WATER RESOURCES RESEARCH, VOL. 33, NO. 4, PAGES 865–876, APRIL 1997

Hydraulics and erosion in eroding rills

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Abstract. Rills often act as sediment sources and the dominant sediment and water transport mechanism for hillslopes. Six experiments were conducted on two soils and a uniform sand using three experimental methodologies. The results of this study challenge the assumption often used in hydrologic and erosion models that relationships derived for sheet flow or larger channel flow are applicable to actively eroding rills. Velocity did not vary with slope, and Reynolds number was not a consistent predictor of hydraulic friction. This result was due to interactions of slope gradient, flow rate, erosion, and the formation of rill roughness, bed structures, and head cuts. A relationship for rill flow velocities was proposed. Stream power was found to be a consistent and appropriate predictor for unit sediment load for the entire data set, while other hydraulic variables were not. The data for stream power and sediment load fit the form of a logistic curve ($r^2 = 0.93$), which is promising relative to recently proposed erosion models which are based on probabilistic particle threshold theory.

Introduction

Where water erosion rates on disturbed upland areas are greatest, rills are active. Rills are small, ephemeral concentrated flow paths which function as both sediment source and sediment delivery systems for erosion on hillslopes. Flow depths in rills are typically of the order of a centimeter or less, and slopes may be quite steep. These conditions constitute a very different hydraulic environment than typically found in channels of streams and rivers. Rills actively erode and thus evolve morphologically over short timescales. For a particular rill morphology at a given time, there are associated roughness features, widths, and depths which are functions of the eroding material, runoff rate, and prior rill structure. On the other hand, as erosion creates the morphology, the rill structure influences also the erosion. Head cuts and sidewall sloughing are sources of sediment, and roughness created by erosion has a direct impact on dissipation of flow energy. During runoff recession, plunge pools become sediment traps and hydraulic roughness decreases. If the runoff were to again increase, those sediment traps become sediment sources. The morphological evolution of rills and the erosion rate in rills both act upon and depend upon each other.

Various studies have investigated hydraulic roughness of shallow overland flows. *Gilley et al.* [1990] measured in field studies the velocities of flow in rills for 10 soils. For each soil, slope was constant, and flow rates were varied. For the entire data set, slope ranged from 4 to 10% and the flow Reynolds number ($Nr = uR/\nu$) ranged from 300 to 10,000. The Darcy-Weisbach friction factor ($f = 8gRS/v^2$) was found to be the negative exponential function of Nr for each soil. Thus, for constant slopes, *Gilley et al.* [1990] found that roughness decreased in the rill as flow rate increased. This suggests that the physical roughness had a lesser influence on the hydraulic resistance as flow depth increased.

The decrease in hydraulic roughness as a function of Nr is well documented [*Chow*, 1959, pp. 10–11]. *Savat* [1980] used sand and loess to investigate hydraulic resistance due to grain friction in a laboratory flume. His results indicate a progressive decrease in hydraulic grain roughness with increased Nr, with a change in the relationship between laminar and turbulent flow regimes. The relationship between f and Nr is made more complicated when depths of flow are of the same order of size as roughness elements. The recent consensus in the literature [*Abrahams et al.*, 1986; *Abrahams and Parsons*, 1994; *Gilley et al.*, 1992; *Prosser and Dietrich*, 1995] appears to indicate that when the depth of water is less than the size of the physical roughness elements, f increases as Nr increases. This process is explained by the fact that as the water depth increases, the roughness elements are progressively inundated, and the wetted area and associated friction are increased. Once the flow depth exceeds the height of the physical roughness, f decreases with increased Nr as is the case for smooth beds and those with grain friction only.

The effect of slope gradient on flow velocities and hydraulic roughness in rills is subject to debate. *Rauws* [1988] showed that for a nonerodible bed with artificial roughness elements glued to a flat bed, the relationship between Nr and f varied as a function of slope. *Rauws'* [1988] data showed that velocity increased with slope on the rough surface but to a somewhat lesser degree than for smooth surfaces [*Govers*, 1992]. *Foster et al.* [1984] and *Abrahams et al.* [1996] used nonerodibile, stabilized rill beds to show that flow velocity increased as a function of slope gradient. However, *Govers* [1992] showed that for an eroding rill, velocity was not influenced by bed slope. The principal differing factor in that experiment was the use of an eroding surface rather than a noneroding surface. *Govers* [1992] suggested that the increase of erosion on the steeper slopes effects an associated increase of bed roughness, thereby slowing the flow velocity.

Various hydraulically based variables and equations have

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Paper number 97WR00013.
0043-1397/97/97WR-00013$09.00

been proposed and used as predictors for soil erosion in rills and by overland flow in general. Most of these variables and equations have been borrowed from research on streams and rivers. Some of the hydraulic variables commonly considered for rill erosion are total discharge [*Meyer et al.*, 1975], bed shear stress [*Foster*, 1982; *Nearing et al.*, 1989], stream power [*Rose et al.*, 1983a, b; *Hairsine and Rose*, 1992a, b], and unit stream power [*Moore and Burch*, 1986]. As is the case discussed above for rill hydraulics, most of the testing for these models has been conducted on data sets with constant slopes for each experimental treatment. The study of *Meyer et al.* [1975], for example, was performed on a 6% slope. *Laflen et al.* [1991] conducted rill erosion experiments on 36 disturbed soils in order to obtain erodibility parameter values for the rill erosion shear stress model (Water Erosion Prediction Project (WEPP)) [*Nearing et al.*, 1989]. The slope used for each soil was a constant, and, in general, the results were consistent for the shear-based erosion model. However, a thorough evaluation of the data by *Elliot* [1988] showed that stream power (termed "potential energy" by *Elliot* [1988, p. 9]) actually was a better independent variable for the data than was hydraulic shear stress.

Govers [1990] conducted a series of experiments in a laboratory flume on the transport of five well-sorted quartz materials in the silt and sand size classes at slopes ranging from 1.7 to 21%. The relationship between sediment transport rate and shear stress was reasonable, though different, for all five material size classes. Effective stream power, which is stream power modified by a critical stream power and hydraulic radius, correlated well with transport rate for the larger three material size classes only. Unit stream power provided a reasonable fit to the data for the three smallest size classes but not the two largest classes. Govers presented a general equation of the form:

$$Tc = Aq^B S^C, \tag{1}$$

where Tc (kg m^{-1} s^{-1}) is transport capacity, q (m^2 s^{-1}) is unit discharge, S (m m^{-1}) is slope, and coefficients A, B, and C vary with the five different grain sizes and between laminar and turbulent flow regimes.

The influence of hydraulic roughness on erosion by surface water runoff is a much debated and much utilized concept. *Einstein and Banks* [1950] and *Einstein and Barbarossa* [1951] presented the idea that hydraulic resistance, as well as hydraulic shear stress, could be divided into that which occurs because of grain friction and that which occurs because of form roughness. *Foster* [1982] suggested that this concept could be applied to the effects of cover in direct contact with the soil surface on rill erosion of cropped land. *Abrahams and Parsons* [1991, 1994] discussed the implications and limitations of the shear-partitioning concept for rough desert pavements. *Foster* [1982] hypothesized that the portion of the shear stress which is available for detaching soil and transporting sediment is that acting on the soil and sediment, τ_s (Pa), which is given by

$$\tau_s = \tau_{tot}(f_s/f_{tot}) = \rho_w g S R(f_s/f_{tot}), \tag{2}$$

where τ_{tot} (Pa) is the total shear acting on the surface, R (m) is hydraulic radius, ρ_w (kg m^{-3}) is water density, and f_s and f_{tot} are Darcy-Weisbach friction factors for the bare soil and composite surface, respectively. *Nearing et al.* [1989] applied this model formulation to the rainfall simulator data of *Dedecek* [1984] with reasonable success to account for the effects of surface litter on erosion. It is important to note here that *Nearing et al.* [1989] considered the f_s term to be friction for a bare soil surface in a rill which included soil form roughness rather than grain roughness alone. The total friction was composed of the bare soil friction and friction due to surface litter and crop residue. Also, the rill erodibility parameters, that is, the coefficients within the shear stress equations used for soil detachment, were calibrated on the data for experiments on bare soil, which is not free of form roughness due to head cuts and irregular soil bed surfaces. Thus the model and calibration process were mutually consistent.

Govers and Rauws [1986] investigated the use of both grain shear and unit stream power for transporting capacity on irregular beds and found both to be good predictors. They recommended the use of unit stream power based on the practical grounds that this flow characteristic is calculated from simple measurements. This is an important consideration, particularly since the errors associated with typical field and laboratory measurements of average flow velocities, wetted perimeters, and hydraulic radii are significant. *Govers and Rauws*' [1986] recommendation for using unit stream power was accepted and used in the European Soil Erosion Model (EuroSEM) model [*Morgan et al.*, 1992; *Morgan*, 1995].

The effect of cover on sediment yield may be more complicated than is described by the shear-partitioning model or unit stream power. *Bunte and Poesen* [1993] demonstrated that local turbulence near rocks can cause local scouring. For their data, when the rock cover was less than 20%, sediment yield actually increased as a function of the rock amount. It is certainly possible that other types of roughness, such as residue cover, might have a similar effect. Head cuts offer another example that illustrates the complexity of the rill morphology. Head cutting has been discussed as a source of sediment [*Nearing et al.*, 1990], and *Elliot and Laflen* [1993] included head cutting as a source term in their rill erosion model. However, while retreating head cuts are certainly sediment sources, plunge pools below head cuts act as drop structures which dissipate flow energy. The relative magnitudes of the two effects have not been measured or analyzed.

The objectives of this study were to investigate rill hydraulics and erosion as a function of slope and discharge rate. Relationships between flow velocity and hydraulic roughness on the one hand and other rill characteristics such as slope, unit discharge, and Reynolds number on the other are reported. The study compares and contrasts the use of various hydraulic variables, such as bed shear stress, grain shear, stream power, effective stream power, and unit stream power as dependent variables relative to sediment production from rills. Also discussed is the validity of the concept of shear partitioning based on relative differences between grain roughness and form roughness for erosion of bare soil in rills. The results of the study are used to show the effect of sample length on sediment production from the rill. Six series of experiments were conducted using two soils and a uniform sand and four different experimental arrangements. The Reynolds numbers covered an interval from 4 to 70,000, and Froude numbers covered an interval from 0.2 to 700. Shear stress, stream power, and unit stream power ranged over 4 to 5 log cycles in value. The results of the study have important implications relative to flow velocities, hydraulic roughness relationships, and sediment production in rills. Because of the wide range of hydraulic conditions represented in the data we see that some of the previously held concepts and relationships for rills, which are based on concepts borrowed from studies on rivers or overland sheet flow, might not best represent rills which are actively eroding. The results thus have implications for both surface water runoff and sediment routing models where rilling is present.

Experiments

Six series of experiments were conducted: laboratory rill and "miniflume" studies on two agricultural soils, a field rilling study on one of the soils, and a flume study on a uniform sand. The soils used were a Miami silt loam (fine–silty, mixed, mesic, Typic Hapludalf) from Indiana and a Cecil sandy loam (clayey, kaolinitic, thermic, Typic Hapludult) from Georgia. The Miami, which contained 84 g kg^{-1} sand, 760 g kg^{-1} silt, and 156 g kg^{-1} clay, was taken from a cornfield at the Purdue University Throckmorton Agricultural Center. The Cecil, which contained 656 g kg^{-1} sand, 196 g kg^{-1} silt, and 148 g kg^{-1} clay, was taken from a pasture. The sand was sorted quartz, with an median diameter of approximately 0.30 mm washed free of silt and clay.

The laboratory rill studies were conducted in a 3 m flume. The soil was packed loosely in the flume over a sand tension table bed with a "V"-shaped surface, the depth of the V being approximately 0.5 cm and the width being 15 cm. The dry bulk densities of the soils were 1.10 and 1.40 Mg/m^3 for the Miami and Cecil, respectively. The soils were prewetted from the bottom through the tension table for a day and then brought to 10 cm tension for a day. Drainage from the tension table was then clamped off for the experiment, except for the end drainage hole which was allowed to drain freely to prevent reemergent flows at the bottom end of the soil bed during the test. The flume bed was then raised to the appropriate slope. Slopes used were 5, 10, 15, and 20%. Water was added to the top of the flume at increasing levels, with discharge, sediment, velocity, and flow width measurements taken in triplicate for each inflow level. Nominal inflow rates were 2.2, 4.4, 8.8, 13.2, and 17.6 L min^{-1}. Mean flow velocities were calculated by measuring the velocity of the leading edge of a fluorescent dye and multiplying by a correction factor [*King and Norton*, 1992; *Gilley et al.*, 1990]. Rill widths were measured with a ruler during the experiments.

The field studies on the Cecil soil were performed on a hillslope greater than 20%. Rill plots were laid out partially across slope to obtain the desired slope for each plot. Slopes used were 3, 6, 10, 15, and 20%. Rill lengths were 3.4 m. The soil was completely disturbed to a seedbed-like condition using a rototiller at field moisture. Inflow rates were the same as those used in the laboratory study. Rill widths were measured with a ruler during the experiments. Velocity measurements were made using a $CaCl_2$ salt-tracing technique [*King and Norton*, 1992].

Laboratory studies were also performed on the two soils using a miniflume technique [*Shainberg et al.*, 1994]. The miniflume was 50 cm long with a V-shaped inlet and outlet. The flume is packed with soil, also in a V shape. Side angles on the V are 45°. The preparation of the soils in the miniflume was similar to that in the larger 3 m laboratory rill flumes. Slopes used were 5, 10, and 20%. Nominal inflow rates varied depending upon the slope and ranged from 0.1 to 0.9 L min^{-1}. Velocity was measured using the $CaCl_2$ salt-tracing technique [*King and Norton*, 1992].

The sand bed experiments were also performed in the 3 m laboratory flume. Treatments for the experiment included bed slope, which ranged from 0.5 to 35%, and sample length, which ranged from 30 to 240 cm. In the case of the sand experiment, the flow rate for each slope treatment was adjusted such that flow depth was held to a constant 1 cm. In order to create the appropriate upstream roughness conditions, flat blocks of wood with sand glued to their surfaces were placed in the flume from the upper end of the sand sample to the water inlet area at the upper end of the flume. The length of the wood block section was varied in order to provide the desired sample length; for example, a 60 cm length of wood block was placed in the 3 m flume to obtain the 240 cm sample length.

Thus, in all of the experiments, measured parameters were total mass discharge rate, sediment discharge rate, flow width, slope, sample length, and flow velocity. It was not necessary to measure velocity in the sand experiment since the depth was held constant. Velocity in this case was calculated from the measurements of flow rate and known cross-sectional flow area.

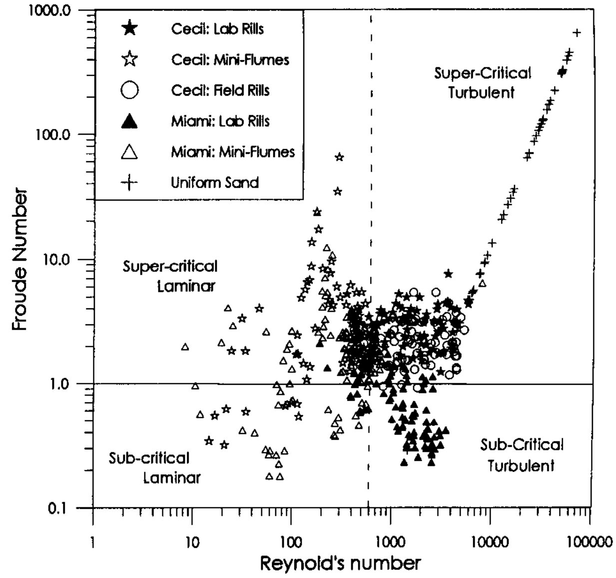

Figure 1. Hydraulic flow regimes for the experimental data.

Hydraulics

Flow Regime

The data represented all four conditions of subcritical laminar, supercritical laminar, subcritical turbulent, and supercritical turbulent flow regimes (Figure 1). The placement of the break in Figure 1 between turbulent and laminar flow is purely arbitrary. *Bunte and Poesen* [1993] aptly show that on surfaces with roughness elements, localized turbulence is present and active in causing scour even when *Nr* is of the order of 250. Figure 1 indicates only the computed flow regimes for average flow conditions. Within a given rill the spatial variability of *Nr* and Froude number *F* will be great. The Miami laboratory rills, for example, experienced significant headcutting, and at the locations of the overfall of a head cut the flow would, of course, not be subcritical, even though the average hydraulic parameters for those data are largely in the subcritical range. Given the importance of local turbulence and localized supercritical flows associated with the rilling process, the interpretation of average flow information such as that shown in Figure 1 is

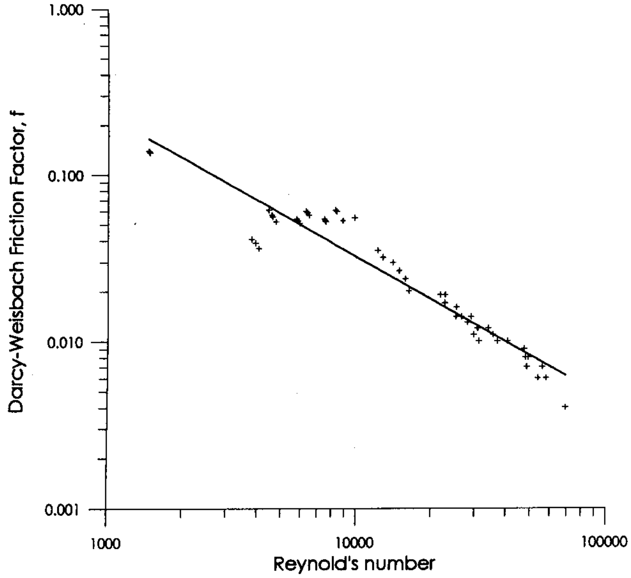

Figure 2. Darcy-Weisbach friction factor f as a function of Reynolds number Nr for the data from the uniform sand experiment.

questionable. It does, however, show, in general, the wide range of flow conditions represented by the data set.

The average measured widths of flows in the rills were 1.1, 6.2, and 10.3 cm for the miniflume, laboratory rill, and field rill experiments, respectively. Flow width for the sand experiment was a constant 20 cm.

Hydraulic Friction

Relationships between the Darcy-Weisbach hydraulic friction factor $f(8gRS/v^2)$ and Reynolds number $Nr(uR/\nu)$ were neither straightforward nor consistent for these data as a whole. For the sand data, however, the relationship was clear: f decreased as Nr increased with increasing slope and constant hydraulic radius (Figure 2). This result is consistent with other experiments where only grain friction is present on otherwise flat beds [*Savat*, 1980] in that log (f) decreased linearly with log (Nr). However, the slope of the relationship between log (f) and log (Nr) was −0.25 as reported by *Savat* [1980] and was −1.10 for the data presented here.

The Cecil soil is sandy and exhibits moderate levels of bed roughness. The overall friction factors are, in general, greater than those for the uniform sand and less than those for the Miami. Figures 3 and 4 show the results of hydraulic friction measurements for the Cecil soil in the laboratory and field rill experiments, respectively. For the case of the laboratory rill studies on the Cecil, the data separated into three statistically distinct groups based on their slope: 5%, 10%, and 15–20% slope. Statistical grouping was based on a general linear test approach for the equivalence of regression lines ($\alpha = 0.05$). The field data separated into two groups: the 3–6% group and the 10–20% group. For each of these five groups there was a consistent trend of decreasing f with increasing Nr, though the fit became poorer at the steeper slopes. This result is consistent with that of *Gilley et al.* [1992]. For both the laboratory and field data there was also a trend of increasing f with increasing slope. The overall results suggest that Nr is not a consistent and unique predictor of hydraulic roughness in rills.

The results for the Miami soil laboratory rill experiment were quite different than for the Cecil soil. For constant slopes and increasing flow rate (and thus Nr), friction increased (Figure 5). For the Miami soil, as was the case for the Cecil, f increased with slope.

Figures 2 through 5 represent a progression from smooth to rough bed conditions. The uniform sand bed was a completely inundated flat bed where grain friction was the dominating roughness. The Miami soil is loess derived, and it forms deep rills with vertical sides and relatively rough beds. Head cuts and plunge pools are important influences on hydraulics in the rills formed in the Miami, whereas the Cecil has relatively less

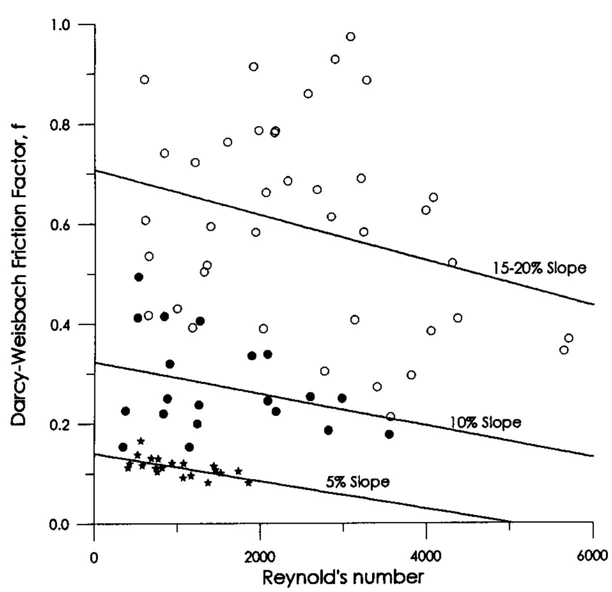

Figure 3. Darcy-Weisbach friction factor f as a function of Reynolds number Nr for the data from the Cecil soil laboratory rill experiments. The data were separated into statistically homogenous slope classifications.

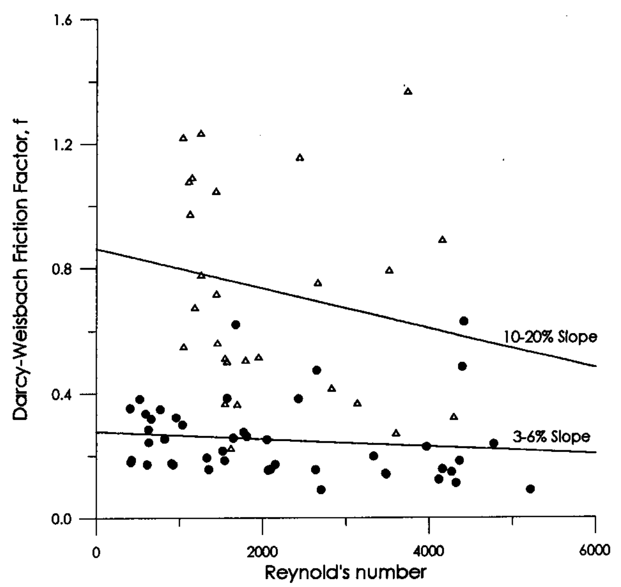

Figure 4. Darcy-Weisbach friction factor f as a function of Reynolds number Nr for the data from the Cecil soil field rill experiments. The data were separated into statistically homogenous slope classifications.

cohesion and does not tend to form the strong physical form roughness found in the Miami soil. The rills' sides were less deep and steep in the Cecil soil. Head cuts were present but with lower overfalls than for the Miami. The Cecil field rills exhibited a slightly rougher bed and higher head cuts than were the case for the Cecil laboratory rills, which is possibly a result of a greater level of disturbance of the laboratory soil in shipment and preparation. Nonetheless, the two Cecil rill experiments resulted in similar trends in the friction data relative to Nr, flow rate, and slope.

Friction factors for the miniflume experiments are shown in Figure 6. Below $Nr = 300$ the friction factors generally decreased with increasing Nr. It appears from the graph that a change in hydraulic conditions in the miniflumes occurs in the range of Nr between 200 and 300. We hypothesize that this region represents a transition from laminar to turbulent flow wherein the turbulence is initiated as a feedback from changes in the bed morphology such as development of head cuts and plunge pools. For the case of immobile rough beds a similar type of transition occurs, albeit at a greater Reynolds number [*Chow*, 1959, pp. 10–11]. If this hypothesis is correct, it indicates an important interaction and feedback between erosion and hydraulics which is not present for the case of the immobile bed. For Nr greater than 300, f levels off to within a range of 0.2–1.0, which creates a smooth transition in the relationship from the miniflumes to the laboratory rill and field rill data.

These data would indicate that Nr is not an appropriate predictor of f for actively eroding rills. This is because form roughness of the rill, including apparent roughness associated with head cuts, is a function of the erosion process, which is itself a function of flow rate and slope. This result is in agreement with that reported previously by *Govers* [1992]. There appear to be two simultaneous offsetting processes taking place with regard to flow velocity in the rills for the turbulent eroding case. Increased flow rate and slope would tend toward greater velocities if the bed shape were not changing. However, increasing flow rate and bed slope also produce a rougher bed

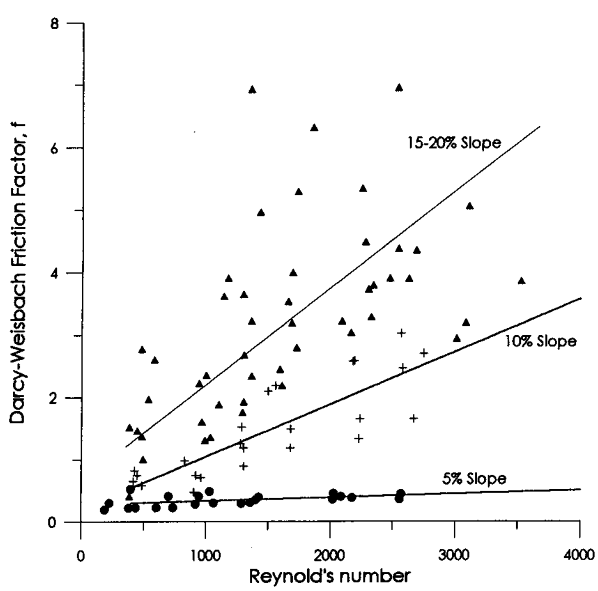

Figure 5. Darcy-Weisbach friction factor f as a function of Reynolds number Nr for the data from the Miami soil laboratory rill experiments. The data were separated into statistically homogenous slope classifications.

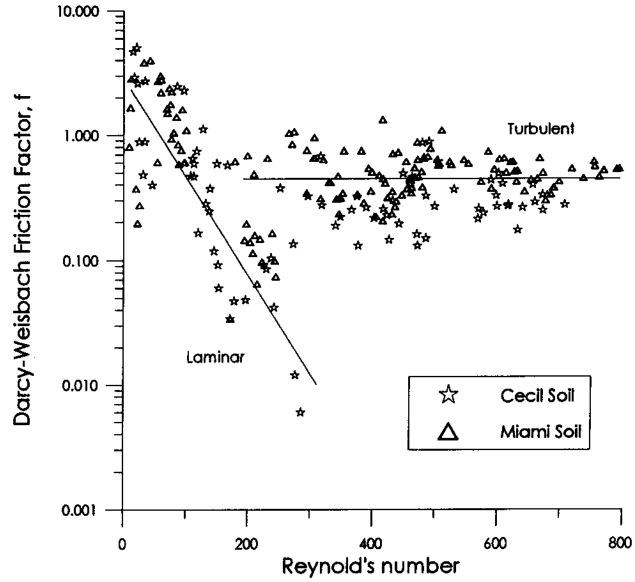

Figure 6. Darcy-Weisbach friction factor f as a function of Reynolds number Nr for the data from the Cecil and Miami soils miniflume experiments.

surface which tends to slow the flow velocity. For the case of the Cecil soil, increased flow rate at constant slope resulted in generally lower f values, while increasing slope created enough surface roughness to net a greater f value. For the Miami soil the results indicated that the f increased with both flow rate and slope.

Implicit in the statement that Nr is not a good predictor of hydraulic roughness in eroding rills is that the use of the Chezy and Manning relationships in rills for velocity and discharge modeling is complicated. Calculation of the friction factor cannot be made for a given soil and hillslope unless a great deal of information is known for the soil (in a condition comparable to field conditions) over the range of appropriate slope gradients. In other words, one would need relationships similar to those shown in Figures 3 through 6 to use the Chezy equation or similar information for n if using the Manning equation.

The Chezy and Manning equations are commonly used to relate average flow velocity U to slope and hydraulic radius of the flow. However, for these data there was no apparent relationship between U and slope for either the Cecil or Miami rills (Figures 7 and 8). Visual observations of the rill development during the experiments indicated a significant increase in both the number and the overfall height of headcuts as rill steepness increased for both soils. The head cutting and associated increase in apparent roughness of the bed at steeper slopes may account for the lack of effect of slope gradient on U in these rills. In other words, it is hypothesized that the erosion process in the rill, which is undoubtedly influenced by slope steepness at the outset of the rill formation, quickly creates bed features such as head cuts which act to decrease U and reduce the effect of the slope on U after the initial stage of the rill cutting.

Because of the limited utility of Nr as a predictor of f and the difficulty in applying the Chezy and Manning velocity relationships, the data from this study were used to evaluate alternative velocity relationships for eroding rills. Flow velocity was best correlated to the unit discharge of water in the rill (Figure 9), though the relationship varied depending on the

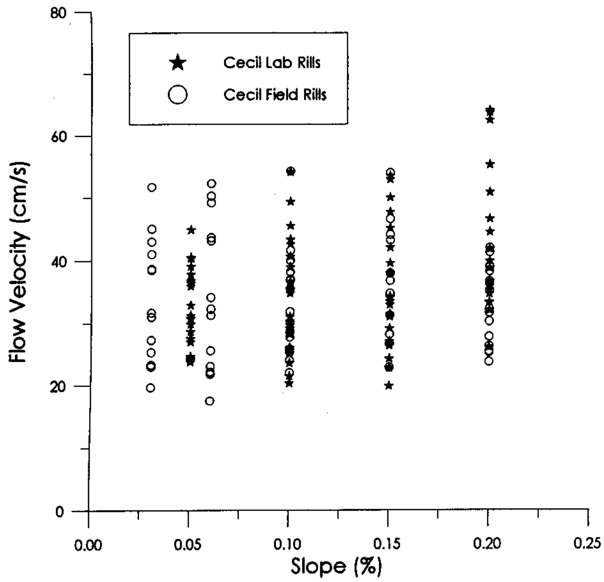

Figure 7. Average flow velocity as a function of slope for the Cecil laboratory and field rill experiments.

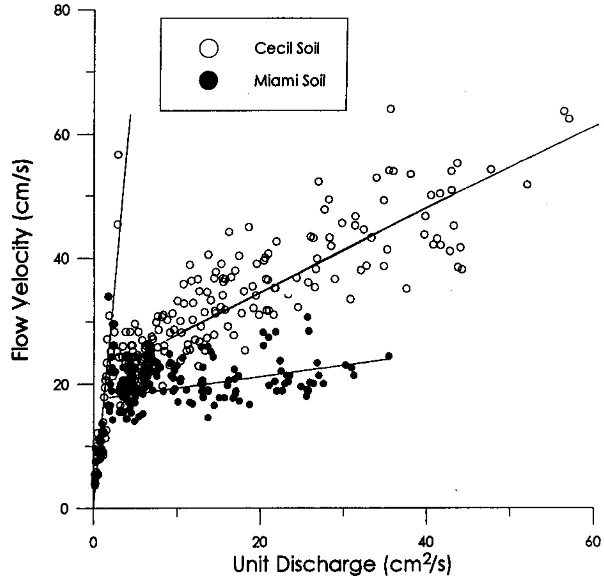

Figure 9. Average flow velocity as a function of unit flow discharge rates for the Miami and Cecil soils, including data from miniflume, laboratory rill, and field rill experiments. The lines of best fit represent the relationship for the Cecil soil in the turbulent flow regime, the Miami soil in the turbulent regime, and the combined data for the Miami and Cecil soils in the laminar flow regime.

soil and the level of discharge. For unit discharge of less than approximately 3 cm^2 s^{-1} the Miami and Cecil soils behaved similarly, and the slope of the U versus discharge line was quite steep ($\Delta U/\Delta q$ = 13.3 cm^{-1}). These data were in the laminar flow regime. In the turbulent regime the Cecil soil exhibited a greater effect of the unit discharge on $U(\Delta U/\Delta q = 0.7 \text{ cm}^{-1})$ as compared to the Miami soil. Again, this is due to the fact that the Miami soil was more susceptible to the development of erosional roughness as a function of increased flow rate than was the Cecil (Figures 3–5). In fact, $\Delta v/\Delta q$ for the Miami soil in the turbulent range was not significantly different from zero, and the velocity of rill flow for the Miami was essentially 20 cm s^{-1}.

For many physically based hydrologic and erosion models the unit discharge of flow may be a difficult parameter to estimate. For that reason, we present here the relationship between unit discharge and total discharge for the Miami and Cecil rill and miniflume experiments (Figure 10). Interestingly, the two soils follow basically the same relationship between q and Q despite the fact that the soils act considerably differently

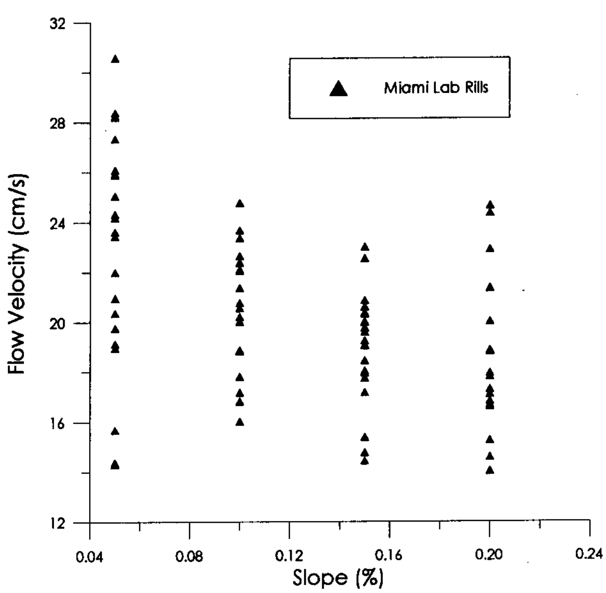

Figure 8. Average flow velocity as a function of slope for the Miami laboratory rill experiments.

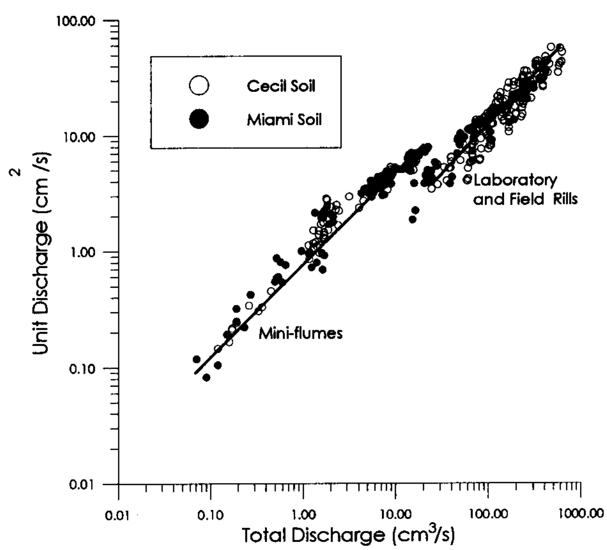

Figure 10. Unit flow discharge rates as a function of total flow discharge rates for the Miami and Cecil soils, including data from miniflume, laboratory rill, and field rill experiments. The two lines of best fit represent the relationship for the combined data for the Miami and Cecil soils from the miniflume experiments (representing the lower discharge rates) and that for the combined data for the Miami and Cecil soils from the rill experiments (representing the greater discharge rates).

in other respects. Also interesting is the marked difference in the q versus Q relationship between the miniflume and rill experiments. Though the two lines of best fit for the miniflume and rill experiments parallel each other, the unit discharge per total discharge for the miniflumes was greater. This result is probably due to the physical configuration of the two experiments. The V notch of the miniflumes was much steeper, and undoubtedly more constrictive, relative to the nearly flat initial bed conditions of the laboratory and rill studies. For field conditions the relationship for the greater flow rates derived from the laboratory and rill data should be used:

$$q = 0.232Q^{0.865}. \tag{3}$$

Erosion

The best predictor of unit sediment load q_s (g s^{-1} cm^{-1}) was stream power ω (g s^{-3}) as calculated by

$$\omega = \rho_w g S q, \tag{4}$$

where ρ_w (g cm^{-3}) is the density of water, g (cm s^{-2}) is the gravitational constant, S is slope, and q (cm^2 s^{-1}) is unit discharge of water (Figure 11). All of the data followed a logistic function curve with the exception of the 15 and 20% slope data from the Cecil field rill experiment, which fall well below the curve. The reason for this could be that these treatments were limited by a relatively less erodible soil layer below the plowed top layer of soil which restricted the sediment source. Sediment source versus transport limitations in the data are discussed below. The form of the best fit logistic curve was

$$\log(q_s) = \alpha + \beta^* e^{[\gamma + \delta \log(\omega)]}/\{1 + e^{[\gamma + \delta \log(\omega)]}\}, \tag{5}$$

where $\alpha = -34.47$, $\beta = 38.61$, $\gamma = 0.845$, and $\delta = 0.412$. The coefficient of determination r^2 between measured and predicted data using (5) was 0.93.

Several types of curves could have been used to give a level of fit to the data equivalent to that given by the logistic curve. However, this curve was chosen for physical considerations.

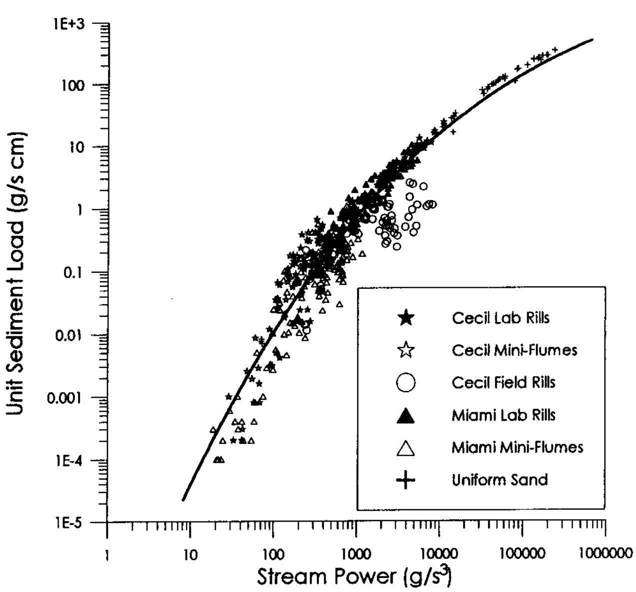

Figure 11. Unit sediment load as a function of stream power ω for the data from all six experiments. The best fit line is in the form of the logistic function given in (5).

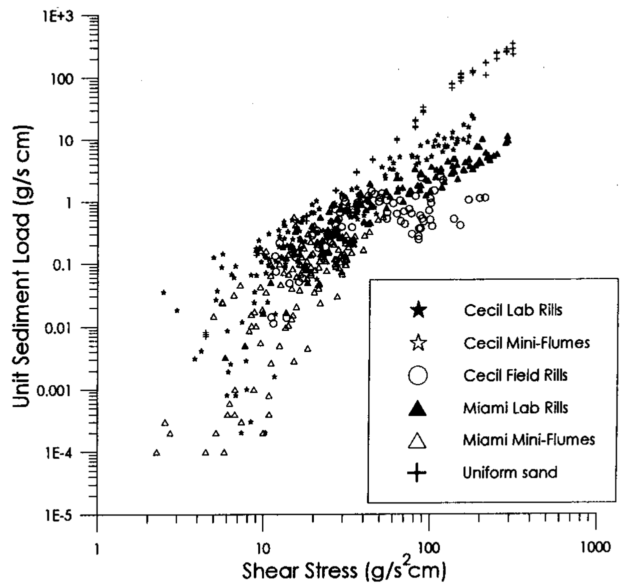

Figure 12. Unit sediment load as a function of average bed shear stress τ for the data from all six experiments.

The logistic curve is often used to describe cumulative distribution functions, particularly those relative to the use of threshold theory [*Neter and Wasserman*, 1974]. Three recent, independently developed soil erosion theories have been proposed which use probabilistically based threshold theories for soil erosion. *Nearing* [1991] proposed a probabilistic detachment model which makes use of overlapping probability density functions, one for the threshold failure strength of individual soil particles on the bed, and one for turbulent "bursting" events which occur at the boundary of the turbulent flow and the soil bed. *Wilson* [1993] proposed another equation wherein exceedance probability distributions were used to describe shear stresses of turbulence at the soil bed. *Larionov and Krasnov* [1994] made use also of a cumulative probability function of velocity fluctuations in turbulent flow to describe erosion in shallow flows. While we do not attempt here to evaluate or apply these models to the data from the study, the reasonable fit of the logistic curve here is encouraging in terms of probabilistically based particle threshold erosion theory.

Other hydraulic parameters often used for erosion and transport equations did not work as well for these data. While the value of r^2 between the logarithm of unit sediment load and the logarithm of stream power was 0.89, the overall fit for logarithm of shear stress to logarithm of sediment load was only 0.76 (Figure 12), and the fit between logarithm of unit stream power and logarithm of sediment concentration was only 0.53 (Figure 13). Also, the potential for improving the sediment load predictions based on shear stress partitioning by using the hydraulic friction factor term did not appear to be promising (Figure 14) since the statistical fit of the logarithms was poorer than that for shear stress alone, with $r^2 = 0.62$. Figures 12, 13, and 14 represent less than desired results for hydraulic parameters commonly used in erosion and sediment transport modeling. These data would suggest that these parameters are probably not the best for modeling rill erosion. *Govers* [1990] tested the possibility of modifying the stream power term as a function of hydraulic radius R (cm) and critical stream power ω_{cr} (g s^{-3}), terming it an "effective stream power" Ω (g$^{1.5}$ s$^{-4.5}$ cm$^{-2/3}$), where

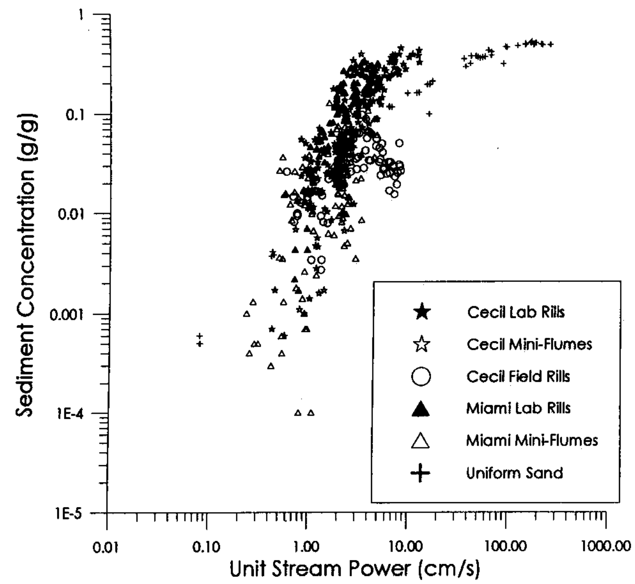

Figure 13. Sediment concentration as a function of unit stream power for the data from all six experiments.

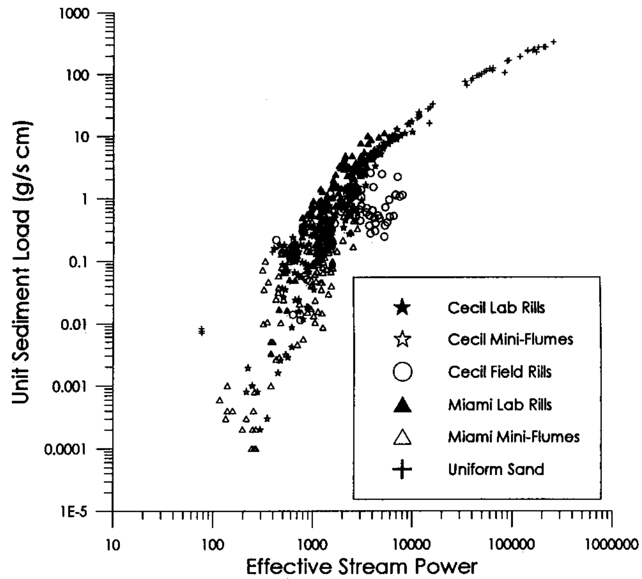

Figure 15. Unit sediment load as a function of effective stream power Ω (without the critical term) for the data from all six experiments.

$$\Omega = (\omega - \omega_{cr})^{1.5}/R^{2/3}. \quad (6)$$

Since the critical stream power values, as judged by the intercept for both the Miami and Cecil data, were similar for both soils, we evaluated the effective stream power term without consideration of the critical stream power value (Figure 15). The overall fit was not as good as the fit for stream power alone, with the r^2 for the logarithm of unit load versus log (Ω) equal to 0.78. Thus the addition of the hydraulic radius term in explaining these data is not warranted.

Govers [1990] proposed the use of (1) for describing the transport of uniform sand. This relationship was evaluated for rills in the nonuniform, aggregated soil used in this study, as well as for the uniform sand (Table 1). The uniform sand used in our experiment was between the sizes C (218 μm) and D (414 μm) used by *Govers* [1990]. For the data of *Govers* [1990] and for the sand data from this experiment the exponent "C" for the slope term was greater in value than the exponent "B" for the unit discharge term. However, for most of the other data (with the exception of the total results for the Cecil) the relative magnitudes of the two exponents were reversed from those for the sand.

Excess shear erosion models, such as the WEPP model [*Nearing et al.*, 1989], might use data such as that shown in Figure 16 for the Cecil and Miami laboratory flume experiments. The scatter of points for each soil around the sediment versus shear relationship is typical for rill erosion studies [*Laflen et al.*, 1991; *Elliot et al.*, 1989; *Norton and Brown*, 1992; *Brown and Norton*, 1994]. For this set of data the use of stream power in place of shear caused two significant differences (Figure 17): (1) the fit between the hydraulic variable and sediment load was better, and (2) the shear versus sediment relationship

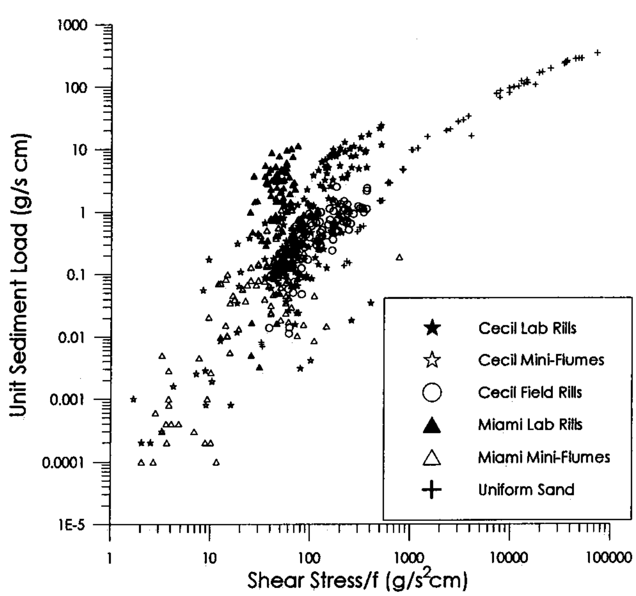

Figure 14. Unit sediment load as a function of average bed shear stress τ divided by the total hydraulic friction factor f for the data from all six experiments.

Table 1. Statistical Regression Results for the Logarithmic Transform of (1) for the Data From This Study and Selected Data of *Govers* [1990]

Soil Material	Experiment	A†	B†	C†	r^2	n
Cecil	miniflume	1.05	2.13	1.90	0.80	80
	laboratory rill	0.60	1.58	1.48	0.94	79
	field rill	0.12	1.55	1.14	0.84	44
	combined data	0.75	1.61	1.64	0.87	203
Miami	miniflume	0.21	2.04	1.55	0.82	142
	laboratory rill	0.62	1.60	1.50	0.88	93
	combined data	0.19	2.01	1.48	0.90	235
Uniform sand	flume	1.12	1.21	1.41	0.96	53
All	combined data	0.64	1.58	1.42	0.89	491
218 μm sand	*Govers*, [1990]*	0.74	1.50	1.96	0.98	69
414 μm sand	*Govers*, [1990]*	0.76	1.24	1.71	0.99	62

*Data taken from *Govers* [1990].

†A, B, and C are fitted coefficients for $\log_{10}(T_c) = \log_{10}(A) + B \log_{10}(q) + C \log_{10}(S)$; r^2 is the coefficient of determination for that equation; and n is the number of data points used.

was different for the two soils while the stream power versus sediment load relationships were statistically similar for the two soils; that is, the intercepts and slopes of the regression lines with stream power for the individual soils were statistically equivalent (within a 95% confidence interval) to those for the regression of the combined data. The values of r^2 for the Cecil and Miami soils using shear stress were 0.83 in both cases, while the value of r^2 when using stream power on the combined data was 0.94.

The difference between stream power and shear stress is flow velocity; that is,

$$\omega = \tau v. \tag{7}$$

This suggests that the additional velocity term was responsible for stream power being a better predictor of sediment load than is bed shear stress. In other words, the variance in velocities, or lack thereof, was correlated to the observed variances, or lack thereof, in the sediment load. Though the stream power versus sediment load relationships were similar for the Cecil and Miami soils, it is not suggested here that the sediment versus stream power relationship would be the same for all soils with model parameters of the scale of these laboratory rills. However, the possibility that the variability of rill soil erodibility parameters might cover a smaller range for the stream power equation than for the excess shear equation is worthy of investigation.

Equation (7) is an alternate form of (4), but it does not imply that one must know the velocity and shear stress of flow in order to apply the stream power model proposed in (6). In fact, an advantage of the use of stream power is that one need not necessarily know the hydraulic radius and velocity of flow in order to apply (4) but rather only the unit flow discharge, that is, the total discharge rate and the rill width.

The miniflume data exhibited a significantly greater level of scatter ($r^2 = 0.70$ for the Cecil and $r^2 = 0.39$ for the Miami) when plotted against stream power than did the data for the larger rills (Figure 18). The reason for this is not immediately clear.

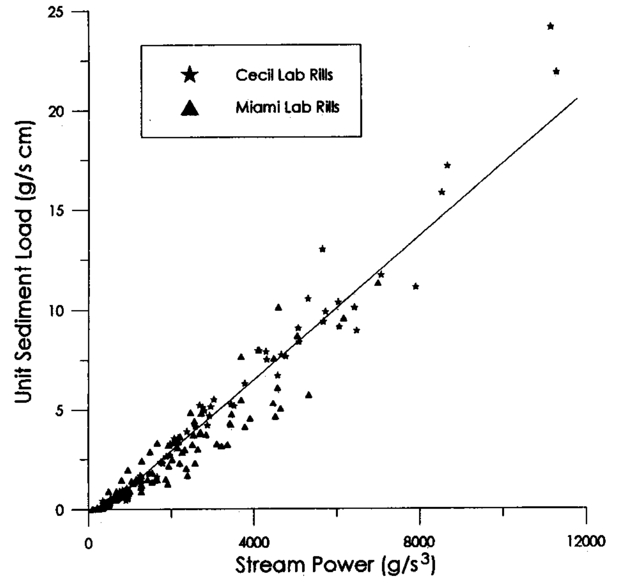

Figure 17. Unit sediment load as a function of stream power ω for the data from the laboratory rill experiments on the Cecil and Miami soils.

Sample Length Effects and Rill Detachment Profiles

Soil detachment, or entrainment, is a term used to identify the process of soil particle removal from the original in situ location within the soil matrix by erosive forces. The term is used to contrast the detachment process with that of sediment transport and the associated reentrainment of previously detached soil particles which may be resting on the bed at a particular point in time. The rate of soil particle detachment caused by flowing surface water has been observed to be a function of the amount of sediment load in the flowing water [*Ellison*, 1947]. It is known that as the sediment concentration

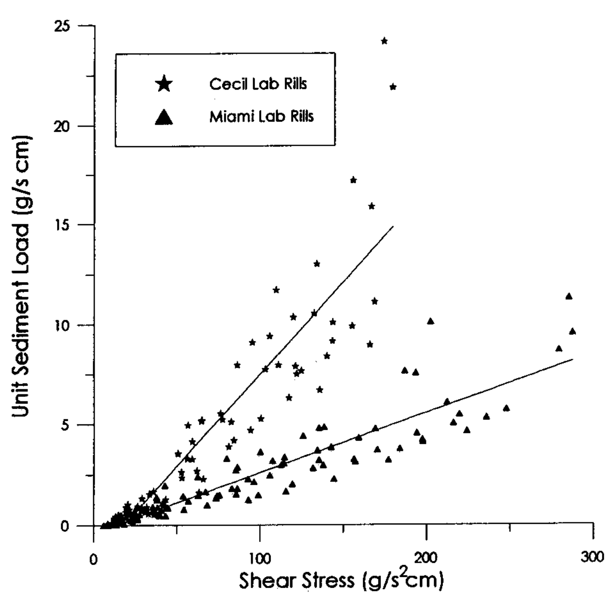

Figure 16. Unit sediment load as a function of average bed shear stress τ for the data from the laboratory rill experiments on the Cecil and Miami soils.

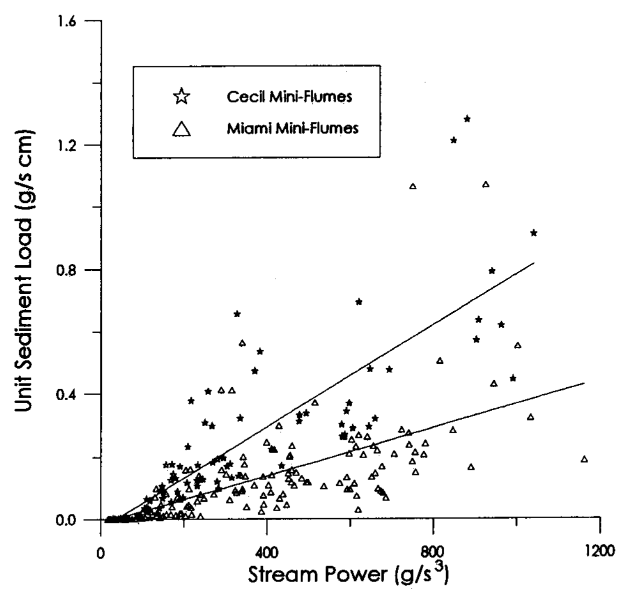

Figure 18. Unit sediment load as a function of stream power ω for the data from the laboratory miniflume experiments on the Cecil and Miami soils.

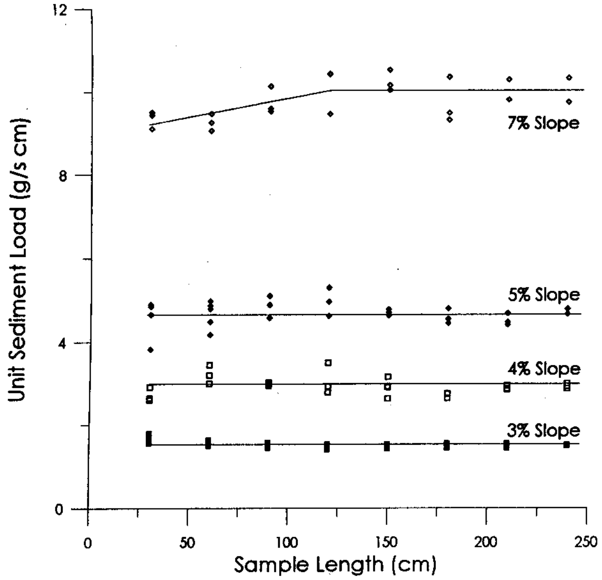

Figure 19. Unit sediment load as a function of sample length for the uniform sand experiment at 3, 4, 5, and 7% slopes.

is increased, the flow of water has less capacity to detach new particles from the soil matrix. This observation has been hypothesized [*Foster and Meyer*, 1972; *Foster*, 1982] to be related to the fact that the flowing water has a finite energy which imparts the ability of the flow to both detach new soil or carry existing sediment. Since the reentrainment and continued movement of sediment requires, in general, less energy than the process of detachment, this energy is used first. Energy in excess of that required for transport may be expended to detach new soil particles [*Foster and Meyer*, 1972]. *Hairsine and Rose* [1992b] hypothesized that this phenomenon may be attributed to a protection of the soil surface from the eroding forces of the water by the sediment which covers a part of the soil bed.

Data from the sample length treatments for the uniform sand illustrate the concept of the interaction between sediment load and detachment (Figures 19, 20, and 21). In general, sediment load washed from the sand bed increased as a function of sample length until it reached a constant value which corresponded to the sediment transporting capacity of the flow. This phenomenon was a function of the slope gradient. At 7% slope the flow reached sediment transport capacity at a critical flow length of approximately 120 cm (Figure 19). Overall, this critical length tended to increase with increased slope, though differences were not detected between 10 and 28% slope where a constant level of sediment load was reached at approximately 180 cm in sample length (Figure 20). Below 7% slope the sediment transport capacity had apparently already been reached with the 30 cm long sample (Figure 19). At 32 and 35% slope the 240 cm flow length was apparently too short to reach a constant level (Figure 21). The implication of these data is that for noncohesive material, surface water flow will reach a transport limiting condition over a very short flow distance.

In order to investigate the flow length effect with the soil materials we measured the eroded cavity volumes of the rills after a sampling of the tests to obtain an overall relative erosion amount as a function of the downslope distance in the rill. The 10% slope for the Cecil (Figure 22) and the 5% slope for the Miami soil are indicative of the results (Figure 23). The Cecil at 10% had a relatively constant volume of eroded rill (approximately 2000 cm^3) up to approximately 150 cm, at which point it began to decrease, reaching a very low volume (less than 200 cm^3) at the end of the rill. The Miami at 5% acted similarly. From 0 to 100 cm length on the sample, the eroded volumes were constant at approximately 500 cm^3 per segment. At lengths greater than 100 cm, the eroded volumes gradually decreased to approximately 50 cm^3 per segment at the end. The very high cavity volume in the first section was common but thought to be due to boundary effects where the water first entered the soil bed. In general, the cavity measurements indicated that the rills had approximately reached a transport limiting state within a flow length of 3 m.

These results have important implications for rill erosion experiments, models, and model parameterization. The studies

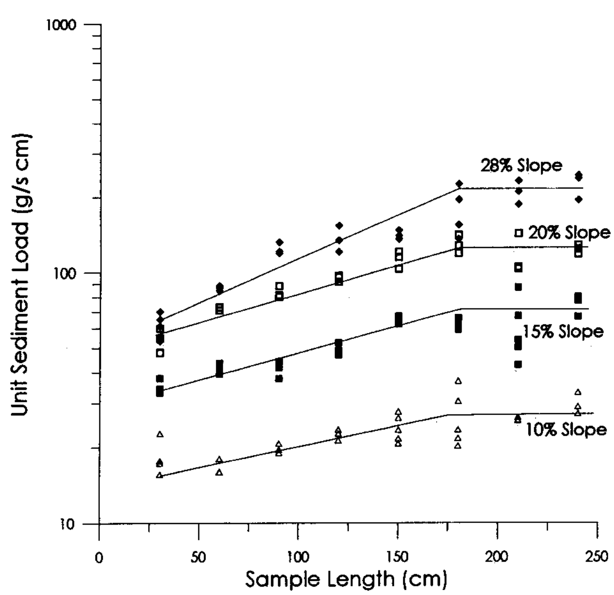

Figure 20. Unit sediment load as a function of sample length for the uniform sand experiment at 10, 15, 20, and 28% slopes.

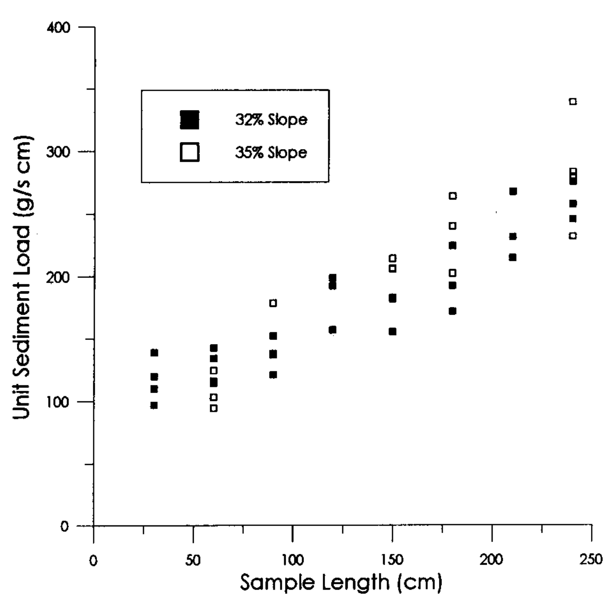

Figure 21. Unit sediment load as a function of sample length for the uniform sand experiment at 32 and 35% slopes.

[*Laflen et al.*, 1991; *Norton and Brown*, 1992; *Brown and Norton*, 1994] utilized rills in disturbed cropland soils which were 9 m long. The question immediately arises in light of the results from the current study as to what exactly these experiments were measuring. The experiments were designed to measure soil erodibility, that is, the detachment rate of the soil as a function of the excess shear stress model in WEPP. The results from this study indicate that these 9 m long rills in the disturbed soil material may be quite close to a transport limiting situation, which confounds the interpretation of the results relative to the definition of rill detachment parameters. In spite of this apparent problem, the application of rill erodibility estimation procedures derived from the experiments for the WEPP model as described by *Laflen et al.* [1991] and *Elliot et al.* [1989] produced model parameters which gave quite reasonable results for long-term natural rainfall erosion plot data [*Zhang et al.*, 1996; *Nearing and Nicks*, 1996]. This may be fortuitous, but more likely it is a result of the fact that the data from the field erodibility experiments were analyzed using the WEPP erosion model, and thus the data analyses and the model were mutually consistent. This idea is not unlike that presented above relative to the discussion of (2) wherein the friction factor term f_s used in the model is based on bare soil friction with its form roughness rather than on grain roughness alone, but the model parameters were calibrated for the results of *Nearing et al.* [1989] with the same assumption. It is important that new erosion models with fewer assumptions be introduced as new information is made available. It is equally important that parameters for models be derived using assumptions which are consistent with the model.

Conclusions

This study investigated rill hydraulics and erosion using data which span a wide range of hydraulic conditions, including slopes, flow rates, Froude numbers, Reynolds numbers, and bed roughness. The results of the study contradict several commonly held concepts and models of hydraulics and erosion of surface water flows which have been developed based on more limited data sets (commonly obtained for constant slope conditions) and/or sheet flow information which is not applicable to actively eroding rills.

The results of the study indicate that Reynolds numbers are not good predictors of hydraulic friction factors in rills even though the Reynolds number has been used for that purpose for both rills and sheet flow. This lack of correspondence is due, in part at least, to the fact that erosion and hydraulics are interactive and that steeper slopes and greater flow rates tend to increase erosion rates which effect rougher bed surfaces. The rougher surface tends to decrease the flow velocity. However, the effect of slope and flow rate are not the same, and the relative effect of the two factors is soil dependent.

Slope gradient, in general, did not influence flow velocities in the rills. This, along with the interactions discussed above relative to estimating hydraulic friction factors, makes the use of velocity equations such as the Chezy and Manning equations difficult for actively eroding rills. Data from these experiments suggested that unit flow discharge was the best predictor of flow velocities in eroding rills. Though the relationship between discharge and velocity appears to be soil dependent in the turbulent regime, it was the simplest and most consistent of relationships for estimating rill flow velocities for these data.

The best overall predictor for unit sediment load from the experiments was stream power. Other variables, including shear stress, shear modified by hydraulic friction, unit stream power, and effective stream power, did not produce as high a level of statistical fit to the sediment load data as did stream power. Our results indicated that a logistic curve between the logarithm of unit sediment load and the logarithm of stream power provided a good fit to the data ($r^2 = 0.93$). This result supports recently developed probabilistic based, particle threshold erosion theories [*Nearing*, 1991; *Wilson*, 1993; *Larionov and Krasnov*, 1994]. Stream power was a better predictor than shear stress for individual data sets as well as for the data as a whole. Stream power showed a better fit to both laboratory and field rill data for each soil, and with less between-soil variability, than did shear stress.

The data from sample length and rill cavity measurements

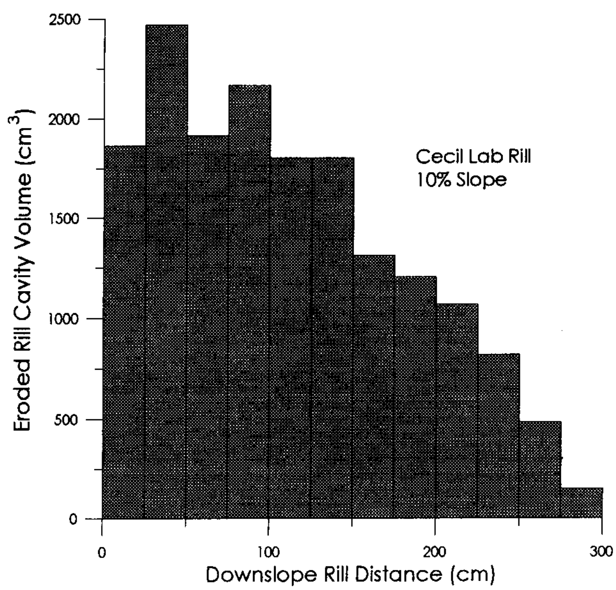

Figure 22. Eroded rill cavity volumes after a Cecil soil laboratory rill experiment run at 10% slope.

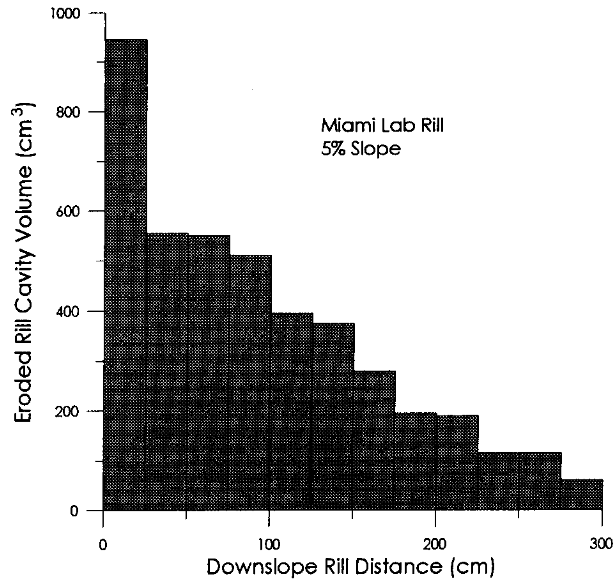

Figure 23. Eroded rill cavity volumes after a Miami soil laboratory rill experiment run at 5% slope.

indicated that these rills were near sediment transport capacity. This result implies that the equation suggested here for sediment load versus stream power (equation (6)) is appropriate for use as a sediment transport equation rather than a sediment source equation, per se. This has important implications for rill erosion experiments used to parameterize erosion models. In the past, it has been common to parameterize rill detachment models using rills in disturbed soil with lengths up to 9 m. According to the results presented here, sediment load from such rills may be more strongly influenced by sediment transport limitations rather than soil detachment.

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(Received June 5, 1996; revised December 17, 1996;
accepted December 31, 1996.)