CENTRE FOR AUSTRALIAN REGOLITH STUDIES

DISCOURSE ON RAINFALL EROSION PROCESSES AND MODELLING ON HILLSLOPES

by

P.I.A. Kinnell

Centre for Australian Regolith Studies
Occasional Publication No. 6
November 2000

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CENTRE FOR AUSTRALIAN REGOLITH STUDIES,
Australian National University
University of Canberra

ISBN 0 85889 831 4

First published 2000

PREFACE

Dr Peter Kinell is an expert on soil erosion, particularly related to rainfall. He worked for 26 years as a scientist with CSIRO Division of Soils, including with the Sediment Transport Group. His early research on raindrop impact contributed to development of the Revised Universal Soil Loss Equation (RUSLE) by the U.S. Department of Agriculture. Following a visit to the USA in 1993 he contributed to a new modification of the USLE called USLE-M. In 1996 he joined the Centre for Resource and Environmental Studies at the Australian National University where he worked on modelling erosion within catchments through linking USLE-M with the Agricultural Non Point Source pollution model (AGNPS). He also has interests in the modelling of climate with respect to hydrology and erosion.

Peter is currently an adjunct Senior Lecturer in the School of Resource, Environmental and Heritage Sciences at the University of Canberra, Australia. He is involved in teaching and research on soil erosion and erosion-related aspects of land management.

This publication reviews components of various models of rainfall erosion.

K.G. McQueen
Co-Director, CARS
November 2000

A Discourse on Rainfall Erosion Processes and Modelling on Hillslopes

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Abstract

Models of rainfall erosion provide a capacity to predict the outcomes of current and alternative land management on onsite erosion and may also provide the basis for predicting offsite impacts. This discourse describes some of the main rainfall erosion modelling approaches that have been developed over the past 40 years. Empirical, conceptual and process-based approaches are described in relation to the modelling of erosion at scales ranging from less than a metre through to the catchment scale.

Introduction

Erosion by water is a major factor in producing land degradation in many parts of the world. An understanding of the processes involved is beneficial to the development of measures to combat erosion and to the development of models capable of examining the impact of alternative farming systems and landuses. In this discourse, descriptions of rainfall erosion processes will be provided and some of the approaches to modelling these processes described.

Forms of Erosion

Water erosion comprises of a number of forms; sheet, rill and interrill erosion, ephemeral gully erosion, gully erosion, bank erosion, snowmelt erosion. Rainfall erosion encompasses all these forms but since snowmelt in not a major problem in most parts of the world, little attention is given to snowmelt erosion in the literature. The various forms, with the exception of snowmelt erosion, can often be seen as a topographic sequence in the landscape. Sheet erosion dominating the areas near the top of a watershed, giving way to rill and interrill erosion, and perhaps gully erosion, further down the hillslope. Rills are channels that can be eliminated by normal cultivation. Gullies are larger channels that cannot. They usually occur in areas where flow concentrates laterally. Lateral concentration of flow in a cultivated area may strip wide strips of soil down to the bottom of the cultivated layer. These areas are known as ephemeral gullies since gullies may develop in these areas if measures are not taken to prevent them. Narrow deep ephemeral gullies can also occur on some soils. Stream bank erosion, as its name implies, occurs in rivers but soil from banks of gullies can contribute to gully erosion. Sidewall collapse can also occur in rills. Each form is usually dominated by particular process that may not necessarily appear important to the development of the other forms. The change of form down a hillslope reflects the change from erosion processes driven by energy derived from raindrops hitting the ground surface to erosion processes driven by flow energy. Because water accumulates on the soil surface in both space and time, the erosional form observed a point in space may also vary in time.

Detachment and Transport Processes Associated with Erosion Driven by Raindrop Energy.

Erosion is a process that involves the detachment of soil particles from within the soil surface followed by the transport of these detached particles away from the site of detachment. When erosion is driven by the energy derived from raindrops impacting the soil surface, raindrop energy is used to overcome the bonds that hold particles in the soil surface and may also be used in the transport of the detached particles away from the site of drop impact. One commonly reported transport mechanism is raindrop splash. Raindrop splash moves detached soil particles radially away from the site of detachment. The raindrop detachment - splash transport (RD-ST) system often operates at the onset of a storm when little or no surface water flow occurs. However, splash transport (ST) is a highly inefficient transport system. If the soil has no slope, material splashed away from the point of impact of one drop is replaced by material splashed by other drops in the surrounding area. If the soil surface has a slope, then material splashed downslope travels further than material splashed upslope resulting in the net down slope migration of detached material. That downslope migration increases as the slope gradient increases but it takes many drop impacts to cause much material to move down slope in most cases. Rainfall erosion is either limited by the detachment or transport capacities associated with raindrop impact or surface water flow. RD-ST is a transport limiting process.

When water flows develop on the soil surface, raindrops penetrate through the flow to detach soil particles which may then be splashed as a result of the breakup of the drop or alternatively may be lifted into the flow where they move downstream as they fall back to the surface. Subsequent drop impacts lift the particles into the flow again and again and they move downstream on each occasion. The resulting transport process involves both raindrop impact and flowing water and, because of this, has been called Raindrop - Induced Flow Transport (RIFT) (Kinnell, 1990). With coarse material, raindrop impact in flowing water may stimulate particles to roll rather than saltate. RIFT is a more efficient transport system than ST. RD-RIFT plays a major role in moving soil material from interrill areas to rills. Splash can also move material from areas not covered by flow to areas where RIFT operates to give RD-ST-RIFT systems. While RIFT is more efficient than ST, it still requires numerous drop impacts to move material downstream and RD-RIFT systems are transport limiting.

In many cases, thin surface water flows have a capacity to move loose material sitting on the surface but may not have the capacity to detach material from the underlying surface. However, raindrops penetrating the flow may be able to do this. As a result, particles detached by drop impacts are transported downstream without the need for raindrops to be involved in the transport process. This raindrop detachment – flow transport (RD-FT) detachment-transport system is more efficient than RD-RIFT. Often, both RD-RIFT and RD-FT occur simultaneously in the same flows, coarse material being transported by RIFT, fine by FT.

Figure 1 provides a schematic representation of how the detachment and transport forms vary with raindrop energy (e) and stream power (Ω) when raindrop energy and stream power are used as measures of erosive forces associated with impacting raindrops and flowing water respectively. In Figure 1, the critical energy required for raindrops to detach soil particles held in the soil surface by cohesion and inter-particle friction is designated e_c. Raindrop detachment

(RD) does not occur unless $e > e_c$. The critical stream power for flow to detach soil particles held in the soil surface by cohesion and inter-particle friction is designated $\Omega_{c(bound)}$. Flow detachment (FD) does not occur unless $\Omega > \Omega_{C(bound)}$. Stream power is a hydraulic parameter that varies with flow discharge and slope gradient. Other parameters such as flow shear stress could have been used but the choice of parameter is academic. Consideration of a critical condition for FD remains the same irrespective of what parameter is used.

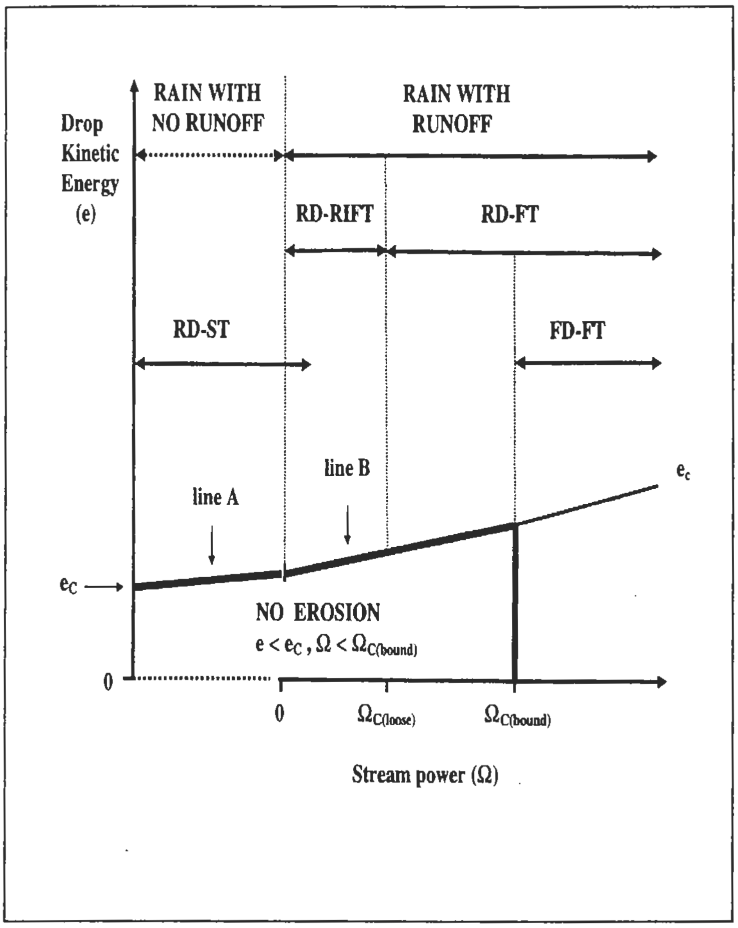

Fig.1. Detachment and transport processes associated with variations in raindrop and flow energies. e_C = critical raindrop energy to cause erosion. Line A = e_C prior to flow (increasing through crust development). Line B= e_C when flow occurs (increasing drop energy used to penetrate flow). $\Omega_{c(loose)}$ = critical stream power for transporting loose material. $\Omega_{c(bound)}$ = critical stream power for detaching soil from surface of soil matrix. RD - ST = raindrop detachment, splash transport. RD – RIFT = raindrop detachment, raindrop induced flow transport. RD – FT = raindrop detachment, flow transport. FD – FT = flow detachment, flow transport

Detachment and Transport Processes Associated with Erosion Driven by Flow Energy

Once $\Omega > \Omega_{c(bound)}$, flows have the capacity to extract particles bound to the soil surface by cohesion and inter-particle friction. Since, when this occurs, $\Omega > \Omega_{c(loose)}$, these particles are transported by the flow so that both detachment and transport processes are driven by flow energy. This is represented as FD – FT in Figure 1. No erosion occurs unless either $e > e_c$ or $\Omega > \Omega_{c(bound)}$. Rills and larger channels are generated only when $\Omega > \Omega_{c(bound)}$.

Factors Influencing the Detachment and Transport Amounts

Factors influencing RD-ST

Splash erosion is the common name given to the RD-ST system. For raindrop detachment (RD) to occur when raindrops impact soil surfaces not covered by water, the kinetic energy of an impacting raindrop (e) must exceed a critical value (Sharma and Gupta, 1989; Sharma et al., 1991). This is conceptualised through e_c in Figure 1 and the equation

$$D = k_D (e - e_c) \qquad \text{for } e > e_c \qquad (1a)$$
$$D = 0 \qquad \text{for } e \leq e_c \qquad (1b)$$

where D is the weight of soil detached by a raindrop, and k_D is the soil detachability coefficient. Considering that soils exhibit differing resistances to detachment, e_c can be expected to vary between soil materials. Loose material sitting on the surface is, in this context, essentially predetached but cohesive forces are involved in holding soil particles within the surface of the soil matrix. In addition to cohesion and inter-particle friction, soil moisture also influences the force holding particles within the soil surface (Trunam and Bradford, 1990). e_c may also increase during a rainfall event (line A, Figure 1) as a result of surface crusting.

Although the notion of soil particles being detached and transported by the impact of falling raindrops may seem simple, RD-ST involves a series of complex processes. 3 stages have been identified for drops impacting soil surfaces not covered by water layers (Terry, 1998):

1. The collision and deformation of a falling raindrop at the soil surface
2. The rupture and collapse of the drop into a thin disk of fluid spraying radially outwards from the point of impact
3. The jetting of daughter ejection droplets in parabolic trajectories away from the original drop landing position

Stages 1 and 2 are involved in the detachment process and in modifying the soil surface characteristics which influence subsequent detachment. Raindrop impact induced changes in the soil surface, such as compaction, may also influence splash trajectories and consequently, splash transport. Splash trajectories are also markedly affected by the presence of water layers on the soil surface. Splash angles between 50° and 70° occur with thin water films (Allen, 1987) and tend to become more vertical as the water depth increases. Raindrop kinetic energy is also absorbed in disturbing water films leaving less energy available for soil detachment. The effect of flow depth on reducing the amount of energy available for detachment is indicated by e_c increasing as Ω increases in Figure 1 (line B).

Small circular containers known as splash cups have been frequently used to study splash erosion. The surface areas of these cups are usually sufficiently small that all or nearly all of material splashed as a result of a drop impact can leave the confines of the cup. Consequently, experiments with splash cups provide data which can be analysed to determine values for factors such as e_c and k_D in Eq.1. However, as the size of the surface increases, a greater proportion of the material splashed by a drop impact falls back to the surface without crossing the boundary. In a large horizontal area, material splashed away from the point of impact of one drop is replaced by material splashed to that point by impacts in the surrounding area and very little of the total material detached by impacting raindrops leaves the area. Under these conditions, the transport efficiency of the system can be considered to be negligible. However, if the soil is sloping, more material is splashed in the down slope direction than the upslope direction so that the transport efficiency of the system increases as the slope gradient increases. There is however, no guarantee that splash will always move soil material downslope on sloping surfaces. Under natural conditions, wind blowing upslope can more than offset the slope gradient effect.

Conceptually, if wind is not a factor that needs to be considered, the concepts incorporated in Eq. 1 can be applied to splash erosion on large surfaces if a transport efficiency term that is dependent on slope gradient is included to give

$$S = k_{sS}\, E_x\, f[s] \qquad (2)$$

where S is the amount of material transported across the downslope boundary in an element of time, k_{sS} is the soil erodibility associated with splash erosion, E_x is the effective rainfall energy applied to the surface by the impacting raindrops during that time, and f[s] is a function of slope gradient (s). It follows from Eq.1 that, given n raindrops impacting the surface during the element of time,

$$E_x = \sum_{i=1}^{n} (e_i - e_c) \qquad \text{for } e_i > e_c \qquad (3)$$

where e_i is the kinetic energy of the ith drop. However, because the pre-detached material sitting on the surface provides some protection to the underlaying soil surface and is also splashed by a drop impact, k_s is not equal to k_D. In effect, two extreme values of k_D need to be considered. The first, $k_{D.M}$, applies when there are no predetached particles on the surface and drops are detaching only material from within the surface of the soil matrix. The second, $k_{D.PD}$, is the value of k_D that applies when the layer of predetached material is too deep for the drop to penetrate it and detach soil material from within surface of the soil matrix. In this case, only predetached material is splashed. If H_R is the degree of protection provided by the predetached material, then

$$k_{sS} = (1 - H_R)\, k_{D.M} + H_R\, k_{D.PD} \qquad (4)$$

H_R will vary in time and space. Given a large flat horizontal surface, H effectively becomes 1 over the whole surface under equilibrium conditions because the transport capacity of the RD-ST system is negligible. However, a large flat tilted surface may also exhibit a k_{sS} value equal to $k_{D.PD}$ under equilibrium conditions despite the fact that H_R is small at the upslope end of the surface. Although a drop impact at the upslope end of the surface may impact a surface that does not have predetached particles on it, any subsequent drop impact downslope of that area will impact a surface which has predetached particles sitting on it. Given that any point in space has an equal probability of being impacted, a large number of drop impacts may contribute to the layer of predetached material on the surface before a particular point on the surface is impacted. Thus, if $k_{D.PD} < k_{D.M}$, which tends to be the case since the predetached particles are held to the surface by essentially nothing more than gravity, upslope drop impacts provide increasing amounts of predetached material to downslope drop impacts as one progresses down the plane. The rate of increase in H_R down the slope depends on the efficiency of the splash transport system, the greater the efficiency, the less the change in H_R in the downslope direction. However, given a long enough slope, $H_R = 1$ will occur in the area close to the downslope boundary where the drop impacts that cause splash to pass across the downslope boundary occur. It is the value of H_R in this downslope zone that controls the rate of erosion from the area as a whole. The distance the zone extends upslope from the downslope boundary depends on splash travel distance.

As noted earlier, RD-ST is a transport limiting system. Although $k_{D.PD} < k_{D.M}$ occurs when the particle size and density characteristics of the material in the surface of the soil matrix and the predetached particles are the same, the size and density characteristics of the predetached material tends to vary in time and space. Coarse material tends to be less easily incorporated into splash and so tends to become more and more concentrated in the predetached material in time. Given sufficient time, the failure to transport coarse material may lead to an erosion pavement. Gravel paths are a prime example of this. Most cultivated agricultural soils do not contain sufficient quantities of coarse material to cause erosion pavements to occur during their productive lifetime. However, erosion pavements occur in some untilled arid and semi-arid lands (Abrahams and Parsons, 1991).

The effect of slope gradient on the net downslope transport of soil material by splash has been observed to be linear in some cases (Moeyersons and DePloey, 1976) and non-linear in others (Qunasah, 1981; Grosh and Jarrett, 1994). When the effect is modelled by

$$Q_{ns} = a\, E^b\, G^c \tag{5}$$

where Q_{ns} is the net downslope splashed material, E is the total kinetic energy of the rainfall, G is percent slope, and a, b, and c are empirically determined constants, Quansah (1981) observed c to vary from 0.7 to 1.0 for sand and loamy sand, and from 1.1 to 1.4 for silt loam, silty clay, clay loam and clay soils. Grosh and Jarrett (1994) observed a value of 2 for c with a silty clay loam. However, considering that H_R varies with size of the eroding area and with the efficiency of the transport system, the assumption that a does not vary with slope gradient may not be correct and uncertainty exists about the reliability of applying Eq. 5 outside the experimental situation used to determine the values of the empirical constants.

RD-ST may be the only detachment and transport system operating for considerable periods of time in some parts of the landscape during a rainfall event or series of rainfall events. At the onset of a rainstorm, runoff is usually absent, and RD-ST alone operates until runoff becomes effective in contributing to the erosion process. Depending on the soil and climate, considerable amounts of soil material can be detached and splashed during this time. Under these circumstances, RD-ST may be causing modification of the soil surface rather erosion *per se*. Drop impacts cause aggregate breakdown and compaction resulting in the development of surface crusts in some soils, particularly silty ones (Moss, 1991a,b). The removal of material from and compaction of microtopographic highpoints and the transport of detached material to microtopographic lowpoints reduces the roughness of the soil surface. The decline in roughness decreases exponentially with the amount of rainfall energy applied to the soil (Romkens and Wang,1987). The changes in surface properties generated by RD-ST are important to the subsequent erosion that occurs when surface water develops. The development of surface crusts encourages runoff but also produces a soil surface that has a greater resistance to detachment. Reductions in roughness also encourage runoff to occur through the reduction in the volume of the depressions which can store water on the surface. How the surface modified is by RD-ST can have a substantial impact of the subsequent development of rills on an eroding surface.

Factors influencing RD-RIFT

Raindrop induced flow transport (RIFT) occurs when flows do not have sufficient energy to entrain soil material unless raindrops impacting the flow disturb the bed underlying the flow. In this situation, the impacting raindrops lift particles up from the bed into the flow and these particles then fall in a downstream direction as they return bed. The process is analogous to wind blowing splash, the difference being that water rather than air is the fluid involved. Like the RD-ST system, the material transported across any given boundary by RIFT results from the impact of raindrops that occur within a limited distance of that boundary. Also, like the RD-ST system, the RD-RIFT system is a transport limited system and material crossing that boundary comes directly from the surface of the soil and loose material detached by previous drop impacts.

Because the material transported across a boundary by RIFT in a unit of time comes from drops impacting within a limited distance upslope, the sediment discharge for a particle of size p being transported by as a result of the impact of a drop of size d is given by

$$q_{sR}(p,d) = M_{pd} [F_d X_{Rpd}] \qquad (6)$$

where $q_{sR}(p,d)$ is the mass discharged in unit time, M_{pd} is the mass of the p sized material lifted into the flow by the drop impact, F_d is the spatially averaged impact frequency for drops of size d, and X_{Rpd} is the effective average particle travel distance. The product of F_d and X_{Rpd} gives the number of drop impacts per unit time that contribute directly to q_{sR} (p,d). X_{Rpd} is linearly influenced by flow velocity (Kinnell, 1990). As noted earlier, water depth influences the amount of the energy of the drop impact that reaches the bed. Thus flow depth influences both M_{pd} and X_{Rpd}. When flows are extremely shallow, such as at the onset of runoff, flow depth restricts the height to which particles can be lifted. Consequently, although M_{pd} is high, q_{sRpd} is restricted by X_{Rpd}. As flow depth increase, so does X_{Rpd} and q_{sRpd}. However, at some stage, impacts no longer have the capacity to lift particles up to water surface so that X_{Rpd} then declines. This decline, together with the reduction in M_{pd} that occurs as more and more of the drop impact energy is absorbed in the flow before particle uplift, causes q_{sRpd} to decline. Eventually, once flows are deep enough, drop impact becomes unable to disturb the bed under the flow. Experiments with drops travelling at or close to their terminal velocity have shown that for loose sand between 0.1 and 0.9 mm (Kinnell, 1993a),

$$q_{sR}(p,d) = a_p I_d u f[h,d] \qquad (7)$$

where a_p is a coefficient that is dependent on particle size, I_d is intensity of rain of drops of size d, u is flow velocity, and f[h,d] is a function that varies with flow depth (h) and drop size (d). This function is given by

$$f[h,d] = h \exp(-0.207h) \qquad ,h < h_c \qquad (8a)$$

$$f[h,d] = h \exp(-0.207h - b_d(h - h_c)) \qquad ,h \geq h_c \qquad (8b)$$

when h is in mm,

$$h_c = 1.017 + 4.111 \ln(d) \qquad (9)$$

$$b_d = \exp(0.585 - 0.387\, d) \qquad (10)$$

when d is in mm. Figure 2 shows how f[h,d] for 1mm to 6mm drops varies with flow depth. In very shallow flows, variations in drop size have negligible influence on the transport of the particles because flow depth restricts X_{Rpd}.

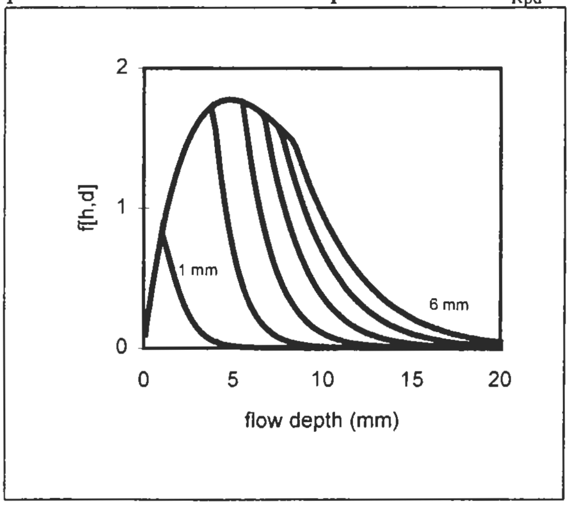

Fig. 2. The effect of drop size and flow depth on f[h.d], the flow depth-drop size function in Eq. 6

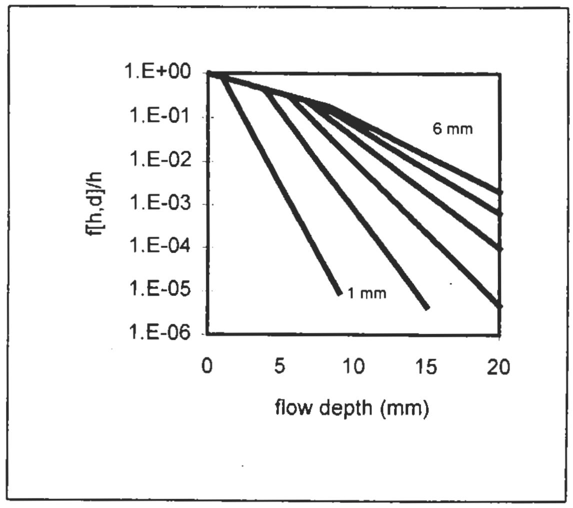

Fig. 3. The effect of drop size and flow depth on the flow depth-drop size function for sediment concentrations produced by RIFT

In some circumstances, it is convenient to model sediment discharge in term of the product of flow discharge (q_w) and sediment concentration (c_s) so that

$$q_{sR}(p,d) = q_w\, c_{sR}(p,d) \qquad (11)$$

Under these circumstances, it follows from Eq. 7 that

$$q_{sR}(p,d) = q_w\, a_p\, I_d\;\; f[h,d]/h \qquad (12)$$

and

$$c_{sR}(p,d) = a_p\, I_d\;\; f[h,d]/h \qquad (13)$$

From Eq. 8, it follows that

$$f[h,d]/h = \exp(-0.207h) \qquad ,h < h_c \qquad (14a)$$

$$f[h,d]/h = \exp(-0.207h - b_d(h - h_c)) \qquad ,h \geq h_c \qquad (14b)$$

Figure 3 shows how f[h,d]/h varies with drop size and flow depth. It follows from these equations that a_p is the exponent of "Y-axis" intercept value obtained when ln $(c_{sR}(p,d)/I_d)$ is regressed against flow depth when flow depths less than h_c are used. Under these circumstances, flow depth restricts X_{Rpd}. However, with cohesive soil material, drop energy influences the amount of material detached from the surface of the soil matrix and hence, M_{pd}. Consequently, a_p and sediment discharge resulting from RD-RIFT systems operating in very shallow flows over cohesive soil surfaces will vary with drop size and velocity.

RD-RIFT systems on natural soil under natural rain involves a range of drops of different size impacting flows over material containing a range of particles of different size. Under these circumstances, it follows from Eq. 7 that

$$q_{sR}(s,r) = k_{sR}\, I_r\, u\, f[h,r] \qquad (15)$$

and from Eq, 13,

$$c_{sR}(s,r) = k_{sR}\, I_r\, f[h,r]/h \qquad (16)$$

where s represents a soil with given soil characteristics and r represents a rain with given rain characteristics, and k_{sR} is the susceptibility of the surface to erosion by the RD-RIFT system. Given N drop sizes in the rainfall, f[h,r] is given by (Kinnell, 1993b)

$$f[h,r] = \sum_{n=1}^{N} (f[h,d]\, I_d)_n \Big/ \sum_{n=1}^{N} (I_d)_n \qquad (17)$$

However, in experiments with artificial rainfall and varying rainfall intensities on short inclined slopes like those used in ridge tillage, f[h,r] is close to 1.0 so that k'_{sR} in

$$q_{sR}(s,r) = uh\, k'_{sR}\, I_r \qquad (18)$$

lies close to k_{sR}. Because

$$q_w = uh \qquad (19)$$

Eq. 16 becomes

$$q_{sR}(s,r) = q_w\, k'_{sR}\, I_r \qquad (20)$$

Eq. 20 has been shown to apply to a number of field and laboratory experiments where raindrop size and velocity characteristics are held constant (Kinnell, 1993b, 2000).

As noted above, RIFT is a transport-limited system. Like the RD-ST system, material transported across the downstream boundary may come from both the surface of the soil matrix and predetached material sitting on the soil surface so that

$$k_{sR} = (1 - H_R)\, k_{sR.M} + H_R\, k_{sR.PD} \tag{21}$$

where $k_{sR.M}$ is the value of k_{sR} that occurs when there is no predetached material in the zone where drop impact contributes to the sediment discharge, $k_{sR.PD}$ is the value of k_{sR} that occurs when predetached material in the zone where drop impact contributes to the sediment discharge completely protects particles in the surface of the soil matrix from being detached and H_R is the degree of protection provided by the layer of predetached material. Modelling of the RD-RIFT system with a single value of X_{Rpd} and without any input from RD-ST shows that H_R increases with distance from the point where RD-RIFT begins, and with time, and decreases with X_{Rpd} and with cohesion in the surface of the soil matrix (Kinnell, 1994). However, in natural soils, a number of values of X_{Rpd} operate simultaneously so that the composition of the layer of predetached material varies in time and space in a dynamic manner depending on rain, flow and soil properties. The reason for this lies in the fact that proportion of a given sized material discharged across a boundary by RIFT is weighted by X_{Rpd}. For example, if 50 % of an eroding surface is made up of particles that have an X_{Rpd} value of 6 mm and 50 % is made up of particles that have an X_{Rpd} value of 18 mm, then, in theory, the proportion of the X_{Rpd} = 6 mm in the sediment discharged will be

$$\frac{q_{sR}(p,d)_{XRpd=6}}{q_s(p,d)_{XRpd=6} + q_s(p,d)_{XRpd=18}} = \frac{F_d\, 0.5M_{pd}\, 6}{F_d\, 0.5M_{pd}\, 6 + F_d\, 0.5M_{pd}\, 18} = 0.25 \tag{22}$$

This results in the material in the layer of predetached material coarsening in time until a steady state is reached. As a consequence of this, the composition of the sediment discharged by the RD-RIFT system also coarsens with time when the particles being transported are stable. Aggregate breakdown while particles are being transported by RIFT delays the rate of change in the composition of the sediment discharged but, in theory, the sediment discharge will be finer than the material being impacted by the raindrops irrespective of the stability of the particles involved. Consequently, since as time goes by, the effect of the layer of predetached material becomes increasingly significant, k_{sR} varies dynamically in time and space. Also, factors such as the development of surface crusts may influence how k_{sR} varies in time. Meyer and Harmon (1989) applied series of artificial storms to soil surfaces under laboratory conditions using a range of slope lengths (150 mm - 600 mm) and gradients (5 % - 30 %). The general procedure adopted was to subject the surface to a 60-min storm of about 76 mm/h on one day and then, on the following day, apply a 30-min storm at the same intensity followed by a series of 15-min storms with very low (e.g. 14 mm/h), low (e.g. 27 mm/h), medium (e.g. 76 mm/h) and high (e.g. 115 mm/h) application rates although not necessarily in that order. Kinnell (2000) observed that with many of the soils used in these experiments, k'_{sR} varied considerably during the first 60-min approx 76 mm/h storm and that it often took the additional rain produced by the 30-min storm before k'_{sR} became relatively stable. However, in some cases, k'_{sR} also varied during the 15-min storms that followed the 30-min storm.

For prediction purposes, Eq. 20 may be combined with functions describing the impact of slope length (L) and gradient (G) on q_{sR} assuming that these factors have no impact on k'_{sR};

$$q_{sR}(s,r) = q_w\, k'_{sR}\, I_r\, f[G]\, f[L] \qquad (23)$$

However, since rainfall energy level (energy per unit quantity of rain) has an impact on k'_{sR} when rain-impacted flow operates on cohesive soil material (Meyer and Harmon, 1992), Eq. 23 should rewritten to include a function describing the impact of rainfall energy level (E_r);

$$q_{sR}(s,r) = q_w\, k'_{sR}\, I_r\, f[E_r]\, f[G]\, f[L] \qquad (24)$$

It follows from Eq. 24 that

$$c_{sR}(s,r) = k'_{sR}\, I_r\, f[E_r]\, f[G]\, f[L] \qquad (25)$$

A number of experiments where slope length (L) and gradient (G) have been varied when rain-impacted flows operate on short slopes have been reported in the literature. Several equations have been proposed to describe the effect of slope gradient on short inclined slopes under rain-impacted flow:

$$f[G] = G^b \qquad \text{(Neal, 1938)} \qquad (26)$$
$$f[G] = 1.05 - 0.85 \exp(-4 \sin \theta\,) \qquad \text{(Elliott et al., 1989)} \qquad (27)$$
$$f[G] = a_s + b_s G \qquad \text{(Kinnell, 1993c)} \qquad (28)$$

where G is slope gradient (m/m), θ is slope angle (degrees) and a_s are b_s coefficients that vary between soils. While Kinnell (1993c) observed that for some soils, f[G] varied linearly with slope gradient with a positive intercept (Eq.28), he observed that for other soils f[G] increased non-linearly with slope gradient. The effects Kinnell observed can be expressed by the general equation

$$f[G] = a_s + b_s G^{c_s} \qquad (29)$$

where a_s, b_s and c_s are coefficients that vary between soils. The reasons why different soils show different forms of response have not been properly identified but must relate to cohesion, aggregate stability and the behaviour of the predetached material sitting on the soil surface. Given increasing slope gradient and length, flow energy will at some stage have the capacity to entrain predetached material by itself so that, at the upslope end of the eroding surface, RD-RIFT is dominant but, at the downslope end, RD-FT has a major impact on the discharge of sediment. Some of the non-linear responses to slope gradient observed with rain-impacted flow result from the development of RD-FT on surfaces where sediment was dominated by RD-RIFT at a lower gradient (Kinnell, 2000).

Factors influencing RD-FT

In many situations, RD-FT and RD-RIFT operate simultaneously within a rain-impacted flow, RD-FT being involved in the discharge of fine material while RD-RIFT controls the discharge of the coarser material. In most experiments with rain-impacted flows, no attempt is made to distinguish the effect of RD-FT from that of RD-RIFT. However, RD-FT systems do not behave exactly like RD-RIFT systems.For sediment discharged by a RD-FT system

$q_{sR}(p,d) = M_{pd} [F_d X_{Fpd}]$ (30)

Eq. 30 is similar to Eq. 6 in that the product of F_d and X_{Fpd} gives the number of drop impact per unit time that contribute directly to q_{sR} (p,d). However, the value of X_{Fpd} is not necessarily equal to the maximum distance particles travel unaided by drop impact after detachment. If $H_R = 1$ does not occur on the eroding surface downstream of the point of detachment, X_{Fpd} is the distance from the downstream end of the eroding surface to the furthest upstream point where the relevant sized particle is detached from the surface of the soil matrix by a drop impact and transported by the flow without aid to the downstream boundary. For clay sized material, that distance is usually the same as the length of flow on the surface. However, if $H_R = 1$ occurs downstream of the point of detachment, then X_{Fpd} is reduced by the distance travelled over that $H_R = 1$ zone. This is because RD only occurs when H does not equal 1. Also, since the material being transported does not in part come from the predetached layer sitting on the soil surface, M_{pd} results only from raindrops detaching material from the surface of the soil matrix and, as such, varies inversely with H_R.

For any given sized particle being transported by RIFT, a change from RD-RIFT to RD-FT may occur at some point along a rain-impacted flow because of the increase in flow energy in the downslope direction. Since the critical level of flow energy required to entrain predetached particles varies with particle size and density, a number of values of X_{Fpd} may operate simultaneously in a rain-impacted flow. Although the particles moving as a result of RD-FT do not become part of the predetached layer, coarser particles will still be discharged through RD-RIFT. Consequently, the X_{Fpd} values will be controlled by the distance of particle travel from the point where the change from RD-RIFT to RD-FT occurs and the presence of any predetached material that provides complete protection against raindrop detachment in the area downstream of that point. On short inclined slopes where sediment discharge is dominated by RD-RIFT, the effect of slope length on sediment concentration is negligible. However, when the dominance changes towards RD-FT, sediment concentrations increase with slope length (Figure 4).

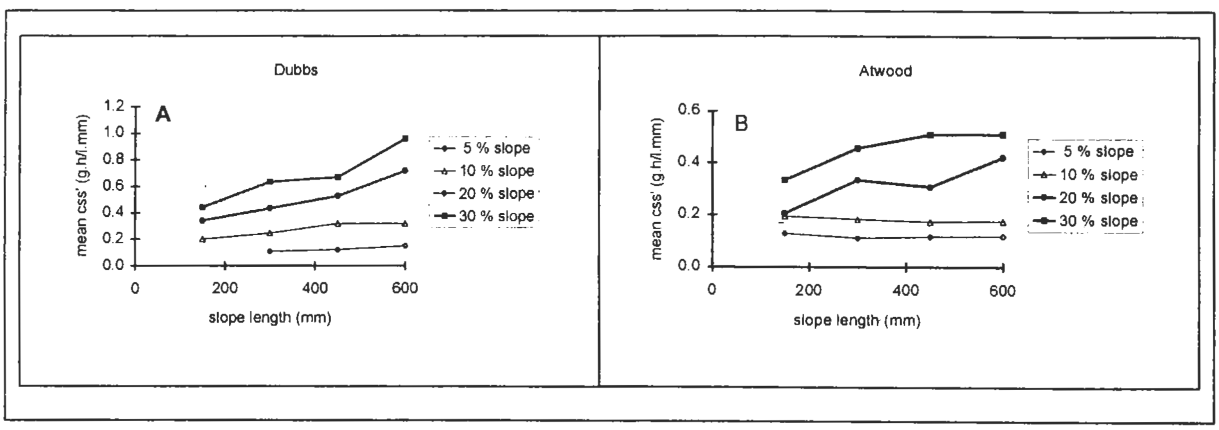

Fig. 4. The effect of slope length on c_{SS}', the sediment concentration per unit rainfall intensity, for (A) the Dubbs and (B) the Attwood soils in the experiments of Meyer and Harmon (1989) (from Kinnell, 2000). Small rills influenced sediment discharges from the Dubbs soil when slopes were 20% and more. RD-FT had a major influence on sediment discharges from the Attwood soil when slopes were 20% and more.

Factors influencing FD-FT

Flow detachment-flow transport (FD-FT) systems operate when flows exert sufficient force to detach soil particles from the soil surface of the soil matrix. Foster and Meyer (1975) proposed that the detachment or deposition rate (D_F) is related to the difference between the transport capacity (T_{cF}) and the sediment load (q_{sF});

$$D_F = \alpha\ (T_{cF} - q_{sF}) \qquad (31)$$

where α is a rate control coefficient. When stream power (Ω)is used as an indicator of flow energetics, detachment by clear water ($D_{F(clear)}$) is given by

$$D_{F(clear)} = k_{sF\omega}\ (\Omega - \Omega_{c(bound)}) \qquad (\Omega > \Omega_{c(bound)}) \qquad (32a)$$
$$D_{F(clear)} = 0 \qquad (\Omega \leq \Omega_{c(bound)}) \qquad (32b)$$

where $k_{sF\omega}$ is a soil related coefficient associated with using stream power as the independent variable in the equation. When shear stress (τ)is used as an indicator of flow energetics, detachment by clear water ($D_{F(clear)}$) is given by

$$D_{F(clear)} = k_{sF\tau}\ (\tau - \tau_{c(bound)}) \qquad (\tau > \tau_{c(bound)}) \qquad (33a)$$
$$D_{F(clear)} = 0 \qquad (\tau \leq \tau_{c(bound)}) \qquad (33b)$$

where $k_{sF\tau}$ is the soil related coefficient associated with using shear stress as the independent variable in the equation. For water transporting sediment, Eq. 33 becomes

$$D_F = k_{sF\tau}\ (\tau - \tau_{c(bound)})\ (1 - q_{sF}/T_{cF}) \qquad (\tau > \tau_{c(bound)}) \qquad (34a)$$
$$D_F = 0 \qquad (\tau \leq \tau_{c(bound)}) \qquad (34b)$$

and Eq. 32 becomes

$$D_F = k_{sF\omega}\ (\Omega - \Omega_{c(bound)})\ (1 - q_{sF}/T_{cF}) \qquad (\Omega > \Omega_{c(bound)}) \qquad (35a)$$
$$D_F = 0 \qquad (\Omega \leq \Omega_{c(bound)}) \qquad (35b)$$

The measurement of RD and FD

There are reports (eg Torri et al. 1987, Parsons et al. 1994) of the use of splashed soil material as a measure of raindrop detachment under ponded conditions. This implies that the relationship between rainsplash and raindrop detachment is not influenced by factors such as water depth. However, this is not supported by observations on the physics of drop impact. As noted earlier, splash trajectories are markedly affected by the presence of water on the soil surface. The rupture and collapse of the drop into a thin disk of fluid spraying radially outwards from the point of impact followed by the jetting of daughter ejection droplets with low trajectories observed when drops impact non-ponded soil surfaces gives way to crown like structures which produce ejection droplets with much higher trajectories as water depth increases (Macklin and Hobbs, 1969; Moss and Green, 1983). The effect of flow depth on splash trajectory influences particle travel distance and hence, the proportion of material lifted from the bed that is collected in splash traps. Further increases in water depth to beyond a critical depth result in the cavity carved by the impact of the drop in the water not reaching the underlying bed and a lack of daughter ejection droplets from crown like structures. However, the subsequent collapse of the cavity produces a vertical jet with a large daughter drop above it and the bed is disturbed when that structure collapses (Moss and Green, 1983). None of the bed material lifted by a drop impact is splashed under these conditions but some will be discharged with surface water flow. Also, in addition to the effect of flow depth on splash, using splash as a surrogate for raindrop detachment in rain impacted flow within large areas ignores the fact that splash transported material comes from not just the surface of the soil matrix but also the layer of loose particles sitting upon it. Since the proportions coming from these two sources are unknown in most experiments, the amount of detached material contained in splash when drops impact a water covered soil is uncertain. In addition, the compositions of splashed material and sediment discharged by rain-impacted flow have been observed to differ markedly (Wan and El-Swaify, 1999)

In terms of splash erosion, splash cups have been widely used to examine the capacity of raindrops to cause splash erosion and the resistances of soil materials to being eroded by raindrops impacting surfaces not covered by water. In many experiments, the materials used have been non-cohesive and, consequently, the results indicative of role of raindrop impact in the transport of predetached material. For cohesive soil materials, experiments using splash cups may provide data on raindrop detachment with respect to RD-ST on surfaces not covered by water if the cups are small enough that detached material falling back on to the eroding surface does not have any significant influence on the amounts being splashed. Although this may appear to be often the case, many experiments use rainfall intensities that, in time, lead to ponding. Irrespective of the effect of ponding on particle travel distances associated with splash, ponding results in some material lifted from the soil surface by a drop impacted travelling very short distances within the water layer. Consequently, ponding encourages the formation of the predetached layer, which interferes with detachment. Schultz et al. (1985) included detached material sitting on the soil surface in 102-mm diameter containers in their measurements of loss when ponding occurred but took no account of the effect of this material on the detachment rate.

In terms of the measurement of raindrop detachment on non-ponded cohesive soil, Al-Durrah and Bradford (1981) developed an apparatus to collect the splash produced by single drop impact under laboratory conditions. This apparatus was central to work showing that a linear relationship exists between raindrop detachment and the ratio of drop kinetic energy to soil shear strength (as measured by the fall cone method) (Al-Durrah and Bradford , 1982a) and that soil shear strength influences splash trajectories (Al-Durrah and Bradford, 1892b). Many experiments on raindrop detachment under non-ponded conditions have used splash cups , rather than the Al-Durrah and Bradford apparatus, and in many cases, cup size and the height of the cup rim relative to the eroding surface have influenced the result. Mathematical corrections have been proposed (Farrell, 1974; Torri and Poesen, 1988) to deal with these issues but the problem of not taking account of the effect of loose material on detachment rate is not considered in the approaches.

A number of studies have been devoted to determining the soil detachment rate produced by unimpacted shallow flow in the laboratory (Nearing et al., 1991; Nearing and Parker, 1991; Parker et al., 1995; Ciampalini and Torri, 1998). The objective of these experiments is to determine detachment for clear water ($D_{F(clear)}$). One technique used involves a test section of soil embedded in the bed of a flume with non-erodible areas upstream and downstream of the test section. Bearing in mind that the clear water condition only occurs within a short distance of the point of introduction of clear water in to a section of flow over soil, and that

$$D_F = D_{F(clear)} (1 - q_{sF}/T_{cF}) \qquad (36),$$

Nearing et al. (1991) used a 12.7-cm long soil surface. Ciampalini and Torri (1998) contended that a surface 10-cm long was too long and that a surface 2.5-cm long was more appropriate. One of the major problems with the technique is that results are influenced by the development of a discontinuity at the upstream end of the test section as soil is eroded during the experiment (Ciampalini and Torri, 1998). Given a capacity to determine the transport capacity of the flow (T_{cF}), an alternative approach is to use surfaces, which are longer and use Eq. 37 to determine $D_{F(clear)}$. This approach is has been used in field experiments with flow in furrows produced by ridge-furrows systems and theoretically derived transport capacities (Elliott et al. 1989). Uncertainty about theoretically derived transport capacities results in uncertainty in the detachment values determined by this approach. Huang et al. (1996) challenge the approached based on Eq. 36 and support an earlier model concept, proposed by Meyer and Wischmeier (1969), in which detachment and transport processes are modelled separately and sediment delivery is limited by the lesser of the two. However, the Meyer-Wischmeier concept was designed to deal with an erosion system which included soil detachment from raindrop impact and there is nothing about the Meyer -Wischmeier concept that does not allow for an effect of transported material on detachment by flow to be considered.

Process-based Approaches to Modelling Rainfall Erosion

The Rose and his associates

Rose et al. (1983) proposed a mathematical framework for modelling erosion by RD-RIFT, RD-FT and FD-FT based on the concept that mass conservation of sediment of size class i requires that

$$\frac{\partial}{\partial x}(q_w c_i) + \frac{\partial}{\partial t}(h c_i) = r_i - d_i + f_i \qquad (37)$$

where h is flow depth, r is the rate of rainfall detachment, d is the sediment deposition rate and f is the sediment entrainment rate. For bare soil surfaces, they proposed that

$$r_i = a C_r I^p / N \qquad (38)$$

where a is a soil related coefficient, C_r is the fraction of the soil that is unprotected, p is an exponent originally thought to be close to 2 but now considered to be approximately 1 (Proffitt et al., 1991), N is the number of sediment size classes being considered. N appears in Eq. 38 as a result of the stipulation that $c_i = c/N$. In terms of Eq. 37,

$$d_i = \alpha_i\, v_{pi}\, c_i \tag{39}$$

where v_{pi} is the mean settling velocity in water of particles in size class i, $\alpha_i\, c_i$ is the sediment concentration close to the bed, and c_i is the depth averaged sediment concentration (Hairsine and Rose, 1991). According to Rose et al. (1983), the steady state solution for c_i in the ordinary differential equations that result from Eqs. 37 to 39 and $f_i = 0$ is

$$c_i = \frac{a\, C_r\, I^p}{N(q'_w + v_{pi})} \tag{40}$$

where q'_w is the water discharged per unit area. The effect of q'_w in Eq. 40 is not in anyway associated with the effect of flow depth on sediment concentration considered earlier in the section on factors influencing RD-RIFT. That is dealt with through the term a (Rose and Hairsine, 1988; Hairsine and Rose, 1991). It simply results from the solution to the differential equations that result from Eqs. 37 to 39 and, according to Rose et al., the effect of q'_w is frequently extremely small because often $v_i >> q'_w$. Although that may be so, it is contrary to the observations that when flow depth is held constant, sediment concentrations associated with RIFT are not influenced by flow discharge because there is a direct relationship between q_{sR} and flow velocity (Kinnell, 1988).

Recognition of the role of predetached material in the discharge of sediment is included through

$$a = (1 - H_R)\, a_M + H_R\, a_{PD} \tag{41}$$

where a_{PD} is the value of a when the layer of predetached material completely protects the underlying surface against detachment (H_R = 1) and a_M is the value of a when there is no predetached material on the soil surface (H_R = 0). According to Hairsine et al. (1999), for shallow rain-impacted flows where $f_i = 0$ and $\alpha = 1$, the time varying solutions to

$$\frac{\partial}{\partial x}(q_w c_i) + \frac{\partial}{\partial t}(h\, c_i) = \frac{a_M\, I}{N}(1 - H_R) + a_{PD}\, H_R - v_i c_i \tag{42}$$

such as those developed by Sander et al. (1996) are both complex and computationally demanding. Steady state solutions, such as those developed by Hairsine and Rose (1991), are less demanding. Although the dynamic nature of the predetached layer is a factor, to some large degree, the complexity in applying the Rose et al. approach to unsteady conditions lies in the involvement of sediment concentration in calculation of the deposition rate since the sediment concentration is a net effect of particle uplift and deposition. The approach adopted by Kinnell (1994) based on X_{Rpd} values and H_R for an element being given by

$$H_R = q_{sR\text{-}in} / q_{sR\text{-}inH1} \qquad q_{sR.in} < q_{sR.inH1} \tag{43a}$$
$$H_R = 1 \qquad q_{sR.in} \geq q_{sR.inH1} \tag{43b}$$

where $q_{sR\text{-}inH1}$ is the value of discharge of sediment in to the element ($q_{sR\text{-}in}$) required to produce H_R = 1 provides a less complex and less computationally demanding approach to modelling

RD-RIFT under unsteady conditions. In this approach, z_{PDp}, the depth of p sized material added to the layer of predetached material is given by

$$z_{PDp} = q_{sRp\text{-}in} / (\rho_p X_{Rpd}) \tag{44}$$

where $q_{sRp\text{-}in}$ is the sediment being discharged across the upslope boundary of a rain-impacted element flow of length ∂x, and ρ_p is particle density. Figure 5 illustrates this assuming that X_{Rpd} varies directly with flow velocity (u) and the settling velocity of the particle in water (v_p). Given knowledge of the sediment input giving rise to the $H_R = 1$ condition enables z_{PD} for the $H_R = 1$ condition to be determined and compared with z_{PD} value obtained for sediment actually entering the element to determine the value of H_R for the element. The Hairsine et al. (1999) approach to determining H_R,

$$H_R = m_{PDt}/m_{PDt.H1} \tag{45}$$

where m_{PDt} is the mass of predetached material per unit area of the bed and $m_{PDt.H1}$ is the mass of predetached material per unit area of the bed when $H_R = 1$, is similar in effect but does not consider particle density.

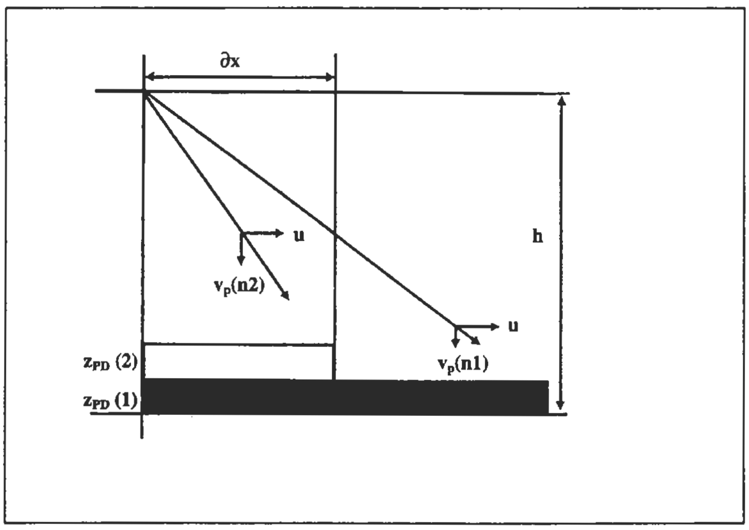

Fig. 5. Schematic diagram illustrating the effect o f travel during deposition on depth of predetached material (zPD) for two non-suspended load materials (n1, n2) of different size or density. (From Kinnell, 1994)

When Eqs.15, 20 and 21 are combined to give

$$q_{sR}(s,r) = ((1 - H_R)\, k'_{sR.M} + H_R\, k'_{sR.PD})\, I_r\, u\, f[h,r] \tag{46}$$

and H_R is determined thought the z_{PD} approach, a relatively simple finite difference model of RD-RIFT can be produced that does not require any consideration of sediment concentration (Kinnell, 1994) because deposition is being considered as a two dimensional rather than a one dimensional process. This enables unsteady conditions to be modelled in a less complex and less computationally demanding way than considered by Sander et al. (1996) and Hairsine et al. (1999).

The direct effect of flow on entraining soil material from the surface of the soil matrix in the Rose et al. approach is considered in terms of excess stream power (Bagnold, 1977),

$$f_{Mi} = (1 - H_F)\frac{\beta}{NJ}(\Omega - \Omega_0) \qquad (\Omega > \Omega_0) \qquad (47a)$$

$$f_{Mi} = 0 \qquad (\Omega \leq \Omega_0) \qquad (47b)$$

where J is the amount of work expended per unit mass of soil in the entrainment process, β is the fraction of the excess stream power $(\Omega - \Omega_0)$ available for the entrainment process, and H_F is the degree of protection provided by the predetached material against detachment from the surface of the soil matrix (Rose and Hairsine, 1988). In the Rose et al. approach, no distinction is made between H_R and H_F despite the fact that the protective effect of predetached material is likely to differ between raindrop-driven erosion and flow-driven erosion. With respect to entraining soil material from the layer of predetached material,

$$f_{PDi} = \frac{H_F\, \beta\, \alpha_i - (\Omega - \Omega_0)\, m_{PDi}}{g\, (\rho_p - \rho)\, h\, m_{PD}} \qquad (\Omega > \Omega_0) \qquad (48a)$$

$$f_{PDi} = 0 \qquad (\Omega \leq \Omega_0) \qquad (48b)$$

where ρ_p is particle density, ρ is the density of water, m_{PDi} is the mass of i sized particles in the predetached layer and m_{PD} is the mass of the predetached layer (Rose and Hairsine, 1988).

$$f_i = f_{Mi} + f_{PDi} \qquad (49)$$

Eq. 47 is similar to Eq. 35 in that it involves an excess stream power term, and a soil dependent term, but differs through the use of H_F rather q_{sF}/T_{cF}. In the RD-RIFT system, H_R results from the fact that RIFT is a transport limiting system and a clearly defined layer of predetached material is visible on the soil surface. In a FD-FT system, the uplift, deposition followed by uplift sequence involved in the transport process results in predetached particles sitting on the surface of the soil matrix briefly and providing a degree of protection to particles within that surface even though the layer of predetached material is not as distinct as in a RD-RIFT system unless net deposition is occurring. $q_{sF} = T_{cF}$ when $H_F = 1$ in a similar way to $q_{sR} = T_{cR}$ when $H_R = 1$. However, if $\Omega_0 = \Omega_{c(bound)}$, the use of the same value of Ω_0 in Eq. 47 as the threshold stream power required to entrain material predetached material is questionable because threshold stream power required to entrain material predetached material is likely to differ from that required to entrain material from the surface of the soil matrix. Rose and Hairsine (1988) do not consider this issue. They asserted that, in many cases, $\Omega >> \Omega_0$ and ignored Ω_0 in obtaining a solution to

$$\frac{\partial}{\partial x}(q_w c) = f_M + f_{PD} - d \qquad (50)$$

to provide an expression for c under equilibrium conditions. Also, Rose and Hairsine gave no particular attention to the case where RD-FT operates when RD-RIFT also occurs in the same area. Under these circumstances, predetached material resulting from RIFT provides the surface upon which material being transported by the RD-FT system is stored briefly. Consequently, there are three layers involved, the soil matrix, the layer of predetached material associated with RIFT and the layer of predetached material associated with FT. If RD-RIFT follows a period of RD-RIFT then the layer associated with RIFT will be coarser than before RD-FT occurred. If flow energy increases, a period of FD-FT may then follow with FD operating on the layer of predetached material associated with RIFT. In terms of Eq. 50, the issues related to f_M in these circumstances are not those associated with the original soil but those associated with the RIFT associated layer. FD may then cease because (a) coarse material remaining in the RIFT associated layer prevents FD from occurring or (b) the forces holding the particles within the soil surface are sufficient to prevent FD once the RIFT associated layer has been removed.

According to Rose (1988), the sediment concentration at the end of a plane of length L at time t is given by (Rose, 1988)

$$c(L,t) = (a\, C_r\, I/q'_w\, N) \sum_{i-1}^{N} (1/\gamma_i) + \rho\, g\, s\, K\, C_e\, (1 - x^*/L) \qquad , (L > x^*) \qquad (51)$$

where $\gamma = 1 + v_i/q'_w$, C_e is the fraction of the soil surface unprotected from entrainment by flow, and x* is the distance down the plane beyond which entrainment by flow takes place, but the larger the runoff event and the better aggregated the soil, the smaller the term associated with raindrop impact driven erosion. Under these circumstances, Rose suggested that that term be neglected to give

$$\begin{aligned} c(L,t) &= \rho\, g\, s\, K\, C_e\, (1 - x^*/L) \\ &= 2700\, s\, \eta\, C_e (1 - x^*/L) \qquad , (L > x^*) \qquad (52) \end{aligned}$$

where η is the efficiency of net entrainment by overland flow. The sediment concentration in Eq. 52 is a function of time only because x* is time-dependent through

$$x^* = \Omega_0 / (\rho\, g\, s\, q'_w) \qquad (53)$$

Misra and Rose (1996) provide theory and parameter sensitivity when the Rose et al. approach for raindrop-driven erosion (Rose et al., 1983; Hairsine and Rose, 1991) and flow-driven erosion (Rose et al., 1983, Hairsine and Rose, 1992a, b) are modelled under steady-state conditions. For erosion on a plane without rills,

$$\frac{dc}{dq_w} = \frac{(1 - H_R)\, aP}{q_w\, q'_w} + \frac{(1 - H_F)\, \beta\, (\Omega - \Omega_0)}{J\, q_w\, q'_w} - \frac{c}{q} \qquad (54)$$

with $H_R = H_F$ applies (Misra and Rose, 1996). For rills, Eq. 54 is modified to take account of rill geometry and sediment inflow from interrill areas. Numerical solutions to these equations were provided by Runge-Kutta methods (Gerald and Wheatly, 1984). However, the form of sediment concentration to stream power relationship at any value of J can be closely approximated by a curve obtained by raising the sediment concentration at the transport limit (c_t) to a power (χ) less than one (Rose, 1993). GUEPS (Griffith University Erosion Prediction System) uses this approach to predict soil loss from sloping lands (Yu et al., 1999). According to Yu et al. (1999), the soil loss for a single event ($M_{e,}$) is given by

$$M_e = \frac{0.01\, q'_{w(tot)}\, a\, I_{10}}{3.6 \times 10^6\, \phi} + 0.01\, q'_{w(tot)}\, \lambda\, {c^*_t}^{\chi^*} \qquad \text{(t/ha)} \qquad (55)$$

where $q'_{w(tot)}$ (mm) is the runoff amount per unit area for the event, I_{10} is the peak 10-min rainfall intensity, ϕ is the mean settling velocity of the sediment determined from

$$\phi = \sum_{i=1}^{N} \frac{v_{pi}}{N} \qquad (56),$$

λ is a binary number having a value of 1 when flow driven erosion is considered and 0 when it is not, and c^*_t is the event sediment concentration (event soil loss divided by event runoff) at the transport limit. ϕ is considered to provide a measure of the depositability of the sediment. For erosion driven by flow only, $a = 0$, $\lambda = 1$.The sediment concentration at the transport limit can be calculated with GUEPS given data on runoff rates and sediment characteristics. The approach is based on (Rose, 1993)

$$c_t = \frac{\beta\, \rho\, \rho_p}{\phi\, (\rho_p - \rho)}\, G\, u \qquad (57)$$

with $\beta = 0.1$.

$$\chi^* = \frac{\ln c^*}{\ln c^*_t} \qquad (58)$$

where c^* is the observed event sediment concentration. In some circumstances, χ^* increases with I_{10} while in others, χ^* decreases with peak runoff rate (Yu et al., 1999). According to Rose (1993), given data on sediment concentrations from short slopes on low slope gradients (G), a_{PD} can be determined from the approximation

$$c = \frac{H_R\, a_{PD}\, I}{\phi} \qquad (59)$$

when $H_R = 0.9$, and a_M can be determined using

$$a_M = \frac{q'_w}{\dfrac{I}{c} - \dfrac{\phi}{a_{PD}}} \qquad (60)$$

However, the approximate value of 0.9 for H under steady state conditions is speculative since it was estimated from "observations" of the soil surface after runoff had ceased in experiments reported by Proffitt et al. (1991). The true value may differ markedly from 0.9 and almost certainly varies with flow depth (since a proportion of drop energy is absorbed by the surface water) and the composition of the sediment.

WEPP

WEPP is a computer model for predicting soil erosion and sediment delivery from fields, farms, forests, rangelands, construction sites and urban areas (Laflen et al., 1997). Sediment delivery is modelled in WEPP with channels and impoundments on or leading from fields. It is a daily simulation model with the hydrologic status of the land computed on a daily basis. Plant growth and its contribution to factors influencing erosion is also modelled. A weather generator is used to provide the rainfall input.

The basic equation used in the WEPP erosion component is a steady state sediment continuity equation (Flanagan and Nearing, 1995)

$$\frac{dq_s}{dx} = D_i + D_r \qquad (61)$$

where D_i is the interrill delivery rate (mass per unit area per unit time) to the rills and D_r is the rill detachment or deposition rate (mass per unit area per unit time). Originally, the interrill delivery rate was modelled using

$$D_i = K_i\, I_e^2\; C_g\, C_c\, S_f \qquad (62)$$

where K_i is the interrill soil erodibility, I_e is the effective rainfall intensity during the period of rainfall excess, C_g is the ground cover effect adjustment factor, C_c is the canopy cover effect adjustment factor and S_f is a function of the interrill slope:

$$S_f = 1.05 - 0.85 \exp(-4 \sin \theta) \quad (63)$$

where θ is the interrill slope angle. K_i values for 18 cropland soils in the western half of the USA were determined experimentally using the I_e^2 based approach with data obtained from experiments with interrill plots under artificial rainfall (Elliott et al., 1989; Liebenow et al., 1990). However, Eq. 62 does not account for the effect of variations in interrill runoff and consequently, Eq. 62 was replaced by

$$D_i = K_i\, I_e\, I_x\, C_g\, C_c\, S_f \quad (64)$$

in the 1995 version of the model (Flanagan and Nearing, 1995). In Eq. 64, the product of I_e and I_x, the excess rainfall rate, replaces I^2. Since I_x and q_w are directly related to each other, Eq 64 is consistent with Eq.20. However, the interrill delivery function goes beyond Eq.64 to involve consideration of sediment characteristics through a term known as the interrill sediment delivery ratio (SDR_{RR}). SDD_{RR} is computed as a function of random roughness of the soil surface, the fall velocity of the individual particle size classes, and the particle size distribution of the sediment.

WEPP uses separate equations to simulate rill detachment and deposition. Rill detachment is predicted to occur when a soil dependent critical shear stress (τ_0) is exceeded and the sediment transport capacity (T_{cF}) exceeds the sediment load (q_s):

$$D_r = K_r (\tau - \tau_0)(1 - q_s/T_{cF}) \quad (65)$$

where K_r is the rill soil erodibility. In WEPP, the sediment transport capacity is predicted by

$$T_{cF} = k_t\, \tau^{1.5} \quad (66)$$

where k_t is the transport coefficient obtained by applying the Yalin equation (Yalin, 1963) at the end of the slope profile (Finkner et al., 1989). τ_0 and K_r vary between soils and can be evaluated through experiments where sediment discharge data is collected from furrows where flow shear stress is varied over a range in a controlled manner (Elliott et al., 1989). When net deposition occurs , i.e when $q_s > T_{cF}$

$$D_r = (0.5\, v_f/q_w)(T_{cF} - q_s) \qquad , q_s > T_{cF} \quad (67)$$

where v_f is the effective settling velocity of the sediment.

Comparison between WEPP and Rose et al. approaches to modelling erosion processes

In WEPP, interrill erosion is dominated by raindrop detachment. It follows from Eq. 64 that if the slope gradient effect on sediment discharged considered in WEPP (S_f) is appropriate, sediment concentrations for rain-impacted flow on bare soil areas are described in WEPP by

$$c_R = K_i\, I_e\, S_f \quad (68)$$

while, from Eq. 59, the Rose et al. approach yields

$$c_R = a\, I\, S_f / \phi \tag{69}$$

as an approximation. The rainfall intensity parameters used in the two equations (I_e and I) are effectively the same but the soil effect in WEPP is dealt with through a single empirical coefficient (K_i) while, in Rose et al. approach, it is dealt with through a measured parameter (ϕ, the mean settling velocity of the sediment) and an empirical coefficient (a) which varies between two extremes according to Eq. 41 ($a = (1 - H_R)\, a_M + H_R\, a_{PD}$)).

As noted earlier, rainfall simulator experiments were conducted on some 18 cropland soils in the USA to determine WEPP soil erodibility parameters (Elliott et al. 1989; Liebenow et al., 1990). In these experiments, interrill plots were subjected to about 1 hour of rain at 63 mm/h and the data during the latter part of the rainfall event used to determine K_i values for these soils when the I_e^2 approach to modelling interrill erosion (Eq. 62) was used. Kinnell (1993b) determined the corresponding interrill erodibilities using the flow discharge-rainfall intensity approach (Eq. 20). In the experiments, 4 of the interrill plots had slope gradients of about 50-60% and were referred to as 'ridged' plots, 2 had slope gradients of 3-8% and were referred to as 'flat' plots. The erosion from the flat plots was dominated by RD-RIFT. Since sediment size data was collected during the experiments, an estimate of ϕ could be obtained so that values of a could be estimated for the 18 soils but, since a/ϕ is as empirical as K_i, there appears little practical gain in doing so. Rose (1993) suggests that a major deficiency in representing detachment processes between rills by Eq. 62 is the inability of the equation to indicate that the sediment coming from interrill areas is finer than that coming from the rills. This is an issue important to estimating chemical transport associated with sediment. However, ϕ provides insufficient information about the sediment to be of use in this context. When it comes to predicting chemical transport , one needs to resort to more detailed modelling of the movement of the various sized particles such as provided by Hairsine et al. (1999). WEPP does this in a simple way through the interrill sediment delivery ratio (SDR_{RR}) which considers 5 particle size classes (primary clay, primary silt, primary sand, small aggregate and large aggregate).

The approach to modelling the discharge of sediment associated with flow detachment and transport in Eq. 55 is based on

$$c^*_F = c^{*\,\chi^*}_t \tag{70}$$

where c^*_F is the sediment associated with flow detachment and transport during and erosion event, c^*_t is calculated from the data obtained from applying Eq. 57 during the erosion event and χ^* is determined from Eq. 58. Since raindrop detachment contributes to the event sediment concentration (c^*), the link between χ and J is not well maintained if raindrop detachment is a major contributor to the sediment discharged. Also, as in WEPP, c_F is, through χ, dependent on the ratio of the sediment being carried by the flow to the sediment carried by the flow when transport-limiting conditions occur. In fact

$$c^{*\,\chi^*}_t = c^*_t\, (c^*/c^*_t) \tag{71}$$

so that there is a less complex mathematical approach to determining c^*_F than adopted by Yu et al. (1999). Since J acts as a coefficient in the relationship between f_{Mi} and effective stream power, it is not surprising that it follows from Eq. 71 that part of the soil effect in the Rose et al. approach can be represented by an empirical coefficient ($K_{F.\chi}$) in the equation

$$q_{sF} = K_{F.\chi} \frac{\beta \rho \rho_p}{\phi (\rho_p - \rho)} G u \qquad (72)$$

when Ω is assumed to be zero.

Modelling Rill Initiation and Growth

Conceptually, the presence or absence of rills has a major impact on rainfall erosion. The WEPP interrill delivery function includes, in addition to the parameters considered in Eq. 64, rill spacing and rill width. These values for these parameters can be readily determined in some forms of cultivation, for example ridge-tillage systems, but are not in others. Cultivation in broadacre small-grain agriculture eliminates existing rills and promotes sheet erosion until rill networks develop in a somewhat self-organised way. Lewis et al. (1994a,b) developed the model PRORIL to provide deterministic estimates of interrill erosion with stochastic estimates of rill erosion. The rill networks in PRORIL are based on probability distributions of the number of rills and the flow rates in rills. The development of probability density functions for the number of rills and flow rates in rills requires experimental data such as that obtained from observing rill networks obtained in experiments with field plots under artificial rainfall (Lewis et al., 1994c). A more dynamic approach has been adopted in the RillGrow model (Favis-Mortlock, 1998). RillGrow is still under development (version 2 is current) and applies simple rules at the millimetre scale to govern the interaction between microtopography, runoff routing and soil loss. The model applies simple rules to the movement of individual runoff "packets" on a grid of microtopographic heights. The model does not discriminate between rill and interrill areas and uses the sediment load – stream power relationship derived experimentally by Nearing et al. (1997) for flow in actively eroding rills to model transport even though its applicability to interrill areas is unproven. Detachment is modelled using the probabilistic approach of Nearing (1991). Deposition is not modelled in versions 1 and 2 but the model is still being refined. Despite the fact that, at this time, RillGrow does not comprehensively consider all the erosion processes at the millimetre scale, the approach provides a reasonable representation of rill systems development under artificial rainfall in laboratory experiments (Favis-Mortlock, pers. comm.).

EUROSEM

The European Soil Erosion Model (EUROSEM) is designed to simulate sediment transport, erosion and deposition over the land surface by rill and interrill processes in both fields and small catchments (Morgan et al., 1998). In contrast to WEPP, which is a continuous simulation model while EUROSEM is directed towards modelling single events. Morgan et al. claim that because of this, the EUROSEM approach simulates the dynamic behaviour of events within a storm and is more compatible with the equations used in process-based modelling of erosion mechanics.

The simulation of erosion in EUROSEM is linked to a water and sediment routing structure such as that provided in KINEROS (Woolhiser et al., 1990). The KINEROS model represents the land surface in a catchment as a series of interlinked uniform sloping planes and channel elements. Soil loss is computed through determination of the sediment discharged passing a given point in a given time. The computation is based on the mass balance equation

$$\frac{\partial (AC)}{\partial t} + \frac{\partial (QC)}{\partial x} - DR - DF = q_s(x,t) \qquad (73)$$

where C= sediment concentration, A = cross-sectional area of the flow, Q = flow discharge, DR = rate of particle detachment by raindrop impact, DF = the net rate of particle detachment by flow (positive for detachment, negative for deposition), q_s = external input or extraction of sediment per unit length of flow, x = horizontal distance and t = time. Traditional concepts of rill and interrill processes are not adopted in EUROSEM. Instead, raindrop and flow processes are modelled on all areas with the distinction between rill and interrill areas being one of geometry. Rills are described as trapezoidal channels, interrill areas as surfaces without orientated roughness. If rills are present, EUROSEM assumes interrill areas to slope towards the rills rather than straight downslope. Detachment by rainfall is modelled using

$$DR = \frac{k}{\rho_s} KE\, e^{-zh} \qquad (74)$$

where k = an index of the detachability of the soil for which values must be obtained experimentally, ρ_s = particle density, KE = the kinetic energy of the raindrops impacting the ground surface, z = an exponent varying with soil texture and h = the mean depth of the water layer. Detachment by flow is modelled by

$$DF = \beta\, w\, v_s\, (TC - C) \qquad (75)$$

where β = flow detachment efficiency factor, w = flow width, v_s =particle settling velocity, and TC is the transport capacity of the flow. TC values for rill flow are calculated from relationships developed for the work of Govers (1990) and for interrill flow from the work of Everaert (1991). With detachment by raindrop impact and flow considered to operate together in the same space, changes in dominance of DR and DF occur with the stream power of the flow, because this determines TC in rill flow, and flow depth because DR decreases with flow depth.

Like most so-called process-based models, the process equations used in EUROSEM can be seen to inadequately represent the processes involved in some cases. For example, the transport capacity of interrill flow is determined without direct consideration of rainfall characteristics when rainfall characteristics are known to influence interrill flow transport capacity (Kinnell, 1990). Also, the coefficients used in Eq. 74 result from measurement of soil material transported by splash under ponded conditions in the experiments of Torri et al. (1987). As noted in the section on the measurement of RD and FD, there is a water depth dependent relationship between splash and RD. Consequently, the approach used in EUROSEM underestimates the amount of material detached by raindrops impacting the soil under the water layer. The resulting bias towards DF means that EUROSEM almost certainly under represents the influence of raindrop impact in the areas where raindrop impact is, in reality, an important contributor to erosion.

The Hillslope Erosion Model

The Hillslope Erosion Model developed by Lane and associates (Shirley and Lane (1978); Lane et al. 1988, 1995) is based on the kinematic wave equations for a plane and the sediment continuity equation

$$\frac{\partial(ch)}{\partial t} + \frac{\partial(cq)}{\partial x} = e_i + e_r \qquad (76)$$

where c is the total sediment concentration resulting from raindrop impact and surface flow, h is flow depth, q is the runoff rate, e_i is the interrill (raindrop impact dominated) erosion rate and e_r is the rill (flow dominated) erosion rate. Lane and associates consider that the transport capacity of shallow flow is simply related to the rainfall excess (R_x) and they assume that e_i is directly related to this transport capacity. Consequently they assume that

$$e_i = b\,R_x \qquad (77)$$

where b is a coefficient. Also, the rate of detachment by flow is assumed depend simply on the difference between the transport capacity (T_c) and the transport rate. These assumptions, and the assumption that $T_c = Bh^{3/2}$, lead to the equation

$$\frac{\partial(ch)}{\partial t} + \frac{\partial(cq)}{\partial x} = K_i R + K_r\,((B/K)q - cq) \qquad (78)$$

where $K = CS^{1/2}$. Here C is the Chezy coefficient for turbulent flow and S is slope gradient. The initial boundary conditions are

$$c(0,x) = c(t,0) = K_i \qquad (79)$$

C_f, the limit of c(t,x) as t approaches infinity at any given x, is given by

$$C_f = B/K + (K_i - B/K)\exp(-K_r x) \qquad (80)$$

The Hillslope Erosion Model is essentially a simplification of the WEPP model. The simplification ignores the effect of shear stress on detachment by flow and ignores the effect of rainfall intensity on interrill erosion.

Empirical and Conceptual Models

The Universal Soil Loss Equation

Originally developed by the US Department of Agriculture in the 1960s, the Universal Soil Loss Equation (Wischmeier and Smith, 1978)

$$A = R\ K\ L\ S\ C\ P \qquad (81)$$

where A = the annual average soil loss, R = the rainfall-runoff (erosivity) factor, K is the soil (erodibility) factor, L = slope length factor, S = slope gradient factor, C = crop and crop management factor, and P = conservation practice factor, in the most widely known and applied erosion prediction equation. The R factor is the average annual sum of the rainfall event EI_{30} values, EI_{30} being the product of event rainfall kinetic energy (E), and I_{30} twice the maximum amount of rain falling a 30-min period during the event. Originally, K was determined experimentally by dividing the average annual soil loss from a 22.13-m long, bare fallow plot on a 9% slope by the R factor. Data obtained this way led to the development of an equation for soils where silt + very fine sand is 70% and less (Wischmeier et al., 1971). For K in SI units, the equation is

$$K = 2.77\ M^{1.14}\ (10^{-7})\ (12\text{-OM}) + 4.28\ (10^{-3})(SS\text{-}2) + 3.29(10^{-3})\ (PP\text{-}3) \qquad (82)$$

where M = (% silt + % very fine sand) (100 - % clay). OM = % organic matter, SS = soil structure code, and PP profile permeability class. Other equations have been developed for volcanic soils (El-Swaify and Dangler, 1976) and Australian soils (Loch et al., 1998).

L, S, C and P are ratios of the annual average soil losses from plots where conditions differ from the control conditions (slope length 22-13 m, gradient 9%, bare fallow and cultivation up and down the slope) to the average annual soil loss from plots where those control conditions occur. Thus

$$L = (\lambda / 22.13)^m \qquad (83)$$

m varies with slope gradient. In the USLE, m = 0.6 when the slope is > 10%, 0.5 when the slope is 5 - 10%, 0.4 when 3 - 5%, 0.3 when 1 - 3%, and 0.2 when < 1%. The USLE was revised in the 1990s and in the revised version (RUSLE)

$$m = \beta / (1 + \beta) \qquad (84)$$

where β is the ratio of rill to interrill erosion. For soils moderately susceptible to rill erosion, β is given by

$$\beta = (\sin\theta / 0.0896) / [3.0\ (\sin\theta)^{0.8} + 0.56] \qquad (85)$$

where θ = angle to horizontal (Renard et al., 1997).

In the USLE,

$$S = 65.4 \sin^2 \theta + 4.56 \sin \theta + 0.654 \qquad (86)$$

where as in the RUSLE,

$$S = 10.8 \sin \theta + 0.03 \qquad \text{,slopes} < 9\% \qquad (87a)$$
$$S = 16.8 \sin \theta - 0.50 \qquad \text{,slopes} \geq 9\% \qquad (87b)$$

Originally, C values were obtained by direct comparison of the soil losses from cropped plots and bare fallow plots on the same soils. However, the cost in time and resources of doing these experiments on all crops and all soils in any country is prohibitive. Consequently, more conceptual means of obtaining C values were developed. The approach adopted involves multiplying short term values of C by short term values of R to take account of the temporal interaction between vegetation and erosive rain during the year. These short term values of C are referred to as soil loss ratios and they depend on prior land use, canopy cover, surface cover and roughness. The P factor was originally used to demonstrate the advantage of changing from cultivation up and down the slope to across the slope. In the RUSLE, P depends on ridge height and furrow grade. Strip cropping, buffer strips, filter strips and subsurface drains have impacts on P values.

The USLE was developed as a tool to help in making management decisions and has been criticised because it is seen an empirical rather than process-based model. The problem with so called process-based models is that often they are require data that is difficult to obtain and this is one of the reasons why the USLE remains, particularly with recent and ongoing revision, a valuable tool.

The USLE-M

The USLE was developed to predict the average annual soil loss. While the long term soil loss is the sum of many individual events, the USLE was not designed to predict erosion produced by individual events and in not well suited to doing so. Figure 6A shows the relationship between event soil losses and EI_{30} values at a site in the USA that was part of the USLE data base. As can be seen from this figure, the EI_{30} index does not perform well with events that produce small soil losses. Foster et al. (1982) observed that using an erosivity index that includes runoff terms with rainfall terms in a product reduced large overestimates of soil loss when runoff is negligible and rainfall amount and rates are great. However, little attention was given to this until more recently when Kinnell and Risse (1998) showed (Figure 6B) the advantage of multiplying the EI_{30} index by the runoff ratio (Q_R). This advantage increases as infiltration capacity of the soil increases (Figure 7). The USLE-M is the name given to the version of USLE that is based on the product of EI_{30} index and Q_R.

While the Q_REI_{30} index is an empirical factor, it has some physical basis. The Q_REI_{30} index is based on the assumption that the maximum 30-minute intensity (I_{30}) parameter used in the USLE provides a useful measure of the impact of the peak rainfall intensity on the sediment concentration for an event, and the assumption that the effect of raindrop energy can be accounted for by the average kinetic energy per unit quantity of rain which is obtained by dividing the total kinetic energy for the rainfall event (E) by the rainfall amount (B_e). The $Q_R EI_{30}$ index is a short form of

$R_{e.UM} = Q_e I_{30} E/B_e$ (88)

that results from the runoff ratio (Q_R) being given by

$Q_R = Q_e/B_e$ (89).

where Q_e is the runoff for the event.

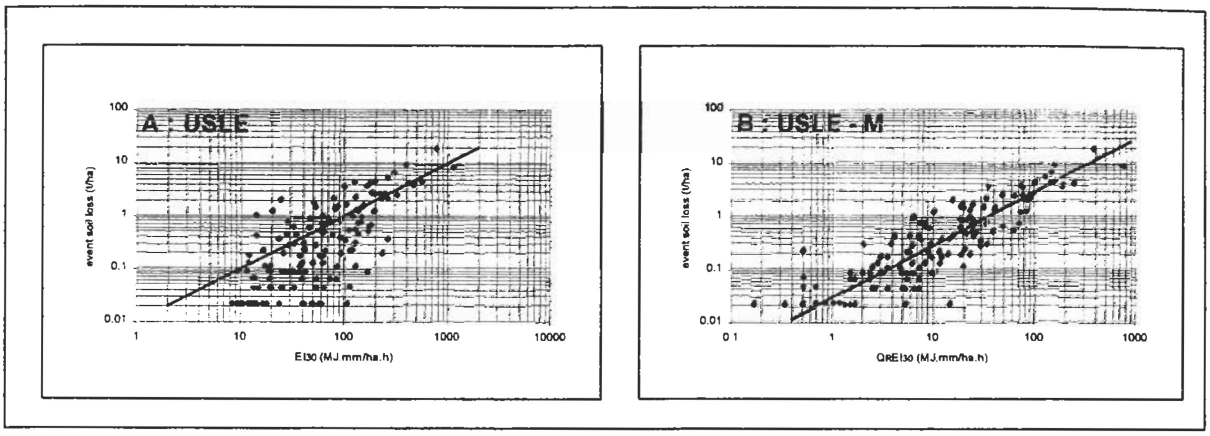

Fig 6. The relationships between event soil loss and the EI_{30} index (A) and the $Q_R EI_{30}$ index (B) on a bare fallow erosion plot at Arnot (Ithaca), NY.

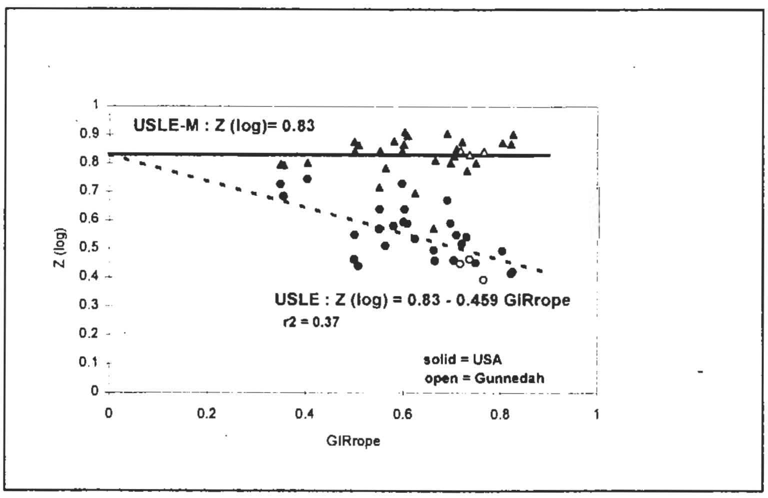

Figure 7. The relationship between the logarithmic form of the Nash-Sutcliffe (1970) model efficiency factor (Z log) and the gross infiltration rate for runoff producing events (GIR_{rope}) for bare fallow plots in the USA and Australia (Gunnedah).

Because the sediment concentration for an event is given (c_e) by

$c_e = A_e/Q_e$ (90)

where A_e is the event soil loss, the K, L, S, C and P factors associated with the USLE-M are directed towards the effects of soil, topography, cropping and conservation practice on sediment concentration alone rather than on both runoff and sediment concentration as in the USLE. Also, while long term average values for these factors exist (eg. K_{UM}, C_{UM}, P_{UM}), the model is event based and can be expressed as

$$A_e = [Q_R EI_{30}]\, K_{eUM}\, L\, S\, C_{eUM}\, P_{eUM} \tag{91}$$

where K_{eUM}, C_{eUM}, and P_{eUM} account for the effect of soil, cropping and conservation practice on sediment concentration during the event and the topographic effects remain as in the USLE. Foster and Wischmeier (1974) considered that the slope could be divided into a number of segments which could be assumed to have uniform slope gradient and soil properties. As a result of this, they developed an equation for calculating the L factor for the ith segment:

$$L_i = \frac{\lambda_i^{m+1} - \lambda_{i-1}^{m+1}}{(\lambda_i - \lambda_{i-1})\,(22.13)^m} \tag{92}$$

where λ_i is the distance from the upslope boundary of the field or hillslope to the lower boundary of the ith segment. One of the assumptions used in the USLE is that runoff is generated uniformly over a hillslope so that, in the calculation of L_i, the segment values of K, S, C and P are considered to apply to the upslope area as well as the segment. However, in many cases runoff is not generated uniformly. For example, soil characteristics may vary down a hillslope in such a way that one part of the hill may yield more runoff than another but Eq. 92 does not have the capacity to deal with this. Eq. 92 was developed by subtracting the sediment yield (proportional to the product of soil loss and slope length) for the hillslope to the bottom of a segment from the sediment yield for the hillslope to the top of the segment assuming the same values for the K,C and P factors as for the segment in both cases, and dividing the result by the area of the segment. Applying a similar approach with the USLE-M gives

$$L_{eUMi} = F\, \frac{Q_{eCi}\,\lambda_i^{m+1} - Q_{eCi-1}\lambda_{i-1}^{m+1}}{Q_{eRi}\,(\lambda_i - \lambda_{i-1})\,(22.13)^m} \tag{93}$$

where

$$F = Q_{eCi-1} + \frac{(1 - Q_{eCi-1})}{(1 + \lambda_{i-1}/(\lambda_i - \lambda_{i-1}))^m} \tag{94}$$

Q_{eCi} is the runoff coefficient (volume of runoff/volume of rain when rain and runoff volumes are associated with the same area) for the hillslope to the bottom of the segment, Q_{eCi} is the runoff coefficient for the hillslope to the top of the segment, and Q_{eRi} is the runoff ratio for the cell. The runoff ratio in this case is the runoff volume discharged from the bottom of the segment divided by the area of the segment. Since this runoff volume includes any runoff from above the segment, Q_{eRi} can have values greater than 1.0 as opposed to Q_{eCi} and Q_{eCi-1} which have values of 1.0 or less. Like the other USLE-M parameters, $L_{i.UMi}$ varies between rainfall events while L_i does not.

While the USLE-M provides an approach that can predict event erosion better than the USLE and also has the capacity to better deal with issues associated with more complex hillslopes, methods for obtaining event parameter values are not widely available yet. Given data on parameters such GIR_{rope} (the gross infiltration rate for runoff producing events) , it is possible to convert USLE K, C, and P values to K_{UM}, C_{UM}, and P_{UM} values (Kinnell and Risse, 1998). K_{UM}, C_{UM} and P_{UM} values differ from USLE K, C and P values because they are directed at comparisons of sediment concentrations rather than soil losses. Procedures for determining K_{eUM}, C_{eUM}, P_{eUM} values have yet to be properly developed.

Other empirical models

The USLE type models (USLE, RUSLE, MUSLE, USLE-M) encapsulate the effect of climate, soil, topography, crops and crop management in the R, K, L, S ,C, and P factors. In the MUSLE and the USLE-M, the R-factor includes runoff so that it depends on an interaction between climate and soil. In the USLE and the RUSLE, R is climate based and the effect of runoff is taken into account through the other terms in the model. That type of approach is not confined to the USLE and the RUSLE. A number of other empirical models use various climate-, soil-, topographic- and vegetation-based indices to estimate erosion. This type of approach may operate at a large scale or a small scale depending on the need. At the hillslope scale, one alternative to the USLE was proposed for use in Southern Africa. The Soil Loss Estimator for Southern Africa (SLEMSA) uses the kinetic energy of rain greater than 25 mm/h (KE>25) in the calculation of R (Elwell and Stocking, 1982). The KE>25 index was originally proposed by Hudson (1965). Hudson observed that when splash cups containing sand were exposed to natural rainfall, the regression between rate of splash loss and rainfall intensity produced a value of zero when the intensity was about 25 mm/h. Following this observation, Hudson suggested that the intensity of 25 mm/h was critical in determining erosion from field plots in tropical Africa despite the fact that came from experiments at a very small scale and with sand, not soil. Kinnell (1978) suggested that the critical intensity should be linked with the infiltration characteristics of the soil, whereas Morgan (1986) considered a critical intensity of 10 mm/h to be more appropriate in temperate areas. However, Kinnell (1983) observed that if one considered rainfall energy as the driving force, the sum of the short-term values of rainfall kinetic energy and the excess rainfall rate (rainfall rate in excess of the infiltration rate) provided a useful storm erosivity index. As the runoff rate is directly linked to the excess rainfall rate at the plot scale, this product can be viewed as indicating that short-term sediment concentrations vary directly with short-term variations in rainfall kinetic energy. Since this view is consistent with field observation, Kinnell et al. (1994) proposed that the sum of the product of short-term values of the excess rainfall rate (I_x) and the rainfall kinetic energy rate (E_A) could be used as an alternative to the EI_{30} index as a storm erosivity index. However, often data on short-term kinetic energies and excess rainfall rates are difficult to obtain and consequently, Kinnell (1997) proposed the $Q_R EI_{30}$ index, the storm erosivity index used in the USLE-M, as an alternative.

One of the major problems that frequently arises in rainfall erosion modelling is the lack of data in relation to running a particular model. This gives rise to the use of alternative indices.

The use of the $Q_R EI_{30}$ index rather than the $I_X E_A$ index mentioned above is a case in point. However, the use of USLE-M, which uses the $Q_R EI_{30}$ index, is constrained by the necessity to obtain different values for the soil, crop and erosion protection factors from those used in the USLE. In some areas, even the data necessary to run the USLE is not directly available. In particular, rainfall intensity data is often lacking and alternative approaches to determining R have to be used. These approaches usually involve correlating USLE R values to some other parameter such as annual precipitation, some monthly weighted value of annual precipitation, or power functions of daily rainfall etc. Obviously, the value of these aproaches needs to be considered in the context of how well they are likely of estimate the value of the particular factor involved and if they actually continue to account for the effects and interactions assigned to the parameter in the original model.

Catchment Scale Models

The MUSLE

Runoff and soil loss plots are small watersheds or catchments. However, as catchment size increases, areas of deposition within the catchment tend to reduce the sediment yield below that predicted from erosion models like the USLE. Under these circumstances, a delivery ratio is used to convert estimates of gross erosion to sediment yield (Williams et al., 1971). The sediment delivery ratio is the ratio of the sediment yield at a specific location in the watershed and the gross erosion upstream of that point. While a sediment delivery ratio is considered necessary to determine sediment delivery from erosion estimated using the USLE in catchments[Ψ], Williams (1975) contended that the delivery ratio is not necessary if the rainfall energy factor in the USLE is replaced by a runoff rate factor because watershed characteristics such as drainage area, stream slope, and watershed shape influence runoff rates and delivery ratios in a similar manner. As a consequence of this, Williams proposed an equation that can be written as

$$SY_e = X_e\, K\, L\, S\, C_e\, P_e \tag{95}$$

where SY_e is the event sediment yield,

$$X_e = \alpha\, (Q_e\, q_p)^{0.56} \tag{96}$$

where α is an empirical coefficient, Q_e is runoff amount and q_p is the peak runoff rate obtained during the erosion event, and K, L, S, are as defined for the USLE with C_e and P_e being event C and P values. This model has become known as the Modified Universal Soil Loss Equation

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[Ψ] The sediment delivery ratio is the ratio of the amount of transported to the amount of material available for transport. It varies from 1.0 only when transport limiting conditions occur. In this context, the USLE is considered to provide an estimate of the amount of material available for transport and a sediment transport model provides an estimate of the sediment transport capacity. Because the sediment delivery ratio is, in this case, based on the values generated by two models, it is an artificial parameter and may not apply if either of the two models is changed.

(MUSLE). The K, L, S, C_e and P_e factors are calculated using weighting factors that are dependent on the drainage area (Williams, 1975). In a comparison on 11 3-acre (1.2 ha) watersheds where delivery ratios were considered to be 1.0, Williams observed that, with the C_e, and P_e set to their average annual values (C, P), Eq. 95 explained 82 % of the variation while the USLE explained 47 %. However, Foster et al. (1982) noted that a major advantage of an erosivity index that includes runoff terms is the elimination of large overestimates of soil loss when runoff is negligible and rainfall amounts and rates are great. The advantage of the Williams index over the USLE is probably more related to the failure of the USLE to consider runoff explicitly as a factor in determining R_e than anything else.

EPIC

In the MUSLE, R_e in the USLE, the EI_{30} index, is replaced by $\alpha\ (Q_e\ q_p)^{0.56}$ while all the other parameters remain as defined for the USLE. In the EPIC model (originally the Erosion/Productivity Impact Calculator but now Environmental Policy Integrated Climate), a continuous simulation model developed by Williams et al. (1984),

$$SY_e = X_e\ K\ L\ S\ C_e\ P_e\ ROKF \tag{97}$$

where ROKF is the coarse fragment factor as defined by Simanton et al. (1984), X_e is selected from one of the following:

$$X_e = EI_{30} \tag{98a}$$

$$X_e = 1.586\ (Q_e\ q_p)^{0.56}\ DA^{0.12} \tag{98b}$$

$$X_e = 0.65\ EI_{30} + 0.45\ (Q_e\ q_p)^{0.33} \tag{98c}$$

where DA is drainage area expressed in ha, Q_e is expressed in mm, q_p in mm/h, EI_{30} in MJ.mm/ha.h and SY_e in t/ha (Williams and Arnold, 1997). Values of K, L, S, C_e, P_e and ROKF are determined independently of the index used for X_e. However, in the USLE, K is, for example, calculated by

$$K = \frac{\sum_{n=1}^{N} (A_e)_n}{\sum_{n=1}^{N} (E\ I_{30})_n} \tag{99}$$

when L = S = C = P = 1.0. Thus it follows that, for EPIC,

$$K = \frac{\sum_{n=1}^{N} (A_e)_n}{N} \tag{100}$$

$$\sum_{n=1} (X_e)_n$$

when L = S = C = P = 1.0, so that the value of K should vary depending which erosivity index is used in EPIC. As the USLE K, C and P factor values are usually used irrespective of which erosivity index is used, concern exists about validity of the MUSLE and EPIC models.

AGNPS

The MUSLE is directed towards the prediction of sediment yield considering a watershed or catchment as essentially a single bulk entity. This does not enable areas of high erosion risk within a catchment to be readily identified. One approach used to overcome this is to use a grid-cell approach to modelling erosion in catchments. The Agricultural Non-Point Source Pollution (AGNPS) (Young et al., 1987) model is an example of this. In AGNPS, the catchment is divided into a number of square primary cells in which factors like soil, slope gradient and land use are considered to be uniform. Erosion is determined by applying the USLE to each cell and a sediment transport model applied to move the sediment from cell to cell giving due consideration to factors such as particle size and deposition. The primary cell can be divided into subcells to deal with nonhomogeniety if necessary. The objective of this approach is to provide a capacity to determine the spatial distribution of erosion risk and sediment sources in addition to the modelling of sediment yield and the impact of pollutants on water quality. However, while the outputs of AGNPS have great potential with respect to providing information to aid in decision making processes related to land and water management, the manner in which the model operates leads to some concern about the validity of those outputs. Firstly, AGNPS is an event model but the USLE is not well suited to predicting erosion caused by individual events. Secondly, with respect to erosion, AGNPS treats each primary cell as a separate hydrologic unit but this approach has serious shortcomings. Naturally occurring hydrologic units are not usually of the same size or square in shape. Consequently, rather than use the primary cell - subcell approach, it is more appropriate to use small grid cells so that the number of cells allocated to a hydrologic unit depends only on the size of the hydrologic unit and the size of the grid cells used. The manner in which data are stored and retrieved in modern Geographic Information Systems (GIS) facilitates the adoption of this approach.

In AGNPS, slope lengths are restricted to a maximum that does not exceed the length of the primary cell or 1000 feet whichever is the least. This rule was developed on the basis that (a), as mentioned earlier, primary cells were separate hydrologic units, and (b), channels or depositions zones which mark the end of slope with respect to the USLE occur within 1000 feet of the upslope boundary of a hillslope. No formal rules were originally set for determining the slope lengths for the subcells. However, Desmet and Govers (1996) extended the method for determining the topographic effect of irregular hillslopes presented in Eq. 92 to the determination of slope length for grid cells that are not independent hydrologic units. For the grid cell with co-ordinates i, j, Desmet and Govers consider the L-factor to be described by

$$L_{i,j} = \frac{(A_{i,j\text{-}in} + D^2)^{m+1} - A_{i,j\text{-}in}^{\ m+1}}{D^{m+2}\, x_{i,j}^{\ m}\, (22.13)^m} \qquad (101)$$

where $A_{i,j\text{-in}}$ is the area contributing to flow into the cell with co-ordinates i,j, D is the length of the sides of the grid cell, and $x_{i,j}$ is the width of the contour over which the flow is discharged. $x_{i,j}$ is dependent on flow direction relative to grid cell orientation. The approach used to develop Eq. 101 involves replacing λ in Eq. 92 by the specific catchment area, the contributing area divided by the width of flow. While applying this approach in AGNPS provides a formal mechanism to determining slope length in the subcells, erosion in a subcell is dependent on the surface water flow that occurs in the subcell. That flow depends not only on the slope length but also on the hydraulic characteristics of the hillslope. The Desmet and Govers method does not account for changes in, for example, variations in runoff from upslope caused by variations in soil hydraulic properties or conditions. In theory, this deficiency, and the problem of the USLE not being suited to predicting event erosion, can be overcome by replacing the USLE by the USLE-M.

AGNPS-UM

AGNPS-UM (Kinnell, 2000) uses the USLE-M rather than the USLE to predict erosion in the grid-cells for an event. As in the case when the USLE-M is applied to hillslope segments, the L factor for the USLE-M differs from that used with the USLE. As noted above, Desmet and Govers (1996) extended Eq. 92 to determine L when the USLE is applied to grid cells. The approach used to develop Eq. 101 involved replacing λ by the specific catchment area, the contributing area divided by the width of flow. Applying a similar approach to give the L factor for applying the USLE-M to grid cells, it follows from Eq. 93 that

$$L_{eUMi,j} = F \, \frac{Q_{eCi,j} \left(A_{i,j\text{-in}} + D^2\right)^{m+1} - Q_{eCi,j\text{-in}} \, A_{i,j\text{-in}}^{\;m+1}}{Qe_{Ri,j\text{-cell}} \, D^{m+2} \, x_{i,j}^{\;m} \, (22.13)^m} \qquad (102)$$

with

$$F = Q_{eCi,j\text{-in}} + \frac{(1 - Q_{eCi,j\text{-in}})}{(1 + A_{i,j\text{-in}} / D^2)^m} \qquad (103)$$

AGNPS-UM uses this approach. In comparison with AGNPS, cell erosion produced by AGNPS-UM is highly dependent on initial soil moisture conditions and position in the landscape. For example, when rain falls on a dry catchment, cell erosion and sediment delivery predicted by the AGNPS-UM is considerably less than that predicted using the AGNPS but the difference diminishes as soil moisture increases (Kinnell, 2000). Cell erosion predicted by AGNPS does not vary with initial soil moisture conditions even though, in reality, it should.

AGNPS98

As noted earlier, despite the USLE not being designed to predict event soil loss, it was used for that purpose in AGNPS. In order to overcome that shortcoming, and take advantage of the new developments associated with the RUSLE, a continuous simulation version of AGNPS called

AGNPS98 was released in 1998. Being a continuous simulation model it uses a weather generator together with plant growth and soil moisture accounting systems but the erosion model still does not explicitly consider the impact of runoff on cell erosion.

Process-based models

As noted earlier, so-called process-based models such as WEPP and EUROSEM can be applied at the small catchment scale but are less applicable in large catchments because of their high data and computational requirements. Both EUROSEM-KINEROS and the WEPP small catchment model use water and sediment routing structures that represent the land surface in a catchment as a series of interlinked uniform sloping planes and channel elements. LISEM (De Roo et al., 1996) uses a EUROSEM based approach in grid cells using dedicated software within a GIS (PCRaster). However, it does not necessarily follow that just because a model is perceived to be process-based, it will predict erosion any better than more empirical one. So-called process-based models include empirical parameters, particularly when modelling the effects of soil on erosion. In relation to an exercise where a number of models were compared in common catchments, Boardman and Favis Mortlock (1996) concluded that it appears that calibration is still essential (or at least desirable) for all current models. They also concluded that runoff was always better simulated than soil loss and consequently, these is a need to improve the conceptual linkages of runoff to soil loss within current models. In addition, more stringent data requirements of event-based models (particularly with regard to initial conditions) render them less able to compete with continuous models when data sets have missing or dubious values.

Where to Now ?

The various modelling approaches described above cover three model types; empirical, conceptual and process-based. The original USLE is considered to be an empirical model since it was developed from experimental data. Process-based models model subprocesses in some detail. WEPP and EUROSEM are considered to be process-based models even though they contain some empirical parameters. The RUSLE is more of a conceptual model than the USLE in that while it includes some of the original empirical formulations, it use conceptual approaches to deal with issues where experimental data does not exist. The C factor is an example. A subfactor approach considering prior land use, canopy cover, surface cover and surface roughness provides a mechanism for determining C factor values for crops and cropping systems for which no plot data exists. The RUSLE 2 goes a step further in modelling deposition by including a sediment transport model.

Each of these types of model has its strengths and weaknesses. Empirical models are usually not seen as being useful outside the data set on which they are based. The USLE was based on data collected largely in the eastern USA and under soil and climate conditions which differ from many other parts of the world where is has been subsequently applied without any particular attempt to validate it. As noted earlier, the USLE was designed to predict the average annual soil loss for a particular crop. Prediction of mean annual values is useful and permitted in areas where annual rainfall tends to follow a normal probability distribution but in arid and semi-arid zones, annual rainfall distributions are often skewed to a large extent. Consequently, averaging techniques, based on normal event probability, are of limited value for describing rainfall recurrences in semiarid areas (Mannaerts and Gabriels, 2000). In view of this, Mannaerts and Gabriels have proposed a probability-based approach where the maximum annual storm, and its associated erosivity, is used as a core element in the assessment of annual rill and interrill erosion rates. Frequency and cumulative soil loss distributions are obtained by combining verified annual and maximum daily rainfall frequency distributions with an erosion algorithm,

$$A_{an} = EI_{24max} \left(1 + \ln\left(\frac{2R - EI_{24max}}{R}\right)\right) K\, Rat \qquad (104)$$

where A_{an} is the annual soil loss, EI_{24max} is the annual maximum 24-h storm erosivity, R is the annual erosivity, K is the soil erodibility and Rat is the soil loss ratios for topographic, surface cover and soil management effects normally involved in determining C is the RUSLE, to provide a statistical representation of erosion. Mannaerts and Gabriels suggest that the method seems applicable to arid and semiarid ecosystems with a high seasonal concentration of precipitation and with rainfall limited to only a few major storm events.

The Mannaerts-Gabriels approach is an attempt to modify the USLE approach towards conditions that prevail in semi-arid and arid areas more than wetter areas. Approaches using more event orientated models, such as the USLE-M, WEPP and EUROSEM, can also be used to better deal with seasonal or low frequency, high impact events. A considerable effort has been made in developing models that are more process based than the USLE. However, lack of data often provides constraints as to what approach can be used. Event based approaches by and large involve runoff prediction which is seen by some to be difficult. However, as noted above, Boardman and Favis Mortlock (1996) observed that many recent models predict runoff better than they do soil loss. Event based models appear to fail more readily than those directed towards predicting long term averages because it is easier to predict that average than the individual values which contribute to the average. Also, models which poorly predict event values may do a very reasonable job in predicting the average. However, models which predict event values well have the potential to predict the long term average well. Even so, lack of good data may be the primary constraint to achieving this.

There are a large number of rainfall erosion models covering the hillslope, catchment and regional scales. The more process-based models operate at the hillslope scale. Empirical and conceptual models operate at all three scales. Regional scale models usually have to work with relatively poor data and produce crude results. However, they are useful for identifying areas where more detailed approaches should be applied. Usually, catchment scale models are, as illustrated by AGNPS (Young et al., 1987), extensions of hillslope models. Different catchment models use different approaches to the modelling of hillslope erosion or landscape hydrology. For example, SWAT (Soil and Water Assessment Tool, Arnold et al, 1995) uses the MUSLE (Williams et al., 1971) while AGNPS uses the USLE to predict erosion. The two models also differ in respect to how they deal with landscape hydrology. SWAT uses a water balance equation while AGNPS uses the Curve Number technique. AGNPS is an event model but AGNPS98 and SWAT are continuous-time models designed to predict longer term values. Given differences in the approaches to modelling erosion and landscape hydrology, applying a number of models in the same situation will, be it a hillslope or a catchment, an event or longer time period, generate a number of differing results. Some testing of models usually occurs during development but many models are applied elsewhere without proper checks on their performance. A lack of distributed runoff and sediment yield data exists and this is a problem that limits all distributed simulation models (Lane et al, 1997). As noted earlier, Boardman and Favis Mortlock (1996) concluded that it appears that calibration is still essential (or at least desirable) for all current models even though most have been developed to produce results without calibration. Uncertainty about the value of results obtained when calibration has not taken place is often ignored when the results are used. The credibility of a model usually results from confidence in the operation of the model components irrespective of whether that confidence is misplaced or not.

Obviously, the current state of rainfall erosion modelling results from a suite of historical events. In the 1960's and 70's, the objective was to provide an approach that would illustrate the impact of land management and land management change on a hillslope within the boundaries of a given farm. The USLE has been an extremely powerful tool in this context. However, its incapacity model event erosion reasonably well and to deal with deposition on non-uniform slopes led to a call for more process-based approaches that had the capacity to overcame such inadequacies. This spawned the USDA Water Erosion Prediction Project (WEPP) in the USA and the development of EUROSEM in Europe. However, from a management decision point of view, these models are too complex and require too much data so that, despite their development, the USLE approach has been, and is being upgraded through the RUSLE (RUSLE 2) and the USLE-M. That upgrade has also been spawned by the need to model erosion and its offsite impacts within catchments (watersheds in US terms) and large basins. However, the distributed modelling approaches, particularly the event base ones, that have resulted at the catchment/basin scale are virtually impossible to validate because of a lack of appropriate data. Some testing of the performance of distributed models producing long term results has occurred using presence of radioisotopes associated with atmospheric nuclear testing as indicators of erosion and deposition but that has not been particularly widespread. Techniques involving radioisotopes are being developed that facilitate the determination of short- and medium-term rates of water erosion on agricultural land (Walling and Blake, 1999) and these could provide useful data for testing event-based distributed models in the future.

Distributed modelling systems tend to deal with erosion and sediment transport through a detachment and transport approach with the amount of soil leaving an area being limited by either detachment or transport capacity. This applies to both empirical-conceptual models and process-based models. In AGNPS, the USLE is used to provide the estimate of the amount of detached soil material that is available to be transported and a sediment transport equation is used to determine the sediment transport capacity of the flow leaving the area of interest. In general, this approach produces a detachment limiting conditions at the top of a hillslope leading to transport limiting conditions at the bottom of a hillslope, particularly where, as is often the case, the lower slopes have a lower slope gradient. Under these conditions, it is the sediment transport model and not the USLE that controls the estimate of the sediment leaving the hillslope. Internally, both models are important in the context of modelling the movement of pollutants. Once again, how the movement of pollutants is modelled in models like AGNPS relies heavily on often not well validated theory. There is considerable scope for improvement in this area.

The sediment-pollutant issue adds a complication in that predicting amounts of soil lost or deposited per se is insufficient since particles of different sizes move at different rates and are not deposited in the same place. Fine material is much more mobile than coarse material. Fine material also tends to be more chemically active than coarse material and thus its movement is of great importance with regard to pollutant movement. While some concern may exits with regard to modelling the movement of sediment of various sizes at the catchment/basin scale, the approaches of Hairsine and et al (1999) and Kinnell (1994) which model the movement of particles at the fine scale provide mechanisms for predicting the movement of pollutants from their initial position on the soil surface to lines of concentrated flow. At this scale, it should be possible to link theory and experiment together better than at larger scales.

Apart from the European effort put into the development of EUROSEM, the USA has been the most active area for water erosion model development. Most other areas do not have sufficient resources to match the US effort. Consequently, US based modelling systems dominate outside the USA. However, since the US effort is directed, as it should be, towards the erosion issues that are dominant in the USA, it will produce approaches which are seen to be less appropriate and less applicable elsewhere. Adaptation and modification of US based technology to operate better outside the USA will continue for the foreseeable future.

References

Abrahams,. A.D., and Parsons, A.J. 1991. Resistance to overland flow on a desert pavement and its implications for sediment transport modelling. *Water Resources Research* **27,** 1827-1836.

Al-Durrah, M, and Bradford, J.M. 1981. New methods of studying soil detachment due to waterdrop impact. *Soil Science Society America Journal* **45**, 949-953.

Al-Durrah, M, and Bradford, J.M. 1982a. Parameters for describing soil detachment due to single waterdrop impact. *Soil Science Society America Journal* **46**, 836-840.

Al-Durrah, M, and Bradford, J.M. 1982b. Mechanism of raindrop splash on soil surfaces. *Soil Science Society America Journal* **46**, 1086-1090.

Allen, R.F. 1987. The mechanism of splashing. *Journal Colloid and Interface Science* **124**, 836-840

Arnold, J.G., Williams, J.R., Maidment, D.R. 1995. Continuous-time water and sediment-routing model for large basins. *Journal of Hydraulic Engineering* **121**, 171-183.

Boardman, J, and Favis-Mortlock, D. 1996. Modelling soil erosion by water: Some conclusions. **In** *Modelling soil erosion by water. Proceedings of the NATO Advanced Research Workshop "Global Change: Modelling Soil Erosion by Water", 11-14 September 1995, University of Oxford, UK.*

Ciampalini, R, and Torri, D. 1998. Detachment of soil particles by shallow flow: Sampling methodology and observations. *Catena* **32**, 37-53.

De Roo, A.J.P., Wesseling, C.G., and Ritesma, C.J, 1996. LISEM: A single event physically based hydrologic and soil erosion model for drainage basins. I. Theory, input and output. *Hydrological Processes* **10**, 1107-1117.

Desmet, P.J.J., and Govers, G. 1996. A GIS procedure for automatically calculating the USLE LS factor on topographically complex landscape units. *Journal Soil and Water Conservation* **51**, 427-433, 1996

Elliott, W.J., Liebenow, A.M., Laflen, J.M., and Kohol, K.D. 1989. A compendium of soil erodibility data from WEPP cropland soil field erodibility experiments 1987 & 88. *Publication No. 3*, ARS-USDA, National Soil Erosion Research Laboratory, Purdue Univ., West Lafayette, IN.

El-Swaify, S.A., and Dangler, E.W. 1976. Erodibilities of selected tropical soils in relation to structural and hydrologic parameters. **In** *Soil Erosion: Prediction and Control*, 105-114, Soil Conservation Society America, Ankeny, Iowa.

Elwell, H.A., and Stocking, M.A. 1982. Developing a simple yet practical method of soil loss estimation. *Tropical Agriculture* **59**, 43-47.

Everaert, W. 1991. Empirical relations for sediment transport capacity of interrill flows. *Earth Surface Processes and Landforms* **16**, 513-532.

Farrell, D.A., Moldenhauer, W.C., and Larson, W.E. 1974. Splash correction factors for soil erosion studies. *Soil Society America Proceedings* **38**, 510-514.

Favis-Mortlock, D.T. (1998). A self-organising dynamic systems approach to the simulation of rill initiation and development on hillslopes. *Computers and Geosciences* **24**, 353-372.

Favis-Mortlock, D.T., Guerra, A.J.T. and Boardman, J., 1998. A self-organising dynamic systems approach to hillslope rill initiation and growth: model development and validation. **In** Summer, W., Klaghofer, E. and Zhang, W. (eds) *Modelling Soil Erosion, Sediment Transport and Closely Related Hydrological Processes*, IAHS Press no. 249, Wallingford, UK., 53-61.

Finkner, S.C., Nearing, M.A., Foster, G.R., and Gilley, J.E. 1989. A simplified equation for modelling sediment transport capacity. *Transactions American Society Agricultural Engineers* **32**, 1545-1550.

Flanagan, D.C., and Nearing, M.A. (Eds). 1995. USDA-Water Erosion Prediction Project: Technical documentation. *NSERL Rep. No. 10.* National Soil Erosion Research Laboratory, West Lafayette, IN.

Foster, G.R., and Meyer, L.D. 1975. Mathematical simulation of upland erosion by fundamental erosion mechanics. In *Present and prospective technology for prediction sediment yields and sources.* ARS-S40.USDA-ARS Southern Region. New Orleans, Louisiana.

Foster, G. R., Meyer, L. D., and Onstad, C. A. 1977. A runoff erosivity factor and variable slope length exponents for soil loss estimates. *Transactions American Society Agricultural Engineers* **20**, 683-7.

Gerald, C.F., and Wheatly, P.O. 1984. *Applied Numerical Analysis (3 edn).* Additon_Wesley, Reading, MA.

Govers, G. 1990. Empirical relationships on the transport capacity of overland flow. *International Association of Hydrological Sciences*, **Pub 189**, 45-63.

Grosh, J.L., and Jarrett, A.R. 1994. Interrill erosion and runoff on very steep slopes. *Transactions American Society Agricultural Engineers* **37**, 1127-1133

Hairsine, P.B., and Rose, C.W. 1991. Rainfall detachment and deposition: Sediment transport in the absence of flow-driven processes. *Soil Science Society America Journal* **55**, 320-324.

Hairsine, P.B., and Rose, C.W. 1992a. Modeling water erosion due to overland flow using physical principles. 1. Sheet flow. *Water Resources Research* **28**, 237-243.

Hairsine, P.B., and Rose, C.W. 1992a. Modeling water erosion due to overland flow using physical principles. 2. Rill flow. *Water Resources Research* **28**, 245-250..

Hairsine, P.B., Sander, G.C., Rose, C.W., Parlange, J. -Y., Hogarth, W.L., Lisle, I, and Rouhipour, H. 1999. Unsteady soil erosion due to rainfall impact: A model of sediment sorting on a hillslope. *Journal of Hydrology* **220**, 115-128.

Huang, C., Bradford, J.M, and Laflen, J.M. Evaluation of detachment-transport coupling concept in the WEPP rill erosion equation. *Soil Science Society America Journal* **60**, 734-739.

Hudson, N.W.H. 1965. The influence of rainfall on the mechanics of soil erosion with particular reference to Southern Rhodesia. Unpublished MSc thesis, Univ. of Cape Town, South Africa.

Kinnell, P.I.A. 1978. A proposal for a new erosive index for soil loss prediction. *Preprint of papers, Conference on Agricultural Engineering 1978, Toowoomba, Queensland.* The Institution of Engineers, Australia.

Kinnell, P.I.A. 1983.The effect of kinetic energy of excess rainfall on soil loss from non-vegetated plots. *Australian Journal Soil Research* **21**, 445-453

Kinnell, P.I.A. 1998. The influence of flow discharge on sediment concentrations in raindrop induced flow transport. *Australian Journal Soil Research* **26**, 575-582.

Kinnell, P.I.A. 1990. The mechanics of raindrop induced flow transport. *Australian Journal Soil Research* **28**, 497-516

Kinnell, P.I.A. 1993a. Sediment concentrations resulting from flow depth/drop size interactions in shallow overland flow. *Transactions American Society Agricultural Engineers* **36**, 1099-1103.

Kinnell, P.I.A. 1993b. Interrill erodibilities based on the rainfall intensity-flow discharge erosivity factor. *Australian Journal Soil Research* **31**, 319-332.

Kinnell, P.I.A., 1993c. Soil/slope gradient interactions in erosion by rain-impacted flow. *Transactions American Society Agricultural Engineers* **36**, 381-387.

Kinnell, P.I.A. 1994. The effect of pre-detached particles on soil erodibilities associated with erosion by rain-impacted flows. *Australian Journal Soil Research* **32**, 127-142.

Kinnell, P.I.A. 1997. Runoff ratio as a factor in the empirical modelling of soil erosion by individual rainstorms. *Australian Journal Soil Research* **35**, 1-13

Kinnell, P.I.A. 2000. The effect of slope length on sediment concentrations associated with side-slope erosion. *Soil. Science Society America Journal* **64**, 1004-1008.

Kinnell, P.I.A. 2000. AGNPS-UM: applying the USLE-M within the agricultural non point source pollution model. *Environmental Modelling & Software* **15**, 331-341

Kinnell, P.I.A., and Risse, L.M. 1998. USLE-M: Empirical modeling rainfall erosion through runoff and sediment concentration. *Soil Science Society America Journal* **62**, 1667-1672.

Kinnell, P.I.A., K.C. McGregor, and C.J. Rosewell. 1994. The I_XE_A index as an alternative to the EI_{30} erosivity index. *Transactions American Society Agricultural Engineers* **37**, 1449-1456.

Laflen, J.J., Elliott, W.J., Flanagan, D.C., Meyer, C.R., and Nearing, M.A. 1997. WEPP-Predicting water erosion using a process-based model. *Journal Soil and Water Conservation* **52**, 96-102.

Lane, L.J., Hernandez, M., and Nichols, M. 1997.Processes controlling sediment yield from watersheds as functions of spatial scale. *Environmental Modelling & Software* **12**, 355-369.

Lane, L.J., Nichols, M.H., and Paige, G.B. 1995. Modelling erosion on hillsopes: Concepts, theory and Data. **In** *Proceedings of the International Congress on Modelling and Simulation (MODSIM'95)* (ed P.Binning, H.Bridgman and B.Willims), Nov 27-30, 1995, Univ of Newcastle, Newcastle, NSW, Australia. pp1-7. Uniprint, Perth, Australia

Lane,L.J., Shirley, E.E., and Singh, V.P. 1988. Chapter 10. Modelling erosion on hillsopes. **In** *Modelling Geomorpholigical Systems* (ed M.G. Anderson). pp 287-308. John Wiley & Sons, Chichester.

Lewis, S.M., Barfield, B.J., and Storm, D.E., and Ormsbee, L.E.. 1994a. PRORIL - An erosion model using probability distributions for rill flow and density. I. Model Development. *Transactions American Society Agricultural Engineers* **37**, 115-123.

Lewis, S.M., Barfield, B.J., and Storm, D.E., and Ormsbee, L.E.. 1994b. PRORIL - An erosion model using probability distributions for rill flow and density. I. Model Validation. *Transactions American Society Agricultural Engineers* **37**, 125-133.

Lewis, S.M., Barfield, B.J., and Storm, D.E. 1994c. A method for developing probability distributions for rill flow and density. *Transactions American Society Agricultural Engineers* **37**, 1879-1887.

Liebenow, A.M., Elliott, W.J., Laflen, J.M., and Kohl, K.D. 1990. Interrill erodibility: Collection and analysis of data from cropland soils. *Transactions American Society Agricultural Engineers* **33**, 1882-1888.

Loch, R.J., Slater, B.K., and Devoil, C.. (1998). Soil erodibility (Km) values of some Australian soils. *Australian Journal Soil Research* **36**, 1045-1055

Maclin, W.C., and Hobbs, P.V. 1969. Subsurface phenomena and the splashing of drops in shallow liquids. *Science* **166**,1107-1108.

Mannaerts, C.M., and Gabriels, D. 2000. A probabilistic approach for predicting rainfall soil erosion losses in semiarid areas. *Catena* **40**, 403-420.

Meyer, L.D., and Harmon, W.C. 1984. Susceptibility of agricultural soils to interrill erosion. *Soil Science Society America Journal* **48**, 1152-1157.

Meyer, L.D., and Harmon, W.C. 1989. How row-sideslope length and steepness affect sideslope erosion. *Transactions American Society Agricultural Engineers* **32**, 639-644.

Meyer, L.D., and Harmon, W.C. 1992. Interrill runoff and erosion: Effect of row-sideslope shape, rain energy and rain intensity. *Transactions American Society Agricultural Engineers* **35**, 1199-1203.

Meyer, L.D, and Wischeier, W.H. 1969. Mathematical simulation of the process of soil erosion by water. *Transactions American Society Agricultural Engineers* **12**, 754-758, 762.

Misra, R.K, and Rose, C.W. 1996. *European Journal Soil Science* **47**, 593-604.

Moore, I.D., and Burch, I.D. 1986. Sediment transport capacity of sheet and rill flow: application of unit stream power theory. *Water Resources Research* **22**, 1350-1360

Morgan, R.P.C. *Soil erosion and conservation.* Longman, Harlow, UK.

Morgan, R.P.C., Quinton, J.N., Smith, R.E., Govers, G., Poesen, J.W.A., Auerswald, K., Chisci, G., Torri, D., and Styczen, M.E. 1998. The European Soil Erosion Model (EUROSEM): A dynamic approach for predicting sediment transport from fields and small catchments. *Earth Surface Processes and Landforms* **23**, 527-544.

Moss, A.J. 1991a. Rain-impact soil crust. 1. Formation on a granite-derived soil. *Australian Journal Soil Research* **29**, 271-289.

Moss, A.J. 1991b. Rain-impact soil crust. 2.Some effects of surface slope, drop size, and soil variation. *Australian Journal Soil Research* **29**, 291-309.

Moss, A.J., and Green, P. 1983. Movement of solids in air and water by raindrop impact. Effect of drop-size and water-depth variations. *Australian Journal Soil Research* **21**, 257-269.

Moyersons, J, and DePloy, J. 1976. Quantitative data on splash erosion simulated on unvegetated slopes *Z.f. Geormorph. N. F.* **Suppl. Bd. 25**: 120-131

Nash, J.E., and J.E. Sutcliffe, 1970. River flow forecasting through conceptual models. Part 1 - A discussion of principles. *Journal of Hydrology* **10**: 282-290

Nearing, M.A. 1991. A probablistic model of soil detachment by shallow turbulent flow. *Transactions American Society Agricultural Engineers* **34**, 81-85

Nearing, M.A., Norton, L.D., Bulgakov, D.A., Larionov, G.A., West, L.T. and Dontsova, K. (1997). Hydraulics and erosion in eroding rills. *Water Resources Research* **33**, 865-876.

Nearing, M.A., and Parker, S.C. 1991. *Soil Science Society America Journal* **58**, 1612-1614.

Neal, J.H. 1938. Effect of degree of slope and rainfall characteristics on runoff and soil erosion. *Agricultural Engineering* **19**: 213-217.

Parker, D.B., Michel, T.G., and Smith, J.L. 1995. Compaction and water velocity effects on soil erosion in shallow flow. J. Irrig. Drain. Eng. 121, 170-178.

Parsons, A.J., Abrahams, A.D, and Wainwright, J. 1994. *Catena* **22**, 215-226.

Proffitt, A.P.B., Rose, C.W., and Hairsine, P.B. 1991. Rainfall detachment and deposition: Experiments with low slopes and significant water depths. *Soil Science Society America Journal* **55**: 325-332.

Quansah, C. 1981. The effect of soil type, slope, rain intensity and their interactions on splash detachment and transport. *Journal Soil Science 32*, 215-224

Renard, K.G., Foster, G.R., Weesies, G.A., McCool, D.K., and Yoder, D.C. 1997. Predicting soil erosion by water: A guide to conservation planning with the Revised Universal Soil Loss Equation (RUSLE). U.S. Dept. Agric., *Agric. Hbk.* **No. 703**.

Romkens, M.J.M., and Wang, J.Y. 1987. Soil roughness changes from rainfall. *Transactions American Society Agricultural Engineers* **30**, 101-107.

Rose, C.W. 1993. Erosion and sedimentation. **In** Bonell, M., Hufschmidt, M.M., and Gladwell, J.S. (Eds). *Hydrology and Water Management in the Humid Tropics - Hydrological Research Issues and Strategies for Water Management.* pp 301-343. Cambridge University Press, Cambridge.

Rose, C.W. 1988. Research progress on soil erosion processes and a basis for soil conservation practices. **In** Lal, R. (Ed). *Soil Erosion Research Methods.* pp 119-139. Soil and Water Conservation Society, Ankeny, Iowa.

Rose, C.W., and Hairsine, P.B. 1988. Processes of water erosion. **In** Steffan, W.L., and Denmead, O.T. (Eds). *Flow transport in the natural environment: Advances and applications.* Spring-Verlag, Berlin.

Rose. C.W., Williams, J.R., Sander, G.C., and Barry, D.A. 1983. A mathematical model of erosion and deposition processes. :I. Theory for a plane land element. *Soil Science Society America Journal* **47**, 991-995.

Sander, G.C., Hairsine, P.B., Rose, C.W., Cassidy, D., Parlange, J. -Y., Hogarth, W.L., and Lisle, I.G. 1996. Unsteady soil erosion model, analytical solutions and comparisons with experimental results. *Journal of Hydrology* **178**, 351-367.

Schultz, J.P., Jarrett, A.R., and Hoover, J.R. 1985. *Transactions American Society Agricultural Engineers* **28**, 1878-1884.

Sharma, P.P., and Gupta, S.C. 1989. Sand detachment by single raindrops of varying kinetic energy and momentum. *Soil Science Society America Journal* **53**, 1005-1010.

Sharma, P.P., Gupta,. S.C, and Rawls, W.J. 1991. Soil detachment by single raindrops of varying kinetic energy. *Soil Science Society America Journal* **55**, 301-307.

Simanton, J.R., Rawizt, E., and Shirley, E.D. (1984). Effects of rock fragments on erosion of semi-arid rangeland soils. **In** *Erosion and Productivity of Soils Containing Rock Fragments.* Soil Science Society Amererica, Madison, WI. 65-67.

Terry, J.P. 1998. A rainsplash component analysis to define mechanisms of soil detachment and transportation. *Australian Journal Soil Research* **36**, 525-542.

Torri, D., and Poesen, J. 1988. The effect of cup size on splash detachment and transport measurements. Part II. Theoretical approach. *Catena Supp.* **12**, 127-137.

Torri, D, Sfalanga, M., and Del Sette, F. 1987. Splash detachment: Runoff depth and soil cohesion. *Catena* **14**, 149-155.

Walling, D.E., and Blake, W. 1999. Use of 7Be and 137Cs measurements to document short- and medium-term rates of water-induced erosion on agricultural land. *Water Resources Research* **35**, 3865-3874.

Wan, Y., and El-Swaify, S.A. 1999. Runoff and soil erosion as affected by plastic mulch in a Hawaiian pineapple field. *Soil and Tillage Research* 52, 29-35.

Williams, J.R. 1975. Sediment-yield prediction with universal equation using runoff energy factor. **In** *Present and Prospective Technology for Predicting Sediment Yield and Sources.* **ARS.S-40**, US Gov. Print Office, Washington, D.C., 244-252

Williams, J.R., and Arnold, J.G. 1997. A system of erosion-sediment yield models. *Soil Technology* **11**, 43-55

Williams, J.R, Hiler, E.A., and Baird, R.W. 1971. Predicting sediment yields from small watersheds. *Transactions American Society Agricultural Engineers* **14**, 1157-1162

Wischmeier, W.H., Johnson, C.B., and Cross, B.V. 1971. A soil erodibility nomograph for farmland and construction sites. *Journal Soil and Water Conservation* **26**, 189-193.

Wischmeier,W.C., and Smith, D.D. 1978. Predictong rainfall erosion losses – a guide to conservation planning. *Agric. Hbk.* **No. 537**. US Dept Agric., Washington, DC.

Yalin, Y.S. 1963. An expression for bed-load transportation. *Journal Hydraulics Division - American Society Civil Engineering* **89**, 221-250.

Young, R.A., Onstad, C.A., Bosch, D.D. and Anderson, W.P. 1987. AGNPS, Agricultural-Non-Point-Source Pollution model; A large watershed analysis tool. *Conservation Research Report* **35**, USDA-ARS, Washington, DC.

Yu, B., Sajjapongse, A., Yin, D., Eusof, D., Anecksamphant, C., Rose, C.W., and Cakurs, U. Application of a physically based erosion model, GUEST, in the absence of data on runoff rates II. Four case studies from China, Malaysia, and Thailand. *Australian Journal Soil Research* **37**, 13-31.