Rainfall Simulation Studies – A Review of Designs, Performance and Erosion Measurement Variability

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INTRODUCTION

The unpredictability, infrequent and random nature of natural rainfall makes difficult the study of its effects on soils while rainfall is occurring. The use of rainfall simulators (RSs) and perhaps runoff simulators for rill erosion can overcome some of these difficulties, enabling a precise, defined storm centrally located over runoff measurement “frames”. RSs are often used to study the effects of various soil factors on rates of infiltration and erosion in the field. Following development of sprinkler and drop-former designs in the 1950s, a variety of RSs have been developed for use in the laboratory and field. Generally, these are associated with smaller plot sizes on the order of 1 m^2 and are directed at assessment of soil cover, tillage or practice treatment effects, determination of soil inter-rill and rill erodibilities for model parameterization, evaluation of pollutant transport or dispersal rates and other applications of particular interest to the research group. That no standardized methodology has been proposed or can be identified in the literature, making comparisons between study results difficult has long been recognized. Such efforts in Europe were represented in part by conferences and meetings resulting in a special journal issue (Parsons and Lascelles, 2006) that detailed some of the efforts of a working group having the goals of cataloging the RSs in use, their specifications and performance characteristics, as well as developing a standard RS evaluation and test methodology for broad use such that data collected by various studies can be compared. Agassi and Bradford (1999) completed a review of inter-rill erosion measurement studies using rainfall simulation methods and categorized the “methodology problems” into inadequate characterization of (a) the type of RS its rainfall intensities, mean drop size and drop-size distribution, and water quality deployed, (b) the soil plot physical and chemical properties, and (c) the type of results obtained and how they are presented. Kinnell (2005, 2006) completed thorough reviews of the processes associated with raindrop impacted erosion and noted that both conceptual models and measurements fail in various respects to adequately characterize observed erosion processes from bare soils. Concerns such as these have also arisen in the Tahoe Basin, because a variety of methods for measurement of infiltration and erosion rates have been deployed, but comparisons between results of different studies are uncertain.

The objective of this paper is to review the more recent literature of the past two decades concerning application of RS techniques in the field and how they might apply to forested, rangeland, and ski-run conditions similar to that found in the Sierra Nevada. As many of the RS-derived erosion measurement efforts are, at least in part, motivated by the historical conceptual view of erosion processes, first, the prevailing descriptions of the erosion processes as they developed from the classic USLE-based interpretation to sediment transport and WEPP-based analyses are considered. Next, as a primary concern of the past has been the ability of RSs to replicate "natural" rainfall characteristics, available studies of these characteristics are reviewed and compared with laboratory analyses of rain drop-sizes, their distributions and kinetic energies (KEs). These reviews set the stage for consideration of RS designs and field methodologies as they may have been affected by attempted definitions of erodibility and "natural" rain characteristics. Following review of RS designs and issues associated with field plot conditions, some of the key issues associated with RS-based erosion measurements; the processes involved in forested landscapes, their interpretation, sources of error or uncertainty and up-scaling plot results to hillslope and catchments are considered. Here, the focus is largely on "portable" RS usable in field studies of these various processes on a range of slopes.

EROSION PROCESSES – USLE and WEPP Development

The rainfall runoff and erosion process is usually considered to be initiated with rain drop impact on bare or nearly bare soils, detaching and splashing soil particles and subsequent downslope transport as part of overland flow (Mutchler et al, 1988). Raindrop momentum or kinetic energy (KE) is a product of raindrop size (mass) and velocity or velocity -squared at impact. Though Wischmeier and others originally found from statistical analyses that rainfall KE alone was insufficient to describe erosivity, Lal (1988) opined that it is a major factor in the soil detachment process, and likewise that the total energy load of a storm is proportional to its erosivity. Net erosion rates (sediment mass/unit area) are a function of both rainsplash and overland flow transport. For shallow slopes, rainsplash is considered the dominant factor in causing erosion; as the slope angle increases, runoff becomes the dominant factor (Kamalu, 1994). Splash erosion alone does not redistribute large amounts of soil, rather it serves to detach soil material for transport by runoff (Evans, 1980). Runoff, as interrill overland flow, carries with it the smaller detached particles, and acts to remove the most erodible silt and very fine sand particles from the soil surface as it flows downhill (Press and Siever, 1986).

When the rainfall intensity exceeds the infiltration rate, surface water excess accumulates on the soil; when the surface depressions are filled, runoff in the form of sheet overland flow can occur. Surface roughness reduces the velocity of overland flow (Evans, 1980); differences in surface roughness can be due to soil textural variations, tillage, residues on the surface, or the presence of living plant stems. Overland flow is a very elusive and difficult process to measure, and as such, relatively little is known about the actual mechanics of soil loss by this process (Emmett, 1980), though classic sediment transport concepts are generally used.

Surface cover, in the form of living vegetation or residues, both reduces the impact energy of the raindrops and prevents them from striking the soil surface. Raindrop impacts both compact the surface and possibly detach surface soil particles from bare soils; together these processes can seal the soil surface, reducing the infiltration rate. At low cover levels, runoff and erosion rates are related to the area of bare ground, increasing as the bare ground area increases. Vegetated soils also have greater structure and aggregation, leading to higher infiltration rates (Evans, 1980). The effect of plant canopy cover on reducing runoff and erosion in natural rangeland environments has been demonstrated to be due primarily to the increased litter cover, soil macro-porosity, and soil structure that occur due to the presence of canopy cover, rather than to the direct interception of rainfall (Simanton et al., 1991). Similarly, rock cover tends to reduce erosion rates proportional to the cover fraction.

Generally with all else equal, erosion rates increase as the slope angle increases (Evans, 1980). As slope increases, overland flow velocities increase (Kloosterboer and Eppink, 1989) such that the greater surface flow velocity increases both the erosive power and the flow competence (i.e. “transport capacity”) to carry suspended sediments (Press and Siever, 1986). The slope angle is also important in the splash erosion process; as the angle steepness increases, more soil is splashed downhill (Evans, 1980). However, the runoff component is the most sensitive to slope change; beyond some threshold inclination, it becomes the dominant erosive process (Kamalu, 1994).

Though only briefly outlined above, it is clear that erosion from soil surfaces involves several inter-related processes that in the field combine in complex spatial and temporal variations such that results from different erosion studies are difficult to compare. These processes can include particle (aggregate) breakdown, particle detachment, related splash effects then particle suspension and transport as part of overland flow or wind, particle filtration by covers or mulch layers, particle movement into the soil profile and so on. Clearly, all these processes are controlled by basic hydrologic phenomena such as precipitation form and rates, soil infiltration rates and capacity and the surface conditions (e.g. cover type and extent, roughness). Thus far, it appears that all water-erosion related research begins at the simplest level of soil condition for analyses; that is, bare soils (no cover/mulch complication) of known textures and bulk densities on mild slopes (<10%) with no infiltration limiting layer. The reality of various tillage, cover and slope conditions in the field resulted in development of comparisons between actual field conditions and that for bare soil in order to derive cover, practice and management factors as simple ratios of the varied condition erosion rate to that from bare soil.

Universal Soil Loss Equation (USLE)

Perhaps one of the first to employ an empirical equation to estimate soil water-erosion, Zingg (1940) developed a regression equation that later served as at least a conceptual basis for the USLE. Zingg's equation took the form

$$\mathbf{A}=CS^{1.4}L^{0.6} \qquad [1]$$

where

A = average soil loss per unit area from a land slope of unit width (lb/ft^2),

C = conversion constant of variation,

S = degree of land slope (%), and

L = horizontal length of land slope (ft).

By 1956, more than 7500 plot-years and 500 watershed-years of agricultural erosion data compiled from 21 states were compiled by Smith and Wischmeier (1958) and developed into a series of empirical equations from which it was possible to estimate rates of erosion eventually forming the more widely known USLE.

The Universal Soil Loss Equation (USLE) was codified of sorts in 1965 (USDA Agriculture Handbook 282) that was revised in 1978 as Agriculture Handbook 537, Wischmeier and Smith (1978). The USLE was derived from statistical analyses of 10,000 plot-years of natural runoff and erosion data and the equivalent of 1000-2000 plot-years of rainfall simulator derived plot data. The authors emphasized that the USLE is an erosion model designed to predict the longtime average annual soil losses from sheet and rill erosion, and from specific field areas in specified cropping and management systems. As noted above, many variables and interactions influence sheet and rill erosion. The USLE groups these variables under six major erosion factors, the product of which, for a particular set of conditions, represents the average annual soil loss (Wischmeier, 1976). The USLE takes the form

$$\mathbf{A} = R \cdot K\ L \cdot S \cdot C\ P \qquad [2]$$

where

A = estimated soil loss (ton/acre-year),

R = rainfall and runoff factor,

K = soil erodibility factor,

L = slope length factor,

S = slope steepness factor,

C = cover and management factor, and

P = supporting practice factor.

One of the key factors of the USLE germane to RS studies is definition of the rainfall erosion index (a value available from the original Isoerodent Map or now in the web-based Soil Survey); the purpose of this parameter is to account for the rainfall KE in the region of interest. For a particular locality it is a function of both the maximum 30-min storm intensity (in/hr), I_{30}, and average storm intensity, I, as given below

$$R = (916+331*\log I)I_{30} \qquad [3]$$

where

I = average annual rainfall intensity (in/h), and

I_{30} = maximum 30-min storm intensity (in/h).

They divided R by 100 and imposed a limit on $I<3$ in/hr based on the finding that median raindrop size did not continue to increase when intensities exceeded 3 in/hr.

At their core, most erosion motivated studies focus on determination of the soil erodibility factor, K, or one of its derivatives. This factor is a measure of the soil susceptibility to erosion. For the USLE, K was defined quantitatively through a soil textural nomograph, or experimentally under the “standard condition” that involved a 22.13 m (72.6 ft) long unit plot with a uniform length-wise slope of 9%. The plot should be bare, tilled up and down the slope, and free of vegetation for more than 2 years. Erosion results from tests conducted on plots that were otherwise “standard”, but at slopes different then 9% could be adjusted by a simple slope equation factor.

The topographic factors, L and S reflect adjustments between hillslopes encountered in the field and the “standard” plot. For example, LS is the expected ratio of soil loss per unit area from a field slope to that from the 22.1 m standard length. L can be calculated from

$$L = (\lambda / 72.6)^{m} \qquad [4]$$

where

λ = field slope length (ft), and

m = 0.5 if slope is >5%, 0.4 on slopes of 3.5-4.5%, 0.3 on slopes of 1-3% and 0.2 on uniform slope <1%.

Similarly, S can be calculated from

$$S = 65.41\sin^2\theta + 4.56\sin\theta + 0.065 \qquad [5]$$

where

θ = angle of slope (%).

The Cover and management factor, C, and the support Practice factor, P, range from near zero to one and rather than process-based factors are equivalently defined as the ratio of soil loss from land cropped under specified cover or practice conditions to that corresponding loss from clean-tilled, bare soil.

Generally, the USLE applies only to determination of average annual soil losses from sheet, rill, and inter-rill erosion from large areas of relatively loose bare soil exposed for 2 or more years. As the USLE uses a long-term averaged annual rainfall index, it can produce misleading soil loss values when applied to seasonal or single storm events (Wischmeier 1976). Other recognized limitations are related to each of the USLE factors. Estimation of K factor may be limited to ranges of soil textures having lower clay contents (Loch, 1984) and those soils from which the nomographs were developed. McCool et al. (1987) suggested that the USLE under-estimates soil loss rates from short slopes, while Weggel and Rustom (1992) suggested that it overestimates soil losses when applied to areas other than large loose farm soil areas such as highway embankments (roadcuts) and small drainage basins. Wischmeier and Smith (1978) indicated that the best estimate range for the S and L factors is 3-18% and 10-100 m. Application of the USLE is not appropriate for flat or steep slopes, small areas, and plots with mixed soil types. Singer and Blackard (1982) noted that slope steepness factor equation (5) has not been validated for slopes >18%. Mutchler and Murphree (1985) found that the USLE greatly over-predicted soil loss on the flatter slopes. Kamalu (1992) reported that the runoff erosion rate becomes dominant on longer or steeper slopes (>9%), while the interactive combination of rainfall and runoff was dominant over other erosive forms on mild slopes (5-7%). He concluded that the runoff rate is the most important contributor to road embankment erosion. Similarly, Huang and Bradford (1993) suggested that the effects of slope steepness on sediment loss rate depended on runoff intensity. McCool et al. (1987) recommended new equations for USLE soil loss estimation from areas at slopes different then 9%:

$$\text{At slopes} < 9\%,\ S = 10.8\sin\theta + 0.03 \qquad [6]$$

$$\text{At slopes} \geq 9\%,\ S = 16.8\sin\theta - 0.50 \qquad [7]$$

For short slopes (length ≤4 m) where all erosion is presumably caused by raindrop impact they suggested

$$S=3.0(\sin\theta)^{0.8} +0.56 \qquad [8]$$

Finally, antecedent soil moisture effects on runoff and erosion rates have been well known when Le Bissonnais, Singer and Bradford (1992) reported that soil drying reduces runoff and sediment concentration, especially for high organic carbon and clay content soils. In part as a result of such limitations, the USLE was modified for broader application into the forms MUSLE and RUSLE.

As erosion rates for individual storms can be better correlated with runoff rather than rainfall rates, Williams (1975) suggested in MUSLE to replace the USLE rainfall energy factor, R, with a runoff rate dependent factor. Incorporation of the runoff factor implicitly attempts to correct the USLE for antecedent soil moisture conditions. MUSLE can be written as

$$S=95(Q\cdot q_p)^{0.56}\cdot K\cdot LS\cdot C\cdot P \qquad [9]$$

where

S = sediment yield in tons,

Q = volume of runoff in acre-feet, and

q_p = peak flow rate in cfs.

Renard et al. (1991, 1994 & 1997) introduced the Revised USLE maintaining the same fundamental structure of USLE, but with new broken down factors developed from additional data. Basically, the RUSLE revisions included:

1) Computerization of calculation algorithms.
2) New R values for western US.
3) Revisions and additions of R values for eastern US.
4) Seasonally variable K factors, (i.e., weighting K-values in proportion to the annual rainfall fraction, rock fragments fraction on and in the soil, and indication of the soil susceptibility to rill erosion relative to interrill erosion).
5) A subfactor calculation approach for C factor determination (e.g. see Dissmeyer and Foster (1980) for forested areas).
6) LS algorithms for varying shape.
7) New P values for different conditions, (e.g., rangelands, stripcrop rotations, contour farming and subsurface drainage.

Water Erosion & Prediction Project (WEPP)

With recognition of the limitation associated with the averaged annualized calculations and empirical basis of the USLE and its modifications, Nearing et al. (1990) claimed that erosion prediction technology needed to move towards development of process-based simulation models. This thinking was reflected in development of the “physically-based”, though continued semi-empirical erosion equations at the core of the

WEPP developed as something of a replacement for the empirically-derived USLE (e.g. Ascough et al., 1997; Baffaut et al., 1996; Liu et al., 1997). To date, physical modeling of soil erosion has involved the mathematical description of soil aggregate breakdown, subsequent particle detachment and their transport to stream channels or deposition on land surfaces (Nearing et al., 1994). Much of this description was taken through extension of knowledge about sediment transport in streams, and may apply reasonably well to either sheet flow over bare soils or gully erosion processes. It is not clear that these same processes apply to developed hillslope soils in which sufficient infiltration capacity exists that particle filtration may be the dominant process rather than particle detachment and transport associated with rainfall/runoff shear stresses exceeding soil, or aggregate strengths. Nonetheless, during the past few decades, there has been considerable research and development into appropriate erosion models for the prediction of soil loss and sediment delivery from bare soils. They are intended to represent the assembly of complex interactions and essential mechanisms affecting runoff and erosion rates and their spatial and temporal variability. Erosion models range in scope and application from relatively simple empirical or lumped parameter models employing primarily statistical relationships, to physically-based process models and distributed-parameter watershed models. Overall, the value of erosion models lies primarily in their predictive capability for assessing soil loss as part of conservation planning, though increasingly they are employed for setting regulatory guidelines and standards.

The basic structure of WEPP reflects its USLE ancestry, with model components for climate, soil, slope and management, but as a process-based model it can be run with a daily time step, and also configured to run in single storm mode. It offers three versions, each suitable for a different scale. The profile version is the replacement of USLE as a predictor of uniform hillslope erosion that now includes possible deposition. The watershed version is applicable at the field scale and incorporates areas where more than one profile version may exist. The grid version can be applied to areas with boundaries that do not match watershed boundaries, or it can be broken into smaller areas where the profile version may be applied (Laflen et al. 1991a). The major determinants of the WEPP erosion processes are soil resistance to detachment, available stream power (transport) and rainfall intensity that, like the USLE, are linked to erosion rates by the soil erodibility, K. The original meaning of K as used in the USLE remained more-or-less the same; that is, a factor representing the relative susceptibility of soil aggregates to breakdown and subsequent particle transport, though there is no further clarification of its precise physical definition. Thus at its soil detachment equation core, WEPP retains a level of empiricism (Owoputi and Stolte, 1995); if K values are otherwise unknown they are determined from soil textural information. Hydrologic processes included in WEPP are climate, infiltration, and a winter component that includes soil frost, snowmelt, and snow accumulation. Plant growth and residue processes estimate plant growth and decay above and below ground. The water balance component uses climate, plant growth, and infiltration to quantify daily potential evapotranspiration, which is necessary to compute soil-water status and percolation. The hydraulic component computes shear forces exerted on soil surfaces assuming turbulent flow and friction factors (a function of surface roughness). Soil processes that are also considered involve various soil parameters such as roughness, bulk density, wetting-front suction, hydraulic conductivity, interrill and rill erodibilities, and critical shear stress. Rather than employing quantifiable factors that could be associated with the soil aggregate stability, shear strength, organic matter or "tilth", WEPP employs USLE-type cover and management factors that account for weathering, tillage, plant growth, residue and biomass development above and below ground. Numerous trial runs, plot runoff, flume and calibration studies were conducted across the USA to expand the range of erodibility values for the WEPP generally from disturbed soils on relatively mild slopes in primarily agriculture but also some rangeland and forest road settings (e.g. see WEPP, 1995 database). Siepel et al. (2002) expanded use of Manning's roughness in determining erosion rates under grass vegetated surface conditions and show that a certain minimal cover is required to trap suspended sediment. Similarly, Grismer and Hogan (2005b) found that less than ~40% grass cover had little effect on reducing erosion rates on Tahoe Basin skiruns, a result echoing earlier work by Blackard and Singer for grass covers and European studies for rock cover fractions. Later research developments have largely focused on expanding capability aspects of WEPP including flow over stony soils (e.g. Li and Abrahams, 1999) and particle sorting (e.g. Flanagan and Nearing, 2000), as well as broadening its application and assessing its performance (e.g. Nearing et al., 1990; Zhang et al., 1996; and Laflen et al., 2004).

Although WEPP may offer more capability than the empirical RUSLE model, to some degree, RUSLE is a relatively simple to apply proven technology, while WEPP is more complex and has not necessarily provided more precise, or realistic estimates of erosion rates (Tiwari et al., 2000; and Laflen et al., 2004). Recent upgrades to the WEPP computer interface have made the program far more accessible to a broader user group.

Assuming dominance of Hortonian and turbulent runoff processes, the WEPP can be used to model both erosion and deposition on a hillslope, and generates sediment mass and runoff particle-size-distributions (PSDs) in terms of fractions of sand, silt and clay. This runoff assumption is more appropriate to highly disturbed areas such as roads than vegetated, less disturbed areas where overland flow is often not observed (Dunne et al., 1991; Croke et al., 1999). Consequently, WEPP does not model saturation excess flow generation thereby limiting its application in shallow slope forested areas of the watershed, though recent improvements better account for subsurface flow processes (Wu and Dunn, 2005). WEPP employs a steady-state sediment continuity equation combining inter-rill and rill soil losses that in turn relies in part on the kinematic wave and Mannings equations relating flow cross-sectional areas to discharge. As a result, there is some ambiguity associated with the applicability of these equations to slopes >10% for which the Mannings equation no longer applies. The inter-rill and rill erosion expressions in the continuity equation are modeled as particle detachment and transport either by raindrops and shallow flows (inter-rill), or concentrated flows (rill), respectively.

$$D_i = K_i \, i \, q \, S_f \, C_v \tag{10}$$

Where D_i = interrill detachment/transport rate (kg m^{-2} s^{-1}),

K_i = interrill erodibility (kg m^{-4} s^{-1}),

i = rainfall intensity (m s^{-1}),

q = runoff rate (m s^{-1}),

S_f = interrill slope factor = $1.05-0.85e^{-4\sin\theta}$ where θ=slope angle, and

C_v = cover adjustment factor ($0< C_v <1.0$).

The interrill slope factor was determined from a best-fit, non-linear regression between slope (%) and the ratio D_i/i^2K_i means from several researchers (Liebenow et al., 1990); nine of the 12 points used were from micro-plots at slopes <20%, one at 30% and two at ~50%, reflecting the very limited availability of erosion rates from more steep slopes. Note that fundamentally D_i could also be expressed in terms of stream power, ***P***, the product of runoff rate and slope (e.g. Zhang, et al. 2002).

$$D_r = K_r\,(\tau-\tau_c)\,(1-Q_s/T_c) \tag{11}$$

Where D_r = rill detachment/transport rate (kg m^{-2} s^{-1}),

K_i = rill erodibility due to hydraulic shear (s m^{-1}),

τ = shear stress (product of unit weight, γ, hydraulic radius & slope, Pa),

τ_c = critical shear stress below which soil detachment does not occur (Pa),

Q_s = rate of sediment flux in rill (kg m^{-1} s^{-1}), and

T_c = rill sediment transport capacity, a power function of τ (kg m^{-1} s^{-1}).

Characterizing Erosion - Continued Developments

While equations (10) and (11) represent an accumulated development of the past several decades, they perhaps originate from Ellison's (1947) observation that "erosion is a process of detachment and transport of soil materials by erosive agents". These "erosion agents", of course, include raindrop impact and overland flow. Subsequent research more-or-less begins with this paradigm of sorts that continues in concept through the soil detachment equation review by Owoputi and Stolte (1995). Foster and Meyer (1972) interpret results of several experiments in terms of Yalin's equation that assumes "sediment motion begins when the lift force of flow exceeds a critical force … necessary to … carry the particle downstream until the particle weight forces it out the flow and back to the bed." Bridge and Dominic (1984) build on this concept and describe the critical velocities and shears needed for single particle transport over fixed rough planar beds. Gilley et al. (1985a & 1985b) include the Darcy-Weisbach friction factor as a measure of the resistance to flow eventually adopted in the WEPP model. Moore and Birch (1986) combine slope and velocity and suggest that particle transport and transport capacity for both sheet (interrill) and rill flows is best derived from the unit stream power. Assuming turbulent flow conditions, stream power, ***P*** can be expressed as

$$\boldsymbol{P} = vS = n^{-0.6}\, q^{0.4}\, S^{1.30} \qquad (12)$$

where n is Manning's roughness, S is slope (m/m) and the other parameters are as defined above. This equation differs only slightly when assuming laminar flow conditions, but without the <10% slope limitation implicit in the Manning's equation assumption, and can be written as

$$\boldsymbol{P} = (\gamma/3\mu)^{0.33}\, q^{0.67}\, S^{1.33} \qquad (13)$$

where μ is the water viscosity. Note that in both equations (12) and (13), slope has a larger effect on stream power, hence detachment rate, than runoff rate. This suggests that some power form of these two parameters should likely be used in equations (10) and (11).

Experimentally, the dependence of stream power on slope between laminar and turbulent flow is probably indistinguishable and the role of stream power on detachment rates is still likely affected by rainfall rates and soil resistance to detachment or aggregate breakdown. In fact, at slopes of 4-12%, McCool et al. (1987) found soil loss rates dependent on $S^{1.37}$ to $S^{1.5}$, rather than ~$S^{1.3}$. In flume studies, Zhang et al. (2002) found that across a range of slopes (3-47%) their detachment data was proportional to $q^{2.04}\, S^{1.27}$ confirming dependence of ***P*** on slope, but suggesting that both equations above may underestimate the effects of runoff rate. At low slopes, detachment rate was more sensitive to q than S, however as S increased, its influence on detachment rate increased. Later, Zhang et al. (2003) found that for undisturbed "natural" soils across a similar slope range (9-47%), detachment was proportional to $q^{0.89}\, S^{1.02}$. In both cases, detachment was a strong power function of q alone for the disturbed and undisturbed soils, that is, $q^{4.12}$ and $q^{3.18}$, respectively, somewhat larger than the $q^{3.0}$ suggested by Eq. (12).

Nearing et al. (1991) noted that hydraulic shear stress can be expressed either in terms of runoff rate or flow depth (a very difficult parameter to measure in practice), but that detachment of different particle-size classes was a logarithmic function of slope, flow depth and particle weight. On the other hand, detachment rates were not unique functions of either stream power or shear stress, but were most dependent on slope, though slopes used were quite flat (1-2%). On the same nearly flat slopes but with deeper flow depths (~10 mm), Nearing and Parker (1994) found that turbulent flow resulted in far greater soil detachment rates than did laminar flow in part as a result of greater shear stresses as suggested by Equation (11). Following Gilley and Finkner (1985), Guy et al. (1987) examined the effects of raindrop impact on interrill sediment transport capacity in flume studies at 9-20% slopes. Assuming laminar flows, they found that raindrop splash accounted for ~85% of the transport capacity, in some contrast to earlier studies indicating that raindrop impact had little or no effect on slopes greater than about 10%. Adding to the possible confusion, Romkens et al. (2001) found that sediment concentrations from lab studies on 3.7 m long plots at slopes of 2, 8 and 17% were practically the same after repeated storms for up to two hours despite a positive relationship between runoff rate and slope. They attributed this lack of slope dependence of erosion rates on the surface roughness of the bare soils as compared to that from a smooth surface. Chaplot and LeBissonnais (2003) found that sediment losses from agricultural loess soils at slopes between 4 and 8% were unaffected by slope at 1 m lengths and was significant at 5 m slope lengths. Sharma et al. (1991, 1993 & 1995) systematically examined rainsplash effects on aggregate breakdown and particle transport in the laboratory. Echoing Singer and Blackard (1982) who suggested that raindrop impact significantly affected erosion rates at slopes up to 35-40%, Fox and Bryan (1990) argued that rain-impacted sheet flow erosion "*increased roughly with the square-root of the slope*" (2 to 40%) and soil losses were correlated with runoff velocities. At greater slopes, Lei et al. (2001) found that both slope and runoff rate were important towards transport capacity up to slopes of about 44%, but that transport capacity increased only slightly at steeper slopes.

Clearly, the original Ellison paradigm of the erosion process continues to direct erosion-process related research. In Owoputi and Stolte's (1995) review, they suggest that semi-empiricism implicit in equations such as (10) and (11) should be replaced by more careful definitions of the forces (including rainfall, runoff and soil resistance to detachment, i.e. erodibility) acting on hypothetical soil particles or aggregates. Presumably from there, the forces or energy needed for aggregate breakdown could be applied (Fristensky & Grismer, 2008) to determine the extent of finer particle liberation and subsequent transport. For example, Sharma et al. (1991) determined that the threshold KE needed to initiate soil detachment depended on soil strength and clay content ranging from 0.2-0.6 mJ. Owoputi and Stolte underscore the need to account for the moisture dependence of soil strength and seepage, though in a rainfall or runoff induced erosion event it is likely that at least the surface soil layers are at or near saturation, that is, their weakest state. Similarly, in a thorough review of raindrop impact induced erosion processes on mildly sloping bare soils, Kinnell (2005) noted that current "models do not represent all of the erosion processes well." None deal with temporal changes in surface properties and all simplify the process descriptions to a planar surface lacking the microtopography variations or surface roughness found in even relatively smooth field soils. Ideally, the soil erodibility would be quantitatively defined as a detachment/transport coefficient relating detachment rates to an appropriate form of stream power. Zhang et al. (2003) found nearly a linear relationship between D_r and $\boldsymbol{P}$, or shear stress at low detachment rates from disturbed and "natural" silt loam cores, however, power functions of $\boldsymbol{P}$ best fit the detachment rates overall (i.e. $\boldsymbol{P}^{1.62}$ and $\boldsymbol{P}^{1.07}$, respectively). It is likely that increasing stream power has a decreasing effect on aggregate disintegration and there may be a practical threshold of stream power effects to consider in detachment modeling (Fristensky & Grismer, 2009). Thus, either the physical process description given by equations (10) and (11) are inadequate, or the concept of erodibility needs greater clarification and evaluation. As Zhang et al. (2002) comment "*a large gap exists between fundamental erosion processes and erosion models … until we are able to fully understand … we are forced to continue using essentially empirical parameters, such as those used by WEPP*". Erosion processes are sufficiently complex that questions of laminar versus turbulent flows in the field, the fundamental applicability of the turbulent flow based shear stress equations at slopes greater than 10%, the discrepancy between measured and modeled soil shear strength (100's *vs.* 1 Pa, respectively), and raindrop effects especially on steeper, relatively undisturbed forest soils remain unresolved, while more precise definition of erodibility remains elusive

(Agassi and Bradford, 1998). They acknowledge that "*erodibility is a dynamic soil property … not a fundamental soil property but is defined by the specific erosion equation … and the conditions under which the value was obtained*." Further, "*erodibility values reported in the literature are often soil properties correlated with soil loss from areas where both rill and interrill processes occur simultaneously*." As such, "*erodibility is not a process-based term in most soil … depending on whether detachment or transport is limiting sediment yield, erodibility can vary between two extremes, and the extreme erodibilities are dominated by different soil factors.*"

This research briefly summarized above and others like it, by necessity is conducted on bare soils and as a result may not apply to mulch/duff matted forest soils in which the dominant sediment "detachment and transport" processes are not characterized by any of the equations above, rather perhaps a filtration process (Grismer, 2007). Such uncertainties in the meaning of basic erosion parameter definitions set the stage for evaluation of RS methods in the field.

NATURAL AND SIMULATED RAINFALL ENERGIES AND INTENSITIES

Before considering the variety of spray nozzle or drop-former type RSs used in field research, the reported characteristics of "natural" and "simulated" rainfall are broadly outlined. The role of raindrop velocity or energy in splash detachment of soil particles has been a concern for decades (e.g. Ellison, 1944; and Bisal, 1960). There has been some debate whether raindrop size, velocity, momentum, kinetic energy (KE), or some combination thereof is/are the key parameters of design concern with respect to RS use for erosion studies. In addition, these parameters need to be considered together with a threshold concept that can account for the limited erosion rates encountered during low intensity storms (for which use of KE alone tends to over-estimate erosivity). Nonetheless, in contrast to many early studies, more recent work generally includes determination of the rainfall KE as a measure of the total energy available for aggregate disintegration, detachment and eventually transport. These estimated KEs depend in part on drop sizes and their distribution. Figure 1 illustrates how median drop size (D_{50}) of natural rainfall varies with intensity from several studies and suggests that drop sizes of ~2.5 mm may be appropriate for simulated rainfall at the intensities often employed when using RSs in the field. Figure 2 illustrates dependence of drop-size distributions expressed as a fraction of the rain event volume on rainfall intensity and underscores that relatively low intensity events are dominated by drop-sizes <1 mm while rainfall intensities between 40 and 120 mm/hr are associated with a median drop size of ~2 mm. Cerda (1997) cautions that a larger data set would be advisable to confirm such findings as shown in Figures 1 and 2, especially under very high rainfall intensities that are extremely rare and highly difficult to measure.

Van Dijk et al. (2002) reviewed studies of the relationship between rainfall drop sizes, intensity and KEs and developed a generalized equation from storm events in SE Australia as summarized in Table 1. Note that in Table 1 when expressed on a per unit depth basis, the overall storm KE decreased to ~19 J/m^2-mm with increasing storm depth class. Generally, KE variability within a small range of overall storm depths was +/-10%. He characterized the relative quality of measured storm KE values from around the world and found that "good" quality data, KE ranged from 11 – 36 J/m^2-mm with maximum values that averaged ~29 J/m^2-mm and minimum values about 12 J/m^2-mm. Particular KE values depended on locations, type of storms and storm patterns or storm hysteresis effects on the measured KEs. For example, Figures 3 illustrates the effects of storm type and rainfall intensity on the KE produced by the event. Again, in part (b) of Figure 3, note that high intensity actual storms typical of RS studies (>40 mm/hr) result in an average KE of 23-24 J/m^2-mm. These latter KEs are similar to that suggested by Renard et al. (1997) for natural rainfall having an intensity of 40 mm/hr.

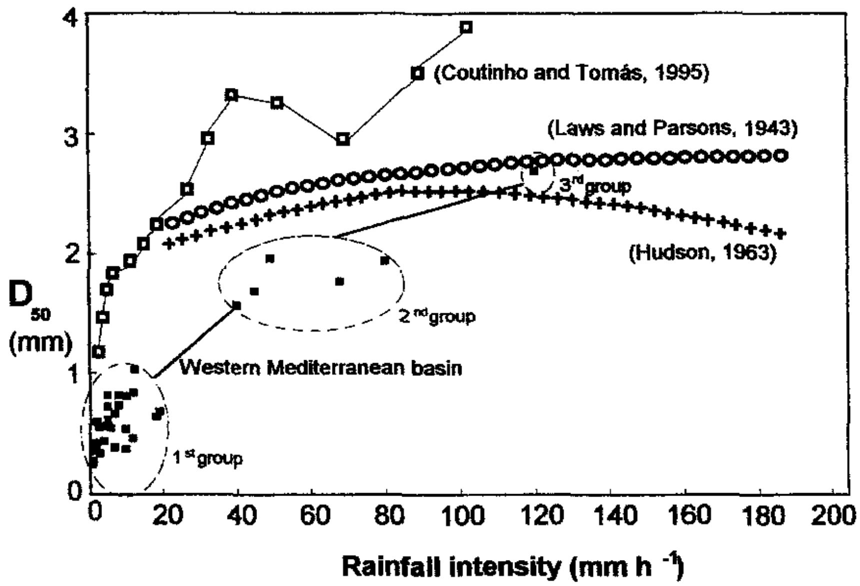

Figure 1. Comparison of natural rain D_{50} drop sizes for storms from the Western Mediterranean basin and that collected by Laws and Parsons (1943) and Hudson (1963) (from Cerda, 1997).

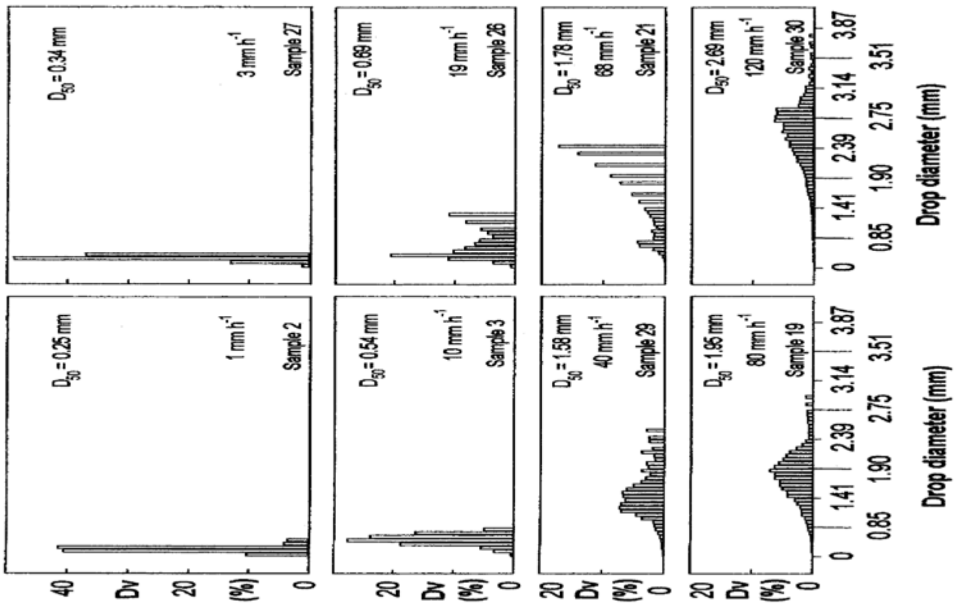

Figure 2. Comparison of natural rain D_{50} drop sizes and drop-size distributions by fraction of rain volume (from Cerda (1997).

Table 1. Summary of measured rain event intensities, overall depths and KEs for NSW, Australia (from van Dijk, 2002).

Storm class		Total depth	Average Intensity	Average KE	Average KE
(mm)	number	(mm)	(mm/hr)	(J/m^2)	(J/m^2-mm)
0-2.5	10	12.4	3.1	31	25.0
2.5-5.0	2	6.2	4.7	79	25.5
5.0-25	8	104.5	6.2	322	24.7
25-50	3	115.6	4.4	730	18.9
>50	1	93	6.3	1770	19.0

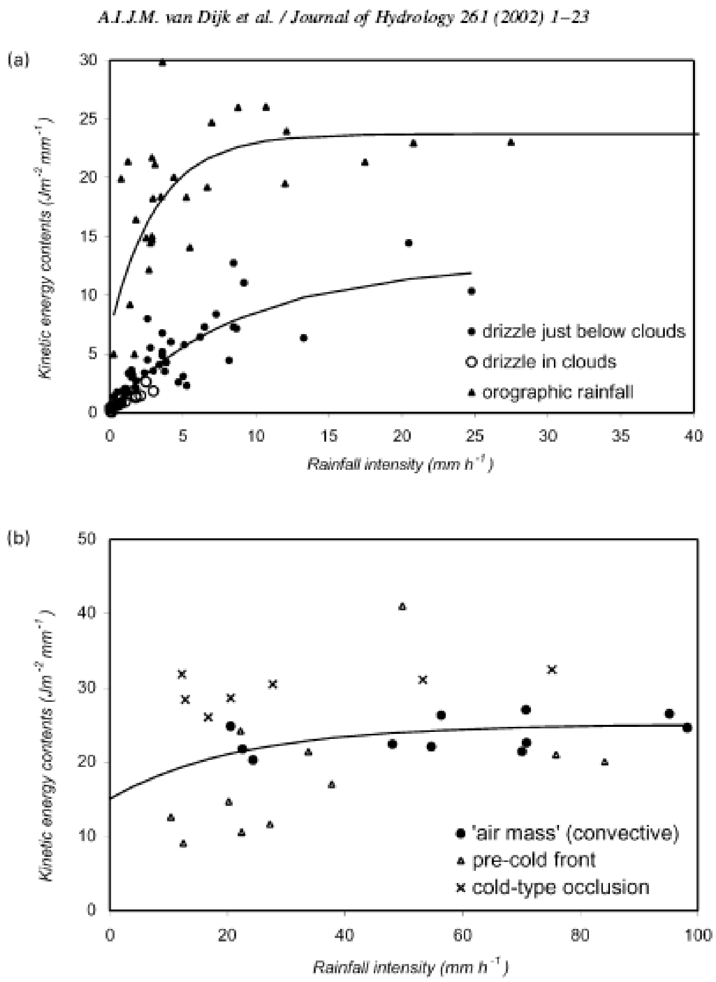

Figure 3. Dependence of raindrop energy on storm type and intensity (van Dijk, 2002).

Overall, van Dijk (2002) commented that

> "*in terms of process-based research, it appears that our knowledge of the distribution of drop size and terminal velocity in natural rainfall is well ahead of our understanding of the way in which these interact to detach and transport soil particles by splash. If rain falling at high intensities is compared to that falling at low intensities, the former appears to be considerably more effective in detaching soil than is to be expected from the difference in KE alone. Although results from*

laboratory studies go some way to explain this phenomenon, such experiments have been fraught with interpretational difficulties. Moreover, the translation of laboratory results to field simulations is not straightforward because of the fundamental differences between the drop size distributions and fall velocities of artificial and natural rainfall.

Dunkerly (2008) laments that the most RS based studies employ extreme rainfall intensities for the region of application and/or durations with an over-emphasis or focus on drop sizes, their distributions and KEs. Considering some 49 different studies, Dunkerly found that the average RS intensity of 103 mm/hr (+/- 81 mm/hr) is often sustained for nearly an hour; a rate 30 times greater than the mean natural rate and when combined with the long duration generates, an event that exceeds that of even extreme events is most locals. Moreover, he speculates that drop arrival rate may be the critical rainfall factor to subsequent "downstream" transport of sediment; however, rarely is such information provided. Finally, Dunkerly (2008) concludes that:

(a) "It is vital to analyze and report the relevant storm properties, whether in natural or simulated rain, when accounting for observed patterns of soil loss, nutrient loss, overland flow, etc. Only in this way can the relative roles of storm and soil properties be disentangled.

(b) One primary reason for adopting the use of rainfall simulation as a research tool is to reproduce in a controlled way the behaviour expected in the natural environment… Less attention appears to have been paid to correctly reproducing other event properties, including duration, mean rain rate, and the temporal pattern and magnitude of rain rate fluctuations. Other properties seem to have received little attention, including the density of droplet impacts per unit area and unit time ('raindrop arrival rate') at the soil surface…However, even where general principles are being explored, the results have diminished value if the imposed rain event properties do not lie within the range commonly experienced at field sites where the results are intended to find application."

Few direct measurements of KE for simulated and natural rainfall exist; rather, KEs are estimated from drop-sizes, assumed distributions and fall heights, or terminal and nozzle velocities. Kinnell developed a distrometer for measurement of raindrop size distribution and energy as a function of rainfall intensity. Overlooked by the review of van Dijk (2002), Madden et al. (1998) used a piezoelectric crystal to directly measure natural and simulated raindrop power (KE/unit time) and found that both rain power and intensity varied greatly within natural events, and that power varied considerably even at any given rain intensity. Simulated rains at intensities of 23 to 48 mm/h developed powers of 200- 1320 J/m^2-hr, while natural rainfall powers for 85 events ranged from ~200 to ~3000 J/m^2-hr at intensities between 1-42 mm/hr, but reached as much as 6000 J/m^2-hr for a short high-intensity storm event. When lacking direct raindrop size measurements, the Marshall-Palmer or gamma (Fox, 2004) size distributions are the most widely assumed, while terminal velocities determined in "rain tunnel" chambers or from theoretical drag considerations are used together with drop masses to determine KE. Though the original drop terminal velocities of Laws (1941) are the most commonly cited, more recent studies that correct for drop "flattening" during fall as they depend on atmospheric pressure and temperature have been developed (Wang & Pruppacher, 1977). In their rain tower experiments, they found that drop size in rainfall is limited to ~ 4 mm and the terminal velocities of the larger drops > 2 mm are limited to about 9 m/s, for 1.4 mm drops terminal velocities are~8 m/s and for small drops ~1 mm about 6 m/s. Of course, the related fall heights necessary to achieve these terminal velocities also decreases with decreasing drop size such that small drop sizes reach near terminal velocities within only a few meters of fall. Figure 4 illustrates the dependence of raindrop power on drop size, rain intensity and fall height developed from the work of Wang & Pruppacher (1977), while Figure 5 compares this work for 2 mm drop sizes to that estimated from equations developed by Wischmeier and Smith (1958) and van Dijk (2002) for natural rain. For rainfall intensities less than ~90 mm/hr, rainfall powers at near terminal velocities (20 m fall height) are less than the relative maximum~3000 J/m^2-hr measured by Madden et al. (1998) for rainfall intensities less than half as great. Moreover, the rainfall powers of the short, high-intensity storm power of ~6000 J/m^2-hr measured by Madden et al. (1998) seem unlikely to be generated by RSs. There is also a question about how rainfall power compares with that needed for aggregate breakdown (Fristensky and Grismer, 2009). The average upper range of rain impact powers between 3000-4000 J/m^2-hr, or approximately 1 W/m^2 is far less than the 4-14 W applied in aggregate stability studies (see Figure 6). In terms of RS erosion related research, Schiotz et al. (2006) summarized the "frequently used" KE relationships for natural rain developed by low intensity (10 mm/hr) storms and questions the broad range in computed values in their Table 6, reproduced here as Table 2. While it is interesting to note that for the natural rainfall events considered by van Dijk (2002) and the generalized KE-intensity curve suggested by Wischmeier and Smith (1958), the ranges in KEs for relatively low intensity storms (from the perspective of RS studies) of ~20 mm/hr ranges from 16-38 J/m^2-mm, while at the range of intensities of 40 – 100 mm/hr often used in RS studies the average KE is ~23-28 J/m^2-mm (see Table 1 & Figures 7). Whatever this range of KE at a given intensity means with respect to evaluation of erodibilities remains unclear.

Table 6: Frequently used kinetic energy (E_K) - rainfall intensity relationships with values for kinetic energy at a rainfall intensity of 10 mm h^{-1}.

Region	No. of observations	Kinetic energy E_K (Jm^{-2} mm^{-1}) with rainfall intensity I (mm h^{-1})	E_k (J m^{-2}) for 10mm h^{-1}	Reference
Washington D.C., USA	95	E_K= 11.87 + 8.73 log I	20.6	Wischmeier & Smith (1958, 1978) based on data from Laws & Parsons (1943)
Denmark	2.5x10^6 drops	E_K = 6.261 ln I+9.771	24.19	Pedersen & Hasholt (1995) based on data from Jensen (1981)
Zimbabwe	n.a.	$E_K = 29.8 - 127.5\ I^{-1}$	17.05	Hudson (1961)
Ottawa, Canada	n.a.	E_K = 8.95+8.44 log I	17.39	Marshall & Palmer (1948)
Southern central United States	496	*$E_K = 429.2+534\ I-122.5\ I^2 +7.8\ I^3$	16.39	Carter et al. (1974)
Miami, USA	n.a.	$E_K = 29- (1- 0.74^{(-0.039I)})$	14.47	Kinell (1980)

Table 2. KE – rainfall intensity relationships as summarized by Schiotz et al. (2006).

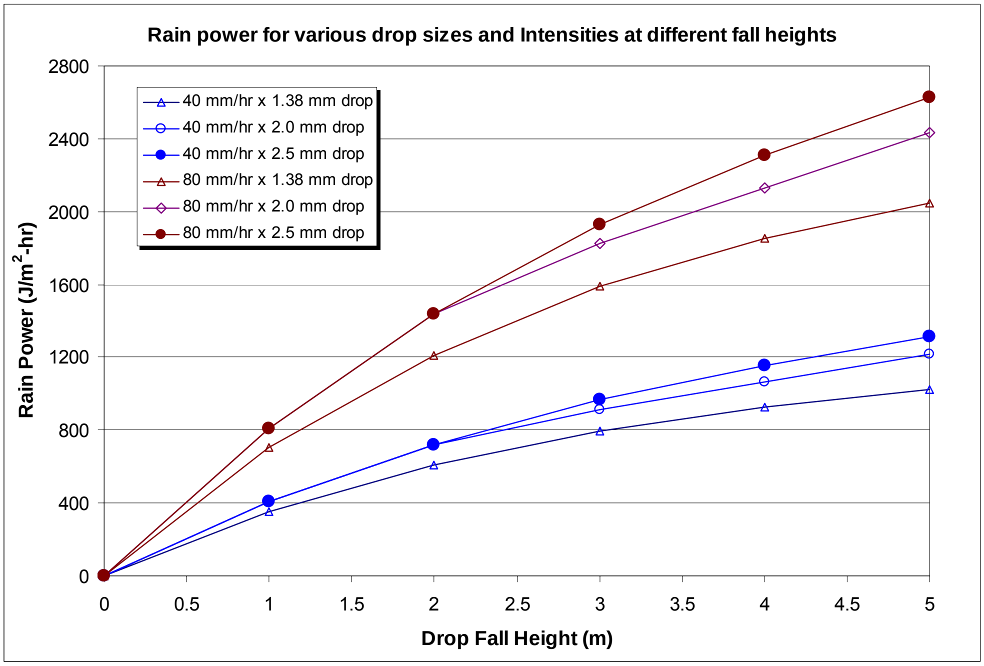

Figure 4. Dependence of raindrop power on drop size, rain intensity and fall height.

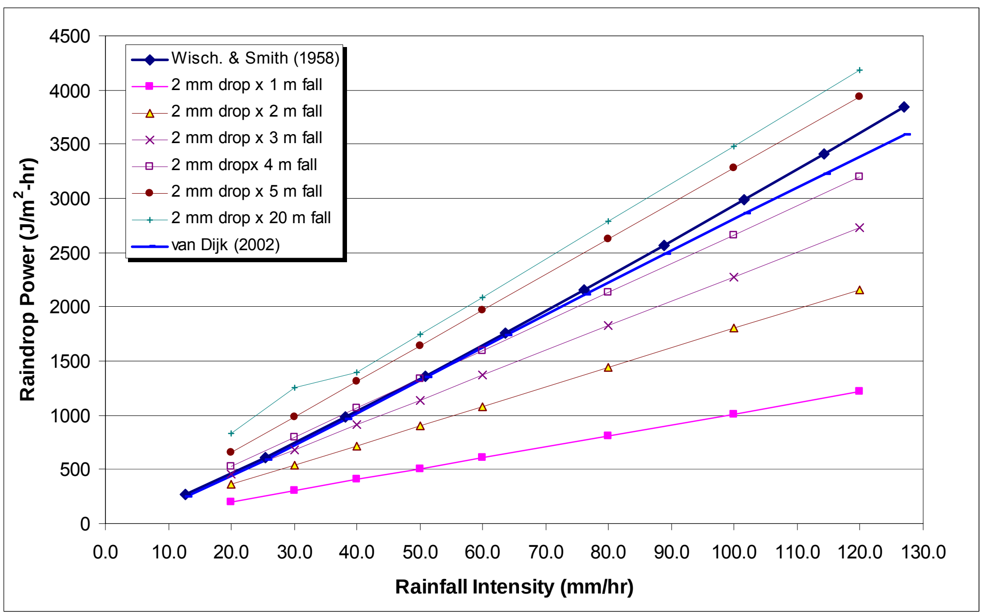

Figure 5. Dependence of raindrop power on rain intensity for a 2 mm drop size and fall heights from 1-20 m as compared to natural raindrop power equations developed by Wischmeier and Smith (1958) and van Dijk (2002).

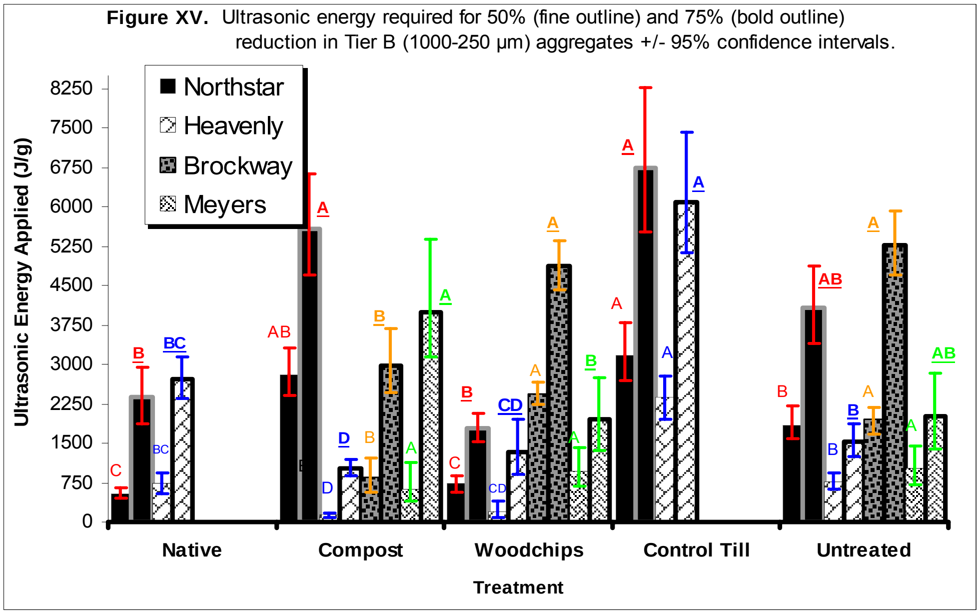

Figure 6. Energy required for disintegration of half of the large aggregate size most closely associated with soil tilth and erosion potential under different soil type and treatments in the Tahoe Basin (Fristensky & Grismer, 2009).

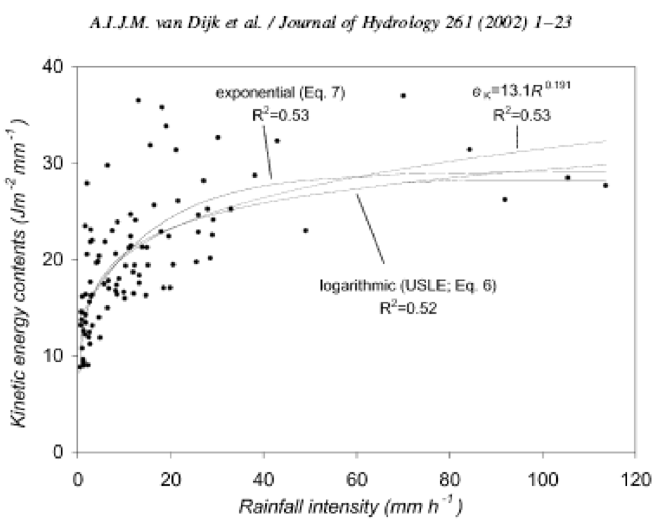

Figure 7. Comparison of different regression fits to the dependence of rainfall KE on natural rain intensity as developed by Wischmeier and Smith (1958) and van Dijk (2002).

RAINFALL SIMULATOR DESIGNS

RS methods to assess various erosion control or treatment technologies have been widely used and comprehensive reviews are available from Sutherland (1998a & 1998b). Their use in erosion studies is not new (Young and Burwell, 1972). Sutherland noted that the "formative years" prior to ~1990 resulted in a mass of information that lacked scientifically creditable, standardized methods or data from actual applications. He argued for standardized evaluation methods that have field applicability and greater emphasis on study of surface, or near surface processes controlling erosion, a matter that has only been slightly addressed in subsequent studies. Relatively portable RSs have been more commonly deployed in the past 2-3 decades with corresponding plot areas of 1-2 m^2 that are well suited to a wide range of field studies, particularly where access is difficult, or if multiple replications are needed across a larger area. They have been used to study runoff and erosion mechanisms in a wide range of environments; however, in practice these RSs tend to compromise natural rainfall characteristics, due to portability, cost design and/or management limitations (Meyer 1988). However, direct field measurements of runoff and erosion rates as well as to some degree modeling approaches capable of predicting these rates from less-disturbed forest and rangeland soils (as compared to bare compacted or tilled soils) remain few. While runoff and erosion rates per unit area from rangeland and forest soils are generally much less than that from more disturbed soils, these soils often comprise substantially larger areas within watersheds and may contribute significant loading to streams. Determination of net erosion mass per unit area as with USLE is no longer adequate and information about the runoff particle-size distribution (PSD), nutrient content and contaminant concentrations from erosion control treatments or soil restoration efforts for particular storm events is needed to evaluate their relative performance (Grismer, 2007). Concerns about lack of standardized RS methodologies or designs and precise determination of the process being measured are not new as Lal (1998) and Agassi and Bradford (1999) suggested there is an inability to compare results between studies, and possibly as a result, generation of unreliable erosion rate predictions. Meyer (1988) contended that the results from simulated rainfall only give relative, rather than absolute, erosion data; and that to correlate the simulation results to that of natural events, data from similar plots subject to long-term natural rainfall events must be available for comparison, such a comparison later reported by Hamed et al. (2002) for example. Nonetheless, RSs in the field continue to be developed and used as there is little replacement available for generating process-based erosion information.

Basic RS Designs - Overview

RS design encompasses two challenges, duplication as closely as possible the physical characteristics of natural rainfall, and to do so with a device that matches the process scale of interest and resources available. The two types of RS mechanisms that have emerged in field research can be broadly categorized as spray/sprinkler nozzle and drop-former types that develop intensities of 10 to 200 mm/hr and drop sizes of 0.1 to 6 mm. Sizes of RSs have ranged from the simple, very small portable infiltrometer with a 15 cm diameter rainfall area (Bhardwaj and Singh, 1992), to the complex Kentucky Rainfall Simulator covering a 4.5 m by 22 m plot (Moore et al., 1983). The design or type of RS has been directed at meeting the often competing demands of "replicating natural" rainfall, ease of portability across remote, difficult or steep terrain, costs of construction and uniformity of simulated rainfall across the test plots in terms of intensity, drop-sizes and KEs. Duplicating both the range of drop sizes and KE of natural rainfall has proven quite difficult; likewise is development of a controllable, uniform, or even distribution of rainfall across the plot. Many of the original laboratory RSs were of the nozzle type, presumably due to ease of construction, with laboratory-based drop former RSs emerging later as a response in part to the uncertainties associated with nozzle-generated drop sizes, distributions and intensities. During the past decade, examples of RSs used in a variety of field environments across a range of slopes for plot sizes on the order of 1 m^2 that have emerged as something of standards include the oscillating veejet nozzle systems, perhaps most completely described by Paige et al., (2003) and the needle drop-former RSs of the type described by Battany and Grismer (2000). Assuming cost and portability are relatively equivalent, the differences between these two types of RSs is related to their simulated rainfall characteristics.

Development of the two primary types of rainfall simulators (i.e. spray nozzle and drop-formers) for field and laboratory research during the past three decades is outlined below and example characteristics of several more recently report RSs are summarized in Table AA. In the past decade alone, use of roughly 40 different RSs in erosion related research has been reported in more than a dozen different types of journals, of which ~80% are of the nozzle type and the remainder variations on drop-former type RSs. Advances in nozzle-type RS have been use of multiple and different spray nozzles and use of computer controlled solenoid switches/valves that rotate, sweep or vibrate the spray nozzles (Norton and Savabi, 2010). Advances in drop-former type RSs include use of greater areal density hypodermic needles in vibrating, or rotating chambers, or use of "screens" below the drop-formers to partially manipulate drop-size distributions. As drop-former type RSs are more difficult to construct and possibly maneuver in the field, nozzle-type RSs are more common, but require additional equipment and power as compared to drop-former type RSs. More sophisticated vehicle-supported designs utilizing capillary drop formers (Onstad et al., 1981), multiple sprayers (Guelph RSII, Tossell et al., 1990a & b), or rotating-disk sprayers (Green and Sawtell, 1992; Thomas and El Swaify, 1989) are appropriate where vehicle access to study sites is possible. Such RSs require truck access, considerable water and have limited mobility and applicability to steeper slopes (e.g. Norton and Savabi, 2010). Simpler drop-former designs are commonly used where access is more difficult, or there is limited water availability (Munn, 1974; Wierda et al., 1989; Robinson and Naghizadeh, 1992; Naslas et al., 1994; Clarke & Walsh, 2007).

Table AA. Summary of example reported RS characteristics from studies between 1990 and 2010.

RS Description **Lab - Field** **NZ**=Nozzle **DF**=Drop former	**Drop fall height (m)**	**Intensity range or used (mm/hr)**	**Median (D_{50}) drop size (mm)**	**Rainfall KE ($J/m^2.mm$) or Power**	**Intensity Distribution Uniformity (CU, %)**	**Plot size (m^2)**	**Reference**
Field-NZ Three screened "F" nozzles	3.0	60-120	3.7	1450-2900 J/m^2-hr		0.6x0.76 m	Designed by Gifford (1968); used by Guerrant et al. (1990)
Field- NZ 1-3 screened "F" nozzles @29 kPa	1.4	2-86		23	87-92	1-3	Miller (1987)
Field-NZ 6.35 mm impact sprinkler nozzle	1.37	12-63	1.8				Designed by Miller & Mahannah (1982); in Guerrant et al. (1990)
Field-NZ 180^o fan nozzle & 6.35 mm impact sprinkler	2.13	80-100	1.6				Guerrant et al. (1990)
Field-DF	2.5					0.9x1.52 m	Freebairn and Gupta (1990)
Field-NZ Guelph RS with full jet nozzles	1.5	18-200			88-90	1.0	Tossell et al. (1990a & b)
Field-NZ	3.0	48 & 58		13.1		2.56	Navas et al., (1990) & Navas (1993)
Field-DF 500-23 gage needles in 1m^2 rotating disk	1.4	80-100	2.5	1060-1330 J/m^2-hr		0.6x0.76 m	Designed by Malekuti & Gifford (1978); used by Guerrant et al. (1991) & Naslas et al (1994)
Field-DF 554-0.56 mm Teflon tubes per m^2	2.7	45	3.0	75% of terminal	91	0.76x0.76 m	Commandeur (1992)
Field-DF	2.0					0.5	Wierda and Veen (1992)

Field-NZ		10-150			>90	1.0	Claassens and Van der Watt (1993)
Field-NZ		54	1.6	23.9		8.0	Parsons et al. (1994)
Field-NZ Northfield		100		28.6			Malinda (1995)
Field-NZ Eight sprinkler heads	1.83	13-300	?			0.9x1.8 m	Byars et al. (1996)
Field-NZ Many sprinkler heads	3.0	25	1.52		91	14.6x42.7 m	Sumner et al. (1996)
Field- NZ 1 to 3 HARDI-1553-10 nozzles @ 144 kPa with diffuser & mesh	2.0	54	2.53	7.1	92.3	0.24 m^2 circular	Cerdá et al. (1997)
Lab-DF		68		18.1		lab	Ben-Hur & Keren (1997)
Field-NZ Spray System ½ HH10 40 nozzle @ 69 kPa		75	2.99	17.25	70	1.0	Morgan et al. (1997)
Field-NZ Rotating boom		60				3x10 m	Frasier, GW et al. (1995)
Lab & Field-DF		23-48		240-1320 J/m^2-hr		Lab sensor	Madden et al. (1998)
Lab-DF (0.8 mm holes in ½" PVC pipe)	1.8	12-120	6.7	212-2124 J/m^2-hr		0.66	Liu et al. 1998
Field-NZ Hollow-cone nozzle @ 200 kPa	2	40	0.75-1			0.28 (circular)	Designed by Calvo-Cases et al. (1988) & Lasanta et al. (1994); used by Cerdá (1999) & Seeger (2007)
Field–DF 49 plastic tubes	1.5	15-130	4.7	12.7			Designed by Irurtia & Mon (1994); Modified by Aoki & Sereno (1999) & Aoki & Sereno (2006)
Field-DF 864-22 gage needles/m^2	3.0	60	2.58	24.2	91.7	0.64	Battany & Grismer (2000)

Field-NZ Spray System 1H106SQ nozzles @ 41.4 kPa	6.58	65	2.4	23.5	78-92	5 x 10 m	Esteves, M. et al. (2000)
Lab-DF 21 gage tubing	14.	12.7 & 51	1.9 & 2.6	95% of terminal		1.0	Regmi, TP and Thompson, AI. 2000.
Field-NZ Oscillating Veejet 80100 nozzle @ 41 kPA	2.4	>40	1-3	29.5		1.5x2 m	Loch (2000a & b))
Field-NZ 3/8 GG20W & 1/3 HH35W nozzles @ 1 bar	3.6	33 & 60	1.05 & 1.85	275 & 1070 J/m^2-hr	89 & 94	4.0	Martínez-Mena et al. (2001)
Lab-NZ Three veejet 80150 oscillating nozzles		15-60		27		3.7 x 0.6 m	Romkens et al. (2001)
Lab & Field-NZ Oscillating Veejet 80100 nozzle @ 41-55 kPa	2-3	13-178 in steps of 13	3.0	25.7-27.1	87-91	2-12	Paige et al. (2003)
Field-DF	1.0	75-120	2.28		91.9	0.64	Grismer & Hogan (2004)
Field-NZ						1.2 x 12 m	Cornelis et al. (2004)
Field-NZ Veejet 80100 nozzles		65, 86 & 115			95		Herngren et al (2005)
Lab- pendant DF Needles & fitted plastic caps	1, 3.6, 11.2	64	2.7 & 5.1				Kinnell (2005)
Field-NZ Laechler nozzle (# 460.608)	3	12-25				1.0	Mathys et al. (2005)
Field-NZ Emani ¼ HH10SQ nozzle	3	90-150				1.0	Mathys et al. (2005)
Field-NZ Laechler nozzle	3.86	70				1.0	Ndiaye et al. (2005)

(# 461.008.30)							
Field-NZ 20 sector sprinklers @ 170 kPa	6	43	1-4.5			288	Designed by Summer et al. (1996); used by Castro et al. (2006)
Lab-DF			4.7	12.7		0.0625	Aoki and Sereno (2006)
Lab-NZ		100				1.1	Pan & Shangguan (2006)
Lab-NZ 4 axial cone-jet nozzles	4.5		1.2	652-2394 J/m^2-hr		0.25	Parsons & Stone (2006)
Field- NZ Nine nozzles	6	76			~80	4 x 8m	Designed by Panini et al. (1993); used by Rulli et al (2006)
Field-DF	2.5	80	2.5				Ramos & Martinez-Casanovas (2006)
Field- NZ	3	10	0.42	1.54	81	1.0	Schiotz et al. (2006)
Field-NZ Five Spray nozzles	4.9	20, 60, 250 & 420	1-2.8				Keim et al. (2006)
Field- NZ Four plate sprinklers	11.0	25-155	1.7-2.4	16.8-25.9	58-73	7 x 14 m	Munster et al. (2006)
Field-NZ		60, 70 & 120				1.0	Designed by Swanson. (1965); used by Bertol et al. (2007)
Field-NZ		60				1.0	Asseline & Valentin (1978); in Le Bissonnais et al (2007)
Field-NZ	1.57	95	2.4	2050 J/m^2-hr		1.0	Designed by Luk et al. (1986); used by Neaver & Rayburg (2007)
Field-NZ oscillating nozzles	3.7	5.1, 29.4 & 6.3		4 & 16		0.16 (circular)	Augeard et al. (2007)
Field-NZ oscillating nozzles	2.5 & 3.7	30-117.5	0.5-1.2	293-1914 J/m^2-hr		0.16 (circular)	Arnaez et al. (2007)
Field- DF	1.35	160 & 200	3.65 & 4.15	2200 J/m^2-hr	87.7 & 91.5	0.1	Clarke and Walsh (2007)

Field-NZ 20 full-cone nozzles	2.2	21	95% < 2	13.5		80	Marques et al (2007)
Field-NZ/DF	0.033–0.054	72	5.9	4		0.0625	Designed by Kamphorst (1987); Jordan & Martinez-Zabala (2008)
Field-NZ	3.5	56.5 & 90				0.23	Designed by Navas et al. (1990) & Lasanta et al. (2000). In Martínez-Zavala et al. (2008)
Field-NZ	3.5	56.5				0.13 (circular)	Designed by Navas et al. (1990) & Lasanta et al. (2000); used by Jordan et al. (2008)
Field-NZ Oscillating veejet 80100 nozzle	2.0	~100		29.5		1.5 x 2.0 m	Designed by Loch (2001) Sheridan et al. (2008)
Field-NZ Operated @ 45 kPa	2.5	20, 30 & 40				0.6	Pappas et al. (2008)
Field-NZ Veejet 80100 nozzles above rotating disks, operated @ 36 kPa	2.3	94-573	1.8 & 2.0	>90% of terminal	81-85	0.7	Sobrinho et al. (2008)
Field-NZ		69				40	Tatard et al. (2008)
Field–mod. DF 216 holes of 0.5 mm diameter	1.5	24.5 & 32	3.6			0.95 x 1.2 m	Vahabi & Nikkami (2008) Vahabi & Mahdian (2008)
Field-NZ Oscillating Veejet nozzles @ 41 kPa	2.5	70	1.05			0.3 (lab)	Designed by Foster et al. (1979), in Rimal & Lal (2009)
Field-NZ Micro-sprinklers	2.2	75		28.1		2.5	Singh & Khera (2009)
Field-NZ Oscillating Veejet 80100 nozzles @ 41-42 kPa	3.0	100				1.0	Folz et al. (2009)
Field-NZ Oscillating jet	3.5	60				1.0	Designed by Asseline & Valentin (1978); used by Blavet et al. (2009)

Field-NZ Oscillating flat fan Veejet 80150 nozzles	2.13	170-200	3.5	22.6	87	1.0	Designed by Meyer & Harmon (1979) as modified by Kato et al. (2009)
Field-NZ Four Fulljet ½ HH 40WSQ nozzles w/ solenoid valves @ 45 bar		47				1.2x3.9 m	Designed by Strauss et al. (2000); Armstrong & Quinton (2009)
Field-NZ TeeJet® TG SS 14W nozzles	1.8	85 & 170	4.5			2.7	Designed by Schiettecatte et al. (2005); in Jin et al. (2009)
Field-NZ 4 full-cone Unijet nozzles	1.8	119-124			~91	1.0 x 2.5 m	Sangüesa et al. (2010)
Field-NZ Fulljet 24WSQ & 50-WSQ nozzle @ 34.5 kPa	3.0	45 & 85	1.0 & 1.6		85-86	2 x 2 m	Dufault & Isard (2010)
Field-NZ Full-cone nozzle with solenoid valve (90-300 kPa)	2.0	21-83	0.5 - 2.8	15.1			Designed by Miller (1987),
Field-NZ Full-cone nozzle with solenoid valve (90-300 kPa)	1.0-1.4	20-80	0.5-2.8	15.1	80-92		Perez-Latorre et al. (2010)
Field-NZ Oscillating flat fan Veejet 80100 nozzles	2.2	10-130	2.2	27	~90	1 x 6m	Norton & Savabi (2010)
Field-NZ 1-3 180º plane-jet NZs @ 20º angle & 100 kPa	1.0-1.4	20 (1 nz) 59 (3 nz)	0.5 - 2	10.1	80-92		Perez-Latorre et al. (2010)
Lab-NZ	1.96	64.3 & 95.6					Designed by Morin et al. (1967); Sepaskhah & Shahabizad (2010)

Basic RS Designs –Drop sizes, Their Distribution & Intensity Uniformity

<u>Drop-former type RSs</u>

Generally, as a result of using a single aperture size, drop-former type RSs without underlying mesh screens generally generate a single drop size dependent on the pressure driving water through the aperture and the aperture size. Drop sizes of 1-6 mm have been developed, but most RS in use generate drops between 1.8 and 2.8 mm. Drop-formers that use yarn strings and plastic tubes operate at very low pressures and generally produce a narrow range of drop sizes whose drop KE depends on the drop-forming mechanism height above the soil plot. Agassi and Bradford (1999) contended that drop-former RSs that produce only one drop size are normally used in fundamental erosion studies and that such simulators should not be used to quantify interrill erosion components of wash and splash (Bradford and Huang, 1993). In addition, Bradford and Huang (1993) showed that erosion rates determined from a nozzle and single-drop-size type RSs at the same intensity can be quite different. While they argue that capillary-tube type RSs with a hanging screen provide a good alternative to the nozzle type simulators, they considered their usefulness limited to the laboratory. Field type RSs of this type, however, have been developed more recently (e.g. Clarke & Walsh, 2007). Early examples of smaller RSs used in the field include Munn (1974) who employed catheter tubing to form drops with a fall height of 2.5 m over a 0.61 m by 0.61 m plot area to evaluate runoff/erosion from seven soils in the Lake Tahoe basin. Painuli et al. (1985) describe a drop former assembly comprised of 20-gauge hypodermic needles capable of producing rainfall intensities of 29-113 mm/hr with Christiansen Uniformities (CU) of 95.9-99.8%. A common shortcoming of field-portable drop-former RS designs has been an inadequate fall height, resulting in low raindrop KE relative to that generated when drops reach terminal velocities presumably characteristic of natural rainfall (Guerrant et al., 1990; Robinson and Naghizadeh, 1992).

Modified drop-formers operating at greater intensities can develop uni-modal drop-size distributions as found by Clarke and Walsh (2007) and shown in Figure 8. Such modified drop-formers type RSs were also developed previously. For example, using a mesh screen placed some distance below the needles, breaks the uni-size drops into a range of smaller and larger drops (Poesen, 1984; Roth and Helming, 1992). The Roth and Helming (1992) RS consisted of 2500 capillaries 0.3 m suspended below which was a screen with a 3 mm wide opening resulting in drop sizes ranging from 0.5-5.0 mm and a median drop size of 2.89 mm that fell from 7 m above the test plot. Their RS produced rainfall with drop velocities approaching ~95% of terminal at intensities of 30 and 60 mm/hr. The drop-former RS uniformity of drops across the designated plot area depends on the relative areal density of drop-formers (e.g. number of needles/m^2), their functional state at the time of measurement (e.g. salt, or sediment clogging) and relative exposure to air currents below the drop former. Measured CUs for drop-former type RSs are generally high, often >90%, and are improved by greater areal density of drop-formers. For example, Figure 9 illustrates the relative rain intensity (ratio of local intensity in sub-plot section to average across plot) distributions across a 1 m^2 plot from the drop-former (needles) type RS developed by Battany (1998). Clarke and Walsh reported similar results with CUs of 87.7 and 91.5% at much greater intensities of 160 and 200 mm/hr and median drop sizes of 3.65 and 4.15 mm, respectively, from a drop-former type RS used in the tropics.

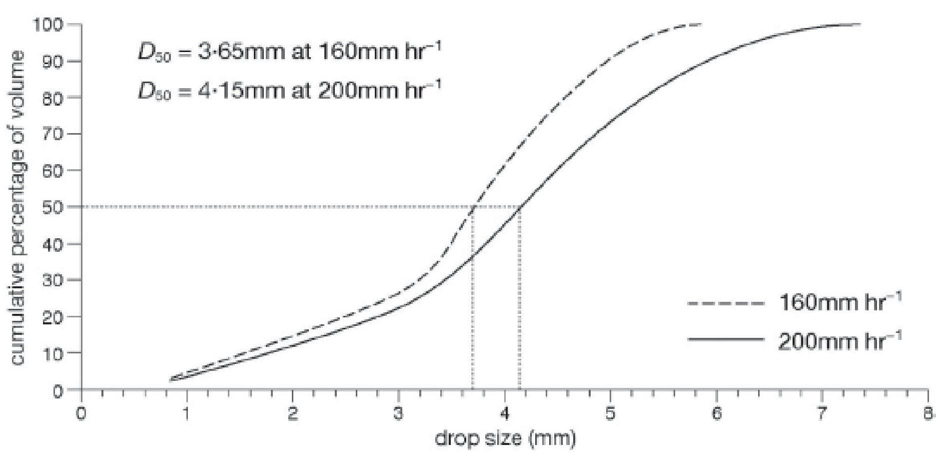

Figure 9. Cumulative drop-size distributions from a modified drop-former RS operating at relatively high rainfall intensities (Clarke & Walsh, 2007).

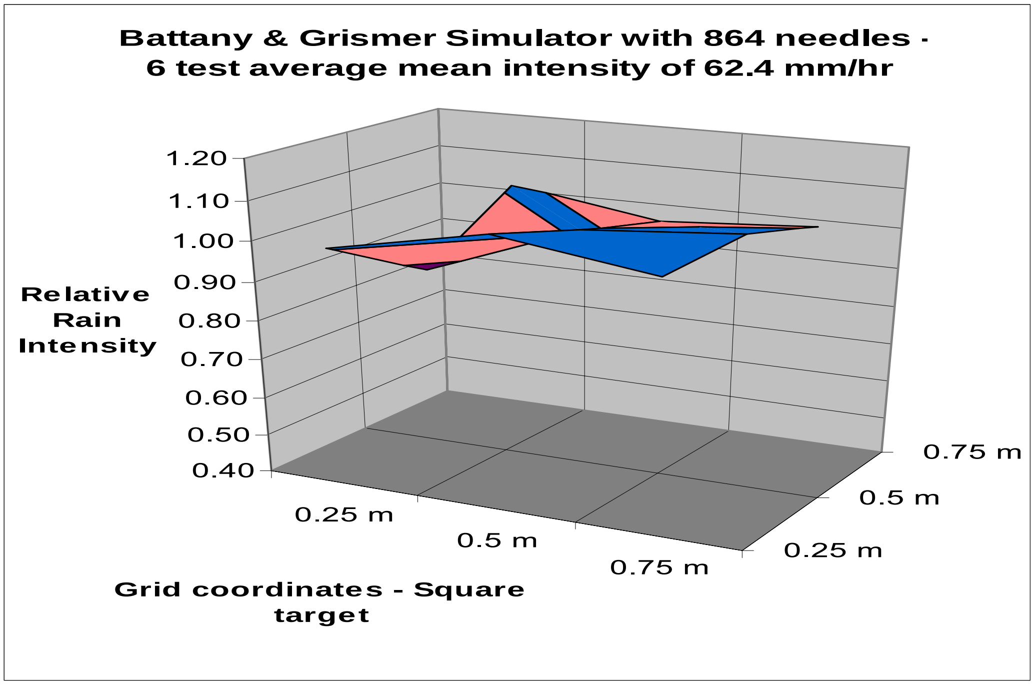

Figure 9. Relative rain intensity distribution surface across 1 m^2 plot from drop-former type RS developed by Battany (1998).

Spray-Nozzle Type RSs

Like drop-former type RSs, simulated rain mean drop size, distributions and intensities for nozzle-type RSs depend on type of nozzle(s) used, applied pressures and how they are arranged or moved albeit in a more complicated fashion. Generally operating at higher pressures than drop-former types, nozzle type RSs develop a wide range of drop sizes, possibly imparting substantial initial velocities to the smaller drops (i.e. in excess of 10 m/s as compared to terminal velocities of natural rainfall between 6-9 m/s) and at initial angles of flight far from vertical. Most nozzle-type RSs operate at pressures ranging from 34-140 kPa; where higher pressures generally develop good drop-size distributions but potentially excessive intensities, and lower pressures give very poor drop-size distribution (drops are too large) and distribution uniformity. Water pressure also affects the area covered by the rainfall: low pressure reduces the application area, high pressure increases it, but at a lower application rate per unit area. A pressure gauge is used to monitor pressure throughout an experiment. Some consider these RSs sensitive pieces of equipment, and their reliability in the field is often affected by their sensitivity to frost and poor handling.

Drop-size distribution across the fan width of spray follows a bell-like distribution with larger size drops more centrally located while smaller drops comprise the fan edges. Agassi and Bradford (1999) underscored that drop velocity for a fan-type veejet nozzle favored by many researchers varies from a maximum vertically above the target area and decreasing toward the target area edges. This velocity differential is reduced by increasing the height of the nozzle above the target area and by decreasing the travel angle (Meyer and Harmon, 1979). Stationary fulljet spray nozzles tend to produce spatially conic drop distributions (Hall, 1970). For example, Cerda et al. (1997) in a thorough evaluation of a portable RS (1 m^2 rain area) evaluated the effects of pressure on intensity, drop size, drop-size distributions and Christiansen Uniformity (CU). Figure 10 illustrates the effects of operating pressure on simulated rainfall intensity and CU, while Figures 11 and 12 illustrate the effects of pressure on mean drop size and KE distribution. For the Cerda et al. (1997) RS, as pressure increased, rainfall intensity, mean drop size and KE decreased (due to smaller drop sizes and intensity) while CU increased to a maximum at ~ 55 mm/hr and then decreased. Similarly, Figures 13 and 14 illustrate relative rain intensity distributions across 1 m^2 plots from single nozzle type RSs as tested by Kinnel (1993). Single-nozzle type RSs tend to generate less evenly distributed intensities as compared to multi-nozzle systems such as that developed by Loch (2001). For example, Dufault and Isard (2010) used two different single-nozzle type RSs and obtained CUs of 85-86% at intensities of 45 and 84 mm/hr (see Figure 15). Multi-nozzle RSs tend to develop localized zones of higher relative rainfall rates associated with overlapping spray patterns, though CU values have improved from ~70% to the mid-80% values. Examples of such rainfall distributions from field type multi-nozzle RSs developed more recently are shown in Figures 16, 17, 18 and 19 for average intensities ranging from 80 - 120 mm/hr.

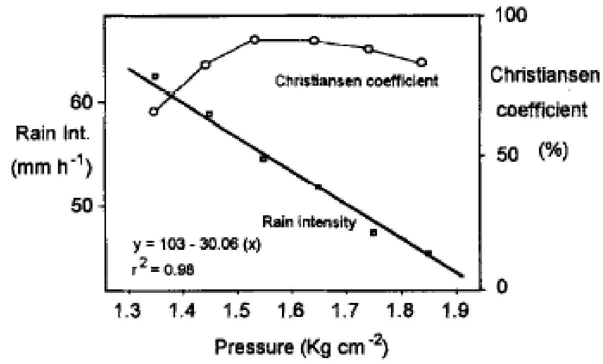

Figure 10. Dependence of simulated rainfall intensity and distribution uniformity across 1 m^2 plot on nozzle pressure (Cerda et al., 1997).

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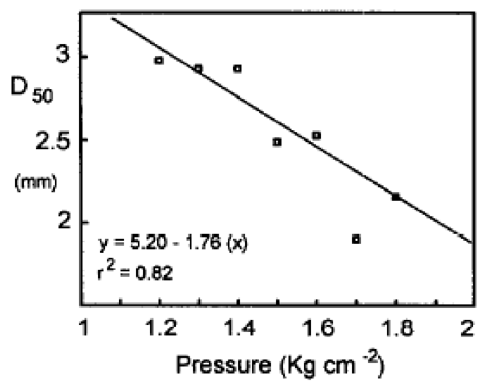


Figure 11. Dependence of simulated rainfall mean drop size on nozzle pressure (Cerda et al., 1997).

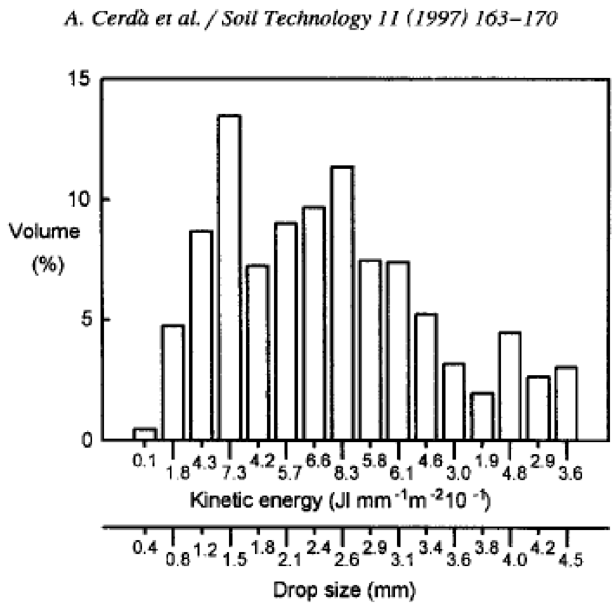


Figure 12. Dependence of simulated rainfall KE on drop size at a 54 mm/hr intensity (Cerda et al., 1997).

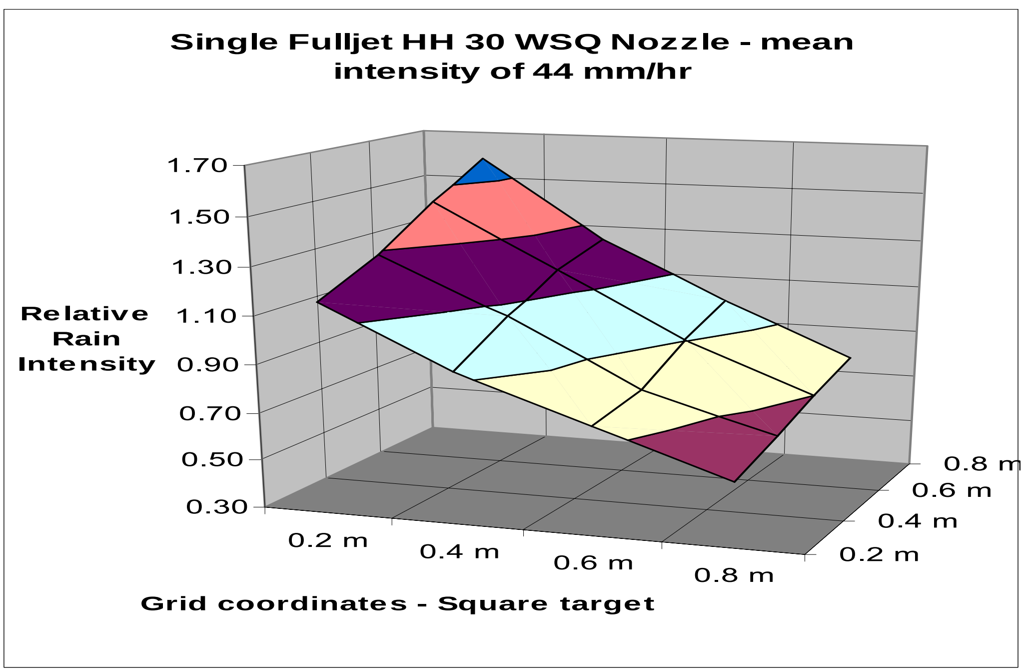


Figure 13. Relative rain intensity distribution surface across 1 m^2 plot from a single nozzle type RS as tested by Kinnel (1993).

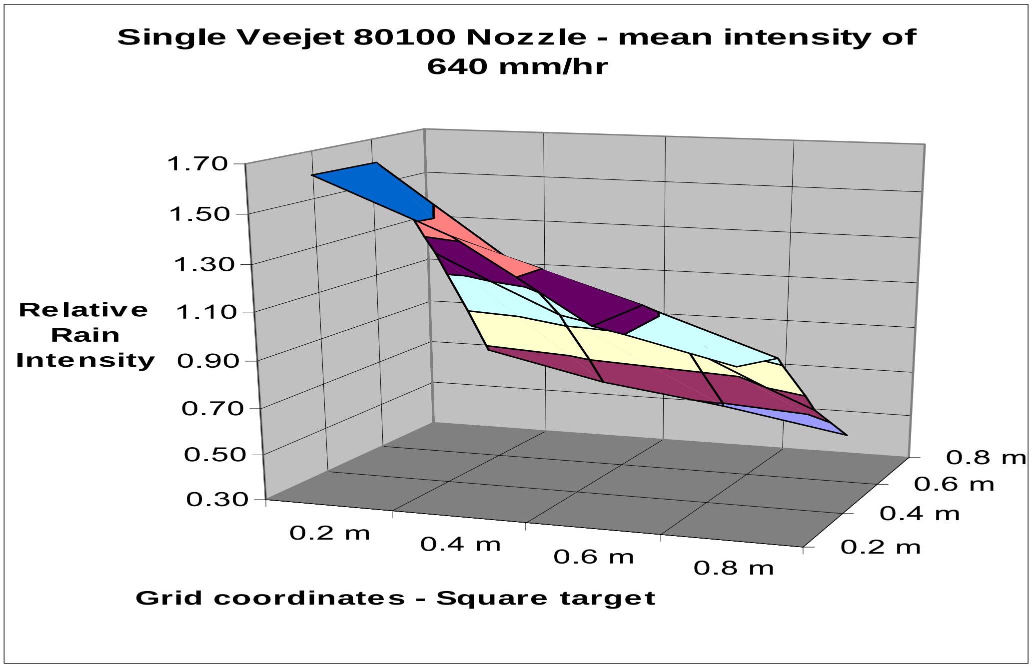


Figure 14. Relative rain intensity distribution surface across 1 m^2 plot from a single nozzle type RS as tested by Kinnel (1993).

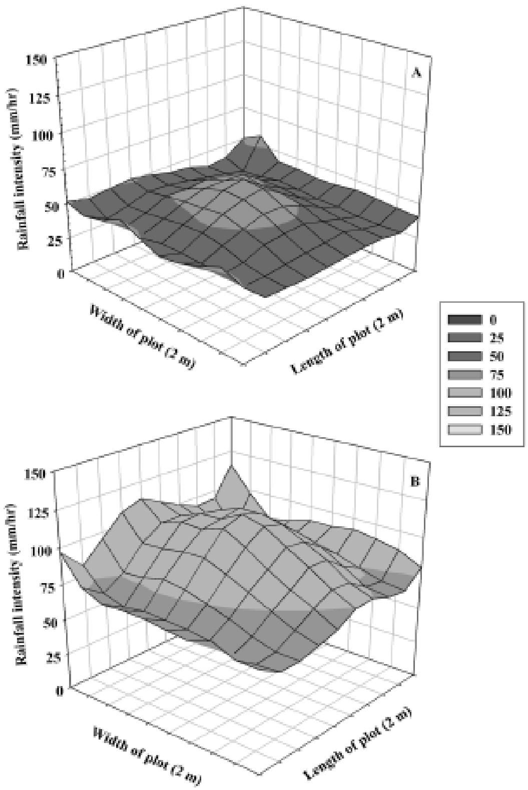


Figure 3. The rainfall intensity distribution of simulated rainfall collected over the 2- × 2-m sample area (5-min duration) for the (A) Fulljet 3/8HH-SS24WSQ nozzle (CU = 86%) and (B) Fulljet 1/2HH-SS50WSQ nozzle (CU = 85%) at 34.5-kPa operating pressure.

Figure 15. Relative rain intensity distribution surface across 4 m^2 plot from a single nozzle type RSs as tested by Dufault and Isard (2010).

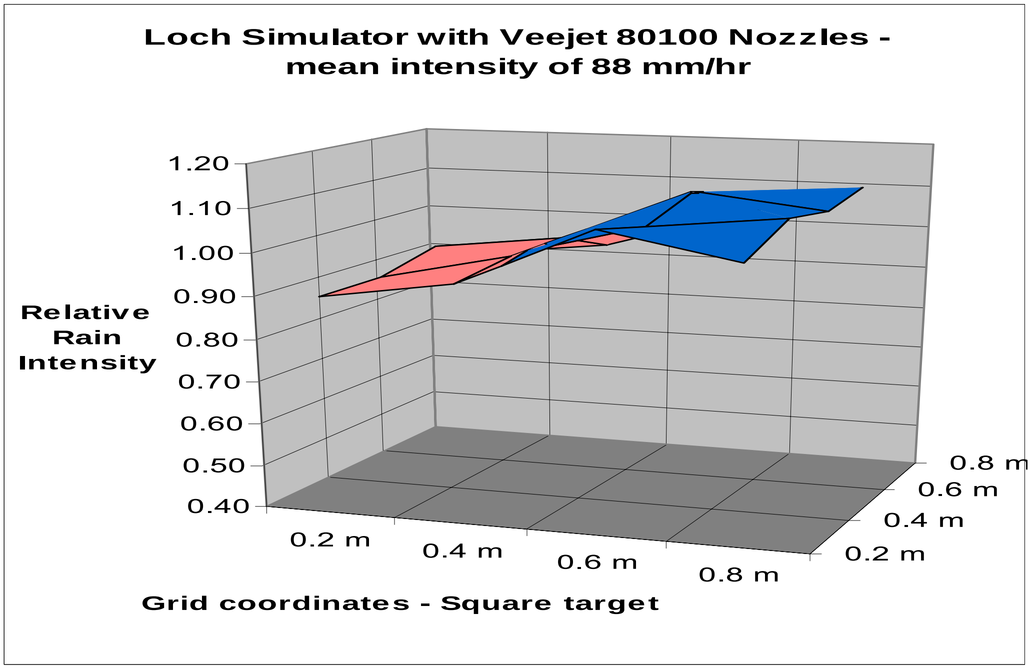


Figure 16. Relative rain intensity distribution surface across 1 m^2 plot from a Loch multiple nozzle type RS as tested by Kinnel (1993).

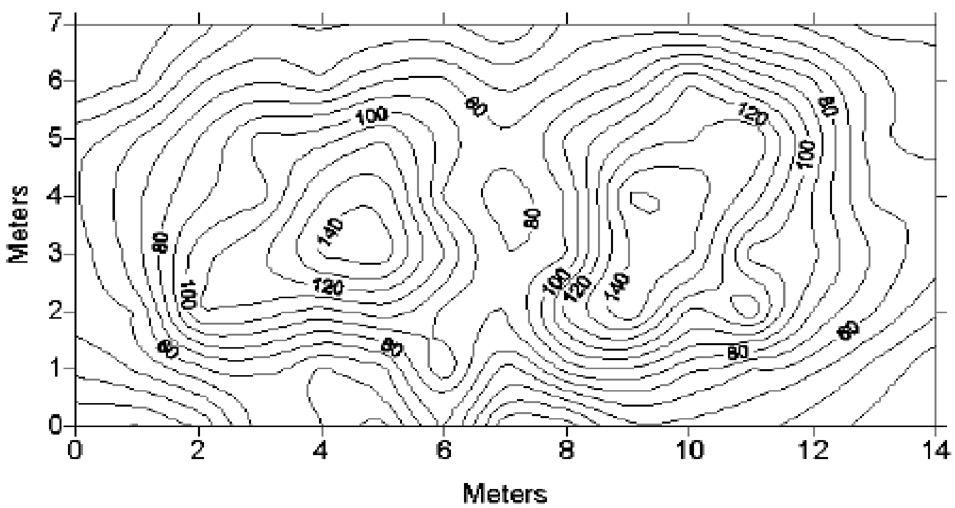
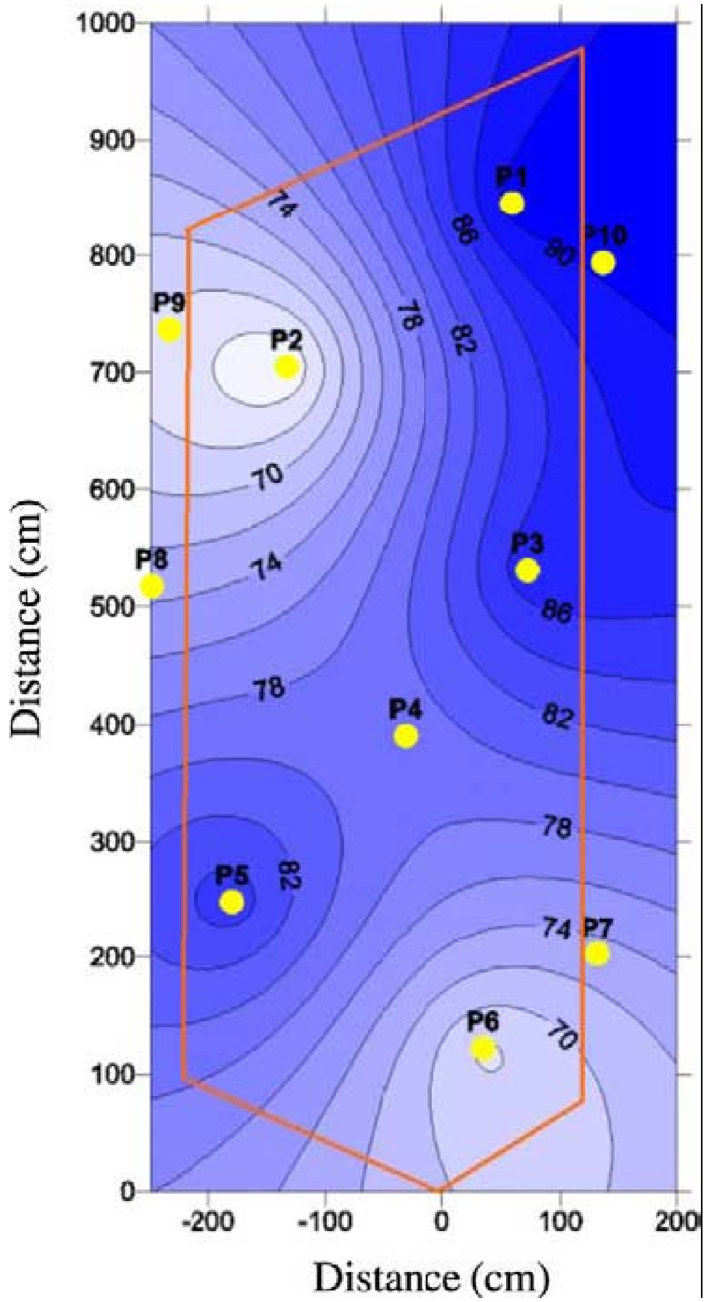


Figure 17. Contour map of simulated rainfall intensity (target intensity of 127 mm/h) taken from the mean of three replicated simulations (Munster et al., 2006).

Figure 18. Spatial distribution of rainfall intensity (mm/hr) across a 2 x 7 m plot at an average intensity of 80 mm/hr (P = pluviometer locations (Rulli et al., 2006).



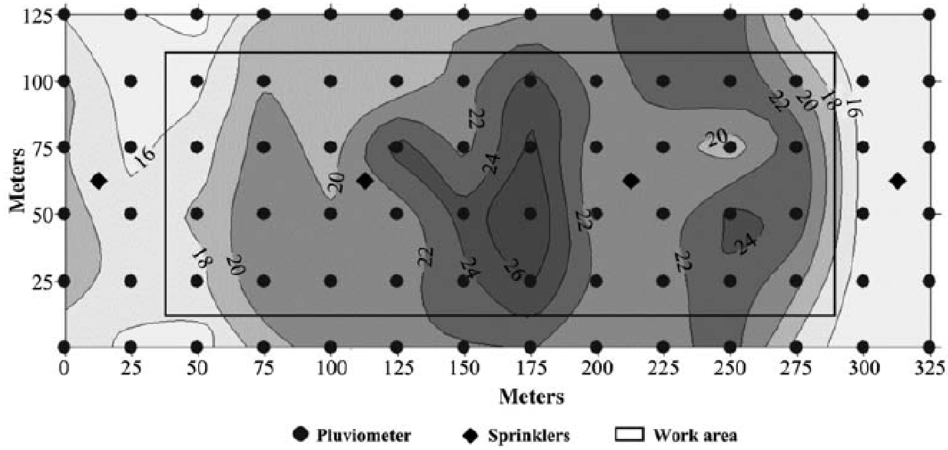


Figure 19. Spatial distribution of simulated rainfall depths (mm) after 10 minutes from four sprinklers (at 100 kPa pressure) across an approximately 0.75 x 2.75 m plot (Sanguesa et al., 2010).

Nozzle-type RSs that use rotating or oscillating spray nozzles have an unavoidable rainfall intensity periodicity (Kinnel, 1990 & 1993) over the plot surface (i.e. rain surges, followed by a period of repose) such that rain intensities and uniformities not only depend on nozzle water pressure, but also on fan sweep oscillation frequency (Paige et al., 2003). Such rain "surges" can result in localized instantaneous intensities as high as 2000 mm/hr as compared to averaged intensities for the plot on the order of 100 mm/hr. Paige et al. (2003) found that veejet nozzles working from a drop height of 2.44 m and at a nozzle operating pressure of 41 kPa results in a median drop size of 2.985 mm, while increasing that pressure to 55 kPa increasing the breadth of the drop-size distribution to a range of 0.29 – 7.2 mm while decreasing the median drop size slightly to 2.857 mm. Increasing nozzle oscillation frequency increases the rainfall intensity and CU, both of which are determined in part by the test plot size considered. For the Paige et al. (2003) RS, at the 55 kPa nozzle pressure to apply a 50 mm/hr rainfall intensity across a 2 m wide plot 1.5 long, the cycle frequency is 15.2% or about 9.1 sec per min of application indicating that the instantaneous application rate is approximately 330 mm/hr at any given location. At a greater average rainfall intensity of 127 mm/hr, the spray time fraction is much greater, about 37.9%, but the instantaneous rate remains about at 335 mm/hr. Of course, a longer plot length requires a greater "sweep" time that results in possibly unacceptable "periods of repose" thereby leading researchers to deploy additional nozzles to sweep each additional 1-2 m lengths. For example, Paige et al. (2003) deployed three nozzles and the maximum rainfall intensity of 175 mm/hr to develop a CU of 91.7% (see Figure 20) with greater rainfall intensities occurring along one edge of the test plot area. Becher (1994) reported that when used in erosion studies, such RSs result in less erosion as compared to that from non-periodic rainfall application, though Kinnel (1993) found otherwise comparing continuous spray versus oscillating systems.

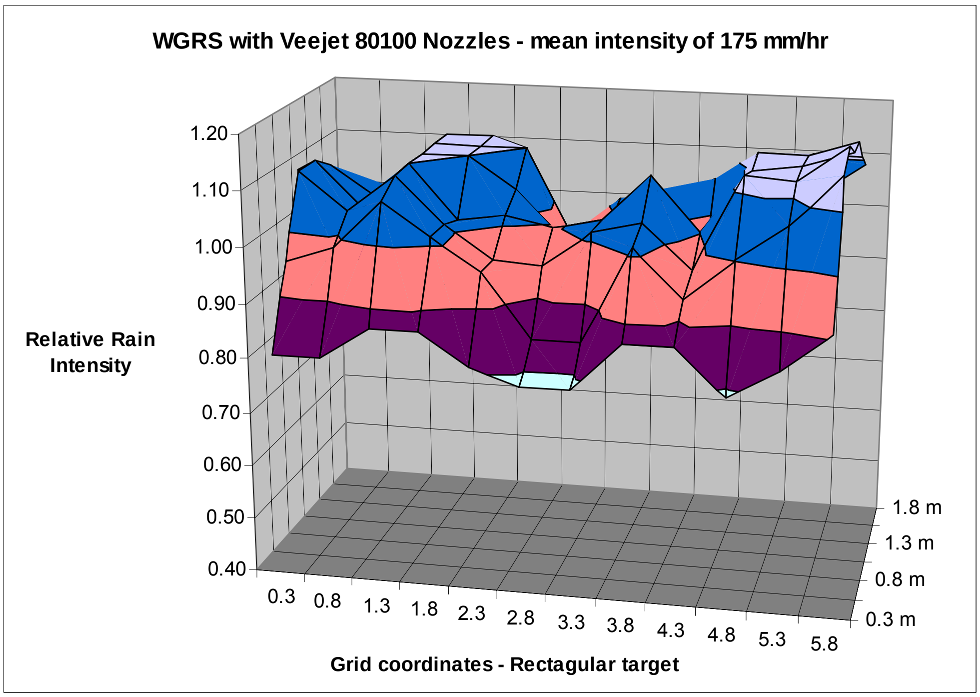


Figure 20. Relative rain intensity distribution surface across 2 by 6 m plot from a multiple-nozzle type Loch RS as tested by Paige et al. (2003).

Basic RS Designs –Drop-size Distribution & Rainfall Intensity Effects on KEs

Drop-size distributions developed in simulated and natural rainfall are important towards estimating storm relative KEs or power (KE/unit time). For example, Clarke and Walsh (2007) found with their drop-former RS that the <1 mm drops, though more abundant (61% of droplets) generated only ~1% of the total storm KE because they represent a much smaller mass whereas 1–5 mm diameter drops (38% of the storm mass) are responsible for most of the KE (75%) due to their magnitude and comparative frequency. Though simulated raindrops >5 mm diameter were rare (1% of storm mass) they contributed 24% of the total KE because of their large mass. Ideally, therefore, storm KE should be calculated by integrating across the drop-size distribution. More often, nozzle-type RSs develop a range of drop-size distributions that depend on nozzle type and applied pressures (rainfall intensities) and measurement method. Marques et al., (2007) noting the range of reported KEs, questions whether these values are method determination dependent and perhaps should be independently measured for each RS experiment. For example, Kincaid et al. (1996) measured drop-size distributions by three different methods for a variety of sprinklers (Figure 21) and found that the dominant drop size as determined by the stain method, while similar to that from the other methods, represented 35% by volume of the drops as compared to ~22% determined by the other methods. Nozzle-generated distributions tend to be somewhat bi-modal, a characteristic not readily apparent in the natural rainfall drop-size distributions such as those illustrated in Figure 2 previously. For example, Kinnel (1993) tested two different nozzles used in RS whose quasi-bimodal drop-size distributions are shown in Figures 22. Erpul et al. (1998) found that drop-size distributions within wind-tunnel experiments also depended on the number of nozzles and wind speeds as illustrated in Figures 23 and 24, respectively. Applied cross-winds tended to shift the drop-size distributions towards the larger drop sizes while also limiting effects of drop "drilling" of the soil surface.

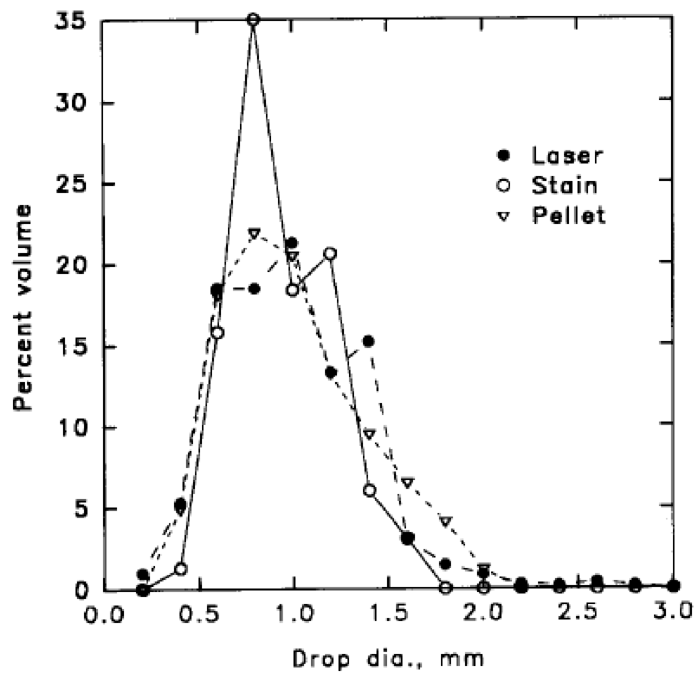


Figure 21. Example drop–size distributions as measured by different methods for a smooth-plate 4.7 mm fan-head sprayer operated at 206 kPa (Kincaid et al., 1996)

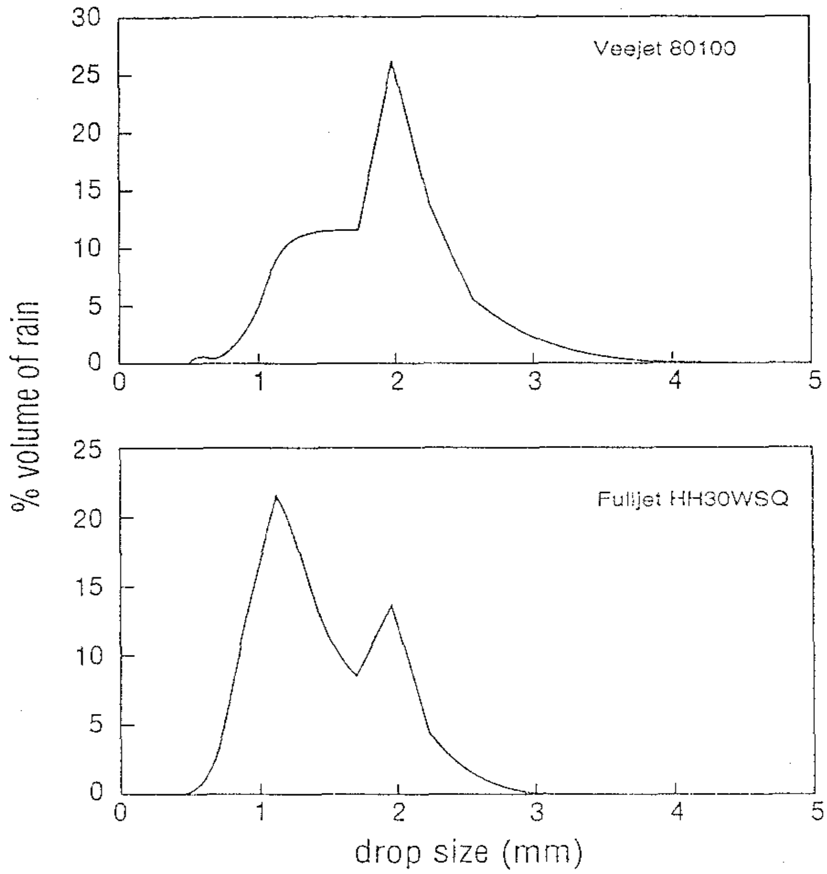


Figure 22. Rain drop-size distributions from two commonly used nozzles in RSs, the Veejet 80100 and the Fulljet HH30 WSQ operating at pressures of 50 and 30 kPa, respectively, as tested by Kinnel (1993).

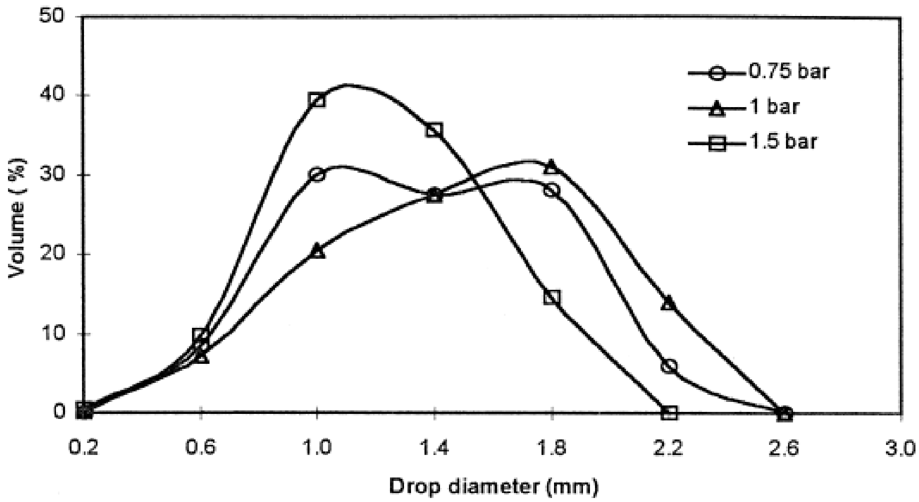


Figure 23. Drop-size distributions without wind from a five-nozzle RS operating at different pressures (143 mm/hr intensity) from Erpul et al. (1998).

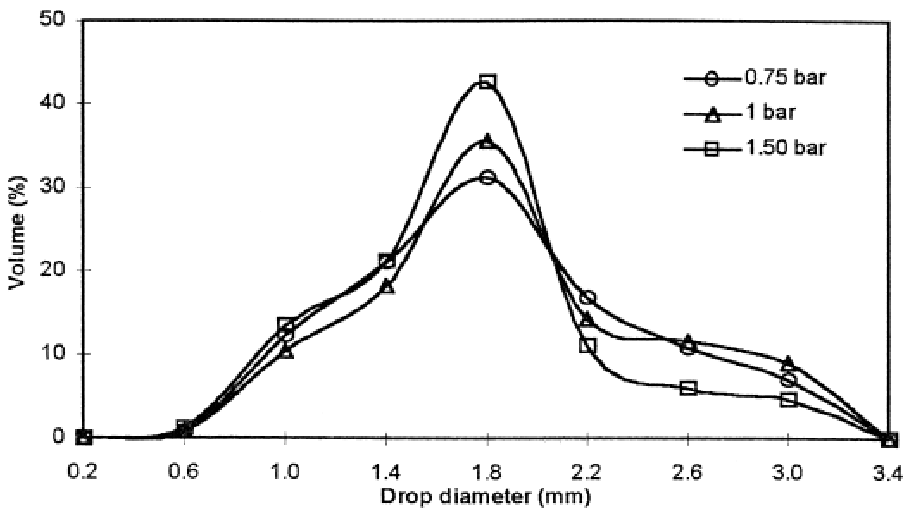


Figure 24. Drop-size distributions with 9.96 m/s cross-wind from a five-nozzle RS operating at different pressures (143 mm/hr intensity) from Erpul et al. (1998).

While changing rainfall intensity from nozzle-type RSs does not necessarily affect initial drop velocities, there are some changes in the corresponding drop-size distribution. The RS described by Poesen et al. (1990) uses a continuous spray Lechler type 460.788 full-cone nozzle that at an intensity of about 35 mm/h, developed drop-size distributions "similar" to that for natural rainfall, but the storm energy for the simulated rainfall was much less. The Guelph RS described by Tossell et al. (1987, 1990a) uses several low- to medium-flow rate, fulljet nozzles in continuous spray. For both nozzle types the rainfall intensity and drop velocity varies from the center to the edges of the impact area. For a target area of one square meter, some technicians located nozzles above each of the four corners of the plot (Bradford and Huang, 1993), while others positioned a single nozzle above the center of the plot. Because of the different impact angles resulting from the overlapping nozzles, KE for the two systems will differ. Prior to conducting interrill erosion experiments, variability in rainfall intensity and possibly KE across the test area should be evaluated.

Few early studies directly evaluated the effects of raindrop KE on relative rates of erosion rather; most studies of this type considered the effects of splash impacts or rainfall intensity (as perhaps something of a surrogate measure for KE) on erosion from bare, re-packed soils on mild slopes in a laboratory environment. For example, considering loess soils subject to crusting, Mermut et al. (1997) found that for clay loam soil repacked into 0.3 m diameter columns relative soil losses were 10 times greater when increasing the rainfall intensity from 40 to 100 mm/hr, though reportedly at the same KE of 27 J/m^2-mm. They attributed the difference to rain splash effects. At very high simulated intensities of 200 mm/hr and direct measurement of splash detachment, Clarke and Walsh (2007) found that splash detachment was independent of slope angle up to 89%, but downslope movement of splash-detached particles was significantly slope dependent between <22% and ~78% slopes where splash erosion from midrange slopes of 22-67% were not distinguishable. Also considering raindrop splash effects directly, Kim and Miller (1995) conducted single and multi-drop splash/detachment tests of 4.1 mm drops falling from 7.0 m on five repacked sandy loam to clay loam agricultural soils in 0.76 m diameter containers. The average weight of splashed soil particles after 75 drops did not show any significant difference between the five soils. Using a nozzle-type

RS at 74.9 mm/hr intensity for 85 min and the same soils, total soil splash losses ranged from 5000-6000 g/m^2 for the finer-textured soils and 3000-4000 g/m^2 for the coarser-textured soils. There were no obvious relationships between soil losses measured from the different experiments (single drop and multiple drop splash tests). Sukhanovskii and Sanzharov (1995) conducted similar experiments using a sprinkler type RS and attempted to develop criteria to evaluate the effect of droplet falling velocity on soil detachment. Legout at al. (2005) found that stronger aggregated silty clay and clay loam soils yield smaller splash dispersal distances from impact as compared to low-strength sandy soils. Splash impacts enriched the relative mass fractions of 250-1000 μm particles on the surface.

As noted previously, many nozzle-type RSs employed in soil erosion studies use oscillating or sweeping nozzles that rely on rain "pulsing" frequency to control the rainfall intensity. Considering only rain intensity effects in a lab study, Kinnel (1993) used 0.2 mm sand repacked into 0.5 x 0.5 m square pans to examine the effects of pulsed versus continuous rainfall at a wide range of intensities. He found, as he had predicted from earlier studies, that sediment losses were strongly dependent on runoff depths between 4-8 mm and type of RS nozzle arrangement. Throughout his experiments (1993 and 2005), he maintained a steady surface sheetflow in addition to that rain-induced so as simulate overland flow while better controlling flow depths. His intended study of the effects of simulated rain "pulsing" in 1993 was inconclusive; in contrast to later findings by Armstrong and Quinton (2009). Armstrong & Quinton (2009) examined the effect of simulated rain pulsing on runoff sediment concentration and size using three different pulse cycles operating at an average intensity of 47 mm/hr (0·45 bar to each of four Fulljet ½ HH 40WSQ nozzles). There was considerable variation in sediment concentration and particle-size distribution through the pulse cycle. The greatest concentration was as much as four times that of the lowest concentration; in addition, the peak median particle size was double the lowest median particle size. The magnitude of differences in sediment concentration and particle size were greater the longer the pulse cycle and these dynamics are likely to vary between RSs. Overall, they suggested the impact of the pulsing on sediment is significant and that high-intensity "pulses" associated with oscillating nozzles have significant effects on measured erosion rates and that pulse periodicity should be as small as possible to eliminate these effects.

Considering field erosion under no tillage, reduced tillage and conventional tillage silt loam plots using a single-nozzle Guelph RS (1 m^2 plots), Nolan et al. (1997) found that total soil loss from 20 minute duration storms at 60 and 140 mm/hr increased from 20 to ~900 kg/ha for the reduced to conventional tillage conditions. Coincidentally perhaps, the soil loss rates from the high intensity RS events matched that measured under natural rainfall conditions without corrections for slope, slope length, and simulated rainfall energies. From the perspective of variability in erosion rates associated with consideration of rainfall intensity effects, Lui et al. (1998) evaluated the soil losses for sand and sand/clay mixtures (repacked in 0.81 x 0.81 m square boxes) at drop-former RS intensities ranging of 12.7, 25.4, 50.8 76.2 and 101.7 mm/hr at very flat slopes of 0.1, 0.5, 1 and 4 %. They found no slope effect on unit sediment loss and a weak relationship between sediment loss and rainfall intensity or net water power (raindrop impact plus surface flow) as illustrated in Figures 25 and 26 (for all three "soil" mixtures). There was little improvement in predictive capability of the linear regressions through inclusion of rainfall intensity, KE and runoff depth effects in the determination of water power between the results summarized in these two figures. However, rather than plotting all of the soil-slope combinations together, Figures 27 and 28 consider the effects of rainfall intensity on erosion from two of the different soils (clay-sand mix and roadcut soil) as segregated by slope groups. An obvious dependence of erosion rate on rainfall intensity is apparent for the re-packed soils, however, the range or variability in erosion rates also appears to increase with increasing rainfall intensity. Variability in erosion rates from the approximately 51 and 101 mm/hr intensities yielded CoVs of ~20% at both intensities for the clay-sand mixture, and 21% and 33% for the roadcut soil, respectively. Perhaps more important is to note the range of sediment loss values in at each rainfall intensity, or power to appreciate something of the variability associated with these type of measurements, even in the laboratory on very flat slopes.

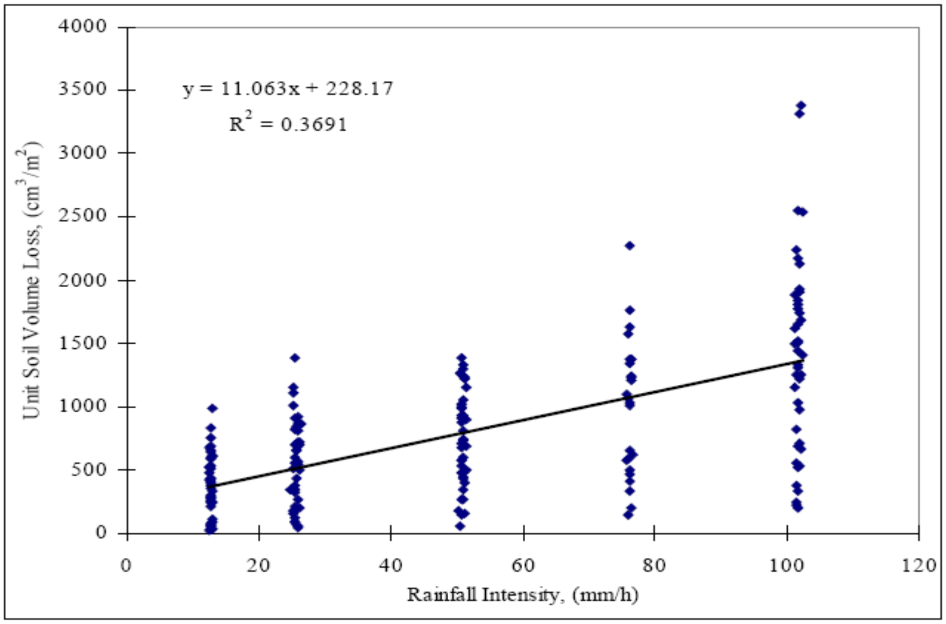


Figure 25. Dependence of soil loss on rainfall intensity for sands, sand-clay mixtures and roadcut soils (from Lui et al., 1998).

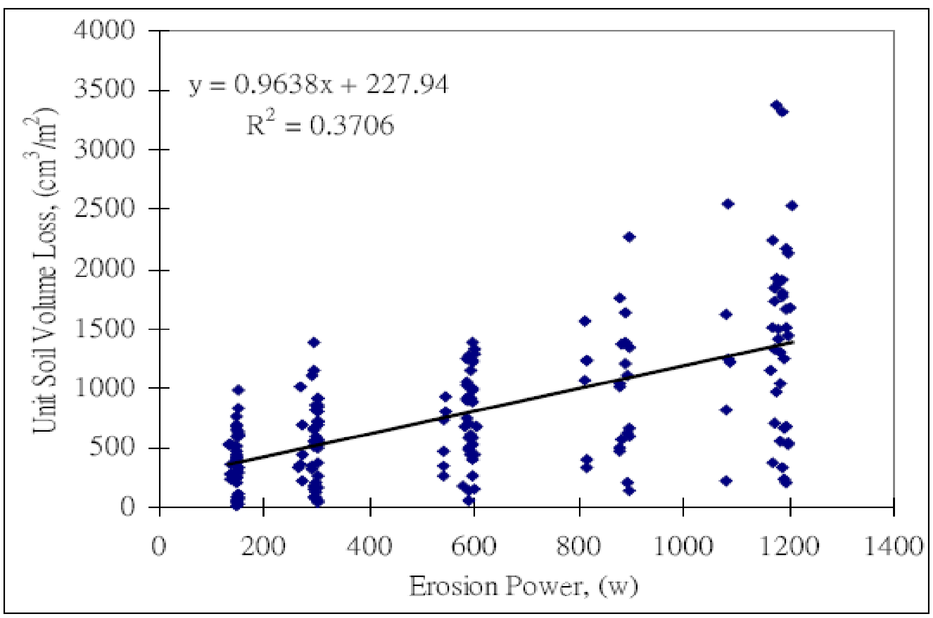


Figure 26. Dependence of soil loss on water erosion power for sands, sand-clay mixtures and roadcut soils (from Lui et al., 1998).

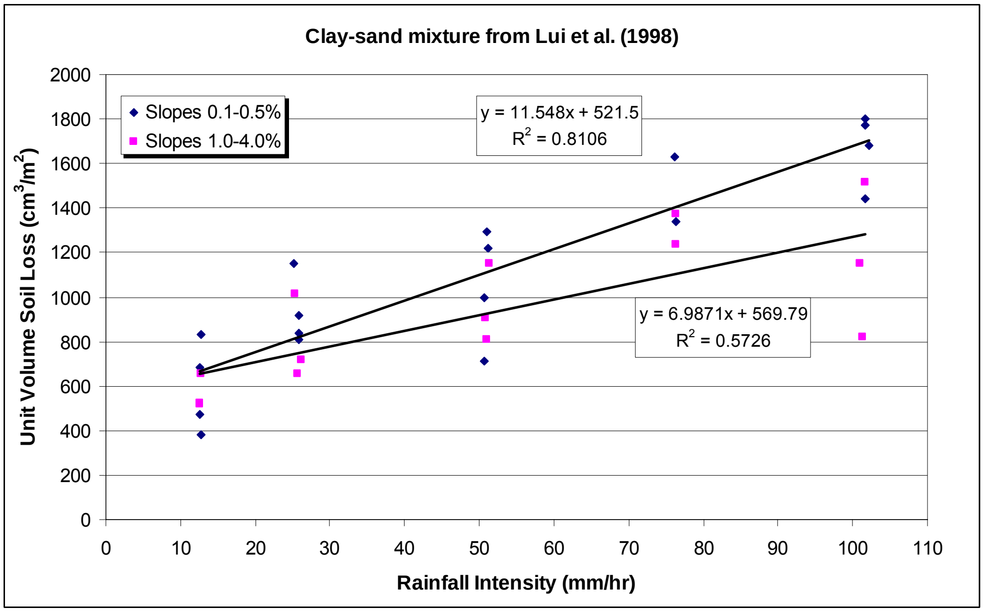


Figure 27. Dependence of soil loss on water erosion power for the sand-clay mixture.

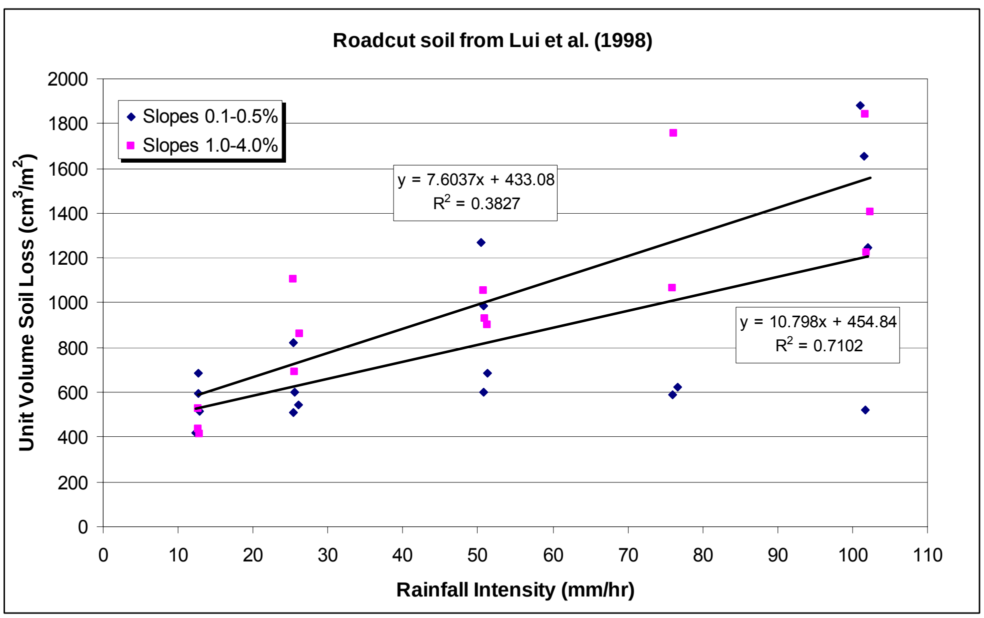


Figure 28. Dependence of soil loss on water erosion power for the roadcut soil.

Kinnell (2005) attempted to attack the KE – erosion rate question directly using two drop-former type RSs generating average drop sizes of 2.7 and 5.1 mm from fall heights of 1.0, 3.6 and 11.2 m to generate erosion of the same 0.2 mm repacked sand used previously at flow depths of 3-14 mm. Sediment discharge rates were linearly related to rainfall power at each flow depth considered such that for the 2.7 mm raindrop size and flow depth of 3 mm, average sediment discharge increased by 3.2 times and 5.5 times when increasing the fall height from 1.0 to 3.0 m and 1.0 to 11.2 m, respectively. The relative dependence (or line slopes) of 0.2 mm sediment discharge on flow depth also increased with increased drop fall height with the effects of fall height diminishing with increasing flow depth for the 2.7 mm drop size. Though similar relationships were obtained in some respects for the 5.1 mm drop size, the relationship between sediment discharge and rainfall power were different such that discharge rates leveled at higher powers rather than linearly increasing as with the 2.7 mm drop size. Effects of slope were either not considered, or had no appreciable effect in these studies of rainfall intensities or energies and erosivity; however, those reported were generally very mild slopes of 1-5 %. Ries et al. (2009) contends that despite numerous studies on drop-size characterization of simulated rainfall, there as yet remains no established technique for its measurement, or a single parameter that can express the drop-size distribution accurately with respect to it impacts on erosion rates. They consider use of volumetric average or median drop diameters as "not specific enough for detailed comparisons of different RSs." It is likely that more fruitful comparative approaches will involve determinations of the net storm energy rates or power for each RS in its application.

Finally, Agassi and Bradford (1999) raised several other concerns that apply to RS studies of erosion processes. They note that the effects of drop impact angle on infiltration and erosion rates has not yet been quantified; possibly an important issue both for nozzle-type RSs, but also in general for RS erosion studies conducted on steep slopes. With respect to nozzle-type RSs, they note that the raindrop energy is constant irrespective of the intensity (Hignett et al., 1995) and that drop size is also constant, rather than the maximum drop size increasing with intensity as under natural rainstorms. At equivalent intensities, runoff and soil loss is possibly greater for oscillating nozzle type RSa using a high delay time between sweeps as compared with RSs with low delay times, particularly for those soils highly susceptible to surface sealing. For equal rainfall intensities, kinetic energy per unit time of drop impact for the intermittent spray nozzles is greater than that for the continuous spray nozzles. Comparisons of the infiltration, runoff and erosion rates between RSs generating multiple drop and single drop sizes though the same KE are lacking; though these factors may be practically insignificant.

Field RS Methodologies – Effects of measurement methods and plot conditions

As the area of simulated rainfall coverage is limited in extent by the RS, slope, available water and possibility of replication, small field plot RS-erosion studies are necessarily compromised by sampling issues relative to the larger landscape whose infiltration, runoff and erosion conditions are to be determined from the study. Variations in methodologies and possible sources of uncertainty relative to comparison of results between studies can be broadly grouped into those associated with;

(a) RS water supply,
(b) simulated rainfall characteristics (e.g. D_{50} drop size, intensity and KE),
(c) plot runoff frame size and installation,
(d) runoff sampling size, frequency and duration,
(e) identification (determination) of plot cover, slope and surface soil conditions,
(f) measurement of interrill or rill erosion,
(g) plot replication, or degree to which plots represent hillslope conditions, and
(h) interpretation of runoff sediment sampling information relative to the local soil, cover and climate conditions.

Each variation or source of uncertainty is considered below in terms of small plot RS studies in forested catchments.

A. Several researchers have noted that the simulated rainwater chemistry may be an important factor to consider in RS experiments (e.g. Levee et al, 1979; Agassi et al. 1981; and Keren & Singer, 1988) as electrolyte and cation (e.g. SAR) concentrations can dramatically affect infiltration rates in some soils. Water with a high electrical conductivity tends to flocculate soil particles, whereas with the low electrical conductivity more typical of natural rain, the same particles may be dispersed and readily eroded. Kim and Miller (1996) concluded that the presence of salts in water used for RS studies may cause serious errors where the intent is to simulate rainwater of low electrical conductivity. The source and chemistry of the water used in the RS study should be reported for possible later comparisons. The volume of water available during the field trials is also an important consideration in field RS studies and is limited by the ability to transport large quantities of water to remote areas, however, the available supply constrains the durations of the simulations.

B. As discussed in the previous section, the range of simulated rainfall intensities and energies used in various erosion related studies has varied as much as that from natural rainfall. There is no single standard intensity or KE that has been identified as applicable to inter-rill and rill erosion studies. As Dunkerly (2008) noted, nearly all RS studies employ relatively large intensities that are typical of more extreme natural events. Each RS-erosion study employs a different intensity as needed so as to exceed the plot infiltration rate such that runoff and erosion occur. Simulated rainfall KEs are typically less than half that of "natural" rainfall as determined by the simulated median drop size and the associated terminal velocity calculated for that drop size. Directly measured natural rainfall powers have a similar span to that simulated, but at typically smaller intensities to generate that same power. Ries et al. (2009) opines that "despite the numerous studies on drop size characterization of simulated rainfall, there is still no established technique for its measurement or data unit to express the drop size distribution accurately." Without accurate characterization of the simulated rainfall, they are concerned parameters such as the median drop size "are not specific enough for detailed comparisons of different RSs." They recommend use of the "Laser Disdrometer as the best measurement method for rainfall characteristics." Given the variability in infiltration, runoff and erosion rates results as will be discussed below, this issue is probably a minor concern with respect to field simulations on small plots. While it is generally understood that low intensity, potentially long duration storms may result in little or no erosion, there is scant information available about what threshold rainfall intensity or power is required to "trigger" an erosion event for a particular set of conditions at any given locale (perhaps with the exception of definition of I_{30} by Wischmeier & Smith, 1978). Nonetheless, RS studies in the past decade have better reported the simulated rainfall characteristics as compared to earlier studies; most contain at a minimum the basic information about the median drop size(s) intensities and associated KEs used in the erosion evaluation.

C. Typically, metal frames are installed to delineate the plot runoff area as a smaller centrally located portion of the simulated rainfall area. By design, for reasons of portability, water use, replication potential and possibly cost; runoff collection "frames" are on the order of ~1 m^2 in many studies (see Table AA). Clearly, the size of the runoff frame should be less than that of the rainfall area so as to have "buffer zones" for rain splash inside and outside the frame and allow for possible wind drift of the simulated rain. Smaller frame enclosed areas of <0.3 m^2 can yield greatly different results from those of 1-2 m^2 or larger (Wang, 1988; Loch & Faley, 1992; Bradford & Huang, 1993). In addition, the length:width ratio of the frame can be important and ratios of ~1 have been suggested, or that the frame width is at least ~1 m (Agassi & Bradford, 1999). Using the nozzle-type (Veejet 80100) RSs, Auerswald et al. (1992) studied the effect of plot size on erosion dynamics in the mildly–sloped agricultural fields and found that narrower plots were not "suitable" for erosion experiments. In their study, effects of plot length could be satisfactorily described with the LS factor of the USLE down to a plot length of ~4.5 m and with the RUSLE for interrill plots of ~0.75 m. Greater slope lengths allow for more development of overland flow, thus surface hydraulic shear, which is expected to become the dominant erosive force as slope increases (Kamalu, 1994). For example, Goff et al, (1993) found that soil loss increased linearly with runoff plot downslope length for bare soils. In contrast to some other studies, Auerswald et al. (1992) found that as their plot size decreased, runoff began later, not only as a result of plot length (r=0.78), but mainly from plot size (r=0.92). Large time to runoff lags on small plots complicated interpretation of their results leading to a recommendation to "disregard rain erosivity" during the time lag for determination of USLE parameters.

Installation of the metal plot "frame" several cm into the soil serves to define the runoff area, limit upslope run-on and enable collection of runoff samples for later sediment analyses. This installation process involves some surface and soil disturbance and the relative success of efforts to "seal" the edges (with possibly bentonite) cannot be evaluated, resulting in non-quantifiable "edge effects" from plot to plot. The plot frame can intercept splash erosion that may leave a layer of soil particles on the plot frame not replenished by particles from outside the frame; thereby reducing the amount of particles available for transport by the overland flow. Overall, the disturbance effect of frame installation is largely unknown and likely increases with the extent and depth of cover across the plot.

D. As with the various RS designs, no single or standard runoff sampling size, frequency and duration has been developed. As with other aspects of the RS methodology, runoff results have been presented as simply the total storm duration sediment mass, the sediment mass per unit area or depth of rain, the average sediment concentration during the simulation period or after steady-state runoff rates are achieved, the sediment mass per unit area and unit runoff, or as a computed erodibility from averaged sediment losses that incorporates the rain intensity and possibly the runoff rate. With the exception of simply reporting the sediment mass per unit area for the simulated rain period, the other values depend on the sampling frequency and when during the simulation the runoff sediment concentrations are selected. This issue can be better illustrated through some examples of data collected from a disturbed bare soil and a less-disturbed adjacent, deep-duff covered forest soil of the same type from the north shore of Lake Tahoe. Both test plot yielded similar runoff rates and runoff sediment concentrations, but different types of results.

Figure 29 illustrates the basic information collected about the infiltration and runoff rates as well as sediment concentrations from continuous sampling of all runoff from the test plot frames for a 59.0 mm/hr simulated average rainfall intensity. Figure 30 is the corresponding graph of cumulative sediment collected in the runoff as a function of the cumulative runoff depth from the data shown in Figure 29. Note that after approximately 16 minutes of simulation, infiltration and runoff rates as well as sediment concentrations stabilize. In this case, however, the interrill erodibility can be calculated from the slope (sediment yield = 12.0 gm/mm) of the linear regression using the compete data set. With a more limited sampling, say every 2-3 minutes (4 samples total), the average SY is 13.1 gm/mm, or using only the last four more "steady' flow samples, the SY is 11.6 gm/mm. These are relatively small differences as compared to those from plot to plot. For example, while all of the adjacent bare soil plots at Brockway had similar results as shown in Figures 29 and 30 and field slopes of 45-50%, they produced SYs that ranged from 6-12 gm/mm. Results from a RS test on the deep duff plots just upslope from the bare plot test area at similar field slopes are illustrated in Figure 31; the corresponding cumulative sediment and runoff information is presented in Figure 32. In this case, steady infiltration and runoff began at about the same time as that for the bare soil plots, though the sediment concentrations were far more variable as is more typical of low runoff/erosion from relatively undisturbed forest soils. Clearly, in Figure 32, the linear regression fits the data poorly and suggests a SY of ~9.6 gm/mm. Using periodic sampling every 2-3 minutes as described for the bare soil plot, or 4 and 8 of the last "steady" flow runoff and sediment data suggests SYs of 7.36, 4.18 and 6.95 gm/mm, respectively; values that differ substantially, with selection of the latter four points from the test seemingly the most appropriate. However, if the test had been terminated earlier after "steady" runoff conditions were achieved, the larger SY value would have likely been used to determine erodibility. Again, plot-to-plot variability was similar to that of the bare soil plots. In either case, continuous sampling is valuable towards interpretation of the collected data and the methodology chosen to select the data used in the determination of “erodibility” should be specified.

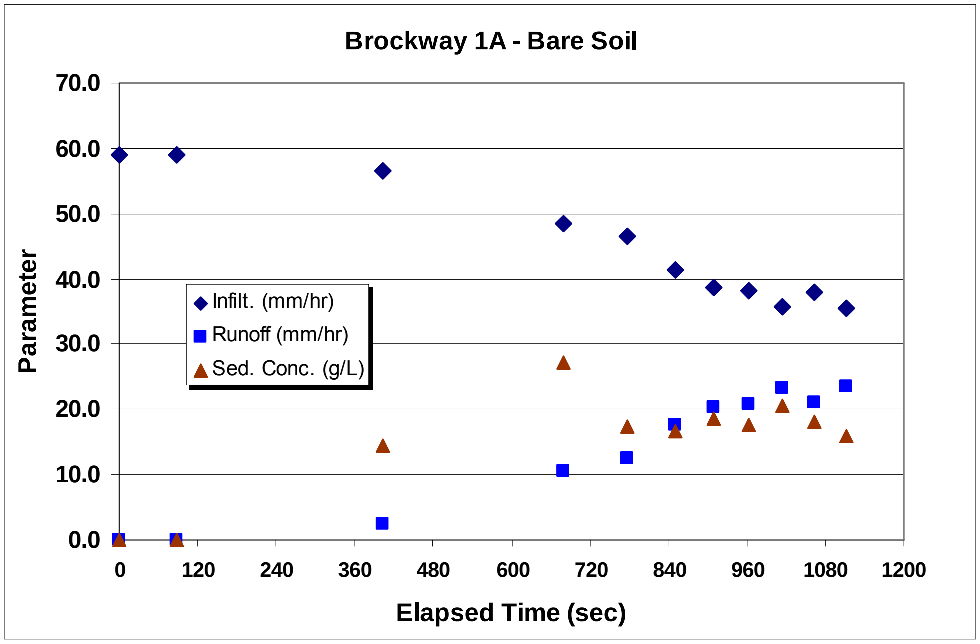


Figure 29. Example RS-derived infiltration, runoff and erosion data from 1 m^2 test plot of volcanic disturbed bare soil plot on a 47.0% slope.

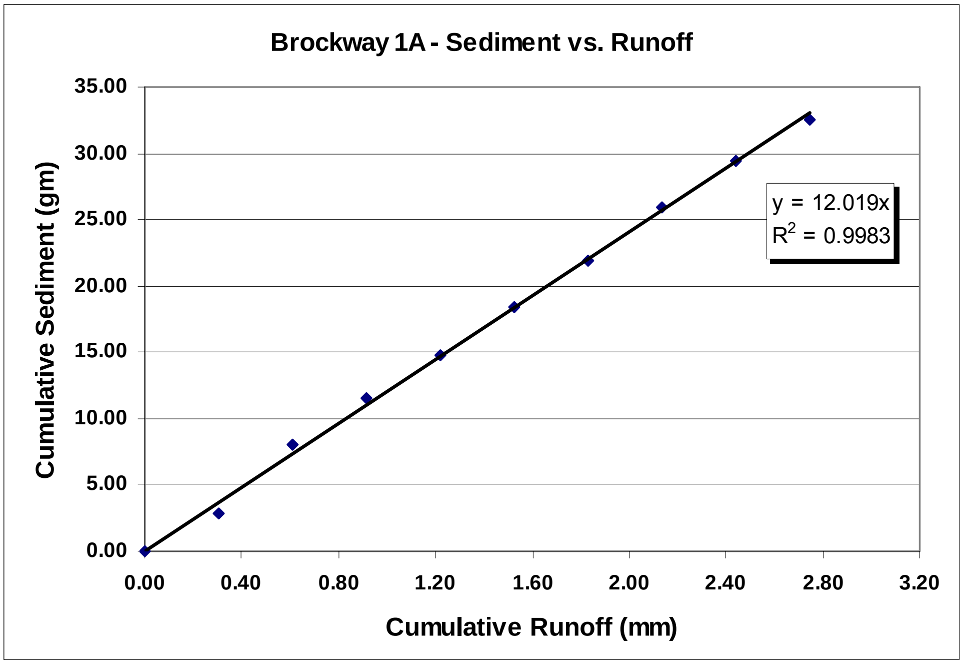


Figure 30. Cumulative sediment as it depends on cumulative runoff from 1 m^2 test plot of Figure 29.

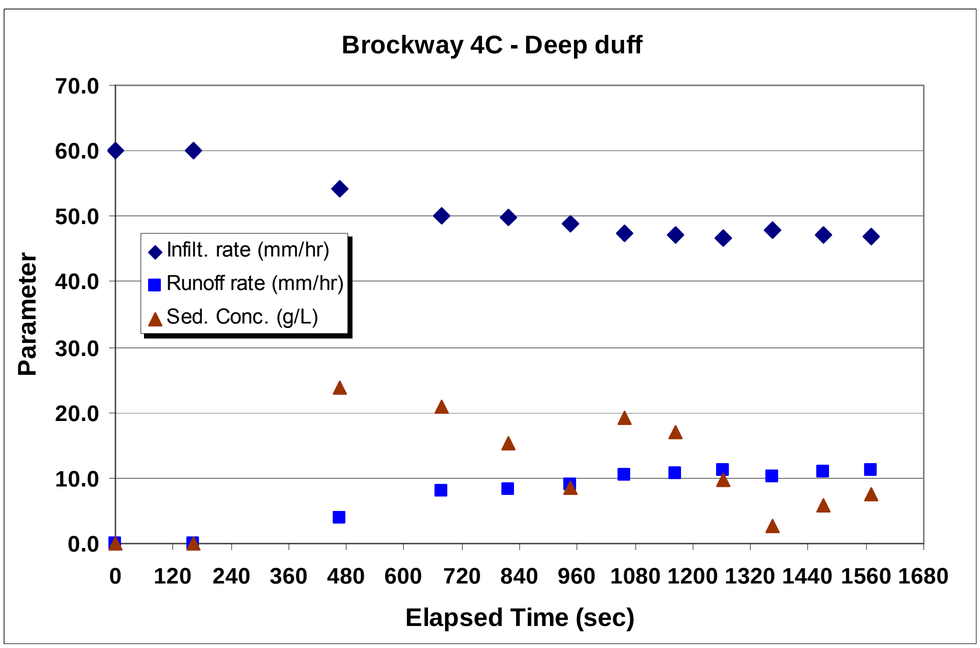


Figure 31. Example RS-derived infiltration, runoff and erosion data from 1 m^2 test plot of volcanic soil with deep duff cover on a 45.4% slope.

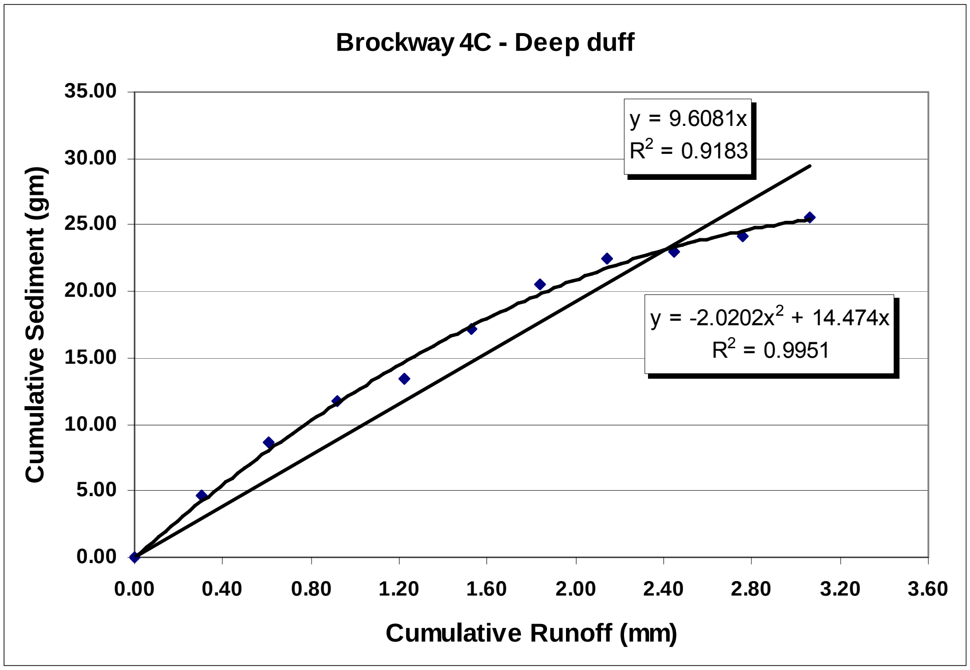
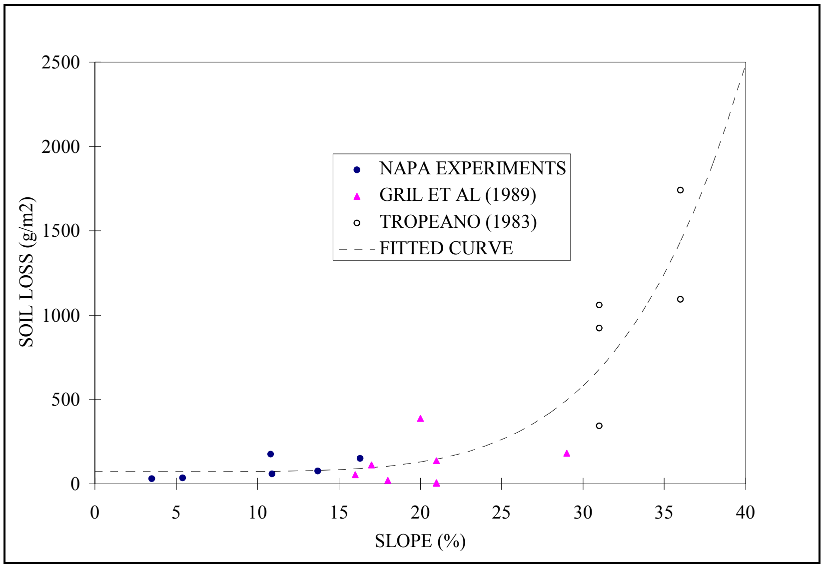


Figure 32. Cumulative sediment as it depends on cumulative runoff from 1 m^2 test plot of Figure 31.

E. The relative surface roughness of bare soils and the extent and type of cover for planted or mulched surfaces can have a significant effect on measured sediment losses and possibly infiltration and runoff rates from the test plots. No standard methods are available for describing or determining the nature of the surface soil and cover conditions. Surface roughness for bare soils has been measured by a variety of methods including use of multiple pin heights across one or more plot transects, or more recently, use of LIDAR methods in a similar fashion. It appears that for small plots, that moderate relative roughness is a minor factor as compared to cover effects with respect to measured sediment losses. Surface cover determinations depend on the method chosen, but usually involve estimation of the areal extent of the coverage and the type of coverage. Cover-point methods taken from the plant sciences have also been used to determine the actual plant or mulch cover with some statistical significance. Such cover fraction estimation methods alone are inadequate to characterize the "cover" conditions and investigators should provide as much detailed information about not only plot fraction covered, but also the type of cover, the materials comprising the cover, the cover thickness and relative age, among others.

Determination of test plot slope is generally straightforward and most methodologies reported involve either simple measurements using long carpenter levels and tape measures or surveying in surface elevations using an auto-level. However, the effects of slope towards measured erosion rates as compared to that of cover/mulch conditions appears to be much smaller. Conflicting results considering the effects of slope have been reported historically; conceptually, however, as slope increases, erosion rates should increase as a result of greater effects on gravity on surface flow rates and downhill particle movement at steeper slopes. This dependence of erosion rates on slope is captured in both USLE and WEPP type equations outlined above. For bare or nearly bare soils, erosion rates tend to increase more rapidly with slope resulting in something of a power relationship between the erodibility and slope, particularly at slopes steeper than ~25%. By way of an example, Figure 33 illustrates results from three different vineyard erosion studies in which the relationship between sediment losses increases exponentially. Grismer and Hogan (2004, 2005) reported similar relationships with the effects of slope on SYs decreasing in importance with increasing restoration effort (varying mulch depth layers, mulch/woodchip incorporation, etc.).

Figure 33. Effect of slope on soil loss for 40 mm rainfall depth and equivalent intensity simulated and natural rainfall storms (Battany, 1998).



In addition to basic slope and cover information, knowledge about surface soil moisture prior to rainfall simulation is helpful towards explaining time lags to initiation of runoff and possible differences in total sediment losses from similar plots. Initial, or antecedent soil-water content also affects aggregate destruction/disintegration. Ward and Bolton (1991), Blum and Gomes (1999) and Duiker et al., (2000) suggested that antecedent soil moisture is “the most efficient factor determining SY”. LeBissonnais and others showed that moist soil erodes less than dry soil because of less aggregate disruption. Historically, erosion studies on agricultural soils have shown that when surface soils are at moisture contents greater than field capacity, soil losses increase considerably over that from comparably dry soils; by as much as five times (Luk, 1985), or much greater sediment concentrations (Benito et al., 2003). On the other hand, previous rain events on a plot may deplete available sediment for transport such that smaller interrill erodibilities are determined after successive rain events over the same plots despite greater initial soil moisture contents. For example, on mildly sloped (~10%) dirt roads in the Tahoe Basin, Foltz et al. (2009) found that erodibilities decreased by a factors of approximately four and two for "brushed-in" and "re-opened" road conditions, respectively, during the third simulated rainfall event. Overall, the antecedent water content effect remains unclear and may not be entirely straight forward.

Finally, it has long been known that many forest soils are susceptible to surface crusting or water repellency (hydrophobicity) that result in unusually large runoff rates, though smaller SYs or inter-rill erodibilities. Hydrophobic soils found after fire events in the forest limit infiltration rates despite the often dry soil conditions and the increased runoff rates result in greater rilling and net sediment losses from the watershed following the first rain event after the fire. Late-summer and early fall dry conditions also result in litter/duff layers developing hydrophobic covers. Where appropriate, investigators should provide some information about the relative hydropbicity of the soil test plots – use of a simple infiltrometer with and without surfactant provides a rapid quantitative assessment of surface hydrophobicity (e.g. Robichaud et al., 2008; and Rice and Grismer, 2010).

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F. As suggested in Table AA, there are a large variety of intensities, plots sizes and shapes used in RS erosion studies increasing the difficulties in comparing data and results between studies as such data may reflect only interrill or a combination of interrill and rill erosion. Typically, when larger plots are used (e.g. Marques et al., 2007 who used 80 m^2 plots) the measured erosion rates are attributed to both rill and interrill processes. These potential sources of erosion data variation have been reported by authors around the world. For example, Loch and Donnollan (1983) and Loch and Thomas (1987) suggested that a 2 m plot downslope length was insufficient to generate rill erosion and that rill erosion could be generated by introducing surface flows at the top of 12 m long plots (Loch (2000a). Similarly, Parsons et al. (2006) again demonstrated the relationship between plot length and SY. Boix-Fayos et al (2006) reported in a review the sources of variation with the use of field plots that *"scale issues, disturbance and the representation of natural conditions (continuity, connectivity and heterogeneity of natural systems) and the complexity of the ecosystem interactions (connectivity, patterns and processes operating across scales) are key-questions when trying to collect representative field data using erosion plots."* Erosion rates from small plot RS studies are assumed to reflect interrill erosion processes and potentially miss the erosion produced in gullies (Hamed et al, 2002) at larger scales. But this distinction in processes is not at all clear in medium and larger field plots (Vahabbi & Nikami, 2008) and even with high RS intensities. It may be important to observe and check which is the dominant erosion processes in the area of study (Martinez-Zabala et al., 2008) and under what specific experimental conditions (Pappas et al, 2008; Sheridan et al, 2008) it applies. Some authors, even using small erosion plots, attribute high rates of erosion or changes in the size distribution of the sediments, to rill development during the experiments such as Jin et al (2009) who applied three different high rainfall intensities (65, 85 and 105 mm h^{-1}) and observed rill formation under high rainfall intensities obtaining smaller fine particle fractions in the eroded sediments. Similarly, Tatard et al (2008) underscored that sometimes rill erosion is the major part of total erosion, even on small plots in short time periods but under high intensities. "*Recent studies based on rare earth elements have shown experimentally that rill erosion can produce 4.3 to 5 times (Song et al., 2003) and even 29 times (Whiting et al., 2001) as much sediment as interrill erosion.*" Even on small plots (1.5 x 3 m), Yang et al. (2006) showed that simulated rainfall at an intensity of 73 mm/hr can cause twice as much rill erosion as interrill erosion after only 13 min of runoff. Tatard et al.'s (2008) results show that supercritical flows are a necessary condition for a rill to emerge from a smooth surface. Yang et al., (2006) suggested that use of radionuclides may be necessary to finally distinguish interrill from rill erosion in practice.

More recently, runoff simulators have been deployed in forested catchments to determine rill erosion rates in the Tahoe Basin and the Pacific Northwest (e.g. Hatchett et al., 2006; and Robichaud et al., 2010). Though designs are not well documented, the runoff simulator is typically a pipe manifold with energy dissipating material downslope that enables application of measured surface flows across a width of 1-2 m. About 2-9 m downslope a metal barrier is placed to funnel and collect runoff samples. With the exception of the rainfall KE issue, many of the same experimental concerns discussed here apply to use of runoff simulators (e.g. flowrates, antecedent soil moisture, replicability). Similar to RS studies, results have been variable, though less-disturbed forest soils yield consistently and significantly smaller erosion rates as compared to disturbed soils (e.g. roads, burned areas, skid trails). For example, Robichaud et al. (2010) found no significant rill erosion rate dependence on forest slopes between 18-79% due in part to highly variable though very small rates. Sediment flux rates decreased with increasing plot length (2 to 9 m) for less-disturbed sites, while they increased for more disturbed sites.

G. In addition to portability and access, a key advantage of small plot RS studies is the ability to more readily replicate plots in an effort to capture something of the hillslope hydrologic dynamics. The need for adequate sampling of erosion rates has plagued erosion studies for decades (Nearing et al., 1999) and the number of plots needed for statistically significant replication is typically quite large and beyond what is practically feasible in the field. Nearing et al. (1999) considered replicated plot variability effects on measured erosion rates for storm, annual and multi-year periods and noted that measured variability decreased as a power function with increasing sediment yields. At the practical scale typical of small plot studies, plot variability may overwhelm other factors leaving interpretation of results ambiguous. For example, in a RS erosion study of a range of arid soil conditions (43 plots) in Spain, Calvo-Cases (1991) found “the relationships between previous conditions and response to simulated rainfall are very variable, with poor correlation between variables." Using a Spearman rank correlation, slope (ranged from 7-43%) had very little influence, while soil moisture had a small positive influence on time to runoff. The dominant correlation was between cover and sediment concentration, with an apparent threshold cover of at least 20% required before sediment losses decreased. He later grouped the various plots more appropriately and underscored the effects of cover or soil cracking on runoff and erosion rates. While such a "blanket" approach to analyses of erosion plot data is probably not warranted, it provides some insight into the plot variability in erosion studies. In forest soils, it appears that disturbances associated with logging or roads result in less variability (spatial) in erosion rates between plots (e.g. Page-Dumrose & Jurgensen, 2006) as compared to less-disturbed forest soils (Arnaez et al., 2004; and Ziegler & Giambelluca, 1997); presumably an effect of soil compaction. Nonetheless, plot-to-plot or spatial variability remains large; Foltz et al. (2009) for 12-15 forest road test plots in Idaho found that the re-opened road erodibilities had a coefficient of variation (*CV*) of ~30% as compared to "brushed-in" (semi-restored) road *CV* of ~77%. They obtained somewhat similar results in the Tahoe Basin for these two road conditions with *CVs* of ~30% from 10 test plots. Grismer and Hogan (date) have found that for low runoff /erosion, less-disturbed forest soils such plot-to-plot variability spans an order of magnitude. (*CV*~100%). For many forest erosion studies, the question of plot replication requirements remains open, typically 3-10 plots are tested; this number ultimately depending not only on available time and resources to conduct the study, but also available land space with similar soil, slope and cover conditions.

H. Outside of disturbance areas associated with logging, trails and roads, forested soils are typically covered with mulch/litter/duff layers that can dramatically influence rates of runoff and sediment losses from the study plots. These layers can be fairly thick, as much as 10 cm, and partially "incorporated" into the surface mineral soil. The meaning of interrill erodibility in these cases of thick surface layers is not clear as some of the assumed processes outlined above may not be present. For example from field observations of RS tests on thickly pine-needle mulched soils, there is no obvious rain splash detachment of mineral particles and some particle filtration may be occurring. Similarly, the effects of slope and runoff rates on “erosion” rates may not be apparent, and at the same time, provide some insight into the plot variability described above by Calvo-Cases (1991). Loose upper layers on some Tahoe Basin hillslopes result in shallow subsurface flows downslope at depths less than 30 cm during RS tests that result in unusually high “apparent” measured infiltration rates. Figure 34 illustrates RS test plot SY as compared to slope results for “treated” granitic soils around the Tahoe Basin; “covers” refer to grass planted or lightly mulched covered soils, while “incorps” refer to “amended” soils in which compost, woodchips or combinations thereof are lightly tilled or incorporated into the upper soil horizons by the snowpack. In this figure, the effects of slope (and runoff rate implicitly) apparently diminish with greater “treatment” such that “incorp” type test plots developed SYs similar to that of less-disturbed “native” soils within the forest canopy. Moreover, increased “treatment” also shifts the collected runoff (if any) sediment sizes to larger particles that may be associated with the greater organic matter concentrations associated with “incorp” or “native” test plots (Grismer and Hogan, 2005b; and Grismer et al., 2008). Such results are not unlike those observed in other semi-arid regions. For example, several investigators (Boix-Fayos, 1999; Cammeraat, 2002; Calvo-Cases et al., 2003; and Boix-Fayos et al., 1998, 2001 & 2005) have described how improvement of such soil properties as organic matter content and aggregation result in greater infiltration capacities and water availability such that soil-microbe-plant organic factors control runoff and erosion rates while developing an organic feedback loop to sustain reduced erosion rates. This has been observed at both slope and patch scales, with the vegetation cover and the organic matter content being the most important parameters controlling soil aggregation processes and runoff generation (Boix- Fayos et al., 2006). Nonetheless, how to interpret runoff "sediment" sampling information relative to the local mineral soils under "native" conditions remains challenging.

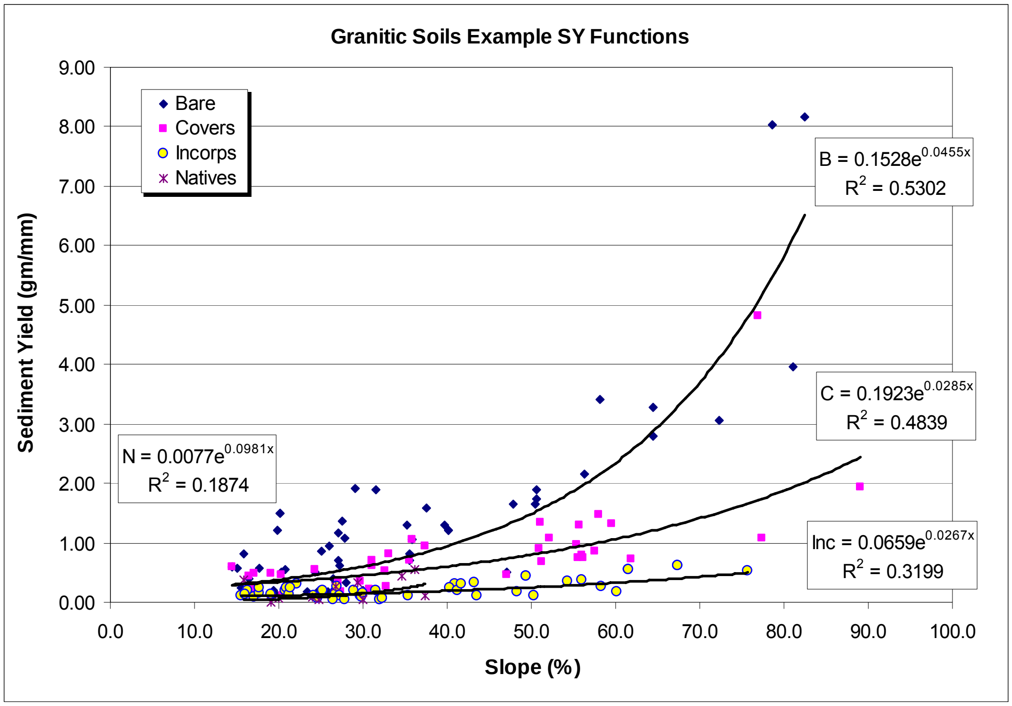

Figure 34. Example SY versus slope functions from bare, treated and "native" RS test plots in the Tahoe Basin (from Drake et al., 2010).

APPLICABILITY OF FIELD RS DERIVED EROSION RATES – Up-scalability?

A primary purpose behind conducting RS erosion evaluations in the field is to guide development of more general assessments of hillslope and catchment runoff and erosion rates that are associated with the different soils and land-use conditions of concern. In some cases, the stated purpose of the RS tests is to determine erodibility parameters for use in the USLE or WEPP from which estimates of runoff and erosion rates from larger areas are calculated. While small plot RS studies are uniquely suited to such determinations, they are also compromised by the limited extent to which the tests represent reality with respect to simulated rainfall characteristics as compared to "natural" rain and the typically small range of plot soil and land use conditions considered in the study as compared to that encountered in the hillslope or catchment. The restricted range of fixed simulated rainfall intensities, invariant drop-size distributions and KEs reproduced by RSs that are not characteristic of the variability found in natural storms implies that natural storm conditions are poorly represented (Wainwright et al., 2000; and Dunkerly, 2008) and that subsequent erosion response is at best simplified. Parsons and Stone (2006) suggest that the present understanding of the processes of interrill soil detachment and transport is inadequate to predict runoff and erosion rates associated with the temporal variability in drop sizes and intensities found in natural rain. In a catchment modeling exercise using a dynamic distributed watershed model, Smith et al. (1999) found that with the exception of very low rainfall events, erosion catchment sediment yields were more sensitive to "to changes in runoff and flow velocity than the splash and hydraulic detachment parameters" that would be determined for bare soils from Small plot RS studies. Agassi and Bradford (1999) suggested that the lack of a uniform coverage across a large area and the lack of a continuous coverage at low rainfall intensity were two of the main problems of RS experiments; however, this is precisely the advantage of RS experiments in that they remove one degree of freedom by keeping rain intensity and drop sizes constant, thereby presumably simplifying the task of discovering relationships between rainfall and runoff or erosion (Lascelles et al., 2000). Some of the issues associated with field variability including that introduced by erosion plot experimental design (Zobisch et al., 1996) were recognized more than a decade ago (e.g. Bagarello and Ferro (1998); and Nearing et al., 1999). Unexplained variability between erosion test plot results (even in apparently homogeneous cultivated fields, Rüttiman et al., 1995) remains perplexing and limits development of more generalized conclusions about runoff and erosion rates (e.g. Wendt et al., 1986; and Gómez et al., 2001). As noted above, within site variability of 30% to 75% between the plots located on a seemingly homogeneous landscape are common. At the same time, a general demand remains for knowledge about the soil erosion processes occurring in field plots across a range of sizes, the threshold limits at which different processes are significant, and of factors that determine natural variability (Bagarello and Ferro, 2004). To establish the influence of plot length on soil loss and meet this need in part, Bagarello and Ferro

(2010) measured soil losses from a high number of replicated, bare plots of different lengths (0.25, 0.4, 1, 2, 5, 11, 22, 33 and 44 m) all on a 14.9% slope maintained continuously fallow, simultaneously operating in the period 1999–2008 south of Palermo Italy. Overall, they found a lack of significant relationship between soil loss and slope length that was associated with an increasing sediment concentration versus plot length relationship and a runoff volume per unit area that decreased or did not vary with plot length. Mean sediment concentration coefficients of variation (*CV*) ranged up to 170% for microplots (up to 0.4x0.4 m) at low values decreasing to less than 50% for larger plots and larger means; a dependency observed by others as discussed above.

Garcia-Ruiz et al. (2010) underscored the importance of considering various spatial and temporal scales since it is well known that geomorphic and hydrological processes are scale-dependent with each scale underpinning certain processes. Rainfall simulation type studies tend to focus only on experimental plots or emphasize processes such as infiltration, splash or runoff generation, but do not consider connectivity with the fluvial channel and the consequences on sediment outputs from catchments and on temporal sediment stores. Similarly, studies at the regional scale can enable sediment balances to be assessed and identify sediment sources for large basins, but cannot contribute to understanding of what is happening “within the slopes”. They advocate a holistic perspective of the hydromorphological functioning of the region that then requires a multiscale approach integrating slopes, small catchments, large basins, and fluvial channels.

The spatial scaling issue from small plots to hillslope/catchment involves at least three components that are beyond the scope of small-plot RS studies; (a) natural heterogeneity of soil conditions (e.g. infiltration and erosion rates) across the hillslope, or plot-to-plot variability as described above, (b) inter-connectivity between measured and non-measured areas, or between eroding and depositional areas, and (c) soil plot disturbance effects as a result of the RS measurements. Le Bissonnais et al. (1998) noted the need to consider catchment spatial structure while Garcia-Ruiz et al. (2010) highlighted that ultimately connectivity with fluvial channels is the important factor linking plot to catchment studies. On the other hand, using the Guelph RS at rates of 60 and 140 mm/hr, Nolan et al. (1997) successfully linked small to large plot scale measurements of erosion rates from different tillage regimes to that from natural rainfall through adjustments for slope length and rainfall KE. Similarly, Hamed et al. (2002) matched the RS measured erosivity to the Wischmeier and Smith R value correcting for slope and rainfall energy and successfully predicted net sediment losses for 2 of 3 storms from a semi-arid, mildly sloped (2-8%) 158 ha catchment in Tunisia. Parsons et al. (2006) asserted that sediment yield from plots in Arizona increased with increasing plot length and then decreased, suggesting some maximum value associated with a plot length between 4 and 14 m. Kinnell (2008) disputed this claim and indicated that the correct interpretation was that the plot sediment yield was runoff rate dependent as described above and that the apparent maximum at plot lengths between 4 and 14 m was an experimental artifact of changing runoff coefficients. Nonetheless, though individual up-scaling issues have been discussed by several researchers, Boix-Fayos et al. (2006) sought to review these issues as posed in the following framework; "*(i) temporal and spatial scales, (ii) representation of natural conditions, (iii) the disturbance of natural conditions and (iv) accounting for the complexity of ecosystem interactions.*" Ultimately, the uncertainties associated with these issues are set aside to a degree such that erosion predictions can be made as part of watershed process modeling to evaluate the effects of changing landscape conditions (i.e. disturbance or restoration) on watershed health and discharge water quality.

Possibly conflicting research has developed relating erosion estimates from plot-based measurements to that of the hillslope or catchment. Unfortunately, actual field data on infiltration and erosion rates at different spatial scales from 1 to beyond 10s of meters is difficult to obtain and little can be found in the literature (Le Bissonnais et al., 1998; Bagarello and Ferro, 2004), since most field measurements have concentrated on water erosion processes operating at the runoff plot scale (Poesen and Hooke, 1997). For example, Boix-Fayos et al. (2006) found that soil loss is underestimated from RS plots as compared to that from natural rain plot experiments (Chaplot and Le Bissonnais, 2000; Hamed et al., 2002; Calvo-Cases et al., 2003) and attributed this difference to the constant intensities and relatively low KEs generated by the RSs used. They recognized that exceptions to this under-estimation can be found, but that these occurred because the simulated rain applied was at extremely high intensities that generated greater than natural runoff rates (e.g. Schlesinger et al., 1999, 2000). In most cases reported, extrapolation of test results on bare soils results in an overestimation of erosion at hillslope and catchment scales (Loughran, 1989; Evans, 1995; and Poesen et al., 2003). Le Bissonnais et al. (1998) estimated a scaling factor of ~2 to relate sediment concentrations between 20 and 1 m^2 plots, and ~0.5 for sediment concentrations from 500 and 20 m^2 plots. Results from the 1 m^2 plots underestimate soil losses as compared to that from the 20 m^2 plots due to smaller surface flow velocity and transport capacity (Chaplot and Le Bissonnais, 2000), while erosion test results from the 20 m^2 plots overestimated soil losses as compared to that from the 500 m^2 plots because of the greater likelihood of variable or preferential infiltration rates with increasing plot size. Of course, soil loss data obtained at the plot scale are difficult to extrapolate to the catchment level because heterogeneity at the catchment scale is always greater than that of a plot. In the experiment conducted by Le Bissonnais et al. (1998), the conditions of their studied catchment were more homogeneous in the winter season, when the response of the catchment was similar to that of the 500 m^2 plot. Grismer (2011) used 1 m^2 erosion test plot information relating SYs to soil type, soil condition and slope developed for a wide range of conditions across a range of 15 land-use categories and two parent soil types to model daily sediment loads from "paired" watersheds ranging from 261 to 530 ha on the Tahoe Basin west shore. Analogous to Le Bissonnais et al. (1998), he found that the scaling factor (SF) need to take the plot level SY function sediment loads per unit of runoff to that of the watersheds to be runoff depth (R, mm) dependent (i.e. $SF=0.1917/R^{0.50}$) across 12 water years of simulation, such that factors of 5-7 result for average runoff depths of 1-2 mm.

Recent Rainfall Simulation Studies in the Tahoe Basin

Studies by Munn (1974) are likely some of the earliest RS-oriented erosion studies in the Tahoe Basin. He evaluated the erosion potential of seven different soil types in the Lake Tahoe Basin, under both natural and disturbed conditions. Munn built and used a highly portable drop-former RS design. Rain occurred over a square (0.71 x 0.71 m) area, employing catheter tubing to form drops with a fall height of 2.5 m; water was supplied by gravity from a 20 liter jug mounted atop the simulator. The square runoff collection frame (0.61 x 0.61 m) channeled runoff into collection jars during the 15-minute duration storms. Overall, Munn (1974) reported greater amounts of erosion from steeper slopes and estimated erodibilities from several soil series found in the Basin, identifying those most likely to present erosion problems.

Later, Guerrant et al (1990) compared four different types of rainfall simulators for use in the Lake Tahoe Basin, consisting of a modular needle-type drop-former, and three nozzle-type simulators. They concluded that the modular needle-type design was the most practical, due to its low labor and water needs, ease of setup, and ability to operate on difficult terrain. Plot condition was found to significantly influence infiltration rates and the effect of each plot conditions was significantly different. Measured infiltration rates ranged from 47-62 mm/hr from rainfall intensities of 80-100 mm/hr. Using the drop-former RS described by Guerrant et al (1990), Guerrant et al (1991) further investigated the effect of three slope ranges (015%, 15-30% & >30%) and four soil conditions (undisturbed with duff, undisturbed without duff, disturbed with duff, disturbed without duff) for the Cagwin soil series. Infiltration and runoff rates similar to earlier rates were found. However, slope was found to have a negligible effect on infiltration and runoff rates, but had s significant positive effect on erosion rates. Though there were some conflicting results from the various plots, generally plot condition had a significant effect on infiltration, runoff and erosion rates. Continuing, Naslas et al (1994) used the same RS to evaluate runoff and erosion as influenced by different soil types, slopes, and cover conditions in the Lake Tahoe Basin. They concluded that a three-way interaction existed between these factors, with greater amounts of runoff and erosion occurring at greater slopes, and less runoff yet increased erosion with increased plot disturbance.

Beginning in 2001, Grismer and others began RS studies using the RS described by Battany and Grismer (2000) at first directed at roadcut slopes around the Basin and later expanded to include other disturbed soil areas of the Basin catchments. They developed a series of papers considering the RS method, the effects of soil type, slope and restoration treatment on erosion rates (SYs) and runoff particle-size distributions (PSDs). Grismer & Hogan (2004) conduct a preliminary assessment of the effectiveness of a variety of erosion control treatments and treatment effects on hydrologic parameters and erosion. The particular goal of this paper was to determine if the RS method could measure revegetation treatment effects on infiltration and erosion. The RS-plot studies were used to determine slope, cover (mulch and vegetation) and surface roughness effects on infiltration, runoff and erosion rates at several roadcuts across the Basin. Measured parameters included time to runoff, infiltration, runoff/infiltration rate, sediment discharge rate and average sediment concentration as well as analysis of total Kjeldahl nitrogen (TKN) and dissolved phosphorus (TDP) from filtered (0.45 μm) runoff samples. Runoff rates, sediment concentrations and yields were greater from volcanic soils as compared to that from granitic soils for nearly all cover conditions. For example, bare soil SYs from volcanic-derived soils ranged from 2 -12 as compared to 0.3-3 gm m^{-2} mm^{-1} for granitic-derived soils. Pine needle mulch cover treatments substantially reduced SYs from all plots. Plot micro-topography or roughness and cross-slope had no effect on sediment concentrations in runoff or SY. Runoff nutrient concentrations were not distinguishable from that in the rainwater used. Grismer & Hogan (2005a) included multiple RS test replications of bare soil plots as well as some adjacent "native", or relatively undisturbed soils below trees where available. Laboratory measurements of PSDs using sieve and laser counting methods indicated that the granitic soils had larger grain sizes than the volcanic soils and that road cut soils of either type also had larger grain sizes than their ski run counterparts. Soil PSD based estimates of saturated hydraulic conductivity were 5-10 times greater than RS determined steady infiltration rates. RS measured infiltration rates were similar, ranging from 33-50 mm/hr for disturbed volcanic soils and 33-60 mm/hr for disturbed granitic soils. RS measured runoff rates and sediment yields from the bare soils were significantly correlated with plot slope with the exception of volcanic road cuts due to the narrow range of road cut slopes encountered. Sediment yields from bare granitic soils at slopes of 28 to 78% ranged from ~1 – 12 g m^{-2} mm^{-1}, respectively, while from bare volcanic soils at slopes of 22 – 61% ranged from ~3 – 31 g m^{-2} mm^{-1}, respectively. As was found in the first study, surface roughness did not correlate with runoff or erosion parameters, perhaps also as a result of a relatively narrow range of roughness values. The volcanic ski run soils and both types of road cut soils exhibited nearly an order of magnitude greater sediment yield than that from the corresponding native, relatively undisturbed sites. Similarly, the granitic ski run soils produced nearly four times greater sediment concentration than the undisturbed areas. Grismer & Hogan (2005b) built upon results from use of the portable rainfall simulator (RS) described in the previous two papers to evaluate cover and revegetation treatment effects on runoff rates and sediment concentrations and yields from disturbed granitic and volcanic soils at road cuts and ski runs in the Basin. The effects of slope on rainfall runoff, infiltration and erosion rates were determined at several revegetated road cut and ski run sites. Runoff sediment concentrations and yields from sparsely covered volcanic and bare granitic soils could be correlated to slope. Sediment concentrations and yields from nearly bare volcanic soils exceeded those from granitic soils by an order of magnitude across slopes ranging from 30-70%. Revegetation, or application of pine needle mulch covers to both soil types decreased sediment concentrations and yields 30-50%. Incorporation of woodchips or soil rehabilitation that included tillage, use of amendments (Biosol®, compost) and mulch covers together with plant seeding resulted in little, or no runoff or sediment yield from both soils. Follow-up measurements of sediment concentrations and yields from the same plots in the subsequent two years after woodchip or soil rehabilitation treatments continued to result in little or no runoff. Revegetation treatments involving use of only grasses to cover soils were largely ineffective due to sparse sustainable coverage (<35%) and inadequate infiltration rates.

As concern over runoff PSDs increased in the Basin, the focus of the RS studies shifted slightly to consider soil, slope and treatment effects on runoff sediment PSDs. Grismer and Ellis (2006) and Grismer et al. (2007) reported that granitic soils had larger particle sizes than volcanic soils in both bulk soil and runoff samples. Later, they made an effort to develop quantified information about erosion rates and runoff PSDs for determining stream and Lake loading associated with land management. They determined the dependence and significance of runoff sediment PSDs and SY on slope and compared these relationships between erosion control treatments (e.g. mulch covers, compost, or woodchip incorporation, plantings) with bare and undisturbed, or "native" forest soils. As granitic soils had larger particle-sizes than volcanic soils in bulk soil and runoff samples, runoff rates, SCs and SYs were greater from bare volcanic as compared to that from bare granitic soils at similar slopes. Generally, runoff rates increased with increasing slope on bare soils, while infiltration rates decreased. Similarly, SY increased with slope for both soil types, though SYs from volcanic soils are 3-4 times larger than those from granitic soils. As SY increased, smaller particle-sizes are observed in runoff for all soil conditions and particle-sizes decreased with increasing slope. Combined soil restoration with pine needle mulch cover treatments substantially reduced SYs as well as increasing average runoff particle size as compared to those from bare soils while very little, if any runoff and erosion occurred from native soil plots at similar slopes.

Grismer et al. (2009) acknowledged that revegetation and soil restoration efforts, often associated with erosion control measures on disturbed soils, are rarely monitored or otherwise evaluated in terms of improved hydrologic, much less, ecologic function and longer term sustainability. Numerous erosion control measures deployed in the Basin during the past several decades have under-performed, or simply failed after a few years and new soil restoration methods of erosion control are under investigation. They outlined a comprehensive, integrated field-based evaluation and assessment of the hydrologic function associated with these soil restoration methods with the hypothesis that restoration of sustainable function will result in longer term erosion control benefits than that currently achieved with more commonly used surface treatment methods (e.g. straw/mulch covers and hydroseeding). The monitoring includes cover-point and ocular assessments of plant cover, species type and diversity; soil sampling for nutrient status; rainfall simulation measurement of infiltration and runoff rates; cone penetrometer measurements of soil compaction and thickness of mulch layer depths. Through multi-year hydrologic and vegetation monitoring at ten sites and 120 plots, they illustrated the results obtained from the integrated monitoring program and describe how it might guide future restoration efforts and monitoring assessments.

As forest dirt roads and trails are some of the greatest sources of sediment loadings to streams per unit land area, Folz et al. (2009) and Copeland & Folz (2009) measured runoff and sediment concentration during simulated rainfall events for a variety of forest dirt road surfaces in Idaho and around the Tahoe Basin. Road slopes were generally on mild grades of~10% or less and from both volcanic and granitic parent materials. Simulated rainfall intensities of 80-100 mm/hr were used for 30-minute durations from a single Veejet 80100 nozzle located 3 m above the soil surface. The runoff rates measured on these roads followed trends typical of native surface forest road runoff hydrographs (Foltz et al., 2009; Sheridan et al., 2008). Measured infiltration rates of ~16 mm/hr were substantially less than those observed in forested areas of 40-50 mm/hr. While more recently opened or used roads generated greater sediment losses or erodibilities as compared to abandoned roads (Folz et al., 2009), Copeland and Folz (2009) found no soil dependence as found by Grismer and Hogan (2005a) for bare disturbed soils on steeper slopes. Copeland and Folz (2009) found that the two granitic-based roads demonstrated sediment concentration trends similar to those reported in other studies; however the volcanic-based roads followed a slightly different trend, beginning with relatively low sediment concentrations early in the rain events, gradually increasing to steady-state concentrations. Soil water repellency on the road running surfaces may have caused the sustained sediment concentrations measured during the rainfall events. They suggested that while shapes of the hydrographs and sedigraphs indicate differences in the hydrologic responses between grantic and volcanic based roads; they do not necessarily affect the model parameters, saturated hydraulic conductivity and calculated interrill erodibilities. Average interrill erodibility ranged from 0.7–1.2 x 10^6 kg s m^{-4}. As discussed above, high plot-to-plot variability in the measured parameters precluded assessment of differences among the different native roads or their parent materials.

Rice & Grismer (2010) found that though often critical towards estimation of runoff and erosion rates, knowledge of soil-water repellency remains over-generalized or anecdotal because few studies isolate and quantify repellency effects. They again employ the RS used in several previous studies, but now with a surfactant solution to investigate the effects of repellency at relatively undisturbed 'native' forested soil sites on slopes of 10-15%. These RS tests were compared with the often, more simply used Mini-Disk Infiltrometer (MDI) measurements of infiltration rates as a means of quantifying repellency effects. Repellency effects on infiltration were evident as all plots with untreated water produced runoff, while only 2 of 12 plots treated with surfactant had runoff. At the volcanic soil sites, MDI measured infiltration rates using surfactant exceeded those with water by 20% when there was little litter cover (Blackwood Canyon), and by factors of 3 with substantial litter cover (Truckee). Similarly, at the granitic soil sites, surfactant-enhanced MDI infiltration rates were 4 times greater with little litter (Bliss SP), and 8 times greater with substantial litter cover (Meyers RC). Infiltration rates differed significantly ($p<0.05$) due to the surfactant treatment for both methods at Bliss SP, and at 3 of 4 sites for the MDI. Post-simulation soil moisture content and wetting depth were greater with the surfactant treatment. Excavations following the RSs indicated that the surfactant treatment entered discontinuities in the highly hydrophobic organic layer and infiltrated preferentially through the mineral soil.

Finally, in an effort to relate RS plot measurement to catchment sediment loads, Grismer (2011a) made an effort to link local-scale field measurements associated with the range of land-uses or soil restoration efforts with the catchment-scale sediment loading. A distributed hydrologic model with locally–derived, slope dependent SY equations developed from RS studies at the 1 m^2 scale across the Basin is employed to determine the runoff-dependent scaling factors (SFs) necessary to predict daily stream sediment loading from the forested uplands comprising some 80% of the Tahoe Basin area. Here, SFs and loadings from three "paired", adjacent west shore Lake Tahoe tributary catchments of 261 (Homewood Cr.), 383 (Quail Cr.) and 530 ha (Madden Cr.) are considered during the period 1994-2004 at time scales ranging from daily to annual. For each of the three watersheds, there was no real dependence of the SF-runoff regression equations on type of water year (e.g. dry or wet), nor on dominant soil parent material (volcanic or granitic), or ranges of different land-use areas. At all time scales (daily, weekly, seasonal and annual), the SF was dependent on runoff (R), particularly at smaller values, but was readily simplified as an inverse square-root function (i.e. $SF=0.1917/R^{0.50}$). Optimized SF-runoff regressions for each watershed were equivalent when modified by ratios of watershed areas. As a result, a single daily SF-runoff equation was determined (through minimization of sediment load prediction errors) that could be successfully applied to all three watersheds with accuracy consistent with that predictive error associated with any one of the watersheds alone. Sensitivity analyses indicated that sediment loading predictions were more sensitive to the SF-runoff equation coefficient rather than the exponent. Annual sediment load prediction errors of ~30% might be expected for low or high runoff years. Grismer (2011b) continued this effort to determine the effect of areal extent of forest fuels reductions on daily sediment loads from the largely forested Madden Creek watershed, presuming only slight temporary degradation of soil function. Similarly in the Homewood Creek (HMR) watershed, the effects of proposed soil restoration (e.g. dirt road removal, skirun rehabilitation) towards daily load reductions were considered. Both modeling efforts were directed at an assessment of the threshold (by fractional area treated and/or soil function) required to obtain measurable changes in sediment loads; a concept not unlike that of threshold ERAs (equivalent roaded areas) used in cumulative watershed evaluations (CWEs). For example, in the Madden Creek watershed fuels management in more than 30% of the basin area was required to result in a detectable increase in daily sediment loads at the >95% confidence level. Similarly, considering substantial dirt road restoration (50% by roaded area) within the HMR watershed reduced mean daily sediment loads by 12-30 kg/day for average daily flows of 99 to 804 L/s, a reduction that could only be assessed with ~78% confidence using the entire 11-year record. However, including restoration of 20% of the skirun area (combined for ~5% of the catchment) further decreases the daily sediment load 15- 37 kg across this range of flowrates, but enables measurement of this reduction with >95% confidence for the 11-year record as well as in 2-3 years following restoration. The modeled daily flows and loads, based on accumulated hourly data, reflected the considerable variability associated with sediment concentration hysteresis in the hydrograph. Examining this problem in detail using continuous monitoring data at the adjacent Blackwood and Ward Creek watersheds to the north suggests that considering only the rising limb of the flow hydrograph reduces the sediment load-flow relationship variability considerably. That is, stream monitoring should focus on measurement of the daily spring snowmelt hydrograph rising limb flowrates and loads and subsequent computation of watershed sediment yields as a function of flowrate. Comparison of pre- and post-project rising limb aggregate catchment SY functions can then be used to determine the relative impacts of the project on daily sediment loads so as to guide TMDL “crediting” for load reduction efforts.

SUMMARY & CONCLUSIONS

This review was directed at developing literature-based information that can guide development of a standard RS methodology for small plot erosion studies in forested hillslopes. Following the style of Kinnel (1993), this information can be summarized as key questions and their associated responses, where possible, concerning conducting RS-erosion studies in forested catchments.

1. What are the characteristics of "natural" rain and how do they compare to simulated rainfall characteristics?

   a. Natural rainfall variability in drop size, their distribution, intensity and temporal patterns in terms of KEs or powers is high and RSs provide only a "snapshot" of "natural" rain.
   b. Natural rainfall powers range from ~0.05 to 1.2 W/m^2 while simulated rainfall powers are generally <0.8 W/m^2, the significance of this difference in terms of aggregate disintegration and particle detachment is unclear as the energies or powers required for either process are highly variable spatially and temporally and thus remain largely unknown.
   c. The relationships between applied rainfall energy, splash impact and the like and the energy/power needed for aggregate disintegration remains unknown.
   d. The connections between rainfall characteristics (e.g. median drop size, drop-size distribution, intensity and temporal patterns in terms of KEs or powers) and erosion rates are not clear, especially as these rates are affected by complicating factors of slope, infiltration rates (e.g. crusting) and of course cover.
   e. For comparative purposes between RS, the total rainfall energy or power applied in the simulated events should be computed by integration across the drop-size distribution and rainfall intensity rather than simply estimating the relative raindrop velocities to their estimated terminal velocities.

2. Which rainfall characteristics are important towards determination of erosion rates, or erodibilities ?

   a. For determination of erodibilities from bare soils, drop-size distribution and associated intensity and KEs are the primary important rain characteristics. Often, larger median or mean drop sizes in natural rains are associated with higher intensities, while in simulated rains this relationship depends on whether nozzles or drop formers are used.

   b. For determination of erodibilities from sloping, litter/duff covered forest soils, the likely key rain characteristic is simply rainfall intensity (runoff rates) as cover conditions limit raindrop impact effects on aggregate disintegration and particle detachment. However, no studies directed at elucidating these effects in the field are available, so the important rainfall characteristics in this case remain largely unknown.

3. Are there soil–related (e.g. aggregate stability or strength) and rainfall intensity, KE, or arrival rate "thresholds" critical to determination of erodibilities? If so, how can they be determined or measured if they are significant?

   a. While there appears to be some information suggesting possible energy related thresholds of aggregate stability that need to be exceeded prior to disintegration/detachment and particle transport from bare soils, the actual values for different soil conditions remain unknown as well as the particular soil factors (e.g. OC, or clay contents) controlling aggregate strength in the field. Moreover, as aggregate stability is a dynamic property, such thresholds, if they exist, are expected to be antecedent moisture and climate dependent.
   b. Under forest litter/duff cover conditions, other factors associated with OM content and hydrophobicity may be of greater importance than aggregate strengths.

4. What ***is*** erodibility in the context of the forested landscape, or deeply mulch/duff covered soils? How can it best be defined or measured in this case?

   a. The definition of erodibility depends on the conceptual equation applied and appears subject to temporal variability associated with surface processes such as crusting, hydrophobicity, and surface roughness.

b. At present, erodibility or erosivity remain realistically undefined for any conditions other than bare soils on mild slopes <10%.
c. Erosion rates expressed as mass per unit area or time alone are inadequate; should be expressed as mass per unit runoff, raindrop energy or power.
d. Information about infiltration rates, antecedent moisture and depth to less permeable layer, or relative level of soil compaction is also required when reporting erosion rates.

5. Given the considerable plot-to-plot variability in measured erosion rates from seemingly homogeneous areas, standard replication and statistical analyses approaches should be promoted. How many replications are sufficient to characterize the sample population of interest (e.g. runoff or erosion rates)?

a. As noted in earlier studies, plot variability effects increase with decreasing measured sediment yields and that the variability is so large in general that the number of “samples” required to approximately characterize the population distribution may be impractically large. Nonetheless, field RS experiments typically involve 3-20 plots and analyses assume normally distributed erosion rates. However, Grismer (date) found that application of ANOVA to test regression models of SY as it depends on runoff rate for bare granitic soils (n=32) resulted in non-normally distributed residuals and lack of variance homogeneity suggesting that use of ANOVA was invalid. Using a log transform of the SY values reasonably corrected the residual non-normality and variance heterogeneity resulting in an ANOVA result suggesting a significant (p=0.05) positive relationship between SY and runoff rate as expected; however, the R^2 values were quite low (~0.25) raising questions about the meaning of such analyses.
b. Similarly, under forest litter/duff cover conditions that typically result in much smaller erosion rates as compared to equivalently sloped bare soils, plot-to-plot variability is expected to be much greater, but may be of less practical importance in watershed planning/TMDL studies.

6. **6.** While erosion rates conceptually increase with increasing slope and associated increased runoff rate for a given rainfall intensity, is there a slope threshold(s) below which slope effects are negligible and above which they are significant?
    a. Some information suggests that plot variability within a given soil condition has a greater affect on measured erosion rates than increased slope at slopes less than ~20% for bare soils.
    b. Similarly, under forest litter/duff cover conditions, it appears that slope effects on erosion rates are greatly diminished up to slopes of ~50%.

7. Is there an implicit slope dependence of erodibility at larger slopes, even when defined as in Eq. 10, where effects of rainfall and runoff rates together with slope are explicitly considered?
    a. Maybe – see above.

8. At what combinations of bare soil slope length, surface runoff rate, slope angle, and surface condition (e.g. roughness) does rill erosion become dominant as compared to interrill erosion?
    a. This is an open question in the field and appears to depend on soil type.

9. While considerable attention has been given to RS rainfall characterization, little, if any, has been given to describing the runoff plot frame installation methods and assurance that they are capturing the surface erosion processes appropriately. There has been no study that quantifies the effects of runoff plot frame installation on measured erosion rates.

10. Plant/mulch/duff covers need careful descriptions and probably have a threshold-based effect that needs further clarification/definition.

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