doi:10.2489/jswc.70.1.36

# Sediment transport capacity and its response to hydraulic parameters in experimental rill flow on steep slope

Z. Wang, X. Yang, J. Liu, and Y. Yuan

**Abstract:** Sediment transport capacity must be considered when developing physical models of soil erosion. The effects of related hydraulic parameters (e.g., flow discharge, slope gradient, and flow velocity), and of force predictors (e.g., shear stress, stream power, and unit stream power) on sediment transport capacity in rill erosion are still poorly known on the farmland of the Loess Plateau in China where rill erosion is common. We conducted a series of experiments to simulate and evaluate the sediment transport capacity of rill flow in a nonerodible rill flume. The test sediment was the loessial soil of the farmland of the Loess Plateau. Five flow discharges ranging from 0.22 to 0.67 × $10^{-3}$ $m^2$ $s^{-1}$ (0.00237 to 0.00721 $ft^2$ $sec^{-1}$) and five slope gradients ranging from 15.8% to 38.4% were tested. Sediment transport capacity increased with both flow discharge and slope gradient, as expected, but was more sensitive to flow discharge than to slope gradient, unlike other similar studies. Mean flow velocity, related to the flow discharge, was strongly correlated with sediment transport capacity ($r^2$ = 0.93). Stream power was the best predictor of sediment transport capacity; shear stress and unit stream power, with critical values of 0.55 W $m^{-2}$ and 0.02 m $s^{-1}$ (0.04 mi $hr^{-1}$) respectively, were poor predictors. An empirical equation of sediment transport capacity of the loessial soil for rill flow was developed. Our results present a different view, compared to previous studies, of the relationship of sediment transport capacity with discharge and slopes, especially with lower discharges, steep slopes, and loessial soil. Further study should be conducted to evaluate the performance of farmland soil at various slopes and discharges.

**Key words:** hydraulic parameters—loessial soil—rill flow—sediment transport capacity—water erosion

**The transport of sediments by runoff can potentially alter river courses and navigability through siltation; it can also contaminate ecosystems with the chemicals carried by the sediments themselves (Lal 1998).** To combat and predict soil erosion and its related problems, several physically based models of soil erosion have been developed, including the Limburg soil erosion model (De Roo et al. 1996), the European soil erosion model (Morgan et al. 1998), and the water erosion prediction project (WEPP) (Flanagan et al. 2001). The efficiencies of the models, however, can be limited by a range of problems that include insufficient evaluation, overparameterization, or the use of parameters that are inappropriate for local conditions, resulting in poor performance (Merritt et al. 2003). A priority for describing soil erosion is to develop a processing-based erosion model that can calculate the rate of sediment detachment and the transport capacity. Foster and Meyer (1972) documented that the rate of sediment detachment could be estimated separately from transport capacity and sediment load. Nearing et al. (1989) suggested that sediment transport capacity could be distinguished from soil-detachment capacity without any deposition. Models of sediment transport capacity thus assumed that the maximum equilibrium sediment could be transported under certain conditions of discharge. Many studies have validated and calibrated these earlier formulations or have developed empirical formulations based on limited laboratory or field investigation of discharges, slopes (gradient and length), grain diameters, and some hydraulic parameters (Beasley and Huggins 1982; Govers and Rauws 1986; Finkner et al. 1989; Govers 1990, 1992; Ferro 1998; Zhang et al. 2009, 2010a, 2010b; Gokmen and Vijay 2012; Ali et al. 2012a, 2012b, 2013).

Beasley and Huggins (1982) reported that sediment transport capacity could be calculated by the equations:

$$T_c = 146\ Sq^{0.5}\ (q \le 0.046), \tag{1}$$

$$T_c = 146\ Sq^{2}\ (q > 0.046), \tag{2}$$

where $T_c$ is the sediment transport capacity (kg $m^{-1}$ $s^{-1}$), $S$ is the slope gradient (m $m^{-1}$), and $q$ is the unit discharge per unit width of slope ($m^2$ $s^{-1}$). These two equations were subsequently used in the Areal Nonpoint Source Watershed Environment Response Simulation (ANSWERS) model. Discharge was the only limiting factor considered for sediment transport even though other models emphasized the need to consider slope gradient and/or mean flow velocity.

However, Julien and Simons (1985) found that the relationship of sediment transport capacity could be expressed as a power function of slope and discharge, with the exponent derived from the actual condition of its use. Based on a series of laboratory experiments with a hydraulic flume, Govers (1990) developed a general equation that described the relationship between sediment transport capacity and both discharge and slope:

$$T_c = \mathrm{A}q^{\mathrm{B}}\ S^{\mathrm{C}}, \tag{3}$$

**Zhanli Wang** is a professor in State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau with the Institute of Soil and Water Conservation at Northwest A&F University in Yangling, China; and the Institute of Soil and Water Conservation at the Chinese Academy of Sciences and Ministry of Water Resources in Yangling, China. **Xiaomei Yang** is a PhD student in the State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau with the Institute of Soil and Water Conservation at the Chinese Academy of Sciences and Ministry of Water Resources in Yangling, Shaanxi, China; and the Soil Physics and Land Management at Wagningen University in Wagningen, Netherlands. **Jun'e Liu** is a PhD student in School of Resources and Environment at Northwest A&F University in Yangling, Shaanxi, China. **Yin Yuan** is a master student in School of Resources and Environment at Northwest A&F University in Yangling, Shaanxi, China.

where A, B, and C are coefficients associated with grain size and with laminar and turbulent-flow regimes. The relationship between sediment transport capacity and discharge and/or slope has been studied, but the dominant contribution to capacity by either discharge or slope can vary. Govers (1990) and Everaert (1991) reported that the effects of slope on transport capacity was greater than that of discharge, while Zhang et al. (2009) concluded that discharge contributed more than did slope gradient. Several studies have been conducted with hydraulic flumes with either nonerodible or erodible beds. The surface roughness of nonerodible beds is substantially different from that of erodible beds (Hu and Abrahams 2006). Using nonerodible beds and sand of variable coarseness, Ali et al. (2012a) developed an empirical equation with exponents and showed that sediment transport capacity was more sensitive to slope than to discharge, where the derived exponents were 2.89 and 1.46, respectively. These different results could all be explained by unit discharge, slope gradient, and test material (diameter and texture), which in various combinations can lead to a large variety of experimental conditions.

Transport capacity has close relationships with several hydraulic parameters other than discharge, slope, and grain diameter. Flow velocity, the only parameter detected directly (Zhang et al. 2010a), is influenced by discharge, slope, and the surface roughness of the flow bed, which are all related to the rate of sediment detachment and transport capacity. Zhang et al. (2009) reported that the relationship between mean flow velocity and sediment transport capacity could be described by a linear function, while many other studies have shown that mean flow velocity was independent of sediment transport capacity (Nearing et al. 1999; Gimenez and Govers 2001).

Another important hydraulic parameter, shear stress, was calculated by the formula of Yalin (1963):

$$\tau = \rho g h S, \tag{4}$$

where $\tau$ is the shear stress (Pa), $\rho$ is the water mass density (kg m$^{-3}$), $g$ is the gravity constant (m s$^{-2}$), and $h$ is the depth of flow (m). This equation was used in WEPP to estimate sediment transport capacity (Nearing et al. 1989), but its reliability remains in doubt (Julien and Simons 1985).

Bagnold (1966) emphasized energy expenditure, and expressed $T_c$ as a function of stream power:

$$\omega = \tau v = \rho g S q, \tag{5}$$

where $\omega$ is the stream power (W m$^{-2}$) and $v$ is the mean flow velocity (m s$^{-1}$). Yang (1972) later used noncohesive sands to develop a load equation with unit stream power:

$$P = vS, \tag{6}$$

where $P$ is the unit stream power (m s$^{-1}$).

The relationship and form of formulation, however, varied between parameters and sediment transport capacity, some of which were even contradictory (Yalin 1963; Bagnold 1966; Yang 1972; Foster 1984; Moore and Burch 1986; Govers and Rauws 1986; Govers 1990; Nearing 1997; Zhang et al. 2009, 2010b; Ali et al. 2012a, 2013). Adapting suitable hydraulic parameters is thus essential either to calculate sediment transport capacity or to evaluate the response relationships among them.

As the most active type of erosion on disturbed uplands, rill erosion is quite different in its characteristics and in its conditions of hydraulic environments compared to typical channels of streams and rivers (Nearing 1989). Rills have small and ephemerally concentrated flow paths and have been considered to be the main source and means of sediment transport from hillslopes (Gilley et al. 1990; Nearing et al. 1997). As determined by long-term monitoring in field plots, rill erosion is considered the main type of soil erosion in the farmland on the Loess Plateau (Zhu 1982; Cai 1998). Lei et al. (2001) developed a reasonable method of understanding sediment transport and its relationships with rill length, discharge, and slope:

$$T_c = -0.03109 + 0.01718S + 0.12703q. \tag{7}$$

Using the maximum sediment load, which is equivalent to sediment transport capacity, Zhang et al. (2009) reported a dual power function of sediment transport capacity:

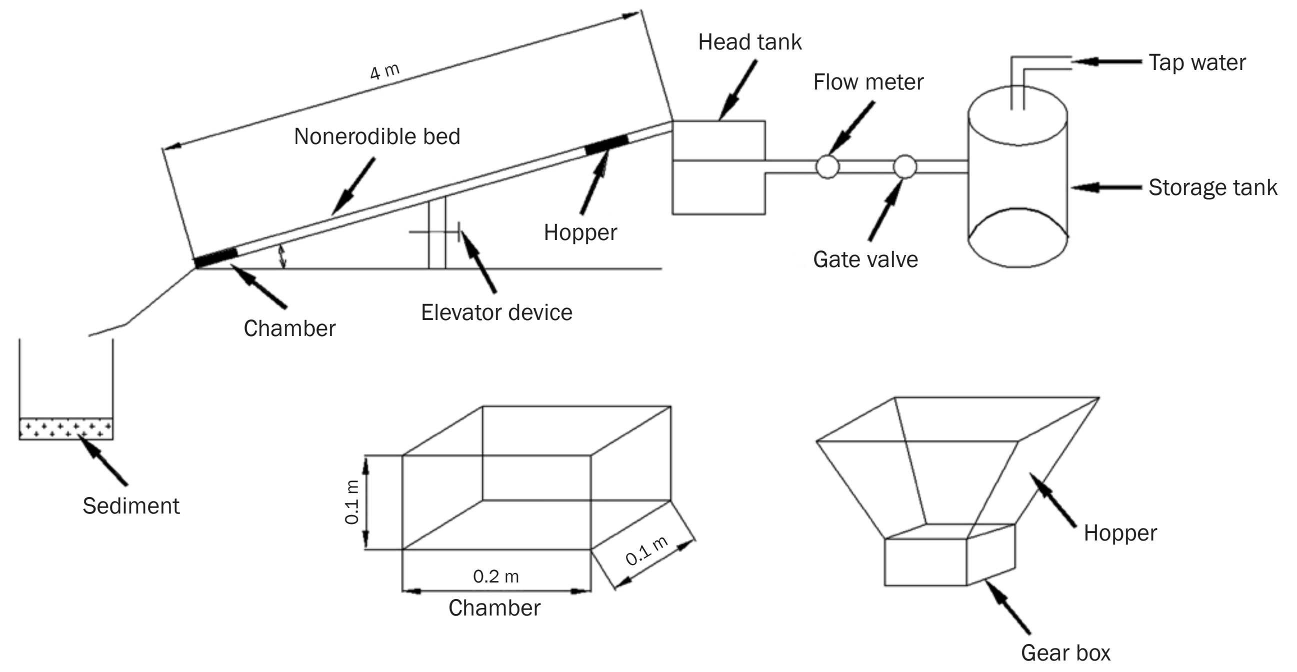
$$T_c = 19{,}831\ S^{1.227}\ q^{1.237}. \tag{8}$$

This function was developed according to 64 combinations of experiments under steep slopes, larger discharges, and riverbed sediments. The grains of loessial soil, though, are obviously different from uniform sand, which is a drawback for using equation 8 to predict sediment transport of rill erosion in the farmland of the Loess Plateau where the integrated interaction of meteorology, topography, land use, and anthropogenic activities affect the erosion processes seriously. Moreover, several functions cannot adequately predict transport capacities, particularly at low flow rates (Ali et al. 2013). According to field studies, flow discharge from erosion-producing rains varies from 0.000 007 m$^2$ s$^{-1}$ to 0.01 m$^2$ s$^{-1}$ (0.00008 ft$^2$ sec$^{-1}$ to 0.11 ft$^2$ sec$^{-1}$). Furthermore, the threshold of slopes on farmland was 46.6% according to the Grain for Green Project. Based on field observation on efficiency of combating soil loss and land degradation, this threshold should be 36.4% demonstrated by Tang et al. (1998). We thus focused on steep slopes, which are typical on the Loess Plateau, in combination with relatively low discharge rates and with loessial soil as the test sediment. The objectives were (1) to determine how the sediment transport capacity of rill flow changed under different conditions of steep slopes and lower discharges and (2) to determine the response of sediment transport capacity to various hydraulic parameters. Most importantly, sediment transport capacity was empirically derived with the native loessial soil rather than an artificial test material such as sand grains of uniform diameter.

## Materials and Methods

***Experimental Facilities.*** The experiments were conducted in a hydraulic flume at the State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau in Yangling, China. Loessial soil was collected from the upper 20 cm (7.9 in) of a cultivated field at the Ansai Experimental Station in Shaanxi Province for use as the test material. Soil samples were air dried, crushed to pass through a 2 mm (0.079 in) sieve, and mixed thoroughly. The particle-size distribution of the soil was 8% fine clay (<0.0001 mm [0.0000039 in]), 5.7% coarse clay (0.001 to 0.002 mm [0.000039 to 0.00004 in]), and 86.3% silt (0.002 to 0.05 mm [0.00004 to 0.002 in]). Lei et al. (2001) had demonstrated that sediment transport capacity would not increase beyond certain values of rill length and slope, so we designed a hydraulic steel flume 4 m (13.1 ft) in length and 0.1 m (0.33 in) in width with double-sided PVC

**Figure 1**
**Experimental setup.**




walls. The sediment transport capacity with these dimensions of rill channel was the same as that reported by Lei et al. (2001). Because bed roughness affects flow hydraulics (Gimenez and Govers 2001), the test soil was glued on the surfaces of the flume walls and bed to simulate the soil surface reported by Lei et al. (2001).

Two special designs were added to ensure that the necessary sediment transport capacity was attained. A 0.8 $m^3$ (2.6 $in^3$) hopper was installed vertical over the flume 0.2 m (0.66 in) from the upper end. A wooden board extended from the outlet of the hopper to allow the rolling test soil adding to flow, thereby avoiding the effects of hydrophobicity and trapped air. Also, a chamber 20 cm (7.9 in) in length, 10 cm (3.9 in) in width, and 10 cm (3.9 in) in depth was inserted at the upper end of the flume. The upper part of the chamber was at the same level as the rill bed. Tap water was used for the experiments, which first entered the storage tank and was pumped to the head tank. The rate of flow into the head tank was controlled and measured with a calibrated flow meter at the inlet pipe. The slope of the flume was adjustable by an elevating device from 0% to 60%. The details of the equipment are shown in figure 1.

***Measurements.*** Prior to the experiments, an uncapped 0.4 cm (0.16 in) thick Plexiglass box 19.2 cm (7.6 in) in length, 9.2 cm (3.6 in) in width, and 9.6 cm (3.9 in) in depth with several 0.5 cm (0.2 in) diameter holes in the bottom was filled with test soil. The box was set in shallow water for 24 hours to saturate the soil before being installed in the chamber. Air-dried soil samples were added to the hopper, and the slope and discharge were adjusted to the desired values. Preliminary experiments were conducted without any test soil to determine the conditions which were necessary for obtaining steady flow discharges in different slope gradients (Zhang et al. 2009). The feed rates of the test soil for each combination of flow discharge and slope gradient were determined and then maintained throughout the tests.

Flow depth is difficult to detect during erosion due to the conditions of the flow and the bed surface. In this study, flow depth was thus determined by an electric probe with 0.01 mm (0.00039 in) resolution along the flow section at 2, 32, and 62 cm (0.8, 12.6, and 24.4 in) above the upper end of the chamber, providing instantaneous detection at each site in triplicate. Nine depths were measured for each combination of flow discharge and slope gradient. The average depth was calculated to be the mean flow depth for that combination of flow rate and slope gradient (table 1). Flow velocity was measured using a dye-tracing technique (Lei et al. 1998) and then the detected velocity could be validated based on the flow regime to eliminate the effect of dye-tracer dispersion in flow (Luk and Merz 1992).

In order to ensure rill flow can reach up to transport capacity—maximum sediment concentration (table 2)—two sediment sources were added (Zhang et al. 2009). One was from the hopper and the other is from the box which can be inserted in chamber carefully. The edges of the box and flume bed were sealed with petroleum jelly to prevent leaks. The box was protected by a thin iron sheet before the feed rate of the sediment and the flow discharge stabilized. As described by Zhang et al. (2009), deposition may occur at the point where the test soil enters the rill flow. The deposition was thus slightly stirred with an iron rod under the hopper during the experiments. Six samples were continuously collected for each combination of flow rate and slope gradient, and the sampling time was recorded.

A series of 25 combinations of discharge (0.22, 0.33, 0.44, 0.56, and 0.67 × $10^{-3}$ $m^2$ $s^{-1}$ [0.00237, 0.00355, 0.00473, 0.006, and 0.00721 $ft^2$ $sec^{-1}$]) and flume-bed slope (15.8%, 21.3%, 26.8%, 32.5%, and 38.4%)

**Table 1**
Flow depth for different slope gradients and flow discharges.

| Flow discharge ($10^{-3}$ m$^2$ s$^{-1}$) | Flow depth in different slope gradient (mm) | | | | |
|---|---|---|---|---|---|
| | 15.8% | 21.3% | 26.8% | 32.5% | 38.4% |
| 0.22 | 1.4 | 1.4 | 1.3 | 1.3 | 1.3 |
| 0.33 | 1.5 | 1.4 | 1.4 | 1.3 | 1.3 |
| 0.44 | 1.7 | 1.5 | 1.5 | 1.4 | 1.4 |
| 0.56 | 1.7 | 1.6 | 1.6 | 1.4 | 1.4 |
| 0.67 | 1.7 | 1.6 | 1.6 | 1.6 | 1.5 |

**Table 2**
Sediment concentration of treatments.

| Flow discharge ($10^{-3}$ m$^2$ s$^{-1}$) | Sediment concentration in different slope gradient (g L$^{-1}$) | | | | |
|---|---|---|---|---|---|
| | 15.8% | 21.3% | 26.8% | 32.5% | 38.4% |
| 0.22 | 188.6 | 227.8 | 269.4 | 342.4 | 451.6 |
| 0.33 | 196.3 | 234.0 | 272.9 | 404.4 | 474.6 |
| 0.44 | 207.5 | 251.8 | 280.5 | 410.9 | 489.1 |
| 0.56 | 211.3 | 251.1 | 297.1 | 427.4 | 488.1 |
| 0.67 | 219.0 | 263.2 | 338.1 | 431.9 | 489.1 |

were tested and each combination was repeated once. All samples from each experiment were allowed to settle for 24 hours. The supernatants were discarded, and the wet sediments were oven dried at 105°C (221°F) for 12 hours.

***Statistical Analysis.*** All data were analyzed by SPSS 16.0 *t*-test for comparing the variations among the different experimental combinations and regression analysis for developing equations of sediment transport capacity. The results were also compared to the transport capacities calculated by equation 1 and 8 for the experimental conditions according to equation form and similar experimental conditions. Shear stress, stream power, and unit stream power were calculated by the correspondence equations mentioned above. In addition, the statistic parameters relative error (RE, %), mean relative error (MRE, %), mean absolute relative error (MARE, %), Nash-Sutcliffe efficiency (NSE), and coefficient of determination ($r^2$) were used to evaluate the performance of our model based on empirical data and the performances of equations 1 and 8. The statistical parameters were given by

$$RE = \frac{(O_i - P_i)}{O_i} \times 100\% \,, \tag{9}$$

$$MRE = \frac{1}{N} \sum_{i=1}^{n} \frac{(O_i - P_i)}{O_i} \times 100\% \,, \tag{10}$$

$$MARE = \frac{1}{N} \sum_{i=1}^{n} \left| \frac{(O_i - P_i)}{O_i} \right| \times 100\% \,, \tag{11}$$

$$r^2 = \frac{\left[ \sum_{i}^{n} (O_i - \bar{O})(P_i - \bar{P}) \right]^2}{\sum_{i=1}^{n} (O_i - \bar{O})^2 \sum_{i=1}^{n} (P_i - \bar{P})^2} \text{, and} \tag{12}$$

$$NSE = 1 - \frac{\sum (O_i - P_i)^2}{(O_i - \bar{O})^2} \,, \tag{13}$$

where $O_i$ is an observation value, $\bar{O}$ is the mean observation value, $P_i$ is the predicted value, $\bar{P}$ is the mean predicted value, and $n$ is the number of samples.

## Results and Discussion

***Effect of Flow Discharge and Slope Gradient on Sediment Transport Capacity.*** Sediment transport capacity increased with increases in both flow discharge and slope gradient (figure 2). For the same level of discharge, the increase in transport capacity was largest when the slope gradient increased from 26.8% to 32.5%. Zhang et al. (2009) also noted obvious changes in transport capacity when the slope gradient increased from 26.8% to 36.4%, which was similar to the change in our study. Furthermore, the variations in our study tended to become nearly parallel to the axis of flow discharge when the slope reached 32.5%, which suggests that sediment transport capacity may not increase significantly above this slope. Similarly, Lei et al. (2001) reported that a slope gradient of 36.4% could be considered a critical slope for the detachment of loessial soil. Upon further analysis, we found that power ($r^2 > 0.92$, $p < 0.01$) and exponential ($r^2 > 0.98$, $p < 0.01$) functions could describe the relationships between transport capacity and flow discharge and slope gradient, respectively.

To determine these relationships between transport capacity and flow discharge and slope gradient, we used multivariate, nonlinear regression analysis to develop the equation

$$T_c = 67.68\ S^{0.98} q^{1.20}\ (r^2 = 0.97, \text{NSE} = 0.99, p < 0.01). \tag{14}$$

Sediment transport capacity calculated by the model developed in this study was similar to the observed values (figure 3). As indicated by the values of the exponents, transport capacity was more sensitive to changes in flow discharge than to changes in slope gradient for the nonerodible beds used in this study. The exponents for discharge and slope gradient were 26.22% and 2.25% lower, respectively, than those obtained by equation 8. The exponents for slope gradient (0.98) and flow discharge (1.20) were also lower than the average ranges for slope gradient of 1.2 to 1.9 and for flow discharge of 1.4 to 2.4 reported by Julien et al. (1985), based on the statistical analysis of data from a series of experiments on transport capacity and the average exponents of 1.4 for both slope gradient and flow discharge reported by Prosser and Rustomji (2000). In terms of constant coefficients (67.68), it was also lower than those obtained from equations 1 and 8. These differences in exponents and coefficients suggested that the relationships with flow discharge and slope gradient were confirmed, but the formulations varied with test bed, test material, and rill width, including the combination of slope gradients and flow discharges. According to the study of Zhang et al. (2009), riverbed sediment was selected as the testing material which particle cohesion was completely different from soil used in this study. Furthermore, with respect to flow discharges and slope gradients, they were all selected based on field observation (Tang et al. 1998) in this study, and the results are meant to reflect the realistic situation rather than the ideal one. Therefore, sediment transport capacity would be overestimated or underes-

**Figure 2**
Sediment transport capacity for different (a) flow discharges and (b) slope gradients.

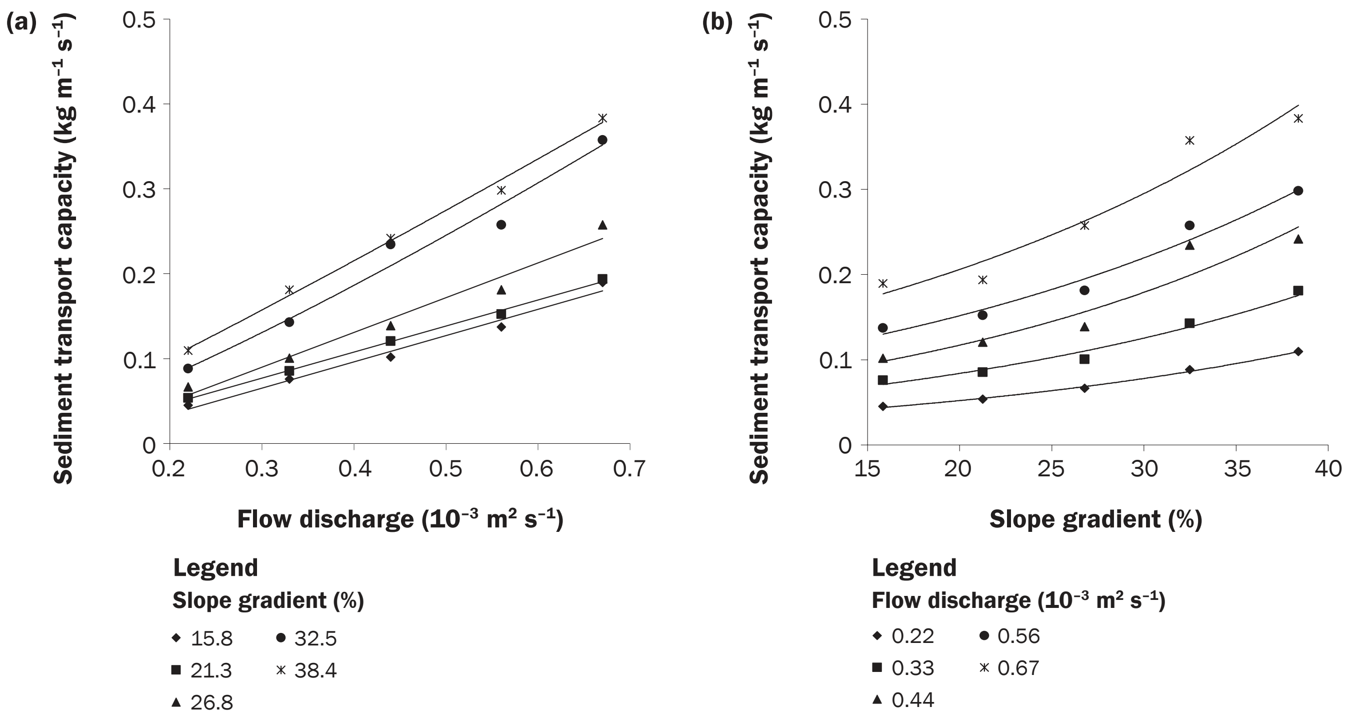
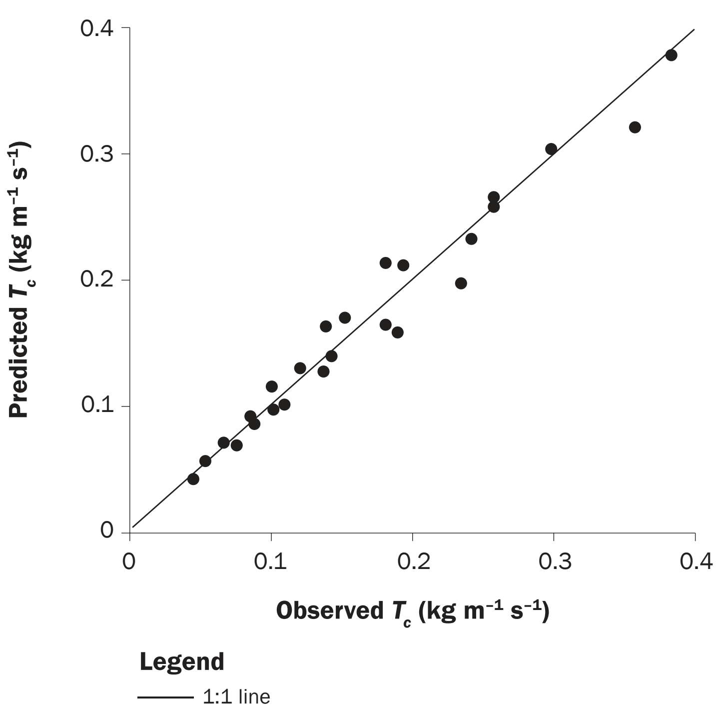



**Figure 3**
Predicted (using equation 14) vs. observed sediment transport capacity ($T_c$).




timated in the farmland of Loess Plateau if the equations 1 and 8 are considered.

Comparative analyses between predicted and observed sediment transport capacities using equations 1 and 8 are presented (table 3; figure 4a and 4b). The predicted transport capacity using equation 1 was lower than the transport capacity from the experimental observations above 0.11 kg $m^{-1}$ $s^{-1}$. Relative error, $r^2$, and NSE were approximately −13.2% to 59.3%, 0.83, and 0.10, respectively. These parameters suggested that the equation of Beasley and Huggins (1982) underestimated transport capacity in our experiment even though the flow discharge of this study was within the range of their analysis. In contrast, all the sediment transport capacities predicted by equation 8 were higher than the observed capacity, indicating that Zhang's model would overestimate the capacity, especially at lower discharges supported by NSE.

***Response of Sediment Transport Capacity to the Hydraulic Parameters of Rill Flow.*** Flow velocity is one of the main factors that directly determines sediment transport capacity. Velocities that are measured directly, however, should be modified based on the flow regime (Nearing et al. 1997; Lei

**Table 3**

**Assessment of models based on the coefficient of relative error (RE), the coefficient of mean relative error (MRE), the coefficient of mean absolute relative error (MARE), the coefficient of determination ($r^2$), and the coefficient of Nash-Suticliffe model efficiency (NSE) between observed and predicted transport capacity using equations 1 and 8.**

| Model | RE (%) | MRE (%) | MARE (%) | $r^2$ | NSE |
|---|---|---|---|---|---|
| $T_c = 146\ Sq^{0.5}$ ($q \leq 0.046$) | −13.2~59.3 | 28.7 | 30.9 | 0.83 | 0.10 |
| $T_c = 19{,}831\ S^{1.227}\ q^{1.237}$ | −99.9~−25.6 | −67.2 | 67.2 | 0.96 | −1.44 |

et al. 1998; Zhang et al. 2010a). The mean flow velocities used to determine the other hydraulic parameters considered in this study were thus obtained by multiplying the measured velocities by 0.7.

Govers et al. (1990) and Nearing et al. (1999) reported that velocity could not increase much with increase in flow discharge and slope gradient in erodible rills. In contrast, in our nonerodible bed, the mean velocity of rill flow increased with higher flow discharges and slope gradients (figure 5a and 5b). For slopes steeper than 21.3%, velocity increased more rapidly, perhaps because the sine component of gravity increased rapidly and affected the flow down the flume. Because flow velocity changed with flow discharge and slope gradient, the sediment transport capacity also varied differently. This study showed that the best equation to describe the relationship between sediment transport capacity and mean flow velocity was a power function (figure 6a):

$$T_c = 3.03\ v^{4.52}\ (r^2 = 0.93,\ \text{NSE} = 0.93,\ p < 0.01). \tag{15}$$

From figure 6a we can learn that transport capacity increased as velocity increased. Zhang et al. (2009) came to a similar conclusion but found that a linear function adequately described the relationship. These authors also reported a critical velocity (0.58 m s$^{-1}$ [1.3 mi hr$^{-1}$]) beyond which sediment could be transported. The grains of the test sediment (with a median diameter of 280 μm [0.01 in]) used in Zhang's experiment seemed to be suspended and/or deposited in the rill flow below the critical velocity. However, our study had no critical velocity, probably due to the test material, loessial soil, rather than to the uniform sediment, test bed, flow discharge, or slope gradients.

Furthermore, Nearing et al. (1989) reported that shear stress was also a good parameter for calculating sediment transport capacity. In our study, the measured sediment transport capacity could be expressed by a power function better than by other estimated functions associated with shear stress (figure 6b):

$$T_c = 0.02\ \tau^{1.65}\ (r^2 = 0.54,\ \text{NSE} = 0.60,\ p < 0.01). \tag{16}$$

With the above function, shear stress performed poorly on sediment transport capacity in our study and in that by Ali et al. (2012a). The form of relationship between shear stress and sediment capacity, however, was similar to that by Nearing et al. (1989) ($T_c = \tau^{3/2}$; WEPP model), Zhang et al. (2009) ($T_c = 0.054\ \tau^{1.982}$), and Ali et al. (2012a) ($T_c = 0.0085\ \tau^{2.06}$). The exponent was about 10% higher than that of WEPP and 17% and 20% lower than that by Zhang et al. (2009) and Ali et al. (2012a), respectively. The variation in these results is likely due to experimental conditions and test materials, lower slope gradients in the WEPP model, different grain size in the erodible bed of Ali et al. (2012a; 2012b), and uniform sand in a nonerodible bed with steep slopes and higher discharges

**Figure 4**

**Predicted ([a]equation 1 and [b] equation 8) vs. observed sediment transport capacity.**

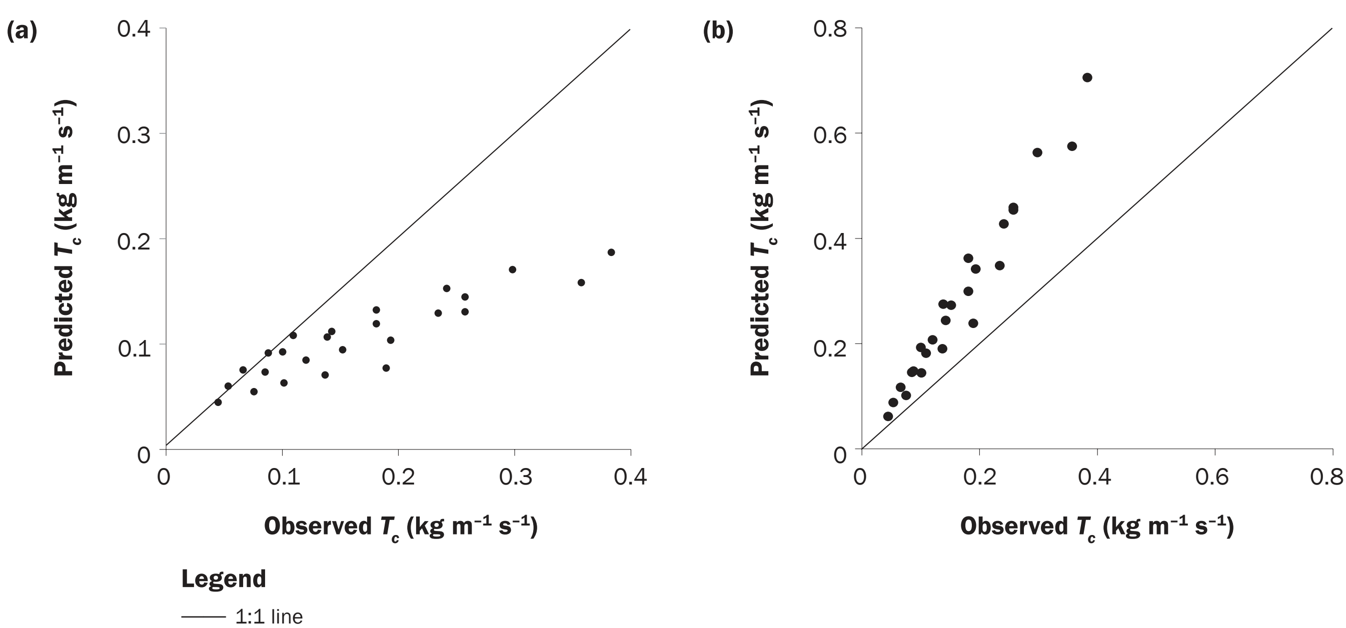
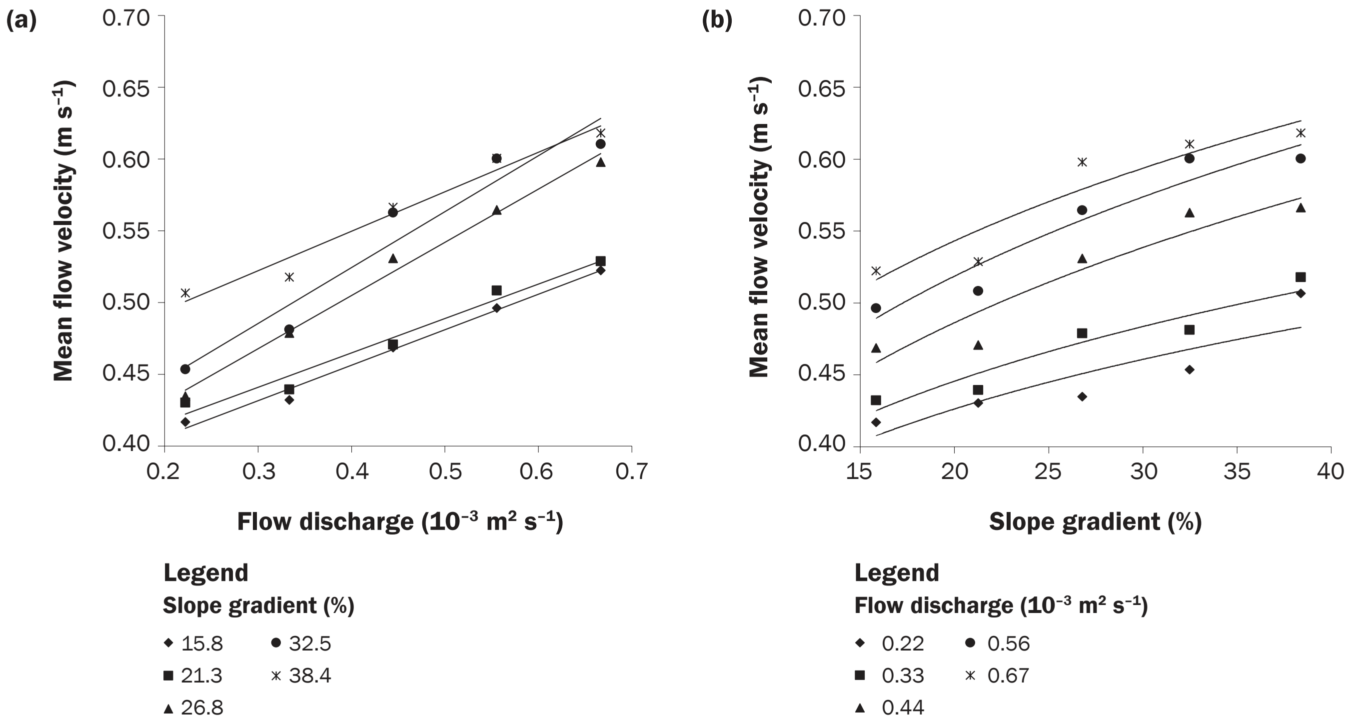


**Figure 5**

Response of mean flow velocity to different discharge and slope.




by Zhang et al. (2009). The prediction of sediment transport capacity under both high and low flow discharges and different test beds thus plays a vital role in the accuracy of the models of soil erosion (Ali et al. 2013).

Stream power or unit stream power has been considered a sensitive hydraulic parameter for developing models of empirical transport capacity (Yang 1972; Bagnold 1966; Moore and Burch 1986). Stream power and unit stream power had significant linear relationships with sediment transport capacity even though the coefficients of determination were 0.79 and 0.59, respectively (figure 6c and 6d). The linear functions for stream power and unit stream power suggested critical values for stream power (0.55 W $m^{-2}$) and unit stream power (0.02 m $s^{-1}$ [0.04 mi $hr^{-1}$]), which are supported by similar studies (Yang 1972; Moore and Burch 1986; Govers 1990 1992; Zhang et al. 2009).

Several statistical parameters were also used to evaluate the performance of the equation developed by hydraulic parameters (table 4). Sediment transport capacity was calculated by these regression equations and was compared with observation. Sediment transport capacity was evaluated well by flow velocity, in agreement with Ali et al. (2012a). However, flow velocity was not a good predictor for estimating sediment transport, due to the intrinsic drawback of direct detection. Stream power was the best hydraulic parameter for predicting sediment transport capacity, in agreement with Zhang et al. (2009). In terms of other parameters, further research is required to validate the equation related to transport capacity.

## Summary and Conclusions

This study investigated the sediment transport capacity of flows within artificial rill channels at various discharge rates and slope gradients using a loessial soil as the sediment source material. Sediment transport capacity notably increased with increases in both the flow discharge and the slope gradient. Flow discharge had a greater effect on transport capacity than did slope gradient. A power function relating sediment transport capacity to discharge and slope was well fitted by the data. By comparison, predicting sediment transport capacity resulted in a poor fit, and the predicted values were lower than the observed values calculated by equation 1 and were higher than that by Zhang et al. (2009), implying that sediment transport capacity requires verification even though several previous models have been developed.

In addition, sediment transport capacity was generally well correlated with the investigated hydraulic parameters. Mean flow velocity, which was also related to the discharge, had a strong power relationship with sediment transport capacity ($r^2$ = 0.93, NSE = 0.93). Of the other hydraulic parameters, stream power correlated best with sediment transport capacity ($r^2$ = 0.79, NSE = 0.78), while weaker relationships were obtained for shear stress ($r^2$ = 0.61, NSE = 0.6) and unit stream power ($r^2$ = 0.59, NSE = 0.59). Notably, both stream-power parameters were better predictors of transport capacity than was shear stress, even though the former are functions of the latter. Furthermore, these parameters had threshold values, which could be termed the critical stream power (0.55 W $m^{-2}$) and the critical unit stream power (0.02 m $s^{-1}$ [0.04 mi $hr^{-1}$]) under the conditions of this study.

Sediment transport capacity under these conditions could thus be accurately predicted by the derived empirical relationship with flow discharge and slope gradient or by the relationship with flow velocity. The differences between the relationships determined by this study and those of other studies can be attributed to the use of a nonerodible bed and loessial soil as the sediment source and to rill width and the ranges of flow discharge and slope gradient, implying

**Figure 6**

**Sediment transport capacity as a function of (a) mean flow velocity ($v$), (b) shear stress ($\tau$), (c) stream power ($\omega$), and (d) unit stream power ($P$) for all combination experiments.**

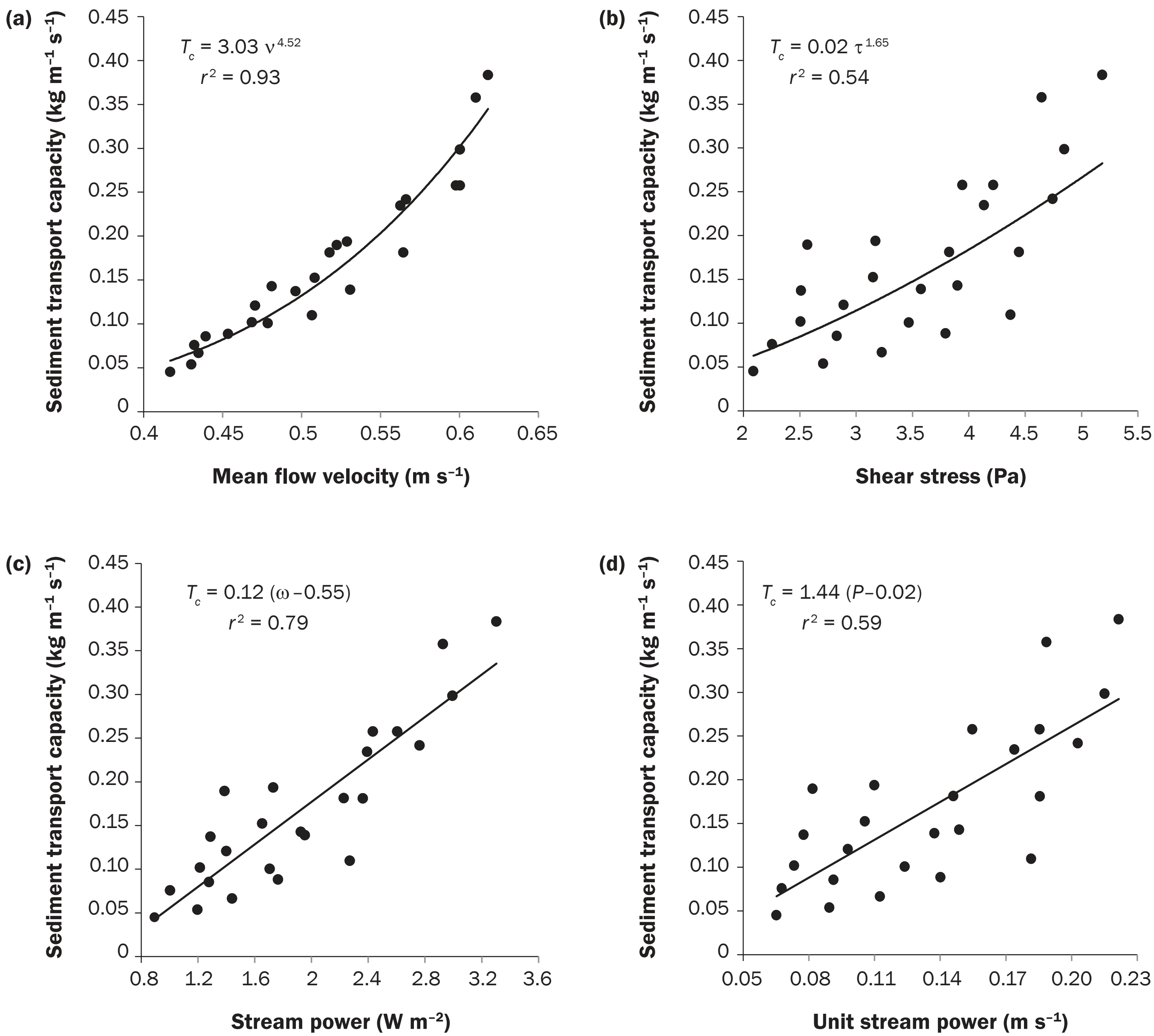



**Table 4**

**The coefficient of relative error (RE), the coefficient of mean relative error (MRE), the coefficient of mean absolute relative error (MARE), the coefficient of Nash-Sutcliffe model efficiency (NSE), and the coefficient of determination ($r^2$) on observed vs. predicted sediment transport capacity by hydraulic parameters.**

| Hydraulic parameters | RE (%) | MRE (%) | MARE (%) | $r^2$ | NSE |
|---|---|---|---|---|---|
| Mean flow velocity | −28.77~22.21 | −1.10 | 13.26 | 0.93 | 0.93 |
| Shear stress | −108.06~50.03 | −15.32 | 35.70 | 0.61 | 0.60 |
| Stream power | −88.46~47.01 | −4.94 | 25.24 | 0.79 | 0.78 |
| Unit stream power | −112.32~53.12 | −12.62 | 35.47 | 0.59 | 0.59 |

that these relationships are often site-specific. Further experiments will thus be required to evaluate sediment transport capacity in other scenarios for estimating total rill erosion over the large scale of the Loess Plateau.

## Acknowledgements

We want to thank workers from simulation rain hall of the State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau. We also want to thank David Warrington, a guest researcher at the Laboratory of Soil

Erosion and Dryland Farming on the Loess Plateau, Institute of Soil and Water Conservation, for giving suggestions on the initial manuscript and two anonymous reviewers for their valuable comments. Funding for this research came from the National Natural Science Foundation of China (No. 41171227 & 41471230) and the Key project of Chinese Academy of Sciences (No. KZZD-EW-04-03).

## References

Ali, M., M. Seeger, G. Sterk, and D. Moor. 2013. A unit stream power based sediment transport function for overland flow. Catena 101:197-204.

Ali, M., G. Sterk, M. Seeger, M. Boersema, and P. Peters. 2012a. Effect of hydraulic parameter on sediment transport capacity in overland flow over erodible beds. Hydrology and Earth System Sciences 16:591-601.

Ali, M., G. Sterk, M. Seeger, and L. Stroosnijder. 2012b. Effect of flow discharge and median grain size on mean flow velocity under overland flow. Journal of Hydrology 452-453:150-160.

Bagnold, R. 1966. A: An approach to the sediment transport problem from general physics, United States Geological Survey Professional Paper: 422–I.

Beasley, D.B., and L.F. Huggins. 1982. ANSWERS user's manual. West Lafayette, IN: Department of Agriculture Engineers, Purdue University.

Cai, Q. 1998. Research of rill initiation condition on loess hillslopes. Journal of Sediment Research (1):52-59.

De Roo, A.P.J., R.J.E Offermans, and N.H.D.T. Cremers. 1996. LISEM: A single event physically based hydrological and soil erosion model for drainage basin II: sensitivity analysis, validation and application. Hydrological Processes 10:1119-1126.

Everaert, W. 1991. Empirical relations for the sediment transport capacity of interrill flow. Earth Surface Processes and Landforms 16:513-532.

Ferro, V. 1998. Evaluating overland flow sediment transport capacity. Hydrological Processes 12:1895-1910.

Flanagan, D.C., J.C. Ascough, M.A. Nearing, and J.M Laflen. 2001. The water erosion prediction project (WEPP) model. *In* Landscape Erosion and Evolution Modelling, eds. R.S. Harmon and W.W Doe, 145-199. New York, NY: Kluwer Academics/Plenum.

Finkner, S.C., M.A. Nearing, G.R. Foster, and J.E Gilley. 1989. A simplified equation for modeling sediment transport capacity. Transactions of the American Society of Agriculture Engineers 32:1545–1550.

Foster, G.R., L. Huggins, and L.D. Meyer. 1984. A laboratory study of rill hydraulics: I. Velocity relationships. Transactions of the American Society of Agriculture Engineers 27(3):790-796.

Foster, G.R., and L.D. Meyer. 1972. Transport of particles by shallow flow. Transactions of the American Society of Agriculture Engineers 19:99-102.

Gilley, J.E., E.R. Kottwitz, and J.R. Simanton. 1990. Hydraulics characteristics of rills. Transaction of the American Society of Agriculture Engineers 33:1900-1906.

Gimenez, R., and G. Govers. 2001. Interaction between bed roughness and flow hydraulics in eroding rills. Water Resource Research 37:791-799.

Gokmen, T., and P.S. Vijay. 2012. Transport capacity models for unsteady and non-equilibrium sediment transport in alluvial channels. Computers and Electronics in Agriculture 86:26-33.

Govers, G. 1992. Evaluation of transport capacity formulae for overland flow. *In* Overland Flow: Hydraulics and Erosion Mechanics, eds. A.J. Parsons and A.D. Abrahams 243-273. London, UK: UCL Press.

Govers, G. 1990. Empirical relationships on the transporting capacity of overland flow, Transport and Deposition Processes. Proceedings of the Jerusalem Workshop, March-April 1987. IAHS 189:45-63.

Govers, G., and G. Rauws. 1986. Transporting capacity of overland flow on plane and on irregular beds. Earth Surface Processes and Landforms 11:515-524.

Hu, S.X., and A.D. Abrahams. 2006. Partitioning resistance to overland flow on rough mobile beds. Earth Surface Processes and Landforms 31:1280-1291.

Julien, P.Y., and D.B. Simons. 1985. Sediment transport capacity of overland flow. Transactions of the American Society of Agriculture Engineers 28(3):755-762.

Lal, R. 1998. Soil erosion impact on agronomic productivity and environment quality. Critical Reviews in Plant Sciences 17:319-464.

Lei, T., M.A. Nearing, K. Haghighi, and V.F. Bralts. 1998. Rill erosion and morphological evolution: A simulation model. Water Resources Research 34(11):3157-3168.

Lei, T., Q. Zhang, J. Zhao, and Z. Tang. 2001. A laboratory study of sediment transport capacity in the dynamic process of rill erosion, Transaction of American Society and Agriculture Engineers 44:1537-1542.

Luk, S., and W. Merz. 1992. Use of the salt tracing technique to determine the velocity of overland flow. Soil technology 5:289-301.

Merritt, W.S, R.A. Letcher, and A.J. Jakeman. 2003. A review of erosion and sediment transport models. Environment Modelling & Software 18:761-799.

Moore, I.D., and G.J. Burch. 1986. Sediment transport capacity of sheet and rill flow: Application of unit stream power theory. Water Resource Research 22:1350-1360.

Morgan, R.P.C., J.N. Quinton, R.E Smith, G. Govers, J. Poesen, K. Auerswald, G. Chisci, D. Torri, and M.E. Styczen. 1998. The European Soil Erosion Model (EUROSEM): A dynamic approach for predicting sediment transport from fields and small catchments. Earth Surface Processes and Landforms 23:527-544.

Nearing, M.A., L.D. Norton, D.A. Bulgakov, and G.A. Larionov. 1997. Hydraulics and erosion in eroding rills. Water Resource Research 33:865-876.

Nearing, M.A., J.R. Simanton, L.D. Norton, S.J. Bulygin, and J. Stone. 1999. Soil erosion by surface water flow on a stony, semiarid hillslope. Earth Surface Processes and Landforms 24:677-686.

Nearing, M.A., G.R. Foster, L.J. Lane, and S.C. Finkner. 1989. A Process-based soil erosion model for USDA-water erosion prediction project. Transactions of the American Society of Agriculture Engineers 32(5):1587-1593.

Prosser, I.P., and P. Rustomji. 2000. Sediment transport capacity relations for overland flow. Progress in Physical Geography 24:179-193.

Tang, K., K. Zhang, and A. Lei. 1998. Demonstration of the upper limit of slope gradient for farmland on Loess hilly and gully region. Chinese Science Bulletin 43(2):200-203.

Yalin, Y.S. 1963. An expression for bed-load transportation. Journal of Hydraulics Division- American Society of Civil Engineers 89:221-250.

Yang, C.T. 1972. Unit stream power and sediment transport. Journal of Hydraulics Division-American Society of Civil Engineers 98:1805-1825.

Zhang, G., B. Liu, Y. Han, and X.C. Zhang. 2009. Sediment transport and soil detachment on steep slopes: I. transport capacity estimation. Soil Science and Society of American Journal 73:1291-1297.

Zhang, G., R. Luo, Y. Cao, R. Shen, and X.C. Zhang. 2010a. Correction factor to dye-measured flow velocity under varying water and sediment discharges. Journal of Hydrology 389:205-213.

Zhang, G., R. Shen, R. Luo, Y. Cao, and X.C. Zhang. 2010b. Effects of sediment load on hydraulics of overland flow on steep slopes. Earth Surface Processes and Landforms 35:1811-1819.

Zhu, X. 1982. Main water erosion types and related factors on Loess Plateau. Bulletin of Soil and Water conservation (1):1-9.