Overland flow
Hydraulics and erosion mechanics

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Modelling long-term soil loss and landform change

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Abstract

EROSION 2D is a physically based computer model for simulating sediment transport on slopes. The model calculates erosion and deposition per unit area, including the resulting changes in slope geometry. The input parameters of the EROSION 2D model are the altitude coordinates of the initial slope profile, the surface and soil properties and the vegetation cover of the slope. The characteristic variables can vary through space (changes in the long profile of the slope) and time. Sediment transport is always calculated on the basis of single precipitation events. These are characterized by the specific soil and canopy-cover conditions at the time of the event, by the duration of rainfall and the temporal variation of rainfall intensity. Several events can be linked to form a sequence representing a month or a year. These sequences can, in turn, be coupled repeatedly, thus extending the simulation period almost indefinitely.

Introduction

A review of landform change over the past millennium in central Europe (Bork 1988) has shown that, with increasing anthropogenic use of the soil, processes of wind and water erosion have increased in intensity and have substantially affected landform change during this period. Soil erosion research has thus become an important field of both historic-genetic and applied geomorphological research. Because of the many parameters involved, early efforts were made to develop mathematical models to describe the interaction of individual factors governing erosion. The first models were purely empirical, the best known being USLE (Wischmeier & Smith 1965). However, the application potential of purely empirical models is very limited. As well as the problem of non-transferability, it is mainly the low spatial and temporal resolution and the neglect of deposition processes that limit the use of empirical models, particularly USLE. For this reason, efforts have been made to develop physically based erosion models. One such physically based model is
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EROSION 2D (Schmidt 1991a). This model describes the detachment, transport and deposition of soil particles, including the resulting changes in slope geometry. Possible applications of EROSION 2D range from purely practical questions – such as planning and calculating soil protection measures (Schmidt 1991b) – to simulating long-term landform change. In addition, the theoretical approach permits the user to view the influence of individual parameters in isolation (sensitivity analyses), thus opening up new possibilities of understanding the complex processes involved.

Description of the model

Because of the complexity of erosion processes it is advisable to split them up into various subprocesses and to represent each of these by an appropriate submodel. EROSION 2D distinguishes the following process components: the detachment of soil particles from the soil surface and the transport of the detached particles by runoff. As well as the complexity of the processes, the model should take into account the great temporal and spatial variations of the parameters governing erosion. To ensure an adequate consideration of parameter behaviour, the model’s equations are applied to small spatial and temporal segments that are in themselves homogeneous. The following sections give an overview of the equations on which the model is based.

Detachment

For an erosional process to occur it is necessary that individual soil particles or small aggregates can be detached from the soil surface. This is only possible when the fluid forces generated by overland flow and raindrops overcome the particle’s cohesion and gravity. Because of the heterogeneous conditions at the soil surface, direct measurement and theoretical description of the forces affecting the soil particles are practically impossible, especially when taking individual soil particles into consideration. For this reason some simplifications cannot be avoided. In particular, it is necessary to move from the microscopic scale of individual particles to a macroscopic view. An expression which summarizes the erosional effects of overland flow and raindrops in that way is the momentum flux exerted by the flow and falling droplets, respectively. The momentum flux \( \varphi_q \) exerted by the flow is defined as:

\[
\varphi_q = w_q \Delta y v_q
\]  

(17.1)

where \( w_q \) is the mass rate of flow; \( \Delta y \) is the width of the slope segment; and
$v_q$ is runoff velocity.

According to Equation 17.2:

$$w_q = q \rho_q$$  \hspace{1cm} (17.2)

the mass rate of flow, $w_q$, is obtained from the volume rate of flow, $q$, and the fluid density, $\rho_q$. To obtain the volume rate of flow, $q$, the following relation is used:

$$q = (r_\alpha - i) \Delta x + q_{in}$$  \hspace{1cm} (17.3)

where $r_\alpha = r \cos \alpha$ is the effective rainfall intensity related to the slope surface, $i$ is the infiltration rate, $\Delta x$ is the length of the slope segment and $q_{in}$ is the inflow into the slope segment from the segment above.

For a sufficiently short time interval the flow velocity, $v_q$, contained in Equation 17.1 can be assumed as uniform. Under this condition the mean velocity of flow may be estimated according to Equation 17.4 from the coefficient of surface roughness, $n$, the slope, $S$, and the depth of flow, $\delta$.

$$v_q = \frac{1}{n} \delta^{2/3} S^{1/2}$$  \hspace{1cm} (17.4)

where

$$\delta = \left[ \frac{q n}{S^{1/2}} \right]^{3/5}. \hspace{1cm} (17.5)$$

Equations 17.4 and 17.5 are based on the Manning equation, which was originally established for turbulent flow in channels. Various experiments have shown that the Manning equation is also valid for overland flow on slopes (Emmett 1970, Pearce 1976), as long as turbulent flow conditions can be assumed. This generally applies in the cases relevant for erosion (Bork 1988, 140). When calculating the flow velocity $v_q$ according to Equation 17.4, it is assumed that runoff is distributed uniformly all over the slope segment. Local variations in flow velocity and flow depth within the slope segment are disregarded.

The momentum flux, $\varphi_{r,\alpha}$, exerted by the falling droplets is defined by analogy to Equation 17.1 as:
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\[
\varphi_{r,\alpha} = w_r A v_r \sin \alpha \tag{17.6}
\]

where \(w_r\) is the mass rate of rainfall, \(A\) is the area of the slope segment, \(v_r\) is the mean fall velocity of the droplets, and \(\alpha\) is the slope angle. The effect of a canopy cover on rainfall impact is taken into account by the use of a groundcover index, defined as \(C_L = A_{leaf}/A\), where \(A_{leaf}\) is the segment area covered by vegetation or plant residues and \(A\) is the entire segment area. The groundcover index may be combined with Equation 17.6 to give:

\[
\varphi_{r,\alpha} = w_r A v_r \sin \alpha (1-C_L) \tag{17.7}
\]

The mass rate of rainfall is determined by:

\[
w_r = r_\alpha \rho_r \tag{17.8}
\]

where \(r_\alpha\) is the effective rainfall intensity related to the slope surface and \(\rho_r\) is the fluid density of rainfall.

Substitution of this expression into Equation 17.7 yields:

\[
\varphi_{r,\alpha} = r_\alpha \rho_r A v_r \sin \alpha (1-C_L) \tag{17.9}
\]

The fall velocity of raindrops contained in Equation 17.9 is very difficult to measure under field conditions. Available data show that the size, and hence the velocity, of the droplets increases with rainfall intensity (Laws 1941, Laws & Parsons 1943). By making use of these data, we obtain the following empirical equation (Eq. 17.10), which provides a simple method of estimating the mean fall velocity of raindrops on the basis of rainfall intensity data:

\[
v_r = 4.5 r^{0.12} \tag{17.10}
\]

Here \(v_r\) is the mean fall velocity of raindrops in m s\(^{-1}\) and \(r\) the rainfall intensity in mm h\(^{-1}\).

In order to characterize the erosional resistance of the soil, use is made of the fact that the occurrence of a measurable rate of erosion presupposes a minimum rate of overland flow, \(q_{crit}\) (Hjulström 1935). Substitution of \(q_{crit}\) in Equations 17.1 and 17.2 yields the critical momentum flux \(\varphi_{crit}\), with which the specific erosional characteristics of the soil can be described, similar to the previously derived equations for the erosional effects of overland flow and raindrops:
\[ \varphi_{\text{crit}} = q_{\text{crit}} \rho q \Delta y v_q. \]  

(17.11)

Here \( q_{\text{crit}} \) is the volume rate of flow at initial erosion (as a function of soil type, state of tillage, etc.). In addition, \( \rho q \) is the fluid density, \( \Delta y \) is the width of the specified slope segment, and \( v_q \) is the flow velocity according to Equation 17.4. \( q_{\text{crit}} \) has to be determined experimentally for a given soil.

Because of their formal conformity, the model concepts describing the erosional effects of raindrops and overland flow can be linked with the concept describing the soil's resistance to give a dimensionless coefficient, \( E \):

\[ E = \frac{\varphi_q + \varphi_{r,\alpha}}{\varphi_{\text{crit}}}. \]  

(17.12)

This coefficient characterizes the ability of a given flow \((q > 0)\) to detach particles from the soil surface. Erosion occurs when \( E > 1 \), which means that the erosional effects of overland flow and raindrops (given by the momentum fluxes \( \varphi_q \) and \( \varphi_{r,\alpha} \)) exceeds the soil's resistance to erosion (given by the critical momentum flux \( \varphi_{\text{crit}} \)). \( E \leq 1 \) characterizes the erosion-free state of flow.

For quantitative results, the coefficient \( E \) is correlated with experimental data. Fifty experiments under simulated rainfall have been performed in a test flume filled with silty soil (Schmidt 1988). The data can be fitted by the following regression equation:

\[ q_s = (1.75E - 1.75) \times 10^{-4} \]  

(17.13)

where \( q_s \) is the sediment discharge of detached particles. Figure 17.1 shows the regression curve and the experimental data on which the curve is based. Because of the theoretical postulate that sediment cannot be eroded when \( E \leq 1 \) the regression curve must intersect the \( x \)-axis at \( E = 1 \). The mean relative deviation of the values calculated from Equation 17.13 from the measured values is \( \pm 20\% \). Taking into account the inaccuracies in the measured values, we may presume that the theoretically derived erosion coefficient adequately describes the main factors governing detachment.

Transport

Due to gravity, the sediment particles suspended in a fluid sink to the bottom with a velocity largely dependent on particle size. This process can only be stopped or delayed when it is counteracted by a sufficiently large, vertical (turbulent) flow component. Hence, the size-dependent settling velocity of the particles and the vertical turbulence component within the flow are decisive for
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![Graph showing sediment discharge vs. erosion coefficient]

**Figure 17.1** Measured sediment discharge, $q_s$, vs. erosion coefficient, $E$.

The suspended transport of particles. The settling velocity of a particle in a stagnant fluid is described by the Stokes equation:

$$v_p = \frac{1}{18} D^2 \frac{(\rho_p - \rho_f) g}{\eta}$$  \hspace{1cm} (17.14)

in which $D$ is particle size, $\rho_p$ particle density, $\rho_f$ fluid density, $g$ acceleration of gravity, and $\eta$ fluid viscosity.

If the settling velocity, $v_p$, is multiplied by the mass rate of the settling particles, $w_p$, and by the segment area, $A$, we obtain the critical momentum flux of the suspended particles, $\varphi_{p,\text{crit}}$ (analogous to $\varphi_{\text{crit}}$), below which the particles are not maintained in suspension:

$$\varphi_{p,\text{crit}} = w_p \ A \ v_p$$  \hspace{1cm} (17.15)

The mass rate of particles, $w_p$, from Equation 17.15 can be expressed as:
\[ w_p = c \rho_p v_p \]  
(17.16)

where \( c \) is the concentration of suspended particles in the fluid, \( \rho_p \) is the particle density and \( v_p \) is the settling velocity of the particles according to Equation 17.14.

The critical momentum flux of the suspended particles, \( \varphi_{p, \text{crit}} \), is counteracted by the vertical momentum flux component of the flow, \( \varphi_{q, \text{vert}} \), which is assumed to be a fraction of the total momentum flux exerted by the flow and the falling droplets respectively.

\[ \varphi_{q, \text{vert}} = \frac{1}{\kappa} (\varphi_q + \varphi_r) \]  
(17.17)

where \( \kappa \) is a factor.

When transport capacity has been reached, the vertical momentum flux component of flow equals the critical momentum flux of the suspended particles as defined in Equation 17.15:

\[ \varphi_{p, \text{crit}} = \varphi_{q, \text{vert}} \]  
(17.18)

Substituting Equations 17.15, 17.16 & 17.17 into Equation 17.18 gives:

\[ c_{\text{max}} \rho_p A v_p^2 = \frac{1}{\kappa} (\varphi_q + \varphi_r) \]  
(17.19)

where \( c_{\text{max}} \) is the concentration of particles at transport capacity.

Rearranging Equation 17.19 yields:

\[ c_{\text{max}} = \frac{1}{\kappa} \frac{\varphi_q + \varphi_r}{\rho_p A v_p^2} \]  
(17.20)

Transport capacity is then determined according to:

\[ q_{s, \text{max}} = c_{\text{max}} \rho_p q. \]  
(17.21)

According to Equation 17.21, it is possible to calculate the transport capacity for any particle size class separately. The transport capacities thus derived specify the maximum mass rate of particles that can be transported within this size class under the given flow conditions (assuming that transport is not limited by detachment). In order to determine the actual mass rate and the particle size...
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distribution of the transported sediment, the following assumptions are made:

(a) The particle size distribution of the detached sediment is the same as in the original soil.

(b) Detachment occurs only if there is excess transport capacity.

This means that the particle size distribution of the transported sediment corresponds to that of the initial soil, as long as the mass rate of the detached particles does not exceed the transport capacity in any of the particle classes considered. If that is not the case, the mass rate of the particles and hence the size distribution of the transported sediment is controlled by transport capacity.

Erosion/deposition

In order to calculate the rate of erosion or deposition for each of the individual slope segments the following simple equation is used:

\[
\gamma = \left( \frac{q_{s,\text{in}} - q_{s,\text{out}}}{\Delta x} \right)
\]  

(17.22)

where \( \gamma \) is the rate of erosion (\( \gamma < 0 \)) or deposition (\( \gamma > 0 \)) per unit area, \( q_{s,\text{in}} \) is the sediment discharge into the segment from the segment above, \( q_{s,\text{out}} \) is the sediment discharge out of the segment and \( \Delta x \) is the length of the slope segment.

Program structure and operation

In order to make the model applicable for practical purposes the equations described above have been transferred to a computer program called EROSION 2D.

Input parameters and data management

EROSION 2D uses three main groups of input parameters: relief, soil and precipitation. These groups consist of the following individual parameters: The program also takes into account a number of other parameters, which are fixed and cannot be influenced by the user. They include fluid and particle density, fluid viscosity, and acceleration due to gravity.

Each group of parameters mentioned in Table 17.1 is organized in a separate file. The relief parameter file contains the geometry data of the initial slope.
Table 17.1 Input parameters.

<table>
<thead>
<tr>
<th>Relief</th>
<th>Soil</th>
<th>Precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope length</td>
<td>Grain size</td>
<td>Precipitation intensity</td>
</tr>
<tr>
<td>Slope geometry</td>
<td>Infiltration rate</td>
<td></td>
</tr>
<tr>
<td>(x-, y- coordinates)</td>
<td>Resistance to erosion</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(crit. momentum flux)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Surface roughness</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Manning’s n)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Canopy cover</td>
<td></td>
</tr>
</tbody>
</table>

profile. The file comprises at least two pairs of coordinates, denoting the top and the bottom point of the slope profile. (However, more than two points are generally necessary to represent the slope geometry adequately.) The soil parameter file contains the soil- and canopy-specific data of any number of homogeneous slope segments $\geq 1$ m, and the precipitation parameter file comprises the duration and the rainfall intensity data in 10 min intervals. Unlike the relief parameters, the soil and precipitation parameters refer to one actual event or to a time section of such an event $\geq 10$ min.

Output parameters

Table 17.2 shows the output parameters supplied by the program.

Table 17.2 Output parameters.

<table>
<thead>
<tr>
<th>Area-related parameters</th>
<th>Point-related parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rate of erosion per unit area</td>
<td>Sediment discharge</td>
</tr>
<tr>
<td></td>
<td>Sediment concentration</td>
</tr>
<tr>
<td>Rate of deposition per unit area</td>
<td>Clay, silt and sand fractions of the transported sediment</td>
</tr>
</tbody>
</table>

The area-related parameters each relate to the pre-set slope segments ($\Delta x = 1$ m, $\Delta y = 1$ m) or to their multiples ($\Delta x = 1, 2, 3, \ldots n$ m, $\Delta y = 1$ m). The point-related parameters refer to a specific, user-selected slope position (e.g. the sediment concentration at the position $x = 156$ m). The time reference basis for all output parameters is the internally determined time interval ($\Delta t = 10$ min) or a corresponding multiple ($\Delta t = 20, 30, 40, \ldots n$ min).
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Long-term simulation
The long-term option of EROSION 2D is based on the coupling of any number of individual erosional events. Two complementary procedures are employed. The first procedure combines several events into a sequence, for example the erosional events of 1 year or of a particular season. The program automatically identifies and processes sequential files (containing the corresponding soil and precipitation data of each event). The sequential procedure is particularly suitable for simulating the seasonal or annual rate of erosion. If a longer period (several years to several hundred years) is to be simulated, an iterative procedure is employed, based on either a single event or on a sequence of events.

In both the sequential and the combined sequential-iterative procedures, the erosion/deposition-dependent changes in slope geometry are taken into account. After processing each of the successive events the slope profile is modified according to the calculated rates of erosion and deposition. At the end of simulation the initial and final slope profiles and the cumulated rates of erosion and deposition are displayed in graphical and numerical form.

Sensitivity analysis
In the following, some hypothetical examples are used to examine the dependence of predicted soil loss upon slope length and geometry, gradient, rainfall intensity and soil properties. First, the impact of slope length and geometry is investigated, using the following variants:

(a) straight slope;
(b) convex slope;
(c) concave slope;
(d) convex-concave slope; and
(e) irregularly shaped slope.

In all cases, total slope length is 130 m and the difference in height between the top point of the slope and the base is 14 m.

Figure 17.2 shows the results for the straight slope. The top half of Figure 17.2a describes the slope profile, the lower half the calculated soil loss
Figure 17.2 Calculated soil loss and sediment discharge on a straight slope.

(negative values) or deposition (positive values). We see that soil loss per unit area increases with slope length. However, the increase in soil loss diminishes as slope length increases. If, instead of soil loss, the sediment discharge is plotted against slope length (Fig. 17.2b), a comparison of both curves shows that the increase in soil loss on the upper slope is linked to an overproportional increase in sediment discharge. On the lower slope, however, sediment discharge increases more or less linearly, and soil loss per unit area is
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Figure 17.3 Calculated soil loss and sediment discharge on a convex slope.

approximately constant.

In the case of the convex slope also, soil loss increases with slope length (Fig. 17.3). In contrast to the straight slope, the increase in soil loss is larger with increasing slope length. The same is true of the sediment discharge.

The results are completely different in the case of the concave slope. Here soil loss per unit area is more or less constant throughout the entire slope (Fig. 17.4a). Accordingly, the increase in sediment discharge with slope length is approximately linear (Fig. 17.4b).

Figure 17.5a shows soil loss at a convex-concave slope. As expected, soil
loss increases with slope length on the upper, convex slope section, reaching a maximum in the middle third of the profile and rapidly decreasing at the transition to the lower, concave slope section. At the base of the slope, soil loss decreases to almost zero. Sediment discharge (Fig. 17.5b) first increases exponentially, then almost linearly with slope length, finally reaching an almost constant value. (As Fig. 17.5a shows, soil loss is then practically zero).

In the case of the irregular slope (Fig. 17.6a), soil loss peaks at the convex slope sections, while the concave part at mid-slope undergoes very little erosion. Sediment discharge increases exponentially in each of the convex
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Figure 17.5 Calculated soil loss and sediment discharge on a convex–concave slope.

slope sections. By contrast, the concave sections have approximately constant values (Fig. 17.6b).

As the preceding figures have shown, the simulated variations in slope geometry affect both the relation of soil loss to slope length and the absolute amount of soil loss. Since slope length and difference in height have been kept constant, it is possible to compare the net amounts of soil loss calculated for each variant. Taking as reference value the soil loss on the straight slope (= 100%), we obtain the diagram shown in Figure 17.7. It is obvious that the
convex slope loses by far the most soil (139%), exceeding even the straight slope. The convex–concave slope loses least soil (57%). This is particularly interesting because most natural slopes belong in this category.

A further decisive factor governing soil loss is slope gradient. The straight slope demonstrates this most clearly (cf. Fig. 17.2). In a series of nine simulation runs, slope inclination was increased in steps of $5^\circ$. In Figure 17.8 the predicted soil loss (related to soil loss at $45^\circ$) is plotted against slope angle. We see that soil loss increases as the slope becomes steeper. However, the increase in soil loss decreases exponentially with increasing slope angle.
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Figure 17.7 Dependence of soil loss upon slope geometry.

Figure 17.8 Soil loss as a function of slope gradient.
Figure 17.9 Soil loss as a function of rainfall intensity.

Figure 17.9 describes the relationship between soil loss and rainfall intensity. Again, the simulations are based on the straight slope shown in Figure 17.2. Rainfall intensity was varied in steps of 0.1 mm min\(^{-1}\), rainfall duration remaining constant. Taking as reference value the soil loss at 1.0 mm min\(^{-1}\), Figure 17.9 shows the predicted soil loss as a function of rainfall intensity. (Here, rainfall intensity relates to the excessive rainfall, i.e. the infiltration rate has already been subtracted.) We see that increasing intensity is always linked to increasing soil loss.

In the model, the soil and surface properties are described by means of the following parameters: grain size, resistance to erosion, surface roughness and canopy cover. Resistance to erosion and surface roughness are, in turn, dependent on grain size and canopy cover. The canopy cover, in particular, is subject to considerable spatial and temporal fluctuations, depending on land use. Because of the interdependence of the various parameters, it is not practical to vary one of the above parameters separately while the others remain constant. Since the simulation examples described in the following section give an idea of the interaction of the various individual parameters, further analyses can be dispensed with here.
Figure 17.10 Single events 1–4 (Profile A).

Figure 17.11 Single events 1–4 (Profile B).
Long-term simulations

Data
Simulations are based on four, sequentially linked events representing different seasonal conditions. Event 1 applies to a bare slope and represents the start of the vegetation period. The events shown in Table 17.3 characterize conditions of increasing cover. The rainfall duration of each individual event is 40 min. Rainfall intensity varies according to Table 17.4. Beyond that, the following variations in agricultural use are taken into account: I, tillage over the entire slope; II, tillage interrupted by strips of permanent vegetation (green strips); III, tillage followed by permanent vegetation (e.g. grassland) at the bottom of the slope.

<table>
<thead>
<tr>
<th>Table 17.3 Soil and canopy data.</th>
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<tbody>
<tr>
<td><strong>Event</strong></td>
</tr>
<tr>
<td>Infiltration rate (mm min⁻¹)</td>
</tr>
<tr>
<td>Resistance to erosion ([kg m] s⁻²)</td>
</tr>
<tr>
<td>Surface roughness (s m⁻¹s)</td>
</tr>
<tr>
<td>Canopy cover (%)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Table 17.4 Rainfall intensity data (in mm min⁻¹).</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Event</strong></td>
</tr>
<tr>
<td>Time</td>
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</tbody>
</table>

Single events
Looking at each of the single events (Figs 17.10 & 17.11), it is easy to see how they differ. Precipitation was clearly less heavy, but more material was redeposited during the first event than during the following ones. In particular, deposition at the slope base was considerably greater than in the following events. The main reasons for this are the lack of plant cover during event 1.

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(start of the vegetation period) and the decreased resistance to erosion offered by the tilled soil.

Long-term simulations
The following long-term simulations are based on the iteration of the sequence described above. Figures 17.12–17.16 each show the initial slope profile and the superimposed curve of the slope profile as calculated after 250 iterations. In addition, the cumulated soil loss and deposition is plotted in the lower half of the diagrams.

![Profile A](image)

Figure 17.12 Long-term simulation (Profile A/I).

Figures 17.12 & 17.13 show conditions when the slope is used for tillage only. As Figure 17.12 shows, a more or less smooth, convex–concave slope (Profile A) undergoes almost parallel backward displacement with no substantial change in geometry. This is not the case when slope shape is more complex (Profile B, Fig. 17.13). Here erosion tends to smooth down the convex parts of the slope overlying the general slope profile. This is particularly noticeable on the lower parts of the slope, owing to the increase of runoff with slope length. In comparison, the concave section mid-slope is subject to considerably less erosion.

The examples below show the geometrical changes of the slope profile when slope use is not homogeneous. Figure 17.14 describes conditions when the
Figure 17.13 Long-term simulation (Profile B/I).

Figure 17.14 Long-term simulation (Profile A/II).

slope is partitioned into three sections, divided from each other by strips of permanent vegetation. The green strips decelerate surface runoff and thus lead
to partial or complete deposition of the soil eroded further up slope. Runoff is therefore almost devoid of sediment when leaving the green strips, and erosion starts again below these strips with increased intensity. The
combination of both these processes — accumulation within the green strips, increased erosion below them — leads to the formation of small steps. As Figure 17.14 shows, these steps are more pronounced in the lower part of the slope than in the upper because runoff increases with slope length.

In the examples depicted in Figures 17.15 & 17.16 the upper slope is tilled and the lower part is covered by grassland. As in the case of the green strips, soil eroded up slope is deposited in the grassland down slope. The curve of sediment deposition shows that sedimentation quickly reaches a maximum when overland flow enters the grassland and then continuously decreases towards the bottom of the slope. The steady accumulation of soil in the transitional area between tilled soil and grassland finally also leads to the formation of a small step; however, unlike the steps described above, this is purely colluvial in origin.

In the last example (Fig. 17.16) it is assumed that the base level of the slope rises by 0.5 mm per iteration step, owing to the deposition of flood sediments. In contrast to the previous examples, the slope foot merges into a completely flat area of sedimentation. In other respects the profile does not differ substantially from the curve in Figure 17.15.

Assessment of results

The results yielded by simulation seem plausible. Hence we may assume that the model accurately describes the processes of erosion by water. To validate the model further it is necessary to compare its results with field data. Such studies are currently underway. Initial results are encouraging: an average deviation of 18% was noted. This is within the desired accuracy limits. However, since comparative data are scarce, it is not yet possible to give a final assessment of the model’s accuracy.

Symbols and units

\[ A \quad \text{area of slope segment} \quad \text{m}^2 \]
\[ C_L \quad \text{ground cover} \quad \text{m} \]
\[ c_{\text{max}} \quad \text{concentration of particles at transport capacity} \quad \text{m}^3 \text{m}^{-3} \]
\[ D \quad \text{particle size} \quad \text{m} \]
\[ E \quad \text{erosion coefficient} \quad \text{m} \]
\[ g \quad \text{acceleration due to gravity} \quad \text{m} \text{s}^{-2} \]
\[ i \quad \text{infiltration rate} \quad \text{m} \text{s}^{-1} \]
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\( n \) roughness coefficient
\( q \) volume rate of flow
\( q_{\text{crit}} \) volume rate of flow at initial erosion
\( q_s \) sediment discharge
\( r \) rainfall intensity
\( S \) slope gradient
\( \Delta t \) time interval
\( v_p \) settling velocity of soil particles
\( v_q \) mean flow velocity
\( v_r \) mean fall velocity of raindrops
\( w_p \) mass rate of particles
\( w_q \) mass rate of flow
\( w_r \) mass rate of rainfall
\( \Delta x \) length of slope segment
\( \Delta y \) width of slope segment
\( \alpha \) slope angle
\( \gamma \) erosion (\( \gamma < 0 \)) or deposition (\( \gamma > 0 \))
\( \delta \) mean flow depth
\( \eta \) fluid viscosity
\( \kappa \) factor
\( \rho_p \) particle density
\( \rho_{q,r} \) fluid density
\( \varphi_{\text{crit}} \) critical momentum flux
\( \varphi_{p,\text{crit}} \) critical momentum flux of suspended particles
\( \varphi_q \) momentum flux exerted by flow
\( \varphi_{q,\text{vert}} \) vertical momentum flux component of flow
\( \varphi_r \) momentum flux exerted by raindrops

\( s \) \( m^{-1/3} \)
\( m^3 (m s)^{-1} \)
\( m^3 (m s)^{-1} \)
\( kg (m s)^{-1} \)
\( m^{-1} \)
\( m m^{-1} \)
\( s \)
\( m s^{-1} \)
\( m s^{-1} \)
\( m s^{-1} \)
\( kg (m^2 s)^{-1} \)
\( kg (m s)^{-1} \)
\( kg (m^2 s)^{-1} \)
\( m \)
\( m \)
\( kg (m^2 s)^{-1} \)
\( m \)
\( kg (m s)^{-1} \)
\( kg m^{-3} \)
\( kg m^{-3} \)
\( kg m s^{2} \)
\( kg m s^{2} \)
\( kg m s^{2} \)
\( kg m s^{2} \)

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