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THE EUROPEAN SOIL EROSION MODEL (EUROSEM): A DYNAMIC APPROACH FOR PREDICTING SEDIMENT TRANSPORT FROM FIELDS AND SMALL CATCHMENTS

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ABSTRACT

The European Soil Erosion Model (EUROSEM) is a dynamic distributed model, able to simulate sediment transport, erosion and deposition over the land surface by rill and interill processes in single storms for both individual fields and small catchments. Model output includes total runoff, total soil loss, the storm hydrograph and storm sediment graph. Compared with other erosion models, EUROSEM has explicit simulation of interill and rill flow; plant cover effects on interception and rainfall energy; rock fragment (stoniness) effects on infiltration, flow velocity and splash erosion; and changes in the shape and size of rill channels as a result of erosion and deposition. The transport capacity of runoff is modelled using relationships based on over 500 experimental observations of shallow surface flows. EUROSEM can be applied to smooth slope planes without rills, rilled surfaces and surfaces with furrows. Examples are given of model output and of the unique capabilities of dynamic erosion modelling in general. © 1998 John Wiley & Sons, Ltd.

KEY WORDS: soil erosion model; soil erosion processes; distributed modelling; dynamic modelling.

INTRODUCTION

An increasing awareness by scientists, governments and the general public that soil erosion is an important problem within the countries of the European Community (Morgan and Rickson, 1990) has drawn attention to the lack of a satisfactory system in Europe for assessing the risk of erosion, predicting erosion rates and designing and evaluating different soil protection strategies. Present technologies for erosion assessment, based on scoring systems for rainfall erosivity, soil erodibility, slope and land use (Auerswald and Schmidt, 1986; Rubio, 1988; Briggs and Giordano, 1992; Jäger, 1994), provide information on the spatial distribution of erosion risk but only limited data on erosion rates. Attempts to use the Universal Soil Loss Equation (USLE) (Wischmeier and Smith, 1978) in Europe as a technique for predicting erosion rates and evaluating different soil conservation practices show that great care is required in the selection of input values for rainfall (R) (Chisci and Zanchi, 1981; Richter, 1983) and soil erodibility (K) (Richter, 1980; De Ploey, 1986; Schwertmann, 1986) factors. Also, the equation is of limited value since it cannot provide information on the fate of sediment once it is eroded. The USLE is not able to predict deposition or the pathways taken by eroded material as it moves from hillslope sites to water bodies. In a European context, where the most important consequences of erosion are pollution and sedimentation downstream rather than loss of productivity on-site, policy-makers need to know more about the location of sediment sources and sinks. Further, the design of strategies to control pollution associated with runoff and erosion on agricultural land requires knowledge of what happens in individual rain

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storms, often on a minute-by-minute basis, in order to predict the size and timing of peak discharges of water and sediment from hillslopes to rivers. The USLE cannot provide this because it predicts only mean annual soil loss.

In order to provide a better representation of erosion processes, American scientists have concentrated in recent years on developing more physically based erosion models such as those used in CREAMS (Knisel, 1980; Foster *et al*., 1981), ANSWERS (Beasley *et al*., 1980), KINEROS (Woolhiser *et al*., 1990) and WEPP (Nearing *et al*., 1989). Similar models are also being developed in Australia (Rose *et al*., 1983; Misra and Rose, 1992).

Whilst both CREAMS and WEPP can be run for individual storms, they simulate only total storm soil loss, and assume a steady surface flow profile. They do not model peak sediment discharge or describe the pattern of events within a storm, or provide a sediment graph showing the pattern of sediment discharge over time, information which is useful for looking at potential pollution loadings from sediment fluxes into water courses. For catchments where one or two storms account for most of the annual soil loss, which is the typical situation in many European countries (Sfalanga and Franchi, 1978; Boschi ahnd Chisci, 1978; Richter, 1979; Raglione *et al*., 1980; Boschi *et al*., 1984; Tropeano, 1984; Chisci *et al*., 1985; Govers, 1987), steady flow is rarely achieved, and the CREAMS and WEPP methodologies may be inappropriate. In order to obtain a good approximation of what is happening over the land surface in such events, a fully dynamic approach to erosion modelling is required.

The need for an alternative approach was recognized at the European Community Workshop held in Brussels, 1986, when Chisci and Morgan (1988) proposed a framework for a European model to be based on the best European research into erosion processes and their control. One of the recommendations of the Workshop was that European scientists should ‘try to develop a new general erosion model for use in the EC countries for erosion risk evaluation and the design of erosion control measures’ (Chisci, 1988). This paper describes the outcome of the resulting research effort which has led to the European Soil Erosion Model (EUROSEM). In addition to describing EUROSEM, emphasis is given to the features which make it different from and an advance upon the American and Australian models referred to above.

APPROACH

Since soil erosion by water is closely related to rainfall and runoff, erosion modelling cannot be separated from the procedures used to model the generation of runoff and its routing down a hillside and through the river channel network. Some models, like WEPP and CREAMS, use continuous simulation to model the generation of runoff. Continuous simulation models require a large amount of input data for weather and land use that are only indirectly related to erosion studies. In addition, they are sensitive to modelling evapotranspiration and soil dynamics, and simulate a large number of small events that may not produce significant runoff or soil loss. It seems therefore more appropriate to focus on the development of an event-based model, considering, as indicated above, that erosion is dominated by only a few events per year which are characterized by a highly dynamic behaviour.

An alternative approach was therefore adopted based on simulating the dynamic behaviour of events within a storm. Within-storm modelling is also more compatible with the equations used in process-based models to describe the mechanics of erosion. These equations are strictly applicable to instantaneous conditions and they cannot be applied to average conditions without loss of accuracy. Applying them to conditions averaged over one minute is thus more acceptable than using them for conditions averaged over one hour or more. It is recognized, however, that this approach creates problems for the determination of certain input data, particularly relating to the starting conditions prior to each storm.

Many of the factors that influence erosion, particularly soil, slope and land use, have considerable spatial variability and cannot be described by a single average value, even over areas as small as one field. Spatially lumped models, which treat an area as a single unit of uniform characteristics, are not appropriate for most natural catchments. If this spatial variability is to be taken into account, a dynamic distributed model must be used. The simulation of erosion contained in EUROSEM is linked to the water and sediment routing structure of the KINEROS model (Woolhiser *et al*., 1990). It has also been linked with the MIKE SHE model (Danish

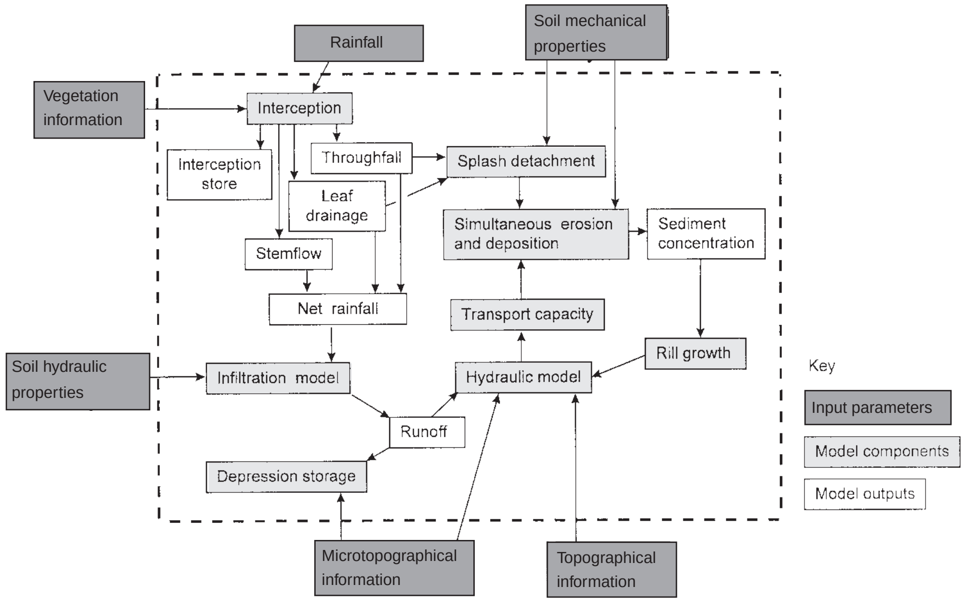

Figure 1. Flow chart of EUROSEM

Hydraulic Institute, 1993) which is a continuous simulation model and an extension of the original Système Hydrologique Européen (SHE) model (Danish Hydraulic Institute, 1985; Abbott *et al*., 1986). The description given in this paper is for the EUROSEM–KINEROS version.

BASIC CONCEPTS

The flow chart for EUROSEM is presented in Figure 1. EUROSEM has a modular structure with each module being developed in as much detail as the existing level of knowledge permits. This structure will enable continuous improvements to be made in the light of new research. EUROSEM uses a dynamic and distributed simulation of erosion processes with explicit simulation of rill and interill flow. Water and sediment are routed over the land surface which is represented, following KINEROS, as a series of interlinked uniform slope planes and channel elements to simulate small catchments. Soil loss is computed as a sediment discharge, defined as the product of the rate of runoff and the sediment concentration in the flow, to give a volume (or mass) of sediment passing a given point in a given time. The computation is based on a numerical solution of the dynamic mass balance equation (Bennett, 1974; Kirkby, 1980; Woolhiser *et al*., 1990):

$$\frac{\partial(AC)}{\partial t}+\frac{\partial(QC)}{\partial x}-e(x,t)=q_s(x,t) \tag{1}$$

where C = sediment concentration (m^3 m^{-3}), A = cross-sectional area of the flow (m^2), Q = discharge (m^3 s^{-1}), q_s = external input or extraction of sediment per unit length of flow (m^3 s^{-1} m^{-1}), e = net detachment rate or rate of erosion of the bed per unit length of flow (m^3 s^{-1} m^{-1}), x = horizontal distance (m), and t = time (s).

For channel flow q_s represents lateral inflows of sediment from the base of adjacent hillsides. When applied to overland flow over hillslopes, q_s becomes zero.

The term e in Equation 1 is composed of two major components:

$$e = DR + DF \tag{2}$$

where DR = the rate of soil particle detachment by raindrop impact (m^3 s^{-1} m^{-1}), and DF = the net rate of soil particle detachment by the flow (positive for detachment and negative for deposition) (m^3 s^{-1} m^{-1}).

Values of surface runoff $Q(x,t)$ and $A(x,t)$ are obtained by numerical solution of the dynamic mass balance equation for water, analogous to Equation 1:

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = w[r_i(t) - f(t)] \quad (3)$$

where $r_i(t)$ = the rainfall rate less the interception (mm min^{-1}), $f(t)$ = the local infiltration rate (mm min^{-1}) and w = flow width (mm).

Equation 3 is solved using a kinematic wave assumption for a fixed relation $Q(A)$ (Woolhiser and Liggett, 1967; Woolhiser, 1969). During flow recession, the calculations of infiltration loss, $f(t)$, following KINEROS, are based on the depth of water ($h(x,t)$, m) and an estimate of the fraction of the surface covered to this depth.

MODEL COMPONENTS

Rainfall interception

Rainfall input to the model is in the form of break-point data giving a depth R (mm) for each time step during a storm. From this input, rainfall intensity (R_i, mm h^{-1}) and rainfall volume (m^3) are calculated. Account is also kept of the cumulative rainfall (R_{cum}, mm) received during the storm.

On reaching the canopy of the vegetation, the rainfall is divided into two parts, namely that falling either on open ground or passing through gaps in the canopy and reaching the soil surface as direct throughfall (*DT*, mm), and that which strikes the vegetation cover. The division is based on the simple relationship:

$$IC = R \cdot COV \quad (4)$$

where IC = the depth of rainfall intercepted by the vegetation (mm), and COV = the fraction of the surface covered with vegetation.

An initial proportion of the intercepted rainfall is stored on the leaves and branches of the vegetation. This is termed the interception store. The rainfall held in this store does not reach the soil surface and therefore is unavailable for infiltration or runoff. In many erosion models, this interception store is either ignored, as in CREAMS, or is considered as a depth which has to be filled before rain is allowed to pass from the vegetation canopy to the ground, as in KINEROS. This last approach means that no rain reaches the soil surface from the canopy until the interception store is full. EUROSEM adopts a more dynamic approach which allows rainfall to pass from the canopy to the ground at the same time as the interception store is being filled. This means that some transfer of water from the canopy will take place right from the start of the storm. The depth of the interception store (IC_{store}, mm) for a time step (t_s, s) is modelled as a function of the cumulative rainfall from the start of the storm, using the exponential relationship proposed by Merriam (1973):

$$IC_{store} = IC_{max}[1 - \exp(R_{cum} / IC_{max})] \quad (5)$$

where IC_{max} = the maximum depth of the interception store for the given crop or vegetation cover (mm).

In most erosion models, it is assumed that the canopy cover protects the underlying soil from erosion and so little attention is given to the fate of intercepted rainfall and its influence on the erosion system. In EUROSEM, however, the importance of leaf drips in detaching soil particles from the soil mass is recognized, so it is necessary to calculate the proportion of the rainfall reaching the ground surface as leaf drainage. Thus, the rainfall which is intercepted by the canopy and not held in the interception store (termed temporarily intercepted throughfall (*TIF*, mm)) is partitioned into stemflow (*SF*, mm) and leaf drainage (*LD*, mm). The depth of stemflow for each time step is modelled as a function of the average acute angle (*PA*, degrees) of the plant stems to the ground surface, using equations developed in laboratory experiments by van Elewijck (1989a,b). These equations have been modified by assuming that a maximum of half the depth of temporarily intercepted throughfall is available for stemflow, to give:

$$SF = 0 \cdot 5\, TIF\, (\cos PA \sin^2 PA) \tag{6}$$

for grasses, and

$$SF = 0 \cdot 5\, TIF \cos PA \tag{7}$$

for other plant species.

Conceptually, Equations 6 and 7 describe the relationship between the diameter of the catching surface (stems and leaves) and the median volume drop diameter of the raindrops. Where, as with grasses, the mean diameter of the catching surface is less than the drop diameter, gravity, expressed by sin *PA*, plays an important role in determining the depth of stemflow. With thicker catching surfaces, stemflow depth depends only on the projected length of the stems or leaves, as expressed by cos *PA*.

The difference between the depth of the temporarily intercepted throughfall and the depth of stemflow for each time step comprises leaf drainage, i.e. that component of the rainfall which reaches the soil surface as individual drips from the leaves. The net rainfall at the ground surface, which is therefore available for infiltration, is the summation of the direct throughfall, stemflow and leaf drainage.

Infiltration

KINEROS–EUROSEM models infiltration using the equation (Smith and Parlange, 1978):

$$f_c = K_s \frac{\exp(F/B)}{\exp(F/B) - 1} \tag{8}$$

where f_c = the maximum rate at which water can enter the soil, which is known as the infiltration capacity (mm min^{-1}), K_s = the saturated hydraulic conductivity of the soil (mm min^{-1}), F = the amount of rain already absorbed by the soil (mm), and B = an integral capillary and water deficit parameter (mm) related to soil fraction <2 mm.

The term B is obtained from:

$$B = G(\theta_s - \theta_i) \tag{9}$$

where G = the effective net capillary drive (mm), θ_s = the maximum value of water content of the soil (m^3 m^{-3}), and θ_i = the initial value of soil water content (m^3 m^{-3}).

The term G is a conductivity-weighted integral of the capillary head of the soil, used in many infiltration equations, and defined as:

$$G = \frac{1}{K_s} \int_{-\infty}^{0} K(\psi)\,\mathrm{d}\psi \tag{10}$$

where ψ = the soil matric potential (mm), and $K(\psi)$ = a hydraulic conductivity function.

G is essentially a property of the soil with units of length and is conceptually equivalent to a value of effective capillary head. The parameters G, K_s and θ_s are measurable and physically related, but guidelines for their estimation based on soil texture are given by Woolhiser *et al*. (1990).

In the initial part of a storm, F is increased by the addition of rainfall, since the value of f_c is very large for small values of F. The model predicts the initiation of runoff as the time when F grows to the point that Equation 8 finds f_c to be equal to or less than the rain rate (r_i), beyond which $f(t) = f_c\,(F)$. The prediction of ponding and subsequent infiltration rate in a unified equation is based on soil physics, and is the main reason behind using F rather than t as the independent variable (Smith, 1983).

When operating EUROSEM, the input values of G, K_s and θ_s are varied according to the soil texture, degree of crusting and management practice. Adjustments to these values are made within the model to account for the effects of rock fragments and the presence of any vegetation or crop cover. Rock fragments affect infiltration in two ways. First, they reduce the effective overall storage in porosity ($\theta_s - \theta_i$) by modifying the parameter B in Equation 8 to account for the presence of rock fragments (ROC) using the relationship (Woolhiser *et al*., 1990):

$$B_{roc} = B(1 - ROC) \tag{11}$$

where B_{roc} = the parameter B modified for rock fragments (mm min^{-1}), and ROC = the fraction of the soil composed of rock fragments, expressed as a volume between 0 and 1.

Second, rock fragments affect infiltration into soils through their position on the soil surface (Poesen and Ingelmo-Sanchez, 1992; Poesen *et al*., 1994). Rocks which are embedded in a surface seal reduce infiltration whereas rocks which sit on the surface or are surrounded by macropores (e.g. those produced as a result of tillage) protect the soil structure and promote infiltration. EUROSEM models the first condition using the equation:

$$K_{sroc} = K_s(1 - PAVE) \tag{12}$$

and the second using the equation:

$$K_{sroc} = K_s(1 + PAVE) \tag{13}$$

where K_{sroc} = a modified value of saturated hydraulic conductivity (mm min^{-1}), and $PAVE$ = the fraction of the surface area covered by rock fragments (between 0 and 1).

Equation 13 should not be used, however, if the effects of tillage on saturated hydraulic conductivity outweigh the effects of rock fragments.

The research base for modelling the effect of vegetation cover on infiltration is rather sparse. Thornes (1990) proposes that infiltration capacity increases exponentially with increasing percentage vegetation cover as a function of increases in organic matter and decreases in the bulk density of the soil. Such a relationship is similar to that developed by Holtan (1961) to express the saturated hydraulic conductivity of the soil as a function of the percentage basal area of the vegetation. Based on his work, the following equation is used in EUROSEM to modify the saturated hydraulic conductivity value of the soil:

$$K_{sveg} = K_s \frac{1}{1 - PBASE} \tag{14}$$

where K_{sveg} = the saturated hydraulic conductivity of the soil with the vegetation (mm min^{-1}), $PBASE$ = the total area of the base of the plant stems expressed as a proportion (between 0 and 1) of the total area of the plane.

Soil surface condition

The soil surface is considered to be composed of form roughness and hydraulic roughness, both of which play a role in erosion as well as water flow. The form roughness of the soil surface determines the volume of water than can be held on the surface as depression storage. The basis for modelling, however, is extremely limited and depression storage is ignored in many hydrological models. However, it is included in EUROSEM where it can be used to describe different surfaces produced by tillage. The form roughness of the soil surface is expressed by a roughness measure (RFR), defined as the ratio of the straight line distance between two points on the ground (X, m) to the actual distance measured over all the microtopographic irregularities (Y, m). RFR is determined in the field by placing a one metre long chain with a 5 mm link on the soil surface and measuring the distance between the two ends of the chain:

$$RFR = \frac{Y - X}{Y} \times 100 \tag{15}$$

This measure is converted into a surface storage depth, D(m), using a regression equation developed from research in southern Germany:

$$D = \exp\,(-6 \cdot 6 + 0 \cdot 27\ RFR) \tag{16}$$

Surface runoff processes

When the net rainfall intensity at the ground surface exceeds the infiltration rate and surface depression storage is satisfied, the excess comprises surface runoff. In KINEROS–EUROSEM, runoff along a slope for a plane element, a rill, or a channel is viewed as a one-dimensional surface flux in which discharge (Q) is related to the hydraulic radius (r). The rating equation, based on the normal flow equation, is:

$$u = \alpha\, r^{m-1} \tag{17}$$

where, based on the Manning equation for flow velocity, u = flow velocity (m s^{-1}), r = hydraulic radius (m), $\alpha = s^{0.5}/n$ (m$^{1/3}$ s^{-1}), s = slope (%), n = Manning roughness value (m$^{-1/3}$ s), and $m = 5/3$.

Assuming discharge $Q = uA$, the general rating equation can be written as:

$$Q = \alpha\ p\ r^{m} \tag{18}$$

where p = the wetted perimeter (m).

Combining Equation 18 with the continuity Equation 3 gives:

$$\frac{\partial A}{\partial t} + \alpha\left[mpr^{m-1}\frac{\partial r}{\partial x} + r^{m}\frac{\partial p}{\partial x}\right] = wq(x,t) \tag{19}$$

where $q = r_i - f$ is the lateral inflow rate (m^3 s^{-1} m^{-1}), or 'rainfall excess'.

Manning's equation for flow velocity is chosen because of its wide use by engineers and availability of input data. It is recognized that this implies turbulent flow but, given the types of storms for which EUROSEM is designed, this does not seem unreasonable. Also, it avoids the need to identify the transition between early laminar and turbulent flow when, as is generally the case, the flow is disturbed by raindrop impacts.

The traditional concepts of rill and interrill processes, where flow erosion occurs in the rills and splash erosion on the interrill area, are not adopted by EUROSEM. Instead, splash and flow processes are modelled on all areas, with the distinction between rill and interrill areas being one of geometry. Rills, which can encompass any form of concentrated flow path, are described as essentially trapezoidal channels with side wall slopes of 0 (vertical) or greater (Figure 2). Interrill areas are surfaces without orientated roughness. The whole area may or may not be rilled, but if rills are present EUROSEM assumes the interrill areas slope towards the rills rather than straight downslope. Erosion and deposition by splash and flow detachment can occur at any point on either the rill or interrill areas (see next two sections). Water and sediment are routed across the interrill areas as well as through any rills until the bottom of the plane is reached.

Within EUROSEM it is possible to specify whether the rills are uniform in depth over the whole length of a plane or whether their depth increases downslope. The depth and width of the rill at each point along its length will change in response to erosion and deposition. EUROSEM assumes that deposition will reduce the depth

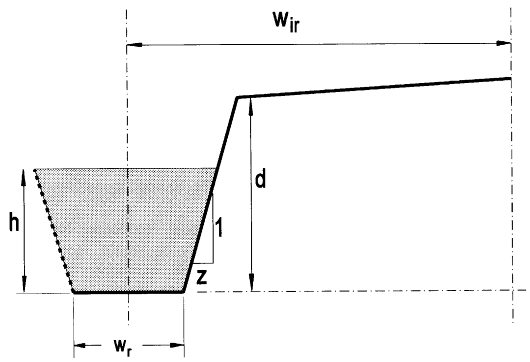

Figure 2. Representation of the hydraulic geometry of a rill channel (W_{ir} = distance between the centre line of the rill and centre line of the interrill area; d = depth of base of the rill below the average height of the interill surface; z = side slope of the rill expressed as the ratio of horizontal to vertical component)

whereas erosion will increase the width and the depth by taking material equally from the two sides and bottom of the rill.

For shallow interrill surface flow, a unit width is used for computations, so that $p=1$ and $r=h$. In EUROSEM, h is a lumped variable representing the mean of a distribution of depths across the slope. The discharge rating Equation 18 becomes:

$$Q = a\,h^m \tag{20}$$

which represents the mean (non-linear) condition for such an averaged concept of h. Equation 19, for a unit width ($p,w=1$) and $r=h$, becomes:

$$\frac{\partial h}{\partial t} + \alpha m h^{m-1}\frac{\partial h}{\partial x} = q(x,t) \tag{21}$$

In KINEROS, the kinematic wave Equations 19 or 21 are solved numerically for a finite difference grid by a four-point implicit method using the Newton–Raphson technique (Pearson, 1983; Woolhiser *et al.*, 1990). The upslope boundary condition for the depth of flow (h) at $x=0$ and $t=0$ is either 0 or equal to the depth of runoff from an upslope contributing plane. The finite difference solution of Equation 1 provides arrays of nodal point values of Q, A and u. These values, along with the array of nodal values of C at the previous time step (C_i, $i=1$, N) (plus the upstream and initial conditions described below), allow an explicit solution of the finite difference form of Equation 1 for new values of C_i, starting with the node just below the upstream boundary.

A similar procedure is adopted for routing flow in rills or channels, where the relevant rating equation is Equation 19. The term $q(x,t)$ in the equation then becomes the unit discharge into the rills from interrill contributions.

Soil detachment by raindrop impact

Modelling this process is based on relationships between detachment and the kinetic energy of rainfall. In contrast with most erosion models which, as seen earlier, ignore the effects of leaf drip, EUROSEM explicitly models soil detachment by raindrop impact for both direct throughfall and leaf drainage as a function of their kinetic energy. This enables the effects of different heights of vegetation and canopy and residue covers to be simulated explicitly.

The rainfall energy reaching the ground surface as direct throughfall ($KE(DT)$, $\mathrm{J\,m^{-2}\,mm^{-1}}$) is assumed to be the same as that of the natural rainfall. It is estimated as a function of rainfall intensity from the equation derived by Brandt (1989) which assumes that the raindrop size distribution follows that described by Marshall and Palmer (1948):

$$KE(DT) = 8\cdot 95 + (8\cdot 44 \log R_{\mathrm{i}}) \tag{22}$$

The energy of the leaf drainage ($KE(LD)$, J m^{-2} mm^{-1}) is estimated from the following relationship developed experimentally by Brandt (1990):

$$KE(LD) = (15\cdot 8 PH^{0\cdot 5}) - 5\cdot 87 \tag{23}$$

where PH = the effective height of the plant canopy (m).

The relationship is considered valid because, for a wide range of plants, the drop-size distribution of leaf drainage (with a median drop diameter of about 4·8 mm) has been found not to vary (Brandt, 1989). This means that the variations in the energy of leaf drainage are solely a function of the impact velocity of the raindrops which depends on the height of fall. The model sets the kinetic energy of leaf drainage to zero when the canopy height is less than 14 cm to avoid the negative values predicted by Equation 23.

The total kinetic energy of the rainfall can be calculated by multiplying the energies obtained from Equations 22 and 23 by the respective depths of direct throughfall and leaf drainage received and summing the two values. This calculation is made in EUROSEM for every time increment of the rainstorm.

Detachment equation for rainfall. Soil detachment by raindrop impact (DR, m^3 s^{-1} m^{-1}) for time step (t_s) is calculated from the equation:

$$DR = \frac{k}{\rho_s} KE\, \mathrm{e}^{-zh} \tag{24}$$

where k = an index of the detachability of the soil (g J^{-1}) for which values must be obtained experimentally, ρ_s = particle density (kg m^{-3}), KE = the total kinetic energy of the net rainfall at the ground surface (J m^{-2}), z = an exponent varying between 0·9 and 3·1, depending on soil texture but for which a value of 2·0 can be used for a wide range of conditions (Torri *et al*., 1987), and h = the mean depth of the surface water layer (m).

The influence of slope on soil particle detachment is neglected in EUROSEM because of the difficulty of characterizing the ‘effective slope’ which needs to be measured over distances of several drop diameters from the point of raindrop impact and is dependent upon local surface roughness and the angle of internal friction of the soil (Torri and Poesen, 1992). It is not the same as the general surface slope, which is generally smaller.

Where non-erodible surfaces, such as rock outcrops, surface rock fragments, concrete and tarmac, occur within the element, the detachment rate is modified by:

$$DR_{\mathrm{pav}} = DR(1 - PAVE) \tag{25}$$

where DR_{pav} = the detachment rate allowing for the non-erodible surfaces (m^3 s^{-1} m^{-1}), and $PAVE$ = the fraction (between 0 and 1) of the soil surface covered by non-erodible surfaces.

Initial condition for sediment concentration. Since, during a rainstorm, splash erosion will already be taking place when runoff begins, the initial sediment concentration in the runoff cannot be taken as zero. Based on an analysis of Equation 1 at the time of ponding (t_p, s) or $x = 0$ and $A = 0$, the sediment concentration (C) at t_p is calculated, as in KINEROS, from:

$$C(t_{\mathrm{p}}) = \frac{DR}{q + v_{\mathrm{s}}} \tag{26}$$

where v_s is the particle settling velocity (m s^{-1}).

This equation is also used to determine the boundary condition at the upper end of a slope plane when there is no input of runoff or sediment from above.

Soil detachment by runoff

Soil detachment by runoff is modelled in terms of a generalized erosion–deposition theory proposed by Smith *et al*. (1995). This assumes that the transport capacity concentration of the runoff (TC, m^3 m^{-3}) reflects a balance between the two continuous counteracting processes of erosion and deposition. It implies that the ability of flowing water to erode its bed is independent of the amount of material it carries and is only a function of the energy expended by the flow, particularly the shear between the water and the bed, and the turbulent energy in the water. The erosion rate of the flow (E_q, m^3 s^{-1} m^{-1}) is continually accompanied by deposition at a rate equal to $w\,C\,v_s$, where w is the width of flow (m). This condition can be expressed as:

$$DF = E_q - w\,C\,v_s \tag{27}$$

where DF = the net detachment rate of soil particles by the flow (Equation 2), which is positive for erosion and negative for deposition.

According to this generalized theory, the transport capacity concentration (TC) represents the sediment concentration at which the rate of erosion by the flow and accompanying rate of deposition are in balance. In this condition, DF is zero and E_q equates to the deposition rate ($w\,TC\,v_s$). A general equation for net soil detachment during flow, expressed in terms of settling velocity and transport capacity, then becomes:

$$DF = w\,v_s\,(TC - C) \tag{28}$$

This equation, however, assumes that the soil particles are loose so that processes are reversible whereas, in reality, detachment will be limited by factors such as soil cohesion. The net pick-up rate for cohesive soil therefore needs to be reduced by a coefficient whenever C is less than TC. Equation 28 becomes:

$$DF = \beta\,w\,v_s\,(TC - C) \tag{29}$$

where β = a flow detachment efficiency coefficient. This coefficient is equivalent to the efficiency functions proposed by Rose *et al*. (1983) and Styczen and Nielsen (1989) in their procedures for modelling of soil detachment by flow.

By definition, $\beta = 1$ when DF is negative (deposition is occurring) and $\beta < 1$ for cohesive soils when DF is positive ($TC > C$). This parameter can be evaluated experimentally by a steady flow experiment in which clear water flows across a uniform bed of soil, and the spatial rate of change of C is measured. Such experiments have in fact been done in furrow irrigation (Trout, 1996). EUROSEM estimates β as a function of the cohesion of the soil (J, kPa) as measured by a torvane under saturated conditions. Soil cohesion has been successfully related to the onset of rilling (Rauws and Govers, 1988), and is recognized by a number of authors (Torri and Borselli, 1991; Brunori *et al*., 1989; Crouch and Novruzi, 1989) as related to erodibility. It is recognized, however, that the relationship between soil cohesion and detachability of the soil by runoff is dependent upon the initial moisture content (Govers and Loch, 1993) and on initial structural conditions (Govers *et al*., 1990), and that no unique relationship exists even for a single soil. However, until a procedure is developed that enables the detachability of soils by flow to be predicted from soil parameters which can be easily measured in the field, an approach based on cohesion seems the most appropriate. For application, EUROSEM assumes at present that when $J < 1$ kPa, $\beta = 1$. For larger values of J, the value of β is reduced exponentially:

$$\beta = 0 \cdot 79\ \mathrm{e}^{-0 \cdot 85 J} \tag{30}$$

When TC is zero and DR has a value due to rainfall energy, a value of C will be obtained such that, using Equation 2 with $e=0$, $DET = w v_s C$. The concentration in flow will be $C = DET/w v_s$. This has been termed 'rain flow transportation' (Moss *et al*., 1979).

Transport capacity of the flow

The capacity of runoff to transport detached soil particles is expressed in EUROSEM in terms of a concentration, TC. For flow in rills, it is modelled as a function of unit stream power, using a relationship based on the work of Govers (1990) which showed that the transporting capacity of overland flow could be predicted from simple hydraulic parameters. For interrill flow, TC is modelled as a function of modified stream power, based on the experimental work of Everaert (1991).

Rill transport capacity. Unit stream power (ω, cm s^{-1}) is the hydraulic variable on which rill TC is based, and is defined as:

$$\omega = 10\, u\, s \tag{31}$$

where s = slope (%), and u = mean flow velocity (m s^{-1}).

Based on this variable, Govers (1990) found that TC could be expressed for any particle size (ranging from 50 to 250 μm) as follows:

$$TC = c(\omega - \omega_{\mathrm{cr}})^{\eta} \tag{32}$$

where ω_{cr} = critical value of unit stream power (= 0·4 cm s^{-1}), and c,η = experimentally derived coefficients depending on particle size.

Further analysis has shown that one can estimate c and η as follows:

$$c = [(d_{50} + 5) / 0 \cdot 32]^{-0 \cdot 6} \tag{33}$$

$$\eta = [(d_{50} + 5) / 300]^{0 \cdot 25} \tag{34}$$

where d_{50} is the median particle size of the soil (μm).

These relationships were derived from over 500 observations in experiments carried out on a range of materials with median grain sizes (d_{50}) from silt to coarse sand, slopes from 1 to 15 per cent and discharges from 2 to 100 cm^3 cm^{-1} s^{-1}. They are valid for sediment concentrations up to 0·32 which seemed to be an upper limit obtained in the experiments beyond which further increases in stream power caused no further increase in sediment concentration. The need to insert a critical value for unit stream power of 0·4 cm s^{-1} means that the equations cannot be used at very low unit stream powers.

Interrill transport capacity. EUROSEM uses the following interrill flow equations based on experimental work on shallow interrill flow by Everaert (1991), using a range of particle sizes from 33 to 390 μm:

$$TC = \frac{b}{\rho_s q}\left[\left(\Omega - \Omega_c\right)^{0 \cdot 7/n} - 1\right]^{\kappa} \tag{35}$$

where $\kappa = 5$, ρ_s = the sediment density (kg m^{-3}), and b = a function of particle size defined by:

$$b = \frac{19 - (d_{50} / 30)}{10^4} \tag{36}$$

Ω = the modified stream power (g$^{1\cdot5}$ cm$^{-2/3}$ s$^{-4\cdot5}$) defined by Bagnold as:

$$\Omega = \frac{\left(U^* u\right)^{3/2}}{h^{2/3}} \tag{37}$$

in which U^* = shear velocity (m s^{-1}). The symbol Ω_c is a critical value of Ω found by using, in the same formula, the Shields critical shear velocity (White, 1970):

$$U_c^* = \sqrt{y_c(\rho_s - 1) g d_{50}} \tag{38}$$

where y_c is the modified Shields' critical shear velocity based on particle Reynolds number (m s^{-1}), and g is the acceleration due to gravity (m s^{-2}).

CALCULATION OF SOIL EROSION

Hillslope erosion

For each time step and each node along the slope plane, the net rate of erosion (e) and the sediment discharge (product QC) are calculated. Combining Equations 2 and 29, e is obtained as:

$$e = DR + \beta w\, v_s (TC - C) \tag{39}$$

When rates of soil detachment by raindrop impact are sufficiently small and the sediment concentration in the flow exceeds the transport capacity, e becomes negative and represents a net deposition rate. This situation arises when DR is very low or when runoff and sediment are routed from one slope plane to another of lower gradient.

Three cases of surface topography can be simulated, based on the geometry of the rill and interrill areas (see section on 'surface runoff processes'):

(a) the surface may contain no rills, but have some surface irregularities;
(b) the surface may be rilled, with interrill flows routed toward the rills as described by Equation 21;
(c) the surface may be furrowed, or have very dense rills, such that interrill routing is illogical due to the short distance traversed by interrill flows.

For case (a), typical of relatively smooth slope planes, interrill flow is assumed over the entire element, and the flow direction is directly down the plane. Interrill splash and transport relations are used. For case (b), the model simulates both shallow flow and rainsplash erosion between the rills, and downslope flow with much larger carrying capacities in the rills. Interrill flow is routed towards the rills along a slope taken as the vector sum of the slope along the rills and the slope of the surface in a direction normal to the rills; rill spacing on the element is assumed to be uniform. When distance of interrill flow is less than 1 m, as in case (c), interrill routing is inappropriate, and rill input rate q is taken as the rainfall excess rate times the interrill flow distance. Rain flow transport concentrations are used for interrill sediment concentrations. By using case (c), EUROSEM can simulate the furrows produced by agricultural implements and also plough-rill erosion.

Channel erosion

Channel flow and erosion are simulated in EUROSEM using the same general approach adopted for rill erosion. The main differences are that soil detachment by raindrop impact within the channel is neglected and that lateral inflows of sediment from the hillsides (q_s in Equation 1) become important. In the same manner as for rills, C may be explicitly computed for each time step in the finite difference scheme, once the array of hydraulic variables, A, Q and u, is found, starting with the first node below the upper boundary. If there is no input of runoff at the upper end of the channel, the transport capacity at the first node is zero and the boundary condition is set as:

$$C(0,t) = \frac{q_s}{Q + v_s BW} \tag{40}$$

where BW=the bottom width of the channel (m). Otherwise, procedures are precisely the same as for calculation of rill sediment transport. Bank collapse is not simulated.

Catchment representation

As in the KINEROS model, a catchment is represented as a network of surfaces and channels of rather arbitrary complexity. Channels may receive distributed inputs from hillslopes on either or both sides or as a concentrated flow from upstream (as for a headwater area). Channels may also receive input from one or two upstream channels. Hillslopes may be represented as heterogeneous along their flowpaths in slope, width, rill density or other properties by using a cascade of adjacent surfaces (Morgan *et al*., in press).

MODEL APPLICATIONS

We acknowledgte that a validation of this model is beyond the scope of an introductory paper, and this section is rather intended to demonstrate briefly some of the unique capabilities of EUROSEM and dynamic erosion models in general. More extensive field applications are intended for subsequent publications. Data collected from the Woburn Erosion Reference Experiment (Catt *et al*., 1994), a series of erosion plots operated jointly by Silsoe College and Rothamsted Experimental Station at Woburn, Bedfordshire, UK, are used for a simulation to demonstrate the output that can be obtained from EUROSEM (Version 3.4). The experiment is sited on a sandy loam soil of the Cottenham Series, a brown sand, as defined by Avery (1980), classified as an Udipsamment (Soil Survey Staff, 1975) and a Cambic Arenosol (FAO–UNESCO, 1974), developed on Cretaceous Lower Greensand (Catt *et al*., 1974). The plot used for the simulation is 40 m long and 22·5 m wide with a mean slope of 9 per cent. Hydrographs and sedigraphs were recorded using the sediment sampling system described by Vivian and Quinton (1993). To take account of under-recording of the pumped samplers, the sediment concentrations were corrected by multiplying them by the ratio of the total soil loss determined in the field to that determined from the pumped samples. The event chosen for simulation, a storm of 9·8 mm on 17 September 1992, is a characteristic short-duration, high-intensity summer storm for southern England. At the time of the storm, the plot was under a crop of sugar beet with a 70 per cent cover. The main pulse of the storm lasted 20 min with a peak intensity of 126 mm h^{-1} for 1 min (Figure 3).

The output hydrograph and sediment graph for the storm, together with observed data, are illustrated in Figure 3. A best fit for the hydrograph was obtained by a trial-and-error calibration, with all parameters constrained to physically realistic values. EUROSEM appears to simulate quite well the time to peak discharge, the peak discharge and the overall shape of the hydrograph. The sediment graph, for which no calibration was attempted, shows that EUROSEM appears to overestimate the sampled sediment concentrations, but the differences are not large.

EUROSEM can also provide information on changes in surface elevation and rill geometry, as shown in Figure 4 where it can be seen that the greatest changes have taken place at the lower end of the plot where discharge is higher. Figure 5 illustrates the dynamic modelling of the interception process within EUROSEM. Under the sugar beet canopy on the plot, most of the water intercepted by the plant canopy reaches the ground as leaf drip with the rest being made up of stemflow and direct throughfall; only a relatively small amount is held in the interception store. Comparing Figure 5 with the rainfall hyetograph in Figure 3, the relative importance of leaf drip is seen to increase with rainfall intensity.

The dynamics of surface water flow during a storm of rapidly varying rainrates are important in erosion and deposition properties. For most field slope lengths, storms rarely, if ever, reach steady flow, as assumed by

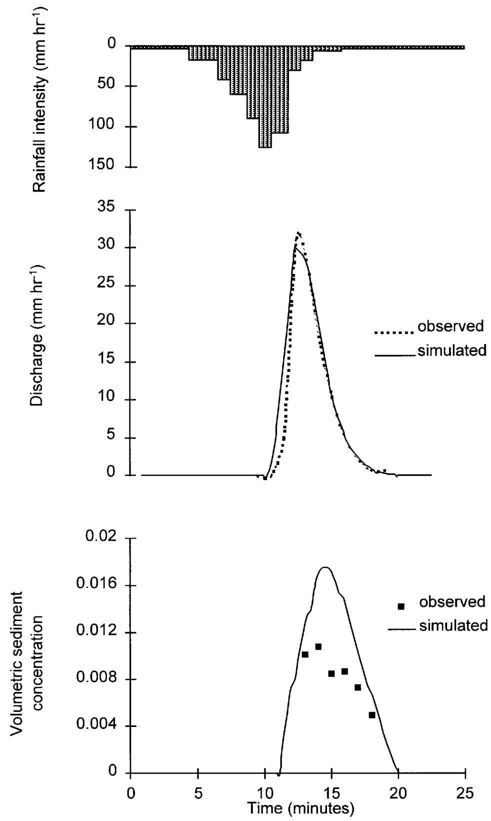

Figure 3. Hyetograph, simulated and observed hydrographs, and simulated and observed sediment graphs for the event of 17 September 1992. (The simulated hydrograph was fitted to the observed data. Although the simulated sediment concentrations appear too high, the simulated total storm soil loss was 57 kg compared with an observed value of 42 kg, suggesting that the measured sediment concentrations are in error.)

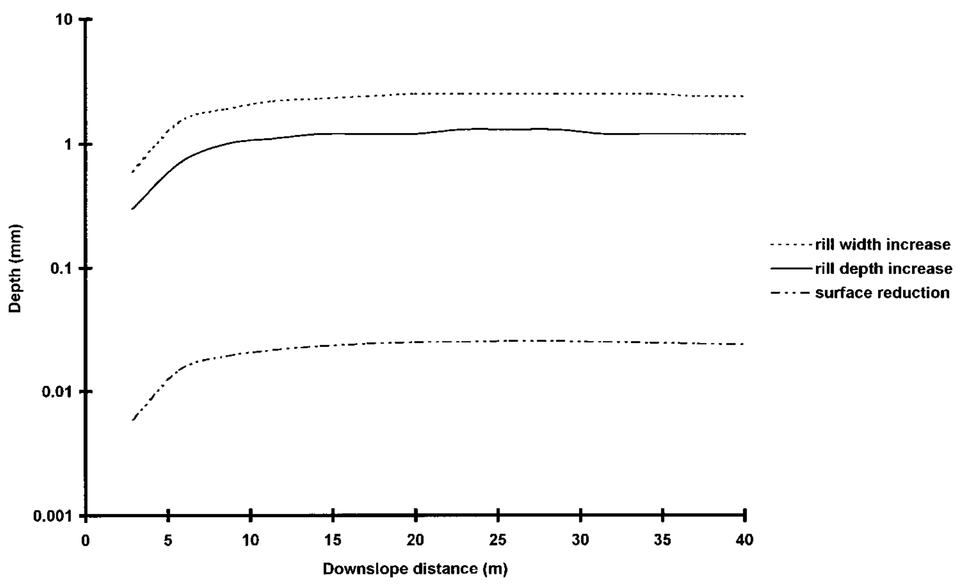

Figure 4. Simulated changes in rill geometry and surface elevation with distance downslope

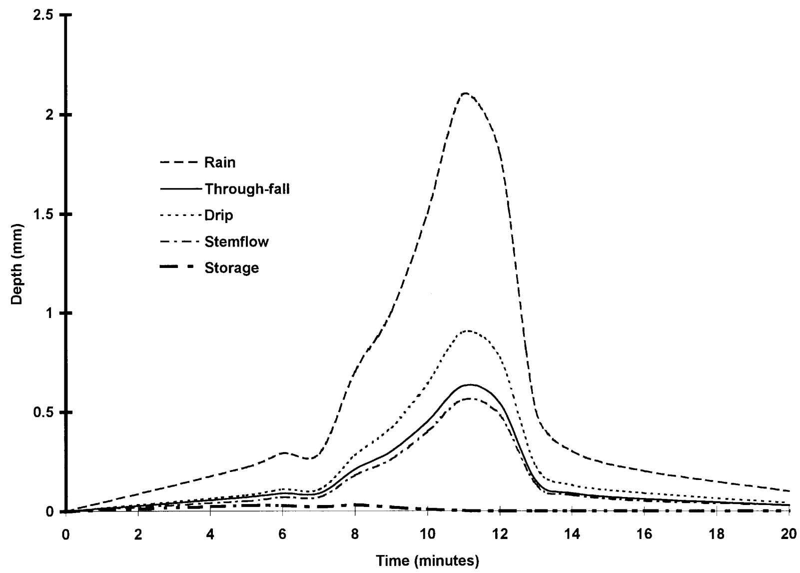

Figure 5. Simulation of the division of rainfall into leaf drip, direct throughfall, stemflow and interception storage

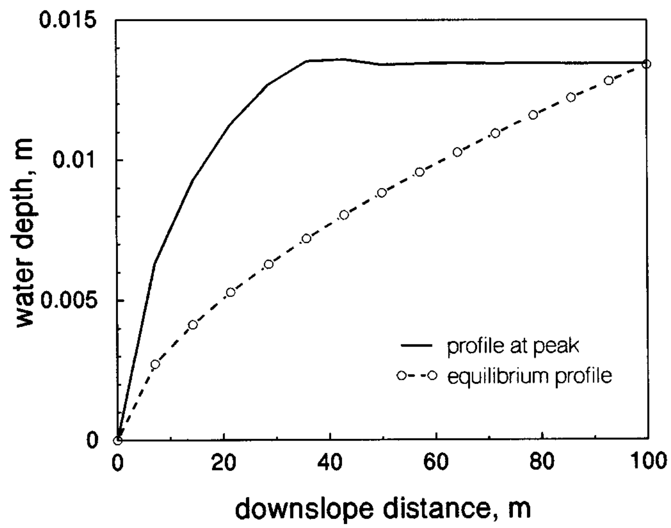

Figure 6. Comparison of the actual water profile with the steady flow profile at peak flow for the storm of Figure 3 on a 100 m slope

several sediment yield models. For the storm of Figure 3, for example, on a 100 m slope, the actual surface water profile at time of flow peak is compared with a steady flow profile in Figure 6. This is a simple slope shape, but it is clear that the steady flow assumption cannot actually predict with much accuracy where erosion and deposition will occur. Since the WEPP methodology finds the peak flow using kinematic water routing, the accuracy of that method will depend critically on the extent to which the sediment concentration at estimated peak flow approximates the overall average *C*. Further, as shown by Smith *et al*. (1995), steady flow assumptions which lie behind the slope length extrapolations in the USLE and RUSLE approaches are considerably in error for longer slopes, for the same reason: steady flow does not occur under erosive storms on longer slopes.

CONCLUSIONS

EUROSEM is a fully dynamic state-of-the-art erosion model, able to simulate sediment transport, erosion and deposition by rill and interrill processes over the land surface in individual storms for both single fields and small catchments. It provides information on total runoff, total soil loss, the storm hydrograph and storm sediment graph. At present, EUROSEM is not able to simulate erosion by ephemeral gullies or by saturation overland flow. Compared with other models, however, EUROSEM provides for explicit consideration of rill and interrill flow, improved simulation of plant cover effects on interception, rainfall energy and flow velocity, transport capacity computations using relationships based on over 500 experimental observations made with shallow surface flows, and specific simulation of the effects of rock fragments (stoniness) on infiltration, flow velocity and splash erosion. The true dynamic nature of the simulation provides advantages over approximate steady flow methods now in use.

Readers interested in obtaining a copy of EUROSEM may visit http://www.silsoe.cranfield.ac.uk/eurosem/eurosem.htm

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