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A process-based model for channel degradation: application to ephemeral gully erosion

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Abstract

Ephemeral gullies are important features of soil erosion, yielding large amounts of sediment and dissecting the landscape. In spite of their agricultural and environmental importance, gully erosion is not usually considered in routine schemes for predicting soil loss. An event-oriented process-based model for stream degradation has been adapted for the description of ephemeral gully erosion. The initial channel is considered as prismatic, formed in a uniform soil profile, defined by the watershed swale, and receives upstream and lateral runoff. The resulting gradually varied flow is computed by the standard step method, whereas the erosion, assumed transport-limited, is attributed first to bed erosion, and the remainder is dedicated to the bank erosion. At the end of each time step, channel shape is redefined accordingly with the computed erosion in each reach. A sensitivity analysis revealed that particle density, particle size and roughness coefficient are the key parameters. A calibration of the model simulating the incision of an ephemeral gully in a small watershed with highly erodible soil allowed a proper estimation of soil loss and the gully cross-section shapes along the channel with realistic values of the calibrated parameters.

Keywords: Soil erosion; Gully erosion; Erosion modelling

1. Introduction

Ephemeral gullies are small incised channels formed in cultivated landscapes by concentration of runoff in small valleys (swales), which are refilled by regular far-

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ming operations frequently during the dry summer season. Even though refilling may be done frequently, they are reformed during the next rainy season, developing an incised channel with accelerated formation of lateral rills (Foster, 1986). Their contribution to soil losses from agricultural fields is very important although not always recognized in regular surveys (e.g. Thomas and Welch, 1988). Ephemeral gullies are common in the loess belt of central Europe and in the southeastern United States, where they are responsible up to 40% of the total soil losses in agricultural areas (Casalí et al., 2000).

There are many processes involved in the formation of ephemeral gullies. Foster (1986) described the main processes of detachment of particles and their transport. More recently, Bennett et al. (2000) and Casalí et al. (2000) have identified the processes as (i) formation and migration of headcuts; (ii) erosion and deposition in the channel bed; and (iii) channel bank erosion and sloughing. The headcut is an abrupt slope change, like a step, appearing upstream of the gully where water concentrates and starts its erosive activity. Headcut occurrence and migration is attracting the attention of researchers because of its importance in ephemeral gully erosion (e.g. Bryan, 1990). Stein et al. (1993) explored the mechanics of the scour downstream of the headcut. De Ploey (1989) proposed a model for headcut retreat in rills and gullies, and Robinson and Hanson (1994), a simple model of mass failure to describe headcut advance. Bennett (1999) and Bennett and Casalí (2001) analyzed the influence of bed slope and headcut height, respectively, on the growth and migration of headcuts. Robinson et al. (2000) found a great similarity of headcut processes in rills and gullies. Other authors have examined the connection between headcut erosion and occurrence of ephemeral gullies (Casalí et al., 1999). An appreciable contribution to bank erosion is the segregation and sliding of blocks, starting a sort of chain reaction since the erosion in banks and bed increases the instability of wall slopes (Alonso and Combs, 1990). Nevertheless, the contribution of wall sliding and slumping to the erosion in ephemeral gullies is not always accepted. Other researchers restrict the importance of these processes to permanent gullies (USDA-SCS, 1992).

Given the spread and importance of soil erosion by ephemeral gullies, it is necessary to develop a tool to assess their magnitude, describing their behavior and predicting their activity. Most of the empirical schemes commonly used to evaluate soil loss in agricultural areas like RUSLE (Renard et al., 1997) do not consider gully erosion. Other models including gully erosion are CREAMS (Knisel, 1980), WEPP (Flanagan and Nearing, 1995), EGEM (USDA-SCS, 1992) or CONCEPTS (Langendoen et al., 1998). None of the above models considers the headcut component. Other approaches include the use of topographic indexes to locate or/and quantify the sediment yield by ephemeral gully erosion (e.g. Vandaele et al., 1996; Vandekerckhove et al., 1998; Casalí et al., 1999). The results of some studies (Montgomery, 1999) are encouraging to integrate gullies in the erosion–deposition pattern in basins with the help of digital elevation models (DEM).

The main purpose of this report is the description of a simple event-based model to study ephemeral gully erosion, developed from the river erosion model proposed by Alonso and Combs (1990).

2. Model description

The model is fully described in Casalí (1997). The main channel is prismatic with a trapezoidal cross-section located in the swale. The longitudinal and lateral slopes are those of the surrounding terrain. The channel has a length L, divided in intervals of length Δx. The duration of the event is divided in time steps Δt. The aim of the model is the analysis of the formation of ephemeral gullies more than the study of the evolution of existent gullies, which is very important for the management of gully-prone areas where a few intense rains may cause large gullies (Casalí et al., 1999).

The basic equations of the model are the conservation of mass and momentum for water and sediment. The mass conservation of water states that the increase in flow rate in the channel Q (L^3T^{-1}) is due only to lateral contributions q_L (L^2T^{-1}):

$$Q(x + \Delta x) = Q(x) + q_L \Delta x \tag{1}$$

where x represents distance (L). The conservation of momentum, assuming one-dimensional, unsteady, gradually varied flow, is written as (Alonso and Combs, 1990):

$$\frac{\partial}{\partial x}\left(\frac{V^2}{2g} + h + z\right) + S_f = 0 \tag{2}$$

where V is the average water velocity (LT^{-1}), h is the water depth (L), z is the bed elevation (L), g is the acceleration due to gravity (LT^{-2}) and S_f is the energy slope averaged within the reach and estimated by the Manning uniform flow equation:

$$S_f = (nQR_h^{-2/3}A^{-1})^2 \tag{3}$$

where n is the roughness coefficient assumed as constant ($L^{-1/3}T$), R_h is the hydraulic radius (L) and A is the cross-section area (L^2).

These equations are solved by an iterative scheme analogous to the standard step method for backwater calculations (Alonso and Combs, 1990; Chaudhry, 1993), assuming that at the downstream end, where flow rate is known, the water depth is normal. Chow (1958) estimates that possible errors, whether by the choice of an inaccurate water depth upstream, or by the use of an improper direction in the computation (upstream under supercritical flow conditions or downstream under subcritical conditions), may not be appreciable.

The mass conservation of sediment is formulated as (Bennett, 1974):

$$\frac{\partial z}{\partial t} + \frac{\partial q_s}{\partial x} = 0 \tag{4}$$

The effective flux density of sediment per unit width q_s (LT^{-1}) is,

$$q_s = \frac{Q_s}{B(1-\lambda)} \tag{5}$$

where Q_s is the total mass flux density of sediments (L^2T^{-1}), B is the active bed width (L), λ is the effective porosity of the sediment layer deposited on the bed and t is the time (T).

The erosion process is assumed as transport-limited (Bennett, 1974; USDA-SCS, 1992) (Figs. 1 and 2). Spomer and Hjelmfelt (1986) and Grissinger and Murphey (1989) observed in Iowa and Mississippi, respectively, that ephemeral gully erosion was a transport-limited process. Flow transport capacity T_c is computed by Yang's equation (Yang, 1973) as recommended by Alonso et al. (1981) for particle sizes over 0.1 mm in diameter, since soil particles move as aggregates during ephemeral gully formation events. Other transport capacity equations could also be considered (Govers, 1990, 1992; Yang, 1996) and easily incorporated to the model. Given the transient character of these events, causing nonequilibrium sediment transport, the sediment load q_s is corrected with the factor proposed by Bell and Sutherland (1983) to a more suitable value for these conditions, q_{sn}:

$$q_{sn}(x,t) = \{1 - \exp[-C(t)(x - x_0)]\}q_s(x,t) \tag{6}$$

where $C(t)$ (L^{-1}) is a time-dependent coefficient, and x_0 (L) is the reference point. $x_0 = 0$ when the reference point is taken at the starting point of the channel. Choudhury (1995) suggested an expression for the time-dependent coefficient, valid for alluvial river degradation processes:

$$C = (1 + \alpha' t)^{-1} \tag{7}$$

where α' (L^1T^{-1}) is a parameter to be calibrated later. Although coefficient C (Eq. (7)), despite several research efforts, remains undefined, curve fitting was carried out by Choudhury (1995) using Bell and Sutherland (1983) data and α' was found to be 0.5 for the case.

The estimation of bed erosion requires the knowledge of the maximum bed shear stress, τ ($ML^{-1}T^{-2}$), related to the mean value by a coefficient C_τ (Olsen and Florey, 1952). τ can be calculated by:

$$\tau = C_\tau \gamma R_h S_f \tag{8}$$

where γ ($ML^{-2}T^{-2}$) is the specific weight of the water. Bed soil erosion $E_{j,n}$ in any reach $j\Delta x$, at any time $n\Delta t$, as volume per unit length, (L^2) is computed by:

$$E_{j,n} = \frac{k(\tau_{j,n} - \tau_c)B_{j,n}\Delta t}{\gamma_s} \tag{9}$$

with k as the erodibility coefficient (T^{-1}); $\tau_{j,n}$ and τ_c are the shear stresses in the same j increment and time step n and the critical value ($ML^{-1}T^{-2}$), respectively; $B_{j,n}$ (L) is the current bed width and γ_s is the specific weight of bed particles ($ML^{-2}T^{-2}$). k and τ_c can be estimated from a number of data sets and different methods (Arulanandan et al., 1980; Osman and Thorne, 1988; Lal and Elliot, 1994; Hanson, 1990a,b; Flanagan and Livingston, 1995).

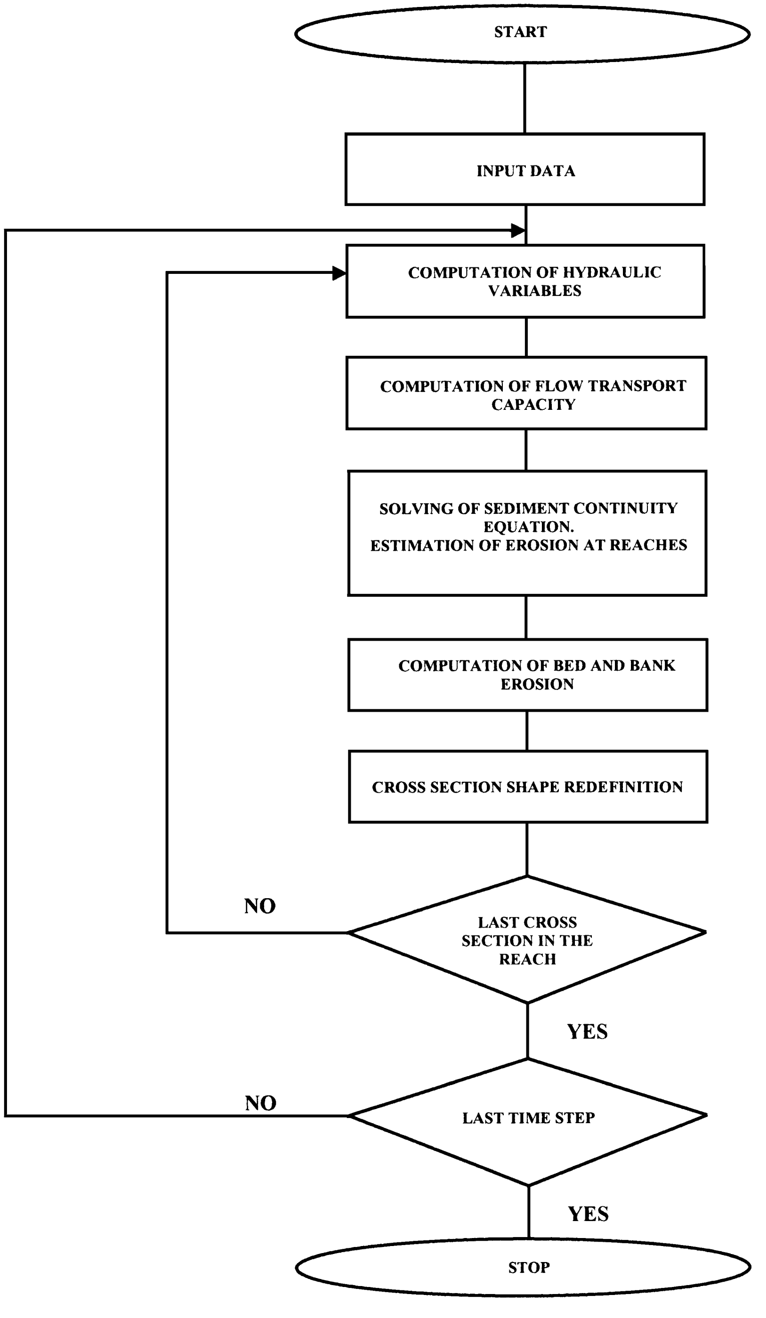

Fig. 1. Flow diagram of the model.

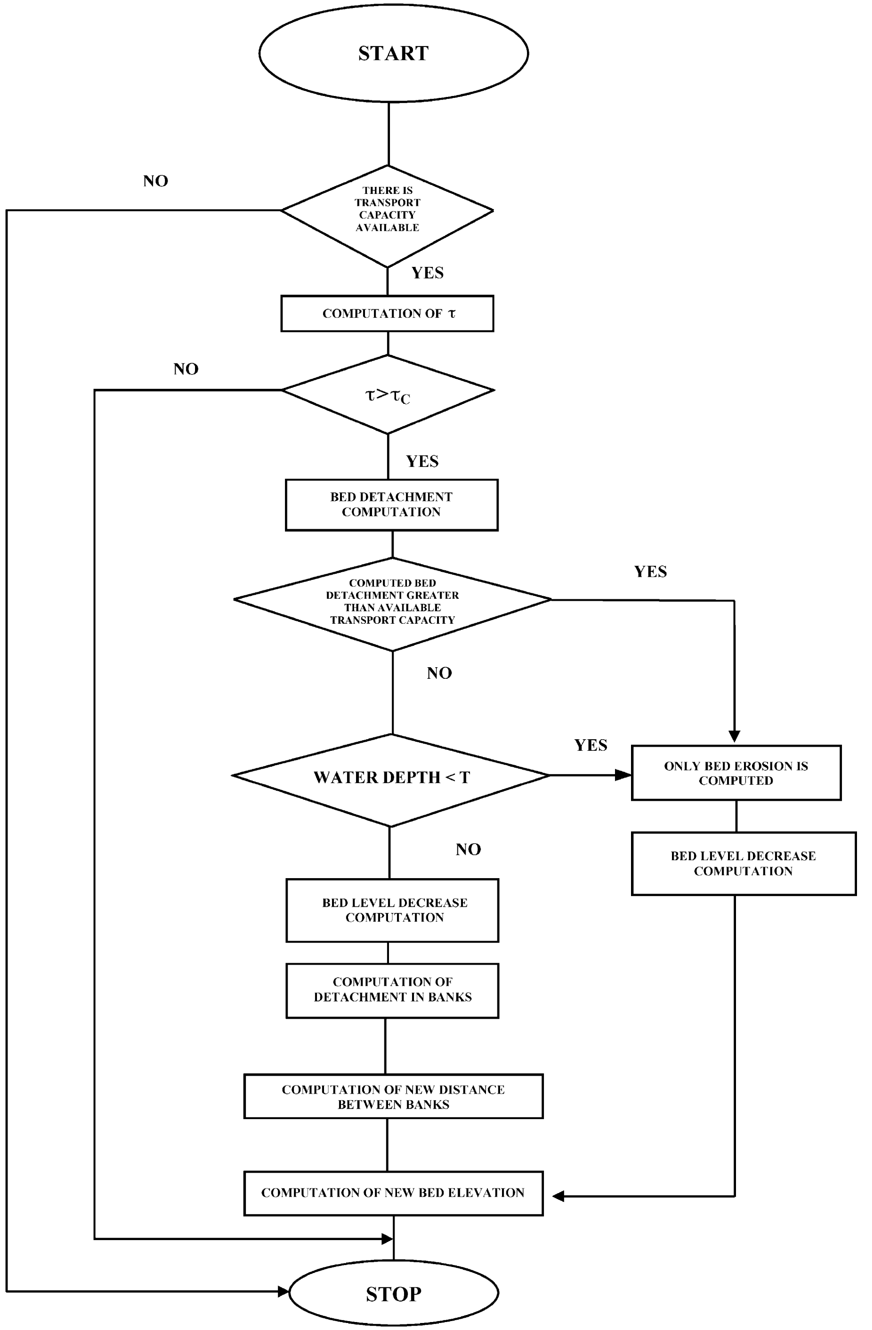

Fig. 2. Flow diagram of the erosion computations.

Eq. (4) is applicable to a channel of any shape assuming the changes of the sediment load affects bed elevation only, as it has been used in bed load studies or descriptions of bed forms. The integration of sediment continuity equation (4) requires small time steps. The Lax scheme as modified by De Vries, with centered finite differences, successfully adopted by Alonso and Combs (1990), is used here:

$$z_{j,n} = \frac{\alpha}{2}(z_{j-1,n} + z_{j,n}) + (1-\alpha)z_{j,n} + \frac{\Delta t}{2\Delta x}(q_{s_{j,n}} - q_{s_{j+1,n}}) \tag{10}$$

where the subindices j and n refer to the position and time, respectively, and α is a time-dependent coefficient. De Vries (1971) showed that the numerical stability condition for the scheme in Eq. (10) is $C_R^2 \langle \alpha \leq 1$, where C_R^2 is the local Courant's net number. The values at both external boundaries are found by linear extrapolation.

Fig. 1 shows the flow diagram of the model. Eq. (4) is solved for z in each time step and channel reach. The product $B\Delta t$ (L^2) represents the areal change in a noncohesive bed as a consequence of the degradation–aggradation process, an estimate of the erosion–deposition. As the more detailed flow diagram (Fig. 2) indicates, if the calculated maximum bed shear stress τ is less than the critical shear stress τ_c, no erosion is computed. As shear stress is larger in the bed than in the walls (Chow, 1958), the available transport capacity in the reach is used firstly to erode soil from the bed (Fig. 2, Eq. (9)), and then completed with soil from the walls. If the calculated bed soil loss during the time step, $E_{j,n}$, exceeds the available transport capacity transformed to equal units, erosion is restricted to the bed. Similarly, if water depth is lower than the height T (Figs. 2 and 3), no wall erosion is considered. To support this assumption, it can be considered that: (1) shear stress is larger in the bed than in the walls (Chow, 1958); (2) if water depth is lower than the height T, flow in the gully is usually shallow and comes after a period of active wall erosion (Bennett et al., 2000); (3) rectangular in shape and deep ephemeral gully cross-sections are frequent (Poesen and Govers, 1990; Casalí et al., 1999). Then, the elevation of the bed is computed as $\Delta z = E/B$ and the bed shape is redefined.

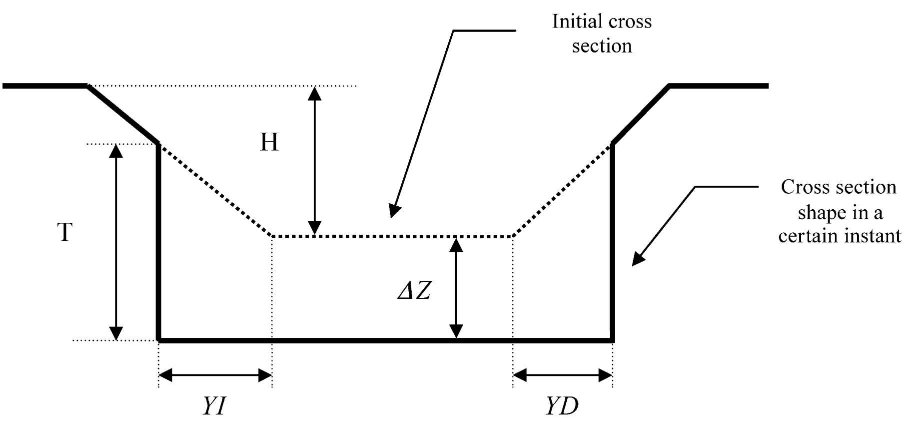

Fig. 3. Cross-section at the initial and final stages of a time step.

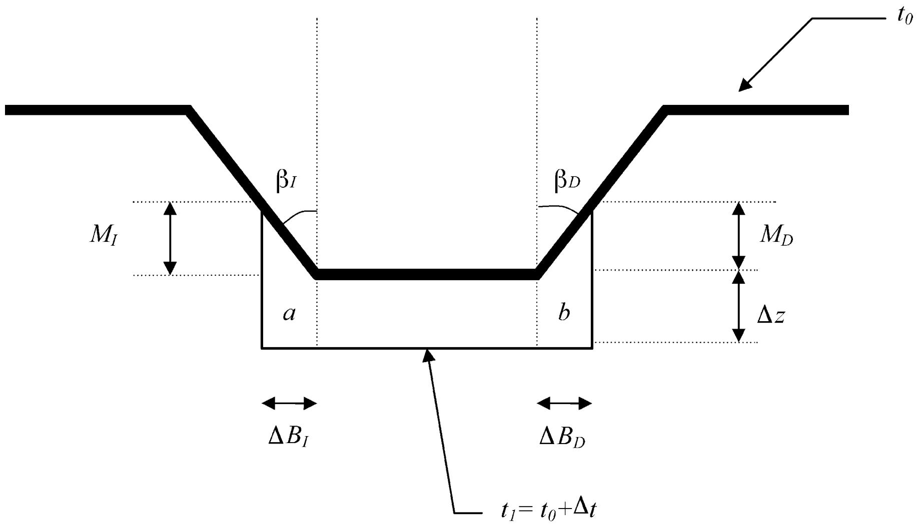

Fig. 4. Scheme of lateral erosion in a cross-section.

The lateral extension of the channel is shown in Fig. 4, where the profile appears with a bold line, corresponding to time t_0, and with thinner line, deepened at a later stage, $t_0 + \Delta t$. The quadrilaterals marked with letters a and b (Fig. 4) disappear by lateral or wall erosion. Their area is determined by the available transport capacity not previously used for bed erosion. A simple area expression:

$$a = \frac{M_\mathrm{I} + 2\Delta z}{2}\Delta B_\mathrm{I} = (\Delta B_\mathrm{I}\cot\beta_\mathrm{I}/2 + \Delta z)\Delta\beta_\mathrm{I} \tag{11}$$

yields a second degree equation in ΔB_I. A similar equation gives the extension of the right side of the channel section.

3. Model evaluation

Once developed, the model needs to be evaluated after a previous sensitivity analysis showing the more important variables.

3.1. Sensitivity analysis

The main parameters of the model are: channel length, channel slope, soil critical shear stress τ_c, the erodibility coefficient k, the coefficient C_τ, the initial bed width, side slopes, Manning's n roughness coefficient, the nonequilibrium transport coefficient α', the specific weight of soil particles, soil porosity, water temperature, size and density of carried particles and the parameters of the settling velocity equation proposed by Dietrich (1982), which are Powers' roundness factor and Corey's shape factor. In order to assess the relative importance of each variable, a sensitivity analysis was performed, studying the effect that a change in any input would cause to the model output. An expression used by McCuen and Snyder (1986) was chosen for that purpose. The sensitivity coefficient s is the ratio of the relative output change and the relative input change. If for any input whose value is I_1, an output O_1 is produced, and for the input I_2 the output is O_2, the sensitivity coefficient is:

$$s = \left(\frac{O_2 - O_1}{O_{\overline{12}}}\right)\left(\frac{I_2 - I_1}{I_{\overline{12}}}\right)^{-1} \tag{12}$$

The normalizing values are the average of output $O_{\overline{12}}$ and input $I_{\overline{12}}$, respectively. This index is a discrete version of the logarithmic sensitivity (e.g. Kabala, 2001). As Baffaut et al. (1997) indicated, the use of this sensitivity index has the disadvantage of not taking into account the interaction between variables but, as these authors suggested, it is a simple and preliminary way to examine the behavior of the model variables.

Model sensitivity was estimated in a common situation of ephemeral gully occurrence in southern Navarre (Spain), considering a 0.355-ha watershed with an average slope of 0.02. The parameters of the model are gathered in Table 1. Flow hydrograph for an erosive winter storm was computed with the model KINEROS (Woolhiser et al., 1990). The rainfall lasted 150 min, with a total depth of 17 mm and a maximum intensity of 57 mm h^{-1}. Simulated upstream inflow was 2×10^{-4} m^3 s^{-1} and lateral uniform inflow of 1.75×10^{-4} m^3 m^{-1} s^{-1} along the 40 m of the channel side during 20 min, that was the estimated time to reach the peak flow rate. Table 1 shows the sensitivity coefficients (s), where total volume of soil lost is the output variable.

Table 1
Values of the main model parameters considered for the model sensitivity analysis and calibration, the corresponding sensitivity coefficient (s) and the ratio $\left|\frac{s}{s_{min}}\right|$, where $s_{min} = -0.06$

Parameter	Sensitivity analysis value	S	$\left\lvert\frac{s}{s_{min}}\right\rvert$	Initial calibration value
Length (m)	40	–	–	80
Longitudinal slope	0.02	1.20	20.00	0.047
Critical shear stress τ_c (Pa)	1.5	–	–	1.5
Erodibility k (min^{-1})	3.11	–	–	3.11
Shear stress coefficient C_τ	1.4	−0.06	1.00	1.4
Initial bed width B_0 (m)	0.25	−0.23	3.83	0.25
Wall slopes	0.07	0.36	6.00	0.125
Manning roughness coefficient n	0.05	−3.30	55.00	0.05
Water temperature (°C)	15	−0.66	11.00	15
Specific weight of soil particles γ_s (kN m^{-3})	15. 0	–	–	15.0
Bed porosity	0.42	–	–	0.42
Corey's shape factor	0.7	−1.09	18.17	0.7
Powers' roundness factor	3.5	−0.07	1.17	3.5
Particle size d (mm)	0.2	−3.62	60.33	0.2
Specific gravity of bed particles	1.8	−4.44	74.00	1.8
Coefficient of nonequilibrium transport α'	0	−1.48	24.67	0.5

The model is very sensitive to variables such as particle density, very important in aggregated soils, particle size, Manning's coefficient, transport coefficient and bed longitudinal slope, as shown in Table 1. These are the key parameters for the calibration process. The influence of coupled parameters as τ_c and k, or porosity and specific weight, is examined through the change of the input when both variables are modified. The model is not very sensitive to the critical shear strength τ_c and to the coefficients k and C_τ. Nevertheless, when the dependent or output variable is width or depth of the channel, the respective sensitive coefficients increase. On the other hand, variations in the initial bed width or wall slope do not yield any appreciable change in either volume of soil lost or cross-section shape. Once the model was calibrated with the parameters density and size of bed particles fitting total soil loss, the parameters critical shear strength and erodibility will be used to fit cross-section shape.

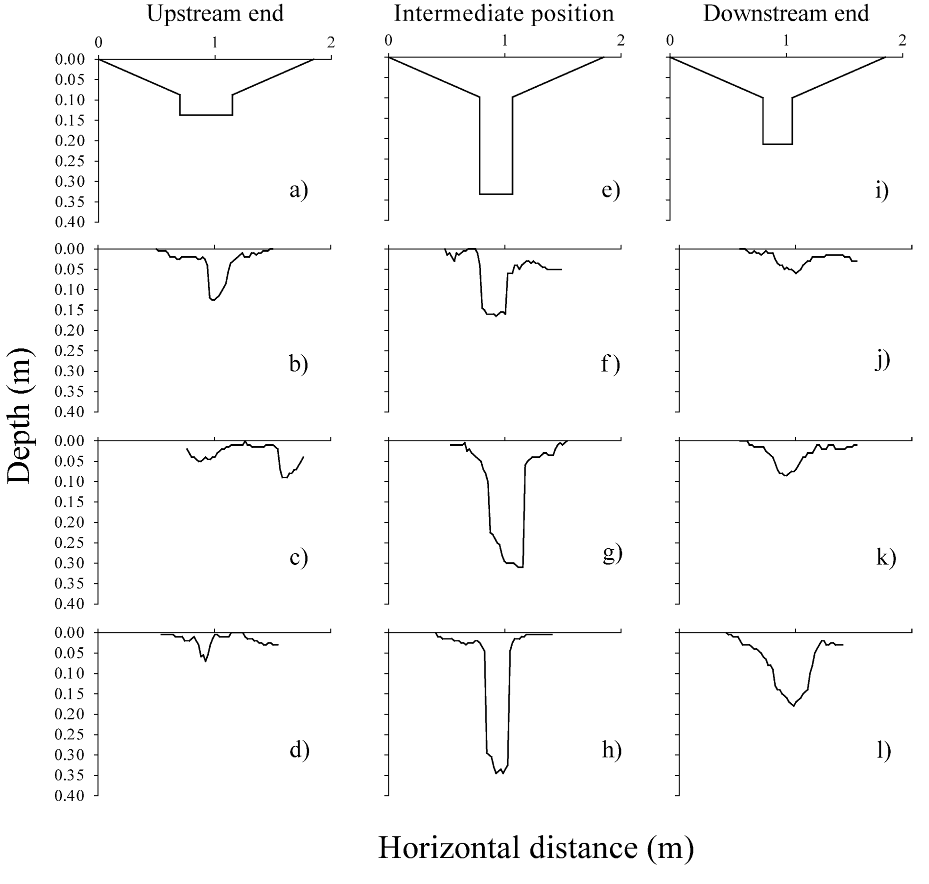

Fig. 5. Comparison of three simulated cross-sections, at the upstream end (a), in the middle of the channel (e) and at the downstream end (i), with three measured cross-sections at the same position: upstream, (b, c, d), in the middle, (f, g, h) and downstream (j, k, l).

3.2. Model calibration

The model was calibrated using data from an ephemeral gully that occurred in the Cobaza I watershed on January 22, 1996. This small watershed of 0.55 ha with an average slope of 5.2% located near the village of Pitillas in southern Navarre was dedicated to winter cereal. The climate is Mediterranean with a continental character. The soils are silty loams and highly susceptible to erosion. The rain event causing the gully was described in the sensitivity analysis. The channel of the gully had a length of 80 m, with a slope of 4.7% and average cross-section area of 0.050 m^2. The total volume of soil lost was estimated as 3.85 m^3 (Casalí et al., 1999). Runoff flow was computed using the model KINEROS. A constant upstream runoff flow of 3.14×10^{-3} m^3 s^{-1}, increased to a downstream flow of 1.54×10^{-2} m^3 s^{-1}, during a period of 20 min. The uniform lateral flow rate was 1.53×10^{-4} m^3 m^{-1} s^{-1}. The initial values of the parameters are shown in Table 1.

The calibration process yielded values of 0.7 Pa for the critical shear stress τ_c, 8.80 min^{-1}, for the erodibility coefficient k, 0.178 mm for particle size and a specific gravity of 1.75. The values for the critical shear stress and erodibility correspond to a soil not too susceptible to the erosion. In Fig. 5, three cross-sections computed with the model for the upstream end, the intermediate part and the downstream end are compared to cross-sections measured in the field. Maximum computed channel depth, 24 cm, is not far from the measured depth of 30 cm. Both computed and measured cross-sections are similar. The observed downstream degrading trend being reversed in the middle of the channel is reproduced with the model. This tendency is due to a decrease in transport capacity since the sediment load is increasing downstream. The good performance of the model was confirmed applying it to posterior events. After one winter rainfall of long duration and low intensity, the model predicted small losses of soil due to bed erosion without modification in the width as observed in the area. Similar observations were made by Poesen and Govers (1990).

4. Conclusions

The proposed model has simulated acceptably both the volume of soil lost and the shape of the eroded ephemeral gully channel with the estimated values of the parameters. This model is a promising tool for the management of ephemeral gully areas. Future refinements will include headcut occurrence and migration.

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