Muddy Sediment Erosion: Insights from Field Studies

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Studies on cohesive sediment erosion can be traced back to the mid-1950s, and since that time many of the benchmark papers describing the response of settled mud beds to fluid stress have been published in the *Journal of Hydraulic Engineering*. Relatively few main players, in particular three professors, have dominated the stage: R. Krone, A. Mehta, and E. Partheniades. Mehta and Partheniades continue to publish, but sadly Krone passed away in 2000. Their work has provided a cornerstone to cohesive sediment research, outlining the fundamental aspects of the erosion process as well as some of the controlling factors. Following recognition of the importance of the thixotropic properties of mud, there has been a move toward direct in-situ experimentation. Quite literally, the laboratory has been taken into the field. This has been possible through the development of field-portable or "benthic" flumes. Many of these developments have been published in the nonengineering marine scientific literature and have not come to the general notice of the engineering fraternity. The aim of this paper is to provide an overview of the knowledge gained from this work, highlighting both advances and shortcomings, and to place it within the contextual framework of the earlier studies of Krone, Mehta, Partheniades, and others.

Three Professors: A Mini Review

Although a number of early relevant studies are available (summarized in Hollick 1976), E. Partheniades' (1962) Ph.D. studies, submitted as a paper to the *Journal of Hydraulic Engineering* in 1965, were probably the first earnest attempt to describe and quantify the erosion process of estuary mud subjected to fluid stress. This work subsequently stimulated the first comprehensive review of the subject (ASCE Task Committee 1968), and this was revisited by Mehta and Parchure (2000). Partheniades also provided an experimental framework in which sediment samples were placed in a laboratory flume (a straight-recirculating type) and a fluid with preselected physicochemical properties (temperature and salinity) was passed over the bed to cause erosion. Erosion was determined from the time-series of suspended sediment concentration (SSC). Partheniades characterized the erosion process through the form of the stress-flux ($\tau_o - \varepsilon$) relationship, where τ_o is the applied bed-shear stress and ε the erosion rate or mass of sediment eroded per unit bed area per unit time, and was able to derive the critical erosion shear stress for *significant* transport (τ_{ocrit}, units Nm^{-2} [$\equiv$Pa]) and time-mean surface erosion rates for specified stresses (units $\mathrm{kgm}^{-2}\,\mathrm{s}^{-1}$). Various different designations of ε have been used, depending upon the amount of time the data are integrated over, for example, the instantaneous erosion rate (measured over short time scales, usually less than a minute), or peak erosion rate (the maximum instantaneous erosion rate) or the rate after some initial response has passed (Aberle et al. 2006). Other indices that could be calculated from the time-series include the depth of scour (z) and the vertical dry density or dry mass concentration profile ($\delta\rho_d/\delta z$), where ρ is the density of the sediment. Investigation of erosion down through the surface sediment layers allows measurement of the vertical bed strength profile, $\delta\tau_s/\delta z$, where τ_s is the shear strength of the sediment bed. The influence of the vertical variation in shear strength on erosion patterns was investigated in a series of research experiments (Mehta and Partheniades 1975, 1982; Mehta et al. 1982; Parchure and Mehta 1985) by creating two types of bed structures: one in which τ_s increases with depth, notionally representative of a freshly deposited layer in estuaries [e.g., as found by Güst and Morris (1989), for Puget Sound]; and one in which τ_s was relatively constant with depth (akin to exposed, deeper layers). The first type of bed was created through natural sedimentation under slowly moving or stationary flow, the second by pouring a thick slurry of mud into the flume, or through manual reconstitution of an existing bed. These studies revealed the close relation between bed structure and erosion. Under a constant applied stress ε remains constant for uniform beds, but is instantly high and then decays exponentially through time to a constant or equilibrium value (the floc erosion rate) for strength-stratified beds. Although various workers have presented other formulations, two formulations based upon the concept of excess stress that represent these different processes and are widely accepted are

$$\ln\left(\frac{\varepsilon}{\varepsilon_f}\right) = \alpha[(\tau_o - \tau_{s(z)})]^\beta \tag{1}$$

the so-called Ariathurai–Partheniades equation, where τ_o = applied bed-shear stress; $\tau_{s(z)}$= bed-shear strength (analogous to τ_{ocrit}) at depth z; ε_f=floc erosion rate; α = rate coefficient; and β is an exponent; and

$$\varepsilon = \varepsilon_M\left(\frac{\tau_o - \tau_{s(z)}}{\tau_{s(z)}}\right)^\delta \tag{2}$$

where ε_M = rate coefficient; and δ is an exponent.

Surface mud layers may be best represented by Eq. (1); buried layers tend to be more compacted and are often better represented by Eq. (2), highlighting a multiple-layer approach to cohesive sediment erosion. Mehta and Partheniades (1982) labeled the erosion characteristics of stratified mud as "type I" surface erosion and that of uniform mud as "type II" surface erosion. Amos et al. (1993) later coined the respective synonyms "benign" and "chronic" to describe these patterns. Mehta and Partheniades (1982) also showed broad increases in both τ_s and ρ_d (density of sediment at a given depth) through time associated with consolidation, and discounted concurrent deposition as erosion occurs. More recently, Sanford and Maa (2001) present a "unified erosion formulation for fine sediments" that incorporates both types of erosion. They derived an expression for ε (E in their paper) for the specific case of a step increase in bed-shear stress, typical for measurements using erosion devices (Sanford and Maa 2001). Their solution shows that a single equation may be used to describe both type I and type II erosion.

A substantial body of work now exists through the research activities of students of the three professors as well as others that provides insight into factors that mediate or govern the erosion of mud (McCave 1984; Mehta and Parchure 2000). There is insufficient space here to describe all of these developments, but Partheniades recently published "Engineering Properties and Hydraulic Behavior of Cohesive Sediments" provides a comprehensive review of erosion of mud (Partheniades 2007). Controlling factors on sediment erosion that have received attention include the physicochemical properties of the sediment–water system, such as the cationic exchange capacity (CEC) and sodium adsorption ratio (SAR) of the clay fraction, rheological properties such as the yield stress, and the bulk physical properties of the sediment (e.g., voids ratio, density, texture. etc.). A major finding is that the utility of bulk physical or engineering soil indices as predictors of erosion were seriously limited, principally because of the scale miss-match between surface erosion shear strength and the depth-integrated nature of these measures (although see Zreik et al. 1998).

Why Field Flumes? Problems Associated with Sampling Mud

Natural muddy sediments are a complex and delicate milieu. Fine-grained marine sediments typically have median deflocculated particle diameters within the medium-coarse silt size range. They occur as a heterogeneous, poorly mixed matrix of silt and cohesive clay particles together with adhesive (predominantly microbial) and detrital organic material, and flora and fauna of various genera. This compositional complexity, from which arises the tendency toward thixotropic behavior [from the Greek *thixis*, meaning "touch," and indicating liquefaction following physical disturbance (Mitchell 1976)], means that it is virtually impossible to sample mud without disruption either to the fabric or to the internal cohesive strength. Historically many devices, such as corers and grabs, have been invented to sample marine muds for erosion testing, but it is doubtful that any of these instruments successfully retrieved samples that were completely undisturbed. For instance, a bow wave usually precedes submerged corers, which can wash away the delicate surface layers, especially in biologically conditioned sediments (Hawley 1991), or force benthic water horizontally through the more permeable surface layers. Horizontal tensile stresses due to wall drag are a feature of most small-scale coring devices (Parker 1991), giving rise to compression and the formation of a small dome in the center of the core. Further, sedimentologically important biological activities such as burrow cleaning and deposit feeding are often observed to be erratic during sample retrieval and transportation with a commensurate impact on sediment fabric and structure. Currently there is no single parameter that can be used to assess structural change pre- and postsampling, and therefore one can never fully know the degree of disturbance. It is this fact that limits the utility of laboratory-derived data. The preferable methodology is to directly measure erosion in the field, a sentiment echoed several decades ago by Young and Southard (1978). They tested the same sediment in situ and in the laboratory and found that the laboratory method overestimated τ_{ocrit} by a factor of two or more—a finding that really sparked the initial concerted interest in field erosion instrumentation. Since then numerous workers have found differences between laboratory and field measurements. Tolhurst et al. (2000a) showed orders-of-magnitude differences between field and laboratory erosion thresholds, where cores were subjected to vibration disturbance and desiccation during transport back to the laboratory. Maa et al. (2007) found τ_{ocrit} was larger in laboratory measurements, which was attributed to differences in the vertical profiles of τ_{ocrit}. The calm conditions in the laboratory allowed greater consolidation to occur at the surface, but the shorter time for consolidation meant that τ_{ocrit} was comparatively smaller deeper in the sediment. This was reflected in the type of erosion observed, with both type I and type II erosion occurring in the laboratory, but only type I in the field. Ravens (2007) showed that measurements of erosion rate were about 5.5 times greater in sediments brought back to the laboratory compared to measurements made in situ. Because the laboratory and field flumes had different dimensions, which may have been affecting the results, Ravens (2008) investigated the influence of the devices test section length on the erosion rate. The measured erosion rates were 35% greater in the short test-section measurements, showing that device differences only contributed a small amount to the differences found between laboratory and field. Physical disturbance during core collection and/or differences in biota were suggested as possible explanations for the differences found.

Comparisons of field and laboratory studies where care was taken to minimize physical disturbance and disruption of biota generally show less differences, with a broad agreement between field and laboratory measurements, especially for similar types of devices (Tolhurst et al. 2000b; Widdows et al. 2007). However, small differences often remain between laboratory and field measurements, which when scaled up to estuary-wide processes would result in considerable inaccuracy in sediment budget estimates. For example, Schaaff et al. (2006) predicted an increase in SPM of 14 mg l^{-1} from laboratory measurements, but the increase in SPM measured in situ was 10 mg l^{-1}.

Field Instrumentation: Benthic Erosion Devices

A variety of portable erosion devices capable of deployment in the field have been developed (Black and Paterson 1997; Güst and Muller 1997). These include a shear pad, continuous and pulsed vertical jets, an erosion bell, a stirring device, a suction-stirring device, and a myriad of differing geometry flumes. Table 1 summarizes some of the flume type systems that have been constructed to date. The flumes have been of the annular or raceway recirculating type, or straight flow-through type. Devices can be placed on the sediment bed either sub-aerially (e.g., on mudflats) or submerged (Fig. 1), and water is then pumped or driven over the bed until erosion occurs. The applied stress is usually increased in a stepwise fashion, and most devices use fast-response sensors to measure the time evolution of suspended sediment and imposed flow speed. With the exception of the raceway flume of Black (1991) and the ASSET flume of Roberts et al. (2003), which incorporate a bed-load trap in the flume base, benthic flumes measure only the suspended sediment flux.

Limitations of Benthic Devices

The major advantage of benthic devices is that the sediment being tested remains nominally undisturbed. However, these instru-

Table 1. Summary of Some of the Data from In-Situ Benthic Flumes, Including Channel Geometry, Deployment Details, and the Range of Published Critical Entrainment Stresses

	Deployment	Flume type	Environment	τ_{ocrit} (Nm^{-2})
Scoffin (1968)	Submerged	Flow through	Carbonate sands	c. 0.16
Young and Southard (1978)	Submerged	Flow through	Shelf seafloor	0.01–0.07
Manzenrieder (1983)	Subaerial	Flow through	Sandy tidal flat	0.10–7.5
Young and Mann (1985)	Submerged	Flow through	Carbonate sand	0.02–1.0
Güst and Morris (1989)	Submerged	Flow through	Shelf seafloor	0.02–0.22
Hawley (1991)	Submerged	Flow through	Great Lakes	0–0.13
Aberle et al. (2006)	Submerged	Flow through	Various riverine and marine	~0.01–0.55
Ravens and Gschwend (1999)	Submerged	Flow through	Shallow bay	0.03–0.12
Lee et al. 2004	Cores	Recirculating	River sediment	0.07–1.14
Houwing and van Rijn (1993)	Subaerial	Recirculating raceway	Estuary mud	0.2
Houwing (1999)				0.11–0.18
Black (1991); Black and Cramp (1995)	Subaerial	Recirculating raceway	Estuary mud	0.0–0.27
Pierce et al. (1970)	Subaerial	Annulus	Estuary mud	1.6 (Mersey) 3.2–5.1 (Severn)
Amos et al. (1993)	Subaerial and submerged	Annulus	Estuary silt-flat	0.5–2.5
Amos et al. (1996)			Subarctic estuary	0.8–2.0
Amos et al. (1997)			Delta slope	0.11–0.5
Amos et al. (1998)			Estuary mud	0.19–1.22
Maa et al. (1993)	Submerged	Annulus	Estuary mud and beach sand	0.1–0.19; 0.22
Maa et al. (1998)			Estuary mud	0.05–0.10
Maa (2008)			Riverine	0.03–0.08
Widdows et al. (1998)	Subaerial	Annulus	Estuary mud	0.02–1.6
Amos et al. (2004) Moreau et al. (2006)	Submerged	Annulus (miniflume)	Subtidal fjord	0.22–3.06
	Subaerial and submerged		Lagoon	0.07–0.48
Black et al. (2003)	Submerged	Suction/stirring	Shelf sea floor intertidal mud	0.004–0.25
Tolhurst unpublished data	Subaerial			~0.1–0.5
Tolhurst et al. 2000a	Subaerial	Stirring propeller	Intertidal mud/sand flats	0.2–1.95
Andersen 2001				0.2–1.9

Note: Erosion rates are omitted as each relates to a specific imposed stress, and the averaging interval within the time-step varies between workers. This makes intercomparison rather fruitless (Tolhurst et al. 2000a; Sanford and Maa 2001; Widdows et al. 2007). Definitions of and methods of computing τ_{ocrit} also vary between researchers, but data tend to be more comparable.

ments are not "off-the-shelf" devices and there are both operational and fundamental differences contributing to the range of τ_{ocrit} measured by the different devices that require further investigation to unravel (Tolhurst et al. 2000b; Widdows et al. 2007).

Channel size is of fundamental importance, as each of the flumes has a different footprint area. There is some evidence suggesting that different instruments provide a different value of τ_{ocrit} due to different footprints, which incorporate different scales of sediment heterogeneity [i.e., larger devices integrate more heterogeneity (Tolhurst et al. 2000b)]. Amos et al. (2004) assigned differences in threshold estimates of 0.19 ± 0.13 Pa ($21\pm13\%$ of the maximum value) to differences in the footprint size of the Sea Carousel and Mini Flume measuring different scales of spatial heterogeneity, although overall the two devices gave comparable results. In contrast, Widdows et al. (2007) found little evidence of differences due to device size between the 0.17 m^2 PML (Plymouth Marine Laboratory) Annular Flume and the 6.5-fold smaller footprint of the PML Mini Annular Flume. Spatial heterogeneity will, however, only be detected if the footprint size is smaller than the scale of the heterogeneity and these two devices have comparatively large footprints. There have been numerous studies showing that sediment stabilizing microphytobenthos are spatially variable at the scale of centimeters and temporally variable on a scale of minutes (Decho and Fleeger 1988; Sandulli and Pinckney 1999; Tolhurst et al. 2003; Murphy et al. 2008a). It is, therefore, not surprising that flume devices with their relatively large footprints and long deployment times are often unable to resolve small scale spatial heterogeneity. The cohesive strength meter (CSM) pulsed jet device (Tolhurst et al. 1999; Vardy et al. 2007) with its 0.0007 m^2 footprint and ~5 min deployment time is currently the only device capable of resolving such small-scale spatial and temporal variability. It induces a primarily vertical stress, however, which may be more representative of the oscillatory stresses from waves rather than the horizontal stresses from tidal flows, and correlates well with shear strength measures (Watts et al. 2003).

Protocol factors such as the length of time sediments are exposed to standing water prior to an experimental run, the magnitude and duration of time-steps, and the rate of increase of flow velocity during stress increases (the ramping period) also have a bearing on the data obtained (Maa et al. 1998). These problems have been addressed for field devices in a series of intercomparison exercises (Amos et al. 1993; Tolhurst et al. 2000b; Widdows et al. 2007). Widdows et al. (2007) considered the much larger erosion thresholds of intertidal sediments measured with CSM and EROMES (a propeller-driven stirring device) compared to flume devices to be primarily caused by fundamental differences in the applied stress in these devices. Tolhurst et al. (2006a) showed, however, that CSM measures of erosion threshold are significantly reduced when the length of time sediments are exposed to standing water prior to and during an experimental run is considered. CSM measurements not corrected for submergence had an erosion threshold of 0.86 $N\,m^{-2}\pm0.1$, while corrected measurements (immersed for 30 a)

b)

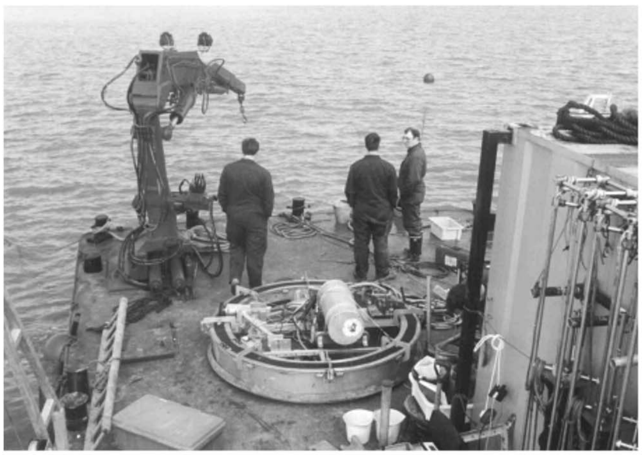


Fig. 1. Two examples of benthic flumes: (a) flume of Widdows et al. (1998) being deployed subaerially in the Humber estuary, U.K., during the U.K. LISP campaign; (b) flume of Carl Amos (Sea carousel) on the barge deck during the U.K. LISP program (Black et al. 1998) (Courtesy of C. Amos, Southampton Oceanography Center)

min) were significantly lower at 0.39 N $m^{-2} \pm 0.07$, compared to a microcosm measurement of 0.21 N m^{-2} (erosion was first noted ~30 min after water was placed in the microcosm chamber). Significant changes in erosion threshold were also measured over a tidal cycle, with increases during exposure and decreases during immersion. These changes cannot be detected by most flume devices because their deployment times are too long (Widdows et al. 2007). Most flumes require a minimum of 30 min. to take a measurement and many require much longer than this—sufficient time for significant changes in the erosion threshold to occur, which cannot be detected when deployment times are longer than ~30 min (Tolhurst et al. 2006a). Device deployment time and footprint size are confounded with each other, as large devices generally require longer deployment times than smaller ones. Thus, the differences noted by Tolhurst et al. (2000b) and attributed to footprint size may be due to deployment time instead. This highlights the importance of submergence/deployment time in measuring erosion threshold and goes some way to explaining why the different erosion devices generate different answers for τ_{ocrit}. This is important for understanding natural processes, because natural tidal/wave applied stresses are usually largest at the very beginning and end of immersion, the times when changes to τ_{ocrit} caused by submergence are at their smallest and greatest.

The distribution of bottom stress within benthic flumes is also of concern. Although natural seafloors are rarely smooth or planar, from an experimental viewpoint it is necessary that flume devices impart a bottom stress that is spatially uniform. This may be mostly true for straight flumes, but severe secondary flows and a radially increasing bottom stress afflict raceway/annular geometries. These radial gradients rise in importance only above a certain device dependent critical flow speed, and thus experiments may confidently be undertaken for flow speeds less than this. The Güst microcosm device (Güst and Muller 1997), which nominally imposes a spatially uniform stress on bottom sediments through a system of opposing hydraulic gradients, and a laboratory carousel built in Chonbuk National University, South Korea (Maa 2008), which allows the top ring and channel bed to rotate in opposite directions to reduce secondary circulation, were invented specifically to solve these problems.

Biological activity in the surface layers of natural mud undoubtedly creates a transitional-rough interface, and yet many flume operators use smooth-walled false floor surfaces to calibrate the applied stress. Not surprisingly, there is often a "hydrodynamic" miss-match between calibration and actual values of τ_o, which has consequences for measured values of τ_{ocrit}, and may also skew the $\tau_o - \varepsilon$ relationship. This problem can be avoided by using a rough calibration boundary—Black and Cramp (1995) used an aluminum plate deliberately pitted by prolonged exposure to seawater, while Aberle et al. (2003) used wood and sandpaper to provide different bed roughness. Prolonged bed erosion invariably produces an increasingly roughened boundary through time and thus the friction experienced by the bed will change (a factor difficult to account for in field experiments).

Device calibrations usually use clean water, yet recirculating devices accumulate suspended sediment in the overlying water column and thus may be susceptible to both interfacial drag reduction and stratification effects (Li and Güst 2000; Cloutier et al. 2006). These processes alter the stress exerted on the bed, increasing error in τ_{ocrit} estimates and potentially invalidating calibrations that do not take such effects into account. These effects do not occur, at least not persistently, in flow-through designs. Of course, natural benthic boundary layers may include these effects (Dyer et al. 2004), or they may not, depending on the particular environment, and it is advisable to take account of this.

One of the biggest oversights of in-situ erosion devices is that they generally create horizontal stresses that mimic tidal flows, yet it is known that natural erosion often correlates better with oscillatory wave activity (Amos et al. 2004). The PES erosion device (Abdelrhman et al. 1996) simulated oscillatory wave action, but in the absence of horizontal flow. Modified laboratory flumes (Paphitis et al. 2001), which incorporate an oscillating plate to simulate wave action, have shown that wave period controls threshold conditions for sand. The laboratory has, again, moved into the field with SEAWOLF (Jepsen et al. 2004), a linear flume based on the SEDflume (McNeil et al. 1996). SEAWOLF simulates wave/tide shear stress and is capable of testing in-situ or laboratory-prepared cores. Jepsen et al. (2004) showed that the shear stress was much larger in undeveloped, oscillatory flow than for fully developed, unidirectional flow at the same flow rate.

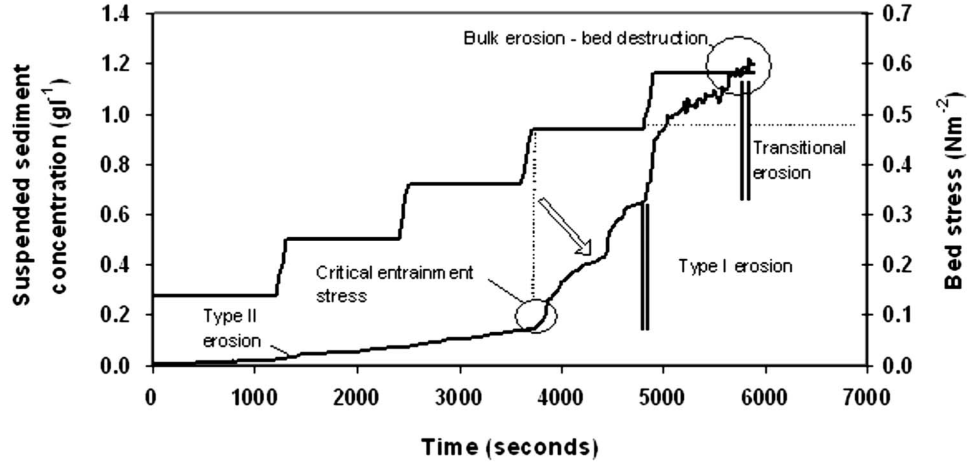


Fig. 2. Example experimental erosion time-series from the Carew estuary (SW Wales, U.K.) showing the applied stepwise stress profile (Nm^{-2}) and the cumulative suspended sediment concentration (gl^{-1}) due to surface erosion

Synthesis of Knowledge from Field Studies

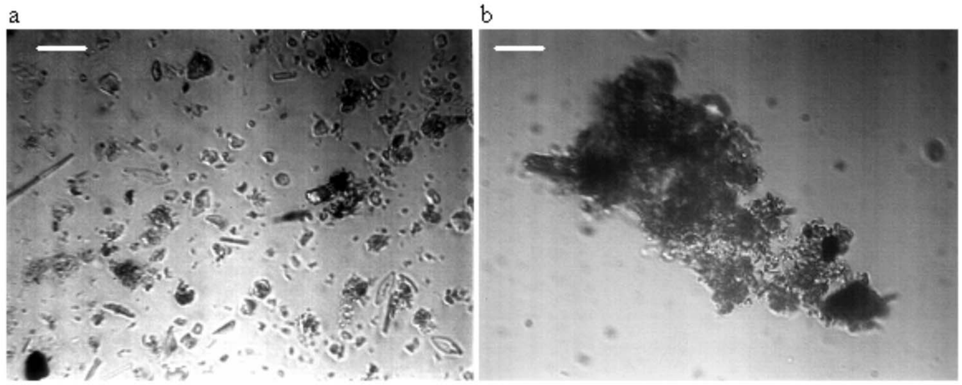
Surficial Phenomena

Benthic flumes have been used in a variety of shallow-water environments from the subarctic to the tropics. The flow-recirculating flumes, for example the Sea Carousel instrument of C. Amos (Southampton Oceanography Centre, U.K.; Amos et al. 1992a), have broadly confirmed the type I/type II surface erosion characterization of Mehta and Partheniades (1982) in the field situation. Fig. 2 shows a typical time-series of increasing SSC with increasing τ_o from the intertidal zone of the Carew estuary, SW Wales, measured with the racetrack flume of Black and Cramp (1995). Qualitatively similar profiles have been reported from a variety of estuarine regions, including the Dutch Wadden Sea (Houwing 1999); the Hudson Bay (Amos et al. 1996); the Humber Estuary (Widdows et al. 2000); the Bay of Fundy (Amos et al. 1992b); and Saguenay Fjord (Moreau et al. 2006). However, the response of natural, heterogeneous, and fine-grained sediment is often more complex. More than two erosion types have been observed and an update to the early classification is therefore necessary. It is important to consider, however, that the response in SSC in flumes to an applied stress is dependent upon whether the system is closed or open [i.e., water is recirculated (closed) or not (open)]. Open flumes do not retain the same sediment-laden water, and therefore SSC can (and usually does) decrease during each step in applied stress. This results in profiles where SSC is largest shortly after the shear stress is increased and then decreases with time. For examples of SSC profiles from open flumes see Ravens and Gschwend (1999) and Aberle et al. (2004); for closed flumes see Amos et al. (1996) and Fig. 4 this paper. There is a need for further experimental data, particularly in relation to the spectrum of hydrodynamic conditions and sediment bed types found in nature and the phase-amplitude characteristics of natural tidal and fluvial flows, before it may be said that the patterns found in flumes represent a universal order to the erosion process.

The following description is an overview in broad process terms of erosion patterns (in terms of SSC) of naturally formed muddy beds (whether intertidal or subtidal) exposed sequentially to low, intermediate, and high applied stress in closed-flume systems. It is accepted that site-specific differences may present departures from the scheme, and that erosion is as much contingent upon the sediment properties as flow stress (for example, erosion induced by a fast current flowing across a hard, dry mud bed will differ substantially from that over a soft, watery deposit).

Low Applied Stress. At low-flow speeds, surface erosion of the organomineral bed matrix occurs, often together with the surface creep (bed-load transport) of pellets and other interfacial debris. Amos et al. (1997) termed this phase "type Ia surface erosion" as the observed trend in SSC was characteristically *asymptotic* in form. By way of contrast, the SSC trend in this phase in the Carew estuary was *linear* (Fig. 2), and may therefore be termed type II surface erosion (Black 1991). Both erosion modes appear characteristic of low-stress surface erosion in the literature; we have also observed type II surface erosion patterns for low values of applied stress (τ_o~0.025 Nm^{-2}) in Loch Creran, Scotland (unpublished data). Black (1991) measured the size spectra of particles entrained into suspension by free-stream azimuthal flow speeds of 5 and 10 cm s^{-1} and confirmed that erosion at low applied stress was through breakage of the weaker particle–particle bonds of surface aggregate networks to yield a fine, mostly primary particle and small particle cluster population (Fig. 3). This is the phenomenon of winnowing, and was obvious from the rapid clouding of the flume water. Mehta and Partheniades (1982) showed that flake-by-flake surface erosion can occur almost indefinitely under low applied stress as there is an almost infinite supply of primary particles and small flocs. The winnowing of the clay fraction by low hydrodynamic stress suggests that there is no true critical entrainment stress, a moot point for sedimentologists (see Lavelle and Mofjeld 1987). The suggestion is that there will always be some (albeit very small) transport for a finite applied flow stress. For intertidal mud, winnowing may also be facilitated or enhanced through clay fraction dispersion, which refers to the detachment of clay particles from the aggregate matrix in the presence of water, and occurs at the moment of tidal submersion.

Intermediate Applied Stress. As flow speed is increased, discernibly larger floc aggregates are entrained into suspension, with the suggestion that the relationship between speed (applied stress) and suspended aggregate size is nonlinear (Fig. 3). This point broadly corresponds to the critical entrainment stress for *significant* erosion (τ_{ocrit}) identified by numerous researchers (Table 1). Thus, τ_{ocrit} for practical purposes is related to the definition of a



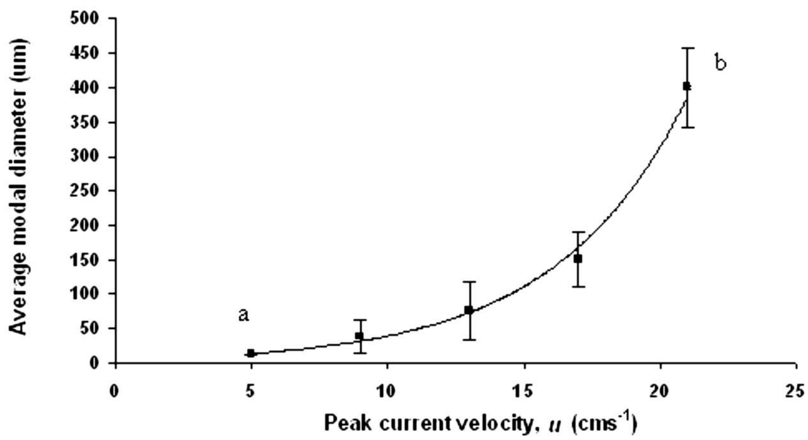


Fig. 3. Relationship of average entrained modal aggregate size ($\bar{M}_o$) to imposed benthic current velocity (u) in the field flume, showing an increase in floc size with current velocity

minimum sediment flux, although the magnitude of this flux remains to be standardized between workers and instruments. The use of a prescribed erosion rate for τ_{ocrit} is currently considered to be of limited use, due to the considerable differences in erosion rate between different devices, even when standardized procedures are used (Tolhurst et al. 2000b; Widdows et al. 2007). Most commonly the trend in SSC for intermediate applied stress is asymptotic to an equilibrium concentration (C_{eq}), although it is not certain from field experiments whether C_{eq} arises from a finite supply of erodible material or a dynamic erosion–deposition equilibrium. Asymptotic erosion patterns have been termed "type I b surface erosion" (Amos et al. 1992a,b), and further increases in stress can give rise to a series of type I b SSC profiles. Type I b erosion thus dominates the central portion of erosion experiments corresponding to intermediate applied stress. Sometimes C_{eq} is not reached (Houwing 2000), which is most probably a function of both the magnitude of the bed stress over and above the shear strength of the surficial aggregate matrix and (more critically) the duration of the applied stress. Maximum or peak type I b erosion rates (ε_p) can approach 10^{-3} kgm^{-2} s^{-1} at the beginning of the time-step (Amos et al. 1996; Maa et al. 1998; Houwing 2000), although these are rarely sustained for more than a minute or so.

High Applied Stress. The penultimate and final high-flow speed time-steps in the majority of field studies reveal a transitional erosion mode wherein erosion is at first rapid (as in type 1) but then decays at a reduced but steady rate within a given time-step (i.e., the trend is nonlinear but *not* asymptotic (Amos et al. 1996). This is apparent in the erosion time-series from the Carew estuary for τ_o=0.58 Nm^{-2} (Fig. 2), and is related to the undercutting and scouring of deeper sediment horizons. Rates of transitional erosion are more variable through time reflecting a more chaotic situation, and mean erosion rates are typically one or two orders of magnitude less than type I erosion rates [~10^{-4} to 10^{-6} kgm^{-2} s^{-1} (Black 1993; Amos et al. 1996; Widdows et al. 1998; Houwing 1999, 2000)]. It is possible that instrument-related factors, such as scouring at flume walls, direct abrasion of saltating and rolling aggregates, and abrasion by particles such as shells (Amos et al. 2000) may also contribute to such behavior.

Confusion has arisen in the literature as to the classification of transitional erosion. On the basis of the trend in SSC, transitional erosion may be described as a nonequilibrium form of type I erosion—if the duration of the time-step were increased then an equilibrium concentration would undoubtedly be achieved either through a balance between erosion and deposition or through a temporal reduction in the supply of bed aggregates. However, in process terms transitional erosion relates to *the onset of destruction of an increasingly rough bed by a highly turbulent flow*, which is indicated through the greater variance in the SSC signal (Fig. 2); it is, therefore, ultimately *not* surface erosion but bulk or mass erosion (Peirce et al. 1970; Amos et al. 1996). Finally, what should strictly be referred to as "type II bulk erosion," in which the trend in SSC is strictly linear through time, is related to the quantum release and entrainment of very large aggregate clusters (mm-cm in size) and is occasionally seen in field studies only at very large boundary stress [> 1.5 Nm^{-2} (Amos et al. 1997)]. It is not evident in the erosion profile from the Carew estuary (Fig. 2) as the maximum applied stress is not that large. Type II bulk erosion is the final stage in bed erosion and probably occurs naturally only under powerful storm or flood currents.

Principal Erosion Patterns

Altogether, field studies have defined five principal different patterns of erosion related to floc detachment and bed behavior, with the inclusion of dispersion for intertidal mud. The initial suspension of surface ultra-low-density layers of organic detritus (fluff) that are invariably associated with muddy sea bottoms could also be included (Jago and Jones 1998). This material has negligible strength, and is washed away by even very slowly moving flows and is unavoidably suspended during flume filling (Tolhurst et al. 2000b; Widdows et al. 2007). Thus, under an increasing applied stress, erosion patterns may be arranged in phenomenological order as (fluff suspension) type 1 a surface erosion or type II surface erosion, type I b surface erosion, transitional erosion, and type II bulk erosion, although the precise sequence will depend on the initial and final applied stress in experiments, the number, duration and magnitude of stress increments, as well as the nature and structure of surface sediment layers. In this phenomenological context, a final but extremely important point should be addressed: the very existence of type I erosion in both laboratory and field studies may, in fact, be an artifact of the stepwise applied stress. Characteristically, peak erosion rates correlate in time to periods when applied stress is incremented (the ramping period; see Fig. 2). Sudden increases in stress of this nature typify perhaps short-duration accelerations at the leading edge of the tidal fringe in macrotidal intertidal environments (Christie and Dyer 1998), during flash floods in fluvial environments, or under storm waves only. Bed response to a smooth, continuously increasing applied stress (characteristic of the major portion of natural tidal and steady fluvial flows) may differ, but this remains to be fully explored in either the laboratory or the field situation. Likewise, type II surface erosion, given sufficient time, will asymptote for a given stress.

Bed-Load Transport (Surface Creep)

The transport of mud has conventionally been viewed in terms of suspension only, principally because the constituent particles are so small it is assumed they will be entrained directly into suspension under most flow conditions. However, there is evidence that indicates that bed-load transportation of cohesive intertidal mud (i.e., rolling along and saltating over the bed surface) does occur (Young and Southard 1978; Amos et al. 1996; Houwing 1999, 2000; Aberle et al. 2004, 2006), and that this may be a function of biogenic pelletization by macrobenthos (Nowell et al. 1981; Andersen 2001; Andersen et al. 2002). This is not a bed load in the classic sense, as there does not appear to be a dynamic and continuous exchange between mobile bed particles and suspended particles, thus bed load transport of mud aggregates or fecal pellets may constitute a different transport regime (Mehta, personal communication, 2000). The term "benthic load" or "surface creep" (as used by Amos) may perhaps be more appropriate.

During field use of the Güst microcosm on estuary mud we have observed the formation of a central cone (due to the flow pattern) due entirely to the rolling of clasts and grain agglomerations over the bed surface. We have also observed ephemeral transport of detritus, biological particles, and loosely bound flocs just as the flooding turbid tidal fringe submerges intertidal muds. Young and Southard (1978) report the appearance of the bed load as "elongate streamers of individual sediment aggregates moving intermittently and parallel with the flow" from Buzzards Bay, which almost exactly matches the descriptions of bed load by Black (1991) from the Carew estuary. Both Grant and Daborn (1994) and Torfs (1997) time-series erosion experiments on retrieved field sediments revealed the rolling of mud clasts over the bed surface. Black (1991) eroded smooth areas of sediment under a series of increasing unidirectional bed-shear stresses (0.14–0.58 Nm^{-2}). Incorporation of a small sediment trap in the flume channel permitted direct measurement of the time-mean surface creep flux. Approximately millimeter-sized (presumably fecal) pellets were eroded in the form of a continuous traction carpet under all applied bed stresses, although the magnitude of the total creep flux (TCF) was negligible in relation to the total suspended sediment flux (TSSF), TSSF:TCF=1.3 - 7.2. Surface creep transport rates varied from 0.005–0.217 g^{-2} s^{-1} over the range of applied bed stresses. There was considerable variability in the creep flux between experiments spaced only meters apart, precluding derivation of any meaningful correlative relation between this and flow stress for the mudflat as a whole. Debnath et al. (2007), using the National Institute of Water and Atmospheric Research (NIWA) in situ Flume II, concluded that bed load may not be negligible and should be included in erosion estimates. The adjustable shear stress erosion and transport flume of Roberts et al. (2003) was specifically designed to allow measurement of both suspended and bed load transport. Visual observation showed that the sediments eroded primarily in aggregate form (size 0.1 to 10 mm) and transported primarily as bed load. These aggregates generally maintained their size and shape as they were transported down the channel.

Spatiotemporal Variability

Knowledge of spatiotemporal gradients in nature comes only from field studies. There is an increasing amount of field information on spatiotemporal gradients in biological and physical properties of sediments from various locations, such as European intertidal mudflats (Black et al. 1998; Dyer 2000; Mehta and McAnally 2001) and Australian mangrove forests/mudflats (Chapman and Tolhurst 2004, 2007; Tolhurst and Chapman 2005, 2007). This work shows that the real world is defiantly variable!

The advent of field flumes has provided an opportunity to examine sediment erosion resistance on a variety of spatial and temporal scales, and much new information has emerged that suggests that single values of τ_{ocrit} (and ε) are not appropriate parameterizations within numerical sediment transport models. Spatiotemporal variability in sediment stability has been documented from several nearshore subtidal environments (Amos 1995, 1996, 1997; Maa et al. 1998) and on intertidal regions where meter-to-kilometer-scale gradients in sediment properties (grain size, organic content, chlorophyll *a*) are more pronounced (Defew et al. 2002; Murphy et al. 2008a). Properties are driven by tidal and subaerial forcings (exposure), bed-form development, drainage channel processes, and biomediation. For example, Widdows et al. (1998) and Amos et al. (1998) both report significant shore-normal gradients in τ_{ocrit} in the Humber estuary, where sediments adjacent to the marsh frontage have the greatest strength (τ_{ocrit}~0.7 Nm^{-2}) but further offshore (2,250 m) this declines (though not monotonically) to ~0.35 Nm^{-2}. Friend et al. (2003a) report a similar gradient in the Ria Formosa tidal lagoon, Portugal, which they attribute to variation in stabilization by different types of microphytobenthos. Exposed estuary mud may also display both spatial and temporal variability on a centimeter and minute scale, due to biological processes. Tolhurst et al.

(2000a, 2006b), for instance, found considerable centimeter scale spatial variation in erosion threshold due to the presence of diatom biofilms. Cyclical vertical migrations of microalgae induced by increasing light intensity during exposure can also give rise to large increases in stability in a few minutes over an exposure period (Paterson 1989, 1997; Tolhurst et al. 2003, 2006b). Diurnal studies have shown significant changes in stability and properties of sediments between day and night (Friend et al. 2003b, 2005). Drainage-related bed morphology on a smaller scale is known to dictate differences in sediment strength, and factor-of-two differences have been reported between the stability of sediment in gullies and pools and that on drier, ridge-like features (Paterson et al. 2000). High shore sediments often have a strong temporal signal in surface cohesive strength as they may be exposed for a period of some 10 h during each 12.5 h spring tide cycle and to at least 4 days of continuous air exposure during neap tides. Widdows et al. (1998) correlated sediment erosion potential in the Humber estuary directly with air exposure, and the authors have observed dehydration of intertidal sediments to beyond the shrinkage limit during summer months. Tolhurst et al. (2006a) showed that there were significant increases in erosion threshold over exposure in both the laboratory and field, but only where sediments were able to drain. These increases were "reset" by a subsequent period of immersion. This type of information is crucial in determining sediment fluxes across intertidal zones (Black 1999).

Due to the considerable logistical limitations, there remains a paucity of information about in-situ changes in erosion potential *during* the flood and ebb of the tide. Are sediments rapidly saturated, or do they retain antecedent gains/losses in cohesive strength? What are the effects of biota on erosion processes during immersion? These are important questions that require an answer if we are to improve and extend our modeling capability through entire tidal cycles.

Rainfall

Periodic rainfall is obviously a factor on exposed intertidal flats, but raindrops can also generate erosive forces on bottom sediments in up to 1-m depth of water by transferring kinetic energy to the bed (Green and Houk 1980). A number of studies on salt marsh sediments (Mwamba and Torres 2002; Torres et al. 2003, 2004; Voulgaris and Meyers 2004) show that rain can cause significant erosion and mobilization of cohesive sediments. Field observations and experiments indicate that rain is also an important variable in the erosion of unvegetated cohesive intertidal sediments (Paterson et al. 2000; Tolhurst et al. 2003, 2006c, 2008a). Generally, rainfall decreases erosion threshold, although erosion and removal of surface sediments can reveal deeper, more compacted layers resulting in a subsequent increase in erosion threshold and reduction in erosion rate (Tolhurst et al. 2008a). Rainfall appears to be particularly good at mobilizing sediments, which can then be transported by tidal/wave action. An interesting finding is that rainfall appears to negate biostabilization by benthic biofilms (Tolhurst et al. 2003, 2008a). In the Northern Hemisphere, biostabilization is reduced in winter and rainfall is more intense (Tolhurst et al. 2008a), which may contribute to patterns of increased erosion in winter. In the Southern Hemisphere, biostabilization is increased in winter and rainfall is more intense, suggesting biophysical mediation of erosion patterns may be different to those in the Northern Hemisphere.

Sediment Deposition

Recirculating benthic flumes can also be used to provide estimates of aggregate settling velocity (ω_s) following resuspension experiments, although disappointingly this aspect is ignored by many workers. Determination of ω_s is achieved by halting the flow following the maximum applied stress and measuring the temporal decrease in suspended sediment concentration. Mostly these have been still-water tests (Fig. 4), although Amos et al. (2004) decreased flow in a series of 10-min. steps with vigorous stirring between each step. Data from these and other experiments (Amos et al. 1996, 1997, 2004; Widdows et al. 1998) have yielded, not surprisingly, ω_s values very similar to those reported from within the natural bottom boundary layer of estuarine waters (typically 0.2–2.0 mm/s^{-1}; van Leussen and Cornelisse 1993). The critical deposition threshold (τ_{cd}) can also be derived by gradually slowing the flow rather than halting it, in the manner of Mehta and Partheniades (1975). Amos et al. (1998), for example, show that τ_{cd} varied between 0.03 and 0.32 Nm^{-2}, (amounting to $38 \pm 16\%$ of τ_{ocrit}) for Humber estuary mudflat sediments. In much the same way as for critical erosion threshold, Winterwerp (2006) questioned the existence of a critical threshold for deposition and secondary circulations in erosion devices (which have an upward velocity component) may confound results (Maa et al. 1993). Laboratory studies by Maa (2008), however, show that τ_{cd} does exist (at least in laboratory flumes on prepared sediment) and that convective transport in the water column has little effect on deposition.

Biological Influence

McCave (1984) brought biology to the fore of sediment transport researchers' attention when he stated, in the first sentence of his review, that "muds are a wonderful medium for life…." Natural muddy sediments do indeed support a high floral and faunal biomass, attributable mostly to enhanced organic and nutrient levels. It is not, therefore, altogether unexpected that the properties and behavior of natural mud may be linked to biological processes. A spectrum of biotic effects is superimposed on the physical and electrochemical backdrop of estuary muds, termed *biomediation*. In terms of sediment erosion, biomediation can be viewed as either positive (*biostabilization*) or negative (*biodestabilization*) (Paterson and Black 1999; Black et al. 2002). Microphytobenthos tend to be net biostabilizers and fauna net destabilizers (Tolhurst et al. 2008b;de Deckere et al. 2001).

There has been a proliferation of work investigating biological processes in flows and muddy sediments, in particular a number of large European Union projects. Bioflow (Flume Facility Co-operation Network for Biological Benthic Boundary Layer Research) was an E.U.-funded project (EVR1-CT-2001–2000) designed specifically to bring together European scientists and research institutes using field and laboratory flumes (Friend and Amos 2007). The project's findings have been published in four different special issues: "Flow measurement techniques," "Exchange processes at the sediment–water interface," "The comparison of laboratory and in situ measuring devices, and the extrapolation of flume and erosion device data to the field," and "Flume design, construction, and innovation." These papers provide an excellent starting point for researchers interested in this field (Friend and Amos 2007).

Changes in the erosion potential of natural intertidal mud may be related to shifts in the balance between key biota acting in contrasting directions—that is, there is a controlling biological overlay on the physical sediment dynamics caused by interactions between stabilizing and destabilizing organisms (Widdows et al.

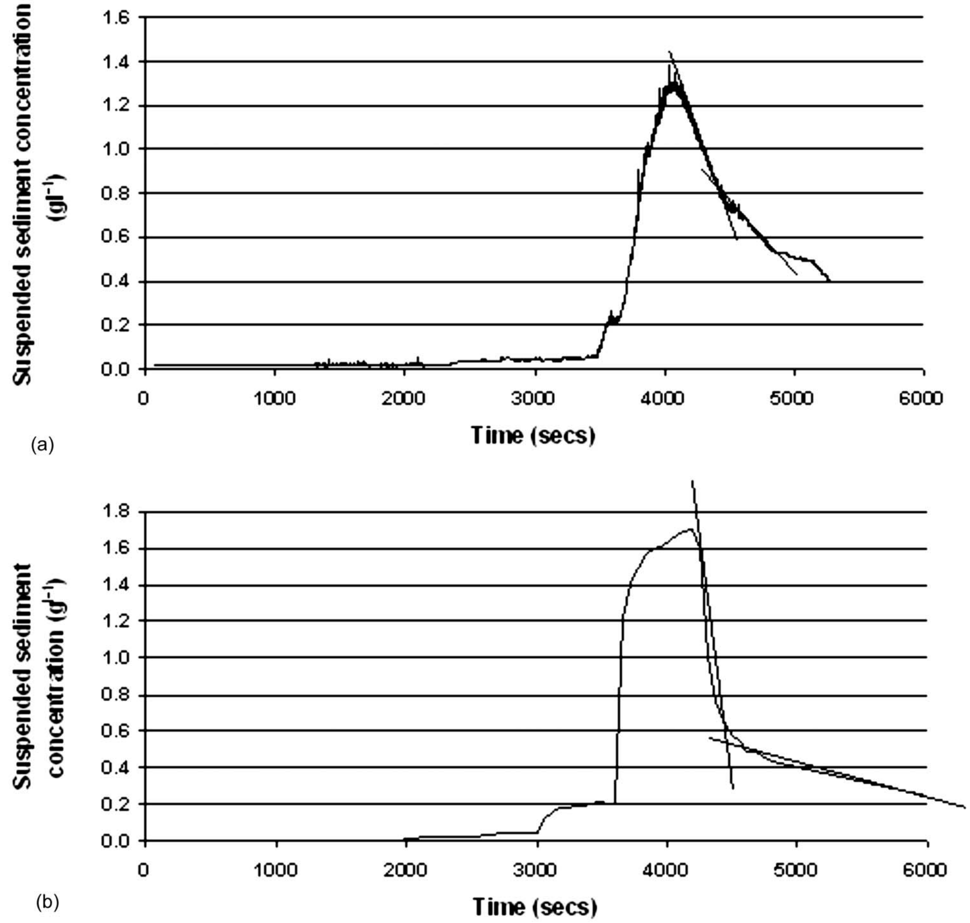


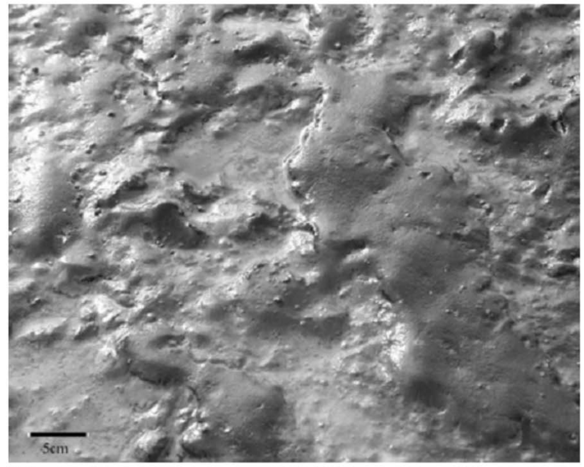
Fig. 4. Time-series of surface erosion from two contrasting sites showing posterosion sedimentation of suspended particles: (a) proposed subtidal artificial reef site in Oban Bay, western Scotland; (b) midshore intertidal site in the Eden estuary, eastern Scotland, that receives both fluvial and marine input

2000; Andersen 2001; Widdows 2001; Widdows and Brinsley 2002). Dealing with this is an ongoing challenge in cohesive sediment research. The problem with such field experimentation is that the "real" world for muddy sediments consists of an overwhelming number of interacting biotic and abiotic variables, making for a very complex system. Data tend to be either a "shotgun" scatter of data points or dominated by large-scale events and processes (such as storms or biofilm blooms), which mask key underlying processes. Friend and Amos (2007) suggest that a carefully planned combination of field monitoring with key laboratory studies would enable a more profound understanding of complex ecosystems, because laboratory experiments represent a subset of reality upon which a distinct cause-effect relationship may be tested under controlled conditions. They state that it is important that these experiments are grounded in an understanding of reality from extensive field observation/experimentation. Care is needed in choosing laboratory studies over field ones. The fact that all variables but the ones being studied are allowed to vary naturally in field studies is their major advantage. Such problems have been recognized in ecological research for a long time. Connell (1974) suggested that laboratory experiments have limitations in ecological investigations precisely because they require every aspect of the environment of an organism to be altered by being held constant and every biological interaction to be excluded. Since laboratory conditions are almost diametrically opposite to those found in natural habitats, it is sensible to assume that organisms (and hence sedimentary processes mediated by organisms) would behave and respond differently under such conditions. Thus, laboratory studies should be supported and verified by field studies.

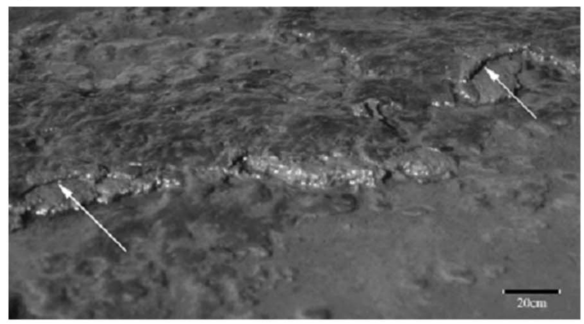
Such an approach seems to be evolving naturally, with researchers using the controlled conditions of the laboratory to assess the influence of a wide range of organisms— littorinid shells (Amos et al. 2000), bacterial extracellular polymeric substances (Tolhurst et al. 2002), algae (Romano et al. 2003), clams (Sgro et al. 2005), snails (Andersen et al. 2002; Orvain et al. 2006), diatom biofilms and cockles (Neumeier et al. 2006), and cockles (Ciutat et al. 2007; Quaresma et al. 2007). This work has provided new insights. Tolhurst et al. (2008b) showed experimentally that microphytobenthos stabilize muddy sediment while simultaneously increasing water content and reducing bulk density; but other work shows that they can also cause local destabilization under certain conditions (Sutherland et al. 1998). Fernandes et al. (2006) showed that *Nereis diversicolor* has multiple antagonistic effects on sediment stability, increasing sediment shear strength and hence τ_{ocrit}, but also increasing subsequent erosion rate. Understanding gained from the laboratory can then be applied to the field and there has been a move toward developing methodologies for manipulating the different components of biota and sediments to determine causative relationships in situ. The most obvious example of this are the various defaunation experiments (de Deckere et al. 2001), but also include using algaecide and shading by roofs to manipulate microphytobenthos (Murphy et al. 2008b).

Within the context of historical laboratory erosion experiments in which biological influence was deliberately limited or absent, biological influence causes differences in the way that laboratory and field mud erode. First, biostabilization of the upper few hundred microns of the sediment–water interface by microalgae and their exudate (extracellular polymeric substances, or EPS) is evident from erosion studies and microsectioning of mud cores (de Brouwer et al. 2002, 2005; Tolhurst et al. 2002, 2003, 2006b; Friend et al. 2003a). In this situation the actual interface mud layers are *stronger* than the underlying layers, in spite (usually) of increased amounts of water and lower bulk density. Often the underlying layers are of the same stability as uncolonized surface layers (Houwing and van Rijn 1993). Thus τ_{ocrit} is elevated and initial erosion rates are low until the biofilm is disrupted at which point underlying layers are rapidly entrained. Parchure (1980) noted this in his laboratory flume studies with cell cultures, but it is now known to be a pervasive phenomenon in the field situation. Note that increasing erosion with depth is contrary to the physically based notion of decreasing erosion with depth due to density increases. It also confounds numerical computations based on bulk density where it is assumed that τ_{ocrit} decreases with decreasing bulk density.

Second, biological communities are inherently patchily distributed within natural sediments and produce distributed biostabilization (Tolhurst et al. 2000a, 2006b; Fig. 5(a)) and distributed microtopography (Andersen et al. 2002). Our observations of the erosion process on mudflats reveal erosion to be a quasi-continuous but often localized phenomenon where surface and subsurface topographic irregularities and obstructions (such as bubbles, worm tubes, and casts) act as nuclei for erosion. Subsequent erosion often proceeds along planes of weakness relating to depositional layers and/or microbial biofilms (Fig. 5(b)). Young and Southard (1978) describe similar localized erosion in Puget Sound, and increased erosion generated by roughness elements have been reported from other marine and fluvial settings (Amos et al. 1996; Andersen et al. 2007; Bouma et al. 2007). This contrasts with observations of a more spatially uniform process reported from laboratory experiments on reformed sediments that are devoid of microtopography and distinct planes of weakness (Mehta and Partheniades 1982; Mehta et al. 1982; Parchure and Mehta 1985). Erosion of natural fine-grained substrata can thus be framed in the context of a point-source phenomenon contingent upon the activities of benthic in- and epi-fauna. This particular observation confounds the notion of sequential, layer-by-layer vertical scour by flows and thus also compromises computation of z, the depth of scour, from the suspension time-series. This is a major departure from the behavior of muddy sediments in historical laboratory studies. The related idea that it is possible to derive the vertical microscale distribution of, say, sedimentary chlorophyll *a* or organic carbon from measurements on the eroded aggregates (construction of so-called "synthetic cores") is equally void.



(a)



(b)

Fig. 5. Diatom biofilm: (a) on the intertidal of the Westerschelde estuary, showing the large amount of small scale variability in the biofilm and associated microtopography (dark raised areas are thick biofilm); (b) on the intertidal of the Ems Dollard estuary, showing erosion along the edge of the biofilm (arrows indicate where thick layers of biofilm were peeling back along subsurface planes of weakness under tidal flow)

Other Methods for Investigating Erosion Processes

New methods for investigating erosion processes are being explored. Laser and electronic holography has been used to investigate erosion and deposition in the laboratory (Black et al. 2001a; Perkins et al. 2004; Sun et al. 2004, 2005). It offers a powerful tool for visualization and quantification of erosion processes, allowing tracking of individual grains and flocs as they move from the bed into the water column. Size class of flocs can be determined for any given applied erosion stress and break-up of suspended particles can be followed.

Spectroradiometry has been successfully applied to the investigation of intertidal microphytobenthos (Murphy et al. 2005). It may represent the Holy Grail of intertidal biosedimentological erosion research, as it offers a quick, nondestructive, surrogate measure of the erosion threshold and erosion rate (Murphy et al. 2008b). Preliminary studies, using multivariate analysis of the spectra, give an R^2 of 0.75 for erosion threshold and 0.87 for erosion rate. The advantage of the technique over other surrogates for stability (such as chlorophyll *a*) is that it provides information on both biological and physical properties of the sediment in a single measurement. It can be mounted on helicopters or aircraft enabling rapid assessment of erosion characteristics of intertidal sediments over wide areas in inaccessible locations.

Future Prospects

The advent of in-situ benthic flumes has opened up a range of exciting experimental possibilities for both cohesive and noncohesive sediment transport. Investigation of the dynamic cycling of natural mud (erosion-settling-erosion) under controlled conditions (Lau et al. 2001) and assessment of the importance of biotic mediation is now feasible in situ. The next major step is to transfer

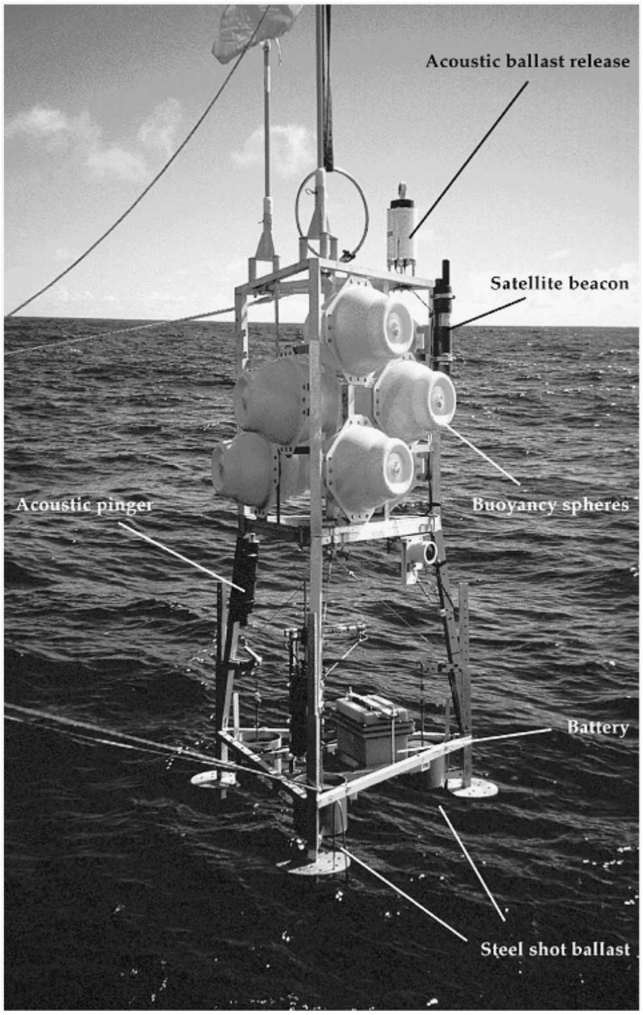


Fig. 6. Free-fall, deep-sea lander system used by Black et al. (2001b) in the northern Atlantic Ocean

benthic flume technology into deeper waters, to the continental shelf and beyond. This would broaden our knowledge to other sediment types and biogeochemical substrata. The deep ocean, for example, comprises calcareous and siliceous ooze with a rather different biological composition (Black et al. 2003). The beginnings of this are just arriving with the marriage of benthic flume technology with benthic lander technology (Tengberg et al. 1995). Currently most benthic flumes are operated in a tethered mode where there is electrical contact between the flume and the surface or shore, but this limits deployments to relatively shallow waters. Benthic landers are autonomous, instrumented oceanographic platforms that can be deployed remotely in the ocean depths (down to 6,000 m). They are equipped with ballast and buoyancy that can be remotely controlled from a surface ship to bring the lander back (Fig. 6). Contemporary benthic landers currently carry payloads to measure sediment-dissolved oxygen, total benthic respiration, and microscale metal concentration profiles (Black et al. 2001b). The benthic flume component would simply be another module capable of being attached to the lander frame. With recent developments in analytical chemistry hardware (e.g., in-situ nutrient autoanalysers and submarine mass spectrometers), coupled in-situ experiments examining the combined particulate and dissolved flux under a variety of moving flows may soon be possible.

Conclusion

Research into the erosion process of cohesive sediments began in the laboratory. The laboratory environment is particularly useful since it presents a controlled environment in which specific phenomena can be observed. Historical laboratory experiments have provided a wealth of information and a firm foundation for this area of research. However, natural muddy sediments are notoriously complex and this factor has limited the usefulness of laboratory-derived data in quantitative terms. This has lead to direct experimentation using specialized in-situ flumes. Although these devices have certain limitations, they have given rise to an entirely new data set that is inherently more useful since it represents the field situation. Much new information has been uncovered (e.g., patterns of natural erosion including bed load transport, magnitude of spatiotemporal gradients, macro- and microbiological effects, etc.). The principal elements of the erosion and deposition process have been confirmed in both qualitative and quantitative terms in the field situation. Muddy sediment erosion has been found to be even more complex than previously thought, resulting in a synergistic approach where field and laboratory studies are combined to improve understanding.

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Notation

The following symbols are used in this paper:

C_{eq} = equilibrium concentration of suspended sediment;
z = depth of scour;
ε = erosion rate;
ε_f = floc erosion rate;
ε_M = rate coefficient;
ε_p = peak erosion rate;
ρ = density;
τ_{cd} = critical deposition shear stress;
τ_s = shear strength of the sediment bed;
$\tau_{s(z)}$ = bed-shear strength at depth z;
τ_o = applied bed-shear stress;
τ_{ocrit} = critical erosion shear stress; and
ω_s = aggregate settling velocity.

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