ELSEVIER

Catena 50 (2003) 507–525

CATENA

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Gully erosion modelling and landscape response in the Mbuluzi River catchment of Swaziland

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Received 13 October 2000; received in revised form 3 October 2001; accepted 17 November 2001

Abstract

In southern African countries soil erosion and the related problems, such as water quality issues or decreasing soil productivity, are the main topics affecting the inhabitants of both rural and urban areas. Therefore, the attention has been recently placed on those problems related to soil erosion. This can also be documented by an increasing number of studies carried out on erosion and by the development and application of erosion models. Nevertheless, gully erosion phenomena have been widely neglected in erosion modelling. This is because the development of erosion models was focused on those regions with an intense agriculture typical of developed countries on the one hand, and because of the spatial and temporal heterogeneity of gully erosion processes on the other hand. This study regards the identification of gully erosion forms and processes in the Mbuluzi River catchment (Kingdom of Swaziland) by using the Erosion Response Units (ERU) concept. The following modelling of gully erosion was done through the stable gully model [Catena 37 (1999) 401]. The input data were obtained through the application of remote sensing techniques (API method) and GIS-analyses. The example from Swaziland shows that the applied methods are able to identify areas affected by gully erosion. Furthermore, it is possible to estimate the amount of soil loss due to gully erosion, which, for example, is not taken into consideration by the USLE-type models.

Keywords: Gully erosion; Erosion modelling; Erosion Response Units; Swaziland; South Africa

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0341-8162/03/$ - see front matter
PII: S0341-8162(02)00123-6

1. Introduction

The improvement of the regional strategic planning of catchment water resources is one of the main tasks for managers and decision makers in developing countries. Therefore basic information on sediment production, transport and deposition is needed. As stated by Botha (1996), Morgan et al. (1997), Felix-Henningsen et al. (1997) or Beckedahl (1998), gully erosion processes are very important in many areas of Southern Africa because they are very effective sediment sources. Consequently the description of the river basin sediment budget has to consider these processes. This study investigates the distribution of gully erosion processes, their related features and their modelling in the Mbuluzi River catchment (Kingdom of Swaziland).

An innovative approach characterizing erosion processes and their integrated dynamics was introduced by Märker et al. (2001) and Flügel et al. (1999) through the concept of Erosion Response Units (ERU). This concept takes into account the active erosion forms and their related processes. Therefore, the ERUs are more than hydrologically defined units such as the erotops introduced by Richter (Kertesz et al., 1995).

The ERUs are defined as: "distributed three-dimensional terrain units, which are heterogeneously structured; they each have homogeneous erosion process dynamics characterized by a slight variance within a unit, if compared to neighbouring ones, and they are controlled by their physiographic properties, and the management of their natural and human environment". When applied as homogeneous erosion modelling entities, ERUs transform the precipitation system input into a corresponding runoff (surface, subsurface) and thereby generate specific erosion and sediment transport as a system output (Märker et al., 2001).

According to this definition, the drainage basin is conceived as an assembly of spatial process entities with different erosion potentials. The latter are in turn determined by the configuration of their natural capital and the respective human management.

The present morphologic features resulting from the erosion process dynamics can be used as a first approximation in the delineation of ERUs (Märker et al., 2001) in order to describe different erosion processes. Once the ERUs have been delineated, these entities can be applied for spatial scale transfer in regional erosion modelling as they conserve their properties. In this study, ERUs were applied to characterise the distribution of gully erosion features and processes and subsequently as a modelling entity. Gully erosion is completely neglected in traditional models such as the USLE (Wischmeier and Smith, 1978). Therefore, a physically based model that has been developed for similar environmental conditions in Australia (Sidorchuk, 1999), was applied. Model validation and verification were carried out by using detailed information on gully system time series and by ground survey.

2. Study area

2.1. Physiography of the Mbuluzi River catchment

The Mbuluzi River originates from the Ngwenya hills in Swaziland and passes through the North-Central part of the country into Mozambique. It runs through all the physiographic regions of Swaziland and drains an area of about 3100 km^2, from the border with Mozambique westwards (Fig. 1). The Highveld area (1500–1066 m a.s.l.) is characterised by steep slopes with average gradients exceeding 18%. The Middleveld altitude ranges from 760 to 610 m m.s.l. with average slopes of 12%. Gentle relief and average slopes of 3% were observed in the Lowveld (364–125 m a.s.l.). The mean annual rainfall ranges from 700 to 1200 mm (905 mm; Kwaluzeni), with the main rainfall in summer (October to March). Kiggundu (1986) calculated a rainfall erosivity of 450 kJ mm/m^2 h (EI_{30} after Wischmeier and Smith, 1978).

The geology of the upper Mbuluzi catchment is dominated by granites, and some areas of Precambrian sediments and volcanic outcrops. Granite and granitic gneisses with outcrops of dolerite and gabbro can be found in the Middleveld. The Lowveld area is composed of sedimentary and volcanic rocks of the Karroo supergroup.

Highveld and Middleveld soils are typically deep, acid and well drained red and yellow ferrisolic and ferralitic soils, often with stone lines indicating old erosion surfaces. In the lower Middleveld grey or red, light textured soils from granite and gneiss are generally found. The Lowveld is characterised by weathered red, brown and black clays originating from basaltic rocks. The land cover in the upper parts of the Mbuluzi River basin is mainly rangeland and bushland with some small-scale farming and subsistence cultivation. The lower part is dominated by intensive sugar cane plantations with irrigation and bush lands in the Lebombo region.

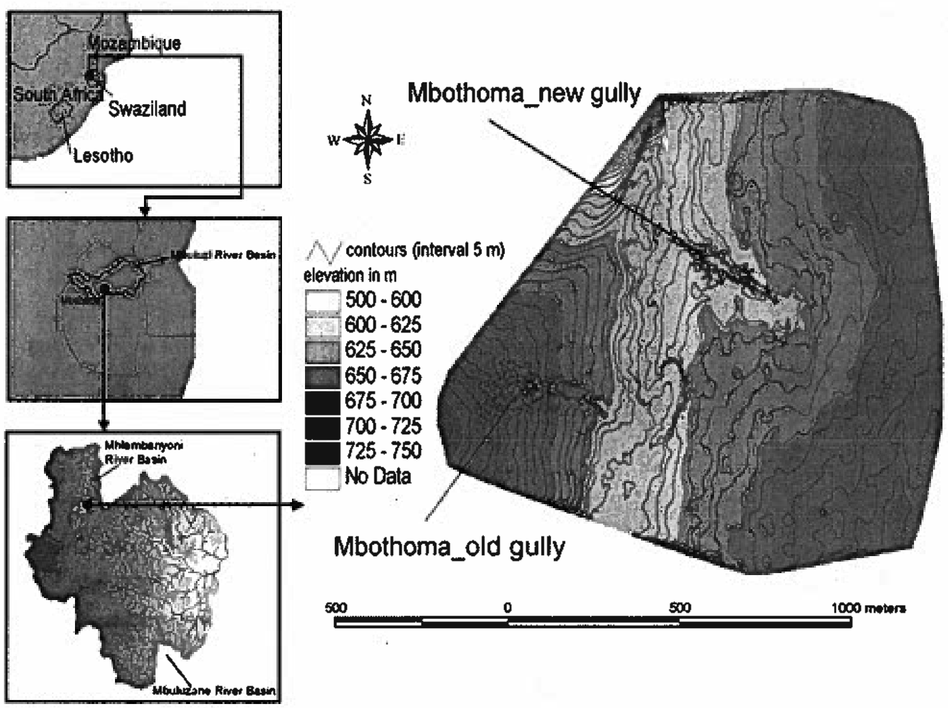

Fig. 1. Study area of the Mbuluzi catchment and DEM and DEM of the Mbothoma Gullies (Swaziland).

The study catchments are drained by the upper Mbuluzane River (contributing area: 221 km^2) and the Mhambanyoni River (contributing area: 42 km^2). Both rivers are tributaries of the Mbuluzi River (Fig. 1). Within the Mhlambanyoni test catchment the Mbothoma area, ca. 15 km north of Manzini (26°20′ S; 31°23′ E) was chosen as a characteristic test site for gully erosion studies. It is a densely populated area and overgrazing is widespread. The dominant land use on this subsistence/small-scale farming land is pasture. The lithology is composed of a thick granodioritic saprolite layer and a system of amphibolite and serpentinite dykes (Hunter et al., 1984).

3. Materials and methods

3.1. Identification of erosion sites and forms

The ERU delineation in the Mbuluzi River catchment was carried out through the analyses of aerial photograph stereo-pairs, orthophotos and GIS. The background for these analyses is provided by a field mapping campaign and photo documentation. The erosion type, its degree and extent, as well as the density of its features were mapped based on 1996 aerial photographs at 1:30,000 scale in order to classify the features and the subsequent delineation of erosion units. For carrying out the analyses the adapted method of Van Zuidam (1985) was applied. The results of these analyses are terrain entities subject to different levels (six classes) of erosional processes (Table 1).

In the following delineation of ERUs these classes are used as reference units (ERefU). These ERefUs consist of a particular erosion process or a combination of processes that are related to certain erosion features or combinations of them.

Once the erosion sites and the erosion processes with the related features have been collected, the soil loss due to the identified erosion processes was estimated.

Fig. 3 shows the ERefUs of the Mhlambanyoni subcatchment where mainly rill–interrill erosion and deep linear erosion (gully erosion) occur. The soil loss due to rill–interrill erosion processes was predicted by the USLE. Gully erosion processes require a separate modelling procedure and cannot be simulated by the USLE. The stable gully erosion model was applied (Sidorchuk, 1999) to estimate the total soil loss due to gully erosion and the maximum area affected by gully erosion.

Table 1
Erosion reference units classification

	Erosion classes	Erosion processes and related features	% Vegetation cover	% Degradation
1	No erosion		>90	<10
2	Slight erosion	Interrill, rill, shallow gully	>75	<25
3	Moderate erosion	Interrill, rill, shallow–medium deep gully	>75	<25
4	Moderate erosion	rill, medium deep gully	51–75	25–49
5	Severe erosion	rill, medium deep, deep gully, landslides	26–50	50–74
6	Severe erosion	rill, deep gully, badlands severe mass movements	<25	>75

3.2. Stable gully model

The model used in this study calculates the final development stage of the gully flowline network. It is based on the assumption that gully bottom and gully walls reach a final morphological equilibrium. Once this equilibrium has been reached the elevations of gully bottom Z and gully bottom width W_b do not change any more:

$$\partial Z/\partial t = 0; \tag{1}$$

$$\partial W_b/\partial t = 0.$$

In this case, the term "stability" is associated with both the erosion of the gully bottom and the sedimentation on it at a negligible rate. This implies that the specific sediment discharge $q_s = QC/W$ (Q is water discharge, C is sediment concentration in the flow, W is the flow width) does not change along the channel length X:

$$\partial q_s/\partial X = 0. \tag{2}$$

This is given in an ideal situation, when

$$V'_{cr} < U < V''_{cr}. \tag{3}$$

The flow velocity U is less than the threshold value for erosion initiation V_{cr}'' and therefore no erosion takes place on the gully bottom. On the other hand, the flow velocity is greater than the critical velocity V_{cr}' of wash load sedimentation. Consequently fine sediments, eroded within the contributing catchment area, are completely transferred through the gully system.

If the Chezy–Manning formula is used for the calculation of the mean critical velocity of erosion initiation, then the stable slope S can be calculated from:

$$S = \frac{(V''_{cr})^{2.67} n^2 (W/D)^{0.67}}{Q^{0.67}} \tag{4}$$

where Q = water discharge; D = flow depth; n = Manning roughness coefficient; W = flow width.

The relative flow width W/D is a function of sediment texture and discharge. The Manning roughness coefficient is assumed to be constant for the gully flow. The depth-averaged critical velocity of erosion initiation V_{cr}'' is related to flow depth D and to the texture of the deposits into which the gully is incised. If the flow is turbulent and characterised by a logarithmic velocity distribution, then, following Mirtskhulava (1988),

$$V''_{cr} = \left(\log_{10}\frac{8.8d}{D}\right) V_{\Delta}. \tag{5}$$

Here $V_{\mathrm{Delta};}$ is the bed critical velocity of erosion initiation and d is the mean diameter of soil particles (m). A flow depth D is related to flow discharge and can be calculated through the formula (Sidorchuk, 1999)

$$D = 0.48Q^{0.45}. \tag{6}$$

Discharge is a function of the contributing catchment area A: $Q = YA$. Here Y is the specific discharge (m^2/s). The contributing catchment area is a function of the flow line length: $A = f_1(X)$. Specific discharge can change at the catchment and therefore is also a function of flowline length: $Y = f_2(X)$.

The general problem in applying regime equations such as formula (4) is given by the choice of the channel forming discharge. It is well known, from fluvial geomorphology, that each discharge produces some transformation in the channel. The magnitude M_i of channel deformation during the flow with discharge Q_i is proportional to the product of sediment discharge Q_{si} (or its surrogate) and its duration or frequency P_i: $M_i = Q_{si}P_i$. To obtain the stable slopes of a gully profile, corresponding to the whole range of water flow, the partial slopes S_{ij} (calculated for every given X_j with the partial discharge Q_i) have to be averaged with the weight equal to $Q_{si}P_i$:

$$S_{0j} = \left(\sum_{i=1}^{N} S_{ij}M_{si}\right) \Big/ \left(\sum_{i=1}^{N} M_{si}\right). \tag{7}$$

According to these assumptions the shape of the stable gully longitudinal profile can be calculated from the first order differential equation:

$$Z_j = Z_{j-1} + \frac{1}{\sum_{i=1}^{N} M_{si}} \sum_{i=1}^{N} \left\{ \frac{a(V''_{\mathrm{cr}})^{2.67} n^2 M_s}{[f_1(X) f_2(X)]^b} \right\}_i \mathrm{d}X. \tag{8}$$

The values of the coefficients a and b depend on the relationship of the relative flow width with the discharge. The depth of the stable gully D_g will then be

$$(D_g)_j = Z_{0j} - Z_j. \tag{9}$$

Here Z_0 is the elevation of the initial surface, which will be cut by the gully.

The bottom width of a stable gully W_b is more than $20W$, where W is the width of the flow. It can be generally calculated from the empirical relationship with the discharge, and averaged for the whole sequence of discharges with the same weight, as the stable slope:

$$W_{bj} = \left(\sum_{i=1}^{N} pQ_{ij}^{f} M_{si}\right) \Big/ \left(\sum_{i=1}^{N} M_{si}\right). \tag{10}$$

The shape of the cross-section of the stable gully is usually trapezoidal. The top width of the stable gully W_t can then be calculated by using the stable bottom width W_b, the stable gully depth D_g and the stable gully wall inclination ϕ

$$W_t = W_b + 2.0\frac{D_g}{\tan\phi}. \tag{11}$$

3.3. Input data for stable gully model

The input information required to run the stable gully model consist of geomorphologic and geological data, derived from terrain morphology, lithologic composition (including vegetation cover), and hydrological information. They are obtained from measurements at the hydrological stations or calculated from meteorological data.

3.3.1. DEM analysis

To obtain the morphological input data, high resolution DEMs were derived from the gullies of the Mbothoma test area ca. 15 km north of Manzini with a horizontal resolution of 1 × 1 m. A photogrammetric stereoanalyser (Planicomp P33, Carl Zeiss Jena) was utilised to obtain digital elevation data from the 1940s, 1960s, 1970s, 1980s and 1990s aerial photograph series. The georeferencing was done with 1:5000 scale orthophoto maps.

DEMs were used for flowline network evaluation. A procedure was drawn up for the filling of closed depressions. One of eight possible flow directions following the maximum gradient was used for flow path estimation. The possibility of setting the preferable direction was provided to allow the estimation of the influence of out-of-scale features such as small roads or ploughed furrows. The terrain gradient at a point or pixel was calculated from two elevations, taking the point itself and the lowest neighbouring point. The catchment area of each point was calculated as the sum of the pixel areas following the gradient or flowline linked to this point upwards. For the Mbothoma test gully basins the following stratigraphic and morphometric features were delineated for each flowline from the DEM:

- subcatchment area;
- subcatchment length;
- profile of the initial surface of the basin (distance and altitude) along the flowline;
- contributing drainage areas for all points along the flowline (distance and area);
- initial profile of the surface (distance and altitude) of each lithological layer along the flowline.

These parameters are not only used in static gully erosion model but also in dynamic gully modelling (Sidorchuk and Sidorchuk, 1998; Sidorchuk, 1996, 1998, 1999). Fig. 4 shows the DEM of the Mbothoma gully system (initial surface) and the calculated drainage lines. The lithological layers were delineated by the triangulation of field-measured sample spots and by taking into account ground truth information. The lithologic information for each layer (Table 2) was obtained from our own analyses and from data in

Table 2
Physical parameters of the lithological layers at Mbothoma gully catchments

Physical parameters	Topsoil (~ 0.2 m deep)	Subsoil (up to 1.5 m deep)	Saprolite (more than 60 m deep)
Sand (%)	30	40	50
Silt (%)	20	30	40
Clay (%)	50	30	10
Bulk density (g/cm^3)	1.1–1.5	1.1–1.5	1.2
Cohesion (kPa)	4.5–9.0	4.5–9.0	3.14
Saturated hydraulic conductivity (cm/day)	0.5–1.8	0.5–1.0	1.0–4.4

the literature (Hunting Technical Services, 1983; Murdoch, 1970; Mushala et al., 1994; Scholten et al., 1995; WMS Associates, 1988). The V''_{cr} value generally has to be changed along the flowlines during the calculations at the points of texture change. As the geological structure of the Mbothoma gully catchment is not very complicated (saprolite), the soil erodibility and the V_{cr}'' value do not change along the flowlines or through the depth of incision.

3.3.2. *Runoff estimation*

As the stable gully model uses the entire range of the discharges, forming the gully cut, annual maximum runoff values and their probability have to be estimated. Therefore, runoff was simulated for a period of 45 years with the "agrohydrological modelling system" (ACRU) (Smithers and Schulze, 1995). The ACRU model uses an adapted SCS procedure (USDA, 1985; Schulze et al., 1993), designed to use daily rainfall input as the driving mechanism. The SCS stormflow routine is based on the principle that runoff potential is considered as an inverse function of the soil's relative wetness (Smithers and Schulze, 1995). The modelled runoff data have a daily time scale resolution and have been provided (Department of Agricultural Engineering, University of Pietermaritzburg) for three subcatchments of the Mbuluzi River. For the Mbuluzane River catchment (at GS3 runoff gauging weir), the regression between observed and simulated runoff shows a r^2 of 0.84. Finally by using these data, the annual maximum runoff depths (mm) and their probabilities (%) were calculated for the smallest subcatchment (the Mhlambanyoni River basin with an area of 41.6 km^2).

The empirical probability density of annual maximum runoff was approximated with the help of the Maximum Likelihood Method as gamma-distribution:

$$P(Y) = \frac{100}{b\Gamma(c)}\left(\frac{Y}{b}\right)^{c-1}\exp\left(-\frac{Y}{b}\right). \tag{12}$$

Here Γ is the Gamma function, b is the scale parameter, and c is the so-called shape parameter. This approximation was not only carried out for the whole flow (Y=base flow+quick flow; b=1.411; c=6.06) but also for the quick flow itself (Y=quick flow; b=1.042; c=7462). This procedure permits to estimate the probabilities of runoff depths, which are greater than the observed maximum (28.6 mm).

These values (Table 3) have to be used in formulas (8) and (10). The upper part of the slopes (above 650–655 m) is under Hortonian-type runoff, and only quick flow was used for these areas. For the lower part of the slopes, where base flow is formed by ground water seepage, the whole runoff was used to calculate stable gully morphology. In both cases the maximum of the magnitude M_i of channel deformation (which is proportional to the product of the surrogate of sediment discharge $Y_i^{1.5}$) and its frequency P_i: $M_i \sim Y_i^{1.5}P_i$, correspond to the lowest runoff values with the highest probability.

3.3.3. Evaluation of empirical relationships

In the stable gully model two main empirical relationships are used: width/depth ratio (in the flow) versus discharge and stable bottom width versus discharge. The measurements of the flow width and depth in the gullies during runoff events, which are rather usual in areas of snowmelt erosion, are rarely available in semiarid environments with storms. In these conditions the measurements in rivers and creeks with perennial flow can be carried out. The comparison of empirical data from different authors shows a vast variability. For example, Bray (1982) obtained a rather significant

Table 3
Annual maximum runoff depth (mm) and their probabilities (%) for the Mbulzi catchment empirical (P_{em}) and calculated through the gamma-distribution (P_c)

Annual maximum runoff depth (mm)		Probability (%) of the whole runoff (base flow + quick flow)		Probability (%) of quick flow	
Lower limit	Upper limit	P_{em}	P_c	P_{em}	P_c
0	2	9.8	13.8	19.6	21.8
2	4	21.6	16.9	23.5	17.7
4	6	19.6	15.1	11.8	13.8
6	8	7.8	12.5	7.8	10.7
8	10	11.76	9.9	9.8	8.3
10	12	5.9	7.8	9.8	6.4
12	14	5.9	6.0	0.00	4.9
14	16	0.00	4.6	0.00	3.8
16	18	2.0	3.5	5.9	2.9
18	20	5.9	2.6	3.9	2.2
20	22	3.9	2.0	2.0	1.7
22	24	0.00	1.5	0.00	1.3
24	26	2.0	1.1	2.0	1.0
26	28	0.00	0.8	0.00	0.8
28	30	3.9	0.6	3.9	0.6
30	32	0.00	0.4	0.00	0.5
32	34	0.00	0.3	0.00	0.4
34	36	0.00	0.2	0.00	0.3
36	38	0.00	0.12	0.00	0.2
38	40	0.00	0.1	0.00	0.2
40	42	0.00	0.1	0.00	0.1

increase in the width/depth ratio with 2-year flood discharge (m^3/s) for the rivers of Alberta (Canada):

$$W/D = aQ_2^{0.2}.$$

The empirical coefficient a varies from 3.4, for channels with silt and clay, to 4.8 for channels with sand and gravel. In the discussion of Bray's paper, Parker (1982) showed the regressions of W and D versus bankfull discharge (in dimensionless form) for four sets of data. These regressions give significantly different relationships of W/D versus dimensionless discharge $Q_d = Q/(d^{2.5}\sqrt{g(s-1)})$: $W/D = 15.8Q_d^{-0.007}$ Britain, single-channel; $W/D = 22.5Q_d^{0.043}$ Alberta, single-channel; $W/D = 24.6Q_d^{0.036}$ Alberta, braided anabranches; $W/D = 164.6Q_d^{-0.23}$ laboratory model, braided anabranches.

For the gullies in the Yamal peninsula, eroded in frozen loams, Sidorchuk (1999) obtained a regression of $W/D = 6.0Q^{-0.08}$ For the gullies in the Russian Plain, eroded in silty loams, Zorina (1979) uses a constant value of a width/depth ratio equal to 10. Hannam (1983) estimated W/D ratios for the lower part of the stable Poor Man's gully (New South Wales, Australia), which, was incised into ancient alluvial deposits. He obtained values in the range 3.0–26.0, with the mean equal to 11. In the gullies of the Snowy River basin (Victoria, Australia), eroded in granite saprolite (very similar to the ones in the Mbothoma area), the mean W/D ratio, measured by Sidorchuk and Fogarty for five gullies, is 8.4 (unpublished data). Due to the great difference in regime formulas and the absence of local empirical data, the latter value will be used for the calculations of the stable Mbothoma gully. According to the conditions of the Mbothoma gully system, formula (6) can be written in the following form:

$$Z_j = Z_{j-1} + \frac{8.4\cdot}{\sum_{i=1}^{N} M_{si}} \sum_{i=1}^{N} \left\{ \frac{(V''_{cr})^{2.67} n^2 M_s}{[f_1(X) f_2(X)]^{0.67}} \right\}_i \mathrm{d}X. \tag{13}$$

The empirical relationship between stable gully bottom width and discharge can be obtained from the morphology and hydrological data of the Old Mbothoma gully, which has had a stable morphology, at least for the last 30 years. The stable bottom width W_b (m) of this gully increases with the contributing area A (m^2) as follows:

$$W_b = 0.5A^{0.3}. \tag{14}$$

3.3.4. Evaluation of empirical parameters

There are three main empirical parameters in the stable gully model: critical flow velocity V_{cr}'', flow bed roughness coefficient in Manning formula n, and inclination of the stable gully walls ϕ. The accurate estimation of these parameters is important, as the model is very sensitive to their values. Generally, these characteristics can all be evaluated through the relevant formulas or tables (Sidorchuk, 1998, 1999). Here, the presence of a stable gully system within the modelled basin (Mbothoma area) implies the possibility of estimating (calibrate) these values through the backward solution of the stable gully model for the existing stable Old Mbothoma gully or of measuring them directly in the field or on the map.

3.3.4.1. Measurement of stable gully walls inclination. Based on the case of the Mbothoma gully system, the inclinations of the stable gully side walls (angle of repose) were empirically estimated from the measurements of the slopes in the old stable Mbothoma gully. At the lower 200 m of the gully, the mean inclination of the walls is $\phi = 49.2°$ (with standard deviation 7.4° for 35 measurements).

3.3.4.2. Calculation of critical velocity of erosion initiation. The whole procedure of stable gully morphology calculations was applied to the basin of Old Mbothoma gully. For each flowline, distance and altitude of all the points along the flowline were determined by using the HRDEM of the observed initial surface. In addition, the distances between the various points and the gully mouth and the contributing drainage area of each point were also determined. With the runoff values from Table 3, the stable longitudinal profiles were calculated for each flowline by using formula (13). The value of the expression $(V_\Delta)^{2.67}n^2$ was chosen to fit the minimal difference between observed and calculated stable longitudinal profiles. It is equal to 6.35×10^{-5}. A critical bed velocity can be calculated through the formula of Mirtskhulava (1988)

$$V_\Delta = 1.25\sqrt{\frac{2m_1}{2.6\rho n_1}[(\rho_s - \rho)gd + 1.25C_f^n K_0]}. \tag{15}$$

Here m_1 is equal to 1.0 for clean water flows, and is equal to 1.4 for the flows with a colloidal particle content of more than 0.1 kg/m^3; the parameter of turbulence n_1 is usually about 4; ρ_s and ρ are sediment and water density (kg/m^3); d is the mean diameter of soil particles (m); K_0, the coefficient of variability of soil mechanical pattern, is usually 0.5; C_f^n is soil fatigue strength to rupture and it is the function of soil cohesion C_h (Pa): after Mirtskhulava (1988) $C_f^n = 0.035C_h$. For granodioritic saprolite (see Table 2), the bed critical velocity is equal to 0.18 m/s. The corresponding value of roughness Manning coefficient n is 0.078, which is more than the usual value for small channels (0.03–0.04), but it was often measured in gully systems.

4. Results

The distribution and intensity of the different erosion processes and the related features are shown in Fig. 3 (map of ERefUs). A severe gully erosion was identified mainly in the upper part of the Mbuluzane River catchment, especially in the Mhlambanyoni subcatchment. The Mbothoma gullies belong to the highest erosion class and they are clearly visible at this scale (1:50,000). About 8% of the Mhlambanyoni basin is directly affected by severe deep gully erosion (classes 4 and 5), whereas 40% of the area shows signs of erosion (deep linear and rill–interrill erosion: classes 2–5). It should be noted that the zone of intensive erosion is situated along a north to south running system of amphibolite/

serpentite and dolerite/granophyre dykes. The main lithology consists of highly erodible saprolites (Mushala et al., 1994; Scholten et al., 1995) It is a densely populated area with a high livestock concentration. Consequently overgrazing occurs, especially on communal land like the Mbothoma area. Cattle tracks and pathways are visible in the aerial photographs and the analyses of different time series show that gullies often develop along these pathways and tracks (see also WMS Associates, 1988; Mushala, 2000).

Two main gully systems have been identified in the Mbothoma area (Fig. 1): a fossil gully system (Old Mbothoma) that had developed before the 1940s (it appears on the 1947 aerial photos) and a recent gully that has been formed since the 1960s (New Mbothoma) (Fig. 2).

The Old Mbothoma gully, with a catchment area of 4.3 ha, was formed on the convex initial slope, presumably due to linear erosion along the foot paths. The westerly exposed gully system has a dendritic pattern, with the main trunk about 230 m long and three main tributaries 180–210 m long. The entire gully length (about 440 m) is 80% of the gully catchment length, and gullies now occupy more than 60% of the catchment area. The main gully and its tributaries are 7–10 m deep, and have a trapezoidal cross-section with a flat bottom, 8–16 m wide, and steep (more than 45°) side walls. The gully has been stable for at least the last 60 years, and its morphology can be used for the stable gully model calibration. The main reason of the gully stability is the occurrence of the broad terrace induced by an amphibolite dyke at its mouth; this is working as a local erosion base and prevents the contemporary downcutting of the Old Mbothoma gully.

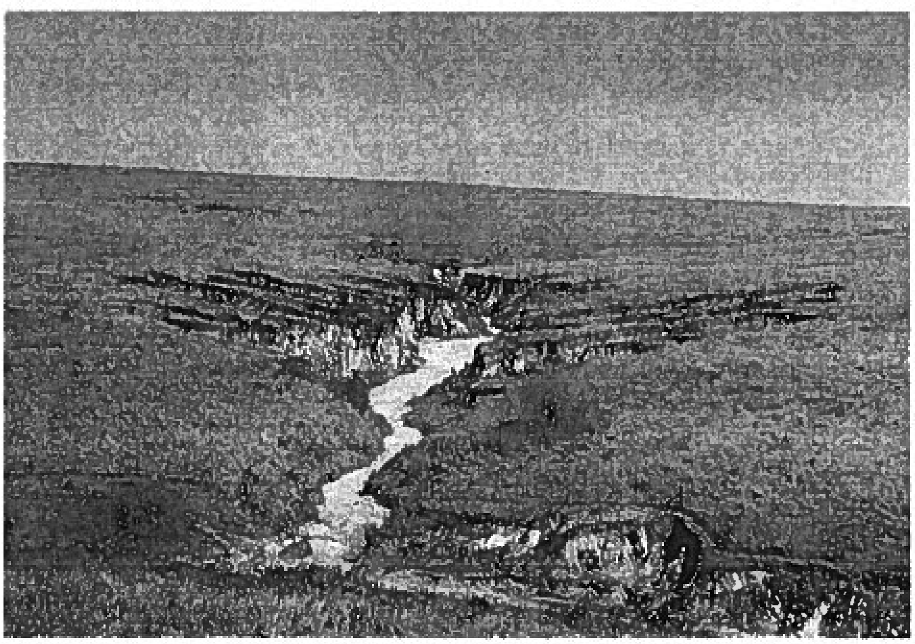

Fig. 2. Photography of the new Mbothoma gully (facing eastwards).

The New Mbothoma gully, with a catchment area of 41.8 ha, was formed on the convex initial slope (Figs. 2 and 3) and it is easterly exposed. The recent gully has a uniform main trunk 440 m long, with short side tributaries. The entire gully length is about 44% of the gully catchment length, and the gully now (1998) occupies only 1.5% of the catchment area. The main gully is 7–13 m deep and has a U-shaped cross-section with bottom widths of 13–20 m and steep sidewalls. The gully is at the first stage of rapid back cutting: this gully was absent in the 1961 aerial photographs, in 1990 it was already 400 m long and in 1998 it was 440 m long. Consequently in the first 30 years the average growth rate was about 14 m a year and it decreased to 5 m a year from 1990 to 1998. Fig. 2 shows the New Mbothoma gully. This gully, on pastureland, is extremely exploited along a small creek, which drains the wetland area situated in a hollow in the middle part of the slope. The side valley gullies are growing along cattle paths crossing the creek. Part of the gully catchment area was fenced in the late 1980s. However, the entire upper part of the slopes is still under pasture and heavily overgrazed.

The New Mbothoma gully (Fig. 2) has developed in granodioritic saprolite. Soil cover varies from less than 15 cm depth at the top of the gully contribution area to more than 1.50 m in the colluvial parts of the slopes.

The gully erosion in the Mbothoma area seems to be induced by an abrupt lowering of the erosion base. Indeed the aerial photo series show that the recent gullies in the

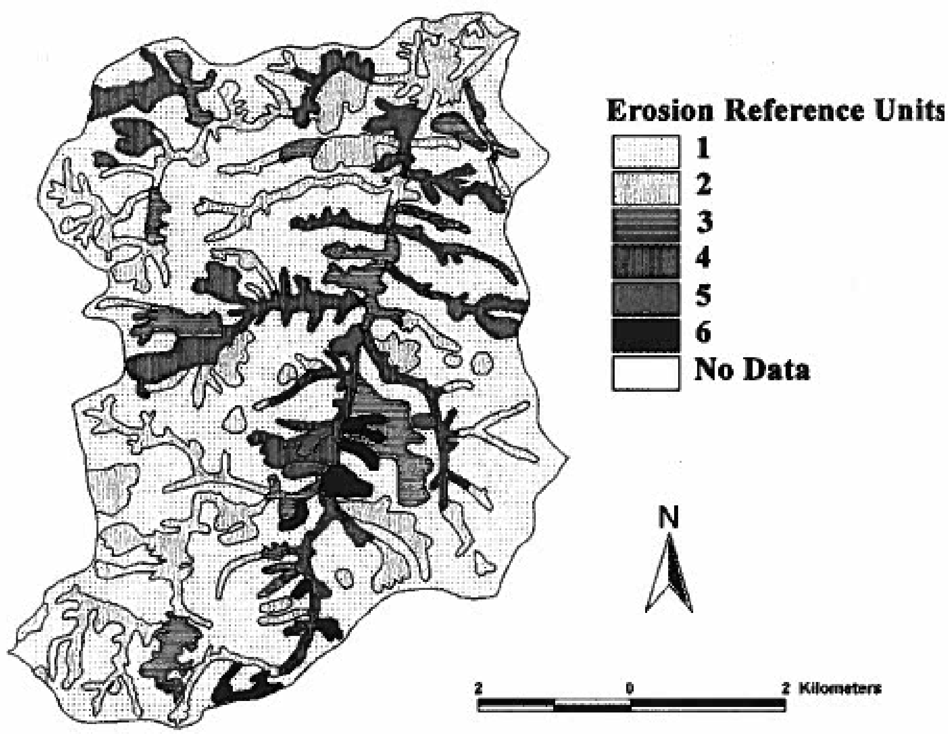

Fig. 3. Distribution of erosion classes (ERefUs) in the Mhlambanyoni River basin.

Mbothoma area have been developing since the 1960s/1970s. The longitudinal profile of the Mhlambanyoni River has step-pool pattern, with pools developed in granodioritic saprolite, and steps formed in correspondence to amphibolite dykes. In 1960s the well marked steps occured only at the headwater of the river. Up to 1990 the significant erosion of the riverbed highlighted the dykes at the lower reaches of the river. The erosion of granodioritic saprolite was generally more intensive (0.77 m/year), than the erosion of amphibolite dykes (0.59 m/year). Consequently, new steps in the longitudinal profile were formed. Furthermore, a structure indicating a recent collapse of an amphibolite dyke, which was crossing the river further down stream from the erosion sites, was observed in the field (year 1999). This collapse might have been caused by a big flood event. For the study river with a basin area of 42 km^2, the averaged 30 years rate of incision equals 0.68 m/year with local maxima of up to 1 m/year, which is close to catastrophic. So recent gully erosion processes are influenced by very high river dynamics and controlled by ancient tectonic structures.

The evolution of the erosion base level (Mhlambanyoni River channel and consequently also the mouth of the new Mbothoma gully) was calculated through the dynamic model. The elevation of the gully mouth decreases exponentially during the last 150 years

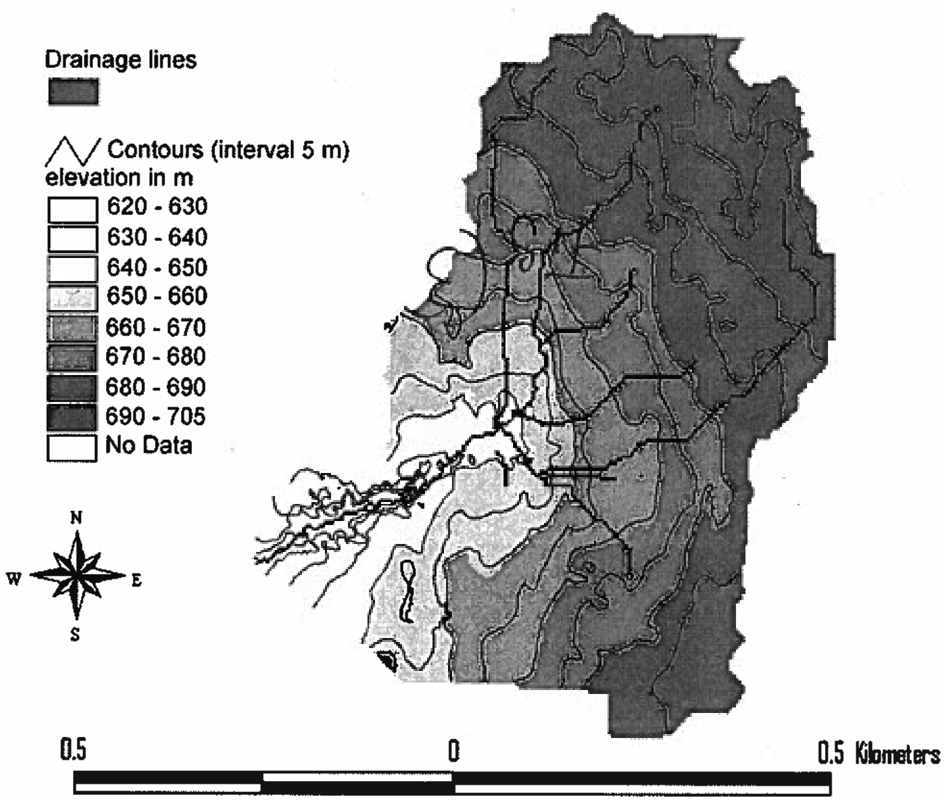

Fig. 4. DEM of the initial surface of the new Mbothoma gully system.

(from 633.7 to 615.8 m), with a rapid incision at the beginning of the calculation period and a nearly stabilisation of the riverbed at its end. Consequently for the new Mbothoma gully, the application of the stable gully model is suitable for the stage of the erosion base stabilisation, corresponding to an elevation of about 616 m. The stable longitudinal profiles, the stable gully bottom and the top width were calculated for the basin of the New Mbothoma gully through formulas (11), (13) and (14) for all the flowlines over 100 m in length. The empirical relationships and parameters, calibrated for the case of stable old Mbothoma gully, were used in those estimations. The calculated altitudes of the basin surface were added to the basin DEM (Fig. 5). The new stable Mbothoma gully system will have a dendritic pattern, with the main trunk more than 1000 m long, and three main tributaries of 270–560 m in length. The maximum gully length is calculated to be more than 95% of the gully catchment length along this flowline. About 54% of the entire contribution area will be affected by gully incision in the final situation. The tributaries and the main gully will be up to 23 m deep in their central sections and will have a trapezoidal cross-sections with flat bottoms of 25–30 m widths and steep (more than 45°) side walls. The main trunk of the stable gully will overlap the existing New Mbothoma gully (1998), which has already been in a stable morphologic condition since 1990. The

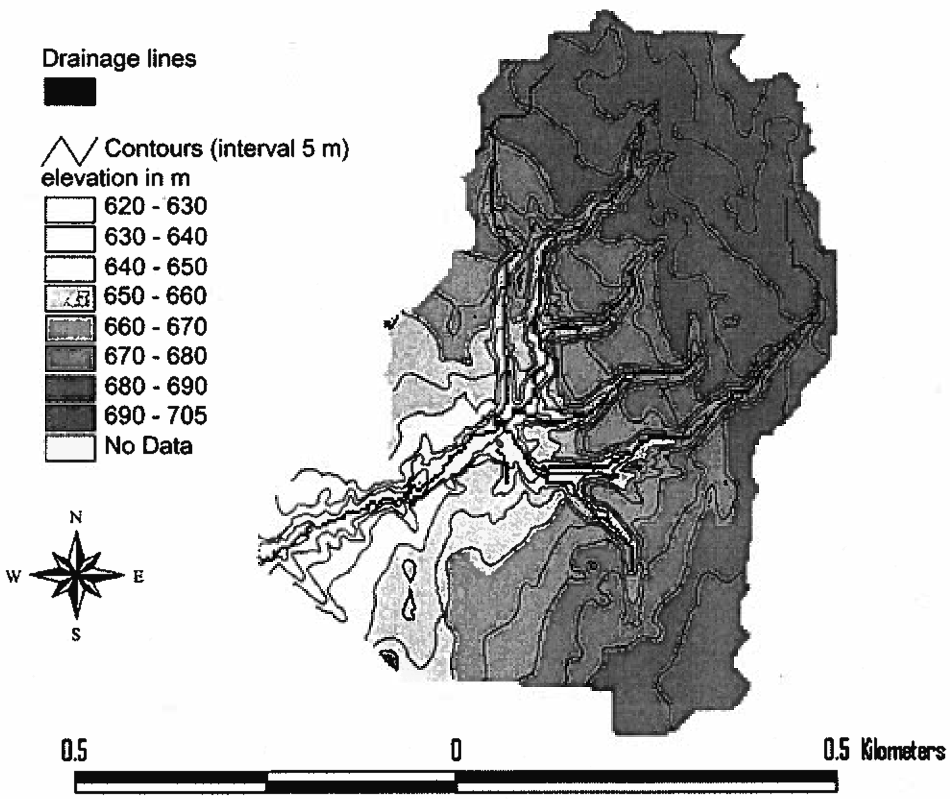

Fig. 5. DEM of the final stage of gully development (new Mbothoma gully).

most extensive erosion will occur at the upper part of the basin above the existing gully head. The stable gully volume will be 1,900,400 m^3 for the gully catchment (area 316,740 m^2) and the erosion layer will be 6.0 m deep.

Fig. 4 shows the initial surface, used as the starting point in the modelling process. Each computed flowline forms the high-resolution digital elevation models and it is separately processed by the model with its particular parameters. The final positions of the gully flowlines are provided as spatial x, y, and z data. This information was then transformed into a digital elevation model showing the final stage of gully development (see Fig. 5).

5. Discussion and conclusions

In this study, the concept of ERUs is used to identify areas subject to different erosion processes. The distribution of gully erosion in the study area clearly shows that gully erosion must be included in the calculation of sediment yield, especially where the lithology is highly vulnerable to erosion (saprolites). Nevertheless, in traditional models such as the USLE, gully erosion is almost completely neglected. The application of the ERU concept gives the possibility of identifying areas affected by different types of erosion. This can be utilised in the regionalisation of erosion processes and dynamics because it allows different models for different erosion types to be applied in order to obtain more exact information on the total amount of erosion within a catchment. The parts of a catchment affected by rill–interrill erosion for example could be modelled through a USLE type model, whereas gully erosion has to be modelled with a specific model. In this study the volume of sediments provided by gully erosion was calculated by using the stable gully model. The stable 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 greater than the critical velocity of wash load sedimentation. This criterion of stability means that the main expression for the estimation of gully stable slope is the well-known reverse relationship between the slope and the discharge (or its surrogates), now intensively used in gully erosion investigations (see, for example, Desmet et al., 1999). The main problem of the practical application of Eq. (13) or its simplified versions lies in the great instability of the complex

$$(V''_{\mathrm{cr}})^{2.67} n^2.$$

This is due to a variability of controlling, in some cases, inherent (implicit) factors: flow hydraulics and eroded deposits composition. The values of critical velocity and the roughness coefficient can usually be estimated from semi-empirical formulas and tables with a rather low degree of accuracy. High exponents (2.67 for velocity and 2 for the Manning coefficient) significantly increase these errors.

The whole range of discharges, with different return periods, must be used in the calculation of the stable gully longitudinal profile. The magnitude M_i of channel deformation during the flow with specific discharge (or runoff depth) Y_i was assumed to be proportional to the product of sediment discharge Y_{si} and its frequency P_i: $M_i = Y_{si}P_i$. In order to obtain the elevations of the stable profile, corresponding to the whole range of water flow, the partial elevations Z_i, calculated for every given point with partial runoff depth Y_i, have to be averaged with the weight equal to $Y_{si}P_i$. This procedure of averaging through the weight is completely accurate only in the special case of linear or power low decrease of sediment discharge (Sidorchuk, 1984). In other cases the described approach results in longitudinal profiles, which will differ from the profiles, calculated through the dynamic gully model for the same sequence of discharges.

A careful calibration of lithological and hydrological erosion factors is necessary to minimize the organic errors of the static gully model. The most practical way is the backward calculation of

$$(V''_{\mathrm{cr}})^{2.67}n^2$$

complex for a stable gully system with the same hydrology and lithology, as the modelling basin. In most cases, the lower parts of existing active gullies have already reached the stage of stable morphology, and these lower, stable sections can be used for model calibration. These procedures were applied for the proposed modelling of the new stable Mbothoma gully system. As it is carefully calibrated, the static gully model is a powerful tool for a very quick estimation of the final gully morphology (length, depth, area and volume).

In order to achieve more sophisticated conclusions the dynamic gully erosion model has to be applied. This has been done for the new Mbothoma gully system (Sidorchuk et al., in press). The application of the dynamic gully model shows that a complete stabilisation of the gully system (fulfilling the conditions of Eqs. (1) and (2)) takes several thousand years. For shorter periods (100–150 years) of gully evolution the stable model over predicts the main morphological characteristics, mainly in the upper part of the catchment.

A comparison of both models shows that in the year 2100, the gully length along the main trunk is predicted to 100% of its stable value whereas the area will be 95% and the volume will be 65% of the stable gully value. This is completely corresponding to the dynamics of gully evolution as found by Kosov et al. (1978) and has to be taken into account in the applications of the stable gully model.

Acknowledgements

This study has been carried out with the financial support from the Commission of the European Communities, INCO-DC, Contract number IC18-CT97-0144, "The development of an innovative computer based 'Integrated Water Resources Management System (IWRMS)' in semiarid catchments for water resources analyses and prognostic scenario planning". It does not necessarily reflect its view and in no ways anticipates the Commission's future policy in this area. We also like to thank Dennis Dlamini and Roland E. Schulze (Dep. Agric. Eng. Uni Natal, Pietermaritzburg, RSA) for supplying runoff data. Also, thanks to the referees.

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