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Testing a Mechanistic Soil Erosion Model for Three Selected Soil Types from Iran

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ABSTRACT

Hairsine and Rose (1991) developed a process-based soil erosion model which described the erosion transport of multiparticle sizes in sediment for rain-impacted flow in the absence of entrainment in overland flow. In order to test this model laboratory experiments were carried out in a detachment tray using simulated rainfall. Three contrasting soil types were subjected to simulated rainfall at different rates (25-110 mm h^{-1}) in a 35×30 ×10 cm detachment tray. Rainfall was applied using a rainfall simulator with a single scanning nozzle located four meters above the soil surface that emitted drops with a mean diameter (volumetric D_{50}) of 1.5 mm. Results showed that the Hairsine and Rose model can clearly describe the sensitivity of different soils to erosion by introducing three terms of detachability, re-detachability and settling velocity, though the model is unable to describe aggregate breakdown which takes place in one of the soil at higher rainfall rates. The experimentally observed relationship between ponding water depth and soil detachability parameters did not agree with previously proposed theories. In addition, the results showed that the Hairsine and Rose model tends to over-predict values at the lower end of the scale, and under-predict values at the upper end, although the average sediment concentration predicted for the entire data set is not greatly different from the average measured values.

Keywords: Rainfall detachment, Rain erosion, Sediment concentration, Soil erosion model.

INTRODUCTION

Soil erosion caused by water is a serious problem in many parts of the world and it stems from a combination of agricultural intensification, soil degradation, and intense rainstorms. Many planning and management theories and formulas have been developed to help reduce soil loss from basins and sediment transport to hydrologic drainage networks.

In recent decades, models have been developed (empirical, conceptual, or physically based) in order to represent and to quantify the processes of detachment, transport, and deposition of eroded soil, with the aim of implementing assessment tools for educational, planning, and legislative purposes (Renschler and Harbor, 2002). Since the phenomena are complex and depend on many parameters, the calibration of models is difficult.

Hairsine and Rose (1991) developed a physically based model of rain erosion on a hillslope. This model addresses the situation in which the shear forces of overland flow are insignificant and surface runoff merely transports sediment entrained into the flow by the energy of falling raindrops. The Hairsine and Rose model provides a basis for

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understanding the interaction of rainfall detachment and deposition of cohesive soils composed of a range of particle and aggregate sizes and densities. Central and unique to the Hairsine and Rose model is the development of a deposited layer over some fraction of the original soil surface. This model has been fully described in many articles (Hairsine and Rose, 1991; Rose *et al*., 1994; Misra and Rose, 1996; Sander *et al*., 1996; Hairsine *et al*., 1999; Parlange *et al*., 1999). Hairsine and Rose (1991) presented the following equation for calculating of sediment concentration at equilibrium for water depths (D) of less than the critical water depth (D_0):

$$c = \frac{a\, a_d P}{a_d Q + a\phi_e} \quad (1$$

where c is the equilibrium sediment concentration due to rainfall related processes (kg m^{-3}), α and α_d are detachability parameters for the soil and deposited layer (kg m^{-3}), respectively, P is the rainfall rate (m s^{-1}), Q is the runoff rate per unit area (m^3 m^{-2} s^{-1}), and ϕ_e is the effective depositability (m s^{-1}) equal to the mean settling velocity of soil ($\approx \frac{1}{I}\sum_{i=1}^{I} v_i$ in which I is the arbitrary number of settling velocity classes of original soil with equal mass in each class and v_i is the settling velocity of *i*th class).

Hairsine and Rose (1991) assumed that the value of α and α_d varies from their maximum value, α_0 and α_{d0} obtained with a shallow critical water depth, D_0, and will decrease as water depth D increases:

$$\alpha=\alpha_0 \text{ and } \alpha_d=\alpha_{d0} \quad \text{for } D \le D_0 \quad (2$$

$$\frac{a}{a_0} = \frac{a_d}{a_{d0}} = \left(\frac{D_0}{D}\right)^b \quad \text{for } D > D_0 \quad (3$$

where *b* is a positive exponent. Proffitt *et al*. (1991) experimentally found *b* to be 0.66 for a shallow flow with depths 3 and 10 mm.

The fraction of the surface covered by the deposited layer is shown by H. Observation made in the flume experiments by Proffitt *et al*. (1991) for constant rainfall rates indicated that $H \approx 0.9$. Values of H even closer to unity are supported by fitting a time-varying solution to the same data (Sander *et al*., 1996). Using $H \approx 0.9$, Misra and Rose (1996) derived the following equations for calculating α and α_d:

$$a_d = \frac{c\phi_e}{0.9P} \quad (4$$

$$a = Q\left(\frac{P}{c} - \frac{\phi_e}{a_d}\right)^{-1} \quad (5$$

Proffitt *et al*. (1991) performed a set of laboratory experiments with low slopes (0.1-1%) and significant water depths (2, 5 and 10 mm) that demonstrated the model's conceptual basis and its ability to predict sediment delivery at the bottom of a hillslope. Heilig *et al*. (2001) used a simple experiment to verify visually and analytically the conceptual basis of the Hairsine and Rose model with special attention paid to the development of the deposited layer. Their experiments visually and quantitatively displayed the formation of a shield during rain impact erosion. They also obtained good quantitative agreement between the observations and the model predictions. The effect of ponding water depth on soil detachability for the Hairsine and Rose model was investigated by Gao *et al*. (2003). Their experimentally observed relationship between ponding water depth and soil detachability agreed well with the proposed theories (equations [2] and [3]). Soil detachability was constant for ponding water depths below a critical depth (10 mm) and dramatically decreased above the critical depth (3).

In soil erosion studies, the results of experiments can be influenced by many factors include rainfall simulator type, soil preparation methods, and rainstorm characteristics (Bryan and De Ploey, 1983; Agassi and Bradford, 1999). Since the Hairsine and Rose model has only been tested against a limited amount of experimental data (Proffitt *et al*., 1991; Heilig *et al*., 2001; Gao *et al*., 2003), the objective of our research was to perform additional experiments using a different rainfall simulator, three contrasting soil types and a different experimental set up to test further the underlying physical principles of the model and its ability for predicting soil loss.

MATERIALS AND METHODS

The laboratory experiments were carried out to test the Hairsine and Rose (1991) model for sediment transport in the absence of flow driven processes. Three contrasting soil types were subjected to simulated rainfall at different rates using a drainable detachment tray.

Soil Types

The soils used in this study were sampled from these different basins: a calcareous Inceptisols from the Research Station at Quin; an Alfisols from the rainforests of Guilan Province ('Forest soil'); and a dispersive Entisols from the sandy marl hills of Eshtehard ('Sandy soil').

Soil samples were collected from the upper 20 cm of the soil profile. Primary and secondary particle size distributions of the soils were measured using the hydrometer method and a wet sieving machine, respectively. Soil chemical properties such as pH, EC, organic matter, and equivalent calcium carbonate were determined using the standard methods. Some of the physical and chemical properties of the soils are given in Table 1. Particle size distributions of the soils are also presented in Figure 1.

The Quin and Forest soils had the same texture (clay loam) but a contrasting aggregate size distribution (Table 1 and Figure 1). Quin soil was a calcareous clay loam with a loose aggregate stability represented by a mean weight diameter (MWD) of 0.53 mm. In contrast, the Forest soil was a clay loam soil with very stable aggregates and MWD of 2.04 mm. The third soil was a slightly dispersive sandy soil (sandy loam) without

Table 1. Some physical and chemical properties of the soils.

Soil properties	Sand (%)	Silt (%)	Clay (%)	MWD (mm)	SP (%)	OM (%)	$CaCO_3$ (%)	pH	EC (dS m^{-1})
Quin soil	31	31.5	37.5	0.53	54	0.95	18	7.9	0.5
Forest soil	35.5	29	35.5	2.04	80	14	2	7.7	0.9
Sandy soil	78	12	10	0.205	28	0.09	12	8.0	2.6

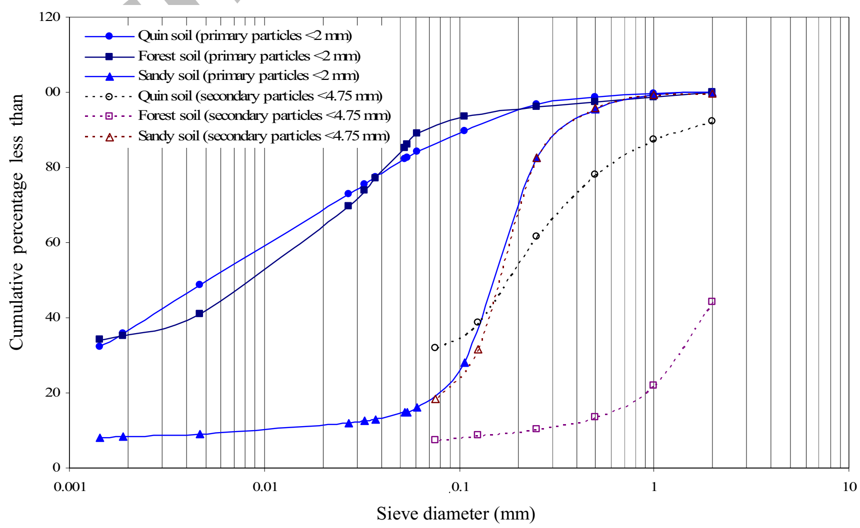

Figure 1. Primary and secondary particle size distribution of the soils.

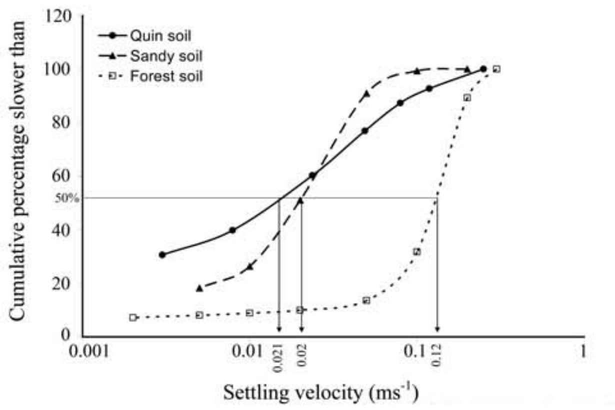

Figure 2. Mean settling velocity distribution of the soils.

any defined stable aggregates. Mean settling velocity distributions of the original soils were calculated by GUDPRO 3.1 (Lisle *et al*., 1996) using the wet sieving data. Figure 2 shows the mean settling velocity distribution of the soils investigated. The effective depositability (ϕ_e) is also calculated by GUDPRO 3.1 for an appropriate (measured) water depth for each experiment.

Experimental Tray and Sample Preparation

The detachment tray used in the experiments was same as the tray described by Misra and Rose (1989) with certain modifications. The tray is a 35×30×10 cm drainable tray and consists of three parts. On two sides of the detachment tray, a splash guard is provided so that soil is not only lost by splash from the central area of the tray from which sediment is collected, but it can also be returned. A photo of the tray is given in Figure 3.

Soil samples were air dried and sieved using a 4.75 mm sieve. Soil was packed into the tray and raked to produce a soil bed of uniform depth with a surface as level as possible for Quin and Sandy soils. In order to reduce the high infiltrability of the soil and to generate runoff, experiments were carried out on the Forest with a compacted sub layer. Soil was spread uniformly in the tray to a depth of 6 cm and, after wetting, compacted using a wooden plate and a weight (about 5 kg). An additional 2 cm layer of soil was then spread on the top of the compacted layer and leveled.

Soil was saturated from the bottom using a

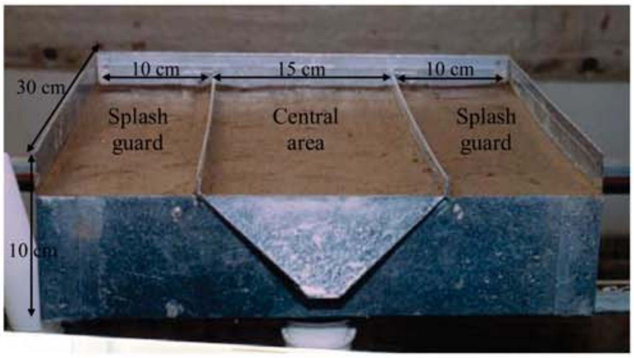

Figure 3. A photo of the detachment tray used in the study.

drainage outlet tube connected to a water reservoir during the night. Gravimetric water was allowed to move out of the soil before each experiment, and the drainage outlet remained open during the experiment. The tray was set up at a slope of 0.5 percent.

Rainfall Simulation and Measurements

Rainfall was applied using a rainfall simulator with a single scanning nozzle that emitted drops with a mean diameter (volumetric D_{50}) of 1.5 mm as measured by the flour pellet method. Simulated rain was applied to the detachment tray from the scanning nozzle located 4 m above the soil bed in the tray with negligible wind effects. The rainfall simulator works with a constant pressure of 0.04 MPa and the rainfall rate increases or decreases by changing the frequency of spray oscillation. Overall, 40 experiments were carried out for three soil types at different rainfall rates (25-110 mm h^{-1}); 17, 14, and 9 experiments for the Quin, Forest and Sandy soils respectively. Each run was carried out using a fresh soil sample. Rainfall duration was 30-40 minutes, depending on the soil type and rainfall rate for achieving a steady state.

Accumulated runoff from the detachment tray was collected at different time intervals. At the end of the experiments, the volume of runoff (V, m^3) and mass of sediment (M_d, kg) were determined after drying the samples at 105 °C for 24 h. Sediment concentrations of runoff samples (c, kg m^{-3}) were estimated as M_d/V. The runoff rate per unit area (Q, m^{-3} m^{-2} s^{-1}) was also calculated for each interval. A rain gauge was placed adjacent to the detachment tray during each experiment for the measurement of total rainfall volume. Average rainfall rate (P, mm h^{-1}) was estimated from measurements of the volume and duration of rainfall at the end of each experiment. Water depth was measured at the outlet when a steady state condition was achieved. Steady state is defined as a condition in which sediment concentration and runoff rate do not change significantly with time.

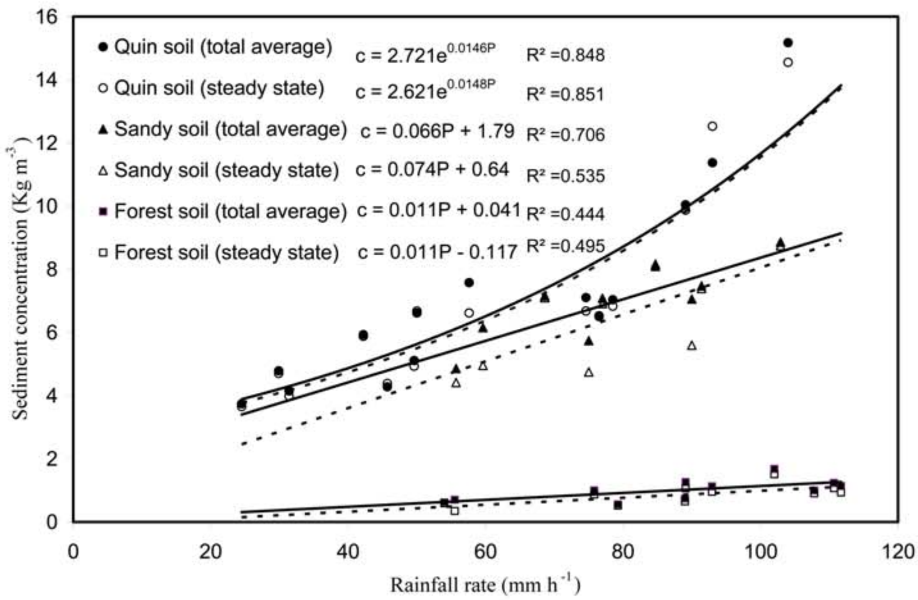

Figure 4. Changes with rainfall rate in sediment concentration at the steady state (—) and time-average sediment concentration of the event (- - -) for three soil types.

RESULTS AND DISCUSION

Sediment Loss

Changes in the steady state sediment concentration and time-average sediment concentration with the rainfall rate for the all three soil types are presented in Figure 4. Sediment concentration increased as the rainfall rate increased for all soils. The rate of increase was faster for the Quin and Sandy soils than for the Forest soil. Sediment concentration was similar for the Sandy and Quin soils at a rainfall rate of less than 60 mm h^{-1}, but increased exponentially for the Quin soil at rates higher than 60 mm h^{-1}. At 100 mm h^{-1} rainfall, sediment concentration for the Quin soil was 1.7 times than the Sandy soil.

Sediment loss was very much lower for the Forest soil than for the other two soils and increased linearly with the rainfall rate. For the Sandy soil, sediment concentration was also increased linearly with the rainfall rate but 6 times faster than for the Forest soil. In contrast, soil loss increased exponentially at rainfall rates higher than 60 mm h^{-1} for the Quin soil. The reason for these differences in behaviour is due to the different characteristics of the soils used in the experiments. The soil referred to as the Sandy soil was a very loose sandy soil with 78 per cent sand particles and no aggregation. The mean weight diameter (MWD) of particles for this soil was 0.205 mm (Table 1), and 50 percent of the particles lie within the size range of 0.125-0.250 mm (Figure 1). The other two soils had a similar texture (Table 1 and Figure 1) but with a different degree of aggregation and aggregate size. The Forest soil was very well aggregated and contained 14 percent of organic mater and MWD of 2.04 mm, in which 56 percent of aggregates were greater than 2 mm and only 7.5 percent of the particles were less than 0.075 mm. The Quin soil was a calcareous soil with 18 percent equivalent calcium carbonate and MWD of 0.53 mm. Sediment concentration at the steady state was slightly lower than the time-average sediment concentration for all three soils, with no significant differences.

Detachability and Re-detachability of the Soils

Detachability (α) and re-detachability (α_d) of the soils were calculated as described in equations [4] and [5]. The measured P, Q and c were used for these calculations. The effective depositability (ϕ_e) for each ex-

Table 2. Detachability and re-detachability of soils and their statistics calculated using (A) steady state sediment concentrations and runoff rates, (B) time-average sediment concentrations and runoff rates.

Parameters		Quin soil		Sandy soil		Forest soil	
Number of runs		17		9		14	
Rainfall rate (mm h^{-1})		25-104		56-103		54-112	
		A	B	A	B	A	B
Soil detachability; (α, kg m^{-3})	Min.	14	9.5	33	32	0.4	0.7
	Max.	107	98	70	66	10	11
	Mean	48	43	49	48.5	5.6	6.1
	SD	28	23	11.8	9.5	3.03	3.42
Soil re-detachability; (α_d, kg m^{-3})	Min.	2150	2930	5770	6980	156	313
	Max.	13670	14000	9405	9485	2622	2844
	Mean	7456	7917	7590	8220	1176	1117
	SD	3206	3114	1222	888	833	679

periment was predicted using the GUDPRO 3.1 program (Lisle *et al*., 1996) based on the measured water depth at the steady state condition. Table 2 summarizes the results of these calculations. Mean detachability (α) of the Quin, Sandy and Forest soils were 48, 49 and 5.6 kg m^{-3} and the mean re-detachability (α_d) of the soils were 7,456, 7,590 and 1,176 kg m^{-3}, respectively.

Hairsine and Rose model (1991) clearly described the sensitivity of soils to erosion, introducing special terms such as detachability, re-detachability and effective depositability. α and α_d are significantly higher for the Quin and Sandy soils than for the Forest soil. The differences in α as well as α_d for the soils investigated in conjunction with effective depositability (or, in other words, settling velocity characteristics of the soils) agree well with changes in sediment concentration as shown in Figure 4. The Forest soil has smaller α and α_d and a higher effective depositability compared with the other two soils, resulting in a lower sediment concentration as shown in Figure 4. The detachability parameters (α and α_d) are slightly higher for the Sandy soil than for the Quin soil (Table 2), but its sediment concentration (Figure 4) is slightly lower than the sediment concentration of the Quin soil (in the low rainfall rates). This can be interpreted considering the effective depositability (ϕ_e) of these two soils (Table 3), which is higher for the sandy soil than for the Quin soil at the low water depths (low rainfall rates).

As shown in Figure 4, sediment concentration increased exponentially for the Quin soil as rainfall rate increased beyond 60 mm h^{-1} and compared to the other two soils. This can not be interpreted using α, α_d and ϕ_e terminology since Quin soil has a lower α and α_d and higher ϕ_e at higher water depths or higher rainfall rates resulting in lower sediment concentration. The reason for this behaviour is probably related to aggregate breakdown at the higher rainfall rates. Mean settling velocity distribution and effective depositability are calculated using the particle size distribution of the original wetted soil in which aggregate breakdown is not taken into account.

The detachability and re-detachability of soils were also calculated using the same method but using the time-averages of sediment concentration and runoff rate. There was no significant difference between α and α_d of the three soils when they were calculated based on steady state concentration or time-averaged sediment concentration (Table 2).

Changes with the measured water depth in the detachability and re-detachability of the Quin, Sandy and Forest soils are given in Figures 5, 6, and 7, respectively. For the Quin soil (Figure 5), detachability (α) and re-detachability (α_d) increased with water depth, but remained relatively constant for the Sandy soil (Figure 6). Changes with water depth in detachability parameters (α and α_d) of the Forest soil led to a different interpretation: α and α_d increased with water depth up to a critical water depth and remained almost constant for water depths grater than the critical depth (Figure 7). This critical water depth was higher for re-detachability, α_d (1.75 mm) than for detachability, α (0.75mm). It seems that the relationship between detachability parameters (α and α_d) and water depth and thus soil loss by raindrop impact is soil type dependent. Variation in α and α_d as shown in Figures 5, 6, and 7, is, at least in part, due to the sensitivity of these parameters to the value of H as shown by Proffitt *et al*. (1991). The greater variation of detachability than re-detachability (Figures 5-7) is because of

Table3. Effective depositability [a] (ϕ_e, ms^{-1}) of the soils for three selected water depths.

Soil type	Selected water depth (mm)		
	1	2	3
Quin Soil	0.0170	0.0220	0.0249
Sandy soil	0.0224	0.0227	0.0230
Forest soil	0.0317	0.0675	0.0855

[a] Calculated by GUDPRO 3.1 software.

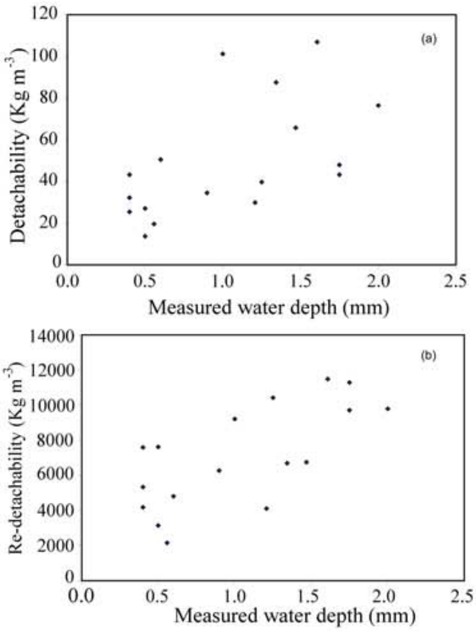

Figure 5. Changes with water depth in detachability (a) and re-detachability (b) of the Quin soil.

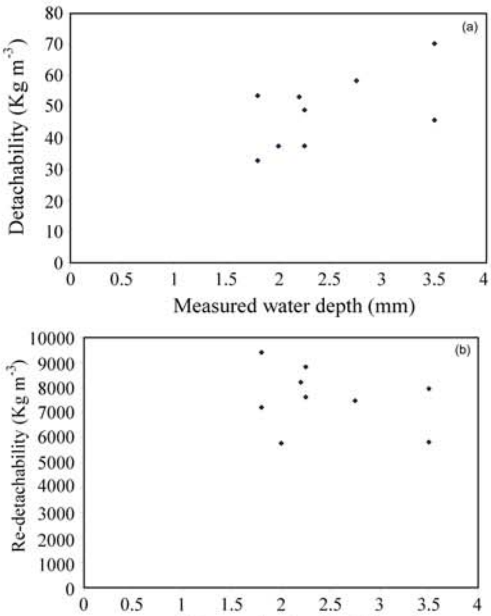

Figure 6. Changes with water depth in detachability (a) and re-detachability (b) of the Sandy soil.

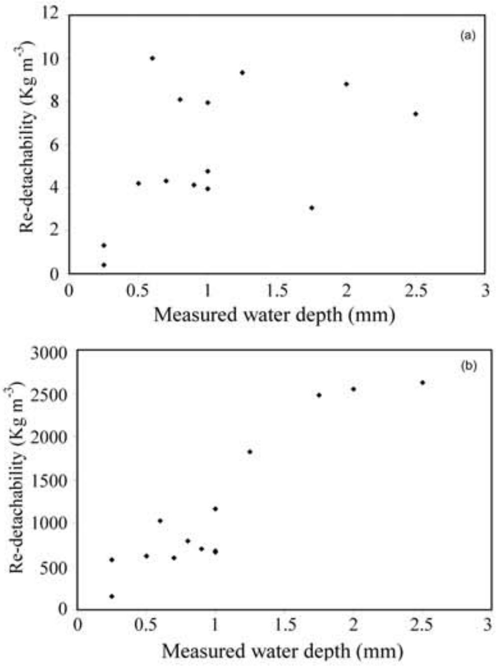

Figure 7. Changes with water depth in detachability (a) and re-detachability (b) of the Forest soil.

the greater sensitivity to H for detachability than for re-detachability, especially around H=0.9 which was used for the calculation of α and α_d in this study. The variation of α and α_d may also, in part, be due to uncertainty in measuring water depth especially in low water depths.

Hairsine and Rose (1991) assumed that α and α_d are constant when the water depth (D) is below a critical or breakdown depth (D_0), and that they are reduced for $D>D_0$, Equations [2] and [3]. Proffitt *et al*. (1991) examined three water depths (2, 5, and 10 mm) for two soil types and showed that both α and α_d reduce with an increasing depth of water on the soil surface, indicating the critical water depth is equal or less than 2 mm. The study of Gao *et al*. (2003) using a man-made soil (hydrous Kaolin clay particles) also showed that soil detachability is constant for ponding water depths below a critical depth (10 mm) and dramatically decreases above the critical depth. Moss and Green (1983) showed that the rain-flow transportation rates increased with water depth until a water depth with a 2 to 3 drop diameter and reduced after this critical water depth for drops of 0.81 and 1.27 mm diameter. For raindrops of 2.7 and 5.1 mm diameter, they showed that rain-flow transportation rates are constant until the critical water depth of 2-3 drop diameter and then reduce with water depth. In our study, all water depths were less than 3 drop diameter in which the drop diameter was 1.5 mm. Thus all water depths investigated appear to be less than the critical water depth.

Evaluation of the Model

A jack knifing method (Shao and Tu, 1995) was used for evaluation of the Hairsine and Rose model. In this method, for obtaining each data point (predicted vs. measured value), measured values of sediment concentration were drawn out one by one from the data set for each soil type. Detachability parameters (α and α_d) for a given experiment were then calculated using the other measured values in two different ways. For the first, water depth was assumed to be less than the critical (D_0) and α and α_d are constant for all water depths, as proposed by the model. Therefore, the detachability parameters (α and α_d) of other experiments were averaged and considered as for the given experiment. For the second, first the relationship between detachability parameters (α and α_d) and water depth as described earlier was drawn. Then the related α and α_d were selected from the regression curve for the measured appropriate water depth. Finally sediment concentration for the given experiment was predicted using the average α and α_d of other experiments (way I) and selected α and α_d from the regression curve (way II).

Evaluation was done separately for each soil type. The regression lines between predicted (P_c) vs. measured (M_c) sediment concentration for all three soil types and the two different ways used are given in Table 4. In this Table, m is the regression slope, n is the y-intercept value, R^2 is the coefficient of determination of the regression line and MSE is the mean square error. For all regression lines, the regression slope is less than one and the intercept (n) is positive which shows that the model tends to over-predict the lower values of sediment concentration, and under-predict the higher values of sediment concentration, though the average sediment concentration predicted for the entire data set is not greatly different from the average measured values. Results from the testing of USLE (Riss *et al*., 1993), Revised USLE (Rapp, 1994) and WEPP (Zhang *et al*., 1996) have produced similar results. Nearing (1998) has concluded that a limitation of the current erosion models is

Table 4. Regression parameters, average measured sediment concentration (M_c), and average predicted sediment concentration (P_c) for the three soils and two ways of evaluation.

Regression parameter	Quin soil Way I	Quin soil Way II	Sandy soil Way I	Sandy soil Way II	Forest soil Way I	Forest soil Way II
m^a	0.24	0.54	0.45	0.50	0.93	0.47
n^b (kg m^{-3})	5.2	2.9	3.5	3.2	0.45	0.42
$R^{2\,c}$	0.13	0.53	0.46	0.45	0.13	0.44
MSE^d	9.16	4.48	1.19	1.22	0.99	0.09
$E_{Nush}{}^f$	0.04	0.53	0.46	0.44	-5.65	0.42
Average P_c	6.68	6.37	6.44	6.44	1.27	0.84
Average M_c	1.34		6.45		0.88	

[a] The regression slope.
[b] The y-intercept value.
[c] The coefficient of determination of the regression line.
[d] Mean square error.
[f] Nush--Sutcliffe efficiency coefficient.

their deterministic nature since the model is not capable of capturing the natural variation in the measured value, whereas his results do not suggest that this factor is necessarily the only one at work to create this bias.

Table 4 also provides the values for Nash-Sutcliffe efficiency coefficients, E_{Nush} (Nash and Sutcliffe, 1970). For the given number of model output/observation comparisons, E_{Nush} equal to 1 indicates perfect agreement between model and observation, and $E_{Nush} \geq 0.6$ is commonly regarded as acceptable for flow simulation models (Chiew and McMahon, 1993), which was not achieved for any case in this study.

Considering the relationship between the detachability parameters and water depth, the second way (way II) improved prediction for all three soils especially for the Quin and the Forest soils (Table 4). The average predicted sediment concentration was also improved for the Quin and Forest soils by using the second way. All results of evaluation are also presented in Figure 8.

CONCLUSION

It was concluded that secondary particle/aggregate size distribution in soil plays a very important role in soil erosion processes. The Quin and Forest soils present approximately the same texture (primary particle size distribution) but with contrasting aggregate size and stability. As indicated, the overall sediment lost from the Quin soil was seven times more than from the Forest soil, showing the greater significance of aggregate size distribution as a soil erodibility index than the primary particle size distribution.

Despite the model validation by Proffitt *et al*. (1991) and Gao *et al*. (2003), changes with water depth in detachability parameters (α and αd) indicated that the relationship between detachability parameters and water depth is soil type dependent and did not agree with previously proposed theories. The reasons for the disagreement between our results and previous finding may be related to the fact that Proffitt *et al*. (1991) examined relatively high water depths (2, 5

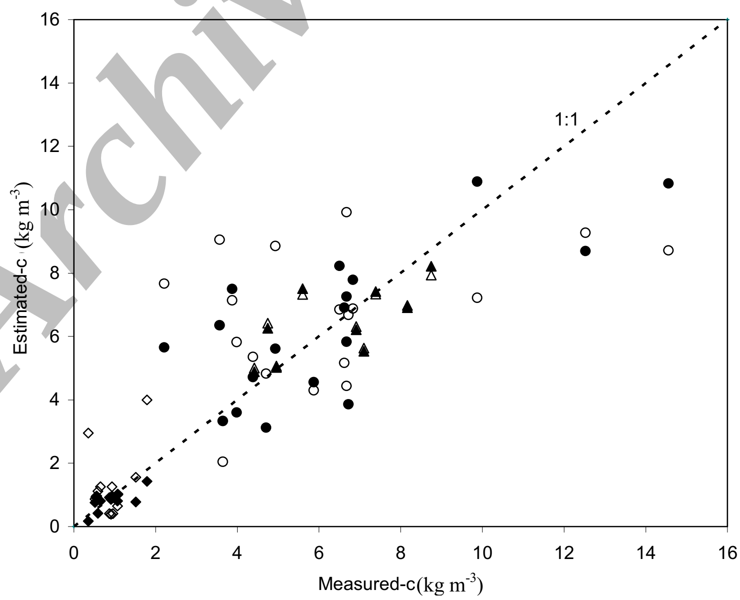

Figure 8. Estimated vs. measured sediment concentration. The square, triangle and circle symbols are for the Forest, Sandy and Quin soils respectively. Empty symbols are the results of Way I, and filled symbols are results of Way II.

and 10 mm) compared with the water depths of less than 3.5 mm used in this study. Gao *et al*. (2003) also used a man-made soil consisting of just one particle size (hydrous Kaolin) in which settling velocity is almost zero and therefore deposition and redetachment play negligible roles. Splash creep by raindrop impact reported by Terry (1998) seems to have an important role in sediment transport and affects the relationship between water depth and sediment transport by raindrop impact which is not considered by the Harisine and Rose model as well as other models. This mechanism which was observed in this study needs to receive more attention and further investigation.

There was no significant difference between detachability parameters (α and α_d) for the three soils when they were calculated based on sediment delivery at steady state concentration or time-averaged sediment concentration for the whole duration of the experiment. Therefore, time-average sediment concentration could be used for calculating detachability parameters in the model where the steady state values are not available, even though the Hairsine and Rose model was developed for steady state condition.

This study also indicated that the Hairsine and Rose model tends to over-predict the lower values of sediment concentration, and under-predict the higher values of sediment concentration. Sensitivity of the detachability parameters to value of H as shown by Proffitt *et al*. (1991) and the lack of a practical method for determining the actual value of H could be the reasons for this weakness. Uncertainty in measuring water depth, especially at low water depths, may be another reason for the poor prediction of the model.

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ارزیابی یک مدل فرآیندی فرسایش خاک برای سه خاک انتخابی از ایران

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چکیده

هیرساین و رز (۱۹۹۱) برای فرسایش ناشی از اثر بارندگی مدلی فرآیندی ارائه داده‌اند. مدل آنها برای شرایطی ایجاد شده که نیروی برشی جریان سطحی ناچیز بوده و رواناب تنها نقش انتقال ذرات جدا شده توسط برخورد قطرات باران را برعهده دارد. در تحقیق حاضر به منظور بررسی برخی مفاهیم و کارآیی این مدل، سه نوع خاک با خصوصیات کاملاً متفاوت با استفاده از یک سینی پاشمان ۱۰×۳۰×۳۵ سانتی‌متری زهکش‌دار تحت اثر بارندگی با شدت‌های مختلف (۱۰۰–۲۵ میلی‌متر در ساعت) قرار گرفت. شبیه‌سازی

باران با استفاده از یک دستگاه شبیه‌ساز با نازل منفرد جارویی انجام شد که قطر قطرات آن ۱/۵ میلی‌متر و انرژی جنبشی آن مشابه باران‌های طبیعی است. نتایج نشان داد که مدل هیرساین و رز با معرفی ضرایب جدایش‌پذیری α، جدایش‌پذیری مجدد α_d و ترسیب پذیری ϕ به خوبی می‌تواند حساسیت خاک‌های مختلف را به فرسایش توصیف نماید. هرچند که این مدل قادر نیست شکستن خاکدانه‌ها و افزایش نمایی غلظت رسوب را که در مورد یکی از سه خاک مورد بررسی در شدتهای بالای بارندگی رخ داد، توصیف نماید. همچنین بررسی رابطه بین ضرایب جدایش‌پذیری با عمق آب رابطه فرضی مطرح شده در مدل را تأیید نکرد. ارزیابی مدل نیز نشان داد که مدل مورد بررسی تمایل به بیش‌برآورد مقادیر کوچک و کم‌برآورد مقادیر بزرگ دارد، هرچند که میانگین غلظت رسوب برآوردی تطابق نسبتاً خوبی با میانگین غلظت رسوب اندازه‌گیری شده برای مجموعه داده‌های هر خاک داشت.