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Critical shear stress for mass erosion of organic-rich fine sediments

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ABSTRACT

In shallow lakes of Florida laden with low-strength organic-rich sediments, wind-induced water movement is believed to actuate bed surface erosion as well as mass erosion. Experiments in hydraulic flumes to measure the critical shear stress for mass erosion tend to be lengthy and require large quantities of sediment. For bottom sediment from Lake Okeechobee at naturally occurring values of the floc volume fraction, a comparison of the viscoplastic yield stress, readily obtained from rheometry, with the mass erosion critical stress from flume tests indicates that it may be permissible to consider the yield stress as a surrogate for the critical stress. This inference appears to be supported by ancillary observations from Lake Apopka and Newnans Lake. Interestingly enough, the variation of yield stress with the floc volume fraction of the organic-rich bed is found to conform to fractal characterization commonly invoked for mineral sediment flocs, consistent with a representative constant value of 2.55 of the fractal dimension. Pending fuller investigations with a wide range of organic-rich sediments, recourse to rheometry in lieu of flume experiments holds promise as a means to simplify testing requirements for estimating the mass erosion critical stress.

1. Introduction

Surface erosion and mass erosion are the two main modes by which aquatic fine sediment beds scour when subjected to flow. Surface erosion, a gradual process with a time-scale of minutes to hours, occurs when the bed shear stress exceeds the critical shear stress τ_{cs}. Mass erosion is a catastrophic process in the sense that bed failure takes place almost instantaneously at some plane below the surface, where the induced shear stress exceeds the bed shear strength. The critical shear stress for mass erosion τ_{cm} is commonly measured in conventional hydraulic flumes or in high-velocity ducts or other devices (Lick, 2009). A limitation is that extensive test preparation, long run times and large quantities of required sediment make it cumbersome to carry out these tests. As a consequence, the viscoplastic yield stress τ_{ym} of the sediment, easily measured in a rheometer or a viscometer, has been suggested as a surrogate for τ_{cm}, albeit without systematic supportive evidence (Mehta, 1991). Inasmuch as substitution of erosion tests for τ_{cm} by rheometric tests for τ_{ym} would mean a substantial reduction in the experimental effort, there is much interest in exploring the physical underpinning of the relationship between τ_{cm} and τ_{ym}.

In order to assess if τ_{ym} can be a reasonable measure of τ_{cm}, data from preliminary field and laboratory studies of bottom sediments from three shallow lakes in Florida are used. This assessment is described here; details of the investigations have been reported elsewhere (Hwang, 1989; Jiang and Mehta, 1992; Mehta et al., 2009; Jain et al., 2005).

2. Physical framework

2.1. Stresses in bottom sediment

For the physical basis of assessment, consider the framework in Fig. 1. A viscoplastic bed with a fluid mud layer is subject to uniform flow over a mildly sloping bed of angle θ. Particle concentration is expressed by the solids volume fraction $\Phi = \rho_D/\rho_s = 1 - n$, where ρ_D is the dry bulk density, ρ_s is the material density and n is the porosity. The shear stress curve $\tau(\Phi)$ intersects the yield stress curve $\tau_y(\Phi)$, where $\Phi = \Phi_i$, and $\tau = \tau_i$. Above the level of intersection fluid mud is in motion, whereas below that level it is stationary. When τ_i increases to the shear stress τ_b at the bed, motion extends down to the bed surface. For the present analysis, we will assume

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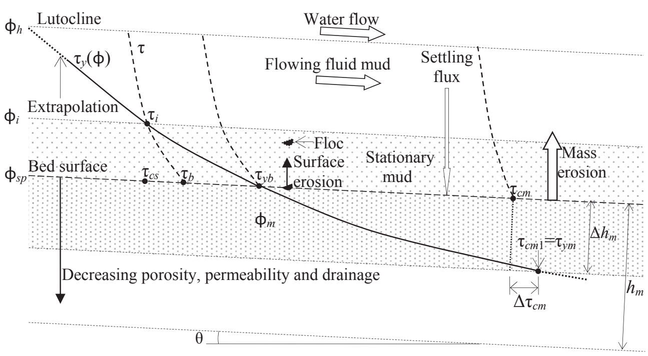

Fig. 1. Stresses in a layered bottom sediment, and mass settling and erosion fluxes.

that the effect of fluid mud density on τ_b is minor.

At the lutocline the sediment settling mass flux has its maximum value $\rho_s \Phi_h w_s$, where w_s is the settling velocity of flocs. The volume fraction Φ_h at the lutocline marks the onset of hindered settling inasmuch as w_s rapidly decreases with depth as Φ increases above Φ_h (Mehta, 1991). This suspension, or fluid mud, is sustained until Φ reaches its space-filling value Φ_{sp} marking the transition from fluid mud, in which the effective normal stress σ' is nil, to the bed in which σ' is greater than zero. Thus, below the bed surface at which $\Phi = \Phi_{sp}$, the total normal stress σ is equal to the sum of σ' and the pore water pressure σ_u.

With increasing elevation τ_y decreases together with Φ; however, as Φ approaches Φ_h, τ_y becomes too small to be detected in standard shear-rheometry (Mezger, 2006). We will conveniently set $\tau_y(\Phi_h) = 0$ (extrapolated curve in Fig. 1), and understand that the rate of increase in τ_y downward from the lutocline depends on sediment density and composition.

At the bed surface τ_y is denoted as τ_{yb}. Within the bed, consolidation due to self-weight causes both Φ and σ' to increase with depth, which in turn implies increasing sharing of the total normal stress (load) by the particle matrix and decreasing sharing by water. As a result, the rate of increase in τ_y with Φ tends to be more rapid in the bed than in fluid mud.

2.2. Critical shear stress for surface erosion τ_{cs}

The surface floc layer is arguably in the so-called drained porewater condition (Bardet, 1997). Within this layer, penetrating turbulent flow fluctuations can cause an instantaneous rise in the excess pore-water pressure. However, due to high permeability the excess pressure tends to dissipate rapidly relative to the time-scale over which the floc is detached from the bed and erodes. Under this condition, the critical shear stress τ_{cs} is on the order of cohesion c defined by the Mohr–Coulomb force balance over a soil element (Winterwerp et al., 2012).

The floc volume fraction Φ_f is equal to Φ/Φ_{sp}. When the surface floc layer erodes, overburden is removed and the newly exposed flocs swell due to elastic rebound accompanied by downward entrainment of water (Mehta, 1991). Swelling decreases Φ_f as well as σ' with the outcome that, as the bed continues to erode, a newly exposed layer nearly acquires properties of the eroded surface layer. As a consequence, τ_{cs} shows only weak dependence on Φ_f (Gowland et al., 2007).

2.3. Mass erosion shear stress τ_{cm} and yield stress τ_{ym}

Deeper in the bed, perhaps no more than a few floc diameters, due to low permeability the rate of dissipation of pore-water pressure becomes very slow and the bed remains practically undrained during its rapid shearing at the instant of failure in a rheometer or a viscometer used to measure τ_y (Mezger, 2006). When the bed shear stress τ_b increases to the mass erosion stress τ_{cm}, the stress at the plane of failure just exceeds the undrained shear strength s_u, which is nearly independent of changes in the pore pressure because the water pressure remains practically hydrostatic (Bardet, 1997). At that plane, for weak beds $\tau_y = \tau_{ym}$ is found to be a convenient proxy for s_u (Winterwerp et al., 2012).

2.4. Relationship between τ_{cm} and τ_{ym}

Referring to Fig. 1, at the depth Δh_m along the failure plane the shear stress τ_{cm1} is

$$\tau_{cm1} = \tau_{cm} + \Delta\tau_{cm} \tag{1}$$

where $\Delta\tau_{cm}$ is obtained by equating the down-slope component of the buoyant weight of the layer, a solid at the instant of failure, and the opposing drag resistance. Thus

$$\Delta\tau_{cm} = \rho_w g \phi_{sp} \Delta h_m (s-1) \phi_{f\Delta} \sin\theta \tag{2}$$

where g is the acceleration due to gravity, $s = \rho_s/\rho_w$ is the sediment specific weight, ρ_w is water density and $\Phi_{f\Delta}$ is the mean value of Φ_f over the layer Δh_m made mobile by failure. Therefore,

$$\tau_{cm1} = \tau_{cm} + \rho_w g \phi_{sp} \Delta h_m (s-1) \phi_{f\Delta} \sin\theta \tag{3}$$

We may express τ_{cm} in terms of a near-bed reference water velocity u_{bm}, i.e.

$$\tau_{cm} = C_D\left(\frac{1}{2}\rho_w u_{bm}^2\right) \tag{4}$$

in which C_D is the bed surface resistance coefficient. Substituting in Eq. (3) gives

$$\tau_{cm1} = C_D\left(\frac{1}{2}\rho_w u_{bm}^2\right) + \rho_w g \phi_{sp} \Delta h_m (s-1) \phi_{f\Delta} \sin\theta \tag{5}$$

For a given bed and flow defined by u_{bm}, by measuring Δh_m in a flume erosion test τ_{cm1} can be determined.

Now, following Mazurek et al. (2003), at depth Δh_m we may conveniently represent the yield stress τ_{ym} as

$$\tau_{ym} = K\left(\frac{1}{2}\rho_w u_{bm}^2\right) \tag{6}$$

where K is a proportionality coefficient dependent on cohesion. Thus,

$$\frac{\tau_{ym}}{\tau_{cm1}} = \frac{K\left(\frac{1}{2}\rho_w u_{bm}^2\right)}{C_D\left(\frac{1}{2}\rho_w u_{bm}^2\right) + \rho_w g \phi_{sp} \Delta h_m (s-1) \phi_{f\Delta} \sin\theta} \tag{7}$$

When the bed is horizontal in non-uniform flow, i.e. $\theta = 0$, $\Delta\tau_{cm}$ is nil, i.e. $\tau_{cm1} = \tau_{cm}$. In this case,

$$K = C_D \frac{\tau_{ym}}{\tau_{cm}} \tag{8}$$

In the event τ_{ym} can be a substitute for τ_{cm}, K would be equal to C_D. To make this assessment, laboratory experiments were carried out in a flume to evaluate τ_{cm} and τ_{ym} using sediment from Lake Okeechobee in peninsular Florida (Fig. 2).