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Understanding Fluid Mud in a Dynamic Environment

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Abstract. The generation, transportability and dewatering of fluid mud under current- and wave-induced forcing is a multi-variate problem that requires knowledge not only of mud density and composition, but also of rheology, which ultimately characterizes mud dissipative properties. It is, therefore, self-evident that the predictive accuracy of mathematical modeling, a powerful tool for simulating mud dynamics, will depend strongly on the reliability of rheological description. Defining the state of mud solely in terms of density, as done commonly, only leads to ambiguities in quantifying the rates of transport of fluid mud.

Introduction

The significance of fluidized mud relative to its transportability in coastal and estuarine sedimentation zones has been considered for decades beginning with the issue of mud underflows in quiescent waters, followed by transport due to current- and wave-induced forcing (Einstein 1941; Inglis and Allen 1957; Krone 1962; Owen 1976; Nichols 1984–1985). For operational convenience it is common to define the fluid state of mud in terms of a density range alone, while tacitly recognizing that the coupling between the state of mud and the fluid forces makes density an ambiguous descriptor of this state. In that context it is not surprising, therefore, that the lower density limit proposed for fluid mud has varied from 1.003 to 1.12 g/cm^3, and the upper one from 1.11 to 1.30 g/cm^3 (Ross and others 1987). Inherent as well to this significant variability in the density range is the additional influence of the composition of mud. In order to highlight these matters, the effect of hydrodynamic forcing on the nature of fluid mud response is here revisited; followed by reference to some tentative conclusions based on recent observations on the effects of mud composition.

Mud–Water Interaction

Because the dynamic behavior of a given sediment–water mixture is contingent upon the flow regime and the suspension concentration, the transition from particulate behavior in dilute, low concentration suspensions under turbulent flows to soil mechanical behavior in hyperconcentrated, viscous flows occurs through wide ranges of flows and concentrations (Fig. 1). In reference to these ranges, noted here only in the qualitative sense, the distinctly non-Newtonian behavior of fluid mud occurs over comparatively high concentrations under flows in which turbulence is damped to varying degrees. The mixed particulate and fluid (continuum) behavior occurs over concentrations that may be considered to be moderate, and poses challenging problems in modeling the horizontal transport of sediment over the water column (Owen 1976).

The layering of sediment–water mixture, that occurs by virtue of the phase transitions (Fig. 1) and the character of flow, identifies the fluid mud layer in terms of the water depth as shown in Fig. 2, wherein fluid mud is designated as a mobile hyperpycnal layer (Wright and others 1988); mobility here implying horizontal motion. The upper bound of this layer is characterized by the suspension-induced pycnocline or lutocline, and the strong flow shear zone that is associated with the rapid change in concentration that occurs over a comparatively short vertical distance. Shear production is often associated with interfacial instabilities and mixing that occurs in this layer, but by-and-large the lutocline itself is stabilized by the hindered nature of settling below the lutocline (Ross and Mehta 1989a). The lower bound of the hyperpycnal layer is specified by the level at which the horizontal flow velocity becomes zero; hence the mud below this layer is stationary in the horizontal sense of movement.

The difference between stationary mud and the cohesive bed below is that the latter is characterized by the occurrence of a measurable effective stress, while in the stationary mud this stress is zero; hence both the mobile and stationary hyperpycnal layers have been correctly called "fluid-supported particle assemblages" or slurries (Kirby 1986; Parker 1989). On an instantaneous basis, stationary mud that has, for example, a distinct Bingham plastic character, and one which has not dewatered sufficiently to form a fully "particle-supported assemblage," or matrix, will offer measurable

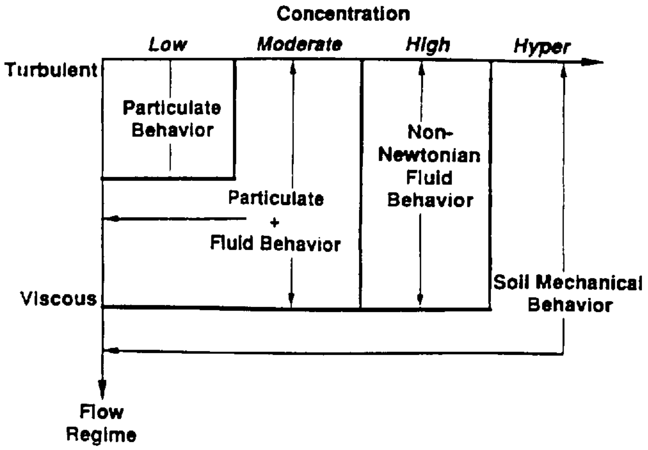

Figure 1. Qualitative plot showing transitions between particulate behavior and soil mechanical behavior associated with sediment concentration and flow regime.

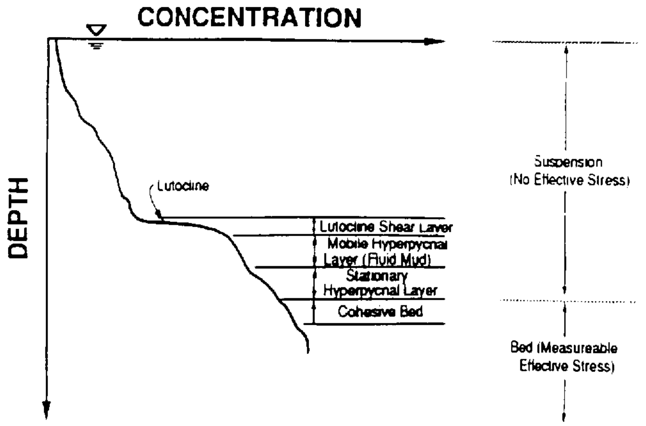

Figure 2. Horizontal layering of the vertical sediment concentration profile. The horizontal axis should be considered to represent the logarithm of concentration, implying a rapid change of concentration with depth.

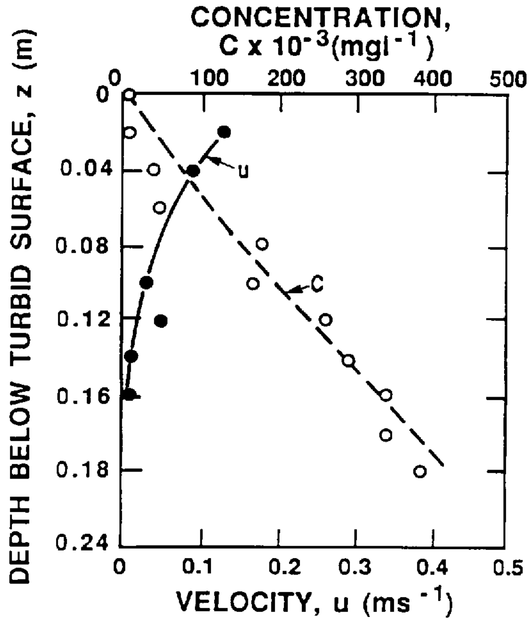

Figure 3. Concentration and velocity profiles below lutocline in the Avon River, UK (after Kendrick and Derbyshire 1985).

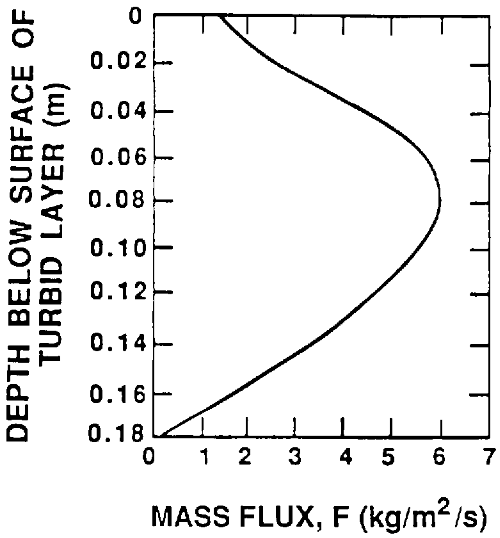

Figure 4. Horizontal sediment mass flux corresponding to Figure 3 (after Ross and Mehta 1989b).

resistance to deformation; hence it may neither be a fluid nor a cohesive bed. Because, however, it is not mobile, it does not contribute to instantaneous horizontal sediment transport.

It is evident that for calculation of the horizontal transport rate of sediment mass, the boundary between mobile and stationary hyperpycnal layers must be identified on an instantaneous basis. In that connection it is noteworthy that, at relatively low velocities at which there is little vertical mass transport due to erosion or deposition, the horizontal flux of fluid mud can be quite large. Figure 3 shows measured instantaneous concentration and velocity profiles below the lutocline (turbid layer surface) in the Avon River estuary, UK (Kendrick and Derbyshire 1985), and Figure 4 shows the corresponding sediment mass flux. The total rate of transport was 0.7 kg/m/sec per unit width, which is not negligible, particularly because at the time of measurement, about one-half hour before low water, erosion of the turbid surface or its growth by deposition from settling sediment were not significant. Ross and Mehta (1989b) simulated the velocity profile of Figure 3 using the principle of momentum diffusion into the turbid layer by an applied stress at the surface (lutocline) of this layer. It was found that a stress of only 0.02 Pa was necessary to cause the measured velocity profile within ~20 min of the application of stress. It was further found that mud viscosity and its variation with sediment concentration critically controlled the velocity magnitude.

Growth and Persistence of Fluid Mud

Because the thickness of the fluid mud layer and its relative elevation respond continuously to hydrodynamic forcing, tracking this response through time requires an understanding of the phase transitions that lead to changes in the thicknesses of the various layers mentioned. In that context, the behavior of the fluid mud is evidently easier to examine under purely depositional or purely erosional conditions, particularly the former, because in that case the upper bound is specified by the maximum rate of the settling sediment flux, as exemplified in Figure 5. In this figure, the settling flux, F_s, is plotted against suspension concentration, C, for sediment from Lake Okeechobee, Florida (Hwang 1989). The region to the right of the peak value of F_s (4.5 g/m^2/sec) is characterized by hindered settling, hence in terms of the vertical concentration profile, the value of C = 4.4 g/liter associated with the peak flux corresponds to the depth at which fluid mud is first encountered. Consequently, this depth represents the approximate level at which the lutocline occurs.

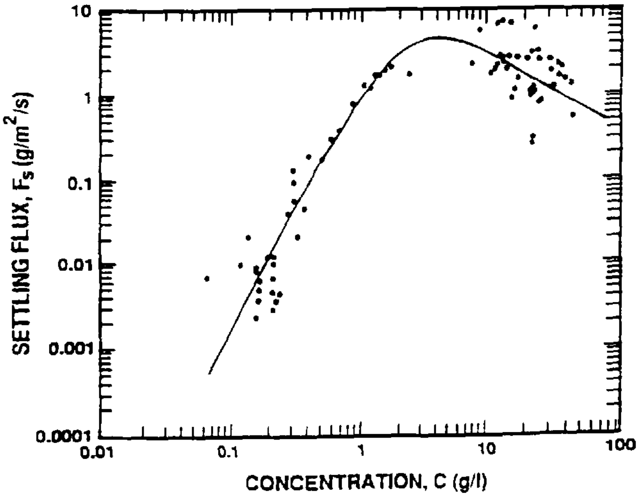

Figure 5. Settling flux as a function of sediment concentration for mud from Lake Okeechobee, Florida.

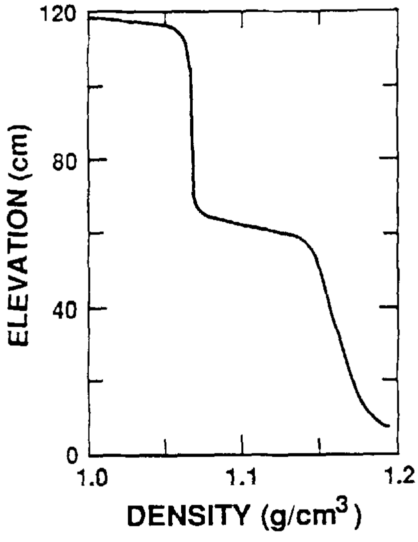

Figure 6. Instantaneous density profile during settling of a silty clay in tap water (after Sills and Elder 1986).

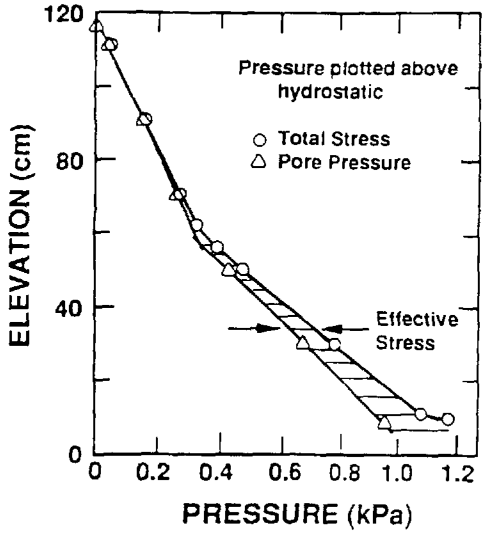

Figure 7. Total and pore water profiles corresponding to Figure 6 (after Sills and Elder 1986).

The transition from a hyperpycnal layer to a cohesive bed is revealed in settling column experiments in which the effective stress is profiled vertically at various times by measuring the corresponding profiles of total and pore pressures. An example of instantaneous profiles during settlement of estuarine silty clay is shown in Figures 6 and 7, 4.75 hr after test initiation starting with a uniformly mixed suspension having a density of 1.09 g/cm^3 (Sills and Elder 1986). The level dividing the suspension from the bed can be considered to be practically at 60 cm, which is close to, but does not uniquely correspond to, the instantaneous elevation of the lutocline. This level rose to about 80 cm at the end of the test several hours later. It is interesting to note, however, that during this settling process, no unique relationship between the onset of effective stress and mud density was found, and that the transition from suspension to structured phase occurred gradually over the density range of 1.09 to 1.13 g/cm^3.

Under wave action, for example, the reverse process of transition from a bed of fluid mud can be documented in the same way as during settlement. This process is essentially one of resuspension, because the fluidized mud becomes amenable to horizontal transport if and when a steady flow is superimposed on wave oscillation. Figure 8 shows the gradual dissipation of effective stress in an estuarine mud bed subject to wave action in a flume. The 1 Hz waves were 6 cm high in 12 cm deep water column above the bed. A relevant observation associated with this process was that while the fluid mud thickness increased with depth, no significant change in bottom mud density took place (Ross and Mehta 1989a). This in turn means that mud fluidization by wave action cannot usually be parameterized uniquely by density.

In general, when erosion and deposition occur simultaneously, comprehension of the vertical and the

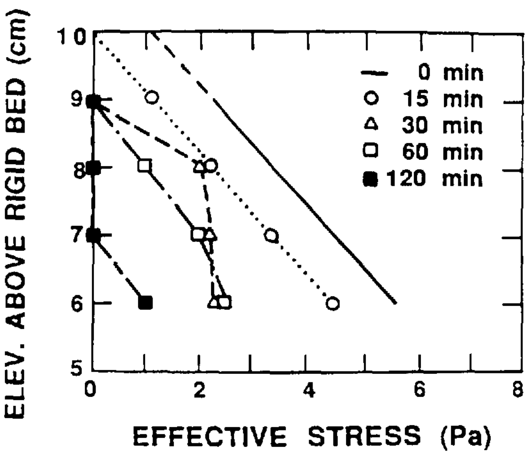

Figure 8. Dissipation of effective stress with time in an estuarine mud bed subjected to wave action in a laboratory flume (after Ross and Mehta 1989a).

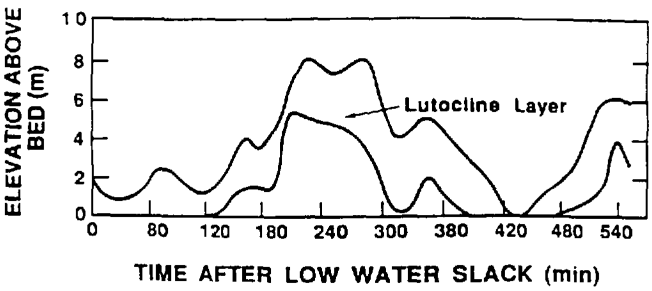

Figure 9. Measured tidal variation of the lutocline shear layer in the Severn Estuary, UK (after Kirby 1986). The mean water depth was 21 m.

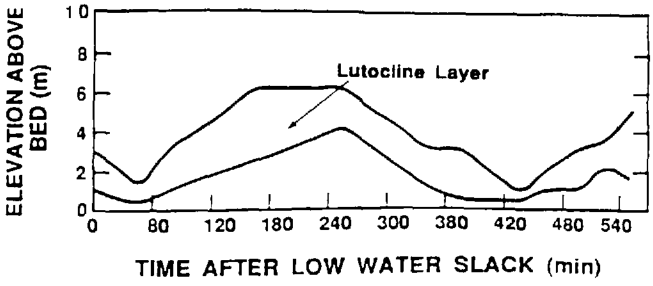

Figure 10. Numerical simulation of the layer shown in Figure 9 using data obtained at only five instances during the measurement period (after Ross and Mehta 1989a).

horizontal motions of fluid mud is significantly facilitated by mathematical modeling, even though it may be relatively simple. Figure 9 for example shows measured tidal variation of the lutocline shear layer, that is, the layer within which the suspension concentration changes rapidly with depth, in the macrotidal Severn Estuary, UK (Kirby 1986). Figure 10 shows the same layer approximately simulated by a simple one-dimensional model (Ross and Mehta 1989a). Observe that the layer rises and falls over an elevation on the order of 6–8 m but, in spite of the 7 m tidal variation that occurred during the measurement period, this layer was quite persistent owing to the strong buoyancy stabilization imparted by the relatively high density of mud.

Influence of Sediment Composition

While the effect of sediment composition on the behavior of mud is generally well known, the case of Lake Okeechobee in the south-central part of Florida is noteworthy, particularly since the mud there contains about 40 percent by weight of organic matter of peaty origin. It appears that the presence of such a large fraction of organic material leads to the occurrence of an open sedimentary structure in the top 10–20 cm layer of the bottom mud, which fluidizes quite readily and, once fluidized, does not dewater easily as a result of the strength provided by the organic matrix itself. A further consequence is that this layer possesses a uniform bulk strength that, when exceeded by the applied stress, results in structural failure and an almost instantaneous fluidization of the bed layer. Thus it was found that, for example, at a density of 1.12 g/cm^3, the bed became fluidized when the applied stress exceeded 0.75 Pa in laboratory flume experiments (Hwang 1989).

The easily fluidizable nature of mud has significant implications for the dynamics of Lake Okeechobee, which experiences seasonally high trophic levels due to episodic release of phosphorus and other nutrients associated with the bottom sediment. The diffusion of nutrients into the water column is also an important mechanism of phosphorus loading in this lake, particularly during intra-storm calms when wave action is quite mild. It was therefore a matter of interest to determine whether the top layer of bottom mud, with its very weak structure, in a practically fluidized state even at low wave-induced stresses, moves measurably and thereby conceivably increases the rate of nutrient diffusion.

An example of the surface wave energy spectrum obtained under moderate breeze conditions (20 km/hr)

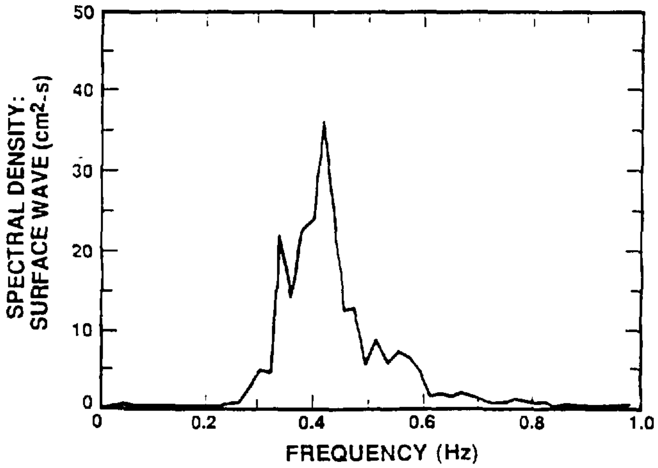

Figure 11. Measured surface wave energy spectrum in Lake Okeechobee, Florida (after Jiang and Mehta 1991).

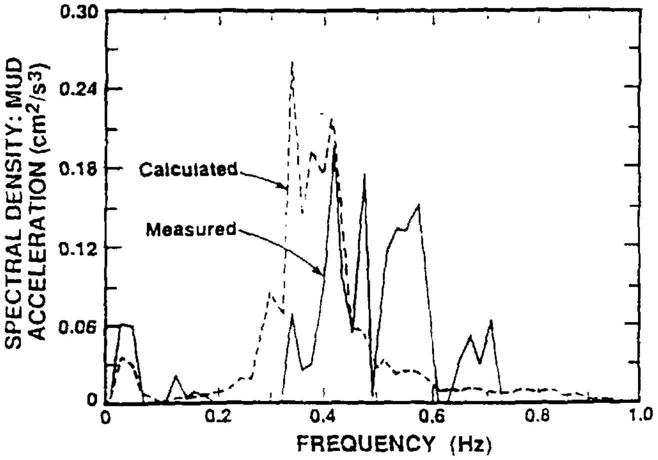

Figure 12. Measured and calculated horizontal mud acceleration spectra corresponding to the surface wave condition shown in Figure 11 (after Jiang and Mehta 1991).

in the shallow margin of the southeastern part of the lake is shown in Figure 11. The still water depth at the test site was 1.43 m. The spectrum corresponds to a significant wave height of about 8 cm at a dominant frequency of 0.42 Hz. Figure 12 shows the corresponding spectrum of the horizontal, wave-induced mud acceleration. The measurements were obtained with a biaxial accelerometer embedded 20 cm below the mud–water interface, where mud density was 1.18 g/cm^3. Comparing this spectrum with the one in Figure 11 indicates the wave-coherent nature of the mud oscillation, which amounted to a maximum horizontal displacement of about 2 mm (Jiang and Mehta 1991). The data have been compared with results from a simple, shallow water wave model that assumed the water layer to be inviscid and mud to be a viscous fluid. Agreement between the data and model calculated spectrum is only marginal, partly because the model could not fully account for the occurrence of non-shallow water frequencies above the dominant one (0.42 Hz). It is believed that an important additional cause of error in simulation may have been the overly simplified constitutive behavior of the mud as a fully fluid-supported slurry at a density as high as 1.18 g/cm^3, even under the influence of continued wave action.

Notwithstanding these sources of error, however, the data and calculation in Figure 12 demonstrate that under mild wave action, which is persistent in the lake, fluid-like bottom mud tends to undergo measurable oscillations. The impact of these oscillations requires further examination, not only in relation to nutrient exchange, but also for the generation and release of gas bubbles, which occur most commonly in this lake, thereby presumably contributing as well to lake nutrient dynamics.

Concluding Remarks

Inherent in understanding the behavior of fluid mud is knowledge not only on the density and the composition of the material, but also on the nature of granular packing, including the state of flocculation, as related to mud rheological properties. A brief review of mathematical models dealing, for example, with wave–mud interaction (Maa and Mehta 1989) reveals that modelers have chosen a range of constitutive relationships to characterize mud rheology in terms of its elastic and dissipative properties. Furthermore, model applications clearly show that the results are strongly contingent upon the selected relationship. It is concluded, therefore, that because appropriately representing rheology is demonstrably of critical importance to understanding mud behavior, efforts must be directed toward improved understanding of mud rheology and instrumentation for its characterization, in order to reliably quantify the rate of transport of fluid mud under dynamical conditions.

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