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Modelling sediment transport capacity of rill flow for loess sediments on steep slopes

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ABSTRACT

Sediment transport is an important aspect of soil erosion, and sediment transport capacity (T_c) is a key to establishing process-based erosion models. A lot of studies exist that have determined T_c for overland flow, however, few studies have been conducted to determine T_c for loess sediments on steep slopes. Experimental data for this region are thus needed. The objectives of this study are to formulate new equations to describe T_c and evaluate the suitability of these equations for loess sediments on steep slopes. The slope gradients in this study ranged from 10.51% to 38.39%, and flow discharges per unit width varied from 1.11×10^{-3} m^2 s^{-1} to 3.78×10^{-3} m^2 s^{-1}. Results showed that T_c increased as a power function with flow discharge and slope gradient, with $R^2 = 0.99$ and Nash–Sutcliffe model efficiency (NSE) = 0.99. T_c was more sensitive to flow discharge than slope gradient. T_c increased as a power function with mean flow velocity, which was satisfied to predict T_c with $R^2 = 0.99$ and NSE = 0.99. Shear stress ($R^2 = 0.89$, NSE = 0.88) was also a good predictor of T_c, and stream power ($R^2 = 0.96$, NSE = 0.96) was a better predictor of T_c than shear stress. However, unit stream power was not a good predictor to estimate T_c in our study, with $R^2 = 0.63$ and NSE = 0.62. These findings offer a new approach for predicting T_c for loess sediments on steep slopes.

1. Introduction

Soil erosion has become an important environmental problem worldwide (Lal, 1998; Ali et al., 2011; Heathcote et al., 2013), and it often occurs in hilly and mountainous areas (Ali et al., 2011). The Loess Plateau in northwest China has suffered from serious soil erosion in recent decades (Shi and Shao, 2000; Liu et al., 2012; Zhao et al., 2013). Several process-based erosion prediction models (Smith et al., 1995; De Roo et al., 1996; Morgan et al., 1998; Flanagan et al., 2001) have been established to help predict the intensity of soil erosion and assess the rate of erosion in a particular area. In the Loess Plateau of China, a process-based erosion model must be established to aid in the decision making concerning soil erosion control in the area. Soil erosion involves the processes of detachment, transport and deposition of soil particles (Nearing et al., 1997). Predicting the transport capacity of overland flow (T_c) can help in understanding the soil erosion processes for developing process-based erosion prediction models (Julien and Simons, 1985; Finkner et al., 1989; Govers, 1990; Ferro, 1998). A number of equations that are credible in their representation of T_c have been proposed to estimate T_c (Beasley et al., 1982; Finkner et al., 1989; Nearing et al., 1989; Govers, 1990; Govers, 1992; Prosser and Rustomji, 2000; Flanagan et al., 2007; Zhang et al., 2008; Zhang et al., 2009; Ali et al., 2013; Mahmoodabadi et al., 2014). However, Govers (1992) suggested that using existing formula developed from observations in channels and alluvial rivers to predict the T_c of overflow is questionable because of the different hydraulic conditions. Govers (1992) tested a number of formulae using an experimental dataset obtained under laboratory conditions that simulated rill flow. The tested slopes ranged from 0.017 to 0.21. Five well-sorted quartz materials were used with a median grain size ranging from 58 μm to 1100 μm, and unit discharges were in the intermediate to high range (2×10^{-4}m^{2}s^{-1}–150×10^{-4}m^{2}s^{-1}). Govers (1992) found that no existing formula performs well over the whole range of available data. Thus far, very little data on T_c is available for loess sediments in combination with steep slope gradients, and this situation is very relevant for the Chinese loess areas. Govers (1992) also found that simple empirical equations based on shear stress, unit stream power and effective stream power, as well as the shear stress-based formula of Low (1989), can be used to predict the T_c of overland flow, at least in some cases. Thus, evaluating the relationship of T_c with the hydraulic parameter for loess sediments in combination with

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steep slope gradients is essential. Overall, obtaining accurate estimates of the T_c of rill flow for loess sediments on steep slope gradients is key to establishing a reliable soil erosion model in the Loess Plateau in China.

In some past studies, different unit flow discharges and slope gradients were set up to analyse the relationship of T_c with flow discharge and slope gradient, such as

$$T_c = k_1 q^{\beta} S^{\gamma}, \tag{1}$$

where T_c is the sediment transport capacity per unit width of slope (kg m^{-1} s^{-1}); q is the discharge per unit width (m^2 s^{-1}); S is the local energy gradient (m m^{-1}), approximated here as the surface gradient; and k_1, β and γ are empirical or theoretically derived constants (Prosser and Rustomji, 2000). Most of these equations were set up on a gentle slope. Beasley and Huggins (1982) reported that slope gradient and flow discharge strongly influenced T_c and proposed equations derived from extensive research and data analysis:

$$T_c = 146 S q^{0.5} \quad q \le 0.046 \tag{2}$$

and

$$T_c = 14600 S q^2 \quad q > 0.046, \tag{3}$$

where T_c is the sediment transport capacity (kg m^{-1} min^{-1}), S is the slope gradient (m m^{-1}) and q is the flow discharge (m^2 min^{-1}). Eqs. (2) and (3) belong to the erosion part of the ANSWERS model, whose the slope gradients were <10%. Mahmoodabadi et al. (2014) reported that a regression equation is provided as a function of unit flow discharge and final slope gradient:

$$T_c = 8590.1 q^{0.855} S^{1.872}, \tag{4}$$

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 flow discharge (m^2 s^{-1}). In this experiment, 27 experiments on three soils with three constant inflow rates (50, 75 and 122 mL s^{-1}) on three slope gradients (2%, 4% and 6%) were carried out. The mean weighted diameters of the three soils were 0.77, 0.33 and 0.19 mm, respectively. Zhang et al. (2009) suggested that flow discharge is more important than slope on steep sandy slopes and derived the following equation:

$$T_c = 19831 q^{1.237} S^{1.227}, \tag{5}$$

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 flow discharge (m^2 s^{-1}). In this experiment, the slope gradients were from 8.8% to 46.6%, flow discharge ranged from 0.625×10^{-3} m^2 s^{-1} to 5.000×10^{-3} m^2 s^{-1} and well-sorted sand with a median diameter of 0.28 mm was used. However, the test materials were not the typical soil that comes from the Loess Plateau in northwest China.

In addition, many researchers investigated new algorithms to estimate T_c with hydraulic parameters and analysed the influence of different hydraulic parameters on T_c, such as mean flow velocity, shear stress, stream power and unit stream power.

Foster and Meyer. (1972) used experimental data to obtain T_c and found that the Yalin equation estimated the T_c of overland flow well. Alonso et al. (1981) tested nine equations based on the T_c of rivers and sinks and considered the Yalin (1963) equation the most suitable for application to overland flow. The Water Erosion Prediction Project (WEPP) model used a modified Yalin equation to calculate T_c. In WEPP, T_c is determined using the shear stress, which is calculated as

$$\tau = \rho g h S, \tag{6}$$

where τ is the shear stress (Pa), ρ is the water mass density (kg m^{-3}), g is the gravitational constant (m s^{-2}), h is the hydraulic radius (m) and S is the sine of the bed slope (m m^{-1}). The modified Yalin equation used in WEPP is as follows:

$$T_c = k \tau^{1.5}, \tag{7}$$

where T_c is the sediment transport capacity (kg m^{-2} s^{-1}) and k is a transport coefficient (m$^{0.5}$ s^2 kg$^{-0.5}$). Abrahams et al. (2001) found that T_c is a function of shear stress, and that shear stress predicts it well from non-erodible flume experiments:

$$T_c = a \tau^{1.5} \left(1 - \frac{\tau_c}{\tau}\right)^{3.4} \left(\frac{u}{u*}\right)^{c} \left(\frac{w_i}{u*}\right)^{-0.5}, \tag{8}$$

where T_c is the dimensionless sediment transport rate, τ is the dimensionless shear stress, τ_c is the critical dimensionless shear stress, u/u^* is the resistance coefficient, w_i is the inertial settling velocity of the sediment, a and c are coefficients calculated respectively as log $a = -0.42 C_r / D_r^{0.20}$ and $c = 1 + 0.42 C_r / D_r^{0.20}$, where C_r is the roughness concentration and D_r is the roughness diameter.

Various studies have demonstrated the relationship between T_c and stream power. Bagnold (1966) suggested that T_c is related primarily to the stream power. Aziz and Scott (1989) found that the power relationship is a good fit for T_c and stream power according to their analysis of the behaviour of well-sorted sand with four median diameters (0.285, 0.508, 0.718, and 1.015 mm) at slopes of 3%–10%. Li and Abrahams (1999) further established this relationship based on 384 sets of flume experiments. Li et al. (2011) analysed the behaviour of well-sorted sand with a median diameter of 0.74 mm in flumes at slopes of 5%–17.6% and reported that the new sediment transport capacity equation is a function of stream power. The main hydraulic variable is the stream power in the GUEST(Griffith University Erosion System Template). The stream power is calculated as (Misra and Rose, 1996) follows:

$$\Omega = \tau V, \tag{9}$$

where V is the mean velocity (m s^{-1}), Ω is the stream power (W m^{-2}) and τ is the shear stress (Pa). The equivalent concept of T_c in the GUEST is the sediment concentration at the transport limit (C_t), which is calculated as (Misra and Rose, 1996):

$$C_t = \frac{R_1 F}{V_a} \left(\frac{\sigma}{\sigma - \rho}\right) \left(\frac{\Omega - \Omega_0}{f_r g D}\right) \tag{10}$$

where C_t is the sediment concentration at the transport limit (kg m^{-3}), R_1 is the ratio of sediment layer width to the wetted perimeter, F is the fraction of stream power effective in entrainment and re-entrainment, V_a is the weighted average settling velocity (m s^{-1}), σ is the wet density of the sediment (kg m^{-3}), ρ is the water density (kg m^{-3}), Ω_0 is the threshold stream power (W m^{-2}), f_r is a dimensionless parameter calculated through the sidewall slope of rill, and D is water depth (m). Mahmoodabadi et al. (2014) found that the performance of GUEST in predicting T_c can be further improved using the proposed value of F = 0.15.

Unit stream power became another frequently used hydraulic variable after Yang (1972, 1973) used it to develop a total load equation. The unit stream power is calculated as follows:

$$P = VS \tag{11}$$

where P is the unit stream power (m s^{-1}), V is the mean velocity (m s^{-1}) and S is the sine of the bed slope (m m^{-1}). Based on Govers (1990), the European Soil Erosion Model (Morgan et al., 1998) and the Limburg Soil Erosion Model (De Roo et al., 1996) modelled T_c as a function of unit stream power:

$$T_c = m(P - P_c)^n \text{ or } T_c = d_s m (P - P_c)^n \tag{12}$$

where T_c is the sediment transport capacity in EUROSEM ($m^3\ m^{-3}$) or in LISEM ($kg\ m^{-3}$); d_s is the mass density of the test soil ($= 2650\ kg\ m^{-3}$); m and n are coefficients calculated as $m = [(d_{50} + 5)/0.32]^{-0.6}$ and $n = [(d_{50} + 5)/300]^{0.25}$, where d_{50} is the median particle diameter of the test soil (μm) and P is the unit stream power ($cm\ s^{-1}$); and P_c is the critical unit stream power ($cm\ s^{-1}$).

The Loess Plateau in northwest China contains steep slopes and experiences high rain intensities. Most equations established previously to predict T_c were under the condition of gentle slope, and the test materials were well-sorted sand with different median diameters. Additional work is necessary on materials that have a grain size distribution similar to real soils, and few studies have determined the T_c for loess sediments on steep slope. Hence, experimental data for this region are needed.

The objective of this study is to understand the effect of flow discharge, slope gradient and mean flow velocity on the T_c of rill flume to examine the relationships of the T_c of rill flow with the hydrodynamic parameters for loess sediments on steep slopes.

2. Materials and methods

2.1. Experiment soil sample

Test materials were loess soil collected from Ansai, Shaanxi in China. Ansai County is located in a typical loess region on the Loess Plateau. The test soil was air dried and sieved through a 2 mm sieve. The test soil consisted of 6.04% fine sand (0.1–0.25 mm), 30.53% very fine sand (0.05–0.1 mm), 53.41% silt (0.002–0.05 mm) and 10.02% clay (<0.002 mm). The median diameter of the test soil was 0.04 mm. The contents of water stable aggregates of the test soil were measured using a wet sieve method. The contents of water stable soil aggregates consisted of 1.88% (1–2 mm), 2.50% (0.5–1 mm), 3.62% (0.25–0.5 mm) and 91.97% (<0.25 mm).

2.2. Experiment set-up

The experiment was conducted in the rainfall simulation laboratory of the State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau in Yangling, China. T_c was measured in a 4 × 0.1 × 0.1 m (L × W × H) rill flume (Fig. 1). The length of our experiment flume was similar to those reported by Zhang et al. (2009), who measured T_c in a 5 m-long flume, and Ali et al. (2013), who selected a 3 m-long flume. Lei et al. (2002) suggested that T_c will not increase beyond a certain rill length on the Loess Plateau and that the critical rill length is 4 m. They also reported that the rill depth is from 0.05 m to 0.15 m. Therefore, a 4 × 0.1 × 0.1 m (L × W × H) rill flume was selected in our experiment. The slope of the flume could be adjusted to within 0.05% of a desired slope. The elevation of the upper end of the flume was adjusted using a stepping motor that allowed adjustment of the bed gradient to 57.73%. Flow discharge was controlled by a series of valves installed on a flow diversion box and measured directly by a calibrated flow meter. The test soil was evenly and smoothly glued to the surface of the flume bed to maintain a constant roughness throughout all experiments.

Two sediment sources were designed so that T_c was reached for each combination of flow discharge and slope gradient. A hopper was installed above the flume 0.6 m from the top. The sediment feeding rate was controlled by the rotational speed of rotors installed within the hopper and was calibrated to measured data. The rotational speed of the rotors was controlled by adjusting the drives. The second sediment source was a box embedded in the lower end of the rill flume. The box was filled with the test soil and was covered with a very thin iron sheet while the hydraulic characteristics of the flow were being measured.

2.3. Experiment procedures

In the experiment, the slope of the flume bed and the flow discharge were adjusted to the desired values before feeding the sediment (i.e. test soil). After the flow discharge stabilised, flow depth measurements were taken using a level probe with an accuracy of 0.2 mm across the flow section at the left, middle and right at 0.8, 0.5 and 0.2 m from the lower end of the rill flume. Nine depths were measured for each combination of flow discharge and slope gradient, and each combination was repeated once. The average of 18 depths was considered to be the mean flow depth for that combination of flow discharge and slope gradient. Flow velocity was measured using $KMnO_4$ as a tracer. The time during which the tracer was required to traverse a marked distance (0.6 m)

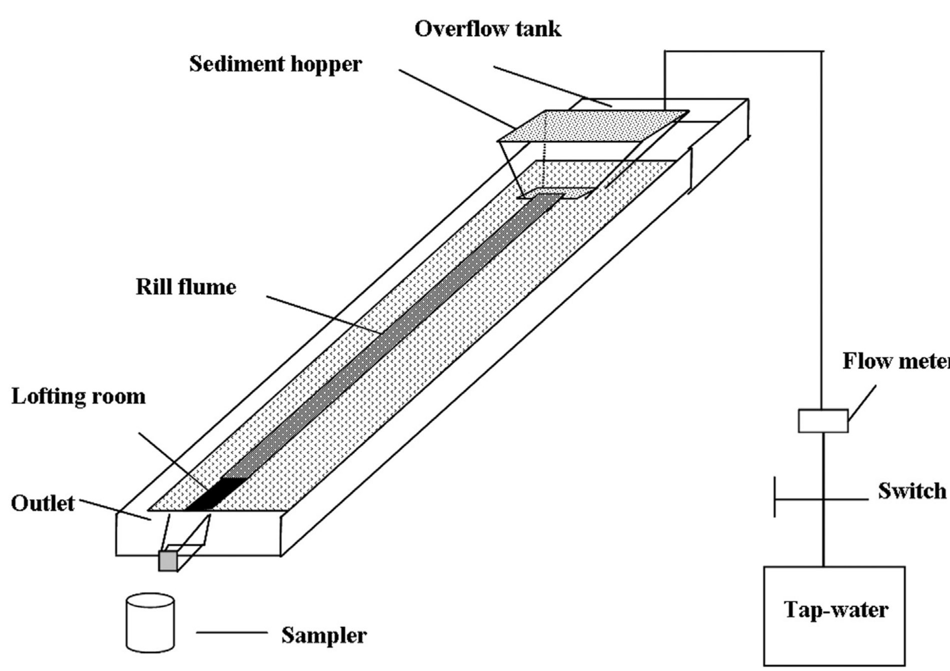

Fig. 1. Experimental rill flume used to measure sediment transport capacity.

was determined based on the colour-front propagation using a stop watch. Flow velocity was measured twice, and three flow velocity values were obtained from the left, middle and right of the experiment flume, which was 0.1 m wide every time. A total of six flow velocity values were used to calculate the mean flow velocity. The mean flow velocity was used to calculate the stream power (ω) and unit stream power (P).

The hopper began feeding the test soil to the flow of rill flume after the hydraulic characteristics were measured. The sediment feeding rate was adjusted gradually until the feeding sediment could not be carried completely and deposition occurred at the position of sediment feeding to flow, at which point T_c was assumed to be reached and the feeding rate was set. The iron sheet was then removed and measurements began. If the transport capacity was not reached due to insufficient sediment fed from the hopper, the deficit was added via sediment entrainment from the slot (i.e. the second sediment source) to reach T_c. Five samples were collected as quickly as possible for each combination of flow rate and slope gradient to avoid excessive erosion in the slot. The second sediment source was refilled with test soil if scouring occurred during the test. A new test was then started with another combination of flow rate and slope gradient.

The collected samples were allowed to settle for 24 h. The clear supernatant was decanted from the containers, and the wet sediment was oven dried at 105 °C for 12 h. The weight of the dry sediment was divided by the sampling time and flume width to obtain T_c. Sampling time was adjusted according to the flow discharge (longer for low flow rates and shorter for high flow rates). The average of the five samples was used as the measured equilibrium T_c for that combination of flow discharge and slope gradient. A series of 42 combinations of flow discharge (1.11, 1.56, 2.00, 2.44, 2.89, 3.33 and 3.78 × 10^{-3} m^2 s^{-1}) and slope gradients (10.51%, 15.84%, 21.26%, 26.79%, 32.49% and 38.39%) were tested.

2.4. Data calculation and analysis

The shear stress, stream power and unit stream power were calculated using Eqs. (6), (9) and (11), respectively. The value of V_s was used to estimate the mean flow velocity (V) through the formula:

$$V = kV_s, \tag{13}$$

where V_s is the surface flow velocity (cm s^{-1}); V is the mean flow velocity (cm s^{-1}); and k is the correction coefficient that is 0.67 for laminar flow, 0.7 for transitional flow and 0.8 for turbulent flow (Li et al., 1996; An et al., 2012). However, some researchers used other methods to obtain the mean flow velocity. Govers (1992) used Savat's algorithm to calculate the mean flow velocity for clean water, while doing velocity measurements using dye in sediment-laden flows, Savat's algorithm (1980) is as followed:

$$V = \frac{7Q}{11}R \tag{14}$$

where V is the mean flow velocity (cm s^{-1}); Q is flow discharge per unit width (cm^2 s^{-1}); R is hydraulic radium(cm). Govers (1990) observed that the flow velocities which were measured using dye tracing were significantly higher than the flow velocities predicted using Savat's algorithm. Gut et al. (1990) found that mean surface velocities of sediment-laden sheet flow were 12% higher than those measured in clear water. Aziz and Scott (1989) suggested that sediment load has a negative effect on flow velocity because of an increase in flow depth when sediment is present in the flow.

The dataset have been split in two parts at random and the size of each dataset (n) was 21. One part of the dataset was carried out to derive new equations that can describe the relationship of T_c with flow discharge and slope gradient, and hydraulic parameters by regression analysis, and derive statistical parameters R^2 and NSE values. Another part of dataset was used to the equation validation by generating statistical parameters R^2, RE and NSE values, which were used to evaluate the performance of new equations. When power equations were tested in the regression analysis, log transform before testing was conducted first to derive the coefficients and powers of power equations accurately. R^2, RE and NSE values were calculated as follows:

$$RE = \frac{(O_i - P_i)}{O_i} \times 100, \tag{15}$$

$$R^2 = \frac{\left[\sum_{i=1}^{n}\left(O_i - \overline{O}\right)\left(P_i - \overline{P}\right)\right]^2}{\sum_{i=1}^{n}\left(O_i - \overline{O}\right)^2 \sum_{i=1}^{n}\left(P_i - \overline{P}\right)^2}, \tag{16}$$

$$NSE = 1 - \frac{\sum (O_i - P_i)^2}{\sum \left(O_i - \overline{O}\right)^2}, \tag{17}$$

where RE is the relative error; R^2 is the coefficient of determination; O_i are the observed values; P_i are the predicted values; $\overline{O}$ is the mean of the observed value; $\overline{P}$ is the mean of the predicted value; and NSE is

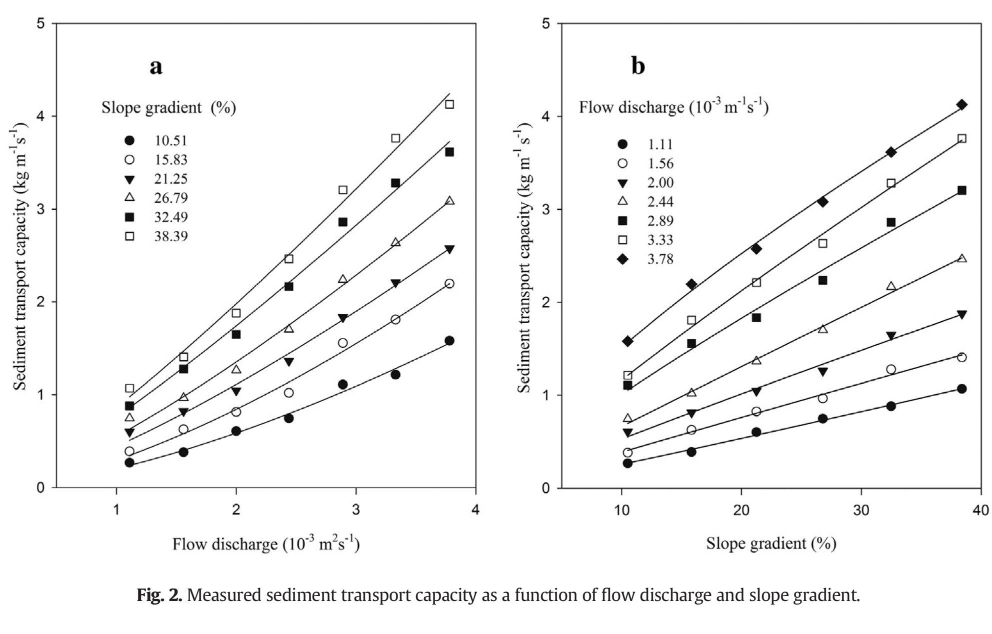

Fig. 2. Measured sediment transport capacity as a function of flow discharge and slope gradient.

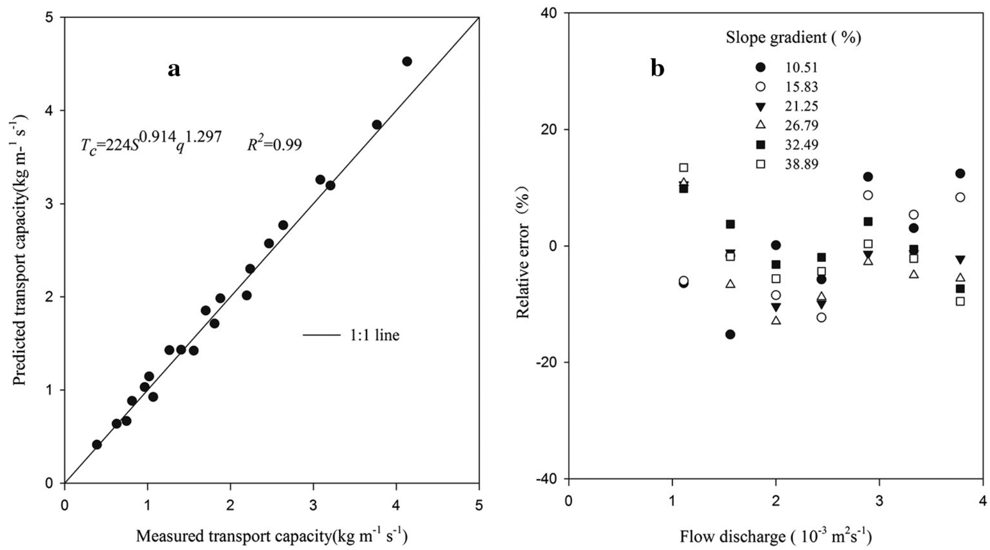

Fig. 3. Measured vs. predicted sediment transport capacity and simulated relative error vs. flow discharge across a range of slopes (using Eq. (18)).

the Nash–Sutcliffe model efficiency (Nash and Sutcliffe, 1970), a normalised statistic that reflects the relative magnitude of the residual variance compared with the variance of the observed data [good (NSE > 0.7), satisfactory (0.4 < NSE ≤ 0.7) and unsatisfactory (NSE ≤ 0.4)].(Moriasi et al., 2007; Ahmad et al., 2011; An et al., 2012).

3. Results

3.1. T_c estimated by flow discharge and slope gradient

Fig. 2 shows the measured T_c varied with different flow discharges and slope gradients. Evidently, flow discharge and slope gradient strongly influenced the measured T_c. In particular, the measured T_c increased with increasing flow discharge and slope gradient. Under the same discharge level, the increase in T_c was the largest when the slope gradient increased from 26.79% to 32.49%. However, when the slope gradient increased from 32.49% to 38.39%, the increase in T_c was reduced. This finding could be attributed to the fact that a critical slope >32.49% for transporting loess sediments existed.

To evaluate the relationship of T_c with flow discharge and slope gradient, multivariate regression analyses were conducted to derive the following relationship using one part of dataset:

$$T_c = 224S^{0.914}q^{1.297} \quad \left(R^2 = 0.99, NSE = 0.99, P<0.01, n = 21\right) \tag{18}$$

where T_c is the sediment transport capacity (kg m^{-1} s^{-1}), S is the slope gradient (%), and q is the flow discharge (m^2 s^{-1}). It was showed that Eq. (18) could be used to predict T_c well by $R^2 = 0.99$ and NSE = 0.99. The validation of Eq. (18) was further done using another part of dataset. The result of validation further showed that Eq. (18) could be used to predict T_c well by derived statistical parameters ($R^2 = 0.99$, NSE = 0.99, $P < 0.01$, $n = 21$). Fig. 3(a) also shows that the predicted T_c is extremely close to the measured values. Fig. 3(b) illustrates that relative error (RE) changes with flow discharge across a range of slopes. Most RE values were nearly 0. This result could explain the expression in Fig. 3(a).

To evaluate the suitability of the *ANSWERS* and Zhang models in our experiment, the relationship between measured T_c and the values

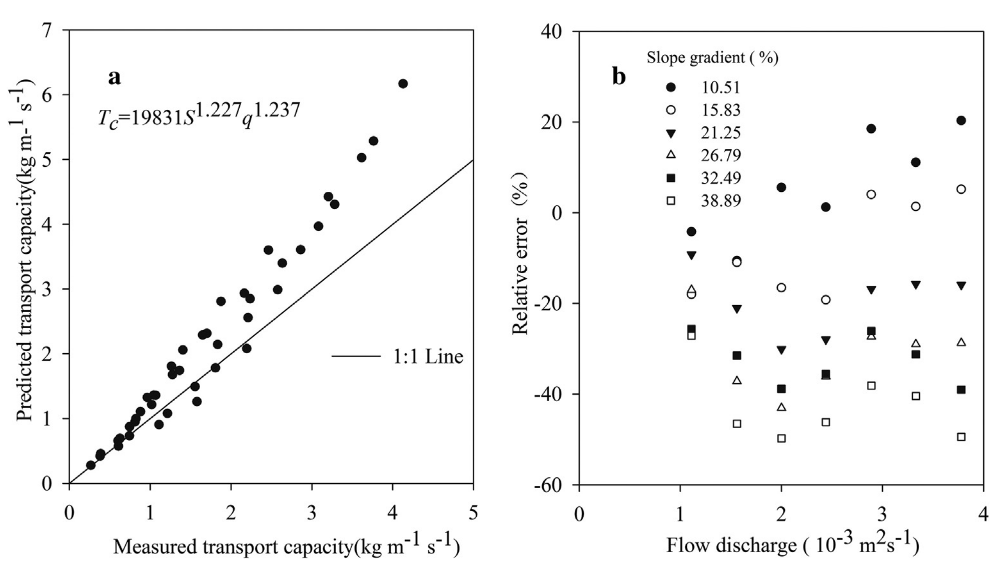

Fig. 4. Measured vs. predicted sediment transport capacity and simulated relative error vs. flow discharge across a range of slopes (Using Eq. (5)).

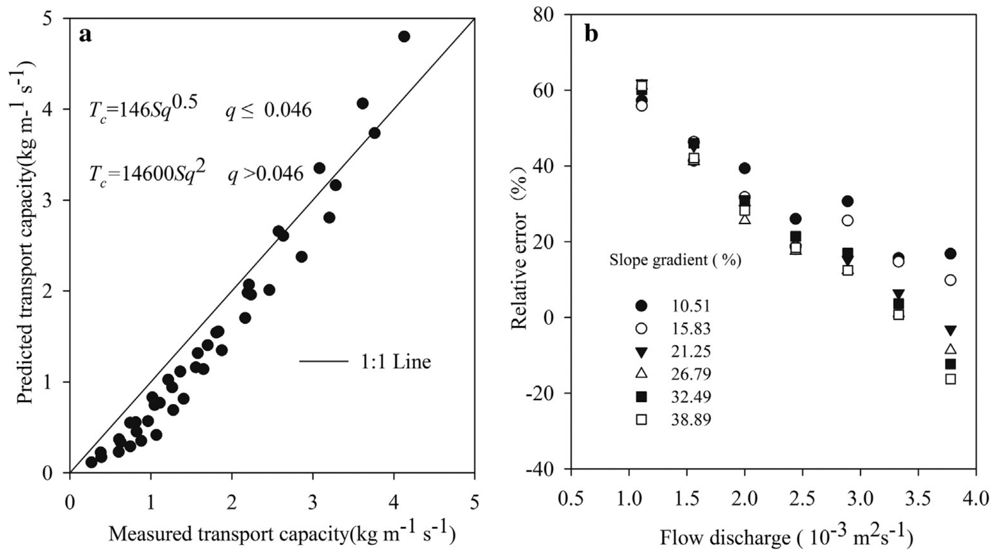

Fig. 5. Measured vs. predicted sediment transport capacity and simulated relative error vs. flow discharge across a range of slopes (Using Eq. (2) and (3)).

predicted by these models are shown in Figs. 4 and 5, respectively. Fig. 4(a) shows that the T_c calculated by the Zhang model is greater than the measured values when these values are higher than 1 kg m^{-1} s^{-1}. The prediction capability of the Zhang model decreased with increasing measured T_c. Fig. 4(b) shows that most of the predicted values are higher by 20% than the measured values; moreover, the greatest RE is nearly 50%. Fig. 5(a) indicates that only four predicted values calculated by the *ANSWERS* model are greater than their observed counterparts. Fig. 5(b) illustrates that RE decreases with an increase in flow discharge. This result indicated that the simulated T_c determined by the *ANSWERS* model was highly sensitive to flow discharge under the experiment data used in this study.

To clearly evaluate the suitability of the *ANSWERS* and Zhang models in predicting T_c in this study, Table 1 presents the results of the assessment when these models are used to predict T_c. The Zhang model was used to predict T_c in this study with $R^2 = 0.96$ and NSE = 0.54. However, RE varied from −49% to 20%, and the mean RE (MRE) was −21.3%. This result revealed that the T_c predicted by the Zhang model was 21.3% greater than the measured value, which is in accordance with Fig. 4(b).

The *ANSWERS* model (Beasley and Huggins, 1982) was also used to predict T_c under the conditions used in this study with $R^2 = 0.96$ and NSE = 0.87. However, RE varied from −16% to 62%, and MRE was 25.4%. Overall, Eq. (18) was the best option for predicting T_c with $R^2 = 0.99$, NSE = 0.99 and MRE = −1.38% under the conditions used in this study.

3.2. Effect of mean flow velocity on T_c

Fig. 6 illustrates that the mean flow velocity is as a function of flow discharge and slope gradient. Mean flow velocity clearly increased with increasing flow discharge and slope gradient. Fig. 7 shows that a power function exists between mean flow velocity and T_c using one part of dataset. The equation of this function is given as follows:

$$T_c = 2.054V^{3.976} \quad \left(R^2 = 0.97, \text{NSE} = 0.97, P{<}0.01, n = 21\right) \tag{19}$$

where T_c is the sediment transport capacity (kg m^{-2} s^{-1}), and V is the mean velocity (m s^{-1}). Eq. (19) could predict T_c satisfactorily with $R^2 = 0.97$ and NSE = 0.97. The validation of Eq. (19) was further done using another part of dataset. The result of validation further showed that Eq. (19) could be used to predict T_c well by derived statistical parameters ($R^2 = 0.99$, NSE = 0.99, $P < 0.01$, $n = 21$).

3.3. Response of T_c to hydrodynamic parameters

Fig. 8 shows the relationship between shear stress and T_c using one part of dataset. A power function existed between shear stress and T_c, i.e.

$$T_c = 0.06\tau^{1.72} \quad \left(R^2 = 0.88, \text{NSE} = 0.87, P{<}0.01, n = 21\right) \tag{20}$$

where T_c is the sediment transport capacity (kg m^{-2} s^{-1}), and τ is the shear stress (Pa). Eq. (20) indicated that shear stress was a good predictor when the measured T_c was simulated by shear stress with $R^2 = 0.88$ and NSE = 0.87. The validation of Eq. (20) was further done using another part of dataset. The result of validation further showed that Eq. (20) could be used to predict T_c well by derived statistical parameters ($R^2 = 0.89$, NSE = 0.88, $P < 0.01$, $n = 21$).

The relationship between stream power and T_c using one part of dataset is shown in Fig. 9. In our study, a linear function existed between stream power and T_c, i.e.

$$T_c = 0.3(\omega - 1) \quad \left(R^2 = 0.96, \text{NSE} = 0.96, P{<}0.01, n = 21\right) \tag{21}$$

where T_c is the sediment transport capacity (kg m^{-1} s^{-1}), and ω is the stream power (W m^{-2}). Eq. (21) showed that the measured T_c was simulated well by stream power with $R^2 = 0.96$ and NSE = 0.96. The validation of Eq. (21) was further done using another part of dataset. The result of validation further showed that Eq. (21) could be used to predict

Table 1
Assessment of models based on the coefficient of relative error (*RE*), the coefficient of mean relative error (*MRE*), the coefficient of determination (R^2) and the coefficient of Nash-Suticliffe model efficiency (*NSE*) between observed and predicted transport capacity.

Model	Equation	*RE* (%)	*MRE* (%)	R^2	*NSE*
Zhang	$T_c = 19831S^{1.227}q^{1.237}$	−49–20	−21.3	0.96	0.54
ANSWERS	$T_c = 146Sq^{0.5}$ q ≤ 0.046 $T_c = 14600Sq^2$ q > 0.046	−16–62	25.4	0.96	0.87
This study	$T_c = 224S^{0.914}q^{1.297}$	−15–11	−1.38	0.99	0.99

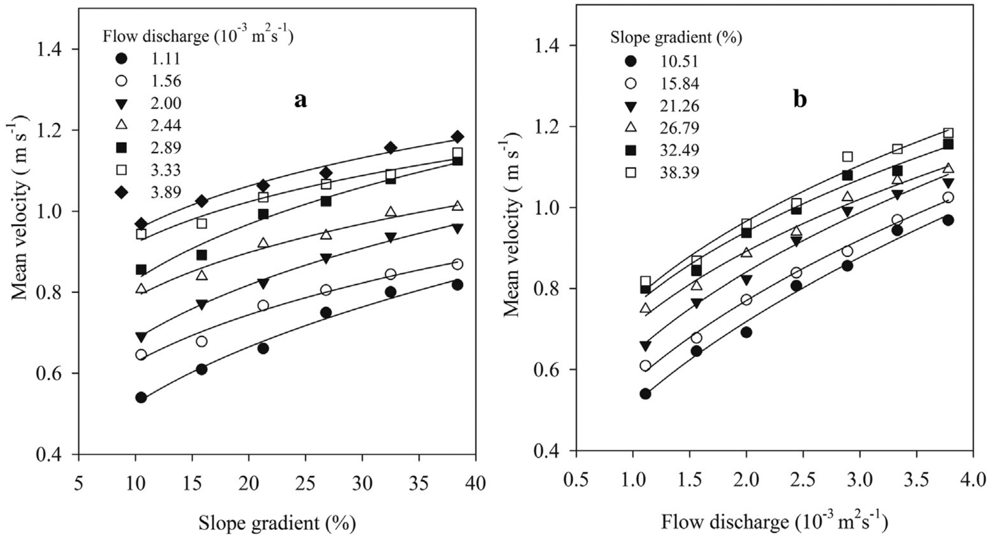

Fig. 6. Mean flow velocity as a function of flow discharge and slope gradient.

T_c well with $R^2 = 0.96$, NSE = 0.96 and MRE = −0.13%. This result indicated that stream power was a valuable hydrodynamic parameter for predicting T_c.

Fig. 10 shows that the measured T_c is a function of unit stream power using one part of dataset. Unit stream power was not a good predictor of T_c according to further validation used anther part of dataset in our study with $R^2 = 0.63$, NSE = 0.62 and MRE = −23.8%.

4. Discussion

In this study, the measured T_c increased with increasing flow discharge and slope gradient. This result is consistent with those reported by Zhang et al. (2009) and Ali et al. (2011).

Prosser and Rustomji (2000) suggested that the values of $1.0 \leq \beta \leq 1.8$ and $0.9 \leq \gamma \leq 1.8$ were recommended for sediment transport modelling when Eq. (1) was used. If a single combination was chosen, then selecting the median value of each parameter ($\beta = \gamma = 1.4$) would be appropriate, the exponents of flow discharge (0.914) and slope gradient (1.297) were within these ranges, however, both were <1.4. The slope gradient index (0.914) was 41.9% smaller than the flow discharge index (1.297), so T_c was more sensitive to flow discharge than slope gradient. The results are consistent with those of Zhang (2009) and Beasley and Huggins (1982).

The comparison of slope exponent and flow exponent for different sediment transport capacity models was shown in Table 2. Compared with the *ANSWERS* model (Beasley and Huggins, 1982), the slope exponent in this study was 9.4% smaller than the slope exponent of *ANSWERS* and the flow exponent was 54.2% smaller than the flow exponent of *ANSWERS*, so the T_c predicted by the *ANSWERS* model was 25.4% smaller than the measured T_c. This result could be attributed to the fact that slope gradient was an important factor that influenced T_c. The steep slopes selected in this study ranged from 10.51% to 38.39%, which were different from the gentle slopes used in the *ANSWERS* model. Compared with the Zhang et al. (2009) model, the slope exponent in this study was 34.2% smaller than the slope exponent of Zhang model and the flow exponent was 7.8% larger than the flow exponent of Zhang model, so T_c predicted by the Zhang model was 21.3% greater than the measured value. This result could be attributed to the differences in soil texture and median particle diameter. In our study, the slope gradients and flow discharges were similar to their counterparts in the Zhang model. The non-erodible bed used in our study was also the same as its counterpart in the Zhang model. However, the test soil in our study was

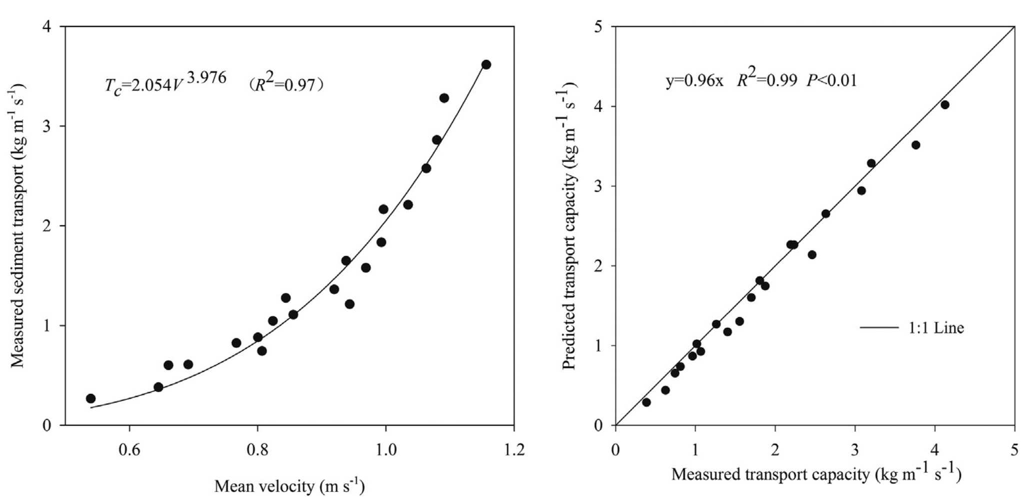

Fig. 7. Sediment transport capacity as a power function of mean flow velocity.

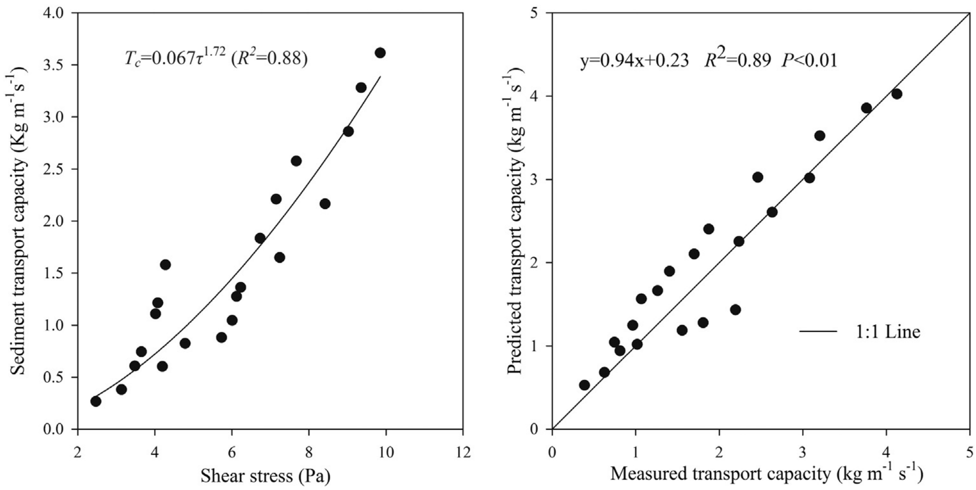

Fig. 8. Sediment transport capacity as a power function of shear stress.

loess sediments obtained from Ansai, Shaanxi in China. The test material reported by Zhang et al. (2009) was well-sorted sand collected from the bed of the Yongding River near Beijing. Moreover, the median particle diameter (0.04 mm) in our study was 600% smaller than its counterpart (0.28 mm) in the study of Zhang et al. (2009). Compared with the equation reported by Mahmoodabadi et al. (2014), the slope exponent in this study was 115.8% smaller than the slope exponent of the equation reported by Mahmoodabadi et al. (2014) and the flow exponent was 31.8% larger than the flow exponent of the equation reported by Mahmoodabadi et al. (2014) because the experiment slopes of the equation reported by Mahmoodabadi et al. (2014) were from 2% to 6% which were much smaller than the slopes in this study and the median diameter of test soil was much larger than the median diameter of our test soil. Overall, the differences of the exponents and equations in the present study and those in the published literature could be mainly attributed to the differences in slope gradients and sediment-size distributions used in the aforementioned models. On the one hand, loess sediment, which was different from sand, was selected as the test soil. On the other hand, steep slope played an important role in the result.

Mean flow velocity was a highly important hydraulic parameter that affected T_c and depended on slope gradient, flow discharge, flow depth and bed geometry (Ali et al., 2011; Zhang et al., 2009). Several researchers have suggested that slope gradient or flow discharge slightly influences mean flow velocity when the experiment bed is erodible; moreover, bed geometry is the main factor that affects mean flow velocity (Govers, 1992; Nearing et al., 1997; Takken et al., 1998; Nearing et al., 1999; Giménez and Govers, 2001; Ali et al., 2011). However, a non-erodible bed was selected in the present study, but flow discharge and slope gradient still strongly influenced mean flow velocity. This result is consistent with that in the study of Zhang et al. (2009). However, the Eq. (19) was different from its counterpart in the report of Zhang et al. (2009), who determined that a linear function existed between mean flow velocity and T_c. This finding could be attributed to the fact that soil particles could influence mean flow velocity.

Many studies have suggested that shear stress is generally not a good predictor of T_c for overland flow on erodible beds (Govers and Rauws, 1986; Govers, 1992). However, Nearing et al. (1989) and Zhang et al. (2009) found that their measured T_c was simulated well by shear stress. In our study, shear stress was a good predictor of T_c. The result is consistent with the assertions of Zhang et al. (2009)and Nearing et al. (1989). Compared with the equation ($T_c = k\tau^{1.5}$), which was reported by Nearing et al. (1989), the exponent of Eq. (20) (1.72) was approximately 13% greater than that of the *WEPP* model (1.5) reported by Nearing et al. (1989). The variation in result was likely

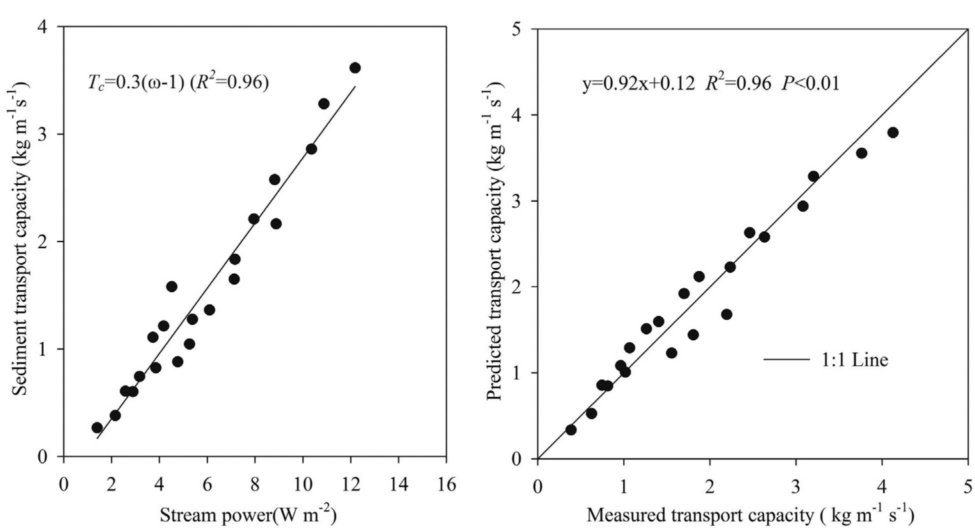

Fig. 9. Sediment transport capacity as a linear function of stream power.

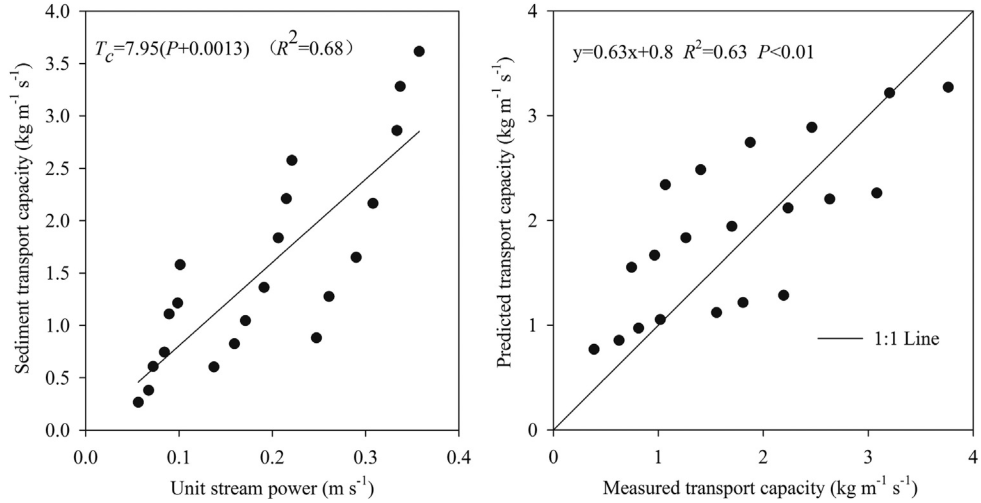

Fig. 10. Sediment transport capacity as a linear function of unit stream power.

ascribed to slope gradient and bed surface conditions (smooth or rough); a steep slope and a non-erodible bed were used in the present study. The exponent of 1.72 of Eq. (20) was approximately 13% smaller than the exponent of 1.982 reported by Zhang et al. (2009). This variation could be attributed to the different test materials used in the two studies. Loess sediments were selected in our work, whereas uniform sand was used as the test material in the study of Zhang et al. (2009).

Several researchers have indicated that stream power or excess stream power is a valuable hydrodynamic parameter for predicting T_c (Govers, 1990, 1992; Li and Abrahams, 1999; Abrahams et al., 2001). In this study, stream power is a very good predictor of T_c, which agreed with the findings of previous researchers (Govers, 1990, 1992; Li and Abrahams, 1999; Abrahams et al., 2001).

Unit stream power is also an important hydrodynamic parameter that affects T_c. The response of T_c to this parameter was also analysed in the present study. In the Limburg Soil Erosion Model (Govers, 1990; de Roo et al., 1996), T_c with the unit of kg m^{-3} was described by unit stream power. In the present study, T_c with the unit of kg^{-1} m^{-1} s^{-1} was used instead of T_c with the unit of kg m^{-3}, which could be obtained from the unit of kg^{-1} m^{-1} s^{-1} divided by the per unit width sediment-laden flow discharge with the unit of m^2 s^{-1}. This T_c was described by unit stream power such that the relationships between T_c and different hydraulic parameters could be compared in terms of R^2 and NSE based on the same unit of sediment T_c.

For erodible beds, numerous studies have indicated that unit stream power exhibited the greatest potential to estimate T_c (Govers and Rauws, 1986; Moore and Burch, 1986; Govers, 1990; Ali et al., 2011). However, for non-erodible beds, the unit stream power in our study was not a good predator of T_c and this result agreed with the finding of Zhang et al. (2009), who also used a non-erodible bed to measure T_c. Such result could be attributed to the fact that theoretical concepts derived from erodible beds did not necessarily reflect the conditions for non-erodible beds (Ali et al., 2011).

5. Conclusion

In this study, graphical and statistical analyses using one part of dataset were conducted to establish sediment transport capacity equations and another part of dataset was used to the equation validation to evaluate the suitability of these equations. The results of this study showed that T_c increased as a power function with increasing flow discharge and slope gradient. A critical slope >32.49% for the transport of loess sediments was observed. T_c was more sensitive to flow discharge than to slope gradient. This result is similar to those of the studies of Zhang et al. (2009) and Mahmoodabadi et al. (2014). The *ANSWERS* and Zhang models were used to calculate T_c. The results showed that the T_c predicted by the *ANSWERS* model was smaller than the measured value because steep slopes ranging from 10.51% to 38.39% were used in the present study, which were different from the gentle slopes used in the *ANSWERS* model. By contrast, the predicted T_c used by the Zhang model was larger than the measured value probably because of the differences in soil texture and median particle diameter. Compared with the *ANSWERS* and Zhang models, Eq. (18) provided the best formula for predicting T_c with $R^2 = 0.99$ and NSE = 0.99 under the condition of the present study.

In addition, a power function was observed between mean flow velocity and T_c. Eq. (19), which was established with mean flow velocity and T_c, could predict T_c satisfactorily with $R^2 = 0.99$ and NSE = 0.99.

Based on the results of this study, when loess sediments and steep slopes were selected in the experiment, both shear stress and stream power were good predictors of T_c for a non-erodible bed. A power function existed between shear stress and T_c. Eq. (20) showed that shear stress was a good predictor when the measured T_c was simulated by shear stress with $R^2 = 0.89$ and NSE = 0.88. A linear function was observed between stream power and T_c. Eq. (21) indicated that stream power was a valuable hydraulic parameter for predicting T_c with $R^2 = 0.96$ and NSE = 0.96. Unit stream power was not a good predictor

Table 2
The comparison of model parameters of slope exponent and flow exponent.

Model	Equation	Slope exponent	Flow exponent	Slope (%)	D_{50}(mm)
Zhang (Zhang et al., 2009)	$T_c = 19831S^{1.227}q^{1.237}$	1.227	1.237	8.8–46.6	0.28
ANSWERS (Beasley and Huggins, 1982)	$T_c = 146Sq^{0.5}$ q ≤ 0.046	1	0.5	<10	–
	$T_c = 14600Sq^2$ q > 0.046	1	2		
Mahmoodabadi (Mahmoodabadi et al., 2014)	$T_c = 8590.1S^{1.972}q^{0.855}$	1.972	0.855	2–6	0.19–0.77
This study	$T_c = 224S^{0.914}q^{1.297}$	0.914	1.297	10.51–38.39	0.04

of T_c in our study with $R^2 = 0.63$ and NSE = 0.62. This result was different from those of many other studies (Govers and Rauws, 1986; Moore and Burch, 1986; Govers, 1990; Ali et al., 2011), because research performed on erodible beds did not necessarily reflect the same conditions in research conducted on non-erodible beds (Ali et al., 2011). Therefore, the derived relationships should be evaluated further using loess sediments on an erodible bed with steep slopes.

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