# Length-slope factors for the Revised Universal Soil Loss Equation: Simplified method of estimation

Ian D. Moore and John P. Wilson

*ABSTRACT: The combined length-slope (LS) factor in the Universal Soil Loss Equation (USLE) is a measure of the sediment transport capacity of overland flow. A dimensionless sediment transport capacity index that is a non-linear function of specific discharge and slope was derived by considering the transport capacity limiting sediment flux in the Hairsine-Rose, WEPP, and catchment evolution erosion theories. For a two-dimensional hillslope, the index is equivalent to the combined LS factors in the Revised Universal Soil Loss Equation (RUSLE), but it is simpler to use and conceptually easier to understand. A major advantage of the index is that it can be easily extended to three dimensional terrain.*

The Universal Soil Loss Equation (USLE) was empirically derived from over 10,000 plot-years of data *(24)* and has recently been revised *(12, 13, 18)*. The USLE and the Revised Universal Soil Loss Equation (RUSLE) can be written as:

$$A = RKLSCP \qquad [1]$$

where A is the soil loss, R is the rainfall-runoff erosivity factor, K is a soil erodibility factor, L is a slope-length factor, S is a slope-steepness factor, C is a cover-management factor and P is a supporting factor. Land use and management are represented by CP and can, with some difficulty, be inferred by remote sensing combined with ground-truthing. Climate erosivity is represented by R and can be computed directly from a knowledge of rainfall intensities and amounts; it varies on a regional scale. Soil erodibility is represented by K, and in the United States values of K have been interpolated or measured for all mapped soil series as part of the Soils-5 database that is derived from County Soil Surveys. Soil series are mapped at scales of 1:15,000 to 1:20,000 in these surveys. The effects of topography and hydrology on soil loss are characterized by the combined LS factor. Table 1 presents expressions for the L and S factors in both USLE and RUSLE. Soil loss predictions are more sensitive to slope steepness than slope length. The combined LS factors for the USLE-LS and the RUSLE-LS are compared in Figure 1. Only values of slope-length <100m, slope-steepness <25%, and USLE-LS and RUSLE-LS <5 are included in this figure.

*Ian D. Moore is the Jack Beale Professor of Water Resources with the Centre for Resource and Environmental Studies, The Australian National University, GPO Box 4, Canberra, ACT. 2601, and John P. Wilson is director of the Geographic Information and Analysis Center and associate professor, Department of Earth Sciences, Montana State University, Bozeman, 59717.* *This study was funded in part by grant No. 90/6334 from the Bilateral Science and Technology Program of the Department of Industry, Technology and Commerce (Australia) and by the Water Research Foundation of Australia.*

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Estimation of the LS factor poses more problems than any of the other factors in the USLE *(18, 23)* and is a particular problem in applying it to real landscapes as part of a Geographic Information System (GIS). Some of these problems follow from a number of implicit assumptions concerned with runoff generation and sediment transport imbedded in the equation, notably that runoff is generated uniformly over a catchment, runoff occurs via the infiltration excess mechanism (i.e., Hortonian overland flow) and ignores saturation overland flow, and sediment deposition is not represented, which represents a major practical problem because the model does not differentiate those parts of landscapes experiencing net erosion and those areas experiencing net deposition (i.e., the lower ends of concave slopes). The USLE has been adapted to variable hillslopes (4, 5), but only applies to those areas experiencing net erosion. Other problems and limitations follow from the implicit division of landscapes into hillslopes. Soil loss is best estimated for points or small areas (i.e., grid cells) when USLE is applied to large areas rather than fields and hillslopes *(5)*. This one-dimensional structure means that the equation cannot handle converging and diverging terrain (i.e., real 3-D landscapes).

A simplified method of estimating the LS factors in RUSLE is presented that can be easily extended to estimating soil loss in complex 3-D terrain. It may also help distinguish areas experiencing net erosion and those experiencing net deposition. The method is derived by considering the steady-state, sediment transport limiting case predicted by three erosion theories.

## Erosion theory

Three approaches that represent the "state-of-the-art" in erosion modeling are described here. They include sophisticated physically-based or process-oriented models, as well as both dynamic and steady-state models and cover simulation time scales ranging from seconds to thousands of years (geological time scales).

***Water Erosion Prediction Project (WEPP) theory.*** The United States Department of Agriculture (USDA) is developing improved, process-based erosion-prediction models aimed at replacing the USLE by 1995. WEPP is to be delivered in three versions: profile, watershed, and grid, with the profile version being the direct replacement of the USLE *(10)*. WEPP is based on Foster and associates *(2)* concept that divides erosion into an interrill component representing detachment and transport by raindrops and very shallow flows and a rill component that represents net erosion or deposition in rills. Rill detachment is modeled as a function of excess hydraulic shear *(9)*. WEPP per- forms its internal calculations on a per rill area basis.

The theory is encapsulated by the following steady-state sediment continuity equation for a rill:

$$dq_s / ds = D_f + D_i \qquad [2]$$

where $q_s$ is the sediment flux ($kgm^{-1}s^{-1}$), $D_i$ is the interill sediment delivery rate to the rill ($kgm^{-2}s^{1}$) and $D_f$ is the net erosion or deposition rate in the rill ($kgm^{-2}s^{-1}$) *(1, 3)*. The shallow flow hydraulics in the interrill areas are not directly modeled, but their effects on delivering sediment to the rill are lumped with the rainfall kinetic energy term and modified by a land slope adjustment factor in the expression for $D_i$,


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which can be written as:

$$D_i = K_i I_e^2 C_g C_c S_f \,(R_s/w) \qquad [3]$$

where $K_i$ is the interrill erodibility ($kgsm^{-4}$), $I_e$ is the effective rainfall (m $s^{-1}$)$C_g$ is a ground cover adjustment factor, $C_c$ is a canopy cover adjustment factor, $S_f$ is a slope adjustment factor (= $1.05 - 0.85\ e^{-4\sin a}$), a is the slope of the land surface towards the rill, $R_s$ is the spacing between rills (m per rill) and w is the rill width (m) *(2,11)*. The net erosion or deposition rate in the rill, $D_f$ is:

$$D_f = \phi\,(T_c - q_s) \qquad [4]$$

where $T_c$ is the sediment transport capacity in the rill ($kgm^{-1}s^{-1}$) $\phi = ßv_s/q$ for net deposition in the rills (when $T_c < q_s$) and $\phi = D_c/T_c$ for net soil detachment in the rills (when $T_c > q_s$), q is the water flux ($m^3m^{-1}s^{-1}$), $v_s$ is the sediment settling velocity ($ms^{-1}$) and $D_c$ is the detachment capacity of rill flow ($kgm^{-2}s^{-1}$). For rainfall conditions ß=0.5 and for non-rainfall or snowmelt conditions ß=1.0. $D_f$ is positive when there is net erosion and negative when there is net deposition. The detachment capacity of rill flow can be expressed as:

$$D_c = K_r\tau\,(1-\tau_o/\tau) \text{ for } \tau > \tau. \qquad [5]$$

$$\text{and } D_c = O \text{ for } \tau < \tau_o$$

where $\tau$ is the flow shear stress acting on the soil particles (Pa), $\tau_o$ is a threshold shear stress (Pa) and $K_r$ is a rill erodibility parameter ($sm^{-1}$). The sediment transport capacity is represented by an approximation of the Yalin sediment transport equation:

$$Tc = kt\tau 3/2 \qquad [6]$$

where $k_\tau$ is a transport coefficient ($m^{1/2}s^2kg^{-1/2}$).

***Hairsine-Rose theory.*** The Hairsine-Rose theory is a process-based approach that recognizes raindrop impact and surface flow as the agents causing erosion of surface soils. Rainfall detachment, entrainment (detachment by overland flow), rainfall re-detachment and re-entrainment of deposited sediment and deposition are modeled as separate processes *(6, 7, 8)*. The theory is an outgrowth of concepts originally developed by Rose *(20, 21)*. The theory is encapsulated by the following one-dimensional sediment continuity equation:

$$\partial q_{si} + \partial(C_i h)\partial s \partial L = \qquad [7]$$
$$r_i + r_{di} + e_i + e_{di} + r_{gi} - d_i$$

where $q_{si}$ (=$qC_i$) is the sediment flux ($kgm^{-1}s^{-1}$) in the direction of flow (s), q is the water flux (specific discharge), $C_i$ is the sediment concentration ($kgm^{-3}$), h is the depth of overland flow (m), $r_i$, $r_{di}$ $e_i$, $e_{di}$, and $d_i$ are the rainfall detachment, rainfall re-detachment, entrainment, re-entrainment and deposition, respectively (kg $m^{-2}s^{-1}$) and subscript i refers to each of N sediment settling velocity classes (i) with an equal mass of soil in each class. The gravity process rate is $r_{gi}$ and represents contributions from headcut collapses and slumping of rill walls *(7)*.

Rainfall detachment, entrainment and deposition can be expressed as follows:

*Rainfall Detachment:*

$$r_i = (1-H)\ k\ C_e\ i^p/N \qquad [8a]$$

$$r_{di} = H\ k_d C_e i^p\ f_{di} \qquad [8b]$$

*Entrainment:*

$$e_i = (1-H)\ n/NE\ (\omega - \omega_o) \text{ for } \omega > \omega_o \qquad [9a]$$

$$edi = H\eta\alpha_i f_{di}/g(A/fs-f)(\omega - \omega_o/h) \text{ for } \omega > \omega_o \qquad [9b]$$

*Deposition:*

$$d_i = \alpha_i v_{si} C_i \qquad [10]$$

where H is the fraction of the soil shielded by a deposited layer, k and $k_d$ are measures of the detachability (kgs m-4) of the original and deposited soil, respectively, $C_e$ is the fraction of soil surface exposed to raindrop impact, i is the rainfall rate (m $s^{-1}$), p is a nondimensional exponent, $f_{di}$ is the fraction

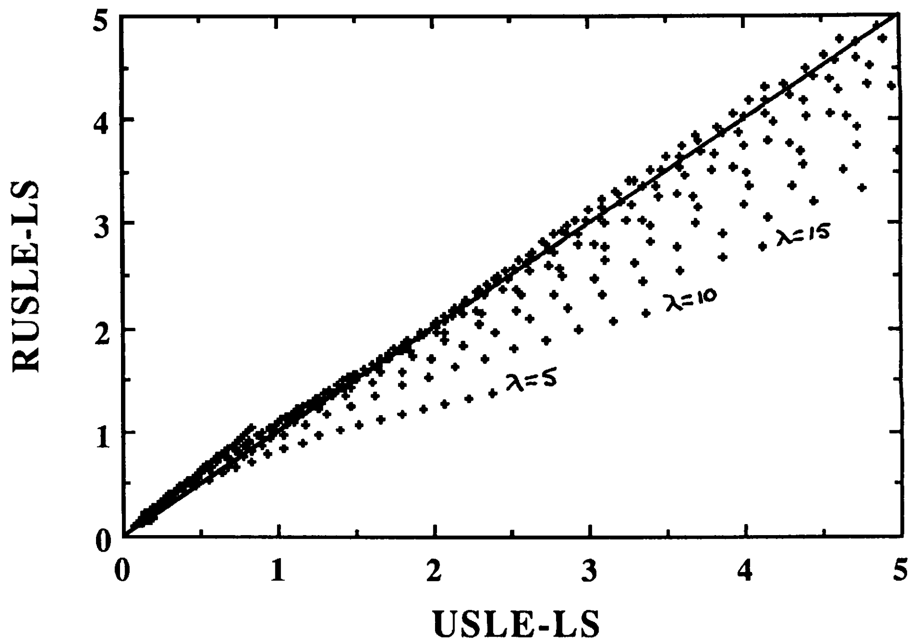


**Figure 1. Comparison of the USLE-LS and RUSLE-LS factors, where λ is the slope length (m).**

**Table 1. LS factors in the Universal Soil Loss Equation (USLE) and Revised Universal Soil Loss Equation (RUSLE) (λ=slope length in meters; β=slope angle in degrees).**

| | LS Factors | | | |
|---|---|---|---|---|
| | S | | L=(λ/22.13)$^m$ | |
| USLE (22) | 65.4sin²β+4.56sinβ+0.0654 | | m=0.5 | tanβ>0.05 |
| | | | m=0.4 | 0.03<tanβ≤0.05 |
| | | | m=0.3 | 0.01<tanβ≤0.03 |
| | | | m=0.2 | tanβ≤0.01 |
| RUSLE (12,13) | 10.8sinβ+0.03 | tanβ<0.09 | m=F/(1+F)wher | |
| | 16.8sinβ-0.50 | tanβ≥0.09 | F=(sinβ/0.0896)/(3sin$^{0.8}$β+0.56)* | |
| | 3sin$^{0.8}$β+0.56 | λ≤4m | or F=0 when there is deposition | |
| | (sinβ/0.0896)$^{0.6}$ | thawing soils with tanβ≥0.09 | when λ=4m λ≤4m equal to 4m. | |
| Moore & BURCH† (14,15) | LS=$A_s$/22.13)$^m$(sinβ/0.0896)$^n$ where m=0.4,n=1.3, and $A_s$=specific catchment area | | | |

* Assumes a moderate rill/interrill ratio (13).
† Derived from unit stream power theory.

of particles (on a mass basis) is settling velocity class i in the deposited layer, $\eta$ is the fraction of the available stream power for entrainment, E is the energy required to entrain a unit mass of soil or specific energy of entrainment ($Jkg^{-1}$) $\omega$ is the stream power (watts $m^{-2}$), $\omega_o$ is the threshold stream power, $\alpha_i$ is the ratio of the sediment concentration next to the bed to the mean concentration across the entire depth ($C_i$) so that $\alpha i \geq 1$, $\rho$ is the density of the sediment laden water ($kgm^3$) ($\rho$=1.000 + 0.615C), $\rho_s$ is the density of detached soil or soil aggregates ($kgm^3$), h is the depth of flow (m), and $v_s$ is the sediment settling velocity ($ms^{-1}$). The depositabilty of the sediment is equal to $\Sigma v_{si}/N$.

Equations 8 and 9 assume that rainfall detachment and entrainment are nonselective whereas equation 10 shows that deposition is highly selective. The re-entrainment process represented by equation 9b assumes that the deposited soil has no cohesive strength. The stream power used in equations 9a and 9b is the stream power per unit wetted area *(19)*, $\omega = \rho gqs\sin\beta$, where $\rho g$ is the unit weight of water and $\beta$ is the slope of the energy grade line, which is assumed equal to the land slope. An equivalent expression for stream power is $\omega = \tau v$, where $\tau$ is the shear stress and v is the flow velocity. In equation 9b the $\omega/h$ term is equivalent to $\rho gv\sin\beta$, where $v\sin\beta$ is the unit stream power (watt $N^{-1}$ or $ms^{-1}$), defined as the stream power per unit weight of water. In both equations 9a and 9b $e_i$=0 and $e_{di}$=0 when $\omega \leq \omega_o$. The fraction of available stream power for entrainment ($\eta$) is about 0.1, although it increases to 0.2 for low stream powers, and typical values of E in equation 9a are 20-30 $Jkg^{-1}$ for cultivated soils and 100-150 $Jkg^{-1}$ for rangeland soils (Hairsine, personal communication). Also, the exponent p in equation 8a is usually assumed to to be 1. The soil rainfall detachability terms in equations 8a and 8b can be written as functions of the maximum detachability ($k_o$ and $k_{do}$) and water depth (h):

$$k = k_o(h_o/h)^b \text{ for } h>h_o \qquad [11a]$$

and $k = ko$ for $h \leq h_o$

$$k_d = k_{do}(h_o/h)^b \text{ for } h>h_o \qquad [11b]$$

and $k_d = k_{do}$ $h \leq h_o$

where $h_o$ is a threshold water depth.

***Catchment evolution theory.*** Willgoose and associates *(22)* have recently proposed a hillslope and catchment evolution model that explicitly differentiates between the sediment transport behavior in channels and on hillslopes via coupled flow and sediment continuity equations in the hillslope and channel. Both diffusive (function of slope only, e.g., raindrop splash, soil creep and rock slide) and fluvial (function of slope and discharge) sediment transport processes and tectonic uplift are represented. Channel initiation is modeled as a threshold process that is nonlinearly related to slope and discharge. The governing sediment continuity equation and channel indicator function can be expressed, respectively, as:

$$\partial z = c_o(x,y,) + 1/\rho_s(1-n)\ (\partial_q+\partial_q\ /\ \partial x\ \partial y) + D_z\ (\partial^2 z+\partial^2 z\ /\ \partial x^2\ \partial y^2) \qquad [12a]$$

$$\partial Y/\partial t = d_t\ [0.0025a/a^t - 0.1Y + Y^2/ 1+9Y2] \qquad [12b]$$

where z is elevation, $c_o$ is the rate of tectonic uplift, $\rho_s$ is the density of eroded material, n is the porosity of material before erosion and after deposition, $D_z$ is a diffusivity constant, Y is a channel indicator function (=0 hillslope; =1 channel; $0 \leq Y \leq 1$), $d_t$ is a rate constant for channel growth, a is a channel initiation function [= $\phi_1 q^{m1}(\sin\beta)^{n1}$], $a_t$ is a channel initiation threshold, $\beta$ is the slope angle in the direction of steepest descent, and $\phi_1$, m1, and n1 are constants. The sediment flux, $q_s$, is a function of the water flux, q, and the land surface slope $\beta$:

$$q_s = \phi_2 q^m (\sin\beta)^n \qquad [13]$$

where m and n are constants and $\phi_2$ is a rate constant for sediment transport, that is different for hillslopes and channels. In equation 12b Y=0 for $a<a_t$, goes into a transition when $a=a_t$, increasing to Y=1 at a speed dependent on the channel growth rate constant, $d_t$, and once it reaches a value of 1 remains there. Most of the hillslope evolution models developed in the last 20 years solve a sediment continuity equation similar to equation 12a.

***General sediment transport equations.*** Using dimensional analysis,

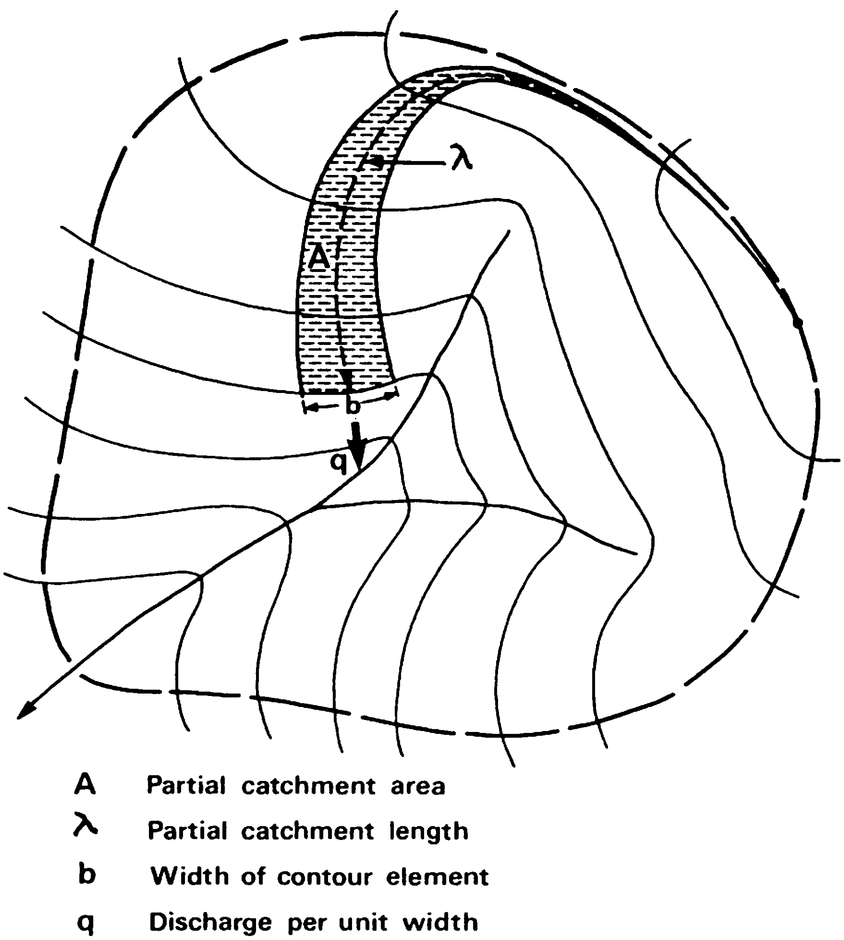


**Figure 2. Schematic representation of specific catchment area, $A_s$ - A/b [adapted from (15)].**

Julien and Simons *(9)* have shown that most sediment transport equations can be expressed in the following general form:

$$q_s = \phi_2 \, q^m \, (\sin\beta)^n i^\delta \left(1-\tau_o/\tau\right)^\varepsilon \quad [14]$$

where i is the rainfall intensity, $\phi_2$, n, m, $\delta$ and $\varepsilon$ are experimental or physically-based coefficients, and the other terms are as previously defined. The first three terms ($\sin\beta$, q, i) represent the potential erosion or transport by flow, which is reduced by the last term (the shear stress term) reflecting the soil resistance to erosion *(9)*. When $\tau_o$ is small compared to $\tau$, the shear stress term can be neglected. The rainfall intensity term is also ignored in many sediment transport equations (i.e., $\delta$ = 0), but this is only strictly true for turbulent flows in deep channels.

If we assume rainfall excess is generated uniformly over a catchment then $q = A_s i_e$, where $A_s$ is the upslope contributing area per unit width of contour (or rill) or the specific catchment area ($m^2\ m^{-1}$) and $i_e$ is the rainfall excess rate ($m\ s^{-1}$). A schematic representation of the specific catchment area is presented in Figure 2. For a 2-D hillslope where there is no flow convergence or divergence $A_s = \lambda$, the slope length.

**Figure 3. Comparison of different forms of the dimensionless sediment transport capacity index versus the length-slope factors in the Universal Soil Loss Equation (USLE) and Revised Universal Soil Loss Equation (RUSLE).**

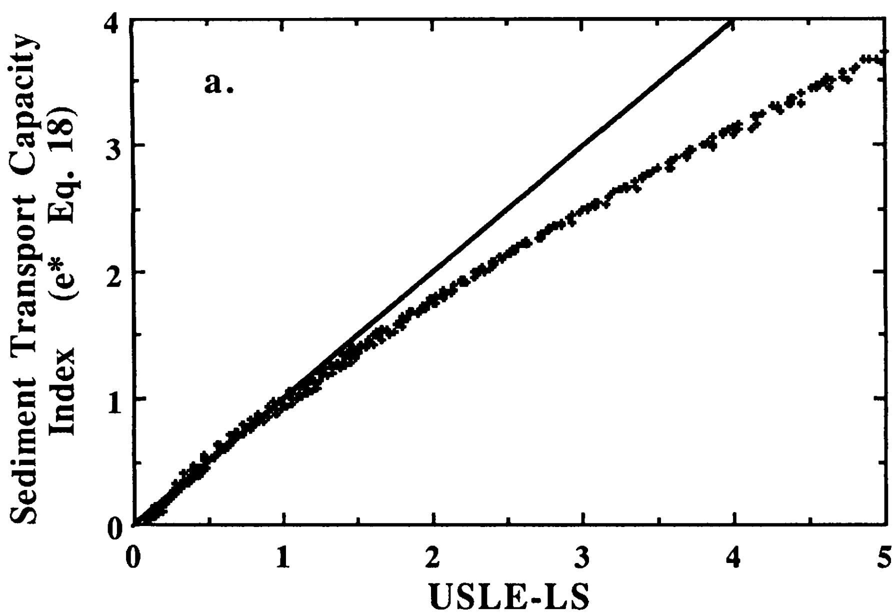


**Figure 3a. Hairline-Rose theory-based index versus USLE-LS.**

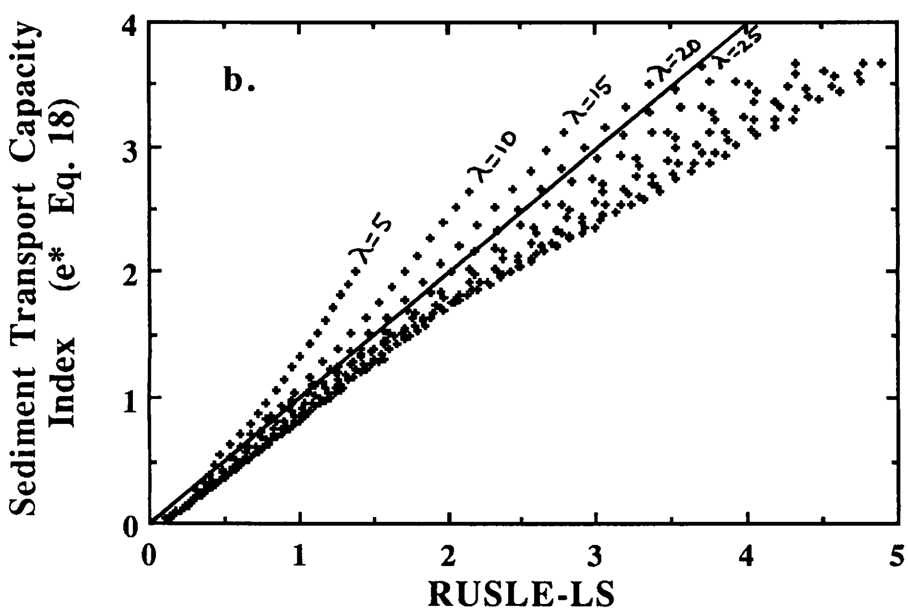


**Figure 3b. Hairsine-Rose theory-based index versus RUSLE-LS (λ = slope length in m).**

## Transport limiting case

For large runoff and erosion events the "transport limiting" case, where the sediment flux is limited only by the ability of the flow to carry the sediment, is likely to be the dominant influence on the pattern of erosion in landscapes. With the WEPP theory this transport limiting case occurs when $q_s = T_c$. We can represent the overland flow hydraulics as uniform turbulent flow using Manning's equation. The WEPP theory assumes that sediment is transported from a site by concentrated flow in rills. By approximating the relationship between hydraulic radius, R, and rill cross sectional area, A, by $R = UA^{1/2}$, where U is a rill shape factor *(16)*, equation 6 can be written in terms of the specificlc discharge, q (discharge per unit width of catchment, not the discharge per unit width of rill), and slope angle, $\beta$:

$$T_c = k_t(\rho g)^{1.5} (R_s n U^2)^{0.56} q^{0.56} (\sin\beta)^{1.22} \quad [15a]$$

and with $q = i_e A_s$

$$T_c = k_t(\rho g)^{1.5} (R_s i_e n U^2)^{0.56} A_s^{0.56} (\sin\beta)^{1.22} \quad [15b]$$

where n is Manning's roughness coefficient and $R_s$ is the rill spacing (m per rill). If we write equation 15 in a dimensionless form so that $T_c^*$, the dimensionless sediment transport capacity, is unity when $A_s = 22.13\ m^2\ m^{-1}$ and $\tan\beta = 0.09$ (as with the LS factor in the USLE), then:

$$T_c^* = (A_s/22.13)^{0.56} (\sin\beta/0.0896)^{1.22} \quad [16]$$

where the exponents 0.56 and 1.22 are equivalent to the slope-length and slope-angle exponents, m and n, respectively, in the LS factor in the USLE. If shallow sheet flow were assumed rather than concentrated flow in rills then the exponents in equation 16 would be 0.9 and 1.05, respectively.

In the Hairsine-Rose theory the equivalent "transport limiting" case under the steady-state sediment flux occurs when $\partial(C_i h)/\partial t = 0$ and H=1 in equations 7 to 10 *(7)*, which corresponds to the condition where there is a layer of deposited sediment over the entire hillslope. If the threshold term, $\omega_o/h$ in equation 9b is small compared to $\omega/h$, and can therefore be neglected, then the re-entrainment rate given by equation 9b can be rewritten as: [17]

$$e_{di} = H\eta\, \rho\alpha_i f_{di}/_n 0.6\, (\rho_s/\rho-\rho)q^{0.4} (\sin\beta)^{1.3} = H\eta\, \rho\alpha_1 f_{di}/_n 0.6\, (\rho_s/\rho_s\text{-}\rho)\, i_e^{0.4} A_s^{0.4} (\sin\beta)^{1.3}$$

Again, writing equation 17 in dimensionless form like equation 16, a dimensionless re-entrainment rate, e*, can be derived: [18]

$$e^* = (A_3/22.13)^{0.4}\, (\sin\beta/0.0896)^{1.3}$$

which is the unit stream power based length-slope factor proposed by Moore and Burch *(15)* (Table 1).

Equation 13, used in the catchment evolution model, can also be reduced to a dimensionless form having the same structure as both equations 16 and 18. The S factor in the RUSLE for thawing soils is also expressed in this form, but with the slope exponent n=0.6 (Table 1).

## Results and discussion

The dimensionless re-entrainment rate derived from the Hairsine-Rose theory and derived independently by Moore and Burch *(14, 15)*, e*, is compared to the LS factors in the USLE (USLE-LS) and RUSLE (RUSLE-LS) (Table 1) in Figures 3a and 3b, respectively, for the case where $A_s = \lambda$, and $\lambda$ is the slope-length. There is a strong monotonic function relating USLE-LS to e*, with e* < USLE-LS for USLE-LS values >1.5, which is consistent with the widely held view that the USLE overpredicts LS values at higher slopes and longer slope-lengths. Figure 3b shows that there is considerable scatter in the RUSLE-LS versus e* relationship. However, for $A_s=\lambda=22.13$ m$^2$ m$^{-1}$ (i.e., $A_s/22.13=1$) there is very close agreement, indicating that the theoretical exponent of 1.3 on the slope term of e*, is quite accurate. Only values of slope-length < 100 m, slope-steepness < 25% and USLE-LS and RUSLE-LS<5 are included in Figure 3.

Figure 3c shows that there is also good agreement between the dimensionless sediment transport capacity, $T_c$*, that is derived from the WEPP theory and RUSLE-LS. To a large degree this is expected as the LS factors developed for the RUSLE by McCool et al. *(12, 13)* were derived in part by applying the Foster and Meyer *(1)* theory, which is the basis of the WEPP model.

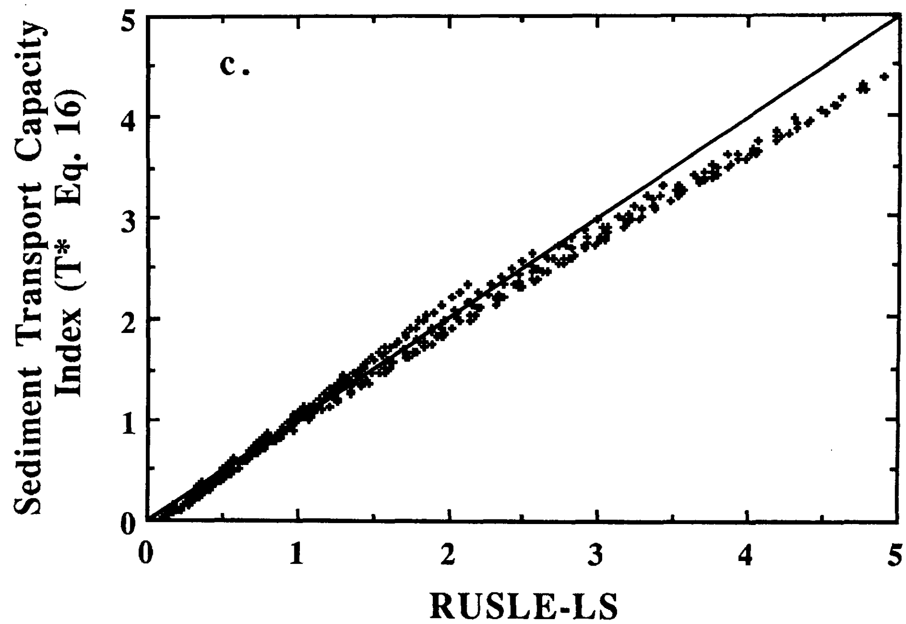


**Figure 3c. WEPP theory-based index versus RUSLE-LS.**

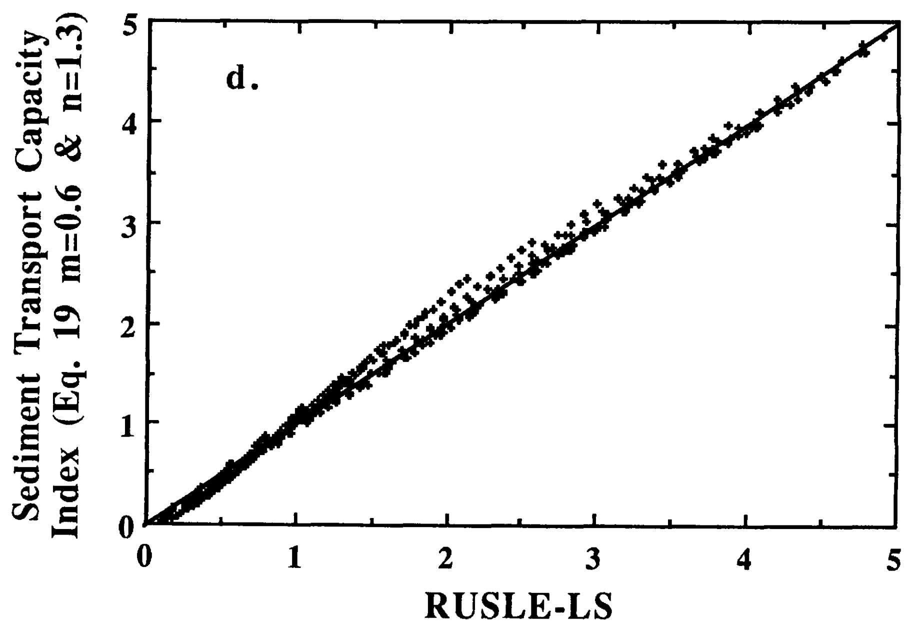


**Figure 3d. Generalized dimensionless sediment transport equation with m=0.6 and n=1.3 (Eq.) versus RUSLE-LS.**

However, equation 18 is functionally simpler and easier to use than the RUSLE-LS factors.

The best fit between RUSLE-LS and an equation of the form of equations 14 and 18 occurs when the area and slope exponents (m and n) are 0.6 and 1.3, respectively (Figure 3d). These results suggest that the combined LS factor in the USLE and RUSLE are measures of the sediment transport capacity of the flow. Furthermore, they show that a sediment transport equation of the form of equation 14 or written in dimensionless form as: [19]

$$T_c^* = (A_s/22.13)^m\, (\sin\beta/0.0896)^n = LS$$

with m=0.6 (0.4 to 0.6) and n=1.3 (1.2 to 1.3) can be used to map the effects of hydrology, and hence 3-D terrain, on soil erosion in natural landscapes. The $A_s$ term can characterize the effect of converging and diverging terrain on soil erosion, unlike the $\lambda$ term in the USLE and RUSLE, which is only applicable to 2-D, non-converging and non-diverging hillslopes. For predicting erosion at a point, equation 19 should by multiplied by (m+1), as proposed by Griffin and associates *(5)*.

The change in sediment transport capacity across a grid-cell, $\Delta T_c$, provides a possible measure of the erosion or deposition potential in each cell *(14)*. Conceptually, $\Delta T_c$ is related to the $dq_s/ds$ term in Equations 2, 7, and 12a when the sediment transport limiting case is considered (i.e., when $q_s = T_c$). The change in sediment transport capacity can be written as:

$$\Delta T_{cj} = \phi[A_{sj-}^{m}(\sin\beta_{j-})^n - A_{sj}^{m}(\sin\beta_j)^n] \qquad [20]$$

where $\phi$ is a constant, subscript j signifies the outlet of cell j and j- signifies the inlet to cell j. Moore and associates *(17)* have hypothesized that a positive value of $\Delta T_{cj}$ indicates net deposition and that a negative value indicates net erosion. However, further testing is required to determine whether or not this relationship applies across most landscapes.

## Conclusions

By considering the transport limiting entrainment rate (H=1) in the steady-state version of the Hairsine-Rose erosion theory, the sediment transport limiting case in the WEPP theory and the general forms of the sediment flux expressions in a model of catchment evolution, a dimensionless sediment transport capacity index can be derived. For a two-dimensional hillslope, the index is equivalent to the combined length-slope factor (LS) in RUSLE, but it is simpler to use and conceptually easier to understand. RUSLE provides lower estimates of the LS factor for longer slope-lengths and steeper slope-angles than in the original USLE. These results lend support to the concept of the LS factor as a measure of the sediment transport capacity of overland flow.

A major advantage of the index is that it can be easily extended to three-dimensional terrain. Further work is required to determine whether or not it can also differentiate areas experiencing net erosion from those experiencing net deposition. The specific discharge is a function of the specific drainage area, soil properties and rainfall intensity. Therefore, the index can be estimated as a function of primary terrain attributes and soil properties and can be readily implemented within an appropriately scaled GIS.

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