EROSION MODELLING AS SUPPORT FOR LAND MANAGEMENT IN THE LOESS BELT OF FLANDERS

EROSIEMODELLERING ALS ONDERSTEUNING VOOR LANDBEHEER IN DE VLAAMSE LEEMSTREEK

door

ir. Jan Biesemans

Thesis submitted in fulfilment of the requirements for the degree of Doctor (Ph.D.) in Applied Biological Sciences

Proefschrift voorgedragen tot het bekomen van de graad van Doctor in de Toegepaste Biologische Wetenschappen

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PREFACE

In January 1996 I started this IWT-granted Ph.D. research. For me, this was the start of an excellent and absorbing period. Many people helped me write this work – some, often without knowing it. I had the luck to work in a department (Department of Soil Management and Soil Care), having all the facilities a modern scientist can dream of: specialists in every field (not only the scientific...), modern information technology, and an extensive laboratory.

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It is my hope that this work may have a contribution in the conservation of the quality of our most important natural resources, soil and water.

April 2000
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Chapter 1

General Introduction

Although soil erosion by surface runoff is not that spectacular in Belgium as in some other parts of the world, the soil loss amounts may not be underestimated. Most sediment that causes off-site silting problems in watersheds within the Belgian loess belt, is produced by small rill erosion on agricultural fields. Severe gully erosion is rather rare. Because rill and sheet erosion is seemingly invisible and the reduction in crop productivity is almost neglectable, soil erosion is generally considered as only a marginal problem. However, the sometimes severe flooding effects of high intensity rains in the summer or long-lasting low intensity rains in the winter months illustrate frequently the opposite. Therefore, it can be stated that on-site sediment control remains one of the most important factors to ensure a good drainage capacity of ditches, brooks and rivers.

Water erosion has both on-site and off-site consequences. As on-site consequences in the Belgian loess belt, one can quote: the removal of the agricultural versatile loamy/silty soil layers and the possible outcrop of Tertiary sands (often rich on stones and pebbles) or heavy Tertiary clays. An example of off-site erosion problems is given in figure 1.1, illustrating the most frequent off-site consequence of erosion. In the winter period, many fields remain bare (e.g. after the harvest maize or beets). During this period there are many long-lasting low intensity rains. The topsoil becomes totally saturated and from then on, every rain event (even very low intensity rains) produces a gentle sheet and small rill runoff, inducing the silting of the drainage ditches and culverts. To solve the problem, the ditches are dredged. But this maintenance operation has destroyed the bank vegetation, resulting in accelerated sediment transport processes.
As figure 1.1 shows, the only way to transport the water was by putting drain-pipes in the ditches.

Figure 1.1: *Examples of off-site silting problems in the community of Heuveland (West-Flanders, Belgium, 5 January 1995).*

The off-site problems of water erosion are much more important than the on-site consequences. Therefore, governmental organizations feel the need for an environmental action programme. In order to establish a regional soil conservation policy, an ‘erosion control workgroup’ was founded in January 1995 as an initiative of the regional land management board ‘Regionaal Landschap Westvlaamse Heuvels’ and financed by the province of West-Flanders. The purpose of the workgroup was to make the farmers and governmental organizations sensible for the erosion problems, through erosion demonstration plots and workshops. Another purpose of the workgroup was to evaluate the alternative land management scenarios, generated by the expert systems described in this work. These alternatives must be acceptable for both farmers and environmental policy makers. The erosion control workgroup was an important link between the pure scientific research and the practice. Therefore, the applicability of the developed methodology was always a central guideline throughout this research.
As can be derived from the problem definition, the main purpose of the study presented here, was to develop an erosion expert system that can be used to support a soil and water conservation policy. The expert system must be capable to map the problem areas on a parcel level, so that the farmers with problem fields can be directly located. For these fields, the expert system must also be usable to design soil conservation measures and to evaluate the efficiency of all possible scenarios.

Before an erosion control policy can be made, the problem areas must be mapped. For these tasks the empirical RUSLE (Revised Universal Soil Loss Equation) (Renard et al., 1996) model was chosen. The RUSLE is basically a 2D model: it calculates the soil loss along a hillslope. A methodology was developed to adapt it to a 3D reality. Because modelling environmental processes is very complex, and the model itself and the model input factors are inherently associated with a certain amount of uncertainty, attention was focused on the error propagation techniques in the calibrating/validating stage of the model. The expert system for the regional mapping of long-term soil loss and sediment accumulation is available as an ANSI C program, which act directly on input maps of IDRISI (Eastman, 1997), a popular Geographical Information System (GIS). Because this program uses only standard C functions, it can be executed on every computer platform.

The design of alternative land management scenarios focused on the reduction of the sediment transport towards the drainage system involves a model, a uniform methodology, for all possible measures which one wants to evaluate on their efficiency. Because the RUSLE does not account for all erosion control measures (e.g. grassed riparian buffer zones), an event-based physical model was constructed. This model was based on the EROSION-2D/3D model, developed by Schmidt (1996). However, during the implementation the hydrological component and sediment transport module was totally redesigned. The total amount of input parameters and the general structure was maintained. This new physical model was given the general name ‘Sediment Transport Model’ (STM-2D/3D). The research resulted in an erosion expert system (a 2D and 3D version) for the mapping of the risk areas and the design of erosion control measures. The 2D hillslope version was translated into a Java applet and is accessible for the policy makers through the World Wide Web (WWW). The 3D watershed version was translated into a platform independent Java program.
Chapter 1: General Introduction

The methodology used to develop, calibrate and validate the erosion expert systems is described in the following five chapters:

1. *Chapter 2*: Physical Geography of the Study Area
2. *Chapter 3*: Water Erosion Processes
3. *Chapter 4*: Predicting Long-term Sediment Transport
4. *Chapter 5*: Predicting Event-based Sediment Transport
5. *Chapter 6*: Erosion Control Measures

The source code of the STM-2D Java applet is given in the Appendix. This applet illustrates the general concept of the physical erosion model. The source codes of the STM-3D and the RUSLE model are also available but not listed in this work and are free of use and modification on the condition that its source is cited.
Chapter 2

Physical Geography of the Study Area

2.1 Introduction

The experimental sites for this research are located in the very south of the province of West-Flanders (Belgium), in the community of Heuvelland and a small part of the community of Ieper. Figure 2.1 gives a geographical overview.

The calibration and validation phase of the RUSLE model (Renard et al., 1996), used to predict long-term off-site sediment accumulation and on-site soil loss (chapter 4), was done in the Kemmelbeek watershed. This watershed of 1075 ha feeds a drinking-water reservoir of the city of Ieper. The reservoir has a mean water production of 4000 m³·day⁻¹ and long-term recordings of sediment deposition are known. Because the sediment trapping efficiency is close to 100 %, this watershed was suitable to validate the RUSLE model. Also, the brook system intersects the landscape deeper and deeper throughout the decades. Therefore, it might be expected that all sediment loss from the bordering agricultural fields into this brook network will finally be transported towards the reservoir.

The test site for the development of the physical STM-2D/3D model, used to predict soil loss during single rainfall events (chapter 5), is also shown in figure 2.1. It is a small catchment of 142 ha, located inside the Kemmelbeek watershed. In this watershed, precipitation and discharge were measured at the outlet. This data was indispensable to test and validate the concepts of the hydrological submodel.
Figure 2.1: Geographical position of the community of Heuvelland in Belgium. The experimental watersheds and some important hilltops are indicated (elevation in m), together with the river system.

2.2 Short historic overview

Due to the specific landscape, the southern part of West-Flanders always had a special attractive force on people. The hilly ridge that intersects the landscape had a very strategic function: when weather conditions are good the range of vision from the hilltops is almost 30 to 50 km. It is therefore not surprising that this region was the scene of many battles. The communities of Heuvelland, Ieper and Passendale are worldwide known as ‘Flanders Fields’, the battle fields of World War I, the Great War. The dozens of cemeteries scattered throughout the landscape are also the first spots that attract the attention when taking a walk. Within the community of Heuvelland there are more than thirty British military cemeteries. Every year, even today, farmers are ploughing-up dozens of still active bombs and shells. The many pools in the landscape
are impact craters from that period. An example is shown in figure 2.2, a small extract of an orthophoto.

Figure 2.2: On this small extract of an orthophoto, four large impact craters dating from WW I are visible (Wijtschate, Heuvelland).

The earliest signs of human activity date from the Neolithic period (Hellegat, Westouter). First permanent settlements date from the pre-romanic period (450–300 BC). On the top of the Kemmelberg (151 m, see figure 2.1) the Celt built a fortification. However, most of this site was destroyed during the battle of Kemmel (25 April 1918). Traces of human activity during the Roman period are found, but most homestead areas in the community of Heuvelland originate from the Merovingic period (end of the 5th century). These areas originated as settlements of stock-farmers. Agriculture was since then and is up to now the main activity of the autochtone population. Only in the Medieval times there were some industrial activities, focused on the production of textile.

Because of the central position within Europe, its progressive industry, economy and culture, Flanders was many times a factor of dispute between the political centers of Europe during the last millennium. Throughout the centuries Flanders was alternately controlled by the French, Spanish, Austrian and Dutch. However, the absolute focus point in the history of the community of Heuvelland is situated at the beginning of the 20th century. Between October 1914 and October 1918, the frontline between
the German and allied troops was located across the community of Heuvelland. The chain of hills of West-Flanders was a defense for the important ports of Calais and Dunkerque, in the very north of France. These ports were vital for the allied supplies. Because of this strategic importance, there was a constant battle in order to control the hilltops.

2.3 Climate

The climate in the study area is humid with a moderate temperature regime. Due to the oceanic influences, the summers are cool and the winters are mild. Table 2.1 gives an overview of the most important average climatological parameters of the weather station in Vlamertinge (community of Ieper, 3–5 km from the community of Heuvelland). Because precipitation is the most important climatological parameter with regard to water erosion, this section concentrates on the rainfall properties.

Table 2.1: Some average climatological parameters (station Vlamertinge) of the study area (Poncelet and Martin, 1947).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>mean yearly air temperature</td>
<td>10.0 °C</td>
</tr>
<tr>
<td>mean air temperature of the coldest month (January)</td>
<td>2.9 °C</td>
</tr>
<tr>
<td>mean air temperature of the warmest month (July)</td>
<td>17.5 °C</td>
</tr>
<tr>
<td>frost-free period</td>
<td>195 ± 22 days</td>
</tr>
<tr>
<td>average first day of frost</td>
<td>30 October ± 14 days</td>
</tr>
<tr>
<td>average latest day of frost</td>
<td>25 April ± 15 days</td>
</tr>
<tr>
<td>mean yearly rainfall</td>
<td>825 mm</td>
</tr>
<tr>
<td>mean rainfall in the driest month (February)</td>
<td>50 mm</td>
</tr>
<tr>
<td>mean rainfall in the wettest month (October)</td>
<td>85 mm</td>
</tr>
</tbody>
</table>

There is a clear effect of the elevation on the annual precipitation in Belgium. A regression analysis based on 361 weather stations (Alexandre et al., 1992) resulted in the following equation:

\[ P = 715 + 0.8 \cdot E \]  \hspace{1cm} (2.1)

wherein, \( P \) is the average annual precipitation in mm and \( E \) the elevation in m. The
vertical gradient is 80 mm per 100 m increase in elevation. The coefficient of determination \(R^2\) of this regression was 0.76. Similar regressions between rainfall and elevation were made for every month. This revealed that the influence of the elevation is twice as much in the winter than in the summer or early autumn: in the winter months (December, January) the vertical gradient is approximately 100 mm per 100 m and in the summer months (July, August, September) around 50 mm per 100 m elevation difference. The coefficient of determination, which indicates the regularity of the topographical influence, is also much higher in the winter (0.8) than in the summer (0.5). Most of the rainfall during the summer months originates from unstable cloud formations, with a limited spatial coverage. Due to this effect there is only a weak relationship with the topography. Rainfall during the winter period is mainly produced by cyclonic centra, occluded fronts or double fronts, wherein stratus-shaped clouds dominate (60%, around 40% in the summer). Then, the areas under precipitation are much more extensive so that the effect of the elevation is more pronounced.

However, the amount of precipitation does not give information about the intensity of the water erosion processes. More important are the number of precipitation events during a year, the intensities and total amount of rainfall of the single precipitation events. This type of analysis comprises detailed time series. In Belgium, only one station provides such detailed information: Ukkel. Because the total yearly rainfall and the distribution of the rainfall during the year observed in that weather station is quite similar as in the study area, the observations of Ukkel are used for a more detailed analysis of the precipitation events.

For Ukkel, precipitation time series of 27 years (1967-1993) were available. Every 10 minutes the precipitation amount was registered. Precipitation events were defined to be separated by a time gap of at least 6 hours. This time gap is artificial and is derived from the RUSLE methodology to calculate the rainfall erosivity (Renard et al., 1996). Table 2.2 gives an overview of the properties of single rainfall events. From these rainfall characteristics it can be seen clearly that the mean intensity of summer precipitation events is 2 to 3 times higher than in the winter. This will have important consequences for the erosivity distribution during the year. However, the number of rainfall events and their mean duration is higher in the winter months. Because of the low evapotranspiration in this period, this results in many areas in a total saturation
of the topsoil for long periods. Therefore, the agricultural parcels are more vulnerable to soil erosion during the winter months. The soil cover is then also only a fraction of the soil cover in the summer months. A more detailed study about the intensity-duration-frequency relations (IDF) for Ukkel can be found in Verhoest et al. (1996).

Table 2.2: Some rainfall characteristics of Ukkel, based on a time series of 27 years (1967–1993).

<table>
<thead>
<tr>
<th>month</th>
<th>mean number of events (—)</th>
<th>mean event duration (h)</th>
<th>mean maximum intensity (mm/h)</th>
<th>mean rainfall per event (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>10.1</td>
<td>6.9</td>
<td>8.5</td>
<td>3.4</td>
</tr>
<tr>
<td>Februari</td>
<td>7.5</td>
<td>7.2</td>
<td>6.8</td>
<td>3.5</td>
</tr>
<tr>
<td>March</td>
<td>9.2</td>
<td>7.0</td>
<td>9.2</td>
<td>4.0</td>
</tr>
<tr>
<td>April</td>
<td>8.4</td>
<td>6.4</td>
<td>9.0</td>
<td>3.0</td>
</tr>
<tr>
<td>May</td>
<td>8.6</td>
<td>5.5</td>
<td>16.0</td>
<td>4.1</td>
</tr>
<tr>
<td>June</td>
<td>8.5</td>
<td>5.5</td>
<td>18.7</td>
<td>4.5</td>
</tr>
<tr>
<td>July</td>
<td>7.7</td>
<td>5.5</td>
<td>20.5</td>
<td>4.6</td>
</tr>
<tr>
<td>August</td>
<td>7.9</td>
<td>4.8</td>
<td>24.5</td>
<td>3.9</td>
</tr>
<tr>
<td>September</td>
<td>7.9</td>
<td>5.7</td>
<td>15.4</td>
<td>3.7</td>
</tr>
<tr>
<td>October</td>
<td>8.5</td>
<td>6.1</td>
<td>9.9</td>
<td>3.5</td>
</tr>
<tr>
<td>November</td>
<td>9.4</td>
<td>8.1</td>
<td>9.3</td>
<td>4.4</td>
</tr>
<tr>
<td>December</td>
<td>9.0</td>
<td>7.5</td>
<td>8.4</td>
<td>4.0</td>
</tr>
</tbody>
</table>

2.4 Geology

Figure 2.3 gives an overview of the depth of the Paleozoic (245 – 570 million years) sediments in Belgium (Maréchal, 1992). These deposits have a prominent slope towards the north—north-east. This had very important consequences for later earth surface processes. The marine transgressions throughout the eons always came from the north-east. This can be seen on a geological map of Belgium, where the Tertiary outcrops have a clear north-west—south-east orientation. The vertical profile given in figure 2.4 explains the banded Tertiary outcrops on the surface. Due to this north-east exposure of the Tertiary deposits, the groundwater flows along the transition zones between the
different geological layers in northern directions. Therefore, there are also more springs and wells along the northern flanks of the hills.

Figure 2.3: Depth in meter of the Paleozoic socle (Maréchal, 1992). The legend categories on this map are: (1) Paleozoic outcrops, (2) Paleozoic level (m) and (3) Paleozoic fracture-lines.

At the beginning of the Eocene (a geological time scale of the Tertiary and Quaternary sediments can be found in table 2.3) the whole of Flanders was marine. During this marine period, the Ypresian (named after the city of Ieper or Ypres in English), a clay layer of 100 to 200 m was deposited. The Ypresian sediments are gray-green clays. At the end of the Ypresian period more and more sand was deposited, indicating a regression of the sea level. From then on there were constant fluctuations in the sea level, resulting in alternating continental and marine phases. A proof for this can be found in the fossil fauna and flora. The Ypresian sediments are rich in shark teeth and mollusks; in the Lutetian sediments there are alternating layers where marine fossils
can be found and layers locally rich in petrified wood (Lutetian deposits of Beernem and Aalter). The Lutetian sediments are coarse sands, indicating a shallow sea level. During the Eocene, there was a last important transgression in the Bartonian age. These sediments are more rich in clay. In the west of Flanders there are no deposits of the Oligocene and the early and middle Miocene, indicating a continental phase of approximately 30 million years. During this phase the Eocene deposits were partially eroded. Bartonian and Lutetian deposits are therefore only found in the north of the province of West-Flanders, and on the local hilltops in the south of this province (in the community of Heuvelland).

The transgression in the Messinian (this corresponds with 'Diestiaan' in the Belgian nomenclature) was the last marine phase which overflowed the total area of Flanders. Because of the preceding continental phase of 30 million years, erosion processes had produced a hilly landscape: a ridge of small hills was formed starting in France, throughout the community of Heuvelland and up to Hasselt (see figure 2.3). This hilly chain formed the coastline, where they formed sand banks in the Messinian sea.
Table 2.3: *International geological time scale of the Tertiary and Quaternary (Geological Society of America Map and Chart Series MC-50).*

<table>
<thead>
<tr>
<th>Period</th>
<th>Epoch</th>
<th>Age</th>
<th>Time (million years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quaternary</td>
<td>Holocene</td>
<td></td>
<td>0.01 – present</td>
</tr>
<tr>
<td></td>
<td>Pleistocene</td>
<td>Calabrian</td>
<td>1.0 – 0.01</td>
</tr>
<tr>
<td>Tertiary (Neogene)</td>
<td>Pliocene (Late)</td>
<td>Piacenzian</td>
<td>3.4 – 1.0</td>
</tr>
<tr>
<td></td>
<td>Pliocene (Early)</td>
<td>Zanclean</td>
<td>5.3 – 3.4</td>
</tr>
<tr>
<td>Miocene (Late)</td>
<td>Messinian</td>
<td></td>
<td>6.5 – 5.3</td>
</tr>
<tr>
<td>Miocene (Middle)</td>
<td>Tortonian</td>
<td></td>
<td>11.2 – 6.5</td>
</tr>
<tr>
<td>Miocene (Early)</td>
<td>Serravallian</td>
<td></td>
<td>15.1 – 11.2</td>
</tr>
<tr>
<td></td>
<td>Langhian</td>
<td></td>
<td>16.6 – 15.1</td>
</tr>
<tr>
<td></td>
<td>Burdigalian</td>
<td></td>
<td>21.8 – 16.6</td>
</tr>
<tr>
<td></td>
<td>Aquitanian</td>
<td></td>
<td>23.7 – 21.8</td>
</tr>
<tr>
<td>Tertiary (Paleogene)</td>
<td>Oligocene (Late)</td>
<td>Chattian</td>
<td>30.0 – 23.7</td>
</tr>
<tr>
<td>Oligocene (Early)</td>
<td>Rupelian</td>
<td></td>
<td>36.6 – 30.0</td>
</tr>
<tr>
<td>Eocene (Late)</td>
<td>Priabonian</td>
<td></td>
<td>40.0 – 36.6</td>
</tr>
<tr>
<td>Eocene (Middle)</td>
<td>Bartonian</td>
<td></td>
<td>43.6 – 40.0</td>
</tr>
<tr>
<td>Eocene (Early)</td>
<td>Lutetian</td>
<td></td>
<td>52.0 – 43.6</td>
</tr>
<tr>
<td></td>
<td>Ypresian</td>
<td></td>
<td>57.8 – 52.0</td>
</tr>
<tr>
<td>Paleocene (Late)</td>
<td>Selandian</td>
<td></td>
<td>63.6 – 57.8</td>
</tr>
<tr>
<td>Paleocene (Early)</td>
<td>Danian</td>
<td></td>
<td>66.4 – 63.6</td>
</tr>
</tbody>
</table>

These sand banks periodically emerged and, exposed to air, the iron of the abundant glauconite clay in the sediments was transformed to limonite. This resulted in a cementation process, where the loose coarse sands and gravels formed a hardened layer. This cementation process continued after the regression of the Messinian sea.

The formation of the present hills is the result of a process called ‘differential’ erosion. This is erosion at different rates. At locations where hardened Messinian layers can be found, the erosion rate was much lower. On top of a few hills within the community of Heuvelland (those above 130 m), these Messinian sand–gravel deposits can still be found. This hardened layer was an excellent shield against fluvial and fluviatile erosion. The elevation difference of 130 m between the present sea level and the occur-
rence of the Messinian layers can be partially explained by the Alpine uplift which still goes on at present (De Moor and Pissart, 1992) and the climatological changes which resulted in a constant fluctuation of the sea level.

The climatological fluctuations during the Quaternary period had an important impact on the present soil formation, and the characteristics of the topsoil. During the ice-ages, Belgium was never covered with ice, but the climate was periglacial. During the youngest glacial epoch, the Weichselian (70000–13000 years ago), the sea level was approximately 70 m lower than at present, resulting in a land-bridge between Belgium and the UK. A massive aeolian transport occurred, due to the dominant northern winds. The coarser the wind-blown particles, the faster they were deposited. Therefore, the Quaternary cover of Flanders has three main belts, almost with a perfect west–east direction: a sandy to loamy sand belt in the north, a loamy belt in the middle and a silty belt in the south. The sandy particles were transported by saltation, the loess particles in suspension. This Weichselian deck features now large differences in depth. In the valley areas and the flat plateaus the mean thickness is several meters (locally up to 30 m). On steep slopes and on a convex topography these layers are very shallow (a few decimeter) or sometimes absent or eroded. In hilly areas these sediments are thick on the gentle slopes with a northern exposure and significantly thinner on the slopes faced towards the south. The transition from the loamy belt towards the silty belt is also visible in the community of Heuvelland (see figure 2.6).
2.5 Pedology

A short overview of the Belgian nomenclature concerning soil classification is necessary. This system classifies soils according to a morphogenetical classification. The soil classes were basically defined by top-soil texture, natural drainage condition and the nature of the profile differentiation. These soil classes were indicated by a three letter symbol (Tavernier and Maréchal, 1962).

The first symbol gives an indication of the soil texture. Figure 2.5 shows the Belgian textural triangle. The textural classes are: Z (zand – sand), S (lemig zand – clayey and silty sand), P (licht zandleem – light sandy silt), L (zandleem – sandy silt), A (leem – silt), E (klei – clay), U (zware klei – heavy clay). These classes are not optimal: the intra-class variability is very high compared with the international accepted USDA textural triangle. This results in large uncertainties if one wants to convert the qualitative data of the Belgian soil map to quantitative variables.

The second symbol gives an indication of the soil drainage. A description of these drainage symbols is given in table 2.4. The third symbol represents an indication of the profile development.

![Textural Triangles](image)

Figure 2.5: The Belgian (left) and USDA (right) textural triangles.

Figure 2.6 gives an overview of the major textural groups (Belgian nomenclature) within the community of Heuvelland. From this figure it can be clearly seen that there are two major soil types: loamy soils (zandleem, licht zandleem) in the north and silty soils (leem) in the south. The most southern part of the community of Heuvelland
Table 2.4: Definition of the drainage classes on the Belgian soil map. Note that in the description a difference is made between the heavy (L, A, E, and U) and the light (Z, S and P) soil textures.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Drainage</th>
<th>Depth of gley or reduction features (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>L,A,E,U</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Gley</td>
</tr>
<tr>
<td>a</td>
<td>very dry</td>
<td>to strong</td>
<td></td>
</tr>
<tr>
<td>b</td>
<td>dry</td>
<td>good</td>
<td></td>
</tr>
<tr>
<td>c</td>
<td>fairly dry</td>
<td>moderate</td>
<td>80–125</td>
</tr>
<tr>
<td>d</td>
<td>fairly wet</td>
<td>insufficient</td>
<td>50–80</td>
</tr>
<tr>
<td>e</td>
<td>wet</td>
<td>fairly bad</td>
<td>30–50</td>
</tr>
<tr>
<td>f</td>
<td>very wet</td>
<td>bad</td>
<td>0–30</td>
</tr>
<tr>
<td>g</td>
<td>extremely wet</td>
<td>very bad</td>
<td></td>
</tr>
<tr>
<td>h</td>
<td>wet</td>
<td>fairly bad</td>
<td>30–50</td>
</tr>
<tr>
<td>i</td>
<td>very wet</td>
<td>bad</td>
<td>0–30</td>
</tr>
</tbody>
</table>

is situated in the alluvial plain of the river Leie. There, the soils have a large clay content with local peaks up to 60 %. Near the highest hilltops, the Messinian sands (see table 2.3) are exposed.

The drainage condition of the soil depends on a large number of factors, such as soil texture, soil structure, the profile development, agricultural management, topography. However, due to artificial tile drainage, most soils have good drainage conditions for agriculture.

Except for the soils in the direct neighborhood of the hills, the soils of the community of Heuvelland have a very high agricultural potential, with almost optimal yields. The locations where Tertiary material is exposed are more demanding: they are more difficult to till, are sometimes very rich in stones and silex pebbles, and consequently yields are variable. The clayey soils in the alluvial plain of the Leie are difficult to till and in some wet years they are hardly accessible. However, it are very good soils for beets and wheat.
Figure 2.6: Soil map of the community of Heuvelland, generalized for the major soil textural groups (Belgian nomenclature).

2.6 Agriculture

Agriculture has always been the major activity within the community of Heuvelland and the present land occupation by agriculture is approximately 83% of the total area. There is a slight depopulation by lack of industrial investments and by the evolutions in modern agriculture. Figure 2.7 shows the demographic evolution during the last two centuries. After WW I the population never reached the pre-war density. Due to the distance from the economic centers in Belgium (Brussel, Gent, Antwerpen and Brugge) the population is still slightly declining.

The agricultural statistics of 1992 (data provided by the National Institute of Statistics, NIS) indicate that there were 422 farms, with a mean area of 18.8 ha. Only 25% of the agricultural land is the property of the farmers, meaning that most exploitations were done on rented parcels. At present, approximately 15% of the professional active people
is employed in the agricultural sector. The mean age of the farmers is relatively young, so that it can be expected that the present situation will continue in the following decades.

After WW II agriculture changed drastically. In 1950 the mean area of the 1958 farms was only 4 ha. The distribution of land area of the farms was as follows: less than 1 ha: 65 %, 1–5 ha: 10 %, 5–10 ha: 10 %, 10–20 ha: 9 % and 20–50 ha: 6 %. In general, it can be concluded that farms with a mean area larger than 5 ha, survived the recent mechanization and rationalization (25 % or 486 enterprises). This evolution also explains the small average parcel area (approximately 1.4 ha) which can still be observed.

Since 1910, the area for cereals declined with more than one third. This is compensated with an increase in pasture, sugar-beets, beets and potatoes. Since 1950 there was an important substitution of beets by maize as fodder for cattle.

Table 2.5 presents a short overview of the recent evolutions in stock-farming. It can be concluded that cattle- and pig-breeding are the most important sectors in stock-farming. It is the specialization in respectively 77% and 44% of the farms. Around 1985, the number of pig-breeding farms was strongly reduced but the number of animals increased. In 1985, there were 229 pigs per farm and in 1992 this increased to 448 animals per farm, showing an increase in the importance of the bio-industry. Sheep breeding is only marginal in the Heuvelland community and is limited to 31 farms (7%). Although the number of animals is very high, only 4% of the farms are specialized in
poultry. As a consequence of the mechanization after WW II, the number of horses (and thus also the cultivation of oat) diminished strongly.

Table 2.5: Stock-farming: recent evolution within the community of Heuvelland (data provided by the National Institute of Statistics, NIS).

<table>
<thead>
<tr>
<th></th>
<th>1950</th>
<th>1985</th>
<th>1992</th>
</tr>
</thead>
<tbody>
<tr>
<td>horses</td>
<td>1024</td>
<td>33</td>
<td></td>
</tr>
<tr>
<td>cattle</td>
<td>9472</td>
<td>20202</td>
<td>21779</td>
</tr>
<tr>
<td>pigs</td>
<td>10465</td>
<td>69303</td>
<td>91331</td>
</tr>
<tr>
<td>sheep</td>
<td>1280</td>
<td>2220</td>
<td>2145</td>
</tr>
<tr>
<td>goats</td>
<td>325</td>
<td>12</td>
<td></td>
</tr>
<tr>
<td>poultry</td>
<td>83970</td>
<td>148745</td>
<td>229873</td>
</tr>
</tbody>
</table>
Chapter 3

Water Erosion Processes and Erosion Models

3.1 Introduction

It is necessary to make a definition of the concepts erosion and soil loss, because these terms are often mixed. There will be referred to erosion to indicate the amount of soil transported over a terrain element. Soil loss is the total amount of soil, lost from an agricultural field towards the drainage system.

Before a soil conservation policy can be developed, one has to know the location of the problem areas and the extent of the problems. If larger areas have to be evaluated, the use of soil erosion models is indispensable. In this chapter an overview and short description is given of the available water erosion models. In order to better understand the structure of an erosion model and how to calibrate and validate it, a description is presented of the fundamental processes of water erosion.

3.2 Forms of water erosion

The properties of the overland flow are very important for the resulting soil loss towards the drainage system. Experimental data show that the soil detachment rate decreases as the runoff depth increases (Torri et al., 1987b). This indicates that the detachment power of the raindrops is partially dispersed by the water layer. It is therefore important to make a distinction between unconcentrated (interrill erosion) and concentrated flow
(rill erosion).

Interrill erosion is also called sheet erosion. There will be referred to sheet erosion to indicate these land areas where the amount of soil transport by overland flow is influenced by raindrop impact. If a bare field is considered, one might expect that the major part of the soil surface is subject to sheet erosion.

Rill erosion is the concentrated flow in small ephemeral channels. When these channels reach dimensions where they cannot be erased by simple tillage operations, the term gully erosion is used. Smaller channels are called rills. Severe gully erosion is rather rare in the study area (community of Heuvelland). In central Belgium, gullies are more frequently occurring, due to the more prominent topography. Rill erosion on agricultural land is mostly induced by the micro-topographical structures due to tillage, which creates small linear furrows. While the position of the rills in a field alter according to the tillage operations, large gullies often occur on the same spot. This is due to the local cumulative drainage area and the local soil properties.

A last form of water erosion is bank erosion. This form of erosion is induced by the shear forces of the water. However, this is in natural environments and in moderate climate types a slow process. However, many banks in the study area suffer stability problems. This is almost always induced by an improper management. Too many farmers till their lands too close to the borders of the drainage ways to have the largest possible effective cultivating area.

### 3.3 Geological and accelerated erosion

Soil erosion in agricultural watersheds can be subdivided in a natural (geological) component and an accelerated (human-induced) component. Geological erosion is generally a very slow process. The soil loss rates are in most areas usually not more than 0.1 to 0.4 ton·ha⁻¹·year⁻¹, which is equivalent to a removal of 1 meter soil loss per 25000 to 100000 years.

Because of the magnitude of the present world population, it becomes difficult to find undisturbed watersheds to measure geological erosion. Only few reports deal with the subject of geological soil loss rates. Rawat and Rawat (1994) reported soil loss rates
resulting from a monitoring campaign of several years for watersheds under natural vegetation in the Indian central Himalayas. They found average soil loss rates of 0.02 to 0.04 mm·year⁻¹ (0.28–0.56 ton·ha⁻¹·year⁻¹) in a watershed with an almost undisturbed mixed oak-pine forest vegetation.

Whitmore et al. (1994) used paleolimnology techniques to study the sediments of lakes in the Yunnan province of China. The objective of these authors was to assess the historical trends concerning soil loss rates. Lake sediments are a source of information about historical changes in lakes, as well as long-term patterns of soil and nutrient erosion from watersheds. Watersheds on the Yunnan Plateau have a long history of human occupation. Neolithic settlements and rice agriculture were widespread by 4000 BP, and terracing was introduced into the region by 1740 BP (Whitmore et al., 1994). Figure 3.1 gives an overview of the sediment dry mass composition, and in table 3.1 soil loss rates are given for two lakes. The appearance of the anthropogenic zone in the sediment of Xingyun Hu and Qilu Hu coincides with the period of acute environmental deterioration that began in China between 500 and 250 years ago because of increased agricultural activity. The accumulation rates of non-carbonate inorganic clastic sediment (carbonate and organic-rich deposits are overlain by iron-rich, red clays and silts) increased dramatically in these lakes following anthropogenic impact. Note that the bulk sediment rates given in table 3.1 are of the same order as the results of Rawat and Rawat (1994).

<table>
<thead>
<tr>
<th>Lake</th>
<th>Bulk sediments</th>
<th>Non-carbonate, clastic sediments</th>
<th>Phosphorus</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(ton·ha⁻¹·year⁻¹)</td>
<td>(ton·ha⁻¹·year⁻¹)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>LD</td>
<td>R</td>
<td>LD</td>
</tr>
<tr>
<td>Qilu Hu</td>
<td>0.20</td>
<td>1.42</td>
<td>0.08</td>
</tr>
<tr>
<td>Xingyun Hu</td>
<td>0.28</td>
<td>2.07</td>
<td>0.16</td>
</tr>
</tbody>
</table>

Geological erosion is usually compensated by the weathering of the parent material. Therefore, most undisturbed areas will remain a good and well structured soil pro-
file. On sloping agricultural land, the soil loss rates are much higher. Even in the province of West-Flanders, with a very gentle sloping topography, mean soil loss rates in agricultural watersheds of 3 to 5 ton·ha\(^{-1}\)·year\(^{-1}\) are common (see Chapter 4), with local peaks of 10 to over 100 ton·ha\(^{-1}\)·year\(^{-1}\), but the latter are rare events. In some locations, a few decades are enough for the removal of 1 meter soil.

When average soil loss rates for the human-induced accelerated erosion process are given in literature, all kinds of figures can be found. It is advisable not to copy blindly these soil loss rates. Pimentel (1993) suggests that soil erosion is a major threat to the sustainability of agriculture, and asserts that, “soil loss rates in Europe range between 10 and 20 ton·ha\(^{-1}\)·year\(^{-1}\)”, but does not give a source. However, Arden-Clarke and Evans (1993) state that water erosion rates in lowland Britain vary from 1 to 20 ton·ha\(^{-1}\)·year\(^{-1}\), but the latter are rare and localized events. They suggest that, “most rates of erosion are less than 1 to 2 m\(^{3}\)·ha\(^{-1}\)·year\(^{-1}\)”. The monitoring of 86 fields in central Belgium (Govers, 1991) revealed an average soil loss rate of 3.6 ton·ha\(^{-1}\)·year\(^{-1}\), a value close to the one found in the Kemmelbeek watershed.

Figure 3.1: Organic matter, carbonate, and non-carbonate, inorganic-clastic content of sediments, expressed as the percent of dry mass, in long cores of two Chinese lakes (Whitmore et al., 1994).
during this work (Chapter 4). These figures suggest that Pimentel’s average soil loss rate for Europe is far too high (Boardman, 1998) for countries with a moderate climate type (like Belgium, the UK and the Netherlands). Nevertheless, the observed soil loss rates within agricultural watersheds are magnitudes higher than the average rates of geological erosion.

Most sediments that cause off-site silting problems in watersheds within the Belgian loess belt, are produced by sheet and small rill erosion on agricultural fields. Because this type of erosion is more difficult to spot than severe gully erosion, and the reduction in crop productivity caused by these forms of erosion is almost negligible, soil erosion in Belgium is generally considered as only a marginal agricultural problem. However, the sometimes severe flooding effects of long-lasting low intensity winter rains illustrate frequently the opposite. Given the observed soil loss rates and the frequently re-occurring off-site silting problems, it can be stated that on-site sediment control remains one of the most important factors to ensure a good drainage capacity of ditches, brooks and rivers. A more thoughtful agriculture and management of our natural resources is a necessity to ensure a good soil and water quality for future generations.

3.4 Water erosion processes

The soil erosion process comprises two phases: (1) detachment and transport by raindrop impact and (2) runoff (Ellison, 1947). When water is supplied to a soil surface, either by rainfall or irrigation, some quantity might penetrate through the soil profile, while the remaining part can produce runoff. The part of the runoff water that cannot be retained by the surface micro-topography flows freely over the soil surface where it can detach and transport particles (soil particles, organic material, small stones). Along the flowpath of the runoff there is a constant translocation of particles, in other words, there is a constant erosion (detachment and/or transport of already detached particles) and deposition of transported particles. When the runoff water reaches the drainage system it is loaded with sediment, producing a net soil loss.

A good modelling approach of the infiltration processes and the runoff routing are basic to estimate the erosion losses and to delineate the erosion risk areas. These processes describe the amount and the properties of the overland flow respectively. The basic
physical and chemical soil characteristics concerning soil stability will determine the amount of infiltration and eventually also the amount of runoff and sediment loss. This indicates the importance of a good hydrological model when erosion processes are studied on a regional scale.

### 3.4.1 Overland flow processes

Overland flow is the flow of water over the land before becoming channelized. Four major types of overland flow can be distinguished: (1) Hortonian overland flow occurs when rainfall intensity exceeds the infiltration rate, (2) delayed Hortonian overland flow occurs when rainfall intensity exceeds the infiltration rate only after some delay, during which changes in the soil occur under the influence of wetting or raindrop impact, (3) topsoil saturation overland flow takes place on soils where a relative permeable topsoil layer overlies less permeable material, (4) saturation overland flow is produced when the storage capacity of the soil is completely filled, so that all subsequent additions of water at the surface, irrespective of their rate of application, are forced to flow over the surface.

The detailed scientific study of overland flow started around the 1930’s with the work of Horton (1933). He described overland flow as follows: “Neglecting interception by vegetation, surface runoff is that part of the rainfall which is not absorbed by the soil by infiltration. If the soil has an infiltration capacity \( f \), then when the rain intensity \( i \) is less than \( f \), the rain is all absorbed and there is no surface runoff. It may be said as a first approximation that if \( i \) is greater than \( f \), surface runoff will occur at the rate \( (i - f) \).” Horton termed this difference \( i - f \) “rainfall excess”. Horton considered surface runoff to take the form of a broad sheet flow, with a measurable flow depth uniform over the terrain. As the flow accumulates going downslope, the depth increases until the runoff water discharges into a stream channel.

However, field observations demonstrated that Hortonian surface flow is rather rare in humid regions. In these regions, the infiltration capacity of the soil exceeds the observed rainfall intensities for all except the most extreme events (Troch et al., 1994). Hursh and Brater (1944) doubted the concept of overland flow as defined by Horton (1933), however, while surface storm runoff was not observed during rainfall, the characteristic
flood hydrographs had still the shape as produced by heavy rain. Betson (1964) put forward the “partial area concept”, denoting that infiltration excess runoff occurs from a relative small part of the catchment area. By analyzing a series of rainfall events, Ragan (1968) showed that only a small portion of the study catchment ever contributed to the storm hydrograph.

From the work of Hewlett and Hibbert (1967) the concept of subsurface storm flow was created. However, subsurface flow velocities are normally so low that subsurface flow alone cannot contribute a significant amount of storm precipitation directly to streamflow. A second mechanism, derived from field observations, is the “saturation overland flow”. Saturation overland flow is produced when subsurface flow saturates the soil near the bottom of a slope and overland flow then occurs as rain falls onto saturated soil. Saturation overland flow differs from Hortonian overland flow in that in Hortonian overland flow the soil is saturated from above by infiltration, while in saturation overland flow it is saturated from below by subsurface flow. Dunne et al. (1975) suggested that the partial contributing areas were likely to be wetlands with locations controlled by the topographic and hydrogeological configuration of the basin. These contributing areas are likely to be located adjacent to stream channels, in swamp areas and on shallow soils and expand during a rainfall event and contract thereafter. This dynamical character earned them the term “variable source areas”. This overland flow process was supported by a series of computer simulation experiments, using a fully integrated surface–subsurface flow model (Freeze, 1972, 1974). The simulations carried out with rainfall events on hypothetical hillslope areas showed that significant contributions to stormflow from the subsurface runoff could only occur under the restrictive conditions of a convex hillslope feeding a deeply incised channel, and then only if saturated soil conductivities are very large. On concave slopes, with lower permeabilities, and on all convex slopes, hydrographs are dominated by direct runoff through very short overland flow paths from near-channel areas.

Flow routing is the procedure to determine the time and magnitude of flow (the hydrograph) at a point on a watercourse from known or assumed hydrographs at one or more points upstream. In general, flow routing may be considered as an analysis to trace the flow (overland flow and/or channel flow) through a hydrological system. The difference between lumped and distributed system routing is that in a lumped system
model the flow is calculated as a function of time alone at a particular location, while in a distributed system routing the flow is calculated as a function of space and time throughout the system (Chow et al., 1988). The lumped models of routing are sometimes called hydrologic routing, and the distributed models are sometimes referred to as hydraulic routing. An overview of the basic concepts and equations of overland flow routing and channel routing for both methodologies is given below.

3.4.1.1 Lump ed flow routing

In a hydrological system, the input $I(t)$, output $Q(t)$, and the storage $S(t)$, are related by the following continuity equation (Chow et al., 1988):

\[
\frac{dS}{dt} = I(t) - Q(t).
\]

If only the inflow hydrograph $I(t)$ is known, equation (3.1) cannot be solved directly to obtain the outflow hydrograph, $Q(t)$, because both $Q$ and $S$ are unknown. A second relationship, known as a storage function, is necessary to relate $S$, $I$ and $Q$.

3.4.1.2 Distributed flow routing

The flow of water over the land, through the soil and in stream channels of a watershed is a distributed process because the flow rate, flow velocity and flow depth vary in space throughout the watershed. Estimates of flow rate and/or water level at important locations in the channel can be obtained using a distributed flow routing model. This type of model is based on the partial differential equations (the Saint-Venant equations for one-dimensional flow) that allow the flow rate and water level to be computed as function of space and time, rather than time alone, as in the lumped models. The flow processes in natural hydrological systems vary in all three space dimensions. However, for many practical situations, the spatial variation of velocity across the channel and with respect to the depth can be ignored, so that the flow process can be approximated by a one-dimensional model. The Saint-Venant equations (de Saint-Venant, 1871) describe a one-dimensional unsteady open channel flow. Because the Saint-Venant equations are the fundamentals of almost all physical flow routing models (also for the
STM-2D/3D model, see chapter 5), the mathematical background of these equations is given here.

![Diagram of water erosion processes and erosion models](image)

**Figure 3.2: Elementary channel segment.**

Consider an elementary channel segment (figure 3.2), with a length $dx$ and a small time interval $dt$. Then, based on the principle of mass conservation, one can write:

$$\frac{\partial S}{\partial t} \cdot dt = Q \cdot dt - \left( Q + \frac{\partial Q}{\partial x} \cdot dx \right) \cdot dt. \tag{3.2}$$

In this equation is $S$ the storage in the control volume, $Q(x,t)$ the segment inflow and $Q + ((\partial Q/\partial x) \cdot dx)$ the segment outflow. Consider $A$ the cross-sectional area, then is $S = A \cdot dx$ and consequently:

$$\frac{\partial S}{\partial t} = \frac{\partial A}{\partial t} \cdot dx \tag{3.3}$$

thus:

$$\frac{\partial A}{\partial t} \cdot dx \cdot dt = -\frac{\partial Q}{\partial x} \cdot dx \cdot dt \tag{3.4}$$

the continuity equation then becomes:

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = 0. \tag{3.5}$$
The second basic equation is based on the energy balance. Applying the Bernoulli equation on the elementary channel section gives:

\[
\Delta f^2 \left( z + \frac{p}{\rho \cdot g} + \alpha \cdot \frac{v^2}{2 \cdot g} + \frac{1}{g} \int_{x}^{x+dx} \frac{\partial v}{\partial t} \cdot dx + F \right) = 0 \tag{3.6}
\]

wherein, \(z\) is the elevation of the water level (m), \(p\) the pressure (N·m\(^{-2}\)), \(\rho\) is the fluid density (kg·m\(^{-3}\)), \(g\) the gravitational constant (m·s\(^{-2}\)), \(\alpha\) the Coriolis coefficient \((-\)), \(v\) the mean velocity of the flow (m·s\(^{-1}\)), and \(F\) the hydraulic head (m). Consider a constant hydrostistical pressure in every cross-section (\(z + p/\rho g\) is constant) and for a point at the surface is \(p = p_{atmospherical} = 0\). Consider also a turbulent flow, so that \(\alpha \approx 1\). Equation (3.6) can then be rewritten as:

\[
\Delta f^2 \left( z + \frac{v^2}{2 \cdot g} + \frac{1}{g} \int_{x}^{x+dx} \frac{\partial v}{\partial t} \cdot dx + F \right) = 0. \tag{3.7}
\]

By dividing with \(x\), using the \(\lim_{dx \to 0}\) operator and by considering a constant \(t\):

\[
\frac{\partial z}{\partial x} + \frac{\partial}{\partial x} \left( \frac{v^2}{2 \cdot g} \right) + \frac{1}{g} \cdot \frac{\partial}{\partial x} \left( \int_{x}^{x+dx} \frac{\partial v}{\partial t} \cdot dx \right) + \frac{\partial F}{\partial x} = 0 \tag{3.8}
\]

or:

\[
\frac{\partial z}{\partial x} + \frac{\partial}{\partial x} \left( \frac{v^2}{2 \cdot g} \right) + \frac{1}{g} \cdot \frac{\partial v}{\partial t} + S_f = 0 \tag{3.9}
\]

with \(S_f = \partial F/\partial x\) the head loss per unit length. Introducing the water level \(h\) in replacement of the elevation \(z\), with \(z = y + h\), results in:

\[
\frac{\partial z}{\partial x} = \frac{\partial y}{\partial x} + \frac{\partial h}{\partial x} \tag{3.10}
\]

and with:

\[
\frac{\partial y}{\partial x} = -sin(\delta) \approx -S_0 \tag{3.11}
\]

wherein \(S_0\) is the local bed slope, with \(\delta\) having small values. Equation (3.9) can then be written as:
\[-S_0 + \frac{\partial h}{\partial x} + \frac{\partial}{\partial x} \left( \frac{v^2}{2 \cdot g} \right) + \frac{1}{g} \cdot \frac{\partial v}{\partial t} + S_f = 0. \tag{3.12}\]

The latter equation can also be written in terms of the dependent variables \(Q\) (discharge) and \(h\) (water level). Taking into account that \(Q = v \cdot A\), and \(v\) perpendicular on \(A\) (thus, for small bed slopes \(S_0\)), equation (3.12) becomes:

\[
\frac{\partial Q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{Q^2}{A} \right) = A \cdot \left( \frac{\partial v}{\partial t} + \frac{1}{2} \cdot \frac{\partial v^2}{\partial x} \right) \tag{3.13}\]

\[
S_f = S_0 - \frac{\partial h}{\partial x} - \frac{1}{g \cdot A} \cdot \frac{\partial Q}{\partial t} - \frac{1}{g \cdot A} \cdot \frac{\partial Q^2}{\partial x} A. \tag{3.14}\]

The momentum equation can then be written as:

\[
\frac{1}{A} \cdot \frac{\partial Q}{\partial t} + \frac{1}{A} \cdot \frac{\partial}{\partial x} \left( \frac{Q^2}{A} \right) + g \cdot \frac{\partial h}{\partial x} - g \cdot \frac{(S_0 - S_f)}{\text{Dynamic wave}} = 0. \tag{3.15}\]

Equation (3.15) forms together with equation (3.15) the Saint-Venant equations. In equation (3.15) the terms are respectively: the local acceleration term, the convective acceleration term, the pressure force term, the gravity force term and the friction force term. The latter equation also exists in various simplified forms: the diffusion wave and the kinematic wave (for more explanation see section 5.3.3).

### 3.4.2 Infiltration processes

Infiltration is the process of water penetrating from the ground surface into the soil. Many factors influence the infiltration rate, including the condition of the soil surface and its vegetation cover, the properties of the soil, such as its porosity and hydraulic conductivity, and the initial moisture content of the soil. Soil strata with different physical properties may overlay each other, forming horizons. Also, soil exhibits a large spatial variability even within relatively small areas such as a field. As a result of these spatial and time variations in soil properties, infiltration is a very complex process. Consequently, it can be described only approximately with mathematical
equations (Chow et al., 1988). These mathematical equations can be subdivided into two main groups: the Green-Ampt (Green and Ampt, 1911) and the Richards equations (Richards, 1931). Remark that these references might look ‘old’. However, it is only now, with the processing capacity of modern computers, that the power of these equations can be fully exploited inside distributed models.

### 3.4.2.1 Green-Ampt based models

The Green-Ampt based infiltration models start from an approximated physical theory that has an exact analytical solution. The wetting front in this approach is supposed to be a sharp boundary dividing soil with an initial moisture content below from saturated soil above the wetting front. Furthermore, the soil profile is supposed to be homogeneous. The original Green-Ampt equation (Green and Ampt, 1911) describes only the situation where the soil column is initially ponded. Mein and Larson (1973) developed a method for modelling infiltration during a steady rain, where the soil surface is not ponded at the initial stage. Mein and Farrell (1974) described how to determine the wetting front pressure head for use in the Green-Ampt equations. Finally, Chu (1978), extended the methodology of Mein and Larson (1973) for describing the infiltration during an unsteady rain event, with more than one stage of ponding and non-ponding. This model was also preferred for the description of the infiltration processes inside the physical erosion model because every parameter has a physical meaning, and the model can be solved analytical (i.e. fast). A complete outline of this methodology can be found in chapter 5. The Green-Ampt infiltration model can also be extended for non-homogeneous soil profiles (Chow et al., 1988).

### 3.4.2.2 Richards based models

The Richards based models are a second group of infiltration models. Like the Green-Ampt based models, they are fully physical, but with less approximations. Because of the importance of this methodology in soil physics and hydrology, an overview of the governing equations will be given. Darcy’s law for flow in a porous medium can be written as:
\[
\frac{Q}{A} = q = K \cdot S_f
\]  
(3.16)

where \(Q\) is the discharge (m\(^3\)-s\(^{-1}\)), \(A\) is the cross-sectional area (m\(^2\)), \(q\) is the volumetric flux (m\(^3\)-s\(^{-1}\)), \(K\) is the hydraulic conductivity (m-s\(^{-1}\)) of the medium and \(S_f\) is the head loss per unit length of medium (m-m\(^{-1}\)). Consider flow in the vertical direction and denote the total head of the flow by \(h\), then \(S_f = -\partial h/\partial z\) where the negative sign indicates that the total head is decreasing in the direction of the flow because of friction. Darcy’s law is then:

\[
q = -K \cdot \frac{\partial h}{\partial z}
\]  
(3.17)

This law can be applied in the case of stationary flow processes (constant flux in time and space dimensions) and saturated soil profiles. In the case of an unsaturated soil, the hydraulic conductivity is a function of the moisture content, \(\theta\):

\[
q = -K(\theta) \cdot \frac{\partial h}{\partial z}
\]  
(3.18)

In unsaturated soil profiles, stationary flow is rare. In order to describe a non-stationary flow, it is necessary to introduce a continuity equation, based on the law of conservation of mass. Consider a soil volume \(V = \Delta x \cdot \Delta y \cdot \Delta z\), and the flow direction along the \(x\)-axis. The continuity equation can then be written as:

\[
q \cdot \Delta y \cdot \Delta z - \left( q + \frac{\partial q}{\partial x} \cdot \Delta x \right) \Delta y \cdot \Delta z = \frac{\partial \theta}{\partial t} \cdot \Delta x \cdot \Delta y \cdot \Delta z.
\]  
(3.19)

Thus, the net inflow into the soil volume \(V\) equals the moisture change of the soil volume. The continuity equation is then:

\[
\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial x}
\]  
(3.20)

Or, for a flow in a three dimensional space:

\[
\frac{\partial \theta}{\partial t} = -\left( \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} \right) = -\nabla \tilde{q}.
\]  
(3.21)
Chapter 3: Water Erosion Processes and Erosion Models

The general Richards equation to predict the moisture contents during unsaturated flow can now be found by combining equations (3.18) and (3.21):

\[
\frac{\partial \theta}{\partial t} = \nabla (K(\theta) \nabla h) = - \left( \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} \right)
\]  

with:

\[
q_x = -K_x(\theta) \cdot (\partial h/\partial x),
\]
\[
q_y = -K_y(\theta) \cdot (\partial h/\partial y),
\]
\[
q_z = -K_z(\theta) \cdot (\partial h/\partial z).
\]

Suppose a homogeneous and isotrope soil, then:

\[
K_x(\theta) = K_y(\theta) = K_z(\theta) = K(\theta).
\]

The equation (3.22) can now be written as:

\[
\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( K(\theta) \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( K(\theta) \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left( K(\theta) \frac{\partial h}{\partial z} \right).
\]  

(3.23)

The head \( h \) of the water is measured in dimensions of height, but can also be thought of as the energy per unit weight of the fluid. In an unsaturated porous medium, the part of the total energy possessed by the fluid due to soil suction forces is referred to as the pressure head \( \psi \). The total head \( h \) is the sum of the pressure and gravity heads \( (h = \psi + z) \). Equation (3.23) can then be written as (in the horizontal direction there is no gravity head):

\[
\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( K(\theta) \frac{\partial \psi}{\partial x} \right) + \frac{\partial}{\partial y} \left( K(\theta) \frac{\partial \psi}{\partial y} \right) + \frac{\partial}{\partial z} \left( K(\theta) \frac{\partial \psi}{\partial z} \right) + \frac{\partial K(\theta)}{\partial z}
\]  

(3.24)

In equation (3.24) there are two dependent variables: the pressure head \( (\psi) \) and the moisture content \( (\theta) \). The water retention characteristic gives the relation between both variables. Let us define the differential watercapacity, \( C(\theta) = \partial \theta/\partial \psi \), with \( 1/C(\theta) \) the slope of the water retention characteristic at moisture level \( \theta \). Equation (3.24) can then be written as:

---

33
\[ \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( K(\theta) \frac{\partial \psi}{\partial \theta} \frac{\partial \theta}{\partial x} \right) + \frac{\partial}{\partial y} \left( K(\theta) \frac{\partial \psi}{\partial \theta} \frac{\partial \theta}{\partial y} \right) + \frac{\partial}{\partial z} \left( K(\theta) \frac{\partial \psi}{\partial \theta} \frac{\partial \theta}{\partial z} \right) + \frac{\partial K(\theta)}{\partial z} \] (3.25)

and supposing a unique water retention characteristic (neglecting the hysteresis effects):

\[ \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( \frac{K(\theta)}{C(\theta)} \frac{\partial \theta}{\partial x} \right) + \frac{\partial}{\partial y} \left( \frac{K(\theta)}{C(\theta)} \frac{\partial \theta}{\partial y} \right) + \frac{\partial}{\partial z} \left( \frac{K(\theta)}{C(\theta)} \frac{\partial \theta}{\partial z} \right) + \frac{\partial K(\theta)}{\partial z}. \] (3.26)

In these equations, \( \frac{K(\theta)}{C(\theta)} \), is called the soil water diffusivity, \( D(\theta) \):

\[ \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( D(\theta) \frac{\partial \theta}{\partial x} \right) + \frac{\partial}{\partial y} \left( D(\theta) \frac{\partial \theta}{\partial y} \right) + \frac{\partial}{\partial z} \left( D(\theta) \frac{\partial \theta}{\partial z} \right) + \frac{\partial K(\theta)}{\partial z} \] (3.27)

or:

\[ \frac{\partial \theta}{\partial t} = \nabla (D(\theta) \nabla \theta) + \frac{\partial K(\theta)}{\partial z}. \] (3.28)

Because in many applications it is not necessary to know the moisture flux in three dimensions, some authors propose simplified solutions of Richards diffusion equation. The most important simplified models are those of Horton (1933, 1939) and Philip (1957).

**Horton’s equation**

Horton observed that infiltration begins at some rate \( f_0 \) and exponentially decreased until it reaches a constant rate \( f_c \):

\[ f(t) = f_c + (f_0 - f_c) \cdot e^{-kt}. \] (3.29)

In this formula, \( k \) is a decay constant (s\(^{-1}\)), and is dependent of the soil properties. Raudkivi (1979) has shown that Horton’s equation can be derived from the Richards equation by assuming that \( K \) and \( D \) are constants, independent of the soil moisture content. Under these conditions, equation (3.28) can be reduced to:

\[ \frac{\partial \theta}{\partial t} = D \cdot \frac{\partial^2 \theta}{\partial z^2}. \] (3.30)
This equation may be solved to yield the moisture content \( \theta \) as a function of time and depth.

**Philip’s equation**

Philip (1957) solved Richard’s equation under less restrictive conditions than Horton, by assuming that \( K \) and \( D \) can vary with the moisture content \( \theta \). Philip employed the Boltzmann transformation, \( B(\theta) = z \cdot t^{-1/2} \), to convert the Richards nonlinear partial differential equation for vertical flow into an ordinary differential equation in \( B \), and solved this equation to yield an infinite series for cumulative infiltration \( F(t) \), which is approximated by:

\[
F(t) = S \cdot t^{1/2} + K \cdot t \tag{3.31}
\]

where \( S \) is a parameter called sorptivity, which is a function of the soil pressure head, and \( K \) is the hydraulic conductivity. By differentiating, the infiltration equation can be written as:

\[
f(t) = \frac{1}{2} \cdot S \cdot t^{-1/2} + K. \tag{3.32}
\]

The two terms in Philip’s equation represent the effects of the soil pressure head and the gravity head, respectively. For a horizontal column of soil, the soil pressure head is the only force drawing water into the column, and Philip’s equation reduces to \( F(t) = S \cdot t^{1/2} \).

### 3.4.2.3 Infiltration during unsteady rain events

Most rain events, if not all, have an unsteady character: there are multiple periods of infiltration during surface ponding and infiltration without surface ponding. Under a ponded surface the infiltration process is independent of the effect of time distribution of rainfall. The rate of infiltration reaches its maximum capacity and is referred to as the infiltration capacity. Without surface ponding, all the rainfall infiltrates into the soil. The infiltration models described in the previous sections describe infiltration during ponding. If the time that separates the two stages can be determined, the difficulty
involved in modelling infiltration during an unsteady rainfall event is reduced, since
the infiltration for different stages can be treated separately.

In section 5.3.1 the methodology of Chu (1978) is described to determine the different
infiltration stages. This methodology is integrated in the Green-Ampt infiltration
equations, and is used in the development of the physical based erosion model. The
WEPP (Water Erosion Prediction Project) model also uses this infiltration model
(Flanagan and Nearing, 1995).

Another, but similar methodology is the ‘time compression approximation’ (Ibrahim
and Brutsaert, 1968), which can be used in combination with any infiltration model.
However, this method implies the knowledge of the infiltration characteristic of the
soil under zero surface head. It should be noted that although this soil infiltration
characteristic is easy to determine for a given soil in situ or in the laboratory, the form
of the curve depends on the initial moisture content of the soil (Troch, 1993). Ibrahim
and Brutsaert (1968) and Reeves and Miller (1975) described methods to incorporate
also soil hysteresis in the infiltration simulations.

Figure 3.3 shows the general concept of the time compression approximation. If the
soil infiltration characteristic is superimposed with the first part of cumulative rainfall
curve, then the portion which will infiltrate can be estimated graphically: at time \( t_1 \)
a portion \( I_1 \) will infiltrate. The second part of the cumulative rainfall curve should be
superimposed at \( I_1 \). The third part of the cumulative rainfall curve is superimposed at
\( I_2 \). Finally, the fourth part is superimposed at \( I_3 \). During the second and fourth time
interval, no runoff is generated.

The soil specific infiltration characteristic is not commonly occurring in Belgian soil
data bases. Therefore, the ‘time compression approximation’ methodology was not
selected for the development of the physical erosion model (chapter 5). A methodology
to estimate the infiltration characteristic based on soil texture, organic material and
the Green-Ampt concept is given in section 5.2.2.

3.4.3 Transport of soil particles, detachment and deposition

It has been widely supposed that the soil erosion processes occurring during overland
flow are very similar to those occurring during streambed erosion (Hainsline and Rose,
1992). This has led to the use of sediment transport equations derived for deep flow conditions to describe the movement of sediment in the relatively shallow flows characteristics of soil erosion on a field scale. The best known research is that of Shields (1936). He made a fundamental analysis of the processes involved and the forces at
work to determine the critical conditions for initiating particle movement over relatively gentle slopes in rivers in terms of the dimensionless shear stress (ζ) of the flow and the particle roughness Reynolds number (Reₚ), defined respectively by:

\[
\zeta = \frac{\rho_w \cdot u_s^2}{g \cdot (\rho_s - \rho_w) \cdot D}
\]  \hspace{1cm} (3.33)

\[
Re_p = \frac{u_s \cdot D}{\nu}
\]  \hspace{1cm} (3.34)

where ζ is known as the Shields number, ρₜ is the density of water, uₚ is the shear velocity of the flow, g is the gravitational acceleration, ρₛ is the density of the sediment and D is the diameter of the particles. The shear velocity is expressed as:

\[
u_s = \sqrt{\frac{\tau_0}{\rho_w}}
\]  \hspace{1cm} (3.35)

\[
\tau_0 = \rho_w \cdot g \cdot h \cdot S_0
\]  \hspace{1cm} (3.36)

where τ₀ is the shear force, h is the water level of the overland flow and S₀ is the bed slope. When the value of Reₚ is greater than 40 (turbulent flow), the critical value of ζ, ζₑ, for particle movement assumes a constant value of 0.05.

Another important research work about the sediment transport in rivers is that of Schoklitsch (1950). This work and research reports on the initiation of particle movement by overland flow (Govers, 1987; Guy and Dickinson, 1990; Torri et al., 1987) indicate that the Shields number consistently overpredicts the hydraulic requirements for particle movement. This implies that the initiation of movement is not solely a fluid shear stress phenomenon — depending on particle density and particle diameter — but is enhanced by other factors, like rainfall impact, slope steepness, soil cohesion, changes in density of the fluid as the sediment concentration increases (Morgan, 1996). An overview of the influence of soil properties on the detachment and transport processes is given in section 3.4.4.

Only initial removal is dependent upon the cohesive resistance of the original soil. Foster and Meyer (1975) introduced two terms, detachment limiting and transport limiting, to describe the sediment flux when either the resistance to the original soil to release
sediment, or the ability of overland flow to move sediment are respectively considered limiting. To describe the transport limiting case, streambed sediment transport equations are used. For the detachment limiting case, the concept of ‘soil erodibility’ was created. The most important erodibility equation is that of the (R)USLE model (Wischmeier and Mannering, 1969).

### 3.4.4 Soil characteristics and soil stability

The physical and chemical soil characteristics determine the soil structure and therefore the infiltration rate. They will also determine the response of the soil surface when it is subjected to rainfall. The aggregate stability can be defined as the soil structural response to rainfall. It is closely related to primary soil characteristics, but there is no single relationship between aggregate stability and primary soil characteristics. Many attempts were made to find these relationships, but none were suitable for all soils (Le Bissonnais, 1996). As an example, Al-Durrah and Bradford (1982) tried to correlate splash weights to the soil properties (texture, organic matter, surface area, pH, soluble salts, CEC, exchangeable cations, bulk density and matric potential). A maximum coefficient of determination of 0.61 was found from many combinations of independent variables in a multiple regression model. These regression models were most influenced by (in order of importance) the percent clay, organic matter, exchangeable sodium and the amount of soluble salts. This indicates that besides texture (percent clay) there is no dominant parameter affecting soil erodibility. The interaction — concerning soil erodibility — between the different physical and chemical soil characteristics causes the lack of a direct relationship between the measured soil physical and chemical properties and the actual forces that determine soil detachability and erodibility. Numerous studies have reported about this topic, with sometimes contradictory results. A review will be given here concerning the effects of soil texture, clay mineralogy, organic matter, cations, Fe and Al oxides, and CaCO₃. From this short review, it will be clear that it is not possible to construct a general and universal empirical relationship between soil erodibility and the basic soil properties. Due to the large amount of soil parameters which determine soil erodibility and their mutual interaction, one can state that the reported relationships in literature are only valid for the region wherein they were derived and tested. This is also the case for the nomograph Wischmeier and Smith
(1978) of the so called *Universal* Soil Loss Equation.

### 3.4.4.1 Soil texture

It is generally considered that as the silt (2 – 50 μm) or the silt and very fine sand (50 – 100 μm) fraction increases and the clay content decreases, soil erodibility increases. This is because of (a) the aggregation and bonding effect of clay, (b) the transportability of fine and nonaggregated particles, and (c) the detachability of sand and silt. This knowledge is translated in the erodibility nomograph of Wischmeier and Smith (1978). This nomograph is empirical and was derived from numerous erosion plot measurements. Gollany et al. (1991) found that aggregate stability increases with clay content, and that the effect was more pronounced at a higher water content. The same results were found by Le Bissonnais (1988), for a range of silty soils. However, these results are not consistent throughout all research results. Le Bissonnais and Singer (1993), as well as Pierson and Mulla (1990) did not find a significant correlation between the clay content and the aggregate stability.

### 3.4.4.2 Clay mineralogy

Levy et al. (1993) state that micro-aggregate stability depends strongly on the clay mineralogy. When analyzing the dry-soil aggregate stability, Skidmore and Layton (1992) found that the clay content was a good predictor of the mean aggregate stability. However, the effect is difficult to assess because soils most often contain a mixture of clay minerals (Le Bissonnais, 1996). Using pure clay minerals, Emerson (1964) found that swelling clays like montmorillonite were less subject to slaking than kaolinite or illite (the latter is the most important clay mineral in the clay textural fraction of Belgian soils) because the pressure which is developed by entrapped air is released by swelling. Young and Mutchler (1977) found that the montmorillonite content was highly correlated with aggregate stability and low erodibility. However, Stern et al. (1991) found that smectite soils are dispersive and susceptible to seal formation and soil erosion. In general, one can conclude that the effect of the clay type depends on the interaction with other soil physical and chemical properties: clay with a low cation exchange capacity (CEC) — like kaolinite — induces less strong aggregates
than smectite clays, because of its smaller surface area. However, kaolinite is often associated with iron oxides, which may induce very strong aggregates.

3.4.4.3 Organic matter

Organic matter is one of the most important and well-known aggregate stabilizing agents in soils. It was considered by Wischmeier and Mannering (1969) as the second most influential property affecting soil erodibility after soil texture. However, this finding is not always consistent in the literature (Le Bissonnais, 1996). Rogers (1939), for example, found no significant relationship between the total organic matter and the degree of aggregation or observed erodibility. Recently, Roberson et al. (1991) showed that the aggregate stability to slaking increases as the heavy fraction of carbohydrates (polysaccharides) increases, although the total organic matter content did not change. Therefore, the total organic matter which is usually measured, is probably not the best measurement for predicting the aggregate stability or soil erodibility (Le Bissonnais, 1996). However, in most studies it was found that the organic matter reduces soil erodibility. It is a bonding agent between mineral soil particles. The relation is probably nonlinear and depends on the interactions between organic matter and other soil properties. For very high organic matter contents, hydrophobicity may be a factor increasing runoff and consequently erosion. Furthermore, the different organic matter fractions may have different effects: the litter has a clear physical effect (mulching results in a reduced detachment), but also increases biological activity when it becomes degraded. The micro-organisms and their by-products increase the aggregate stability by physicochemical reactions.

3.4.4.4 Cations

The nature and the amount of exchangeable cations influence soil erodibility through their effect on the clay dispersion/flocculation processes. These effects are also closely related to clay mineralogy. Several authors have shown the effect of increasing exchangeable sodium percentage (ESP) on the decrease of the infiltration rate and the subsequent erosion. Dispersion results from reduction of the attractive forces between colloidal particles while wetting. However, some soils are already affected at very low
ESP levels, while other soil types need higher ESP levels to initiate the dispersion processes, and some are not affected at all. Levy and Van Der Watt (1988) found that dispersion of a kaolinitic soil was not significantly affected in the ESP range of 1% to 9%, while two other soils (1) with mixed kaolinite, illite and montmorillonite and (2) with illite and interstratified minerals, were significantly affected at an ESP level of 4.3%. Stability or dispersion depends on the cation size and valence. Therefore, the cations can be classified in the following order, according to their stability inducing properties: $Ca^{2+} > Mg^{2+} > K^+ > Na^+$. Kazman et al. (1983) showed that by the increase of the electrolyte concentration, the dispersive effect of sodium and the chemical dispersion may be prevented by spreading phosphogypsum or another readily available electrolyte source on the soil surface. The effect of ESP could not be observed in tropical soils with high levels of iron oxides, due to the cementing action of iron and the nonswelling nature of kaolinite.

### 3.4.4.5 Fe and Al oxides

As already mentioned above, sesquioxides have generally a good positive correlation with the stability of the soil aggregates. Besides the empirical results, there are also different theories explaining the mechanisms of sesquioxides effect (Le Bissonnais, 1996): (1) Iron and aluminium in solution act as flocculants in lowering the zeta-potential of clays, preventing dispersion in about the same way that calcium does in other soils. (2) Sesquioxides play a role in clay particle–organic polymer interactions. (3) Sesquioxides can precipitate as gel on clay surfaces, hereby altering the properties of the clay particles.

Iron and aluminum can behave differently under different conditions such as pH, clay minerals, climate, organic matter and soil solution composition. There is a debate whether iron or aluminum is more important in aggregating soil particles (Le Bissonnais, 1996). Aluminum seems to be more efficient because of its higher solubility over a wide range of pH, but it is usually found in smaller amounts than iron in soil. A more important concern is the form of the soil sesquioxides. Two basic forms exist: (1) iron present in primary minerals and (2) free iron, which includes crystalline and amorphous iron oxides and organically associated iron. Total free iron correlates good
with the soil erodibility (Le Bissonais and Singer, 1993).

### 3.4.4.6 Carbonate

Research reports about the effect of CaCO$_3$ are not that abundant. However, CaCO$_3$ can have important effects on the aggregate stability. Muneer and Oades (1989) studied the role of Ca-organic interactions in soil aggregate stability, using field experiments with 14C-labeled straw, CaCO$_3$ and CaSO$_4$·2H$_2$O (gypsum). They found that the addition of 14C-labelled wheat straw (with or without gypsum or CaCO$_3$) increased the stability of aggregates greater than 2 mm diameter. The addition of Ca during the decomposition of straw resulted in a synergistic stabilization of aggregates greater than 1 mm. Moreover, the stabilization was prolonged in the presence of Ca. However, Ben-Hur et al. (1985), comparing calcareous and non-calcareous soils, observed no effect of CaCO$_3$ on the infiltration rate under simulated rainfall. From a chemical point of view, carbonates in soil should be favorable to aggregate stability and infiltration rate because of the Ca cations. The effect probably depends on the size distribution of the CaCO$_3$ particles and on the clay content. When CaCO$_3$ particles have the size of silt and the clay content is low, the soil may behave as a silty soil (i.e., as a very unstable soil). For beneficial aggregating effect of CaCO$_3$ sufficient clay must be present (Le Bissonnais, 1996).

### 3.5 Water erosion models

Up to now, there are many soil erosion models. However, the USLE (Universal Soil Loss Equation) (Wischmeier and Smith, 1978), was and still is the most popular one. It is an empirical model, derived from more than 10,000 plot-years. One plot year is equivalent to the measurement of the soil erosion during one year on a standard erosion plot of 22.13 m long and 1.83 m width, with a uniform slope steepness of 9%. This model became so popular because of the relative ‘small’ amount of necessary data and because of it’s simplicity. Because the USLE is an empirical model, derived from observations in the USA, this model can not be directly used in other parts of the world with different climatological features and different soil properties. Bollinne (1982) adjusted the model
using data from erosion plots — for the Belgian climatological boundary conditions and soil properties.

Only recently, due to the development in modern computer technology, the attention in soil loss prediction is focused on the physical and process-based description of the soil erosion processes. The advantage of these types of models is that they have to be calibrated only once. This is a huge advantage because the calibration phase of a model is expensive, both in time and money.

Before going into details about the available models, a short outline is given about the different types of models and their general structure.

### 3.5.1 Model structures

A model is a description of the physical reality using mathematical equations. The origin of these equations will determine the general model concept: the two basic model structures are *theoretical* and *experimental*.

The theoretical model structure is also called conceptual. In this method, the modelled system is described by the basic physical laws (equilibrium equations, conservation of mass and energy, etc...) which determine the system behaviour. Depending on the complexity of the processes one wants to model, there are almost always simplifications necessary in order to make the calculations practicable. Many systems are not only time-dependent, but also space dependent. Such systems are systems with *distributed* parameters, and their corresponding models are then called distributed models. Usually, these models are simplified by *lumping*: the physical equations are only applied on some points in space.

The experimental model structure is also called black-box or empirical. In this structure, the model is derived from measurements. Starting from a theoretical analysis, the input and output variables are measured. These measurements are used to find mathematical relations between the measured variables. A difficult task is the elimination of disturbing influences, which may result in erroneous measurements.

Both theoretical and experimental models have their advantages and disadvantages. However, a huge advantage of the theoretical model structure is that the mathematical
relation is conserved between the system parameters and the physical parameters of the processes. In experimental models these relations are lost, the system parameters are then just numbers without a physical meaning. When the system becomes more complex, it becomes more and more difficult to implement a theoretical model. This is a drawback not occurring in the experimental methodology.

In both theoretical and experimental methods, one can make a distinction between stochastic and deterministic models. A model is stochastic if one or more system parameters are random variables. This means that the exact value of that parameter can not be determined. In natural systems such variables are abundant (e.g. the daily amount of rainfall, the length of the daily queues on the motor ways, etc...). Probably, almost all physical, social and economic parameters have a stochastic nature. In some cases, these fluctuations are not important and are not taken into account. Then, one can say that these parameters are quasi-deterministic. Consider the output variable $y(t)$ of a dynamical system with as input variable $u(t)$. When the value of $y(t)$ can be exactly derived from the system behaviour and from the input variable $u(t)$, then we can say that $y(t)$ is a deterministic system. It is then possible to make the following division concerning the model structure (Clarke, 1973):

1. stochastic-theoretical
2. stochastic-experimental
3. deterministic-theoretical
4. deterministic-experimental

These four groups can then be further extended in:

1. linear systems
2. nonlinear systems

When the output signal $y_1(t)$ corresponds with the input $u_1(t)$ and $y_2(t)$ corresponds with the input $u_2(t)$, then the system is called linear when $y_1(t) + y_2(t)$ corresponds with $u_1(t) + u_2(t)$ (superposition principle). This linearity may not be confused with the linearity in statistical (regressions) sense.
3.5.2 Short overview of the available erosion models

The subdivision of models concerning their structure, outlined above, is idealized. Most, if not all water erosion models have a mixture of theoretical and experimental components. This is quite logical because water erosion and soil loss is the result of many subprocesses. There exist fully experimental models, but there is no pure theoretical water erosion model. The water erosion models described below are subdivided according to their major structure. Note that the list of water erosion models is far from exhaustive.

3.5.2.1 Experimental models

*Universal Soil Loss Equation (USLE)*

The first attempt to develop a soil loss equation for hillslopes was that of Zingg (1940), who related erosion to slope steepness and slope length, according to:

$$ E \propto \tan^n(\theta) \cdot L^n $$  \hspace{1cm} (3.37)

where $E$ is the soil loss per unit area, $\theta$ is the slope angle and $L$ is the slope length. In a study of data from five experimental stations of the United States Soil Conservation Service, Zingg (1940) found that $m = 1.4$ and $n = 0.6$.

Since the values for the exponents have been confirmed in respect of $m$ by Musgraves (1947) and $m$ and $n$ by Kirkby (1969), there is some evidence to suggest that equation (3.37) has general validity. However, working with data from experimental stations in Zimbabwe, Hudson and Jackson (1959) found that $m$ was close to 2.0, indicating that the effect of slope is stronger under tropical conditions where rainfall is heavier. The effect of particle size is illustrated by the short slope laboratory experiments of Gabriels et al. (1975). These authors show that the value of $m$ increases with the grain size of the material, from 0.6 for particles of 50 $\mu$m to 1.7 for particles of 100 $\mu$m.

Further developments led to the addition of a climatic factor based on the maximum 30-minute rainfall total with a 2-year return period (Musgrave, 1947), a crop factor, to take into account the soil protection of different crops (Smith, 1958), a conservation factor and a soil erodibility factor (Wischmeier and Mannering, 1969). Changing the
climatic factor to the rainfall erosivity index (R) ultimately yielded the Universal Soil Loss Equation (USLE) which, following considerable experience with its use, was later modified and up-dated (Wischmeier and Smith, 1978). The general equation is:

\[ A = R \cdot K \cdot L \cdot S \cdot C \cdot P \]  \hspace{1cm} (3.38)

where A is the mean annual soil loss, R the rainfall erosivity factor, K is the soil erodibility factor, L is the slope length factor, S is the slope steepness factor, C is the crop management factor and P is the erosion-control practice factor.

**Modified Universal Soil Loss Equation (MUSLE)**

There were many different MUSLE models developed. All these models are identical to the USLE except for the rainfall erosivity factor, R. The modified R-factor is then not only a representation of the energy of the rain events, but also a quantification of the hydrological runoff. An example of a modified R-factor is given by Kinnell et al. (1994). The E\textsubscript{130} index used for the rainfall-runoff factor (R) in the USLE modelling environment was originally developed from the empirical observation that soil loss increases with rainfall amount and the intensity of the rainfall event. In an analysis of erosion data from nonvegetated plots at Holly Springs, Mississippi, the I\textsubscript{X}E\textsubscript{A} index, an index that is based on the product of the excess rainfall rate (I\textsubscript{X}) and the rate of expenditure of rain kinetic energy (E\textsubscript{A}), was shown to be superior to the E\textsubscript{130} index (Kinnell et al., 1994). Apart from accounting better for the processes of detachment and transport, the index provides a possibility to consider the effects of hydrology more directly within the USLE environment than is currently possible. However, including more direct consideration of hydrology within R is likely to affect a number of other USLE factors as well. These effects have yet to be evaluated and for this reason, the MUSLE models are almost never applied in practical studies.

**Revised Universal Soil Loss Equation (RUSLE)**

Under impulse of new erosion-plot observations and the intensive use of the USLE in all parts of the world, a revision of the original USLE was necessary. For this research study the reference manual of Renard et al. (1996) was used. The RUSLE retains the
six factors of the USLE, denoted in equation (3.38), for the calculation of the mean annual soil loss. Major changes were made for each of the factors. The algorithm for calculating the erosivity factor (R) was altered. The erodibility factor (K) was changed to take into account frost periods. The slope length and slope steepness factors (LS) were changed to reflect the ratio of rill to interrill erosion. The C factor is now calculated as the product of terms reflecting prior land use, surface cover, crop canopy, and surface roughness. The support practice factor (P) was expanded to consider conditions for contouring, stripcropping and terracing, as well as to address conditions for rangelands, and pastures.

*Agricultural Nonpoint Source (AGNPS)*

The AGNPS model, developed by Young et al. (1985), is an event-based, watershed-scale model developed to simulate runoff, sediment, chemical oxygen demand (COD), and nutrient transport in surface runoff from ungaged agricultural watersheds. Sub-surface transport processes are not considered at present, but a ground water loading version of the model is planned. Nutrients considered include nitrogen and phosphorus. The model operates on a square cell basis, which facilitates data base creation. All model inputs are defined on the cell level. Pollutants are routed from the source cell through intervening cells to the watershed outlet. Model output may be viewed at any cell, a capability that allows identification of critical source areas and evaluation of targeting alternatives. Runoff volume is simulated using the SCS curve number method (Soil Conservation Service, 1972) and the peak runoff rate equation used in CREAMS. Erosion and sediment transport are calculated with modified forms of the USLE and Bagnold’s stream power equation (Bagnold, 1966). Nutrient yield in the sediment-bound phase is calculated as a function of the nutrient content of the field soil, the sediment yield, and an enrichment ratio which is a function of soil texture and sediment yield. Soluble nutrient loss is a simple function of the soil nutrient level and an extraction coefficient. The model considers only losses of total nitrogen and phosphorus and does not consider nutrient transformations. The model also allows for inputs from feedlots, sewage treatment plants, and other point sources. Data file creation is time consuming, since 22 parameters must be specified for each cell.
3.5.2.2 Theoretical models

*Water Erosion Prediction Project (WEPP)*

The objective of the Water Erosion Prediction Project is: “To develop new generation water erosion prediction technology for use by the USDA–Soil Conservation Service, USDA–Forest Service, and USDA-Bureau of Land Management, and other organizations involved in soil and water conservation and environmental planning and assessment” (Foster and Lane, 1987). Currently there exists a hillslope profile model (Lane and Nearing, 1989) and a watershed model (Flanagan and Nearing, 1995). The WEPP erosion model is intended to be executed primarily as a continuous simulation model, although it can be run on a single-storm basis. By continuous simulation it is meant that the model mimics the processes which are important to erosion prediction as a function of time, and as affected by management decisions and climatic environment. The WEPP model can be subdivided into a series of conceptual components: climate generation, frozen soils, snow accumulation and melt, irrigation, infiltration, overland flow hydraulics, water balance, plant growth, residue decomposition, soil disturbance by tillage, consolidation, and erosion and deposition. The model is too extensive to give a description of the different components. Although this model was intended to replace the (R)USLE, it will never reach the status of the (R)USLE because of the enormous amount of input parameters required.

*European Soil Erosion Model (EUROSEM)*

EUROSEM (Morgan, 1996) is an event-based model designed to compute the sediment transport, erosion and deposition over the land surface throughout a storm. EUROSEM simulates rill and interrill erosion explicitly, including the transport of water and sediment from interrill areas to rills, thereby allowing for deposition of material en route. However, EUROSEM is a very complex model, with a lot of difficult to assess parameters (e.g.: number of rills in a field, rill width . . . ), which reduces strongly the applicability of the model in real modelling applications.

*Areal Nonpoint Source Watershed Environment Response Simulation (ANSWERS)*

ANSWERS (Beasley et al., 1980) is an event-oriented, watershed-scale model developed to describe the impact of existing and proposed agricultural management practices
on water quality in unaged watersheds. Recent versions of ANSWERS include an extended sediment detachment/transport model allowing prediction of sediment yield and concentrations for mixed particle size distributions, a phosphorus transport model (Storm et al., 1988), and a nitrogen transport model. ANSWERS subdivides the watershed into a uniform grid of square cells. Land use, slopes, soils and management practices are assumed uniform within each cell. Typical cell sizes range from 0.4 to 4 ha, with smaller cells providing more accurate simulations. Eight to 10 parameter values must be provided for each cell. The extended sediment model uses a modification of Yalin’s equation (Yalin, 1963) similar to that in CREAMS.

Chemicals, Runoff and Erosion from Agricultural Management Systems (CREAMS)
CREAMS (Knisel, 1980) is a physically-based, field-scale model developed for comparing pollutant loads from alternate management practices. Although it is intended for use as a continuous simulation model, it can also be used as an event-oriented model. The model estimates runoff volume, peak runoff, infiltration, evapotranspiration, soil moisture, percolation, sediment yield, particle-size distribution of eroded sediment, and losses of dissolved and adsorbed nitrogen, phosphorus, and pesticides in surface runoff and percolate. The primary limitation of the model is that as a field-scale model, it is limited to areas with uniform soils and cropping and does not consider pollutant transport to receiving waters. CREAMS has been found to underestimate runoff volumes. It is more accurate in representing average annual runoff volumes than daily or monthly runoff volumes (Knisel, 1980).

Erosion Productivity Impact Calculator (EPIC)
The EPIC model is developed by Williams et al. (1984). Later versions (Williams et al., 1989) were more focused on plant productivity than on erosion. The model describes soil erosion, economic yield, amount of drainage water and resolution of nitrogen in dependence on the type of crop, the type of soil, the amount of water and the use of fertilization either in form of fertilizer or liquid manure. The model assumes a homogeneous area of 1 ha. The soil is vertically divided in up to 10 layers. The model handles the following fields: water, nitrogen and phosphorbalance in the soils and the crops, cropgrowth, weather, soil temperature.
Griffith University Erosion and Sediment Simulator (GUESS)

GUESS is a mathematical model which simulates the processes of erosion and deposition along a hillslope (Rose et al., 1983). The difference with other erosion models is that in the GUESS model the soil is separated into two parts: that which is the original soil and possesses a certain degree of cohesion and that comprising recently detached material with no cohesion.

Kinematic Runoff and Erosion Model (KINEROS)

The KINEROS model is developed by Woolhiser et al. (1990) and describes the processes of interception, infiltration, surface runoff and erosion for small rural watersheds. A watershed has to be represented by a cascade of planes (overland flow) and channels (channel flow). The runoff routing is solved by a finite difference scheme of the kinematic wave equation.

EROSION-2D/3D

The EROSION-2D/3D model developed by Schmidt (1991), is an almost pure physical model. It is based on calculating the momentum flux of the raindrop impact \( \phi_r \) and the overland flow \( \phi_q \). For every hillslope segment the erosion number, \( E \), can be determined by:

\[
E = \frac{\phi_r + \phi_q}{\phi_{critical}}
\]  

(3.39)

wherein \( \phi_{critical} \) is the critical momentum flux, below which no particles can be transported. Once the erosion number is known, the following transport equation can be used to determine the sediment transport, \( q_s \) (kg·m\(^{-1}\)·s\(^{-1}\)):

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

(3.40)

This regression equation was derived from laboratory rainfall simulations.
3.5.3 Variability and uncertainty

Almost all data a model is fed with, is subject to some kind of uncertainty. This can be caused by measurement errors, measuring device accuracy, the way we generalize spatial data (interpolation methods). All these errors propagate throughout the modelling process. The results are then also subject to uncertainty. In order to support decision making, it is important to have an idea of the degree of uncertainty on the end results. Using error propagation techniques (Heuvelink, 1998) the amount of uncertainty on the end products can be determined. If one wants to decrease the uncertainty on the end result, these techniques provide a methodology to pinpoint the input parameters which should be measured more accurately. In chapter 4 much attention is given on the error propagation techniques.
Chapter 4

Predicting Long-Term Sediment Transport

This chapter is published in Biesemans et al. (2000), and some parts of this chapter appeared also in Biesemans et al. (1998).

4.1 Introduction

Abundant rains in the winters of 1993–94 and 1994–95, resulted in substantial on-site and off-site erosion problems in Belgium. Therefore, the Regional Land Management Board of the hilly region in the south of West-Flanders requested a scientific study of these problems. This Board was installed by Ministerial Decree to preserve and protect the landscape of this area, which is considered to be of an exceptional value. The motivation for this request came from the need for information to support ‘land management agreements’ between farmers and this Board.

Although process-based erosion models, such as the Water Erosion Prediction Project (WEPP) (Flanagan and Nearing, 1995) are being developed to replace the empirical models (Laflen et al., 1991), the Revised Universal Soil Loss Equation (RUSLE) (Renard et al., 1996) was selected as a basis to develop an erosion expert system. The latter should be able to assess both on-site soil losses and off-site sediment accumulations. This system was implemented in ANSI C (American National Standards Institute), intended to be used inside a Geographic Information System (GIS) environment. The
RUSLE was chosen mainly based on the limited amount of data needed to perform a field-scale erosion analysis for larger areas, compared to the process-based models.

To predict the off-site sediment accumulation in rivers or water reservoirs, the equations of the erosion model describing the hydrological processes must be a measure of the transport capacity of the overland flow. Only when this is realized, can the amount of sediment leaving a field be estimated. The RUSLE is a factor-based erosion model designed to predict long-term average soil losses carried by runoff from specific field slopes in specified cropping and management systems (Renard et al., 1996). The factor that summarizes the hydrological components of the water erosion process is the topographic factor, LS. Foster and Wischmeier (1974) stated that the LS equation, derived from unit field plots, with a length of 22.13 m and 1.80 m width, applies to situations where detachment limits the sediment load, and is not a transport-capacity equation. However, recent rill erosion experiments conducted by Nearing et al. (1997) indicate that transport capacity in eroding rills is already reached within a sample length of 2.5 m for slopes ranging between 3 to 28%. These results support the statement of Moore and Wilson (1992) that the equations of the topographic factor in the RUSLE model are also a measure of the sediment transport capacity of overland flow. This implies that the RUSLE can estimate the sediment actually leaving a field and does not account for deposition as colluvium (intrabasinal storage).

The end product of a model is always the result of operations and computations performed on uncertain data. In GIS studies every layer of information has its associated uncertainty caused by different sources of variance (Heuvelink et al., 1989). Some of these sources can be unavoidable (e.g., the intrinsic variability of the climate or the uncertainty associated with interpolation methods). If the results of a model are not in agreement with the field observations it is important to know if this is mainly due to the model itself or mainly due to the uncertainty of the model input. This can be evaluated with the Monte Carlo error propagation technique (Wesseling and Heuvelink, 1995). With this technique the model output is generated at least a few hundred times, but instead of using the parameter values, their stochastic distributions are used. This allows a modeler to determine the stochastic properties of the model output.

The factor of the RUSLE model that poses the most problems in the error propagation process is the LS factor. Desmet and Govers (1996) developed a method to calculate
the LS factor on topographically complex landscapes within a GIS, based on the unit contributing area. For this study, an alternative method to calculate the LS of a field was developed. This alternative method is capable to perform a Monte Carlo analysis and is more closely related to the linear structure of the RUSLE model and the linear micro-topography that can be observed in the field. The linear micro-topography is created by the cultivation techniques: crops are cultivated in rows and the usage of heavy machinery creates linear furrows, which initiates rill erosion. This induces a parallel flow pattern and prevents the concentration of the Horton runoff into bigger rills and gullies. These features can not be captured by the unit contributing area because the resolution of most Digital Elevation Models (DEM) is too low to describe these linear micro-topography.

This chapter has two main objectives: (a) to evaluate if the RUSLE can be used to predict long-term off-site sediment accumulation, which is equivalent to checking if the RUSLE LS equations are a measure of the transport capacity, and (b) to perform a Monte Carlo error propagation to determine the uncertainty of the calculated on-site soil losses and off-site sediment accumulation.

4.2 Assessment of the RUSLE variables

The procedure described below was applied to the Kemmelbeek watershed (figure 4.1), located in the south of West-Flanders, within the loess belt of Belgium. It covers an area of 1075 ha and feeds a drinking-water reservoir of the city Ieper. The highest elevation is 151 m which drops to 23 m at the reservoir inlet. The average slope steepness is 4.6 %, with a maximum value of 71 %, although 99 % of the slopes are less then 30 %. This watershed was chosen as a pilot test area to determine the on-site soil losses and off-site sediment accumulation in the reservoir, using the adapted RUSLE expert system. A validation of the predictions was possible, since data on the sediment input in the reservoir are available. The sediment trapping efficiency of this reservoir is close to 100 %. This is based on the fact that (1) most of the incoming sediment is deposited directly behind the reservoir inlet, (2) the daily amount of water used for drinking-water production is a volume of approximately 4000 m³, which corresponds with a waterlayer of only 1.15 cm for a total reservoir area of 34.88 ha and (3) excessive
water is only pumped into the downstream drainage system when the reservoir exceeds a certain critical level (the pumps are located 700 m from the reservoir inlet).

![Image of a 3D view of the Kemmelbeek river and watershed.]

Figure 4.1: 3D view of the Kemmelbeek river and watershed.

In a GIS environment a model can be written as (Wesseling and Heuvelink, 1995):

\[ M = F(Z_1, Z_2, \ldots, Z_n, a_1, a_2, \ldots, a_m) \]  \hspace{1cm} (4.1)

where the resulting map \( M \) is obtained by applying expression \( F \) on input maps \( Z_i \) and model coefficients \( a_i \). Because most model input is subject to uncertainty, not all parameters of \( F \) are exactly known. Therefore the parameters of \( F \) must be represented by probability distributions rather than by deterministic quantities. A Monte Carlo error propagation technique can then be used to determine the distribution of the model output. This technique repeatedly (e.g. 500 times) runs the model with input values that are sampled from their distributions. If the number of runs is sufficiently large (depending on the model complexity) the distribution obtained from the runs will approximate the true distribution of the model output.

Before an error propagation can be performed the following must be determined (Wesseling and Heuvelink, 1995):

1. the properties of function \( F \);
2. which parameters are stochastic and which are deterministic;
3. the probability distribution of the stochastic parameters;
4. the correlation between the different parameters at the same location;
5. the correlation between the spatial parameters at different locations.

In this study, the function $F$ in equation (4.1) is the RUSLE model. To test if the elaborated methodology of the RUSLE expert system is capable to predict the off-site sediment accumulation, it is hypothesized that the RUSLE model itself induces no error. The uncertainty of the output is then only induced by the uncertainty of the model input parameters: LS, R, K, C and P. Therefore, the stochastic properties of these input parameters must be determined.

### 4.2.1 The topographic factor (LS)

The RUSLE topographic factor describes the combined effect of slope length (L) and slope steepness (S). Because a terrain element downslope gets more runoff water than a terrain element near the water divide, Foster and Wischmeier (1974) subdivided a slope into a number of uniform segments. The LS of a flowline in the landscape can be calculated by:

$$LS = \sum_i \left[ \frac{(L_i^{M_i+1} - L_{i-1}^{M_i+1}) \cdot S_i}{L_{tot} \cdot 22.13^{M_i}} \right]$$

(4.2)

where $L_{tot}$ is the total length of a flowline (m), $L_i$ is the length from the top of the slope to the foot of segment $i$ (m), $L_{i-1}$ is the length from the top of the slope to the top of segment $i$ (m), $S_i$ is the slope steepness factor for segment $i$ (-) and $M_i$ is the slope-length-exponent (-). The slope-length-exponent can be written as (McCool et al., 1989; Renard et al., 1996):

$$M = \frac{\beta}{1 + \beta}$$

(4.3)

$$\beta = \left[ \frac{\sin(\alpha)/0.0896}{3.0 \cdot (\sin(\alpha))^{0.8} + 0.56} \right] \cdot r$$

(4.4)
where $\beta$ is the rill/interrill ratio (\%) and $\alpha$ is the slope steepness (radians). When field conditions favor rill erosion $r = 2$ (e.g. on ridged potato fields), when field conditions favor sheet erosion $r = 0.5$ (e.g. a field that remains bare for a long time), and for in-between conditions $r = 1$. The main crops in the study area are beets, maize, potatoes and wheat. The field conditions for these crops favor rill erosion. Therefore, an $r$-value of 2 was used, except for pasture fields and forested areas where $r$ was set to 1.

The slope steepness factor (for a slope length longer than 4.56 m) is given by (McCool et al., 1987; Renard et al., 1996):

$$ S = \begin{cases} 
10.8 \cdot \sin(\alpha) + 0.03 & \text{if } slope < 9\% \\
16.8 \cdot \sin(\alpha) - 0.50 & \text{if } slope \geq 9\% 
\end{cases} \quad (4.5) $$

Due to the small parcels in the study area (the average field area is only 1.40 ha) and the dense ditch system, every field can be considered as a separate hydrological unit concerning Horton flow. Consequently, the flowlines in the landscape start at the upper field boundaries and end at the lower field edges, where the runoff water flows into the drainage system. The calculation process for the LS factor is shown in figure 4.2. Suppose a square field or hydrological unit with a length and a width of 80 m. In a 10 m resolution DEM, this results in a grid with 8 rows and 8 columns. Figure 4.2A gives the drainage directions for each elementary cell of the grid. These drainage directions are used to construct linear flowlines in the field (figure 4.2B). The flowline matrix ($FLM$), given in figure 4.2C, indicates the number of flowlines running through a cell. The representative area for each segment of a flowline is the cell area divided by the $FLM$ value for that segment. The actual erosion (ton-year$^{-1}$) along a flowline, $A_f$, can then be written as:

$$ A_f = \sum_{i=1}^{N} \left[ \frac{(L_{i+1}^{M_{i+1}} - L_{i+1}^{M_{i+1}})}{L_{tot} \cdot 22.13^{M_{i}}} \cdot a_i \cdot R_i \cdot K_i \cdot C_i \cdot P_i \right] \quad (4.6) $$

where $N$ is the number of segments in a flowline, $a_i$ is the representative area of a segment in a flowline (ha), and $R_i$, $K_i$, $C_i$, and $P_i$ are the other RUSLE factors, respectively, the rain erosivity (MJ:mm:ha$^{-1}$:h$^{-1}$:year$^{-1}$), the soil erodibility (ton:ha:h:ha$^{-1}$:MJ$^{-1}$:mm$^{-1}$), the cover-management factor (\%) and the support practice factor (\%). If $dx$ and $dy$ are the cell dimensions (m), the representative area (ha)
Figure 4.2: Illustration of the calculation process for the topographic factor. Figure A shows the flow directions for each elementary cell of the field or hydrological unit. Figure B shows the possible flow lines in the field. Figure C shows the flowline matrix (FLM) for the field, indicating the number of flow lines passing through a cell.

The actual erosion ($\text{ton-field}^{-1}\cdot\text{year}^{-1}$) of a segment in a flowline can be calculated by:

$$a_i = \frac{dx \cdot dy}{10000 \cdot FLM_i}.$$  \hspace{1cm} (4.7)

The total actual erosion ($\text{ton-field}^{-1}\cdot\text{year}^{-1}$) of a field or hydrological unit, $A_{\text{field}}$, results from:

$$A_{\text{field}} = \sum_{f=1}^{F} A_f$$  \hspace{1cm} (4.8)

where $F$ is the number of flowlines in a field or hydrological unit.

There are two sources of variance in the LS algorithms which influence the uncertainty on the predicted sediment loss:

1. The uncertainty on the elevations in the DEM. For every iteration in the error propagation process, an error surface was created and added to the original DEM. If the error between the DEM elevation and the real elevation for a certain point is high, the error will also be high for positions in the neighborhood of that point. Therefore the error surfaces must be autocorrelated over short distance. This requires the construction of a random generator which can create autocorrelated error surfaces.
2. The rilling pattern that can be observed in a field is never the same for every storm event. To simulate this randomness of the Horton flow a stochastic flow routing model was used to determine the flow directions.

4.2.1.1 Construction of autocorrelated surfaces

The fractional Brownian motion \((fBm)\) and fractional Gaussian noise \((fGn)\) serves as the basis for many models for natural fractal shapes such as landscapes (Polidori, 1991; Peitgen et al., 1992). In its one-dimensional form, the fractional Brownian motion model is defined as a continuous function, \(f(x)\), of the independent spatial variable, \(x\), having the following properties (Molz and Liu, 1997; Peitgen et al., 1992):

1. The increments of \(f\) are stationary. This means that for all values of \(x\) and a fixed increment, \(h\):

\[
E[f(x + h) - f(x)] = C_1(h)
\]  
(4.9)

\[
E[(f(x + h) - f(x))^2] = C_2(h) = \gamma(h)
\]  
(4.10)

where \(E[X] \equiv \text{expected value of the random variable } X\), \(\gamma(h)\) is the variogram, and \(C_1\) and \(C_2\) are functions of \(h\). Since the increments of \(f(x)\) are defined to be stationary, one can define:

\[
n(x, h) = f(x + h) - f(x)
\]  
(4.11)

with the statistical properties of \(n\) depending only on \(h\).

2. The variable \(n\) has a Gaussian distribution with:

\[
E[n(x, h)] = 0 \text{ and } E[(n(x, 1))^2] = \sigma^2.
\]  
(4.12)

Thus in equations (4.9) and (4.10), \(C_1(h) = 0\) for all \(h\), and \(C_2(1) = \sigma^2\).

3. The increments, \(n(x, h)\), are statistically invariant with respect to an affine transformation. This means that the random variable \(n(x, rh)\) and \(r^n n(x, h)\), with \(r\) and
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$H$ constants ($0 \leq r < \infty$ and $0 < H < 1$), have the same Gaussian distribution. Thus:

$$E[n(x, rh)] = E[r^H n(x, h)] = 0$$  \hspace{1cm} (4.13)

$$E[(n(x, rh))^2] = E[(r^H n(x, h))^2] = r^{2H} E[(n(x, 1))^2].$$  \hspace{1cm} (4.14)

Using equations (4.12), (4.13) and (4.14), one can write:

$$E[(n(x, h))^2] = h^{2H} E[(n(x, 1))^2] = h^{2H} \sigma^2 = \gamma(h).$$  \hspace{1cm} (4.15)

In order to define $fBm = f(x)$, it was necessary to define the properties of its increments, $n(x, h)$. This last function is known as $fGn$. The concepts of $fBm$ and $fGn$ are generalizations of the classical concepts of Brownian motion and Gaussian noise, denoted as $cBm$ and $cGn$. These functions are obtained by setting $H = 0.5$. The parameter, $H$, is the so-called Hurst coefficient and is an indicator of the surface complexity. $H$ is related to the fractal dimension of the surface and can be estimated by fitting a power function, $y = a \cdot x^b$, on the variogram. According to equation (4.15), $a \approx \sigma^2$, $b \approx 2H$, when $y$ is the semivariance and $x$ is the lag distance. The $fBm$ can be divided into three distinct categories: $H < 0.5$, $H = 0.5$ and $H > 0.5$. The case $H = 0.5$ is the ordinary Brownian motion, which has independent increments. For $H > 0.5$ there is a positive correlation between the increments. For $H < 0.5$ the opposite is true. In the latter case there is a negative correlation between the increments and the curves or surfaces seem to oscillate more erratically. As an illustration, figure 4.3 gives three $fBm$ series with a Hurst coefficient of 0.2, 0.5 and 0.8 with their respective variograms. Random surfaces with a given Hurst coefficient can be created by the ‘random midpoint displacement’ algorithm (Peitgen et al., 1992).

### 4.2.1.2 Flow routing algorithm

Several flow-routing models exist to determine the flow directions in a grided DEM: the deterministic eight neighbors method (D8) (O’Callaghan and Mark, 1984; Jenson and Domingue, 1988), the stochastic eight neighbors method (Rho8) (Fairfield and Leymarie, 1991), multiple-direction methods (Quinn et al., 1991; Freeman, 1991) and DE-
Figure 4.3. Three fBm traces generated with the random midpoint displacement method and with a specified Hurst coefficient ($H$). To estimate the Hurst coefficient ($<H>$) a power function is fitted at the variogram.

MON (Costa-Cabral and Burges, 1994). For this application the Rho8 method, which is basically a stochastic extension of method D8, was chosen based on its stochastic character to determine the flow directions.

In method D8, each pixel discharges into one of its eight neighbors. The drainage direction is determined by the direction of the largest weighted elevation drop ($LD$), calculated by:

$$LD = (H_c - H_n) \cdot \rho$$

(4.16)

where $H_c$ and $H_n$ is the elevation of the center cell and the neighbor cell respectively, and $\rho$ is the weight factor for the direction and equals: $1/dx$ in the $x$ direction, $1/dy$ in the $y$ direction and $1/\sqrt{dx^2 + dy^2}$ in the diagonal directions. If $dx$ equals $dy$, $\rho$ is
equivalent to 1 for the cardinal directions and 1/\sqrt{2} for the diagonal directions. This method gives maximum errors in the flow direction of 22.5 degrees in planar areas which are not aligned with the grid orientation (Fairfield and Leymarie, 1991). To solve this problem, method Rho8 gives the weight factor \( \rho \) a stochastic character. For the cardinal directions \( \rho \) equals 1 and for the diagonal directions \( \rho \) ranges between [0.5,1.0] with a mean value of 1/\sqrt{2}, according to its cumulative distribution function (cdf):

\[
P(\rho \leq x) = \begin{cases} 
0 & x < 0.5 \\
2 - 1/x & 0.5 \leq x \leq 1 \\
1 & x > 1
\end{cases}
\]  

(4.17)

Let \( r \) be an uniformly distributed random variable between [0, 1], then the inverse of the cdf, \( \rho = 1/(2 - r) \), generates random weight factors for the diagonal directions, with a mean value of 1/\sqrt{2}. Using this flow routing model, every iteration in the Monte Carlo error propagation process an other flow pattern is generated.

For the calculation of the LS factor, a DEM of the study area was interpolated from the contour lines, digitized from the Belgian topographic maps with a 1/100000 scale. The grid resolution was chosen as 10 m. The field boundaries were digitized from the 1/10000 orthophotos. The random midpoint displacement method, used in the Monte Carlo simulations, requires two parameters: the Hurst coefficient, indicating the complexity of the DEM error surfaces and a standard deviation, indicating the dimension of the errors. In order to determine the Hurst coefficient, the elevation was measured along a transect down a hill slope. These measurements were compared with the elevations in the DEM.

Figure 4.4A gives the trace of the errors along the transect and figure 4.4B shows the respective variogram of this error trace. Fitting a power function to the variogram results in a Hurst coefficient of 0.83, indicating a positive correlation between the errors at successive points. The interval of the digitized contour lines, used to interpolate the DEM, was 2.5 m. Consequently, the maximum error of the elevations in the DEM is 2.5 m, which can be considered to be normally distributed. The standard deviation, used in the random midpoint displacement method, is then approximately one-sixth of the contour interval, or 0.42 m, according to (with \( f(x) \) the normal probability density
function): 

\[ \int_{-3\sigma}^{+3\sigma} f(x) \, dx = 0.997. \]  

\[ (4.18) \]

Figure 4.4: Measured and DEM elevations along a transect (figure A) and the variogram (figure B) of the error trace. Fitting a power function at the variogram results in a Hurst coefficient of 0.83.

Figure 4.5 gives the results of the Monte Carlo error propagation: the mean LS value of a field and the uncertainty of these estimations, expressed by the standard deviation.
Figure 4.5: The mean LS value of a field and the standard deviation of these estimations.
4.2.2 The rain erosivity (R)

Precipitation over a 27-years span was used to calculate the rain erosivity. Over this period, rain intensity was recorded every 10 minutes. The equations to calculate the erosivity can be found in Renard et al. (1996). For the Kemmelbeek watershed, the mean yearly rain erosivity is 724 (MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·year\(^{-1}\)). The natural variability of the precipitation characteristics (intensity and amount) is very large. Consequently, the standard deviation of the R-factor is also very large: 224 (MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·year\(^{-1}\)). Because the RUSLE model predicts the mean soil loss over long time periods (a few decades) it is not necessary to take this uncertainty into account in the model calculations. Therefore the R-factor was considered to be deterministic, with a value of 724 (MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·year\(^{-1}\)). This value is in agreement with the erosivity map of Bollinne et al. (1980), given in figure 4.6.

![Kemmelbeek watershed map](image)

Figure 4.6: Erosivity map of Belgium (Bollinne et al., 1980). Units on the original map are converted to S.I. units (MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·year\(^{-1}\)).
4.2.3 The soil erodibility (K)

The soil erodibility factor can be calculated by (Renard et al., 1996):

\[
K = \left[ \frac{2.1 \cdot (S \cdot (100 - C))^{1.14} \cdot 10^{-4} \cdot (12 - OM)}{100} \right] \cdot 0.1317
\]  

(4.19)

where \( K \) is the soil erodibility (ton·ha·h·ha\(^{-1}\)·MJ\(^{-1}\)·mm\(^{-1}\)), \( S \) is the textural fraction between 2 and 100 \( \mu \text{m} \) (%), and \( OM \) is the organic material content (%).

The Belgian soil map contains only qualitative information, so it is not appropriate to convert this into quantitative information. Therefore within the watershed 154 locations were sampled to determine the textural fractions and the organic material content. Using equation (4.19), the erodibility of the samples was calculated and block-kriged (Webster and Oliver, 1992; Van Meirvenne, 1991) — using the variogram given in figure 4.7 — for blocks of 10x10 m, resulting in an erodibility grid (figure 4.8A) of the same resolution as the DEM. The uncertainty of this interpolation is expressed by the kriging variance (figure 4.8B). The \( K \) factor can be considered to be normally distributed with the kriged value as mean and the kriging variance as a measure of the spread of the estimation error.

![Figure 4.7: Variogram used to block-krige an erodibility grid. Fitting a spherical model at the experimental variogram results in a nugget of 4.286\( \cdot 10^{-5} \), a sill of 2.092\( \cdot 10^{-1} \) and a range of 832 meter.](image)
Figure 4.8: Block-kriged erodibility map (figure A). The darker colors indicate the loamy soils, the lighter colors indicate the (partial) denudation of Tertiary sands. Figure B gives the kriging standard deviation. The standard deviation is directly proportional with the distance from the sample positions.
4.2.4 The cover-management factor (C)

The dimensionless C factor, which has a range between 0 and 1, expresses the degree of protection of the soil surface by the crops or vegetation. The information for calculating the C factor was obtained from a detailed inquiry with the farmers (Ghekiere, 1997). However, there was not enough information (e.g. soil biomass, crop residues, soil consolidation, ...) to use the RUSLE methodology. Instead, the USLE methodology was used to calculate the C factor for every crop rotation.

For every crop the growing season was subdivided in 6 stages. However, not every crop cultivated in Western Europe can be found in the USLE crop database. The C value for these crops must be estimated using the data of similar crops. Therefore, for every growing stage a possible minimum and maximum C value was chosen from the USLE C factor table — table 5 of Agriculture Handbook No 537 (Wischmeier and Smith, 1978) — and weighted with the erosivity value of that period. This resulted in two maps (figure 4.9) representing the minimum C factor and the maximum C factor for a field. The C-factor of a field can be considered to be uniformly distributed between this minimum and maximum C value.
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Figure 4.9: The minimum and maximum C values of a field.
4.2.5 The support practice factor (P)

The dimensionless support practice factor, P, takes into account the effect of special management practices, such as strip cropping and terraces. Because no farmer applied such soil conservation practices, the P-factor has a deterministic value of 1 for the entire study area.

4.2.6 The actual erosion (A)

Because the stochastic RUSLE parameters (LS, K and C) are mutually independent, there are no correlation terms that should be taken into account in the error propagation. To assess the actual soil loss of every field or hydrological unit and the uncertainty on these results, a Monte Carlo simulation, with 500 runs per field, was performed. The RUSLE model was represented by equations (4.6) and (4.8). Every run a new DEM was created for the field and the K and C values for every pixel in the field were randomly sampled according to their probability distribution. If the number of iterations is high enough (at least a few hundred) the actual soil loss will be normal distributed. This resulted in two maps (figure 4.10) showing the actual soil loss per field and the uncertainty of these estimates, expressed by the standard deviations.
Figure 4.10: The mean actual soil loss of a field (ton·field$^{-1}$·year$^{-1}$) and the standard deviation of these estimations.
4.3 Validation

Using the actual soil erosion map and the standard deviation map on the actual soil loss one can calculate (using a Monte Carlo simulation) the average yearly sediment input in the water reservoir and the uncertainty on that value. The calculated yearly sediment input was 4376 ton·year$^{-1}$ with a standard deviation of 75 ton·year$^{-1}$. Given the uncertainty of the model input parameters, the predicted average yearly sediment input, and its 68% confidence interval, was 4376 ± 75 ton·year$^{-1}$. In 1982, 204 141 ton sediment was dredged out of the reservoir (Gabriels, 1985). The previous dredging dated from 1936. This means that in the 46 years between the two dredging operations 204 141 ton sediment was deposited into the reservoir, which represents a yearly input of 4438 ton·year$^{-1}$. This value lies within the 68% confidence interval, the model output and field truth differed only 1.4%. It can be concluded that the RUSLE model in combination with the methods presented in this chapter, in particular the adaptation of the linear RUSLE model to a 3D reality, might be capable to predict both on-site soil losses and off-site sediment accumulation within an acceptable accuracy. Note that the RUSLE can not be used to estimate the intrabasinal sediment storage as colluvium.

4.4 Discussion and conclusions

Extending the RUSLE with an error propagation was very important in evaluating the difference between what can be predicted by a model and what exists in the field. If field measurements coincides reasonable with the model output, then the difference between the model output and the field truth can be explained mainly by the uncertainty of the model input. Otherwise the difference must be mainly due to the model itself. For example in regions where gully erosion is very prominent, the RUSLE will underestimate the soil losses and sediment accumulation. Also, an incorrect assignment of the r-value of a field in equation (4.4) can be responsible for considerable deviations between model output and field truth. However, it was not possible to estimate the amount of error induced by the model itself. This requires supplementary statistical information about the model regression equations (confidence and prediction interval equations) from the model designers. Scheinost et al. (1997) and Sinowski et al. (1997) investigated the error contribution of a model (a pedotransfer function to predict soil
water retention) and the error contribution of the spatial interpolation. Probably due to error self-compensation, the overall error was substantially smaller than the sum of both single components. The standard deviation of 75 ton-year$^{-1}$ is a fraction of only 1.7% of the total off-site sediment accumulation (4376 ton-year$^{-1}$), while the standard deviation of the soil loss per agricultural field represents a mean fraction of 14.6% (with a maximum of 54.6% on the steepest slopes) of these soil loss estimates. This indicates that, when the soil loss values are aggregated on a watershed level, it is possible to make predictions with a narrow confidence or prediction interval.

Computing sources of error can be done by pretending that all parameters have no error except for the parameters that are traced. Comparing these results with the original model output shows the (relative) error contribution of that particular parameter. Percentile error maps of each parameter can be calculated by dividing the variance by the total variance and multiplying by 100 (Spiegel and Meddis, 1982; Wesseling and Heuvelink, 1995). Figure 4.11 gives the percentile error maps for the LS, K and C factors. In the case that the uncertainty induced by the data is too large, one can select the input parameters and locations which must be sampled more precise or at a higher spatial resolution. The major error contribution is from the LS factor, indicating the need of high-quality DEM data in erosion studies. This also might indicate that the LS factor is assigned a to important influence in the RUSLE model (see also section 5.6).

This study indicates the possible power of the RUSLE model, when applied in agricultural watersheds, and when used within the boundary limits of the model. Because the calculated and measured mean yearly sediment input in the water reservoir differed only by 1.4% and gully erosion is not significant in the study area, it can be concluded that the RUSLE LS equation is also a measure of the sediment transport capacity of the overland sheet flow and small rill flow.

Modelling environmental processes is very complex, and a model — how complex it might be — is always a generalization of the real processes, and the output of a model is always the result of computations and operations on uncertain input data. Therefore, it should be encouraged to perform the model calculations using the probability distributions of the input parameters rather than their deterministic values.
Figure 4.11: Percentile error maps for the LS, K and C factor.
Chapter 5

Predicting Event-Based Sediment Transport

Parts of this chapter are published in Biesemans et al. (1999).

5.1 Introduction

The preliminary erosion risk mapping for the study area was done with the RUSLE (Revised Universal Soil Loss Equation) (chapter 4). From the results (RUSLE soil loss estimations accumulated at a watershed scale) presented in chapter 4, it can be concluded that the RUSLE might predict the long-term soil loss very accurately. However, all the erosion plot measurements used in the development of the model are measurements on hydrologically isolated areas: the plots are isolated from surrounding influences. This might have important consequences when applying the RUSLE at a watershed scale. Therefore it was decided to use a physical based model to compare the generated erosion patterns.

Furthermore, the RUSLE does not account for all possible erosion control measures (e.g.: riparian buffer zones). In order to have a uniform methodology to design alternative land management scenarios and to analyze their effects, an alternative model was required.

Many physical or process-based models are available (see chapter 3), but most are very complex and very gluttonous for data (e.g. WEPP, EUROSEM), what makes it
difficult to apply them in practice. Therefore the EROSION-2D/3D model (Schmidt, 1991, 1992, 1996) was selected, since this model needs only input about the most significant parameters governing the erosion processes. The necessary data to run the EROSION-2D/3D model is almost the same as the data set needed for the RUSLE model.

Due to recent findings in the experimental and physical description of soil detachment and soil transport (e.g.: Nearing et al., 1997) the necessity was felt to change and add some components. Some major changes were introduced in the hydrological submodel and the soil transport function. Therefore, it was decided to identify this modified EROSION-2D/3D model as STM-2D/3D (Sediment Transport Model, which is implemented in a 2D hillslope and 3D watershed version). Because the general layout of the model and the necessary data set to run the model was maintained, an overview is also given of the governing equations of the original EROSION-2D/3D model.

Because of the complexity of the erosion processes, a full quantitative validation of the STM-2D/3D physical model was not the objective. It was the intention to calibrate and validate the most important component of the model: the hydrological submodel. A reasonable prediction of the runoff contributing areas is essential for the delineation of the erosion risk areas. Also, based on laboratory rainfall simulations, a new soil transport model was developed which links the sediment concentration of the overland flow to its physical characteristics. Besides a quantitative validation of the concepts of the hydrological submodel, a qualitative evaluation of the model was possible based on the simulated and observed occurrence of deposition zones and the plausibility of the simulated erosion patterns.

5.2 EROSION-2D/3D: original model layout

5.2.1 General concept

Erosion is a complex natural phenomenon. It is therefore necessary to use a modular (object oriented) approach to model and simulate the soil loss processes. This means that the different subprocesses are described separately, where the output of one subprocess is used as input for another subprocess. In the EROSION-2D/3D model, water erosion is modelled as the result of the following subprocesses:
1. infiltration/runoff generation,

2. calculation of the momentum flux of the raindrop impact,

3. calculation of the momentum flux of the runoff,

4. calculation of the sediment detachment and transport (erosion/deposition).

Erosion occurs when the forces induced by the raindrop impact and the surface runoff exceed the cohesion of the soil particles. These forces are taken into account by calculating the momentum fluxes of the raindrop impact and overland flow. The sediment transport is limited by the transport capacity of the flow. On very cohesive surfaces, sediment transport is also limited by the erodibility of the substrate.

The factors controlling the detachment processes are the shear forces induced by the surface runoff and the raindrop impact and the reciprocal shear forces due to the cohesion of the soil particles and gravity.

### 5.2.2 Infiltration model for runoff generation

The infiltration model of the original EROSION-2D/3D model is based on the Green-Ampt concept: the wetting front is a sharp boundary dividing the soil of initial moisture content ($\theta_i$, m$^3$·m$^{-3}$) below from saturated soil with moisture content ($\theta_s$, m$^3$·m$^{-3}$) above. The basic infiltration equation is based on the Darcy equation (Schmidt, 1996):

\[
i = -K_s \cdot \frac{\Delta(\Psi_m + \Psi_g)}{x_{wf}(t)}
\]  

(5.1)

wherein, $i$ is the infiltration rate (kg·s·m$^{-3}$), $K_s$ is the saturated hydraulic conductivity (kg·s·m$^{-3}$), $\Psi_m$ the matric potential (J·kg$^{-1}$), $\Psi_g$ the gravitational potential (J·kg$^{-1}$) and $x_{wf}(t)$ the depth of the wetting front (m) at time $t$. This infiltration equation can be further simplified by assuming:

\[
\Delta \Psi_m \approx \Psi_{mi}
\]  

(5.2)

with, $\Psi_{mi}$ the matric potential at the initial water content $\theta_i$. The depth of the wetting front can then be estimated from:
\[ x_{wf} = -\left( \frac{K_s \cdot g \cdot t}{\rho_w \cdot \Delta \theta} + \sqrt{\frac{2 \cdot K_s \cdot \Psi_{mi} \cdot t}{\rho_w \cdot \Delta \theta}} \right) \] (5.3)

wherein, \( g \) is the gravitational constant (m·s\(^{-2}\)), \( \rho_w \) is the water density (kg·m\(^{-3}\)), and \( \Delta \theta = \theta_s - \theta_i \).

To apply this infiltration model in practice, the following data are necessary:

1. the saturated hydraulic conductivity
2. the water retention characteristic of the soil.

### 5.2.2.1 Estimation of the saturated hydraulic conductivity

The saturated hydraulic conductivity \( (K_s) \) can be estimated if the soil texture and soil density of the soil are known. A frequently used equation to estimate \( K_s \) is (Campbell, 1985):

\[ K_s = 0.004 \cdot \left( \frac{1.3}{\rho_s} \right)^{1.3 \cdot b} \cdot e^{-6.9 \cdot m_c - 3.7 \cdot m_s} \] (5.4)

wherein \( m_c \) is the clay (0–2 \( \mu \)m) mass fraction (kg·kg\(^{-1}\)), \( m_s \) is the silt (2–50 \( \mu \)m) mass fraction (kg·kg\(^{-1}\)), and \( \rho_s \) is the density of the soil (ton·m\(^{-3}\)). In this equation, \( K_s \) is expressed in (kg·s·m\(^{-3}\)). To convert this to (m·h\(^{-1}\)), equation (5.4) must be multiplied with 35.28. The parameter \( b \) is a function of the ‘air entry’ water potential, \( \psi_{es} \) (J·kg\(^{-1}\)), and the standard deviation of the geometric diameter of the soil particles, \( \sigma_g \), according to:

\[ b = (-2 \cdot \psi_{es}) + (0.2 \cdot \sigma_g) \] (5.5)

The air entry water potential of a certain soil can be estimated with:

\[ \psi_{es} = -0.5 \cdot (d_g)^{-0.5} \] (5.6)

where \( d_g \) is the geometric diameter of the soil particles (mm). The geometric diameter and the geometric standard deviation can be calculated using (Shirazi and Boersma, 1984):
\[
d_d = \exp \left( \sum_n F_n \cdot \ln(d_n) \right) \\
\sigma_d = \exp \left( \sqrt{\left( \sum_n F_n \cdot (\ln(d_n))^2 \right) - \left( \sum_n F_n \cdot \ln(d_n) \right)^2} \right)
\]

wherein, \( n \) is the number of textural classes, \( F_n \) is the mass fraction of textural fraction \( n \) \((\text{kg}\cdot\text{kg}^{-1})\), and \( d_n \) the mean diameter of textural fraction \( n \) \((\text{mm})\).

### 5.2.2.2 Estimation of the water retention characteristic

The knowledge of the water retention characteristic is essential to describe water movement through a soil profile. The determination of the water retention characteristic consists of measuring the soil water content at given pressure levels. A continuous curve is then fitted through the measurements. The best known and most frequently used curve, is the Van Genuchten model (Van Genuchten, 1980):

\[
\theta = \theta_s + \frac{\theta_s - \theta_r}{\left(1 + (\alpha \cdot |h|)^m\right)^n}
\]

In this equation \( \alpha \), \( n \) and \( m(=1-1/n) \) have no physical meaning and determine the general form of the curve. \( \theta_s \) is the moisture content at saturation \((\text{cm}^3\cdot\text{cm}^{-3})\), \( \theta_r \) is the residual moisture content at a pF of 4.2 \((\text{cm}^3\cdot\text{cm}^{-3})\) and \( h \) is the pressure head \((\text{cm} \text{H}_2\text{O})\). Because \( m \) and \( n \) are not independent, \( m \) can be set to equal 1, such that equation (5.9) can be simplified to a function of four parameters \( (\theta_s, \theta_r, \alpha, n) \). Rewriting equation (5.9) results in:

\[
h(\theta) = \left(\left(\frac{\theta_s - \theta_r}{\theta - \theta_r}\right) - 1\right) \cdot \frac{1}{\alpha^n}^{\frac{1}{n}}
\]

The four parameters of this equation \( (\theta_s, \theta_r, \alpha, n) \) can be estimated using the pedotransfer functions of Vereecken et al. (1989):

\[
\theta_s = 0.81 - (0.283 \cdot \rho_s) + (0.001 \cdot C\ell) \tag{5.11}
\]

\[
\theta_r = 0.015 + (0.005 \cdot C\ell) + (0.014 \cdot C) \tag{5.12}
\]
\[
\alpha = \exp(-2.486 + (0.025 \cdot Sa) - (0.351 \cdot C) - (2.617 \cdot \rho_s) - (0.023 \cdot Cl))
\]  \quad (5.13)

\[
n = \exp(0.053 - (0.009 \cdot Sa) - (0.013 \cdot C) + (0.00015 \cdot Sa^2)
\]  \quad (5.14)

wherein, \(\rho_s\) is the soil density (g\textcdot cm}^{-3}), \(Cl\) is the clay content (%), \(Sa\) is the sand content (%) and \(C\) is the organic carbon content (%). These regression equations were derived from Belgian soil types and have an adjusted R squared value of respectively: 84.8, 70.3, 68.0 and 56.0. With these equations the water retention characteristic can be estimated based on the texture, organic carbon content and the soil density.

### 5.2.3 Momentum flux of the overland flow

The hillslope, along which the net erosion/deposition is calculated, is subdivided in a number of segments. This number of segments must be large enough to describe the general form of the topography. The momentum flux generated by the overland flow, \(\phi_q\) (N\textcdot m}^{-2}), can be calculated with:

\[
\phi_q = \frac{w_q \cdot v_q}{\Delta x}
\]  \quad (5.15)

wherein \(w_q\) is the mass flux of the overland flow (kg\textcdot m}^{-1}\textcdot s}^{-1}), \(\Delta x\) the length of a slope segment (m) and \(v_q\) the mean velocity of the overland flow (m\textcdot s}^{-1}). The mass flux of the overland flow, \(w_q\), can be calculated with:

\[
w_q = q \cdot \rho_q
\]  \quad (5.16)

wherein \(q\) is the unit discharge of the overland flow (m\textsuperscript{3}\textcdot m}^{-1}\textcdot s}^{-1}) and \(\rho_q\) is the density of the fluid (kg\textcdot m}^{-3}). The discharge results from:

\[
q = q_{\text{runoff}} + q_{in}
\]  \quad (5.17)

with \(q_{\text{runoff}}\) the runoff volume for a given segment, calculated with the Green-Ampt model (see section 5.2.2), and \(q_{in}\) the inflow of runoff water from the upstream segment. The mean velocity of the overland flow can be calculated using the Manning formula:
\[ v_q = \frac{1}{n} \cdot R^{2/3} \cdot S^{1/2} \]  

(5.18)

where \( n \) is the Manning roughness coefficient (s\(\cdot\)m\(^{-1/3}\)), \( R \) is the hydraulic radius (m) and \( S \) is the slope steepness of a segment (m\(\cdot\)m\(^{-1}\)). For a broad sheet flow, the hydraulic radius can be approximated by the level of the sheet flow. This level can be calculated with:

\[ R \approx \delta = \left( \frac{q \cdot n}{S^{1/2}} \right)^{3/5} \]  

(5.19)

The determination of the Manning roughness coefficients in the original EROSION-2D/3D model is based on the Garbrecht formula (Garbrecht, 1961):

\[ n = \frac{(D_{90})^{1/6}}{26} \]  

(5.20)

where \( n \) is the Manning roughness coefficient (s\(\cdot\)m\(^{-1/3}\)) and \( D_{90} \) the particle diameter at 90 % of the cumulative soil aggregational composition (m). Table 5.1 gives an overview of some tillage practices and the corresponding Manning coefficients according to formula (5.20).

<table>
<thead>
<tr>
<th>tillage operation</th>
<th>( D_{90} ) (mm)</th>
<th>Manning coefficient (s (\cdot)m(^{-1/3}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>plowed</td>
<td>45</td>
<td>0.023</td>
</tr>
<tr>
<td></td>
<td>35</td>
<td>0.022</td>
</tr>
<tr>
<td>harrowed</td>
<td>27</td>
<td>0.021</td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>0.017</td>
</tr>
<tr>
<td>waltzed</td>
<td>4</td>
<td>0.015</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>0.013</td>
</tr>
</tbody>
</table>

Table 5.1: Manning roughness coefficients for some tillage practices according to Garbrecht (1961).
5.2.4 Momentum flux of the raindrop impact

The velocity of the raindrops depends on the rainfall characteristics. From the analysis of natural precipitation events a relationship was found between the fall velocity $v_r$, (m·s$^{-1}$) and the rainfall intensity $I_r$, (mm·h$^{-1}$), (Laws, 1941; Laws and Parsons, 1943):

$$v_r = 4.5 \cdot I_r^{0.12} \quad (5.21)$$

The mass flux of the precipitation, $w_r$ (kg·m$^{-2}$·s$^{-1}$), can then be calculated using:

$$w_r = I_\alpha \cdot \rho_{fluid} \quad (5.22)$$

wherein $\rho_{fluid}$ is the fluid density (kg·m$^{-3}$) and $I_\alpha$ is the effective rainfall intensity (mm·h$^{-1}$), according to:

$$I_\alpha = I_r \cdot \cos(\alpha) \quad (5.23)$$

with $\alpha$ the slope steepness of a slope segment (radians). The momentum flux exerted by the impact of raindrops, $\phi_r$ (N·m$^{-2}$), is then:

$$\phi_r = w_r \cdot v_r \cdot \sin(\alpha) \cdot (1 - C) \quad (5.24)$$

wherein $C$ is the soil cover by the vegetation (%).

5.2.5 Transport equation

When the erosive forces (represented by the momentum flux of the overland flow and the momentum flux of the raindrop impact) exceed the forces which counteract the erosion processes (cohesive forces between soil particles and the downward component of the turbulent flow) then erosion will occur. In the EROSION-2D/3D model this concept is represented by the following dimensionless erosion index ($E$):

$$E = \frac{\phi_q + \phi_r}{\phi_{crit}} \quad (5.25)$$
This coefficient characterizes the ability of the combined effect of surface flow and raindrop impact to detach and transport soil particles. Erosion occurs when $E > 1$. When $E \leq 1$, there is no transport of soil particles. When the rain intensity is too low to initiate runoff ($\phi_q = 0$), there will be no net erosion ($E = 0$). The net transport of soil particles by the raindrop impact is neglectable.

The critical momentum flux, $\phi_{\text{crit}}$, is the result of the forces which work against the erosion forces. This critical momentum flux can be deduced from the measurable shear stress of the soil. Experiments (Torri, 1987; Torri et al., 1987a) indicated that the critical momentum flux is in relation with the measurable shear stress of the soil ($\phi_{\text{soil}}$):

$$\phi_{\text{crit}} = 0.0001 \cdot \phi_{\text{soil}}$$  \hspace{1cm} (5.26)

with $\phi_{\text{crit}}$ and $\phi_{\text{soil}}$ expressed in Pa (= N m$^{-2}$). Torri et al. (1987b) found that the mean ratio of $\phi_{\text{crit}}/\phi_{\text{soil}}$ has only a small variation between the different agricultural soil types (0.00011 to 0.00014). The measurable shear stress might therefore be a good parameter to describe the soil erodibility.

The critical momentum flux in the original EROSION-2D/3D model is determined using an iterative method (Schmidt, 1996). The model results were compared with the soil loss measurements of the laboratory rainfall simulations. The critical momentum flux was then iteratively changed until the model results were in agreement with the measurements. Table 5.2 gives an overview of the resulting values for the critical momentum flux for some soil types (Schmidt, 1996).

Table 5.2: Critical momentum flux, $\phi_{\text{crit}}$, for some soil types (Schmidt, 1996).

<table>
<thead>
<tr>
<th>soil type</th>
<th>$\phi_{\text{crit}}$ (N m$^{-2}$)</th>
<th>number of experiments</th>
</tr>
</thead>
<tbody>
<tr>
<td>sandy soils</td>
<td>$&lt; 0.0005$</td>
<td>4</td>
</tr>
<tr>
<td>silty soils ($\geq$ 70 % silt)</td>
<td>0.0005 - 0.0025</td>
<td>21</td>
</tr>
<tr>
<td>silty soils ($&lt; 70$ % silt)</td>
<td>0.0025 - 0.0050</td>
<td>13</td>
</tr>
<tr>
<td>loamy soils</td>
<td>$&gt; 0.0050$</td>
<td>7</td>
</tr>
</tbody>
</table>

Using laboratory rainfall simulations, the following relation between the dimensionless
erosion index, $E$, and the observed sediment transport, $q_s$ (kg·m$^{-1}$·s$^{-1}$), was found (Schmidt, 1991):

$$q_s = (1.75 \cdot E - 1.75) \cdot 10^{-4} \quad (5.27)$$

### 5.2.6 Transport capacity

The factors which influence the transport capacity are the settling velocity of the soil particles and the vertical component of the turbulent overland flow. The settling velocity, $v_p$ (m·s$^{-1}$), can be calculated using the law of Stokes, according to:

$$v_p = \frac{1}{18} \cdot \frac{D^2 \cdot (\rho_p - \rho_f) \cdot g}{\eta} \quad (5.28)$$

wherein $D$ is the diameter of the soil particles (m), $\rho_p$ is the density of the soil particles (kg·m$^{-3}$), $\rho_f$ is the density of the fluid (kg·m$^{-3}$), $g$ is the gravitational acceleration (m·s$^{-2}$) and $\eta$ is the dynamical viscosity of the fluid (kg·m$^{-1}$·s$^{-1}$). Multiplying the settling velocity with the mass flux of the particles, $w_p$, results in a critical momentum flux, $\phi_{p, crit}$, of the particles in suspension:

$$\phi_{p, crit} = w_p \cdot v_p \quad (5.29)$$

When the momentum flux of the overland flow is lower than the critical momentum flux, the soil particles cannot remain in suspension. The mass flux, $w_p$, can be obtained from:

$$w_p = c \cdot \rho_p \cdot A \cdot v_p \quad (5.30)$$

wherein $c$ is the concentration of the suspended particles (m$^3$·m$^{-3}$), $\rho_p$ is the density of these particles (kg·m$^{-3}$) and $A$ is the area of a hillslope segment (m$^2$). The critical momentum flux of the suspended particles is counter-acted by the vertical component of the overland flow, $\phi_{q, cert}$, which is supposed to be a fraction, $1/\chi$, of the total momentum flux exerted by the overland flow and raindrop impact, according to:
Chapter 5: Predicting Event-Based Sediment Transport

\[ \phi_{q, vert} = \frac{1}{\chi} \cdot (\phi_q + \phi_r) \]  

(5.31)

Measurements and calculations showed that the factor $\chi$ has a constant value of approximately 1000 (Schmidt, 1991). At transport capacity, the vertical component of the momentum flux equals the critical momentum flux of the suspended particles:

\[ \phi_{p, crit} \equiv \phi_{q, vert} \]  

(5.32)

Substituting equations (5.29), (5.30) and (5.31) in equation (5.32) and rearranging, results in:

\[ c_{max} = \frac{1}{\chi} \cdot \frac{\phi_q + \phi_r}{\rho_p \cdot A \cdot v_p^2} \]  

(5.33)

wherein $c_{max}$ is the maximum concentration of soil particles at transport capacity. The transport capacity, or in other words, the maximum amount of soil particles which can be transported, can be calculated with:

\[ q_{s, max} = c_{max} \cdot \rho_p \cdot q \]  

(5.34)

### 5.2.7 Estimation of the net soil loss and deposition

To calculate the rate of erosion or deposition for each of the individual slope segments, the following equation can be used:

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

(5.35)

wherein $\gamma$ is the rate of erosion ($\gamma < 0$) or deposition ($\gamma > 0$) per unit area, $q_{s, in}$ is the sediment discharge into the segment from the segment above, $q_{s, out}$ is the sediment discharge out of the segment and $\Delta x$ is the length of the slope segment. Remark that the calculated sediment discharges, using the empirical equation (5.27), have an upper boundary limit given by equation (5.34).
5.2.8 Runoff routing

The recent versions of the EROSION-2D/3D model have no hydrological model. The estimation of overland flow discharge is based on the Manning-Strickler formula. Because the model might be used for design applications (e.g. dimensioning a sediment trapping construction), it is important to make good predictions of the hydrological properties (water level, discharge, water velocity) of the overland flow and channel flow. Therefore, this component of the EROSION-2D/3D model has to be modified (see section 5.3.3).

5.3 STM-2D/3D: modifications of the EROSION-2D/3D

During the development of the software for the 2D and 3D versions and the calibration/validation phase, there were some major changes introduced in the original EROSION-2D/3D model, which resulted in the STM-2D/3D model: (1) the original infiltration model was changed in order to predict the infiltration for unsteady rain events, (2) an implicit flow routing finite difference scheme of the kinematic wave was developed and (3) an alternative soil transport model was introduced, to eliminate a difficult to determine parameter, namely the critical momentum flux. The following sections deal with these modifications.

Flow charts of the EROSION-2D/3D model and the STM-2D/3D model are given in figure 5.1 and figure 5.2 respectively. Because the soil transport functions of the STM-2D/3D model are based on the stream power concept (section 5.3.4), wherein discharge is an important parameter, it was decided to convert the lumped runoff routing model towards a distributed model. Because the stream power based soil transport function is valid for all agricultural soil types in the Belgian loess belt, the critical momentum flux in the EROSION-2D/3D model could be avoided. The Green-Ampt infiltration model was extended to deal with unsteady rainfall events (multiple periods of ponding and non-ponding) and two additional input parameters were introduced in the infiltration process description: the vegetation interception capacity and the soil retention capacity, which are important parameters to design erosion control measures concerning vegetation and tillage management.

Because a sensitivity analysis of the EROSION-2D/3D model (Schmidt, 1996) indicated
Figure 5.1: Flow chart of the original EROSION-2D/3D model.

Figure 5.2: Flow chart of the STM-2D/3D model. Remark that if the user has information about the outlet hydrographs, this can be used to calibrate (with the initial moisture content as calibration parameter) the hydrological submodel.
that the initial moisture content was the most sensitive parameter concerning runoff production, it was preferred to use this parameter as calibration parameter in the validation phase of the STM-2D/3D model.

To validate the STM-2D/3D model for a certain area, the user needs some time series: precipitation time series and time series of the discharge at the watershed outlet. The temporal resolution of these time series must be small enough to capture the characteristics of the rainfall events and the related channel hydrographs. The runoff amounts are generated with the Green-Ampt model and are routed over the land and through the channels using a linear implicit finite difference scheme of the kinematic wave equation. Additional information for this distributed flow routing model is a digital terrain model (DTM), the channel widths, and the Manning roughness coefficients. From the DTM, the flow direction and cumulative drainage area can be determined for every pixel. In the validation phase of the model, there is a feedback loop: if the simulated peak flow does not correspond with the measured peak flow, the map of the initial moisture content is adapted using the intermediate model results, and the whole model is recalculated. If the simulated hydrograph corresponds with the measured hydrograph at the watershed outlet, the overland flow hydrographs of every pixel are used together with information about the soil cover in a soil transport function to generate a map of the net soil loss and net soil deposition.

5.3.1 Runoff generated by infiltration excess: modified Green-Ampt infiltration model for unsteady rain events

During a rainfall event, the intensity is not constant. During high intensity periods, ponding will occur. During low intensity periods, infiltration occurs without surface ponding. Consequently, during a rainfall event there are two states in the infiltration process: (1) infiltration without surface ponding, where all the rainfall penetrates in the soil profile, and (2) infiltration with surface ponding, where not all rainfall water can penetrate the soil profile, what might induce surface runoff. This model represents the infiltration excess type of runoff. However, by changing the initial moisture content of the soil (figure 5.2), this infiltration model can be forced to represent also the saturation excess type of runoff.
The original Green-Ampt equation (Green and Ampt, 1911), is formulated to describe infiltration when the soil surface is ponded. Chu (1978) made some modifications in order to adapt the Green-Ampt model for unsteady rainfall events. This author developed a method to determine the time of transition between the two stages of the infiltration process. His methodology will be described in the following paragraphs.

The water balance during a precipitation event can be formulated as follows:

\[
P(t) = F(t) + G(t) + R(t)
\]

wherein \( P \) is the cumulative rainfall (m), \( t \) is the time (h), \( F \) is the cumulative infiltration (m), \( G \) is the water level on the surface (m) and \( R \) is the rainfall excess (m) which might result in runoff. A distinction is now made between infiltration during ponding \((F_p, G > 0)\) and infiltration without \((F_u, G = 0)\) ponding. For a given soil surface there exists a maximum amount of water which can be retained by the soil surface without causing runoff. This is the retention capacity. Consequently, the level of the ponded water ranges between:

\[
0 \leq G \leq D
\]

with \( D \) the retention capacity (m). An unsteady rainfall event can be written as:

\[
i = f + \frac{dG}{dt} + r
\]

wherein \( i \) is the rainfall intensity \((m\cdot h^{-1})\), \( f \) is the infiltration rate \((m\cdot h^{-1})\), and \( r \) is the rainfall excess rate \((m\cdot h^{-1})\). There is a rainfall excess \((r > 0)\) when the ponding level exceeds the soil surface specific retention capacity and when rainfall intensity becomes larger than the infiltration rate. Mathematically, this can be written as, with \( f_p \) the infiltration capacity \((m\cdot h^{-1})\):

\[
r = i - f_p \quad \text{for } G = D, \ i > f_p
\]
\[
r = 0 \quad \text{for } G < D \text{ or } i \leq f_p
\]

The cumulative rainfall excess is by definition:
\[ R = R(t) = \int r(t) \cdot dt \]  

(5.40)

From equations (5.39) and (5.40) follows, with \( t' \) a time prior to \( t \) (h):

\[
R(t) > R(t') \quad \text{for } G = D, \quad i > f_p \\
R(t) = R(t') \quad \text{for } G < D \text{ or } i \leq f_p
\]  

(5.41)

The original Green-Ampt equation describes infiltration for the situation where the soil surface is ponded. This model gives the relation between the infiltration capacity and the associated variables. The equation can be written as (Mein and Larson, 1973):

\[
f_p = f_p(t) = \frac{dF}{dt} = K \cdot \left(1 + \frac{\Delta h_{aw} \cdot M}{F}\right)
\]  

(5.42)

where \( f_p \) is the infiltration capacity (m·h\(^{-1}\)), \( K \) is the mean hydraulic conductivity of the wetted zone (m·h\(^{-1}\)), \( \Delta h_{aw} \) is the difference in average pressure head before and after wetting (m), \( M \) is the difference in average soil moisture before and after wetting (m\(^3\)·m\(^{-3}\)), and \( F \) is the cumulative infiltration (m).

The initial conditions must now be examined. Consider the instant where ponding starts. Let the ponding time be represented by \( t_p \). Prior to the ponding time, there was a period without surface ponding (\( G = 0 \)). From equations (5.36) and (5.41), the cumulative infiltration at ponding time is:

\[
F(t_p) = P(t_p) - R(t_p) = P(t_p) - R(t') = F_0
\]  

(5.43)

with \( t' \) an instant where ponding was not occurring, thus prior to \( t_p \), and \( F_0 \) the cumulative infiltration on the time when ponding initiates. The limit condition for integration is therefore: \( F = F_0 \) for \( t = t_p \). The integration of equation (5.42) with \( t = t_p \) to \( t \) and with \( F = F_0 \) to \( F_p \) results in:

\[
\frac{F_p}{\Delta h_{aw} \cdot M} - \ln \left(1 + \frac{F_p}{\Delta h_{aw} \cdot M}\right) - \frac{F_0}{\Delta h_{aw} \cdot M} + \ln \left(1 + \frac{F_0}{\Delta h_{aw} \cdot M}\right) = K \cdot \frac{t - t_p}{\Delta h_{aw} \cdot M}
\]  

(5.44)
In order to reduce the length of equation (5.44), let the limit constant $F_0/\Delta h_{av} \cdot M - ln(1 + F_0/\Delta h_{av} \cdot M)$ be represented by $t_s$, such that:

$$\frac{F_0}{\Delta h_{av} \cdot M} - ln \left(1 + \frac{F_p}{\Delta h_{av} \cdot M}\right) = K \cdot \frac{t_s}{\Delta h_{av} \cdot M}$$  \hspace{1cm} (5.45)

Here, $t_s$, can be interpreted as a shift of time scale due to the effect of cumulative infiltration at the ponding time. This parameter is also called the ‘pseudotime’. Substituting $t_s$ in equation (5.44) results in:

$$\frac{F_p}{\Delta h_{av} \cdot M} - ln \left(1 + \frac{F_p}{\Delta h_{av} \cdot M}\right) = K \cdot \frac{t - t_p + t_s}{\Delta h_{av} \cdot M}$$  \hspace{1cm} (5.46)

This equation (Chu, 1978) is the modified version of the traditional Green-Ampt equation. However, equation (5.46) is still not applicable in practice unless the time parameters $t_p$ and $t_s$ can be determined during an unsteady rain. At the ponding time $t_p$, the rainfall intensity equals the infiltration capacity, so that:

$$i(t_p) = f_p$$  \hspace{1cm} (5.47)

Substituting equations (5.43) and (5.47) in equation (5.42), results in:

$$P(t_p) - R(t') - \frac{K \cdot \Delta h_{av} \cdot M}{i(t_p) - K} = 0 \hspace{1cm} for \hspace{0.5cm} i > K$$  \hspace{1cm} (5.48)

Equation (5.48) is the basic formula used to determine the ponding time $t_p$. This equation is an implicit function in $t_p$, thus demanding an iterative technique to solve for $t_p$. Once $t_p$ is determined, the pseudotime can be calculated by combining equations (5.43) and (5.45):

$$\frac{K \cdot t_s}{\Delta h_{av} \cdot M} = \frac{P(t_p) - R(t')}{\Delta h_{av} \cdot M} - ln \left(1 + \frac{P(t_p) - R(t')}{\Delta h_{av} \cdot M}\right)$$  \hspace{1cm} (5.49)

With these equations, the two time parameters $t_p$ and $t_s$ can be determined for an unsteady rainfall event. For the special case, where the total duration of the event is subdivided in a number of short periods where the rainfall intensity can be supposed to be constant (e.g. when the precipitation is measured using a tipping bucket pluviograph), we can write:
\[ i(t) = \frac{P(t_n) - P(t_{n-1})}{t_n - t_{n-1}} = I = \text{constant} \]  

wherin \( n \) is a time period indicating index, \( t_n \) is the terminal time of period \( n \), \( t_{n-1} \) is the initial time of of period \( n \) and \( I \) is the constant rain intensity within period \( n \). The cumulative precipitation falling within period \( n \) is then:

\[ P(t) = \int_{t_{n-1}}^{t} i(t) \cdot dt = P(t_{n-1}) + (t - t_{n-1}) \cdot I \]  

When combining equations (5.48), (5.50) and (5.51), and for \( t = t_p \) and \( t' = t_{n-1} \), one can obtain the following equation for the ponding time:

\[ t_p = \left( \frac{K \cdot \Delta h_{ac} \cdot M}{I - K} - \frac{P(t_{n-1}) + R(t_{n-1})}{I} \right) + t_{n-1} \quad \text{for} \quad I > K \]  

Equation (5.52) is an explicit function in \( t_p \) and therefore more appropriate to use in practice than equation (5.48). The infiltration process during an unsteady rainfall event, can be summarized as follows:

1. **no ponding for a period from \( t' \) to \( t \), with:**

   \[
   R(t) = R(t') \\
   F(t) = F_u(t) = P(t) - R(t') \\
   f(t) = i(t) < f_p \\
   r(t) = 0
   \]

2. **ponding for a period from \( t' \) to \( t \), with:**

   \[ F(t) = F_p(t) \]

wherein \( F_p \) is described by the implicit function:

\[ \frac{F_p}{\Delta h_{ac} \cdot M} - \ln \left( 1 + \frac{F_p}{\Delta h_{ac} \cdot M} \right) = K \cdot \frac{t - t_p + t_s}{\Delta h_{ac} \cdot M} \]

and \( t_p \) can be determined using equation (5.48) or (5.52) and \( t_s \) with (5.49).
However, a precipitation event can feature multiple periods of high and low intensity. During the periods of high intensity surface ponding occurs, and disappears during the periods of low intensity. Consequently, there can be more than one ponding time during such an event. Each time when the surface ponding begins, the parametric values of \( t_p \) and \( t_s \) have to be calculated. Therefore, two surface condition indicators are introduced, which help the determination of \( t_p \) and \( t_s \). Suppose a short period where there is no ponding at the beginning of the period. When there is still no ponding at the terminal time of this period, then by definition and by equation (5.39):

\[
G(t_n) = G(t_{n-1}) = 0 \\
R(t_n) = R(t_{n-1})
\]

and from equation (5.36):

\[
I < f_p = K \cdot \left(1 + \frac{\Delta h_{av} \cdot M}{F(t_n)}\right)
\]

or:

\[
F(t_n) < \frac{K \cdot \Delta h_{av} \cdot M}{I - K} \quad \text{for } I > K
\]

Substituting equations (5.56) and (5.58) in equation (5.36) results in:

\[
C_u = P(t_n) - R(t_{n-1}) - \frac{K \cdot \Delta h_{av} \cdot M}{I - K} < 0 \quad \text{for } I > K
\]

wherein \( C_u \) is defined as the ‘surface condition indicator’ when there is no surface ponding at the initial time. If surface ponding occurs at the terminal time, then:

\[
G(t_n) > 0 \\
R(t_n) \geq R(t_{n-1}) \\
F(t_n) = F_p
\]

and:

\[
I < f_p = K \cdot \left(1 + \frac{\Delta h_{av} \cdot M}{F_p}\right)
\]
or:

\[ F(t_n) > \frac{K \cdot \Delta h_{\text{ave}} \cdot M}{I - K} \]  

(5.62)

Substituting equations (5.60) and (5.62) in equation (5.36) results in:

\[ C_u = P(t_n) - R(t_{n-1}) - \frac{K \cdot \Delta h_{\text{ave}} \cdot M}{I - K} > 0 \quad \text{for } I > K \]  

(5.63)

Likewise, consider a short period when there is surface ponding at the initial time. If surface ponding also occurs at the terminal time, then equations (5.60) are valid for this situation as well. Substitute these equations into (5.36) to obtain:

\[ C_p = P(t_n) - F_p(t_n) - R(t_{n-1}) > 0 \]  

(5.64)

wherein \( C_p \) is defined as the surface condition indicator when there is surface ponding at the initial time. If there is no surface ponding at the terminal time, then by definition:

\[
\begin{align*}
G(t_n) &= 0 \\
F(t_n) &< F_p(t_n) \\
R(t_n) &= R(t_{n-1})
\end{align*}
\]

(5.65)

Substitute these equations into equation (5.36) to obtain:

\[ C_p = P(t_n) - F_p(t_n) - R(t_{n-1}) < 0 \]  

(5.66)

In summary, by determining the sign of the surface condition indicators, it can be determined when the soil surface is ponded or not.

To apply the Green-Ampt infiltration model in practice, the following data is necessary:

1. the initial moisture content of the soil
2. the unsaturated hydraulic conductivity
3. the pressure head at the wetting front
4. the bulk density
5. the retention capacity

6. the saturated hydraulic conductivity (see section 5.2.2.1)

7. the water retention characteristic of the soil (see section 5.2.2.2)

With this information the parameters of the Green-Ampt model can be determined: the mean hydraulic conductivity (K) of the wetted zone, the difference in mean pressure head before and after wetting ($\Delta h_{\text{aw}}$) and the difference in mean moisture content before and after wetting (M).

The problem of data assessment for regional modelling applications is a limiting factor for the applicability of a model. For example, the determination of the water retention characteristic is a very time consuming task, especially if this characteristic must be assessed for areas of a multiple thousand hectares (regional scale in Belgium). Therefore, pedotransfer functions have a very important role in environmental studies based on models. Based on the basic soil parameters (texture, density and organic material), the necessary parameters of the Green-Ampt model can be estimated.

### 5.3.1.1 Estimation of the initial moisture content

If the STM-2D/3D model will be used for design properties (e.g. the determination of the location for a possible sediment trap in a river or brook and the necessary dimensions of the sediment trap) the initial moisture content can be set to any value between the residual and saturated moisture content of a given soil type. The residual and saturated moisture values can be calculated using equations (5.11) and (5.12).

For the calibration/validation of the STM-2D/3D model, a good estimation of the initial moisture content is indispensable. However, it is an almost impossible task to monitor the soil moisture content continuously at a watershed scale (remote sensing techniques (Verhoeest et al., 1998) may provide a solution). This because of the large spatial and temporal variability. Therefore, the initial moisture content was the (only) calibration parameter during the validation phase of the hydrological component of the STM-2D/3D model (see figure 5.2).

The first iteration of an event simulation, the initial moisture content, $\theta$, was set to the following value, with $n$ the iteration counter:
\[ \theta_{i,n+1} = \theta_r + \frac{\left( \theta_s - \theta_r \right)}{2} . \]  

(5.67)

If no correlation was obtained between the measured and calculated hydrograph at the watershed outlet, the initial moisture content was changed. If the calculated peak flow was lower than the observed peak flow, the following equations are used:

\[ \theta_v = \theta_{i,n-1} \]
\[ \theta_{i,n} = \theta_r + \frac{\left( \theta_s - \theta_r \right)}{2} . \]  

(5.68)

And in the case that the calculated peak flow was higher than the observed one:

\[ \theta_s = \theta_{i,n-1} \]
\[ \theta_{i,n} = \theta_r + \frac{\left( \theta_s - \theta_r \right)}{2} . \]  

(5.69)

### 5.3.1.2 Estimation of the unsaturated hydraulic conductivity

When the soil desaturates, the driving force for water flow becomes the gradient in matric and gravitational potential. As the largest pores in the soil drain, the hydraulic conductivity reduces rapidly. This leads to an apparent paradox where increasing the driving force for flow across a soil column by lowering the water potential on one end, may actually decrease water flow. This is because the hydraulic conductivity is reduced more than the driving force is increased. The narrow capillary tubes conduct water much slower and the tortuosity of the capillars is also much higher than the largest soil pores. Therefore, the time and distance the soil water has to travel for a unit depth displacement becomes larger the smaller the capillary tubes. The unsaturated hydraulic conductivity, \( K(\theta) \), can be estimated using the following empirical equation (Campbell, 1985):

\[ K(\theta) = K_s \cdot \left( \frac{\theta}{\theta_s} \right)^m \]  

(5.70)

\[ m = 2 \cdot b + 3 \]  

(5.71)

wherein \( m \) is a ‘pore interaction’ term and \( b \) a parameter which can be calculated using equation (5.5). The equations (5.5) and (5.6) indicate that soils with a high clay
content, by their small geometric diameter, have lower (more negative) values for the air entry potential and consequently, have higher values for the parameter b compared to soils with a coarser texture. The unsaturated hydraulic conductivity decreases more rapidly with decreasing moisture contents in a sandy soil than in a clay soil.

5.3.1.3 Estimation of the pressure head at the wetting front

Using equations (5.10) and (5.70), a relation can be constructed between the pressure head and the relative conductivity. The latter is needed to calculate the mean pressure head at the wetting front. The relative conductivity can be defined as (Mein and Farrell, 1974):

$$K_r = \frac{K}{K_s} \quad (5.72)$$

The wetting front is the transition zone between the upper soil volume with a saturated moisture content (and a zero matric potential) and the lower soil volume at a certain initial moisture content (with a much higher matric potential). The difference in average pressure head, $\Delta h_{av}$, can be estimated by integrating over the moisture range $\theta_i - \theta_s$ (with $\theta_i$, the initial moisture content). This is equivalent with the area under the $h - K_r$ curve between $K_r = 0$ and $K_r = 1$, according to:

$$\Delta h_{av} = \int_0^1 h \cdot dK_r \quad (5.73)$$

Because the pressure head becomes very high if $K_r$ is close to zero, in practice the mean pressure head is calculated over the interval 0.01–1 (Mein and Farrell, 1974). Figure 5.3 gives an example of a $h - K_r$ relation for a soil with 12% clay, 76% silt, 12% sand, 1% organic carbon and a density of 1400 kg·m$^{-3}$.

5.3.1.4 Estimation of the soil bulk density

Soil bulk density, $\rho_{soil}$, is a parameter almost always required in soil related models. Direct measurements of $\rho_{soil}$ are based on undisturbed samples, what makes these measurements labor-intensive and time-consuming. Therefore, soil bulk density is not
Figure 5.3: The matric potential in function of the relative conductivity for a soil with 12% clay, 76% silt, 12% sand, 1% organic carbon and a density of 1400 kg·m$^{-3}$.

routinely analyzed and is not a standard parameter in every soil database.

From the soil database generated during the Belgian soil survey (Van Orshoven et al., 1988), the bulk density of soils in northern Belgium can be estimated using two pedotransfer functions published in Van Hove (1969) and Van Ruymbeke et al. (1991). Van Ruymbeke et al. (1991) estimated $\rho_{soil}$ as being 1.45 g·cm$^{-3}$ for topsoil horizons and 1.50 g·cm$^{-3}$ for underlying horizons. The pedotransfer function of Van Hove (1969) is more refined and based on the Belgian soil textural triangle (see tabel 5.3).

Table 5.3: Pedotransfer function of Van Hove (1969) for the soil bulk density ($\rho_{soil}$). Classification according to the Belgian textural triangle.

<table>
<thead>
<tr>
<th>soil textural type</th>
<th>textural symbol</th>
<th>$\rho_{soil}$ (g · cm$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>Z</td>
<td>1.37</td>
</tr>
<tr>
<td>Clayey and silty sand</td>
<td>S</td>
<td>1.42</td>
</tr>
<tr>
<td>Light sandy silt</td>
<td>P</td>
<td>1.47</td>
</tr>
<tr>
<td>Sandy silt</td>
<td>L</td>
<td>1.44</td>
</tr>
<tr>
<td>Silt</td>
<td>A</td>
<td>1.43</td>
</tr>
<tr>
<td>Clay</td>
<td>E</td>
<td>1.34</td>
</tr>
<tr>
<td>Heavy clay</td>
<td>U</td>
<td>1.20</td>
</tr>
</tbody>
</table>
Boucneau et al. (1998) analyzed several pedotransfer functions for estimating the soil bulk density: the functions of Van Hove (1969), Van Ruymbeke et al. (1991), Rawls (1983) and Manrique and Jones (1991). Using the Belgian soil database Boucneau et al. (1998) found that the model of Manrique and Jones (1991) had the best overall result. This pedotransfer function can be written as:

\[
\begin{align*}
\rho_{\text{soil}} &= 1.704 - 0.005 \cdot CI - 0.127 \cdot OC \quad \text{for Alfisols} \\
\rho_{\text{soil}} &= 1.861 - 0.452 \cdot \sqrt{OC} \quad \text{for Entisols} \\
\rho_{\text{soil}} &= 1.720 - 0.403 \cdot \sqrt{OC} \quad \text{for Inceptisols} \\
\rho_{\text{soil}} &= 1.587 - 0.259 \cdot \sqrt{OC} \quad \text{for Spodosols}
\end{align*}
\]  

(5.74)

with \( CI \) the clay percentage and \( OC \) the organic carbon percentage. If the organic carbon content is not known, the pedotransfer function of Van Hove (1969) can be used.

### 5.3.1.5 Estimating the retention capacity

The retention capacity is an important parameter for the determination of the amount of effective runoff. If the soil surface is not completely flat, not all excess rainfall might produce runoff. A certain amount of excess rainfall is captured in the microtopographical features (the random roughness).

The determination of the retention capacity for the STM-2D/3D model is based on two easy to determine parameters: the random roughness height, and the Hurst coefficient. If this Hurst coefficient is lower than 0.5 there is a negative autocorrelation, for a Hurst coefficient greater that 0.5 there is a positive correlation and a Hurst coefficient equal to 0.5 indicates a Brownian process (for a more detailed outline about the Hurst coefficient and the related fractal dimension see section 4.2.1.1).

For a freshly tilled soil surface the Hurst coefficient will be close to 0.2 – 0.3. The longer a surface is exposed to raindrop impact the smoother it will become and thus the higher the Hurst coefficient will be. Using the random midpoint displacement method (Peitgen et al., 1992) a random surface can be generated to simulate the random roughness with the same spatial autocorrelation as indicated by the Hurst coefficient. For every pixel of the digital terrain model (DTM) a random surface with a very high spatial resolution...
can be calculated, with an overall slope equal to the pixel slope. It is not necessary that the random surface is as large as a pixel of the DTM, an area of 1 m\(^2\) is usually sufficient to capture the microtopographical features. Using these random surfaces it is possible to determine the amount of water that can be retained. Figure 5.4 gives an example (a numerical simulation) of the retention capacity in function of the Hurst coefficient for a horizontal surface. The steeper the slopes, the less water that will be retained.

![Graph showing retention capacity vs Hurst coefficient](image)

Figure 5.4: The retention capacity in function of the Hurst coefficient, for a non-sloping surface.

5.3.2 Runoff generated by saturation excess

Field observations demonstrated that Hortonian surface flow, or infiltration excess, is rather rare in humid regions. In these regions, the infiltration capacity of the soil exceeds the observed rainfall intensities for most rainfalls (Chow et al., 1988; Troch et al., 1994). Hursh and Brater (1944) doubted the concept of overland flow as defined by Horton (1933), however, while surface storm runoff was not observed during rainfall, the characteristic flood hydrographs had still the shape as produced by heavy rain. Betson (1964) put forward the “partial area concept”, denoting that infiltration excess
runoff occurs from a relative small part of the catchment area. By analyzing a series of rainfall events, Ragan (1968) showed that only a small portion of the study catchment ever contributed to the storm hydrograph.

From the work of Hewlett and Hibbert (1967) the concept of subsurface storm flow was created. However, subsurface flow velocities are normally so low that subsurface flow alone cannot contribute a significant amount of storm precipitation directly to streamflow (Chow et al., 1988).

A second mechanism, derived from field observations, is the saturation overland flow. Saturation overland flow differs from Hortonian overland flow in that in Hortonian overland flow the soil is saturated from above by infiltration, while in saturation overland flow it is saturated from below by subsurface flow (Chow et al., 1988). Dunne et al. (1975) suggested that the partial contributing areas were likely to be wetlands with locations controlled by the topographic and hydrogeological configuration of the basin. These contributing areas are likely to be located adjacent to stream channels, in swamp areas and on shallow soils and expand during a rainfall event and contract thereafter. This dynamical character earned them the term “variable source areas”.

The TOPMODEL (Beven and Kirkby, 1979; Beven, 1997) is the best known hydrological model build around the concept of variable source areas. In this model, total runoff is calculated as the sum of two flow components: saturation excess overland flow from variable contributing areas and subsurface flow from the saturated zone of the soil. For the saturated zone, it is assumed that: (1) its dynamics can be represented by a series of steady states, (2) its hydraulic gradient can be approximated by the tangent of the local slope of the ground surface (tanβ). On the basis of these two assumptions a relationship between mean storage deficit, \( \overline{D} (m) \), of a catchment and the local storage deficit, \( D_i (m) \), at any point within the catchment can be derived:

\[
D_i = f(\overline{D}, T_{bi}, I_i, \overline{I}, m) \tag{5.75}
\]

wherein \( T_{bi} \) (m²·h⁻¹) is the local lateral transmissivity when the saturated zones reaches the ground surface, \( I_i \) and \( \overline{I} \) are the local and the catchment averaged value of the topographic index \( a/tan\beta \) (Kirkby, 1975), where \( a \) (m) denotes the upslope contributing area per unit contour length. The parameter \( m \) (m) defines the variation of saturated
hydraulic conductivity with depth. Different formulations presented by Ambroise et al. (1996) can be selected in this model (linear or exponential decline with soil depth or constant value with step function decline to zero). Areas with $D_t \leq 0$ are contributing areas for saturation excess overland flow. Precipitation or snowmelt on saturated areas contributes completely and immediately to runoff.

Literature reviews (Moore and Thompson, 1996; Coles et al., 1997; Güntner et al., 1999) about the TOPMODEL indicate that the delineation of the saturated zones based on the topographic index alone does not capture the complete picture concerning the infiltration-runoff process. These authors conclude that soil spatial variability and certain geological and pedological features (fractures, strata boundaries) can have an important effect on the occurrence of saturated areas. Therefore, the topographic index alone will not be the ideal parameter to predict the runoff contributing areas.

Also, from field observations in the test watershed, the concept of the topographic index to determine the saturated areas can be questioned. Within the test watershed, there are no saturated zones after a long dry period in the summer months. After such period, the moisture content of the upper soil horizon will be more related to its texture and organic carbon content. In the winter period, the topsoil of larger parts of the catchment are saturated or close to saturation due to the almost omnipresent plough pans. This can be illustrated with figures 5.5 and 5.6. The (gravimetric) moisture profiles of locations M1, M2, M3, M4, M5 and M7 are positioned very close to the Kemmelbeek brook, locations M6, M8 and M9 closer to the water divide. The moisture profiles were measured on 7 January 2000 and the previous rainfall event was on 5 January 2000. All moisture profiles were measured between the wheel tracks. The moisture content of the topsoil (0–15 cm or 0–30 cm, depending on the depth of the plough layer) was for most profiles (6 profiles of 9) still higher than that of the soil just below the plough layer (15–30 cm or 30–45 cm) (M1, M2, M3, M5, M7, M8). For the 6 profiles close to the brook, 5 had higher moisture contents in the topsoil, indicating a saturation from above (rainfall), instead of saturation from below (ground water).
Figure 5.5: Map of the position of the measured soil moisture profiles (7 January 2000).
Figure 5.6: Soil moisture profiles at 7 January 2000. The last rain event was on 5 January 2000.
Surface sealing increases the probability of runoff and thus soil erosion, and occurs on a variety of soils worldwide (Ewing and Gupta, 1994). Filtration (washing-in) and compaction have been reported as mechanisms of surface sealing. Ewing and Gupta (1994) showed that filtration was more effective than compaction in reducing soil hydraulic conductivity during seal formation, even though compaction reduced seal porosity more than filtration. Most of the particles available for filtration were deposited on the soil-pore network surface rather than being washed into the soil-pore network. This suggests that surface deposition (runoff deposition, splash deposition) may be more important than compaction or filtration in determining the seal hydraulic conductivity.

It is clear that the physical description of infiltration/runoff processes is a very complex problem given the large spatial and temporal variability of soil physical properties. Figure 5.7, a photograph of an agricultural field inside the Kemmelbeek watershed taken in a dry period of September 1997, shows a complex color pattern. The soil moisture content of the darker patches is higher than the lighter patches. The most important reason for this pattern is a textural differentiation from sandy loam (lighter patches) to silty loam (darker patches). Locally the thin Quaternary silty surface layer is mixed by tillage with the underlying Tertiary sands. These spots can also be recognized by the occurrence of silex pebbles at the surface.

To ensure the applicability of the model, a strong generalization of the infiltration/runoff processes is a necessity (to reduce the amount of model input data). Therefore, it was decided to estimate the initial moisture content based only on the most important soil physical characteristics, namely soil texture and soil organic material. By applying the methodology presented in section 5.3.1.1, the general soil related and seasonal pattern of initial moisture content could be simulated (see section 5.4): in the summer months, when rainfall intensities are in general high to very high, infiltration excess is the dominant simulated infiltration process, and in the winter months characterized with long lasting rains with low intensities, saturation excess is the dominant simulated infiltration process. However, the ‘saturation excess’ concept must be extended. On fields with plough/traffic pans and surface seals, the soil is not only saturated from below, but also from above: the thin compacted plough layer or sealed surface layer blocks the infiltration what can result in an almost saturated topsoil. Be-
cause plough/traffic layers are almost omnipresent in Belgian agricultural watersheds, there was no need to create an additional data layer. It was assumed that the maps of soil texture and soil organic material provide enough information to capture the spatial variability in runoff production.

5.3.3 Runoff routing model

The chosen runoff routing model is based on a simplification of the Saint-Venant equations (see section 3.4.1.2), namely the kinematic wave equation. This methodology enables to calculate the surface water level as a function of time and space.

The Saint-Venant equations (de Saint-Venant, 1871) have various simplified forms, each defining a one dimensional distributed routing model. The momentum equation consists of terms for the physical processes that govern the flow momentum. These terms are: (1) a local acceleration term, which describes the change in momentum due
to the change in velocity over time, (2) a convective acceleration term, which describes the change in momentum due to change in velocity along the channel, (3) a pressure force term, proportional to the change in water depth along the channel, (4) a gravity force term, proportional to the bed slope $S_0$, and (5) a friction force term, proportional to the friction slope $S_f$. The local and convective acceleration terms represent the effect of inertial forces on the flow.

When the water level or flow rate changes at a particular point in a channel carrying a subcritical flow, the effects of these changes propagate back upstream. These backwater effects can be incorporated into distributed routing methods through the local acceleration, convective acceleration, and pressure terms. Lumped routing models may not perform well in simulating the flow conditions when backwater effects are significant and the river slope is mild, because these methods have no hydraulic mechanisms to describe upstream propagation of changes in flow momentum.

The simplest distributed model is the kinematic wave model, which neglects the local acceleration, convective acceleration and pressure terms in the momentum equation. It assumes $S_0 = S_f$ and the friction and gravity forces balance each other. The diffusion wave model neglects the local and convective acceleration terms but incorporates the pressure term. The dynamic wave model considers all the acceleration and pressure terms in the momentum equation. Kinematic waves govern flow when inertial and pressure forces are not important. Dynamic waves govern flow when these forces are important, such as in the movement of a large flood wave in a wide river. Because the mean slope in the test watershed (section 5.4) is 7.5 %, the kinematic wave approximation could be applied for describing the overland flow properties. For a kinematic wave motion, the Saint-Venant equations reduce to (Chow et al., 1988):

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = q \quad \text{continuity equation} \tag{5.76}$$

$$S_0 = S_f \quad \text{momentum equation} \tag{5.77}$$

wherein $S_f$ is the friction slope and assumed to be equal to the bed slope, $S_0$ (m$\cdot$m$^{-1}$), $q$ is the lateral inflow in a channel segment per unit width (m$^3$·s$^{-1}$·m$^{-1}$), $A$ is crosssectional flow area (m$^2$) and $Q$ is the discharge (m$^3$·s$^{-1}$). The momentum equation can also be represented by (Chow et al., 1988):

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\[ A = \alpha \cdot Q^\beta. \]  

(5.78)

Based on the Manning equation, with \( S_f = S_0 \) and \( R = A/P \) (R is the hydraulic radius and P the wetted perimeter) the following relation can be found:

\[ Q = \frac{\sqrt{S_0}}{n \cdot P^{2/3}} \cdot A^{5/3} \]  

(5.79)

with \( n \) the Manning roughness coefficient (\( \text{m}^{-1/3} \)). This last equation can be solved for \( A \):

\[ A = \left( \frac{n \cdot P^{2/3}}{\sqrt{S_0}} \right)^{3/5} \cdot Q^{3/5}. \]  

(5.80)

The parameters \( \alpha \) and \( \beta \) in equation (5.78) are then:

\[ \alpha = \left( \frac{n \cdot P^{2/3}}{\sqrt{S_0}} \right)^{0.6} \]  

(5.81)

\[ \beta = 3/5 = 0.6 \]  

(5.82)

For a broad overland flow, \( P \) equals the hillslope segment width, \( B \). Equation (5.76) has two dependent parameters: \( A \) and \( Q \). \( A \) can be eliminated by differentiating equation (5.78) and by substituting the term \( \partial A/\partial t \) in equation (5.76). This results in:

\[ \frac{\partial Q}{\partial x} + \alpha \cdot \beta \cdot Q^{\beta-1} \cdot \frac{\partial Q}{\partial t} = q \]  

(5.83)

In order to calculate the net erosion and deposition along a hillslope, it is necessary to know the hydrographs of every hillslope segment. To determine these hydrographs, equation (5.83) can be solved using a finite difference numerical method (figure 5.8). The objective of the numerical solution is to solve equation (5.83) for \( Q(x, t) \) at each point on the \( x-t \) grid, given the channel parameters \( \alpha, \beta \), the lateral inflow \( q(t) \) and the initial and boundary conditions. In particular, the purpose of the solution is to determine the outflow hydrograph of a certain hillslope segment.
The numerical solution of the kinematic wave equation is more flexible than the analytical, it is more easy to handle variations in the channel properties and in the initial and boundary conditions, and serves as an introduction to numerical solution of the dynamic wave equation (which can be implemented in future versions of the STM-2D/3D model). Equation (5.83) can be written as, with $i$ and $j$ variables indicating respectively the hillslope or channel segment and the time step:

$$
\frac{Q_{i+1}^j - Q_{i}^j}{\Delta x} + \alpha \beta \left( \frac{Q_{i+1}^j + Q_{i}^j}{2} \right)^{\beta - 1} \left( \frac{Q_{i+1}^j - Q_{i}^j}{\Delta t} \right) = \frac{q_{i+1}^j + q_{i}^j}{2} \tag{5.84}
$$

![Diagram showing implicit linear finite difference scheme for solution of the kinematic wave equation](image)

Figure 5.8: Implicit linear finite difference scheme for solution of the kinematic wave equation (Chow et al., 1988).

Equation (5.84) can now be solved for the unknown $Q_{i+1}^j$, with $\Delta x$ the length of a segment $i$ (m):
\[ Q_{i+1}^{j+1} = \frac{\Delta t}{\Delta x_i} Q_{i+1}^{j+1} + \alpha \beta Q_{i+1}^{j} \left( \frac{Q_{i+1}^{j+1} + Q_{i+1}^{j}}{2} \right)^{\beta - 1} + \Delta t \left( \frac{Q_{i+1}^{j+1} + Q_{i+1}^{j}}{2} \right) \]

(5.85)

To solve equation (5.85) three boundary conditions must be known: (1) the maximum length of a time step to ensure accurate calculations (applies only to explicit schemes), (2) the upstream boundary condition \( Q(0, t) \) and (3) the lateral inflow \( q(t) \).

For a given precipitation event and soil parameters, the Green-Ampt runoff volume, \( Q(GA) \), can be calculated. The water level, \( h \) (m), and the wave celerity, \( Ck \) (m·s\(^{-1}\)), can be calculated using the following equations:

\[ h_i^j = \left( \frac{n_i \cdot Q(GA)_i^j}{\sqrt{S_{0i} \cdot B_i}} \right)^{3/5} \]

(5.86)

\[ Ck_i^j = \left( \frac{5}{3} \right) \cdot \left( \frac{\sqrt{S_{0i}}}{n_i} \right) \cdot \left( h_i^j \right)^{2/3} \]

(5.87)

wherein \( B \) is the width of the flow (m), and in the case of a broad overland flow (sheet flow) \( B \) equals the width of a segment.

A necessary condition for the stability of the numerical algorithms is that the time step of an explicit finite difference routine must be small enough (the Courant condition), according to (Courant and Friedrichs, 1948):

\[ \Delta t \leq \text{minimum} \left( \frac{\Delta x_i}{Ck_i^j} \right). \]

(5.88)

This condition requires that the time step is smaller than the time for a wave to travel the distance \( \Delta x_i \). If \( \Delta t \) is so large that the Courant condition is not satisfied, then there is an accumulation or piling up of water. The Courant condition does not apply for implicit schemes, like the one shown in figure 5.8. Figure 5.9 shows two hydrographs at the outlet of a straight rectangular shaped channel with a uniform slope of 5%, and 10 segments of each 30 meter. The hydrograph at the upper boundary is known and the hydrographs are calculated using an implicit linear finite difference scheme of the kinematic wave, with time steps of 76 s (the Courant condition) and 300 s (a timestep 4 times larger than the Courant condition). From this figure, it can be concluded that the general shape of the outlet hydrograph is not dependent on the time step of the
calculations in an implicit scheme. However, the larger the time step, the less accurate the peak discharge is determined. Therefore, it is advisable not to chose the time step too large. Therefore, the Courant condition was also taken into account in the implementation of the implicit finite difference scheme.

![Courant condition](image)

Figure 5.9: Two hydrographs at the channel outlet for a given upstream wave. The channel has a uniform slope of 5 % and consists of 10 segments, each 30 m long. The hydrographs were calculated using a linear implicit finite difference scheme of the kinematic wave equation. The time step of the calculations was 76 s (the Courant condition) and 300 s.

### 5.3.3.1 Boundary conditions for the overland flow

The upstream (at segment \( i = 1 \)) boundary condition for an overland runoff flowline can be written as:

\[
Q_i^t = Q(GA)_i^t
\]  
(5.89)

with \( Q(GA) \) the runoff amount determined with the Green-Ampt infiltration model. The lateral inflow can then be quantified with:
\[ q_i^j = \frac{Q(GA)_i^j}{\Delta x_i} \]  

(5.90)

The methodology to calculate the overland flow and the resulting hydrographs for every terrain element in a 3D application is shown in figure 5.10. Suppose a rectangular parcel with a total length and width of 80 m. In a DEM with a resolution of 10 m this results in a grid of 8 rows and 8 columns. Figure 5.10A displays the drainage directions, which are used to create linear flow paths over the land (figure 5.10B). The flowline matrix, given in figure 5.10C, is indicative for the number of flowpaths running through the pixels of the DEM. The respective flow widths, \( B \), used in equations (5.81), (5.85) and (5.86) can then be calculated, with \( \Delta y_i \), the width of a pixel:

\[ B_i = \frac{\Delta y_i}{FLM_i} \]  

(5.91)

---

**Figure 5.10:** Methodology to calculate the kinematic wave hydrographs over a 3D terrain. Figure A displays the drainage directions of each cell in the DEM, figure B shows the resulting flow lines and figure C displays the flow line matrix (FLM).

wherein, \( FLM_i \) is the flow line matrix value (see figure 5.10) of the \( i \)th flow line segment. The net erosion and net deposition is then calculated along every overland flow path. If there are multiple flow lines running through a given pixel, the net sediment inflow and outflow is summed up in order to obtain the erosion and deposition amounts per pixel.
5.3.3.2 Boundary conditions for the channel flow

This section only applies to the 3D version of the STM-2D/3D model. Before the channel flow can be calculated, the complete overland flow must be processed. The outflow hydrographs of these overland pixels which drain into a channel are stored into memory (this can be a huge amount of data for high resolution DEM's and a small time step). Using this data the time dependent lateral inflow into the channel, \( q(x, t) \), can be determined for all channel pixels.

Another problem is the construction of the upstream boundary condition, \( Q(1, t) \), for every first order channel section. Therefore the channel network is first classified according to the Shreve (1967) system (see figure 5.11). Every channel segment is assigned an order-number which is also the order of the calculation process (first the flow is routed through all first order segments, then through all second order elements, etc ...). If the hydrograph at the watershed outlet is known, then \( Q(1, t) \) can be calculated by dividing the discharge value at the watershed outlet at the beginning of the precipitation event (= base flow) by the number of first order channel segments. This base flow discharge is kept constant during the entire event.

![Diagram of a channel network](image)

Figure 5.11: Classification of a channel network according to the Shreve system.
5.3.3.3 Estimating the Manning roughness coefficient

In most hydrological modelling applications, the Manning coefficients are kept constant during the entire simulation. However, field experiments indicate that the Manning roughness coefficients vary with changing flow properties. Gilley and Finkner (1991) found that the hydraulic roughness coefficients rapidly decrease when the flow rate increases, even for those tillage treatments with relative large random roughness values. Once roughness elements are submerged, their influence on overland flow is less pronounced as the depth of overland flow becomes greater.

On recently cultivated soils, the type of cultivation can induce a large variation in the Manning roughness coefficients. Gilley and Finkner (1991) found for a moldboard plow and a planter operation, Manning roughness coefficient ranges of respectively 0.1–0.6 and 0.01–0.08 (s·m$^{-1/3}$). However, the random roughness microtopographical structures are rapidly destroyed by rainfall. This induces a high spatial and temporal variation in the values of the hydraulic roughness coefficients.

In order to incorporate this high degree of spatial and temporal variability, the methodology of Vieux and Farajalla (1994) is used to assign Manning roughness coefficients. From detailed field studies, these authors found that the Manning roughness coefficient featured a fractal dimension of 1.4 (= a Hurst coefficient of 0.6; see section 4.2.1.1). This indicates that the measurements of hydraulic roughness have Brownian properties. In these field studies, rainfall simulators were set up on hillslopes with a length of 12 to 14 m and flow rates and depths were measured at 0.6 m intervals. The sampling interval that captures the essential spatial variability of the Manning roughness coefficients does not seem to matter due to its Brownian variation in the field. The measurements followed a nearly uniform random distribution. Consequently, to incorporate the spatial and temporal variability of the hydraulic roughness coefficients, a map of the hydraulic roughness can be created with a random number generator, generating numbers from a uniform distribution (Vieux and Farajalla, 1994).

Chow (1959) reported mean values of the Manning roughness coefficients for pasture, field crops, light brush and weeds, dense brush, and dense trees with respective values of: 0.035, 0.040, 0.050, 0.070 and 0.100. These values are in the same order of magnitude as the ones proposed by Garbrecht (1961), which can be calculated using equation (5.20). Therefore, event simulations in the experimental watershed (see sec-
tion 5.4) were done by sampling the Manning coefficients from the uniform interval [0.01, 0.1] with a random number generator.

### 5.3.4 Soil transport model

Schramm and Prinz (1993) carried out rainfall experiments on large plots (4.0 m width and 22.5 m length) in order to determine the critical momentum flux necessary in the original transport equation of the EROSION-2D/3D model (equation 5.25). However, these authors were not able to find simple relationships between the critical momentum flux and the basic soil parameters. Because the critical momentum flux is very important in the original EROSION-2D/3D model, and difficult to assess, it was decided to select another transport model to eliminate this parameter.

For the erosion processes in rills and gullies, Nearing et al. (1997) reported a simple but very accurate transport function. Based on laboratory flume experiments (soil in V-shaped flumes was exposed to a known discharge) using silt loam and sandy loam soils, these authors found a single logistic relationship ($R^2 = 0.93$) between the unit sediment load and the stream power of the overland flow, for all soil types (see also figure 5.13). This indicates that for all unconsolidated agricultural soils in the Belgian loess belt the same transport function can be used. The stream power can be calculated with:

$$\omega = \rho_w \cdot g \cdot S \cdot q$$  \hspace{1cm} (5.92)

wherein $\rho_w$ is the density of water (g·cm$^{-3}$), $g$ is the gravitational constant (cm·s$^{-2}$), $S$ is the slope (m·m$^{-1}$) and $q$ is the discharge per unit width (cm$^2$·s$^{-1}$). The resulting rill transport equation can be written as:

$$\log_{10}(q_w) = a + b \cdot \frac{e^{c+d\log_{10}(\omega)}}{1 + e^{c+d\log_{10}(\omega)}}$$  \hspace{1cm} (5.93)

with parameters $a = -34.47$, $b = 38.61$, $c = 0.845$, $d = 0.412$. From their experiments it was found that for slopes lower than 30%, transport capacity was already reached within the first 180 cm of rill length.

Although the most important transport of soil particles towards the drainage system
is governed by rill flow, sheet erosion processes in the interrill areas are an important source of sediment transport towards the rills. Because rills are also fed with sediment from interrill areas, we might therefore expect that transport capacity will be reached very rapidly, also on very steep sloping areas. To have an idea of the sediment delivery toward the rilling system on an agricultural field, a similar type of transport function was developed for the sheet erosion processes.

This was done using the results of 140 laboratory rainfall simulations. From these 140 laboratory experiments, 133 experiments were carried out in the period from 1973 to 1998 by Pauwels (1973), Gabriels (1974), Verdegem (1979), De Beus (1983), Goossens (1987) and Gabriels et al. (1998). All these experiments were done using sandy, loamy and silty soils. In addition, 7 rainfall experiments were performed on an alluvial clay soil (42 % clay). The textural composition of the soils used in all these experiments is given in figure 5.12.

![Figure 5.12: Textural composition of the soil used for the rainfall simulations.](image-url)

All laboratory rainfall experiments were performed on a smoothed surface to prevent possible rilling and to ensure a broad sheet flow during the simulated rain. The intensities of the simulated rain was held constant during the experiments and ranged from 20 to 128 mm·h⁻¹. The width of the soil pans was 20 to 30 cm and the length 30 to 90 cm. The slope of the soil pans ranged between 4 and 33 %. The duration of the experiments was between 60 and 120 minutes, and the soil loss was measured at 5 or 10-minute intervals, depending on the intensity of the simulated rain. This resulted in
672 observations of discharge and soil loss.

The measurement of the discharge, \( Q \) (cm\(^3\)-s\(^{-1}\)), is very straightforward: depending on the simulated rainfall intensity, every 5 or 10 minutes the amount of runoff is caught in a graduated cylinder. Because a broad sheet flow is established during the simulations, the width of the flow equals the sample width and the volumetric water flux per unit plane width, \( q \), is directly computable. If the measured unit sediment load, \( q_s \) (g-s\(^{-1}\)-cm\(^{-1}\)), is log-log plotted against the stream power of the overland flow (figure 5.13), a linear relationship can be identified in the data. This relationship is also a function of the clay content. The higher the clay content, the lower the unit sediment load. On figure 5.13, it can be clearly seen that the unit sediment load is between 1 and 3 log cycles higher for the laboratory rainfall experiments than for the flume experiments of Nearing et al. (1997). This can be explained by the raindrop impact. Because sheet flow is broader and more shallow than rill flow, the momentum of a raindrop impact can act directly upon the soil surface. Consequently, raindrop impact can be considered as the most important soil detachment process in interrill areas. The parameters of the regression power equations, with there corresponding correlation coefficient are also given in figure 5.13. Remark that the exponents of the regression equations are all (except one) around 1.3. The intercept of the regression equations is a measure of the erodibility of the soil. In general, the higher the clay content, the higher the cohesion and the lower the erodibility.

The transport functions derived from laboratory rainfall experiments are only valid if there is direct raindrop impact. Therefore, in the STM-2D/3D model, these transport equations are used to describe the sediment transport during rainfall and when the soil surface is not shielded by a vegetation cover. The transport equation of Nearing et al. (1997) is used when there is runoff without rainfall, or for these parts of the land where there is a surface cover.

The transport functions are valid for all non-cohesive agricultural soil types, and cannot explain the different observable erosion rates between for example sandy and silty soils. However, in section 5.4 it will be proved, using summer and winter event simulations, that the STM-2D/3D model can predict these different erosion rates based on differences in runoff generation.

Because the clay content in the test watershed is on every location lower than 25 %,
Figure 5.13: Soil transport function: stream power as predictor for the unit sediment load. Note that the laboratory flume experiments of Nearing et al. (1997) had a much wider stream power range than the laboratory rainfall experiments.
it was preferred to use one single transport equation for the estimation of the unit sediment load during sheet and micro-rill flow under raindrop impact. For this reason, the observations made on a clayey soil (41% clay) were eliminated from the data set. A regression analysis using the remaining observations resulted in the following equation, with an $R^2$ of 0.89:

$$q_s = 0.00016357 \cdot \omega^{1.31}$$ \hspace{1cm} (5.94)

wherein $q_s$ is the unit sediment load (g s$^{-1}$ cm$^{-1}$) and $\omega$ the stream power (g s$^{-3}$).
5.4 Validation of the STM-2D/3D model

The validation of the STM-2D/3D model was focused on the hydrological submodel. The results of the Green-Ampt infiltration model combined with the kinematic wave routing algorithm were confronted with measured hydrographs and the observed erosion pattern in a small test watershed with a total area of 142 ha. To test the validity of the physical concept of the model, the model was calibrated using only one parameter: the soil initial moisture content, which is also the most sensitive parameter in conceptual erosion models (Schmidt, 1996). All event calculations simulate the potential erosion risk (worst case scenario): it was assumed that there was no vegetation. The interception capacity, the soil cover and soil retention capacity were therefore set to zero. The output maps of the event simulations indicate then the location of the potential partial contributing areas for runoff production. These partial contributing areas are also the erosion risk areas. Using this methodology, the potential erosion maps of both the RUSLE and the STM-2D/3D model could be compared.

5.4.1 Data generation

From the flow chart of the STM-2D/3D model (see figure 5.2) the following data is necessary to run the model: (1) soil properties (sand, silt, clay and organic carbon content), (2) (micro)-topographical features (general topography, field boundaries, retention capacity, cumulative drainage area, drainage directions, the channel width and Manning coefficients), (3) vegetation parameters (interception capacity and soil cover). To calibrate and validate the hydrological model additional information was necessary: time series of rainfall and discharge.

5.4.1.1 Topography and topography related parameters

Figure 5.14 shows the topographical features and land use within the watershed (see also figure 2.1 for the position of this test watershed inside the community of Hevel-land). The brook system starts at the spring levels on the Kemmelberg hill. The elevation of this hill is 151 m, while the elevation at the watershed outlet is 35 m. The spring levels are located at the transition of Tertiary sands towards Tertiary clays.
boundaries of the agricultural fields were digitized from the 1:10,000 orthophotos.

![Watershed boundary with land use classification](image)

**Figure 5.14:** Geographical location and land use of the watershed used to test the hydrological submodel of the EROSION-2D/3D model.

Figure 5.15 shows the digital elevation model (DEM) and the cumulative drainage area and watershed boundary as derived from the DEM. The DEM was interpolated using the digitized contour lines on the 1:10,000 topographical maps. A 10 m resolution grid was created using an inverse distance interpolation routine. The DEM delivers a considerable amount of input data for the STM-2D/3D model (see figure 5.2): drainage directions, the local slope in these drainage direction, cumulative drainage area, the position of the channels and the spatial extent of the watershed. This information was derived from the DEM using the stochastic eight-neighbours method, described in Fairfield and Leymarie (1991). A map of the topographic index is given in figure 5.16. However, this map was not used for the determination of the initial moisture content.

In the model, the layout of the cross-sectional channel area is supposed rectangular over the whole route of the brook because the brook-system is deeply incised into the
topography. The flow width ranges everywhere between 60 and 120 cm. In the model, the hydrological characteristics are calculated using a mean flow width in the channels of 90 cm.

Figure 5.15: Digital elevation model (elevation in m above sea level) and cumulative drainage area (Natural logarithm of the number of pixels).

Figure 5.16: Topographical index, ln(α/tan(β)), of the test watershed, where α is the cumulative drainage area and β the local slope.
5.4.1.2 Soil texture, soil organic carbon content and drainage condition

The soil map of Belgium gives an indication of the soil drainage condition. Figure 5.17 shows the drainage condition of the test watershed (for a more detailed description of the drainage symbols, see section 2.5). However, this map was created around 1963 (Hubert and T’Jonck, 1963). Because of tile drainage, the drainage classes on the soil map are not up-to-date. A measuring campaign in East-Flanders (Blancquaert et al., 1997) revealed that under tile drainage, only 33 % of the measurements of groundwater depth correspond with the soil map. For those areas there was a clear shift of one drainage class. Also, the use of heavy machinery has created almost everywhere very compact plough pans. Therefore, the observed moisture condition of the upper soil horizon is not indicative for the depth of the ground water table. In many cases, the soil directly beneath the plough pan is significantly dryer than the upper soil horizon.

Figure 5.17: Drainage condition according to the Belgian soil map.

To generate maps of the textural composition and organic carbon content, the same
data set was used as for the generation of the soil erodibility map (see section 4.2.3). A total of 154 samples were analyzed for 8 textural fractions and organic carbon content. The sampling grid resolution was 325 m, with 9 transects where the sampling interval was 42m to determine the short distance variability. Maps of the sand, silt and organic carbon content were created using block-kriging. The descriptive statistics of the soil characteristics can be found in table 5.4. The respective variograms and the kriging results (kriging estimates and kriging standard deviation) are displayed in figures 5.18, 5.19 and 5.20. The map of the clay content was created by applying the formula: \(100(\%) = Silt(\%) - Sand(\%)\).

<table>
<thead>
<tr>
<th>Table 5.4: Descriptive statistics of the soil parameters.</th>
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<tbody>
<tr>
<td>number of samples</td>
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<tr>
<td>-------------------</td>
</tr>
<tr>
<td>mean</td>
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<td>standard dev.</td>
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<td>skewness</td>
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<td>coefficient of variation</td>
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Figure 5.18. Spherical variogram (nugget = 13, sill = 170, range = 826m), block-kriged estimates and the kriging standard deviation of the sand content in the Kemmelbeek watershed.
Figure 5.19: Spherical variogram (nugget = 22, sill = 168, range = 938m), block-kriged estimates and the kriging standard deviation of the silt content in the Kemmelbeek watershed.
Figure 5.20: *Exponential variogram (nugget = 0.060, sill = 0.384, range = 350m), block-kriged estimates and the kriging standard deviation of the organic carbon content in the Kemmelbeek watershed.*
Figure 5.21 gives a more detailed overview of the textural composition (percent sand, silt and clay) and the organic carbon content of the soils inside the watershed.

Figure 5.21: Textural composition and organic carbon content of the soils inside the test watershed.

5.4.1.3 Time series: discharge and precipitation

In the period of 31 May 1997 to 24 January 1998 discharge and precipitation were logged at the watershed outlet at a time resolution of 10 minutes. Precipitation was measured
using a tipping bucket rain gauge. Discharge was measured using a commercial flume (figure 5.22) calibrated in the range of 0.2 to 60 l·s⁻¹.

Figure 5.22: Automated discharge measurement device at the watershed outlet: flume calibrated in the range 0.2 to 60 l·s⁻¹.

It is clear that in a watershed of 142 ha, it is not possible to measure all peak flows for high intensity rain events, due to the limited measuring range of this instrument. One could argue, why this flume was not placed further upstream in order to enable the measurement of all rain events. The position was chosen in such a way that, the smallest possible watershed was created which boundaries enclose more than one soil type. This was done to verify if the STM-2D/3D model can explain the different observable erosion rates between different soil textures, although the soil transport function (equation 5.93) applies for all agricultural soil in the loess belt of Flanders. The brook system has always a certain base flow, even in very dry periods. The observed base flow was never lower than 1 l·s⁻¹.

Figure 5.23 shows the resulting precipitation and discharge time series of the experimental watershed in the period 31 May 1997 – 24 January 1998. Remark that hydrographs of these events where the peak flow is higher than 60 l·s⁻¹ cannot be used, due to the specific measuring range of the flume. The discharge is measured using a pressure transducer at the base of the measuring tube. During many rains, there was a consid-
erable transport of sediment. This sediment was trapped inside the measuring tube, creating an over-pressure on the pressure transducer, leading to erroneous discharge measurements. Therefore, not all rain events could be used to test the accuracy of the combined overland flow and channel flow algorithms. However, there were still enough summer and winter events to test the numerical algorithms.

Figure 5.23: Precipitation and discharge time series (raw data, 10 minute time resolution) of the experimental watershed in the period 31 May 1997 – 24 January 1998.
Figure 5.24 gives the cross-correlation between the observed rainfall and discharge data for summer and winter events. From this graph it can be observed that the concentration time of the runoff for winter events is larger than the concentration time for summer events. This can be explained by the rainfall properties and the land management parameters.

Figure 5.24: *Cross-correlation between the rainfall and discharge data for summer and winter events.*

As mentioned in section 2.3, in the summer period most of the rainfall originates from unstable cloud formations with a limited spatial coverage. The rainfall has a clear peak intensity, much higher than the observed peak intensities in the winter period. During winter, rainfall is mainly produced by cyclonal centra, occluded fronts or double fronts wherein stratus-shaped clouds dominate (60%, around 40% in the summer). The areas under precipitation are then much more extensive and the rain event itself does not feature such high peak intensities as in summer (see figure 5.23 and table 2.2), resulting in more uniform distributed intensities during the rainfall event.

Another important reason is that most agricultural fields are roughly ploughed before the winter period, inducing a higher random roughness. This has some effects when
modelling the processes during summer and winter events. For summer events the
Manning coefficients were supposed to be uniformly distributed between a minimum of
0.01 and a maximum value of 0.07, for winter events between a minimum of 0.01 and a
maximum of 0.10. These Manning coefficients have almost the same range and mean
value as the ones obtained by Dingman and Sharma (1997) in their study of discharge
equations in natural channels.

5.4.2 Results and discussion

Figures 5.25 to 5.31 give an overview of the results for some summer, autumn and
winter rainfall events. In each figure, maps of the initial moisture content, the runoff
production, the net soil loss and net soil deposition are shown.

The initial moisture content was the (only) calibration factor during the simulations.
Using pedotransfer functions (5.11) and (5.12) the saturated ($\theta_s$) and residual moisture
content ($\theta_r$) was determined for each pixel respectively. These values define the range
of the initial moisture content. Initially (the first iteration), the soil moisture is set to:
$\theta_r + ((\theta_s - \theta_r)/2)$. For the next iterations the procedure described in section 5.3.1.1
was used to alter the initial moisture content during calibration. The maps of the
initial moisture content clearly indicate the general observable evolution of the topsoil
moisture content during the year: more dry in the summer period, and wet (close to
saturation) in the winter period.
Figure 5.25: Simulation of the precipitation event of 17 July 1997.
Figure 5.26: Simulation of the precipitation event of 27 August 1997.
Figure 5.27: Simulation of the precipitation event of 30 August 1997.
Figure 5.28: *Simulation of the precipitation event of 4 November 1997.*
Figure 5.29: Simulation of the precipitation event of 20 November 1997.
Figure 5.30: Simulation of the precipitation event of 9 December 1997.
Figure 5.31: Simulation of the precipitation event of 18 December 1997.
As an objective function of model performance, the Nash-Sutcliffe model efficiency value, $E_{\text{eff}}$ (Nash and Sutcliffe, 1970), was used. This evaluates mainly high flow conditions. The efficiency calculated with logarithm values of discharge, $E_{\text{log}}$, lays a stronger stress on the performance of low flow simulations. The Nash-Sutcliffe efficiency value, $E$, can be written as:

$$E = 1 - \frac{\text{Var}[R]}{\text{Var}[Q]}$$

(5.95)

with $\text{Var}[R]$ the variance of the residuals and $\text{Var}[Q]$ the variance of the observed discharges. These variances can be calculated with:

$$\text{Var}[R] = \frac{SS_R}{n}$$

(5.96)

$$\text{Var}[Q] = \left(\frac{SS_Q}{n} - E[Q]^2\right)$$

(5.97)

wherein, $SS_R$ is the sum of squared errors (with error = observed - simulated), $SS_Q$ is the sum of squared observed discharges, $E[Q]$ is the mean observed discharge and $n$ is the number of observations. In order to balance between the importance of both high and low discharges for the evaluation of model performance, a combined model efficiency, $E_{\text{com}}$ can be used. $E_{\text{com}}$ results from a multiplication of $E_{\text{eff}}$ and $E_{\text{log}}$. Compared to an additive combination, this requires both values to be large for an overall good model performance. The Nash-Sutcliffe model efficiency values are given in table 5.5.

In the calibration phase of most hydrological models, the input parameter set is optimized by using an indicator such as $E_{\text{com}}$ or equivalent, to obtain the optimal model performance (Güntner et al., 1999). However, if multiple parameters of a model can be tuned, a reasonable fit between model output and field measurements is always possible (Beven, 1997). It is then difficult to validate the general concept of a model and false assumptions in the model concept are then less restrictive (Beven, 1997). Therefore, it was decided to use only one calibration parameter, the initial moisture content. The general reasoning behind this assumption was: (1) The moisture content of the upper soil horizon governs the amount of runoff; (2) The moisture content of the upper soil horizon is determined by the soil physical parameters.
From table 5.5 and the figures 5.25, 5.26, 5.27, 5.28, 5.29, 5.30 and 5.31 the following can be deduced:

1. Simulations of summer events with the STM-2D/3D model, result in better overall efficiency values ($E_{\text{com}}$) than winter events.

2. The rising limbs of the simulated hydrographs of the winter event simulations (figures 5.29, 5.30 and 5.31) react slower on rainfall than the observed hydrographs.

3. The model efficiency values stressed on the performance of the lower discharges ($E_{\text{log}}$) are lower for the winter event simulations (figures 5.29, 5.30 and 5.31).

Table 5.5: The Nash-Sutcliffe model efficiency values for some summer and winter event simulations with the STM-2D/3D model.

<table>
<thead>
<tr>
<th>event date</th>
<th>$E_{\text{eff}}$</th>
<th>$E_{\text{log}}$</th>
<th>$E_{\text{com}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>17 July 1997</td>
<td>0.90</td>
<td>0.90</td>
<td>0.81</td>
</tr>
<tr>
<td>27 August 1997</td>
<td>0.89</td>
<td>0.85</td>
<td>0.76</td>
</tr>
<tr>
<td>30 August 1997</td>
<td>0.86</td>
<td>0.80</td>
<td>0.69</td>
</tr>
<tr>
<td>4 November 1997</td>
<td>0.87</td>
<td>0.75</td>
<td>0.65</td>
</tr>
<tr>
<td>20 November 1997</td>
<td>0.73</td>
<td>0.67</td>
<td>0.49</td>
</tr>
<tr>
<td>9 December 1997</td>
<td>0.75</td>
<td>0.57</td>
<td>0.43</td>
</tr>
<tr>
<td>18 December 1997</td>
<td>0.81</td>
<td>0.76</td>
<td>0.62</td>
</tr>
</tbody>
</table>

From these results the following can be concluded focused on the general concept of the hydrological submodel of the STM-2D/3D model:

1. The concept of determining the initial moisture content solely based on soil physical characteristics (soil texture, soil density and organic matter content) appears to be an acceptable approximation for summer and early autumn rainfall events.

2. Because of the slow reaction on rainfall of the rising limb of the simulated hydrograph for winter events there can be concluded that not all partial runoff contributing areas are determined by the initial moisture concept of the STM-2D/3D model. In the winter months (November, December, January, February)
there will be some small partial contributing areas close to the river/brook system (necessary for a fast response of the rising limb) due to the moisture redistributing effect of surface topography (diverging/converging topography elements) and subsurface topography of low conductivity soil layers (e.g.: plough/traffic pans, Tertiary clay layers, areas affected by surface sealing).

3. Subsurface flow is much more intense during the winter months than in the summer period. Because the STM-2D/3D model has no subsurface flow component and the baseflow was kept constant during the entire event, the model efficiency values stressed on the performance of the lower discharges ($E_{log}$) are lower in winter than in summer. However, this component should only be taken into account if one wants to use the model for continuous simulations. For single event simulations, a subsurface flow component does not provide additional information regarding to sediment transport.

The simulations presented in figures 5.25 to 5.31 are all low intensity rainfall events. Only for these kind of events the peak discharges of the hydrograph could be measured, due to the limited measuring range of the flume. To have an idea of the runoff producing areas during rainfall events with a higher intensity, figures 5.32 and 5.33 show the results for a summer storm. The initial moisture content used for the generation of figure 5.32 was: $\theta_r + (\theta_s - \theta_r) \cdot 0.5$. For figure 5.33 the initial moisture content was set to: $\theta_r + (\theta_s - \theta_r) \cdot 0.9$. 
Chapter 5: Predicting Event-Based Sediment Transport

Figure 5.32: Simulation of the precipitation event of 25 August 1997 (no measured hydrograph available, initial moisture content 50% of saturated value).
Figure 5.33: Simulation of the precipitation event of 25 August 1997 (no measured hydrograph available, initial moisture content 90% of saturated value).
For a validation of the sediment transport component the following validation methods are possible: a fully quantitative validation, a comparison of soil transport simulations with observations of indirect erosion indicators (e.g.: occurrence of deposition areas, Cs-137 tracer technique) and a plausibility test (does the model generate plausible results).

A comparison between the soil loss results obtained with the STM-2D/3D model and the field truth on a fully quantitative basis was not possible. This requires an infrastructure inside the watershed which can record the sediment transport on a continuous and preferable also on a distributed basis. Erosion plots measurements were elaborated inside the Kemmelbeek watershed, but the temporal resolution of these measurements was not fine enough and the spatial extent (1 m width, 10 to 30 m long) not large enough to use these observations for a quantitative validation.

A qualitative (visual) comparison is possible, but is rather difficult to interpret. The erosion rate in the study area is not spectacular, and gullies or bigger rills are rarely formed. However, deposition areas can be very clearly recognized in the field. Figure 5.34 gives an overview of the major observed deposition areas in the winter period (January 2000) of the test watershed. If the deposition images of figures 5.32 and 5.33 are visually compared with figure 5.34 a correspondence between the predicted and observed deposition areas can be noticed.

From field observations it is known that silty soils are the most erodible soil types. Figures 5.25 to 5.31 and figure 5.21 indicate that this phenomenon can be simulated with the STM-2D/3D model. The higher the silt content, the higher the runoff production, thus the higher the potential sediment transport. It can be concluded that the STM-2D/3D model generates plausible results. For a further analysis of the plausibility of the results generated with the STM-2D/3D model reference is made to section 5.5 and chapter 6, where the effect of the different erosion control techniques are simulated and where some of these simulations are qualitatively compared with erosion plots observations inside the Kemmelbeek watershed.

However, an intensive measuring campaign (continuous logging of precipitation, discharge and sediment transport at the outlet of the test watershed combined with large erosion plot measurements) is necessary for a fully quantitative validation of the sediment transport component of the STM-2D/3D model.
5.5 Comparison of the STM-2D/3D with the RUSLE model

Figure 5.35 shows the potential soil loss per agricultural parcel, calculated with the RUSLE model. Potential soil loss, is the soil loss when a field remains bare for the whole year (RUSLE vegetation factor, C = 1) and there are no special erosion control measures (RUSLE management factor, P = 1).

When figure 5.35 is compared with figure 5.15, there can be concluded that according to the RUSLE model, the steep sloping areas are the high risk areas. This can be illustrated by figure 5.36, wherein the RUSLE predicted potential soil loss is plotted against the RUSLE topographic factor (LS). A linear relationship between the RUSLE potential erosion and the RUSLE-LS factor was found with a high R² value of 0.96. Because the RUSLE potential soil loss is the result of the multiplication of the LS factor with the rainfall erosivity factor (R) and the soil erodibility factor (K) and because the RUSLE-R factor is constant over the watershed, this regression indicates that the
RUSLE-K factor has only an influence of 4% in the resulting potential soil loss. This is a rather low value given the large differentiation in soil types within the watershed (silty to sandy soils, see figure 5.21).

As mentioned before, the erosion plot measurements, which formed the basis of the RUSLE model, were initially designed to measure soil erodibility (Wischmeier and Mantering, 1969). Consequently, one might expect that a relation must exist between the RUSLE-K factor and the runoff amount predicted with the STM-2D/3D model. Figure 5.37 shows this relation for the two event simulations given in figures 5.32 and 5.33. These two simulations concern the same rainfall event but with a different initial soil moisture content. From this figure it can be clearly seen that a nonlinear relationship exists between the RUSLE-K factor and the STM-2D/3D predicted runoff volume, but this relationship is strongly dependent on the initial soil moisture content:

Figure 5.35: Soil loss data generated with the RUSLE model for the test watershed.
the higher the initial moisture content, the less differentiation between the different soil types concerning runoff production. For the simulation where the initial moisture content was 90% of the saturated value, the difference in estimated runoff production between the lowest soil erodibility (±0.015) and the highest soil erodibility value (±0.072) is only 1 log cycle (factor 10). For the simulation where the initial moisture content was 50% of the saturated value, this difference is around 5 log cycles (factor 100000).

Figure 5.37 is an important figure to prove the plausibility of the STM-2D/3D model, because all the findings of Bajracharya and Lal (1992) can be derived from it. The paper of Bajracharya and Lal (1992) is one of the very few dealing with seasonal soil loss and erodibility variation. The aim of their study was to make a preliminary assessment of the trend and magnitude of seasonal variations in erosion and erodibility for a silt loam soil, using standard Wischmeier runoff plots (22 by 4 m). Soil loss and erodibility varied significantly among seasons. Erodibility was high under wet soil
conditions during the winter and spring. Summer erodibilities were lowest in each year despite high soil loss from erosive rains. Kinnell and Risse (1998) tried to enhance the USLE for better event-based predictions. These authors found that the efficiency of the USLE decreases as the proportion of the rain infiltrating increases. Kinnell and Risse (1998) proposed new K and C values, but stated that the explicit consideration of runoff means that values of many of the other factors used in the USLE have to be re-evaluated because the effects of variations in factors like soil and crop management on runoff are now being treated separately from the effects on sediment concentration.

Figure 5.37 indicates the importance of the soil moisture regime throughout the year concerning runoff production and eventually the observed sediment transport. However, this is not taken into account in the current versions of the RUSLE model. This problem is also recognized by Renard et al. (1996): “K-values are difficult to estimate mainly because of antecedent soil-water and soil-surface conditions and because of seasonal variations in soil properties”. These authors propose the introduction of seasonal
varying K values. However the development of regression equation to incorporate these seasonal effects is hampered by the interdependency between the different RUSLE factors and the need for site-specific long-term field measurements (Renard et al., 1996). Up to now, only temperature effects (frost periods) are taken into account in the RUSLE-K factor (Renard et al., 1996).

A fully quantitative comparison between the RUSLE model and the STM-2D/3D model on a yearly basis was not possible, since this comprises the availability of good discharge/precipitation data during at least one year. However, a comparison on an event basis is possible because the RUSLE-R factor results from the sum of the rainfall energies of all individual rainfall events throughout a year (Kinnell and Risse, 1998). The RUSLE erosivity value of a single storm is in the RUSLE model concept invariant of other parameters.

Figure 5.38 compares the potential erosion losses per agricultural field calculated with the RUSLE and the STM-2D/3D model. The RUSLE potential erosion losses are calculated with the respective erosivity values of the events. For a good fit between the results of both models, the linear regression equations (through the origin) need a slope value of 1.0 (bisector line) and a high $R^2$ value. However, the slope values of the events during dry soil moisture conditions (17 July and 27 August) are considerable larger than 1. For these events a larger portion of rainfall could infiltrate than in wet soil moisture conditions. The slope values of the regression equations of the events during wet soil moisture conditions (30 August, 4 November, 20 November and 9 December) are smaller than 1. During these events a smaller portion of the rainfall could infiltrate than in dry soil moisture conditions. This means that the RUSLE overestimates the soil erodibility during dry soil moisture conditions (usually in summer) and underestimates soil erodibility during wet conditions. The $R^2$ values of the fits are also very low, indicating that the spatial erosion pattern generated by both models is substantially different for the low intensity events in figure 5.38. These findings are in agreement with the conclusions of Kinnell and Risse (1998) stating that the efficiency of the USLE decreases as the proportion of the rain infiltrating increases.
Figure 5.38: Comparison between the RUSLE potential erosion and the STM-2D/3D potential erosion for 6 rainfall events. The overland flow estimated with the STM-2D/3D model was calibrated using discharge measurements. In each plot the RUSLE erosivity value \((MJ \cdot mm \cdot ha^{-1} \cdot h^{-1} \cdot event^{-1})\) of the event is given.
Figures 5.39 and 5.40 compare the potential erosion generated by the RUSLE and the STM-2D/3D model for a high intensity rainfall event (RUSLE-R = 123.8 MJ·mm·ha⁻¹·h⁻¹·event⁻¹). When the soil is initially dry (figure 5.39), there is only a poor correlation (R² = 0.26). However, with initially wet soil conditions (figure 5.40), a better correlation (R² = 0.50) can be observed, indicating that the spatial erosion patterns generated by both models are comparative.

Figures 5.37 to 5.40 indicate the strong interaction between the rainfall erosivity, soil erodibility and antecedent moisture content. Overpredicting the soil loss during summer and underpredicting the soil loss during winter, compensates errors when the RUSLE is used on a yearly, or long-term (decades) basis. This is why still a good correlation can be found between the RUSLE-predicted soil loss accumulated for the whole Kemmelbeek watershed (4376 ± 75 ton·year⁻¹) and the long-term observed sediment input (4438 ton·year⁻¹) in a water reservoir (see Chapter 4). Because the efficiency of the (R)USLE decreases as the proportion of the rain infiltrating increases, the spatial erosion pattern generated with the RUSLE can be questioned. The lower the total rainfall amount during an event the higher the necessity of a good description of the infiltration/runoff processes for the delineation of the runoff producing areas and thus the erosion risk areas inside a watershed. To indicate the importance of these moderate rainfall events, Bollinne (1983) found from erosion plot observations that about 40 % of the soil loss in Belgium is the result of low intensity precipitations (with a total rainfall amount lower than 12.7 mm). This comparison analysis of both models indicated that the RUSLE can not be used on a time resolution shorter than a year, which is an important consequence when analyzing the effect of seasonal erosion control measures (e.g. off-season cover crops, winter mulching) is the objective.
Figure 5.39: Relation between the potential erosion estimated with the RUSLE and the STM-2D/3D model for a high intensity rainfall event (25 August 1997, RUSLE-R = 123.8 MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·event\(^{-1}\)) on an initially dry soil (initial moisture content was 50\% of saturated value).

Figure 5.40: Relation between the potential erosion estimated with the RUSLE and the STM-2D/3D model for a high intensity rainfall event (25 August 1997, RUSLE-R = 123.8 MJ·mm·ha\(^{-1}\)·h\(^{-1}\)·event\(^{-1}\)) on an initially wet soil (initial moisture content was 90\% of saturated value).
5.6 Conclusions

The original EROSION-2D/3D model was modified to integrate a physical based hydrological model. The general model layout and the necessary amount of data was maintained, but due to the integration of new algorithms and methodologies, the original model underwent important transformation. Therefore, it was decided to indicate this modified EROSION-2D/3D model with the name STM-2D/3D (Sediment Transport Model, which is implemented in a 2D hillslope and 3D watershed version).

It was the intention to build a model wherein the user can manipulate only a very limited number of hydraulic parameters. All critical parameters (saturated and residual moisture content, hydraulic conductivity, matric potential) were derived using pedo-transfer functions (based on the basic soil physical characteristics texture, organic carbon content and soil density) and cannot be changed. The only parameters the user can manipulate during event simulations are the initial moisture content and the Manning roughness coefficients (of the latter only the minimum and maximum value inside the whole catchment). By doing so, it could be checked if the general reasoning corresponds to what can be observed in the field. If every variable can be changed, the model can be tuned such that always good event simulations are possible.

The 2D and 3D versions are written in Java, what makes this software platform-independent. The 2D hillslope version can be integrated in an Internet environment and the full source code of this program can be found in the Appendix.

From an extensive data set of laboratory rainfall simulations, a sediment transport function based on the stream power concept was created for sheet and micro-rill flow with a high R² coefficient of 0.89. For rill and gully flow, a similar equation was available from literature (with an R² of 0.93). Using these functions, the unit sediment load can thus accurately be estimated if the discharge of the flow is known. Therefore, the attention of this research was focused on the calibration and validation of the hydrological component (estimation of the overland flow and channel discharge) of the STM-2D/3D model.

From calibrated event simulations the following can be concluded regarding the general concept of the hydrological submodel of the STM-2D/3D model: (1) The concept of determining the initial moisture content solely based on soil physical characteristics
(soil texture, soil density and organic matter content) appeared to be an acceptable approximation for summer and early autumn rainfall events. (2) Because of the slow reaction on rainfall of the rising limb of the simulated hydrograph for winter events there can be concluded that not all partial runoff contributing areas are determined by the initial moisture concept of the STM-2D/3D model. In the winter months (November, December, January, February) there will be some partial contributing areas close to the river/brook system (necessary for a fast response of the rising limb), due to the moisture redistributing effect of surface topography (diverging/converging topography elements) and subsurface topography of low conductivity soil layers (e.g.: plough/traffic pans, Tertiary clay layers, areas affected by surface sealing). (3) Subsurface flow is much more intense during the winter months than in the summer period. Because the STM-2D/3D model has no subsurface flow component and the baseflow was kept constant during the whole rainfall event, the model efficiency values stressed on the performance of the lower discharges ($E_{low}$) are lower in winter than in summer. However, this component should only be taken into account if one wants to use the model for continuous simulations. For single event simulations, a subsurface flow component does not provide additional information regarding to sediment transport.

To further refine the hydrological component of the STM-2D/3D model research is necessary to account for surface and subsurface topographical effects in the moisture redistribution (only relevant for winter event simulations).

It was found that the STM-2D/3D model generates plausible and qualitatively good results based on the position of the predicted and observed deposition areas and the general erosion pattern for moderate (both in intensity and rainfall amount) rainfall events. The generated erosion pattern for these events corresponded with the soil erodibility pattern of the RUSLE model (figure 5.37). For a further analysis of the plausibility of the results generated with the STM-2D/3D model reference can be made to chapter 6, where the effect of the different erosion control techniques are simulated and where some of these simulations are qualitatively compared with erosion plot observations inside the Kemmelbeek watershed.

On an event basis, there was no correlation between the RUSLE and the STM-2D/3D model both in absolute potential soil loss per agricultural parcel and the spatial erosion pattern. This is because the antecedent moisture content is not taken into account in
the RUSLE model. The current version does not provide a method which takes into account the rainfall and soil moisture regime. Overpredicting the soil loss during dry soil moisture conditions and underpredicting the soil loss during wet soil moisture conditions, compensates errors when the RUSLE is used on a yearly, or long-term (decades) basis. This is the reason why still a good correlation is found (Chapter 4) between the RUSLE-predicted soil loss aggregated for the whole Kemmelbeek watershed (4376 ± 75 ton·year⁻¹) and the long-term sediment input (4438 ton·year⁻¹) in a water reservoir. The comparison of both models indicated that the RUSLE cannot be used on a time resolution shorter than a year, which is an important consequence when analyzing the effect of seasonal erosion control measures (e.g. off-season cover crops, winter mulching) is the objective.

In section 5.5 it is shown that a seasonal varying erodibility value which is strongly related with the soil moisture regime is a necessity to enhance seasonal predictions of the RUSLE model. However, developing such site specific empirical relations are very costly both in time and money and including the soil moisture regime within the model is likely to affect a number of other RUSLE factors as well (Renard et al., 1996). Therefore, if the time resolution of the erosion analysis is shorter than a year and to evaluate seasonal effects, the use of physical based models is advisable. This type of models is more flexible in use, model parameters still have a physical meaning and a model analysis results in much more information (soil loss, sediment deposition, channel hydrographs, overland flow hydrographs, degree of turbulence, water levels, water velocities, ...) with the same efforts.
Chapter 6

Erosion Control Measures

6.1 Introduction

In general, the aim of soil conservation is to obtain the maximum sustained level of production from a given land area whilst maintaining soil loss below a threshold level which, theoretically, permits the natural rate of soil formation to keep pace with the rate of soil erosion (Morgan, 1996). In addition, there may be a need to reduce erosion, to control the loss of nutrients, pesticides and herbicides from agricultural land, to prevent pollution of water bodies, to decrease the rates of sedimentation in reservoirs, rivers, canals, ditches and harbors. However, since erosion is a natural process, it cannot be prevented but it can be reduced to a maximum acceptable rate or soil loss tolerance.

In this chapter a short overview is given of the different soil conservation measures and in particular how the hillslope version of the physical event based sediment transport model (STM-2D) can be used to design these erosion control measures. Some of the simulations are also qualitatively compared with erosion plot observations inside the Kemmelbeek watershed. The program source code of this model can be found in the Appendix.

6.2 Soil loss tolerance

As mentioned in the introduction, the theoretical soil loss tolerance level is the rate at which soil loss equals soil formation. Because soil formation is so slow, measuring
this rate becomes practically impossible. Zachar (1982) estimated the average soil formation rate as 0.1 mm·year⁻¹. However, this definition of the soil loss tolerance is based solely on agricultural considerations. In many parts of the world, problems of pollution and sedimentation outweigh those of lost farming potential (Morgan, 1996).

In Flanders, a general figure about the soil loss tolerance is not relevant. Severe crop damage by soil erosion is exceptional and therefore farmers consider erosion as being only a marginal problem. Due to the modern cultivation techniques the biomass production is good to very good even on fields which should have been considered as marginal in earlier decades. In general, the farmers are not unwilling to change practices but will do so only if substantial benefits arise and the investment costs can be recovered.

For the community, the sediment transport towards the water bodies should be as low as possible to reduce the dredging costs and the possible flooding risk. Because most sediments in larger rivers are enriched with pollutants during transport, it is not allowed to dump this sediment again on the agricultural fields. The sediment is considered as a waste material, which increases the general cost for the community.

For the Doutebeek watershed (2727 ha, 23 km long brook-system inside the community of Heuvelland) approximately 8500 m³ (≈ 11 050 ton) sediment is dredged every three years. This maintenance operation represents a cost of 2760 000 BEF (€68 419) each three years (D. Cuvelier, 1999, Regional Land Management Board, personal communication). This is equivalent with a cost of approximately 250 BEF (€6.20) per ton of dredged sediment (storage costs excluded). From the viewpoint of the community, the sedimentation should be as low as possible. Therefore, on-site sediment control is a necessity.

### 6.3 Erosion control measures

Soil conservation strategies can be classified in three main groups: agronomic measures, soil management measures and mechanical methods (Morgan, 1996). Agronomic measures are: mulching (with natural or synthetic products), high-density planting, crop rotations, strip cropping (contains also riparian buffer zones) and cover cropping. As specific soil management measures one can quote: contour tillage, ridging, minimum
tillage or no-till. The most important mechanical methods are: terracing, waterways, tile drainage and structures. However, many of these cited erosion control measures are not suitable because of the agricultural boundary conditions in Flanders, such as: small fields, heavy machinery and very intensive production methods.

Generally, preference is always given to agronomic measures. These are less expensive and deal directly with reducing raindrop impact, increasing infiltration, reducing the runoff volume and decreasing water velocities. They are also more easily fitted into an existing farming system. Mechanical measures are largely ineffective on their own because they cannot prevent the detachment of soil particles. Their main role is in supplementing agronomic measures, being used to control the flow of any excess water that may arise. Many mechanical works are costly to install and maintain.

It is important to mention that an economical evaluation is necessary to introduce a sound erosion control policy in an existing farming system. Farmers have no substantial economic benefit: it introduces more labor, more investments and sometimes there is a loss of effective cultivating area (e.g. grassed buffer strips). An economical analysis can then be helpful to construct a kind of subsidizing system to promote soil and water conservation measures. Therefore, land planners and policy makers have an important role when a land area is restructured. If erosion control measures (like riparian buffer strips) are not taken into account in the planning phase, it becomes very difficult to promote them afterwards.

To analyze the effect of the different erosion control measures on a field scale, the STM-2D model can be used. This model was implemented in the Java program language, which makes the program platform independent and runnable in an Internet environment. Figure 6.1 shows a snapshot of the initial screen. The program uses a simple GUI (Graphical User Interface) for the definition of the hillslope and rainfall parameters and for a query of the model results (soil loss, net sediment transport and deposition along the hillslope, hydrographs, water velocity, water levels and Reynolds number). The figures displayed in the following sections about the effect of the different erosion control measures are generated with this program.

To evaluate the effect of the erosion control measures, as calculated with the STM-2D model, a reference simulation was necessary. The reference field was chosen to be 100 m long and 100 m wide, with a uniform slope of 10 %. The clay, silt, sand and organic
carbon content was chosen to be respectively: 20%, 60%, 20% and 1.2%. The soil density was set to 1400 kg·m⁻³ and the initial moisture content of the soil was assigned a very high value of 95% of the saturated value to simulate winter conditions. As reference precipitation, a winter event (January 1998) was chosen from the available time series (see figure 5.23). The intensity distribution in mm·h⁻¹ for this event was (with a 10 min pluviophase length): 5.0, 5.0, 2.7, 1.3, 0.0, 2.7, 2.7, 5.0, 7.0, 5.0, 2.7, 1.3. Figure 6.2 gives the cumulative soil loss along this reference hillslope and the net soil loss per segment. The total soil loss towards the drainage system was for this reference situation 1.483 ton. The STM-2D program subdivides a hillslope in segments of 1 meter. The time step of the finite difference scheme of the kinematic wave is then also very small (seconds). Figure 6.2 also shows the hydrograph and stream power
evolution of the last hillslope segment.

Figure 6.2: Reference simulation: cumulative erosion (upper left) and net sediment transport (upper right) along the reference hillslope (a bare field of 100 m long, 100 m wide, 10 % slope, without vegetation); hydrograph (lower left) and stream power (lower right) at the end of the hillslope.

6.3.1 Agronomic measures

Agronomic measures for soil conservation use the protective effect of plant covers to reduce erosion. Because of differences in their density and morphology, plants differ in their ability to protect the soil. The description below is focused on the agricultural conditions in Flanders.
6.3.1.1 Rotation

Particular problems are associated with maize and beets. These crops are harvested in the period from late September to November. In this period, the moisture condition of the fields is usually wet, resulting in a structural decay of the soil and the creation of a severe traffic/plough pan at a depth of 20–35 cm. Also, because of the late harvest, no cover crop can develop a dense cover to give a reasonable soil protection during the winter period, which is the highest risk period because of the wet soil moisture conditions. Monocultures of these crops in runoff contributing areas will often feature a high soil loss (when the soil loss is evaluated on a yearly basis) even in areas with moderate slopes.

A thoughtful crop rotation is therefore important to decrease the long-term soil loss on high risk parcels. However, some problems of the last decade in the agricultural economic sector have favored the development of the bio-industry, which induces a higher degree of specialization and thus also the introduction of monocultures.

6.3.1.2 Cover crops

Cover crops are grown as a conservation measure during the off-season. Typical crops are: grasses (in particular Lolium multiflorum, Lolium multiflorum var. weaterwoldicum, Lolium perenne and Secale cereale), Phacelia tanacetifolia, Raphanus sativus subsp. oleiferus and Sinapis alba. For these crops, early sowing is essential because a good crop cover must be obtained by the end of October/November since the climate is too cold for growth during the late autumn and winter months. Table 6.1 gives an overview of the biomass production (measured at the end of November and for Belgian climatological conditions) of some cover crops as a function of the sowing date (Ninane et al., 1995).

Grasses are excellent cover crops. They develop a dense root zone, have a high leaf area index (less direct raindrop impact), leaves give a direct soil cover (less area available for detachment by overland flow) and they suppress the growth of weeds. Winter cereal crops, like Secale cereale are relative cheap, other grasses, like Lolium sp., are more expensive. Secale cereale can be sown until the end of November (when the soil is not too wet) and is therefore the 'latest' cover crop. However, Secale cereale
demands lighter soil textures. Results from a demonstration field (see Table 6.2) in the Kemmelbeek watershed illustrate the importance of a grass cover crop. This table summarizes the results of November 1996. This was a very wet month with a total rainfall of 120.5 mm, where the normal value for November is 69 mm. For the bare erosion plots a mean soil loss of 6.54 ton·ha⁻¹ was measured, while on the plot with Lolium multiflorum this soil loss was only 0.07 ton·ha⁻¹. A very dense grass vegetation

<table>
<thead>
<tr>
<th>Crop</th>
<th>Sowing date</th>
<th>N-fertilization (kg·ha⁻¹)</th>
<th>Biomass (ton dry matter·ha⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sinapis alba</td>
<td>13-Aug-1990</td>
<td>60</td>
<td>5.8 ± 0.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>120</td>
<td>6.6 ± 0.4</td>
</tr>
<tr>
<td>Lolium sp.</td>
<td>27-Jul-1990</td>
<td>60</td>
<td>6.4 ± 0.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>120</td>
<td>6.2 ± 0.5</td>
</tr>
<tr>
<td>Sinapis alba</td>
<td>29-Aug-1991</td>
<td>0</td>
<td>3.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>5.0 ± 0.5</td>
</tr>
<tr>
<td>Lolium sp.</td>
<td>06-Aug-1991</td>
<td>0</td>
<td>6.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>6.8 ± 0.4</td>
</tr>
<tr>
<td>Phacelia</td>
<td>06-Aug-1991</td>
<td>0</td>
<td>3.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>5.8 ± 0.5</td>
</tr>
<tr>
<td>Sinapis alba</td>
<td>27-Aug-1992</td>
<td>0</td>
<td>3.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>6.9 ± 0.5</td>
</tr>
<tr>
<td>Lolium sp.</td>
<td>29-Jul-1992</td>
<td>0</td>
<td>4.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>8.8 ± 0.3</td>
</tr>
<tr>
<td>Phacelia</td>
<td>29-Jul-1992</td>
<td>0</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>8.3 ± 0.5</td>
</tr>
<tr>
<td>Sinapis alba</td>
<td>17-Aug-1993</td>
<td>0</td>
<td>3.2 ± 0.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80</td>
<td>5.5 ± 0.3</td>
</tr>
<tr>
<td></td>
<td>30-Aug-1993</td>
<td>0</td>
<td>2.4 ± 0.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>60</td>
<td>2.9 ± 0.3</td>
</tr>
<tr>
<td></td>
<td>13-Sep-1993</td>
<td>0</td>
<td>0.7 ± 0.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>50</td>
<td>1.3 ± 0.2</td>
</tr>
</tbody>
</table>
has thus an efficiency value of more than 99%.

Table 6.2: Results of some Lolium multiflorum and bare demonstration plots with a width of 1 m, on a loamy field in the Kemmelbeek watershed. These results are the soil losses for the month of November 1996 (total rainfall: 120.5 mm).

<table>
<thead>
<tr>
<th>Plot</th>
<th>Slope length (m)</th>
<th>Rainfall (l/plot)</th>
<th>Runoff (l/plot)</th>
<th>Runoff (%)</th>
<th>Soil loss (ton/ha)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lolium sp.</td>
<td>20</td>
<td>2410</td>
<td>184</td>
<td>7.6</td>
<td>0.07</td>
</tr>
<tr>
<td>Bare</td>
<td>10</td>
<td>1205</td>
<td>254</td>
<td>21.1</td>
<td>7.69</td>
</tr>
<tr>
<td>Bare</td>
<td>20</td>
<td>2410</td>
<td>452</td>
<td>18.8</td>
<td>5.78</td>
</tr>
<tr>
<td>Bare</td>
<td>30</td>
<td>3615</td>
<td>626</td>
<td>17.3</td>
<td>6.14</td>
</tr>
</tbody>
</table>

Phacelia develops a very good root system and a relative good soil cover. This crop is not related to any family of production crops and is therefore an excellent off-season crop. Phacelia is very sensitive to frost: a soil temperature of -5 °C is fatal, but the remains give a good cover during the winter period. Phacelia can be used on all soil types until the end of August. Raphanus sativus develops a very good soil cover when sown before September and is very frost sensitive. Sinapis alba has only a moderate soil coverage, can be sown until mid-September and is also frost sensitive.

Because some crops are not earlier harvested than the end of September to November (e.g. beets and maize), some experiments are currently being performed in the Kemmelbeek watershed of undersowing grasses (different Lolium species) in maize. Although there are some references (Buchner, 1988; Goeck and Geisler, 1988) which indicate that good to very good results can be obtained by undersowing grasses in sugar beets and maize, it is not a popular technique because it is labor intensive. With the experiments in the Kemmelbeek watershed, the Flemish governmental agricultural organizations try to reproduce these good results to elaborate a subsidizing system to promote this erosion control measure in its initial phase.

Figure 6.3 shows the cumulative and net soil erosion for the reference field mentioned in section 6.3 with a Sinapis alba cover crop. This crop has only a limited leaf coverage. In the simulation with the STM-2D model, the leaf coverage was set to a for this crop high value of 50%. The mulching effect (direct soil cover) of Sinapis alba is very low and was set to 0 % in this simulation. The calculated soil loss towards the drainage
system for the reference rainfall event (see section 6.3) was 0.822 ton. Without any vegetation the calculated soil loss was 1.483 ton. By protecting 50 % of the parcel area from direct raindrop impact, the efficiency value of the Sinapis alba cover crop was in this simulation 45 %.

Figure 6.4 shows the effect of a dense grass cover crop. For this simulation the leaf coverage was set to 100 % and the soil coverage (mulching effect) to 50 %. The simulated net soil loss towards the drainage system was 0.160 ton. The efficiency value of this cover crop for the reference boundary conditions is thus 89 %.

Figure 6.3: Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with Sinapis alba having a vegetation cover of 50 %.

Figure 6.4: Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with Lolium sp. having a vegetation cover of 100 % and a soil cover of 50 %.
6.3.1.3 Strip-cropping and buffer zones

With strip-cropping, row crops and protection-effective crops are grown in alternating strips aligned on the contour. Erosion is largely limited to the row-crop strips and soil removed from these is trapped within and behind the next strip downslope which is generally planted with a leguminous or grass crop. Guidelines for widths are between 15 and 45 m depending on the erosion hazard (Troeh et al., 1980). Although strip-cropping can be very effective, Emama and Morgan (1995) warn that erosion rates can be higher with contour strips than without. Strip-cropping is best suited to well-drained soils because the reduction in runoff velocity, if combined with a low rate of infiltration on a poorly-drained soil, can result in waterlogging and standing water.

Unfortunately, this technique is not compatible with the highly-mechanized agriculture on small-scale fields in Flanders (the average field area within the community of Heuvelland is only 1.4 ha). Furthermore, the protection-effective crops are usually of limited economical value.

However, riparian grassed buffer strips can be very important in an integrated soil and water conservation policy. This erosion control measure only reduces the off-site consequences of erosion. Most of the sediment can be trapped before the runoff water discharges towards the drainage system, but on-site erosion still occurs.

There are not many references about the efficiency of buffer strips. A good overview of the importance of buffer strips or riparian buffer zones is given in Haycock et al. (1996). Riparian buffer zones have many functionalities: (1) on-site sediment control, (2) ditch, brook and river bank stabilization and (3) they prevent or decrease direct input of fertilizers and crop protection products. Raffaele Jr. et al. (1997) found that the effectiveness of grass strips is very high. Strips of 0.6 m wide at the end of a standard Wischmeier erosion plot (22 m long, 1.8 m width) reduced soil loss from 50 to 80 % of that from an unprotected bare plot. From observations on standard Wischmeier erosion plots with a slope steepness of 7 % and 12 %, Robinson et al. (1996) found that almost all sediment is deposited in the first 6 m of the strip. Beyond the first 6 m, the additional sedimentation is only marginal. The first 3 m was the most effective with a sedimentation always (high and low intensity events) higher than 70 % of the total sediment deposition in the buffer strip.
These results can be reproduced with the STM-2D program. Figure 6.5 shows the sediment transport and deposition in a grass buffer strip of 5 m. The vegetation cover of this buffer strip was set to 90 % and the soil cover to 90 %, which represents a dense grass vegetation. The soil loss towards the drainage system reduced to 0.292 ton. Compared with the reference simulation this is an efficiency of 80 %.

Figure 6.6 illustrates the importance of a dense vegetation coverage to prevent direct raindrop impact. In this simulation, the soil cover of the 50 m wide mulch buffer zone was set to 100 %. The results indicate that a much wider buffer zone is needed for the deposition of sediment carried in the runoff water when there is raindrop impact than when the runoff flow is shielded from raindrop impact by a vegetation cover.

When the simulation presented in figure 6.5 is repeated, but on a steep hillslope of 35 % (about the maximum slope occurring in the Kemmelbeek watershed) a soil loss of 2.779 ton is found. Compared with a soil loss of 8.307 ton for the same field but without a buffer strip, this represents an efficiency of 67 %. A 5 m buffer zone, with a very dense grass vegetation might thus be wide enough for most agricultural fields to reduce the soil loss at least with 50 %.

![STM-2D Results](image)

**Figure 6.5:** Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with a riparian grass buffer zone of 5 m width.

To convince the farmers of the importance of riparian buffer zones, some demonstration fields are currently monitored in the Kemmelbeek watershed. The first observations seem to support the simulation results, but indicate that it should not be the only control measure: the precipitation amount in December 1999 was around 172 mm,
where the normal precipitation for this month is approximately 70 mm. After this very wet period, the sediment storage capacity of the total buffer area (5 m width) was completely filled. Sediment transported by subsequent runoff events could then freely flow towards the drainage system. This indicates that the necessary width of the buffer zone must be calculated based on its sediment storage capacity and the deposition potential of the agricultural field. Although 5 m will be enough for most fields, some fields or some spots at the end of a runoff concentration zone will need considerable wider buffer widths. To evaluate this the STM-2D/3D model should be extended to enable continuous simulations.

6.3.1.4 Mulching

Mulching is the covering of the soil with crop residues such as straw, maize stalks or composts. Overall, incorporated mulches are less effective than surface mulches. Because (in Flanders) the nutrient balance of the fields is now controlled by governmental organizations, and because addition of extra mulches or compost must be taken into account in this nutrient balance, this erosion control technique is not often applied.

Figure 6.7 shows the cumulative soil loss and the net sediment transport along the reference hillslope, when the soil if for 50 % covered with a mulch. The soil loss towards the drainage system is in this case 1.482 ton, which represents an efficiency of only 0.1 %. When the soil would be covered for 90 % (figure 6.8), the simulated soil
loss is further reduced to 1.306 ton, which represents an efficiency of 12%.

These results obtained with the STM-2D model indicate that the erosion reduction effect of mulches is only a fraction of that of the vegetation cover. This can be confirmed with results of some erosion plots in the Kemmelbeek watershed. Table 6.3 summarizes the observations on some erosion plots on a loamy field in the Kemmelbeek watershed. From that table it is clear that the mulch (green compost, 30 ton·ha⁻¹) had no effect in reducing the erosion from a potato field.

Figure 6.7: Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with a mulch having a soil cover of 50 %.

Figure 6.8: Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with a mulch having a soil cover of 90 %.

Mulching can have an indirect long-term effect on the erosion reduction because of the increase in organic material. This technique can be used in combination with other erosion control measures, but not as the only measure.
Table 6.3: Results of some plots with a mulch (green compost, 30 ton·ha\(^{-1}\)) and untreated demonstration plots inside a potato field on a loamy field in the Kemmelbeek watershed. These results are the soil losses observed in the period from 11 May 1997 to 27 September 1997. The total rainfall in this period was 347 mm.

<table>
<thead>
<tr>
<th>Plot</th>
<th>Slope length</th>
<th>Rainfall (l/plot)</th>
<th>Runoff (l/plot)</th>
<th>Runoff (%)</th>
<th>Soil loss (ton/ha)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mulch 1</td>
<td>10</td>
<td>4159</td>
<td>185</td>
<td>4.45</td>
<td>2.50</td>
</tr>
<tr>
<td>Mulch 2</td>
<td>10</td>
<td>4159</td>
<td>128</td>
<td>3.08</td>
<td>1.74</td>
</tr>
<tr>
<td>Control 1</td>
<td>10</td>
<td>4159</td>
<td>115</td>
<td>2.77</td>
<td>1.51</td>
</tr>
<tr>
<td>Control 2</td>
<td>10</td>
<td>4159</td>
<td>154</td>
<td>3.70</td>
<td>2.90</td>
</tr>
</tbody>
</table>

6.3.2 Soil management measures

The aim of a sound soil management is to maintain the fertility and structure of the soil. A good fertility and structure can be assured by appropriate tillage by maintaining the organic matter content at a high level and by soil stabilizers.

6.3.2.1 Tillage practices

Wheel tracks are frequently zones of concentrated erosion. The pattern of compaction depends upon tyre pressure, the width of the wheels and the speed of the tractor. The latter controlling the contact time between the wheel and the soil. To reduce the compaction and thus the resulting infiltration problems some tillage control measures can be applied: no tillage, minimum tillage and strip tillage.

Random roughness is directly linked with tillage operations. If the soil is roughly ploughed before the start of the autumn–winter period the random roughness features induce a very high retention capacity: the possible runoff water is trapped. Figure 6.9 gives simulated results if the random roughness Hurst coefficient (see also section 5.3.1.5) is set to 0.1 and the maximum height of the roughness features is 0.12 m. The simulated soil loss was in this example 0.124 ton, which is a soil loss reduction of 92 %. However, the random roughness features are smoothed by raindrop impact and the water trapping efficiency becomes gradually lower. Figure 6.10 shows an example when the Hurst coefficient is reduced to 0.6. The simulated soil loss was in this case
1.311 ton, which is an efficiency of 12 \%.

At the beginning of the winter period, erosion can be strongly reduced when a field
is very roughly tilled. However, this should not be the only soil conservation tech-
nique because the retention capacity is gradually diminished by the destruction of the
aggregates by raindrop impact.

Figure 6.9: Cumulative erosion (left) and net sediment transport (right) on a field
(100 m long, 100 m wide and 10 \% slope) with a random roughness Hurst coefficient
of 0.1 and a maximum height of the roughness features of 0.12 m.

Figure 6.10: Cumulative erosion (left) and net sediment transport (right) on a field
(100 m long, 100 m wide and 10 \% slope) with a random roughness Hurst coefficient
of 0.6 and a maximum height of the roughness features of 0.12 m.
6.3.2.2 Organic matter

Table (6.4) and table (6.5) give reference values of the organic carbon content for Belgian agricultural and pasture fields respectively (Vanongevel et al., 1995). Organic matter improves the cohesiveness of the soil, increases its water retention capacity and promotes a stable aggregate structure. However, increasing the organic matter content of a soil is a long-term process (Van Meirvenne et al., 1996), consequently the effects of an increase in organic carbon content will only be visible after some decades.

Table 6.4: Organic carbon content (%) of Belgian agricultural fields (Vanongevel et al., 1995).

<table>
<thead>
<tr>
<th></th>
<th>Sandy soils</th>
<th>Silty soils</th>
<th>Clay soils</th>
</tr>
</thead>
<tbody>
<tr>
<td>very low</td>
<td>&lt; 1.00</td>
<td>&lt; 0.70</td>
<td>&lt; 0.90</td>
</tr>
<tr>
<td>low</td>
<td>1.00 – 1.30</td>
<td>0.70 – 0.90</td>
<td>0.90 – 1.10</td>
</tr>
<tr>
<td>rather low</td>
<td>1.30 – 1.55</td>
<td>0.90 – 1.05</td>
<td>1.10 – 1.40</td>
</tr>
<tr>
<td>normal</td>
<td>1.55 – 2.50</td>
<td>1.05 – 1.50</td>
<td>1.40 – 2.30</td>
</tr>
<tr>
<td>rather high</td>
<td>2.50 – 4.00</td>
<td>1.50 – 2.70</td>
<td>2.30 – 4.00</td>
</tr>
<tr>
<td>high</td>
<td>4.00 – 8.60</td>
<td>2.70 – 6.00</td>
<td>4.00 – 8.60</td>
</tr>
<tr>
<td>peaty</td>
<td>&gt; 8.60</td>
<td>&gt; 6.00</td>
<td>&gt; 8.60</td>
</tr>
</tbody>
</table>

Table 6.5: Organic carbon content (%) of Belgian pasture fields (Vanongevel et al., 1995).

<table>
<thead>
<tr>
<th></th>
<th>All soils except silty soils</th>
<th>Silty soils</th>
</tr>
</thead>
<tbody>
<tr>
<td>very low</td>
<td>&lt; 1.70</td>
<td>&lt; 1.30</td>
</tr>
<tr>
<td>low</td>
<td>1.70 – 2.60</td>
<td>1.30 – 1.80</td>
</tr>
<tr>
<td>rather low</td>
<td>2.60 – 3.10</td>
<td>1.80 – 2.25</td>
</tr>
<tr>
<td>normal</td>
<td>3.10 – 4.80</td>
<td>2.25 – 3.70</td>
</tr>
<tr>
<td>rather high</td>
<td>4.80 – 6.10</td>
<td>3.70 – 5.70</td>
</tr>
<tr>
<td>high</td>
<td>6.10 – 8.60</td>
<td>5.70 – 7.80</td>
</tr>
<tr>
<td>peaty</td>
<td>&gt; 8.60</td>
<td>&gt; 7.80</td>
</tr>
</tbody>
</table>

The organic matter content of the reference field used in the simulations with the STM-2D program was set to a normal (for Belgian agricultural fields on silty soils) value of 1.2 %. The erosion reduction when the organic carbon content would be 4 % is given in
figure 6.11. The simulated soil loss towards the drainage system was 1.175 ton, which is an erosion reduction of 21 % when compared with the reference simulation. This indicates the importance of maintaining the organic matter content at a high level.

![Graphs showing cumulative erosion and net sediment transport](image)

Figure 6.11: Cumulative erosion (left) and net sediment transport (right) on a field (100 m long, 100 m wide and 10 % slope) with an organic carbon content of 4 %.

### 6.3.2.3 Soil stabilizers

Improvements in soil structure can be achieved by applying soil conditioners (Gabriels, 1974). These may take the form of organic by-products, polyvalent salts and various synthetic polymers. Polyvalent salts such as gypsum bring about flocculation of the clay particles while organic by-products and synthetic polymers bind the soil particles into aggregates. Synthetic polymers however are not relevant in agricultural applications.

### 6.3.3 Mechanical methods

The mechanical erosion control measures are used to control the movement of the overland flow. As mechanical measures one can quote: contouring, terraces, waterways, stabilization structures and geotextiles.
6.3.3.1 Contouring

Ploughing, planting and cultivation on the contours can reduce soil loss from sloping land compared with cultivation up-and-down the slope. In the public media, this is often considered as the ultimate erosion control measure. However, De Ploey (1989) warns that this technique is only effective during storms of low rainfall intensity and might induce severe gully erosion for high intensity rainfalls on fields where the runoff water can concentrate along the contours. Also, contour ridging does not provide sufficient drainage under wet conditions, thus resulting in a higher runoff coefficient for the subsequent rainfall events. Due to the use of heavy machinery and the often wet conditions at harvest time, it is not advisable to promote this technique.

6.3.3.2 Terraces

This erosion control technique is not compatible with heavy machinery. Furthermore, environmental organizations tend to protect the landscape. Therefore, terraces as an erosion control measure is not relevant in Flanders.

To illustrate how the sediment transport along complex slopes is calculated with the STM-2D model, figure 6.12 shows the cumulative transport along a convex, concave and convex-concave hillslope. From this figure it is clear that fields with a concave lower part will feature a much lower soil loss than convex hillslopes. The simulated soil losses for the convex, concave and convex-concave hillslope presented in figure 6.12 are respectively: 6.270 ton, 0.1752 ton and 0.0684 ton. Each of these hillslopes had a mean slope of 10 %. 
Figure 6.12: Cumulative erosion (right) along a convex (upper plot), concave (middle plot) and convex-concave (lower plot) hillslope (100 m long, 100 m width). The mean slope of each hillslope is 10 ‰.
6.3.3.3 Waterways

The purpose of waterways in a conservation system is to convey runoff at a non-erosive velocity to a suitable disposal point. In larger fields, where the runoff water can concentrate considerably, there is a high risk for gully development. These natural waterways must be protected with a permanent grass vegetation. Grassed waterways are often applied in the USA, where most agricultural fields are huge, compared with the fields in Flanders. In most agricultural watersheds in Flanders, the first order waterways in the river system, are incorporated in the ditch system. However, most farmers, if not all, cultivate their land too close to the waterways. This causes often a considerable amount of bank erosion. This can be prevented by the introduction of permanent grassed riparian buffer zones.

6.3.3.4 Stabilization structures

Stabilization structures play an important role in gully reclamation and gully erosion control. These stabilization structures are usually small dams which are built across gullies to trap sediment and thereby reduce channel depth and slope. This erosion control measure is only necessary when a field suffers from severe on-site gully erosion, which is rare in Belgium.

6.3.3.5 Geotextiles

A geotextile is any permeable textile material used with foundation, soil, rock, earth or geotechnical engineering-related material. It can be in the form of a mat, sheet, grid or web of either natural fiber, such as jute or coir, or artificial fiber, such as nylon (Morgan, 1996). This erosion control measure is not relevant in agricultural application.

6.4 Conclusions

From the analysis of the different erosion control measures, it can be concluded that a dense vegetation cover is the most effective erosion control measure. If possible, the cultivation of off-season cover crops should therefore always be considered. After the
harvest of some crops, in particular maize and beets, no cover crop can develop a dense cover before the start of the autumn-winter period. Grasped riparian buffer zones are then the most promising permanent erosion control measure.

From the modelling results and the literature reviews, it can be concluded that dense grassed buffer strips with a width of 5 m are effective enough to decrease the sediment transport towards the drainage network with 50 to 80 %. However, this width of 5 m is dimensioned for single rainfall events, and is therefore a minimum. The necessary width of the buffer zone to give sufficient protection during the whole off-season (late autumn, winter and early spring) must be calculated based on its sediment storage capacity and the erosion/deposition risk of the agricultural field. Some field experiments are now performed in the Kemmelbeek watershed to convince farmers of the importance of permanent buffer zones. The first observations seem to support this buffering capacity of 50 to 80 % for single rainfall events, but also indicate that 5 m gives not enough storage capacity in very wet years. However, given the small field dimensions in Flanders it is not realistic to propose wider permanent buffer zones. The use of permanent grassed buffer zones should therefore be promoted, always in combination with other — more on-site related — erosion control measures like cover crops, undersowing of cover crops in late harvested production crops and adjusted tillage techniques (creating a high random roughness, avoiding soil compaction) to reduce the runoff amount.

Given the actual agricultural boundary conditions in Flanders (small fields and the use of heavy machinery), erosion control measures like strip cropping or terracing are economically nor technically possible. Due to the heavy machinery and the often not optimal moisture conditions when a field is tilled or harvested and because of the possible risk to initiate gully erosion, it is not advisable to promote contouring. Ensuring a high random roughness is a more adequate measure to block the runoff water than contouring: the possible runoff is retained in local cavities, such that concentration of runoff water is prevented. The risk of gully formation is then strongly reduced.
Modelling is an indispensable tool in all sciences to predict the behavior of natural or man-made systems. It enables the evaluation of the response of the system itself and to perform scenario analysis. Modelling is sometimes the only tool we have if the scenarios can not be tested in practice, because of too expensive, too risky and/or too time-consuming. This work was focused on the modelling of erosional processes, to support soil and water conservation planning in Flanders.

In a first stage, the erosion risk mapping was done using the RUSLE (Revised Universal Soil Loss Equation) model (Renard et al., 1996). This model estimates the long-term soil loss. Basically, the RUSLE is a 2D hillslope model, but it was extended with a new routing algorithm and with the Monte Carlo error propagation technique, to adapt it to a 3D reality and to incorporate the uncertainty of the input parameters in the model analysis. The Kemmelbeek watershed in the south of West-Flanders (community of Heuvelland), was the pilot catchment to test the RUSLE modelling results. This watershed, with a total area of 1075 ha, drains in a drinking-water reservoir, of which the long-term sediment input was known.

The calculated yearly sediment input in the reservoir was 4 376 ton·year$^{-1}$ with a standard deviation of 75 ton·year$^{-1}$. Given the uncertainty of the model input parameters, the predicted average yearly sediment input, and its 68 % confidence interval was 4 376 ± 75 ton·year$^{-1}$. The average observed sediment input between 1936 and 1982 was 4 438 ton·year$^{-1}$ (Gabriels, 1985). This value lies within the 68 % confidence interval, the model output and field truth differed only 1.4 %. From these results, there might be concluded that the RUSLE model in combination with the methods presented in chapter 4, in particular the adaptation of the linear RUSLE model to a 3D reality, might be capable to predict both long-term on-site soil losses (the sediment leaving each field, the RUSLE can not be used to estimate the intrabasinal sediment storage.
as colluvium) and long-term off-site sediment accumulation within an acceptable level of accuracy.

However, the RUSLE does not account for all possible erosion control measures. To have a uniform methodology to design alternative land management scenarios and to analyze their effects, an alternative model was a necessity. Many physical or process-based models are available, but most are very complex and very gluttonous for data what makes it difficult to apply them in practice. Therefore the EROSION-2D/3D model (Schmidt, 1996) was selected, since that model uses only the most significant parameters governing the erosional processes. The necessary data to run the EROSION-2D/3D model is almost the same as the data set needed for the RUSLE model.

The second stage of this research work was focused on the elaboration of a physical based erosion model. Due to recent findings in the experimental and physical description of soil detachment and soil transport (e.g.: Nearing et al., 1997) the necessity was felt to change and add some components. Some major changes were introduced in the hydrological submodel and the soil transport functions. Therefore, it was decided to identify this modified EROSION-2D/3D model as STM-2D/3D (Sediment Transport Model, which is implemented in a 2D hillslope and 3D watershed version). The hydrological algorithms are based on the combination of a modified version of the Green-Ampt equation (which is capable to predict runoff for unsteady rain events) and an implicit finite difference scheme of the kinematic wave. These algorithms were tested in a small catchment of 142 ha, located inside the Kemmelbeek watershed. At the outlet of this watershed, rainfall and discharge were measured at a 10 minute interval.

It was the intention to build a model wherein the user can manipulate only a very limited number of parameters. All critical parameters (saturated and residual moisture content, hydraulic conductivity, matric potential) are derived using pedotransfer functions (based on the basic soil physical characteristics: texture, organic carbon content and soil density). The only parameters the user can modify during event simulations are the initial moisture content and the Manning roughness coefficients (only the minimum and maximum value inside the whole catchment). By doing so, it could be checked if the general reasoning corresponds to what can be observed in the field (outlet hydrographs, erosion patterns). If every variable can be changed, the model
can be tuned such that always good event simulations are possible.

From an extensive data set of laboratory rainfall simulations, a sediment transport function based on the stream power concept was created for sheet and micro-rill flow with a high $R^2$ coefficient of 0.89. For rill and gully flow, a similar equation was available from literature (with an $R^2$ of 0.93). Using these functions, the unit sediment load can thus accurately be estimated if the discharge of the flow is known. Therefore, the attention of this research was focused on the calibration and validation of the hydrological component (estimation of the overland flow and channel discharge) of the STM-2D/3D model.

To calibrate the model, the initial moisture content was considered as the calibration parameter because this parameter is the most sensitive parameter which governs the amount of infiltration and runoff. From calibrated event simulations the following can be concluded regarding the general concept of the hydrological submodel of the STM-2D/3D model: (1) The concept of determining the initial moisture content solely based on soil physical characteristics (soil texture, soil density and organic matter content) appeared to be an acceptable approximation for summer and early autumn rainfall events. (2) Because of the slow reaction on rainfall of the rising limb of the simulated hydrograph for winter events there can be concluded that not all partial runoff contributing areas are determined by the initial moisture concept of the STM-2D/3D model. In the winter season (November to February) there will be some partial contributing areas close to the river/brook system (necessary for a fast response of the rising limb), due to the moisture redistributing effect of surface topography (diverging/converging topography elements) and subsurface topography of low conductivity soil layers (e.g.: plough/traffic pans, Tertiary clay layers, areas affected by surface sealing). (3) Subsurface flow is much more intense during the winter months than in the summer period. Because the STM-2D/3D model has no subsurface flow component and the baseflow was kept constant during the whole rainfall event, the model efficiency values stressed on the performance of the lower discharges are lower in winter than in summer. However, this component should only be taken into account if one wants to use the model for continuous simulations. For single event simulations, a subsurface flow component does not provide additional information regarding to sediment transport.
It was found that the STM-2D/3D model generates plausible and qualitatively good results based on the position of the predicted and observed deposition areas and the general erosion pattern for moderate (both in intensity and rainfall amount) rainfall events. The generated erosion pattern for these events corresponded with the soil erodibility pattern of the RUSLE model. The plausibility of the results generated with the STM-2D/3D model was also tested for the different erosion control techniques and some of these simulations are qualitatively compared with erosion plots observations inside the Kemmelbeek watershed.

On an event basis, there was no correlation between the RUSLE and the STM-2D/3D model both in absolute potential soil loss per agricultural parcel and the spatial erosion pattern. This is because the antecedent moisture content is not taken into account in the RUSLE model. The current version does not provide a method which takes into account the rainfall and soil moisture regime. Overpredicting the soil loss during dry soil moisture conditions and underpredicting the soil loss during wet soil moisture conditions, compensates errors when the RUSLE is used on a yearly, or long-term (decades) basis. This is why still a good correlation is found (Chapter 4) between the RUSLE-predicted soil loss aggregated for the whole Kemmelbeek watershed (4376 ± 75 ton-year⁻¹) and the long-term sediment input (4438 ton-year⁻¹) in a water reservoir. The comparison analysis of both models indicated that the RUSLE can not be used on a time resolution shorter than a year, which is an important consequence when analyzing the effect of seasonal erosion control measures (e.g. off-season cover crops, winter mulching) is the objective.

From the analysis of the different erosion control measures, it can be concluded that a dense vegetation cover is the most effective erosion control measure. If possible, the cultivation of off-season cover crops should therefore always be considered, in particular grasses (Lolium sp.). After the harvest of some crops, especially maize and beets, no cover crop can develop a dense cover before the start of the autumn-winter period. In these situations, a combination of several erosion control techniques is necessary to obtain the same results as dense cover crops. As a permanent erosion control measure, the introduction of riparian buffer zones with a width of at least 5 m and a dense grass vegetation might be considered. However, field observations indicate that 5 m gives not enough sediment storage capacity in very wet years. Unfortunately, given the small
field dimensions in Flanders it is not realistic to propose wider permanent buffer zones. Therefore, land planners and policy makers have an important role when a land area is restructured. If erosion control measures (like riparian buffer zones) are not taken into account in the planning phase, it becomes very difficult to promote them afterwards.
Samenvatting en Besluiten

Binnen alle wetenschappen is modelleren een onvervangbaar hulpmiddel om het gedrag van een natuurlijk of artificieel systeem te voorspellen. Het laat toe de respons van het systeem zelf te evalueren en om scenario-analyses uit te voeren. Modellering is soms het enige hulpmiddel als deze scenario’s niet kunnen uitgetest worden omdat ze ofwel te duur zijn, te riskant en/of te tijdrovend. Dit werk was gericht op het modelleren van erosieprocessen, ter ondersteuning van het bodem- en waterbeleid in Vlaanderen.

In een eerste fase van het onderzoek werd het RUSLE (Revised Universal Soil Loss Equation) model (Renard et al., 1996) gebruikt ter kartering van het erosieisico. Met dit model kan een schatting gemaakt worden van het lange termijn bodemverlies. De RUSLE is origineel ontworpen als een tweedimensionaal model. Om het model aan te passen aan de driedimensionale werkelijkheid en om de invloed op het eindresultaat na te gaan als gevolg van de onzekerheden van de inputparameters, werd het respectievelijk uitgebreid met een nieuw stromingsalgoritme en met de Monte Carlo foutenpropagatie techniek. Het Kemmelbeek stroombekken in het zuiden van de provincie West-Vlaanderen (gemeente Heuvelland) was het testgebied ter evaluering van de resultaten gegenereerd met de RUSLE. Dit stroombekken met een totale oppervlakte van 1075 ha mondt uit in een drinkwater reservoir waarvan de lange termijn sedimentdepositie gekend was.

De jaarlijkse berekende sedimentdepositie in het reservoir was 4 376 ton·jaar⁻¹ met een standaard afwijking van 75 ton·jaar⁻¹. Gegeven de onzekerheid op de inputparameters, was de jaarlijkse voorspelde sedimentdepositie en het 68 % betrouwbaarheidsinterval 4 376 ±75 ton·jaar⁻¹. De gemiddelde waargenomen sedimentdepositie tussen 1936 en 1982 was 4 438 ton·jaar⁻¹ (Gabriels, 1985). Deze waarde ligt dus binnen het 68 % betrouwbaarheidsinterval. De modelresultaten en de geobserveerde werkelijkheid tonen een verschil van slechts 1.4 %. Aan de hand van deze resultaten zou besloten
kunnen worden dat met het RUSLE model in combinatie met de methoden beschreven in hoofdstuk 4, in het bijzonder de aanpassing van het model aan een 3D werkelijkheid, het mogelijk is om het lange termijn bodemverlies (het sediment dat de velden daadwerkelijk verlaat, de RUSLE kan niet gebruikt worden om de depositie als colluvium binnen een stroombekken te schatten) en de lange termijn sedimentaccumulatie met een aanvaardbare nauwkeurigheid te voorspellen.

Het is echter niet mogelijk om met het RUSLE model alle bestaande erosiebeteugelingsmaatregelen te evalueren. Een alternatief model was dus noodzakelijk omdat het ontwerpen van alternatieve beheersmaatregelen een uniforme methodologie vergt. Er zijn veel fysische of proces-gebaseerde modellen beschikbaar, maar velen zijn zeer complex en zijn veelal te gulzig wat betreft inputgegevens. Hierdoor wordt het dan ook moeilijk om dergelijke modellen op een regionale schaal toe te passen. Daarom werd voor het EROSION-2D/3D model (Schmidt, 1996) geopteerd, daar dit model is opgebouwd enkel rond de meest bepalende parameters betreffende de erosieprocessen. De hoeveelheid en de aard van gegevens nodig om een analyse uit te voeren met het EROSION-2D/3D model is ongeveer identiek aan de gegevens nodig voor het RUSLE model.

De tweede fase van dit onderzoek was dus gericht op de ontwikkeling van een model gebaseerd op fysische principes. Door recente ontwikkelingen in de beschrijving van de erosieprocessen (bv: Nearing et al., 1997), werden enkele belangrijke wijzigingen doorgevoerd in het originele model. Deze wijzigingen betreffen het hydrologische submodel en de sediment transportvergelijkingen. Daarom werd er besloten om dit gewijzigde EROSION-2D/3D model als STM-2D/3D (Sediment Transport Model, geïmplementeerd in een 2D hellings- en 3D stroombekken versie) aan te nemen. Het hydrologische submodel is gebaseerd op een gemodificeerd Green-Ampt infiltratiemodel (met de mogelijkheid om de runoff te berekenen tijdens onregelmatige regenbuien) en op een impliciet eindig differentieschema van de kinematische golf. Deze algoritmes werden getest in een klein stroombekken van 142 ha, gelocaliseerd binnen het stroombekken van de Kemmelbeek. Aan de monding van dit stroombekken werd continu (tijdsinterval van 10 minuten) de neerslag en het hieraan gekoppelde debiet in de beek gemeten.

Het was de bedoeling om een model te bouwen waarin de gebruiker maar een beperkt aantal parameters kan manipuleren. Alle voor het model critische parameters (verzadigd
en residuoel vochtgehalte, hydraulische conductiviteit, matrix potentiaal) worden geschat
aan de hand van pedotransferfuncties (gebaseerd op de elementaire bodemfysische
gegevens: textuur, organisch materiaal en bodemdichtheid). De enige parameters die
de gebruiker kan wijzigen is het initiële vochtgehalte en de Manning ruwheidscoefficien-
ten (enkel de minimum en maximum waarde binnen het gehele stroombekken). Daar
dus enkel het initiële vochtgehalte als calibratieparameter dient, kon gecontroleerd wor-
den of de algemene fysische beschrijving van de processen in het model overeenkwam
met de veldwaarnemingen (hydrogrammen, erosiepatronen). Indien elke parameter
can gewijzigd worden, kan het model zodanig afgesteld worden dat altijd een goede
simulatie mogelijk is.

Uitgaande van een uitgebreide dataset van labo-regenvalssimulaties, kon een sediment
transportfunctie voor ‘sheet’ en micro-rill stromingen bepaald worden met een hoge
R$^2$ van 0.89. Voor stromingen in rills en geulen was een gelijkaardige transportfunc-
tie beschikbaar vanuit de literatuur (met een R$^2$ van 0.93). Aan de hand van deze
transportfuncties kan de sedimentlading nauwkeurig bepaald worden indien het debiet
de stroming gekend is. Daarom werd de aandacht gericht op het calibreren en
valideren van de hydrologische component (het schatten van debieten) van het STM-
2D/3D model.

Ter calibratie van het model werd het initiële vochtgehalte als calibratieparameter
beschouwd daar deze parameter de meest gevoelige is betreffende de hoeveelheid infil-
tratie en runoff. Aan de hand van gecalibreerde simulaties kon het volgende besloten
worden in verband met het algemene concept van de hydrologische component binnen
het STM-2D/3D model: (1) Het concept om het initiële vochtgehalte volledig te laten
afhangen van bodemfysische parameters (textuur, dichtheid en organisch stofgehalte)
bleek een aannembare vereenvoudiging voor regenbuizen in de zomer en vroege herfst.
(2) Door de drage reactie op regenval van de stijgende tak van het gesimuleerde hy-
drogram voor regenbuizen in de winterperiode kan besloten worden dat niet alle tot
de runoff bijdragende gebieden kunnen voorspeld worden door het in het STM-2D/3D
model gehanteerde concept betreffende het initiële vochtgehalte. In het winterseizoen
(November tot en met Februari) zullen er enkele verzadigde gebieden zijn die dicht
bij de beek gelegen zijn (nodig voor een snelle reactie van de stijgende tak van het
hydrogram), door het vocht hervorderende effect van de topografie van het oppervlak
(divergerend/convergerend reliëf) en de topografie van ondergrondse lagen met een lage conductiviteit (ploegzolen, Tertiaire kleilagen). (3) De ondergrondse waterstroming is veel intensiever in de winter dan in de zomer. Daar het STM-2D/3D model geen component bezit ter beschrijving van dit proces en daar het basisdebiet constant gehouden werd tijdens de simulaties, was de model efficientie voor de lage debieten minder goed in de winter dan in de zomer. Echter, met deze component moet slechts rekening gehouden worden indien continue simulaties wenselijk zijn. Voor simulaties van enkelvoudige regenbuien, geeft een submodel dat de ondergrondse waterstroming inschat geen extra informatie betreffende het bodemtransport.

Er werd gevonden dat het STM-2D/3D model plausibele en kwalitatief goede resultaten kan genereren. Deze kwalitatieve validatie was gebaseerd op de positie van de geobserveerde en berekende depositiezones en het algemene erosiepatron voor gematigde (zowel in intensiteit als regenhoeveelheid) regenbuien. In hoofdstuk 6 werd aangetoond dat het STM-2D/3D model plausibele resultaten kan genereren voor de verschillende erosiebeteugelingsmaatregelen. Sommige van deze simulaties werden tevens kwalitatief vergeleken met observaties van enkele proefvelden binnen het Kemmelbeek stroombekken.

Wanneer het RUSLE en STM-2D/3D model op een tijdschaal van afzonderlijke regenbuien vergeleken werden, werd geen correlatie gevonden tussen beide modellen noch betreffende het potentiële bodemverlies noch het ruimtelijke erosiepatroon. De reden hiervoor is dat het vochtgehalte van de bodem niet in rekening gebracht wordt binnen het RUSLE model. De huidige versie van dit model heeft geen methode beschikbaar om het regionale bodemvochtregime in rekening te brengen. Door het bodemverlies tijdens droge bodemvochttoestanden te overschatten en het bodemverlies tijdens natte bodemvochttoestanden te onderschatten, worden de fouten die de RUSLE maakt gecompenseerd wanneer het model op een jaarlijkse basis of lange termijn (decaden) tijdsniveau gebruikt wordt. Daarom werd toch een goede correlatie gevonden tussen het door het RUSLE voorspelde bodemverlies geaggregeerd over het gehele Kemmelbeek stroombekken (4376 ±75 ton·jaar⁻¹) en de lange termijn sedimentatie (4438 ton·jaar⁻¹) in een waterreservoir. Deze vergelijking van beide modellen wees erop dat de RUSLE dus niet kan gebruikt worden op een tijdsniveau korter dan een jaar. Dit heeft als gevolg dat het erosiebeteugelend effect van seizoenale beheersmaatregelen
zoals buiten-seizoen groenbemesters met de RUSLE niet kan begroot worden.

Uit de analyse van de verschillende erosiebeteugelingsmaatregelen kon besloten worden dat een dichte vegetatie het meest efficiënt is. Indien mogelijk moet de teelt van een groenbemester altijd overwogen worden, vooral grassen (Lolium sp.). Na de oogst van sommige gewassen, voornamelijk mais en bieten, kan geen enkele groenbemester een nog voldoende dichte bedekking ontwikkelen voor de start van het herfst-winter seizoen. In dergelijke situaties is een combinatie van verschillende erosiebeteugelingsmaatregelen noodzakelijk om dezelfde graad van bescherming te leveren als een groenbemester. Als een meer permanente maatregel, kan de introductie van buffer-zones met een minimum breedte van 5 m en een dichte grasvegetatie langs de grachten, beken en rivieren overwogen worden. Veldobservaties toonden echter aan dat 5 m niet voldoende opslagruimte biedt voor het sediment tijdens regenrijke seizoenen. Het is echter weinig realistisch om bredere zones voor te stellen gegeven de kleine perceelsoppervlakte in Vlaanderen. Daarom hebben landschapsplanners en beleidsverantwoordelijken een belangrijke taak wanneer het beschikbare land geherstructureerd wordt. Wanneer erosiebeteugelingsmaatregelen niet in acht genomen worden tijdens de planningsfase (zoals buffer zones), is het zeer moeilijk om deze achteraf te promoten.
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Appendix A

STM-2D Program

In this appendix the JAVA source code is given of the 2D hillslope version of the STM-2D/3D model. This JAVA applet can be integrated in an Internet environment (WWW) and can be used to design erosion control measures (terracing, buffer zones, contour ploughing, altered random roughness, strip cropping, changing organic matter content, etc ...). Graphs can be created of the segment hydrographs, water level, runoff velocity, runoff Reynolds number, runoff stream power, the cumulative erosion/deposition along a hillslope and the net segment erosion/deposition.

One of the objectives in the development of the STM-2D program code was to make the general layout of the model object oriented, according to the structure of modern program languages (C++, Java): the different model components (the infiltration/runoff model, the overland flow routing model, the sediment transport functions, the data input routines and the component for result display) form separate objects. The physical model presented in this work is not universal. For example, if one has to evaluate the erosion risk and alternative land management scenarios in a region where the main soil types are Vertisols, the Green-Ampt infiltration model will not generate good results. An infiltration model which can deal with crack formation is in such case necessary. Because of the object oriented program structure and because the model is based on physical principles, model components can be replaced without violating the model structure. Using this approach, the end-user can change or add model objects according to his/her expert knowledge.

Disclaimer: The source code is free of use and modification as on the condition that the author is cited. In no event the author shall be liable to anyone for any delays, inaccuracies, errors or omissions with respect to the use of this software, for any damage arising therefrom or occasioned thereby, or for the results obtained. The entire risk as to the quality and performance of the software and the accuracy, adequacy, completeness, validity and quality of any information is with the user.
STM_2D.java

This is the main program. STM stands for ‘Sediment Transport Model’. The program uses ‘threads’ (light-weight processes controlled by the parent process, a concept developed to enhance multi-tasking) to manage the input, output and calculation modules.

```java
/**
 * This applet calculates the hydrological properties of the runoff during a user specified storm event and a user defined hillslope and segment soil properties. Using the hydrological properties, the net erosion and deposition along the hillslope can be estimated.
 */
import java.applet.*;
import java.awt.*;

public class STM_2D extends Applet {
    /************
    /** model data */
    /************
    static public STM_2D_Globaldata data = new STM_2D_Globaldata();

    /************
    /** GUI data */
    /************
    private Button define_slope;
    private Button define_rainfall;
    private Button about;
    private Panel applet_panel;

    /************
    /** Applet constructor */
    /************
    public STM_2D() {
        data.segments = 0;
        data.rseed = 0;

        define_slope = new Button("Enter/Change hillslope data");
        define_rainfall = new Button("Enter a precipitation event");
        data.calculate = new Button("CALCULATE");
        about = new Button("About");

        define_slope.setBackground(Color.gray);
        define_rainfall.setBackground(Color.gray);
        data.calculate.setBackground(Color.gray);
        about.setBackground(Color.gray);
    }

    public void paint(Graphics g) {
        /************
        /** paint output */
        /************
        data.pprint(g);

        /************
        /** update the output */
        /************
        data.update();
    }

    public void stop() {
        /************
        /** dispose the applet */
        /************
        data.tidyUp();
    }
}
```
public void init()
{
    this.resize(250,150);
}

public boolean mouseDown(Event e, int x, int y)
{
    data.rseed++;
    return true;
}

public boolean mouseDrag(Event e, int x, int y)
{
    return true;
}

public void mouseClicked(Event e)
{
    data.rseed++;
}

public boolean action(Event event, Object arg)
{
    if(event.target == define_slope)
    {
        if(data.lock != true)
        {
            Thread th1 = new Thread(new STM_2D_GetHillslopeData());
            th1.setPriority(Thread.NORM_PRIORITY);
            th1.start();
            return true;
        }
        else return false;
    }
}
if(event.target == define_rainfall)
{
    if(data.lock != true)
    {
        Thread th2 = new Thread(new STM_2D_GetPrecipitationData());
        th2.setPriority(Thread.NORM_PRIORITY);
        th2.start();
        return true;
    }
    else return false;
}

//---------------------------------------------------------------------------
if(event.target == data.calculate)
{
    if(data.calclock != true)
    {
        data.calclock = true;
        data.lock = true;
        Thread th3 = new Thread(new STM_2D_Calculate());
        th3.setPriority(Thread.NORM_PRIORITY);
        data.calculate.setBackgroundResource(Color.red);
        th3.start();
        return true;
    }
    return true;
}

//---------------------------------------------------------------------------
if(event.target == about)
{
    if(data.lock != true)
    {
        Thread th4 = new Thread(new STM_2D_About());
        th4.setPriority(Thread.NORM_PRIORITY);
        th4.start();
        return true;
    }
    else return false;
}

//---------------------------------------------------------------------------
else return super.action(event, arg);
}

//**********************************************************************************/
STM_2D_Globaldatadata.java
This program defines all data that must be visible for all subcomponents of the STM_2D applet.

import java.awt.*;

public class STM_2D_Globaldatadata
{
    public Button calculate  ;// Calculate button
    public boolean lock = false ;// Thread lock
    public boolean calclock = false ;// Calculating thread lock
    public int segments ;// segment counter
    public double[] dx, dy ;// segment dimensions [m]
    public double[] S ;// segment slope [%]
    public double[] sand, silt, clay ;// soil texture [%]
    public double[] corg ;// organic carbon [%]
    public double[] soil_d ;// soil density [kg/m**3]
    public double[] MI  ;// moisture content [vol %]
    public double[] Intercep  ;// interception capacity [m]
    public double[] cover ;// soil cover [%]
    public double[] pcover ;// vegetation cover [%]
    public double[] RWF ;// ridge wave period [m]
    public double[] RWH ;// ridge wave height [m]
    public double[] Hurst ;// random roughness hurst coeff [-]
    public double[] RRH ;// random roughness height [m]
    public double[] Ieff ;// effective precipitations [mm/h]
    public double[] retc ;// retention capacity [m]
    public double pds  ;// pluviophase time step [s]
    public long rseed ;// random number generator seed value
}
STM_2D_Calculate.java

This class does the actual calculating. After finishing the calculations, a GUI (Graphical User Interface) is created to query the results per segment.

```java
public class STM_2D_Calculate extends Frame implements Runnable {
    // ...
}
```

import java.awt.*;
import java.math.*;
import java.util.*;
import java.text.*;
import java.awt.event.*;
```java
private boolean running = true ;// thread control
private boolean stop ;// loop control variable
private int ss,SS,tt,ii,i,r ;// segment and time step counter
private int event_steps ;// number of time steps
private int wdt, TT ;// timestep length in seconds [s]
private int SEGMENTS ;// number of 1m segments
private double total_time ;// total calculation time
private double h1,h2,h3,h4,help=0.0 ;// help variables
private double h5,h6,h7,h8,h9 ;// other help variables
private double viscos=0.0010091802 ;// dynamic viscosity
private double SOIL_LOSS ;// total soil loss [kg]
private double[] X, Y ;// elevation data
private double[] plotX,plotY,plotY2 ;// plot data
private double[] Runoff ;// cumulative runoff amount [m]
private double[] Mann ;// Manning coefficients
private double[] a, b ;// channel parameters
private double[] slope ;// slope data
private double[] EROSION ;// erosion/deposition [kg/segment]
public double[] ERO, DEPO ;// erosion/deposition [kg/m**2]
private double[] q ;// Green-Ampt runoff [m**3/s]
public double[] level ;// overland flow water level [m]
private double[] Ck ;// overland flow wave celerity [m/s]
private double[] diam, pperc ;// particle diameter, textural %
private double[] no_depos ;// particles kept in suspension [%]
private double[] erosion ;// sediment transport
public double[] Q, QQ ;// discharge [m**3/s]
```
public double[][] Re; // Reynolds number [-]
private double[][] stream_power; // Stream power of runoff [g/s**3]
private Random RAN = new Random(); // reference to java.util.Random
private STM_2D_GreenAmpt G_A; // reference to Green_Ampt class

/******************************************
 *** GUI components
 ******************************************/
private Choice program_choices;
private Panel p1, p2;
private Button done, plot;
private STM_2D_DoubleTextField segmentDTF;
public STM_2D_List clist = new STM_2D_List();

/******************************************
 *** Constructor
 ******************************************/
public STM_2D_Calculate()
{
    //-------------------------
    //--- Add a window listener ---
    //-------------------------
    addWindowListener(new WindowAdapter()
    {
        public void windowClosing(WindowEvent e)
        {
            STM_2D.data.calculate.setBackground(Color.gray);
            STM_2D.data.calclock = false;
            STM_2D.data.lock = false;
            dispose();
        }
    });
}

/******************************************
 *** Thread run method
 ******************************************/
public void run()
{
    if((STM_2D.data.segments > 0) && (STM_2D.data.pdt != 0.0))
    {
        //-------------------------
        //--- Make the GUI components ---
        //-------------------------
        this.setBackground(Color.darkGray);
        this.setForeground(Color.white);
        this.setTitle("STM-2D Processing and Results");
        p1 = new Panel();
        clist.setBackground(Color.black);
        clist.setForeground(Color.white);
        p1.add(clist);
        add(p1, "North");
        setSize(540,280);
        setVisible(true);
    }
}
//-- create segments of 1m length and allocate memory --
//-------------------------------------------------------------
SEGMENTS = 0;
for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
    {SEGMENTS += STM_2D.data.dx[ss] ;}

RAN.setSeed(STM_2D.data.rseed);
Mann = new double[SEGMENTS];
a = new double[SEGMENTS];
b = new double[SEGMENTS];
slope = new double[SEGMENTS];
Y = new double[SEGMENTS+1];
X = new double[SEGMENTS+1];
q = new double[15][SEGMENTS];
level = new double[15][SEGMENTS];
Ck = new double[15][SEGMENTS];

//-----------------------------
//-- set segment parameters --
//----------------------------
tt = 0;
for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
    {
        for(SS = 0 ; SS < STM_2D.data.dx[ss] ; SS++)
            {
                Mann[tt] = 0.01+(RAN.nextDouble()*(0.1-0.01));
b[tt] = 3.0/5.0;
                if(STM_2D.data.S[ss] == 0) slope[tt] = 0.0001;
                else slope[tt] = STM_2D.data.S[ss];
                help = Mann[tt]*Math.pow(STM_2D.data.dy[ss],(2.0/3.0));
a[tt] = Math.pow((help)/(Math.sqrt(slope[tt]/100.0)),b[tt]);

                tt++;
            }
    }

//-------------------------------
//-- recalculate slopes: smooth to avoid sharp segment slope breaks --
//---------------------------------------------------------------------
for(ii = 0 ; ii < 5 ; ii++)
    {
        for(ss = 0 ; ss < (SEGMENTS-1) ; ss++)
            {
                if(slope[ss] != slope[ss+1])
                    {
                        help = (slope[ss]+slope[ss+1])/2.0;
slope[ss] = help;
slope[ss+1] = help;
ss++;
                    }
            }
    }
//---------------
//-- create elevation data --
//---------------
Y[SEGMENTS] = 0.0 ;
X[0] = 0.0 ;
for(ss = (SEGMENTS-1) ; ss >= 0 ; ss--)
{
    Y[ss] = Y[ss+1] + (1.0*(slope[ss]/100.0)) ;
}
for(ss = 0 ; ss < SEGMENTS ; ss++)
{
    X[ss+1] = X[ss] + 1.0 ;
}

//---------------
//-- determine precipitation event end point --
//---------------
i = 14 ;
stop = false ;
while(stop != true)
{
    if(STM_2D.data.Ieff[i] > 0.0)
    {
        stop = true ;
        continue ;
    }
    i-- ;
    if(i < 0) stop = true ;
}
event_steps = i+1 ;
total_time = event_steps*STM_2D.data.pdt ;

//---------------
//-- calculate segment retention capacity --
//---------------
contour_ret_cap() ;
random_roughness_ret_cap() ;

//---------------
//-- calculate the Green-Ampt infiltration --
//---------------
clist.addItem("Calculate the Green-Ampt infiltration") ;
G_A = new STM_2D_GreenAmpt() ;
Runoff = new double[20] ;
ii = 0 ;
for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
{
    G_A.green_ampt(STM_2D.data.clay[ss],
                   STM_2D.data.silt[ss],
                   STM_2D.data.sand[ss],
                   STM_2D.data.corg[ss],
                   STM_2D.data.soil_d[ss]/1000.0,
                   STM_2D.data.MI[ss],
                   Runoff[

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STM_2D.data.retc[ss],
15,
STM_2D.data.pdt/3600,
STM_2D.data.Ieff,
Runoff,
STM_2D.data.Intercep[ss]) ;

for(SS = 0 ; SS < STM_2D.data.dx[ss] ; SS++)
{
    for(tt = 0 ; tt < event_steps ; tt++)
    {
        if(tt == 0) q[tt][ii] = (Runoff[tt]*1.0*STM_2D.data.dy[ss])/
STM_2D.data.pdt ;
        else q[tt][ii] = ((Runoff[tt]-Runoff[tt-1])*1.0*
        STM_2D.data.dy[ss]) / STM_2D.data.pdt ;
    }
    ii++ ;
}

//---------------------------------------------------------------
//-- Calculate finite difference time step: Courant criterium --
//---------------------------------------------------------------
wdt = 10000000000 ;
ii = 0 ;
for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
{
    for(SS = 0 ; SS < STM_2D.data.dx[ss] ; SS++)
    {
        for(tt = 0 ; tt < event_steps ; tt++)
        {
            //water level
            //--------
            help = Math.sqrt(slope[ii]/100.0)*STM_2D.data.dy[ss] ;
            level[tt][ii] = Math.pow(((Mann[ii]*q[tt][ii])/help),(3.0/5.0)) ;
            //wave celerity
            //--------
            help = (Math.sqrt(slope[ii]/100.0))/Mann[ii] ;
            Ck[tt][ii] = Math.pow(level[tt][ii],(2.0/3.0))*(3.0/5.0)*help ;
            //time step
            //--------
            if(Ck[tt][ii] > 0)
            {
                if((1.0/Ck[tt][ii]) < wdt) wdt = (int)(1.0/Ck[tt][ii]) ;
                if(wdt > (STM_2D.data.pdt/2.0)) wdt = (int)(STM_2D.data.pdt/2.0) ;
            }
        }
        ii++ ;
    }
}
wdt *= 4 ;
clist.addItem("calculation time step: "+Integer.toString(wdt)+" seconds") ;
if(wdt < 10000000000)
{
    //-------------------------------
    //--- Create boundary conditions ---
    //-------------------------------
    TT = (int)((3*total_time)/wdt);
    Q = new double[TT][SEGMENTS];
    QQ = new double[TT][SEGMENTS];
    Re = new double[TT][SEGMENTS];
    Ck = new double[TT][SEGMENTS];
    level = new double[TT][SEGMENTS];
    stream_power = new double[TT][SEGMENTS];
    for(tt = 0 ; tt < TT ; tt++)
    {
        for(ss = 0 ; ss < SEGMENTS ; ss++)
        {
            r = (int)(Math.floor((tt*wdt)/STM_2D.data.pdt));
            if(r > 14)
            {
                Q[tt][ss] = 0.0;
                QQ[tt][ss] = 0.0;
            }
            else
            {
                Q[tt][ss] = q[r][ss];
                QQ[tt][ss] = q[r][ss]/1.0;
            }
            if(ss == 0)
            {
                if(Q[tt][ss] <= 0.0) Q[tt][ss] = 0.00000001;
            }
        }
    }
    //--------------------------------------------
    //--- Implicit linear finite difference scheme of the Kinematic wave. ---
    //--- Change the following code if you want to use the diffusion wave ---
    //--- or dynamic wave model. ---
    //--------------------------------------------
    c1list.addNum("Kinematic wave implicit finite difference scheme");
    help = 0.0;
    for(tt = 0 ; tt < (TT-1) ; tt++)
    {
        for(ss = 0 ; ss < (SEGMENTS-1) ; ss++)
        {
            h1 = Q[tt+1][ss];
            if(h1 < 1E-10) h1 = 0.0;
            h2 = Q[tt][ss+1];
            if(h2 < 1E-10) h2 = 0.0;
            h3 = QQ[tt+1][ss+1];
            if(h3 < 1E-10) h3 = 0.0;
            h4 = QQ[tt][ss+1];
        }
    }
```c
if(h4 < 1E-10) h4 = 0.0;

h5 = (wtdt/1.0)*h1;

if((h1 == 0.0)&&(h2 == 0.0)) h6 = 0.0;
else h6 = Math.pow((h2+h1)/2.0,(b[ss]-1));
h7 = a[ss]*b[ss]*h2*h6;
h8 = wtdt*((h3+h4)/2.0);
h9 = (wtdt/1.0)+(a[ss]*b[ss]*h6);

Q[tt+1][ss+1] = (h5+h7+h8)/h9;
if(Q[tt+1][ss+1] > help) help = Q[tt+1][ss+1];
}

//Calculate Reynolds number, water level, velocity, stream power --
clist.addItem("Calculate water level, velocity, stream power");
for(tt = 0; tt < TT; tt++)
{
  ii = 0;
  for(ss = 0; ss < STM_2D.data.segments; ss++)
  {
    for(SS = 0; SS < STM_2D.data.dx[ss]; SS++)
    {
      //level in [m]
      //----------
      h1 = Mann[ii]*Q[tt][ii];
h2 = Math.pow(slope[ii]/100.0,0.5)*STM_2D.data.dy[ss];
      level[tt][ii] = Math.pow(h1/h2,3.0/5.0);
      //velocity in [m/s]
      //----------
      h1 = (Math.pow(slope[ii]/100.0,0.5))/Mann[ii];
      Ck[tt][ii] = h1*(5.0/3.0)*Math.pow(level[tt][ii],2.0/3.0);
      //Reynolds number [-]
      //----------
      Re[tt][ii] = (4*Ck[tt][ii]*level[tt][ii])/(0.000001311);
      //Stream power [g/s^3]
      //----------
      stream_power[tt][ii]=9.81*(slope[ii]/100.0)*998200*
                        (q[tt][ii]/STM_2D.data.dy[ss]);
      ii++;
    }
  }
}

//Calculate cumulative particle distribution --
//------------------------------------------
diam = new double[STM_2D.data.segments][3];
```

pperc = new double[STM_2D.data.segments][3];
double x1, x2, y1, y2;

for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
{
diam[ss][0] = 1.0; pperc[ss][0] = STM_2D.data.clay[ss];
diam[ss][1] = 26.0; pperc[ss][1] = STM_2D.data.silt[ss];
diam[ss][2] = 75.0; pperc[ss][2] = STM_2D.data.sand[ss];
}

//-------------------------------
//-- Calculate sediment deposition --
//-------------------------------
clist.addItem("Calculate sediment transport");
EROSION = new double[SEGMENTS];
ERO = new double[SEGMENTS];
DEP0 = new double[SEGMENTS];
erosion = new double[TT][SEGMENTS];
no_depos = new double[TT][SEGMENTS];

ii = 0;
for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
{
for(SS = 0 ; SS < STM_2D.data.dx[ss] ; SS++)
{
for(tt = 0 ; tt < TT ; tt++)
{
no_depos[tt][ii] = 0.0;
r = 0;

for(r = 0 ; r < 3 ; r++)
{
  //settling velocity (Stokes) [m/s]
  //-------------------------------
  h3=(1.0/18.0)*((Math.pow((diam[ss][r]/1000000.0),2.0)*
                   (2650-1000)*9.81)/viscos);

  //time necessary to settle [s]
  //-------------------------------
  h4 = (level[tt][ii] /(h3/1000.0));

  //distance needed to settle [m]
  //-------------------------------
  h4 = Ck[tt][ii]*h4;

  //particles kept in suspension [%]
  //-------------------------------
  if(h4 <= 1.0) h5 = 0.0;
  else h5 = 1.0-(1.0/h4);

  no_depos[tt][ii] += (pperc[ss][r]/100.0)*h5;
}

level[tt][ii] = level[tt][ii]*(1000.0); //level in [mm]
Ck[tt][ii] = Ck[tt][ii]*100.0; //velocity in [cm/s]
Appendix A: Source code of the STM-2D model

Q[tt][ii] = Q[tt][ii]*1000.0 ; //discharge in [l/s]

ii++; }
}

//--------------------
//-- Calculate sediment transport --
//--------------------
STM_2D_SedimentTransportModel st = new STM_2D_SedimentTransportModel();

for(tt = 0 ; tt < TT ; tt++)
{
    ii = 0;
    for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
    {
        for(SS = 0 ; SS < STM_2D.data.dx[ss] ; SS++)
        {
            h1 = h2 = h3 = h4 = h5 = h6 = h7 = h8 = h9 = 0.0;
            r = (int)(Math.floor((tt*wdt)/STM_2D.data.pdt));
            h9 = stream_power[tt][ii];

            // avoid NaN floating point exceptions
            // --------------------
            if(h9 < 0.00001)
            {
                h3 = 0.0;
                h4 = 0.0;
            }
            else
            {
                // If direct raindrop impact on soil surface
                // --------------------
                if(r < event_steps)
                {
                    if(STM_2D.data.leaf[r] > 0.0)
                    {
                        // probability of cover overlap by mulching and vegetation
                        // --------------------
                        h5 = (STM_2D.data.pcover[ss]/100.0);
                        h7 = (STM_2D.data.cover[ss]/100.0);
                        h8 = (h6*h7);

                        // transport function (Biesemans et al., 2000) [g/(s*cm)]
                        // --------------------
                        h1 = st.sheet(h9);
                        h4 = h1*(1.0-h6);
                        h1 = h1*1.0-((h6+h7)-h8));

                        // transport function (Nearing et al., 1997) [g/(s*cm)]
                        // --------------------
                        h2 = st.rill(h9);
                        h4 = h4+(h2*h6);
                        h2 = h2*(1.0-(STM_2D.data.cover[ss]/100.0));

                    } catch (ParseException e) {
                        e.printStackTrace();
                    }
                } catch (FileNotFoundException e) {
                    e.printStackTrace();
                } catch (IOException e) {
                    e.printStackTrace();
                }
            }
        }
    }
}
h3 = h1 + h2;
}

// If no direct raindrop impact on soil surface
// -----------------------------
if(STM_2D.data.Ieff[x] <= 0.0)
{
    h3 = st.rill(h9);
    h4 = h3;
    h3 = h3*(1.0-(STM_2D.data.cover[ss]/100.0));
}

// If no direct raindrop impact on soil surface
// -----------------------------
else
{
    // transport function (Nearing et al., 1997) [g/(s*cm)]
    // -----------------------------
    h3 = st.rill(h9);
    h4 = h3;
    h3 = h3*(1.0-(STM_2D.data.cover[ss]/100.0));
}

// if soil covered, how many sediment remains in suspension
// -----------------------------
if((ii > 0)&&(tt > 0))
{
    h3 += (no_depos[tt][ii]*erosion[tt-1][ii-1]);
    if(h3 > h4) erosion[tt][ii] = h4;
    else erosion[tt][ii] = h3;
}
else erosion[tt][ii] = h3;

ii++;
}

//-----------------------------
// Calculate total segment erosion --
//-----------------------------
ii = 0;
for(ss = 0; ss < STM_2D.data.segments; ss++)
{
    for(SS = 0; SS < STM_2D.data.dx[ss]; SS++)
    {
        EROSION[ii] = 0.0;
        for(tt = 0; tt < TT; tt++)
        {
            EROSION[ii] += (erosion[tt][ii]/100.0)*wrt*
                            (STM_2D.data.dy[ss]*100.0);
        }
        ii++;
    }
Appendix A: Source code of the STM-2D model

```java
}
}

// ---------------
//-- Calculate soil loss towards drainage system --
// ---------------
SOIL_LOSS = EROSION[SEGMENTS-1] ;

// ---------------
//-- Calculate cumulative erosion/deposition along hillslope --
// ---------------
for(ss = 0 ; ss < SEGMENTS ; ss++)
{
    if(ss == 0)
    {
        ERO[ss] = EROSION[ss] ;
        DEP[ss] = 0.0 ;
    }
    else
    {
        help = EROSION[ss]-EROSION[ss-1] ;
        if(help >= 0.0)
        {
            ERO[ss] = ERO[ss-1]+(help) ;
            DEP[ss] = DEP[ss-1] ;
        }
        else
        {
            ERO[ss] = ERO[ss-1] ;
            DEP[ss] = DEP[ss-1]-(-help) ;
        }
    }
}
// --------
//-- Done --
// --------
STM_2D.data.calculate.setBackground(Color.gray) ;
clist.removeAll() ;
clist.setForeground(Color.yellow) ;
DecimalFormat df = new DecimalFormat("0.0000") ;
clist.addItem("TOTAL SOIL LOSS TOWARDS DRAINAGE SYSTEM: "+
                  df.format(SOIL_LOSS/1000.0) + " TON") ;
clist.addItem("There are " +
                  Integer.toString(SEGMENTS) + " hillslope segments") ;

// ---------------
//-- Make the GUI components --
// ---------------
p2 = new Panel() ;
p2.setLayout(new GridLayout(3,2)) ;
p2.setBackground(Color.darkGray) ;

p2.add(new Label("Enter a segment to plot the properties: ")) ;
segmentDTF = new STM_2D_DoubleTextField(1.0,10) ;
```
p2.add(segmentDTF);

p2.add(new Label("Select a property to plot: "));
program_choices = new Choice();
program_choices.addItem("Draw hillslope");
program_choices.addItem("Draw segment hydrograph");
program_choices.addItem("Draw segment stream power");
program_choices.addItem("Draw segment water velocity");
program_choices.addItem("Draw segment water level");
program_choices.addItem("Draw segment Reynolds number");
program_choices.addItem("Draw erosion/deposition");
program_choices.addItem("Draw net transport");
program_choices.setBackground(Color.gray);
p2.add(program_choices);

p2.add(new Label("Click this button to terminate: "));
done = new Button("Done/Cancel");
p2.add(done);

add(p2, "South");

setSize(540,350);
setVisible(true);

//-----------------------------------------------
//-- Wait for the user to terminate this thread --
//-----------------------------------------------
while(running)
{
    try{Thread.currentThread().sleep(50);}
    catch(InterruptedException e) {}
}
} //end if

STM_2D.data.calculate.setBackground(Color.gray);
STM_2D.data.calclock = false;
STM_2D.data.lock = false;
}

*******************************************************************************/
/** Calculate the retention capacity of contour tillage  
*******************************************************************************/
public void contour_ret_cap()
{
    double theta, alpha, gamma, Sa, alpha1, alpha2; // angles
    double C, G, E, F, area; // some help variables

    for(ss = 0; ss < STM_2D.data.segments; ss++)
    {
        STM_2D.data.retc[ss] = 0.0;

        if((STM_2D.data.RWP[ss] > 0.0)&&(STM_2D.data.RWH[ss] > 0.0))
        {
            theta = Math.atan((STM_2D.data.RWP[ss]/2.0)/STM_2D.data.RWH[ss]) ;
            C = STM_2D.data.RWH[ss]/Math.cos(theta) ;
        }
alpha = 2.0*theta;
gamma = (Math.PI-alpha)/2.0;
Sa = Math.atan(STM_2D.data.S[ss]/100.0);
alpha1 = Math.PI-(Math.PI/2)-(gamma-Sa);
alpha2 = alpha-alpha1;
G = C*Math.cos(alpha1);
F = G*Math.tan(alpha1);
E = G*Math.tan(alpha2);
area = ((1.0/2.0)*G*E)+((1.0/2.0)*G*F);
STM_2D.data.retc[ss] = area*(1.0/STM_2D.data.RF[ss]);
}

*******************************************************************************/
*** Calculate the retention capacity of the random roughness using the ***
*** fractal "random midpoint displacement" methodology. ***
*******************************************************************************/
public void random_roughness_ret_cap()
{
    int i,j,k,l,ss ; //counter variables
    int Q = 5 ; //surface resolution
    double res, minY ; //surface variables
    double[][] Z ; //random surface
    int D, d, z, N ;
    double sigma, Sigma ;
    double help1, help2 ;

    for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
    {
        if((STM_2D.data.Hurst[ss] > 0.0)&&(STM_2D.data.Hurst[ss] < 1.0)&&(STM_2D.data.RRH[ss] > 0.0))
        {
            N = (int)(Math.pow(2.0,Q));
            res = (1.0/N);

            Z = new double[(int)(N+1)][(int)(N+1)];

            D = (int)(N);
            d = (int)(N/2.0);

            Z[0][0] = 0.0;
            Z[0][(int)(N)] = 0.0;
            Z[(int)(N)][0] = 0.0;
            Z[(int)(N)][(int)(N)] = 0.0;

            sigma = STM_2D.data.RRH[ss]/6.0;

            for(z = 0 ; z < Q ; z++)
            {
                Sigma = sigma*Math.sqrt(Math.pow((1/Math.pow(2,z)), 2*STM_2D.data.Hurst[ss]));

                for(i = d ; i <= (int)(N-d) ; i += D)
{ 
    for (j = d; j <= (int)(N-d); j += D) 
    { 
        Z[i][j] = Z[i+d][j+d] + Z[i+d][j-d] + Z[i-d][j+d] + Z[i-d][j-d];
        help1 = Ran.nextDouble();
        help2 = Ran.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                 (Math.cos(2.0*Math.PI*help2));
        Z[i][j] = (Z[i][j]/4.0)+help1;
    }

    for (i = d; i <= (int)(N-d); i += D) 
    { 
        Z[i][0] = Z[i+d][0]+Z[i-d][0]+Z[i][d];
        help1 = Ran.nextDouble();
        help2 = Ran.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                 (Math.cos(2.0*Math.PI*help2));
        Z[i][0] = (Z[i][0]/3.0)+help1;

        Z[i][N] = Z[i+d][N]+Z[i-d][N]+Z[i][N-d];
        help1 = Ran.nextDouble();
        help2 = Ran.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                 (Math.cos(2.0*Math.PI*help2));
        Z[i][N] = (Z[i][N]/3.0)+help1;

        Z[0][i] = Z[0][i+d]+Z[0][i-d]+Z[d][i];
        help1 = Ran.nextDouble();
        help2 = Ran.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                 (Math.cos(2.0*Math.PI*help2));
        Z[0][i] = (Z[0][i]/3.0)+help1;

        Z[N][i] = Z[N][i+d]+Z[N][i-d]+Z[N-d][i];
        help1 = Ran.nextDouble();
        help2 = Ran.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                 (Math.cos(2.0*Math.PI*help2));
        Z[N][i] = (Z[N][i]/3.0)+help1;
    }

    if (D < N) 
    { 
        for (i = d; i <= (int)(N-d); i += D) 
        { 
            for (j = D; j <= (int)(N-d); j += D) 
            { 
                Z[i][j] = Z[i][j+d]+Z[i][j-d]+Z[i+d][j]+Z[i+d][j];
                help1 = Ran.nextDouble();
                help2 = Ran.nextDouble();
                help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                         (Math.cos(2.0*Math.PI*help2));
                Z[i][j] = (Z[i][j]/4.0)+help1;
            }
        }
    }
Appendix A: Source code of the STM-2D model

for(i = D ; i <= (int)(N-d) ; i += D)
{
    for(j = d ; j <= (int)(N-d) ; j += D)
    {
        Z[i][j] = Z[i][j+d]+Z[i][j-d]+Z[i+d][j]+Z[i-d][j] ;
        help1 = RAN.nextDouble();
        help2 = RAN.nextDouble();
        help1 = Sigma*(Math.sqrt(-2.0*Math.log(help1)))*
                   (Math.cos(2.0*Math.PI*help2));
        Z[i][j] = (Z[i][j]/4.0)+help1 ;
    }
    D = D/2 ;
    d = d/2 ;
}

for(i = 0 ; i < (N+1) ; i++)
{
    for(j = 0 ; j < (N+1) ; j++)
    {
        Z[i][j] = Z[i][j]+(10.0-(res*j*(STM_2D.data.S[ss]/100.0))) ;
    }
}

for(i = 1 ; i < N ; i++)
{
    for(j = 1 ; j < N ; j++)
    {
        minY = 1000000.0 ;
        for(k = -1 ; k <= 1 ; k++)
        {
            for(l = -1 ; l <= 1 ; l++)
            {
                if(!((k==0)&&(l==0)))
                {
                    if(Z[i+k][j+l] < minY) minY = Z[i+k][j+l] ;
                }
            }
        }
        if(Z[i][j] < minY)
        {
            help1 = (minY-Z[i][j])*Math.pow(2.0*res,2.0) ;
            STM_2D.data.retc[ss] += help1 ;
        }
    }
    //END
    //---
}
}
if(event.target == done)
{
    STM_2D.data.calclock = false;
    STM_2D.data.lock = false;
    this.dispose();
    running = false;
    return true;
}
//----------------------------------------------------------------------------------------
if(event.target == program_choices)
{
    Double help = new Double(segmentDTF.getText());
    int seg = help.intValue();

    //----------------------------------------------------------------------------------------
if(arg.equals("Draw hillslope"))
{
    Thread th1 = new Thread(new STM_2D_Plot(X,Y,Y,SEGMENTS+1,0,
        "Hillslope (X in [m], Y in [m])");
    th1.setPriority(Thread.NORM_PRIORITY);
    th1.start();
    return true;
}
//----------------------------------------------------------------------------------------
if((arg.equals("Draw segment hydrograph"))&&(seg <= SEGMENTS)&&(seg > 0))
{
    plotX = new double[TT];
    plotY = new double[TT];

    for(i = 0 ; i < TT ; i++)
    {
        plotX[i] = (i*wdt)/60.0;
        plotY[i] = q[i][seg-1];
    }

    Thread th1 = new Thread(new STM_2D_Plot(plotX,plotY,plotY,TT,0,
        "Hydrograph (X in [min], Y in [liter/s])");
    th1.setPriority(Thread.NORM_PRIORITY);
    th1.start();
    return true;
}
//----------------------------------------------------------------------------------------
if((arg.equals("Draw segment stream power"))&&(seg <= SEGMENTS)&&(seg > 0))
{
    plotX = new double[TT];
    plotY = new double[TT];

    for(i = 0 ; i < TT ; i++)
    {
        plotX[i] = (i*wdt)/60.0;
        plotY[i] = stream_power[i][seg-1];
    }
}
Thread thi = new Thread(new STM_2D_Plot(plotX,plotY,plotY,TT,0,
        "Stream power (X in [min], Y in [g/s**3])");
thi.setPriority(Thread.NORM_PRIORITY);
thi.start();
return true;
}---------------------------------------------------------------------
if((arg.equals("Draw segment water velocity")&&seg <= SEGMENTS)&&(seg > 0))
{
    plotX = new double[TT] ;
    plotY = new double[TT] ;

    for(i = 0 ; i < TT ; i++)
    {
        plotX[i] = (i*wdt)/60.0 ;
        plotY[i] = Ck[i][seg-1] ;
    }

Thread thi = new Thread(new STM_2D_Plot(plotX,plotY,plotY,TT,0,
        "Overland flow velocity (X in [min], Y in [cm/s])");
thi.setPriority(Thread.NORM_PRIORITY);
thi.start();
return true;
}---------------------------------------------------------------------
if((arg.equals("Draw segment water level")&&seg <= SEGMENTS)&&(seg > 0))
{
    plotX = new double[TT] ;
    plotY = new double[TT] ;

    for(i = 0 ; i < TT ; i++)
    {
        plotX[i] = (i*wdt)/60.0 ;
        plotY[i] = level[i][seg-1] ;
    }

Thread thi = new Thread(new STM_2D_Plot(plotX,plotY,plotY,TT,0,
        "Overland flow water level (X in [min], Y in [mm])");
thi.setPriority(Thread.NORM_PRIORITY);
thi.start();
return true;
}---------------------------------------------------------------------
if((arg.equals("Draw segment Reynolds number")&&(seg <= SEGMENTS)&& (seg > 0))
{
    plotX = new double[TT] ;
    plotY = new double[TT] ;

    for(i = 0 ; i < TT ; i++)
    {
        plotX[i] = (i*wdt)/60.0 ;
        plotY[i] = Re[i][seg-1] ;
    }
Thread th1 = new Thread(new STM_2D_Plot(plotX,plotY,plotY,T,0,
   "Overland flow Reynolds number (X in [s], Y in [-])")
   );
th1.setPriority(Thread.NORM_PRIORITY);
th1.start();
return true;
}

// if(arg.equals("Draw erosion/deposition"))
{
  plotX = new double[SEGMENTS+1];
  plotY = new double[SEGMENTS+1];
  plotY2 = new double[SEGMENTS+1];

  for(i = 0 ; i < SEGMENTS+1 ; i++)
  {
    plotX[i] = i;

    if(i == 0)
    {
      plotY[i] = ERO[i] ;
      plotY2[i] = DEP0[i] ;
    }
    if(i > 0)
    {
      plotY[i] = ERO[i-1] ;
      plotY2[i] = DEP0[i-1] ;
    }
  }

  Thread th1 = new Thread(new STM_2D_Plot(plotX,plotY,plotY2,SEGMENTS+1,1,
   "Cumulative erosion(+) / deposition(-) (X in [m], Y in [kg])")
   );
th1.setPriority(Thread.NORM_PRIORITY);
th1.start();
return true;
}

// if(arg.equals("Draw net transport"))
{
  plotX = new double[SEGMENTS+1];
  plotY = new double[SEGMENTS+1];
  plotY2 = new double[SEGMENTS+1];

  i = 0 ;
  plotX[0] = 0.0 ;
  plotY[0] = 0.0 ;
  plotY2[0] = 0.0 ;
  for(ss = 0 ; ss < STM_2D.data.segments ; ss++)
  {
    for(SS = 0 ; SS < STM_2D.data.dz[ss] ; SS++)
    {
      plotX[i+1] = i+1;

      if(i > 0)
      {

if((EROSION[i] - EROSION[i-1]) >= 0)
{
    plotY[i+1] = (EROSION[i] - EROSION[i-1])/STM_2D.data.dy[ss];
    plotY2[i+1] = 0.0;
}
else
{
    plotY2[i+1] = (EROSION[i] - EROSION[i-1])/STM_2D.data.dy[ss];
    plotY[i+1] = 0.0;
}
}
if(i == 0)
{
    plotY[i+1] = EROSION[i]/STM_2D.data.dy[ss];
    plotY2[i+1] = 0.0;
}
i++;
}

Thread th1 = new Thread(new STM_2D_Plot(plotX, plotY, plotY2, SEGMENTS+1, 1, "Net erosion(+) / deposition(-) (X in [m], Y in [kg/m**2])");
    th1.setPriority(Thread.NORM_PRIORITY);
    th1.start();
    return true;
}
else return true;
}

//*******************************************************************************/
else return super.action(event, arg);

*******************************************************************************/
}
STM_2D_GreenAmpt.java

This class contains the code for the modified Green-Ampt infiltration model, to predict the runoff during unsteady rainfall events.

```java
passport 1: Source code of the STM-2D model

public class STM_2D_GreenAmpt {
    public void green_ampt(
        double C, /* clay content [%] */
        double L, /* silt content [%] */
        double S, /* sand content [%] */
        double O, /* organic carbon content [%] */
        double Ds, /* soil density [g/cm^3] */
        double Ws, /* percent of saturated [%] */
        double D, /* retention capacity [m] */
        int N, /* number of time steps */
        double dt, /* length of the time steps [h] */
        double[] Ueff, /* effective rainfall [m/h] */
        double[] R, /* cumulative runoff [m] */
        double ic /* rainfall interception [m] */
    )

    int n, i; /* time step counter, counter */
    double help, help1, help2; /* help variables */
    double Ws, W; /* soil moisture: saturation, pH 4.2 [vol%] */
    double Wmin, W; /* moisture corresponding Kmin, moisture [vol%] */
    double a, nn; /* non-physical Van Genuchten parameters [-] */
}
```

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Appendix A: Source code of the STM-2D model

double B ;    /* parameter to calculate Ks (Campbell) [-]/
double P_es ;    /* air entry water potential [J/kg]/
double dc, dl, ds ;    /* mean of particle class diameters [mm]/
double dg, sg ;    /* mean particle-size diameter + st. dev. [mm]/
double Ks ;    /* saturated hydraulic conductivity [m/h]/
double K ;    /* unsaturated hydraulic conductivity [m/h]/
double[] Kr = new double[100] ; /* relative conductivity [-]/
double Kmin ;    /* minimum value of K [m/h]/
double M ;    /* moisture deficit [vol%]/
double m ;    /* parameter to calculate Sa (Campbell) [-]/
double SM ;    /* capillary suction*moisture deficit [m]/
double[] CS = new double[100] ; /* capillary suction [cm water]/
double Sa ;    /* average capillary suction [m]/
double[] T ;    /* shifted time [h]/
double[] P ;    /* cumulative rainfall [m]/
double[] Cu, Cp ;    /* surface condition indicators [-]/
double[] tp, ts, tn ;    /* ponding time, pseudo time, event time [h]/
double[] F, Fo, F2 ;    /* cumulative infiltration [m]/

P = new double[N+1] ; T = new double[N+1] ;
Cu = new double[N+1] ; F = new double[N+1] ;
tn = new double[N+1] ; tp = new double[N+1] ;
ts = new double[N+1] ; Fo = new double[N+1] ;
Cp = new double[N+1] ; F2 = new double[N+1] ;

for(n = 0; n <= N; n++)
{
    tn[n] = n*dt ;
}

//--------------------------------------------------------------------------------------------------------------------------
//--- Calculate VAN GENUCHTEN PARAMETERS [Vereeck et al. (1989)]. The Water
//--- Retention Curve (WRC) is estimated using the VAN GENUCHTEN model. The WRC
//--- is used to determine the average capillary suction.
//--------------------------------------------------------------------------------------------------------------------------
Ws = 0.81 - (0.283*Dm)+(0.001*C) ;    /* Water content at saturation */
Wr = 0.015 + (0.005*C) + (0.014*OC) ;    /* Water content at pH 4.2 */
W1 = Wr+((W1/(100.0))*(Ws-Wr)) ;
a = Math.exp((-2.486)+(0.025*S)-(0.351*OC)-(2.617*Dm)-(0.023*C)) ;
nn = Math.exp((0.053)-(0.009*S)-(0.013*C)+(0.00015*(S*S))) ;

//--------------------------------------------------------------------------------------------------------------------------
//--- Calculate the saturated hydraulic conductivity based on the
//--- soil properties using the methodology described in CAMPBELL (1985)
//--------------------------------------------------------------------------------------------------------------------------
dc = (0.000 + 0.002)/2.0 ;    /* mean diameter of clay particle class [mm]/
dl = (0.002 + 0.050)/2.0 ;    /* mean diameter of loam particle class [mm]/
ds = (0.050 + 2.000)/2.0 ;    /* mean diameter of sand particle class [mm]/

/* mean diameter [mm] */
/* ------------------- */
help1 = 0.01*((C*Math.log(dc))+(L*Math.log(dl))+(S*Math.log(ds)));
dg = Math.exp(help1) ;
/* standard deviation */
/* ---------------------- */
help2= 0.01*((C*Math.pow(Math.log(dc),2.0))+(L*Math.pow(Math.log(dl),2.0)) +
    (G*Math.pow(Math.log(ds),2.0)));  
sg = Math.exp(Math.sqrt(help2-Math.pow(help1,2.0))); 

P_es = -0.5*Math.pow(dg,-0.5); 
B = (-2.0*P_es)+0.2*sg;

/* saturated hydraulic conductivity [m/h] */
/* ------------------------------- */
Ks = (0.004*Math.pow(1.3/Ds,1.3*B)*Math.exp((-6.9*(C/100.0))-
    (3.7*(L/100.0))))*35.28;

/* minimum unsaturated conductivity [m/h] and corresponding moisture */
/* --------------------------------------------------------------- */
Kmin = 0.01*Ks; 
m = (2.0*B) + 3.0; 
Wmin = Math.pow(Kmin/Ks,1.0/m)*Ws; 
if(Wmin < W) Wmin = W; 

/* relative conductivity and capillary suction [cm water] */
/* -------------------------------------------------------- */
for(i = 0 ; i < 100 ; i++)
{
    W = Wmin+(((Ws-Wmin)/99.0)*i);  
    K = Ks*Math.pow(W/Ws,m);  
    CS[i]= Math.pow(((Ws-Wr)/(W-Wr))-1.0)*((1.0/Math.pow(a,mn)), 1.0/nn);  
    Kr[i]= K/Ks;  
}

/* average capillary suction [m] */
/* ----------------------------- */
Sa = 0.0; 
for(i = 1 ; i < 100 ; i++) Sa += CS[i]*(Kr[i]-Kr[i-1]);  
Sa /= 100.0;

/* moisture deficit */
/* --------------- */
M = Ws-Wi;

/* average capillary suction * moisture deficit */
/* -------------------------------------- */
SM = Sa*M;

//---------------------------------------------------------------------------------
//-- Main loop: Calculate infiltration and runoff
//---------------------------------------------------------------------------------
for(n = 0 ; n < N ; n++)
{  
    if(n == 0) P[n] = (Ieff[n]*dt);  
    else P[n] = (Ieff[n]*dt)+P[n-1];  
}
for(n = 0 ; n < N ; n++)


```c
{  if(ic > P[n]) P[n] = 0.0 ;
    else P[n] = P[n] - ic ;
}

for(n = 0 ; n < N ; n++)
{
    /* --------------------------- */
    /* Calculate first time step */
    /* --------------------------- */
    if(n == 0)
    {
        /* there is no surface ponding at initial time */
        Cu[n] = P[n] - ((Ks*SM)/((P[n]/dt)-Ks)) ;
        if(1eff[n] <= Ks) Cu[n] = 0.0 ;

        /* no surface ponding at terminal time of this period */
        if(Cu[n] <= 0.0)
        {
            F[n] = P[n] ; /* cumulative infiltration = cumulative rainfall */
            R[n] = 0.0 ; /* no runoff */
        }
    }

    /* surface ponding at terminal time */
    if(Cu[n] > 0.0)
    {
        /* the ponding time */
        tp[n] = ((Ks*SM/((P[n]/dt)-Ks))/(P[n]/dt)+tn[n] ;

        /* at the ponding time is the cumulative rainfall */
        Fo[n] = tp[n]*(P[n]/dt) ;

        /* the pseudotime */
        help = Fo[n]/SM ;
        ts[n] = (help-Math.log(1.0+help))*(SM/Ks) ;

        for(i = n ; i < N ; i++) {tp[i] = tp[n] ; ts[i] = ts[n] ;}

        /* Time on the shifted time scale */

        /* cumulative infiltration */
        if(((Ks*T[n])/SM) < 0.0)
        {
            System.out.print("\n\nam error occured during root calculation.\n\n") ;
            System.exit(1) ;
        }
        else
        {
            F2[n] = root((Ks*T[n])/SM) ;
            F[n] = F2[n]*SM ;
        }

        /* runoff */
        R[n] = P[n]-F[n] ;
        if(R[n] < 0.0) R[n] = 0.0 ;
    }
}
```
% Calculate the other time steps */
if(n > 0)
{
    /* 1. surface not ponded at initial time */
    Cu[n-1] <= 0.0
{
    /* surface condition indicator */
    Cu[n] = P[n] - R[n-1] - ((Ks*SM)/((P[n]-P[n-1])/dt)-Ks));
    if(((P[n]-P[n-1])/dt) <= Ks) Cu[n] = 0.0 ;

    /* no surface ponding at terminal time of this period */
    if(Cu[n] <= 0.0)
    {
        F[n] = P[n] ;/* cumulative infiltration = cumulative rainfall */
        R[n] = R[n-1];/* no runoff */
    }

    /* surface ponding at terminal time */
    if(Cu[n] > 0.0)
    {
        /* the ponding time */
        tp[n]=((Ks*SM/(((P[n]-P[n-1])/dt)-Ks)-P[n-1]+R[n-1])/
                ((P[n]-P[n-1])/dt)+tn[n];
        /* at the ponding time is the cumulative rainfall */
        Fo[n] = P[n-1] + ((tp[n]-tn[n])*((P[n]-P[n-1])/dt));

        /* the pseudotime */
        help = (Fo[n]-R[n-1])/SM ;
        ts[n] = (help-Math.log(1.0+help))*(SM/Ks) ;
        for(i = n ; i < N ; i++) {tp[i] = tp[n] ; ts[i] = ts[n] ;}

        /* Time on the shifted time scale */

        /* cumulative infiltration */
        if(((Ks*T[n])/SM) < 0.0)
        { System.out.print("\n\n \n\n\n"; System.exit(1) ;
        } else
        { F2[n] = root((Ks*T[n])/SM) ;
        F[n] = F2[n]*SM ;
    }
}
    /* runoff */
    R[n] = F[n]-F[n] ;
if(R[n] < 0.0) R[n] = 0.0 ;
}
}

/* 2. surface ponded at initial time */
/*--------------------------------------*/
if(Cu[n-1] > 0.0)
{
    /* the shifted time is */

    /* cumulative infiltration */
    if(((Ks*T[n])/SM) < 0.0)
    {
        System.out.print("\n\n\nerror occurred during root calculation.\n\n\n");
        System.exit(1) ;
    }
    else
    {
        F2[n] = root(((Ks*T[n])/SM) ;
        F[n] = F2[n]*SM ;
    }
    /* surface condition indicator */

    /* no ponding at terminal time */
    if(Cp[n] <= 0.0)
    {
        /* adjust cumulative infiltration */
        F[n] = P[n] - R[n-1] ;
        R[n] = R[n-1] ;
        Cu[n] = -100.0 ;
    }

    /* ponding at terminal time */
    if(Cp[n] > 0.0)
    {
        Cu[n] = 100.0 ;
        R[n] = F[n]-F[n] ;
        if(R[n] < 0.0) R[n] = 0.0 ;
    }
}
}
for(n = 0 ; n < N ; n++)
{
    if(D > R[n]) R[n] = 0.0 ;
    else R[n] = R[n] - D ;
}

/***************************************************************************/
/*** Class Method: calculate root of the implicit function: ***********/
/*** b - ln(1+b) - a = 0 **********************************************/
/***************************************************************************/
public double root(double a)
double b=0.0, dif;
double max, min, between;
double n1, n2, n3;

min = 0.0; max = 1E30; dif = 100000.0;
while(dif > 0.000000001)
{
    between = (max + min)/2.0;

    n1 = max - Math.log(1.0+max) - a;
    n2 = min - Math.log(1.0+min) - a;
    n3 = between - Math.log(1.0+between) - a;

    if(n1*n3 < 0){b = between; min = between; max = max;}
    if(n2*n3 < 0){b = between; min = min; max = between;}

    n1 = b - Math.log(1.0+b);
    dif = Math.abs(a-n1);
}
return(b);
}
**STM_2D_SedimentTransportModel.java**

This class contains the equations for the sediment transport. These equations are based on the hydraulic properties of the overland flow and are only valid for non-cohesive sandy, loamy and silty soil types. The clay content of the soil must be lower than ±35%. Consolidated soils, cohesive soils, and soils with a clay content more than ±35% need an additional erodibility factor, which can be for example critical shear stress, critical momentum flux. Because the erosion risk zones in Belgium are almost all located in the loess belt, the sediment transport equations listed here are sufficient to evaluate the different land management scenarios within these risk zones.

```java
import java.math.*;

public class STM_2D_SedimentTransportModel
{
    /**
     * Rill equation (Nearing et al., 1997)
     */
    public double rill(double stream_power /* [g/s^3] */) {
        double sediment_transport ; //sediment transport in [g/(s*cm)]
        double help;

        help = 0.845+0.412*blog(10.0,stream_power);
        sediment_transport = Math.pow(10.0,(~34.47+(38.61*(help/(1+help)))));
        return sediment_transport;
    }

    /**
     * Sheet flow equation (Bissemans et al., 1999)
     */
    public double sheet(double stream_power /* [g/s^3] */) {
        //Implementation of sheet flow equation
    }
}
```

References:


double sediment_transport; //sediment transport in [g/(s*cm)]

    sediment_transport = 0.0001635701*Math.pow(stream_power,1.314);
    return sediment_transport;
}
/**************************************************************************/
/**** Base 'a' logarithm  ****/
**************************************************************************/
double blog(double base, double x)
{
    return(Math.log(x)/Math.log(base));
}
/**************************************************************************/
STM_2D_GetHillslopeData.java

This class defines the GUI (Graphical User Interface) components to enter the hillslope related data.

```java
/**
   *** STM_2D_GetHillslopeData.java
   ***
   *** This class can be started as a user thread and asks for the number of ***
   *** segments, and for each segment the topographical, soil, vegetation and ***
   *** land management parameters. ***
   ***
   *** Programmed by:  ir. Jan Biesemans ***
   ***
   *** Dept. Soil Management ***
   ***
*/
import java.awt.*;
import java.awt.event.*;

class STM_2D_GetHillslopeData extends Frame implements Runnable {
    /**
      *** General data ***
      ***
    */
    private int s, segments;
    private boolean running;

    /**
      *** GUI components ***
      ***
    */
    private Button store_button;
    private Button init_button;
    private Button forward;
    private Button back;
    private Button done;
    private Panel p1, p2, p3;
    private int DTFW;
    private Label segelabel;
    private Label initlabel;
    private Label[] soillabel;
    private STM_2D_DoubleTextField segsDTF;
    private STM_2D_DoubleTextField segnDTF;
    private STM_2D_DoubleTextField[] soilDTF;

    /**
      *** Constructor ***
      ***
    */
    STM_2D_GetHillslopeData() {
        int i;

        running = true;
    }
```
// Width of a Double Text Field (DTFW)
// -------------------------------
DTFW = 10;

// Window colors
// -------------
this.setBackground(Color.darkGray);
this.setForeground(Color.white);
this.setTitle("Hillslope Data");

// Upper panel
// -------------
p1 = new Panel();
p1.add(new Label("Hillslope Input Parameters"));
p1.setForeground(Color.red);
add(p1, "North");

// Center panel
// -------------
p2 = new Panel();
p2.setLayout(new GridLayout(21, 2));
p2.setBackground(Color.darkGray);

segsDTF = new STM_2D_DoubleTextField(STM_2D.data.segments, DTFW);
segsLabel = new Label("Number of segments: ") ;
initLabel = new Label("Click this button if not initialized: ") ;
init_button = new Button("INITIALIZE AND ERASE MEMORY") ;
back = new Button("go to previous hillslope segment") ;
forward = new Button("go to next hillslope segment") ;
init_button.setForeground(Color.yellow);
init_button.setBackground(Color.gray);

p2.add(segslabel) ; p2.add(segsDTF) ;
p2.add(initlabel) ; p2.add(init_button);
p2.add(back) ; p2.add(forward) ;

soillabel = new Label[18] ;
soillabel[0] = new Label("Segment number: ") ;
soillabel[1] = new Label("Segment slope [%]: ") ;
soillabel[2] = new Label("Segment length [m]: ") ;
soillabel[3] = new Label("Segment width [m]: ") ;
soillabel[4] = new Label("Clay content [%]: ") ;
soillabel[5] = new Label("Silt content [%]: ") ;
soillabel[6] = new Label("Sand content [%]: ") ;
soillabel[8] = new Label("Soil density [kg/m**3]: ") ;
soillabel[9] = new Label("Initial moisture [% of saturated]: ") ;
soillabel[10] = new Label("Vegetation interception capacity [m]: ") ;
soillabel[12] = new Label("Mulch soil cover [%]: ") ;
soillabel[13] = new Label("Contour ridge wave period [m]: ") ;
soillabel[14] = new Label("Contour ridge height [m]: ") ;
soillabel[15] = new Label("Random roughness Hurst coeff. [0-1]: ") ;
soillabel[16] = new Label("Random roughness height [m]: ") ;
soillabel[17] = new Label("Store this segment data: ") ;
Appendix A: Source code of the STM-2D model

```java
segmDTF = new STM_2D_DoubleTextField(1,DTFW) ;
soilDTF = new STM_2D_DoubleTextField[16] ;

if(STM_2D.data.segments == 0)
{
    soilDTF[0] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[1] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[2] = new STM_2D_DoubleTextField(100,DTFW) ;
    soilDTF[3] = new STM_2D_DoubleTextField(20,DTFW) ;
    soilDTF[4] = new STM_2D_DoubleTextField(60,DTFW) ;
    soilDTF[5] = new STM_2D_DoubleTextField(20,DTFW) ;
    soilDTF[6] = new STM_2D_DoubleTextField(1.2,DTFW) ;
    soilDTF[7] = new STM_2D_DoubleTextField(1400,DTFW) ;
    soilDTF[8] = new STM_2D_DoubleTextField(95,DTFW) ;
    soilDTF[9] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[10] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[11] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[12] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[13] = new STM_2D_DoubleTextField(0,DTFW) ;
    soilDTF[14] = new STM_2D_DoubleTextField(1,DTFW) ;
    soilDTF[15] = new STM_2D_DoubleTextField(0,DTFW) ;
}
else
{
    soilDTF[0] = new STM_2D_DoubleTextField(STM_2D.data.S[0] ,DTFW) ;
    soilDTF[1] = new STM_2D_DoubleTextField(STM_2D.data.dx[0] ,DTFW) ;
    soilDTF[2] = new STM_2D_DoubleTextField(STM_2D.data.dy[0] ,DTFW) ;
    soilDTF[3] = new STM_2D_DoubleTextField(STM_2D.data.clay[0] ,DTFW) ;
    soilDTF[4] = new STM_2D_DoubleTextField(STM_2D.data.silt[0] ,DTFW) ;
    soilDTF[5] = new STM_2D_DoubleTextField(STM_2D.data.sand[0] ,DTFW) ;
    soilDTF[6] = new STM_2D_DoubleTextField(STM_2D.data.corg[0] ,DTFW) ;
    soilDTF[7] = new STM_2D_DoubleTextField(STM_2D.data.soil_d[0] ,DTFW) ;
    soilDTF[8] = new STM_2D_DoubleTextField(STM_2D.data.MI[0] ,DTFW) ;
    soilDTF[9] = new STM_2D_DoubleTextField(STM_2D.data.intercep[0] ,DTFW) ;
    soilDTF[10] = new STM_2D_DoubleTextField(STM_2D.data.pcover[0] ,DTFW) ;
    soilDTF[11] = new STM_2D_DoubleTextField(STM_2D.data.cover[0] ,DTFW) ;
    soilDTF[12] = new STM_2D_DoubleTextField(STM_2D.data.RWP[0] ,DTFW) ;
    soilDTF[13] = new STM_2D_DoubleTextField(STM_2D.data.RWH[0] ,DTFW) ;
    soilDTF[14] = new STM_2D_DoubleTextField(STM_2D.data.Hurst[0] ,DTFW) ;
    soilDTF[15] = new STM_2D_DoubleTextField(STM_2D.data.RRH[0] ,DTFW) ;
}

store_button= new Button("Store Segment Data");
store_button.setForeground(Color.yellow) ;
store_button.setBackground(Color.gray) ;

for(i = 0 ; i <= 17 ; i++)
{
    p2.add(soillabel[i]) ;
    if(i == 17) p2.add(store_button) ;
    if(i == 0 ) p2.add(segmDTF) ;
    if((i != 17)&&(i != 0)) p2.add(soilDTF[i-1]) ;
}

add(p2, "Center") ;
```
/** Bottom panel */

p3 = new Panel();
done = new Button("Done/Cancel");
p3.add(done);
add(p3, "South");

/* Set visibility */

setSize(550, 750);
setVisible(true);

/******************************/

/** Run Method */

public void run()
{
    while(running)
    {
        try{Thread.currentThread().sleep(50);}
        catch(InterruptedException e) {} 
    }
}

/******************************/

/** Define button actions */

public boolean action(Event event, Object arg)
{
    if(event.target == init_button)
    {
        Double help = new Double(segsDTF.getValue());
        STM_2D.data.segments = help.intValue();
        segments = STM_2D.data.segments;
    }

    STM_2D.data.dx      = new double[segments];
    STM_2D.data.dy      = new double[segments];
    STM_2D.data.S       = new double[segments];
    STM_2D.data.sand    = new double[segments];
    STM_2D.data.silt    = new double[segments];
    STM_2D.data.clay    = new double[segments];
    STM_2D.data.corg    = new double[segments];
    STM_2D.data.soil_d  = new double[segments];
    STM_2D.data.MI      = new double[segments];
    STM_2D.data.Intercep = new double[segments];
    STM_2D.data.pcover  = new double[segments];
    STM_2D.data.cover   = new double[segments];
    STM_2D.data.RWP     = new double[segments];
    STM_2D.data.RWH     = new double[segments];
    STM_2D.data.Hurst   = new double[segments];
    STM_2D.data.RRH     = new double[segments];
    STM_2D.data.retc    = new double[segments];

    return true;
}
if(event.target == store_button)
{
    Double help = new Double(segnDTF.getValue());
    s = help.intValue();

    if(((s-1) >= 0)&&(s <= STM_2D.data.segments))
    {
        STM_2D.data.S[s-1] = soilDTF[0].getValue();
        STM_2D.data.dx[s-1] = soilDTF[1].getValue();
        STM_2D.data.dy[s-1] = soilDTF[2].getValue();
        STM_2D.data.clay[s-1] = soilDTF[3].getValue();
        STM_2D.data.silt[s-1] = soilDTF[4].getValue();
        STM_2D.data.sand[s-1] = soilDTF[5].getValue();
        STM_2D.data.corg[s-1] = soilDTF[6].getValue();
        STM_2D.data.soil_d[s-1] = soilDTF[7].getValue();
        STM_2D.data.MI[s-1] = soilDTF[8].getValue();
        STM_2D.data.Intercep[s-1] = soilDTF[9].getValue();
        STM_2D.data.pcover[s-1] = soilDTF[10].getValue();
        STM_2D.data.cover[s-1] = soilDTF[11].getValue();
        STM_2D.data.RWP[s-1] = soilDTF[12].getValue();
        STM_2D.data.RWH[s-1] = soilDTF[13].getValue();
        STM_2D.data.Hurst[s-1] = soilDTF[14].getValue();
        STM_2D.data.RRH[s-1] = soilDTF[15].getValue();
    }

    return true;
}

//-------------------------------------------------------------------------------------
if(event.target == forward)
{
    segments = STM_2D.data.segments;

    if(segments > 0)
    {
        Double help = new Double(segnDTF.getValue());
        s = help.intValue();
        s++;
        if(s > segments) s = 1;

        segnDTF.setText(Integer.toString(s));
        soilDTF[0].setText(Double.toString(STM_2D.data.S[s-1]));
        soilDTF[1].setText(Double.toString(STM_2D.data.dx[s-1]));
        soilDTF[2].setText(Double.toString(STM_2D.data.dy[s-1]));
        soilDTF[3].setText(Double.toString(STM_2D.data.clay[s-1]));
        soilDTF[4].setText(Double.toString(STM_2D.data.silt[s-1]));
        soilDTF[5].setText(Double.toString(STM_2D.data.sand[s-1]));
        soilDTF[6].setText(Double.toString(STM_2D.data.corg[s-1]));
        soilDTF[7].setText(Double.toString(STM_2D.data.soil_d[s-1]));
        soilDTF[8].setText(Double.toString(STM_2D.data.MI[s-1]));
        soilDTF[9].setText(Double.toString(STM_2D.data.Intercep[s-1]));
        soilDTF[10].setText(Double.toString(STM_2D.data.pcover[s-1]));
        soilDTF[11].setText(Double.toString(STM_2D.data.cover[s-1]));
        soilDTF[12].setText(Double.toString(STM_2D.data.RWP[s-1]));
        soilDTF[13].setText(Double.toString(STM_2D.data.RWH[s-1]));
        soilDTF[14].setText(Double.toString(STM_2D.data.Hurst[s-1]));
        soilDTF[15].setText(Double.toString(STM_2D.data.RRH[s-1]));
    }
soilDTF[14].setText(Double.toString(STM_2D.data.Hurst[s-1]));
soilDTF[15].setText(Double.toString(STM_2D.data.RHH[s-1]));
}

return true;
}

// ==========================================================================
if(event.target == back)
{
segments = STM_2D.data.segments;

if(segments > 0)
{
Double help = new Double(segnDTF.getValue());
s = help.intValue();
s--;
if(s < 1) s = segments;

segnDTF.setText(Integer.toString(s));
soilDTF[0].setText(Double.toString(STM_2D.data.S[s-1]));
soilDTF[1].setText(Double.toString(STM_2D.data.dx[s-1]));
soilDTF[2].setText(Double.toString(STM_2D.data.dy[s-1]));
soilDTF[3].setText(Double.toString(STM_2D.data.clay[s-1]));
soilDTF[4].setText(Double.toString(STM_2D.data.silt[s-1]));
soilDTF[5].setText(Double.toString(STM_2D.data.sand[s-1]));
soilDTF[6].setText(Double.toString(STM_2D.data.corg[s-1]));
soilDTF[7].setText(Double.toString(STM_2D.data.soil_d[s-1]));
soilDTF[8].setText(Double.toString(STM_2D.data.MI[s-1]));
soilDTF[9].setText(Double.toString(STM_2D.data.Intercep[s-1]));
soilDTF[10].setText(Double.toString(STM_2D.data.pcover[s-1]));
soilDTF[11].setText(Double.toString(STM_2D.data.cover[s-1]));
soilDTF[12].setText(Double.toString(STM_2D.data.RWP[s-1]));
soilDTF[13].setText(Double.toString(STM_2D.data.RWH[s-1]));
soilDTF[14].setText(Double.toString(STM_2D.data.Hurst[s-1]));
soilDTF[15].setText(Double.toString(STM_2D.data.RHH[s-1]));
}

return true;
}

// ==========================================================================
if(event.target == done)
{
this.dispose();
running = false;
return true;
}

// ==========================================================================
else return super.action(event, arg);

/*************************************************************************/
STM_2D_GetPrecipitationData.java

This class defines the GUI (Graphical User Interface) components to enter the precipitation data.

import java.awt.*;
import java.awt.event.*;

public class STM_2D_GetPrecipitationData extends Frame implements Runnable
{
    /************************************************************/
    /*** STM_2D_GetPrecipitationData.java                         ***/
    /************************************************************/
    /*** This class can be started as a user thread and asks for the rainfall ***/
    /*** characteristics.                                        ***/
    /************************************************************/
    /*** Programmed by:  ir. Jan Biesemans                        ***/
    /*** Ghent University                                       ***/
    /*** Dept. Soil Management                                  ***/
    /************************************************************/
    private int    i    ;
    private boolean running ;

    /************************************************************/
    /*** GUI components ***/
    /************************************************************/
    private Button done    ;
    private int DTFW    ;
    private Label[] rainlabel ;
    private STM_2D_DoubleTextField[] rainDTF ;
    private Panel   p1, p2, p3 ;

    /************************************************************/
    /*** Constructor ***/
    /************************************************************/
    public STM_2D_GetPrecipitationData()
    {
      running = true ;

      //Width of a Double Text Field (DTFW)
      //-------------------------------
      DTFW = 10 ;

      //Window colors
      //------------
      this.setBackground(Color.darkGray) ;
      this.setForeground(Color.white) ;
      this.setTitle("Precipitation Data") ;

      //Upper panel
      //--------
      p1 = new Panel() ;
p1.add(new Label("Rainfall Event Input Parameters"));
p1.setForeground(Color.red);
add(p1, "North");

//Center panel
//-----------------
p2 = new JPanel();
p2.setLayout(new GridLayout(17,2));
p2.setBackground(Color.darkGray);

rainlabel = new Label[16];
rainlabel[0] = new Label("Pluviophasic time step [s]:");
rainlabel[1] = new Label("Intensity pluviophasic 01 [mm/h]:");
rainlabel[2] = new Label("Intensity pluviophasic 02 [mm/h]:");
rainlabel[3] = new Label("Intensity pluviophasic 03 [mm/h]:");
rainlabel[4] = new Label("Intensity pluviophasic 04 [mm/h]:");
rainlabel[5] = new Label("Intensity pluviophasic 05 [mm/h]:");
rainlabel[6] = new Label("Intensity pluviophasic 06 [mm/h]:");
rainlabel[7] = new Label("Intensity pluviophasic 07 [mm/h]:");
rainlabel[8] = new Label("Intensity pluviophasic 08 [mm/h]:");
rainlabel[9] = new Label("Intensity pluviophasic 09 [mm/h]:");
rainlabel[10] = new Label("Intensity pluviophasic 10 [mm/h]:");
rainlabel[12] = new Label("Intensity pluviophasic 12 [mm/h]:");
rainlabel[13] = new Label("Intensity pluviophasic 13 [mm/h]:");
rainlabel[14] = new Label("Intensity pluviophasic 14 [mm/h]:");
rainlabel[15] = new Label("Intensity pluviophasic 15 [mm/h]:");

rainDTF = new STM_2D_DoubleTextField[16];
rainDTF[0] = new STM_2D_DoubleTextField(600.0,DTFW);
rainDTF[1] = new STM_2D_DoubleTextField(5.0,DTFW);
rainDTF[2] = new STM_2D_DoubleTextField(5.0,DTFW);
rainDTF[3] = new STM_2D_DoubleTextField(2.7,DTFW);
rainDTF[4] = new STM_2D_DoubleTextField(1.3,DTFW);
rainDTF[5] = new STM_2D_DoubleTextField(0.0,DTFW);
rainDTF[6] = new STM_2D_DoubleTextField(2.7,DTFW);
rainDTF[7] = new STM_2D_DoubleTextField(2.7,DTFW);
rainDTF[8] = new STM_2D_DoubleTextField(5.0,DTFW);
rainDTF[9] = new STM_2D_DoubleTextField(7.0,DTFW);
rainDTF[10] = new STM_2D_DoubleTextField(5.0,DTFW);
rainDTF[11] = new STM_2D_DoubleTextField(2.7,DTFW);
rainDTF[12] = new STM_2D_DoubleTextField(1.3,DTFW);
rainDTF[13] = new STM_2D_DoubleTextField(0.0,DTFW);
rainDTF[14] = new STM_2D_DoubleTextField(0.0,DTFW);
rainDTF[15] = new STM_2D_DoubleTextField(0.0,DTFW);

for(i = 0 ; i < 16 ; i++)
{
    p2.add(rainlabel[i]);
p2.add(rainDTF[i]);
}
add(p2, "Center");

//Bottom panel
//-----------------
p3 = new Panel() ;
done = new Button("Done/Cancel") ;
p3.add(done);
add(p3, "South") ;

//Set visibility
//-----------------------
setSize(500, 650) ;
setVisible(true) ;
}

/***************************************************************************/
/*** Run Method ***************************************************************************/
/***************************************************************************/
public void run()
{
    while(running)
    {
        try{Thread.currentThread().sleep(50) ;}
        catch(InterruptedException e) {}
    }

/***************************************************************************/
/*** Define button actions ***************************************************************************/
/***************************************************************************/
public boolean action(Event event, Object arg)
{
    //----------------------------------------------------------------------------
    if(event.target == done)
    {
        STM_2D.data.Ieff = new double[20] ;
        STM_2D.data.pdt  = rainDTF[0].getValue() ;

        for(i = 1 ; i <= 15 ; i++)
        {
            STM_2D.data.Ieff[i-1] = (rainDTF[i].getValue())/1000.0 ;
        }

        this.dispose() ;
        running = false ;
        return true ;
    }
    //----------------------------------------------------------------------------
    else return super.action(event, arg) ;
}
/******************************************************************************/
}
STM_2D_DoubleTextField.java

This class is a subclass of the JAVA TextField class. The latter is modified this way that only double formatted numbers can be entered.

```java
import java.awt.*;
import java.awt.event.*;

class STM_2D_DoubleTextField extends TextField implements TextListener {
    private String lastValue;
    private double DefaultValue;
    private int lastCaretPosition;

    public STM_2D_DoubleTextField(double defval, int size) {
        super(""+defval,size);
        DefaultValue = defval;
        addTextListener(this);
        this.setBackground(Color.blue);
        this.setForeground(Color.yellow);
        addKeyListener(new KeyAdapter()
            {
                public void keyTyped(KeyEvent evt)
                {
                    char ch = evt.getKeyChar();
                    if('0' <= ch && ch <= '9' || ch == '.' || Character.isISOControl(ch))
                        evt.consume();
                    else lastCaretPosition = getCaretPosition();
                }
            });
        lastValue = ""+defval;
    }

    public void textValueChanged(TextEvent evt) {
        checkValue();
    }
}```
/***********************************************
/*** Check entered characters                     ***/
 ***********************************************
private void checkValue()
{
    try
    {
        Double.valueOf(getText().trim()+"0").doubleValue();
        lastValue = getText();
    }
    catch(NumberFormatException e)
    {
        setText(lastValue);
        setCaretPosition(lastCaretPosition);
    }
}

/***********************************************
/*** Return the double value                    ***/
 ***********************************************
public double getValue()
{
    checkValue();
    try{return Double.valueOf(getText().trim()).doubleValue();}
    catch(NumberFormatException e){return DefaultValue;}
}
STM_2D_List.java

This class is a subclass of the JAVA List class, used to display some text results.

```java
public class STM_2D_List extends List
{
    // Override some methods of the parent class
    public Dimension minimumSize()
    {
        return new Dimension(500,200) ;
    }
    public Dimension minimumSize(int rows)
    {
        return new Dimension(500,200) ;
    }
    public Dimension preferredSize()
    {
        return new Dimension(500,200) ;
    }
    public Dimension preferredSize(int rows)
    {
        return new Dimension(500,200) ;
    }
}
```
STM_2D_Plot.java

This class defines the GUI (Graphical User Interface) components to draw simple XY graphs of the results.

```java
/* *******************************************************************/
/* *** STM_2D_Plot.java                                            ****/
/* *******************************************************************/
/* *** This class can be started as a user thread and plots an XY graph or an ****/
/* *** erosion/deposition graph.                                   ****/
/* *******************************************************************/
/* *** Programmed by:  ir. Jan Biesemans                          ****/
/* *** Ghent University                                            ****/
/* *** Dept. Soil Management                                       ****/
/* *******************************************************************/
import java.awt.*;
import java.text.*;
import java.math.*;
import java.awt.event.*;

public class STM_2D_Plot extends Frame implements Runnable
{
    private boolean running;
    private double[] x, y1, y2;
    private int N, mode;
    private String Info;

    /***********************************************************/
    /*** Run method                                           ****/
    /***********************************************************/
    public void run()
    {
        while(running)
        {
            try{Thread.currentThread().sleep(50);} 
            catch(InterruptedException e) {}
        }
    }
    /***********************************************************/
    /*** Constructor                                          ****/
    /***********************************************************/
    public STM_2D_Plot(double[] XX, double[] YY1, double[] YY2, int NN, int mmode, 
            String IInfo)
    {
        int i;

        //--------------------
        //-- Add a window listener --
        //--------------------
        addWindowListener(new WindowAdapter()
        {
            public void windowClosing(WindowEvent e)
            {
                
            }
        };
```
dispose();
}
}

//------------------
//-- Set frame dimensions --
//------------------
this.setTitle("STM-2D Results");
setSize(500,400);
setVisible(true);

//------------------
//-- allocate data --
//------------------
N = NN ;
mode = mmode ;
Info = IInfo ;

x = new double[N] ;
y1 = new double[N] ;
y2 = new double[N] ;

for(i = 0 ; i < N ; i++)
{
    x[i] = XX[i] ;
    y1[i] = YY1[i] ;
    y2[i] = YY2[i] ;
}

if(mode == 0) XY_Plot(x, y1, N, Info) ;
if(mode == 1) Bar_Plot(x, y1, y2, N, Info) ;

/*******************************************************************************/
/** XY graph plotting function  ***
/*******************************************************************************/
public void XY_Plot(double[] X, double[] Y, int N, String info)
{
    double maxY, maxX, minY, minX ; // min,max values
    double rangeX, rangeY ; // range values
    double help ; // general purpose
    int i, HE, WI ; // Font string Height and Width
    int helpix, helpiy ; // general purpose
    int dheight, dwidth ; // drawing area
    int wbx1=60,wbx2=20,wby=30 ; // distance between graph and window
    String fxy ; // axis label
    Font AxesFont = new Font("Helvetica", Font.PLAIN, 10) ;
    Graphics g = this.getGraphics() ;

    //----------------------------------
   //-- Determine data maximum and minimum value --
    //----------------------------------
    maxY = -1E30 ; minY = 1E30 ;
    maxX = -1E30 ; minX = 1E30 ;
    for(i = 0 ; i < N ; i++)
```javascript
{  
  if(X[i] > maxX) maxX = X[i] ;  if(X[i] < minX) minX = X[i] ;  
  if(Y[i] > maxY) maxY = Y[i] ;  if(Y[i] < minY) minY = Y[i] ;  
}  
rangeX = maxX - minX ;  
rangeY = maxY - minY ;

//----------------------------------------------------------
//-- Get frame dimensions --
//----------------------------------------------------------
Dimension d = getSize() ;
g.translate(getInsets().left, getInsets().top) ;
g.clearRect(0,0,d.width,d.height) ;

if((d.width < 300) || (d.height < 300))
{
  setSize(500,400) ;
  setVisible(true) ;
}
d = getSize() ;
dwidth = d.width - getInsets().left - getInsets().right ;
dheight = d.height - getInsets().top - getInsets().bottom ;

//----------------------------------------------------------
//-- Draw info string --
//----------------------------------------------------------
this.setBackground(Color.white) ;
g.setColor(Color.blue) ;
g.drawString(Info,wbx1,20) ;

//----------------------------------------------------------
//-- Draw grid lines --
//----------------------------------------------------------
g.setColor(Color.LightGray) ;
helpyi = (dheight - (2*wby))/6 ;
for(i = 0 ; i < 7 ; i++)
{
  g.drawLine(wbx1-3,
            wby+(i*helpyi),
            wbx1+(dwidth-(wbx1+wbx2))+3,
            wby+(i*helpyi)) ;
}

helpx = (dwidth - (wbx1+wbx2))/10 ;
for(i = 0 ; i < 11 ; i++)
{
  g.drawLine(wbx1+(i*helpx),
            wby-3,
            wbx1+(i*helpx),
            wby+(dheight-(2*wby))+3) ;
}

g.setColor(Color.black) ;
g.drawRect(wbx1,wby,helpx*10,helpyi*6) ;
```
---
//-- Draw axis labels --

DecimalFormat df = new DecimalFormat("0.00");
FontMetrics FM = g.getFontMetrics(AxesFont);

for(i = 0; i < 7; i++)
{
    help = maxX - (i*(rangeY/6.0));
    fxy = df.format(help);
    WI = FM.stringWidth(fxy);
    HE = FM.getHeight();
    g.drawString(fxy,wbx1-4-WI,wby+(i*helpy)+(HE/2));
}

for(i = 0; i < 11; i++)
{
    help = minX + (i*(rangeX/10));
    fxy = df.format(help);
    WI = FM.stringWidth(fxy);
    HE = FM.getHeight();
    g.drawString(fxy,wbx1-(WI/2)+(i*helpx),dheight-wby+4+HE);
}

//-- Draw data --

for(i = 0; i < (N-1); i++)
{
    g.drawLine((int)(wbx1+((dwidth-(wbx1+wbx2))/rangeX)*X[i]),
            (int)(dheight-wby-((dheight-(2*wby))/rangeY)*Y[i]),
            (int)(wbx1+((dwidth-(wbx1+wbx2))/rangeX)*X[i+1]),
            (int)(dheight-wby-((dheight-(2*wby))/rangeY)*Y[i+1]));
}

/**
 ** Bar chart plotting function
 **
 */
public void Bar_Plot(double[] X, double[] Y1, double[]Y2, int N, String Info)
{
    double maxY, maxX, minY, minX; // min,max values
    double rangeX, rangeY; // range values
    double help; // general purpose
    int i, HE, WI; // Font string Height and Width
    int helpix, helpiy; // general purpose
    int dheight, dwidth; // drawing area
    int wbx1=60,wbx2=20,wby=30; // distance between graph and window
    String fxy; // axis label
}
Font AxesFont = new Font("Helvetica", Font.PLAIN, 10);
Graphics g = this.getGraphics();

//------------------------
//-- Determine data maximum and minimum value --
//------------------------
maxY = -1*E30; minY = 1*E30;
maxX = -1*E30; minX = 1*E30;

for(i = 0; i < N; i++)
{
  if(X[i] > maxX) maxX = X[i]; if(X[i] < minX) minX = X[i];
  if(Y1[i] > maxY) maxY = Y1[i]; if(Y1[i] < minY) minY = Y1[i];
  if(Y2[i] > maxY) maxY = Y2[i]; if(Y2[i] < minY) minY = Y2[i];
}

if(Math.abs(minY) > maxY) maxY = Math.abs(minY);
else
  minY = -maxY;

rangeX = maxX - minX;
rangeY = maxY - minY;

//------------------------
//-- Get frame dimensions --
//------------------------
Dimension d = getSize();
g.translate(getInsets().left, getInsets().top);
g.clearRect(0,0,d.width,d.height);

if((d.width < 300) || (d.height < 300))
{
  setSize(500,400);
  setVisible(true);
}
d = getSize();
dwidth = d.width - getInsets().left - getInsets().right;
dheight = d.height - getInsets().top - getInsets().bottom;

//-- Draw info string --
//--
this.setBackground(Color.white);
g.setColor(Color.blue);
g.drawString(Info, wbx1, 20);

//-- Draw grid lines --
//--
g.setColor(Color.lightGray);
helpiy = (dheight - (2*wby))/6;
for(i = 0; i < 7; i++)
{
  g.drawLine(wbx1-3,
             wby+(i*helpiy),
             wbx1+(dwidth-(wbx1+wbx2))+3,
wby+(i*helpiy));
}

helpix = (dwidth - (wbx1+wbx2))/10;
for(i = 0 ; i < 11 ; i++)
{
    g.drawLine(wbx1+(i*helpix),
        wby-3,
        wbx1+(i*helpix),
        wby+(dheight-(2*wby)+3);
}

g.setColor(Color.black);
g.drawRect(wbx1,wby,helpix*10,helpiy*6);

//----------------------
//-- Draw axis labels --
//----------------------
g.setColor(Color.black);
g.setFont(AxesFont);

DecimalFormat df = new DecimalFormat("0.000");
FontMetrics FM = g.getFontMetrics(AxesFont);

for(i = 0 ; i < 7 ; i++)
{
    help = maxY - (i*(rangeY/6.0));
    fxy = df.formatString(help);
    WI = FM.stringWidth(fxy);
    HE = FM.getHeight();
    g.drawString(fxy,wbx1-4-WI,wby+(i*helpiy)+(HE/2));
}

df = new DecimalFormat("0.00");
for(i = 0 ; i < 11 ; i++)
{
    help = minX + (i*(rangeX/10));
    fxy = df.formatString(help);
    WI = FM.stringWidth(fxy);
    HE = FM.getHeight();
    g.drawString(fxy,wbx1-(WI/2)+(i*helpix),dheight-wby+4+HE);
}

//------------
//-- Draw data --
//------------
g.setColor(Color.orange);
g.setClip(wbx1+1,wby+1,(helpix*10)-1,(helpiy*6)-1);

for(i = 1 ; i < N ; i++)
{
    help = (dheight-(2*wby))/rangeY;

    g.fillRect3DRect((int)(wbx1+(((dwidth-(wbx1+wbx2))/rangeX)*X[i-1])),
        (int)(wby+((maxY-Y1[i])*help)),
        (int)(((dwidth-(wbx1+wbx2))/rangeX)*(X[i]-X[i-1]))+1,
(int)(help*Y1[i]),
true);
g.fill3DRect((int)(wbx1+(((dwidth-(wbx1+wbx2))/rangeX)*X[i-1])),
(int)(wbx+(maxY*help)),
(int)(((dwidth-(wbx1+wbx2))/rangeX)*(X[i]-X[i-1]))+1,
(int)(help*Math.abs(Y2[i])),
true);
}

/*******************************************************************************/
/******** Paint method: called when the Frame is manipulated (resized, etc...) ******/
/*******************************************************************************/
public void paint(Graphics g)
{
  if(mode == 0) XY_Plot(x, y1, N, Info);
  if(mode == 1) Bar_Plot(x, y1, y2, N, Info);
}

/*******************************************************************************/
STM_2D_About.java

This class is used to give some general information.

```java
import java.awt.*;
import java.awt.event.*;

public class STM_2D_About extends Frame implements Runnable {
    /***********************
    /// General data ///
    ***********************
    private boolean running;
    private Runtime rt = Runtime.getRuntime();
    private long mem_tot = rt.totalMemory();
    private long mem_free = rt.freeMemory();

    /**************
    /// GUI components ///
    **************
    private Button OK;
    private Panel p1, p2;

    /*********************
    /// Constructor ///
    *********************
    public STM_2D_About() {
        running = true;

        //Window colors
        //-----------
        this.setBackground(Color.black);
        this.setForeground(Color.white);
        this.setTitle("STM-2D About");

        //Center panel
        //--------
        p1 = new Panel();
        p1.setLayout(new GridLayout(14,1));
        p1.add(new Label("STM-2D Java Applet",Label.CENTER));
        p1.add(new Label("Sediment Transport Model - 2D version",Label.CENTER));
        p1.add(new Label(" "));
        p1.add(new Label("Developed and Programmed by ir. Jan Biesemans",Label.CENTER));
        p1.add(new Label("Jan.Biesemans@rug.ac.be",Label.CENTER));
```
p1.add(new Label("Ghent University",Label.CENTER));
p1.add(new Label("Department of Soil Management",Label.CENTER));
p1.add(new Label("Coupure Links 653, 9000 GENT, BELGIUM",Label.CENTER));
p1.add(new Label(" ") );
p1.add(new Label(" ") );
p1.add(new Label("Your Java Version: "+System.getProperty("java.version"), Label.CENTER));
p1.add(new Label("Total Allocated Memory (kB): "+Long.toString(mem_tot/1024), Label.CENTER));
p1.add(new Label("Free Memory (kB): "+Long.toString(mem_free/1024), Label.CENTER));
add(p1, "Center") ;

//Bottom panel
//--------
p2 = new Panel() ;
OK = new Button(" OK ") ;
p2.add(OK);
add(p2, "South") ;

//Set visibility
//--------
setSize(450,400) ;
setVisible(true) ;
}

="/****************************** Run Method ****************************/
public void run() {
while(running) {
try{Thread.currentThread().sleep(50) ;}
catch(InterruptedException e) {}
}

="/****************************** Define button actions ****************************/
public boolean action(Event event, Object arg) {
//------------------------------
if(event.target == OK) {
this.dispose() ;
running = false ;
return true ;
//-------------------------------
else return super.action(event, arg) ;
}

="/****************************** */
Appendix B

Publications

Parts of this thesis were published in:

