Sediment Problems: Strategies for Monitoring, Prediction and Control (Proceedings of the Yokohama Symposium, July 1993). IAHS Publ. no. 217, 1993.

Examining the transition from sediment transport in water to mass movement

C. W. ROSE, A. L. PRESBITERO & R. K. MISRA
Faculty of Environmental Sciences, Griffith University, Nathan Campus, Brisbane, Queensland, Australia 4111

Abstract Recent physical-based theory of soil erosion by water provides a theoretical expression for the upper limit sediment concentration produced by rainfall and overland flow driven processes. This theory thus provides the capacity to indicate whether or not additional erosion processes such as rapid mass movement, have occurred in any measured erosion event. The paper outlines such theory and a simplified form which is applicable when data on total soil loss and runoff rate per unit area from runoff plots are measured. This simplified theory is applied to 31 sets of field measurements on 12 m long bare plots up to 70% in slope in the Philippines. For this set of data mass movement was indicated to be a relatively minor contributor to soil loss even on these steep slopes. The theory and methodology to reach such conclusions is general in its application.

NOTATION

c sediment concentration (kg m^{-3})
$\bar{c}$ average value of c for erosion event (kg m^{-3})
c_t sediment concentration at the transport limit (kg m^{-3})
D depth of overland flow of water (m)
F fraction of streampower used in eroding soil (dimensionless)
I arbitrary number of settling velocity classes into which a sediment settling velocity characteristic is divided
J specific energy of entrainment (J kg^{-1})
r_e rate of entrainment (kg m^{-2} s^{-1})
S land slope (the sine of the angle of inclination)
V velocity of overland flow of water (m s^{-1})
v_i settling velocity of sedimentary units in size range i (m s^{-1})
β soil erodibility parameter defined in equation (4) (dimensionless)
ρ density of water (kg m^{-3})
σ density of sediment (kg m^{-3})
Ω streampower (W m^{-2})
Ω_o threshold value of Ω(W m^{-2})

INTRODUCTION

Soil erosion research has been dominated by agricultural objectives. Hence such research has commonly been limited to slopes of about 20%, steeper slopes being assumed marginal for cultivation (Lal, 1988). However population pressure and other factors have led to increasing dependence on cultivated steeplands, for example in the Philippines (Garrity & Sajise, 1990), with consequent issues of sustainability and offsite environmental damage.

In steeplands, whether or not cultivation is practised, the possibility of gravity-driven mass movement always exists. The question then arises as to whether or not the products of water erosion research developed assuming no contribution from rapid mass movement are applicable at steep slopes. Also, do such products provide an indication as to whether or not the assumptions on which they are based are abrogated?

In this paper these questions will be addressed using the physically-based model of erosion described by Rose & Hairsine (1988); Rose *et al.* (1990); Hairsine & Rose (1992a,b). Because they are based entirely on experimentation, methods such as the Universal Soil Loss Equation (Wischmeier & Smith, 1978) will not be considered here.

This paper also considers the results when the physically based model referred to above is applied to data collected from field runoff plots at a range of slopes from 10 to 70%.

Theoretical background

Serious erosion on steep slopes which does not involve mass movement is dominantly due to the rate of working of the sheer stress between the soil surface and water flowing over it (i.e. to the streampower as defined by Bagnold (1977). Rose & Hairsine (1988) assumed that some fraction (F) of this streampower Ω is used in eroding soil, the remaining fraction being dissipated into heat. The effective streampower ($F\Omega$) can be used to do work against the cohesive forces in the original soil matrix in the process of soil entrainment. By definition of terms the rate at which this energy is expended is the product of the mass rate of entrainment, r_e (kg m^{-2} s^{-1}) and J (J kg^{-1}), the specific energy of entrainment. Thus:

$$r_e J = F(\Omega - \Omega_o) \qquad (1)$$

where Ω_o is a threshold value of Ω, below which entrainment is negligibly small.

Unless it is very fine, sediment removed from the soil surface by any mechanism including entrainment will settle and return again to the soil surface where it forms a deposited layer of loose aggregates which can also be eroded, a process called "re-entrainment." Since the loose aggregates in this deposited layer have little opportunity to re-establish cohesive bonds to neighbouring aggregates before they are re-entrained, work available from the effective streampower will be expended chiefly in lifting the aggregates into the water layer against their immersed weight. Thus the work expended per unit mass in re-entrainment can be much less than that in entrainment. It follows that the maximum concentration of sediment in the water layer will be achieved if the soil surface is completely covered by a deposited layer of sediment, so that all the effective streampower is used in the process of re-entrainment.

This theoretical maximum in sediment concentration can be identified with the experimental observation of an upper limit, previously called the "transport limit" (c_t) by Foster (1982).

Based on these concepts Rose & Hairsine (1988) derived the following expression for c_t applicable to sheet flow:

$$c_t = \frac{F\rho}{\Sigma v_i / I} \left[\frac{\sigma}{\sigma - \rho} \right] SV \quad (\text{kg m}^{-3}) \tag{2}$$

where symbols are given in the Notation section. If downslope flow is carried in rills then equation (2) still applies with minor modification reflecting changed geometry.

The product SV in equation (2) has been called the "unit streampower" by Yang (1972). The mean settling velocity of sedimentary units in water ($\Sigma v_i/I$) has been termed the "depositability" of the sediment (Rose *et al.*, 1990). In practice an upper limit to this summation $\Sigma v_i/I$ is set by the size of aggregate that can be said to "settle" in the limited depth of overland flowing water. Misra & Rose (1991) have used a sediment size equal to water depth for this limit. Retaining the term "depositability" for the situation where this summation is not limited by water depth, the term "effective depositability" is used for the sum truncated at an upper settling velocity limit. This corresponds to a sediment size equal to the average water depth at exit from the runoff plot during the erosion event.

Figure 1 shows c_t, calculated using equation (2) plotted against the effective streampower ($\Omega - \Omega_o$) of the overland flow, where:

$$\Omega = \rho g S D V \quad (\text{W m}^{-2}) \tag{3}$$

and Ω_o (see Notation) is a constant. The solid curve in Fig. 1 is where $\Sigma v_i/I$ is assumed independent of flow depth (D), and the dashed curve where this dependence is recognised. Both curves show a dependency of c_t on Ω at low values of Ω. However the relative constancy in c_t with Ω when recognition is given to the dependency of effective depositability on V (and thus D, to which it is related) indicates that there is an important range of Ω for which the ratio $[V/(\Sigma v_i/I)]$ in equation (2) is approximately constant.

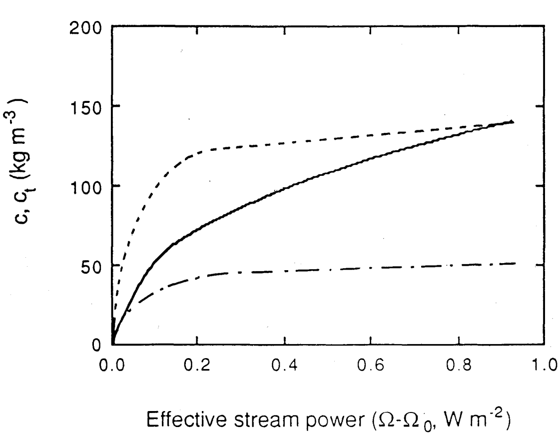

Fig. 1 Variation of c_t with ($\Omega - \Omega_o$) for the same assumed variables as in Rose *et al.* (1990), for constant $\Sigma v_i/I$ (—), and with $\Sigma v_i/I$ increasing with depth of flow (---). Also shows c calculated as c_t^β with c_t computed using variable $\Sigma v_i/I$ (_._) and $\beta = 0.8$.

There is an upper limit to $\Sigma v_i/I$ (equal to the depositability of the sediment), but V increases indefinitely with Ω (for any given slope). Thus c_t must increase significantly with Ω beyond some high value of Ω where the water depth exceeds the maximum size of sedimentary units (i.e. the largest aggregates). It is possible that at high flow velocities large sedimentary units move by rolling rather than by the successive hops involved in the series of entrainment or re-entrainment events followed by deposition which is assumed in the theory of Rose & Hairsine (1988). Which of these alternate mechanisms is more effective in sediment transport is unclear and requires further examination.

Even cultivated soil has some strength, and this limits the sediment concentration at any given value of Ω to a value less than the maximum value c_t. As illustrated in Rose *et al.* (1990), as soil strength (and thus J) increases, the theory of Rose & Hairsine (1988) predicts that sediment concentration will be related to Ω by a series of curves corresponding to different values of J, one of which is illustrated in Fig. 1.

It has been shown (Misra & Rose, 1989) that such relationships, which arise from more complex theory, can be very well represented using the following simple relationship involving a single parameter β (≤ 1):

$$c = c_t^{\beta} \tag{4}$$

At the current stage of soil erosion research technology, measurement of sediment concentration as a function of time during an event is not common. Accepting this experimental limitation, the only sediment data available is the mean sediment concentration $\bar{c}$ calculated from the total soil loss and effective runoff for the complete erosion event (defined as runoff for which $\Omega > \Omega_o$). Misra & Rose (1989) have shown how equation (4) can be applied to the event as a whole, where c is replaced by $\bar{c}$ and c_t by the appropriate average value of c_t during this event, which can be calculated as described by Rose (1993) if data on runoff rate as a function of time is available.

With c experimentally measured, and the appropriate average value of c_t calculated from theory using data on runoff rate and rill geometry and frequency (if rilling occurs), then the erodibility parameter β (equation 4) can be calculated from:

$$\beta = \ln \bar{c} / \ln c_t \tag{5}$$

In Fig. 1 the calculated variation of c with $(\Omega - \Omega_o)$ for $\beta = 0.8$ is shown. The possible family of curves for β correspond to a family of curves for J (Rose *et al.*, 1990).

From equation (5), if $\beta = 1$, then $\bar{c} = c_t$, the maximum sediment concentration at the transport limit. Hence if experimental data from runoff plots yields a value of β greater than unity, then this indicates that some other process or processes not taken into account in the theory must be contributing to the sediment concentration. The processes taken into account in the theory are rain-driven as well as flow-driven processes. Thus if $\beta > 1$ is obtained, this could be an indication that mass movement of some kind has contributed to soil loss, through the mechanism leading to a value of $\beta > 1$ is not defined.

An application of the above theory to data form field runoff plots will be described in the next section.

FIELD EXPERIMENT: RESULTS AND DISCUSSION

The experiment investigated the effect of various farm practices on soil loss from instrumented runoff plots at the Visayas State College of Agriculture (VISCA) at Baybay, Leyte, the Philippines (latitude 10°44'N, longitude 124°48'E at 30 m above sea level, mean annual rainfall 2000-4000 mm. Results will be reported for two years of measurement of loss of soil (classified as Oxic Dystropepts) from runoff plots kept bare of cover by intermittent manual cultivation and weeding.

Plot lengths were 12 m, and slopes approximately 10, 50, 60 and 70%. Mass movement was expected to be unlikely at 10%, but possible at the higher slopes. Details of the experiment are given by Escalante *et al.* (1993), using methods of analysis described by Ciesiolka *et al.* (1993) and Rose (1993), using methods of analysis described by Ciesiolka *et al.* (1993) and Rose (1993). Results of this experiment, including a presentation and full discussion of the variation in β with cultivation and time is given by Presbitero *et al.* (1993). The data presented were derived from measurements of total soil loss for each erosion event during which runoff rate was measured as a function of time using tipping bucket technology (Bonnell & Williams, 1986) and electronic data loggers. Data was handled using the suite of data analysis programs described by Ciesiolka *et al.* (1993) and Rose (1993).

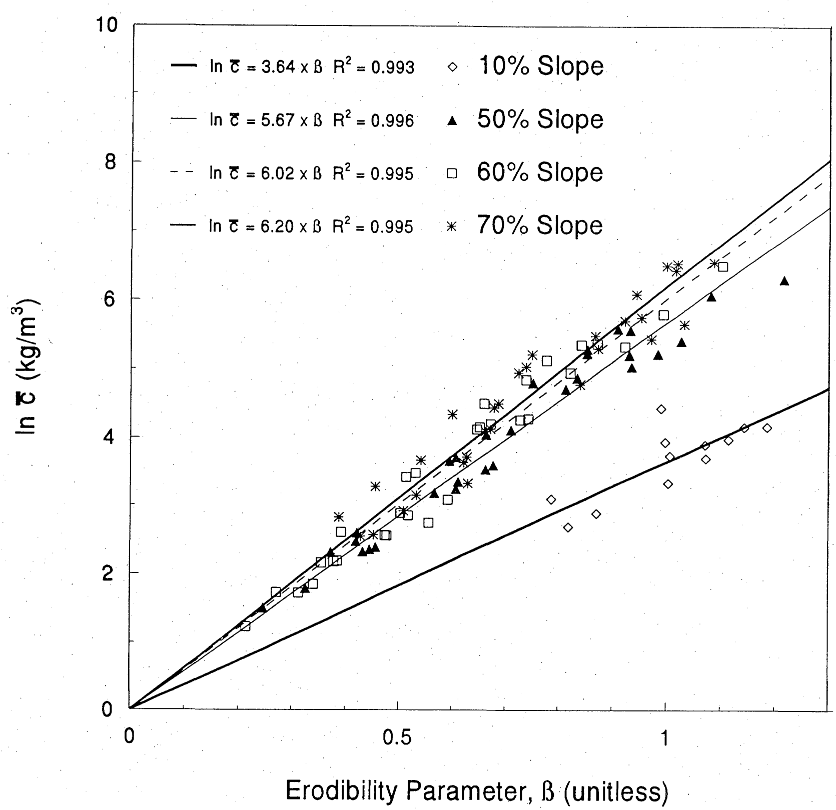

Fig. 2 Plot of ln$c$ against β. Data on c and β are from Presbitero *et al.* (1993) for 12 m long bare soil runoff plots of the different slopes indicated in the figure.

Figure 2 shows ln$c$ plotted against β, with each data point representing an erosion event. It can be seen that the data from each plot of different slope can be quite well fitted by a linear relationship. A later start in recording is the reason why there is less data for the 10% plot compared with the plots of steeper slope. The slope of the relationship in Fig. 2 increases with the slope (S) of the experimental plot.

It follows from Fig. 2 and equation (5) that for each plot with its particular slope, $\ln\bar{c}_t$ is approximately constant despite the substantial range in runoff streampowers represented in the data in Fig. 2. That c_t (and hence $\ln\bar{c}_t$) should increase with plot slope S is in accord with equation (2) (whether or not rilling occurred). That *ct (*and hence $\ln\bar{c}_t$) can be approximately constant across a range of streampowers was demonstrated from theory in Fig. 1.

There was visual observation of occasional minor surface "mudflows" on the steep slope plots under some conditions. Thus that there were some values of $\beta > 1$ shown in Fig. 2 were not unexpected. Though no visual observations during erosion events are so far available for the 10% plot, no mass movement or processes other than those recognised in the theory of Rose & Hairsine (1988) would be likely. However, Fig. 2 shows values of $\beta > 1$ for this low slope plot also.

In the calculations used to obtain Fig. 2 the value of the fraction F introduced in the Theory section was assumed to have the value 0.1, based on data by Proffitt (1988). Evaluation of the value of F is ongoing, and there is some evidence that F can be greater than 0.1. It is therefore possible than an explanation of the values $\beta > 1$ for the 10% slope plot is due to use of too low a value of F in the analysis. However, if this explanation is accepted, it applies equally to the high slope plots, in which case there would be little evidence of $\beta > 1$ on these plots also.

This current minor uncertainty in the appropriate value to use for F is reflected in uncertainty in β, in the direction that calculated values of β are unlikely to be too low, but may be too high. Despite this uncertainty in F, Fig. 2 suggests that even on steep slopes of up to 70%, mass movement of soil or other processes directly aided by gravity does not appear to be a major contributor to soil loss. This conclusion is no doubt location and soil type specific.

CONCLUDING COMMENTS

Whilst the conclusion reached that mass movement is a relatively minor contributor to soil loss even on these steep plots may be limited to this soil-type and environment, the importance of the theory and methodology outlined is in its generality of application. This theory also has the advantage that it provides a definite indication as to whether or not erosion processes other than those due to rainfall impact or the action of overland flow are contributing to soil loss. Such an indication can be useful, for example in considering or evaluating management options designed to limit soil erosion or sediment loss to lowland areas, streams or impoundments.

Acknowledgements The authors wish to acknowledge the support of the Australian Centre for International Agricultural Research under ACIAR Project No. 8551. Data in Fig. 2 was collected with the collaboration of Dr M. C. Escalante.

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