SEDIMENT TRANSPORT RESEARCH IN SHALLOW OVERLAND FLOW – A PHENOMENOLOGICAL DESCRIPTION

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ABSTRACT

Shallow overland flow on agricultural land has distinctly different sediment transport characteristics than flow in large channels and streams. The high flow velocities on steep land coupled with often large variations in sediment concentrations stand in contrast with the sub-critical flow regimes and relatively low sediment concentrations in stable stream systems. Yet, erosion prediction models for upland areas use the same sediment transport relationships based on critical shear stress or stream power concepts that are used in large stream flows. There is increasing evidence that greater consideration must be given to the micro-mechanic nature of sediment movement in shallow overland flow that substantially affects the sediment transport capacity and bedform development. This article discusses the results of a highly controlled sediment transport study that shows the sediment transport capacity limiting effect due to a high degree of sediment particle interaction. The key parameters measured were particle velocity and the sediment concentration. The study was conducted in a steady state flow regime to which sediment was added at a controlled rate at the upstream of a 7 m long and 10 m wide channel of about 1^{o} slope steepness. As the sediment addition rate is increased, sediment movement by saltation gives way to an organized structure consisting of a strip that transitions into a meandering bedform. The analysis is based on a two-phase flow model involving the St. Venant equations of shallow water flow and granular flow.

1. 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^{o} to 38^{o}. 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

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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.

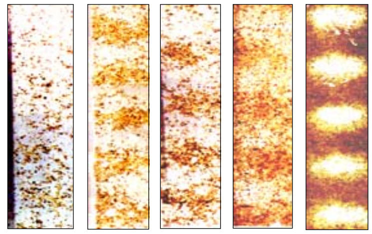

Figure 1. Granular organization with increasing sediment seeding rates.

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.

2. GRANULAR GRAVITY 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 downslope 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 downslope. This flow does

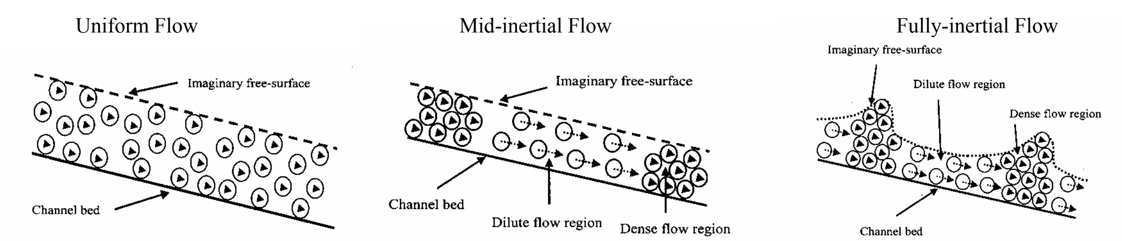

Figure 2. Shallow granular flow mode.

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.

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 the sediment movement was followed.

Table 1. Properties of Materials and Flow Characteristics.

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

3. SHALLOW FLOW STUDIES

Briefly, the experiments were concluded 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 Fig. 3.

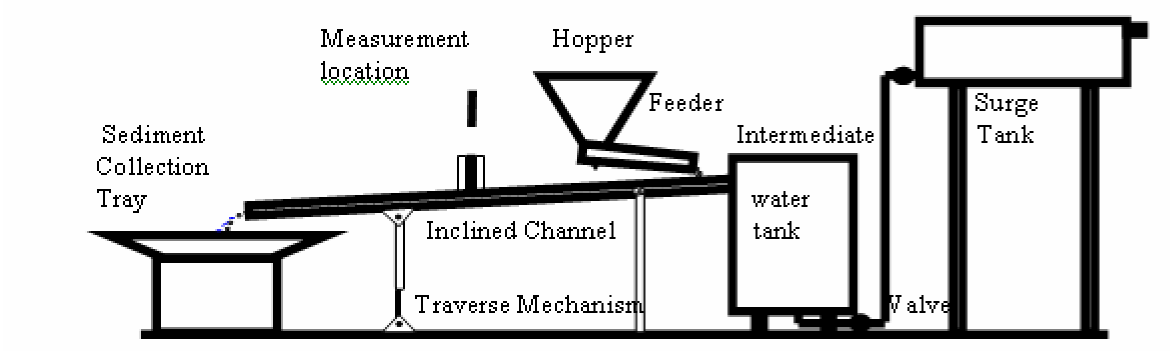

Figure 3. Experimental setup of sediment transport in shallow flow.

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. Depending on the flow rate and particle size, the sediment transport rate at a given addition rate critical transitions into a highly organized mode of transport due to the increased frequency of particle collision in the flow in which kinetic energy is lost and agglomeration of particles into clusters, followed by domains and clouds and eventually wave packets, takes place.

When this occurs the transport rate seems to stabilize. Further increases in the addition rate do not significantly change the 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. Above a certain addition rate, the sediment in the channel transitions from a wave packed to a meandering bed (Fig. 4). For a flow regime of 21.6 ℓ/min with Froude number 1.92, a feed rate of 170.3 g/min of coarse sand this change takes place in approximately 20 min. Fig. 5 shows the observed relationship of particle addition rate

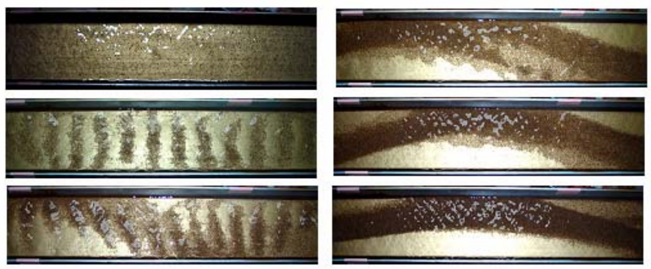

Figure 4. Successive stages of sediment movement in shallow flow in a steady state flow regime and a constant sediment addition rate.

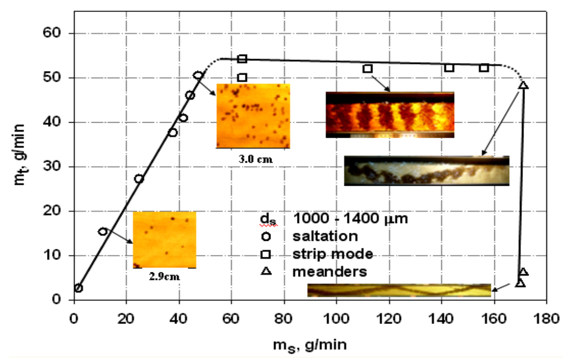

Figure 5. Sediment transport rate in relation to sediment addition rate.

and sediment transport rate. Similar observations were made with other flow rates (Fr. = 1.45) and particle sizes.

From these observances, 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 were noted with further increases in the particle concentrations (Fig. 6). 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.

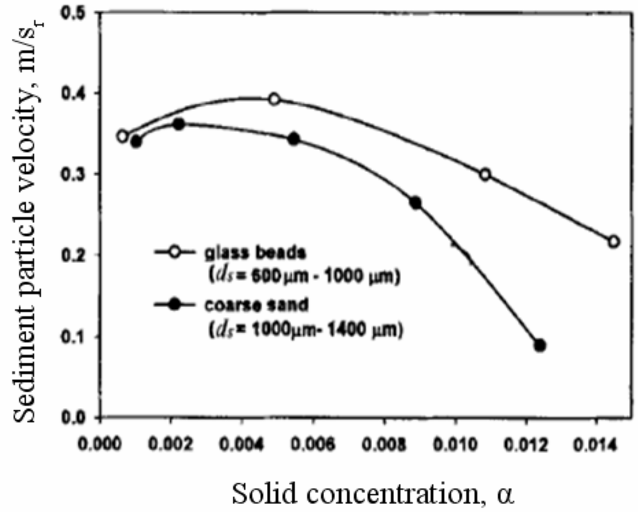

Figure 6. Observed relationship between sediment particle velocity and solid concentration for different glass beads and coarse sand and a flow rate of Fr = 1.92.

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. 7). The momentum equation for the

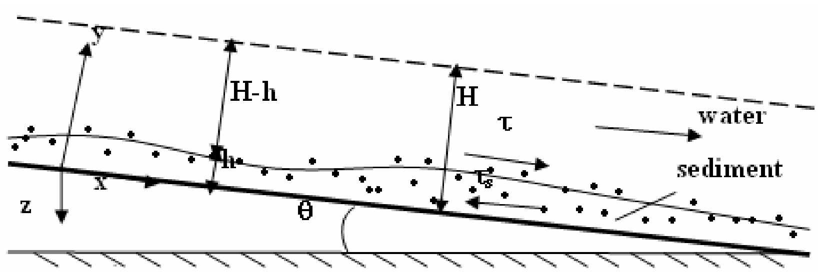

Figure 7. Sediment transport in water over an inclined channel. τ_s - unit width dispersive stress in the sediment ,τ - unit width tractive hydro-dynamic stress on the sediment, h - saltation height, H - water depth (flow), θ - bed slope and x,y,z are coordinates

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:

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

$$\frac{\partial}{\partial t}(\rho_m hu) + \frac{\partial}{\partial x}\left(\rho_m hu^2\right) = \frac{\partial p}{\eta \partial x} + \rho_m g\, h \sin\theta - h\left(\tau_s - \tau\right) \tag{2}$$

and for the sediment free zone are:

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

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

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

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 pressure on the solid particles, consists of two components the hydrodynamic pressure p_h of the overland flow, and the dispersive pressure p_d through dilatational effects while the hydrodynamic pressure is readily accounted for by the relationship:

$$p_h = \rho_w\ g(H - y) \tag{5}$$

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, this information the concentration profile $\alpha(X)$ can be determined in terms of its spatial derivative (Prasad et al; 2005):

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

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 slope 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. The hydrodynamic stress between fluid and solids was derived as:

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

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

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

Fig. 8 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.

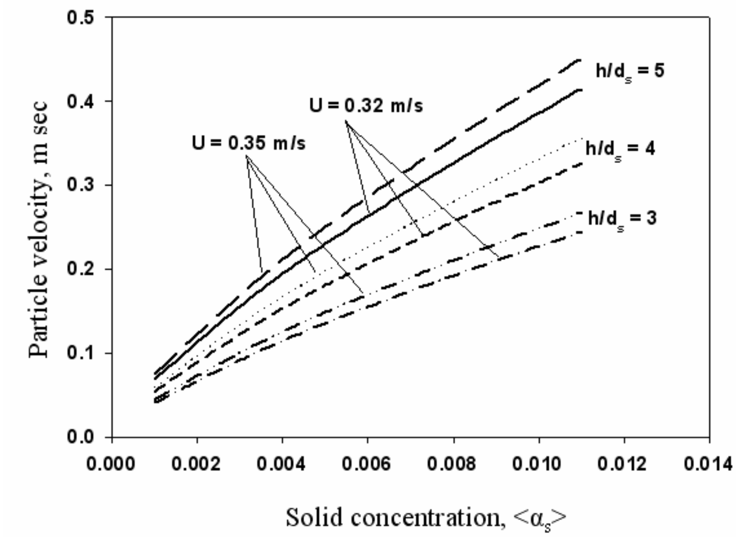

Figure 8. Particle velocity predictions relationships as a function of the solid concentration for h/d_s ratios.

The transition from saltation to a slip mode transport capacity is obtained for the condition $d\alpha/dx = 0$ in Eq. (6). This lead, to the following relationship between the solid concentration:

$$\frac{\alpha^6}{\left(1+\alpha^2\right)} = \frac{3q_m^2}{48\left(\dfrac{h}{d_s}\right)c_r^2 o\rho_s\rho_u\,\nu U} \tag{9}$$

SUMMARY

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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