Controls on Erodibility in a Partially Mixed Estuary, York River, Virginia

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Controls on Erodibility in a Partially Mixed Estuary: York River, Virginia

A Thesis
Presented to
The Faculty of the School of Marine Science
The College of William and Mary

In partial fulfillment
of the requirements for the degree of
Master of Science

by
Patrick J. Dickhudt
2008
Approval Sheet

This thesis is submitted in partial fulfillment of the requirements for the degree of Master of Science

July 2008

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Abstract

A time-series of erodibility measurements and physical and biological sediment properties was used to evaluate spatial and temporal variability in cohesive bed erodibility and controls on erodibility in the York River estuary, VA. Two sites near Clay Bank displayed dramatic seasonal variations in bed erodibility while a third near Gloucester Point displayed a more consistent level of low erodibility. Total bed solids fraction and cohesive sediment grain size were not correlated with bed erodibility. The surficial sediments were characteristically composed of 2% to 50% sand supported in a mud matrix. The total solids fraction of the bed was shown to be a function of sand fraction, complicating its use as a measure of bed consolidation. The solids fraction of the mud matrix was calculated to evaluate changes in consolidation independent of sand content. An evaluation of data from multiple studies available in the literature demonstrated that when a large range of data was considered, the solids fraction of the mud matrix provided a better correlation to erodibility than did the total solids fraction. However, the range in solids fraction of the mud matrix in York River sediments was quite small and no significant relationship was found locally between this parameter and bed erodibility.

Common proxies for bioadhesion including colloidal carbohydrate concentration, extracellular polymeric substance concentration, and organic content were measured to evaluate the influence of biostabilization on bed erodibility. In contrast to the results of many published works studying erodibility on mudflats, these bioadhesion proxies did not exhibit systematic variation in either space or time and were not correlated to bed erodibility. The seasonal timing of erodibility maxima and presence of laminations during periods of highest erodibility indicated that bioturbation was not responsible for increases in erodibility during the winter and spring at Clay Bank. Although not quantified in this study, anecdotal evidence from sediment cores and seabed camera images suggest that biological reworking and repackaging may alter bed cohesivity and structure potentially influencing the relationship between erodibility and bed solids fraction.

Periods of rapid deposition, as indicated by the presence of thick sequences of laminated sediments, suggest that periodic sediment flux convergence processes are responsible for the observed seasonal cycle in bed erodibility at Clay Bank. High bed erodibility was associated with rapid deposition in the likely presence of the mid-estuary turbidity maximum while low erodibility was associated with sediment bypassing and decreased deposition in the absence of a local turbidity maximum. Finally, a consistent, low erodibility condition was observed at the Gloucester Point site for the majority of the study period and at Clay Bank in the summer and fall. A comparison of these data to other published data from the Chesapeake Bay indicated that the low erodibility background state may be characteristic of an “equilibrium” bed condition often present in many regions of the Chesapeake Bay. In the absence of recent, rapid deposition, the data set presented from the York River reasonably characterizes both the range and the profile information of the bed critical erosion stress for mid-depth, muddy regions in other similar estuaries.
Controls on Erodibility in a Partially Mixed Estuary: York River, Virginia
Chapter 1

Introduction
1 Introduction

Understanding the transport and fate of cohesive sediments is of great importance in coastal waters. In net depositional areas, such as most estuaries, fine sediments rapidly fill in navigation channels resulting in the need for repeated dredging operations to maintain navigable waterways. Additionally, cohesive sediment deposition can alter benthic communities influencing local ecology (Thrush et al., 2003; Lohrer et al., 2004) and altering sediment-water exchange processes. When suspended in the water column, cohesive sediments attenuate light, limiting primary productivity, and often degrading the health of ecosystems (e.g. Dennison et al., 1993; Jassby et al., 2002). In particular, high turbidity, likely attributed to suspended cohesive sediments, has been implicated as a primary factor hindering the recovery of sea grasses in the Chesapeake Bay, Florida Bay, and many other systems (e.g. Dennison et al., 1993; Hall et al., 1999; Koch, 2001).

Additionally, many contaminants including, PCBs, dioxins, and heavy metals adsorb to cohesive sediments (e.g. Valette-Silver et al., 1996; Chapman and Wang, 1999). The fate and distribution of these contaminants are directly linked to fine sediment dynamics.

While the dynamics of cohesive sediments have been studied for decades, in both the seabed and the water column, significant uncertainties remain in virtually every aspect of their transport and fate. This project focuses on the thin veneer of sediment on the surface of the seabed that is actively and regularly resuspended and deposited by tidal
currents. Of particular interest are the seabed properties that control the erodibility of this active surface layer.

The erodibility of the seabed is quantified in terms of the critical shear stress necessary to initiate movement of particles. Near the surface of the seabed, the critical stress increases rapidly with depth into the seabed (Parchure and Mehta, 1985; Piedra-Cueva and Mory, 2001; Sanford and Maa, 2001). As critical stress rapidly increases, the properties of the seabed responsible for controlling critical stress must also exhibit a corresponding gradient. In terms of erodibility, a number of physical variables have been investigated including water content, temperature, salinity, mineralogy, and grain size (e.g. Kandiah, 1974; Kelly and Gularte, 1981; Parchure and Mehta, 1985; Roberts et al., 1998). Additionally, it is recognized that in highly productive environments such as estuaries and mudflats, benthic biota may play a significant role in influencing seabed erodibility both by destabilizing the bed through their burrowing and feeding activities (expressed as an increase in water content) and stabilizing the bed with sticky secretions (e.g. Widdows et al., 2000; Black et al., 2002; Tolhurst et al., 2002; Widdows and Brinsley, 2002).

This work focused on cohesive bed erodibility in the York River estuary, VA, USA. Close proximity to the Virginia Institute of Marine Science made the York River a convenient field area, while extensive previous works characterizing the hydrodynamics, benthic biology, and sediment transport processes of this system provided an excellent framework to build upon. Three sites on the York River were sampled monthly to bi-
monthly over a 19-month period to provide a range in cohesive sediment erodibility and
bed conditions as well as investigate spatial and temporal variations in bed erodibility.
The results of this effort are presented in the following chapters. Chapter Two
investigates variations in the physical properties of the bed sediment and their
relationship to bed erodibility. Chapter Three evaluates seasonal and spatial variations in
bed erodibility and addresses the dominant processes resulting in the observed variability.

Chapter Two presents the results of a 14-month time-series of bed erodibility
measurements and numerous related physical bed properties. Bed erodibility at two sites
in the Clay Bank region of the York River was observed to vary dramatically over the
course of the time-series. A third site, near Gloucester Point, VA generally exhibited a
consistent and low level of erodibility over the 14-month period.

Other published works have reported that the solids fraction of cohesive
sediments was a representative measure of the degree of consolidation of the bed with
erodibility decreasing with increasing solids fraction. The work presented in Chapter
Two demonstrated that the solids fraction of the top centimeter of the bed was quite
variable but not related to the erodibility of the bed. The observed variations in solids
fraction were shown to be a function of the sand fraction of the bed with solids fraction
increasing with sand fraction. Sand fraction was found to vary between 2% and 50%
indicating a cohesive bed with sand grains supported by a mud matrix. As the sand in the
bed was characteristically fine, the critical stress for erosion of the sand was similar to
and generally less than that observed for mud.
It was proposed in Chapter Two that for cohesive beds with varying sand fraction, solids fraction was not a good measure of the degree of consolidation of the bed as related to erodibility. Instead, the solids fraction of the mud matrix was calculated to represent the degree of consolidation of the cohesive portion of the bed responsible for controlling erodibility. However, data from the York River indicated no relationship between bed erodibility and solids fraction of the mud matrix. Published results from the literature were surveyed to evaluate the relationship between solids fraction and erodibility and solids fraction of the mud matrix and erodibility in a range of settings. The results from this study in the York River were included with seven published works, greatly expanding the range in data relative to that from the York River alone. The resulting comparisons demonstrated a weak positive relationship ($R^2=0.16$) between solids fraction and the initial critical stress for erosion. Impressively, this comparison provided a strong relationship ($R^2=0.68$) between solids fraction of the mud matrix and initial critical stress indicating that across a range of environments the solids fraction of the mud matrix was a better measure of the degree of bed consolidation than total solids fraction. Comparing the data from the York River to other published data indicated that the range in solids fraction of the mud matrix observed in the York was quite small. It was concluded that the range in solids fraction of the mud matrix in the York River was sufficiently small that its relationship to erodibility was likely overwhelmed by other factors.
Chapter Two has been submitted for publication as a manuscript co-authored by Carl T. Friedrichs and Lawrence P. Sanford.

Chapter Three provides a detailed investigation of the observed spatial and temporal variability in bed erodibility as well as evaluates the relative contributions of physical and biological influences on bed erodibility. A strong seasonal cycle in erodibility was observed at two sites in the Clay Bank region of the York River with consistent low erodibility in the summer and fall and elevated erodibility in the winter and spring. A third site, near Gloucester Point, VA had consistently low erodibility at a level similar to that found in the Clay Bank region in summer and fall.

Colloidal carbohydrate concentration and extracellular polymeric substance (EPS) concentration have proven to be proxies for bioadhesion in mudflat environments where biofilms are prevalent on the sediment surface. However, at the three sites sampled in the York River, little light reached the sediment surface and biofilms have not been observed. Measurements of colloidal carbohydrates and EPS from sediment samples collected in coordination with the samples from this study indicated that colloidal carbohydrate and EPS concentrations were relatively consistent in space and time and were not correlated to the observed variations in erodibility. Additionally, while no measures of biodestabilization were presented, the seasonal timing and presence of laminated sediments during the period of highest erodibility indicated that biodestabilization could not explain the elevated erodibility in the winter and spring.
While biostabilization and biodegradation were not clearly linked to the observed variations in bed erodibility, it is hypothesized in Chapter Three that biological modifications to surficial sediments may confound the expected relationships between bulk measures of sediment consolidation and bed erodibility. Pelletization may alter the cohesivity of sediments as well as the fabric of the seabed, effectively altering the relationship between solids fraction of the mud matrix and bed erodibility. Additionally, burrowing activities may introduce heterogeneity into the seabed resulting in high values for bulk measures of solids fraction of the mud matrix that do not represent the true consolidation state of the bed on the small scale relevant to surface erosion.

Previous studies have characterized the Clay Bank region as physically dominated with high suspended sediment concentrations resulting from an ephemeral secondary turbidity maximum (STM). These studies demonstrated that the presence of an STM in the Clay Bank region was associated with elevated river flow enhancing stratification and particle trapping. The associated convergent sediment transport periodically resulted in high suspended sediment concentrations in the region. The coincidence of high river flow, low salinity, and high erodibility presented in Chapter Three indicates that this process translates to the sediment bed with convergent sediment transport resulting in enhanced erodibility. Further, thick laminated sequences observed in digital X-radiographs, revealed that greater than 15 centimeters of sediment was deposited between November 2006 and April 2007. This rate of deposition greatly exceeded the long-term
accumulation rate of this region indicating an ephemeral deposit likely migrating with the STM.

Erodibility data from the York River was compared to other published results from the Chesapeake Bay. The York River data was split into three categories, low erodibility, transitional, and high erodibility based on the level of erodibility and seasonal timing. A striking similarity was found in both the ranges and mean conditions of the York River low erodibility data, representing nearly all measurements at the Gloucester Point site as well as the Clay Bank sites in the summer and fall of 2006 and 2007, and published data from Baltimore Harbor, MD and the upper portion of the Chesapeake Bay main-stem. Conversely, the transitional and high erodibility cases from the York River were found almost entirely at the Clay Bank sites and fell well outside the range of the data from Baltimore Harbor and the upper-Chesapeake main-stem. This comparison suggests a common state of “low” erodibility in many muddy areas in the Chesapeake Bay outside of turbidity maximum zones. This low erodibility condition likely represents an equilibrium state resulting from a balance between physical resuspension and consolidation and biological reworking and repackaging. Overall, the cumulative York River data set provides a range in bed erodibility characterizing and contrasting mid-depth muddy regions both inside and outside of turbidity maximum zones.

Chapter Three will be submitted for publication as a manuscript co-authored by Carl T. Friedrichs, Linda C. Schaffner, Lawrence P. Sanford, and Robert J. Diaz. Measurements of colloidal carbohydrate and EPS presented in Chapter Three were
provided by Dr. Schaffner. Sediment images presented in Chapter Three were provided by Dr. Diaz.
Chapter 2

Mud matrix solids fraction and bed erodibility in the York River, USA, and other muddy environments

Mud matrix solids fraction and bed erodibility in the York River, USA, and other muddy environments

Abstract

A 14-month time series of sediment cores from the bed of the York River estuary, Chesapeake Bay, USA, were sampled with a Gust erosion microcosm and further analyzed to evaluate variability in a variety of physical bed properties. Variation in sediment solids volume fraction did not relate to variability in bed erodibility. However, solids volume fraction was found to be highly dependent on the sand fraction of the bed. The solids volume fraction of the mud matrix was calculated to evaluate changes in bed compaction not related to the sand fraction of the bed. The range of variability in solids volume fraction of the mud matrix was found to be significantly less than the variability of the total solids volume fraction. Reevaluation of erodibility data from the literature combined with that from this study revealed a strong correlation between solids volume fraction of the mud matrix and the initial critical stress for erosion when a large range in sand fraction and solids volume fraction were included. These results suggest that compaction within the cohesive portion of the bed is better related to erodibility than compaction of the bed as a whole (mud and sand). The poor correlation found within the
York River data alone likely resulted from the relatively small range observed in the solids volume fraction of the mud matrix.

2.1 Introduction

Sediment erodibility is generally quantified as the erosion rate or mass of sediment available for erosion at a given stress applied to the bed. While bed erodibility can be measured by a number of in situ or laboratory devices (McNeil et al., 1996; Gust and Mueller, 1997; Tolhurst et al., 2000; Widdows et al., 2007), it is typically time consuming and expensive. As a result, it is often impractical to resolve bed erodibility at adequate temporal or spatial scales. To improve resolution and understand the controlling factors causing changes in erodibility, it is desirable to relate the erosion rate and/or the critical stress for erosion to the properties of the deposited sediments that can be more easily measured at sufficient spatial and temporal scales.

When working with non-cohesive sediments, the critical stress for erosion is controlled by the gravitational resistance to motion and is a relatively simple and known function of grain size (Shields, 1936). Cohesive sediments on the other hand are notably more complicated. Cohesive sediments form weak inter-particle bonds the strength of which are a function of both the fluid properties (e.g., salinity, temperature) (Kelley and Gularte, 1981; Parchure and Mehta, 1985; Lau, 1994) and bed properties (e.g., fine fraction grain size, porosity, mineralogy) (Kandiah, 1974; Roberts et al., 1998; Torfs et al., 2001). As muddy seabeds are commonly found in biologically productive areas, the
relationship between the physical properties of the bed and fluid and erodibility are often further modified by a variety of biological influences (e.g., bioturbation, pelletization, EPS) (Widdows et al., 2000; Black, 2002; Andersen et al., 2005; and many others) that can act to either enhance or reduce erodibility.

Sediment bulk density (a term functionally interchangeable with dry density, porosity, solids fraction, and water content) is a common physical bed property linked to erodibility. Laboratory studies have shown a clear relationship between sediment bulk density and erodibility when other physical properties of the sediment are controlled (particularly grain size and mineralogy; Jepsen et al., 1997; Roberts et al., 1998) and in the absence (to the extent it is possible) of biological influence. As the bulk density increases, cohesive sediment grains are more tightly packed, enhancing the inter-particle bonds and increasing the critical stress for erosion. However, in natural sediments this relationship has often proven difficult to resolve (Houwing, 1999; Riethmueller et al., 2000; Mahatma, 2004; Tolhurst et al., 2006; Stevens et al., 2007). While this may be largely due to biological influences, variations in sediment grain size and mineralogy may also obscure this relationship.

Laboratory studies have shown clear variations in the cohesive strength of different clay minerals (Kandiah, 1974; Torfs et al., 2001). When evaluating erodibility in different geographical regions, the mineralogy of the seabed may vary due to different source materials. Within a given region on the other hand, mineralogy may be effectively constant or vary slowly over distances on the order of hundreds of kilometers. Within the
Chesapeake Bay for instance, Feuillet and Fleischer (1980) found the mineralogy of the seabed to vary only gradually along the estuarine gradient. In this case, both the freshwater reaches and Atlantic Ocean provided sediments that mixed within the James River estuary. The relative proportion of marine and freshwater source minerals were controlled by estuarine circulation. This suggests that at a particular site, and likely for a particular region, mineralogical composition will be effectively constant and will not be responsible for temporal or relatively small-scale (10s of kilometers) spatial variations in bed erodibility.

In most muddy coastal environments, the seabed is a mixture of non-cohesive (sand) and cohesive (mud) sediment. As a result, the erodibility of a cohesive bed may be modified by the presence of sand in the bed. While a few studies have addressed this problem, there is still no clear relationship to explain the modification in critical shear stress resulting from the presence of sand in the bed. In a laboratory study, Torfs et al. (2001) measured the critical stress for erosion of artificial and natural sediments of a constant bulk density but with varying sand versus mud fraction. As mud fraction increased above 3-13%, the bed began to behave cohesively, and the critical stress for erosion increased as a function of increasing mud fraction of the bed. A related study by Panagiotopoulos et al. (1997) using estuarine mud mixed with quartz sand showed that as mud fraction increased from 0-50%, the critical stress for erosion also increased. Similar to the results of Torfs et al. (2001), the bed began to behave cohesively at a mud fraction
of 11%-14%. In Panagiotopoulos et al. (1997), bulk density was not held constant, and they noted that as the mud fraction increased, the bulk density decreased.

Similar to Panagiotopoulos et al. (1997), a number of other researchers have noted that as sand fraction increases, bulk density also increases. Flemming and Delafontaine (2000) and Riethmüller (2000) found a positive relationship between water content and mud fraction, which Flemming and Delafontaine (2000) approximated as a power law function with site specific coefficients. Mahatma (2004) also found water content to vary with mud fraction and proposed that changes in mud fraction must be accounted for if water content is to be used as a proxy for bed compaction. To evaluate changes in the degree of compaction between sites or temporally at a given site, Mahatma (2004) derived a generalized relationship between mud fraction and water content using data from four of six sites. The ratio of the actual water content to that predicted from this relationship gives the normalized water content. Values of normalized water content less than 100% indicate more compacted sediments while values more that 100% indicate less compacted sediments.

Rather than using a site-specific empirical relationship, Sanford (2008) represented the influence of sand on sediment solids fraction using a simple analytical relationship of the form

$$\phi_{sdw} = \frac{1}{f_s + \frac{1-f_s}{\phi_{sm}}}$$  \hspace{1cm} (Eq. 1)

where:
\( \phi_{\text{tot}} = \text{bed solids volume fraction} = \frac{\text{volume solids}}{\text{volume total}} \)

\( \phi_{\text{sm}} = \text{solids volume fraction of mud matrix} = \frac{\text{volume mud}}{\text{(volume mud plus water matrix)}} \)

\( f_s = \text{mass fraction of sand per mass total solids} \)

This approach treats the bed as a mixture of sand and mud-water matrix. The sand with a bulk density of 2650 kg m\(^{-3}\) is suspended in the mud and water matrix of lower bulk density. As the sand fraction is increased, lower density mud matrix material is displaced by higher density sand, raising the bulk density (or solids fraction) of the bed. At a sand fraction of zero, \( \phi_{\text{tot}} = \phi_{\text{sm}} \).

2.2 Methods

2.2.1 Study Site

The field component of this study was conducted on the York River estuary, a subestuary of the Chesapeake Bay, USA. Although its mean tidal range is only about 80 cm, the York River is characterized by strong tidal currents reaching magnitudes of \(~1\text{m s}^{-1}\) at the surface during spring tide (Schaffner et al., 2001). Three sites were chosen in the York River to provide variation in erodibility and associated bed properties (Figure 2-1). Two sites were chosen near Clay Bank, about 30 km from the mouth of the York in a region characterized by strong tidal currents, an ephemeral secondary estuarine turbidity
maximum, and suspended sediment concentrations often reaching 100s of mg l\(^{-1}\) (Lin and Kuo, 2001). The strong physical processes at this site make it a generally unfavorable environment for benthic biota, leading to a depauperate benthic community (Schaffner et al., 2001). One site at Clay Bank was located on the flank of the main channel at \(\sim 11\) m depth (CC), while the other site was about \(1\) km away in a secondary channel at \(\sim 6\) m depth (CS). The third site, near Gloucester Point (GP) is located about 10 km from the mouth of the York at a depth of \(\sim 8\) m. Relative to the Clay Bank sites, the Gloucester Point site experiences less physical disturbance to the seabed and lower suspended sediment concentrations (10s of mg l\(^{-1}\)), leading to more favorable conditions for benthic communities and potentially less physical dominance in sediment transport processes (Schaffner et al., 2001).

2.2.2 Sampling

Each of the three sites was sampled monthly to bimonthly from April 2006 to July 2007. The dates and times of sampling were determined primarily for logistical reasons, and sampling took place randomly with respect to tidal phase and spring/neap cycle. Whenever possible, the three sites were sampled on consecutive days and following the same phase of the tide.

Each time a site was visited, two cores were collected for erodibility measurements, three slabs were collected for digital X-radiography, and sediment samples sliced at 1-cm intervals were collected to be analyzed for water content,
Figure 2-1  a. Map of Chesapeake Bay, USA.  b. Map of York River estuary. Circles represent locations of study sites. GP is Gloucester Point site, CC is Clay Bank channel site, and CS is Clay Bank secondary channel site.
clay/silt/sand fraction, and organic content. All sediment samples were sub-sampled from an Ocean Instruments Gomex box corer (surface area 625 cm$^2$) deployed from a small vessel (~8 m) on anchor at or near slack tide. To minimize errors associated with core collection and sub-sampling, particular attention was paid to preservation of the sediment-water interface. Due to the soft nature of these sediments, the weight of the corer allowed sufficient penetration (~20-40 cm). As a result, the Gomex could be slowly lowered to the seabed, allowed to penetrate, and then slowly retrieved. Lids on the box corer, which automatically closed following penetration, prevented sloshing of the water overlying the sediment sample, further minimizing disturbance of the sediment-water interface. Box cores and sub-cores with turbid water overlying the sediment surface were assumed to be disturbed and were rejected. Due to the number of sub-samples collected and the relatively small size of the box corer, numerous box cores were collected from each site. As a result of collecting numerous cores while swinging on anchor, all results likely incorporate spatial variation within a radius of roughly 15 to 25 m.

2.2.3 Erodibility Measurements

Seabed erodibility was measured with a dual core Gust erosion microcosm system constructed at the University of Maryland Center for Environmental Science – Horn Point Laboratory. This device uses a rotating disc with central suction to impose a nearly uniform shear stress on the surface of a 10 cm sediment core (Gust and Mueller,
The individual erosion heads were calibrated in the lab of Dr. Gust using hot film anemometry.

Cores chosen for erosion testing were carefully selected to ensure uniform, level surfaces and minimal disturbance of the sediment-water interface. After collection, cores were carefully transported to the Virginia Institute of Marine Science, located adjacent to the GP site on the banks of the York River. Erodibility tests were generally underway within about two hours of core collection to minimize consolidation effects. Erodibility measurements consisted of a sequence of seven steps of approximately 20-min duration, each with a consecutively increasing stress (0.01 Pa, 0.05 Pa, 0.1 Pa, 0.2 Pa, 0.3 Pa, 0.45 Pa, 0.6 Pa) applied to the sediment surface.

Each erodibility measurement included two cores from a given site eroded simultaneously. Surface water, collected from the sampling site, was circulated through the cores, generating the radial component of the applied bed stress, and flushing suspended sediment from the core. The effluent containing eroded sediment was then passed through a Hach 2100AN turbidimeter equipped with a flow-through cell and collected. The sediment-laden effluent was filtered using 0.7-μm GFF filters to determine the total mass of sediment eroded during each step and to calibrate the turbidimeters. The product of the calibrated turbidimeter data and the flow rate of water circulated through the cores provided a time series of erosion rate. Additionally, the filtrate from each step provided the actual mass eroded during each step, the sum of all steps providing the total mass eroded during the experiment.
Results were analyzed with the erosion formulation of Sanford and Maa (2001). This erosion formulation uses a linear erosion rate expression and allows for depth-varying critical stress and “erosion rate constant.” The initial critical shear stress, final critical shear stress, erosion rate constant (M), and the initial erosion rate (E₀) were calculated for each step. For each step, E₀ was calculated as the erosion rate at time t = 0 from an exponential fit to the erosion rate time series. The erosion rate constant M was calculated using this value of E₀.

2.2.4 Determination of Physical Bed Properties

Samples for sediment water content, sand/silt/clay fraction, and organic content were collected from cores sub-sampled from the Gomex box corer while on anchor and simultaneous to collecting cores for erosion with the Gust erosion microcosm. These sub-cores where then sliced at 1-cm intervals and saved for analysis. Whenever possible, two X-ray slabs were also sliced at 1-cm intervals after being X-rayed providing a total of three samples and allowing estimation of the standard error for each of these physical properties of the bed.

Sediment water content was determined using standard wet weight/dry weight analysis. By assuming the density of water to be 1015 kg m⁻³ and the density of sediment to be 2650 kg m⁻³ the solids volume fraction was calculated from the water content. Sand fraction was measured by sieving sediments with a 63-µm mesh sieve. The sand fraction was calculated as the mass of sand per mass dry weight of sediment. The size
distribution of the sand fraction was determined using a Rapid Sediment Analyzer. Percentages of silt and clay were determined by standard pipette method for each of the three sites over the entire 14-month period. Prior to the pipette test, a dispersant was added and samples were placed in a sonicator bath to disaggregate the sample.

2.3 Results

2.3.1 Erodibility

Erodibility measurements conducted with the Gust Microcosm system produced profiles of the critical shear stress for erosion and erosion rate. The results of erodibility measurements conducted on 60 cores (20 from each site) are presented here. With the exception of a small amount of sediment eroded at a low shear stress in some cores, erosion was entirely Type 1 depth-limited erosion. As a result, we used the critical shear stress for erosion to evaluate sediment erodibility. As comparing 60 critical shear stress profiles would be overly complicated and potentially confusing, these data have been simplified and are presented as the total erodible mass at a shear stress of 0.4 Pa. This is similar to the results presented in Widdows et al. (2000), where erodibility data were presented as the total mass eroded at a given current velocity. A critical shear stress of 0.4 Pa was chosen as this represents a typical tidal maximum shear stress within the York River system. Additionally, a relatively large mass of sediment (1 - 8 mm of depth) was eroded from the bed at this stress. As these results are compared to the properties
Figure 2-2 Time series of eroded mass at 0.4 Pa at three sites on York River, VA. Bars represent mean eroded mass of two cores eroded. Stars represent mass eroded from each individual core.
(e.g., solids fraction and sand fraction) of the top centimeter of the bed, it is desirable to have erodibility data and bed property data represent the same material as much as possible. This is in contrast to measuring the initial critical shear stress, as is common, in which case the material eroded likely represents less than the top millimeter of sediment. Resolution of bed properties at this scale was not feasible, so a larger scale of erosion was chosen.

The erodibility results are summarized in Figure 2-2. As only two replicate measurements were performed each time a site was sampled, it was not possible to evaluate statistical significance associated with these measurements. To illustrate the amount of variation between replicates, the results of each replicate core are presented as stars in Figure 2-2, while the average of both cores is presented as the bars in this plot.

In all cores eroded, the initiation of motion occurred at an applied stress of 0.05 Pa or 0.1 Pa. Erosion behavior was typical of depth-limited erosion in this type of stepwise measurement. A spike in erosion rate occurred at the beginning of a step followed by an exponential decay in erosion rate throughout the step. While erosion rate had decreased significantly by the end of each step, it was not uncommon for erosion to continue until the end of a 20-minute step, particularly at the higher applied stresses. The erodible mass at 0.4 Pa was found to vary by roughly an order of magnitude at all sites (Figure 2-2). A total erodible mass at 0.4 Pa of ~ 0.2 kg m$^{-2}$ was typical in cases of low erodibility, representing about 1 mm of sediment eroded from the seabed. An erodible
mass of 1.5 – 2 kg m\(^2\) was representative of periods of high erodibility, corresponding to about 8 mm of sediment eroded from the seabed.

2.3.2 Erodibility and Solids Volume Fraction

The results of a 14-month time series of total solids (sand + mud) volume fraction (\(\phi_{\text{sol}}\)) for each of the three sites are presented in Figure 2-3a. Three samples for \(\phi_{\text{sol}}\) were collected each time a site was sampled. The mean \(\phi_{\text{sol}}\) ranged from 0.08 to 0.18, and the error bars in Figure 2-3a represent two times the standard error associated with these data. The GP and CS sites displayed the largest variation in \(\phi_{\text{sol}}\), and both covered approximately the entire range. The CC site showed less variation and tended to have lower values of \(\phi_{\text{sol}}\) than either the GP or CS site. The large error bars in Figure 2-3a illustrate a high degree of variability on a relatively small spatial scale (10s of meters associated with core collection on anchor). Additionally, the approximately month-to-month variation at a given site suggests this property of the bed can vary significantly on relatively short time scales.

No significant correlation was found between the eroded mass at 0.4 Pa and \(\phi_{\text{sol}}\) of the top cm of the bed (Table 2-1). Cases of low erodibility \(\sim 0.2\) kg m\(^2\) spanned the entire range of \(\phi_{\text{sol}}\), suggesting that large changes in \(\phi_{\text{sol}}\) can result in virtually no change in bed erodibility. While it would be expected that the highest erodibility would coincide with the lowest values of \(\phi_{\text{sol}}\), the observed correlation was opposite (i.e., positive; Table 2-1), and the cases of highest erodibility fell toward the middle of the range of \(\phi_{\text{sol}}\).
Figure 2-3 Time series of properties of upper cm of bed: a. total solids (sand + mud) volume fraction ($\phi_{\text{tot}}$), b. solids volume fraction of mud matrix ($\phi_{\text{sm}}$), c. sand fraction of total solids, d. clay fraction, e. clay : silt ratio. In each case, error bars represent two standard errors.
<table>
<thead>
<tr>
<th></th>
<th>Eroded Mass @ 0.4 Pa</th>
<th>$\phi_{\text{tot}}$</th>
<th>$\phi_{\text{sm}}$</th>
<th>Sand Fraction</th>
<th>Clay Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi_{\text{tot}}$</td>
<td>0.1906&lt;sup&gt;a&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\phi_{\text{sm}}$</td>
<td>0.0667&lt;sup&gt;a&lt;/sup&gt;</td>
<td>0.8116&lt;sup&gt;b**&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sand Fraction</td>
<td>0.1306&lt;sup&gt;a&lt;/sup&gt;</td>
<td>0.6395&lt;sup&gt;b**&lt;/sup&gt;</td>
<td>0.0785&lt;sup&gt;b&lt;/sup&gt;</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clay Fraction</td>
<td>-0.2909&lt;sup&gt;a&lt;/sup&gt;</td>
<td>-0.6007&lt;sup&gt;b**&lt;/sup&gt;</td>
<td>-0.2720&lt;sup&gt;b**&lt;/sup&gt;</td>
<td>-0.6769&lt;sup&gt;b**&lt;/sup&gt;</td>
<td></td>
</tr>
<tr>
<td>Clay:Silt ratio</td>
<td>-0.3106&lt;sup&gt;a*&lt;/sup&gt;</td>
<td>-0.3041&lt;sup&gt;b**&lt;/sup&gt;</td>
<td>-0.2521&lt;sup&gt;b**&lt;/sup&gt;</td>
<td>-0.1895&lt;sup&gt;b&lt;/sup&gt;</td>
<td>-0.7992&lt;sup&gt;b**&lt;/sup&gt;</td>
</tr>
</tbody>
</table>

**Table 2-1** R-values of Pearson’s correlation test amongst erodibility and bed property data. <sup>a</sup> indicates n=30. <sup>b</sup> indicates n=72. * indicates significant at p<0.10, ** at p<0.05.
2.3.3 Solids Volume Fraction, Sand Fraction, and Solids Volume Fraction Mud Matrix

Similar to $\phi_{s\text{tot}}$, a large variation in sand fraction was found both within a given site and between sites (Figure 2-3c). The GP and CS sites exhibited the largest range in sand fraction, varying between 0.02 and 0.42, while the CC site typically had less sand, ranging between 0.02 and 0.15. In all cases, the sand fraction was sufficiently small that the bed was a mud-dominated cohesive mixture supported by a mud matrix. The sand at all sites was typically very fine to fine sand with a D50 of between 100 and 150 μm. The D90 of the sand size distribution was almost always less than 250 μm.

Consistent with the results of Panagiotopoulos et al. (1997), Flemming and Delafontaine (2000), Paterson et al. (2000), Riethmueller (2000), and Mahatma (2004), $\phi_{s\text{tot}}$ was found to increase with increasing sand fraction of the bed (Table 2-1; Figure 2-4). Additionally, the analytical relationship (Eq. 1) between sand fraction and solids volume fraction presented in Sanford (2008) was tested, the results of which are also presented in Figure 2-4. Similar to Flemming and Delafontaine (2000) and Mahatma (2004), a constant state of consolidation was assumed regardless of sand fraction to prepare Figure 2-4. This was implemented by assigning a constant value of 0.11 to the solids volume fraction of the mud matrix ($\phi_{sm}$).
Figure 2- 4 Total solids volume fraction (\(\phi_{\text{sol}}\)) of the upper cm of bed as a function of the sand fraction of total solids. Solid line represents relationship from Eq. 1 with \(\phi_m = 0.11\). Diamonds represent outliers not included in \(R^2\) calculation.
Clearly there is a good relationship between the data presented in Figure 2-4 and Eq. 1 when \( \phi_{sm} \) is assumed to be constant. Even more useful, however, if sand fraction and \( \phi_{tot} \) are known, Eq. 1 can be rearranged to

\[
\phi_{sm} = \frac{\phi_{tot} - f_s \phi_{tot}}{1 - f_s \phi_{tot}}
\]

(Eq. 2)

and used to calculate \( \phi_{sm} \). Doing so eliminates the necessity of assuming a constant state of consolidation with varying sand fraction.

2.3.4 Erodibility and Solids Volume Fraction Mud Matrix

The solids volume fraction of the mud matrix (\( \phi_{sm} \)) was calculated using Eq. 2 for the top 1 cm of sediment from each of the three sites over the 14-month period sampled. This time series of \( \phi_{sm} \) is presented in Figure 2-3b. The range of \( \phi_{sm} \) for all three sites (excluding one data point) is 0.085 – 0.145, about half that of the range in \( \phi_{tot} \) (Figure 2-3a). The range in \( \phi_{sm} \) was similar for all sites, eliminating the distinction between the CC site and the other two sites, and suggesting that much of the between site variation in \( \phi_{tot} \) was a result of variation in sand fraction between sites. The standard errors associated with \( \phi_{sm} \) (Figure 2-3b) are also considerably diminished in comparison to \( \phi_{tot} \) (Figure 2-3a) suggesting that much of the variability observed within a site was a result of local variation in the sand fraction of the bed. Figure 2-5 further emphasizes that consideration of \( \phi_{sm} \) in place of \( \phi_{tot} \) significantly reduces variability between sites as well as down
Figure 2-5  a. Depth profiles of $\phi_{\text{tot}}$ collected over a 14-month period at each of the three York River sites.  b. Depth profiles of $\phi_{\text{sm}}$ collected over a 14-month period at each of the three York River sites. $\phi_{\text{sm}}$ was calculated using data from Fig. 2-5a and sand fraction profiles.
cores. The overall standard deviation of $\phi_{sm}$ for all samples from all sites is only 0.021, whereas the standard deviation of $\phi_{sot}$ for all samples is 0.040.

Similar to the comparison of $\phi_{sot}$ to eroded mass, there was no correlation between $\phi_{sm}$ and the eroded mass at 0.4 Pa (Table 2-1). While eroded mass varied by roughly an order of magnitude, there was no clear distinction in $\phi_{sm}$ between periods of high and low erodibility. In fact, values of $\phi_{sm}$ for periods of low erodibility ranged from 0.085 to 0.145, representing the entire range observed in $\phi_{sm}$ data.

2.3.5 Fine Fraction Grain Size

The fraction of silt and clay in York River sediments was quite variable between sites and over the course of the 14-month sampling period. A time-series of clay fraction from each site is presented in figure 2-3d. While the observed clay fraction ranged from 0.30 to 0.76, the correlation between sand fraction and clay fraction reported in Table 2-1 indicates that much of the variation in clay fraction results from variations in sand fraction. Similarly, the correlation between clay fraction and $\phi_{sot}$ (Table 2-1) results from the correlation of clay fraction to sand fraction and the dependence of $\phi_{sot}$ on sand fraction.

The ratio of the clay fraction to silt fraction was determined and presented in Figure 2-3e to evaluate changes in fine fraction grain size not resulting from variations in
sand fraction. Results from pipette analysis indicated that the clay to silt ratio was variable over the 14-month period, with a mean clay to silt ratio of approximately 1.5 (Figure 2-3e). While there was no significant correlation between clay fraction and eroded mass, a significant, negative correlation \((p<0.10; \text{Table 2-1})\) was found between the eroded mass at 0.4 Pa and the clay to silt ratio. This suggests that periods of higher erodibility coincided with siltier, less cohesive bed sediments. However, a negative correlation, significant at \(p<0.05\), was also found between the clay to silt ratio and \(\phi_{\text{sm}}\). This may indicate that an increase in the clay to silt ratio both inhibited consolidation and increased cohesivity, resulting in two counteracting influences on erodibility.

2.4 Discussion

2.4.1 Limitations of \(\phi_{\text{slot}}\)

The erodibility of muddy beds is largely a function of inter-particle cohesion. As the bed becomes more compacted, the degree of inter-particle cohesion will increase, presumably resulting in decreased erodibility. The magnitude of \(\phi_{\text{slot}}\) of the bed is assumed to reasonably represent the degree of compaction of the bed and, thus, is often compared to measures of erodibility. Numerous studies in both laboratory and natural environments have compared \(\phi_{\text{slot}}\) of the bed to erodibility with varying results. While some studies have shown a strong relationship between erodibility and \(\phi_{\text{slot}}\) (Amos et al., 1997; Jepsen et al., 1997; Roberts et al., 1998), it has been equally as common to find
little (Riethmueller et al., 2000) or no relationship (Houwing, 1999; Mahatma, 2004; Tolhurst et al., 2006; Stevens et al., 2007).

While both $\phi_{sol}$ and erodibility were found to be variable in this study, no clear relationship was found between the two. However, we and others (Flemming and Delafontaine, 2000; Panagiotopoulos et al., 1997; Paterson et al., 2000; Riethmueller, 2000; and Mahatma, 2004) have illustrated that $\phi_{sol}$ is highly dependent on the sand fraction of the bed. The relationship shown in Eq 2. and data shown in Figure 2-4 clearly illustrate that the addition of sand to the bed results in an increase in $\phi_{sol}$. As $\phi_{sol}$ is assumed to be an estimate of the degree of compaction of the bed and, thus, inter-particle cohesion, it stands to reason that the addition of sand to the bed may complicate this assumption. In the simplest sense, comparing $\phi_{sol}$ to erodibility for mixed mud/sand beds of varying sand fraction assumes that mud and sand are equally cohesive. As this is known to not be true, it becomes apparent that an increase in $\phi_{sol}$ resulting from increasing sand fraction should not result in the same decrease in erodibility as would be expected for a similar increase in $\phi_{sol}$ in a bed composed entirely of mud. This distinction may explain the lack of correlation between erodibility and $\phi_{sol}$ found in many studies.

2.4.2 Removing the Effect of Sand on Solids Fraction

Flemming and Delafontaine (2000) and Mahatma (2004) recognized the influence of sand fraction on $\phi_{sol}$ and attempted to account for it by deriving empirical relationships between the two. Flemming and Delafontaine (2000) found the relationship to be site-
specific and developed site-specific relationships to account for changes in $\phi_{\text{mud}}$ with sand fraction. Mahatma (2004) used data from four of six sites to derive an empirical relationship and used this relationship to normalize data from all sites. This “normalized water content” was used to compare the degree of consolidation. While strong relationships were shown in both cases, this approach requires the assumption that the true degree of compaction does not vary with sand fraction. While this assumption may be correct, deriving site-specific relationships makes comparison between sites difficult, while normalizing data is only possible when all sites exhibit a similar degree of compaction.

Rather than deriving empirical relationships between $\phi_{\text{mud}}$ and sand fraction, we propose the analytical solution given in Eq. 2. This solution derives the solids volume fraction of the mud matrix ($\phi_{\text{sm}}$) supporting the bed and is independent of variation in sand fraction. Further, Eq. 2 eliminates the necessity of deriving site-specific relationships and effectively normalizes solids fraction data without making the assumption that $\phi_{\text{sm}}$ does not vary with sand fraction. Calculating $\phi_{\text{sm}}$ from $\phi_{\text{mud}}$ and sand fraction provides a way of comparing erodibility to the degree of compaction of the muddy matrix supporting the bed. While eliminating the apparent increase in compaction and cohesion associated with the addition of sand, this approach (as well as that of Mahatma, 2004) makes the assumption that the sand fraction of the bed has no influence on bed erodibility. At first glance, this assumption appears to be contradicted by the results of Torfs et al. (2001).
2.4.3 Re-evaluating the results of Torfs et al. (2001)

A laboratory study presented in Torfs et al. (2001) shows a clear increase in the critical stress for erosion with increasing mud fraction or, conversely, that as sand fraction increases, the critical stress for erosion decreases. However, Torfs et al. (2001) maintained a constant bulk density while varying the mud fraction. As was shown here and by others (Flemming and Delafontaine, 2000; Riethmueller, 2000; and Mahatma, 2004), as sand fraction increases, bulk density (or $\phi_{\text{slot}}$) also increases. Conversely, from Eq. 2 it is apparent that maintaining a constant bulk density while increasing sand fraction must also cause a decrease in $\phi_{\text{sm}}$. Thus the relationship shown in Torfs et al. (2001) illustrating a decrease in critical shear stress with increasing sand fraction also represents a decrease in critical shear stress with decreasing $\phi_{\text{sm}}$.

Equation 2 predicts a monotonic negative relationship between sand fraction and $\phi_{\text{sm}}$ for sediments of a constant bulk density that approaches linearity as $\phi_{\text{sm}}$, $\phi_{\text{slot}}$, and $f_s$ decrease. Thus, reevaluating the data of Torfs et al. (2001) in terms of both sand fraction and $\phi_{\text{sm}}$, and including only data with a sand fraction of 0.96 or less, demonstrates that either sand fraction or $\phi_{\text{sm}}$ can be used to predict the initial critical stress for erosion equally well (Figures 2-6a and 2-6b). A power law regression between $\phi_{\text{sm}}$ and critical stress for montmorillonite and for a combination of kaolinite and natural mud yield $R^2$ values of 0.740 and 0.794, both of which are significant at $p<0.05$. While the resulting relationships between $\phi_{\text{sm}}$ and critical stress are dependent on sediment type, it allows a
single type of equation (i.e., critical stress versus $\phi_{sm}$) to represent the data, rather than requiring both a relationship between bulk density and critical stress and a relationship between sand fraction and critical stress.

Similar to the reevaluated results of Torfs et al. (2001), the results of this study also suggest that sand fraction does not have an appreciable influence on the relationship between $\phi_{sm}$ and erodibility. The CC and CS sites both exhibited similar patterns in erodibility including both the amount of material eroded and the seasonal timing of major changes in erodibility. Additionally, these two sites had very similar values of $\phi_{sm}$, both exhibiting a small and similar range. However, the sand fraction was generally quite different between these sites. Sand fraction at the CC site ranged from 0.02 to 0.15 while the CS site was always sandier ranging between 0.08 and 0.32. Despite differences in sand fraction of as much as 0.30, the two sites displayed similar erodibility.

While neither this study nor that of Torfs et al. (2001) was designed to evaluate the influence of sand fraction on erodibility for a specific value of $\phi_{sm}$, this new approach could easily be evaluated with controlled laboratory experiments. It should also be noted that this approach is only appropriate for sand that erodes at critical shear stresses similar to those of the supporting mud matrix. If coarser sand with higher critical stress was present, the eroded mud mass might be limited by bed armoring (Wiberg et al., 1994) as opposed to consolidation of the mud matrix, which would confound interpretation of the
Figure 2-6  a. Initial critical stress for erosion as a function of fine fraction for a mixed mud/sand bed. Results re-plotted from Torfs et al. (2001). b. Initial critical stress for erosion as a function of solids fraction mud matrix. Reevaluated results from Torfs et al. (2001). Solids fraction mud matrix was calculated from bulk density and sand fraction using Equation 2.
results. As the sand present in our study was about 120 µm in diameter, the critical stress for erosion of the sand was about 0.1 Pa, similar to that of the eroded mud.

2.4.4 Relationship Between Erodibility and $\phi_{sm}$ over a broader range

While the results of this study did not reveal a significant relationship between erodibility and $\phi_{sm}$, we still believe this approach to be promising. To further evaluate its potential, a literature search was performed in an attempt to apply this relationship to a greater range of data from mud matrix supported mixed mud/sand beds. $\phi_{sm}$ could only be calculated from published data that included both sand fraction and $\phi_{sat}$ (or bulk density, water content, etc.), greatly limiting the previous work that could be included in this comparison. As the initial critical stress for erosion proved to be the most common measure of erodibility included in these published works, the results of this study were also presented in this form. The erodibility measurements presented in this study were originally designed to measure profiles of the critical stress for erosion with depth into the seabed. As a result, the initial critical stress was not well resolved. For every core eroded, initiation of motion occurred at an applied stress of either 0.05 Pa or 0.1 Pa, and these data are presented accordingly in Figure 2-7.

The results of nine published works were found to include information on sand fraction, $\phi_{sat}$, and critical stress for the initiation of motion and were considered for this comparison. Of these nine, Torfs et al. (2001) and Kamphius and Hall (1983) proved to
be outliers in that the values of critical stress presented were an order of magnitude or more higher for a given value of $\phi_{sm}$ and were thus excluded. As there are no accepted standards for measuring erodibility or interpreting erodibility results (Sanford, 2006), the relatively high critical stress values presented in these papers may represent differences in the definition of initial critical stress, erosion device behavior, calibration, or data analysis techniques. Data from the remaining seven and this study (Table 2-2) were included in Figure 2-7. Critical stress data with sand fractions greater than 90% were excluded from these results as the work of Panagiotopoulus et al. (1997) and Torfs et al. (2001) demonstrated that sediment with sand fraction greater than approximately 90% did not behave cohesively. The remaining data represent a wide range of sand fraction (2 - 90%), $\phi_{sed}$ (0.08 – 0.5), and $\phi_{sm}$ (0.06 – 0.39).

Figure 2-7a presents a comparison of $\theta_{tot}$ to the initial critical stress for erosion. While there is a weakly significant positive relationship overall ($R^2 = 0.16$), some of the individual data sets showed a negative relationship, while others show no relationship at all. In contrast, the comparison of $\phi_{sm}$ to initial critical stress in Figure 2-7b provides a much stronger and consistently positive relationship. Given the number of devices used and likely inconsistencies in identification of initial critical stress, there is a surprisingly strong relationship between $\phi_{sm}$ and initial critical stress ($R^2 = 0.68$). This suggests that, provided a sufficient range in $\phi_{sm}$, it may be possible to identify a characteristic relationship between $\phi_{sm}$ and initial critical stress.
Figure 2-7 Critical stress for initiation of erosion ($\tau_{c0}$) from this study and a number of published works (see Table 2-2), as a function of a. total solids volume fraction ($\phi_{\text{tot}}$) and b. solids volume fraction of mud matrix ($\phi_{\text{sm}}$).
<table>
<thead>
<tr>
<th>Reference</th>
<th>Location</th>
<th>Sediment Source</th>
<th>Instrument</th>
</tr>
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<tbody>
<tr>
<td>Panagiotopoulos et al. (1997)</td>
<td>Laboratory - remoulded</td>
<td>Natural estuarine mud</td>
<td>Recirculating Flume</td>
</tr>
<tr>
<td>Williamson and Ockenden (1996)</td>
<td>In situ</td>
<td>Intertidal mudflat-Blue Anchor Bay</td>
<td>ISIS - bell</td>
</tr>
<tr>
<td>Houwing (1999)</td>
<td>In situ</td>
<td>Intertidal mudflat-Dutch Wadden Sea</td>
<td>ISEF – in situ flume</td>
</tr>
<tr>
<td>Sanford and Maa (2001)</td>
<td>In situ</td>
<td>Chesapeake Bay</td>
<td>Sea Carousel</td>
</tr>
<tr>
<td>Widdows et al. (2007)</td>
<td>In situ</td>
<td>Intertidal mudflat-multiple sites</td>
<td>In situ flume</td>
</tr>
<tr>
<td>Widdows et al. (1998)</td>
<td>In situ</td>
<td>Intertidal mudflat-Humber Estuary</td>
<td>In situ mini-flume (flume also similar results but not shown)</td>
</tr>
<tr>
<td>Amos et al. (1997)</td>
<td>In situ</td>
<td>Fraser River Delta</td>
<td>Sea Carousel</td>
</tr>
<tr>
<td>This study</td>
<td>Ex situ – undisturbed cores</td>
<td>River, Chesapeake Bay</td>
<td>Gust microcosm</td>
</tr>
</tbody>
</table>

Table 2-2 Sources and relevant information for data included in Figure 2-7.
However, the scatter about the fit between critical stress and $\phi_{sm}$ in Figure 2-7b is still considerable, and if considering a subset of these data with a relatively small range in $\phi_{sm}$ (0.2 to 0.25 for instance), $\phi_{sm}$ may be a relatively poor predictor of initial critical stress. This suggests that when the range in $\phi_{sm}$ is small, other factors may overwhelm this relationship. Consistent with this, the relatively small range in $\phi_{sm}$ shown in Figures 2-3b and 2-5 may explain the lack of relationship found in this study. It seems likely that a similar parallel may be drawn to other bed properties commonly related to erodibility in natural sediments. For example, studies that have found a clear and strong relationship between biological parameters and erodibility, such as colloidal carbohydrate concentration (Lucas et al., 2003) or macrofauna abundances (Widdows et al., 2000; Andersen et al., 2005), typically include a large range in these biological parameters, often spanning orders of magnitude.

The relatively small range in $\phi_{sm}$ and the lack of a relationship between $\phi_{sm}$ and erodibility in this study suggest that seasonal changes in $\phi_{sm}$ are not responsible for the observed variability in bed erodibility at the three York River sites included in this study. In a highly productive environment such as the Chesapeake Bay, it seems likely that biologically induced changes in cohesion may be largely responsible for the observed variation in bed erodibility. Analysis of seasonal variations in the concentration and composition of organic matter at each of the sites visited in this study is ongoing. We also plan to further analyze the significance of clay versus silt fraction as a potential control on erodibility using a Multisizer 3 Coulter Counter. Nonetheless, $\phi_{sm}$ may still be
the dominant variable distinguishing our York River data as a whole from the erodibility results from the other systems in Figure 2-7b.

2.5 Conclusions

Seabed erodibility at three mid-depth muddy sites within the York River was found to be highly variable over a 14-month time period. The eroded mass at 0.4 Pa varied by an order of magnitude, ranging from about 0.2 kg m\(^{-2}\) to 2.0 kg m\(^{-2}\). The total solids volume (sand plus mud) fraction (\(\phi_{sot}\)) of the top centimeter of sediment also varied extensively but was not significantly related to observed changes in erodibility. Consistent with the findings of other studies, variations in \(\phi_{sot}\) were largely a function of the sand fraction of the bed. The analytical relationship between \(\phi_{sot}\), sand fraction, and solids volume fraction of the mud matrix alone (\(\phi_{sm}\)) provided in Sanford (2008) was shown to reasonably represent observed changes in \(\phi_{sot}\). Further, this relationship was modified to allow determination of \(\phi_{sm}\) provided \(\phi_{sot}\) and sand fraction were known.

The variability of \(\phi_{sm}\) was found to be significantly less than the variability of \(\phi_{sot}\), a finding consistent with the inference that changes in \(\phi_{sot}\) driven by sand fraction do not control erodibility in muddy environments. Although the York River data presented in this study demonstrated no clear relationship between \(\phi_{sm}\) and erodibility, we believe this to be a result of the small range in \(\phi_{sm}\) observed at the three sites included in this study. In spite of a lack of correlation within the surficial sediments at selected sites in the York
River, we believe $\phi_{\text{sm}}$ to be a better parameter to relate to erodibility than $\phi_{\text{slot}}$. Relating $\phi_{\text{sm}}$ to erodibility eliminates apparent increases in compaction resulting from the addition of sand to a muddy bed.

Including available published data from other systems when evaluating the relationship between $\phi_{\text{sm}}$ and initial critical stress for erosion effectively increased the range in observed $\phi_{\text{sm}}$ and revealed a surprisingly strong correlation across eight of ten studies considered, including the York. Furthermore, the correlation with $\phi_{\text{sm}}$ was significantly stronger than that between $\phi_{\text{slot}}$ and initial critical shear stress. This suggests that across a range of settings it may be advantageous to use $\phi_{\text{sm}}$ as a proxy for the initial critical stress for erosion. However it is also apparent that other influences may result in significant deviations from this general relationship, including fine fraction grain size, mineralogy, biological influences, and other effects.
Chapter 3

Seasonal and spatial variation in cohesive sediment erodibility in the York River estuary: physical deposition versus biological reworking

Dikhuudt, P.J., Friedrichs, C.T., Schaffner, L.C., Sanford, L.P., Diaz, R.J. Seasonal and spatial variation in cohesive sediment erodibility in the York River estuary: physical deposition versus biological reworking. To be submitted to Continental Shelf Research.
3 Seasonal and spatial variation in cohesive sediment erodibility in the York River estuary: physical deposition versus biological reworking

Abstract

Sediment erodibility was measured at three sites on the York River estuary monthly to bimonthly from April 2006 through October 2007. Erodibility at the three sites was similar during the summer and fall. One site near Gloucester Point maintained this level of erodibility greater than 90% of the time while the two sites in the more physically dominated Clay Bank region exhibited a consistent and pronounced increase in erodibility in the late winter and spring. Weak to non-existent correlations between bed erodibility, solids volume fraction, and surficial concentrations of organic matter, colloidal carbohydrate, and extracellular polymeric substances, were not sufficient to explain the observed seasonal pattern at Clay Bank. Digital X-radiographs revealed thick sequences (10 – 20+ cm) of laminated sediments at the surface at Clay Bank coincident with the period of highest erodibility and more biologically reworked sediment during the rest of the year, suggesting that periodic rapid deposition introduced new sediment that was seasonally easy to erode. The finding that seasonal deposition influenced erodibility in the Clay Bank region is consistent with previous results indicating the occasional
presence of a secondary turbidity maximum. Comparison of the biologically reworked, but still “low” erodibility condition in the York to other published Chesapeake Bay erodibility data revealed a consistent critical shear stress range and profile, suggesting this equilibrium critical stress profile may be representative of other similar estuarine environments in the absence of rapid deposition. At relatively low stresses and in the absence of rapid deposition, we speculate that burrowing and/or pelletization may play a role in maintaining high equilibrium bulk water content without reducing the strength of the surface of the seabed.

3.1 Introduction

Cohesive sediments in marine and coastal environments are responsible for degrading water quality, influencing the cycling and availability of particle bound contaminants, and filling in navigable waterways. In spite of their significance to coastal environments, there are still major gaps in our understanding of many of the fundamental processes governing cohesive sediment transport and deposition. In particular, seabed erodibility, which facilitates sediment exchange between the bed and water column, is under-resolved in field studies and is often treated as a tuning parameter in numerical models.

The predominant physical influences on the erodibility of muddy sediments are the degree of consolidation and the distribution of grain sizes making up the bed. (e.g., Kandiah, 1974; Jepson et al., 1997; Roberts et al., 1998). The critical stress for the
initiation of motion of a pure sand bed increases with grain size and is independent of time. Inter-particle bonds in abiotic, cohesive beds result in a very different relationship, with critical stress for erosion increasing with decreasing grain size and increasing consolidation time. Laboratory studies have shown that even at mud to sand ratios as low as 0.05 to 0.2, a mixed bed begins to act as a cohesive bed (Panagiotopoulos et al., 1997; Torfs et al., 2001). Thus, even for a bed with a relatively small mud fraction, the degree of consolidation of the bed, often represented as bulk density or solids fraction, can play an important role in determining bed erodibility. In the absence of biological effects, the degree of consolidation of the bed is largely a function of the deposition rate and consolidation time (Parchure and Mehta, 1985; Jepsen et al., 1997; Roberts et al., 1998; Sanford, 2008). As a result, spatial or temporal variations in deposition may greatly affect local patterns in bed erodibility.

In shallow environments, sufficient light reaches the seabed that biofilms are common. In this setting, biological influences have generally been characterized as biostabilizing or biodestabilizing, acting to respectively decrease or increase sediment erodibility. Biofilms on mudflats, often quantified by measuring chlorophyll-a (Chl-a) or extracellular polymeric substances (EPS), are secreted by benthic diatoms and have been shown to stabilize the seabed and decrease sediment erodibility (Andersen, 2001; de Brouwer et al., 2005). On mudflats, Chl-a and EPS levels often vary seasonally with minima in winter and maxima in late spring and early summer (de Brouwer et al., 2000; Widdows et al., 2004). Potentially counteracting biostabilizing influences, the feeding
and bioturbation rates of benthic organisms also reach a maximum during the warmer months, which has been shown to result in a seasonal maximum in biodestabilization (de Brouwer et al., 2000; Widdows et al., 2000; Andersen, 2001). In a deeper water, estuarine environment such as the Chesapeake Bay, benthic communities and their influence on the estuarine seabed can vary significantly over space and time due to variations in salinity, hydrodynamic processes, and sediment dynamics (Schaffner et al., 1987a; Schaffner et al., 2001). In either mudflat or deeper water, estuarine settings, the net biological influence on erodibility may represent the integration of numerous biostabilizing and destabilizing influences each of which may be variable in space and time.

Measurements of cohesive sediment erodibility in coastal and marine environments are typically costly and time consuming (Sanford, 2006). As a result, seabed erodibility is often poorly resolved in both space and time. While numerous exceptions do exist (e.g., Sanford and Maa, 2001; Maa and Kim, 2002; Stevens et al., 2007), cohesive bed erodibility studies which attempt to systematically resolve seasonal and spatial gradients have most often been conducted in mudflat environments. Fewer published works address temporal and spatial variations of erodibility in deeper, turbid environments were little light reaches the seabed. As the stabilizing effects of biofilms are likely to be less well developed in these environments it is difficult to draw parallels as to the dominant influences on seabed erodibility.

The goal of this work was to evaluate variation in seabed erodibility over several seasons in a cohesive estuarine environment and gain insight into the dominant physical
Figure 3-1 a. Map of Chesapeake Bay, USA. Black box indicates region included in Figure 3-1b. b. Map of York River estuary including sampling station locations. GP = Gloucester Point station, CC = Clay Bank channel station, CS = Clay Bank secondary channel station.
and biological processes influencing these variations. Following a description of our field
site, methodology, and main results, our findings are discussed in terms of the dominant
seasonal cycle and the main differences between the more biologically and physically
influenced regions of the estuary. Next we discuss the impact of key biological and
physical interactions with regard to equilibrium and disequilibrium bed conditions.
Finally, we close with a conceptual model for the dominant controls on spatially and
temporally evolving bed erodibility in the York River estuary.

3.2 Regional Setting

This study was conducted on the York River estuary, a sub-estuary of the
Chesapeake Bay, USA (Figure 3-1). The York River, with a mean tidal range of about 80
cm, is classified as a microtidal estuary. However, this system is tidally energetic with
strong tidal currents reaching magnitudes of ~ 1m s\(^{-1}\) at the surface during spring tide
(Schaffner et al., 2001) and capable of resuspending large amounts of sediment
(Friedrichs et al., 2000). Sediment supply to the York River is a combination of riverine
input (~55%), shoreline erosion (~12%), and influx from the Chesapeake Bay main stem
(~33%; Nichols, 1991). However, the ratio of annually introduced sediment to internally
recycled sediment is small. Based on \(^{210}\text{Pb}\) accumulation rates, mixing depths, and
sediment supply, Dellapenna et al. (1998, 2003) estimated the sediment residence time
prior to burial to be about 70 to 200 years.
A persistent estuarine turbidity maximum (ETM) has been reported just upstream of West Point in both the Pamunkey and Mattaponi Rivers, the York’s main tributaries. Additionally, an ephemeral secondary turbidity maximum (STM) has been reported in the Clay Bank region, about 25 km from the York River mouth (Lin and Kuo, 2001). Near bottom suspended sediment concentrations in the STM commonly exceed 100 mg l⁻¹ at slack tide (Lin and Kuo, 2001) and may reach concentrations as high as 2 g l⁻¹ during strong tidal currents (Friedrichs et al., 2000). Using ⁷Be as a tracer for recently deposited sediment, Romine (2004) observed a muddy deposit migrating with the location of the Pamunkey River ETM suggesting a mobile pool of cohesive sediment actively transported in association with evolving hydrodynamic conditions.

Radioisotope geochronology, used in previous studies in this region, has indicated an active seabed with physical reworking depths of as much as 1 meter (Dellapenna et al., 1998; Kniskern and Kuehl, 2003). Additionally, stair-stepped ²¹⁰Pb profiles and X-radiographs with laminated sequences, bioturbated sequences, and hialtal surfaces suggest a history of active erosion and deposition (Schaffner et al., 1987b; Dellapenna et al., 1998; Kniskern and Kuehl, 2003). However, the mechanism and time-scale associated with deep physical mixing and erosion and deposition “events” remain unclear. Severe fetch limitation results in waves seldom exceeding heights of 0.5 m and periods of 2 to 3 seconds (Vandever, 2007). In this weak wave climate, cohesive sediment resuspension is dominated by tidal currents (Boon, 1996). However, suspended sediment stratification has been observed to limit tidal resuspension (Friedrichs et al., 2000) at concentrations on
the order of 2 g l\(^{-1}\) indicating a limit to direct resuspension by tidal currents on the order of 1 to 2 centimeters.

With regards to benthic biology, Schaffner et al. (2001) reported a strong estuarine gradient in ecological diversity from the ETM region of the York into the main stem of the Chesapeake Bay. Both the ETM and STM regions were characterized by strong tides, high suspended sediment concentrations, and evidence of physical bed reworking which resulted in unfavorable conditions for benthic biota. The resulting benthic community in these regions were typically impoverished and consisted of predominantly short lived, shallow dwelling, opportunistic species. In contrast, the main stem of the Chesapeake Bay had weaker tides, the occasional influence of local wind waves, and lower suspended sediment concentrations. The resulting benthic community was diverse, longer lived, and capable of higher bioturbation rates than those found in the ETM and STM regions of the middle to upper York.

Three sites were chosen in the York River (Figure 3-1) to provide variation in erodibility and the associated physical and biologic influences. Two of these sites were located near Clay Bank, about 25 km from the York River mouth, with a typical salinity of 10 to 20 ppt. This region is commonly occupied by a STM and is characterized by strong tidal currents and high suspended sediment concentrations (100s of mg l\(^{-1}\)). It falls toward the physically dominated end of the biologic gradient described by Schaffner et al. (2001). One site at Clay Bank was located on the flank of the main channel at \(\sim 11\) m depth (CC), while the other site was about one kilometer away in a secondary channel at
~6 m depth (CS). A third site was chosen near Gloucester Point (GP), about 10 km from the York River mouth and at a depth of 8 m and typical salinity of 13 to 23 ppt. Suspended sediment concentrations near Gloucester Point (10s of mg l\(^{-1}\)) are typically much less than those found at Clay Bank. This site falls between the physically and biologically dominated end members of the estuarine gradient described by Schaffner et al. (2001). Conditions are generally more favorable for benthic fauna down-estuary of the Clay Bank region and potentially result in less physical dominance in sediment transport processes at the GP site.

3.3 Materials and Methods

3.3.1 Sampling

Three sites in the York River were sampled monthly to bimonthly from April 2006 through October 2007 (Table 3-1). The dates and times of sampling were determined primarily by logistical concerns and to coincide with slack tide. As a result, sampling took place randomly with respect to tidal phase (ebb/flood) and the spring/neap cycle. Whenever possible, the three sites were sampled on consecutive days.

All sediments were collected from a small vessel (~8 meters), on anchor, at or near slack tide, with an Ocean Instruments Gomex box corer (surface area 625 cm\(^2\)). Individual samples were sub-sampled from the box corer. Each time a site was visited, two cores were collected for erodibility measurements, three cores were collected for
digital X-radiography, and numerous (12-24) samples of the top 5 mm of sediment were collected for analysis of organic components. Due to the large number of samples required and relatively small size of the box corer, numerous box cores were collected while swinging on anchor. As a result, all data presented here likely incorporated spatial variation within a radius of roughly 15 to 25 m at each site.

Due to the interest of this study in the surface of the seabed and easily eroded material, particular attention was paid to minimize disturbance of the sediment-water interface during core collection and sub-sampling. The soft nature of these sediments allowed collection of 20 to 40 cm long box cores using only the weight of the corer. As a result, the Gomex could be slowly lowered to the seabed, allowed to penetrate under its own weight, and then slowly retrieved. Lids on top of the box corer automatically closed upon retrieval, preventing agitation to the sediment surface due to sloshing of the overlying water. Box cores and sub-cores with turbid water (exceeding the ambient conditions) overlying the sediment surface were assumed to be disturbed and were rejected.

3.3.2 Erodibility measurements

Seabed erodibility was measured with a dual core Gust erosion microcosm system. This device uses a rotating disc with central suction to impose a nearly uniform
<table>
<thead>
<tr>
<th>Site</th>
<th>GP</th>
<th>CC</th>
<th>CS</th>
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</thead>
<tbody>
<tr>
<td></td>
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<td>5/3/06</td>
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<tr>
<td></td>
<td>7/14/06</td>
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<td>8/23/06</td>
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<td></td>
<td>9/25/06</td>
<td>9/29/06</td>
<td>9/28/06</td>
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<td></td>
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<td>11/17/06</td>
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<td>10/15/07</td>
<td>10/17/07</td>
</tr>
</tbody>
</table>

**Table 3-1** Sampling dates for erodibility measurements at the three York River sites. GP = Gloucester Point, CC = Clay Bank channel, and CS = Clay Bank secondary channel.
shear stress on the surface of two 10 cm diameter sediment cores (Gust and Mueller, 1997). Cores chosen for erosion testing were carefully selected to ensure uniform, level surfaces and minimal disturbance of the sediment-water interface. After collection, cores were carefully transported to the Virginia Institute of Marine Science located on the York River adjacent to the GP site. Erodibility measurements were generally underway within about two hours of core collection to minimize consolidation effects.

Erodibility measurements consisted of a sequence of seven steps of approximately 20-min duration, each with a consecutively increasing applied shear stress (0.01 Pa, 0.05 Pa, 0.1 Pa, 0.2 Pa, 0.3 Pa, 0.45 Pa, 0.6 Pa) applied to the sediment surface. Each erodibility measurement included two cores from a given site eroded simultaneously. Results were analyzed with the erosion formulation of Sanford and Maa (2001). For a more detailed description of these erodibility measurements see Chapter 2 of this thesis.

Sediment laden effluent from the Gust microcosm system was collected and then filtered with 0.7 μm GF/F filters. After being dried and weighed to determine total suspended solids, these filters were muffled at 550 degrees Celsius to determine the loss on ignition organic content for each applied stress step of each core eroded.

3.3.3 Digital X-radiography

Slabs for digital X-radiography measuring ~12 cm in width, 2.5 cm thick, and 10 to 20 cm in length were sub-sampled from the Gomex box corer. Upon returning to
VIMS, digital X-radiographs of the slabs were collected with a coupled X-ray generator and Paxscan 4030R Amorphous Silicon Digital X-ray detector system. This system was connected to a desktop computer providing nearly instantaneous high quality X-ray images. The grayscale of digital X-radiographs was indicative of the relative density of the sediment. X-ray negatives are presented here where lighter shades of gray indicate higher density while darker shades indicate lower density. Images were adjusted to optimize contrast and aid in identification of sedimentary structures. As a result, in this paper grayscale levels vary between images and the relative grayscale are not compared between images to evaluate changes in bed density.

3.3.4 Solids volume fraction mud matrix

Solids volume fraction of the mud matrix ($\phi_{sm}$), averaged over the top centimeter of the bed, was calculated from measurements of the total solids volume fraction ($\phi_{soc}$) and the sand fraction ($f_s$). Sediment water content was determined by standard wet weight/dry weight analysis. Total solids volume fraction ($\phi_{sol}$) was calculated from the water content by assuming the density of water to be 1015 kg m$^{-3}$ and the density of solids to be 2650 kg m$^{-3}$. Sand fraction was determined by sieving sediments with a 63 \textmu m sieve with the sand fraction equal to the mass of sand (retained on sieve) per total mass dry weight sediment. Following the method described in Chapter 2, solids volume fraction of the mud matrix ($\phi_{sm}$) was calculated as
Figure 3-2 Time series of eroded mass at 0.4 Pa at a. Gloucester Point (GP) site, b. Clay Bank channel (CC) site, c. Clay Bank secondary channel (CS) site. Stars are values for each of the two individual cores eroded; bars indicate the mean value. d. Time-series of water temperature in the York River estuary near Clay Bank. Data from Virginia Estuarine and Coastal Observing System (VECOS, http://www2.vims.edu/vecos). e. Time-series of Pamunkey River flow (solid black line) measured at Hanover, VA. Data
3.3.5 Colloidal Carbohydrates and EPS

Colloidal carbohydrate and extracellular polymeric substance (EPS) concentrations were determined by Dr. Schaffner’s lab. Sediment samples were collected for analysis from multiple box cores using 5 ml syringes. Twelve to twenty-four sediment samples were collected from the top 5 mm of sediment each time a site was visited. Sediment colloidal carbohydrate concentrations were determined following the methodology described in Underwood et al. (1995). Sediment EPS concentrations were determined following the methodology described in Smith and Underwood (1998, 2000).

3.3.6 Other environmental data

River flow data for the Pamunkey River at Hanover, VA was acquired from the USGS Water Data website at http://waterdata.usgs.gov/nwis. Water temperature and salinity data near Clay Bank were acquired from the Virginia Estuarine and Coastal Observing System (VECOS) website at http://www2.vims.edu/vecos. These data were collected near the seafloor in about 1.5 meters of water on the northeast shoal of the York River just outside the main channel.
3.4 Results

3.4.1 Erodibility

Erodibility measurements from the Gust microcosm system were analyzed using the erosion formula of Sanford and Maa (2001) to produce profiles of the critical stress for erosion vs. eroded mass. Rather than compare the 72 critical stress profiles derived in this study, these data have been simplified and presented as the total eroded mass at 0.4 Pa in Figures 3-2a, 3-2b, and 3-2c. 0.4 Pa was chosen as this integrates the results of the majority of an erodibility measurement and represents the typical maximum tidal shear stress in the York River (Scully and Friedrichs, 2003).

The two sites in the Clay Bank region (CC and CS) exhibited a similar and pronounced seasonal cycle in erodibility (Figures 3-2b and 3-2c). The lowest erodibility was found in the summer and fall. These sites had a relatively low level of erodibility from July 2006 through November 2006 and from July 2007 through October 2007. Erodibility was quite consistent during these periods, ranging from 0.10 to 0.32 kg m\(^{-2}\) at 0.4 Pa. The combined data from both sites over these two periods of low erodibility had a mean eroded mass at 0.4 Pa of 0.22 kg m\(^{-2}\) with a standard deviation of 0.074 (data from 28 eroded cores).

In contrast, the highest levels of erodibility were found in the late winter and spring. The CC site in April and May 2007 and the CS site in March, April, and May...
2007 had eroded mass at 0.4 Pa ranging from ~1.2 to 2.0 kg m\(^2\) representing a five to eight fold increase over the observed conditions of the previous summer and fall. Moderate erodibility (~0.5 to 0.7 kg m\(^2\) at 0.4 Pa) observed in January 2007 at CS and January and March 2007 at CC appear to represent transitional periods, as erodibility had increased from the summer/fall low erodibility condition but had not yet reached the very high erodibility observed in the spring. Similar erodibility at both sites in May 2006 may be another transitional period. In this case, erodibility may have been decreasing from a period of higher erodibility prior to the onset of this time series.

Erodibility at the GP site (Figure 3-2a) was generally low and did not exhibit the pronounced seasonal pattern found at the two Clay Bank sites. Excluding May 2007, the eroded mass at 0.4 Pa at GP ranged from 0.05 to 0.32 kg m\(^2\) with a mean eroded mass of 0.18 kg m\(^2\) and standard deviation of 0.089 (data from 22 eroded cores). Interestingly, the range in eroded mass and mean eroded mass found at GP nearly year-round was quite similar to that measured at the Clay Bank sites in the summer and fall of 2006 and 2007. The exception to this was May 2007, when the eroded mass at 0.4 Pa at GP was considerably elevated relative to all other data from this site. The eroded mass increased by roughly an order of magnitude from 0.19 kg m\(^2\) in April to 1.62 kg m\(^2\) in May and then decreased again to 0.16 kg m\(^2\) in June. In contrast to the Clay Bank sites, there did not appear to be any transitional period of increasing or decreasing erodibility at GP, suggesting that the increased erodibility in May 2007 may have been short lived.
Figure 3-3 Digital X-radiograph negative images from the York River, VA. Month and year of sample collection is indicated above each X-radiograph. Site associated with each sample is located in lower right hand corner of each image. GP = Gloucester Point, CS = Clay Bank secondary channel, and CC = Clay Bank channel.
3.4.2 Digital X-radiography

Digital X-radiographs were collected coincident with cores for erodibility measurements to provide insight into the condition of the seabed and the relative influence of biological and physical processes operating on the seabed. The presence of laminated bedding highlighted times when physical processes dominated bioturbation due to either a lack of abundant bioturbators or rapid deposition and/or reworking overwhelming the ability of benthic biota to mix the seabed. Mixed or mottled sediments were indicative of times with either very active bioturbation or low rates of sediment deposition (Schaffner et al., 1987b).

Digital X-radiographs from the GP site (Figures 3-3a, 3-3b, 3-3c) reveal little temporal variability in the structure of the seabed. Sediments from this site appeared moderately bioturbated when compared to other published results from the Chesapeake Bay main stem and York River (Schaffner et al., 2001; Dellapenna et al., 1998). Fine scale laminations were rare but occasionally observed in the top 3-5 cm of the bed. Coincident with a substantial increase in erodibility in May 2007, a 1 to 2 cm thick layer of more X-ray transparent material was present in May 2007 (Figure 3-3b) but not observed in April or June 2007 (Figures 3-3a and 3-3c), suggesting recent deposition of an ephemeral, thin layer of highly erodible material.

Similar to the GP site, digital X-radiographs from the CS site often appear mixed with few laminations (Figure 3-3d). However, a more uniform surface layer, occasionally
containing thin laminations and ranging in thickness from 1 to 15 cm thick, was occasionally observed (Figures 3-3e and 3-3f). This layer was most distinct in March, April, and May 2007.

Digital X-radiographs from the CC site revealed the most dramatic seasonal variability in bed structure. From May through September 2006 the bed appeared mottled and relatively featureless (Figure 3-3g). In November, a surface layer, approximately 2 cm thick and containing many fine laminations appeared (Figure 3-3h). Similar fine laminations were then observed in January 2007 (Figure 3-3i) down to a depth of about 10 cm. By March and April 2007, the laminations spanned the entire length of the X-ray cores (~15 – 18 cm) with many thick laminations (~1 cm) observed throughout (Figures 3-3j and 3-3k). In May 2007 only traces of laminations remained and by June 2007 laminations were absent from X-radiographs and the cores appeared generally mottled and featureless once again (Figure 3-3l).

3.4.3 Solids fraction, sand fraction, and organic fraction

Total solids fraction ($\phi_{soc}$), solids fraction of the mud matrix ($\phi_{sm}$), and sand fraction ($f_s$) were determined for sediments from the top centimeter of the bed to evaluate the degree of consolidation of the bed and its relationship to erodibility. Based on the results presented in Chapter 2, $\phi_{sm}$ is expected to be the most direct proxy for erodibility among these variables. Including the data from all three York River sites, $\phi_{sm}$ ranged from about 0.08 to 0.14 (Figure 3-4a) suggesting poorly consolidated sediments. No
Figure 3-4 Eroded mass at 0.4 Pa as a function of: a. solids volume fraction mud matrix of top centimeter of bed (modified from Chapter 2), b. organic fraction of eroded material, from the first 0.3 kg m$^{-2}$ eroded, c. colloidal carbohydrate concentration, d. extracellular polymeric substance (EPS) concentration. Units in c. and d. are milligrams dry weight glucose equivalents per gram dry weight sediment. X’s in b indicate cores in which less than 0.3 kg m$^{-2}$ total sediment was eroded during erosion measurement. EPS and colloidal carbohydrate data provided by Dr. Schaffner’s lab.
significant correlation was found between the eroded mass at 0.4 Pa and $\phi_{sm}$, $\phi_{slot}$, or $f_s$ (Table 3-2; see also Chapter 2). Values of $\phi_{sm}$ during periods of low erodibility spanned the entire range, while $\phi_{sm}$ associated with periods of highest erodibility fell roughly in the middle of the range.

The organic fraction (by weight) of the eroded sediment was determined by loss on ignition for each applied stress step of each core eroded. There was a tendency for the material eroded at lower stresses to be enriched in organic matter relative to the material eroded at higher stresses. The organic fraction at an applied stress of 0.05 Pa was often 0.2 or greater although the total mass of material eroded was generally quite small at this low applied stress. In general, the organic fraction decreased with depth into the bed (as measured by eroded mass) and the total organic fraction decreased with total mass eroded. An aim of this study was to evaluate the influence of organic material on erodibility, making it necessary to prevent biasing of the organic fraction by its correlation with the total mass eroded. To do so, we always analyzed the same portion of the upper seabed for organic matter to allow comparison of the same mass depth interval of sediment.

The organic fraction of eroded sediment as shown in Figure 3-4b represents the organic fraction of the first 0.3 kg m$^{-2}$ of sediment eroded from each core. A weak but statistically significant ($p<0.01$) negative correlation was found between the eroded mass at 0.4 Pa and the organic fraction of eroded material to a mass depth of 0.3 kg m$^{-2}$ (Figure
<table>
<thead>
<tr>
<th></th>
<th>$\phi_{stot}$</th>
<th>$\phi_{sm}$</th>
<th>$f_s$</th>
<th>Organic fraction eroded sediment</th>
<th>Colloidal Carbohydrate</th>
<th>EPS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eroded mass @ 0.4 Pa</td>
<td>R 0.191</td>
<td>0.0667</td>
<td>0.131</td>
<td>-0.445</td>
<td>-0.341</td>
<td>0.0526</td>
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<tr>
<td></td>
<td>p 0.313</td>
<td>0.726</td>
<td>0.492</td>
<td>&lt;0.01</td>
<td>0.130</td>
<td>0.821</td>
</tr>
<tr>
<td># samples</td>
<td>30</td>
<td>30</td>
<td>30</td>
<td>57</td>
<td>21</td>
<td>21</td>
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<td>4/06 – 10/07</td>
<td>9/06 – 6/07</td>
<td>9/06 – 6/07</td>
<td></td>
</tr>
</tbody>
</table>

**Table 3-2** Correlation table relating eroded mass at 0.4 Pa to selected measured sediment properties. \(R\) is Pearson’s rho. \(p\) is p-value. \# of samples refers to the number of samples used to determine \(R\). For $\phi_{stot}$, $\phi_{sm}$, $f_s$, colloidal carbohydrates, and EPS, this represents the number of mean values of each property that was related to eroded mass. For organic fraction eroded sediment, this value represents the number of eroded cores included in this comparison.
3.4.4 Colloidal carbohydrates and extracellular polymeric substances

The concentrations of colloidal carbohydrates and extracellular polymeric substances (EPS) were measured as a proxy for biostabilization by benthic and/or particle associated organisms. In contrast to the erodibility results and digital X-radiograph comparisons, EPS concentrations were relatively similar for all three sites (Figure 3-4d). Colloidal carbohydrate concentrations tended to be higher at CC than the other two sites (Figure 3-4c). Including all three sites, colloidal carbohydrates concentration varied by about a factor of 3, ranging from about 4 to 13 milligrams dry weight glucose equivalents per gram dry weight sediment (mg dw glu eq / g dw sed). EPS concentration varied by approximately an order of magnitude, ranging between 0.3 and 2.5 mg dw glu eq / g dw sed. However, the majority of samples fell within a smaller range, with approximately two-thirds of the samples having colloidal carbohydrates concentrations of 5 to 10 mg dw glu eq / g dw sed and EPS concentrations of 1 to 2 mg dw glu eq / g dw sed. There were no apparent seasonal variations in colloidal carbohydrate or EPS concentrations.
Comparison of colloidal carbohydrates and EPS concentrations and eroded mass at 0.4 Pa revealed no significant correlation (Figure 3-4c and 3-4d and Table 3-2).

3.5 Discussion

3.5.1 Seasonal cycle in erodibility: The role of biology versus deposition

Two sites (CS and CC) located in the Clay Bank region of the York River estuary exhibited a pronounced seasonal cycle in seabed erodibility (Figure 3-2). Both CS and CC had consistent and relatively low erodibility in the summer and fall of 2006 and 2007 and elevated erodibility in the late winter and spring of 2007. CS and CC also exhibited what appear to be transitional periods of moderate and presumably increasing erodibility prior to the periods of highest erodibility found in late winter and spring of 2007. Similar levels of moderate erodibility in May 2006 may represent a transitional period of decreasing erodibility after a period of elevated erodibility prior to the onset of sampling. Further evidence suggesting this may be a recurring seasonal pattern was presented in Maa and Kim (2002). They presented results of erodibility measurements conducted with a Sea Carousel in the Clay Bank region about 4 kilometers up-estuary from the CS site (also in the secondary channel) in winter, spring, summer, and fall of 1995. They found the lowest critical stress for initiation of motion and by far the highest erosion rate to occur in May. In fact, their reported erosion rates in May were 80 to 400 times what they reported for summer and winter.
With respect to biological influences on erodibility, it would be expected that the magnitude of bioturbation would be in phase with seasonal variations in water temperature (Bender and Davis, 1984; Gerino et al., 1998). Thus, we expected a bioturbation maximum when water temperature was high (Figure 3-2d) resulting in enhanced erodibility in summer and fall (Widdows et al., 2000; de Brouwer et al., 2000; Andersen, 2001). Contrary to this, the highest measured erodibility occurred in the late winter and spring when the water temperature was low. Further indication that bioturbation was not the cause of elevated erodibility was found in the X-radiograph images from CC (Figures 3-3j and 3-3k) depicting strongly laminated sediments coincident with periods of highest erodibility and indicating a seabed sculpted by physical processes with relatively little disruption by bioturbators.

Potentially counteracting the influence of biodestabilization, biostabilizing influences can bind sediment together and decrease erodibility. In mudflat environments, where biofilms are common, colloidal carbohydrate and extracellular polymeric substance concentrations have proven to be proxies for bioadhesion (Paterson et al., 2000; Widdows et al., 2004). In these settings, erodibility has been negatively correlated to colloidal carbohydrate and/or EPS concentrations. However, it was not clear that similar results should be expected at depths of 5 - 12 meters in the Chesapeake Bay, an environment where little light reaches the seabed and that lacks the diatom mats found on tidal mudflats.
Contrasting the results from mudflat settings, colloidal carbohydrate and EPS concentrations sampled at the GP and Clay Bank sites did not correlate well to variations in bed erodibility (Figures 3-4c and 3-4d and Table 3-2). This suggests that either biostabilization was not responsible for variations in bed erodibility or that colloidal carbohydrates and EPS were not the appropriate measure of biostabilization in this environment. Further, colloidal carbohydrate and EPS concentrations at our sites were relatively consistent in both space and time. This indicates that while EPS was a potential and likely influence on sediment cohesivity in the York River, it did not vary significantly and thus did little to influence seasonal and spatial variations in bed erodibility.

Previous studies in the York River have indicated that the Clay Bank region is generally physically dominated, with strong tidal currents, high suspended sediment concentrations (Lin and Kuo, 2001), records of erosion and deposition in the seabed (Dellapenna et al., 1998) and a depauperate benthic community dominated by short lived opportunistic species (Schaffner et al., 2001). Lin and Kuo (2001) found this region to be seasonally occupied by an ephemeral secondary turbidity maximum (STM), the location of which was related to river flow and salinity. Their results demonstrated that in the Clay Bank region, high suspended sediment concentrations associated with a STM occurred in winter and/or spring following periods of elevated river flow, decreased salinity, and increased stratification (Figure 3-5). While the present study did not include hydrographic surveys documenting the location of the STM, the very high seabed
Figure 3-5  a. Time-series of Pamunkey River flow (solid gray line) measured at Hanover, VA, and depth-averaged salinity (dashed black line) in the York River near Clay Bank. Squares indicate timing of surveys from Lin and Kuo (2001).  b. Time-series of bottom – surface salinity along York River.  c. Time-series of near bottom total suspended solids (TSS) along York River. In Figures b and c, Gloucester Point, Clay Bank and West Point are ~ 10, 25 and 55 km from the York River mouth, respectively. Arrows highlight time-lags between episodes of high river flow, salinity stratification, and high sediment concentration in the Clay Bank region. River flow data from USGS Water Data website at http://waterdata.usgs.gov/wvix. All other data from Lin and Kuo (2001).
erodibility which we observed seasonally in the Clay Bank region also followed high river discharge and low salinity (see Figure 3-2), and we conclude that high erodibility was most likely associated with the seasonal STM.

While relatively low levels of erodibility were observed near the onset of high river flow conditions in November, it is expected that there should be some lag period while conditions favorable for convergent sediment transport develop, and sufficient convergence occurs to result in a large increase in bed erodibility. The gradual increase in erodibility at Clay Bank from November through March (Figures 3-2b and 3-2c) indicates this process may occur on a time scale of months. Additionally, the lack of a similarly clear seasonal pattern at GP (Figure 3-2a), outside the region where the STM has normally been observed, further supports this to be the process causing seasonal increases in erodibility at Clay Bank. Finally, the deposition of a thick sequence of laminated sediments at CC, beginning in November 2007 and reaching a thickness of greater than 15 cm in March and April 2007 (Figure 3-3h - 3-3k), indicates periodic rapid deposition in this area. The deposition of more than 15 cm of sediment in about 6 months greatly exceeds the long-term accumulation rate in this area of about 0.2 cm yr\(^{-1}\) (Dellapenna et al., 1998), indicating the deposit itself may be ephemeral, presumably migrating seasonally with the location of the STM. Further indication of periodic rapid deposition occurring near Clay Bank in the spring was presented in Kniskern and Kuehl (2003) who found thick laminated sequences in the surface sediments of the Clay Bank

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region in March through May of 1999. Additionally, they found the depth of maximum $^{137}$Cs penetration to increase by about 20 cm from February to April of 1999.

3.5.2 Limited spatial variation in erodibility: Indications of a background equilibrium condition

Our measurements from the York River (Figure 3-1), as well as comparison to other published works, provide insight on spatial variation in bed erodibility in the York River estuary and Chesapeake Bay. Small-scale variation on the order of meters to tens of meters was evaluated by comparing erodibility at a given site from duplicate cores during a single sampling period and from multiple cores at a given site during extended periods of relatively constant bed erodibility. Variation on the order of a kilometer was evaluated by comparing the two Clay Bank sites, while variation over 10s to hundreds of kilometers was addressed by comparing the results from the Clay Bank sites to the GP site as well as to other published data from the Chesapeake Bay.

Cores were collected for erodibility measurements while swinging on anchor, and no two erodibility cores were collected from the same box core. Consequently, the variation in results from a given site and sampling period incorporate spatial variation on the order of meters to tens of meters, providing an indication of relatively small-scale variation in bed erodibility. As some of this variation may be due to core collection and handling, this comparison may overestimate small-scale variations. For individual pairs of cores run simultaneously, the eroded mass at 0.4 Pa had an average deviation of only
9% from the mean. On many occasions, the two cores are nearly identical, indicating consistency in the sampling and measurement procedure and negligible small-scale spatial variation. Additionally, the regular occurrence of a consistent level of relatively low erodibility (for example August, September, November 2006 at CC) suggests that even spatial variations associated with attempting to re-occupy a given site (~50 to 100 meters) have minor effects on the measured erodibility.

Comparing the time-series data from the two Clay Bank sites indicates that even on a spatial scale of kilometers there may not be substantial variability in bed erodibility within a given environment (in this case mid-depth and muddy). The two Clay Bank sites were approximately 1 kilometer apart and at different depths (6 m vs. 11 m). Despite this, both sites exhibited quite similar behavior both in terms of the mass of sediment eroded and the timing of seasonal changes in bed erodibility (Figures 3-2b and 3-2c). Only upon comparing the results from the Clay Bank sites to GP were large spatial differences in bed erodibility occasionally observed. GP did not exhibit the seasonal pattern found at Clay Bank and five of the twelve times sampled, eroded mass at the Clay Bank sites was two to ten times that at GP. However, during the remaining seven times each site was sampled, all between the months of June and November in both 2006 and 2007, the eroded mass was quite similar for all three sites.

Erodibility was quite consistent for nearly half of each year, on a spatial scale of tens of kilometers along the axis of the estuary, and at depths ranging from about 6 to 11 meters. Impressively, combining all the GP data (except for May 2007) with
**Figure 3-6** Profile of critical stress for erosion versus eroded mass including: **a.** data and power law best fits from Sanford and Maa (2001), Sanford (2006), and York River “low” erodibility periods, **b.** data from Figure 3-6a as well as York River “transitional” and “high” erodibility data and power law best fits. YRI = York River low erodibility, YRt = York River transitional erodibility, YRh = York River high erodibility, SM01 = Sanford and Maa (2001), S06 = Sanford (2006), YRIf = fit to YRI, YRtf = fit to YRt, YRhf = fit to YRh, SM01f = fit to SM01, and S06f = fit to S06.
data from both Clay Bank sites from July, August, September, and November of 2006 and June, August, and October of 2007 yielded a mean eroded mass at 0.4 Pa of 0.21 kg m\(^{-2}\) with a standard deviation of only 0.082. This included 50 of the 72 total cores eroded and indicated a surprisingly consistent level of erodibility within the York River much of the time. This may represent an equilibrium background condition characteristic of mid-depth muddy sites in this system provided the absence of recent deposition.

The results of several other published erodibility measurements from elsewhere in the Chesapeake Bay region are quite similar to the results presented here from the York River estuary. Sanford and Maa (2001) presented the results of erodibility measurements conducted with a Sea Carousel at four sites in Baltimore Harbor, MD. Additionally, Sanford (2006) presented erodibility measurements conducted with a Gust microcosm system from one site in the main stem of the upper Chesapeake Bay. Data from the present study representing the “low” erodibility condition as discussed in the previous paragraph are presented along with data from Sanford and Maa (2001) and Sanford (2006) in Figure 3-6a. Best fit power law regressions in the form of \(\tau_c = am^b\), where \(\tau_c\) = critical stress for erosion, \(m\) = cumulative eroded mass, and \(a\) and \(b\) are coefficients, were applied to the Sanford and Maa (2001), and Sanford (2006), and York River data (Figure 3-6; Table 3-3). Interestingly, the range and best-fit power law regression for the four Baltimore Harbor sites from Sanford and Maa (2001) were nearly identical to the range and a best-fit power law regression for the “low” erodibility condition presented in this study. Additionally, the data from the upper-Chesapeake Bay main stem site presented in
Sanford (2006) fell roughly in the middle of the range of data from both the York River and Baltimore Harbor.

Figure 3-6b contains the data presented in Figure 3-6a as well as the higher erodibility data from the York River. The entire York River data set displayed in Figure 3-6b was grouped into 3 categories based on erodibility. Low erodibility cases had a mean eroded mass at 0.4 Pa of less than 0.4 kg m\(^2\), transitional cases had a mean eroded mass at 0.4 Pa of 0.4 to 0.8 kg m\(^2\), and high erodibility cases had a mean eroded mass at 0.4 Pa of greater than 0.8 kg m\(^2\). All periods in the transitional group were from the CC or CS sites and immediately preceded an extended period of either low or high erodibility. Clearly, the largest variations in bed erodibility were observed seasonally at Clay Bank, indicating that turbidity maximum zones may experience seasonal changes in sediment flux convergence resulting in a dynamic seabed and highly variable bed erodibility. Additionally, it appears that muddy areas outside turbidity maximum zones generally exhibit a somewhat consistent level of bed erodibility in both space and time.

The data presented in Figure 3-6a and relationship in Table 3-3 (for either Sanford and Maa (2001) or York River low erodibility cases) may be a good approximation of the range and mean of this apparently common equilibrium condition of bed erodibility in muddy areas of moderate depth in the Chesapeake Bay.
<table>
<thead>
<tr>
<th>Name</th>
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<th>b</th>
<th>$R^2$</th>
<th># of samples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sanford and Maa (2001)</td>
<td>0.862</td>
<td>0.526</td>
<td>0.88</td>
<td>6</td>
</tr>
<tr>
<td>Sanford (2006)</td>
<td>0.590</td>
<td>0.349</td>
<td>0.95</td>
<td>2</td>
</tr>
<tr>
<td>York River (low erodibility)</td>
<td>0.832</td>
<td>0.486</td>
<td>0.87</td>
<td>50</td>
</tr>
<tr>
<td>York River (transitional)</td>
<td>0.540</td>
<td>0.590</td>
<td>0.91</td>
<td>10</td>
</tr>
<tr>
<td>York River (High erodibility)</td>
<td>0.280</td>
<td>0.620</td>
<td>0.84</td>
<td>12</td>
</tr>
</tbody>
</table>

Table 3-3 Power law fit parameters relating critical stress for erosion to eroded mass. Fit was in the form $\tau_c = am^b$, where $\tau_c$ = critical stress for erosion (Pa) and $m$ = cumulative eroded mass (kg m$^{-2}$). For Sanford and Maa (2001), # of samples represents the number of individual deployments of the VIMS Sea Carousel. For the remaining data, # of samples represents the number of individual cores eroded with a Gust erosion microcosm.
3.5.3 Erodibility and biological impact on solids fraction

We can safely conclude that the seasonal pattern in bed erodibility observed in the Clay Bank region of the York River estuary was ultimately due to temporal and spatial variability in convergent sediment transport processes. However, neither this study nor a related work (see chapter 2) was able to relate changes in erodibility in the York River to bulk physical or biological properties of the bed (Figure 3-4). For the most part, colloidal carbohydrates and EPS proved to be consistent spatially and temporally, suggesting that while these proxies for bioadhesion may still have influenced the overall cohesivity of the bed, they did so more-or-less evenly over the duration and areal extent of our study. Thus, it does not appear that standard biochemical proxies for bioadhesion were primarily responsible for the observed variations in bed erodibility.

Bulk geophysical proxies for bed cohesion / consolidation in the York River estuary (e.g., water content, sand fraction) likewise did not correlate sensibly with bed erodibility (also see Chapter 2). Particularly interesting was the relative consistency in the bulk solids fraction of the mud matrix ($\phi_{sm}$) in the York River seabed in time and space. In contrast, when the range was expanded by including data from seven diverse cohesive environments, $\phi_{sm}$ was shown in Chapter 2 to resolve changes in the bed consolidation state much better than the total solids fraction ($\phi_{tot}$), since $\phi_{tot}$ is often confounded by changes in sand fraction. In the York River, however, even during periods of rapid deposition and high erodibility, there was not a resolvable decrease in
This indicates either a change in cohesivity unrelated to $\phi_{sm}$, or that bulk $\phi_{sm}$ as determined from slices of the top centimeter of sediment was not sufficient to resolve millimeter-scale changes in bed consolidation responsible for changes in erodibility.

The observed variations in erodibility must be reflected in some property of the sediment bed. A possible explanation may be the degree of pelletization of the sediment by biota. Previous studies in the region have observed fecal pellets to represent an important component of spatially varying surficial sediments (Schaffner et al, 1987b; Schaffner et al., 2001) and of suspensions tending to remain near the seabed (Wright et al., 1997). Off the coast of Peru, Reimers (1982) similarly observed changes in the fabric near the surface of the seabed to be associated with the presence of pelletized sediment. Anecdotal observations of eroded material and sediment core surfaces from this study in the York River estuary indicated that pelletized sediment can represent a large fraction of the surficial sediments and eroded material (Figure 3-7). Ongoing observations are focusing on the distribution and nature of these pellets. During periods of low erodibility, dense, fast-settling pellets appeared to dominate the eroded material. Conversely, during periods of higher erodibility, suspensions within the Gust microcosm appeared to be composed primarily of finer material. In the York River estuary, pelletization represents a significant repackaging of muddy sediment, potentially altering the cohesion of the surficial seabed during periods of elevated activity by benthic biota and modifying the relationship between $\phi_{sm}$ and erodibility.
The presence of burrows may be another factor altering the structure of the seabed. However, the net effect of burrows on bulk geophysical measures is not entirely clear. For example, Jones and Jago (1993) found that the presence of burrows in sandy sediment altered seabed structure as indicated by geophysical measures but was not reflected in bulk measures of sediment porosity from sediment samples. Conversely, Rowden et al. (1998) found a positive correlation between the abundance of burrowers and the bulk water content of cohesive sediment. But Murray et al. (2002) point out that some burrowers may increase the shear strength of freshly deposited cohesive sediment by increasing permeability and, thus, hastening dewatering. Images from a benthic camera deployed across the sediment-water interface (Figure 3-8a) in the York River estuary near Gloucester Point in March 2008 revealed heterogeneity in the seabed resulting from burrows. Although we recognize that glass barriers may cause burrowers to congregate, burrowing still represents a potential mechanism for altering the seabed structure and confounding the expected relationship between $\phi_{sm}$ and bed erodibility.

In a physically dominated seabed the structure of the bed is overwhelmed by depositional processes, often resulting in a laterally homogenous bed as illustrated in Figure 3-8b. In this case, the total solids fraction as measured by a bulk sediment sample ($\phi_s$) is equal to the solids fraction of the sediment matrix ($\phi_m$), and $\phi_t$ accurately portrays the microstructure and degree of consolidation of the bed. Conversely, in a biologically altered seabed, burrows and/or pellets may introduce heterogeneity into the bed,
Figure 3-7 Image from sediment profile camera illustrating highly pelletized surficial sediments. Image collected near Gloucester Point in May 2008 following methods of Diaz and Cutter (2001). Camera pixel size is ~69 μm. Sediment image provided by Dr. Diaz.
Figure 3-8: a. Image from sediment profile camera illustrating abundance of burrows in surficial sediments, collected near Gloucester Point in March 2008. Conceptual images depicting: b. homogeneity in a physically dominated seabed, c. heterogeneous seabed containing burrows. $\phi_t = \phi_1$ and $\phi_t = \phi_1 < \phi_2$.

$\phi_t$: bulk solids volume fraction, $\phi_1$: solids volume fraction of sediment matrix, and $\phi_2$: solids volume fraction of sediment matrix. Sediment image provided by Dr. Diaz.
complicating the validity of bulk measures of solids fraction such as the centimeter-thick sediment slices used in this study. The solids fraction of a burrowed bed is illustrated conceptually in Figure 3-8c. In this case, the idealized bed is composed of a non-burrowed sediment matrix (average solids fraction = $\phi_2$) and numerous dissecting burrows (solids fraction = 0). A mixture such as this may have an apparent solids fraction, as measured by a bulk sediment sample, nearly equal to the physically dominated case ($\phi_1 \approx \phi_1$). However, in this case, the solids fraction of the sediment matrix would have to be greater than in the physically dominated bed ($\phi_2 > \phi_1$).

In the presence of burrows or pellets, the relationship between the solids fraction of the sediment matrix found between voids ($\phi_2$) and the solids fraction of the bulk sediment sample ($\phi_1$) can be represented by

$$\phi_2 = \frac{\phi_1}{1 - \phi_1}$$  
(Eq. 2)

where $\phi_1$ is the volume fraction of voids (burrows or space between pellets). If the concentration of voids increases, but $\phi_1$ as measured by bulk sediment samples remains nearly constant, then $\phi_2$, representing either the solids fraction of the matrix between burrows (Figure 3-8c) or the solids fraction of pellets, must have increased. Equation 2 indicates that in a bed composed of 10% burrows, for example, $\phi_2$ will be about 11% greater than $\phi_1$. If $\phi_2$ is actually the more accurate representation of surficial erodibility, then the lower value provided by the bulk solids fraction may mask the true relationship between the fine-scale solids fraction and the critical erosion stress. Figure 3-9 highlights
Figure 3-9 Depth profiles of bulk solids volume fraction of the mud matrix from each of the three sites in the York River. Data was grouped and averaged by erodibility. Profiles named “low” indicate samples collected when eroded mass at 0.4 Pa was less than 0.4 kg m$^2$. Profiles named “high” indicate samples collected when eroded mass at 0.4 Pa was greater than 0.4 kg m$^2$. 
how small the mean differences are in the bulk solids fraction among cores collected in
the York River estuary. The differences in the unsampled microstructure-scale solids
fraction (which accounts for the presence or absence of burrows or density of pellets)
may have been significantly larger.

Since the Gust erosion microcosm quantifies small-scale surface erosion
associated with typical resuspension by individual tidal cycles, controls on our resulting
measurements of erodibility are likely dominated by microstructure-scale bed properties.
We believe that as a result, the 1-cm core slices used in this study may have been too
large to resolve changes in bed consolidation on scales small enough to explain the
observed variations in bed erodibility. It is therefore likely that changes in seabed
structure associated with biologic pelletization and/or burrowing confounded the
observed relationship between $\phi_{\text{sm}}$ and erodibility. It is important to note that while we
believe the microstructure scale may be dominant in controlling surface erosion, it is not
clear how this translates to larger scales of erosion. For instance, the presence of burrows
likely weakens the bed and enhances erodibility when considering erosion on the scale of
several centimeters. In such a case, it is likely that the heterogeneity associated with
burrows may provide a tendency for mass erosion rather than surface erosion.

3.6 Conceptual model and conclusions

The combined results of this and previous studies of the York River estuary
suggest an active system with dynamic and variable sediment transport processes. A
After period of high river flow:
- High suspended sediment concentrations
- Stratified
- Sediment flux convergence
- STM
- High erodibility

After period of low river flow:
- Low suspended sediment concentrations
- Little or no stratification
- No sediment flux convergence
- No STM
- Low erodibility
Figure 3-10 Conceptual diagram depicting sediment transport processes in the York River estuary. Symbols courtesy of the Integration and Application Network (ian.umces.edu/symbols/), University of Maryland Center for Environmental Science.
persistent estuarine turbidity maximum is located near the limit of salt just up-estuary from West Point. Following periods of high river flow, a secondary turbidity maximum (STM) tends to form in the middle estuary near Clay Bank (Figure 3-10). The STM is found near a transition from stratification in the lower estuary to more mixed-conditions up-river, triggered in part by a localized along-estuary decrease in channel depth (Lin and Kuo, 2001). The transition in stratification favors a convergence in up-estuary near-bed sediment flux due to resulting spatial gradients in tidal asymmetries and gravitational circulation.

The presence of the STM near Clay Bank results in high suspended sediment concentrations dominated by fines and flocs, rapid deposition of an ephemeral deposit 10’s of centimeters thick, a physically dominated seabed, and high bed erodibility (Figure 3-10). The lower estuary, including our Gloucester Point site, lies outside of the region typically occupied by the STM. The result at Gloucester Point is lower suspended sediment concentrations, highly pelletized surficial sediment, a more actively burrowed seabed, and low erodibility throughout the year. After an extended period of low river flow, stratification throughout the estuary breaks down, and the STM either moves up-estuary or dissipates, resulting in a period of divergent sediment transport and/or sediment bypassing. During this period, the conditions near Clay Bank more closely resemble Gloucester Point, with lower suspended sediment concentrations, increasingly pelletized sediment, more intense bioturbation near the sediment water interface, and lower erodibility (Figure 3-10).
A similar spatial gradient with regards to suspended particle properties was observed by Fugate and Friedrichs (2003). At Clay Bank in May 2000, Fugate and Friedrichs (2003) documented relatively high concentrations (up to 230 mg l\(^{-1}\)) of fragile suspended flocs that settled at 0.6 mm s\(^{-1}\). At a biologically dominated site in the adjacent Chesapeake Bay, they observed lower concentrations (up to 75 mg l\(^{-1}\)) of resilient suspended pellets that settled at 1.3 mm s\(^{-1}\).

The largest scale variations in bed erodibility appear to be mainly a function of physical processes facilitating periodic sediment flux convergence and rapid deposition of easily erodible material. The commonly observed condition of “low” erodibility likely represents a more biologically influenced bed with “low” erodibility resulting from a dynamic equilibrium between (i) physical resuspension and consolidation and (ii) biological reworking and repackaging. This “equilibrium” magnitude and range of erodibility was observed throughout most of the year at both Clay Bank and Gloucester Point and also in historical data from other Chesapeake Bay sites, suggesting the low erodibility, biologically mediated scenario may be broadly representative of other cohesive estuarine environments in the absence of rapid deposition.

Standard geophysical and biochemical proxies for consolidation and bioadhesion provided no explanation for the observed variations in bed erodibility. The surprising lack of systematic variation in observed colloidal carbohydrate and EPS concentrations suggests that the forms of biostabilization highlighted in erodibility studies in inter-tidal areas may not translate to an estuarine environment where little light reaches the seabed.
The lack of correlation between erodibility and the bulk solids fraction of the mud matrix suggests that the presence of biologic structures may have confounded the expected relationship. In physically dominated cases, a laterally homogeneous bed may allow bulk samples to adequately reflect the relevant microstructure. In contrast, the presence of burrows may cause the bulk solids fraction to be lower than the solids fraction of the millimeter-scale surficial sediment most relevant to erosion by moderate strength tidal currents.
Chapter 4

Conclusions and Future Work
4 Conclusions and Future Work

A time-series of bed erodibility and physical and biological properties of the surficial sediment was used to evaluate seasonal and spatial scales of variability in bed erodibility and the associated bed properties. Erodibility was found to vary substantially in both time and space but this variation was not clearly correlated to the measured properties of the bed.

Chapter 2 specifically focused on relating key physical bed properties to bed erodibility. The total solids fraction of the bed, commonly measured to represent the degree of consolidation of a cohesive sediment bed, was not correlated to bed erodibility. However, a strong correlation between bed sand fraction and total solids fraction indicated that variation in bed solids fraction was largely a function of sand fraction. This suggested that total solids fraction may not be an ideal proxy for bed consolidation and its relationship to cohesive sediment erodibility may be complicated by variations in sand fraction.

To better represent the degree of consolidation of the muddy matrix supporting the bed, an analytical relationship from Sanford (2008) was used to calculate the solids fraction of the mud matrix. This metric of consolidation did not correlate to the observed variations in bed erodibility in the York River. However, when the range was expanded using other published data, the solids fraction of the mud matrix demonstrated a strong
correlation to the initial critical stress for erosion for a range of environments. This suggests that while this approach is valid, the range of solids fraction of the mud matrix at our sites in the York was insufficient to adequately evaluate this parameter as a proxy for consolidation and develop a relationship between solids fraction of the mud matrix and the critical stress for erosion.

The small range in surficial solids fraction of the mud matrix in the York River estuary could be expanded considerably by eroding deeper into the bed. To more adequately assess using the solids fraction of the mud matrix as a proxy for bed consolidation, future work should include larger scales of erosion not possible with the Gust erosion microcosm device used in this study.

Chapter 3 focused on the strong seasonal cycle in erodibility at the Clay Bank sites and contrasted the results from Clay Bank to those at the more biologically influenced Gloucester Point site.

Traditional measures of biostabilization on mudflats were not correlated to variations in bed erodibility. Additionally, observations of seabed structure from digital X-radiographs indicated that bioturbation was not responsible for the elevated erodibility measured at Clay Bank in the winter and spring. However, observations from an in situ seabed camera indicated an abundance of burrows and anecdotal observations from sediment cores often indicated highly pelletized sediment. This and previous works (e.g., Schaffner et al., 1987b; Wright et al., 1997; Schaffner et al., 2001) suggest that biological
reworking and repackaging may have significantly altered particle and seabed properties relevant to sediment erodibility but their influence was not directly resolved in this study.

Future work should attempt to quantify these likely biological influences on bed geophysical properties and erodibility. The abundance of pellets can be directly measured in association with measures of bed erodibility. In the presence of burrows, measuring bed solids fraction at a sufficiently small scale presents challenges but may be possible by decreasing the individual sample volume with a greater number of smaller bulk samples or by using probes, such as a resistivity probe.

The highly variable bed erodibility condition measured near Clay Bank is in agreement with previous studies in this area describing dynamic sediment transport processes associated with an ephemeral secondary turbidity maximum (STM). An extended period of high river discharge in the winter and spring resulted in the development of a STM near Clay Bank that caused rapid deposition and high bed erodibility. This was followed by an extended period of low river discharge in which stratification broke down and the STM was absent resulting in low erodibility. The Gloucester Point site, outside of the region occupied by the STM, was considerably less dynamic than the Clay Bank sites. Bed erodibility near Gloucester Point was typically low and consistent throughout the year.

This work studying bed erodibility in the York River could be complimented with hydrographic surveys measuring salinity and total suspended solids. Along-channel profiles of these parameters could provide insight on time scales and specific conditions
for development of the high erodibility condition as well as the distribution of sediment
during the low erodibility condition. It remains unclear whether the sediment associated
with high erodibility near Clay Bank in the winter and/or spring simply moves up-
estuary, possibly to the primary estuarine turbidity maximum near West Point, or in the
absence of particle trapping near Clay Bank, is distributed throughout the upper estuary.

A comparison of bed erodibility measurements from the York River to historical
data from other sites in the Chesapeake Bay showed remarkable similarities. Critical
stress versus eroded mass profiles from the frequently observed “low” erodibility
condition in the York River were in agreement with similar measurements from
Baltimore Harbor, MD and the upper-Chesapeake Bay main stem. Due to a paucity of
data, bed erodibility is often used as a tuning parameter in numerical models. However,
these data suggest that it may be possible to characterize erodibility over similar,
relatively large areas, making it possible to constrain this parameter to a reasonable range
even with a limited data set.

The findings of this study were complimented by the contributions of past
researchers in the York River estuary. Combining the findings of this study with
previous work in this setting provides a conceptual model of sediment transport processes
in the York River estuary. A persistent estuarine turbidity maximum is found near West
Point associated with the limit of salt. Following an extended period of low river
discharge, minimal stratification in the rest of the estuary results in sediment bypassing
near Clay Bank, relatively low suspended sediment concentrations, a low rate of sediment
deposition, and low bed erodibility. These conditions are favorable for benthic organisms resulting in a bioturbated seabed and pelletized suspended sediments. Following a period of high river discharge, stratification develops in the Clay Bank region facilitating strong sediment flux convergence, high suspended sediment concentrations, rapid deposition, and high bed erodibility. These conditions are generally unfavorable for benthic organisms. As a result, physical processes dominate, the seabed is laminated, and suspensions are composed primarily of fines and flocs. Down estuary of Clay Bank, conditions remain similar to periods of low river discharge with low suspended sediment concentrations consisting of pelletized sediments and low bed erodibility.
Appendix I
Station Log
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Appendix II
Physical sediment data from top centimeter of seabed
Explanation of Appendix II data table headings

**Site #** – Site number. CHSD lab consecutive station number. See appendix I for location information.

**Sample** – Sample identifier. If none given, sample was specifically for determination of physical sediment properties. If identifier contains “x”, sample was X-ray slab subsampled for sediment property analysis after being X-rayed.

**Min Depth** – Minimum depth. Depth from sediment surface of top of extruded and sliced sediment subsample.

**Max Depth** – Maximum depth. Depth from sediment surface of bottom of extruded and sliced sediment subsample.

**Clay** – Clay fraction. Mass of clay per mass total sediment as determined by pipette test grain sizing. Clay fraction defined as sediment with diameter of < 4 µm.

**Silt** – Silt fraction. Mass of silt per mass total sediment as determined by pipette test grain sizing. Silt fraction defined as sediment with diameter of > 4 µm and < 63 µm.

**Sand** – Sand fraction. Mass of sand per mass total sediment as determined by retention on a 63 µm sieve.

**φ_{sot}** – Solids volume fraction. Water content determined by wet weight / dry weight analysis. φ_{sot} determined water content by assuming density of solids to be 2650 kg m\(^{-2}\) and density of water to be 1015 kg m\(^{-2}\).

**φ_{sm}** – Solids volume fraction of mud matrix. Determined from φ_{sot} and sand fraction using Equation 2 from Chapter 2.
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Appendix IV

Gust erosion microcosm time-series plots and stepwise data tables
Explanation of Appendix IV data table headings

**Step** – Consecutive step number.

\( \tau_b \) – Applied shear stress (Pa).

\( \tau_{ce} \) – Critical erosion stress at end of step (Pa)*. If \( \tau_{ce} > \tau_b \) no erosion occurred.

\( m \) – Mass eroded during current step (kg m\(^{-2}\)).

\( M \) – Erosion rate “constant” (kg m\(^{-2} \) s\(^{-1} \) Pa\(^{-1} \))*.

\( E_0 \) – Initial erosion rate (kg m\(^{-2} \) s\(^{-1} \))*. Erosion rate at beginning of each step as determined by fit to erosion rate time series.

\( \lambda \) - Erosion rate decay rate parameter (s\(^{-1} \))*.

* Parameters determined following methods described in Sanford and Maa (2001).
### Applied Stress and Erosion Rate

- **Erosion Rate** \( \text{[kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)
- **Applied Stress** \((\text{Pa})\)

#### Table

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

 Applied stress (Pa)
 erosion rate
fit to erosion rate

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Erosion Rate, \[ \text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \]

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, \( [\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)

- **Applied stress (Pa)**
- **erosion rate**
- **fit to erosion rate**

### Table

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Erosion Rate, [kg \cdot s^{-1} \cdot m^{-2}] 

Applied stress (Pa) vs. time, [s]

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Erosion Rate, \( \text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \)

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)

Erosion rate

Fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

![Graph showing erosion rate over time](image)

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[\text{kg \cdot s}^{-1} \cdot \text{m}^{-2}]$

 Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

![Graph showing erosion rate and applied stress over time](image)

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Erosion Rate, \( \text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \)

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, \(\text{[kg} \cdot \text{s}^{-3} \cdot \text{m}^{-2}\text{]}\)

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

- **Applied stress (Pa)**
- **erosion rate**
- **fit to erosion rate**

<table>
<thead>
<tr>
<th>Step</th>
<th>$\tau_b$ (Pa)</th>
<th>$\tau_{ce}$ (Pa)</th>
<th>$m$ (kg m$^{-2}$)</th>
<th>$M$ (kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$)</th>
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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)

Erosion rate

fit to erosion rate

Step | $\tau_b$ (Pa) | $\tau_{ce}$ (Pa) | $m$ (kg m$^{-2}$) | $M$ (kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$) | $E_0$ (kg m$^{-2}$ s$^{-1}$) | $\lambda$ (s$^{-1}$) | length (s) | organic fraction
--- | --- | --- | --- | --- | --- | --- | --- | ---
1  | 0.0105 | 0.0400 | 0.00439 | NaN | NaN | 8.408e-04 | 1.649e+03 | 0.2329
2  | 0.0505 | 0.0428 | 0.00300 | 5.439e-06 | 5.202e-04 | 4.916e-04 | 6.440e+02 | 0.2598
3  | 0.1004 | 0.0685 | 0.01258 | 1.388e-05 | 2.412e-04 | 4.912e-04 | 1.199e+03 | 0.2122
4  | 0.1995 | 0.1303 | 0.01565 | 1.772e-05 | 1.352e-04 | 5.340e-04 | 1.199e+03 | 0.1753
5  | 0.2994 | 0.2655 | 0.03764 | 6.313e-05 | 3.733e-04 | 1.341e-03 | 1.199e+03 | 0.1444
6  | 0.4492 | 0.4266 | 0.07933 | 1.581e-04 | 8.610e-04 | 1.748e-03 | 1.199e+03 | 0.1166
7  | 0.5979 | 0.4911 | 0.11893 | 1.225e-04 | 7.147e-04 | 3.881e-04 | 1.219e+03 | 0.1098
Erosion Rate, \( \text{[kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)

Applied stress (Pa) vs. erosion rate vs. time, [s]

**Table:**

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<th>Step</th>
<th>( \tau_b ) (Pa)</th>
<th>( \tau_{ce} ) (Pa)</th>
<th>( m ) (kg m(^{-2}))</th>
<th>( M ) (kg m(^{-2}) s(^{-1}) Pa(^{-1}))</th>
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Erosion Rate, \( \text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \)

Applied stress (Pa)

erosion rate

fit to erosion rate

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<tr>
<th>Step</th>
<th>( \tau_b ) (Pa)</th>
<th>( \tau_{ce} ) (Pa)</th>
<th>( m ) (kg m(^{-2}))</th>
<th>( M ) (kg m(^{-2}) s(^{-1}) Pa(^{-1}))</th>
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<th>organic fraction</th>
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Erosion Rate, \([\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]\)

Applied stress (Pa)

erosion rate

fit to erosion rate

<table>
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<tr>
<th>Step</th>
<th>(\tau_b) (Pa)</th>
<th>(\tau_{ce}) (Pa)</th>
<th>(m) (kg m(^{-2}))</th>
<th>(M) (kg m(^{-2})s(^{-1}) Pa(^{-1}))</th>
<th>(E_0) (kg m(^{-2})s(^{-1}))</th>
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Erosion Rate, [kg m⁻² s⁻¹]

Applied stress (Pa)

time, [s]

Erosion Rate, [kg m⁻¹ s⁻¹]

Step | \( \tau_b \) (Pa) | \( \tau_{ce} \) (Pa) | \( m \) (kg m⁻²) | \( M \) (kg m⁻² s⁻¹ Pa⁻¹) | \( E_0 \) (kg m⁻² s⁻¹) | \( \lambda \) (s⁻¹) | length (s) | organic fraction
---|---|---|---|---|---|---|---|---
1 | 0.0105 | 0.0400 | 0.00304 | NaN | NaN | 4.983e⁻⁰⁴ | 1.739e⁺⁰³ | 0.1595
2 | 0.0501 | 0.0408 | 0.00589 | 5.051e⁻⁰⁶ | 5.006e⁻⁰⁴ | 6.678e⁻⁰⁵ | 1.214e⁺⁰³ | 0.1515
3 | 0.1002 | 0.0512 | 0.01163 | 1.066e⁻⁰⁵ | 1.794e⁻⁰⁴ | 1.605e⁻⁰⁵ | 1.199e⁺⁰³ | 0.1541
4 | 0.1994 | 0.1481 | 0.04250 | 5.632e⁻⁰⁵ | 3.800e⁻⁰⁴ | 8.665e⁻⁰⁴ | 1.224e⁺⁰³ | 0.1088
5 | 0.2991 | 0.2443 | 0.04747 | 6.299e⁻⁰⁵ | 4.170e⁻⁰⁴ | 8.453e⁻⁰⁴ | 1.199e⁺⁰³ | 0.1045
6 | 0.4485 | 0.4100 | 0.15953 | 2.737e⁻⁰⁴ | 1.340e⁻⁰³ | 1.392e⁻⁰³ | 1.199e⁺⁰³ | 0.0784
7 | 0.5972 | 0.5391 | 0.16098 | 2.187e⁻⁰⁴ | 1.168e⁻⁰³ | 9.368e⁻⁰⁴ | 1.249e⁺⁰³ | 0.0784
The diagram illustrates the relationship between erosion rate and applied stress over time. The erosion rate is shown in the y-axis, with values ranging from $10^{-6}$ to $10^{-4}$ kg·s$^{-1}$·m$^{-2}$, and the applied stress in the x-axis, ranging from 0 to 8000 s.

The table below provides data for each step:

<table>
<thead>
<tr>
<th>Step</th>
<th>$\tau_b$ (Pa)</th>
<th>$\tau_{ce}$ (Pa)</th>
<th>m (kg m$^{-2}$)</th>
<th>M (kg m$^{-2}$·s$^{-1}$·Pa$^{-1}$)</th>
<th>$E_0$ (kg m$^{-2}$·s$^{-1}$)</th>
<th>$\lambda$ (s$^{-1}$)</th>
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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

![Graph showing erosion rate and applied stress over time.](image)

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<th>Step</th>
<th>$\tau_b$ (Pa)</th>
<th>$\tau_{ce}$ (Pa)</th>
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<th>$M$ (kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$)</th>
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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, \([\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]\)

Applied stress (Pa) vs. time, [s]

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

![Graph showing erosion rate and applied stress over time.]

### Table

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Erosion Rate, \[ \frac{\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}}{\text{t}} \]

Applied stress (Pa)

Erosion rate

Fit to erosion rate

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Erosion Rate, \([\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]\)

Applied stress (Pa) vs. time, [s]

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**Diagram**: Plot of Erosion Rate vs. Time with parameters showing applied stress and erosion rate trends.
Erosion Rate, \( [k g \cdot s^{-1} \cdot m^{-2}] \)

Applied stress (Pa)

erosion rate

fit to erosion rate

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The image contains a graph and a table related to erosion rate. The graph shows the erosion rate as a function of time, with applied stress and fit to erosion rate plotted. The table provides data for each step, including values for τ_b, τ_ce, m, M, E_0, λ, length, and organic fraction.
Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$
Erosion Rate, \( [\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)

Applied stress (Pa)

Erosion rate

Fit to erosion rate

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<th>organic fraction</th>
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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)

erosion rate

fit to erosion rate

<table>
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<tr>
<th>Step</th>
<th>$\tau_b$ (Pa)</th>
<th>$\tau_{ce}$ (Pa)</th>
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<th>$M$ (kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$)</th>
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### Erosion Rate

Erosion Rate, \( \text{[kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

### Table

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<th>Step</th>
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<th>( \tau_{ce} ) (Pa)</th>
<th>( m ) (kg m(^{-2}))</th>
<th>( M ) (kg m(^{-2}) s(^{-1}) Pa(^{-1}))</th>
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### Table

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The figure shows a graph of erosion rate as a function of time. The graph includes a plot of applied stress (Pa) and erosion rate over time, with a fit to the erosion rate data. The table below lists the data for each step:

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<tr>
<th>Step</th>
<th>$\tau_b$ (Pa)</th>
<th>$\tau_{ce}$ (Pa)</th>
<th>$m$ (kg m$^{-2}$)</th>
<th>$M$ (kg m$^{-2}$ s$^{-1}$ Pa$^{-1}$)</th>
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Erosion Rate, [kg m⁻² s⁻¹ Pa⁻¹]

Applied stress (Pa)

erosion rate

fit to erosion rate

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<th>( \tau_b ) (Pa)</th>
<th>( \tau_{ce} ) (Pa)</th>
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<th>M (kg m⁻² s⁻¹ Pa⁻¹)</th>
<th>( E_0 ) (kg m⁻² s⁻¹)</th>
<th>( \lambda ) (s⁻¹)</th>
<th>length (s)</th>
<th>organic fraction</th>
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## Table

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<th>Step</th>
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Erosion Rate, \( [\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}] \)

Applied stress (Pa) vs. time, [s]

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Erosion Rate, \[ \text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \]

**Graph: S3932-10e**

- **Applied stress (Pa)**
- **Erosion rate**
- **Fit to erosion rate**

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, [kg·s⁻¹·m⁻²]

Applied stress (Pa)

Time, [s]

Step | $\tau_b$ (Pa) | $\tau_{ce}$ (Pa) | $m$ (kg m⁻²) | $M$ (kg m⁻² s⁻¹ Pa⁻¹) | $E_0$ (kg m⁻² s⁻¹) | $\lambda$ (s⁻¹) | length (s) | organic fraction
--- | --- | --- | --- | --- | --- | --- | --- | ---
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2 | 0.0503 | 0.0447 | 0.00958 | 1.079e⁻⁵ | 1.045e⁻⁵ | 5.135e⁻⁴ | 1.184e+03 | 0.1583
3 | 0.1005 | 0.0595 | 0.10108 | 9.333e⁻⁵ | 1.674e⁻⁵ | 2.455e⁻⁴ | 1.259e+03 | 0.0767
4 | 0.2002 | 0.1255 | 0.35099 | 3.749e⁻⁴ | 2.665e⁻³ | 5.013e⁻⁴ | 1.264e+03 | 0.0682
5 | 0.3003 | 0.2315 | 0.36374 | 4.479e⁻⁴ | 2.563e⁻³ | 7.470e⁻⁴ | 1.249e+03 | 0.0650
6 | 0.4502 | 0.3031 | 0.24257 | 2.451e⁻⁴ | 1.121e⁻³ | 3.308e⁻⁴ | 1.199e+03 | 0.0647
7 | 0.5996 | 0.5400 | 0.30650 | 5.091e⁻⁴ | 1.717e⁻³ | 1.327e⁻³ | 1.209e+03 | 0.0670
Erosion Rate, \[ \text{[kg \cdot s^{-1} \cdot m^{-2}]} \]

Applied stress (Pa)

Erosion rate

fit to erosion rate

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**Erosion Rate**

![Graph showing erosion rate over time](image)

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Erosion Rate, \([\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]\)

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, [kg·s⁻¹·m⁻²]

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $\tau$, [kg·s$^{-1}$·m$^{-2}$]

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, [kg · s⁻¹ · m⁻²]

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, \([\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]\)

**Graph:**
- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

**Table: Erosion Rate Data**

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa)  
Erosion rate  
fit to erosion rate

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Erosion Rate, \( \text{kg m}^{-2} \cdot \text{s}^{-1} \cdot \text{m}^{-2} \)

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

![Graph showing erosion rate over time with applied stress (Pa) and erosion rate fit.](image)

## Table

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[kg \cdot s^{-1} \cdot m^{-2}]$

Applied stress (Pa)

erosion rate

fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Erosion Rate vs. time, [s]

- Applied stress (Pa)
- Erosion rate
- Fit to erosion rate

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Erosion Rate, $[\text{kg} \cdot \text{s}^{-1} \cdot \text{m}^{-2}]$

Applied stress (Pa) 

erosion rate 

fit to erosion rate

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Appendix V
X-radiographs
Gloucester Point X-radiographs

S3931
4/18/07

S3937
5/21/07

S3940
6/25/07

S3947
10/16/07
Clay Bank Secondary Channel X-radiographs

S3907
5/02/06

S3913
7/19/06

S3917
8/23/06

S3920
9/28/06
Clay Bank Secondary Channel X-radiographs

S3923
11/17/06

S3927
1/12/07

S3930
3/23/07

S3933
4/20/07
Clay Bank Channel X-radiographs

S3908
5/03/06

S3912
7/17/06

S3916
8/22/06

S3921
9/29/06
Clay Bank Channel X-radiographs

S3922
11/15/06

S3926
1/11/07

S3929
3/22/07

S3932
4/19/07
Clay Bank Channel X-radiographs
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Vita

Patrick John Dickhudt