Erosion by Water: Erosivity and Erodibility

Peter I. A. Kinnell

School of Resource, Environmental and Heritage Sciences, University of Canberra, Holt, Australian Capital Territory, Australia

Published online: 29 May 2013


To link to this chapter: http://dx.doi.org/10.1081/E-EEM-120046415

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Erosion by Water: Erosivity and Erodibility

Peter I.A. Kinnell
School of Resource, Environmental and Heritage Sciences, University of Canberra, Holt, Australian Capital Territory, Australia

Abstract
Conceptually, rainfall erosivity is the capacity of rain to produce erosion, whereas soil erodibility is the susceptibility of the soil to be eroded. However, no absolute numerical values can be determined for them because rainfall erosion results from various forms of erosion (splash erosion, sheet erosion, rill erosion, interrill erosion) that are driven by different forces. Originally, the terms erosivity and erodibility were associated with the rainfall factor \( R \) and the soil factor \( K \) in the Universal Soil Loss Equation where the event erosivity factor was given by the product of the kinetic energy of the storm \( E \) and the maximum 30 min rainfall intensity \( I_{30} \). Soil loss at the scale at which the USLE applies results from the discharge of sediment with runoff, but because the USLE does not include any direct consideration of runoff in the event erosivity factor, it is not good at accounting for event soil loss at some geographic locations. Erodibility values have units of soil loss per unit of the erosivity index and cannot be used with erosivity indices other than the one that was associated in their determination. Some models ignore this fact and so ignore certain fundamental rules that should apply to the modelling of water erosion. So-called process-based models attempt to account for the effects of the various forms of erosion and do consider the effect of runoff on erosivity associated with each form. Particles travel across the soil surface at virtual velocities that vary from the velocity of the flow to near zero. Little regard is given to this fact in the determination of soil erodibility values.

INTRODUCTION
Conceptually, rainfall erosivity is the capacity of rain to produce erosion, whereas soil erodibility is the susceptibility of the soil to be eroded. Historically, the terms erosivity and erodibility were originally associated with the \( R \) and \( K \) factors in the Universal Soil Loss Equation (USLE),

\[
A = R K L S C P
\]

where \( A \) is the long-term (e.g., 20 years) annual average soil loss per unit area from sheet and rill erosion, \( R \) is the rainfall (erosivity) factor defined as the average annual value of the product of the total storm kinetic energy \( E \) and the maximum 30 min intensity \( I_{30} \), twice the maximum amount of rain that falls in any 30 min period during a storm, \( K \) is the soil (erodibility) factor, \( L \) is the slope length factor, \( S \) is the slope gradient factor, \( C \) is the crop (vegetation) and crop management factor, and \( P \) is the conservation support practice factor.\(^1\) Numerically, soil erodibility is the mass of soil eroded per unit of the erosive index. This means that numerical values of \( K \) can only be used when \( R \) is as it was originally defined, the average annual value of the product of \( E \) and \( I_{30} \).

The USLE was designed to predict sheet and rill erosion from field sized areas and only \( R \) and \( K \) have units. The \( L \), \( S \), \( C \), and \( P \) factors each have values of 1.0 for the so-called “unit” plot, a bare fallow area 22.1 m long on a 9% slope with cultivation up and down the slope. Consequently, the soil loss for the “unit” plot \( (A_1) \) is given by

\[
A_1 = R K
\]

and, for any other situation,

\[
A = A_1 L S C P.
\]

Although, traditionally, \( K \) is calculated by dividing \( A_1 \) by \( R \), basically \( K \) can be perceived as the regression coefficient in the direct relationship between event soil loss on the unit plot \( (A_1) \) and \( EI_{30} \).

To reduce the need to run long-term experiments to determine \( K \) values for soils where \( K \) is unknown, Wischmeier\(^2\) developed a nomograph for determining \( K \) from soil properties for soils with less then 70% silt in the United States. Alternatively, \( K \) values in customary U.S. units for soils where the nomograph can be used may be obtained using

\[
K = (2.1 X_1^{1.14} 10^{-4} (12 - X_2) + 3.25 (X_3 - 2) + 2.5 (X_4 - 3)) / 100,
\]

where \( X_1, X_2, X_3, X_4 \) are\(^'\)s.

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where $X_1$ is % silt multiplied by 100 – % clay, $X_2$ is % organic matter, $X_3$ is the soil structure code used in the U.S. soil classification, and $X_4$ is the profile permeability code.

A number of other equations have been developed for soils at various geographic locations (e.g., El-Swaify [3] Young and Mutchler [4], and Zhang et al. [6]), but Eq. 4 is frequently used outside the United States without being validated for the soils involved. The two-staged mathematical approach shown by Eqs. 2 and 3 results from the fact that the USLE is an empirically based model. It was developed in the 1960s and 1970s from more than 10,000 plot years of data. Mathematically, the Revised Universal Soil Loss Equation (RUSLE [7]) uses Eqs. 2 and 3 in the same way as the USLE, but changes were made to how some of the factors in the model are calculated. Originally, in the USLE, the events used to calculate $R$ were restricted to those that produced more than 12.5 mm of rain or at least 6.25 mm of rain in 15 min. That rule was abandoned in the RUSLE when $R$ values were calculated for the western part of the United States because it was argued that the rule had no appreciable effect on $R$ values. Yu [8] noted that changing the threshold to zero increased the $R$ factor by 5% in the tropical region of Australia.

**VARIANTS OF THE USLE**

Eq. 2 operates on the assumption that a direct linear relationship exists between event erosion ($A_e$) and $EI_{30}$. Although this assumption is appropriate at some geographic locations, it is not appropriate at others (Fig. 1). Soil measured as a loss from the plots used to develop the USLE

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**Fig. 1** The relationships between event erosion ($A_e$) and $EI_{30}$ obtained on bare fallow plots in the United States: (a) Plot 5 Experiment 3 at Holly Springs, Mississippi, and (b) Plot 5 Experiment 1 at Morris, Minnesota. $\text{Eff(ln)}$ is the Nash–Sutcliffe [9] efficiency factor for logarithmic transforms of the data.
was discharged in runoff. When runoff occurs, the amount of soil discharged ($Q_s$) in the runoff can be considered to be the product of the amount of water discharged ($Q_w$) and the sediment concentration ($c_s$), the amount of soil per unit quantity of water,

$$Q_s = Q_w c_s$$  \hspace{1cm} (5)

It follows from Eq. 5 that, in the USLE, the sediment concentration is directly proportional to $EI_{30}$ divided by runoff amount. However, it has been observed\(^\text{[10]}\) that the sediment concentration for an event at some geographic locations is correlated to the product of the kinetic energy per unit quantity of rain and $I_{30}$ (Fig. 2). This results in an event erosivity index ($R_e$) that is given by

$$R_e = Q_Re EI_{30}$$  \hspace{1cm} (6)

where $Q_Re$ is the runoff ratio (runoff amount divided by rainfall amount) for the event. This index is more effective in accounting for event soil losses at some locations where the $EI_{30}$ index does not work well (Fig. 3).

A number of other event erosivity indices have been proposed. Williams\(^\text{[11]}\) proposed a modification of the USLE to predict event sediment yield ($SY_e$),

$$SY_e = 11.8 \left( q_e q_{p_e} \right)^{0.56} K L S C_e P_e$$  \hspace{1cm} (7)

where $q_e$ is the volume of runoff (m$^3$) for the event; $q_p$ is the peak flow rate (m$^3$/sec); $K$, $L$, and $S$ are standard USLE factors; and $C_e$ and $P_e$ are event $C$ and $P$ factors. This model is commonly known as the Modified Universal Soil Loss Equation (MUSLE). In APEX,\(^\text{[12]}\)

$$SY_e = X_e K L S C_e P_e \text{[RKOF]}$$  \hspace{1cm} (8)

$X_e$ is selected from

$$X_e = EI_{30}$$  \hspace{1cm} (9a)

$$X_e = 1.586 \left( q_e q_{p_e} \right)^{0.56} DA^{0.12}$$  \hspace{1cm} (9b)

$$X_e = 0.65 EI_{30} + 0.45 \left( q_e q_{p_e} \right)^{0.33}$$  \hspace{1cm} (9c)

$$X_e = 2.5 \left( q_e q_{p_e} \right)^{0.5}$$  \hspace{1cm} (9d)

$$X_e = 0.79 \left( q_e q_{p_e} \right)^{0.65} DA^{0.009}$$  \hspace{1cm} (9e)

$$X_e = b_5 q_e^{b_6} q_{p_e}^{b_7} DA^{b_8}$$  \hspace{1cm} (9f)

where $DA$ is drainage area expressed in hectares, $b_6$–$b_8$ are user-selected coefficients,\(^\text{[12]}\) and RKOF is the coarse fragment factor as defined by Simanton et al.\(^\text{[13]}\) However, Eqs. 7 and 8 with Eqs. 9b–9f all use USLE $K$ factor values and not ones that are associated with the different erosivity indices used and so do not conform with the mathematical modeling rules upon which the USLE model is based. Consequently, the MUSLE and APEX models are not valid variations of the USLE. In addition, any model that uses event erosivity index values that are calculated using runoff from a vegetated area or any area that is not cultivated up and down the slope will violate the mathematical rules if it uses USLE $C$ and $P$ factor values.

Mathematical models like the USLE and its derivatives are largely designed to aid management decisions and operate at a level where spatial and temporal variations in the various forms of erosion (splash erosion, sheet erosion, rill erosion, interrill erosion) are not considered in any appreciable detail. However, rill erosion is driven by flow energy while sheet and interrill erosion are associated more closely with rainfall kinetic energy. In order to better account for this, Onstad and Foster\(^\text{[14]}\) used the equation

![Fig. 2](image-url) The relationship between event sediment concentration ($c_e$) and the product of $I_{30}$ and the kinetic energy per unit quantity of rain ($E/r_e$) for Plot 5 Experiment 1 at Morris, Minnesota.
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\[ R_e = 0.5EI_{30} + 0.5\alpha(q_e/q_p)^{0.333} \]  \hspace{1cm} (10)

Eq. 10 indicates that the sediment concentration for an event \( (c_e) \) in the data set considered by Onstad and Foster\cite{14} is given by

\[ c_e = 0.5[EI_{30}/q_e] + 0.5[\alpha(q_p^{0.333}/q_e^{0.666})]. \]  \hspace{1cm} (11)

The value of \( \alpha \) in Eq. 10 was set so that the long-term average value of \( R_e \) produced by Eq. 10 was the same as the long-term average value produced when \( R_e = EI_{30} \). Given that \( K \) is, by definition, the amount of soil loss per unit of \( R \) when \( R_e = EI_{30} \), this enabled Eq. 10 to be used with USLE \( K \) values to predict soil loss from bare fallow areas. Although Williams et al\cite{15,16} indicated that Onstad and Foster\cite{14} was the source of Eq. 9c, there is no provision in APEX to do the same.

Bagarello et al\cite{17} observed that

\[ R_e = (Q_{Re}EI_{30})^\beta \]  \hspace{1cm} (12)

with \( \beta > 1.0 \) applied to soil losses from a number of simultaneously operating plots of different length \( (\lambda) \) established at the experimental station of Sparacia, Sicily. Subsequent analysis\cite{18} (in review) established that \( \beta > 1.0 \) appeared to be the result of an interaction between event runoff \( (Q_e) \) and slope length \( (\lambda) \) on sediment concentration so that

\[ R_e = Q_{Re}EI_{30}Q_e^{0.0207\lambda} \]  \hspace{1cm} (13)

when \( \lambda \) is in meters.

Arguably, from a physical viewpoint, since slope length and gradient influence the erosive stress, they can be considered to be factors that influence erosivity but the USLE model is not designed to model the physical processes themselves. It is designed to predict rainfall erosion based on climate, soil, topography, crops, and management factors. Normally, in the USLE model, the effect of slope length on soil loss is expressed through the \( L \) factor, but in the case of Eq. 13, \( L \) would remain at 1.0 irrespective of variations in slope length. The positive effect of runoff on sediment concentration may have been associated with the fact that the data were collected on a 15% slope. Rainfall erosion on much lower slope gradients is not associated with as high an erosion stress from runoff as likely to have been the case at the Sparacia site. In fact, the RUSLE makes provision for the cushioning effect of water depth on the energy of raindrop impact on low slopes.

**RAINFALL KINETIC ENERGY**

Data on storm rainfall kinetic energy are not measured at many geographic locations, and often, kinetic energy values are obtained indirectly. If data on storm rainfall intensities are available, then it is common for storm kinetic energies to be estimated from rainfall kinetic–intensity relationships. These relationships are based on raindrop-size data collected during rainfall events at a geographic location assumed to have rainfall characteristics that are consistent with those at the geographic location of interest. Various techniques have been employed to do this, and these have generated rainfall intensity–kinetic relationships which are often used at locations that lie outside the climatic zones where the measurements were originally made. In the USLE, the relationship recommended between the kinetic energy per unit quantity of rain \( (e) \) and rainfall intensity \( (I) \) for the United States is expressed by
\[ e = 916 + 331 \log_{10} I, \]
\[ \text{for } I < 3 \text{ in./hr} \]  
\[ e = 1074, \]
\[ \text{for } I \geq 3 \text{ in./hr} \]  
\[ \text{(14a)} \]
\[ \text{(14b)} \]
where \( e \) is in units of foot-tons per acre per inch and \( I \) is in inches per hour, following analysis of drop-size data collected by Law and Parsons.\[^{19}\] In the RUSLE, Eq. 14 is replaced by
\[ e = 1099 \left(1 - 0.72 \exp\left[-1.27I\right]\right) \]  
\[ \text{(15)} \]
whose metric equivalent is
\[ e_m = 0.29 \left(1 - 0.72 \exp\left[0.05I_m\right]\right) \]  
\[ \text{(16)} \]
where \( e_m \) has units of megajoules per hectare per millimeter and \( I_m \) has units of millimetres per hour. Eqs. 15 and 16 use the mathematical form proposed by Kinnell.\[^{20,21}\] Yu et al.\[^{8}\] observed that replacing Eq. 14 by Eq. 15 reduced the \( R \) factor by about 10% in the tropical region of Australia. A number of other mathematical equations have been reviewed by van Dijk et al.\[^{22}\] Some of these produce negative values of rainfall kinetic energy at low rainfall intensities. Eq. 16 produces a response curve that increases non-linearly from low values at low intensities with \( e_m \) values remaining close to 0.29 MJ/ha/mm for intensities beyond 75-100 mm/hr. However, short-term values measured during rainstorms vary greatly about the values produced by Eq. 16 or any other rainfall intensity–e relationship. Also, although Eqs. 15 and 16 were developed for use in the United States, they have been used in other parts of the world without validation. In many cases, that results in erroneous values. In effect, storm kinetic energies calculated from rainfall intensity–kinetic energy relationships produce numerical values that are biased towards high-intensity rainfall at the expense of low-intensity rainfall.

In many geographic locations, there is a lack of rainfall intensity data so that it is not possible to determine storm kinetic energies using rainfall kinetic energy–intensity relationships. Event or daily rainfall amounts are more commonly recorded. One approach that has been used in a number of geographic areas such as Canada,\[^{23}\] Finland,\[^{24,25}\] Italy,\[^{26}\] and Australia\[^{27}\] considers that \( EI_{30} \) is related to a power of event rainfall amount (\( X_p \)),
\[ EI_{30} = a_1 X_p^{b_1} \]  
\[ \text{(17)} \]
where \( a_1 \) and \( b_1 \) are empirical constants. The value of \( a_1 \) may show seasonal variation.\[^{25,26}\] Given that, in the context of the criteria associated with the USLE/RUSLE model, daily rainfall provides a reasonable proxy for storm rainfall amount\[^{28}\]; Eq. 17, using daily rainfall amount, provides a practical approach to extending observed \( EI_{31} \) values to areas where appropriate rainfall intensity data are lacking.

**MORE PROCESS-BASED MODELS**

Although the USLE/RUSLE is the most widely used method of predicting soil losses from the land worldwide, rainfall erosion results from various forms of erosion (splash erosion, sheet erosion, rill erosion, interrill erosion) that are driven by different forces so that there is no absolute measure of either rainfall erosivity or soil erodibility. Consequently, more process-based models have been developed in order to predict the contributions of the various forms of erosion more directly. Often, the forms vary in a topographic sequence with splash erosion dominating erosion at the upper end of a slope and sheet erosion further down before areas of rill and interrill erosion. Particles detached at the top of the slope may be transported by a number of different transport mechanisms before being finally discharged from the eroding area. The Water Erosion Prediction Program in the United States generated the WEPP model, a more process-based model than the USLE/RUSLE designed to model the spatial contributions of rill and interrill erosion in agricultural landscapes.\[^{30}\] Flow shear stress (\( \tau \)) is used as the erosivity factor in rill erosion model,
\[ D_r = k_r (\tau - \tau_r) (1 - q_{sd} T_c) \]  
\[ \text{(18)} \]
where \( D_r \) is rill detachment, \( k_r \) is the rill erodibility factor, \( \tau_r \) is the critical shear stress that has to be exceeded before detachment occurs, \( q_{sd} \) is the sediment load in the flow, and \( T_c \) is the sediment load at the transport limit. For erosion to occur, particles must be plucked from within the soil surface where they are held by cohesion and interparticle friction. Detachment is the term used to refer to this process. For detachment in rills to occur, the flow must have shear stress that exceeds \( \tau_r \). Also, for erosion to occur, detached particles must be transported away from the site of detachment. Flows are known to have a limited capacity to transport soil material and that limit is represented by \( T_c \) in Eq. 18. Consequently, the term \( 1 - q_{sd} T_c \) causes \( D_r = 0 \) when the sediment load in the rill equals the transport limit. As a result, \( D_r \) may vary along the length of a rill. Interrill erosion contributes to \( q_{sd} \) so that rill erosion may be completely suppressed if the discharge of sediment from the interrill areas is high enough.

Originally, the erosivity factor in the WEPP interrill model was assumed to be the square of rainfall intensity (\( I \)) so that basically
\[ D_i = k_i I^2 \]  
\[ \text{(19)} \]
where \( D_i \) is interrill detachment and \( k_i \) is the interrill soil erodibility factor. A series of rainfall simulation experiments\[^{31}\] was undertaken to determine \( k_i \) values for soils in the United States. Subsequent analysis\[^{32}\] of the data generated by these experiments established that it was more appropriate to use
The coefficients used in Eq. 22 result from measurement of soil material transported by splash under ponded conditions in the experiments of Torri et al.\textsuperscript{[134]} where increases in water depth were observed to reduce splash erosion. The effect of water depth on splash erosion results from 1) dissipation of raindrop energy in the water layer and 2) the effect of water depth on splash trajectories. Considering that splash does not transport 100% of the material detached by raindrops impacting water, Eq. 22 does not actually model detachment. Also, erosion by rain-impacted flows where sediment is transported by rolling, saltation, and suspension in the flow is the real focus of Eq. 22, and the effect of flow depth on erosion by rain-impacted flows is quite different to its effect on splash erosion.\textsuperscript{[35]}

As a general rule, the values of soil erodibility factors have to be determined experimentally, although there are cases where they are predicted from soil properties. In areas where detachment results from raindrop impact, erodibility is affected by modification of the soil surface generated by the impacts. Raindrop impacts on surfaces not covered by water may break soil aggregates and generate surface crusts that affect soil erodibility. In areas where soil crusts occur, particles are held more tightly within the soil surface than in areas that are not crusted and this reduces detachment. With splash erosion, loose particles sit and wait on the soil surface between drop impacts. The transport efficiency of splash increases with slope gradient but, over time, a layer of loose particles builds up on the soil surface and energy has to be expended in moving them before detachment can occur. Consequently, this layer provides a degree of protection against detachment. When raindrops impact a soil surface covered by water, the protection provided by loose particles sitting on the surface is in addition to that provided as the result of the dissipation of raindrop energy in the water layer.

Although splash erosion may dominate erosion for considerable periods during a rainfall event, rain-impacted flows are usually more important in moving soil material across the soil surface because the transport mechanisms in rain-impacted flows are much more efficient than splash transport. In rain-impacted flows, particles move across the soil surface by rolling, saltation, and complete suspension. Shallow low-velocity flows often do not have the capacity to cause particles to move by rolling and saltation by themselves but rolling and saltation can be stimulated to occur when raindrops impact the soil surface through the flow. Under these circumstances, each rolling or saltation event is of limited duration and is associated with individual raindrop impacts. Fig. 4 illustrates how raindrop and flow factors influence the detachment and transport processes associated with the erosion of fine particles, silt, and sand by rainfall. Particles moving by raindrop-induced rolling and saltation move across the surface at rates that depend on raindrop size, impact frequency, particle size, and density and the velocity of the flow. Consequently, depending on the rain and flow conditions, particles in rain-impacted flows travel across the surface at virtual velocities that vary
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from near zero to the velocity of the flow. In effect, particles are winnowed from the soil surface at various rates. Large gravel particles may not move at all and so, over time, soils that have high gravel contents may become highly resistant to erosion through the formation of erosion "pavements." Particles moving by raindrop-induced rolling and saltation sit on the soil surface between drop impacts and so provide a degree of protection against detachment as described in the case of splash erosion. As a result, soil resistance to erosion may vary considerably during a rainfall event. If $H$ is the degree of protection provided by the loose material, then all the material discharged from the soil surface comes from the layer of loose material when $H = 1.0$. Consequently, it follows from Eq. 20 that, for any given rainfall event,

$$D_i = (k_{iqU} (1 - H) + H k_{iqP}) q_i I$$

(24)

where $k_{iqU}$ is the interrill erodibility factor when no loose material exists on the surface and $k_{iqP}$ is the soil erodibility factor when loose material completely protects the soil surface from detachment. As noted earlier, detachment of soil particles from a cohesive surface varies with cohesion and interparticle friction so that factors such as the development of surface crusts can cause $k_{iqU}$ to vary with time. The soil erodibility term in Eq. 24 is $k_{iqU} (1 - H) + H k_{iqP}$, and values of $k_{iq}$ obtained in rainfall simulator experiments like those undertaken by Elliot et al.\[31\] lie between $k_{iqU}$ and $k_{iqP}$. Where exactly they do lie is unknown because $H$ is unknown.

**EFFECT OF PARTICLE TRAVEL RATES ON SEDIMENT COMPOSITION AND ERODIBILITY**

As noted above, erodibilities have units of soil loss per unit of the erosivity index used in the model that is being considered in the analysis of the data. Factors such as cohesion, particle size, and aggregate stability have been observed to influence these erodibilities. Data on the physical and chemical nature of the soil involved may, in some cases, be used to predict erodibility, but often little attention is given to the composition of the sediment discharged in experiments undertaken to determine erodibility.

The composition of the soil transported from an eroding area by rain-impacted flow during an experiment tends to be finer than the original soil (e.g., Meyer et al.\[36\] Miller and Baharuddin\[17\] Palis\[38\] and Parsons\[39\]), and particle travel rate is one of the factors influencing sediment composition. Fast-moving particles detached at the top of
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Fig. 5  Enrichment factor curves for sediment discharged at 2 min and 60 min when beds of sand were eroded by rain-impacted flows in experiments undertaken by Walker et al.\textsuperscript{[40]} using 2.7 mm raindrops. 
\textbf{Source:} Extracted from Fig. 3 of Walker et al.\textsuperscript{[40]}

an eroding area arrive at the downstream boundary well before slow-moving particles detached at the same place at the same time. Consequently, once sediment transport in rain-impacted flow starts, the fine material dominates the early discharge of sediment, but the composition of the sediment becomes coarser with time as more of the slower-moving particles reach the downstream boundary. If the particles are stable, then at the steady state, the composition of the sediment discharge must be the same both physically and chemically as the original soil. This is demonstrated by the results of laboratory experiments on erosion of sand by rain-impacted flows undertaken by Walker et al.\textsuperscript{[39]} In these experiments, 3 m long sloping beds of sand were eroded for 1 hr by rain-impacted flows generated by artificial rain at three rainfall intensities (45, 100, and 150 mm/hr). The data shown in Fig. 5 were generated by rain made up of a single drop size (2.7 mm) falling on rain-generated flows over beds of sand with two different slope gradients (5% and 0.5%). The least erosive situation is the case where 45 mm/hr rain fell when the slope gradient was 0.5%. The most erosive situation is the case where 150 mm/hr rain fell when the gradient was 5%. The enrichment factor is the ratio of the proportion of the material in the sediment discharge to the proportion in the original material. In all cases, the discharge of sediment was dominated by fine material at 2 min and, in some cases, coarse material at 1 hr (Fig. 5). In the most erosive situation (150 mm/hr rain falling on 5% slope), the sediment composition at 1 hr was close to the composition for the steady state, the composition that occurs when the enrichment factor for all particles sizes equals 1.0. Sediment composition generated by the rain-impacted flows varies with the intensity of the rain, the slope gradient, slope length, and time because particles of different size and density travel at different rates. The differential rate of transport of particles in rain-impacted flows has consequences with respect to erodibility because the actual area contributing to the soil loss does not stabilize until the slowest-moving particles detached at the farthest point from the downslope boundary at the start of the rainfall event are discharged. This fact is seldom considered when experiments are undertaken to determine erodibility.

Eq. 24 is essentially targeted at situations where raindrop-induced saltation (RIS) controls sediment discharge in situations like that illustrated in Fig. 6a. Particles of silt and sand may travel over the soil surface in more than one mode before being discharged. For example, as may be perceived from Fig. 6b, they may leave the point of detachment traveling in splash (ST), then move further downslope by RIS, and finally by flow-driven saltation (FDS) as flow energy increases down the slope. Transitions between RIS and FDS may vary in time and space during a rainfall event and will depend on the intensity of the rain, the infiltration characteristics of the soil, the length of the slope, and the slope gradient. The effect of the transition can have an appreciable effect on both soil loss and the composition of the sediment discharged. Fig. 7b shows modeled flow velocities at the brink (discharge boundary) of planar bare soil surfaces of various length on a 9% slope resulting from the rainfall event shown in Fig. 7a. Fig. 8 shows the loss of materials of various size
Fig. 7  Brink (downstream boundary) flow velocities (b) for bare soil of various lengths inclined at 9% produced by the rainfall–runoff model described by Moore and Kinnell[41] and the rainfall intensities recorded during a rainfall event at the Ginninderra Experiment Station, Canberra, Australia (a).

Fig. 8  Amounts of coal, sand, and fine particles discharged for the rainfall and runoff conditions shown in Fig. 7 when a mechanistic model of erosion by rain-impacted flow was used by Kinnell[42].
and density from those bare soil surfaces when a mechanistic model of erosion by rain-impacted flow was used by Kinnell.\footnote{Kinnell, P.I.A. Event soil loss, runoff and the Universal Soil Loss family of models: A review. J. Hydrol. 2010, 385, 384–397.} For the surfaces up to 15 m in length, particles larger than 0.1 mm in size traveled over the whole length by RIS. Under these conditions, more of the 0.11 mm coal was lost during the event than the 0.46 mm coal. On the 20 m long area, the 0.46 mm coal, which had been traveling slower than the 0.11 sand on shorter areas, traveled for a short period of time by FDS during the high-intensity bursts of rain, and, as a result, more of the 0.46 mm coal was lost than the 0.11 mm sand. On areas 25 m long and more, FDS also contributed to the discharge of the 0.11 mm sand, but the amount lost was always less than the amount of the 0.46 mm coal. Little regard is given to the effect of temporal and spatial changes in transport mechanism when experiments are undertaken to determine values for soil erodibility factors in sheet and interrill erosion areas even though these changes may have a major impact on the amount of soil lost from an area.

**CONCLUSION**

Although conceptually, rainfall erosivity is the capacity of rain to produce erosion, whereas soil erodibility is the susceptibility of the soil to be eroded, the factors controlling the erosive stress applied to the soil surface and the factors influencing the resistance of the soil to them vary in time and space in complex ways. In all existing predictive models of rainfall erosion, numerous simplifications and assumptions have to be made for practical reasons, and, as a result, these models do not have the capacity to deal with such complexity. Consequently, erosivity and erodibility values are specific to the model in which they are parameterized and to the scale that the model operates within.

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