SEDIMENT TRANSPORT RESEARCH IN SHALLOW OVERLAND FLOW

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Abstract: Hydraulic induced sediment movement is a highly complex process influenced by many factors of a hydraulic and sediment property nature. This process is even more complicated in shallow overland flow, where the hydraulic regime as well as the sediment characteristics can be highly variable. Research to better understand the micro-mechanical nature of sediment movement in shallow flow has been underway for several years. Studies were conducted to examine this process both in the absence (gravity flow) and presence of water. Experiments consisted of measurements of particle velocity and particle concentration on the mode of transport and the corresponding associated transport rates. Results show that the sediment movement is not a random phenomenon but occurs in a highly organized manner ranging from saltation of sediment particles at very low concentrations to movement in a sediment wave like pattern. These waves may transgress into meanders depending on the channel bed conditions and the grain addition rate into the stream. As a consequence, the sediment transport capacity of shallow flow is severely impacted. This paper discusses the experimental findings from a steady state flow regime to which sediment was added at a controlled rate at the upstream of a 7 m long and 10 cm wide channel of about 1° slope steepness. The analytical interpretations are based on a two-phase flow model involving the St. Venant equations of shallow water flow and granular flow.

INTRODUCTION

At the Seventh Federal Interagency Sedimentation Conference results of a study were reported (Pal et al., 2001) that concerned the development of organizational structures in granular gravity flow, when sand grains and glass beads were dropped at a constant rate into an inclined plane of 30° to 38°. That paper focused on possible similarities and dissimilarities between the origins of structured flow of granular material in a gravity flow field in air or water. The study was motivated by the observations that grains, dropped into a constant flow regime, changed their mode of transport when the concentration increased and exceeded the transportation capacity. This mode consisted of the development of clusters, which subsequently grew larger to become domains, and finally formed grain waves (Fig. 1). Similar experiments were conducted with different granular materials: glass beads between 200-250 μm, and two classes of sand of 200-250 and 300-350 μm. In all cases, sediment waves developed and a critical concentration threshold value was reached.

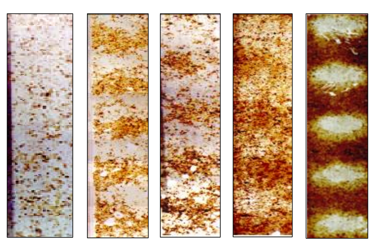

Figure 1 Granular organization with increasing sediment seeding rates.

The conventional view is that detachment and subsequent transport of sediment particles in overland flow exclusively depends on the sheer stress generated by the velocity profile. Incipient motion requires a minimum of critical shear stress at the bed (Foster and Meyer, 1975). Others (Rose et al., 1983) prefer to use the stream power concept and its critical value in describing sediment detachment and subsequent transport. It is also routinely assumed that variations in the free surface profiles in channel flow are the cause of the evolution of bed features. However, no adequate explanation has been given for the regularity in length and time scales of these bed features. In effect, the issue whether the surface waves in shallow flow are the sole driving mechanism for the organized mode of sediment transport is very much an open question. This issue motivated granular gravity flow experiments without the presence of water.

GRAVITY DRIVEN GRANULAR FLOW

Our granular flow studies (Prasad et al., 2000; Pal et al., 1999) have indicated that the particles themselves show under certain conditions, a strong tendency of a high degree of organization in terms of identifiable waves with distinct density characteristics. These waves can move upslope or downslide depending on the prevailing conditions of slope, particle size, etc. In these experiments several modes of sediment movement were noted. They have been schematically illustrated in Fig. 2. They are: (1) Uniform flow, in which the granular material moves in a near uniform concentration, downslide. This flow does not show any evidence of differences in the density of the granular material. This is the prevalent condition in low concentration regimes where the dominant movement is by saltation. (2) A mid-inertial flow regime, where flow exhibits zones of higher densities or waves in which the individual particle velocity has a smaller velocity than that in the rarefied zone between waves. These waves have a higher volumetric solid fraction though the flow depth does not change in the longitudinal flow direction. (3) A fully-inertial flow regime, in which the waves move faster than the individual particles. The waves are zones with a higher density and the free surface varies substantial in the flow direction.

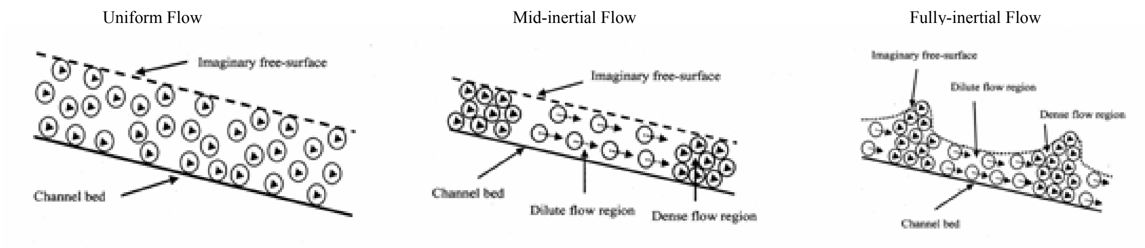

Figure 2 Shallow granular flow mode.

In light of these findings of a micro-mechanical nature of sediment particle interactions and energy flow changes in gravity flow it was postulated that the sediment dynamics in channel with fluid flow must be governed, at least to some degree, by similar kinetic processes, though the presence of water may have a strong moderating influence on the organizational nature of the particulate matter. Open channel flow exhibits a wide variety of velocity details with turbulent scales in the range of a fraction of a mm to several cms. It is assumed that flow with supercritical conditions with waves has sufficiently small turbulent scales. Thus the transport of sediments with diameter larger than 200 µm will tend to preserve the mode and mechanisms observed in gravity dominated granular flows. To address these issues, detailed laboratory tests were conducted in which sediment was added at a known rate at an upstream point into a constant flow regime and its movement was followed.

Table 1 Properties of Materials and Flow Characteristics.

Material	Diameter (d_s)	Density (ρ_s)	Packing Factor	Flow Rate	Froude Number
	µm	kg m^{-3}	(-)	l min^{-1}	(-)
Coarse Sand	1000-1400	2.52	0.64	21.6; 15.7	1.92 & 1.45
Medium Sand	600-850	2.67	0.61	21.6; 15.7	1.92 & 1.45
Glass Beads	600-1000	1.52	0.68	21.6; 15.7	1.92 & 1.45

SEDIMENT TRANSPORT IN SHALLOW FLOW

Briefly, the experiments were conducted in a 7 m x 10.7 cm x 4.4 cm deep rectangular open aluminum channel with an inclination < 1^o. A known, but controlled rate of water entered at the upstream end of the channel and also sediment particles of a desired size range, were seeded to the flow at a constant rate at the upstream end of the channel. Sediment movement was followed by a set of Fotonic probes located about 4 m from the upstream end. Both particle velocity and the solid concentration at the point of observation were determined. Two flow rates and three particle size ranges were studied (Table 1). Details of the velocity and solid concentration measurement have been given by Suryadevara et al. (2004) and Prasad et al. (2004). A schematic of the experimental set up is given in Figure 3.

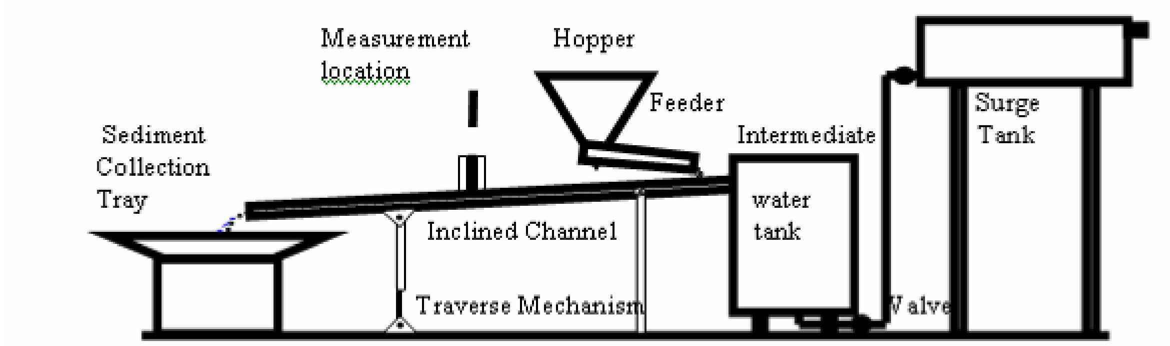

Figure 3 Experimental setup of sediment transport in shallow flow.

Experiments show that at low feed rates, the sediment transport rate measured at the downstream end of the channel equals the addition rate thus suggesting that the transport-capacity of the flow has not yet been reached. Also, visual observations indicate that the mode of sediment transport is by saltation. While the grain addition rate is gradually increased until a critical value beyond which solid grains started clustering. These clusters gradually organized by a natural selection process into particle wave structures of distantly spaced stripes. While the stripes are gradually spreading in the entire channel the measured transport rates started declining thus exhibiting an asymptotic (Fig. 4) behavior. Further increases in the addition rate while the particles are in stripe mode do not significantly change the

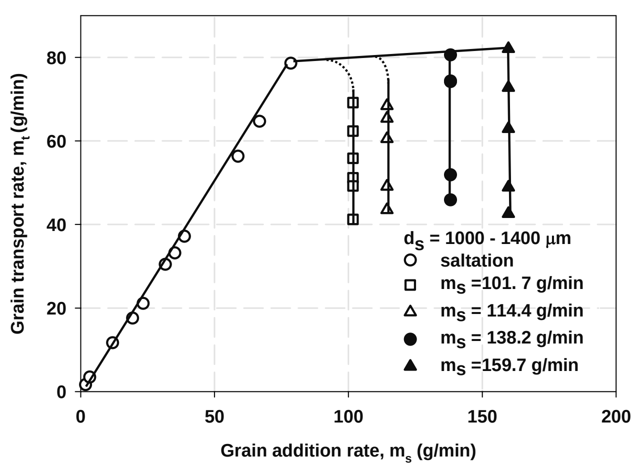

Figure 4 Sediment transport rates in relation to sediment addition rates. Coarse sand (d_s = 1000-1400 μm) water flow rate q_l = 15.7 liters/min (Fr_l = 1.45).

transport rate. The added material is stored in the channel bed and the flow regime in terms of the free surface boundary is materially impacted. Continuing the experiment for longer duration results in a wave pocketed bed into a meandering bed (Fig. 4). For a flow regime of 15.7 ℓ/min with Froude number 1.45 and feed rates beyond 101.7 g/min of coarse sand these changes takes place. Though the bed patterns seem to be similar with different grain addition rates the time scales in developing the structures seem to vary largely with the grain addition rates. As the grains organized into meanders the associated transport mass rate seem to be reduced largely. Meander formation was observed with four values of grain addition rates, 101.7, 114.4, 138.2 and 159.7 g/min and the corresponding transport rates were denoted by four fork shaped trends (Fig. 4). Similar observations were made with other flow rates (Fr_l = 1.92) and particle sizes.

In reality, when the sediment transport is forming into the wave propagation mode by meanders the transport rate associated with a given grain addition rate varies with time. Such a transient nature of the transport rate for a grain addition value of 143.6 g/min for coarse sand is observed in Figure 5. Visual observations indicated that until 6min stripe packets are prevalent and until 14min both stripe and meanders existed together and the later periods the prevalent sediment waves are only meanders. In all these modes the minima indicated by solitary grain transport by saltation, sliding or creeping when the tail of the wave packet is barely touching the tip of the transport end. Maxima occur when the wave packets are in to the measurement sample.

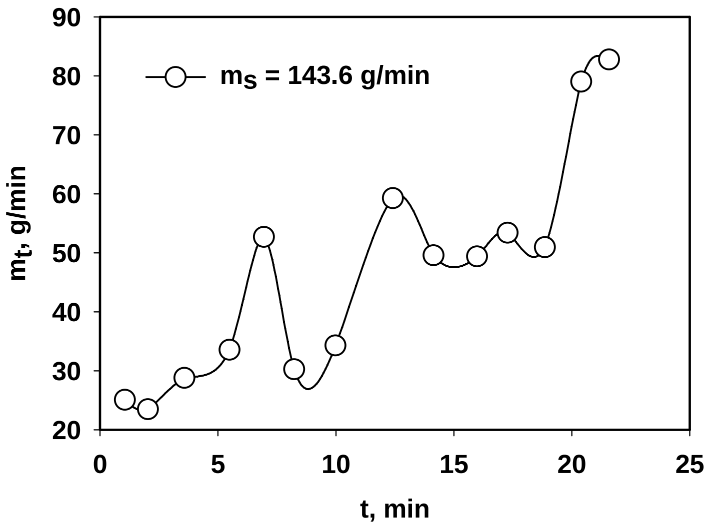

Figure 5 Transient variation of measured transport rates (t = 0 corresponds to addition of grains into the water stream). Solids addition rate is kept constant, m_s = 143.6 g/min. Coarse sand (d_s = 1000 - 1400 μm) and Froude number, Fr_l = 1.45.

In our experiments we have captured the meander evolution and its growth by placing camera at a fixed location. Various stages of a meander formation for coarse sand are shown in Figure. 6. When grains are introduced into the

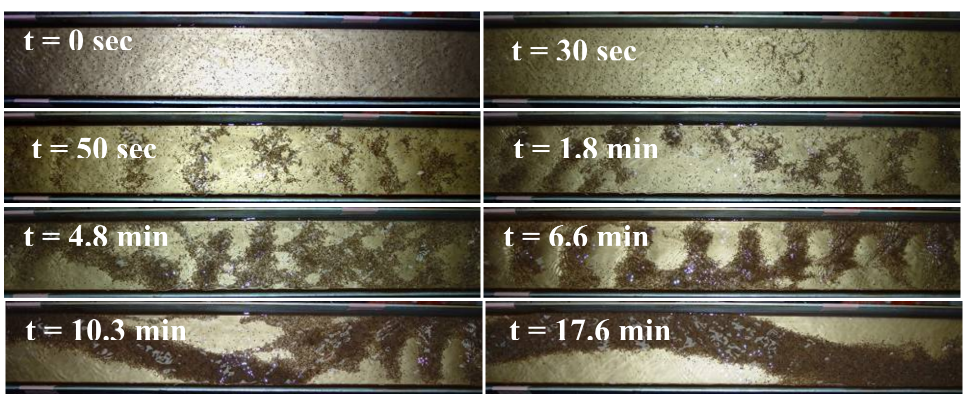

Figure 6. Meander formation in an open channel transport of water-sand system (all the figures correspond to a fixed camera location). Particle size, d_s = 1000 – 1400 μm; water flow rate = 21.6 l/min and the solids addition rate, m_s = 170.3 g/min.

clear water stream at 170.3 g/min, initially saltating grains quickly starts clustering in less than a half minute and in less than a minute they organize into distinctly spaced stripes. Further evolution of the channel bed is noticed by widening and multiplication of these strips with more grain joining into them. This strip wave propagation continues for about 18-20 min, at which a clear meander is formed. A clear picture on transport rates in these modes is till lacking and needs further investigation.

From these observations, it was concluded, that particle interactions during transport have a major impact on the mode and velocity and thus on the transport capacity of the flow regime. Therefore, information about the solid concentration and the velocity of the individual particles in the flow field is essential in developing an understanding of the observed phenomena. To that end, two closely spaced optical probes aligned in the direction of flow with an 8 mm diameter sensor hooked up to a signal analyzer were placed in the flow 4.3 m from the upstream end of the channel. This arrangement allowed for concentration and velocity measurements. The measurements indicated that at very low concentration, the velocity of the saltating particle consistently increased with increasing particle concentrations reaching fairly quickly a maximum value. Thereafter, a rapid decrease in the particle velocity was noted with further increases in the particle concentrations (Fig. 7). The initial increase is attributed to a redistribution of the streamline pattern in the neighborhood of sediment particles. However, it is well known that the boundary layer is distorted due to the sediment near the bed. The subsequent decrease is the result of kinetic energy loss through collisions of individual particles. Thus increasing the sediment addition rate leads to more frequent collisions and reduced particle velocities. Just, as we have seen in the case of gravity flow, a "pile-up" occurs at the upstream end of the wave packet while at the downstream end, particles are swept up again by the flow, gain momentum until the next series collisions lead to a new wave packet development. It was also observed that in the flow, the sediment particles were mainly concentrated in a shallow layer near the channel bottom of which the thickness appears to be decreasing with increasing concentrations. This phenomenon was also observed in the gravity flow experiments. The remainder of the shallow flow layer was particle free.

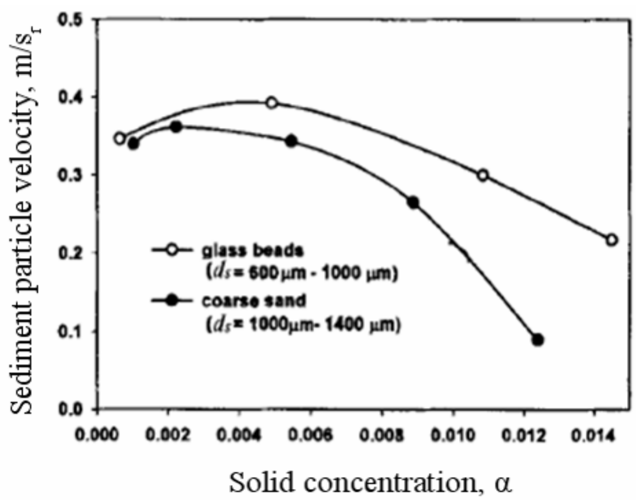

Figure 7 Observed relationship between sediment particle velocity and solid concentration for different glass beads and coarse sand and a flow rate of Fr_l = 1.92.

The observation of two zones of flow, the bottom zone with sediment and the upper one without sediment was the impetus of formulating a mathematical model, in which to each zone the mass and momentum balance equation were applied (Fig. 8). The momentum equation for the sediment some includes components for pressure, gravitation, and the dispersive stress between particles and the water phase, while the momentum equation for the sediment free zone includes the gravity and flow resistance effect. The details of this treatise has been given by Prasad et al. (Submitted). Briefly, the relationships for the sediment zone are:

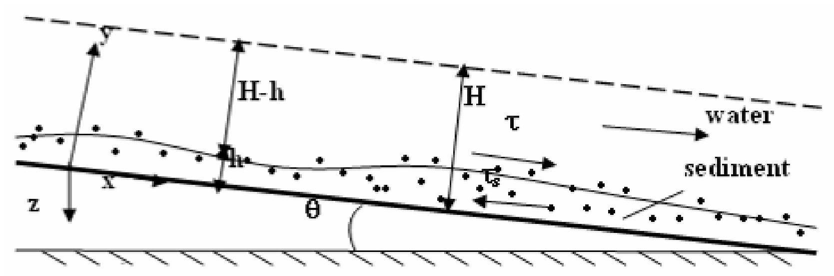

Figure 8 Sediment transport in water over an inclined channel. τ_s - unit width dispersive stress in the sediment ,τ - unit width tractive hydro-dynamic stress on the sediment, h - saltation height, H - water depth (flow), θ - bed slope and x,y,z are coordinates.

$$\frac{\partial}{\partial \mathbf{t}}(\rho_{\mathbf{m}}\mathbf{h}) + \frac{\partial}{\partial \mathbf{x}}(\rho_{\mathbf{m}}\mathbf{hu}) = \mathbf{0} \tag{1}$$

$$\frac{\partial}{\partial \mathbf{t}}(\rho_{\mathbf{m}}\mathbf{hu}) + \frac{\partial}{\partial \mathbf{x}}(\rho_{\mathbf{m}}\mathbf{hu}^2) = -\mathbf{h}\frac{\partial \mathbf{p}}{\partial \mathbf{x}} + \rho_{\mathbf{m}}\mathbf{g}\,\mathbf{h}\sin\theta - \mathbf{h}(\tau_{\mathbf{s}} - \tau) \tag{2}$$

and for the sediment free zone are:

$$\frac{\partial \eta}{\partial \mathbf{t}} + \mathbf{U}\frac{\partial \eta}{\partial \mathbf{x}} + \eta\frac{\partial \mathbf{U}}{\partial \mathbf{x}} = \mathbf{0} \tag{3}$$

$$\frac{\partial \mathbf{U}}{\partial \mathbf{t}} + \mathbf{U}\frac{\partial \mathbf{U}}{\partial \mathbf{X}} + \mathbf{g}\frac{\partial \eta}{\partial \mathbf{x}} - \mathbf{g}\sin\theta = -\mathbf{T} \tag{4}$$

where ρ_m is the effective continuum density of the solid, h is the thickness of the sediment zone or saltation height, u is the grain velocity, p is the pressure on the sediment particle and consists of the hydrodynamic pressure p_h of the overland flow plus the dispersive pressure p_d of the sediment, θ is the channel slope, T is the effective flow resistance, U is the depth-average flow velocity, τ is the shear stress, and $\eta = H - h$ with H being the flow depth.

The dispersive pressure p_d is ascribed to dilatational effects while the hydrodynamic pressure is readily accounted for by the relationship:

$$\mathbf{p}_{\mathbf{h}} = \rho_{\mathbf{w}}\,\mathbf{g}(\mathbf{H} - \mathbf{y}) \tag{5}$$

Where ρ_w is the water density. The dispersive stress is quantified based on Bagnold's work (1954), where it is assumed that for small shear rates, the case behaves like a Newtonian fluid, in which the normal (P_d) and tangential (τ_s) stresses are linearly proportional to the fluid dynamic viscosity μ, the shear rate γ and the volumetric solid factor α_v. Here the stresses varied with the solid concentration as $\alpha^{3/2}$, where α is the linear concentration which is defined as the ratio of the grain diameter to the mean radial separation distance.

From the above information, the concentration profile α(X) can be determined in terms of its spatial derivative (Prasad et al., 2005):

$$3\rho_o \frac{(u-c)^2}{\alpha^2}\left(1+\frac{1}{\alpha}\right)^{-4}\frac{d\alpha}{dX} - \frac{5h}{\rho_s d_s}\mu\gamma\alpha^{1/2}\frac{d\alpha}{dX} = \frac{\rho_w hg}{\rho_s d_s}\frac{d\eta}{dX} - \rho_o g\left(1+\frac{1}{\alpha}\right)^{-3}\sin\theta + \frac{h}{\rho_s d_s}\left[2.25\mu\gamma\alpha^{3/2} - \tau\right] \quad (6)$$

where the moving coordinate X = (x-ct), c is the velocity of the solid density wave.

In evaluating this model, the shear rate γ must be known and this can be derived from particle velocity gradients. Given the difficulty of obtaining this information, an alternative method was used to obtain an analytical expression for the solid fraction based on the hydrodynamic stress with the drag coefficient and slip velocity being the difference between free water velocity of the sediment free zone and the water velocity moving through the solid particles in the saltation layer. The slip velocity is a measure of the mean fluid thrust on the sediment layer on the saltation layer (Eames *et al*., 2004). The hydrodynamic stress between fluid and solids was derived as:

$$\tau = \frac{24\rho_w \nu C_o U\alpha^3}{hd_s} \quad (7)$$

and the particle velocity u relationship in terms of α is given by:

$$u = 10.7 C_m C_o U\alpha^{3/2}\frac{h}{d_s} \quad (8)$$

Where C_o is the maximum possible volumetric concentration (~0.78 for spheres) and C_m is the added mass coefficient due to the particles matrix (Eames *et al*., 2004).

Figure 9 shows the calculated relationship between the particle velocity and the solid concentrations for different ratios of saltation height over particle diameter and flow velocities for low solid concentrations.
The transition from saltation to a strip mode transport capacity is obtained for the condition $d\alpha/dx = 0$ in Eq. (6).

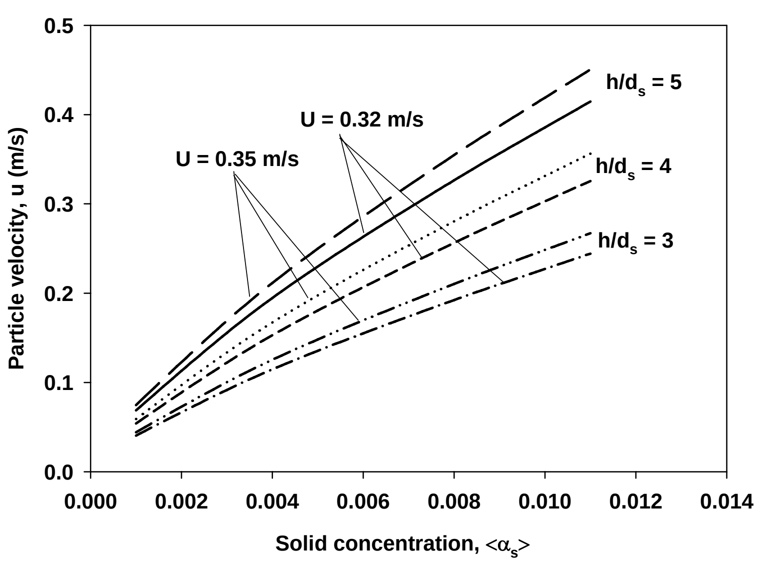

Figure 9 Particle velocity predictions relationships as a function of the solid concentration for h/d_s ratios.

This leads to the following relationship for the solid concentration.

$$\frac{\alpha^6}{\left(1+\alpha^2\right)} = \frac{3 q_m^{\ 2}}{48\left(\frac{h}{d_s}\right) C_0^{\ 2} \rho_s \rho_w \nu U} \qquad (9)$$

SUMMARY

Sediment movement in a laboratory scale demonstrated several features of the overland shallow flow in a controlled manner. Existence of density waves has a tremendous impact on the transport rates of sediment in the channel. The measured transport rates with meander formation are transient in nature and a detailed investigation is required. A small amount of solids in water are initially uniformly dispersed and their movement is by saltation. However a gradual increase in the grain addition to the water stream caused the development of a large dispersive stress component due to intense grain collisions up to a critical value beyond which the flow developed into organized mode. A mathematical model was developed and formulated based on observations of sediment movement in shallow flow which provide a better understanding of the mechanism of sediment transport modes from saltation to wave packets.

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