Dynamic and Static Models of Gully Erosion.

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Abstract

The main causes of gully formation are anthropogenic factors: the clearing of native forests, tilling of fallow lands and associated change of the hydrological conditions in the rainfall-runoff system. Gully channels formation is very rapid during the period of gully initiation, when morphological characteristics of a gully (length, depth, width, area, and volume) are far from stable. This period is relatively short, about 5 % of a gully's lifetime. The most part of a gully's lifetime its size is near stable, maximum value. These two stages of gully development led to two types of gully erosion models: 1) dynamic models to predict rapid changes of gully morphology at the first period of gully development; 2) static models to calculate final morphometric parameters of stable gullies.

The dynamic gully model is based on the solution of the equations of mass conservation and gully bed deformation. The model of straight slope stability was used for prediction of gully side walls inclination.

The static gully model is based on the assumption of final morphological equilibrium of a gully, when averaged for several years, elevations and width of gully bottom does not change. This stability is associated with a negligible rate both of erosion and sedimentation at the gully bottom. That means, that flow velocity is less than threshold value for erosion initiation, but is more than the critical velocity of wash load sedimentation.

The dynamic and static gully models were verified on the data on gullies morphology and dynamics from Yamal peninsula (Russia) and New South Wales (Australia).

Keywords: gully erosion; gully's morphology; model; Australia; Yamal peninsula

1. Introduction

The significance of gully erosion has been well documented. The volume of now existing gullies on the Russian Plain is about 4 10^9 m^3, i.e. about 4 per cent of the volume of soil erosion by the water since 1700 AD (Sidorchuk, 1995). In Australia with mainly pasture land the volume of gully erosion amounts to 14 10^9 m^3 (Wasson et al., 1996). At the Western Europe the part of ephemeral gully erosion can measure 30-40% and up to 80% of the total soil loss (Poesen et al., 1996). The main causes of gully formation are anthropogenic factors: the clearing of native forests, tilling of fallow lands and associated change of the hydrological conditions in the rainfall-runoff system. The main period of these processes is the last half of the 19th century in the USA, Australia and Russia, and to this period most of the gully systems initiation is attributed. The gullies destroy completely the fertile topsoil layer, and the surrounding lands are damaged with more severe sheet and rill erosion.

There are two main stages of gully development, which are controlled by different sets of geomorphic processes. At the first stage of gully initiation hydraulic (and combined thermal and mechanic action of the water on the soil at the areas with permafrost, so called thermoerosion) erosion is predominant at the gully bottom and rapid mass movement occurs on the gully sides. Gully channel formation is very intense during the period of gully initiation, when the morphological characteristics of the gully (length, depth, width, area, and volume) are far from stable. At the last stage of the stable gully sediment transport and sedimentation are the main processes at the gully bottom, its width increases due to lateral erosion, and slow mass movement transforms the gully sides. The experiments of Kosov, Nikol'skaya and Zorina (1978) on the gully formation in sands shows, that the first stage is relatively short and takes about 5 per cent of the gully's lifetime, but >90 per cent of the gully's length, 60 per cent of the gullied area and 35 per cent of the gully's volume are formed at this period. During the largest part of a gullies lifetime (last stage) it is morphologically nearly stable (fig.1). These two stages of gully development led to two types of gully erosion models: 1) a dynamic model to predict rapid changes of gully morphology at the first period of development; 2) a static model to calculate final morphometric parameters of the stable gully.

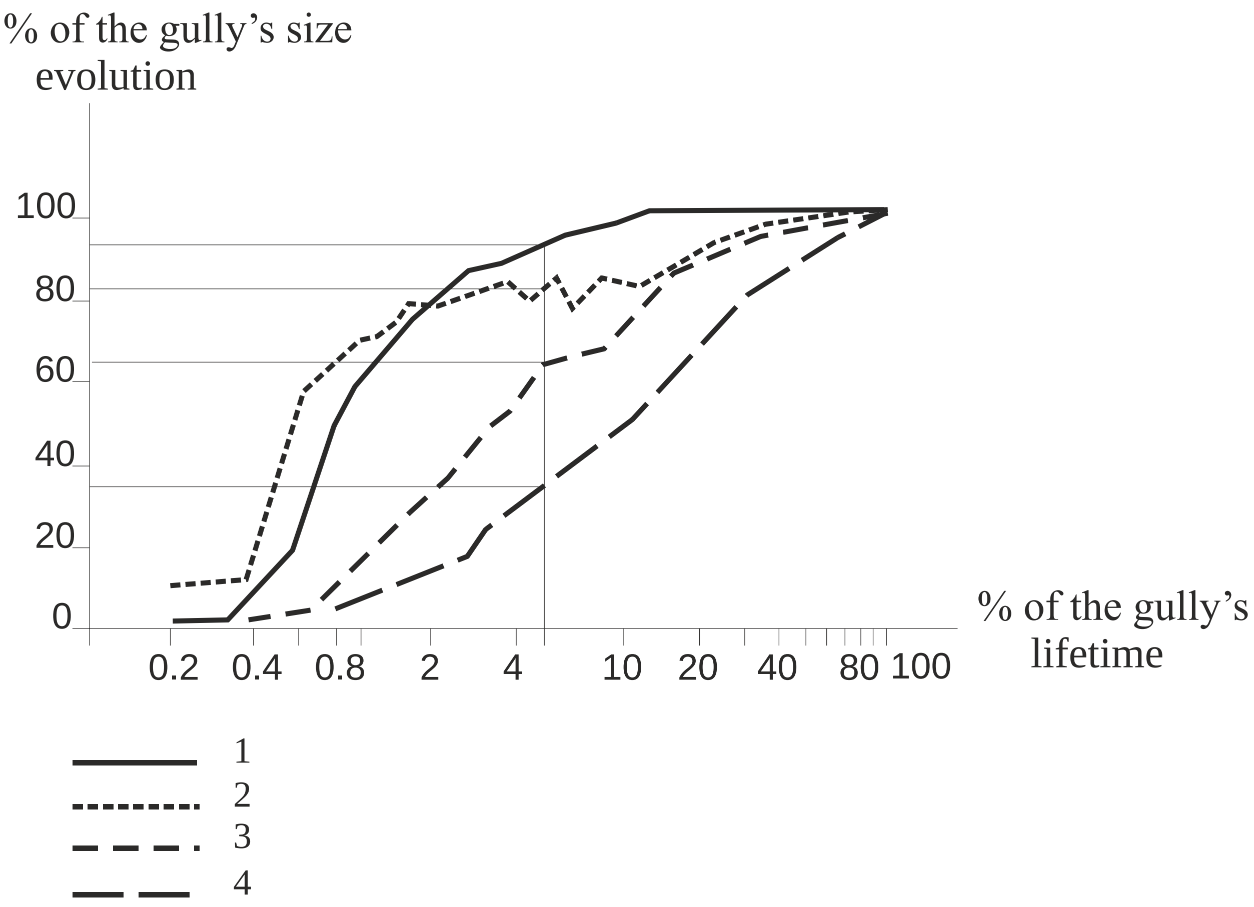

Fig.1. Evolution of the gully morphology during its lifetime (after Kosov et al, 1978). 1 = length; 2 = depth; 3 = area; 4= volume.

2. Methods

2.1 The dynamic gully erosion model

2.1.1. General description.

The model describes the first, quick stage of gully development. At this stage the following main processes occur:

a) During the snowmelt or rainstorm event the flowing water erodes a rectangular channel in the topsoil or at the gully bottom if the flow velocity is more than critical for erosion initiation.

b) The vertical walls of this trench can be unstable. Then shallow landslides transform a rectangular gully cross - section shape to trapezoidal along the period between adjacent water flow events.

The rate of gully incision is controlled by water flow velocity, depth, turbulence, temperature, and by soil texture, soil mechanical pattern, level of protection by vegetation. These characteristics are combined in equations of mass conservation and deformation, which can be written in the form

$$\frac{\partial Q_s}{\partial X} = C_w q_w + M_0 W + M_b D - C V_f W \qquad (1)$$

$$\left(1-\varepsilon\right) W \frac{\partial Z}{\partial t} = -\frac{\partial Q_s}{\partial X} + M_b D + C_w q_w \qquad (2)$$

Here $Q_s = Q\,C$ is sediment discharge (m^3/s), Q =water discharge (m^3/s); X= longitudinal co-ordinate (m); t=time (s); C= mean volumetric sediment concentration; C_w= sediment concentration of the lateral input from the contributing catchment; q_w= specific lateral discharge (m^2/s); M_0 = detachment rate of the soil particles at the gully bottom (m/s); M_b = detachment rate of the soil from the channel banks (m/s); Z=gully bottom elevations (m); W = flow width (m); D = flow depth (m); V_f = sediment particles fall velocity in the turbulent flow (m/s), ε= soil porosity. The left part of equation of mass conservation (1) defines the sediment budget in the channel reach. The right part of (1) defines the sediment flux (specific volumetric sediment discharge): the first term is lateral flux from the catchment, the second one is upward flux from the bottom, the third one is sediment flux from the banks, and the last one is downward flux (sedimentation). The equation of deformation (2) defines the change of gully bottom elevation according the sediment budget. The solution of these equations depends on form of the terms, which describe sediment fluxes.

2.1.2 Upward sediment flux

The analysis of the experiment results in the gullies of Yamal peninsula (north of the Western Siberia, Russia) and of New South Wales (Australia) shows, that in the conditions of steep bottom (with the slope 0.06 - 0.6) and cohesive soils, common for gullies, the rate of soil particles detachment is linearly correlated with the product of bed shear stress $\tau = g\rho DS$ and mean flow velocity U:

$$M_0 = kU \frac{\tau}{\tau_{cr}} \qquad (3).$$

Here S is gully bottom slope, D is flow depth, and g is acceleration due to gravity. Experiments show, that for loams and clays with the cohesion 20-40 kPa the coefficient k equals to 1.9 10^{-6}. If bed shear stress in the flow is less than its critical value for erosion initiation τ_{cr} , then M_0 =0.

Mirtskhulava (1988) showed, that critical shear stress τ_{cr} is mainly controlled by the forces of friction and cohesion:

$$\tau_{cr} = 1.2\lambda \ (m_1/n_1)\left[(\rho_s - \rho)gd + 1.25C_f^n K_0\right] \qquad (4).$$

Here λ is the coefficient of flow resistance: $\lambda = 0.18*(d/D)^{1/3}$; m_1 is equal to 1.0 for clean water flows, and is equal 1.4 for the flows with colloidal particles content more than 0.1 kg/m^3; parameter of turbulence n_1 is usually about 4; ρ_s and ρ are sediment and water density (kg/m^3); d - mean diameter of soil aggregates (m); K_0 - coefficient of variability of soil mechanical pattern, usually it is 0.5; C_f^n is soil fatigue strength to rupture and it is the function of soil cohesion C_h (Pa): $C_f^n = 6.7*10^{-7} C_h^2$ after our experiments, or $C_f^n = 0.035C_h$ after Mirtskhulava (1988).

The first term in square brackets of (4) represents the influence of friction on particle stability, and is of the main importance for noncohesive soils, the second term represents the influence of cohesion on particle stability, and is of the main importance for cohesive soils.

One of the sufficient factors of soil cohesion is the content of grass roots and of vegetation remnants in the soil, and τ_{cr} increase rapidly with grass root content in the topsoil. Thin (less that 1 mm in diameter) living and dead roots gather soil aggregates to each other and increase the soil cohesion. The field and laboratory experiments show that the bulk soil cohesion C_h increase rapidly with the content of thin roots R_0 (kg m^{-3}) in top 5 centimetres of the soil:

$C_h = C_0 \exp\left(0.05 R_0\right)$ (5).

Here C_0 is cohesion of the same soil, but without vegetation roots.

For the case of gully erosion in frozen soil or in soil with the permafrost (so called thermoerosion) the main factor of erosion became water temperature. Field and laboratory experiments of Poznanin, Malinovskiy and Dan'ko (see Sidorchuk, 1996) showed, that as a first approximation the soil detachment rate due thermoerosion M_{0t} is equal to the rate of soil thawing and linearly related with water temperature T°C:

$M_{0t} = k_{te}\mathrm{T}$ (6).

Formula (6) have to be used in (1) in the conditions, when $M_{0t}<M_0$ in the same flow. In this case the thaw layer washes out completely and the frozen soil is always exposed to melting action of the water flow. In the conditions $M_{0t}>M_0$ the thaw layer protects the frozen soil and ordinary erosion process occur, described by formula (3).

The coefficient of thermoerosion k_{te} value is about 5.2 10^{-5} for thin sands and 0.55 10^{-5} for loams, but its variability is rather high due to changes in soil cryogenic texture and ice content (Sidorchuk, 1996).

2.1.3. Gully bank erosion

The process of bank erosion by bottom flow in the gullies has not been satisfactorily investigated. It is assumed that rate of bank erosion dW_b/dt is equal to sediment flux from the banks M_b. Using an analogy with estimations of bank erosion in the rivers the expression

$M_b = M_0 V / U$ (7)

can be suggested as the first approximation. Here V is lateral velocity. For a curved channel Rozovskiy (1957) obtained a simple formula:

$V = 11.0 UD / R.$ (8)

The investigations in the gullies of Yamal peninsula show, that at the narrow incised gully bottom with gully bottom width $W_b < 10.0W$ the radius R of confined channel bends decrease when W_b increases: $R=50.0W(W/W_b)$. When W_b increases due to banks erosion and becomes > $10.0W$ the flow forms free meanders with $R=5.0W$. At the same time curved flow can wash only part of side walls and this part P_e decreases when the relative bottom width W_b/W increases:

$P_e = W/W_b$ when $W_b < 20.0W$

and

$P_e \cong 0$ when $W_b > 20.0W$.

After combining all these formulas the expression for calculation of gully bank erosion rate takes the form:

$$\frac{dW_b}{dt} = k_b M_0. \quad (9)$$

Here $k_b = 0.22D/W$ when $W_b < 10.0W$, $k_b = 2.2D/W_b$. when $W_b > 10.0W$ and $k_b = 0$ when $W_b > 20.0W$.

2.1.4. Downward sediment flux

The expression for downward flux (sedimentation rate) is rather simple and includes the product of fall velocity in turbulent flow and depth-averaged sediment concentration in flow. The fall velocity in the turbulent flow is less than Stokes fall velocity V_{st} in steady water or in laminar flow due to turbulent flow oscillations. As the first approximation the modificated formula of Hwang (1983) can be used for V_f calculation.

$$V_f = V_{st} \Big/ \left\{1 + \left[0.5U \big/ (9.0V_{st})^2\right]\right\} \quad (10).$$

In the case of thin particles and high turbulence V_f can be 0.

2.1.5. Equations for gully incision solution

After substitution of equations (1) and (3) into (2) it takes the form of transport equation for bottom elevations Z for the case of erosion process:

$$\frac{\partial Z}{\partial t} - a\frac{\partial Z}{\partial x} - V_f C = 0. \qquad (11)$$

Here $a = kg\rho q$, and $q=UD$ is specific discharge. The equation (11) can be solved numerically, for example with the aim of explicit predictor- corrector scheme of Lax-Wendroff:

$$Z_i^{j+1/2} = (1-\beta\ )Z_i^j + \beta\ Z_{i+1}^j - \alpha\ \frac{\Delta t}{\Delta x}\left[\frac{(aq)_{i+2}^j + (aq)_{i+1}^j}{2} Z_{i+1}^j - \frac{(aq)_{i+1}^j + (aq)_i^j}{2} Z_i^j\right]$$

$$Z_i^{j+1} = Z_i^j - \frac{\Delta t}{2\alpha\ \Delta x}\{\ (\alpha - \beta\ )\frac{(aq)_{i+2}^j + (aq)_{i+1}^j}{2} Z_{i+1}^j - (2\beta - 1)\frac{(aq)_{i+1}^j + (aq)_i^j}{2} Z_i^j +$$

$$(1-\alpha-\beta\ )\frac{(aq)_i^j + (aq)_{i-1}^j}{2} Z_{i-1}^j + \frac{(aq)_{i+1}^j + (aq)_i^j}{2} Z_i^{j+1/2} - \frac{(aq)_i^j + (aq)_{i-1}^j}{2} Z_{i-1}^{j+1/2}\ \} + V_f C_i \Delta t$$

The symbol 'i' represents the change by the length, symbol 'j ' - in time. The best fit values of net numbers α and β are: β = 0.75--1.0; α =0.25--0.5. For the explicit scheme stability the Courant number must be less than 1.0: $aq\,\Delta t/\Delta x \le 1$.

For the case of thermoerosion (with formula (6) instead of (3)) the equation for bottom elevations Z is much simpler:

$$\frac{\partial Z}{\partial t} + k_{te}T - V_f C = 0. \qquad (12)$$

For the sediment concentration C_i the solution of (1) on the flow reach with length Δx have to be used for erosion process:

$$C_i = \left(C_{i-1} - \frac{(k+k_b)Q_{i-1}S}{q_w(Y+1)} - \frac{C_w}{Y}\right)\left(\frac{Q_{i-1}}{Q_i}\right)^Y + \frac{(k+k_b)Q_i\ S}{q_w(Y+1)} + \frac{C_w}{Y} \qquad (13)$$

or, for thermoerosion process:

$$C_i = \left(C_{i-1} - \frac{k_b Q_{i-1}S}{q_w(Y+1)} - \frac{C_w}{Y} - \frac{k_{te}T}{q_w Y}\right)\left(\frac{Q_{i-1}}{Q_i}\right)^Y + \frac{k_b Q_i\ S}{q_w(Y+1)} + \frac{C_w}{Y} + \frac{k_{te}T}{q_w Y} \qquad (14)$$

Here Q_{i-1} and C_{i-1} are discharge and sediment concentration at the beginning of the reach with the length X_i-X_{i-1} , $Y = \left(q_w + V_f W\right)/q_w$.

The width and depth of the flow in gullies can be calculated with the empirical formulas:

$$W = 3.0 * Q^{0.4} \quad (15)$$

and

$$D = 0.48 Q^{0.45} \quad (16)$$

based on data from Yamal peninsula.

Water discharge Q and temperature T in the gully and lateral flux have to be calculated with a suitable hydrological and catchment erosion model, or obtained from measurements.

2.1.6. Gully side walls transformation

The side walls of the gully become practically straight after rapid sliding, following the incision. In this case a model of straight slope stability can be used for prediction of gully sides inclination. If the depth of incision D_v becomes more than critical value

$$D_{vcr} = \frac{2.0 C_h}{g \rho_s} \cos(\varphi) \Big/ \sin^2 \frac{1}{2}\left(\varphi + \frac{\pi}{2}\right) \qquad (17),$$

then gully walls inclination ϕ can be calculated with the help of the formula:

$$\frac{C_h}{g \rho_s D_v} = \frac{\rho_s - w\rho}{\rho} \tan(\varphi) \cos^2(\phi) - \frac{\sin(2\phi)}{2} \qquad (18).$$

Here w is volumetric water content in the soil, φ is the angle of internal friction.

When the bottom width, wall inclination and the whole volume of incision V_0 are known, the shape of the gully cross-section can be transformed into a trapezium with bottom width W_b, depth $D_t = \left(\sqrt{W_b^2 + \frac{4V_0}{\tan(\phi)}} - W_b\right) \frac{\tan(\phi)}{2}$ and top width $W_t = W_b + 2.0 D_t / \left[\tan(\phi)\right]$.

2.2. The static gully model.

The static gully model is based on the assumption of final morphological equilibrium of a gully's bottom and walls. When averaged for several years elevations of gully bottom Z and gully bottom width W_b do not change, then

$$\partial Z / \partial t = 0; \quad \partial W_b / \partial t = 0. \qquad (19)$$

For the case of gully erosion, stability is associated with a negligible rate both of upward sediment flux (erosion of the gully's bottom) and downward sediment flux (sedimentation on the gully's bottom). That means, that specific sediment discharge $q_s=QC/W$ of the flow does not change along the channel length X:

$$\partial q_s / \partial X = 0. \qquad (20)$$

This can be in ideal situation, when

$$V'_{cr} < U < V''_{cr} \qquad (21)$$

The flow velocity U is less than the threshold value for erosion initiation V''_{cr} and it is no erosion at the gully bottom. Together the flow velocity is more than the critical velocity V'_{cr} of wash load sedimentation, and thin sediments, eroded on the contributing catchment, are completely transferred through the gully.

This criterion of stability is widely used for irrigation channel design. For stable gully morphological characteristics estimation Zorina (1979) first used this criterion.

If Chezy-Manning formula is used for calculation of critical velocity of erosion initiation

$$V''^{2}_{cr} = \frac{SD^{4/3}}{n^2} \quad (22),$$

then stable slope S can be calculated from:

$$S = \frac{\left(V''_{cr}\right)^{2.67} n^2 \ \left(W/D\right)^{0.67}}{Q^{0.67}} \qquad (23)$$

Here Q- water discharge, D - flow depth, n- Manning coefficient.

The relative channel width W/D is the reverse function of a discharge: for the gullies of Yamal peninsula $W/D = 6.0Q^{-0.08}$. Manning roughness coefficient assumed to be constant for the gully flow and is near the value for small streams: n=0.03-0.04. The critical velocity of erosion initiation V''_{cr} is related to a texture of the deposits, into which the gully is incised (tab.1). It means that the main expression for gully stable slope estimation is reverse relation between slope and discharge:

$$S = \frac{K}{Q^m} \qquad (24).$$

Discharge is the function of contributing catchment area A: $Q=Y_0A$. Here Y_0 is specific discharge per unit area ($m^3s^{-1}km^{-1}$). A contributing catchment area is the function of the channel length: $A=f(X)$. For simple shapes of a gully basin (for example rectangular, triangular, leaf- shaped, etc.) this function can be analytical, but in general have to be tabulated. In this conditions the shape of the stable gully longitudinal profile can be calculated from the first order differential equation:

$$\frac{\partial Z}{\partial X} = \frac{6.0\left(V''_{cr}\right)^{2.67} n^2}{\left[Y_0 f(X)\right]^{0.75}} \qquad (25)$$

If the geological structure of the gully catchment is complicated and soil erodibility can change by the length of the flow and by the depth of incision, the value of V''_{cr} must be changed during the calculations at the points of texture change. The empirical versions of formulas (24)-(25) are well known from the works of Leopold et al (1964), Schumm et al. (1984) and others, where different functions for discharge (or channel length, basin area) were used.

The general problem in applying regime equations such as formula (25) is the choice of the channel forming discharge. Leopold et al. (1964) used bankfull discharge, Bray (1982) suggested to use flood discharge with 2 year return periods. Sidorchuk (1984) showed that the whole range of discharges with different return periods must be used in regime equations. It is well known from fluvial geomorphology, that each discharge produces channel transformation. The magnitude M_i of channel deformation during the flow with specific discharge Y_i is proportional to the product of specific sediment discharge Y_{si} and its duration or frequency P_i: $M_i = Y_{si}P_i$.

Using these assumptions the elevations of longitudinal profile of the stable gully, controlled by the whole range of discharges, are calculated in several steps:

1) The whole range of the specific discharges is divided into N intervals.

2) For each Y_i, related to the centre of i-th interval, the frequency P_i and specific sediment discharge Y_{si} is estimated.

3) The gully bottom elevations $Z (X)$ are calculated with the formula (25) for each of i-th specific discharges Y_i;

4) The range of N elevations Z_i for every given X is averaged with the weight equal to $Y_{si}P_i$:

$$Z_{0j} = \left(\sum_{i=1}^{N} Z_{ij} M_{si} P_i \right) \Bigg/ \left(\sum_{i=1}^{N} M_{si} P_i \right)$$

A case study of the longitudinal profile of Whitehead Creek gully (near Goulburn, NSW, Australia) shows, that averaged longitudinal profile, which takes into account whole range of gully forming discharges is most close to the measured one, and that non of the partial discharges can be used to calculate the longitudinal profile of the same shape (fig.2).

The shape of the cross section of the stable gully can be calculated using the same assumptions, as in dynamic gully model with $W_b = 20.0W$.

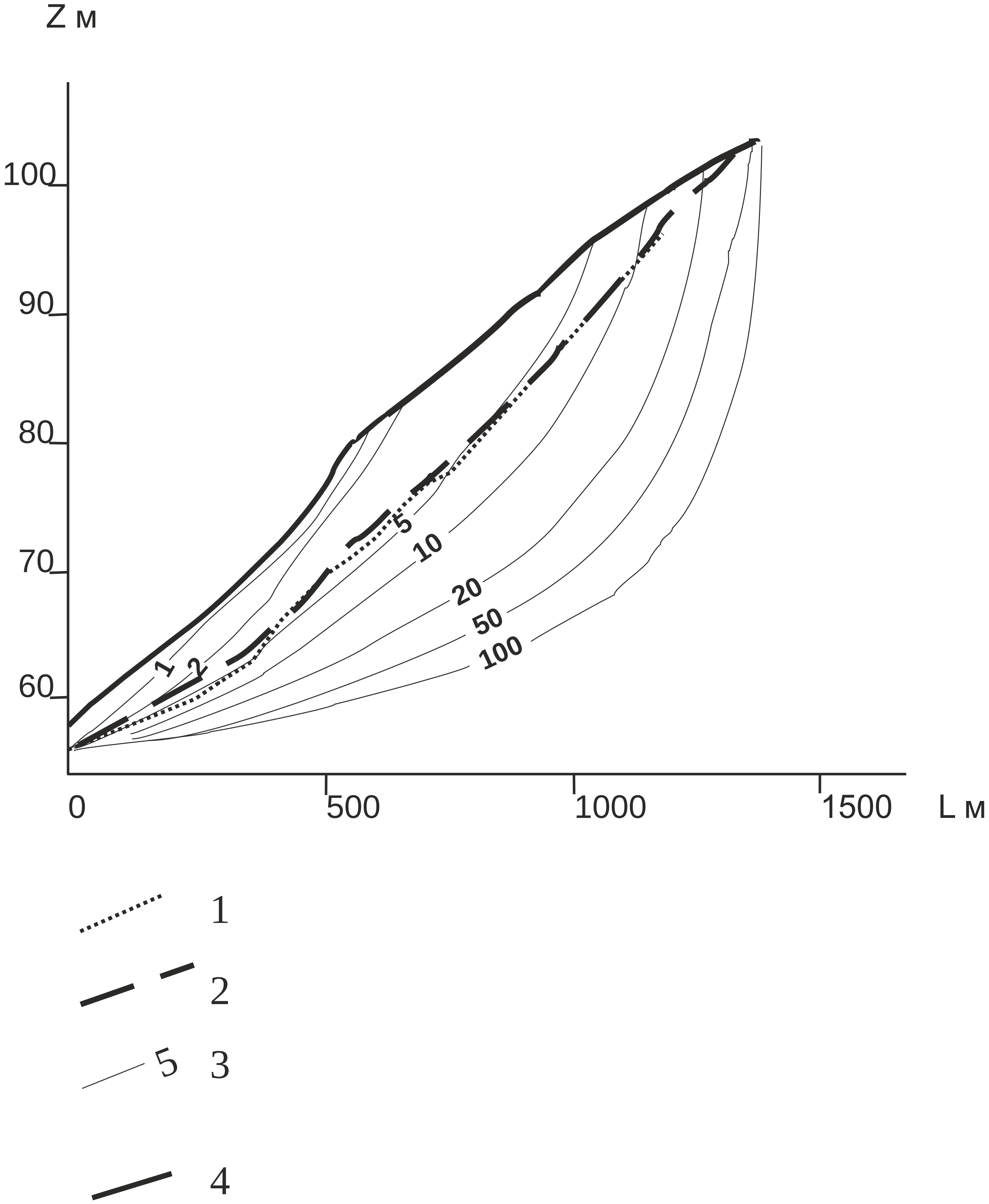

Fig.2. Measured longitudinal profile (1) of stable Whitehead Creek gully (near Goulburn, NSW, Australia) is close to averaged calculated longitudinal profile (2), which takes into account the whole range of gully forming discharges. Non of the stable profiles, calculated for partial discharges with different return periods (3) have the same shape, as measured one (in 1992). Here: 4 = initial profile of the slope.

3. Results and analyses

3.1. The dynamic gully erosion model verification.

The dynamic gully model was verified using data about gully development on the Yamal Peninsula, North Western Siberia. One of these gullies, for which both initial and actual longitudinal profiles are available, is situated at the right bank of Se-Yakha River. Before 1986 there was shallow linear depression with dense vegetation cover and ephemeral flow. In 1986 an exploitation camp was built in the upper part of the 0.3 km^2 large basin. Surface destruction and increase of melt water flow lead to intensive gully erosion. An 840-m long gully (measured along the gully valley) was formed (fig.3.). In 1991 and 1995 the longitudinal profile of the gully was

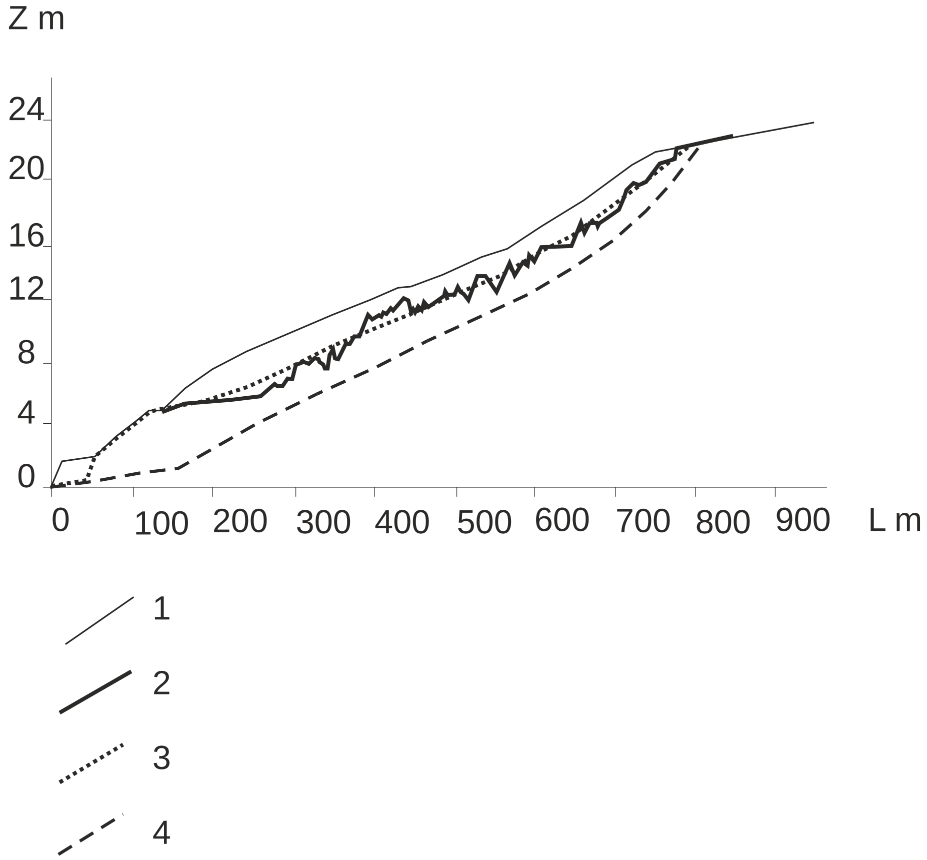

Fig.3 Dynamic gully erosion model verification: case study of longitudinal profile evolution and prediction for the gully on the Yamal peninsula (Russia). Here: 1 = initial profile of the slope; 2 =actual longitudinal profile in 1995; 3 = calculated longitudinal profile for 1995; 4 = prognoses of the longitudinal profile for year 2030.

investigated. The initial profile was available from the large-scale map. The depths of runoff for thaw and rainfall periods for 1986-1995 were calculated on the basis of meteorological data from the nearest station Marre Salye (90 km to south-west from gully site). Some corrections to these data were based on the meteorological and hydrological measurements of 1992-1993 at the gully basin. The coefficients k in formula (3) and k_{te} in formula (6) were calibrated with the data of 1986-1991 period. The calculated and observed altitudes of the gully bottom in 1995 are rather close. Numerical experiments were run to examine the model sensitivity to initial conditions. The solution is mainly controlled by the value of τ_{cr}, which is function of the soil aggregates size; soil cohesion and grass cover density. The next important factor is water discharge.

3.2. The static gully erosion model verification.

Calculations to verify the stable gully model were produced for several stable gullies in New South Wales, Australia. The case of Whitehead Creek gully was already mentioned. The other one, Keepit gully (near Guanahda, NSW) has basin area 0.46 km^2. The initial 1500 m long profile is straight (fig.4). The gully is incised into pebbly loams (V''_{cr}=1.5 m/s) up to 450 m from its mouth. At the short upper section the gully cuts schist with V''_{cr}=2.7 m/s. The terrace with relative altitude up to 1.5 m was formed in the gully. Beer bottles with the dates 1927- 1934 y. at their bottoms are abundant at the base of the terrace, so the deposition occurs in early 30-th. Now this terrace is eroded through the whole thickness of the sedimentation layer. That shows the dynamic equilibrium of the gully longitudinal profile. The discharges with 1, 2, 5, 10, 20, 50 and 100 year return periods were estimated from the maps, used by Soil Conservation Survey of New South Wales. The value of the critical velocity of erosion initiation was calibrated by fitting the calculated profile with stable sections of the measured one. The calculated longitudinal profile of the stable Keepit gully consists of two sections: more gentle in the lower section and more steep in the upper. The gully still has a potential to erode its middle part (fig.4).

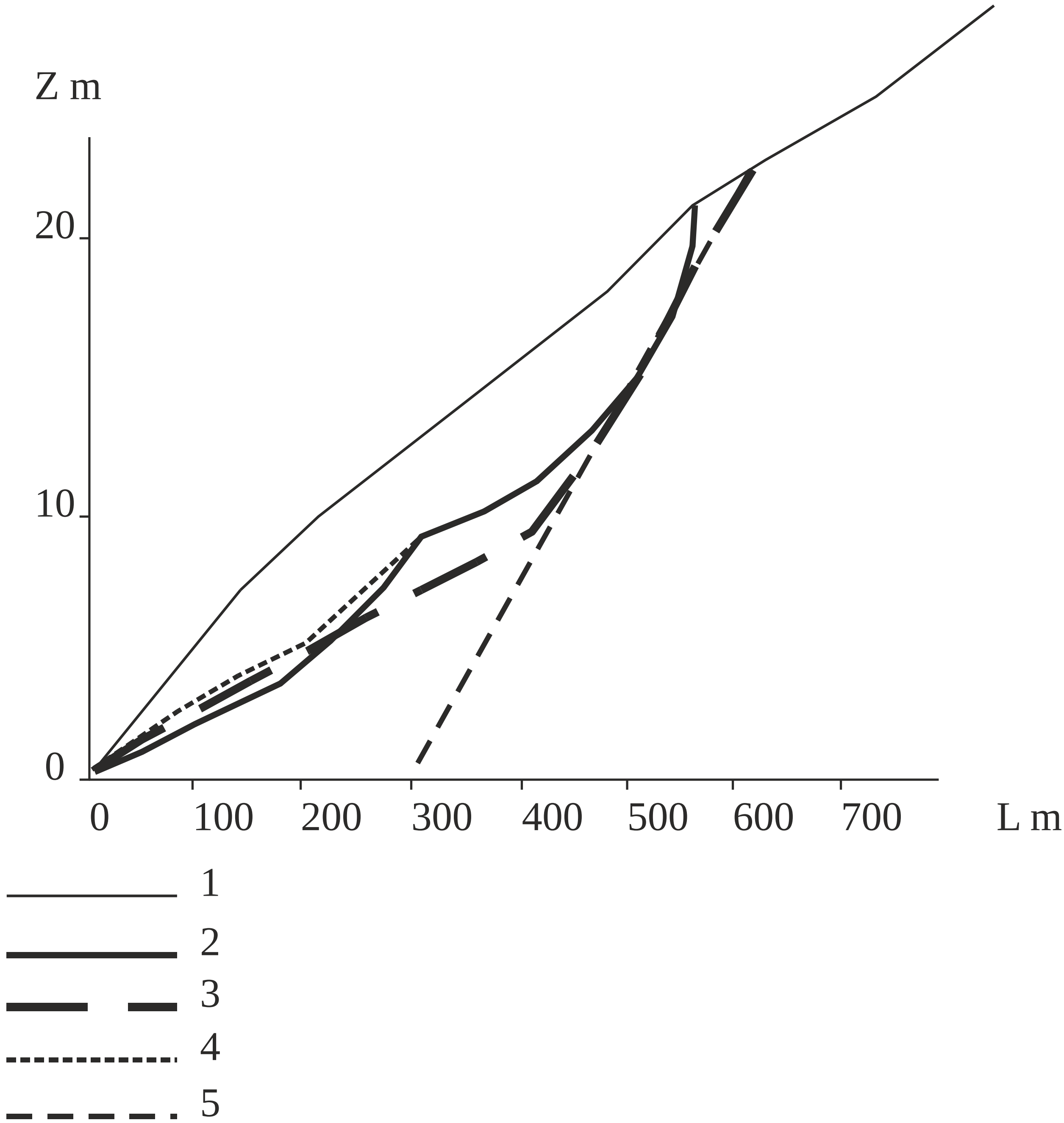

Fig.4 Static gully erosion model verification: case study for stable gully Keepit, in New South Wales, (Australia). Here 1 = initial profile of the slope; 2 =actual longitudinal profile in 1992; 3 = averaged stable longitudinal profile; 4 = terrace; 5 = surface of the rocks.

Numerical experiments were also run to examine the static model sensitivity to changes of parameters. The solution is mainly controlled by V''_{cr}, which is function of soil texture and by Manning coefficient (tabl.1). A very important factor is discharge values and distribution.

Table 1. Critical velocities of erosion initiation for the soils of different texture (after Bogomolov and Mikhaylov, 1972)

Soil texture	V''_{cr} (m/s)	Soil texture	V''_{cr} (m/s)
large boulders	4.00	loamy sand	0.60
small boulders	3.20	soft sandy loam	0.70
coarse gravel	1.10	hard sandy loam	1.00
medium gravel	0.90	soft loam	0.75
fine gravel	0.75	medium hard loam	1.00
very coarse sand	0.65	hard loam	1.15
coarse sand	0.60	soft clay	0.80
medium sand	0.57	medium hard clay	1.20
fine sand	0.32	hard clay	1.40
silt	0.55	very hard clay	1.70

4. Conclusion

These few examples show that dynamic and static gully models can be used for prediction of the evolution and of the finite morphology of the gullies. At the same time verification of dynamic and static gully erosion model showed, that their sensitivity to lithological and hydrological factors is rather high. Field investigations of gullies morphology and dynamics and careful calibration of the models are necessary for accurate prediction of gully erosion.

The dynamic gully erosion model describes the first, quick stage of gully development, which last about 5% of the gully lifetime. During the snowmelt or rainstorm event the flowing water erodes a rectangular channel in the topsoil or at the gully bottom. Change of the gully bottom elevations is controlled mainly by upward detachment of the particles from the bed and by sedimentation on the gully bottom. The analysis of experimental results shows, that the rate of soil particles detachment is linearly correlated with the product of bed shear stress and mean flow velocity: In this case basic equations can be written as transport equation and numerically solved. The vertical walls of this channel are unstable. At the period between water flow events shallow landslides transform quickly gully cross - section shape to trapezoidal. Numerical experiments show, that the model in whole describes the real process of gully longitudinal and cross-section profiles evolution in time and space. It is sensitive to change of the soil erodibility; so field investigations and careful calibration of the model are necessary for accurate prediction of gully erosion.

Static gully model describes the final morphology of the gully. It is based on the assumption that the flow velocity is less than the threshold value for erosion initiation, but is more than the critical velocity of wash load sedimentation. This criterion of stability means that the main expression for estimation of gully stable slope is the well-known reverse relation between slope and discharge. As the discharge is a function of the channel length, the shape of the stable gully longitudinal profile can be calculated from the first order differential equation. The whole range of discharges with different return periods must be used in calculation of the stable gully longitudinal profile. The magnitude M_i of channel deformation during the flow with specific discharge Y_i was assumed to be proportional to the product of sediment discharge Y_{si} and its frequency P_i: $M_i = Y_{si}P_i$. To obtain the elevations of the stable profile, which corresponds to whole range of water flow, the partial elevations Z_i, calculated for every given X with partial discharge Y_i, have to be averaged with the weight equal to $Y_{si}P_i$.

The static gully model verification shows that the model can be used for prediction of the finite morphology of gullies after careful calibration of lithological and hydrological erosion factors.

These two models were realised as programs for PC’s, which are available from GCTE soil erosion Network (Ingram et al, 1996).

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