Earth Surface Processes and Landforms
Earth Surf. Process. Landforms **29**, 1293–1302 (2004)
Published online in Wiley InterScience (www.interscience.wiley.com). DOI: 10.1002/esp.1096

A CONCEPTUAL MODEL FOR DETERMINING SOIL EROSION BY WATER

ANTHONY J. PARSONS,[1]* JOHN WAINWRIGHT,[2] D. MARK POWELL,[1] JÖRG KADUK[1] AND RICHARD E. BRAZIER[3]

[1] *Department of Geography, University of Leicester, Leicester, UK*
[2] *Environmental Monitoring and Modelling Research Group, Department of Geography, King's College London, London, UK*
[3] *Department of Geography, University of Sheffield, Sheffield, UK*

Received 18 January 2002; Revised 18 December 2003; Accepted 5 February 2004

ABSTRACT

Current estimates of rates of soil erosion by water derived from plots are incompatible with estimates of long-term lowering of large drainage basins. Traditional arguments to reconcile these two disparate rates are flawed. The flux of sediment leaving a specified area cannot be converted to a yield simply by dividing by the area, because there is no simple relationship between flux and area. Here, we develop an approach to the determination of erosion rates that is based upon the entrainment rates and travel distances of individual particles. The limited available empirical data is consistent with the predictions of this approach. Parameterization of the equations to take account of such factors as gradient and sediment supply is required to proceed from the conceptual framework to quantitative measurements of erosion. However, our conceptual model solves the apparent paradox of the sediment delivery ratio, resolves recent discussion about the validity of erosion rates made using USLE erosion plots, and potentially can reconcile erosion rates with known lifespans of continents. Our results imply that previous estimates of soil erosion are fallacious.

KEY WORDS: soil erosion; erosion rates; sediment delivery

INTRODUCTION

Knowledge of rates of soil erosion by water is important for two principal reasons. First, it is essential to our understanding of landform development. Secondly, on agricultural land, these rates determine the long-term sustainability of agricultural practices and have profound economic consequences (Pimental, 1995). In turn, these consequences influence policy on erosion control. Determination of these rates is typically obtained from short-term measurements made on run-off plots (of a few square metres) or via equations that predict soil erosion, which are themselves calibrated using data from such plots. However, both to understand the role of soil erosion in landform development and to use these rates for erosion control requires that the plot data be extrapolated both to longer time-scales than the period of measurement and to larger areas than those of run-off plots. Such extrapolation has proved problematic.

All data on soil erosion derived from plots yields rates that are incompatible with long-term rates of continental lowering (Summerfield and Hulton, 1994). Using such data, Pimental *et al*. (1995) claim that, globally, the lowest rates of soil erosion are found in the United States and Europe, for which locations a rate of 17 Mg ha^{-1} a^{-1} is cited. Extrapolating this rate over geological time-scales leads to the conclusion that the two locations with the lowest rates of soil erosion would be eroded to sea level, by this process alone and under the present climatic conditions, in about 1 million and 500 000 years, respectively (assuming approximate average elevations for the United States and Europe of 1000 m and 500 m, respectively, and average soil bulk density of 1·6 Mg m^{-3}). This calculation ignores both other agents of denudation and rates of continental uplift. However, even for the Central Alps the rate of uplift over the last 35 Ma is estimated at no more than 1 mm a^{-1} (Clark and Jäger, 1969) which is equivalent to the claimed rate of erosion by water (*c*. 1 mm a^{-1}). Consequently, even discounting the fact that much higher rates of denudation by mass movements and glacial activity are likely in

* Correspondence to: A. J. Parsons, Department of Geography, University of Leicester, University Road, Leicester, LE1 7RH, UK. E-mail: ajp16@le.ac.uk

a mountain environment, the highest parts of Europe could not have come into existence if these rates of erosion were valid under past climatic conditions. Boardman (1998) has discredited the estimate for the average rate of soil erosion in Europe and Crosson (1995) has disputed the figure for the United States. However, in neither case do these commentators argue for rates to be one to two orders of magnitude less, which they would need to be to become compatible with the evident longevity of the continents and the existence of mountains. Boardman (pers. comm.) suggests that a value of 3–4 Mg ha^{-1} a^{-1} is a more realistic value for Europe, and Crosson merely points to more recent data indicating a reduction of 25 per cent in the estimate for the United States. More recently, Yang *et al*. (2003) have used the Revised Universal Soil Loss Equation (Renard *et al*., 1997) in a GIS-based model to determine global rates of soil erosion. This approach has yielded a global average rate of 10·2 Mg ha^{-1} a^{-1} and values for Europe and North America of 11·1 and 9·3 Mg ha^{-1} a^{-1}, respectively.

Traditionally, two arguments have been put forward in attempts to explain away the difference between measured rates of soil erosion and the longevity of continents. First, the present rates of soil erosion are affected by human activity, in particular clearing of natural vegetation, and are believed to be significantly higher than those that obtain under natural vegetation. However, global estimates (Dedkov and Moszherin, 1992; Walling and Webb, 1996) put the increase in erosion rates due to human activity at no more than 2–5 times, so that even at a lower rate of about 3·4 Mg ha^{-1} a^{-1} (0·2 mm a^{-1}), the lives of the United States and Europe would be extended only to 5 and 2·5 Ma, respectively (again assuming no other denudational processes were active and that the rates of erosion under present climatic conditions remained valid). Thus this factor appears relatively insignificant.

Secondly, it has been found that sediment yield from catchments is typically lower than gross erosion within the catchment (Walling, 1983). The disparity between the two values (known as the sediment-delivery ratio – Glymph, 1954) is explained by the fact that many of the measures of erosion do not take into account deposition of eroded material within the catchment which serves to lengthen the lives of land masses. An estimate of an average global sediment-delivery ratio is difficult to obtain because of the observed inverse relationship between it and the size of the area under investigation (Walling, 1983). Nevertheless, estimates for the 10 largest river basins in the United States (Robinson, 1977) gives an average sediment-delivery ratio of 4·7 per cent, which would be sufficient to extend the life of the United States to over 60 Ma – apparently sufficient to reconcile observed rates of erosion by water and long-term landform evolution, especially if the effects of human activity on erosion rates are also taken into account. However, there is a significant problem with using the concept of the sediment-delivery ratio to reconcile short-term measurements of erosion rates with landscape development at a geological time-scale. As Graf (1988) has pointed out, over geological time-scales the sediment-delivery ratio must equal 100 per cent, for otherwise the quantity of sediment deposited within catchments would increase indefinitely. Even for much shorter periods of time, if measured rates of present erosion are at least approximately valid for the period over which agriculture has been practised, the depths of accumulated sediment within catchments would far exceed that which is observed (Trimble and Crosson, 2000). For Europe, where agriculture has been practised extensively over at least the last 4000 years, a sediment-delivery ratio of about 5 per cent and an erosion rate of 17 Mg ha^{-1} a^{-1} would, if we generously assume valley floor areas over which sedimentation takes place to be equal to 20 per cent of catchment areas, yield average depths of sediment deposition of more than 19 m, assuming that all the soil in Europe had not long since disappeared. Likewise, given the difference in the length of agricultural practice in Europe and the United States, if these rates of erosion were valid, in similar settings the erosional and depositional landscapes of the two continents would look vastly different, whereas they do not. Thus the sediment-delivery ratio argument also fails to reconcile rates of erosion of the order generally claimed with the longevity of continents. We suggest that the problem lies with the way in which such rates of erosion have been calculated.

Rates of soil erosion are obtained from run-off plots by dividing the quantity of soil eroded from the plots (M) by the time (T) over which the measurement is made to obtain the flux of eroded soil ($M\ T^{-1}$), and then area of the plot (L^2) to obtain the erosion rate ($M\ L^{-2}\ T^{-1}$). Implicit in this approach is the assumption that the quantity of soil eroded is directly proportional to the area over which erosion is measured. Three lines of evidence suggest that this is not the case. First, Evans (1995) has observed that, at the scale of field-sized areas, rates of erosion typically become smaller as the area over which they are measured becomes larger. Secondly, measurements of travel distances of individual particles during run-off events show that these travel distances to be small (Parsons *et al*., 1993), inversely related to particle size (Parsons and Stromberg, 1998; Wainwright and Thornes, 1991) and have a negative exponential or gamma distribution. Thus, only the smallest eroded particles, or a fraction of larger ones, are likely to reach the base of a hillslope, after even a very large run-off event. Thirdly, it is generally accepted that erosion is limited by detachment and, where detachment is due to raindrop impact (interrill erosion), the rate of detachment declines with distance downslope (Gilley *et al*., 1985; Abrahams *et al*., 1991). Fundamentally, what these lines of evidence imply is that the notion that the quantity of soil eroded from an area is directly proportional to the size of the area is flawed. There is no simple relationship between the flux of sediment measured at a particular location and the area that contributes to that flux. Thus sediment yield ($M\ L^{-2}\ T^{-1}$) cannot be derived from sediment flux ($M\ T^{-1}$) by dividing by area (L^2). Consequently, when such erosion rates are extrapolated to large catchments or continents the results will be invalid. However, if erosion rates are to be determined for large catchments and continents a relationship between measurable flux and contributing area is required. In this paper we will develop an alternative approach to the determination of erosion rates.

THE FLUX OF SEDIMENT TRANSPORTED BY WATER

For any process of erosion by water, the one-dimensional flux φ_d of sediment ($M\ T^{-1}$) of a given size d (L) is given by the product of its entrainment rate E_d ($M\ L^{-1}\ T^{-1}$) and its travel distance L_d (L) (see Kirkby, 1991). In the simplest case, assuming that particles of only one size are present, entrainment is spatially uniform and that travel distances are all the same (but not infinite), then away from the influence of boundary conditions sediment flux will be spatially uniform as particles are entrained and travel a finite distance before being deposited. Division of this flux by any length L_n (where $L_n > L_d$) will yield a smaller erosion rate ($M\ L^{-1}\ T^{-1}$) as L_n becomes larger with respect to L_d. A corollary of this statement is that the sediment-delivery ratio will approach 100 per cent as the travel distances of particles approaches infinity, as they may do for very small particles. Not surprisingly, therefore, sediment delivery ratios are closest to this value for catchments in loess (e.g. Walling, 1983).

Of course, the simplified case presented above is unrealistic. However, the assumption that particle size is uniform is inconsequential. Integrating the sediment flux over a range of particle sizes would affect only the critical value of L_n. For any value of $L_n > L_{dmin}$, where *dmin* is the travel distance of the particle that travels least far, the erosion rate will decline as L_n becomes larger, but at a slower rate than in the case of a single particle size until $L_n = L_{dmax}$, where *dmax* is the maximum travel distance of any particle. The other two assumptions, however, have greater implications for erosion rates.

In reality, entrainment and travel distances are not spatially uniform, so that the sediment flux ($M\ T^{-1}$) of particles size d, $\varphi_d(x)$ will vary with distance downslope x such that

$$\frac{d\varphi_d}{dx}(x) = E_d(x) - D_d(x) \tag{1}$$

where $E_d(x)$ is the rate of entrainment ($M\ T^{-1}$) and $D_d(x)$ the rate of deposition ($M\ T^{-1}$) of particles of size d. (Note: we have written φ_d, E_d, D_d explicitly as functions of distance downslope x, to emphasize our assumption of non-steady-state fluxes.) If it is assumed that all particles of size d that are entrained at a given point travel the same distance L downslope before being deposited, then $D_d(x)$ in Equation 1 can be replaced by $E_d(x - L_d)$ to give

$$\frac{d\varphi_d}{dx}(x) = E_d(x) - E_d(x - L_d). \tag{2}$$

Equation 2 can be regarded as a general model for sediment flux under erosion by water. However, to examine further the significance of spatial variation in entrainment and travel distances, it is necessary to examine separately the two forms of soil erosion by water: interrill erosion, where sediment entrainment is by raindrop impact, and transport is by flow; and rill and gully erosion, where both entrainment and transport are due to flow.

Interrill sediment flux

For interrill erosion, entrainment is generally accepted to decrease with depth of overland flow (Gilley *et al*., 1985). A more formal relationship for interrill entrainment E ($m^3\ s^{-1}\ m^{-1}$) is given by Torri *et al*. (1987) as

$$E = E_0 e^{-bh^{0\cdot67}} \tag{3}$$

where E_0 is the entrainment rate at zero depth, h is water depth, and b is an exponent (≈2, according to Torri *et al*., 1987, and Morgan *et al*., 1998). Unfortunately, Torri *et al*. do not give units for h, although the diagrammatic representation (Torri *et al*., Figure 2) indicates that it is measured in millimetres. Morgan *et al*., however, state that it is measured in metres. Testing the equation shows that, if metres are used, detachment at 0·1 m depth is still 65 per cent of that at zero depth, whereas using millimetres shows it to be down to 14 per cent at 1 mm depth. Neither result is consistent with research on the effect of flow depth on soil detachment by raindrops (see, for example, Moss and Green, 1983; Kinnell, 1991). If h is measured in centimetres, however, results are reasonably consistent with such research. For this paper, therefore, the units of h will be taken to be centimetres.

Studies by Meyer (1981) have shown that that

$$E_0 \approx kR^2 \tag{4}$$

where R is rainfall energy ($J\ m^{-2}\ s^{-1}$) and k is a constant that varies with soil type. Consequently, entrainment can be approximately expressed as a function of rainfall energy and flow depth:

$$E = kR^2 e^{-2h} \tag{5}$$

Investigation of travel distances of particles in shallow, rain-impacted flow by Parsons *et al*. (1998) produced

$$L = 0{\cdot}525 R^{2\cdot35} \times F^{0\cdot981} \times M^{-1} \tag{6}$$

for the median travel distance, where L is travel distance (cm/min), F is flow energy ($J\ m^{-2}\ s^{-1}$) and M is particle mass (g). These experiments, however, were conducted using particles in excess of 2·88 mm diameter, which is larger than those typically transported in interrill flow. To test the validity of this equation on particles more commonly transported in interrill flow, the data on travel distance of magnetite particles of mean size of 80 μm in a 56-minute storm (Parsons *et al*., 1993) have been used. The result is an estimate of mean travel distance for the magnetite particles of 1·43 m compared to an actual value of 1·74 m (based upon measurements of change in magnetic susceptibility). Thus extrapolating Equation 6 to particles two orders of magnitude smaller than those on which it was developed yields an estimate that is well within the same order of magnitude as that observed. This is a reassuringly good result. Although this is a limited evaluation, it does appear that Equation 6 can be used to estimate travel distances of small particles, at least to a first approximation.

Initially, consider interrill flow on a plan-planar hillslope of uniform gradient, infiltration and hydraulic roughness (such as may characterize an erosion plot). If it is assumed that spatially uniform rain falls onto this hillslope at a rate greater than the infiltration capacity of the soil then, once equilibrium is attained, overland flow discharge q ($m^3\ s^{-1}$) will increase linearly with distance downslope x (m). If the kinematic model (Stephenson and Meadows, 1986) and the Manning friction equation are accepted, then flow depth may be assumed to increase with $x^{2/3}$. Correspondingly, flow velocity will increase with $x^{1/3}$ and, hence, flow energy ($0{\cdot}5\ mv^2$) will increase with $x^{5/3}$. Consequently, for particles of a given mass and assuming $F^{0\cdot981} \approx F^1$, Equations 5 and 6 can be rewritten as

$$E = K_1 e^{-2x^{4/9}} \tag{7}$$

and

$$L = K_2 x^{5/3} \tag{8}$$

where K_1 and K_2 subsume the rainfall terms in Equations 3 and 4. Since L strictly increases with x, it follows (assuming uniform travel distances) that all particles entrained upslope of any given x will be deposited upslope of $x + L(x)$, and that all particles entrained downslope of x will be deposited downslope of $x + L(x)$. Hence from Equation (1)

$$\frac{d\varphi_d}{dx}(x) = E_d(x) - D_d(x) \tag{9}$$

giving

$$\varphi(x+L) = \int_0^{x+L} E_d(u)du - \int_0^{x+L(x)} D_d(u)du = \int_0^{x} E_d(u)du + \int_x^{x+L(x)} E_d(u)du - \int_0^{x+L(x)} D_d(u)du = \int_x^{x+L(x)} E_d(u)du \tag{10}$$

since, as was argued before: $\int_0^{x} E_d(u)du = \int_0^{x+L(x)} D_d(u)du$. Consequently:

$$\varphi(x+L) = \int_x^{x+L(x)} K_1 e^{-2u^{4/9}} du \tag{11}$$

Unfortunately, the integral in Equation 11 is not analytically solvable. However, we can run numerical simulations for different particle sizes. Assuming a particle density of 2·65 Mg m^{-3}, a rainstorm of 30 mm/h (with a Marshall–Palmer drop size distribution), a value for K_1 of 0·000121 and values for K_2 of $1{\cdot}586 \times 10^{-5}$, $1{\cdot}586 \times 10^{-2}$ and 15·86 for particles 1 mm, 0·1 mm and 0·01 mm diameter, respectively, produces estimates of the downslope pattern of fluxes of these particles as shown in Figure 1. In all cases, sediment flux rises to a maximum at a relatively short distance downslope, and declines slowly thereafter. The flux rate varies directly with particle size and maximum flux rate is displaced progressively downslope for increasing particle size. Thus, even on a uniform gradient, the flux of sediment that is recorded varies as a function of the distance downslope. This equation predicts that, contrary to the assumption of a direct relationship between the length of the contributing area and the sediment flux, both direct and inverse relationships may exist, and whether the relationship is direct or inverse is a function of total contributing area and the particle size.

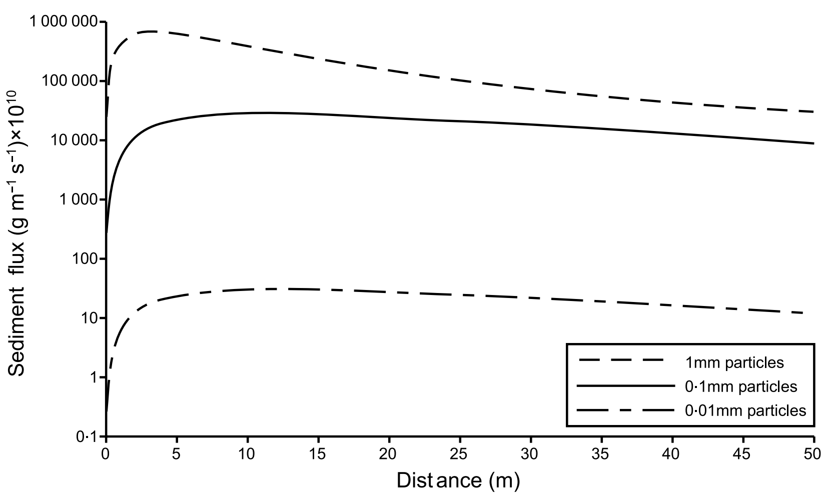

Figure 1. Relationship between interrill sediment flux and distance downslope on hillslopes of uniform gradient, assuming spatially uniform rainfall and infiltration, and unlimited sediment supply

Rill and gully sediment fluxes

For rill and gully erosion both sediment entrainment and travel distance are functions of flow. Flow will increase with catchment area rather than linearly with distance, so that a suitable relationship between flow Q ($m^3\ s^{-1}$) and distance x (m) is that developed by Hack (1957):

$$Q \propto x^{1\cdot67} \tag{12}$$

Experimental data on equilibrium sediment transport rates under conditions of an unlimited supply of entrainable sediment can be used to estimate sediment entrainment in rills and gullies. From Yalin (1977), entrainment may be approximated by

$$E \propto \tau^{1\cdot5} \tag{13}$$

where τ is bed shear stress. Since

$$\tau = \rho g d s \tag{14}$$

where ρ ($Mg\ m^{-3}$) is density of water, g ($m\ s^{-2}$) is acceleration due to gravity, d is flow depth (m) and s is slope ($m\ m^{-1}$), on uniform gradients

$$\tau \propto d \tag{15}$$

Abrahams *et al*. (1996) have inferred that for rills on agricultural land

$$d \propto Q^{0\cdot4} \tag{16}$$

Conveniently, the exponents in the above set of equations lead to

$$E \propto x \tag{17}$$

Using Hassan *et al*. (1992), travel distance can be expressed by the relationship

$$L \propto (\omega - \omega_0)^{1\cdot31} D_g^{-0\cdot94} \tag{18}$$

where L is mean distance of travel (m), $(\omega - \omega_0)$ is excess stream power ($W\ m^{-2}$) and D_g is the geometric mean size (mm) of the particles in transit. For particles of known size on a uniform gradient and where stream power is well in excess of the critical stream power ω_0, $\omega - \omega_0$ can be approximated by ω (which, in turn, is a function of discharge Q) so that we may approximate Equation 18 to

$$L \propto Q^{1\cdot31} \tag{19}$$

Consequently:

$$L \propto x^{2\cdot18} \tag{20}$$

As for interrill erosion, because L increases with x, it follows (assuming uniform travel distances) that all particles entrained upslope of any given x will be deposited upslope of $x + L$, and that all particles entrained downslope of x will be deposited downslope of $x + L$. Hence for rill and gully erosion:

$$\varphi(x + L) = \int_x^{x+L(x)} K_3 u\, du \tag{21}$$

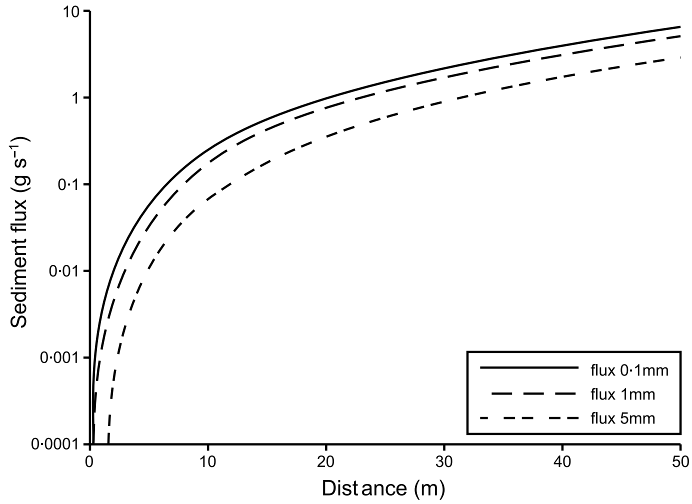

Figure 2. Relationship between sediment flux and distance in rills and gullies of uniform gradient, assuming spatially uniform rainfall and infiltration, and unlimited sediment supply

which yields immediately

$$\varphi(x+L) = 1/2K_3(x+L)^2 - K_3x^2 \tag{22}$$

and then

$$\varphi(x+L) = 1/2K_3(2xL(x) + L(x)^2 \tag{23}$$

Using Equation 23 and assuming the same rainstorm of 30 mm/h delivering all rainfall into a rill or gully of uniform gradient (0·1 m m^{-1}) and with Manning's n of 0·03, estimates of downslope patterns of sediment fluxes for particles 0·1, 1 and 5 mm are shown in Figure 2. Sediment flux increases very rapidly with distance but, as in the case of interrill erosion, the flux decreases with particle size. In this case, assuming a uniform gradient, sediment flux does vary directly with contributing area, but the relationship is not a linear one.

EROSION RATES

For both interrill and rill and gully erosion, sediment flux exhibits pronounced spatial variability. Consequently, whenever erosion rates are calculated by dividing measured sediment flux by the distance upslope of the measuring point, the rate will depend on the point of measurement (Figure 3). For interrill erosion, measurements made over short run-off plots will grossly overestimate erosion for whole hillslopes. For rill erosion, the converse is the case: the greater the distance downrill that the measurement is made, the greater the apparent erosion rate.

EMPIRICAL EVIDENCE

Data on sediment fluxes at different distances along the same slope are few. Because it has been assumed that sediment leaving the bottom of run-off plots is directly proportional to the size of the plot, few authors have attempted to measure sediment fluxes within plots, or to examine the relationship between sediment flux and plot size within reasonably homogeneous areas. However, measurements obtained simultaneously at 6, 12 and 20·5 m from the top of a large run-off plot in Arizona and at its outlet (Parsons *et al.*, 1996) gave fluxes of 2·92, 11·32, 14·01 and 16·13 g m^{-1} min^{-1}, corresponding to areal erosion rates of 0·49, 0·94, 0·68 and 0·66 g m^{-2} min^{-1}.

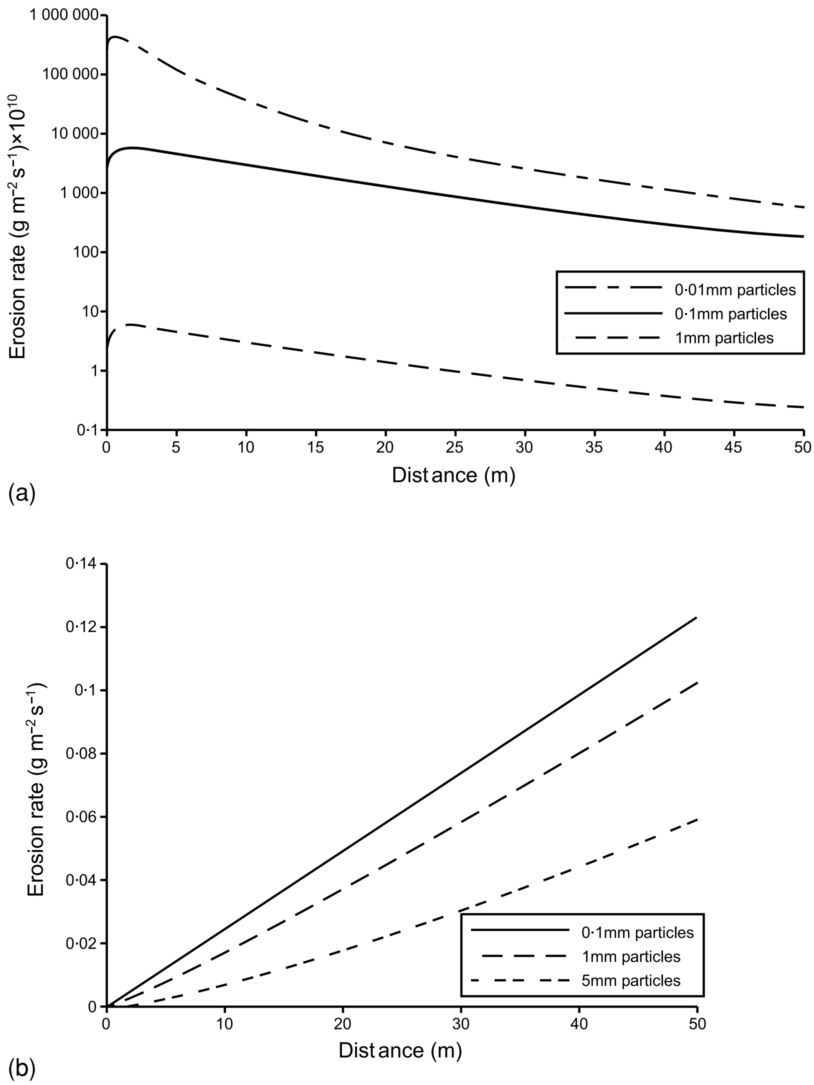

Figure 3. Relationship between erosion rate and distance for (a) interrill areas and (b) in rills and gullies, under the same assumptions as those of Figures 1 and 2

Although sediment flux increased throughout the plot length, the rate of increase declined in the lower part so that erosion rate peaked somewhere between 12 and 20·5 m downslope from the top of the plot. Elsewhere, Rejman *et al*. (1999) have determined erosion rates of 4·875, 4·325 and 2·233 kg m^{-2} for plots of 5, 10 and 20 m length on a 12 per cent slope. Additionally, Wilcox *et al*. (1996) observed that where channel erosion was not present, erosion decreased as slope length increased, but conversely data from sites with channels showed increased rates of erosion with slope length. All of these data provide empirical support for the theoretical model presented above, even though the model takes no account of the spatial variability present in natural overland flow (see, for example, Abrahams *et al*., 1986; Parsons *et al*., 1990).

CONCLUSION

The model presented here has considered only simple cases of erosion under spatially uniform rainfall and on slopes of uniform infiltration and gradient. Furthermore, it has taken no account of supply limitations to sediment flux, nor of sediment transport by other processes than overland flow. (Splash transport, for example, will mean that sediment transport will not tend to zero as downslope distance tends to zero.) Likewise, it has considered interrill erosion and rill and gully erosion separately, whereas on many hillslopes both processes may operate. Nevertheless, the analysis is sufficient to demonstrate that erosion rates calculated in the manner hitherto adopted have no value for extrapolation because they depend on the areas over which they were made. It also provides an explanation for the claimed European erosion rate of 17 Mg ha^{-1} a^{-1} and its incompatibility with the known long-term landform evolution. The data on which this rate is based was obtained from interrill run-off plots 22 m long (Bollinne, 1985).

In contrast, a framework based upon the concepts of entrainment and travel distances of particles, and hence on sediment flux, is not dependent on area of measurement. In demonstrating the viability of this framework, we have employed the limited data in the literature to yield relationships between sediment flux and downslope distance. More specific parameterization of the equations to take account of such factors as gradient and sediment supply is required to proceed from the conceptual framework to using it to provide quantitative measurements of erosion. Our conceptual approach solves the apparent paradox of the sediment-delivery ratio, resolves recent discussion about the validity of erosion rates made using USLE erosion plots (Trimble and Crosson, 2000), and potentially can reconcile erosion rates with known lifespans of continents. Our results imply that previous estimates of soil erosion are fallacious. Those that are typically based on measurement of interrill erosion are likely to be significant overestimates. On the other hand, the significance of travel distance increases the importance of rill and gully erosion and of the transport of fine particles to which nutrients and pollutants preferentially adhere. As noted above, empirical data against which to test this model are sparse. What is required is data from a range of scales, from small plots to hillslopes and catchments against which to test its predictions.

ACKNOWLEDGEMENT

This research has been funded by the Natural Environment Research Council (grant GR3/12754).

REFERENCES

Abrahams AD, Parsons AJ, Luk S-H. 1986. Resistance to overland flow on desert hillslopes. *Journal of Hydrology* **88**: 343–363.

Abrahams AD, Parsons AJ, Luk S-H. 1991. The effect of spatial variability in overland flow on the downslope pattern of soil loss on a semi-arid hillslope, southern Arizona. *Catena* **18**: 255–270.

Abrahams AD, Li G, Parsons AJ. 1996. Rill hydraulics on a semiarid hillslope, southern Arizona. *Earth Surface Processes and Landforms* **21**: 35–47.

Boardman J. 1998. An average soil erosion rate for Europe: myth or reality? *Journal of Soil and Water Conservation* **53**: 46–50.

Bollinne A. 1985. Adjusting the universal soil loss equation for use in Western Europe. In *Soil Erosion and Conservation*, El-Swaify SA, Moldenhauer WC, Lo A (eds). Soil Conservation Society of America: Ankeny, IA; 206–213.

Clark SP, Jäger E. 1969. Denudation rate in the Alps from geochronologic and heat flow data. *American Journal of Science* **267**: 1143–1160.

Crosson P. 1995. Soil erosion estimates and costs. *Science* **269**: 461–464.

Dedkov AP, Moszherin VI. 1992. Erosion and sediment yield in mountain regions of the world. *International Association of Hydrological Sciences Publication* **209**: 29–36.

Evans R. 1995. Some methods of directly assessing water erosion of cultivated land: a comparison of measurements made on plots and in fields. *Progress in Physical Geography* **19**: 115–129.

Gilley JE, Woolhiser DA. McWhorter DB. 1985. Interrill soil erosion. Part II: Testing and use of model equations. *Transactions of the American Society of Agricultural Engineers* **28**: 154–159.

Glymph LM. 1954. Studies of sediment yield from watersheds. *International Association of Hydrological Sciences Publication* **36**: 261–268.

Graf WL. 1988. *Fluvial Processes in Dryland Rivers*, Springer-Verlag: Berlin.

Hack JC. 1957. Studies of longitudinal stream profiles in Virginia and Maryland. *US Geological Survey Professional Paper* 294B.

Hassan M, Church M, Ashworth PJ. 1992. Virtual rate and mean distance of travel of individual clasts in gravel-bed channels. *Earth Surface Processes and Landforms* **17**: 617–627.

Kinnell PIA. 1991. The effect of flow depth on sediment transport induced by raindrops impacting shallow flows. *Transactions of the American Society of Agricultural Engineers* **34**: 161–168.

Kirkby MJ. 1991. Sediment travel distance as an experimental and model variable in particulate movement. *Catena Supplement* **19**: 111–128.

Meyer LD. 1981. How rain intensity affects interrill erosion. *Transactions of the American Society of Agricultural Engineers* **24**: 1472–2475.

Morgan RPC, Quinton JN, Smith RE, Govers G, Poesen JWA, Auerswald K, Chisci G, Torri D, Styczen ME. 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 AJ, Green P. 1983. Movement of solids in air and water by raindrop impact. Effects of drop-size and water-depth variations. *Australian Journal of Soil Research* **21**: 257–269.

Parsons AJ, Abrahams AD. Luk S-H. 1990. Hydraulics of interrill overland flow on a semi-arid hillslope, southern Arizona. *Journal of Hydrology* **117**: 255–273.

Parsons AJ. Stromberg SGL. 1998. Experimental analysis of size and distance of travel of unconstrained particles in overland flow. *Water Resources Research* **34**: 2377–2381.

Parsons AJ, Stromberg, SGL, Greener M. 1998. Sediment-transport competence of interrill overland flow. *Earth Surface Processes and Landforms* **23**: 365–375.

Parsons AJ, Wainwright J, Abrahams AD. 1993. Tracing sediment movement in interrill overland flow on a semi-arid grassland using magnetic susceptibility. *Earth Surface Processes and Landforms* **18**: 721–732.

Parsons AJ, Wainwright J, Abrahams AD. 1996. Runoff and erosion on semi-arid hillslopes. In *Advances in Hillslope Processes*, Anderson MG, Brooks SM (eds). Wiley: Chichester; 1061–1078.

Pimental D, Harvey C, Resosudarmo P, Sinclair K, Kurz D, McNair M, Crist S, Shpritz L, Fitton L, Saffouri R, Blair R. 1995. Environmental and economic costs of soil erosion and conservation benefits. *Science* **267**: 1117–1122.

Rejman J, Usowicz B, Debicki R. 1999. Source of errors in predicting silt soil erodibility with USLE. *Polish Journal of Soil Science* **32**: 13–22.

Renard KG, Foster GR, Weesies GA, McCool DK, Yoder JC. 1997. *Predicting Soil Erosion by Water: A Guide to Conservation Planning with the Revised Universal Soil Loss Equation (RUSLE)*. Handbook 703, US Department of Agriculture.

Robinson AR. 1977. Relationship between soil erosion and sediment delivery. *International Association of Hydrological Sciences Publication* **122**: 159–167.

Stephenson D, Meadows ME. 1986. *Kinematic Hydrology and Modelling*. Elsevier: Amsterdam.

Summerfield MA, Hulton NJ. 1994. Natural controls of fluvial denudation rates in major world drainage basins. *Journal of Geophysical Research* **B7**: 13871–13863.

Torri D, Sfalanga M, del Sette M. 1987. Splash detachment: runoff depth and soil cohesion. *Catena* **14**: 149–155.

Trimble SW, Crosson P. 2000. US soil erosion rates: myth and reality. *Science* **289**: 248–250.

Wainwright J, Thornes JB. 1991. Computer and hardware modelling of archaeological sediment transport on hillslopes. In *Computer Applications and Quantitative Techniques in Archaeology 1990*, Rahtz S, Lockyear K (eds). BAR International. Series 565: Oxford; 183–194.

Walling DE. 1983. The sediment delivery problem. *Journal of Hydrology* **65**: 209–237.

Walling DE, Webb BW. 1996. Erosion and sediment yield: a global view. *International Association of Hydrological Sciences Publication* **236**: 3–19.

Wilcox BP, Newman BD, Allen CD, Reid KD, Brandes D, Pitlick J, Davenport DW. 1996. Runoff and erosion on the Pajarito Plateau: observations from the field. *New Mexico Geological Society Guidebook, 47th Field Conference, Jemez Mountain Region*. New Mexico Geological Society: Socorro, NM; 433–439.

Yalin MS. 1977. *Mechanics of Sediment Transport*. Pergamon Press, Oxford.

Yang D, Kanae S, Oki T, Koike T, Musiake K. 2003. Global potential soil erosion with reference to land use and climate changes. *Hydrological Processes* **17**: 2913–2928.

