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

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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. Re-evaluation 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.

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1. Introduction

Sediment erodibility is usually quantified as the critical stress for erosion or the erosion rate 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) (Kelly and Guilarte, 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, relationships between physical properties and erodibility are often modified by a variety of biological influences (e.g., bioturbation, pelletization, EPS) (Widdows et al., 2000; Black et al., 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; Riethmuller 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.

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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 was controlled by estuarine circulation. In such a system, where mineralogy varies gradually over relatively large spatial scales (100s of kilometers), the mineralogical composition at a particular site, and likely for a particular region, 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) sediments. 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 issue, the critical shear stress resulting from the presence of sand in the bed. A related study by Van Ledden et al. (2004) provided a conceptual framework to describe the transition from non-cohesive to cohesive behavior as mud is added to a sand matrix supported bed. They explained that the transition from non-cohesive to cohesive is best described by the clay content and occurs when the clay content exceeds only ~5–10%. In a laboratory study, Torfs et al. (2001) measured the critical stress for erosion of artificial and natural sediments at a constant bulk density but with varying sand vs. mud content. As mud content increased above 3–13% (~5% clay), the bed began to behave cohesively, and the critical stress for erosion increased as a function of increasing mud content of the bed. In another study by Panagiopoulos et al. (1997) using estuarine mud mixed with quartz sand showed that as mud content increased from 0% to 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 clay content of ~11–14%. In Panagiopoulos et al. (1997), bulk density was not held constant, and they noted that as the mud content increased, the bulk density decreased.

Similar to Panagiopoulos et al. (1997), a number of other researchers have noted that as sand fraction increases, bulk density also increases. Flemming and Delafontaine (2000) and Riethmuller et al. (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 than 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

\[ \phi_{\text{stot}} = \frac{1}{f_s + 1 - \frac{1}{\phi_{\text{solids}}}} \]  

where \( \phi_{\text{solids}} \) is the bed solids volume fraction=volume solids/volume total, \( \phi_{\text{in}} \) the solids volume fraction of mud matrix=volume mud/(volume mud plus water matrix), \( f_s \) the 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 density of 2650 kg m\(^{-3}\) is suspended in the mud and water matrix of lower bulk density. At \( f_s = 0 \), \( \phi_{\text{stot}} \approx \phi_{\text{in}} \). 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. For a given value of \( \phi_{\text{in}} \), as \( f_s \rightarrow 1 \) the sand matrix begins to support the bed. \( \phi_{\text{stot}} \) reaches a maximum value for \( f_s \) in the range ~0.8–0.97, then decreases to its pure sand value at \( f_s = 1 \) (Sanford, 2008). When \( f_s \) is sufficiently high that the bed is supported by a sand matrix, Eq. (1) clearly does not apply. A similar approach presented in Le Hir et al. (2008) and Waeles et al. (2008) used the relative mud concentration (mud mass/volume not occupied by sand), which is equivalent to \( \phi_{\text{in}} \) multiplied by sediment solids density, as a proxy for compaction of the non-sand portion of the bed. Le Hir et al. (2008) and Waeles et al. (2008) applied this approach to both laboratory erosion experiments and modeling of bed erosion and deposition.

2. Methods

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 m s\(^{-1}\) at the surface during spring tide (Schaffner et al., 2001). Three sites were chosen in the York River estuary to provide variation in erodibility and associated bed properties (Fig. 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 ~11 m depth (CC), while the other site was about 1 km away in a secondary channel at ~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 ~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. 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, clay/silt/sand fraction, and organic content. All sediment samples were subsampled from an Ocean Instruments Gomex box corer (surface area 625 cm²) deployed from a small vessel (24 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 15–25 m.

2.3. Erodibility measurements

Seabed erodibility was measured with a dual core Gust erosion microcosm system (Fig. 2a) 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, 1997). The 10 cm microcosms were calibrated in the lab of Dr. Gust using hot film skin friction sensors.

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 estuary. 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, 0.05, 0.1, 0.2, 0.3, 0.45, 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 (m) 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)

\[ E(m, t) = M(m)\tau_c(t) - \tau_c(m) \]  

(2)

where \( E \) is the erosion rate, \( M \) the depth-varying erosion rate "constant", \( \tau_c \) the stress applied to the bed, and \( \tau_c \) the depth-varying critical stress for erosion. If one locally assumes \( \tau_c \) increases linearly with the depth of erosion, this formulation allows a simple exponential decay of \( E \) with time to be fit to eroded mass observations for each period of constant applied stress. It is then straightforward to solve for \( \tau_c(m) \). For additional background on this analysis method, see Sanford and Maa (2001) and Sanford (2006).

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 were 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 for each 1 cm interval and allowing estimation of the standard deviation for each of the 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. Organic content was determined by loss on ignition at 550 °C after determining 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 (4–63 μm) and clay (<4 μm) 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 them.

3. Results

3.1. Erodibility

Erodibility measurements conducted with the Gust Microcosm system produced profiles of the critical shear stress for erosion and erosion rate constant. The results of erodibility measurements conducted on 60 cores (20 from each site) are presented in Fig. 2b as profiles of the critical stress for erosion (Pa) vs. eroded mass (kg m\(^{-2}\)). 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 (Sanford and Maa, 2001). As a result, we used the critical shear stress for erosion to evaluate sediment erodibility.

An additional ramification of strongly depth-limited erosion is that the data analysis becomes relatively insensitive to the specific erosion rate formulation chosen. Although the Sanford and Maa (2001) relationship for erosion rate was applied here, the controlling factor in determining eroded mass at a given bed stress for our York River data is ultimately the critical shear stress profile.

To evaluate correlations between erodibility and various physical bed properties, the critical stress profiles presented in Fig. 2b have been simplified and are presented as the total erodible mass at a shear stress of 0.4 Pa in Fig. 3 and used in Table 1. This is analogous to the approach 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 (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 more common, in which case the material eroded represents less than the top millimeter of sediment for our cores. (Note that the sediment height in the Gust erosion microcosm is not adjusted during individual experiments to account for the erosion of ~1–8 mm. Specifications presented by Gust (1990) indicate that the stress applied to the bed is not sensitive to the relatively small changes in sediment level occurring in our experiments.)

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. It was not uncommon for erosion to continue until the end of a 20 min step, particularly at the higher applied stresses. Nonetheless, the erosion rate always decreased significantly over the 20 min time step. Since 20 min is short compared to tidal time scales, our high erodibility cases do not contradict depth-limited erosion. The erodible mass at 0.4 Pa was found to vary by roughly an order of magnitude across all sites (Figs. 2b and 3). 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.

3.2. Erodibility and solids volume fraction

A scatter plot of the eroded mass at 0.4 Pa as a function of total solids (sand + mud) volume fraction (φ\(_{\text{tot}}\)) of the top centimeter of sediment is presented in Fig. 3a. Three samples for φ\(_{\text{tot}}\) were collected each time a sample was sampled. The mean φ\(_{\text{tot}}\) ranged from 0.088 to 0.16, and the total standard deviation (SD) from all samples was 0.026. The GP and CS sites displayed the largest variation in φ\(_{\text{tot}}\), and both covered approximately the entire range. The CC site showed less variation (SD 0.021) than either the GP or CS site (SD 0.023 and 0.030, respectively) and tended to have lower values of φ\(_{\text{tot}}\). Variation in φ\(_{\text{tot}}\) for a given set of samples (same site, same day) demonstrated a high degree of variability (mean SD 0.015) 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 \( \phi_{\text{stat}} \) of the top cm of the bed (Table 1). Cases of low
erodibility (\( \approx 0.2 \text{ kg m}^{-2} \)) spanned the entire range of \( \phi_{\text{stat}} \).

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**Table 1**

<table>
<thead>
<tr>
<th>Property</th>
<th>Eroded mass @ 0.4 Pa</th>
<th>( \phi_{\text{stat}} )</th>
<th>( \phi_{\text{sm}} )</th>
<th>Sand fraction</th>
<th>Clay fraction</th>
<th>Clay : silt ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \phi_{\text{stat}} )</td>
<td>0.1906*</td>
<td>0.8116**</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \phi_{\text{sm}} )</td>
<td>0.0667*</td>
<td></td>
<td>0.6395**</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sand fraction</td>
<td>0.1306*</td>
<td>-0.6007**</td>
<td>-0.2720**</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clay fraction</td>
<td>-0.3041**</td>
<td>-0.3041**</td>
<td>-0.2521**</td>
<td>-0.1895**</td>
<td>-0.7992**</td>
<td></td>
</tr>
<tr>
<td>Clay : silt ratio</td>
<td>0.3397*</td>
<td>-0.5142**</td>
<td>-0.3770**</td>
<td>-0.4952**</td>
<td>0.2957</td>
<td>-0.0247</td>
</tr>
<tr>
<td>Organic content</td>
<td>0.1167*</td>
<td>-0.7674**</td>
<td>-0.3845**</td>
<td>-0.7738**</td>
<td>0.5426**</td>
<td>-0.0037</td>
</tr>
</tbody>
</table>

* Indicates \( n=30 \).
** Indicates \( n=72 \).
* Indicates \( n=28 \) and corresponds to organic content data where outliers were excluded as indicated by (+) in Figs. 3f and 4b.
* Indicates significant at \( p < 0.10 \).
** At \( p < 0.05 \).
suggesting that large changes in $\phi_{\text{stot}}$ 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 $\phi_{\text{stot}}$, the observed correlation was opposite (i.e., positive; Table 1), and the cases of highest erodibility fell toward the middle of the range of $\phi_{\text{stot}}$.

3.3. Solids volume fraction, sand fraction, and solids volume fraction mud matrix

Similar to $\phi_{\text{stot}}$, a large variation in the sand fraction of the top cm of the bed was found both within a given site and between sites (Fig. 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 such 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), Riethmuller et al. (2000), and Mahatma (2004), $\phi_{\text{stot}}$ was found to increase with increase in sand fraction of the bed (Table 1; Fig. 4a). Additionally, the analytical relationship (Eq. (1)) between sand fraction and solids volume fraction presented in Sanford (2008) was in agreement with our data, as demonstrated in Fig. 4a. Similar to Flemming and Delafontaine (2000) and Mahatma (2004), and to demonstrate the applicability of Eq. (1), a constant state of consolidation was assumed regardless of sand fraction in order to prepare the fit shown in Fig. 4a. This was implemented by assigning a constant value of 0.11 to the solids volume fraction of the mud matrix ($\phi_{\text{sm}}$).

Clearly there is a good relationship between the data presented in Fig. 4a and Eq. (1) when $\phi_{\text{sm}}$ is assumed to be constant. Even more useful, however, if sand fraction and $\phi_{\text{stot}}$ are known, Eq. (1) can be rearranged to

$$\phi_{\text{stot}} = \frac{\phi_{\text{stot}} - \phi_{\text{sm}}}{1 - \phi_{\text{sm}}}$$

(3)

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

3.4. Erodibility and solids volume fraction of the mud matrix

The solids volume fraction of the mud matrix ($\phi_{\text{sm}}$) was calculated using Eq. (3) for the top 1 cm of sediment from each of the three sites over the 14-month period sampled. A scatter plot of eroded mass at 0.4 Pa as a function of $\phi_{\text{sm}}$ is presented in Fig. 3b. The range of $\phi_{\text{sm}}$ for all three sites is 0.085–0.144, about 20% less than the range in $\phi_{\text{stot}}$ (Fig. 3a). The range in $\phi_{\text{sm}}$ was similar for all sites, eliminating the distinction between the CC site and the other two sites, and suggesting that much of the variability in erodibility (Fig. 3b). In fact, values of $\phi_{\text{sm}}$ varied by roughly an order of magnitude, there was no clear distinction in $\phi_{\text{sm}}$ between periods of high and low erodibility (Fig. 3b). In fact, values of $\phi_{\text{sm}}$ for periods of low erodibility ranged from 0.085 to 0.144, representing the entire range observed in $\phi_{\text{sm}}$ data. Thus for the range of values seen in the York River estuary, $\phi_{\text{sm}}$ reduces variability relative to $\phi_{\text{stot}}$, but $\phi_{\text{sm}}$ does not provide an explanation for the observed variability in erodibility.

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 scatter plot relating clay fraction of the top centimeter of sediment to the eroded mass at 0.4 Pa is presented in Fig. 3c. While the observed clay fraction ranged from 0.30 to 0.76, the correlation between sand fraction and clay fraction reported in Table 1 indicates that much of the variation in clay fraction results from variations in sand fraction. Similarly, the correlation between clay fraction and $\phi_{\text{stot}}$ (Table 1) results from the correlation of clay fraction to sand fraction and the dependence of $\phi_{\text{stot}}$ on sand fraction.
While correlation between competing fractions is common and expected, the correlation between sand and clay illustrates that for beds with a mud supported matrix, clay fraction may not provide a good estimate of cohesivity. Instead, the ratio of the clay fraction to silt fraction was determined and presented in Fig. 3d 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 1.5 and a standard deviation of 0.48 (Fig. 3d), a ratio and range commonly seen in estuarine and intertidal environments (Flemming, 2000). While there was no significant correlation between clay fraction and eroded mass (Table 1), a significant, negative correlation \((p < 0.10; \text{Table } 1)\) 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 mud. However, a negative correlation, significant at \(p < 0.05\), was also found between the clay to silt ratio and \(\phi_{\text{stot}}\). 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.

### 3.6. Organic content

The loss on ignition (LOI) organic content of 30 sediment samples from the top centimeter of the York River seabed ranged from about 0.045 to 0.12. Organic content was shown to have a weak but significant positive correlation to the eroded mass at 0.4 Pa (Table 1, Fig. 3f) and a significant negative correlation to sand fraction (Table 1, Fig. 4b). However, 2 data points, depicted by (+) in Figs. 3f and 4b, appeared to be outliers. Eliminating these points from correlation tests resulted in no significant correlation between eroded mass at 0.4 Pa and organic content and a strong negative correlation between sand fraction and organic content (Table 1). The strong negative correlation to sand fraction and positive correlation to clay fraction suggests that organic matter tends to be associated with clays and silts. As indicated in Fig. 4b, a pure mud bed tends to have an organic content of 0.05–0.10. As the sand fraction is increased, the proportion of mud and the associated organic content decreases.

The lack of a strong correlation between organic content of sediment from the top centimeter of the bed and eroded mass at 0.4 Pa suggests that variations in the organic content of York River sediments are not responsible for modifying bed erodibility. A more detailed evaluation of the influence of organic material on erodibility, presented in Dickhudt et al. (2009), found a weak negative correlation between the organic content of sediments eroded by the Gust microcosm (sampled from effluent) and erodibility, and no correlation between the concentration of common biostabilization proxies (extracellular polymeric substances and colloidal carbohydrate) and bed erodibility. Dickhudt et al. (2009) concluded that while organic material in York River sediment likely contributed to cohesiveness, variations in the concentration of this material were not controlling the observed variability in bed erodibility.

### 4. Discussion

#### 4.1. Limitations of \(\phi_{\text{stot}}\)

The erodibility of muddy beds is largely a function of interparticle cohesion with more compacted beds presumably being less erodible. The magnitude of \(\phi_{\text{stot}}\) of the bed is assumed to reasonably represent the degree of compaction of the bed and is often compared to measures of erodibility. While classic attempts to relate \(\phi_{\text{stot}}\) to erodibility were found to be variable in this study, no clear relationship was found between the two. However, we (Eq. (1) and Fig. 4a) and others (Flemming and Delafontaine, 2000; Panagiotopoulos et al., 1997; Paterson et al., 2000; Riethmuller et al., 2000; Mahatma, 2004; Tolhurst et al., 2006; Stevens et al., 2007) have illustrated that \(\phi_{\text{stot}}\) is highly dependent on the sand fraction of the bed.

While both \(\phi_{\text{stot}}\) and erodibility were found to be variable in this study, no clear relationship was found between the two. However, we (Eq. (1) and Fig. 4a) and others (Flemming and Delafontaine, 2000; Panagiotopoulos et al., 1997; Paterson et al., 2000; Riethmuller et al., 2000; Mahatma, 2004) have illustrated that \(\phi_{\text{stot}}\) is highly dependent on the sand fraction of the bed. This greatly complicates attempts to relate overall bulk density to erodibility. In the simplest sense, comparing \(\phi_{\text{stot}}\) to erodibility for mixed mud/sand beds of varying sand fraction assumes that mud and sand are equally cohesive. However, it is apparent that an increase in \(\phi_{\text{stot}}\) resulting from an increase in sand fraction should not result in the same decrease in erodibility as would be expected for a similar increase in \(\phi_{\text{stot}}\) in a bed composed entirely of mud. This distinction may explain the lack of correlation between erodibility and \(\phi_{\text{stot}}\) found in many studies on mixed sediments.

#### 4.2. Removing the effect of sand on solids fraction

Flemming and Delafontaine (2000) and Mahatma (2004) recognized the influence of sand fraction on \(\phi_{\text{stot}}\) and attempted...
to account for it by deriving empirical relationships between the two. However, these empirical relationships were specific to the authors’ study sites, complicating comparison to other settings. Rather than deriving empirical relationships between $\phi_{\text{stot}}$ and sand fraction, we recommend an analytical solution such as Eq. (3) or the relative mud concentration in Waeles et al. (2008). These analytical relationships directly provide the solids volume fraction of the mud matrix ($f_{sm}$) supporting the bed while accounting for variation in sand fraction. Further, it is unnecessary to derive site-specific relationships, and the solids fraction is effectively normalized. Calculating $f_{sm}$ from $\phi_{\text{stot}}$ 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).

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 increase in fine fraction (Fig. 6a) 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 fine fraction. From Eq. (3), it is apparent that maintaining a constant bulk density while increasing sand fraction must also cause a decrease in $f_{sm}$. Thus the relationship shown in Torfs et al. (2001) illustrating a decrease in critical shear stress with increase in sand fraction also represents a decrease in critical shear stress with decrease in $f_{sm}$.

Eq. (3) predicts a monotonic negative relationship between sand fraction and $f_{sm}$ for sediments of a constant bulk density that approaches linearity as $f_{sm}$, $\phi_{\text{stot}}$, and $f_s$ decrease. Thus, re-evaluating the data of Torfs et al. (2001) in terms of both fine fraction and $f_{sm}$ and including only data with a sand fraction of 0.96 or less, demonstrates that either fine fraction or $f_{sm}$ can be used to predict the initial critical stress for erosion equally well.

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**Figure 6.** (a) Initial critical stress for erosion as a function of fine fraction for a mixed mud/sand bed, re-plotted from Torfs et al. (2001). (b) Initial critical stress for erosion as a function of solids fraction mud matrix ($f_{sm}$), re-evaluated from Torfs et al. (2001). Solids fraction of mud matrix was calculated from bulk density and sand fraction using Eq. (3). (c) Critical stress for initiation of erosion ($\tau_{ic}$) from this study and a number of published works (see Table 2) as a function of total solids volume fraction ($\phi_{\text{stot}}$). (d) Critical stress for initiation of erosion ($\tau_{ic}$) from this study and a number of published works (see Table 2) as a function of solids volume fraction of mud matrix ($f_{sm}$).
for a specific value of designed to evaluate the influence of sand fraction on erodibility. Despite differences in sand fraction of as much as 0.30, the while the CS site was always sandier, ranging between 0.08 and these sites. Sand fraction at the CC site ranged from 0.02 to 0.15 However, the sand fraction was generally quite different between similar patterns in erodibility and had very similar values of \( \phi_{sm} \). 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. Two sites differed 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 approach could easily be evaluated with controlled laboratory experiments. It should also be noted that this approach is only appropriate for mud matrix supported beds with sand that erodes at critical shear stresses similar to or lower than 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 results. As the sand present in our study was about 120 \( \mu \)m in diameter, the critical stress for erosion of the sand was about 0.1 Pa, similar to that of the eroded mud.

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 \( \phi_{stot} \) (or bulk density, water content, etc.), greatly limiting the previous works that could be included in this comparison. As the initial critical stress for erosion proved to be the most common measure of erodibility reported in these previous publications, the results of the present 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 experimental design was not optimized to provide high-resolution measurements of the initial critical stress for erosion. However, the initiation of motion was always constrained by a lower stress step with no erosion and a higher stress step where the initiation of motion occurred. The initial critical stress for our York River data is presented in Figs. 6c and d as the mean of these two applied stresses with bars to indicate the total range of possible values.

The results of eight previous studies were found to include information on sand fraction, \( \phi_{stot} \), and critical stress for the initiation of motion and were considered for this comparison. Of these eight, Torfs et al. (2001) (Fig. 6b) 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 from Figs. 6c and d. Data from the remaining seven and this study (Table 2) were included in Figs. 6c and d. Critical stress data with sand fractions greater than 90% were excluded from these results, as the work of Panagiotopoulos et al. (1997) and Torfs et al. (2001) demonstrated that sediment with sand fraction greater than approximately 90% may not behave cohesively. The remaining data represent a wide range of sand fraction (2–90%), \( \phi_{stot} \) (0.08–0.5), and \( \phi_{sm} \) (0.06–0.39).

Fig. 6c presents a comparison of \( \phi_{stot} \) to the initial critical stress for erosion. While there is a weakly significant positive relationship overall (R² = 0.179), 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 Fig. 6d provides a much stronger and consistently positive relationship. Given the number of devices used and likely inconsistencies in identification of initial critical stress, as well as other potential influences on erodibility (e.g. biology, mineralogy, and grain size), there is a surprisingly strong relationship between \( \phi_{sm} \) and initial critical stress (R² = 0.691). 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.

However, the scatter about the fit between critical stress and \( \phi_{sm} \) in Fig. 6d is still considerable, and if considering a subset of these data with a relatively small range in \( \phi_{sm} \) (0.2–0.25 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 Figs. 3b and 5 may explain the lack of relationship found in this study.

5. Conclusions

Seabed erodibility at three mid-depth muddy sites within the York River estuary 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 to 2.0 kg m⁻². The observed variations in erodibility could not be attributed to changes in clay fraction or organic content. However, both clay fraction and organic content were significantly correlated to sand
Relating studies considered, including the York. Furthermore, the correlation between surficial sediments at selected sites in the York River, we believe included in this study. In spite of a lack of correlation within the general relationship, including fine fraction grain size, mineralogy, biological influences, and other factors. The total solids volume (sand plus mud) fraction ($\phi_{\text{sm}}$) of the top centimeter of sediment varied extensively but was not significantly related to observed changes in erodibility. Consistent with the findings of other studies, variations in $\phi_{\text{stot}}$ were largely a function of the sand fraction of the bed. An analytical relationship between $\phi_{\text{sm}}$, sand fraction, and solids volume fraction of the mud matrix alone ($\phi_{\text{sm}}$) (Eq. (1)) was shown to reasonably represent observed changes in $\phi_{\text{sm}}$. Further, this relationship was modified (Eq. (3)) to allow determination of $\phi_{\text{sm}}$ provided $\phi_{\text{stot}}$ and sand fraction were known.

The variability in $\phi_{\text{sm}}$ thus calculated was found to be significantly less than the variability of $\phi_{\text{stot}}$, a finding consistent with the inference that changes in $\phi_{\text{sm}}$ 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 $\phi_{\text{sm}}$ and erodibility, we believe this to be a result of the small range in $\phi_{\text{sm}}$ 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 $\phi_{\text{stot}}$. 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 nine studies considered, including the York. Furthermore, the correlation with $\phi_{\text{sm}}$ was significantly stronger than that between $\phi_{\text{stot}}$ 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 factors.

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