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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

$$T_c = k_1 q^{\beta} S^{\gamma}, \quad (1)$$

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:

$$T_c = 146Sq^{0.5} \quad q \le 0.046 \quad (2)$$

and

$$T_c = 14600Sq^2 \quad q > 0.046, \quad (3)$$

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:

$$T_c = 8590.1q^{0.855} S^{1.872}, \quad (4)$$

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:

$$T_c = 19831q^{1.237} S^{1.227}, \quad (5)$$

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

$$\tau = \rho g h S, \quad (6)$$

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:

$$T_c = k\tau^{1.5}, \quad (7)$$

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:

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

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 $\log a = -0.42C_r/D_r^{0.20}$ and $c = 1 + 0.42C_r/D_r^{0.20}$, 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:

$$\Omega = \tau V, \quad (9)$$

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):

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

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:

$$P = VS \quad (11)$$

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:

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