \) and \( d_p^{1.58} \) for flocs formed in freshwater and seawater, respectively, while Gibbs (1985) found settling velocity to be proportional to \( d_p^{0.78} \). Others have reported values for the exponent to be 1.53 (Mikkelsen and Pejrup, 2001), 1.137 – 1.352 (Manning and Dyer, 1999) and 0.29 – 0.87 for flocs formed in seawater and 0.26 – 2.1 for flocs formed in freshwater (Burban et al., 1990). Some of the variation reported by Burban et al. (1990) is due to the differing condition in which the flocs were formed, specifically the level of turbulence, characterized as the fluid shear rate, \( G \), and the SSC.

As expressed in Stoke’s Law, \( w_s \) is also proportional particle density in excess of the fluid density, \( \Delta \rho = \rho_p - \rho_f \) [kg m\(^{-3}\)]. For flocs the solid particle density \( \rho_p \) is replaced by the floc or aggregate density \( \rho_a \). There is also experimental evidence to suggest floc density changes as the floc diameter changes. The floc density in excess of the fluid density has been related to the floc diameter as:

\[
\Delta \rho = \rho_a - \rho_f = ad_p^{-m}
\]

where \( a \) and \( m \) are empirical constants. Measurements of settling velocity and direct application of Stoke’s Law find that values of \( m \) generally fall between 1 and 2 (Burban et al., 1990; Gibbs, 1985; Lick, 1994; Manning and Dyer, 1999; van Leussen, 1994a). Kranenburg (1994), taking into account the fractal nature of flocs, proposed the excess density of the floc or aggregate \( \Delta \rho_a \) [kg m\(^{-3}\)] is:

\[
\Delta \rho_a = (\rho_p - \rho_f) \left( \frac{d_p}{d_a} \right)^{3-n_f}
\]

where \( d_p \) [m] and \( \rho_p \) [kg m\(^{-3}\)] are the diameter and density of the primary particles of the floc, respectively, \( n_f \) [-] is the fractal dimension and \( d_a \) [m] is the diameter of the floc. Krishnappan (2007) assumed an empirical relationship of the form:

\[
\Delta \rho = \rho_p \exp (-ad_p^b)
\]

with \( a \) and \( b \) found empirically. Krishnappan (2007) used this formulation for \( \Delta \rho \) in Stoke’s Law and treated \( a \) and \( b \) as calibration parameters. Similarly, Winterwerp (2002) incorporated the Equation 2.20 into Stoke’s Law and included a dependence upon the particle Reynolds number to arrive at:
\[ w_s = \frac{\alpha}{18\beta} \left( \frac{\rho_s - \rho_f}{\mu} \right) d_p^{3-n_f} \frac{D^{n_f-1}}{1 + 0.15Re_p^{0.687}} \]  

(2.22)

\( \alpha \, [-] \) and \( \beta \, [-] \) are shape parameters often assumed equal to one. The expression reverts back to Stoke’s Law if spherical solid particles are assumed such that \( \alpha = \beta = 1 \) and \( n_f = 3 \) and \( Re_p \) is much less than 1 as required in the derivation of Stoke’s Law.

Observations have also found that particle concentration affects the settling velocity. In the literature this relationship is often expressed as \( w_{50} \propto C^m \) where \( w_{50} \) is the settling velocity of particles of the median diameter and the value of \( m \) has been found to vary between 0.5 and 3.6 based upon a literature review (Pejrup and Mikkelsen, 2010).

For suspensions of particles, the settling velocity is dependent upon the level of turbulence, usually expressed as the fluid shear rate, \( G \), under which the flocs were formed. As discussed in Section 2.2.1 low levels of turbulence can help particle flocculation resulting in larger flocs with greater settling velocity but high levels of turbulence can cause floc breakup decreasing floc size and settling velocity. Based upon this reasoning van Leussen (1994a) proposed:

\[ w_s = w_{50} \frac{1 + aG}{1 + bG^2} \]  

(2.23)

\( a \, [s] \) and \( b \, [s^2] \) are empirical constant. For kaolinite \( a = 12 \, s \) and \( b = 3 \, s^2 \) while for mud from the Ems Estuary \( a = 0.3 \, s \) and \( b = 0.09 \, s^2 \). Arguing that the above does not accurately capture the increased settling velocity due to the flocculation effects of low levels of turbulence Pejrup and Mikkelsen (2010) proposed the following relationship:

\[ w_{50} \propto C^p [k_1 + (k_2 + G^n)q^{-n}] \]  

(2.24)

\( k_1, k_2, p \) and \( q \) are empirical constants.

Experimental results show that settling velocity is affected by \( C, G, d_p, \Delta \rho \) and \( n_f \). No single expression has successfully combined these parameters to predict settling velocity and the opinion of Pejrup and Mikkelsen (2010) is that there is little hope of ever doing so. However, observations of flocs both from the field and from the laboratory find that settling velocities generally fall within a known range. In laboratory observations, reported values are \( 0.1 \leq w_s \leq 1.0 \, mm \, s^{-1} \) (Burban et al., 1990; Lick, 1994; Lick et al., 1993; Manning and Dyer, 1999). Using a different flocculator design, Lick et al. (1993) reported settling velocities as high as 10 mm s\(^{-1}\) for flocs with diameters greater than 1 mm.

Measurements of settling velocities in situ find a similar lower limit for settling velocity when compared to the values above. In situ measurements of large flocs have found settling velocities approaching 10 mm s\(^{-1}\), in agreement with the observations of Lick et al. (1993). For flocs collected one meter below the surface by Pejrup and Mikkelsen (2010), settling velocities were \( 0.1 - 0.6 \, mm \, s^{-1} \) while for flocs collected one meter above the bed, settling velocities were \( 0.2 \leq w_s \leq 2.5 \, mm \, s^{-1} \) was observed. Fluid shear rates ranged from \( 0.5 - 3.8 \, s^{-1} \) and \( 1.9 - 14.6 \, s^{-1} \) at the sampling locations, respectively. In the Ems Estuary, van Leussen (1994a) measured the settling velocity for flocs with diameters of 200 – 700 \( \mu \)m to be \( 0.5 - 8 \, mm \, s^{-1} \) while in the Dutch coastal zone flocs with diameters less than 80 \( \mu \)m were observed to settle at \( 0.5 - 2 \, mm \, s^{-1} \).
Manning and Bass (2006) found that during periods of high suspended sediment concentrations flocs with diameters greater than 480 μm formed and had settling velocities of 4 – 8 mm s\(^{-1}\) while during periods of high turbulence the microflocs (d\(_p\) < 160 μm) settled with velocities of 1.45 mm s\(^{-1}\) while the macroflocs (d\(_p\) > 160 μm) settled slower at 1.1 mm s\(^{-1}\).

The measurement of turbulence and SSC time series in an estuarine setting make it possible to estimate the settling velocity. By basing the estimates on changes in concentration over time this method in unable to resolve the size or settling velocity of a single floc but rather produces average values for all the flocs found within the measurement volume. This method requires significantly less effort for data collection allowing measurements to be made for long periods of time. Fugate and Friedrichs (2002) made field measurements in the Chesapeake Bay, Virginia, of suspended sediments and analyzed the data assuming a balance between gravitational settling and turbulent diffusion for the sediments. In general form the advection – dispersion equation for sediments is:

\[
\frac{\partial C}{\partial t} + \frac{\partial (uC)}{\partial x} + \frac{\partial (vC)}{\partial y} + \frac{\partial (wC)}{\partial z} + w_s \frac{\partial C}{\partial z} = \frac{\partial}{\partial x} \left( D \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left( D \frac{\partial C}{\partial y} \right) + \frac{\partial}{\partial z} \left( D \frac{\partial C}{\partial z} \right)
\]  
(2.25)

The first term is the unsteadiness in the suspended sediment concentration, C [kg m\(^{-3}\)], in time. Terms 2 – 4 are advection with velocities u, v and w [m s\(^{-1}\)] along the x, y and z axes, respectively. The last term on the left hand side is gravitational settling with velocity w\(_s\) [m s\(^{-1}\)], z positive upwards. The terms on the right hand side are diffusion in the x, y and z directions. Limiting this to two dimensions in the x and z directions and neglecting advection in the vertical direction and diffusion in the horizontal direction and then applying Reynolds averaging to the remaining terms gives:

\[
\frac{\partial \bar{C}}{\partial t} + \frac{\partial (\bar{u}C)}{\partial x} + w_s \frac{\partial \bar{C}}{\partial z} = - \frac{\partial \left( \bar{w'}C' \right)}{\partial z} + \frac{\partial}{\partial z} \left( D \frac{\partial \bar{C}}{\partial z} \right)
\]  
(2.26)

Over bars denote the mean components and primes denote the fluctuating components from the Reynolds average. The diffusion coefficient, D [m\(^2\) s\(^{-1}\)], and w\(_s\) are assumed constant. The eddy viscosity in the vertical direction, K\(_z\) [m\(^2\) s\(^{-1}\)], is defined as:

\[
K_z = - \frac{\bar{w'C'}}{\bar{C}}
\]  
(2.27)

Rearranging and substituting Equation 2.27 into Equation 2.26:

\[
\frac{\partial \bar{C}}{\partial t} + \frac{\partial (\bar{u}C)}{\partial x} + w_s \frac{\partial \bar{C}}{\partial z} = \frac{\partial}{\partial z} \left( K_z \frac{\partial \bar{C}}{\partial z} \right) + \frac{\partial}{\partial z} \left( D \frac{\partial \bar{C}}{\partial z} \right)
\]  
(2.28)

Finally, for the conditions under consideration here, D is much smaller than K\(_z\). Substituting Equation 2.27 for K\(_z\) and neglecting the last term on the right hand side of Equation 2.28 gives:
\[
\frac{\partial \tilde{C}}{\partial t} + \frac{\partial (u \tilde{C})}{\partial x} + w_s \frac{\partial \tilde{C}}{\partial z} = - \frac{\partial (w'C')}{\partial z}
\] (2.29)

Fugate and Friedrichs (2002) assumed that gravitational settling was more important than unsteadiness or advection. This assumption will be qualified later. Dropping the first and second terms then yields:

\[
w_s \frac{\partial \tilde{C}}{\partial z} = - \frac{\partial (w'C')}{\partial z}
\] (2.30)

which can be integrated with respect to z to give:

\[w_s \tilde{C} = -\bar{w'C'}\] (2.31)

\(w_s\) is considered a constant and can be estimated as:

\[w_s = -\frac{\bar{w'C'}}{\bar{C}}\] (2.32)

For settling to balance the turbulent diffusion of sediments two conditions need to be met:

\[
\frac{\frac{\partial \tilde{C}}{\partial t}}{w_s \frac{\partial \tilde{C}}{\partial z}} \ll 1
\] (2.33)

\[
\frac{\frac{\partial (u \tilde{C})}{\partial x}}{w_s \frac{\partial \tilde{C}}{\partial z}} \ll 1
\] (2.34)

From their data in the Chesapeake Bay, Fugate and Friedrichs (2002) identified three different particle populations: fine particles, which did not settle out over the timescale of the tides, a easily resuspended fraction with settling velocity of approximately 1 mm s\(^{-1}\) and a slow settling population, which, after resuspension, settled out over roughly 12 hours.

### 2.3 Cohesive Sediment Beds

The processes of deposition, erosion and consolidation are highly dependent upon one another. Deposition is caused by the settling of sediments out of the water column. Those sediments may be immediately resuspended back into the water column or may spend some amount of time as part of the sediment bed. Those sediments that are not immediately resuspended will start to undergo consolidation where the open structure of the flocs and sediments begin to collapse. The consolidation process increases the shear strength of the bed making it less likely to erode.
2.3.1 Deposition

The rate of deposition, \( D_s \), typically as mass per area per time, may be expressed most simply as \( D_s = w_s C_b \), where \( w_s \) is the settling velocity and \( C_b \) is the near bed concentration. This simple expression is typically modified to incorporate the effects of the bed shear stress on the rate of deposition as (Lumborg, 2005; Sanford and Halka, 1993):

\[
D_s = \begin{cases} 
  w_s C_b \left(1 - \frac{\tau_b}{\tau_d}\right); & \tau_b < \tau_d \\
  0; & \tau_b \geq \tau_d 
\end{cases}
\]  

(2.35)

\( \tau_b \) [Pa] is the bed shear stress and \( \tau_d \) [Pa] is the critical shear stress for deposition. Lumborg (2005) gives empirical values for \( \tau_d \) of 0.05 – 0.1 Pa while Berlamont et al. (1993) reports values of 0.05 – 0.2 Pa but notes that for larger flocs the value may be as high as 1.0 Pa. The above formulation precludes erosion while the bed shear stress is below the critical value.

Data and analysis presented by Sanford and Halka (1993) indicated that deposition began shortly after the bed shear stress began to decrease but was still greater than the critical shear stress required for erosion to occur (see Section 2.3.2). Modeling results indicated that the experimental data were better represented assuming continuous deposition proportional to the near bed concentration. The near bed concentration is an ambiguous parameter. For a numerical model, that concentration is likely to be the concentration of the grid cell closest to the bed. In a field or laboratory setting this parameter is less well defined.

McDonald and Cheng (1997), in a depth averaged model of the San Francisco Bay used Equation 2.34, modified to account for the depth averaging, but only for concentrations below a critical value of \( C_{cr} = 0.30 \) kg m\(^{-3}\). Above this critical value Equation 2.35 was modified further, replacing \( C_b \) with \( C^{5/3} C_{cr}^{2/3} \) in which \( C \) is the depth averaged concentration. This additional modification increased the rate of deposition even further when the concentration exceeded the critical value.

2.3.2 Erosion

Sediment erosion has been separated into Type I and Type II behavior. In Type I erosion, the critical shear stress for erosion, \( \tau_c \) [Pa], increases with depth into the sediment bed. For a given bed shear stress the erosion depth into the bed is then limited to the depth at which the bed shear stress and the critical shear stress become equal. Conversely, for Type II erosion, the critical shear stress is not a function of depth. Once the bed shear stress exceeds the critical shear stress, erosion of the sediment bed will continue indefinitely or until the bed shear stress decreases below the critical value.

Several formulations for the sediment erosion rate, \( E_s \), as mass per area per time, can be found in the literature. The three most prominent relate \( E_s \) to the bed shear stress, \( \tau_b \), as linear, exponential and power law expressions. The linear expression (McDonald and Cheng, 1997; Sanford and Halka, 1993; Sanford and Maa, 2001) is:

\[
E_s = M \left(\frac{\tau_b}{\tau_c} - 1\right); \quad \tau_b > \tau_c
\]  

(2.36)

\( M \), in this formulation and all subsequent erosion formulas, is the erosion rate constant which is determined empirically. The units and values for \( M \) vary between erosion equations. For \( \tau_b \) less
than \( \tau_c \), \( E_s \) is zero. \( \tau_c \) is generally assumed to be independent of depth into the sediment bed and therefore is often used to describe Type II erosion. Equation 2.36 is also known as the Ariathurai-Partheniades equation, and may also be expresses as \( E_s = M'(\tau_b - \tau_c) \), subject to the same constraints with \( M' = M \cdot \tau_c^{-1} \).

The power law expression for erosion takes the form (Lumborg, 2005; Sanford and Maa, 2001):

\[
E_s = M \left( \frac{\tau_b}{\tau_c} - 1 \right)^n \quad (2.37)
\]

\( \tau_c \) may also be a function of depth and both \( M \) and \( n \) are empirical parameters. Several modified power law expressions also exist. \( E_s = M \tau_b^n \) assumes no critical value for erosion (Lavelle et al., 1984; MacIntyre et al., 1990; Sanford and Halka, 1993). A review of the literature by Lavelle et al. (1984) found that, for the modified expression, \( n = 1.2 - 5.0 \). Lick and McNeil (2001) include a power law dependence upon the sediment bulk density, \( \rho_b \), as well to give \( E_s = M \tau_b^n \rho_b^m \), in which \( m, n \) and \( M \) are all constants and \( E_s \) is the erosion rate in cm s\(^{-1} \) which was found to decrease with increasing values of \( \rho_b \).

The exponential expression is (Kuijper et al., 1989; Sanford and Halka, 1993; Sanford and Maa, 2001):

\[
E_s = M e^{\alpha [\tau_b - \tau_c(z)]} \quad (2.38)
\]

where \( \alpha \) is determined empirically and \( n \) is typically 0.5. The exponential formulation explicitly allows the a priori definition of a critical shear stress profile into the sediment bed, \( \tau_c(z) \), and is often used to describe Type I erosion. A depth dependent \( \tau_c \) in the power law formulation (Equation 2.37) could also be used for Type I erosion.

Amos et al. (1992), using \textit{in situ} flume measurements in the Bay of Fundy, found that Type I and Type II erosion behavior could exist in the same location with Type I erosion occurring near the surface and Type II erosion occurring at some lower depth. Sanford and Maa (2001) developed an expression which could incorporate Type I and Type II erosion:

\[
E_s = M(z)(\tau_b - \tau_{c,0})e^{-[\gamma \beta(t-t_0)]} \quad (2.39)
\]

where \( M(z) = \rho_s \phi_s(z) \beta \) in which \( \rho_s \) is the density of the solid particles, \( \phi_s(z) \) is the solid fraction as a function of depth, \( z \), into the sediment bed and \( \beta \) is an empirical constant. \( \gamma = d\tau_c/dz \) and \( \tau_{c,0} \) is the critical shear stress when \( \tau_b \) is first applied at time \( t=t_0 \). Equation 2.39 can also be expressed as a function of mass rather than depth. Sanford and Maa (2001) found that the type of erosion behavior was as much a function of the time rate of change of the forcing as it was a function of the depth rate of change of \( \tau_c \).

In an effort to better simulate the beginning of erosion when the bed shear stress is nearly the same at the critical value, Van Prooijen and Winterwerp (2010) developed a stochastic erosion equation:

\[
E_s = M \rho_w \sigma_u^2 \int_{\tau_c}^{\tau_b} (T_b - T_c) r(T_b) dT_b \quad (2.40)
\]
\( T_b, T_c \) and \( \sigma_u^2 \) are non-dimensional values for bed shear stress, critical shear stress and the standard deviation of the near bed velocity, respectively. \( r(T_b) \) is a non-dimensional distribution of the bed shear stress. Equation 2.40 was solved for a particular distribution of \( r(T_b) \) but the solution was inconvenient to incorporate into numerical models and instead a parameterization was used. The stochastic formulation was only used when the bed shear stress was within roughly a factor of two of the critical shear stress. When \( \tau_b > 1.7 \tau_c \) Equation 2.35 was used.

The values and dimensions for the parameters that appear in each of the above erosion formulae are specific to that formulation. Thus, it is not possible to use the parameters to compare erosion behavior between sediments. Sanford and Halka (1993) found that for their data the erosion rate, \( E \), was nearly the same regardless of the erosion model used. Their observations only measured small values of the bed shear stress, \( \tau_b < 0.2 \) Pa. Above that level of bed shear stress the erosion rate for the linear model diverged significantly from the erosion rates for the power law and exponential formulations. Presumably the erosion formulation chosen would have become more significant at higher bed shear stresses.

### 2.3.3 Consolidation

Once particles become part of the sediment bed via sedimentation they undergo the process of consolidation. Consolidation results in the expulsion of pore water from the flocs as they compress. Over time the excess water makes it way to the sediment surface and is thus removed from the sediment bed system. The expulsion of the water results in a decrease in porosity and an increase in inter-particle contacts. In an estuarine system the effect is a denser sediment bed which is more resistant to erosion. Consolidation is also relevant to dredged sediment disposal, determining whether sediments are quickly dispersed or consolidate and remain in place.

Rapid changes occur to the sediments during the first few hours and days of consolidation but smaller changes may be observed for months. Current consolidation theory is based upon the work of Gibson et al. (1967), which used the conservation of mass and assumes Darcian flow of water through the sediments. Equivalent relationships using different variables and coordinate systems have also been derived by Toorman (1996) and Merckelbach and Kranenburg (2004). Using the volume fraction of solids, \( \phi_s \) [-], equal to one minus the porosity, as the dependent variable, consolidation may be modeled as (Merckelbach and Kranenburg, 2004):

\[
\frac{\partial \phi_s}{\partial t} - \frac{\rho_s - \rho_f}{\rho_s} \frac{\partial}{\partial z} \left[ k \phi_s^2 \right] - \frac{1}{\rho_f g} \frac{\partial}{\partial z} \left[ k \phi_s \frac{\partial \sigma'}{\partial z} \right] = 0
\]  

(2.41)

Equation 2.41 is also commonly presented with the void ratio, \( e \) [-], as the dependent variable:

\[
-\frac{\partial e}{\partial t} = \left( \frac{\rho_s}{\rho_f} - 1 \right) \frac{\partial}{\partial e} \left[ k(e) \frac{\partial e}{1 + e} \right] + \frac{\partial}{\partial z} \left[ \frac{k(e)}{\rho_f g (1 + e)} \frac{\partial \sigma'(e)}{\partial z} \right]
\]  

(2.42)

and is known as the Gibson equation (Gibson et al., 1967). In the above equations \( k \) [m s\(^{-1}\)] is the hydraulic conductivity, \( \sigma' \) [Pa] is the effective stress and the void ratio, \( e \), is defined as:

\[
e = \frac{\text{Volume voids}}{\text{Volume solids}} = \frac{\phi}{1 - \phi}
\]  

(2.43)
where $\phi$ [-] is the porosity. The effective stress is related to the total vertical stress, $\sigma$ [Pa], as $\sigma = \sigma' + p = \sigma' + p_e + p_h$ in which $p$ [Pa] is the total pore water pressure equal to the sum of the excess pore water pressure and the hydrostatic water pressure, $p_e$ [Pa] and $p_h$ [Pa], respectively. Numerical solution is needed for either Equation 2.41 or Equation 2.42 and a relationship between $k$ and either $e$ or $\phi$ must be assumed or determined from experimental data. Hawlader et al. (2008) and Bartholomeeusen et al. (2002) summarize many of the equations that have appeared in the literature.

Laboratory studies of consolidation usually allow a well mixed suspension of sediments to settle quiescently. The height of the sediment water interface is monitored with time and measurements of the pore water pressure can be made with pressure transducers or standpipes. X- and $g$-rays have proven to be useful for non-destructive measurements of density over time (Been, 1981; Been and Sills, 1981; Krinitzsky, 1972; Merckelbach and Kranenburg, 2004). The measurements of Been and Sills (1981) show that the excess pore water pressure decreases over time throughout the sediment column.

In an estuarine setting, the disposal of dredging spoils might be represented as a well mixed suspension of sediments but under other conditions the deposition of sediments is likely to be less uniform. Sills and Thomas (1984) found that the consolidation process is dependent upon the rate of deposition of sediments onto the bed. Slower deposition resulted in a less compact sediment bed even after 50 days of consolidation. It was not clear from the data if, upon the dissipation of the excess pore water pressure and completion of consolidation, sediment beds of similar height would result despite differing rates of deposition.

A number of numerical solutions have been developed to predict the consolidation process. Bartholomeeusen et al. (2002) compared the predictions made by several models, all of which were calibrated by their respective developers based upon the same data. Compared against a similar consolidation experiment as was used for the calibrations, all of the models over predicted the rate of consolidation at early times when the strain rate was large. After seven days of consolidation, most of the models predicted the final height to be less than was observed in the laboratory experiment they were modeling. Bartholomeeusen et al. (2002) indicated that a time- or rate-dependent correlation between the void ratio and the hydraulic conductivity needs to be developed to better predict consolidation at early times.

The time required for numerical solutions of Gibson’s equation make it impractical for many predictions, particularly large scale long time period models as would be needed for an estuary wide sediment model. de Boer et al. (2003) presented a parameterized version of the Gibson equation. By limiting the problem to thin layers with short consolidation times the Gibson equation was simplified to a diffusion equation for the excess pore water pressure. An analytical solution exists under certain initial conditions. The model required constants to be derived from experimental results and performed as well as some of the numerical solutions to Gibson’s equation. The authors note that the parameterized model over predicts the density in deeper sediment layers and under predicts the density near the surface.

Merckelbach and Kranenburg (2004) developed analytical solutions of consolidation during the initial and final stages of consolidation by simplifying Gibson’s equation. Power law relationships were assumed to describe the permeability and effective stress as functions of the volume fraction of particles. Starting from Equation 2.41 they argue early in the consolidation process the effective stress is negligible allowing them to drop the last term on the left hand side, producing an equation with an analytical solution. During the final stages of consolidation Merckelbach and Kranenburg (2004) argue that a steady state solution for the solid fraction is
justifiable, again leading to an equation with an analytical solution. The solutions allowed for the determination of the parameters needed for the permeability and effective stress equations. It is unclear from their data how well the models developed can make predictions of consolidation.

2.4  Beryllium-7

Beryllium (Be) is an alkaline earth metal with one stable isotope, $^9\text{Be}$, and multiple radiogenic isotopes (Holden, 2004). Two isotopes, $^7\text{Be}$ ($T_{1/2} = 53.12$ days) and $^{10}\text{Be}$ ($T_{1/2} = 1.5 \times 10^6$ years), are similarly generated in the atmosphere but their half lives make them useful for scientific measurements on widely different timescales (Jaeger et al., 1996; Kaste et al., 2002). $^7\text{Be}$ has been used to study sediment processes in many aquatic environments due to its ease of measurement, well defined source term and a half-life of similar length to the timescale of sediment processes, namely days to months (Allison et al., 2000; Baskaran and Swarzenski, 2007; Corbett et al., 2007; Dail et al., 2007; Dibb and Rice, 1989; Feng et al., 1999a; Feng et al., 1999b; Fitzgerald et al., 2001; Griffin and Corbett, 2003; Kaste et al., 2002; Olsen et al., 1986; Woodruff et al., 2001).

2.4.1  Beryllium-7 in the Atmosphere

$^7\text{Be}$ is the product of cosmic ray spallation of oxygen and nitrogen primarily in the stratosphere, though some is produced in the troposphere and at the earth surface (Kaste et al., 2002). The rate of production is dependent upon the flux of cosmic rays which varies with latitude, altitude, the 11 year solar cycle and large solar events (Kaste et al., 2002; Megumi et al., 2000; Yoshimori, 2005).

As $^7\text{Be}$ readily attaches electrostatically to particles in the atmosphere the behavior of atmospheric particles determine in large part the fate of atmospheric $^7\text{Be}$. With a residence time six times longer than the half-life $^7\text{Be}$, activities in the stratosphere are assumed to be in equilibrium while in the troposphere residence times are much shorter and more variable due to rapid particle washout (Kaste et al., 2002). A review of the literature by Kaste et al. (2002) found $^7\text{Be}$ in the stratosphere to range from 0.16 – 0.58 Bq m$^{-3}$ while troposphere activities ranged from 0.001 – 0.02 Bq m$^{-3}$. Doering and Akber (2008) measured near surface $^7\text{Be}$ activity for four years in Australia, measuring activities from 0.0006 – 0.0103 Bq m$^{-3}$ while in Japan Kikuchi et al. (2009) measured activities as high as 0.02 Bq m$^{-3}$ over an eight year period. Mean near surface activities generally fall within the 0.002 – 0.01 Bq m$^{-3}$ range (Daish et al., 2005; Doering and Akber, 2008; Duenas et al., 2009; Kaste et al., 2002; Kikuchi et al., 2009; Megumi et al., 2000).

A number of authors also report seasonal variations in the near surface $^7\text{Be}$ activities. Yoshimori (2005) found that in Tokyo $^7\text{Be}$ activities peaked in spring and late summer with activities roughly 3 times higher than those measured during summer and fall. A 15 year record of near surface activities in Osaka, Japan show a similar pattern (Megumi et al., 2000). From October to April $^7\text{Be}$ activities varied from 0.005 – 0.01 Bq m$^{-3}$ while from July to August activities decreased to 0.002 – 0.004 Bq m$^{-3}$. Doering and Akber (2008) attributed higher near surface activities to the injection of stratospheric air into the near surface during the early spring and to higher rates of convective circulation in the troposphere during the summer.

Seasonal variations in the atmospheric activities of $^7\text{Be}$ are also reflected in the fluxes of $^7\text{Be}$ to the earth surface. Olsen et al. (1985) found $^7\text{Be}$ fluxes ranged from $1 \times 10^{-5}$ – $18 \times 10^{-5}$ Bq m$^{-2}$ s$^{-1}$ at Oak Ridge, Tennessee and from $3 \times 10^{-5}$ – $14 \times 10^{-5}$ Bq m$^{-2}$ s$^{-1}$ at Norfolk, Virginia.
the highest fluxes occurring during March through May at both sites. The average daily fluxes over the two years of measurement were $6.4 \times 10^{-5} \text{Bq m}^{-2} \text{s}^{-1}$ and $6.9 \times 10^{-5} \text{Bq m}^{-2} \text{s}^{-1}$ at Oak Ridge and Norfolk, respectively. Baskaran and Swarzenski (2007) found the depositional flux of $^7\text{Be}$ over the course of a year in St. Petersburg, Florida was $2 \times 10^{-5} - 28 \times 10^{-5} \text{Bq m}^{-2} \text{s}^{-1}$. Dry fluxes in St. Petersburg varied from $7 \times 10^{-7} - 1.4 \times 10^{-5} \text{Bq m}^{-2} \text{s}^{-1}$. Dry deposition is generally found to contribute < 10% of the $^7\text{Be}$ inventory (Baskaran and Swarzenski, 2007; Kaste et al., 2002) however Yamamoto et al. (2006) found that seasonal differences in activities could not be fully explained by wet deposition variations.

2.4.2 Beryllium-7 in Estuarine and Coastal Environments

The $^7\text{Be}$ that reaches the terrestrial environment does not appear to be a significant source of $^7\text{Be}$ to estuaries. Olsen et al. (1986) found that freshwater supplied less than 5% of the $^7\text{Be}$ to the James Estuary in Virginia, indicating the drainage basin vegetation and soil sequestered most of the precipitation derived $^7\text{Be}$. Deposition to the estuarine environment should primarily be from direct wet or dry deposition to the water surface.

When $^7\text{Be}$ enters the aquatic environment, it is solubilized and then may be scavenged by particles (Feng et al., 1999b). In the estuarine environment, $^7\text{Be}$ is found dissolved in the water column, attached to suspended sediment particles and in the sediment bed with the partitioning coefficient ($K_d; \text{Bq kg}^{-1} (\text{Bq m}^{-3})^{-1}$) between the water column and the suspended sediments found to vary between $10^{2.8} - 10^{6.0}$ (Kaste et al., 2002). The distribution of $^7\text{Be}$ between suspended particles and the water column has been found to be significantly influenced by the suspended sediment concentration and the length of time particles remain in suspension. Feng et al. (1999a; 1999b) found that the water column activity was affected by the particle abundance, surface area and residence time in the water column. In the Hudson River Estuary the residence time for $^7\text{Be}$ was <1 – 13 days. In most cases >80% of the $^7\text{Be}$ supplied to the water column was removed by scavenging and deposition of particles. Water column activities were $0.3 – 1.7 \text{Bq m}^{-3}$ and $1.2 – 20 \text{Bq m}^{-3}$ in the surface and bottom waters, respectively. Suspended sediment collected near the surface and near the bottom had activities of $50 – 180 \text{Bq kg}^{-1}$ and $50 – 120 \text{Bq kg}^{-1}$, respectively. The average surface sediment activity in the Hudson River estuary was $33 \pm 22 \text{Bq kg}^{-1}$. In a number of estuaries along the Texas coast Baskaran and Santschi (1993) observed similar behavior for $^7\text{Be}$ in the water column. Within one hour 74 – 86% of $^7\text{Be}$ supplied by wet deposition was associated with particles and more than 70% was removed from the water column within 1.5 days. As a result of low SSC, 60 – 90 % of water phase $^7\text{Be}$ was not associated with particles in estuaries and the coastal ocean along the eastern Unites States (Olsen et al., 1986). The highest activities on suspended particles were found in high energy areas where fine particles were continually resuspended into the water column. The lowest activities on suspended particulates were collected from sites where fine particle deposition rates were greater than 3 cm yr$^{-1}$.

2.4.3 Sediment Studies Using Beryllium-7

Activities of $^7\text{Be}$ in sediment beds have also been used to examine sediment deposition and resuspension. Particles in the sediment bed with measurable activities of $^7\text{Be}$ must have scavenged it from the water column and then settled to the bed within the past few months. Repeated sampling, typically as sediment cores, can be used to estimate short term deposition and erosion rates and to estimate the depth of the sediment mixed layer.
Dibb (1989) measured $^{7}$Be in the sediments of Chesapeake Bay from 0 – 6 cm and from 6 – 12 cm. More than 90% of the measured $^{7}$Be was in the upper 6 cm. The estimated bay wide average $^{7}$Be inventory was 540 Bq m$^{-2}$, which agreed well with the estimated 510 Bq m$^{-2}$ supported by atmospheric deposition. At some locations, measured activities were higher than could be supported by atmospheric deposition. Dibb (1989) interpreted this as sediment focusing, the deposition of sediments that originated from other locations in the estuary.

At two locations in the Fox River, Wisconsin, Fitzgerald et al. (2001) sampled sediments at 1 cm intervals. Sediment $^{7}$Be ranged from non-detect to $3\times10^4$ Bq m$^{-3}$ and total inventories in the sediment bed were 320 – 1380 Bq m$^{-2}$. At one site the peak activity occurred at a depth of 4 – 5 cm into the sediment bed indicating recent rapid deposition or mixing to at least that depth. The sediment core collected one month later from the same site showed a nearly uniform activity profile down to 4 cm and an increase in total activity by a factor or 2.4, indicating mixing and deposition of up to 4 cm of $^{7}$Be rich sediments. Subsequent cores showed decreasing $^{7}$Be inventories consistent with decay only but changes in the profiles with depth suggest the top 1 – 2 cm of sediment were still being reworked. The loss of $^{7}$Be was quicker than could be attributed to decay, which is indicative of net erosion of sediments between sampling periods. At the second site in the Fox River the $^{7}$Be inventories indicated net erosion occurred prior to the collection of the last sediment core. This period coincided with hydrologic conditions known to produce erosion at the site. Short term rates of sediment deposition derived from the $^{7}$Be were 0 – 65 cm yr$^{-1}$. $^{137}$Cs profiles from the same sediment cores, which average deposition rates over a much longer time period, estimated rates of 0.3 – 0.5 cm yr$^{-1}$.

Sediment cores collected in the Hudson River estuary by Woodruff et al. (2001) had a mean $^{7}$Be activity of $53 \pm 33$ Bq kg$^{-1}$ and a range of 7 – 110 Bq kg$^{-1}$. The depth of measurable activity varies greatly, from only 1 – 2 cm in some locations to 20 cm in one core corresponding to the known location of the estuarine turbidity maximum where rapid deposition occurred. Seasonal variations in the amount and location of $^{7}$Be containing sediments indicated that sediments were originally deposited at the seaward end of the estuary during the spring freshet in March and April, 1999. By June, 1999 those deposits had eroded away but deposits not previously measured occurred up estuary. $^{7}$Be profiles in these new deposits confirmed the sediments had been in contact with the water column in the last five months. A core from the same location two months prior showed almost no measurable $^{7}$Be. Based upon estimated sediment supply masses Woodruff et al. (2001) concluded that sediments deposited in the estuary prior to the 1999 spring freshet must have also been included in the new deposits measured in July 1999 and that significant seasonal reworking of the sediments occurs.

2.5 Investigations of the San Francisco Bay Estuary Sediments

Inhabitation of the areas surrounding the San Francisco Bay is known to have occurred at least 5,000 years ago, coincident with declines in the rate of sea level rise which originally brought waters through the Golden Gate 10,000 years ago (Atwater, 1979). Discovery of the San Francisco Bay by European explorers did not occur until 1769 and scientific exploration of the Bay and Delta did not occur until 1824 when Otto von Kotzebue and Dr. Ivan Eschscholtz made trips by boat as far south as Santa Clara and as far north as Rio Vista in the Delta making observations of natural history and agricultural potential (Hedgepeth, 1979). Extensive scientific investigation of the Bay did not begin until well into the 20th century. Hedgepeth (1979) recounts much of the early scientific research on the Bay. Here only relevant findings will be presented.
<table>
<thead>
<tr>
<th></th>
<th>Suspended Sediments</th>
<th>Bottom Sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2 – 20 μm</td>
<td>2 – 20 μm</td>
</tr>
<tr>
<td></td>
<td>North</td>
<td>South</td>
</tr>
<tr>
<td>Montmorillonite</td>
<td>21-27</td>
<td>19-25</td>
</tr>
<tr>
<td>Illite</td>
<td>37-40</td>
<td>43-49</td>
</tr>
<tr>
<td>Kaolinite + Chlorite</td>
<td>36-40</td>
<td>32-34</td>
</tr>
</tbody>
</table>

Table 2.2: Clay minerals identified in San Francisco Bay sediments as percent of total by mineral for each size class and location. From Knebel et al. (1977).

2.5.1 Sediment Minerals

Mineral analysis of suspended and bed sediments by Knebel et al. (1977) indicate varying amounts of montmorillonite, illite and chlorite + kaolinite (Table 2.2). Six locations were sampled along the estuary in the North Bay and another six locations in the South Bay. In general montmorillonite made up no more than one third of the sediments. In the water column of the North Bay, illite and kaolinite + chlorite occurred in roughly the same amounts, 36 to 40%, and Knebel et al. (1977) found that the dominant fraction changed with location. In the South Bay water column, illite was the dominant mineral. In the bottom sediments, kaolinite + chlorite was the dominant mineral present at most locations in the North and South Bays for sediments in the 2 – 20 μm size fraction. However in the 0.2 – 2 μm size fraction, illite was more prevalent than kaolinite + chlorite and in the North Bay montmorillonite was also a greater fraction of the sediments that illite. Krone (1963) also identified feldspar and quartz in the < 10 μm fraction of sediments collected from the Mare Island Strait, located along the eastern edge of San Pablo Bay (Figure 1.1).

2.5.2 Sediment Flocs

*In situ* observation of sediment flocs have been made at three sites within the San Francisco Bay estuary. Kineke and Sternberg (1989) used a floc camera to measure floc size and settling velocity in San Pablo Bay near San Pablo Strait. Concentrations during the 12 hours of observation were 83 – 328 mg L⁻¹ and the shear velocity, \( u_* \), was 0.4 – 5.3 cm s⁻¹. The mean floc diameter observed was around 100 μm but flocs as large as 450 μm were observed. Estimated settling velocities were as high as 20 mm s⁻¹. Krank and Milligan (1992), using data from the same site, made several other observations about the flocs. Estimated floc density was 1,077 kg m⁻³ with a range of 1,032 – 1,270 kg m⁻³. Disaggregated particle diameters were generally 1 – 10 μm though some were in the 10 – 100 μm range. The modal floc diameter observed was between 100 – 512 μm. The lower limit of the floc camera was 40 μm. The largest particles observed corresponded with the high concentrations of the flood tide but at high levels of turbulence particles larger than 370 μm were not observed. Krank and Milligan (1992) posited that two sizes of particles existed at their site: a fine fraction, \( d_p < 100 \) μm, that are relatively stable in suspension such that they do not settle out over the tidal cycle. The second fraction is comprised of larger flocs that undergo settling and resuspension over the tidal cycle causing the observed changes in SSC.

Manning et al. (2010) made observations of floc size and settling velocity in the Central Bay and within the turbidity maximum zone (TMZ). The TMZ in the North Bay at the time was in the Carquinez Strait, which is located between San Pablo and Suisun Bays (Figure 1.1). The lower limit of their video system was 6.3 μm. Within the TMZ the mean floc size was 144 μm.
but many low density macroflocs with diameters up to 400 μm were observed. Over half of the mass of suspended sediments was contained within the macroflocs but they made up only one third of the total floc population. Observed settling velocities were 0.4 – 10 mm s⁻¹. Similar settling velocities were observed in the Central Bay but the maximum floc size was only 270 μm. Microflocs comprised 86% of the flocs observed and contained 70% of the floc mass.

2.5.3 Sediment Bathymetry

Since hydraulic mining began regions of the Bay have undergone periods of sediment deposition and erosion as the hydraulic mining material works its way through the estuary and out to sea. Extensive reviews of historic bathymetric surveys have estimated changes in bathymetry and sediment volume changes in the San Pablo Bay (Jaffe et al., 2007) and the Central Bay (Fregoso et al., 2008) between 1856 – 1983 and 1855 – 1979, respectively. In San Pablo Bay, there has been an annual net deposition of sediments from 1856 – 1951, though the rate of deposition decreased substantially after 1887. From 1951 – 1983, an annual net erosion of sediments was observed (Jaffe et al., 2007). In the Central Bay, defined as the region south of San Pablo Bay, north of Hunter’s Point in San Francisco and east of Point Bonita and Lands End in the Golden Gate (Figure 1.1), net deposition occurred from 1895 – 1947 while net erosion was observed in the periods 1855 – 1895 and 1947 – 1980 (Fregoso et al., 2008). Bathymetric surveys for the South Bay, defined as the area south of Hunter’s Point in San Francisco, are more limited, with surveys available only for 1956, 1983 and 2005 (Jaffe and Foxgrover, 2006). The South Bay was found to be more depositional from 1983 – 2005 than from 1956 – 1983. Jaffe and Foxgrover (2006) also determined net erosion occurred in the shallows on the eastern shore north of the Dumbarton Bridge, net deposition occurred in the region south of the Dumbarton Bridge and net deposition occurred in the channels.

2.5.4 Numerical Models

Predicting the response of the Bay to changing environmental conditions relies upon numerical models. Such models must sufficiently resolve the hydrodynamics and must also accurately model sediments in the water column and the sediment bed if predictions of sediment movement and contaminant fates are to be accurately predicted. The SUNTANS model is currently being used to model the Bay with a resolution on the scale of meters. The incorporation of a sediment model is planned. SUNTANS has previously been applied to the Snohomish River estuary (Wang et al., 2009). The inclusion of a sediment model requires, however, parameterization of sediment behaviors which ideally would be based upon measurements of Bay sediments.

Existing models of the Bay aim at either predicting sediment concentrations in the water column or predicting the long term fate of contaminants within the water column and sediment bed. These models are appealing due to simplicity and small computational costs but are in many ways very limited.

The depth averaged 2-dimensional models of the South Bay (Lee et al., 2003) and the North Bay (McDonald and Cheng, 1997) provide the best treatment of the sediment bed. It is modeled as a series of layers with increasing density and shear strength with depth into the bed. McDonald and Cheng (1997) noted that significant limitations to the model include being limited to one particle size, no incorporation of resuspension due to wind driven waves and no capacity for incorporating stratification. Changes made to the South Bay model (Lee et al., 2003) include the addition of an easily erodible layer of sediments in the shallow regions and the effects of a
uniform wind field on the bed shear stress. Most notable of both models is that they were calibrated based on suspended sediment data collected from the channels of the Bay. The broad shallow shoals are a large, highly dynamic reservoir of sediments for the Bay that responds differently to environmental forcing than do the channels. There is little indication that these models properly model sediment dynamics on the shoals. Finally, these models focus upon suspended sediment concentrations (SSC) not the sediment bed. The models relied upon parameterized expressions for deposition and erosion which did not address time varying sediment bed properties.

Drawing upon the same data which was used for this work, a one-dimensional water column sediment model was developed by Brand et al. (2011). Two particle size classes were included in the model, a fast settling coarse fraction ($w_{s,\text{fast}} = 8.2 \times 10^{-4}$ and $5.1 \times 10^{-4}$ m s$^{-1}$, for the spring and fall deployments, respectively) and a slow settling fine fraction ($w_{s,\text{slow}} = 1 \times 10^{-7}$ and $1.1 \times 10^{-5}$ m s$^{-1}$, for the spring and fall deployments, respectively). Particles settling out at the sediment bed boundary were lost from the model while the erosive flux of particles into the water column was forced by an amplified value of the turbulent vertical sediment flux measured 0.36 m above the bed. The amplification factors were $\text{amp} = 2.6$ and $\text{amp} = 1.7$ for the spring and fall deployments, respectively. The parameterized flux of sediments into the water column was split into the fine and coarse fractions by a calibration parameter. The model showed that the fine fraction was a significant contributor to the total sediment contained within the water column despite being only a small percentage of the sediments introduced at the sediment bed interface.

Contaminant fate modeling in the Bay has included PCBs (Davis, 2004), PAHs (Greenfield and Davis, 2005) and mercury (Macleod et al., 2005). These models are used to evaluate the long term, region-wide removal of contaminants from the ecosystem. The water column and the sediment bed are each treated as well-mixed boxes into and out of which contaminants may move via several pathways. Inputs include atmospheric deposition and fluxes from the surrounding watersheds while losses may include degradation and loss to the ocean via advection. Sediments also move between the two boxes as erosive and depositional fluxes. By design the sediment bed is modeled simplistically, assuming an active sediment layer 0.15 m thick overlying a buried sediment layer which does not interact with the water column but may serve as a sink for sediments and contaminants in the active sediment layer. Contaminant concentrations in the active layer are estimated from sediment samples collected throughout the Bay. Fluxes of sediments between the sediment bed and water column are handled by rate constants.

These models are perceived as useful for assessing regional responses to management decisions due to their simplicity but they have significant limitations. The distribution of contaminants is highly heterogeneous both with depth in the sediment bed and across the spatial extend of the estuary (Hunt et al., 1998; SFEI, 2007; SFEI, 2009). Davis (2004) notes that the peak PCB contamination is associated with sediments dating to the 1950s and 1960s and that if net erosion occurs in the Bay, the buried sediments may reenter the ecosystem, delaying the response of the Bay to loading reductions. Greenfield and Davis (2005) found that the timescales for contaminant removal were highly dependent upon the height of the active layer, a quantity which all of the authors noted to be poorly defined.
3 Experimental Methods

As one of the largest estuaries in North America, the San Francisco Bay estuary is important both ecologically and economically. Understanding the role that sediments play is one facet of ensuring that the Bay is able to continue to adapt successfully to the pressures which threaten it, including climate change and legacy contaminants. As each estuary is unique, the best understanding of Bay sediment properties is derived from in situ sediment measurements. The measurements can be used to quantify the sediment processes outlined in the previous chapter and ultimately used to improve our understanding of sediment processes in the Bay and to improve the models of the Bay which are being developed.

For these reasons two instrument deployments were carried out in the South San Francisco Bay from February 24th to March 16th, 2009 and September 9th to October 7th, 2009. The location of the study area is shown in Figure 1.1. This chapter contains the field deployment instrumentation and calibration and the laboratory experimental setup.

3.1 Field Site
The field site was located in the South Bay, roughly two kilometers south of the San Mateo Bridge. The site was chosen because previous informal observations indicated differences in the water masses of the shoals and the channel, particularly with regards to SSC. These SSC differences as well as the bathymetry of the region, including access to a broad area of the shoals as well as the shoal – channel transition, indicated that the area chosen would be ideal for supporting the broader instrument deployment studying shoal – channel sediment exchanges and the vertical sediment bed – water column exchanges to which this work belongs.

The same ten stations were instrumented during both deployments and were laid out along the eastern side of the Bay, spanning the channel – shoal transition. The station locations are shown in Figure 3.1. Stations are identified by bathymetric location (Channel, Slope or Shoal) and relative position (North, Middle or South). Exceptions to the naming scheme are the Channel Station and the Benthic Station. The Channel Station refers then to the Channel Middle Station. The Benthic Station was located one kilometer landward (northeast) from the Shoal stations. This station was outfitted with additional instrumentation for measuring sediment bed properties and water column suspended sediments. Coordinates for all of the stations are listed in Table 3.1.

3.2 Water Column Measurements
Measurements in the water column were taken using a variety of in situ instrumentation with additional water samples collected for further laboratory analysis. Type, location and measurement settings of the instruments are listed in Table 3.2. Descriptions of the instruments and calibration procedures are included below followed by the water sampling methods.

3.2.1 Acoustic Doppler Velocimeter
Acoustic Doppler velocimeters (ADV) provide a three dimensional velocity vector at a single point in the water column. The measurement is based upon the Doppler shift of an acoustic signal backscattered from particles or bubbles within the measurement volume, which is
Figure 3.1: Detail of station layout in the South Bay
roughly one cubic centimeter. The instruments were programmed to make high frequency (4 – 10 Hz) measurements in bursts (8 – 10 minutes) at a selected frequency (every 12 minutes or hourly). Exact settings for all the ADVs are listed in Table 3.2. The acoustic frequencies of the ADV were 10 MHz (Sontek) or 6 MHz (Nortek).

The strength of the acoustic signal returned to the ADV is proportional to the number of particles in the measurement volume. The particle size, a, at which an ADV is the most sensitive depends upon the acoustic wavelength, \( \lambda \), at which it operates according to \( a = \lambda / 2\pi \). A 10 MHz instrument is most sensitive to a 45 \( \mu \)m particle and a 6 MHz instrument that value is 70 \( \mu \)m.

### 3.2.1.1 Calibration of ADV for SSC Measurements

Acoustic backscatter (ABS) measurements collected by the ADVs were used to measure suspended sediment concentrations (SSC) collocated with velocity measurements (Kawanisi and Yokosi, 1997; Voulgaris and Meyers, 2004). Field and laboratory calibrations were carried out to obtain the SSC – ABS calibration relationships for the ADVs used during the field deployments. Only calibration and SSC data for the 10 MHz (Sontek) ADVs are presented and analyzed here. Laboratory calibration of the ADVs was carried out using bed sediments collected at the field site from the foot pad of the Channel Station instrument tripod. Within the measured range of 0 – 0.35 kg m\(^{-3}\) the ABS – \( \log(\text{SSC}) \) relationships were linear (Figure 3.2a). The three ADVs showed nearly identical slopes during the laboratory calibrations but exhibited different intercepts. The ADVs were also calibrated against the filtered water samples collected during the field deployment (see Section 3.2.5). Given the results of the laboratory calibrations, field calibrations for the ADVs were developed by fitting a single slope to all three of the ADVs but the intercept was set independently for each ADV (Figure 3.2b). The slope of the field calibration relationship was smaller than that found during the laboratory calibration. This might be explained by the use of deflocculated sediments in the laboratory compared to flocculated sediments in the field (Brand et al., 2010). The response from single frequency acoustic instruments is known to be particle size dependent. The field derived calibration relationships were used to calibrate the ADV signals measured during the instrument deployments as flocs rather than disaggregated sediments were expected in the field.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (° N)</th>
<th>Longitude (° W)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Benthic</td>
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<td>122.2098</td>
</tr>
<tr>
<td>Shoal North</td>
<td>37.5803</td>
<td>122.2204</td>
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<tr>
<td>Shoal Middle</td>
<td>37.5792</td>
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<tr>
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<tr>
<td>Slope North</td>
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</tr>
<tr>
<td>Slope Middle</td>
<td>37.5753</td>
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</tr>
<tr>
<td>Slope South</td>
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<td>Channel North</td>
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<td>Channel</td>
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</tr>
<tr>
<td>Channel South</td>
<td>37.5798</td>
<td>122.2322</td>
</tr>
</tbody>
</table>

Table 3.1: Coordinates for experiment stations. Locations are accurate to within 10 m and utilized the NAD 83 standard.
Table 3.2: Instrument details for the field deployments. Not all of the instruments listed were used in this work.

<table>
<thead>
<tr>
<th>Station</th>
<th>Instrument†</th>
<th>Location</th>
<th>Manufacturer</th>
<th>Measurement Frequency</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Benthic</td>
<td>ADV</td>
<td>0.72 m above bed</td>
<td>Sontek Hydra</td>
<td>8 min burst at 10 Hz every 12 minutes</td>
<td></td>
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<tr>
<td></td>
<td>ADV</td>
<td>0.36 m above bed</td>
<td>Sontek Hydra</td>
<td>8 min burst at 10 Hz every 12 minutes</td>
<td></td>
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<tr>
<td></td>
<td>LISST 100 Type B</td>
<td>0.55 m above bed</td>
<td>Sequoia Scientific</td>
<td>2 min burst at 1 Hz every 12 minutes</td>
<td>Fouled after ~5 days during both deployments</td>
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<td></td>
<td>Imaging sonar</td>
<td>0.30 cm above bed</td>
<td>Imagenex</td>
<td>3x 4 min scans every hour &amp; 2x 10 min scans every 6 hr</td>
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</tr>
<tr>
<td></td>
<td>Profiling sonar</td>
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<td>Imagenex</td>
<td>3x 4 min scans every hour</td>
<td></td>
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<td></td>
<td>OBS</td>
<td>0.36 m above bed</td>
<td>D&amp;A Instruments</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.72 m above bed</td>
<td>D&amp;A Instruments</td>
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<tr>
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<td>SeaBird</td>
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</tr>
<tr>
<td></td>
<td>CTD</td>
<td>0.40 m above bed</td>
<td>RBR</td>
<td>10 second average every 6 minutes</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD</td>
<td>2 m below surface</td>
<td>BRB</td>
<td>15 second average every 3 minutes</td>
<td>No data for second deployment</td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.40 m above bed</td>
<td>D&amp;A Instruments</td>
<td>6 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>2 m below surface</td>
<td>D&amp;A Instruments</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td>Channel North</td>
<td>Pressure</td>
<td>Near bed</td>
<td>RBR</td>
<td>10 second</td>
<td></td>
</tr>
<tr>
<td>Channel South</td>
<td>Pressure</td>
<td>Near bed</td>
<td>RBR</td>
<td>10 second</td>
<td></td>
</tr>
</tbody>
</table>
Table 3.2 (continued): Instrument details for the field deployments. Not all of the instruments listed were used in this work.

<table>
<thead>
<tr>
<th>Station</th>
<th>Instrument†</th>
<th>Location</th>
<th>Manufacturer</th>
<th>Measurement Frequency</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Shoal North</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>PCADP</td>
<td>1.61 m above bed</td>
<td>RD Instruments</td>
<td>10 min burst at 1 Hz every 12 min</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ADV</td>
<td>0.52 m above bed</td>
<td>Sontek Hydra</td>
<td>8 min burst at 4 Hz every 12 min</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CT</td>
<td>0.65 m above bed (T) 0.72 m above bed (C)</td>
<td>SeaBird</td>
<td>12 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.50 m above bed</td>
<td>D&amp;A Instruments</td>
<td>12 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pressure</td>
<td>1.38 m above bed</td>
<td>ParoScientific</td>
<td>12 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pressure</td>
<td>1.24 m above bed</td>
<td>ParoScientific</td>
<td>12 minute</td>
<td></td>
</tr>
<tr>
<td><strong>Shoal Middle</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>PCADP</td>
<td>1.66 m above bed</td>
<td>RD Instruments</td>
<td>10 min burst at 1 Hz every 12 min</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ADV</td>
<td>0.36 m above bed</td>
<td>Sontek Hydra</td>
<td>10 min burst at 8 Hz every 12 min</td>
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</tr>
<tr>
<td></td>
<td>ADV</td>
<td>0.70 m above bed</td>
<td>Nortek Vector</td>
<td>10 min burst at 8 Hz every 12 min</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD</td>
<td>1.66 m above bed</td>
<td>Microcat</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD</td>
<td>0.47 m above bed</td>
<td>RBR</td>
<td>5 second average every 3 minutes</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.26 m above bed</td>
<td>D&amp;A Instruments</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.72 m above bed</td>
<td>D&amp;A Instruments</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.47 m above bed</td>
<td>D&amp;A Instruments</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pressure</td>
<td>1.40 m above bed</td>
<td>Paroscientific</td>
<td>3 minute</td>
<td></td>
</tr>
<tr>
<td><strong>Shoal South</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>ADV</td>
<td>0.25 m above bed</td>
<td>Sontek Hydra</td>
<td>10 min burst at 8 Hz every hour</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ADV</td>
<td>0.50 m above bed</td>
<td>Nortek Vector</td>
<td>10 min burst at 8 Hz every hour</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD</td>
<td>0.50 m above bed</td>
<td>Microcat</td>
<td>12 minute</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS</td>
<td>0.50 m above bed</td>
<td>D&amp;A Instruments</td>
<td>12 minute</td>
<td></td>
</tr>
</tbody>
</table>
Table 3.2 (continued): Instrument details for the field deployments. Not all of the instruments listed were used in this work.

<table>
<thead>
<tr>
<th>Station</th>
<th>Instrument</th>
<th>Location</th>
<th>Manufacturer</th>
<th>Measurement Frequency</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope North</td>
<td>ADCP 0.20 m above bed</td>
<td>RD Instruments</td>
<td>1 Hz</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 0.5 m above bed</td>
<td>BRB</td>
<td>6 minute</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Turbidity 0.18 m above bed</td>
<td>Seapoint</td>
<td>6 minute</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 1 m below surface</td>
<td>RBR</td>
<td>15 second average every 3 minutes</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS 1 m below surface</td>
<td>D&amp;A Instruments</td>
<td>6 minute</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope Middle</td>
<td>ADCP 0.72 m above bed</td>
<td>RD Instruments</td>
<td>1 Hz</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 3 m below surface</td>
<td>RBR</td>
<td>2 second average every 3 minutes</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS 3 m below surface</td>
<td>D&amp;A Instruments</td>
<td>3 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 0.6 m below surface</td>
<td>RBR</td>
<td>2 second average every 6 minutes</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS 0.6 m below surface</td>
<td>D&amp;A Instruments</td>
<td>6 minute</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CT 0.72 m above bed</td>
<td>Seacat</td>
<td>3 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Turbidity 0.43 m above bed</td>
<td>Seapoint</td>
<td>3 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 0.54 m above bed</td>
<td>RBR</td>
<td>6 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope South</td>
<td>ADCP 0.45 m above bed</td>
<td>RD Instruments</td>
<td>1 Hz</td>
<td></td>
<td>No data for first deployment</td>
</tr>
<tr>
<td></td>
<td>CTD 0.5 m above bed</td>
<td>RBR</td>
<td>6 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Turbidity 0.3 m above bed</td>
<td>Seapoint</td>
<td>6 min</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CTD 1 m below surface</td>
<td>RBR</td>
<td>10 second average every 6 minutes</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>OBS 1 m below surface</td>
<td>D&amp;A Instruments</td>
<td>6 minute</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.2: ADV calibration relationships developed (a) in the laboratory and (b) in the field. The laboratory experiments indicated the calibrations slopes were the same but the intercepts were different. For the field calibrations the intercepts were allowed to vary but a single value for the slope was fit to the data.
3.2.2 Acoustic Doppler Current Profiler

Acoustic Doppler current profilers (ADCP) measure three dimensional velocities in the water column at multiple distances from the instrument based upon the Doppler shift of an acoustic signal backscattered to the instrument by particles or bubbles in the water column. Slope stations were instrumented with upward looking 1200 kHz ADCP measuring at 1 Hz providing velocity measurements from near the bed to the water surface at 0.25 m intervals. The Channel Station was instrumented with an upward looking 600 kHz ADCP making velocity measurements at 0.5 m intervals at 1 Hz.

3.2.2.1 Calibration of ADCP for SSC Measurements

The strength of the returned acoustic signal is recorded by the ADCP and is a function of the amount of scatterers in the water column. Typical scatterers are suspended sediments and air bubbles. As first presented by Deines (1999), the strength of the returned signal can be calibrated to provide an estimate of suspended sediment concentrations (SSC), assuming that suspended sediments are the only source of acoustic scattering.

The strength of the return signal is the echo intensity (EI). The backscatter coefficient, $S_v$ in dB, is a measure of how the source signal is backscattered from scatterers in the water and can be related to the SSC by:

$$SSC = 10^{AS_v+B} \tag{3.1}$$

where A and B are coefficients that need to be determined by comparing known values of SSC against values of $S_v$. The equation for $S_v$ developed for this work draws upon the works of several authors (Deines, 1999; Flammer, 1962; Gartner, 2004; Schulkin and Marsh, 1962; Urick, 1948; Wall et al., 2006) to give:

$$S_v = 10\log_{10} \left( \frac{\sum_{i=1}^{n} \frac{(EI(i) - Er(i))K_c}{10}}{n} \right) + \psi 20 \log_{10}(R) + 2\alpha R + 2\alpha_s R \tag{3.2}$$

The development of Equation 3.2 and a description of all of the parameters appearing therein are detailed in Appendix A. The first term on the RHS is the relative backscatter and each of the following terms is a correction. Additional corrections to $S_v$ can include transmit-power, transmit-length and beam normalization (Deines, 1999; Wall et al., 2006). These corrections were omitted because the required data were not available. Wall et al. (2006) estimated the errors associated with these omissions to be 5.1%, 0.3% and 6.5%, respectively.

The first term on the RHS of Equation 3.2 is an average of the signal strength measured by the ADCP. The second term is a correction for the spreading of the signal and the third term accounts for the attenuation of the signal as it travels through the water column. The final term is a correction for the attenuation of the signal due to scattering and absorption by particles in the water column.

3.2.2.2 The Calibration Process

Estimating SSC from the ADCP data requires values for a number of variables which must either be measured ($S$, $T$, $a_p$, $a_r$, $z$, $EI$, $Er$, $P$, $SSC$) or otherwise assumed ($K_c$, $\rho_a$). In this
calibration some parameters were assumed to be known exactly (\(f, R, \rho_f, \nu\) and the constants \(A, B, C\)). The values for the parameters used for these calibrations are listed in Table 3.3.

The first part of the calibration process requires the calculation of \(S_v\) by Equation 3.2. Data collected at 1 Hz was averaged over 50 s time periods and across each of the four transducers. The averaging reduced the amount of data and therefore computational time but also provided a single EI value and standard deviation for each ADCP bin every 50 s. Fluctuations of EI within the averaging period may be associated with changing SSC, instrument noise, air bubbles entrained near the surface by waves and possibly even aquatic organisms.

The background noise level, \(E_r\), for each transducer was taken as the minimum value measured during each experiment (Gartner, 2004). Salinity and temperature time series data were taken from the instrument nearest the ADCP. In most cases this was a conductivity, temperature, depth instrument (CTD) located less than 1 meter from the ADCP. Constant values for depth and pressure were used rather than time series for computational simplicity. The two variables appear in the calculation for the velocity of sound in water and the attenuation of sound due to water, respectively. At the depths of the instruments used here, both the values and any variations of these parameters due to tidal stage have negligible impacts. Measurements from a Laser In Situ Scattering and Transmissometry (LISST) instrument provided five days of particle size distributions during each deployment. The mean particle size and standard deviation were calculated directly from the PSD data. Aggregate density, \(\rho_a\), was assumed to be 1,200 kg m\(^{-3}\).

Equation 3.1 is normally presented in the literature as the form used for determining SSC from \(S_v\). For reasons which will be discussed during the error analysis (Section 5.2.3.1) \(\log_{10}(SSC)\) was used as the dependent variable in the linear regression with \(a\) and \(b\) as the slope and intercept, respectively. Rearranging the regression equation into a form similar to Equation 3.1 gives:

\[
SSC = 10^{\left(S_v - b\right) / a}
\]

Independent measurements of SSC, from filtered grab samples or from other instruments, should be made at or as close as possible to the measurement field of the ADCP in order to determine the values of \(a\) and \(b\) in Equation 3.3. A linear regression of known values of \(\log_{10}(SSC)\) against values of \(S_v\) measured at the same time and location is used to determine these parameters and to convert the rest of the \(S_v\) values into SSC measurements.

In most cases an optical backscatter (OBS) instrument was located within the ADCP measurement field, typically 1 to 2 meter below the surface. Bio-fouling of the OBS often limited the amount of data available for the calibration. The OBS sensors were calibrated in the laboratory using sediments collected from the sediment bed at the field site and checked against the water column grab samples (Section 3.2.4). For the Channel Station during the September deployment only filtered water column grab samples were available for the calibration.

### 3.2.3 Laser In Situ Scattering and Transmissometer

Laser In Situ Scattering and Transmissometer (LISST) instruments measure particle size distributions (PSD) based upon a laser diffraction pattern generated by particles within the measurement volume. The LISST 100 Type B measures particles with diameters within the 1.25 \(\mu m\) to 250 \(\mu m\) range as particle volume per water volume (\(\mu L L^{-1}\)) in 32 logarithmically spaced bins. The lower limit in \(\mu m\) for each bin is given by 1.25 \(\cdot 1.18^n\) for \(n = 1\) to 32 and the upper limit is 1.18 times the lower limit. Agrawal and Pottsmith (Agrawal and Pottsmith, 2000)
Table 3.3: Parameter values and standard deviations used for ADCP calibration and error analysis

<table>
<thead>
<tr>
<th>Parameter (units)</th>
<th>Value</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Standard Deviation, $\sigma^i$</td>
</tr>
<tr>
<td><strong>ADCP frequency, f, (Hz)</strong></td>
<td>$1228 \times 10^3 / 614 \times 10^3$</td>
<td>Instrument parameter</td>
</tr>
<tr>
<td><strong>Aggregate density, $\rho_{agg}$ (kg m$^{-3}$)</strong></td>
<td>1200</td>
<td>Assumed</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>Assumed $2\sigma = 200$</td>
</tr>
<tr>
<td>Counts to dB conversion factor, $K_c$ (dB/count)</td>
<td>0.45</td>
<td>Typical value</td>
</tr>
<tr>
<td></td>
<td>0.05</td>
<td>$2\sigma = 0.1$. Range of $K_c$ is 0.35 - 0.55</td>
</tr>
<tr>
<td><strong>Depth, z (m)</strong></td>
<td>$9 / 18^\dagger$</td>
<td>Average value from field measurement</td>
</tr>
<tr>
<td></td>
<td>0.01</td>
<td>Instrument accuracy $2\sigma = 0.02$</td>
</tr>
<tr>
<td><strong>Echo Intensity, EI (counts)</strong></td>
<td>Time series</td>
<td>Field measurement</td>
</tr>
<tr>
<td><strong>Echo Intensity Reference Level, Er (counts)</strong></td>
<td>Time series</td>
<td>Standard deviation of field measurements over averaging period</td>
</tr>
<tr>
<td><strong>Fluid density, $\rho_{fluid}$ (kg m$^{-3}$)</strong></td>
<td>1000</td>
<td>Assumed constant</td>
</tr>
<tr>
<td><strong>Fluid viscosity, $n$ (m$^2$ s$^{-1}$)</strong></td>
<td>1.00E-06</td>
<td>Assumed constant</td>
</tr>
<tr>
<td><strong>Particle diameter, $d_p$ (m)</strong></td>
<td>$1.2 \times 10^{-4}$</td>
<td>Estimated from field measurement</td>
</tr>
<tr>
<td></td>
<td>$7.7 \times 10^{-5}$</td>
<td>Estimated from field measurement</td>
</tr>
<tr>
<td>Parameter (units)</td>
<td>Value</td>
<td>Data Source</td>
</tr>
<tr>
<td>--------------------------------------</td>
<td>------------------------------------</td>
<td>---------------------------------------</td>
</tr>
<tr>
<td>Path length, R (m)</td>
<td>Equation 5.12</td>
<td>Bin heights are user defined</td>
</tr>
<tr>
<td></td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Pressure (gauge), P (atm)</td>
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<td>Average values from field measurement</td>
</tr>
<tr>
<td></td>
<td>0.01</td>
<td>Instrument accuracy 2σ = 0.02</td>
</tr>
<tr>
<td>Salinity, S (psu)</td>
<td>Time series</td>
<td>Field measurement</td>
</tr>
<tr>
<td></td>
<td>2.5×10⁻³</td>
<td>Instrument accuracy 2σ = 0.005</td>
</tr>
<tr>
<td>Suspended sediment concentration, SSC (mg L⁻¹)</td>
<td>Mean of time series</td>
<td>Field measurement</td>
</tr>
<tr>
<td></td>
<td>From data</td>
<td>Variance of entire SSC time series</td>
</tr>
<tr>
<td>Temperature, T (°C)</td>
<td>Time series</td>
<td>Field measurement</td>
</tr>
<tr>
<td></td>
<td>2.5×10⁻³</td>
<td>Instrument accuracy 2σ = 0.005</td>
</tr>
<tr>
<td>Transducer radius, aᵣ (m)</td>
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<td>Measured</td>
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<tr>
<td></td>
<td>0.001</td>
<td>Measurement accuracy 2σ = 0.002</td>
</tr>
<tr>
<td>Velocity of sound in water, v (m s⁻¹)</td>
<td>Equation 5.13</td>
<td>Temperature, salinity and depth dependent</td>
</tr>
<tr>
<td></td>
<td>0.05</td>
<td>2σ = 0.1. Reported equation accuracy = 0.1</td>
</tr>
</tbody>
</table>

¹Uncertainties which could be directly determined or estimated, whether from instrument uncertainty or measurement uncertainty, were assumed twice the standard deviation (2σ). The assumed parameter value plus or minus 2 standard deviations creates a range within which the true parameter value should fall most of the time. Parameters for which no standard deviation is listed are assumed to be known exactly.

‡The first value corresponds to the 1200 kHz ADCPs deployed at the slope stations and the second is for the 600 kHz ADCP used at the channel station.
reported measured total particle concentrations were within about 20% of the known value and that the PSD was reproduced satisfactorily for a known distribution of non-flocculated glass spheres.

It is also known that particles larger and smaller than the measurement range of the LISST may produce diffraction patterns which may be measured by the instrument (Agrawal and Pottsmith, 2000; Mantovanelli and Ridd, 2006; Mikkelsen, 2002). Excess diffraction patterns occur primarily at the ends of the measurement range, resulting in an over estimate of particle volumes and a rising tail in the measured PSD (Mikkelsen, 2002).

Mantovanelli and Ridd (2006) and Perdocchi and Garcia (2006) noted several assumptions made in development of the LISST which are not necessarily justified for estuarine flocs. Non-spherical particles, such as estuarine flocs, will produce a different diffraction patterns than spherical solid particles used in the LISST development. Mantovanelli (2006) also notes that the refraction index of the flocs is assumed to be uniform and that it can be approximated by that of spheres. Perdocchi and Garcia (2006) conclude that flocs should not produce large deviations from the spherical theory used in developing the LISST but that the exact size distribution and resolution of the PSD will be degraded. They indicate that caution should be used in drawing conclusions from the measurements, specifically concerning the details of the PSD.

### 3.2.4 Optical Backscatter Sensor

The Optical Backscatter (OBS) sensors used for this study measured suspended particles by emitting a light pulse and measuring the light which is backscattered. The strength of the returned signal is proportional to the area of particles illuminated and hence SSC. However, the signal is also affected by the size, shape and index of refraction of the particles in the water.

Calibrations for the optical backscatter (OBS) sensors were developed in the laboratory using surface layer sediments collected from the field site. Sensor output, in millivolts or nephelometric turbidity units (NTU, determined from a formazin standard), was monitored as the SSC was increased from 0 to as much as 0.7 kg m⁻³. Within this range the sensor response was linear for all of the OBS (Figure 3.3). Sensor calibrations were checked against the filtered water samples. Insufficient samples were available to generate robust field calibrations due to biofouling of the OBS sensors.

### 3.2.5 Water Column Samples for Suspended Sediment Concentration Analysis

Water samples were collected during each instrument deployment with each station being sampled 2 or 3 times on each collection day. At the Benthic and Shoal Stations, samples were collected by pumping approximately 1 L of water into bottles using a submersible sampling station. The station consisted of a weighted platform with an attached vertical rod to which the sampling inlets at 0.25, 0.50 and 0.65 m above the bed were connected. Water samples were pumped to the surface using a peristaltic pump. At each station, the sampled port(s) corresponded to the heights above the bed of the instrumentation deployed at that station. At the Slope Stations and at the Channel Station, samples were collected with a Niskin bottle 1 m below the surface and 1 m above the bed. At the Slope Middle Station, water samples were only collected at 0.6 m and 3 m below the water surface. The water samples were filtered through pre-weighed 0.4 μm filters (Whatman Nucleopore). The filters and filter devices were flushed with deionized water three times to remove any salts and then dried at 60 °C for 24 h. SSC was determined from the mass retained on the filters which was assumed to be due only to sediments.
Figure 3.3: OBS calibration relationships developed using sediments from the field site. OBS measurements were in (a) NTU determined from a formazin standard and (b) volts.
3.3 **Sediment Bed Measurements**

Both sediment grab samples and sediment cores were collected during the field experiments.

### 3.3.1 Sediment Grab Samples

Grab samples of the top layers of bed sediments were collected from the Channel, Slope Middle, Shoal Middle and Benthic Stations concurrently with the water samples using a Ponar grab sampler. The soft fluffy surface layer was separated from the stiffer underlying sediments. Both fractions were retained for grain size analysis. The samples were ultrasonicated and treated with hydrogen peroxide and sodium hexametaphosphate prior to analysis. The grain size distribution was measured using a Beckman Coulter laser diffraction particle analyzer with a measurement range of 0.4 \( \mu m \) to 900 \( \mu m \) using 72 logarithmically spaced intervals.

### 3.3.2 Sediment Cores

Sediment cores were collected the first working day after instrument deployment and the last working day before recovery to establish bed conditions at the start and end of each instrument deployment. At the end of the fall deployment inclement weather delayed instrument recovery for more than a week post sediment core collection. At least two cores were collected from both the Benthic Station and the Shoal Middle Station on February 25th, March 13th, September 10th and September 28th.

#### 3.3.2.1 Collection and Storage

Sediment cores were collected from an anchored boat at or near to low slack water when possible. Water depths at the time of collection varied between 2 – 4 meters. Cores were collected as close as possible to the station surface buoys without endangering the instruments, generally at a distance of about 10 meters.

The cores were collected with a hand driven push corer. Details of the corer design are shown in Figure 3.4. Cores were collected in 95 mm inner diameter clear acrylic tubes (TAP Plastics) approximately 30 cm in length. The coring tube was lowered to the sediment bed using 2 to 4 sections of approximately 1.5 m long ABS plastic pipes attached to the one way valve of the coring device. Holes drilled along the length of the ABS pipes allowed water to flow freely into and out of the tubes.

The coring tubes were easily driven into the sediment bed to produce cores of 20 – 25 cm in length. Cores were discarded if insufficient depth was recovered or extensive disturbance of the sediment was evident.

The entire coring device was kept vertical as it was brought aboard the boat. Once aboard the bottom was capped with a polyethylene cap (US Plastics). A small volume of water was removed from the top of the sediment core to provide room for expansion during freezing. The top of the coring tube was then capped. Sealed cores were placed vertically into an insulated chest and packed with dry ice in order to freeze the cores to minimize disturbance of the sediments and the sediment – water interface due to the motion of the boat and motion of the vehicle during transport. The cores were frozen solid and remained on dry ice until transferred to a -15 °C freezer in the lab.

#### 3.3.2.2 Core Sampling and Processing

Frozen cores were subsampled at 1 cm intervals using a power miter saw. The saw blade
Figure 3.4: Detail of sediment coring device
was 3 mm thick resulting in sections of roughly 7 mm. The plastic coring tube sections were labeled and retained to measure the actual height of each section. Wet sediments were dried in a 105 ± 2 °C oven overnight. Porosity measurements were based upon the mass lost upon drying, the mass of sediment remaining and the volume of the core section. Corrections based upon the salt content of estuary water and the mass of water lost upon drying were made to account for the mass of salts contained within the brackish Bay water.

Sediment core sections selected for isotope analysis were ground into a fine powder by hand using a mortar and pestle. The sections not used for isotope analysis were sealed in plastic bags.

Ground sediment samples were prepared for isotope analysis as follows: 20 grams of sediment were slowly placed into a Prindle vial. The Prindle vials are the standard vials used for isotope analysis at Lawrence Livermore National Laboratory (Livermore, CA). During the filling process the vial was tapped multiple times to encourage uniform sediment packing. A foam plug wrapped in plastic wrap was placed into the vial level with top. Cotton batting was placed on top of the foam plug and the vial sealed with a screw top lid. Vials were sealed within two days of core collection.

3.3.2.3 Radioisotope Analysis

Radioisotope analysis was performed under contract by the Lawrence Livermore National Laboratory. Radionuclide activities were determined using germanium solid-state detectors in a low-level counting facility via gamma spectrometry. Activities of 7Be, 137Cs, 40K, 210Pb, 226Ra, 228Ra, 228Th, 235U and 238U were determined. Counting times were between 24 and 72 hours. The measured activity and the 2σ uncertainty were reported for all samples.

3.4 Consolidation Experiment

Several 20 – 25 cm sediment cores were returned to the laboratory unfrozen and used for the consolidation experiment. After sieving the sediments to remove particles greater than approximately 5 mm, mainly worm tube fragments, the sediments were kept in the laboratory at room temperature. As some settling occurred between the sediment collection and the setup of the consolidation experiments the sediments were agitated by hand to resuspend them and create a uniform mixture.

The uniform mixture was sampled to determine the initial sediment concentration. Acrylic tubes, identical to those used in the coring device (Section 3.3.2.1) were used for the consolidation experiments. The bottom end of the tube was closed with 3.175 mm (0.125 in) thick acrylic sheets (TAP Plastics). The sediment mixture was poured into the acrylic tubes and allowed to settle for 1, 3 and 21 days. A second sample was collected from the bulk sediments after the last experimental tube was filled to check that the sediment concentration had not changed during the time required to prepare all of the consolidation samples. Both of the bulk sediment samples were stored in the freezer until processed with the rest of the consolidation samples. At the end of each consolidation period the cores were carefully moved to a -15 °C freezer. Sediment density measurements were made using the same sectioning, weighing and drying methods detailed in Section 3.3.2.2.
3.5 Determination of Additional Field Conditions from the Experimental Data

Most of the quantities which are used in the data analysis of Chapter 5 were not directly measured by the instrumentation used in the field experiments. Suspended sediment concentrations (SSC) data were determined from optical or acoustic instrument signals calibrated by laboratory and field measurements. Hydrodynamic data including bed shear stresses and turbulence values were extracted from the data collected by the ADVs and pressure sensors. The methods for calibrating and analyzing the raw data are presented here.

3.5.1 Turbulence

Turbulence data at the Benthic Station was extracted from the ADV measurements of velocity. Time series of three dimensional velocity vectors were measured at two elevations, 0.36 and 0.72 m above the bed (mab) as 8 minute measurement bursts at 10 Hz. The measured signal within each burst, \( x \), can be separated into three different components: the mean component, \( \bar{x} \), the fluctuating component due to turbulence, \( x' \), and the fluctuating component due to waves, \( x'' \), as:

\[
\begin{align*}
    x &= \bar{x} + x' + x''
\end{align*}
\]

The mean value is calculated as the mean of the parameter during the measurement burst.

The presence of waves at the water surface creates wave orbitals within the water column. For an instrument aligned perpendicular to the sediment bed the vertical and horizontal components of velocity due to the wave orbital motions would be 90° out of phase, therefore uncorrelated and would not contribute to the Reynolds stress. When the instrument is not aligned perpendicularly the wave orbital signal contaminates the signal and must be removed prior to the calculation of the Reynolds stress. The wave signal was differentiated from the turbulence signal by the method of Shaw and Trowbridge (2001). The method uses two instruments separated at a distance greater than the scale of the turbulent motions such that turbulent motions measured by the two instruments will be uncorrelated. Motions created by the waves will appear in both instrument signals and will be correlated. Therefore within the instrument measurements correlated signals are attributed to the wave motions and can therefore be identified and removed. This method does not work if the wave field is unsteady or if there is wave directional spreading. At the Benthic Station the two ADVs provided spatially separated signals which were used to distinguish \( x'' \) from \( x' \).

With the wave signal removed the shear velocity caused by the current, \( u_* \, \text{[m s}^{-1}] \) was calculated as:

\[
    u_* = \sqrt{\left< u w' \right>}
\]

where \( u' \, \text{[m s}^{-1}] \) and \( w' \, \text{[m s}^{-1}] \) are the fluctuating components of the horizontal and vertical velocity. Based upon a law of the wall profile the turbulent energy dissipation rate, \( \varepsilon \, \text{[m}^2 \, \text{s}^{-3}] \), was calculated as:

\[
    \varepsilon = \frac{u_*^3}{\kappa Z}
\]

48
where \( z \) [m] is the measurement elevation above the bed and \( \kappa \) is von Karman’s constant. Fluid shear rate, \( G \) [s\(^{-1}\)], was calculated from \( \varepsilon \) using Equation 2.5.

### 3.5.2 Wave Parameters

Wave height was calculated from pressure data. Following Wiberg and Sherwood (2008) the significant wave height, \( H_s \) [m], may be approximated as:

\[
H_s = 4 \sqrt{\sum_i S_{n,i} \Delta f_i}
\]  
(3.7)

where \( S_{n,i} \) is the spectral density of surface elevation as a function of frequency, \( f \), and the summation is taken across the \( i \) frequency interval of size \( \Delta f \). From Grace (1978) the spectral density of surface elevation can be determined from a pressure sensor at a height \( z_p \) [m] above the bed as:

\[
S_{n,i} = \left[ \frac{\cosh(k_i H)}{\cosh(k_i z_p)} \right]^2 S_{h,i}
\]  
(3.8)

where \( H \) [m] is the water depth, \( k_i \) [m\(^{-1}\)] is the wave number and \( S_{h,i} \) is the spectral density of the pressure head. The spectral density of the pressure head is related to the spectral density of the pressure, \( S_{p,i} \), as:

\[
S_{h,i} = \frac{S_{p,i}}{(\rho_f g)^2}
\]  
(3.9)

in which \( \rho_f \) [kg m\(^{-3}\)] is the fluid density and \( g \) [m s\(^{-2}\)] is gravity. Finally, the root mean square wave height, \( H_{rms} \) [m], is calculated from \( H_s \) after the substitutions of Equations 3.8 and 3.9 into Equation 3.7 as:

\[
H_{rms} = \frac{H_s}{\sqrt{2}} = \frac{4 \sqrt{2}}{\sqrt{2}} \sqrt{\sum_i \left[ \frac{\cosh(k_i H)}{\cosh(k_i z_p)} \right]^2 S_{p,i} \Delta f_i \left( \frac{\rho_f g}{r_f} \right)^2}
\]  
(3.10)

The wave number, \( k_i \), is solved for iteratively using the method described in Wiberg and Sherwood (2008) since:

\[
k = \frac{\omega^2}{g \tanh(k H)}
\]  
(3.11)

where \( \omega \) is the wave frequency in radians.

### 3.5.3 Bed Shear Stress

Under conditions of both waves and currents the total bed shear stress \( (\tau_{cw}) \) is a non-linear interaction of the bed shear stresses generated by each component and is determined by:
\[ \tau_{cw} = \tau_c + \tau_{wm} \] (3.12)

where \( \tau_c \) [Pa] is the current bed shear stress and \( \tau_{wm} \) [Pa] is the maximum wave bed shear stress. Waves at the surface will create wave-induced bed shear stress when the wave orbital motions extend to the sediment bed. The wave shear stress, \( \tau_w \), can be described as:

\[ \tau_w = \frac{1}{2} \rho_f f_w u_b^2 \] (3.13)

where \( \rho_f \) [kg m\(^{-3}\)] is the water density, \( f_w \) is a wave friction factor and \( u_b \) is the bottom orbital velocity [m s\(^{-1}\)]. According to linear wave theory \( u_b \) depends upon the wave frequency, \( \omega \) in radians, the wave amplitude, \( a \) [m], the wave number, \( k \) [m\(^{-1}\)], and the water depth, \( H \) [m], as (Wiberg and Sherwood, 2008):

\[ u_b = \frac{\omega a}{\sinh(kH)} \] (3.14)

Values for \( u_b \), \( \omega \) and the orbital velocity direction were determined from spectral information of the velocity data following Wiberg and Sherwood (2008).

The combined current-wave shear velocity, \( u_{*,cw} \) is calculated by Grand and Madsen (1986):

\[ u_{*,cw} = \left[ u_{*,w}^4 + 2u_{*,c}^2 u_{*,w}^2 \cos(\varphi_{cw}) + u_{*,c}^2 \right]^{\frac{1}{4}} \] (3.15)

where \( \varphi_{cw} \) (0 < \( \varphi_{cw} < \pi/2 \)) is the angle between the wave and current directions and the relationship between bed shear stress, \( \tau \) [Pa], and shear velocity, \( u_* \) [m s\(^{-1}\)], is:

\[ \tau = \rho_f u_*^2 \] (3.16)

The shear velocity for the wave component in Equation 3.15 is given by:

\[ u_{*,wm}^2 = \lim_{z \to z_0} \left( K \frac{\partial u_w}{\partial z} \right) \] (3.17)

and for the current:

\[ u_{*,c}^2 = \lim_{z \to z_0} \left( K \frac{\partial U}{\partial z} \right) \] (3.18)

in which \( K \) [m\(^2\) s\(^{-1}\)] is the time dependent eddy viscosity, \( u_w \) [m s\(^{-1}\)] is the modulus of the wave velocity in the lower part of the wave boundary layer, \( U \) [m s\(^{-1}\)] is the magnitude of the current velocity, \( z \) [m] is the vertical coordinate, positive upwards from the bed, and \( z_0 \) [m] is the hydraulic roughness (Styles and Glenn, 2000; Styles and Glenn, 2002b). The solution of the non-linear system comprised of Equations 3.15, 3.17 and 3.18 requires profiles of \( K \) as a function of \( z \). The three layer eddy viscosity model of Styles and Glenn (2000) was utilized:
where $\kappa$ is von Karman’s constant, $z_1$ [m] is an arbitrary constant scale height and $z_2 = z_1 \cdot u_{cw} \cdot \omega^{-1}$ [m], derived by matching the eddy viscosities at $z = z_2$. Conceptually $z_1$ and $z_2$ are the lower and upper limits, respectively, of an eddy viscosity transition layer between the wave and current eddy viscosity layers. $z_1$ is calculated as:

$$z_1 = \alpha \left( 1 + \frac{k_b}{A_b} \right) l_{cw}$$

(3.20)

$k_b = 30z_0$ [m] is the Nikuradse equivalent roughness, $A_b$ [m] is the bottom wave excursion amplitude and $\alpha$ and $b$ are closure constants (Styles and Glenn, 2002a). The scale height of the wave boundary layer for combined current and wave flows is $l_{cw} = \kappa \cdot u_{cw} \cdot \omega^{-1}$ [m] (Styles and Glenn, 2002b). Values of $\alpha = 0.3$ and $b = 0.7$, based upon laboratory flume data of Styles and Glenn (2002a) were used.

Solution of Equations 3.15 through 3.20 for $u_{cw}$ requires values for $u_r$, $z_r$, $z_0$, $A_b$, $u_b$ and $\phi_{cw}$ where $u_r$ is the mean current velocity at a known height, $z_r$, above the bed. Styles and Glenn (2000; 2002b) describe in more detail the iterative approach needed to reach a solution.

### 3.5.4 Turbulent Vertical Sediment Flux

Once calibrated the ADVs provide SSC measurements at the same location and frequency as the velocity measurements. The SSC signal may be similarly decomposed into the mean and fluctuating components, similar to Equation 3.4 without the wave component. The turbulent vertical sediment flux is then calculated as:

$$f = \overline{w' \cdot C'}$$

(3.21)

in which $w'$ [m s$^{-1}$] and $C'$ [kg m$^{-3}$], are the fluctuating components of the vertical velocity and SSC, respectively and the over bar indicates the quantity is averaged. For the Benthic Station ADV the averaging period was the 8 minute measurement burst made every 12 minutes.

The turbulent vertical sediment flux is a measure of vertical sediment movement due to turbulence. Assuming a typical sediment profile with high concentration near the bed which decreases upwards towards the water surface, $w' \cdot C'$ will always be positive as positive vertical velocity fluctuations, $w'$, will carry higher sediment concentration water (relative to the average value, $C'$) from below the sensor upwards past the measurement point ($C' > 0$). Similarly, when $w' < 0$ the turbulent motion will carry lower sediment concentration water relative to the average downwards past the measurement point ($C' < 0$). Negative values of $w' \cdot C'$ would indicate an atypical sediment profile. Near bed measurements of $w' \cdot C'$ were used to provide an estimate of the erosional flux from the sediment bed. This approximation assumes steady-state conditions in a one dimensional vertical sediment column and is dependent upon the distance between the sediment bed and the turbulent vertical sediment flux measurement location.
4 Results

The data presented here captures the general conditions during the instrument deployments. Much of it will be referred to or analyzed further in the following chapters. Presentation of some of the data is delayed until Chapter 5 where their inclusion with the analysis and discussion was beneficial. This is case for the $^7$Be results (section 5.1.1.1), the consolidation experiment results (section 5.1.3) and the ADCP SSC results (section 5.2.3.2).

4.1 Time Series Data

General conditions during the spring and fall deployments are presented in Figures 4.1 and 4.2, respectively. With the exception of the wind speed and direction which are from the San Francisco International Airport, the data are from the Benthic Station. Water depth ranged from roughly 2 to 5 m (Figures 4.1a, 4.2a). A similar tidal range was observed at the other stations. Mean water depth at lower low water was 2.2 m at the Benthic Station, 2.6 m at Shoal Middle, 8 m at Slope Middle and 14 m at Channel Station. High energy periods of the tidal cycle occurred in the Bay from DOY 56 to 60 and again DOY 69 to 74 during the spring deployment (Figure 4.1a) and DOY 259 to 264 and DOY 274 to 278 during the fall deployment (Figure 4.2a). Tides during these periods have roughly the same amplitude. The other periods during both deployments are low energy, marked by two weak tides with small depth changes and two larger tides during each day. The tidal range of the largest tide during this period is in some cases greater than the tidal ranges which occur during the high energy period. Horizontal velocities during both deployments follow the same high and low energy pattern. Horizontal velocities during the low energy period are smaller except during the single large tide each day. During the high energy periods horizontal velocities are roughly the same during each tide (Figures 4.1b and 4.2b).

Winds are generally westerly during the spring deployment (Figure 4.1c) but are southerly on DOY 55 and 62 to 63. Peak wind speeds were roughly 15 m s$^{-1}$, and wind speeds of 5 to 10 m s$^{-1}$ occurred on most days. During the fall deployment, winds are primarily westerly (Figure 4.2c). Note that because of the wind direction, the arrows align with the abscissa and point in the positive direction. Regular winds of roughly 10 m s$^{-1}$ occurred daily during the fall deployment. From DOY 270 to 278 two periods of sustained winds of 10 m s$^{-1}$ or greater occurred with maximum wind speeds of nearly 20 m s$^{-1}$ during each event. Due to these periods of sustained high winds and the effect they had upon sediment concentrations and sediment fluxes, the remaining discussion of the fall deployment has been split into the period before DOY 270 (Figures 4.2d-i) and the period of DOY 270 and after (Figure 4.3). The latter period is discussed separately.

Wave height at the Benthic Station was driven in large part by the wind speed with waves of 0.3 m or greater generated when the wind speed was greater than roughly 10 m s$^{-1}$. Periods of sustained wind generated larger waves (Figures 4.1d and 4.2d). Winds combined with strong currents generated regular periods of large waves during the fall deployments, particularly during DOY 256 to 261. During both deployments waves generated during the weaker tidal periods were smaller than those generated during the periods of high tidal energy.
Figure 4.1: Conditions during the spring deployment: (a) water depth, (b) horizontal velocity at 0.36 meters above the bed, (c) wind speed (gray) and direction (black arrows).
Figure 4.1 (continued): Conditions during the spring deployment: (d) wave height, (e) bed shear stress from current (black) and from current and waves combined (red) and (f) SSC at 0.36 mab.
Figure 4.1 (continued): Conditions during the spring deployment: (g) turbulent vertical sediment flux at 0.36 mab, (h) SSC at 0.72 mab and (i) turbulent vertical sediment flux at 0.72 mab.
Figure 4.2: Conditions during the fall deployment: (a) water depth, (b) horizontal velocity at 0.36 meters above the bed and (c) wind speed (gray) and direction (black arrows).
Figure 4.2 (continued): Conditions during the fall deployment: (d) wave height, (e) bed shear stress from current (black) and from current and waves combined (red) and (f) SSC at 0.36 mab. Note the data after DOY 270 are in Figure 4.3.
Figure 4.2 (continued): Conditions during the fall deployment: (g) turbulent vertical sediment flux at 0.36 mab, (h) SSC at 0.72 mab and (i) turbulent vertical sediment flux at 0.72 mab. Note the data after DOY 270 are in Figure 4.3.
The bed shear stress generated by the currents (black) and by the waves and currents combined (red) are shown in Figures 4.1e and 4.2e for the spring and fall deployments, respectively. In both cases waves greater than about 0.1 m resulted in greater bed shear stresses, in some cases 2 to 3 times greater than those generated by the currents alone. When wave height is small the bed shear stress is driven by the currents alone.

Measurements of SSC and the turbulent vertical sediment flux at 0.36 mab and 0.72 mab are shown in Figures 4.1f to 4.1i. Both SSC and turbulent vertical sediment flux measurements follow the same patterns at the two measurement elevations, though both are smaller at the station further from the bed. Peaks in both SSC and flux coincide with the periods of increased bed shear stress due to waves. During the low tidal energy period, in particular DOY 62 to 68 (Figure 4.1f to 4.1i), SSC and the flux remain very small for almost the entire period. The exception at DOY 66 was caused by a large tidal excursion combined with waves greater than 0.3 m which generated large bed shear stress and sediment erosion. During the high energy part of the tidal cycle regular peaks of SSC and flux occur with the tides but the magnitude of these are also increased during periods of high winds and large waves.

The fall deployment exhibits much of the same behavior as the spring deployment with respect to the SSC and turbulent vertical sediment fluxes (Figure 4.2f to 4.1i). The tidal peaks in SSC and flux are more distinct throughout the entire deployment. Greater values of SSC and flux occur whenever there are large winds and waves and this process is enhanced during the high energy tidal periods. SSC and flux decrease noticeably during the low energy tidal period, particularly from DOY 267 to 270.

SSC during the spring deployment is smaller (mean 0.014 kg m\(^{-3}\) and 0.011 kg m\(^{-3}\) at 0.36 and 0.72 mab, respectively) than during the fall deployment (mean 0.030 kg m\(^{-3}\) and 0.027 kg m\(^{-3}\), respectively, prior to DOY 270). The turbulent vertical sediment fluxes, with mean fluxes of 6.1\(\times\)10\(^{-6}\) kg m\(^{-2}\) s\(^{-1}\) and 3.5\(\times\)10\(^{-6}\) kg m\(^{-2}\) s\(^{-1}\) at 0.36 and 0.72 mab, respectively, during the spring versus 8.7\(\times\)10\(^{-6}\) kg m\(^{-2}\) s\(^{-1}\) and 6.4\(\times\)10\(^{-6}\) kg m\(^{-2}\) s\(^{-1}\), respectively, for the fall prior to DOY 270, follow a similar pattern. All of the SSC and turbulent vertical sediment flux data shown in Figures 4.1 and 4.2 are derived from the ADV data.

During the fall deployment a large storm occurred starting around DOY 270, marked by larger and sustained winds (Figure 4.2c), which caused high turbulent vertical sediment fluxes and SSC. This period corresponded to the transition between the low and high energy phases of the tidal cycle (Figure 4.2a). The time series in Figures 4.2d-i were truncated to DOY 270. The data omitted from those figures are shown in Figure 4.3 with a different scale for the ordinate. Waves generated during this period were as large as 0.7 to 0.8 m which caused bed shear stresses as high as 3 N m\(^{-2}\), at least 5 times greater than was generated by the currents alone. Peak values of SSC and turbulent vertical sediment flux occurred when large waves occurred during a strong flood or ebb current. The larger waves of DOY 273 caused smaller increases in SSC and flux because the currents during that period were weak. Peak SSC values during this period were 5 times greater than those measured during the rest of the fall deployment. Note, however, that there is some uncertainty regarding the SSC measurements during these events because the concentrations measured exceeded the values for which the ADVs were calibrated. It is not known if the calibrations developed for the lower SSC values are valid at the concentrations experienced in the field. However, as the OBS sensors had biofouled by this point in the deployment, the ADVs provide the only SSC measurements for this event.
Figure 4.3: Conditions during the fall deployment for DOY 270 and later: (a) wave height, (b) bed shear stress due to current (black) and combined current and waves (red), (c) SSC at 0.36 mab, (d) vertical turbulent sediment flux at 0.36 mab, (e) SSC at 0.72 mab and (f) vertical turbulent sediment flux at 0.72 mab.
4.2 Porosity Profiles

Density profiles in the sediment cores collected from the Benthic and Shoal Middle Stations are shown in Figure 4.4 as porosity. Near surface porosities are within the 0.75 to 0.85 range. All of the sediment cores show a similar decrease in porosity with the first 4 cm and reach a porosity of 0.7 at the deeper depths. The two cores collected at the Benthic Station on February 25, 2009 (filled and empty squares in Figure 4.4a) have porosity profiles which are substantially different. While the #2 core profile agrees well with the other sediment cores from that station, the other core is less dense at a given depth, but does approach the 0.7 porosity at the greatest depth sampled. The uppermost sediment layer sampled from the #1 core (not shown) was abnormally porous, perhaps from disturbance of the sediments during collection or poor demarcation of the sediment – water interface during the core processing. The datum was discarded as anomalous, but it is worth noting, however, that if that porosity profile were shifted upwards by 1 cm it would be similar to the other sediment profiles both near the surface and at the depth at which the porosity approaches 0.7.

There are no discernable patterns between seasons or location and differences between duplicate cores is as great as the variability observed between the cores collected on different days. The shape of the porosity profiles is consistent with sediments which have been consolidating for some time and may be fully consolidated. There is no indication of recent deposition or reworking of the sediments to a depth greater than 1 cm.

4.3 Particle Size Distributions

Particle size distributions (PSD) for in situ water column flocculated sediments were measured with a LISST (Section 3.2.3) and deflocculated bed sediments were measured in the laboratory using a Beckman Coulter laser diffraction particle size analyzer (Section 3.3.1). The measurement range for the LISST is 1.25 μm to 250 μm while the range for Beckman Coulter analyzer is 0.4 μm to 900 μm. The PSD are normalized to the total amount of sediment measured by each instrument. The bed sediments were disaggregated before analysis and exhibit two maxima, one near 6 μm and one near 45 μm (Figure 4.5, gray). The disaggregated PSD indicates that the bed sediments are comprised primarily of clay and silt particles. The Channel Station had a slightly greater percentage of the fine particles less than 10 μm and the Benthic and Slope Middle Stations contained a higher fraction of roughly 45 μm particles compared to the other locations. There is no significant difference between sediments collected at any of the stations. The data shown for the bed sediments are from the surface layer at the middle stations during the fall deployment. No significant differences were found for sediments collected later during fall deployment, sediments collected during the spring deployment or for the sediments which underlay the surface sediments (data not shown).

The LISST measured the PSD for the flocculated particles at 0.55 mab in the water column at the Benthic Station (Figure 4.5, black). Due to biofouling, data were only collected for the first five days of each deployment. For the PSD shown in Figure 4.5, the SSC values as measured by the LISST (μL L⁻¹) and by the ADV 0.36 m above bed (kg m⁻³) as well as the time the PSD were measured are listed in Table 4.1. The PSD shown were measured during the highest and lowest concentration events as measured by the LISST, with most of the other PSD measured falling between the high and low concentration PSD, although there are some PSD for which the modal value occurs at particle diameters greater than those shown in Figure 4.5. The modal values are greater than the majority of the particles measured in the disaggregated bed sediment samples indicating that most of the particles within the water column are comprised of...
Figure 4.4: Density profiles from sediment cores collected at the (a) Benthic and (b) Shoal Middle Stations.
Figure 4.5: Particle size distributions from bed sediments (grey) and the water column (black). Water column PSD are from the period of highest and lowest particle concentration (µL L⁻¹) measured in-situ by the LISST during the spring and fall deployments. Bed sediments are disaggregated sediments from the middle stations measured in the laboratory.

<table>
<thead>
<tr>
<th></th>
<th>Day of Year, 2009</th>
<th>SSC from LISST µL L⁻¹</th>
<th>SSC from 0.36 mab ADV kg m⁻³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring – low</td>
<td>56.11</td>
<td>10.6</td>
<td>0.004</td>
</tr>
<tr>
<td>Spring – high</td>
<td>58.25</td>
<td>208.5</td>
<td>0.051</td>
</tr>
<tr>
<td>Fall – low</td>
<td>254.03</td>
<td>22.4</td>
<td>0.009</td>
</tr>
<tr>
<td>Fall - high</td>
<td>254.48</td>
<td>223.3</td>
<td>0.090</td>
</tr>
</tbody>
</table>

Table 4.1: SSC measurements for the LISST PSD shown in Figure 4.5.
sediment flocs. The modal values for the LISST PSD are greater than 100 μm and during periods of high SSC the modal value of the PSD shift towards higher particle diameters. At the same time the relative contribution of particles with diameters less than 50 μm decreases. It is not clear if this is caused by a decrease in fine particles or an increase in larger diameter particles within the water column.

As the PSD measured by the LISST did not return to the abscissa at the larger particle diameters it is possible that particles greater the upper measurement limit (250 μm) were present. For LISST instruments it is known that particles which fall outside of the measurement range may still be registered by the instrument and falsely counted as belonging to the largest or smallest size classes measured. It is therefore uncertain as to whether the PSD for the smallest and largest size classes are an accurate representation of the PSD actually present in the water column. Differences in the PSD between the spring and fall deployments are small despite the higher SSC during the fall deployment.
5 Data Analysis and Discussion

Analysis of the field data is presented here as two complementary tracts: sediments in the bed and sediments in the water column, with the understanding that the sediments are frequently exchanged between the two. Most of the data for the sediment bed analysis comes from the Benthic Station with the exception of additional sediment cores collected from the Shoal Middle Station. Data from all of the stations are used for the water column sediment analysis, however due to the distribution of instruments the analysis of the water column sediments at the Benthic and Shoal stations is different from those carried out at the Slope and Channel stations.

5.1 Bed Sediments

The bed sediments were investigated using a number of tools to quantify different sediment bed processes. Beryllium-7 analysis was used for determining the sediment mixing depth into the bed, erosion relationships were determined from flux data from the field experiments and lab experiments were used to estimate consolidation rate parameters.

5.1.1 Beryllium-7

The presence of Beryllium-7 ($^7$Be) analysis was used for estimating the depth of the mixing zone within the sediment bed. Changes to the $^7$Be profile in a sediment bed with time can be used to estimate rates of sediment erosion or deposition.

5.1.1.1 Sediment Bed Activities

Detection limits varied between 0.0056 – 0.022 Bq g$^{-1}$ depending upon the counting time. The upper most sediment layers were generally counted longer to achieve a lower detection limit. The upper 5 – 7 cm from a total of 8 sediment cores was sent to Lawrence Livermore National Laboratory (LLNL) for $^7$Be analysis. Only one sample had $^7$Be activity greater than the detection limit. An activity of 0.0064 ± 0.0024 Bq g$^{-1}$ of $^7$Be was measured in the upper most layer of sediment collected from the Shoal Middle Station on March 13, the end of the first instrument deployment.

5.1.1.2 Beryllium-7 Activities in Near Surface Air

The major source for $^7$Be found in the Bay is the atmosphere. If the near surface air activities were smaller than usually reported, this may indicate that the source of $^7$Be to the Bay is also small. An insufficient supply is one possible explanation for why $^7$Be was only detected in one sediment sample. Beryllium-7 activities were monitored at two sites east of the South Bay: at LLNL, 47 km east (Lat: 37.687037° N, Long: 121.706680° W) and at Site 300, 63 km east (Lat: 37.633485° N, Long: 121.503304° W). The measurements were carried out by LLNL and the data are from the LLNL monthly monitoring reports (unpublished).

Monthly averaged near surface air activity data from January 2004 to early 2009 show that the average activity was 0.0028 Bq m$^{-3}$ at LLNL and 0.0056 Bq m$^{-3}$ at Site 300. The higher average activity at Site 300 is caused by four measurements where the $^7$Be activity was an order of magnitude higher than the other measurements. Excluding those four measurements decreases
the average to 0.0036 Bq m\(^{-3}\). The cause of the higher activities is not known. Most of the activities at both sites fall between 0.0015 – 0.005 Bq m\(^{-3}\) with high activities occurring in October or November and low activities occurring in or around March and August. The high activities in the late fall occur regularly each year, however inter-annual variability is as great as the month to month variability obscuring any other trends. The activities measured fall within the range of near surface activities reported in the literature (Section 2.4.1).

5.1.1.3 Discussion

\(^{7}\)Be has been used in a number of estuaries and coastal regions to measure sediment dynamics. However, \(^{7}\)Be was not detectable in most of our samples. There are several possible explanations for this including limited supply, deposition rates, and the amount of time sediments spend in the water column. As the near surface air activities are similar to those reported in the literature, supply of \(^{7}\)Be to the near surface is not likely the cause.

As deposition to the surface is caused primarily by wet deposition, it is possible that the climate of the San Francisco Bay area is a factor. During the fall deployment 7 mm total rain fell on DOY 256. No rainfall was recorded for three months prior to the start of the fall deployment. With little wet deposition it is possible that insufficient quantities of \(^{7}\)Be were deposited into the Bay prior to and during the fall deployment, preventing measurable amounts of \(^{7}\)Be from accumulating in the sediment bed. Without measurements of \(^{7}\)Be in the water column or on particles suspended in the water column it is not possible to know whether little \(^{7}\)Be was present in the Bay or if the \(^{7}\)Be was not being incorporated into the bed sediments.

If sufficient amounts of \(^{7}\)Be were present in the water column, scavenging of \(^{7}\)Be from the water column by suspended sediments should be affected by the concentration and the residence time of the suspended sediments. In the case of SSC of 10 – 100 mg L\(^{-1}\), 60 – 90% of \(^{7}\)Be found within the water column may exist as a dissolved phase rather than associated with suspended particulates (Olsen et al., 1986). Bloom and Crecelius (1983) found that for SSC in the range of 20 – 100 mg L\(^{-1}\) about 50 hours was required for the water column and suspended sediment \(^{7}\)Be distribution to reach equilibrium, though the most significant change occurred within the first 10 hours. Particle dynamics, in particular, the duration of time sediments are suspended in the water column and able to scavenge \(^{7}\)Be, is thought to be the most important factor for determining activities on sediment particles in the water column and in the sediment bed.

Modeling results of the Benthic Station indicate that the quantity of sediments resuspended from the bed during each tidal cycle is small, on the order of mm to 1 cm in depth (see Chapter 6). The 1 cm resolution at which the sediments cores were subsampled for analysis may have been too coarse to distinguish between a fine layer of \(^{7}\)Be rich sediments in a 1 cm interval dominated by sediments devoid of \(^{7}\)Be.

Beryllium-7 activities have been used in a number of other estuaries to determine the mixing or depositional depth of the sediments on the time scale of a few months. As it appears that there is a sufficient supply of \(^{7}\)Be into the near surface air of the Bay, and therefore also the Bay itself, the sediment activity data does not provide any evidence that the mixing depth of the sediments at the study location was greater than 1 cm for the months leading up to and including the instrument deployments.
5.1.2 Erosion Parameters

Erosion of sediments from the bed is the primary pathway for the introduction of sediments into the water column on the tidal timescale. The flux of sediments into the water column is generally modeled as a function of the bed shear stress, $\tau_b$. Several models have been developed for the erosion process (Section 2.3.2). The choice of any particular model depends upon the data being modeled.

The relationship between the bed shear stress, $\tau_b$ (N m$^{-2}$), and the erosion rate from the bed, $E_s$ (kg m$^{-2}$ s$^{-1}$), was determined from the $\tau_b$ and the turbulent vertical turbulent sediment flux data, $\overline{w'C'}$, calculated according to Section 3.5.4. The flux data were measured at the Benthic Station by the ADV 0.36 m above the bed (mab) during both deployments. Both the $\tau_b$ and the $\overline{w'C'}$ data are average values based on 8 minutes of data every 12 minutes. During each ebb and flood the period of time from the minimum $\tau_b$ near high slack tide or low slack tide until the maximum $\tau_b$ at maximum flood or ebb tide was used for the analysis. This assumes that the increase in vertical turbulent sediment flux with increased $\tau_b$ is caused by erosion from the bed and that the vertical turbulent flux of sediments at 0.36 mab is a good approximation for the vertical sediment flux at the bed.

The data plotted in Figure 5.1 has been separated into several subsets for analysis. The data from the spring and fall deployments were analyzed separately and the fall deployment data was further separated into the data prior to and after DOY 270. This separation point is roughly the start of a large storm which caused significantly higher vertical turbulent sediment fluxes and bed shear stresses which completely obscure the rest of the data. Finally, each period of increasing bed shear stress analyzed was characterized as a high tide, a low tide, or a weak tide, as illustrated in Figure 5.2. As the tides of the Bay are mixed, tidal range varied between 1 and 3 m depending on position in the neap – spring cycle. Weak tides represent the smallest tidal range where as the high and low tides are characterized by the greater tidal range and are further separated by water depth. The determination of a weak tide is somewhat arbitrary particularly as the tidal range exists as a continuum within the maximum and minimum values. The data was separated to help explain the range of values measured. All of the data collected during both deployments were also grouped together and analyzed.

The following power law relationship was chosen to model the data:

$$E_s \equiv (\overline{w'C'}) = M \tau_b^n; \quad \tau_b > \tau_c$$
$$E_s = 0; \quad \tau_b \leq \tau_c$$

where $M$ [kg m$^{-2}$ s$^{-1}$] and $n$ [-] are fitting parameters and $\tau_c$ [Pa] is the critical shear stress for erosion, the values for which are in Table 5.1.

5.1.2.1 Discussion

Considering the data shown in Figure 5.1 there is a point in each plot where the relationship between the bed shear stress and $\overline{w'C'}$ changes abruptly. This is clearest in Figure 5.1a at a value of $\tau_b \approx 0.05$ N m$^{-2}$, but it is noticeable in all of the plots. This point was assumed to be the critical shear stress for erosion, $\tau_c$ (Table 5.1). For $\tau_b < \tau_c$ the turbulent vertical sediment flux is assumed to be negligible, with $\overline{w'C'} \approx 10^{-6}$ kg m$^{-2}$ s$^{-1}$. The values for $n$ presented here fall within the range previously reported in the literature (see Section 2.3.2). Other relationships for modeling erosion were considered but the relationship shown in Equation 5.1
Figure 5.1: Determining erosion rate parameters for (a) the spring deployment, (b) the fall deployment prior to DOY 270 and (c) for all of the spring and fall deployment data. Parameters for the power law fits shown are in Table 5.1. The dashed line for All data series in each plot is the same.
Figure 5.1: Determining erosion rate parameters (continued).

Figure 5.2: Division of the data into low tide, weak tide and high tide series for erosion parameter estimation. Each series starts at the minimum $\tau_b$ and ends at the maximum $\tau_b$. 
was found to best match the data for $\tau_b > \tau_c$. Note that the relationship will underestimate the erosional sediment flux at small values of the bed shear stress. Brand et al. (2010) estimated $\tau_c = 0.1$ N m$^{-2}$ for the same spring deployment data presented here, stating that the flux increased significantly above that bed shear stress. For the spring data, particularly during high tides that may be an accurate characterization of the data. For the low tide data during the spring deployment such a distinction is not as easily made, nor for any of the fall data. However, the power law relationships presented here are not incompatible with the conclusions of Brand et al. (2010) as very low fluxes are predicted for bed shear stresses below 0.1 N m$^{-2}$.

The data shows differences both by deployment and as a function of water depth and tidal range. During both deployments, for a given bed shear stress the turbulent vertical sediment flux is greatest for the high tide data, followed by the weak tide data and then the low tide data with the smallest flux.

Large wind events can generate large surface waves, greatly increasing the bed shear stress, as occurred during the fall deployment from roughly DOY 270 until the instruments were recovered. Bed shear stresses were 2 to 3 times larger resulting in vertical turbulent sediment fluxes 5 times greater than observed during any of the previous measurements. As shown in Figure 5.1c the data are well described using the power law relationship of Equation 5.1. In this case the model is fit to all of the data from both deployments but the relationship parameters are set almost entirely by the data from DOY 270 to 278 alone. This All data relationship determined from the data in Figure 5.1c represents the data at the lower shear stresses well, however it generally under predicts the turbulent vertical sediment fluxes for a given bed shear stress during the spring deployment (Figure 5.1a) and over predicts during the fall (Figure 5.1b).

The data demonstrates the difficulty in trying to parameterize a complex process using a simple equation. Any such parameterization neglects the effects of erosion, deposition and consolidation. Under most circumstances, however, such relationships may sufficiently model sediment erosion and also provide a range of reasonable values which may be used for predicting sediment erosion in the Bay.

### 5.1.3 Consolidation Parameters

Consolidation is known to change the porosity and shear strength of cohesive sediments. It is therefore important to incorporate the effects of consolidation when considering estuarine sediments. Consolidation is not a well understood phenomenon and models are more accurate when calibrated with field or experimental data from the site under consideration. Consolidation

<table>
<thead>
<tr>
<th></th>
<th>$M$ (kg m$^{-2}$ s$^{-1}$)</th>
<th>$n$ (-)</th>
<th>$\tau_c$ (N m$^{-2}$)</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring low tide</td>
<td>7.8e-5</td>
<td>1.8</td>
<td>0.04</td>
<td>0.70</td>
</tr>
<tr>
<td>Spring weak tide</td>
<td>3.8e-4</td>
<td>2.4</td>
<td>0.06</td>
<td>0.78</td>
</tr>
<tr>
<td>Spring high tide</td>
<td>7.0e-4</td>
<td>2.5</td>
<td>0.06</td>
<td>0.62</td>
</tr>
<tr>
<td>Fall low tide</td>
<td>6.3e-5</td>
<td>1.5</td>
<td>0.03</td>
<td>0.68</td>
</tr>
<tr>
<td>Fall weak tide</td>
<td>3.9e-5</td>
<td>1.1</td>
<td>0.01</td>
<td>0.77</td>
</tr>
<tr>
<td>Fall high tide</td>
<td>8.0e-5</td>
<td>1.5</td>
<td>0.02</td>
<td>0.72</td>
</tr>
<tr>
<td>All data</td>
<td>6.7e-5</td>
<td>1.5</td>
<td>0.02</td>
<td>0.74</td>
</tr>
</tbody>
</table>

Table 5.1: Power law parameters for Equation 5.1 for the best fit relationships shown in Figure 5.1. The fits to the data exclude the data for which $\tau_b < \tau_c$. The erosion rate is assumed to be zero in those cases.
experiments were conducted with sediments collected at the Benthic Station with density profiles collected after 1, 3 and 21 days.

The MATLAB code used for analyzing the consolidation experiments was developed by Sanford (2008) and was included with the sediment bed model MATLAB code used for the numerical modeling (see Chapter 6). From a user defined initial density profile the model relaxes the density profile towards a user defined final density profile. The consolidation rate determines how quickly the final density profile is approached. Numerical implementation of the consolidation scheme is discussed further in Section 6.1.1 where it was included as part of the sediment bed – water column model developed to model the Benthic Station data.

The final density profile, as porosity, \( \phi_{eq} \), is defined by:

\[
\phi_{eq} = (\phi_0 - \phi_\infty) e^{cz} + \phi_\infty
\]

where \( \phi_\infty \) [-] is the minimum porosity, \( \phi_0 \) [-] the porosity at the surface, \( z \) [m] the distance from the surface and \( c \) [m^{-1}] is a fitting parameter. The parameters for the final density profile were determined by fitting Equation 5.2 to density profiles from four sediment cores collected at the Benthic Station at the start and finish of each instrument deployment. The parameters for the fit to the data shown in Figure 5.3 are: \( \phi_\infty = 0.69 \), \( \phi_0 = 0.765 \) and \( c = 30 \) m^{-1}.

The consolidation model relaxes the current porosity profile towards the final porosity profile according to:

\[
\phi^l = \phi^{l-1} - \Delta t \cdot (\phi^{l-1} - \phi_{eq}) \cdot CR
\]

where the superscripts indicate the current time step (i) and the previous time step (i-1), \( \Delta t \) is the model time step and \( CR \) [s^{-1}] is the consolidation coefficient. The numerical consolidation model was fit to the consolidation experimental data by changing CR. All other parameters were kept constant. The experimental data and model results are shown in Figure 5.4. Porosity measurement errors were estimated to be 0.01. The model represented in Equation 5.3 is a simplification which assumes a constant CR for each modeled time period with each model run beginning at the start of the consolidation period.

5.1.3.1 Discussion

The experimental data for days 1 and 3 have considerable scatter which made it difficult to determine the best model fit parameters for the model from the data. Ultimately the model was fit by minimizing the difference between the modeled and experimental measures of the porosity. It is not clear what caused the variation but it is interesting to note that, as the determination of the porosity was a destructive process, the highly variable density profiles shown in Figures 5.4a and 5.4b are from two separate consolidated sediment columns.

As indicated by the model results, CR is time dependent. A similar time dependence of the rate at which consolidation progresses has also been observed for the more complicated models which solve Equation 2.42. In those cases the time dependence is thought to be introduced via the relationship between the void ratio, \( e \), and the hydraulic conductivity, \( k \) (see Section 2.3.3). This dependence is not currently understood and the added complexity of those models is not warranted here, particularly since the results would still be a time dependent parameterization.

The consolidation coefficients fit to the data are the same range as the consolidation
Figure 5.3: Determination of the final density profile from the density profiles of the sediment cores collected at the Benthic Station during the instrument deployments. The model fit is the final profile described by Equation 5.2.

Figure 5.4: Experimental and model results for the consolidation experiments at (a) one day, (b) three days and (c) 21 days. The model was fit to the data by changing the consolidation coefficient. All other model parameters were kept constant.
Figure 5.4: Modeling sediment consolidation experiment results (continued).
coefficient used by Sanford (2008) to fit their data. Agreement of the model consolidation coefficient with one the consolidation coefficients found here may indicate that consolidation is most important to the sediment processes on that 1, 3 or 21 day timescale. These coefficients will serve as a point of reference for the models developed in Chapter 6.

5.2 Water Column Sediments

Sediment concentrations in the water column and hydrodynamic data from the Benthic Station were used to test the applicability of a previously developed simple flocculation model under field conditions. These data were also used to estimate particle settling velocities. At the deeper Slope and Channel stations acoustic backscatter data from acoustic Doppler current profilers were calibrated for estimating SSC and an analysis was carried out to determine the accuracy of these estimates. These data and the associated uncertainties were then used to estimate water column sediment loads as a function of water column velocity.

5.2.1 Second Order Particle Kinetics

Suspended sediment concentration time series data were measured via calibration of the acoustic backscatter signal from two acoustic Doppler velocimeters at the Benthic Station located 0.36 and 0.73 m above the bed (mab).

The theoretical background for this analysis was originally developed by Hunt (1982b) and is presented in Section 2.2.3. It was shown that under conditions of constant fluid shear rate \( G \) [s\(^{-1}\)] and starting from initially disaggregated particles, the decrease in concentration within the Couette reactor used in the laboratory experiments followed second order in particle concentration after a sufficiently long time period had passed for a quasi-steady state in particle size distribution shape to be reached. Further experiments indicated that the second order rate constant was also proportional to the square root of the fluid shear rate.

The sediments measured in the field data were subject to a time varying fluid shear rate thus it was necessary to incorporate the fluid shear rate dependency into the determination of the rate constant. The incorporation of the \( b \propto G^{0.5} \) dependency into Equation 2.15 gives:

\[
C(t) = \frac{1}{a + kG^{0.5}t} \quad 5.4
\]

and taking the inverse yields:

\[
\frac{1}{C(t)} = a + kG^{0.5}t \quad 5.5
\]

The coefficients \( a \) [m\(^3\) kg\(^{-1}\)] and \( k \) [m\(^3\)·kg\(^{-1}\)·s\(^{-0.5}\)] are determined from linear regression of the inverse of suspended sediment concentration against the new parameter, \( G^{0.5}t \) [s\(^{0.5}\)]. Time, \( t \), in Equations 5.4 and 5.5, is the time elapsed since flocculation began, here taken to be the elapsed time since SSC began to decrease. From Equation 5.5, \( a^{-1} \) is \( C(t = 0) \). Each period of decreasing SSC was evaluated to determine the value of \( k \). An example of the analysis carried out is shown in Figure 5.5.

For comparison the data were also evaluated using a first order model which assumes:
Figure 5.5: Determination of the (a) second order and (b) first order rate constants for a period of decreasing SSC.
The solution to Equation 5.6 is \( C(t) = C_0 \exp(-w_s t) \) in which \( w_s \) [m s\(^{-1}\)] is the settling velocity and \( C_0 \) [kg m\(^{-3}\)] the SSC at \( t = 0 \). Linearization as:

\[
\ln[C(t)] = \ln(C_0) - w_s t
\]

allows for the determination of \( w_s \) and \( C_0 \) using linear regression analysis (Figure 5.5). The two analyses shown in Figure 5.5 use the same data.

The model predictions developed from the analyses of Figure 5.5 are shown plotted against the original data in Figure 5.6. In the example shown the second order model is a better fit to the data. However, in many cases the first order model was a better fit. A range of values for the second order rate constant, \( k \), was generated from the experimental data. These values are summarized in Figure 5.7. The box plots show the 25\(^{th} \), 50\(^{th} \) and 75\(^{th} \) percentiles as well as the range of values for \( k \) from the analysis. The mean value is indicated by the solid diamond while the other symbols are the values determined by Hunt (1982b) and Hunt and Pandya (1984) in laboratory experiments.

### 5.2.1.1 Discussion

The values for \( k \) at both elevations are consistent for each deployment. However, \( k \) is an order of magnitude smaller for the fall deployment when compared to the spring deployment. This potentially could be explained by differences in the sediments between the two deployments. Particles more amenable to flocculation (e.g. particles which are more sticky) would flocculate, and therefore be removed from suspension, with a greater rate constant. Smaller floc sizes during the fall deployment, as measured by the LISST, indicating less sticky particles, might support such a hypothesis, but this is an indirect measure and the behavior is equally well explained by smaller settling velocities due to smaller particles as a result of particle breakup caused by greater turbulence during the fall deployment. The order of magnitude difference in \( k \) between the two deployments in not definitively explainable using the data available.

The large range of measured values for \( k \), covering 3 – 4 orders of magnitude, makes it unlikely that the second order model used here is useful for analyzing the field data. During the duration of each deployment the sediments which were measured should not have undergone any significant changes that would have caused the value of \( k \) to vary so widely. No pattern was found in the variation of \( k \) over time to indicate a slow variation is sediment properties. The conclusion reached is that the second order model as put forth here is insufficient for describing the periods of decreasing sediment concentration observed in the field data. Lateral gradients in SSC (section 5.2.2) and the resuspension and sedimentation of sediments without significant amounts of flocculation (Chapter 6) appear to have a greater affect on SSC than does flocculation.

The extension of this model from the laboratory into the field has highlighted an apparent deficiency of the scientific understanding of flocculation. While most laboratory experiments are carried out under conditions of constant turbulence, the conditions encountered in an estuarine system are subject to tidal forcing and therefore vary with time. The timescale required for flocculation to approach a steady state relative to the timescale of turbulence variation due to the
tidal forcing will determine whether a steady state assumption is justifiable. It is not understood, however, how flocculated particles respond to time varying levels of turbulence. Laboratory experiments conducted by Burban et al. (1989) considered the response of particle flocculation and breakup to a step change in fluid shear rate after an initial equilibrium had been achieved. $G$ ranged from 100 to 400 s$^{-1}$ and after 10 to 40 minutes a new equilibrium mean floc diameter was established. The fluid shear rate encountered during the field experiments considered here were two orders of magnitude smaller. While the laboratory experiments demonstrate that particles will adjust to changing levels of turbulence it is not clear how significant these changes may be given the levels and rate of change of $G$ encountered in the field. It is also not clear how important floc breakup may be for establishing floc diameter in the estuarine environment. If the effects of flocculation and breakup are minor for these field conditions then changing SSC will be driven primarily by resuspension and settling dynamics and lateral transport not flocculation.

### 5.2.2 Settling Velocity

Particle settling is an important part of any estuarine particle dynamics. Measuring particle settling velocity in situ without disturbing the particles is difficult and often labor intensive. Estimates of settling velocity can be made using in situ instrumentation but such estimates require certain conditions to be met and may not be applicable to all situations. The settling velocity was estimated by assuming the vertical settling flux was balanced by the upward flux of sediments due to turbulence (Fugate and Friedrichs, 2002). See Section 2.2.4 for a further discussion of the simplifications.

If the conditions of Equation 2.33 and Equation 2.34 are both met then the simplification of Equation 2.25 to a vertical balance of settling and turbulent diffusion is justified. SSC data
Figure 5.7: Range and 25th, 50th and 75th percentiles for the second order rate constant, k, by deployment and height above bed. The diamond is the mean value of k and the hollow symbols are values from laboratory experiments for comparison.
Figure 5.8: Magnitude of the (a) concentration transient term and (b) the advection term compared to the settling term, Equations 2.34 and 2.35, respectively. Analysis shown is for the spring deployment at the Benthic Station 0.36 meters above the bed.
from the 0.36 mab and the 0.72 mab ADVs at the Benthic Station was used to measure vertical
gradients in SSC. Velocity and SSC data from an ADV 0.5 mab at the Shoal Middle Station,
1,090 m southwest of the Benthic Station (Figure 3.1), was used to calculate horizontal gradients
in velocity and SSC. The magnitude of the terms relative to the settling term are shown in Figure
5.8. The most robust estimate of \( w_s \) would be generated using only the data for which the ratios
plotted in Figure 5.8 are much less than one. If the contributions of the concentration transient
term and the advection term to Equation 2.25 were an order of magnitude smaller than the
settling term their contributions could reasonably be neglected.

At the Benthic Station such restrictive conditions resulted in too few data points for a
reasonable estimate of the settling velocity to be made. The data shown in Figure 5.9 are
therefore those data for which the ratio was less than 1. This restricts the data used to those
where settling is still more important but will include some data for which the contributions of
the concentration transient and advection terms are as important as the contributions of the
settling term. The settling velocity, \( w_s \), is approximated from the slope of the linear relationship
between the turbulent vertical sediment flux, \( \bar{\nu} \bar{C} \) and the SSC. The estimated settling velocities
for the spring and fall deployments are 0.87 and 0.27 mm s\(^{-1}\), respectively. These values are
within the range of settling velocities reported in the literature (Section 2.2.4). However, as the
conditions of Equations 2.33 and 2.34 are not strictly met, the estimates of \( w_s \) presented here
should be considered first approximations.

5.2.3 ADCP Backscatter

Calibration of the ADCP backscatter data for estimating SSC was presented in Section
3.2.2.1 and the development of the calibration equation is included in Appendix A. This section
focuses upon the development and interpretation of error estimates for the SSC values
determined from the ADCP backscatter.

5.2.3.1 Error Analysis Equations

An unknown amount of uncertainty is associated with the SSC values determined from
ADCP backscatter, introduced by uncertainties in the experimental parameters and uncertainties
introduced by the calibration process itself. When the value of a parameter is not known exactly,
the uncertainty can be determined from the experimental data directly (\( E_l, a_t, a_p, SSC \)), from
reported instrument accuracy (\( S, T, P, z \)), or from an accepted range of possible values (\( K_c, \rho_a \)).
Salinity was collected with an instrument with a stated uncertainty of ±0.005 psu. Instrument
uncertainties were taken to be ±2\( \sigma \). The salinity standard deviation is then estimated as
\( \sigma = 0.0025 \) psu. Standard deviations for temperature, depth and pressure were estimated to be \( \sigma = 0.0025 \) °C, 0.01 m, 0.01 atm, respectively, following the same rational. Transducer radii were
measured directly with \( \sigma = 0.001 \) m, the accuracy of the scale used. The SSC value used to
estimate the sediment attenuation factor (Equation A.7) was not well known over the whole
water column for the entire measurement period. Measurements at a single point for only a few
days were the only SSC data available. The SSC standard deviation was set to the standard
deviation of the available record of SSC data. The typical value for \( K_c \) is 0.45 dB count\(^{-1}\) but may
range from 0.35 to 0.55 dB count\(^{-1}\) (Deines, 1999). \( K_c = 0.45 \) dB count\(^{-1}\) was used for all
calibrations and the standard deviation of \( \sigma = 0.05 \) such that the known range of values was
captured within a 2\( \sigma \) variation. The mean value and standard deviations used are listed in Table
3.3.
Figure 5.9: Estimating the settling velocity at the Benthic Station from acoustic Doppler velocimeter measurements of the turbulent vertical sediment flux and SSC. The slope gives the settling velocity in m s\(^{-1}\) during (a) the spring and (b) the fall deployments.
The basis for the error analysis is the propagation of the uncertainties (measurement, instrument, parameter etc.) through the calibration process to establish uncertainties for the estimated SSC. As described by Bevington and Robinson (2003), given variables \( u \) and \( v \) with variances \( \sigma_u^2 \) and \( \sigma_v^2 \), respectively, and given that \( x \) is some function of \( u \) and \( v \), the variance for \( x \) may be approximated as:

\[
\sigma_x^2 \approx \sigma_u^2 \left( \frac{\partial x}{\partial u} \right)^2 + \sigma_v^2 \left( \frac{\partial x}{\partial v} \right)^2 + 2\sigma_{uv} \left( \frac{\partial x}{\partial u} \right) \left( \frac{\partial x}{\partial v} \right) \tag{5.8}
\]

where \( \sigma_{uv} \) is the covariance between \( u \) and \( v \) and variance is the square of standard deviation. When the relationship between \( u \) and \( v \) is non-linear higher order terms in the Taylor series expansion of the relationship would be needed, but have been neglected in developing Equation 5.8. Over the averaging period defined by each set of 50 ADCP measurements, all of the variables with the exception of the echo intensity, EI, are taken as constants as they were measured less frequently than once per minute. The covariance term may be neglected in those cases as the covariance of any parameter and a constant is zero. The covariance between beams within each averaging period was found to be negligible. Over short periods of time variations in EI measurements from one beam are uncorrelated with measurements from a different beam. The application of Equation 5.8 to specific mathematic functions is readily available e.g. Bevington and Robinson (2003). The process of propagating the errors is straightforward up through the calculation of \( S_v \) and the associated variance. The equations are detailed in Appendix B.

Once values of \( S_v \) have been calculated they are plotted against the log10(SSC) and a linear regression analysis performed to determine the values of the slope and intercept as discussed in Section 3.2.2.2. Uncertainties in both \( S_v \) and SSC will contribute uncertainties to the slope and intercept values. However, as long as the uncertainty for the independent variable is much smaller than uncertainty in the dependent variable all of the uncertainty may be ascribed to the dependent value (Bevington and Robinson, 2003). In this case the independent variable is the SSC measured via an OBS or filtered water samples and is considered to be known with greater certainty than the values of \( S_v \) determined from the ADCP data. Treating the SSC values as the independent variable necessitated the use of Equation 3.3 rather than Equation 3.1. With this simplification the variances for the slope, \( a \), and intercept, \( b \), may be determined from (Bevington and Robinson, 2003):

\[
S_v = a \cdot \log_{10}(SSC) + b \tag{5.9}
\]

\[
\sigma_a^2 = \frac{1}{\Delta} \sum_{i=1}^{N} \frac{\left( \log_{10}(SSC_i) \right)^2}{\sigma_{S_v_i}^2} \tag{5.10}
\]

\[
\sigma_b^2 = \frac{1}{\Delta} \sum_{i=1}^{N} \frac{1}{\sigma_{S_v_i}^2} \tag{5.11}
\]

where \( \Delta \) in Equations 5.10 and 5.11 are given by:
\[ \Delta = \left[ \sum_{i=1}^{N} \frac{1}{\sigma_{SV}^2} \right] \left[ \sum_{i=1}^{N} \frac{\log_{10}(SSC_i)^2}{\sigma_{SV}^2} \right] - \left[ \sum_{i=1}^{N} \frac{\log_{10}(SSC_i)^2}{\sigma_{SV}^2} \right] \]  \tag{5.12}

\text{N in Equations 5.10 to 5.12 is the number of points used in the linear regression analysis. As the regression analysis is performed on the subset of S_v data for which values of SSC are also known the final step is to use the slope and intercept values to convert the remaining values of S_v to estimates of SSC. The uncertainties in S_v, slope and intercept are propagated through this last step as before to give an estimate of the uncertainty for the estimated values of SSC.}

The estimated standard error provides a measure of the uncertainty created by the regression analysis used for the calibration (Hines et al., 2009). Rearranging Equation 3.1 gives:

\[ \log_{10}(SSC_i) = AS_{vi} + B \]  \tag{5.13}

On the left hand side is the estimated value of \( \log_{10}(SSC_i) \) based upon the regression coefficients A and B which come from treating \( S_v \) as the independent variable. A and B cannot be related to the previous linear regression coefficients a and b. The mean square error (MSE) is calculated as:

\[ \text{MSE} = \frac{\sum_{i=1}^{N} \left[ \log_{10}(SSC_i) - \log_{10}(\overline{SSC}) \right]^2}{N - 2} = \frac{\sum_{i=1}^{N} \left[ \log_{10}(SSC_i) - (AS_{vi} + B) \right]^2}{N - 2} \]  \tag{5.14}

where \( \log_{10}(SSC_i) \) is the observed value and N is the number of points used in the regression. For each point \( h \) used in the regression the estimated standard error of the \( \log_{10}(SSC) \), \( S^2 \), is:

\[ S^2(\log_{10}(SSC_i)) = \text{MSE} \left[ \frac{1}{N} + \frac{(S_{vh} - \overline{S_v})^2}{\sum_{i=1}^{N} (S_{vi} - \overline{S_v})^2} \right] \]  \tag{5.15}

For points \( j \) not included in the regression analysis the estimated standard error of the difference between the true value and estimated value of \( \log_{10}(SSC) \) is:

\[ S^2[\log_{10}(SSC_j) - \log_{10}(\overline{SSC})] = \text{MSE} \left[ 1 + \frac{1}{N} + \frac{(S_{vj} - \overline{S_v})^2}{\sum_{i=1}^{N} (S_{vi} - \overline{S_v})^2} \right] \]  \tag{5.16}

In the preceding two equations \( \overline{S_v} \) is the mean of the backscatter coefficient data used in the regression.

5.2.3.2 Results

The calibration results are listed in Table 5.2. Values for A and B (Equation 3.1) and \( a^{-1} \) and \( b \cdot a^{-1} \) (Equation 3.3) are both included as the former are reported in the literature but the latter
Table 5.2: $S_v$ to SSC calibration parameters for Equations 5.201 and 5.211 for different calibration schemes

<table>
<thead>
<tr>
<th>Deployment</th>
<th>Station</th>
<th>Case</th>
<th>A</th>
<th>B</th>
<th>a⁻¹</th>
<th>b·a⁻¹</th>
<th>$R^2$</th>
<th>Data used for calibration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring</td>
<td>Channel</td>
<td>Base</td>
<td>0.025</td>
<td>-0.494</td>
<td>21.93</td>
<td>40.02</td>
<td>0.56</td>
<td>4.63 days, 7998 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Channel</td>
<td>Grab 5</td>
<td>0.021</td>
<td>-0.401</td>
<td>38.80</td>
<td>27.72</td>
<td>0.80</td>
<td>5 grab samples</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Base</td>
<td>0.030</td>
<td>-0.810</td>
<td>20.10</td>
<td>40.00</td>
<td>0.61</td>
<td>8.03 days, 10000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Avg SSC</td>
<td>0.030</td>
<td>-0.779</td>
<td>20.39</td>
<td>39.58</td>
<td>0.61</td>
<td>8.03 days, 10000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Max SSC</td>
<td>0.030</td>
<td>-0.789</td>
<td>20.53</td>
<td>39.47</td>
<td>0.62</td>
<td>8.03 days, 10000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>5000 points</td>
<td>0.025</td>
<td>-0.480</td>
<td>23.32</td>
<td>36.07</td>
<td>0.60</td>
<td>4.01 days, 5000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>1000 points</td>
<td>0.017</td>
<td>0.037</td>
<td>20.42</td>
<td>40.63</td>
<td>0.35</td>
<td>0.80 days, 1000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Grab 7</td>
<td>0.043</td>
<td>-1.551</td>
<td>23.10</td>
<td>36.84</td>
<td>0.98</td>
<td>7 grab samples</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope North</td>
<td>Base</td>
<td>0.020</td>
<td>0.006</td>
<td>20.91</td>
<td>35.41</td>
<td>0.42</td>
<td>0.58 days, 1000 data points</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope North</td>
<td>Grab 5</td>
<td>0.011</td>
<td>0.19</td>
<td>40.59</td>
<td>25.50</td>
<td>0.44</td>
<td>5 grab samples</td>
</tr>
<tr>
<td>Fall</td>
<td>Channel</td>
<td>Base</td>
<td>0.025</td>
<td>-0.43</td>
<td>26.66</td>
<td>33.09</td>
<td>0.67</td>
<td>9 grab samples</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Base</td>
<td>0.026</td>
<td>-0.102</td>
<td>26.67</td>
<td>20.97</td>
<td>0.87</td>
<td>5.79 days, 10000 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Avg SSC</td>
<td>0.025</td>
<td>-0.07</td>
<td>27.27</td>
<td>20.13</td>
<td>0.69</td>
<td>5.79 days, 10000 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Max SSC</td>
<td>0.025</td>
<td>-0.066</td>
<td>27.44</td>
<td>19.96</td>
<td>0.69</td>
<td>5.79 days, 10000 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>5000 Points</td>
<td>0.028</td>
<td>-0.19</td>
<td>27.60</td>
<td>18.77</td>
<td>0.77</td>
<td>2.89 days, 5000 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>1000 Points</td>
<td>0.024</td>
<td>0.12</td>
<td>33.08</td>
<td>9.28</td>
<td>0.78</td>
<td>0.58 days, 1000 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Grab 8</td>
<td>0.027</td>
<td>-0.307</td>
<td>30.24</td>
<td>19.21</td>
<td>0.82</td>
<td>8 grab samples</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope North</td>
<td>Base</td>
<td>0.024</td>
<td>-0.182</td>
<td>19.67</td>
<td>33.95</td>
<td>0.47</td>
<td>5.21 days, 7619 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope North</td>
<td>Grab 8</td>
<td>0.023</td>
<td>-0.119</td>
<td>35.46</td>
<td>13.82</td>
<td>0.83</td>
<td>8 grab samples</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope South</td>
<td>Base</td>
<td>0.032</td>
<td>-0.473</td>
<td>19.24</td>
<td>31.49</td>
<td>0.61</td>
<td>9.26 days, 15998 data points</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope South</td>
<td>Grab 9</td>
<td>0.021</td>
<td>-0.119</td>
<td>33.47</td>
<td>16.20</td>
<td>0.69</td>
<td>9 grab samples</td>
</tr>
</tbody>
</table>
are used in the calibrations. For each instrument deployment a Base case is listed in italics and several other cases may follow. The Base case is the calibration based upon the best available data. In most cases this is based upon for 4 – 10 days of data from an OBS located 1 – 2 meters below the surface. The length of OBS record available for the calibration is limited due to bio-fouling. At Slope North during the spring deployment the ADCP did not start recording data until several days into the deployment. The Base calibration was based upon the first 0.6 days of data however bio-fouling of the OBS may have already occurred. The loss of the one meter below surface OBS at the Channel Station during the fall deployment meant that the only data available for the calibration were the water column grab samples collected 1 meter below the surface.

Calibration results from Slope Middle Station Base calibration during the spring deployment will be discussed in detail. The data available at this station was the most complete during both deployments. The plots included are representative of those for the other calibrations. Figure 5.10 shows the regression analyses used to determine the relationship between $S_v$ and $\log_{10}(SSC)$. The plot contains 10,000 data points covering the first 8.03 days of the deployment.

With the calibration parameters determined the remaining SSC values from the calculated values of $S_v$, the results of which are shown in Figure 5.11 for the Slope Middle Station during the spring deployment. Results for the other stations and other calibration schemes are summarized in Table 5.3. The data show periodic resuspension of sediments into the water column during the first several days when the tidal energy is greatest. Towards the end of the sampling period the tides move into the lower energy phase and the water column is shown to clear and sediment resuspension occur less frequently and with smaller amplitudes. During this period, resuspension occurred primarily during the single large ebb tide of the low energy tidal cycle (near DOY 62.75, 63.75 and 64.75 in Figure 5.10). Significant backscatter was measured near the surface, particularly around the slack tides at DOY 61 and 62. The analysis for SSC interprets the high backscatter as suspended sediment. However these features correspond well with a period of high winds (Figure 4.1c) and surface waves of up to 0.5 m at the Benthic Station (Figure 4.1d). The high backscatter is probably due to air bubbles entrained into the water column by the waves. Below these features, which extend into the water column by as much as 3 m, SSC is low but increases towards the sediment bed. These features also show high variability, as measured by the standard deviation, which is not characteristic of suspended sediments, as will be discussed in Section 5.2.3.3. The ADCP measurements clearly capture resuspension and settling over the tidal cycle and a slow clearing of the water column during the low energy portion of the neap-spring cycle. The lowest ADCP measurement, 1.45 m above the bed, are considerably lower than the next higher bin and are deemed to be inaccurate. Similar discrepancies with the bin nearest the transducer were also reported by Wall et al. (2006).

Comparisons between OBS measurements of SSC and those made by the ADCP generally show good agreement (Figure 5.12). The affects of bio-fouling on the 1 m below surface (mbs) OBS are clearly visible starting at DOY 62.5 where the two SSC time-series diverge significantly. The effects of the entrained air bubble are also clearly visible in the ADCP data in the 1 mbs data and also in the 3 mbs data. Despite the 1.3 m distance between the lowest OBS and the lowest ADCP measurement the SSC measurements also agree quite well, although this was not always the case. Comparisons between the laboratory filtered laboratory samples and the ADCP measurements of SSC are shown in Figure 5.13, which also exhibit good agreement.
Figure 5.10: ADCP Backscatter Calibrations for Slope Middle during Spring Deployment

\[ y(x) = ax + b \]
\[ a = 0.03 \]
\[ b = -0.81 \]
\[ R = 0.78 \quad \text{(lin)} \]
Figure 5.11: ADCP calibration results: (a) estimated SSC, (b) standard deviation and (c) square root of estimated standard error for Slope Middle Station during spring deployment. Units are kg m\(^{-3}\). The black circles in panel (a) show the time and location of water column sample collection for laboratory measurement of SSC. The ADCP data compared against the OBS data to determine the calibration relationship have a smaller uncertainty than the rest of the ADCP SSC estimates when determining the estimated standard error. The dark blue line in panel (c) which extends to DOY 64 shows the period of smaller uncertainty where SSC measurements were also made by OBS.
Table 5.3: Average values of SSC and associated uncertainties for different calibration schemes. All values are in mg L⁻¹. Overbars indicate averages.

<table>
<thead>
<tr>
<th>Deployment</th>
<th>Station</th>
<th>Case</th>
<th>SSC</th>
<th>% Error 1σ</th>
<th>% Error ESE</th>
<th>Max % Error 1σ</th>
<th>Min % Error 1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring</td>
<td>Channel</td>
<td>Base</td>
<td>19.9</td>
<td>42</td>
<td>20</td>
<td>132</td>
<td>20</td>
</tr>
<tr>
<td>Spring</td>
<td>Channel</td>
<td>Grab 5</td>
<td>10.8</td>
<td>246</td>
<td>17</td>
<td>330</td>
<td>202</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Base</td>
<td>13.8</td>
<td>38</td>
<td>25</td>
<td>130</td>
<td>11</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Avg SSC</td>
<td>13.7</td>
<td>37</td>
<td>25</td>
<td>128</td>
<td>11</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Max SSC</td>
<td>13.7</td>
<td>37</td>
<td>25</td>
<td>127</td>
<td>10</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>5000 points</td>
<td>13.7</td>
<td>36</td>
<td>18</td>
<td>113</td>
<td>14</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>1000 points</td>
<td>12.3</td>
<td>75</td>
<td>17</td>
<td>147</td>
<td>52</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope Middle</td>
<td>Grab 7</td>
<td>13.1</td>
<td>255</td>
<td>10</td>
<td>375</td>
<td>203</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope North</td>
<td>Base</td>
<td>24.3</td>
<td>73</td>
<td>20</td>
<td>144</td>
<td>54</td>
</tr>
<tr>
<td>Spring</td>
<td>Slope North</td>
<td>Grab 5</td>
<td>8.7</td>
<td>213</td>
<td>25</td>
<td>282</td>
<td>184</td>
</tr>
<tr>
<td>Fall</td>
<td>Channel</td>
<td>Base</td>
<td>27.0</td>
<td>216</td>
<td>41</td>
<td>308</td>
<td>161</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Base</td>
<td>45.4</td>
<td>30</td>
<td>13</td>
<td>96</td>
<td>11</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Avg SSC</td>
<td>45.4</td>
<td>30</td>
<td>13</td>
<td>95</td>
<td>10</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Max SSC</td>
<td>45.3</td>
<td>30</td>
<td>13</td>
<td>95</td>
<td>10</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>5000 Points</td>
<td>47.7</td>
<td>32</td>
<td>11</td>
<td>94</td>
<td>14</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>1000 Points</td>
<td>47.9</td>
<td>46</td>
<td>7</td>
<td>92</td>
<td>31</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope Middle</td>
<td>Grab 8</td>
<td>32.7</td>
<td>203</td>
<td>24</td>
<td>296</td>
<td>149</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope North</td>
<td>Base</td>
<td>40.1</td>
<td>47</td>
<td>17</td>
<td>116</td>
<td>23</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope North</td>
<td>Grab 8</td>
<td>27.0</td>
<td>172</td>
<td>19</td>
<td>240</td>
<td>126</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope South</td>
<td>Base</td>
<td>45.5</td>
<td>40</td>
<td>24</td>
<td>118</td>
<td>12</td>
</tr>
<tr>
<td>Fall</td>
<td>Slope South</td>
<td>Grab 9</td>
<td>24.1</td>
<td>173</td>
<td>25</td>
<td>239</td>
<td>127</td>
</tr>
</tbody>
</table>
Figure 5.12: Comparisons between OBS and ADCP backscatter for SSC at Slope Middle station during spring deployment. In the top panel the deviation of the OBS signal from the ADCP signal starting around DOY 62 was caused by biofouling of the OBS. After that point the OBS data was discarded as unreliable.
Figure 5.13: Comparison of SSC from water column grab samples and ADCP backscatter. Time and location of grab sample collection is indicated in Figure 5.13.
5.2.3.3 Discussion of Error

The bottom two panels of Figure 5.11 show the estimates of error for the SSC measurements at the Slope Middle station during the spring deployment. The middle panel shows the standard deviation, $\sigma$, while the lower shows the error derived from the estimated standard error, $S^2$. Summarized data for other stations and deployments are listed in Table 5.4. The uncertainty associated with the SSC measurements is proportional to the SSC measurements. The high SSC measurements near the surface at days 61 and 62 also have large uncertainties. Some of the uncertainty is due to extrapolating the calibration relationships of Figure 5.10 to concentrations higher than those used in the calibration, which is the source of the errors shown in the lower panel of Figure 5.11, but some uncertainty is due to greater variability of the EI measurements within the 50 s averaging period. Those periods where uncharacteristically high estimates of SSC were made near the surface also exhibit greater variability in the EI measurements.

Percentage error from the error propagation analysis, computed using $1\sigma$ as the amount of error, vary between 10 – 400 % but averaged over the entire water column for each deployment the percentage errors are only 30 – 50 %. The percentage errors determined from the estimated standard error are smaller with averages ranging from 10 – 40 %.

The right hand side of the ADCP calibration equation (Equation 3.2) is comprised of four terms: the relative backscatter (Term I), beam spreading (Term II), water attenuation (Term III) and sediment attenuation (Term IV). The uncertainty in $S_v$ can be attributed to the uncertainties in each of these four terms. The contributions of these terms to the uncertainties of $S_v$ at the Slope Middle Station during the fall deployment are shown in Figure 5.14. Note that the scales are different for each plot. Uncertainty due to water attenuation (Term III) is an order of magnitude smaller than the other uncertainties. Near the transducers, roughly 2 m above the bottom in the plots of Figure 5.14, variation in the relative backscatter term (Figure 5.14a) is the largest source of uncertainty for $S_v$. As the distance from the transducer increases, uncertainty due to spreading of the acoustic beam increases and becomes the dominant source near the water surface. Uncertainty due to sediment attenuation (Term IV) is roughly a factor of 5 smaller than the uncertainty from the beam spreading, but shows the same dependence on distance from the transducer. At higher values of SSC sediment attenuation effects could become more important. The large amounts of uncertainty near the surface, particularly at days 61 and 62, are clearly attributable in part, to variations in EI, as shown in the uncertainties from Term I.

5.2.3.4 Other Calibration Cases

The other cases listed in Tables 5.3 represent an exploration of how the availability of data and assumptions made during the calibration affected the results. Grab samples represent the most direct measurement of SSC available for the calibration process. The low numbers of grab samples means that calibration could be easily skewed by an anomalous measurement or sampling schemes biased towards periods of low SSC. Calibration results from the grab samples are different from those determined from OBS data. In all cases where comparison was possible the use of grab samples resulted in lower average SSC concentrations when compared to the base calibration, in some instances by as much as a factor of 3. When grab samples are used for the calibration the uncertainties from the error propagation analysis increase to approximately 200%. The greater amount of uncertainty is due primarily to the small number of grab samples available for the calibration which greatly increases the uncertainty in the values of $a$ and $b$ in Equation 3.3.
Figure 5.14: Standard deviation contributed by Terms (a) I, (b) II, (c) III and (d) IV of Equation 3.2 for Slope Middle ADCP during spring deployment. All quantities are in dB.
The Slope Middle calibrations for both deployments were selected for a more detailed analysis of what may affect the calibration process. The last term in Equation 3.2 accounted for the attenuation of the acoustic signal due to absorption and scattering by particles within the water column, which is a function of the SSC. For the Base calibrations the SSC value appearing in the sediment attenuation term (see Equation A.7) was assumed to be a constant value representative of the reliable SSC data available. Signal attenuation from sediments is small when SSC $< 0.1$ kg m$^{-3}$ (Gartner, 2004; Wall et al., 2006), so this was thought to be a reasonable approximation. To test this assumption, two SSC times series were derived from the Base calibration results, a depth averaged SSC time series and a maximum SSC time series, and the calibration process repeated using these time series in sediment attenuation calculation. The effect of limited data availability was also examined by decreasing the number of data points used for the calibration process from the 10,000 points used in the Base calibration to 5,000 and 1,000 data points. The results of these analyses are listed in Table 5.3. Use of either the maximum SSC time series or the average SSC time series did not significantly change the SSC estimates. Decreasing the number of points used in the calibration similarly had little effect on the SSC estimates particularly when the uncertainty around these estimates is taken into account.

As a means of visualizing the difference between different calibrations the total sediment mass in the water column, in kg m$^{-2}$, was calculated (Figure 5.15). The variances for each measurement are additive over the water column. To remove the unreliable data discussed previously it was assumed that SSC should decrease with increasing distance from the sediment bed. In the upper water column, due to contamination of the data by bubble entrained by wind driven waves, if SSC measurements increased with greater distance from the bed, the SSC value from the bin directly below was substituted. In this case, however, the original variance was retained as it was greater and would thus better capture the uncertainties introduced by these substitutions. As the SSC for the bin closest to the transducer (bin 1) were consistently lower than those for the next higher bin, the values for bin 2 were substituted for bin 1. The variance for bin 1 was also set to the variance of bin 2 as the variance of bin 2 was greater. Error bars are $\pm 1\sigma$. Points which overlap the 1:1 line indicate that the calibration schemes are not different within the range of uncertainty of the measurements. Of the four calibrations compared to the base calibration, none of them produce values that can be said to be different except at the highest values.

5.2.3.5 Cross Calibration of ADCPs

During each deployment four ADCPs were used: three 1,200 kHz deployed in a line along the slope in roughly the same depth of water at a horizontal separation of 270 m, and one 600 kHz ADCP deployed in the channel 330 m from the Slope Middle ADCP (Figure 3.1). The SSC data and associated uncertainties were used to determine if the calibration relationship from a 1,200 kHz ADCP could be used at another 1,200 KHz ADCP for the same deployment. The data were also evaluated to determine if the calibration relationship developed for a particular instrument during one deployment could be applied to the same instrument during other deployment and whether calibrations developed for the 600 kHz instrument could be used on 1,200 kHz instruments during the same deployment. This was of particular interest given the noted difficulties in calibrating the Slope North ADCP during the spring deployment (bio-fouling of the OBS used for calibration) and the Channel ADCP during the fall deployment (loss of the OBS needed for the calibration).
Figure 5.15: Comparisons between the Slope Middle station spring deployment base calibration water column sediment mass and other calibration schemes for the Slope middle station spring deployment ADCP. Error bars are plus or minus one standard deviation. Units are kg m$^{-2}$. 

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94
Figure 5.16: Comparisons between base calibration total water column sediment mass for Slope Middle station during fall deployment and calibration relationships from other ADCP instruments. Error bars are plus or minus one standard deviation. Units are kg m$^{-2}$. 

95
ADCP data from the Slope Middle station during the fall deployment was selected for the analysis as it provided the richest data set. The results, as total water column sediment mass, are shown in Figure 5.16. The calibration developed from the spring deployment Slope Middle Station data under predicts the concentration when compared to the base fall deployment calibration at the same station. Both the Slope North and Slope South station calibration relationships from the fall deployment over predict the SSC when compared to the base calibration at the Slope Middle station during the fall deployment. Applying the Channel calibration relationship to the Slope Middle data also results in under prediction. Finally, the average of all the base calibration relationships from both deployments was applied to the fall deployment data from the Slope Middle Station. When compared to the base calibration results the SSC was under predicted.

**5.2.3.6 Discussion of ADCP calibration**

The results of these analyses indicate that a unique calibration relationship must be developed for each ADCP every time the instrument is deployed if SSC is to be estimated. Differences between 600 kHz and 1,200 kHz calibrations may be due to the different operating frequencies, which respond differently to particles of a given size, but could also be due to different sediment types found in the channel where the 600 kHz ADCP was deployed, and at the slope stations where the 1,200 kHz instruments were deployed. Results from the grain size analysis of the bed sediments, presented in Section 4.3, do not indicate any difference in deflocculated particle size distributions for sediments collected at the slope or channel stations thus refuting this hypothesis. Differences between the three 1,200 kHz calibrations during a given deployment could be due to differing sediment types but it seems unlikely that the sediments would differ significantly among three similar stations located less than 300 m apart. It is also possible that each instrument makes, receives and interprets the acoustic signals slightly differently or that other instrument factors not corrected for, such as the transmit power, are the cause of the differences. Likewise, calibration differences between deployments of the same instrument may be due to instrument factors not corrected for or may be due to sediment properties which differed between the deployments. However, data from the LISST deployed at the Benthic Station indicated that floc diameters were smaller during the fall deployment than during the spring deployment. As noted by Gartner (2004) a shift in the particle size distribution could affect the relationship between wavelength of the acoustic signal and the particles size requiring recalibration of the instrument. The particle size dependent response is a known limitation of measuring particles with single wavelength acoustic instruments.

Laboratory analyzed water samples are the most reliable method for determining SSC and have been used previously for ADCP calibrations (Wall et al., 2006). However many samples are required to obtain an accurate calibration. Care must be taken to collect samples covering the range of conditions which occur in the field. This may be difficult as the conditions likely to produce high SSC, maximum flood and ebb and periods with significant wave action, are the most difficult conditions under which to collect field samples. Collection of water samples during periods of low SSC is likely to be the cause of the difference between the OBS and water sample calibration relationships developed here.

More common is to use calibrated OBS to provide the required SSC data (Gartner, 2004; Hoitink and Hoekstra, 2005; Kim and Voulgaris, 2003). A reliable calibration of the OBS is still required which should be confirmed using *in situ* water samples, if only to establish the accuracy of a laboratory derived calibration. While less direct, this method is advantageous as data can be
obtained at a high frequency over the range of conditions which occur during the instrument deployment. Conditions which change over long periods of time, such as the general clearing of the water column during the less energetic part of the neap – spring tidal cycle, also need to be taken into consideration when developing calibration schemes. Bio-fouling of most of the OBS sensors precluded such an analysis here, but such changing conditions, particularly if they incorporate a change in particle size distribution, may affect calibration relationships.

5.2.3.7 Application
The SSC time series data derived from the ADCP calibrations were used to determine the sediment resuspension response due to the currents. The ADCPs were mounted above the bed and an interval above the ADCP transducer existed within which no measurements were made. Therefore no measurements of SSC were made by the ADCPs within the 1 to 2 m directly above the sediment bed. In those instances where the backscatter data was contaminated by air bubbles the constant value extrapolation scheme discussed previously was implemented. As a first order approximation, the water column measured by the ADCP can be analyzed as a one-dimensional problem by assuming an increase in the total sediment within the water column is the result of resuspension from the sediment bed. As discussed in Section 2.3.2, resuspension fluxes are typically proportional to the bed shear stress. Velocity measurements were also made by the ADCPs and as such also did not begin until 1 to 2 m above the sediment bed. Without near bed velocity measurements, determination of the bed shear stress must rely upon an assumed velocity profile. Rather than make assumptions about the near bed velocity structure this analysis relies upon the velocities measured by the ADCP. The data were separated into individual flood and ebb events starting from low or high slack water and ending at the following high or low slack water, resulting in roughly 6 hours of data for each event. The maximum and minimum total sediment mass suspended in the water column within each flood or ebb period was determined. In some cases the minimum sediment mass occurred just prior to slack water in which case that value was selected. The difference in mass suspended in the water column must have been resuspended from the bed under a 1-D analysis and should be related to the hydrodynamics. In this case the maximum water column mean velocity which occurred during each flood or ebb period was used to characterize the strength of each tide.

Results from each of the ADCP stations are plotted in Figure 5.17. The errors are 1\( \sigma \) as derived from the error analysis of Section 5.2.3.1. The data are reasonably fit by power law relationships as shown in Figure 5.17 for the combined spring and fall deployment data. The three slope stations respond similarly to the tidal forcing although the total sediment mass measured at the Slope Middle Station is less than measured at either the Slope North or Slope South Stations. This may be due in part to the spring deployment data at the Slope Middle station which shows smaller increases in sediment masses when compared to both the fall deployment data and to the data from the other two stations. Total suspended sediment mass at the Channel Station is lower than that observed at the slope stations despite higher maximum depth averaged current velocities. Particle size analysis of sediments collected from the bed surface did not indicate that this difference could be explained by differing particle sizes. Anecdotally, however, collection of sediment samples at the Channel station was more difficult as the Ponar sampler did not easily penetrate the sediment bed indicating perhaps a stronger, more consolidated sediment bed than at the other stations. Such a bed would be less responsive to strong tidal currents resulting in less material being resuspended into the water column.
Figure 5.17: Increase in water column sediment mass over a tidal cycle at the (a) Channel Station, (b) Slope Middle Station, (c) Slope North Station and (d) Slope South Station. Power law relationships are for the combined spring and fall deployment data. Error bars are one standard deviation.
Figure 5.17: Increase in water column sediment mass (continued). No data from the spring deployment was available at the Slope South Station.
To different degrees, for the three stations where data from the spring and fall is available, there are some differences between flood and ebb tides. At the Channel Station during the fall deployment, ebb tides, the mass of sediments in the water column is greater for a given maximum velocity when compared to the other data, which all tend to follow the same pattern. At the Slope Middle Station the data separate into the spring deployment and the fall deployment with the total mass of sediment greater during the fall deployment. The shorter record of data for the spring deployment may bias the results, however, as data from the greatest tidal range (and presumably velocity) are not available. The tight clustering of the spring data indicate the relationships for the spring and fall are different. There is no discernable difference between flood and ebb tides for either deployment. At the Slope North station there may be some difference between the spring and fall deployment but the spread in the data obscures any definite trend. At the Slope South Station the flood and ebb data are similar. At all of the slope stations there are some data points with much higher sediment masses than for other similar data points at a given velocity. These points have larger errors and correspond well to periods of high wind, indicating the data is likely contaminated by air bubbles generated by surface waves despite efforts to filter out these effects. The data do not appear to have a large affect on the power law relationships shown. The maximum sediment mass resuspended into the water column at all of the stations was on the order of 1 kg m⁻². Assuming a dry sediment density of 2,500 kg m⁻³ and a porosity of 90% the maximum eroded depth into the sediment bed would be about 4 mm.

5.2.3.8 Conclusions

ADCPs have been adopted for measuring suspended sediments in the water column at high frequency with good spatial resolution. Accuracy of the measurements is sacrificed for the ability to make measurements over most of the water column. The analysis presented here has shown that reasonable measurements can still be obtained with an average uncertainty of about fifty percent. Reduction of the uncertainty is dependent upon establishing a good calibration relationship that covers the entire range of sediment concentrations which occur during the instrument deployment. OBSs provided the best calibrations here but such methods are entirely dependent upon accurate conversion of the OBS signal into SSC. Water samples were less useful for calibration here because of the small number of samples collected and because many of the water samples were collected at times of low SSC. In many instances the sediment backscatter signal was obscured by air bubbles entrained into the water column by surface waves which generated anomalously high values of SSC. In such cases the SSC values had significantly higher variances than data representative of suspended sediments. This should make it possible to identify and exclude such data. There is no known method for removing such interference while retaining the suspended sediment acoustic signal. Such difficulties limit the use of ADCPs for measuring SSC near the surface where surface waves occur.

The data gathered during these field experiments were analyzed to determine the mass of sediments resuspended during the tidal cycles. The mass resuspended was shown to be proportional to the maximum depth averaged current velocity. Similar relationships were found at each of the slope stations although in some cases seasonal and tidal stage differences were evident. Total sediment masses in the channel were lower than those observed on the slopes. With the exception of the Channel Station during the fall deployment due to calibration difficulties, the scatter in the data between tidal cycles was greater than the uncertainties associated with any individual measurement.
6 1-D Sediment Bed – Water Column Model

Direct measurements of the sediment bed indicated that the active layer of sediments was no more than 1 cm in thickness. Interpretation of the water column sediment concentrations indicated that the active layer was probably even smaller but that the resolution at which the sediment cores were sampled was too coarse. Furthermore, only porosity of the sediment bed was measured directly. Shear strength of the sediment bed, particularly with depth and how it may change over time, is an important sediment bed property. Again these properties were estimated from the water column sediment concentration measurements, but these measurements were made 0.36 m above the sediment bed and did not allow for any variability over time.

To better understand the changes to the sediment bed over the course of the two instrument deployments a coupled one dimension sediment bed – water column model was developed. The model explicitly modeled the sediment bed including shear strength and allowed for analysis of the depth of active sediments to be estimated at a much finer scale. The model was calibrated for the Benthic Station using the sediment concentration and vertical turbulent sediment flux data measured by the two ADVs at that station.

6.1 Coupled Model

The model developed here utilized two existing models, a sediment bed model developed by Sanford (2008) and a water column model developed by Brand et al. (2011). Both were developed in MATLAB and the original sediment bed model code is available as part of that article. This work coupled the two models via the vertical sediment flux at the interface between the sediment bed and the water column. The code for the model used here is included in Appendix C.

The sediment bed model is an explicit, time-dependent, layered model which can incorporate the processes of erosion, deposition, consolidation and bioturbation or mixing. The model is a recent attempt to create a sediment bed model which incorporates all of the varied processes to which a sediment bed may be subject. The model was used by Sanford (2008) for three simulations based upon idealizations of experimental results from the Chesapeake Bay but no direct comparisons were made between modeled and measured values. The model was chosen because it represented a comprehensive attempt to model the sediment bed and because it was easily modifiable to be interfaced with a water column model. Adaptation of this model to the Bay required values for some of the model parameters to be determined based upon the experimental results of Chapter 5.

The water column model was developed as a different facet of the South Bay research project to which this work also belongs (Brand et al., 2011). It was developed as a standalone water column model but was intended to easily interface with a sediment bed model. The water column is modeled as a number of cells, which change height as the water depth changes with the tides. The water column model resolves two sediment fractions, a fast (coarse) settling sediment fraction and a slow (fine) settling sediment fraction. Sediments introduced into the bottom of the water column model due to erosion from the sediment bed model are split into the
fast and slow settling fractions by a factor \( \alpha \) (0 \( \leq \alpha \leq 1 \)), a fitting parameter, which identifies the fine fraction of sediments. The sediment bed model contained a number of cells which are filled and emptied by the processes of erosion and deposition. Only a single sediment fraction is modeled by the sediment bed model. The coarse and fine sediments which settle out of the water column model are combined to yield a single flux of sediments to the sediment bed surface. The processes of bioturbation and consolidation are also incorporated into the sediment bed model.

Modifications to the original codes, discussed below, were made to both models to facilitate the coupling of the models and to simplify the sediment bed model. The final version of the code used here is included as Appendix C. A conceptual diagram of the two models and how they interact is shown in Figure 6.1.

6.1.1 Sediment Bed Model

Two sediment fractions, nominally a silt and clay (sediment) fraction and a sand fraction, were resolved by the original model. For the development of this model, the sand fraction was removed entirely from the code. Similar results could be achieved with the original code by setting the sand fraction to zero. Those interested in the model treatment of the sand fraction are referred to the Sanford (2008) article. The original model was developed with a simple representation of the water column, a depth averaged cell into which particles could be eroded and out of which particles would settle, with the explicit acknowledgement that this could easily be replaced by a more complicated water column model. The model incorporated here to accomplish that is detailed in Section 6.1.2.

The sediment bed is modeled as a number of discreet sediment layers of user defined number and thickness. The layer thickness is defined on a mass basis, \( \Delta m \) (e.g. as 0.05 kg m\(^{-2}\) thick) rather than a vertical thickness (e.g. in meters) because during consolidation the vertical thickness of a layer may change but on a mass basis the thickness will remain constant. The critical shear stress for erosion, \( \tau_c \) [Pa], and the solid fraction, \( \phi_s \) [-], (equal to one minus the porosity) are defined at the top of each layer and are assumed to change linearly across the layers. An initial and an equilibrium \( \tau_c \) profile needs to be defined by the user. As developed here both profiles take the form of:

\[
\tau_c(m) = \tau_{c1} + \tau_{c2} \left( \frac{m}{1 \text{ kg m}^{-2}} \right)^{c_3}
\]  

where \( \tau_{c1} \) [Pa], \( \tau_{c2} \) [Pa] and \( c_3 \) [-] need to be defined by the user for both the equilibrium, \( \tau_{c,eq} \) [Pa], and initial \( \tau_{c0} \) [Pa], profiles and \( m \) is the depth into the sediment bed in mass coordinates. The original model set the initial critical shear stress profile equal to the equilibrium profile. The initial critical shear stress profile was incorporated into the model to better match the initial conditions encountered in the experimental data.

In modeling consolidation, the process was simplified significantly by assuming first-order relaxation towards the equilibrium state. As noted by Sanford (2008) there is no physical basis for this implementation. It may, however, be used as a first approximation for a process which is not currently well understood. The time rate of change of \( \tau_c \) due to consolidation and swelling is modeled as:

\[
\frac{\partial \tau_c}{\partial t} = r_c \left( \tau_{c,eq} - \tau_c \right) H_f \left( \tau_{c,eq} - \tau_c \right) - r_s \left( \tau_{c,eq} - \tau_c \right) H_f \left( \tau_c - \tau_{c,eq} \right)
\]  

102
Figure 6.1: Schematic showing the coupling between the water column and sediment bed models. Each time step involves running through the water column model and then the sediment bed model. The operations within a time step for each model are shown as they occur moving from left to right. Black arrows indicate sediments entering or leaving each model. White arrows show the processes in each model and the values passed on at the end of each time step.
where $r_c [s^{-1}]$ and $r_s [s^{-1}]$ are the consolidation and swelling coefficients, respectively, and $H_f$ is the Heaviside function defined such that $H_f = 1$ when its argument is $\geq 0$ and zero otherwise. Swelling is included to model the effects of increased erodibility due to wave disturbances, or softening, of the sediments. Both coefficients can be calibration parameters but the default of the swelling coefficient is kept at $10^3$ times smaller than the consolidation coefficient. The time rate of change of the solid fraction is similarly modeled by Equation 6.2, replacing $\tau_c$ with $\phi_s$, except within the Heaviside functions which remain unchanged.

Mixing due to bioturbation is modeled as a diffusion process whereby an amount of sediment mass and its associated properties (e.g. sand fraction and $\tau_c$) are exchanged with the layer above and below. To minimize numerical dispersion due to frequent mixing of small amounts of mass at each time step, mixing is delayed until a threshold of mass to be mixed is exceeded. For the critical shear stress of layer $j$, the change in $\tau_c$ for that layer, $\Delta \tau_{c,j}$, is calculated as:

$$\Delta \tau_{c,j} = \frac{D_m \Delta t'}{\Delta m} \left( \frac{\partial \tau_{c,j}}{\partial m} - \frac{\partial \tau_{c,j-1}}{\partial m} \right) \quad (6.3)$$

where $D_m$ is the bioturbation coefficient with a model default of $3.1 \times 10^{-12} \text{ m}^2 \text{ d}^{-1}$. As mixing does not occur at every model time step, $\Delta t'$ is the total time passed since mixing last occurred. The quantity within the parenthesis is the difference in the gradients of critical shear stress as a function of mass coordinate in layer $j$ and the layer above it, $j-1$. Equation 6.3 is a parameterization of a process which is difficult to quantify. Both bioturbation and the diffusive process used in Equation 6.3 will result in homogenization of the sediment bed, and associated properties, over time. In the numerical code, consolidation via Equation 6.2 and bioturbation via Equation 6.3, are accounted for at the same time. If no bioturbation is required only consolidation occurs.

The interface layer is the layer which interacts with the water column, gaining or losing sediment mass due to net deposition or net erosion, respectively. Over time deposition or erosion will fill entirely or exhaust the interface sediment layer. When the mass in the interface layer exceeds a defined threshold the layer is split into a filled sediment layer overlain by the new interface layer containing the remaining mass. In the case of erosion the mass from the now mostly empty interface layer is combined with the next lower layer and the interface again redefined.

The erosion rate of sediments from the bed, $E_s [\text{kg m}^{-2} \text{ s}^{-1}]$, is modeled by:

$$E_s = \left[ \frac{\Delta}{\alpha} (1-b) H_f(A) + b M (\tau_{b0} - \tau_{c0}) \right] H_f(\tau_{b0} - \tau_{c0}) \quad (6.4)$$

$$b = \frac{1 - \exp(-\alpha M \Delta t)}{\alpha M \Delta t} \quad (6.5)$$

and $M = \rho_p \beta_c [\text{s m}^{-1}]$ where $\rho_p [\text{kg m}^{-3}]$ is the sediment dry density and $\beta_c$ is a mass transfer coefficient assumed constant and set empirically. The default value of $\beta = 1.36 \times 10^{-4} \text{ m s}^{-1} \text{ Pa}^{-1}$ was retained. In Equation 6.4 $\tau_{b0} [\text{Pa}]$ and $\tau_{c0} [\text{Pa}]$ are the bed shear stress and the critical shear stress for erosion at the sediment water – interface, respectively, at the beginning of the model time step. The time rate of change of the bed shear stress is assumed to be constant over a time
step, appearing in Equation 6.4 as \( A = \frac{d\tau}{dt} \). In Equations 6.4 and 6.5 \( \alpha = \frac{d\tau_c}{dm} \) is the derivative of the critical shear stress for erosion with respect to the sediment bed mass and is calculated at each time step. Equation 6.4 is derived from Equation 2.35, modified to allow for both limited sediment supply (Type I) and unlimited sediment supply (Type II) erosion with the type of erosion determined by the parameter \( b \) (Sanford, 2008; Sanford and Maa, 2001). If \( \alpha M \Delta t \) is very large \( b \rightarrow 0 \), and assuming that both Heaviside functions have values of 1, Equation 6.4 simplifies to \( E_s = A/\alpha \), the increase in the bed shear stress with time \( (A) \) is balanced by the increase in the bed shear stress into the sediment bed \( (\alpha) \), or Type I erosion. If \( \alpha M \Delta t \) is very small, \( b \rightarrow 1 \) and Equation 6.4 can be simplified to:

\[
E_s = M(\tau_{b0} - \tau_{c0})
\]  

(6.6)

if the Heaviside functions are equal to one. Equation 6.6 is similar to Equation 2.35 except that this parameterization depends upon the initial bed shear stress and critical shear stress for erosion. For small time steps this is a reasonable approximation.

The final step in the model code is to calculate the rates of deposition and erosion. For the sediment model, deposition \( (D_s) \) is an input from the water column model as a mass per area per model time step. \( E_s \) is calculated via Equation 6.4 and the net deposition rate to the bed, \( D_s - E_s \), is calculated. If \( E_s > D_s \), the net deposition rate would be negative indicating erosion occurred during the time step. The calculated value of \( E_s \) is passed to the water column model as flux of sediments into the water column. Note that \( E_s \) is always positive or zero. The net deposition rate from this time step, determined by \( \Delta t \cdot (D_s - E_s) \), is used at the start of the next sediment model time step as the amount of sediment added to or lost by the interface layer.

### 6.1.2 Water Column Model

Sediment concentration, \( C \) [kg m\(^{-3}\)] was modeled using a one dimensional sediment transport equation for a tidal system:

\[
\frac{\partial C}{\partial t} = -\frac{\partial}{\partial z}(w + w_s)C + \frac{\partial}{\partial z} \left( \varepsilon_s \frac{\partial C}{\partial z} \right) + S_{lat}
\]  

(6.7)

with \( z \) [m] as the vertical coordinate, defined positive upwards. \( w_s \) [m s\(^{-1}\)] is the particle settling velocity and \( w \) [m s\(^{-1}\)] is a vertical velocity due to tidal changes in the water column depth, but is small enough to be neglected here. \( \varepsilon_s \) [m\(^2\) s\(^{-1}\)] is the vertical turbulent sediment diffusivity, often related to the eddy viscosity, \( \nu_T \) [m\(^2\) s\(^{-1}\)], by the turbulent Schmidt number \( Sc_T = \nu_T / \varepsilon_s \). The eddy viscosity is calculated using the parabolic expression for unstratified open channel flow:

\[
\nu_T = \kappa u_* H \left( \frac{z}{H} \right) \left( 1 - \frac{z}{H} \right)
\]  

(6.8)

where \( u_* \) [m s\(^{-1}\)] is the shear velocity, \( \kappa \) is von Karman’s constant and \( H \) [m] is the water depth. A turbulent Schmidt number of one indicates sediment turbulent mixing follows exactly the turbulent eddy mixing while \( Sc_T \) greater than one indicates less efficient vertical mixing of sediments when compared to the vertical mixing of momentum.

In Equation 6.7, \( S_{lat} \) is a source term that arises from the method used to handle changes of the water depth due to tides in the model. The water column is modeled by a user defined number of finite volume cells. Rather than adding or subtracting cells with changes in water
depth, the height of each cell is changed to accommodate the tidal influence. The stretching of cells increases the volume. Since sediment mass and cell volume are tracked separately in the model, as the volume of the cell is increased the mass of sediments in that cell must also be increased if the volume concentration of sediments is to remain constant. The opposite process is required as cells are compressed due to a decreasing water depth. The change in mass in cell \( j \) may be approximated as \( \Delta M_j \approx C_j \Delta h \) where \( \Delta h \) is the change in the volume over one model time step. The required addition or removal of sediments is accounted for in Equation 6.7 as \( S_{lat} \).

Details on the calculation of \( \Delta h_j \) and the grid adaptation algorithm can be found in (Brand et al., 2011).

The first order finite volume formulation of Equation 6.7 for each cell is:

\[
\frac{\partial (h C_j)}{\partial t} = \left( \varepsilon_{s,u} \frac{\partial C_j}{\partial z} \bigg|_u + w_s C_j \bigg|_u \right) - \left( \varepsilon_{s,l} \frac{\partial C_j}{\partial z} \bigg|_l + w_s C_j \bigg|_l \right) + \int_h S_{lat,j} dz \tag{6.9}
\]

The subscripts \( u \) and \( l \) refer to the upper and lower boundary of the cell. As \( C \) is located at the center of the volume represented by the cell, values and gradients of \( C \) are linearly interpolated to the cell boundaries. The top boundary condition at the water surface is a no flux condition. The bottom boundary condition at the sediment – water interface has been modified here to incorporate the erosive flux, \( E_s \), from the sediment bed model as an input and to export the settling flux out of lowest cell as the depositional flux to the sediment bed model. The settling flux is \( D_s = w_s C_1 \) [kg m\(^{-2}\) s\(^{-1}\)] where \( C_1 \) is the concentration in the lowest cell. The boundary condition at the sediment – water interface is \( E_s - D_s \). Initial conditions for sediment concentrations in the water column were initialized using Rouse profiles based on SSC measurements at 0.36 meter above the bed.

The water column model incorporates two sediment fractions, differentiated by their settling velocities, nominally a fast settling (coarse) sediment fraction and a slow settling (fine) sediment fraction. The flux into the water column from the sediment bed model is split between the two fractions by the parameter \( \alpha_f \), the fraction of sediment defined as fine (slow settling) relative to the total amount of sediment. The fraction of coarse (fast settling) sediments is \( 1 - \alpha_f \). Note that the sediment bed model tracks only total sediment mass and does not differentiate between the two sediment fractions. Both settling velocities and \( \alpha_f \) are user defined parameters and serve as fitting parameters. After the two sediment fractions are introduced at the bottom boundary as the erosion flux they are independent of one another in the water column model.

### 6.1.3 Parameterization and Calibration

Each model required a number of input parameters. Some are time dependent parameters derived from the experimental data, some are assumed to be constant and others are fitting parameters. Input parameters, values and sources are listed in Table 6.1. The parameter values arrived at by Brand et al. (2011), which modeled the same set of data but parameterized the sediment flux into the model due to erosion differently (see Section 2.5.4), provided a good initial values for the water column model parameterization. Most of the best fit values from those results were retained for this model because it reduced the number of variables requiring calibration. The exception was \( w_{s,slow} \), for which Brand et al. (2011) found values of \( 1.0 \times 10^{-7} \) m s\(^{-1}\) and \( 1.1 \times 10^{-5} \) m s\(^{-1}\) for the spring and fall deployments, respectively. In the experimental data it was observed that the minimum sediment concentration achieved each day changes over the course of the deployment (on the timescale of several days) in response to the high and low
Table 6.1: Input parameters for the sediment and water column models. Fitting parameters are bolded.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Units</th>
<th>Spring Deployment</th>
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<td>0.46</td>
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</tr>
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</tr>
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<td>0.1</td>
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<sup>*</sup>After DOY 271.7 a 100 s time step was unstable requiring a 10 s time step to model thereafter.
<sup>†</sup>Brand et al. (2011) provided values for the slow settling velocity as well but there were found to not represent the data well so w<sub>s,slow</sub> was used as a fitting parameter.
<sup>‡</sup>Only total SSC was measured. Division between the coarse and fine fractions was estimated based upon the perceived background concentration attributed to the fine fraction.
<sup>¶</sup>The critical shear stress profiles are defined according to Equation 6.1.
energy phases of the neap–spring cycle (Figure 4.1f and 4.2f). With the above settling velocity values it was found that the long term changes in sediment concentration were not accurately captured. The values shown in Table 6.1 better captured the slow decrease in SSC.

The parameters for the initial \( \tau_c \) profile were set to be similar to the equilibrium profile parameters, particularly at depth. To accommodate recent erosion or deposition events the parameters were adjusted to strengthen or weaken the near surface sediments. The parameters were set to match the sediment behavior during the first tidal cycle, thus setting the initial conditions for the sediment bed.

The main fitting parameters were the consolidation coefficient, \( r_c \), the minimum critical shear stress for erosion, \( \tau_{c, \text{min}} \), and \( \tau_{c1} \), \( \tau_{c2} \) and \( c_3 \) for the equilibrium critical shear stress for erosion profile. Model predictions were compared against four parameters measured during the field experiments: sediment concentration and turbulent vertical sediment flux, \( \vec{w} \vec{C} \), at 0.36 and 0.72 meters above the bed (mab). Evaluation of the model predictions was based upon the mean square error (MSE):

\[
MSE(f, x) = \langle (f_i - x_i)^2 \rangle \tag{6.10}
\]

where \( x_i \) is the measured value and \( f_i \) is the model value. The angle brackets indicate averaging over the entire period being modeled. Two metrics based upon the MSE were used for comparing different model runs, the root mean square error (RMSE) and the mean square error skill score (SS):

\[
RMSE(f, x) = \sqrt{\langle (f_i - x_i)^2 \rangle} \tag{6.11}
\]

\[
SS = 1 - \frac{MSE(f, x)}{MSE(g, x)} \tag{6.12}
\]

where \( g \) in Equation 6.12 is the predicted value of a reference model (Murphy and Epstein, 1989). The reference model used here is the average value of \( x_i \) over the observation period, \( \langle x_i \rangle \). The skill score is a measure of the improvement in predicting the chosen parameter (e.g. sediment concentration) via the model when compared to the prediction made by the reference model (e.g. the average sediment concentration measured during the period being modeled). Positive values for the SS indicate an improvement upon the prediction made by the reference model and \( SS = 1 \) is a perfect prediction of measured data.

Concentration at 0.36 mab was the primary parameter upon which the calibration was focused, followed by the concentration at 0.72 mab. Less weight was put upon reproducing the turbulent vertical sediment fluxes at either elevation but reasonable agreement between the measured and modeled values was a model fitting criterion. An automated parameter fitting routine based upon decreasing the value of RMSE or increasing the value of the SS using a non-linear least-squares minimization routine in MATLAB was unsuccessful in finding reasonable model predictions. The routine easily found the local best model fit based upon the initial parameter values but it was not clear that such routines returned the global best model fit. Manual calibration of the model allowed the parameter space to be better explored to identify the best model fit. It was observed that different combinations of the calibration parameters could generate similar model output. This may have been due in part to the sensitivity of the model the calibration parameters.
Computationally, for a 1.83 GHz processor with 1.0 GB of RAM a 20 day model run with a 100 second time step took about 5 minutes. With a 10 second time step a 26 day model run took about 45 minutes. Using the 100 second time step 100 – 200 model runs were completed overnight and the results analyzed the following day. Based upon those results a new parameter space was defined and the process repeated. A wide range of values for the parameters being calibrated were explored in this manner with the objective that the best calibration found is the global best calibration. The longer run times for the 10 second time step made it more difficult to complete numerous model runs in an efficient manner. The short time step was required for modeling the fall deployment after DOY 271.7 for reasons which will be discussed in Section 6.4. The approach taken was to calibrate the model to the first 20 days of the fall deployment using the 100 s time step and then to calibrate the model for the entire 26 day fall deployment record using the 10 s time step with the calibration values from the earlier runs as the starting point for calibration of the entire 26 day model.

6.2 Model Results

Modeled values of SSC and turbulent vertical sediment flux are compared against the values measured by the ADVs during the instrument deployment as a means of validating the model. During much of this discussion the fall deployment will be split into results prior to DOY 270 and post DOY 270. The separation was necessitated by the numerical scheme of the code but is also useful since the latter period experienced significantly greater sediment concentrations and fluxes.

Comparisons of SSC and turbulent vertical sediment flux for the spring deployment are shown in Figure 6.2. In general the modeled values agree with the measured values, both for SSC and for the turbulent vertical sediment flux, although there is scatter around the 1:1 line for both quantities. At the highest SSC values the modeled values are consistently smaller than the measured values at 0.36 and 0.72 mab (Figures 6.2a-b). The fluxes appear to be well reproduced with some scatter around the 1:1 line (Figures 6.2c-d). At the lowest values of the turbulent vertical sediment flux the modeled values are smaller by 1 – 2 orders of magnitude but are closer to the measured values when the fluxes are greater than $10^{-6}$ kg m$^{-2}$ s$^{-1}$. Similar results were attained for the fall deployment models (Figure 6.3), noting that these results are only for data prior to DOY 270. For the fall deployment there is less scatter in the data overall for both SSC and turbulent vertical sediment flux, however the model values for SSC are still small at the highest concentrations.

Time series of SSC for the spring and fall deployments are shown in Figures 6.4 and 6.5, respectively. Only time series from 0.36 mab are shown for both deployments as the data for 0.72 mab was quite similar. In Figures 6.4 and 6.5 the first panel, (a), shows the total modeled sediment concentration in black, which is comprised of the two modeled sediment fractions: a coarse, fast settling fraction and a fine, slow settling fraction, as shown in panels (b) and (c), respectively. Also shown in panel (a) of each figure is the measured SSC in red for comparison. Comparing the modeled and measured time series, the modeled SSC peak during each tidal cycle is smaller than was measured. This was also evident from Figures 6.2 and 6.3 but the time series show that it occurs at many different concentrations, not just the highest SSC values. At other periods, around DOY 66 in Figure 6.4 for example, the SSC peaks in terms of shape and magnitude agrees quite well with the SSC measurements.

Considering the two modeled sediment fractions it is clear in both deployments that most of the variation in SSC on the tidal timescale is due to variation in the coarse sediment fraction.
Figure 6.2: Comparisons of the model results to the measured values for the spring deployment. Values compared are (a) SSC at 0.365 mab, (b) SSC at 0.725 mab, (c) vertical sediment flux at 0.365 mab and (d) vertical sediment flux at 0.725 mab.
Figure 6.3: Comparisons of the model results to the measured values for the fall deployment. Values compared are (a) SSC at 0.365 mab, (b) SSC at 0.725 mab, (c) vertical sediment flux at 0.365 mab and (d) vertical sediment flux at 0.725 mab.
Figure 6.4: Modeled values of SSC for the spring deployment at 0.365 mab as (a) modeled total SSC (black), measured SSC (red), (b) coarse sediment fraction and (c) fine sediment fraction (c). Note the vertical scale for (c) is different.
Figure 6.5: Modeled values of SSC for the fall deployment at 0.365 mab as (a) modeled total SSC (black), measured SSC (red), (b) coarse sediment fraction and (c) fine sediment fraction. Note the vertical scale for (c) is different.
Table 6.2: Metrics of fit for model predictions compared to observational data.

<table>
<thead>
<tr>
<th>Deployment</th>
<th>Concentration at 0.365 mab</th>
<th>Concentration at 0.725 mab</th>
<th>Vertical sediment flux at 0.365 mab</th>
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<td>0.01 kg m^-3</td>
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<td>7.0×10^-6 kg m^-2 s^-1</td>
</tr>
<tr>
<td>Fall (DOY 251.7-277.8)</td>
<td>0.11 kg m^-3</td>
<td>0.08 kg m^-3</td>
<td>5.8×10^-5 kg m^-2 s^-1</td>
<td>4.1×10^-5 kg m^-2 s^-1</td>
</tr>
</tbody>
</table>

Table 6.2: Metrics of fit for model predictions compared to observational data.

(Figures 6.4b and 6.5b). During both deployments the fine sediment fraction is important for determining the total SSC and its contribution changes over the period of several days with the neap – spring tidal cycle (Figures 6.4c and 6.5c). For the spring deployment these results differ from those of Brand et al. (2011) which modeled the fine sediment fraction as being nearly constant over the entire deployment. It was found that the variation in the fine fraction of the SSC over the course of several days was important for reproducing the measured sediment behavior on the timescale of the neap – spring tidal cycle. The fine sediment fraction settling velocity was used as a fitting parameter when calibrating this model for both deployments. The settling velocity fitted for the spring deployment (Table 6.1) is an order of magnitude greater than found by Brand et al. (2011) and much closer to the value found for the fall deployment. For the fall deployment the settling velocity for the fine fraction determined here was within a factor of 2 of that found by Brand et al. (2011). During the low energy part of the tidal cycle (DOY 62 – 66 during the spring and after DOY 267.5 during the fall) the SSC is driven almost entirely by the fine sediment fraction with only very small contributions by the coarse fraction over the course of a tidal cycle. In general this model reproduces the low overall SSC during this period but the lack of coarse sediment fraction contribution neglects entirely the small but measureable changes in SSC which occur on the tidal timescale during the low energy periods. The root mean square error (RMSE, Equation 6.11) and Skill Score (SS, Equation 6.12) values for the spring and fall deployments are listed in Table 6.2. The RMSE and SS values are comparable for all parameters between the two deployments. The values are also similar to those of Brand et al. (2011) when the model was forced by a bed shear stress dependent sediment flux at the bottom boundary (Table 4 in Brand et al. (2011)).

Higher sediment fluxes and SSC during the fall deployment after DOY 270 necessitated a smaller time step when modeling that part of the deployment. Comparisons between the modeled
Figure 6.6: Comparisons of the model results to the measured values for the fall deployment including the storm of DOY 270 and after. Values compared are (a) SSC at 0.365 mab and (b) the vertical turbulent sediment flux at 0.365 mab.
and measured SSC and turbulent vertical sediment fluxes for the entire fall deployment at 0.36 mab are shown in Figure 6.6. There is a clear deviation of the model results from the measured values during part of the deployment but the model does reproduce some of the higher sediment concentrations accurately. The source of significant deviation of the model results from the measured values noted in Figure 6.6 is evident in the time series of SSC for DOY 270 and after (Figure 6.7). Considered alone, the course sediment fraction predicts the total SSC accurately and generally reproduces the highest SSC values and the shape of the peaks, though it would over predict the SSC during the latter half of each of the two high concentration events. The fine sediment fraction contributes a significant amount of suspended sediment to the total SSC, up to 20% during each event. As the fine sediment fraction only settles out over the course of several days the total SSC does not decrease as quickly in the model as did the SCC in the field, hence the over predictions after the initial erosion event. Possible causes for this behavior are discussed in Section 6.5.

These differences between the modeled and measured values for the data after DOY 270 cause the metrics of fit for the model to worsen significantly (Table 6.2). All of the RMSE values for the entire fall deployment are an order of magnitude larger when compared to the values when the data after DOY 270 are excluded. Most of the SS are also negative indicating the reference model of an average value for the SSC over the entire deployment performs better than the model presented here.

6.3 Analysis

The primary interest of this model was to better understand what was happening with the sediment bed during the field deployments. Modeling of the water column was incorporated because that was the location of most of the field measurements against which the model could be calibrated. From the modeling results the depth of the active sediment layer during each deployment and the sediment erosion relationships were determined. The model results were also analyzed to determine if the erodible sediment supply is limited by the sediment shear strength.

6.3.1 Active Sediment Layer

The depth of the active sediment layer is defined here as the depth of sediments, which are actively being reworked by deposition and erosion over the tidal to neap – spring cycle time scales of the field experiments. It is the active sediments which contribute to the SSC in the water column and are responsible for the transport of sediment bound nutrients and contaminants between the sediment bed and water column.

The position of the sediment – water column interface is tracked in the model over time in mass coordinates. The position of the interface is shown in Figure 6.8a for the spring deployment and Figure 6.9a for the fall deployment. For these plots the initial position was set to zero. Deposition causes the position of the interface to become more positive and erosion causes the position of the interface to become more negative. Fluctuations in the interface position during a tidal cycle within the spring deployment are on the order of 0.01 kg m\(^{-2}\) with the largest events eroding 0.1 kg m\(^{-2}\) of sediment. During the fall deployment erosion is roughly a factor of two larger prior to DOY 270. Erosion events after DOY 270 during the fall deployment decreased the position of the interface by nearly 1.5 kg m\(^{-2}\), though the model predicts most of these sediments were redeposit within a day.

Discussion of an active sediment layer is more meaningful in length coordinates rather than mass coordinates but the unit conversion requires the porosity, \(\phi\), of the modeled sediment
Figure 6.7: Modeled values of SSC at 0.365 mab for the fall deployment after DOY 270 during which a storm occurred. Shown in panel (a) is the modeled SSC (black) and measured SSC (red). Total SSC was modeled as (b) a coarse fraction and (c) a fine fraction. Note the vertical scales for (b) and (c) are different.
Figure 6.8: Change in the sediment bed – water column interface location over time during the spring deployment. The interface position was set to zero at the beginning of the model run. Shown are (a) the mass coordinates used in the model and (b) depth coordinates estimated from the mass coordinates.
Figure 6.9: Change in the sediment bed – water column interface location over time during the fall deployment. The interface position was set to zero at the beginning of the model run. Shown are (a) the model results in mass coordinates and (b) depth coordinates estimated from the mass coordinates.
bed as a function of depth to be known or otherwise determined. In the Sanford (2008) sediment bed model the solid fraction is estimated from a relationship between critical bed shear stress, \( \tau_c \) [Pa], and the solid fraction, \( \phi_s = 1 - \phi \), determined by Sanford and Maa (2001) from measurements made in Baltimore Harbor, Maryland. Lacking high resolution porosity profiles from the San Francisco Bay, particularly near the sediment surface, the same relationship:

\[
\phi_s = \left( \frac{\tau_c}{0.3} \right)^{0.93}
\]

was retained for this work as well. Using Equation 6.13 to convert from the mass coordinates to depth coordinates the position of the sediment bed – water column interface in mm is shown in Figures 6.8b and 6.9b. During most of the field experiments during both deployments the depth of the active layer was in the 3 – 5 mm range. During the fall deployment after DOY 270 this value increased to almost 10 mm, although much of that is due to the deposition of highly porous sediments. The eroded depth during that period, relative to the original position of the sediment interface at the beginning of the deployment was only about 5 mm.

6.3.2 Sediment Erosion Relationships

The erosion rate in the model is determined as a function of the bed shear stress and the critical shear stress for erosion, \( \tau_c \), of the upper most sediment layer. This allows the sediment erosion rate to be greater for recently deposited sediments with little shear strength, or smaller for more consolidated sediments with significant shear strength. The modeled erosion fluxes as a function of bed shear stress are shown in Figure 6.10. On each plot three fits to the data are shown: a fit to the model results assuming no critical shear stress for erosion, a fit to the model results assuming a minimum critical shear stress for erosion, \( \tau_{c,min} = 0.07 \text{ N m}^{-2} \) (spring) or \( \tau_{c,min} = 0.08 \text{ N m}^{-2} \) (fall), and, for comparison, the fit developed for all of the experimental data (the All data relationship of Section 5.1.2 shown in Figure 6.10 as the fit to data relationship). The \( \tau_{c,min} \) values assumed for the second fit to the model results are the \( \tau_{c,min} \) values determined as part of the model calibration but for practical applications there is no difference between the two critical shear stresses for erosion. For the relationships with a specified \( \tau_c \) the erosion rate equation takes the form of \( E_s = M \tau_b^n \) for \( \tau_b > \tau_{c,min} \) but is zero otherwise. With this parameterization \( E_s \) could be calculated for \( \tau_b < \tau_{c,min} \) as shown in Figure 6.10, but is forced to zero under those circumstances. The same relationship was used in the erosion parameter analysis based on the experimental data in Section 5.1.2. The values for \( M \) and \( n \) for all of the fits are listed in Table 6.3.

In all of the plots shown in Figure 6.10 there is a distinct change in the slope of the data which occurs when the bed shear stress, \( \tau_b \), is equal to the minimum critical shear stress for erosion and corresponds to an erosion flux of roughly \( 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1} \) in all cases. The inclusion of \( \tau_{c,min} \) in modeling the erosion behavior assumes erosion fluxes of that order of magnitude and smaller are negligible and ignores them when determining the fitting parameters on the rest of the data. The incorporation of \( \tau_{c,min} \) is most significant for the spring deployment where the value of the exponent in the erosion rate equation, \( n \), increases from 1.1 to 2. The increase in \( n \) is the cause of the significantly different slopes of the erosion relationships in Figure 6.10a. The inclusion of \( \tau_{c,min} \) causes the \( R^2 \) value to decreases in all cases but this may be due to the exclusion of a significant amount of data points from the regression analysis. The equations which incorporated a value for \( \tau_{c,min} \) fit the data better at higher values of bed shear stress.
Figure 6.10: Erosive sediment fluxes determined from the model results for (a) the spring deployment, (b) the fall deployment prior to DOY 270 and (c) the entire fall deployment. In each plot two fits to the model results are shown: no critical shear stress for erosion (solid) and the critical shear stress for erosion fit in the model (dotted). The All data (dashed line) is the All data fit determined in Section 5.1.2 from the experimental data. Parameter values for the fits shown are listed in Table 6.3. See the text for the erosion model equation details.
Figure 6.10 (continued): Erosive sediment fluxes determined from the model results.

Table 6.3: Parameter values for the fits to the model results for the erosive sediment fluxes shown in Figure 6.10. The *Fit to data* relationship shown in each plot is the *All data* fit from the experimental data, the parameter values for which are listed in Table 5.1.
6.3.3 Erosion Type

Erosion behavior has been classified into two types, limited sediment supply (Type I) and unlimited supply (Type II). See Section 2.3.2 for further discussion of the erosion types. The contribution of each type to the erosion behavior in the sediment model is determined by the variable $b$, calculated by Equation 6.5, the value of which is shown in Figure 6.11 for the spring and fall deployments. During the spring deployment $b$ is always less than 0.12, indicating the erosion behavior is primarily supply limited. Erosion only occurs when the bed shear stress is increasing. Without continually increasing bed shear stress only a small amount of sediment is eroded before the bed shear strength of the newly exposed sediment matches that of the bed shear stress causing erosion to halt. Under these conditions $E_s$ goes to zero whenever the bed shear stress is less than or equal to the critical shear stress for erosion.

During the fall deployment the erosion behavior is influenced more by unlimited supply erosion, as indicated by the values of $b$ approaching unity. After DOY 270 the sediment erosion behavior is primarily Type II. This is not surprising as the bed shear stress during that time was much higher than the bed shear strength even deep into the sediment bed. Erosion during this period was limited only by the length of time the bed shear stress was applied and the rate at which mass could be transferred from the sediment bed into the water column. Under normal circumstances these results indicate a strong sediment bed relative to the typical (current driven) bed shear stress.

6.4 Discussion

The flux of sediments between the water column and sediment bed impact water quality in the Bay in many ways. A sediment bed model was coupled to a water column model in order to understand how the sediment bed was changing over the course of the instrument but this required added model complexity. Many other models have focused upon sediment concentrations within the water column of the San Francisco Bay, treating the sediment bed simplistically or as a parameterized bottom boundary condition for the water column model (see Section 2.5.4). The modeling results generated here indicate that the sediment bed is highly dynamic at small spatial scales and that sediment bed properties change over a wide range of time scales, from sub-tidal to at least seasonal.

For the conditions which occurred during the field experiments the depth of the active sediment bed was in the range of 1 – 10 mm. Millimeter scale depths of erosion during the tidal cycles have also been reported for the Chesapeake Bay (Sanford et al., 1991). Conversion of the mass coordinates used by the model into more useful depth coordinates incorporates some uncertainty into the active sediment layer depth because the relationship used was not derived based on measurements made in the San Francisco Bay estuary. Measurements of bed properties would be required at a much higher spatial resolution than was achieved during these instrument deployments to generate a relationship for the Bay. The reason for the small active layer depth in the Bay is not currently known. However, it could be explained by a strong, highly consolidated sediment bed. Such a bed would be difficult to erode and would help to explain the small SSC values measured with the Bay, particularly over the last decade.

Within the time scales and environmental conditions of the field measurements, bed sediments that are not located within the active layer are not likely to be reintroduced into the ecosystem. Many of the contaminants of concern within the Bay are currently buried beneath cleaner sediments. Knowing the depth of the active sediment layer allows for better understanding of the environmental conditions those sediments will be exposed to in the future.
Figure 6.11: Type of erosion behavior during the (a) spring and (b) fall deployments. The parameter $b$, calculated by Equation 6.6, modulated the type of erosion behavior modeled by Equation 6.5 in the sediment bed model. As $b \to 0$ erosion is controlled by the rate of increase of the bed shear stress balanced against the increase in bed shear strength into the bed (Type I or limited supply erosion). As $b \to 1$ erosion is proportional to the applied bed shear stress (Type II or unlimited supply erosion).
and helps to assess the risk the contaminants pose to the ecosystem. The models for the Bay which try to estimate the long term fate of contaminants within the sediments have assumed the active layer of sediments to be on the order of 10 cm (Davis, 2004; Greenfield and Davis, 2005; Macleod et al., 2005). Decreasing this value by a factor of 10 will affect the modeled fate of these sediment bound contaminants and the risk they pose to the Bay ecosystem.

Limitations of the sediment bed model became apparent when modeling the fall deployment data after DOY 271.7. The physical forcing during this time period created large sediment fluxes into the water column followed by large sediment fluxes to the bed. Limitations of the sediment bed model resulted in the loss of sediments from the model when the fluxes grew too large. This situation could be avoided by using a smaller time step, hence the use of the 10 second time step when modeling the fall deployment. Sanford (2008) in testing the model used a 5 second time step. The best solution would be to rewrite the sediment bed model to handle large fluxes of sediment to the bed in a single time step or to incorporate a dynamic time step.

The greatest difficulty in implementing the coupled model was the large number of parameters which needed to be calibrated. With the exception of the bed shear stress (or shear velocity) and the water depth, all of the parameters in Table 6.1 need to have values input or must be used for model calibration. Here some of the values were incorporated from (Brand et al., 2011) and were assumed to be correct. In most cases such information is not likely to be available. Other parameters were simply assumed to be the model defaults.

The forcing of a water column sediment model by a simple erosion – bed shear stress expression by Brand et al. (2011) produced SSC model results comparable to those produced here by the coupled model when compared to the SSC values measured during the field experiments. The added insight into the bed sediments was values for better understanding sediment behavior at the field site. However, uncertainty in many of the parameters used within the sediment bed model limits the precision of the modeled behavior. A better understanding of sediment bed properties and processes is required, particularly at fine scales, before the added complexity of a coupled sediment bed – water column model will be justified in most circumstances. Ultimately the user must decide whether the limitations and difficulties encountered are superseded by the addition information such a model can provide.
7 Conclusions

Sediments impact nearly every aspect of the estuarine system. They serve as the habitat for benthic organisms, have direct and indirect impacts upon water quality and are integral for the functioning marshes along estuarine margins. Deposition of sediments in shipping channels and harbors necessitates periodic dredging and the sediment bed may serve as a source or a sink for nutrients and anthropogenic contaminants. Beginning with the impacts of hydraulic mining in the 1850s anthropogenic activities have impacted the sediments in the San Francisco Bay estuary. The salt pond restoration projects currently under way will require large quantities of sediments to reestablish marshes along the Bay margins raising concerns that contaminated sediments currently buried will be exposed and reintroduced into the estuary ecosystem.

Sediment dynamics in the estuarine system is driven in part by the physical and chemical properties of the sediments themselves but is also substantially affected by the fluid mechanics in that system. Advances in instrumentation have made it possible to measure suspended sediment concentrations collocated with measurements of the fluid mechanics, allowing the effects of the latter to be integrated into our understanding of what determine the former. The deployment of a number of instruments in the field provided measurements over the span of several weeks coving a wide variety of environmental conditions. These instrument deployments provided the data needed to assess the current mechanistic models of sediment dynamics in the estuarine system.

This work has resulted in the quantification of several highly important sediment processes within the San Francisco Bay estuary. It was found that a simple flocculation model which took into account a time varying level of turbulence was insufficient for explaining the decreases in suspended sediment concentration measured in the field. Measurements of settling velocity determined from acoustic instruments produced very reasonable values and were easy to measure but under the conditions encountered at the field site, may not have been as reliable as those measurements made by direct observations of settling flocs. Sediment bed properties were found to be an integral part of the bed – water column sediment exchange dynamics but these properties are the most difficult to quantify at the level of detail required for this work. Values for many of the sediment processes developed from the experimental data were used in a sediment bed – water column model. The model estimated the depth of the active sediment layer to be on the order of 1 to 10 mm.

This was a first step towards gathering the required field data and testing our mechanistic understanding of estuarine sediment dynamics. Given the challenges facing estuaries in the near future, including sea level rise, wetland restoration and the emplacement and removal of dams among many others, our actions and management strategies need to be determined by mechanistic models based upon known processes and verified against field measurements. In some cases it was found that our current understanding is sufficient however in many cases considerable work is still required.

7.1 Recommendations

The process of analyzing the experimental data and the model results indicated several areas where further research is warranted. Both field and laboratory investigations are needed to
better understand sediment dynamics within the estuarine system. The results of this work also indicate there is a need to assess longer term sediment transport and to reevaluate the results of the contaminant fate models used in the San Francisco Bay estuary (see Section 2.5.4).

Much of the knowledge of particle flocculation came out of research for water treatment that is not directly applicable to estuarine particle dynamics. There is a lack of data regarding the flocculation of particles found under the conditions such as those found in the San Francisco Bay estuary. Turbulence found within the estuarine water column, the Bay generated fluid shear rates between 0 – 4 s\(^{-1}\), are approximately 2 order of magnitude smaller than used in water treatment. Additionally, almost all studies of particle flocculation use a constant value of turbulence for the duration of the experiment. In an estuary, however, the turbulence varies with time changing from zero to a maximum and back to zero over the span of several hours due to the tidal influence. Research into particle flocculation behavior under time varying amounts of turbulence has been limited to step changes in the turbulence at levels much greater than those found in estuaries. There is no experimental data for particle flocculation under slowly varying levels of turbulence such as would be found in an estuary. Such information is needed to accurately assess and model particle flocculation in the estuarine system.

Within the sediment bed, the centimeter scale of the field measurements made as a part of this work was insufficient to capture the millimeter scale changes to the sediment bed indicated by model. The model, in turn, depended upon a parameterization of the sediment bed at these fine scales. While relationships were developed during the calibration to model the sediment bed there is a clear need to measure the properties of the sediment bed at millimeter or finer resolution. Included in these measurements should be a determination of the decrease in porosity and the increase of shear strength with depth into the sediment bed. Such measurements likely do not need to be extended more than a few centimeters into the bed but it is not clear if the current instruments available are capable of making these measurements at such a fine scale.

Modification of the sediment bed by benthic organisms was not incorporated into this work. During the field experiments worm tubes were noted in some of the sediment cores collected. Such tubes often extended as much as 10 cm into the bed and often extended several cm into the water column. Benthic organisms undoubtedly affect the sediment bed, mixing sediments vertically, increasing porosity and redistributing nutrients (Montserrat et al., 2008; Soares and Sobral, 2009a; Soares and Sobral, 2009b). Quantification of these biologically driven processes is necessary for developing a complete understanding of sediment transport in the estuarine system but more work within the Bay is needed in this area.

There is also a clear need to extend the results of this work over a larger time frame. As the field experiments only covered seven weeks of the year it is difficult to assess how the sediment bed may change over the seasonal or yearly time scales. This becomes particularly important as it appears that periodic events may drive much of the sediment erosion and subsequent redistribution in the South Bay.

Finally, the models for contaminant fate and transport in the Bay need to be reevaluated. Most of those models used a simplistic treatment of the sediment bed and assumed an active layer of sediments an order of magnitude greater than determined by this study. The smaller active layer may decrease the reintroduction of contaminants buried more than a few centimeters into the sediment bed. At the same time this may also decrease the possibility that contaminated sediment will be flushed out of the Bay via the Golden Gate. However, as contaminants in the Bay are known to be impairing the ecosystem it is important to assess what effect the smaller active layer will have on contaminant fate.
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Appendix A

Development of the ADCP Calibration Equation

The theoretical basis for acoustic measurement of SSC is the active sonar equation (Urick, 1967):

\[ SL - 2TL + TS = NL - DI + DT \]  

\[ (A.1) \]

where SL is the source level, TL is the transmission loss, TS is the target strength, NL is the noise level, DI the directivity index and DT is the detection threshold. All quantities have units of dB. From left to right the three terms on the left hand side of the equation describe how a signal emitted at the source level (SL) is altered due to losses as the signal is transmitted through the water, both to the target and back to the receiver (2TL), and how the signal is altered as it is backscattered by the target (TS). The three terms on the right hand side (RHS) incorporate how well the returned signal is measured. NL includes both the background noise and the instrument noise. DI is a measure of how well an instrument can measure the signal and DT is the threshold at which a signal can be distinguished from the noise. For further discussion of these parameters see (Urick, 1967). Equation A.1 describes how an acoustic signal will be altered as it is transmitted through water, scattered off of objects in the water and then measured upon its return.

The translation of Equation A1.1 into parameters relevant to an ADCP was first presented by Deines (1999) as an equation for determining the backscatter coefficient, \( S_v \). Since then the development of the equation has been detailed by several authors, each including or disregarding correction terms as matches their needs and available data (Gartner, 2004; Hoitink and Hoekstra, 2005; Kim and Voulgaris, 2003; Wall et al., 2006). Development of the equation used for calculating \( S_v \) for this work is detailed below. The equation for \( S_v \), Equation 3.2, is reproduced below for convenience.

\[ S_v = 10 \log_{10} \left( \sum_{i=1}^{n} 10^{\frac{(EI(i)-ER(i))Kc}{10}} \right) + \psi 20 \log_{10}(R) + 2\alpha R + 2\alpha_s R \]  

\[ (3.2) \]

The first term on the RHS of Equation 3.2 is an average of the signal strength measured by the ADCP. EI(i) is the echo intensity in counts measured by each beam for each bin. Bins refer to the vertical intervals over which a measurement is taken. A count is the unit used by the ADCPs and is a reflection of the amount of gain applied to the input signal. Less gain is needed for a strong input signal than for a weak input signal. Note that in this discussion of counts the
strength of the input signal is unrelated to the strength of the acoustic signal measured by the
ADCP, but is rather a characteristic of the ADCP instrument itself. The conversion factor
between counts and dB is given by $K_c$ and is instrument and transducer specific.

Each transducer also has an echo intensity reference level, $E_r$ in counts, which is
subtracted from the measured echo intensity. $E_r$ may be determined in the lab (Deines, 1999) or
it may be taken as the minimum value measured over all bins for each transducer during the
deployment (Gartner, 2004). The latter was used for these calibrations. Typical values for $E_r$ are
40 counts (Deines, 1999). Each ADCP used in these experiments had $n = 4$ beams across which
the relative backscatter was averaged for each bin height.

The second term on the RHS of Equation 3.2 is a correction for the spreading of the
signal as it travels a distance $R$, in meters, through the water column. Each transducer sits at an
angle $\theta$ to the vertical such that a bin at a height $\delta$ above the transducer will have a path length,
$R$, of:

$$ R = \frac{\delta + D/4}{\cos(\theta)} \quad \text{A.2} $$

where $D$ is the bin height in meters and is divided by 4 because the ADCP only samples the
upper quarter of each bin (Deines, 1999).

The correction $\psi$ in the second term on the RHS of Equation 3.2 corrects for different
spreading behavior of the acoustic signal near the transducer and is a function of the ADCP
transducer radius, $a_i$ in meters, and the acoustic wavelength of the ADCP, $\lambda$ in meters. The
acoustic wavelength is calculated as $\lambda = v/f$ where $v$ is the speed of sound in water in meters per
second and $f$ is the ADCP frequency in Hz. $v$ is a function of temperature, $T$ in °C, salinity, $S$ in
parts per thousand or practical salinity units (psu), and depth, $z$ in km, and was calculated
according to:

$$ v = 1449.05 + 45.7t - 5.21t^2 + 0.23t^3 + $$

$$ (1.333 - 0.126t + 0.009t^2)(S - 35) + 16.3z + 0.18z^2 \quad \text{A.3} $$

with $t = 0.1T$. The equation is accurate to ±0.1 m s⁻¹ within the limits of 0 < $T$ < 30 °C, 0 < $S$ <
40 psu and 0 < $z$ < 4 km (Coppens, 1981).

The transition between the near and far transducer field is given by $R_{\text{critical}} = \pi a_i^2 \lambda^{-1}$ and is
used to calculated the dimensionless parameter $Z = R/R_{\text{critical}}$. The near field correction is given by (Gartner, 2004; Wall et al., 2006):

$$ Z = \frac{1 + 1.35Z + (2.5Z)^{2.5}}{1.35Z + (2.5Z)^{2.5}} \quad \text{A.4} $$

The third term on the RHS of Equation 3.2 accounts for the attenuation of the signal as it
travels through the water column. $\alpha$ is the water attenuation coefficient and is a function of
salinity, temperature, pressure, $P$ in atm, and the ADCP acoustic frequency. $\alpha$, in dB m⁻¹ is
(Schulkin and Marsh, 1962):
\[
\alpha = 8.686 \left( \frac{SA_\alpha f_T(0.001f)^2}{f_r^2 + (0.001f)^2} + \frac{B_\alpha(0.001f)^2}{f_T} \right) (1 - C_\alpha P) \tag{A.5}
\]

where \( A_\alpha = 2.34 \times 10^{-6}, B_\alpha = 3.38 \times 10^{-6}, C_\alpha = 6.54 \times 10^{-4} \), 8.686 is the conversion factor from nepers (an infrequently used unit) to dB and:

\[
f_T = 21.9 \times 10^{(6-\frac{1520}{T+273})} \tag{A.6}
\]

The final term on the RHS of Equation 3.2 is a correction for the attenuation of the signal due to scattering and absorption by particles in the water column where:

\[
2\alpha_s = 8.686 \cdot \zeta \cdot \text{SSC} \cdot 10^{-3} \tag{A.7}
\]

where \( \alpha_s \) is in dB m\(^{-1} \) and SSC is the suspended sediment concentration in kg m\(^{-3} \). Flammer (1962) notes that in the original work by (Urick, 1948) SSC was in ppm but that 1000 ppm = 0.001. The factor of 10\(^{-3} \) in Equation A.7 is used to convert the SSC into a non-dimensional volume of particles per unit volume. This conversion assumes particles have the same density as water. \( \zeta \), in nepers m\(^{-1} \), is defined as (Flammer, 1962):

\[
\zeta = K(\gamma - 1)^2 \left( \frac{S_p}{S_p^2 + (\gamma + \tau)^2} \right) + \frac{K^4 a_p^3}{6} \tag{A.8}
\]

where \( K = 2\pi \lambda^{-1}, \gamma \) is the aggregate wet density, \( \rho_a \), divided by the water density. \( S_p \) is given by:

\[
S_p = \left( \frac{9}{4\beta a_p} \right) \left( 1 + \frac{1}{\beta a_p} \right) \tag{A.9}
\]

where \( a_p \) is the particle radius in meters and \( \beta = \pi f_v^{-1} \), in m\(^{-1} \), where \( v \) is the viscosity of water in m\(^2\) s\(^{-1} \). For these calibrations a constant radius of \( a_p = 120 \times 10^{-6} \) m was assumed.
Appendix B

ADCP Error Analysis Equation Development

The basis for the ADCP error analysis is error propagation. Errors, here due to uncertainty in input parameters and variability in the ADCP measurements, are propagated through the calibration equation (Equation 3.2) to determine the uncertainty in the final value of the SSC measurement. For the general case, given variables $u$ and $v$ with variances $\sigma_u^2$ and $\sigma_v^2$, respectively, and given that $x$ is some function of $u$ and $v$, the variance for $x$ may be approximated as (Bevington and Robinson, 2003):

$$
\sigma_x^2 \approx \sigma_u^2 \left( \frac{\partial x}{\partial u} \right)^2 + \sigma_v^2 \left( \frac{\partial x}{\partial v} \right)^2 + 2\sigma_{uv} \left( \frac{\partial x}{\partial u} \right) \left( \frac{\partial x}{\partial v} \right) \tag{5.8}
$$

where $\sigma_{uv}^2$ is the covariance between $u$ and $v$ and variance is the square of standard deviation. When the relationship between $u$ and $v$ is non-linear higher order terms in the Taylor series expansion of the relationship have been neglected in developing Equation 5.8.

The development of the error analysis for Equation 3.2 requires the application of several specific applications of Equation 5.21 which are listed below. In the following equations $u$ and $v$ retain their function as the variables of interest and $a$ and $b$ are constants (Bevington and Robinson, 2003):

- **Addition**
  $$
x = au + bv \quad \sigma_x^2 = a^2\sigma_u^2 + b^2\sigma_v^2 + 2ab\sigma_{uv} \tag{B.1}
$$

- **Multiplication**
  $$
x = auv \quad \frac{\sigma_x^2}{x^2} = \frac{\sigma_u^2}{u^2} + \frac{\sigma_v^2}{v^2} + 2\frac{\sigma_{uv}}{uv} \tag{B.2}
$$

- **Division**
  $$
x = \frac{au}{v} \quad \frac{\sigma_x^2}{x^2} = \frac{\sigma_u^2}{u^2} + \frac{\sigma_v^2}{v^2} - 2\frac{\sigma_{uv}}{uv} \tag{B.3}
$$

- **Power**
  $$
x = au^b \quad \frac{\sigma_x}{x} = b \frac{\sigma_u}{u} \tag{B.4}
$$

- **Natural log**
  $$
x = a \ln(bu) \quad \sigma_x = ab \frac{\sigma_u}{u} \tag{B.5}
$$

Note that in Equations B.4 and B.5 the uncertainty is given as the standard deviation and not the variance. With the above equations it is straightforward to develop the error propagation equations. However, at each mathematical operation a corresponding error equation needs to be developed, which in this case required nearly 70 equations. The equations, listed below as each operation in Equation 3.2 and the corresponding error equations, are grouped together by the four terms on the right hand side of Equation 3.2. Those four terms are finally combined to yield the uncertainty for $S_v$, which then appears in Equations 5.10 – 5.12 as $\sigma_{S_v}$.  

142
Term I

\[ Eq01 = \frac{1}{10} E_1 Kc \]
\[ \sigma_{Eq01}^2 = (\text{Eq}01)^2 \left( \frac{\sigma_{E_1}^2}{E_1^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.6)

\[ Eq02 = \frac{1}{10} E_2 Kc \]
\[ \sigma_{Eq02}^2 = (\text{Eq}02)^2 \left( \frac{\sigma_{E_2}^2}{E_2^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.7)

\[ Eq03 = \frac{1}{10} E_3 Kc \]
\[ \sigma_{Eq03}^2 = (\text{Eq}03)^2 \left( \frac{\sigma_{E_3}^2}{E_3^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.8)

\[ Eq04 = \frac{1}{10} E_4 Kc \]
\[ \sigma_{Eq04}^2 = (\text{Eq}04)^2 \left( \frac{\sigma_{E_4}^2}{E_4^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.9)

\[ Eq05 = \frac{1}{10} Er_1 Kc \]
\[ \sigma_{Eq05}^2 = (\text{Eq}05)^2 \left( \frac{\sigma_{Er_1}^2}{Er_1^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.10)

\[ Eq06 = \frac{1}{10} Er_2 Kc \]
\[ \sigma_{Eq06}^2 = (\text{Eq}06)^2 \left( \frac{\sigma_{Er_2}^2}{Er_2^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.11)

\[ Eq07 = \frac{1}{10} Er_3 Kc \]
\[ \sigma_{Eq07}^2 = (\text{Eq}07)^2 \left( \frac{\sigma_{Er_3}^2}{Er_3^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.12)

\[ Eq08 = \frac{1}{10} Er_4 Kc \]
\[ \sigma_{Eq08}^2 = (\text{Eq}08)^2 \left( \frac{\sigma_{Er_4}^2}{Er_4^2} + \frac{\sigma_{Kc}^2}{Kc^2} \right) \] (B.13)

\[ Eq09 = \frac{1}{10} E_1 Kc - \frac{1}{10} Er_1 Kc \]
\[ \sigma_{Eq09}^2 = \sigma_{Eq01}^2 + \sigma_{Eq05}^2 \] (B.14)

\[ Eq10 = \frac{1}{10} E_2 Kc - \frac{1}{10} Er_2 Kc \]
\[ \sigma_{Eq10}^2 = \sigma_{Eq02}^2 + \sigma_{Eq06}^2 \] (B.15)

\[ Eq11 = \frac{1}{10} E_3 Kc - \frac{1}{10} Er_3 Kc \]
\[ \sigma_{Eq11}^2 = \sigma_{Eq03}^2 + \sigma_{Eq07}^2 \] (B.16)

\[ Eq12 = \frac{1}{10} E_4 Kc - \frac{1}{10} Er_4 Kc \]
\[ \sigma_{Eq12}^2 = \sigma_{Eq04}^2 + \sigma_{Eq08}^2 \] (B.17)

\[ Eq13 = 10^{(\text{Eq}09)} \]
\[ \sigma_{Eq13}^2 = (\text{Eq}13)^2 ((\ln (10))^2 \sigma_{Eq09}^2 \] (B.18)

\[ Eq14 = 10^{(\text{Eq}10)} \]
\[ \sigma_{Eq14}^2 = (\text{Eq}14)^2 ((\ln (10))^2 \sigma_{Eq10}^2 \] (B.19)

\[ Eq15 = 10^{(\text{Eq}11)} \]
\[ \sigma_{Eq15}^2 = (\text{Eq}15)^2 ((\ln (10))^2 \sigma_{Eq11}^2 \] (B.20)

\[ Eq16 = 10^{(\text{Eq}12)} \]
\[ \sigma_{Eq16}^2 = (\text{Eq}16)^2 ((\ln (10))^2 \sigma_{Eq12}^2 \] (B.21)

\[ Eq17 = \frac{1}{4} \text{Eq}13 + \frac{1}{4} \text{Eq}14 + \frac{1}{4} \text{Eq}15 + \frac{1}{4} \text{Eq}16 \]
\[ \sigma_{Eq17}^2 = \left( \frac{1}{4} \right)^2 \sigma_{Eq13}^2 + \left( \frac{1}{4} \right)^2 \sigma_{Eq14}^2 + \left( \frac{1}{4} \right)^2 \sigma_{Eq15}^2 + \left( \frac{1}{4} \right)^2 \sigma_{Eq16}^2 \] (B.22)
\[ Eql = \frac{10}{\ln(10)} \ln \left[ \left( \frac{1}{4} \right) 10^\left( \frac{1}{10}E_1Kc - \frac{1}{10}Er_1Kc \right) + \left( \frac{1}{4} \right) 10^\left( \frac{1}{10}E_2Kc - \frac{1}{10}Er_2Kc \right) + \left( \frac{1}{4} \right) 10^\left( \frac{1}{10}E_3Kc - \frac{1}{10}Er_3Kc \right) + \left( \frac{1}{4} \right) 10^\left( \frac{1}{10}E_4Kc - \frac{1}{10}Er_4Kc \right) \right] \]

\[ \sigma_I^2 = \left( \frac{10}{\ln(10)} \right)^2 \frac{\sigma_{Eq17}^2}{(Eq 17)^2} \quad (B.23) \]

**Term II**

\[ Eq18 = 45.7 \left( \frac{T}{10} \right) \]
\[ \sigma_{Eq18}^2 = \left( \frac{45.7}{10} \right)^2 \sigma_T^2 \quad (B.24) \]

\[ Eq19 = 5.21 \left( \frac{T}{10} \right)^2 \]
\[ \sigma_{Eq19}^2 = 2^2(Eq 19)^2 \frac{\sigma_T^2}{T^2} \quad (B.25) \]

\[ Eq20 = 0.23 \left( \frac{T}{10} \right)^3 \]
\[ \sigma_{Eq20}^2 = 3^2(Eq 20)^2 \frac{\sigma_T^2}{T^2} \quad (B.26) \]

\[ Eq21 = 1.333S \]
\[ \sigma_{Eq21}^2 = 1.333^2 \sigma_S^2 \quad (B.27) \]

\[ Eq22 = \frac{0.126}{10} TS \]
\[ \sigma_{Eq22}^2 = (Eq 22)^2 \left( \frac{\sigma_T^2}{T^2} + \frac{\sigma_S^2}{S^2} \right) \quad (B.28) \]

\[ Eq23 = T^2 \]
\[ \sigma_{Eq23}^2 = 2^2 T^2 \sigma_T^2 \quad (B.28) \]

\[ Eq24 = \frac{0.009}{100} ST^2 \]
\[ \sigma_{Eq24}^2 = (Eq 24)^2 \left( \frac{\sigma_T^2}{T^2} + \frac{\sigma_S^2}{S^2} \right) \quad (B.29) \]

\[ Eq25 = \frac{35.0.126}{10} T \]
\[ \sigma_{Eq25}^2 = \left( \frac{35.0.126}{10} \right)^2 \sigma_T^2 \quad (B.30) \]

\[ Eq26 = \frac{35.0.009}{100} T^2 \]
\[ \sigma_{Eq26}^2 = \left( \frac{35.0.009}{100} \right)^2 T^2 \sigma_T^2 \quad (B.31) \]

\[ Eq27 = 16.3z \]
\[ \sigma_{Eq27}^2 = 16.3^2 \sigma_z^2 \quad (B.32) \]

\[ Eq28 = 0.18z^2 \]
\[ \sigma_{Eq28}^2 = (2 \cdot 0.18)^2 z^2 \sigma_z^2 \quad (B.33) \]

\[ Eq29 = v = 1449.05 + Eq18 - Eq19 + Eq20 + Eq21 - Eq22 + Eq24 - 1.333 \cdot 35 + Eq25 - Eq26 + Eq27 + Eq28 \]

\[ \sigma_{Eq29}^2 = \sigma_{Eq18}^2 + \sigma_{Eq19}^2 + \sigma_{Eq20}^2 + \sigma_{Eq21}^2 + \sigma_{Eq22}^2 + \sigma_{Eq24}^2 + \sigma_{Eq25}^2 + \sigma_{Eq26}^2 + \sigma_{Eq27}^2 + \sigma_{Eq28}^2 \quad (B.34) \]

When calculated via Equation B.34 the variance for the velocity of sound in water, \( \sigma_{Eq29}^2 \), was smaller than the reported accuracy of the velocity equation so \( \sigma_{Eq29}^2 \) was taken to be the equation accuracy and Equations B.24 – B.34 were not used.
Eq30 = \lambda = \frac{v}{f} \quad \sigma_{Eq30}^2 = \left(\frac{1}{f}\right)^2 \sigma_{Eq29}^2 \quad (B.35)

Eq31 = a_t^2 \quad \sigma_{Eq31}^2 = 2^2 a_t^2 \sigma_{a_t}^2 \quad (B.36)

Eq32 = R_{crit} = \pi \frac{Eq31}{Eq30} \quad \sigma_{Eq32}^2 = (Eq32)^2 \left(\frac{\sigma_{Eq31}^2}{(Eq31)^2} + \frac{\sigma_{Eq30}^2}{(Eq30)^2}\right) \quad (B.37)

Eq33 = Z = \frac{R}{R_{crit}} \quad \sigma_{Eq33} = (-1)(Eq33) \frac{\sigma_{Eq32}}{R_{crit}} \quad (B.38)

Eq34 = 2.5^{3.2} E_{33}^{3.2} \quad \sigma_{Eq34}^2 = (Eq34)^2 3.2^2 \frac{\sigma_{Eq33}}{(Eq33)^2} \quad (B.39)

Eq35 = 1 + 1.35 \cdot Eq33 + Eq34 \quad \sigma_{Eq35}^2 = 1.35^2 \sigma_{Eq33}^2 + \sigma_{Eq34}^2 + 2(1.35)\sigma_{Eq33,Eq34}^2 \quad (B.40)

Eq36 = 1.35Z + (2.5Z)^{3.2} \quad \sigma_{Eq36}^2 = 1.35^2 \sigma_{Eq33}^2 + \sigma_{Eq34}^2 + 2(1.35)\sigma_{Eq33,Eq34}^2 \quad (B.41)

Eq37 = \psi = \frac{Eq35}{Eq36} \quad \sigma_{Eq37}^2 = (Eq37)^2 \left(\frac{\sigma_{Eq35}^2}{Eq35^2} + \frac{\sigma_{Eq36}^2}{Eq36^2} - 2 \frac{\sigma_{Eq35}^2}{Eq35 \cdot Eq36}\right) \quad (B.42)

Eq41 = (20 \log_{10}(R)) \quad \sigma_{Eq41}^2 = (20 \log_{10}(R))^2 \sigma_{Eq37}^2 \quad (B.43)

Term III

Eq38 = \frac{1520}{T+273} \quad \sigma_{Eq38} = (-1)(Eq38) \frac{\sigma_T}{T+273} \quad (B.44)

Eq39 = 6 - Eq38 \quad \sigma_{Eq39} = \sigma_{Eq38} \quad (B.45)

Eq40 = f_T = 21.9 \times 10^{(Eq39)} \quad \sigma_{Eq40}^2 = 21.9^2 (Eq40)^2 \ln(10)^2 \sigma_{Eq39}^2 \quad (B.46)

Eq41 = S \cdot Eq40 \quad \sigma_{Eq41}^2 = (S \cdot Eq40)^2 \left(\frac{\sigma_S^2}{S^2} + \frac{\sigma_{Eq40}^2}{Eq40^2} + 2 \frac{\sigma_{Eq41}^2}{S \cdot Eq40}\right) \quad (B.47)

Eq42 = Eq40^2 \quad \sigma_{Eq42}^2 = 2^2 (Eq40)^2 \sigma_{Eq40}^2 \quad (B.48)

Eq43 = Eq42 + f^2 \quad \sigma_{Eq43}^2 = \sigma_{Eq42}^2 \quad (B.49)

Eq44 = \frac{Eq41}{Eq43} \quad \sigma_{Eq44}^2 = (Eq44)^2 \left(\frac{\sigma_{Eq41}^2}{(Eq41)^2} + \frac{\sigma_{Eq42}^2}{(Eq43)^2} - 2 \frac{\sigma_{Eq41,Eq43}^2}{Eq41 \cdot Eq43}\right) \quad (B.50)
\[ Eq45 = \frac{Bf^2}{Eq40} \quad \sigma_{Eq45} = (-1)(Eq45) \frac{\sigma_{Eq40}}{f_T} \]  \hspace{1cm} (B.51)

\[ Eq46 = Eq44 + Eq45 \quad \sigma_{Eq46}^2 = \sigma_{Eq44}^2 + \sigma_{Eq45}^2 + 2\sigma_{Eq44,Eq45}^2 \]  \hspace{1cm} (B.52)

\[ Eq47 = 1 - CP \quad \sigma_{Eq47}^2 = C^2\sigma_p^2 \]  \hspace{1cm} (B.53)

\[ Eq48 = \alpha = (Eq46)(Eq47) \]

\[ \sigma_{Eq48}^2 = (Eq48)^2 \left( \frac{\sigma_{Eq46}^2}{(Eq46)^2} + \frac{\sigma_{Eq47}^2}{(Eq47)^2} + 2 \frac{\sigma_{Eq46,Eq47}^2}{Eq46\cdot Eq47} \right) \]  \hspace{1cm} (B.54)

\[ EqlIII = 2\alpha R \quad \sigma_{EqlIII}^2 = (2R)^2\sigma_{Eq48}^2 \]  \hspace{1cm} (B.55)

**Term IV**

\[ Eq49 = K = \frac{2\pi}{\lambda} \quad \sigma_{Eq49} = (-1)(Eq49) \frac{\sigma_{Eq30}}{Eq30} \]  \hspace{1cm} (B.56)

\[ Eq50 = \gamma = \frac{\rho_{agg}}{\rho_{fluid}} \quad \sigma_{Eq50}^2 = \left( \frac{1}{\rho_{fluid}} \right)^2 \sigma_{p_{agg}}^2 \]  \hspace{1cm} (B.57)

\[ Eq51 = \frac{9}{4\beta a_p} \quad \sigma_{Eq51} = (-1)(Eq51) \frac{\sigma_{ap}}{a_p} \]  \hspace{1cm} (B.58)

\[ Eq52 = \frac{1}{\beta a_p} \quad \sigma_{Eq52} = (-1)(Eq52) \frac{\sigma_{ap}}{a_p} \]  \hspace{1cm} (B.59)

\[ Eq53 = 1 + Eq52 \quad \sigma_{Eq53} = \sigma_{Eq52} \]  \hspace{1cm} (B.60)

\[ Eq54 = S_p = (Eq51)(Eq53) \quad \sigma_{Eq54}^2 = (Eq54)^2 \left( \frac{\sigma_{Eq51}^2}{(Eq51)^2} + \frac{\sigma_{Eq53}^2}{(Eq53)^2} + 2 \frac{\sigma_{Eq41,Eq53}^2}{Eq51\cdot Eq53} \right) \]  \hspace{1cm} (B.61)

\[ Eq55 = \tau = 0.5 + Eq51 \quad \sigma_{Eq55} = \sigma_{Eq51} \]  \hspace{1cm} (B.62)

\[ Eq56 = Eq50 + Eq55 \quad \sigma_{Eq56}^2 = \sigma_{Eq50}^2 + \sigma_{Eq55}^2 + 2\sigma_{Eq50,Eq55}^2 \]  \hspace{1cm} (B.63)

\[ Eq57 = (Eq56)^2 \quad \sigma_{Eq57}^2 = (Eq56)^2 \sigma_{Eq56}^2 \]  \hspace{1cm} (B.64)

\[ Eq58 = Eq54^2 \quad \sigma_{Eq58}^2 = (Eq54)^2 \sigma_{Eq54}^2 \]  \hspace{1cm} (B.65)

\[ Eq59 = Eq58 + Eq57 \quad \sigma_{Eq59}^2 = \sigma_{Eq58}^2 + \sigma_{Eq57}^2 + 2\sigma_{Eq58,Eq57}^2 \]  \hspace{1cm} (B.66)

\[ Eq60 = \frac{Eq54}{Eq59} \quad \sigma_{Eq60}^2 = (Eq60)^2 \left( \frac{\sigma_{Eq54}^2}{(Eq54)^2} + \frac{\sigma_{Eq59}^2}{(Eq59)^2} - 2 \frac{\sigma_{Eq54,Eq59}^2}{Eq54\cdot Eq59} \right) \]  \hspace{1cm} (B.66)
\[ Eq61 = (Eq50 - 1)^2 \]
\[ \sigma_{Eq61}^2 = 2^2(Eq61)\sigma_{Eq50}^2 \quad (B.67) \]

\[ Eq62 = Eq49 \cdot Eq61 \cdot Eq60 \]
\[ \sigma_{Eq62}^2 = (Eq62)^2 \left[ \frac{(\sigma_{Eq49})^2}{(Eq49)^2} + \frac{\sigma_{Eq61}^2}{(Eq61)^2} + \frac{\sigma_{Eq60}^2}{(Eq60)^2} \right] + 2 \frac{\sigma_{Eq49,Eq61,Eq60}^2}{Eq49\cdotEq61\cdotEq60} + 2 \frac{\sigma_{Eq60,Eq61}^2}{Eq60\cdotEq61} \]
\[ \sigma_{Eq62}^2 = (Eq62)^2 \left[ \frac{(\sigma_{Eq49})^2}{(Eq49)^2} + \frac{\sigma_{Eq61}^2}{(Eq61)^2} + \frac{\sigma_{Eq60}^2}{(Eq60)^2} \right] + 2 \frac{\sigma_{Eq49,Eq61,Eq60}^2}{Eq49\cdotEq61\cdotEq60} + 2 \frac{\sigma_{Eq60,Eq61}^2}{Eq60\cdotEq61} \quad (B.68) \]

\[ Eq63 = Eq49^4 \]
\[ \sigma_{Eq63}^2 = 4^2(Eq49)^6\sigma_{Eq49}^2 \quad (B.69) \]

\[ Eq64 = a_p^3 \]
\[ \sigma_{Eq64}^2 = 3^2a_p^4\sigma_p^2 \quad (B.70) \]

\[ Eq65 = \frac{Eq63-Eq64}{6} \]
\[ \sigma_{Eq65}^2 = (Eq65)^2 \left[ \frac{\sigma_{Eq63}^2}{(Eq63)^2} + \frac{\sigma_{Eq64}^2}{(Eq64)^2} + 2 \frac{\sigma_{Eq63,Eq64}^2}{Eq63\cdotEq64} \right] \quad (B.71) \]

\[ Eq66 = \zeta = Eq62 + Eq65 \]
\[ \sigma_{Eq66}^2 = \sigma_{Eq62}^2 + \sigma_{Eq65}^2 + 2\sigma_{Eq62,Eq65}^2 \quad (B.72) \]

\[ Eq67 = \alpha_s = \frac{1}{2}8.686 \cdot Eq66 \cdot SSC \]
\[ \sigma_{Eq67}^2 = (Eq67)^2 \left[ \frac{\sigma_{Eq66}^2}{(Eq66)^2} + \frac{\sigma_{SSC}^2}{(SSC)^2} + 2 \frac{\sigma_{Eq66,SSC}^2}{Eq66\cdotSSC} \right] \quad (B.73) \]

\[ EqIV = 2\alpha_s R \]
\[ \sigma_{EqIV}^2 = (2R)^2\sigma_{Eq67}^2 \quad (B.74) \]

**Equation for \( S_v \)**

\[ S_v = 10 \log_{10} \left( \sum_{i=1}^{n} \left( \frac{(y_i - \bar{y}_{i,\bar{r}})}{10} \right)^2 \right) + \psi \cdot 20 \log_{10} R + 2\alpha R + 2\alpha_s R \]

\[ S_v = EqI + EqII + EqIII + EqIV \]

\[ \sigma_{S_v}^2 = \sigma_{EqI}^2 + \sigma_{EqII}^2 + \sigma_{EqIII}^2 + \sigma_{EqIV}^2 + 2\sigma_{EqI,EqII}^2 + 2\sigma_{EqI,EqIII}^2 + 2\sigma_{EqI,EqIV}^2 + 2\sigma_{EqII,EqIII}^2 + 2\sigma_{EqII,EqIV}^2 + 2\sigma_{EqIII,EqIV}^2 + 2\sigma_{EqI,EqIII,EqIV}^2 + 2\sigma_{EqI,EqII,EqIV}^2 + 2\sigma_{EqI,EqII,EqIII,EqIV}^2 \quad (B.75) \]
Appendix C

Coupled Sediment Bed – Water Column Model Code

The sediment bed – water column model was coded in MATLAB. The code presented is modified versions of a water column model (Brand et al., 2011) and a sediment bed model (Sanford, 2008) which were coupled together. The sediment bed code was modified to remove the sand fraction. Lines of code were removed throughout the model as a result of this modification. A similar result would have been achieved by setting the sand fraction to zero. The sediment code was also modified to receive the settling sediment flux from the water column. The water column model was modified to receive as an input the erosional flux of sediments from the sediment model.

The code is written as 6 separate files. The executable file is used to define most of the input and model run parameters (run_coupled_real_data.mat). Note that this file loads a .mat file (InputCompleteCorrectAll.mat) which includes the required time series data derived from the experimental data. The details and structure of this file are included in the .m file at the point where the file is loaded.

The other five files are a series of MATLAB functions called by the executed file and the other functions. The main part of model is contained within the coupled_model_new_01.m file which is constructed as a set of functions which set up the initial conditions for the sediment bed and water column model and then advances the model through time. A modified Heaviside function (H.m) is called by the sediment model. The function phistot.m calculates the total solids volume fraction, and along with the Heaviside function, were developed by (Sanford, 2008). The vertical turbulent sediment fluxes at 0.36 and 0.72 mab are extracted from the model results and compared to the experimental values by the eval_fluxes.m function. The last function (error_est.m) calculates the statistical metrics to determine the model fit the experimental data. In the following code an ellipsis indicates that the line of code is continued on the following line. Comments are preceded by a %.
close all
clear all
clc
tic

load InputCompleteCorrectAll.mat %file contains input data
da='Be364Fall';
b='Be724Fall';
%a.tides    m
%a.fluxes   g/m2.s
%a.shears   m/s
%a.Csed     mg/L
%a.time     s
%a.z0       m
%a.Cini     mg/L

Tmodel=19; %length of time to be modeled, days
%~19.98 days of data for spring, 26.1 for full fall deployment

time=[0:200:Tmodel*86400]'; %units of seconds
DeltaT=100; %timestep, seconds.

%Obligatory Data
TestStruct.tides(:,1)=eval([a '.tides(:,1)']); %tides time series
TestStruct.tides(:,2)=eval([a '.tides(:,2)']); %l:time, 2: elevation
TestStruct.shears(:,1)=eval([a '.shears(:,1)']); %bed shear stress
TestStruct.shears(:,2)=eval([a '.shears(:,2)']);
TestStruct.shearwind(:,1)=eval([a '.ustcw(:,1)']); %wave enhanced bed
TestStruct.shearwind(:,2)=eval([a '.ustcw(:,2)']); %shear velocity, u*
TestStruct.fluxes(:,1)=eval([a '.fluxes(:,1)']); %sediment fluxes
TestStruct.fluxes(:,2)=eval([a '.fluxes(:,2)']);
TestStruct.fluxes(:,2)=TestStruct.fluxes(:,2)/1000; %convert to kg/m2.s

TestStruct.parameters.eff=1.8; %flux amplification factor
TestStruct.parameters.diff_beta=0.462; %beta, inverse shmidt no.
TestStruct.parameters.alpha=9.3e-3; %fine sediment fraction
TestStruct.parameters.ws=[0.51e-3 2.1e-5];%Number of settling velocities
%determines number of modeled sediment fractions
TestStruct.time(:,1)=max(time-720); %time-720)
TestStruct.z0(:,1)=eval([a '.z0']); %z0 in law of wall equation
TestStruct.Cini(:,1)=eval([a '.Cini']); %initial concentration, units of mg/L
TestStruct.Cini(:,1)=TestStruct.Cini(:,1)/1000; %convert to kg/m3

%Used for comparison to water column results.
%Estimated total water column mass kg/m2.
TestStruct.Csed(:,1)=eval([a '.Csed(:,1)']);
TestStruct.Csed(:,2)=eval([a '.Csed(:,2)']); %units of mg/L
TestStruct.Csed(:,2)=TestStruct.Csed(:,2)/1000; %convert to kg/m3

TestStruct.Csed(:,3)=eval([b '.Csed(:,1)']);
TestStruct.Csed(:,4)=eval(['.Csed(:,2)']); %units of mg/L
TestStruct.Csed(:,4)=TestStruct.Csed(:,4)/1000; %convert to kg/m3;
TestStruct.zobs2=0.725;

TestStruct.Tmodel=Tmodel;

%--------------------------------------------------------------------------

%Input structure of values for Sediment Model
%Variables and Initial Conditions.
%All units in kg, m, seconds

%SedVar
SedVar.n=100;                           %number of sediment layers
SedVar.delm=0.05;                       %thickness of sediment layers, kg/m^2
SedVar.Rhos=2500;                       %sediment density, kg/m^3
SedVar.Dmix=2.7e-7/86400;               %Mixing/Bioturbation constant m^2/s
SedVar.beta=11.75/86400;         %mass transfer coeff for mud, m s^-1 Pa^-1
SedVar.conrate=5.5/86400;     %consolidation rate in s^-1
SedVar.exprrate=SedVar.conrate/1000; %expansion (swelling) rate, s^-1
tc1=0.13;                         %tc1+tc2*m^tc3 equilibrium tau-crit profile
tc2=2.3;
tc3=0.4;
SedVar.cl=0.16;           %initial tau-crit profile, same equation as before
SedVar.c2=0.4;
SedVar.c3=0.4;
SedVar.tceqn=([num2str(tc1),'+',... %equilibrium critical shear
num2str(tc2),'*(',num2str(tc3),')])); %stress equation, N/m^2
SedVar.tcmin=0.07;                %minimum critical shear stress, N/m^2
SedVar.phismin=0.1;                     %minimum solids fraction of deposited
mud, unitless

%plot results?
pl=-1; %1 = plot, -1 = R^2 values w/o plot

Results=coupled_model_new_01(TestStruct, SedVar, DeltaT, 0.36, pl);
toc
%Implicit version one sediment fraction calculation of rouse
%new: adaptive grid, reorganized gridstructure to speed up calculations
%New Handling of Gridconcentration assignment: No Interpolation. Withdrawal
%of Sediment during Ebb, Just adding of water during flood
%Input Structure: measured data 1st row is measurement time. Blah.tides:
% n x 2 array, Blah.fluxes: nx2 array,
%blah.shears: nx2 array, C sed= nX2 array

%Tidal Watercolumn with grid variation
% Inputdata: zobs: elevation at which concentration is measured. The value
% is used for the calculation of the rouseprofile as initial value
%DeltaT: time stepsize
%InputStructure: All sorts of input parameters (see also Testrun on how to
%use Inputstructure).

%TestStruct.tides(:,1): Times at which tidal elevations are specified
%TestStruct.tides(:,2): Tidal Elevations [m] The Parameter is obligatory!

%TestStruct.shears(:,1): Times at which shears are specified
%TestStruct.shears(:,2): Shearvelocity u* [m/s] The parameter is
%obligatory (for calculation of the velocity field and Dispersion
%coefficient when getD option is set to 'Rouse'

%TestStruct.fluxes(:,1): Times at which fluxes are specified
%TestStruct.fluxes(:,2): Fluxes observed at certain elevation over the
%sediment value is used to calculate sediment fluxes across sediment-water
%interface The value is used in Function getloweBC

%TestStruct.Csed(:,1): just guess!
%TestStruct.Csed(:,2): Measured Sediment data. VAlue is used for inverse
%fitting in different file Therefore the Value is obsolete for forward
%modeling

%TestStruct.time(1,1): Last time of modelrun (has to be <= times of
%measured values. Obligatory
%TestStruct.z0(1,1): Roughness height. Used for calculation of velocity
%profile used for grid size adaption. Obligatory.

%TestStruct.Cini(:,1): Initial concentrations at z obs used for rouse profile
%inalization. 1 for each settling velocity;

%TestStruct.parameters.eff: Amplification Factor for sediment resuspension.
%neccessary for calculation of sediment resuspension based on observed
%sediment flux in function getlowerBC

%TestStruct.parameters.alpha: Fraction of fine sediment in resuspended
%sediment for calculation of sediment resuspension based on observed
%sediment flux in function getlowerBC one value for each fraction after the
%first one

%TestStruct.parameters.ws: [m/s] Settling velocities for particle size
%fractions.
%Obligatory! (By the way: the code will the detect automatically the number
% of modeled fraction by the number of specified settling velocities.

%SedVar is a structure of variable parameters for the sediment bed model.
%SedVar.n % number of sediment layers
%SedVar.delm % thickness of sediment layers, kg/m^2
%SedVar.Rhos % sediment density, kg/m^3
%SedVar.Dmix % Mixing/Bioturbation constant (kg/m^2)^2/s
%SedVar.beta % mass transfer coeff for mud, m s^-1 Pa^-1
%SedVar.conrate % consolidation rate in s^-1
%SedVar.exp Rate % expansion (swelling) rate, s^-1
%SedVar.tceqn % equilibrium critical shear stress equation, N/m^2
%SedVar.tcmin % minimum critical shear stress, N/m^2
%SedVar.phismin % minimum solids fraction of deposited mud, unitless

% grid is a structure of variables used in the water column model
% grid.numfracs 2
% grid.numcells 60
% grid.Height 4.992
% grid.zlowerbound numcells x 1
% grid.zupperbound numcells x 1
% grid.zcenter numcells x 1
% grid.Deltaz numcells x 1
% grid.Fluxin 1 x numfracs
% grid.Fluxout 1 x numfracs
% grid.Mass1 numcells x numfracs
% grid.Slat numcells x numfracs

%Sediment model needs H.m and phistot.m files to be in same folder in order to run.

function Results=coupled_model_new_01(InputStructure, ...
   SedVar, DeltaT, zobs, pl)
start=clock; % track the time it takes the model to run
plotit=pl; % plot results? 1 = yes.

% Start water column model initialization
% spatial discretization
NGridCell=60; % Number of water column model Gridcells

if isstruct(InputStructure) % Determine Modeled timeinterval and number of fractions
   Timeinterval=InputStructure.time+DeltaT;%19*86400+1;
   numfracs=size(InputStructure.parameters.ws,2);
else
   error('Inputstructure is missing');
end

time=0:DeltaT:InputStructure.time+DeltaT;
umsteps=floor(Timeinterval/DeltaT);
savesteps=max(100,DeltaT); % set times for saving output time array
Outputtimes=savesteps:savesteps:Timeinterval-DeltaT;%Timearray for Dataoutput
Outputindex=floor(Outputtimes/DeltaT)+1; %calculation of indices
%which correspond closest time to outputtimes

grid=generategrid(NGridCell,numfracs); %generate domain
grid_maxind=grid.numcells; %calculate maximum index
grid=getheight(grid, 0, InputStructure); %initialize domain
grid=initialize(grid,'Rousespec',zobs, InputStructure);

Elevations=zeros(grid_maxind+1, size(Outputtimes,2)+1);
Masses1=zeros(grid_maxind+1, size(Outputtimes,2)+1,numfracs);
DeltaZs=Elevations;
Slats=Elevations;
Fluxesin=zeros(numfracs, size(Outputtimes,2)+1);
Fluxesout=Fluxesin;
Forplot=extractgrid(grid); %Store initial conditions
for k=1:size(grid.Mass1,2) %loop through water column concentrations to calculate
    Masses1(:,1,k)=Forplot.Mass1(:,k);
end
Elevations(:,1)=Forplot.elevation;
End Water Column Initialization
%************************************************************************************
%Start Bed Column Initialization
Sed=sed_bed_init(Timeinterval,numsteps,InputStructure, DeltaT,
NGridCell,SedVar, grid, time);
WCm=zeros(1,numsteps); %store water column mass from water column model
WCm(1)=Sed.WCm(1);
End Bed Column Initization
%************************************************************************************
%Start calculations
j=1; % counter for outputtimes
for i=1:numsteps %start main loop
    if mod(i*DeltaT,3600*24)==0
        i*DeltaT/(3600*24)
    end
    gridold=grid;
    grid=steptime(gridold, DeltaT, i, InputStructure,Sed.FintoWC);
    Sed=sed_new(i,Sed,DeltaT,grid, InputStructure);
    if isnan(Sed.m(Sed.int))==1
        Results=NaN;
        return
    end
    WCtot=0;
    for wci=1:NGridCell %loop through water column concentrations to calculate
WCtemp = sum(grid.Mass1(wci,:)*grid.Deltaz(wci)); % total sediment in the water column according to the water column model. units of kg/m2.

WCtot = WCtot + WCtemp;

end

WCm(i+1) = WCtot;

if j <= size(Outputindex,2) && i == Outputindex(j)
    % store values at intervals determined by Outputindex
    Forplot = extractgrid(grid);
    Elevations(:,j+1) = Forplot.elevation;
    DeltaZs(:,j+1) = Forplot.Deltaz;
    for k = 1:size(grid.Mass1,2)
        Fluxesin(k,j+1) = Forplot.Fluxin(1,k);
        Fluxesout(k,j+1) = Forplot.Fluxout(1,k);
        Masses1(:,j+1,k) = Forplot.Mass1(:,k);
        Slats(1:grid_maxind,j+1,k) = Forplot.Slat(:,k);
    end

    j = j + 1;
    end

end % end main loop

clec

timevec = [0, Outputtimes];
endrun = clock;
Results.runstart = start;
Results.runend = endrun;
Results.z = Elevations;
Results.Deltaz = DeltaZs;
Results.Mass1 = Masses1;
Results.time = timevec;
Results.Fluxin = Fluxesin;
Results.Fluxesout = Fluxesout;
Results.Slat = Slats;

% call function to evaluate sediment fluxes
Results.flux365 = eval_fluxes(InputStructure, Results, 0.365, pl);
Results.flux725 = eval_fluxes(InputStructure, Results, 0.725, pl);

Results.RMS_MAPE.f365 = error_est(...
    Results.flux365.Flux4comp(:,2), Results.flux365.Flux4comp(:,1));
Results.RMS_MAPE.c365 = error_est(...
    Results.flux365.C4Comp(:,2), Results.flux365.C4Comp(:,1));
Results.RMS_MAPE.f725 = error_est(...
    Results.flux725.Flux4comp(:,2), Results.flux725.Flux4comp(:,1));
Results.RMS_MAPE.c725 = error_est(...
    Results.flux725.C4Comp(:,2), Results.flux725.C4Comp(:,1));

f365 = Results.RMS_MAPE.f365
 c365 = Results.RMS_MAPE.c365
 f725 = Results.RMS_MAPE.f725
 c725 = Results.RMS_MAPE.c725

154
%variable to define plot number
if plotit==1 %plot figures if true
    for i=1:grid_maxind+1
        timearray(i,:)=timevec;
    end

    for i=1:numsteps+1 %offset tc and fs values for clarity in plot
        Sed.tc(1:Sed.posint(i)-1,i)=Sed.tc(1:Sed.posint(i)-1,i)-1.0;
    end

    figure(pn);
    pn=pn+1;
    subplot(3,1,1); %plot suspended concentration time series
    plot(Sed.t/86400,WCm,'-r');
    hold on
    WCm_int=interp1(InputStructure.Csed(:,1),InputStructure.Csed(:,2),Sed.t);
    plot(Sed.t/86400,WCm_int,'-k');
    ylabel('TSS [kg^m^(Adams et al.)]
    set(gca,'ylim',[0 0.5])
    set(gca,'xlim',[0 InputStructure.Tmodel])
    legend('modeled total water column mass','estimated total water column mass','location','northeast')

    subplot(3,1,2); %plot time series of applied skin friction
    plot(time/86400,Sed.tb);
    ylabel('taub [Pa]')
    set(gca,'ylim',[0 0.6])
    set(gca,'ytick',0:0.2:0.6)
    set(gca,'xlim',[0 InputStructure.Tmodel])
    set(gca,'ydir','reverse')
    colorbar('location','NorthOutside')
    xlabel('time (days)')

    subplot(3,1,3); %image plot of critical stress
    imagesc(Sed.t/86400,Sed.delm*1:Sed.n-.3,Sed.tc,[-.2 0.8]);
    ylabel('m [kg^m^(Allison et al.)]')
    set(gca,'ylim',[-0.25 1.5])
    set(gca,'ytick',0:0.25:1.5)
    set(gca,'xdir','reverse')
    set(gca,'xlim',[0 InputStructure.Tmodel])
    colorbar('location','NorthOutside')
xlabel('time (days)')

    figure(pn)
    pn=pn+1;
    plot(Sed.mwcm,'b'); %plot total water column mud content
    hold on %determined by sediment bed model
    plot(WCm,'r-.') %plot total water column mud content
    legend('Sediment bed model','Water column model')
    ylabel('Water Column Mass (kg/m2)')
    title('Conservation of Mass Plot #1: Water Column Mass Agreement')
figure(pn) %mass conservation according to
pn=pn+1; %sediment bed model
plot(Sed.mwcm,'b')
hold on
plot(Sed.mmbedtot,'r')
plot(Sed.mmtot,'k')
ylabel('Mass (kg/m2)')
legend('Water Column Mass','Sediment Bed Mass','Total Mass')
title(['Conservation of Mass Plot #2: Total Bed Mass ... Conservation. Initialized Mass: ',num2str(Sed.mmtot(1)),' kg/m2'])

k=size(Results.Mass1,3);
m=size(Results.Mass1,2);
ModdMass=zeros(m,k);

for j=1:1:k
  for i=1:m
    ModdMass(i,j)=interp1(Results.z(:,i),Results.Mass1(:,i,j),zobs);
    ModdMass2(i,j)=interp1(Results.z(:,i),...
      Results.Mass1(:,i,j),InputStructure.zobs2);
  end
end

figure(pn)
pn=pn+1;
plot(InputStructure.Csed(:,1)/86400,1000*InputStructure.Csed(:,2),'r')
  %plot 0.36 mab measured SSC timeseries
hold on
plot(Results.time/86400,1000*ModdMass(:,1),'b');
  %plot predicted mass fraction 1 SSC timeseries
if k==2
  plot(Results.time/86400,1000*ModdMass(:,2),'g');
  %if 2 sediment fractions, plot 2nd fraction timeseries
  TotMassMod=ModdMass(:,1)+ModdMass(:,2);
  plot(Results.time/86400,1000*TotMassMod,'k') %and plot total SSC timeseries
else
  TotMassMod=1000*ModdMass(:,1);
end
xlabel('Elapsed time (days)')
ylabel('Suspended Sediment Concentration (mg/L)')
title('SSC at Benthic Station during fall deployment storms')
legend('Measured SSC','Modelled SSC 1','Modelled SSC 2','Modelled SSC Total')
xlim([0 ceil(Results.time(end)/86400)])

figure(pn) %scatter plot of measured and modeled SSC values
pn=pn+1;
TotalMassMod=interp1(Results.time/86400,TotMassMod,InputStructure.Csed(:,1)/86400);
pindex=length(InputStructure.tides);
pnan=isnan(TotalMassMod);
pindex=pindex-sum(pnan);
pl=isnan(InputStructure.Csed(1:pindex,2));
p2=isnan(TotalMassMod(1:pindex));
p3=[];

156
for i=1:pindex
    if p1(i)+p2(i)==0
        p3=[p3 InputStructure.Csed(i,2)];
        p4=[p4 TotalMassMod(i)];
    end
end
scatter(1000*p3,1000*p4,'bo')
f1=ezfit(p3,p4,'affine');
showfit(f1);
hold on
oto=[0 150];
plot(oto,oto,'k')
axis square
xlabel('SSC 0.365 mab from ADV (mg/L)')
ylabel('SSC 0.365 mab modelled (mg/L)')
title('Concentration at 0.365 mab at Benthic Station for spring deployment')

figure(pn)
pn=pn+1;
plot(InputStructure.Csed(:,3)/86400,InputStructure.Csed(:,4),'r')
hold on
plot(Results.time/86400,ModdMass2(:,1),'b');
if k==2
    plot(Results.time/86400,ModdMass2(:,2),'g');
end
plot(Results.time/86400,TotMassMod2,'k')
xlabel('time(days)')
ylabel('Suspended Sediment Concentration kg/m3')
title('SSC at 0.725 mab')
legend('Measured SSC','Modelled SSC 1','Modelled SSC 2','Modelled SSC Total')
xlim([0 ceil(Results.time(end)/86400)])

figure(pn) %scatter plot of measured and modeled 0.72 mab SSC
pn=pn+1;
TotalMassMod2=interp1(Results.time/86400,...
    TotMassMod2,InputStructure.Csed(:,3)/86400);
p1=isnan(InputStructure.Csed(1:pindex,4));
p2=isnan(TotalMassMod2(1:pindex));
p3=[];
p4=[];
for i=1:pindex
    if p1(i)+p2(i)==0
        p3=[p3 InputStructure.Csed(i,4)];
        p4=[p4 TotalMassMod2(i)];
    end
end
scatter(p3,p4,'bo')
f2=ezfit(p3,p4,'affine');
showfit(f2);
```matlab
hold on
oto=[0 0.1];
plot(oto,oto,'k')
axis square
xlabel('SSC 0.725 mab from ADV (kg/m3)')
ylabel('SSC 0.725 mab modelled (kg/m3)')

% calculate R^2 values between modeled and measured SSC values
pindex=length(InputStructure.tides);
k=size(Results.Mass1,3);
m=size(Results.Mass1,2);
ModdMass=zeros(m,k);

for j=1:1:k
    for i=1:m
        ModdMass(i,j)=interp1(Results.z(:,i),Results.Mass1(:,i,j),zobs);
        ModdMass2(i,j)=interp1(Results.z(:,i),Results.Mass1(:,i,j),InputStructure.zobs2);
    end
end
if k==2
    TotMassMod=ModdMass(:,1)+ModdMass(:,2);
else
    TotMassMod=ModdMass(:,1);
end
TotalMassMod=interp1(Results.time/86400,...
                      TotMassMod,InputStructure.Csed(:,1)/86400);
pnan=isnan(TotalMassMod);
pindex=pindex-sum(pnan);
cmeas1=InputStructure.Csed(1:pindex,2);
ccalc1=TotalMassMod(1:pindex);
SSE_csed1=sum((cmeas1-ccalc1).^2);
avg_csed1=mean(cmeas1);
SST_csed1=sum((cmeas1-avg_csed1).^2);
R_Squared_365=1-SSE_csed1/SST_csed1
if k==2
    TotMassMod2=ModdMass2(:,1)+ModdMass2(:,2);
else
    TotMassMod2=ModdMass2(:,1);
end
TotalMassMod2=interp1(Results.time/86400,...
                      TotMassMod2,InputStructure.Csed(:,3)/86400);
cmeas2=InputStructure.Csed(1:pindex,4);
ccalc2=TotalMassMod2(1:pindex);
SSE_csed2=sum((cmeas2-ccalc2).^2);
avg_csed2=mean(cmeas2);
SST_csed2=sum((cmeas2-avg_csed2).^2);
R_Squared_725=1-SSE_csed2/SST_csed2
```
else if plotit == -1 % No plotting, just calculate R^2 values
    pindex=length(InputStructure.tides);
    k=size(Results.Mass1,3);
    m=size(Results.Mass1,2);
    ModdMass=zeros(m,k);

    for j=1:1:k
        for i=1:m
            ModdMass(i,j)=interp1(Results.z(:,i),Results.Mass1(:,i,j),zobs);
            ModdMass2(i,j)=interp1(Results.z(:,i),Results.Mass1(:,i,j),InputStructure.zobs2);
        end
    end

    if k==2
        TotMassMod=ModdMass(:,1)+ModdMass(:,2);
    else
        TotMassMod=ModdMass(:,1);
    end
    TotalMassMod=interp1(Results.time/86400,...
        TotMassMod,InputStructure.Csed(:,1)/86400);
    pnan=isnan(TotalMassMod);
    pindex=pindex-sum(pnan);

    cmeas1=InputStructure.Csed(1:pindex,2);
    ccalc1=TotalMassMod(1:pindex);
    SSE_csed1=sum((cmeas1-ccalc1).^2);
    avg_csed1=mean(cmeas1);
    SST_csed1=sum((cmeas1-avg_csed1).^2);
    R_Squared_365=1-SSE_csed1/SST_csed1

    if k==2
        TotMassMod2=ModdMass2(:,1)+ModdMass2(:,2);
    else
        TotMassMod2=ModdMass2(:,1);
    end
    TotalMassMod2=interp1(Results.time/86400,...
        TotMassMod2,InputStructure.Csed(:,3)/86400);
    cmeas2=InputStructure.Csed(1:pindex,4);
    ccalc2=TotalMassMod2(1:pindex);
    SSE_csed2=sum((cmeas2-ccalc2).^2);
    avg_csed2=mean(cmeas2);
    SST_csed2=sum((cmeas2-avg_csed2).^2);
    R_Squared_725=1-SSE_csed2/SST_csed2
end

Results.rsqrd=[Results.flux365.RSqf ...
    Results.flux725.RSqf R_Squared_365 R_Squared_725]
end

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
% End Plotting
% End main program
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
function grid = generategrid(numcells, numfracs)

    Deltaz=1;

    grid.numfracs=numfracs; %number of sediment fractions
    grid.numcells=numcells; %number of gridcells

    %Values set by getHeight:
    grid.Height=0; %Current height of grid
    % (will be set in function getHeight)
    grid.zlowerbound(1:numcells,:)=[0:1:numcells-1]'*Deltaz;
    %lower boundary of each gridcell
    grid.zupperbound(1:numcells,:)=[1:1:numcells]'*Deltaz;
    %upper boundary of each gridcell
    grid.zcenter(1:numcells,:)=(grid.zlowerbound+grid.zupperbound)/2;
    %Center of each gridcell
    grid.Deltaz(1:numcells,:)=Deltaz; %Size of each gridcell

    %Fluxes and scalars calculated at each timestep (stored separately
    %for each sedimentfraction
    for k=1:numfracs
        grid.Fluxin(k)=NaN; %influx across sediment water interface
        % (calculated by function getlowerBC)
        grid.Fluxout(k)=NaN; %settling flux calculated by steptime after
        %each implicit timestep
        grid.Mass1(1:numcells,k)=NaN; %Sediment Concentration
        grid.Slat(1:numcells,k)=NaN; %lateral influx due to handling of
        %tides. Calculated by function getHeight
    end
end

function initgrid = initialize (grid, How, zobs, InputStructure)
array=grid.zcenter;
if strcmp(How, 'Normal')==1 %Starts with a normal distribution
    M1initvalue(1:grid.numcells,1,1:grid.numfracs)=...
    40*exp(-(array-mean(array)).^2/0.4);
elseif strcmp(How, 'Rouse') %initializes all fractions with
    %the same rouseprofile
    height=grid.Height;
    z0=zobs;
    c0=100;
    ws=0.001;
    ustar=0.010;
    M1initvalue(1:grid.numcells,1,1:grid.numfracs)= c0*(((height... 
        -array)./array)*(z0/(height-z0))).^(ws./(0.41*ustar));
elseif strcmp(How, 'Rousespec') %individual rouseprofiles for each
    %sedimentfraction -> recommended
    height=grid.Height;
    z0=zobs;
    c0=interp1(InputStructure.Csed(:,1), InputStructure.Csed(:,2), 0);
ustar=max(0.001, get_ustar(0, InputStructure));
alpha=[InputStructure.parameters.alpha, 0];
cumprod=1;
for i=1:size(alpha,2)
    fractions(i)=cumprod*(1-alpha(i));
    cumprod=cumprod*alpha(i);
end
values=InputStructure.Cini;
for k = 1:1:grid.numfracs
    if k==3
        mult=50;
    else
        mult=1;
    end
    Concentration=values(k);%fractions(k)*mult*c0;
    ws=getWs(Concentration, k, InputStructure);
    if i==1 %assumes that first fraction is fastest settling one
        beta=min(InputStructure.parameters.diff_beta,...
            0.63*exp(-247.6*ustar)+0.54);
    else
        beta=InputStructure.parameters.diff_beta;
    end
    M1initvalue(1:grid.numcells,k)=Concentration*(((height-
        array)./array)*(z0/(height-z0))).^(ws./(0.41*beta*ustar));
end
elseif strcmp(How,'Zero')
    M1initvalue(1:grid.numcells,1,1:grid.numfracs)=0;
else
    error('exception in function initialize');
end
initgrid=grid;
initgrid.Mass1(:,1:grid.numfracs)=M1initvalue;
end

function ExtData=extractgrid(grid) %Extracts Data from Grid for storage
    for k=1:grid.numfracs
        ExtData.Mass1(:,k)= grid.Mass1(:,k);
        ExtData.Slat(:,k)=grid.Slat(:,k);
        ExtData.Fluxin(1,k)= grid.Fluxin(k);
        ExtData.Fluxout(1,k)= grid.Fluxout(k);
    end
    ExtData.elevation(:,1)=grid.zcenter;
    ExtData.Deltaz(:,1)=grid.Deltaz;
    %Values at top of domain. Neccessary for nice concentration plots
    ExtData.Deltaz(grid.numcells+1,1)=0;
    ExtData.Mass1(grid.numcells+1,1:k)=0;
    ExtData.elevation(grid.numcells+1,1)=grid.Height;
end

function gridnew=steptime(gridold,DeltaT,stepnum, InputStructure,FintoWC)
gridnew = getheight(gridold, stepnum*DeltaT, InputStructure);  % set new waterlevel
gridold = gridnew;
for i = 1:gridold.numfracs % Number of different classes
  % Right Hand Side
  RHS = assembleRHS(gridnew, gridold, DeltaT, stepnum, ... 
    i, InputStructure, FintoWC);

  % Left Hand Side
  LHS = assembleLHS(gridnew, gridold, DeltaT, stepnum, ... 
    i, InputStructure);

  % Solve
  gridnew.Mass1(:,i) = (RHS / LHS)';

  % Influx from sediment
  gridnew.Fluxin(i) = getlowerBC(stepnum*DeltaT, i, ... 
    InputStructure, FintoWC); % from Sediment

  % Calculate settling Flux
  wslow = getWs(gridold.Mass1(1,i), i, InputStructure);
  gridnew.Fluxout(i) = wslow * gridnew.Mass1(1,i);
end
end

function RHS = assembleRHS(gridnew, gridold, DeltaT, stepnum, ... 
  Whichmass, InputStructure, FintoWC)
  RHS = gridold.Deltaz.*gridold.Mass1(:,Whichmass);
  qlow = getlowerBC((stepnum)*DeltaT, Whichmass, InputStructure, FintoWC);
  qup = getupperBC((stepnum)*DeltaT, Whichmass);
  RHS(1,:) = RHS(1,:) - qlow*DeltaT;
  RHS(gridold.numcells,:) = RHS(gridold.numcells,:) - qup*DeltaT;
  RHS = RHS';
end

function LHS = assembleLHS(gridnew, gridold, DeltaT, stepnum, ... 
  Whichmass, InputStructure)
  % Get Dispersion Coefficients and Settling Velocities
  ustar = get_ustar(DeltaT*stepnum, InputStructure);
  Dup_arr = getD(gridnew.zupperbound, DeltaT*stepnum, gridnew.Height, ... 
    ustar, Whichmass, InputStructure);
  Dlow_arr = getD(gridnew.zlowerbound, DeltaT*stepnum, gridnew.Height, ... 
    ustar, Whichmass, InputStructure);

  wsup_arr(1:gridnew.numcells,1) = NaN;
  wsup_arr(1:gridnew.numcells-1) = getWs((gridnew.Mass1(1:gridnew.numcells ... 
    -1,Whichmass).*gridnew.Deltaz(1:gridnew.numcells-1) ... 
    +gridnew.Mass1(2:gridnew.numcells,Whichmass) ... 
    numcells)+gridnew.Deltaz(2:gridnew.numcells)), ... 
    Whichmass, InputStructure);
  wslow_arr(1:gridnew.numcells,1) = NaN;
  wslow_arr(2:gridnew.numcells) = getWs((gridnew.Mass1(2:gridnew.numcells,...
Whichmass).*gridnew.Deltaz(2:gridnew.numcells)+gridnew.Mass1... 
(1:gridnew.numcells-1,Whichmass).*gridnew.Deltaz(1:gridnew.numcells... 
-1))./(gridnew.Deltaz(2:gridnew.numcells)+gridnew.Deltaz... 
(1:gridnew.numcells-1)),Whichmass,InputStructure);

%Calculate distances between gridcellcenters (remember, we have cells 
%with variable size)
DeltazUpF(1:gridnew.numcells,1)=NaN;
DeltazUpF(1:gridnew.numcells-1)=(gridnew.Deltaz...
(1:gridnew.numcells-1)+gridnew.Deltaz(2:gridnew.numcells))/2;
DeltazLowF(1:gridnew.numcells,1)=NaN;
DeltazLowF(2:gridnew.numcells)=(gridnew.Deltaz...
(1:gridnew.numcells-1)+gridnew.Deltaz(2:gridnew.numcells))/2;

%set up main diagonal for C(i)
AF(2:gridnew.numcells,1)=gridnew.Deltaz(2:gridnew.numcells)... 
+(Dup_arr(2:gridnew.numcells)./DeltazUpF(2:gridnew.numcells)... 
+Dlow_arr(2:gridnew.numcells)./DeltazLowF(2:gridnew.numcells))*DeltaT;
AF(1)=gridnew.Deltaz(1)+(Dup_arr(1)./DeltazUpF(1)+wsup_arr(1)/2)*DeltaT;
%FreedrainBC already taken into consideration:... +wsup/2 otherwise (no 
%free drainage): A=...-wsup/2)*DeltaT;
AF(gridnew.numcells,1)=gridnew.Deltaz(gridnew.numcells)+...
(Dlow_arr(gridnew.numcells)/DeltazLowF(gridnew.numcells)+...
wslow_arr(gridnew.numcells))/2)*DeltaT;

%set up first side diagonal for C(i+1)
CF(1:gridnew.numcells-1,1)=(-Dup_arr(1:gridnew.numcells-1)./DeltazUpF...
(1:gridnew.numcells-1)+wsup_arr(1:gridnew.numcells-1)/2)*DeltaT;

%set up second side diagonal for C(i-1)
BF(1:gridnew.numcells-1,1)=(-Dlow_arr(2:gridnew.numcells)./DeltazLowF...
(2:gridnew.numcells)-wslow_arr(2:gridnew.numcells)/2)*DeltaT;

%Create tridiagonal matrix
LHS=diag(AF)+diag(CF,+1)+diag(BF,-1);
end

function Influx=getlowerBC(steptime, Whichmass, InputStructure,FintoWC)
%firstgridcell already removed
%ModeM1=0-> no influx
ModeM1=1;
%evaluate efficiency and alphas
%first step: convert alpha values in distributions. The use of alpha is
%necessary to ensure that the overall cumulative distribution equals
%one when fitting routines are used. The calculation of alphas from
%distributions is found e.g. in the file ttestalphafinding.m (For all
%who deal with less than three sedimentfractions: Don't bother!)
if isstruct(InputStructure)
    %calculate sediment input as Jin=eff*J sed measured. Jin,
    %first = Jin*alpha, second Jin, second= Jin(1-alpha)
    alpha=[InputStructure.parameters.alpha, 0];
    cumprod=1;
end
for i=1:size(alpha,2)
    fractions(i)=cumprod*(1-alpha(i));
    cumprod=cumprod*alpha(i);
end
fract=fractions(Whichmass);
else
    amplify=1;
end
if ModeM1==0
    Influx=0;
else
    if nargin==4 && isstruct(InputStructure)==1;
%meastime=InputStructure.fluxes(:,1);
%measflux=InputStructure.fluxes(:,2);
    SWIFlux=-FintoWC*fract;%-amplify*interp1...
        (meastime, measflux, steptime);
    elseif nargin==2 || isstruct(InputStructure)==0;
        SWIFlux=-(0.1*amplify*cos(2*pi()/43200*steptime+43200/2).^2);
    else
        error('Invalid number of arguments in function getlowerBC');
    end
    Influx=(SWIFlux);
end
end

function Outflux=getupperBC(steptime, Whichmass)
    Outflux=0;
end

function D=getD(z,time,height,ustar,Whichmass,InputStructure)
ModeD='Rouse';
kappa=0.41;
if strcmp(ModeD,'Rouse')
    if Whichmass==1 %assumes that fist fraction is fastest settling one
        beta=min(InputStructure.parameters.diff_beta,...
            0.63*exp(-247.6*ustar)+0.54); %see JGR Paper
    else
        beta=InputStructure.parameters.diff_beta;
    end
    thresh=0.0001;
D=kappa*ustar*beta*z.*(1-z/height);
%If oscillations occur, it might be helpful to uncomment following
%lines:
    ind=find(D<thresh);
a=-1/height;
b=1;
c=-tresh/(kappa*ustar);
dist=real((-b+sqrt(b^2-4*a*c))/(2*a));

D(ind)=max(D(ind), tresh/2+D(ind)/2.*min(z(ind),
    height-z(ind))/dist);

%step neccessary to keep Grid Peclet u*Deltax/D<=2

else
    D(1:size(z,1),:)=0.01;%0.01;%0.01;
end
end

%concentration is currently not used. This might change if ws=kCSed may be
%interesting
function ws=getWs(Concentration, Whichmass, InputStructure)
    if isstruct(InputStructure)
        ws=InputStructure.parameters.ws(Whichmass);
        %currently only first order
    else
        ws=0.001;
    end
end

%get shear velocity
function ustar=get_ustar(time, InputStructure)
    %Input=0;
    Mode='Whatever';

    if nargin==2 && isstruct(InputStructure)==1
        meastime=InputStructure.shears(:,1);
        ustarmeas=InputStructure.shears(:,2);
        ustar=interp1(meastime,ustarmeas,time,'linear');
    elseif nargin==1 || isstruct(InputStructure)==0
        if strcmp(Mode,'Sine')==1
            ustar=abs(0.015*cos(2*pi()/43200*time+43200/2));
        else
            ustar=0.01;%0.01;
        end
    end
end

%this function sets the tidal elevation at each timestep and handles the
%grid stretching
function gridnew=getheight(gridold, time, InputStructure)
    gridnew=gridold;
    numcells=gridold.numcells;
if nargin==3 & isstruct(InputStructure)==1
    meastime=InputStructure.tides(:,1);
    heightmeas=InputStructure.tides(:,2);
    actualelev=interp1(meastime, heightmeas, time);
    z0=InputStructure.z0;
elseif nargin==2 || isstruct(InputStructure)==0
    %SPECIFY ELEVATION DEVELOPMENT HERE!
    actualelev=5-2.5/2+2.5/2*cos(2*pi()/43200*time);
    z0=2E-5;
    InputStructure=0; %For ustar determination
end
ustar=get_ustar(time, InputStructure);

ChangeHeight=actualelev-gridold.Height;

%calculate flow velocity profile
U=ustar/0.41*(log(gridold.zcenter)-log(z0));

%Add "volume" to gridcells
if ustar==0
    weight=U.*gridold.Deltaz/(sum(U.*gridold.Deltaz));
else
    weight=gridold.Deltaz/sum(gridold.Deltaz);
end
AddDz=weight*ChangeHeight;
if time==0
    gridnew.Deltaz=gridold.Deltaz.*AddDz;
else
    gridnew.Deltaz=gridold.Deltaz+AddDz;
end

%recalculate grid parameters
gridnew.zupperbound=cumsum(gridnew.Deltaz);
gridnew.zlowerbound=[0; gridnew.zupperbound(1:gridnew.numcells-1,1)];
gridnew.zcenter=(gridnew.zlowerbound+gridnew.zupperbound)./2;

%recalculate Masses
if time==0
    %NEW VERSION:
    for k=1:gridnew.numfracs
        gridnew.Mass1(:,k)=max(interp1(gridold.zcenter, gridold.Mass1(:,k), gridnew.zcenter, 'spline', 'extrap'),0);
        %gridnew.Mass1(:,k)=gridold.Mass1(:,k);
        %gridnew.Slat(:,k)=AddDz.*gridold.Mass1(:,k);
    end
end
gridnew.Height=actualelev;
end
function sed=sed_bed_init(Timeinterval, numsteps, InputStructure,...
        DeltaT, NGridCell, SedVar, grid, time)

%Variables that are defined in the imported SedVar Structure
%These variables are changeable
sed.n=SedVar.n;
        sed.delm=SedVar.delm;
        sed.Rhos=SedVar.Rhos;
        sed.Dmix=SedVar.Dmix;
        sed.beta=SedVar.beta;
        sed.conrate=SedVar.conrate;
        sed.exprate=SedVar.exprate;
        sed.tceqn=SedVar.tceqn;
        sed.tcmin=SedVar.tcmin;
        sed.phismin=SedVar.phismin;

sed.t=0:DeltaT:Timeinterval;              %define time vector, days
sed.int=floor(sed.n/2);                     %index of the interface layer
sed.posint=sed.int*ones(1,numsteps+1);  %initialize vector of
                                          %interface position
sed.m=sed.delm*ones(sed.n,1);              %initialize vector of layer masses
sed.m(1:sed.int-1,1)=0;                     %zero mass in layers above the
                                          %interface
sed.mcum=min(cumsum(sed.m),circshift(cumsum(sed.m),1));
                                          %define vector of accumulated sediment masses
sed.mint=sed.m(sed.int,1)*ones(1,numsteps+1);
                                          %initialize time series of interface layer mass
sed.mintm=sed.mint;
                                          %initialize mass of mud in int layer time series
sed.dlayer=(sed.m(1:sed.n-1)+sed.m(2:sed.n))/2;
                                          %define vector of mass differences between layers
sed.mmix=sed.delm*ones(sed.n,1);
                                          %initialize vector of mixing induced mass transport
sed.Dmix=sed.Dmix*(0.075*sed.Rhos)^2;
                                          %Mixing/Bioturbation coefficient in (kg/m^2)^2/s
                                          %for phis=0.075
sed.tmix=0;                             %Initialize Mixing time counter
sed.mixcntr=0;                          %Initialize mixing event counter
sed.tceq=eval(sed.tceqn); %equilibrium critical shear stress equation, N/m2
sed.tc=ones(sed.n,numsteps+1);
       %initialize tauc matrix to one
sed.tc(:,1)=sed.tceq.*sed.tc(:,1);
       %set first tauc profile to equilibrium profile

c1=SedVar.c1;              %calculate initial critical shear stress
c2=SedVar.c2; %for erosion profile
c3=SedVar.c3;
mt=0;

for tc_cntr=0:1:50
    sed.tc(sed.int+tc_cntr,1)=c1+c2*(mt^c3);
    mt=mt+0.05;
end

sed.tc(1:sed.int-1,1)=0; %define tau-crit above the interface as 0

sed.alpha=zeros(sed.n,numsteps+1); %initialize alpha (gradient of tau_c) matrix
sed.alpha(sed.int:sed.n-1,1)=diff(sed.tc(sed.int:sed.n,1))... 
./sed.m(sed.int:sed.n-1,1);
    %alpha profile is forward diff of tau_c
sed.alpha(sed.n,1)=sed.alpha(sed.n-1,1);
    %alpha at last point equal to value at next to last point
sed.alpha(1:sed.int-1,1)=0;

sed.b=ones(1,numsteps+1); %initialize erosion mode parameter for mud

sed.Dm=zeros(1,numsteps+1); %mud deposition array
sed.Em=zeros(1,numsteps+1); %mud erosion array
sed.DMEm=sed.Dm-sed.Em; %mud net deposition array
sed.mwcm=zeros(1,numsteps+1); %suspended mud mass

WCtot=0;
for wci=1:NGridCell %loop through water column concentrations
    WCtemp=sum(grid.Mass1(wci,:)*grid.Deltaz(wci)); % to calculate total
    WCtot=WCtot+WCtemp; %sediment in the water column according
    %to the water column model.units of kg/m2.
end

sed.mwcm(1)=WCtot; %initialize water column mud mass of sediment model
sed.WCm=WCtot; %to match water column model.

sed.mmtot=zeros(1,numsteps+1); %initialize total mud mass in wcol and seds
sed.mmtot(1)=sum(sed.m(:,1))+sed.mwcm(1); %initialize large buffer layer at bottom

sed.mbot=3*(sed.mmtot(1)); %"distance" between bottom layer and buffer layer centers

meastime=InputStructure.shearwind(:,1); %get time series of bed shear stress, tb N/m2
ustarmeas=InputStructure.shearwind(:,2);

sed.ustar=interp1(meastime,ustarmeas,time,'linear');
sed.tb=sed.ustar.*sed.ustar.*1000;

sed.A=max(diff(sed.tb)/DeltaT,0); %calculate positive dtaub/DeltaT values
sed.A(numsteps+1)=sed.A(numsteps);

sed.phis=max(((sed.tc(sed.int,1)+sed.tc(sed.int+1,1))/2/9.3)^0.93,...
    sed.phismin)*ones(1,numsteps+1);
Calculate a reasonable profile of mud solids fraction

```
sed.Rhod = sed.Rhos * phistot(0, sed.phis(1,1)) * ones(1, numsteps+1);

% Sediment dry density in interface layer

sed.Mmud = sed.Rhod * sed.beta;
```

Sediment dry density in interface layer

```
% Calculate initial mud erosion rate and deposition rate
if (sed.alpha(sed.int,1) * sed.Mmud(1,1)) == 0
    % no vertical gradient in tauc, simple linear erosion
    sed.Em(1) = sed.Mmud(1,1) * (sed.tb(1) - sed.tc(sed.int,1)) * H(sed.tb(1) - sed.tc(sed.int,1));
else
    % otherwise, use erosion function derived from Sanford and Maa (2001)
    sed.b(1) = (1 - exp(-sed.alpha(sed.int,1) * sed.Mmud(1,1) * DeltaT)) / (sed.alpha(sed.int,1) * sed.Mmud(1,1) * DeltaT);
    sed.Em(1) = (sed.A(1) / sed.alpha(sed.int,1) * (1 - sed.b(1)) + sed.b(1) * DeltaT);
    sed.Mmud(1,1) = (sed.tb(1) - sed.tc(sed.int,1)) * H(sed.tb(1) - sed.tc(sed.int,1));
end
```

```
% initial erosive flux. store to pass to water column model.
sed.FintoWC = sed.Em(1);
```

```
% initial flux of sediments out of water column model to bed. Set initial condition as zero
sed.Dm(1) = 0;
```

```
% calculate difference between present tc profile and equilibrium tc profile
sed.difftc = sed.tceq - sed.tc(:, q-1);
```

Bed mixing section (due to bioturbation)

```
sed.tmix = sed.tmix + DeltaT;
% calculate time since last layer exchange due to mixing
sed.mmix(1:sed.int-1) = 0;
```

```
sed.mmix(sed.int:sed.n) = sed.Dmix * sed.tmix ./ sed.dlayer(sed.int:sed.n);
```

```
if sed.mmix(sed.int) >= 0.3 * sed.m(sed.int)
    % mix layer masses if potential mixed mass is larger than a significant fraction of the interface layer mass
    sed.tmix = 0;
    sed.mixcntr = sed.mixcntr + 1;
end
```

```
sed.mintm(q) = sed.mintm(q) + sed.dmm;
```

```
sed.m(sed.int) = sed.mint(q);
```

```
sed.mintm(q) = sed.mintm(q);  % total mud mass in interface layer
```

```
sed.dm = sed.dmm;  % total mass in model
```

```
sed.dmm = sed.DMEm(q-1) * DeltaT;  % change in mud mass
```

```
sed.dm = sed.dmm;  % change in total mass
```

```
sed.difftc = sed.tceq - sed.tc(:, q-1);  % calculate difference between present tc profile and equilibrium tc profile
```

```
sed.mmix = sed.mmix + DeltaT;
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sed.tmix = sed.tmix + DeltaT;
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sed.mmix = sed.mmix + DeltaT;
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sed.mmix = sed.mmix + DeltaT;
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sed.mmix = sed.mmix + DeltaT;
```

```
sed.mmix = sed.mmix + DeltaT;
```
%Adjust masses due to sediment mixing throughout sediment bed
% mix tc profile and relax it towards equilibrium profile for layers
% beneath interface layer; note different rates of relaxation for
% consolidation and expansion, N/m^2
sed.tc(sed.int+1:sed.n,q)=sed.tc(sed.int+1:sed.n,q-1)...
+sed.mmix(sed.int:sed.n-1).*sed.alpha(sed.int+1:sed.n,q-1)...
-sed.mmix(sed.int:sed.n-1).*sed.alpha(sed.int:sed.n-1,q-1)...
+DeltaT*sed.difftc(sed.int+1:sed.n).*{(sed.conrate*...
H(sed.difftc(sed.int+1:sed.n))...
+sed.exprate*H(-sed.difftc(sed.int+1:sed.n)))};

else  %if no mixing, just relax tc profile, N/m^2
sed.tc(sed.int+1:sed.n,q)=sed.tc(sed.int+1:sed.n,q-1)...
+DeltaT*sed.difftc(sed.int+1:sed.n).*{(sed.conrate*...
H(sed.difftc(sed.int+1:sed.n))...
+sed.exprate*H(-sed.difftc(sed.int+1:sed.n)))};
end

%adjust interface tc to account for net deposition or erosion and relax it
%towards equilibrium
% net mud deposition means we have to account for both relaxation of
% previous tc and relaxation of added material with low tc
% computation is done to keep tc(int+1) at its set value and also
% conserve the interface layer average tc, N/m^2
if sed.dmm > 0
    sed.tcintrelax=sed.tc(sed.int,q-1)+DeltaT*sed.difftc(sed.int).*{(...
        (sed.conrate*H(sed.difftc(sed.int)))...
        +sed.exprate*H(-sed.difftc(sed.int)));
    sed.difftcmin=sed.tceq(sed.int)-sed.tcmin;
    sed.tcminrelax=sed.tcmin+DeltaT/2*sed.difftcmin...
        *(sed.conrate*H(sed.difftcmin));
    sed.tc(sed.int,q)=sed.tcintrelax-sed.dmm/sed.mintm(q)...
        *(sed.tcintrelax+sed.tc(sed.int+1,q)-2*sed.tcminrelax);
else  %net mud erosion
    sed.tc(sed.int,q)=sed.tc(sed.int,q-1)-sed.dm*...
        (sed.alpha(sed.int,q-1) +DeltaT*sed.difftc(sed.int)...*
            (sed.conrate*H(sed.difftc(sed.int))...) +sed.exprate*H(-sed.difftc(sed.int)));
end
sed.tc(sed.int:sed.n,q)=max(sed.tcmin,sed.tc(sed.int:sed.n,q));
%limit minimum value of all taucs
sed.alpha(sed.int:sed.n-1,q)=diff(sed.tc(sed.int:sed.n,q))...
    ./sed.m(sed.int:sed.n-1); %estimate new alphas
sed.alpha(sed.n,q)=sed.alpha(sed.n-1,q);

%Now change interface layers if necessary
if sed.m(sed.int) > 1.9*sed.delm  %split interface layer if too thick
    if sed.int==1
        sed.m(sed.int)=NaN;
        er=('interface = 0')
        return
    end
    sed.int=sed.int-1;
    sed.m(sed.int)=sed.m(sed.int+1)-sed.delm;
    sed.mint(q)=sed.m(sed.int);
end
sed.tc(sed.int,q)=sed.tc(sed.int+1,q);
sed.alpha(sed.int,q)=sed.alpha(sed.int+1,q);
sed.mintm(q)=sed.mint(q);
sed.m(sed.int+1)=sed.delm;

sed.tc(sed.int+1,q)=sed.tc(sed.int+1,q)+sed.m(sed.int)*sed.alpha(sed.int,q);
if sed.m(sed.int)>sed.delm
    sed.m(sed.int)=NaN;
    error('flux to bed to large')
    return
end

if sed.m(sed.int) < 0.1*sed.delm %combine interface layer with the one below if it is too thin
    sed.int=sed.int+1;
    sed.m(sed.int)=sed.m(sed.int)+sed.m(sed.int-1);
    sed.mint(q)=sed.m(sed.int);
    sed.tc(sed.int,q)=sed.tc(sed.int-1,q);
    sed.alpha(sed.int,q)=(sed.tc(sed.int+1,q)...
        -sed.tc(sed.int,q))/sed.mint(q);
    sed.mintm(q)=sed.mint(q);
    sed.m(sed.int-1)=0;
    sed.tc(sed.int-1)=0;
end

%ensure that layers above interface are zeroed
sed.tc(1:sed.int-1,q)=0;
sed.alpha(1:sed.int-1,q)=0;
sed.m(1:sed.int-1)=0;

sed.posint(q)=sed.int; %record the interface position
sed.mcum=min(cumsum(sed.m),circshift(cumsum(sed.m),1)); %recalculate cumulative mass vector
sed.dlayer(1:sed.n-1)=(sed.m(1:sed.n-1)+sed.m(2:sed.n))/2;
sed.dlayer(sed.n)=(sed.m(sed.n)+sed.mbot)/2; %redefine vector of mass differences between layers

sed.tceq=eval(sed.tceqn);

%Now that all layer adjustments have been made, %calculate erosion and deposition of mud for this time step
sed.mwcm(q)=sed.mwcm(q-1)-sed.dmm; %calculate change in water column mud mass
sed.phis(q)=max(((sed.tc(sed.int,q)+sed.tc(sed.int+1,q))... 
    /2/9.3)^0.93,sed.phismin); %Calculate a reasonable value of mud solids fraction in int layer

sed.Rhod(q)=sed.Rhos*phistot(0,sed.phis(q)); %Sediment dry density in interface layer
sed.Mmud(q)=sed.Rhod(q)*sed.beta; %calculate value of mud erosion rate const, s m-1

if (sed.alpha(sed.int,q)*sed.Mmud(q)) == 0 %calculate mud erosion rate, based on Sanford and Maa (2001)
sed.Em(q)=sed.Mmud(q)*(sed.tb(q)-sed.tc(sed.int,q))... 
    *H(sed.tb(q)-sed.tc(sed.int,q));
else
    sed.b(q) = (1 - exp(-sed.alpha(sed.int,q)*sed.Mmud(q)*DeltaT)) ... 
        /(sed.alpha(sed.int,q)*sed.Mmud(q)*DeltaT);
    sed.Em(q) = (sed.A(q)/sed.alpha(sed.int,q)*(1 - sed.b(q)) ... 
        + sed.b(q)*sed.Mmud(q)*(sed.tb(q) - sed.tc(sed.int,q)) ... 
        * H(sed.tb(q) - sed.tc(sed.int,q)));
end

% Store sediment fluxes
sed.FintoWC = sed.Em(q);  % Flux into water column, kg m^-2 s^-1
sed.Dm(q) = sum(grid.Fluxout);  % all fractions from water column model
sed.DMEm(q) = max(sed.Dm(q) - sed.Em(q), -0.99*sed.mintm(q)/DeltaT);
    % limit net erosion to 1st layer mud content

sed.mmbedtot(q) = sum(sed.m(:,1)) + sed.mbot;
    % sum up total masses to check mass conservation
sed.mmtot(q) = sum(sed.m(:,1)) + sed.mwcm(q) + sed.mbot;
end
H.m
function y=H(x)
% more stable implementation of the heaviside function than in MATLAB
% defined as H=1 for x(i)>0, not just for x(i)>0
% x may be either a scalar or a vector
y=(x>=0);

phistot.m
function y=phistot(fs,phis)
% calculate total solids volume fraction for an arbitrary mixture of sand
% with phis, sand=0.55 and phis, mud input in pairs with the sand fraction
ii=find(fs>0.6);
y=1./(fs+(1-fs)./phis);
y(ii)=min(y(ii),0.55./fs(ii));
eval_fluxes.m

function dist=eval_fluxes(InputStructure, Modelresults, elevation, pl)
    %Index: Index of cell used for evaluation e.g. 75 cm = 8, 55 cm = 5
    %cm =3, 35 cm = 4 at 10 cm cellsize.
    %function compares calculated fluxes
    %next steps on monday: interpolate model and measurements for
    %comparison! This will be neccessary for the fitting routine.
    %Index=4;
    plotit=pl;
    plotselection=0;
    z=elevation;%Modelresults.z(Index,1);
    DeltaZ=0.05;  %resolution for gradient calculation.
    %Values will be interpolated
    height=getheight(Modelresults.time, InputStructure);
    KDs=getmixing(Modelresults, InputStructure, height,z)';
    [C, DeltaC]=getgradientandC(Modelresults, elevation, DeltaZ);
    fluxind=-KDs.*DeltaC.Ind;
    flux=sum(-KDs.*DeltaC.Ind,1);
    Concentration=C.all;
    % for k=1:size(Modelresults.Mass1,3)
    %       Concentration=Concentration+Modelresults.Mass1(Index,:,k);
    % end
    Timemeas=InputStructure.Csed(:,1);
    Cmeas=InputStructure.Csed(:,2);
    Fluxmeas=InputStructure.fluxes(:,2);
    %avoids NaNs in distance
    lastmeas=find(Timemeas'<max(Modelresults.time),1,'last');
    Ccalc=interp1(Modelresults.time, Concentration, Timemeas(1:lastmeas)')';
    Fluxcalc=interp1(Modelresults.time, flux, Timemeas(1:lastmeas)')';
    Fluxdist=Fluxmeas(1:lastmeas)-Fluxcalc;
    Cdist=Cmeas(1:lastmeas)-Ccalc;
    Weight=sum(Cmeas)/((sum(Fluxmeas)*2));
    dist.Time=Timemeas(1:lastmeas);
    dist.Flux=Fluxdist;
    dist.Csed=Cdist;
    dist.Weight=Weight;
    dist.CInd=C.Ind;
    dist.Jind=fluxind';
    dist.C4Comp=[Ccalc Cmeas(1:lastmeas)] ;
    dist.Fluxval=flux;
    dist.Flux4comp=[Fluxcalc Fluxmeas(1:lastmeas)];
    measflux=interp1(InputStructure.fluxes(:,1)/(3600*24),...
\begin{verbatim}
InputStructure.fluxes(:,2), Modelresults.time/86400); 
fmeas=measflux; 
fcalc=flux; 
SSE_fsed=sum((fmeas-fcalc).^2); 
avg_fsed=mean(fmeas); 
SST_fsed=sum((fmeas-avg_fsed).^2); 
R_Sq_f=[1-SSE_fsed/SST_fsed elevation]

dist.RSqf=R_Sq_f(1); 

if plotit==1
  h=figure; 
  %subplot(2,1,1); 
  plot(Modelresults.time/(3600*24), flux,'-k', 
       InputStructure.fluxes(:,1)/(3600*24), InputStructure.fluxes(:,2),'-r'); 
  title(['Fluxes and Gradients ',elevation,' mab']); 
  legend('modeled','measured'); 
  ylabel('Fluxes (kg/m2.s)'); 
  xlabel('Time (days)') 
  FluxAx(1)=gca; 
end
end

function height=getheight(time, InputStructure)
  % numcells=size(gridold,1); 
  %Algorithm was getting instable for heightdifferences 
  if nargin==2
    meastime=InputStructure.tides(:,1); 
    heightmeas=InputStructure.tides(:,2); 
    height=interp1(meastime, heightmeas, time); 
  end
end

function KDs=getmixing(Modelresults, InputStructure, height,z)
  time=Modelresults.time; 
  meastime=InputStructure.shears(:,1); 
  ustarmeas=InputStructure.shears(:,2); 
  ustar=interp1(meastime,ustarmeas,time,'linear'); 
  for k=1:size(Modelresults.Mass1,3)
    if k==1 %assumes that first fraction is fastest settling one 
      beta=0.460; 
      %beta=min(1, 0.63*exp(-247.6*ustar)+0.54); 
    else
      beta=1; 
    end
    kappa=0.41; 
    tresh=0.0005; 
    %ustar=get_ustar(time,InputStructure); 
    D=kappa*ustar*z.*(1-z./height); 
  end
end
\end{verbatim}
if D<thresh
    a=-1/height;
    b=1;
    c=-thresh/(kappa*ustar);
    dist=real((-b+sqrt(b^2-4*a*c))/(2*a));
    D=max(D, tresh/2+D/2*min(z, height-z)/dist);
end

KDs(:,k)=D.*beta;
end

function [C, DeltaC]=getgradientandC(Modelresults, z, DeltaZ)
    xi=[z-DeltaZ, z, z+DeltaZ]';
    Values=zeros(3,size(Modelresults.Mass1,2));
    Gradient=zeros(size(Modelresults.Mass1,2),size(Modelresults.Mass1,3));
    CI=Gradient;
    for k=1:size(Modelresults.Mass1,3)
        for i=1:size(Modelresults.Mass1,2)
            Values(:,i)=interp1(Modelresults.z(:,i),Modelresults.Mass1(:,i,k),xi,...
            'spline');
            Gradient(:,k)=(Values(3,:)-Values(1,:))/(2*DeltaZ);
            CI(:,k)=Values(2,:);
        end
        Totalgradient=sum(Gradient,2);
        TotalC=sum(CI,2);
    end
    C.Ind=CI;
    C.all=TotalC;
    DeltaC.Ind=Gradient';
    DeltaC.all=Totalgradient';
end
```matlab
function RMS_MAPE=error_est(xmeas, xmodel)
    MAPE_temp=abs((xmeas-xmodel)./xmeas); %calculate the absolute percentage error at each model point
    p=isnan(MAPE_temp);
    r=isinf(MAPE_temp);
    MAPE_sum=0;
    M=length(p);
    if sum(p)+sum(r)~=0
        for i=1:length(p)
            if p(i)==0 && r(i)==0
                MAPE_sum=MAPE_sum+MAPE_temp(i);
            else
                M=M-1;
            end
        end
    end
    MAPE_sum=sum(MAPE_temp);
    RMS_MAPE.MAPE=MAPE_sum/M; %remove any NaN in order to calculate the mean absolute percentage error (MAPE)
    RMS_MAPE.M=M;
    RMS_temp=(xmeas-xmodel).^2; %calculate square error between measured and modeled values
    q=isnan(RMS_temp);
    s=isinf(RMS_temp);
    RMS_sum=0;
    N=length(q);
    if sum(q)+sum(s)~=0
        for i=1:length(q)
            if q(i)==0 && s(i)==0
                RMS_sum=RMS_sum+RMS_temp(i);
            else
                N=N-1;
            end
        end
    end
    RMS_sum=sum(RMS_temp);
    RMS_MAPE.RMS=(RMS_sum/N)^0.5; %removed NaN in order to calculate root mean square error
    RMS_MAPE.N=N;
    mxmeas=mean(xmeas); %mean value, model 'g' for skill score
    MSEg=(xmeas-mxmeas).^2; %square error for model 'g'
    t=isnan(MSEg);
    u=isinf(MSEg);
    MSEg_sum=0;
    L=length(t);
    if sum(t)+sum(u)~=0
        for i=1:length(t)
            if t(i)==0 && t(i)==0
                MSEg_sum=MSEg_sum+MSEg(i);
            else
                L=L-1;
            end
        end
    end
    MSEg_sum=sum(MSEg); %remove NaN. calc mean square error for 'g'
```
RMS\_MAPE.\text{SS}=1-\frac{(\text{RMS\_sum}/N)}{(\text{MSE\_g\_sum}/L)}; \ %\text{calculate skill score}
RMS\_\text{MAPE.}\text{L}=L;
end