2 THEORY OF SOIL EROSION MODELLING

2.1 Introduction

In chapter 1 some characteristics peculiar to catchments on the Chinese Loess Plateau were identified:

1. Slope angles are steep. This can have consequences for both velocity and transport capacity of the flow.
2. Sediment concentrations in runoff can be very high.
3. Large permanent gullies are common.

This chapter will discuss the erosion processes that are operating on the Loess Plateau, and especially those that are relevant to the characteristics mentioned above. The chapter will also explore to what extent these processes are at present being taken into account in soil erosion modelling. Chapters 5 to 8 will then, in turn, discuss the implications of each of these characteristics for the study area.

2.2 Flow velocity and flow routing

Two dimensionless numbers are used in hydrology to classify flow type. These are the Reynolds number (Re) and the Froude number (Fr). The Reynolds number is used to determine whether a flow is laminar, turbulent or something in between (transitional). The Froude number is used to determine whether flow is sub-critical or super-critical.

The Reynolds number is given by:

\[
Re = \frac{4 \cdot V \cdot R}{\nu}
\]  

(2.1)

Where:

- \( V \) = flow velocity (m/s)
- \( R \) = hydraulic radius (m)
- \( \nu \) = kinematic viscosity (m²/s)

At what Reynolds number flow will be laminar, transitional or turbulent is not clearly defined. However, the boundary between laminar and transitional flow is usually placed between 1500 and 2000 (e.g. Emmett, 1970, Abrahams et al., 1986, Ven Te Chow et al., 1988, Li & Abrahams, 1997), while the same authors placed the boundary between transitional and turbulent flow at 6000 to 10000. The main difference between laminar flow and turbulent flow is the velocity distribution over depth that results from different degrees of vertical mixing. In laminar flow the flow can be envisioned as parallel layers that move over each other, but that do not mix. This results in a clear velocity gradient from bottom to top of the flow, where average velocity of the flow is theoretically 0.67 times the surface velocity (e.g. Emmett, 1970). In turbulent flow eddies are formed that disturb the velocity distribution of the flow and that cause energy loss. The result is that the velocity gradient is less than for laminar flow, and the average velocity is about 0.8 times the surface velocity (Emmett, 1970).
The Froude number is given by:

\[ Fr = \frac{V}{\sqrt{g \cdot h}} \]  \hspace{1cm} (2.2)

Where: \( g \) = gravitational acceleration (m/s²)
\( h \) = water depth (m)

For Froude numbers below 1 flow is sub-critical, while for values above 1 it is super-critical. In super-critical flow any disturbances of the flow can only propagate in the downstream direction. Using the Reynolds and Froude number flows can be classified as being laminar-sub-critical, laminar-super-critical, turbulent-sub-critical or turbulent-super-critical.

Water velocity in erosion models is usually calculated with empirical formulae such as the Manning equation or the Darcy-Weisbach equation. Such velocity equations might or might not be applicable to some kinds of flow. It is, for example, sometimes stated that the Manning equation may only be applied to turbulent flow (e.g. Ven Te Chow et al., 1988). The velocity equations incorporate the slope angle, but were not developed for such steep slope angles as present in the Loess Plateau catchments. Since the equations are of empirical nature one should be careful with using them for conditions outside those for which they were developed. Therefore, the applicability of these equations will be evaluated in chapter 6.

To route flow to the catchment outlet a continuity equation as well as a flux-concentration equation are needed (Singh, 2002). For water flow the velocity equation (Manning, Chezy, Darcy-Weisbach) is the flux-concentration equation. Either of these equations is a specific form of the following equation:

\[ A = \alpha \cdot Q^\beta \]  \hspace{1cm} (2.3)

Where \( A \) is the cross-sectional area of the flow, \( Q \) is the discharge, \( \alpha \) is a parameter and \( \beta \) a coefficient. Both \( \alpha \) and \( \beta \) are often assumed constant, but might in reality vary spatially and with flow conditions (Singh, 2002). When the Manning equation is used \( \beta \) would be 0.6 and \( \alpha \) would be:

\[ \alpha = \left( \frac{n \cdot P^{2/3}}{S^{1/2}} \right)^{0.6} \]  \hspace{1cm} (2.4)

Where \( P \) is the wetted perimeter, \( S \) the energy slope and \( n \) Manning’s \( n \).

The continuity equation, in its basic 1-dimensional form, is given by:

\[ \frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = q \]  \hspace{1cm} (2.5)
Where \( q \) is lateral inflow, which in case of overland flow would be defined as rainfall minus infiltration. Equations 2.3 and 2.5 can be combined to give (Ven Te Chow et al., 1988):

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

(2.6)

This is the kinematic wave equation. According to Singh (2002) the kinematic wave can be applied to both overland flow and channel flow. According to him, kinematic waves are dominant for Froude numbers below one, while for higher Froude numbers dynamic waves are more dominant. The assumption behind the kinematic wave is that the friction slope is equal to the bed slope. This assumption is more realistic for steeper slopes (Fread, 1985; 1993; Singh, 2002). If the assumption is not realistic, or if Froude numbers are high, more complete versions of the Saint Venant equation, such as the diffusion wave and dynamic wave, should be used. The kinematic wave equation is usually solved by numerical methods. These methods transform the governing partial differential equation into a set of finite-difference equations by using a Taylor series expansion (Ven Te Chow et al., 1988). This transformation introduces several types of error:

- Truncation error. The higher order derivatives are dropped from the Taylor series expansion.
- Rounding error. Only a certain number of significant digits are used.
- Numerical errors that are generated because the continuous partial differential equation is transformed into a set of finite difference equations that are only valid for the grid points in the \( x-t \) plane. Between grid points, values are obtained by linear interpolation.

If these errors do not amplify during successive time steps the solution is stable (Ven Te Chow et al., 1988). Stability of the solution, however, does not guarantee that the solution is also accurate. Although the kinematic wave equation does not allow for wave attenuation, attenuation will occur because of the numerical errors associated with the finite difference solution of the kinematic wave (Fread, 1993).

### 2.3 Sediment transport

#### 2.3.1 Introduction

Sediment transport is an important process in studies on soil erosion. Through this process eroded sediment is removed from the catchment. The ratio between sediment transported out of the catchment and sediment eroded in the catchment is called the sediment delivery ratio. The sediment delivery ratio usually decreases with increasing catchment area because in larger catchment there is more opportunity for sediment storage, e.g. in floodplains. By far the most important transporting agent on the hilly part of the Chinese Loess Plateau is flowing water, which can also be a major cause of erosion. Flowing water exerts a force on its bed that, in terms of stress, can be expressed as:
\[ \tau = \rho_f \cdot g \cdot R \cdot S \]  

(2.7)

Where:  
- \( \tau \) = shear stress (kg m\(^{-1}\) s\(^{-2}\) = N/m\(^2\))  
- \( \rho_f \) = fluid density (kg m\(^3\))  
- \( g \) = gravitational acceleration (m s\(^{-2}\))  
- \( R \) = hydraulic radius (m)  
- \( S \) = energy slope (m m\(^{-1}\))

Another important parameter of the flow with respect to sediment transport is stream power. It can be expressed in many different ways (see Rhoads, 1987). The stream power per unit wetted area (or mean stream power, Rhoads, 1987) is given by:

\[ \Omega = \rho_f \cdot g \cdot R \cdot S \cdot V = \tau \cdot V \]  

(2.8)

Where:  
- \( \Omega \) = stream power per unit wetted area (kg s\(^{-3}\))  
- \( V \) = flow velocity (m/s)

The product of \( S \) and \( V \) is called unit stream power. It represents the power per unit weight of water. The energy slope \( S \) is equal to the sine of the slope angle (Rhoads, 1987; Ven Te Chow et al., 1988; Flanagan et al., 2001), and can only be equated with tangent for gentle slopes.

Water can transport sediment in several ways. The total sediment load of flowing water is usually subdivided into bedload and suspension load. Suspended load is sometimes subdivided into suspended bed material load and wash load. Hsieh Wen Shen & Julien (1993) give the following characteristics of these types of load:

**Bedload:** sediment particles moving along the streambed in the processes of rolling, sliding, and/or hopping.

**Suspended load:** sediment particles that are supported by the turbulent motion of the flow. Suspended load has a vertical distribution in the flow. Concentrations are larger at the bed than at the surface. This distribution is caused by a balance between falling due to gravity and upward transport due to turbulence. The finer the particles and the more turbulent the flow, the more evenly the distribution of particles over depth will be.

**Suspended bed material load:** suspended load in which the particles are large enough to be seen on the streambed.

**Wash load:** suspended load in which the particles are so small that they cannot be easily seen individually on the streambed. Wash load does not depend directly on flow conditions, but more on supply rate. Wash load is almost uniformly distributed over depth.

The concentration of suspended sediment in water can be defined in several different ways. First, the amount of sediment can be expressed as volume of sediment or as mass of sediment. Second, the amount of water can be defined as the total amount of fluid or as the amount of clear water. Hence, the following definitions may be used:
\[ C_f = C_{vf} \cdot \rho_s = \frac{V_s}{V_f} \cdot \rho_s = \frac{V_f}{V_s + V_w} \cdot \rho_s \] (2.9)

\[ C_w = C_{vw} \cdot \rho_s = \frac{V_s}{V_w} \cdot \rho_s \] (2.10)

Where:
- \( C_f \) = fluid concentration (g/l) = dirty water concentration
- \( C_{vf} \) = volumetric fluid concentration
- \( C_w \) = clear water concentration (g/l)
- \( C_{vw} \) = volumetric clear water concentration
- \( V_s \) = volume of solids
- \( V_f \) = volume of fluid
- \( V_w \) = volume of water
- \( \rho_s \) = density of solids (kg/m³, which is numerically equal to g/l)

Most authors use \( C_f \), but \( C_w \) is also sometimes used. Unfortunately, authors do not always state which definition they use. This can have large implications when concentrations are large, as on the Chinese Loess Plateau. The amount of bedload that is moving is usually not expressed as a concentration, but as a sediment flux:

\[ q_b = q_s \cdot \rho_s \] (2.11)

Where \( q_b \) is volumetric bedload transport per unit width of flow (m²/s) and \( q_s \) is the sediment transport rate in kg m⁻¹ s⁻¹. Thus, the amount of sediment is explicitly linked to the amount of water for suspended load, but not for bedload. For bedload the link is more implicit, since \( q_b \) will be determined by shear stress or stream power of the flow, which depends on discharge.

2.3.2 Concept of transport capacity

Present day process based soil erosion models like WEPP (Flanagan et al., 2001), KINEROS2 (Smith et al., 1995), EUROSEM (Morgan et al., 1998a,b) and LISEM (Jetten & De Roo, 2001) use the concept of transport capacity to determine sediment transport rates in overland flow and streamflow. Smith et al. (1995) defined transport capacity as the amount of sediment that a given flow can carry at steady state conditions in equilibrium with a loose bed. Detachment and deposition are then functions of transport capacity:

\[ D = a \cdot (TC - C) \] (2.12)

Where:
- \( TC \) = transport capacity
- \( C \) = concentration
- \( a \) = rate control constant
- \( D \) = detachment or deposition
If transport capacity exceeds concentration net erosion will occur at rate \( a \). If concentration exceeds transport capacity net deposition will occur at rate \( a \). The rate control constant might be different for erosion and deposition, e.g. because it will depend on soil cohesion in the case of detachment. If equation 2.12 is used, transport capacity becomes the controlling factor in determining whether net erosion or net deposition occurs. There are several potential limitations to this concept:

1) Huang et al. (1999) argued that equation 2.12 is based on the assumption that there is a coupling between transport and detachment. It can be easily seen from equation 2.12 that if \( TC-C \) increases \( D \) will increase. They stated that this approach does not do justice to the fact that there might be a limit to \( D \) that depends not on \( TC-C \) but on some other factor like cohesion or soil strength. In equation 2.12 the rate control constant \( a \) might depend on cohesion, so that \( D \) is also cohesion dependent, but despite that \( D \) will always increase if \( TC-C \) increases. Therefore, equation 2.12 cannot cope with situations were transport is detachment limited instead of transport limited. Huang et al. do not seem to have taken into account that \( D \) is a net rate (Morgan et al., 1998a) and cannot be equated with either deposition or detachment rate since both will occur simultaneously in reality. Therefore, in the case of net erosion, \( D \) might continue to increase with increasing \( TC-C \) even if erosion rate has reached its upper limit. Nevertheless, a method that calculates erosion and deposition independently, as suggested by Huang et al. is conceptually clearer. Rose (1985) described such a method. In his method there is no need for an a priori definition of transport capacity, instead a ‘transport capacity’ will automatically emerge when detachment equals deposition. Such an approach, however, does not allow a check for impossible concentrations. Rose (1985) found that in practice the prediction is not much different from the results obtained with a method that explicitly uses transport capacity.

2) Transport capacity is a predefined number that depends on flow and sediment characteristics, but not on sediment concentration. It will be shown in chapter 5 that in the case of the Loess Plateau transport capacity might depend on concentration. One way around this might be to treat both erosion and deposition independently (as described above) and to calculate transport as the sum of both. Another option would be to redefine transport capacity to incorporate effects of concentration. Huang et al. (1999) found that transport capacity also depends on surface hydrologic conditions such as drainage and seepage.

3) The concept of transport capacity might not be useful for wash load (grain size below about 50 µm), since the concentration of wash load depends mainly on availability of material and not on flow conditions (e.g. Hsieh Wen Shen & Julien, 1993, Reid et al., 1997). Wash load can apparently be transported in almost limitless quantities (Van Rijn, 1993). Using separate sediment classes with different transport capacity might circumvent this problem. The finest material could then be given a very high transport capacity. This also results in a shift of grain size distribution of transported sediment in comparison to the original soil. The Rose (1985) and WEPP models currently use different sediment size classes.
Thus, the concept of transport capacity is not without problems. Nevertheless the approach of using transport capacity as the controlling factor in net erosion and net deposition seems valid, though it might be necessary to apply transport capacity equations that take local circumstances into account.

### 2.3.3 Transport equations

Many, mostly empirical, equations have been developed to predict sediment transport from flow characteristics, slope and material characteristics. These equations often use a threshold value for stream power, shear stress or discharge. Below this threshold no sediment transport will take place. A distinction is often made between channel flow and overland flow. Overland flow usually has much steeper slopes and much lower discharge than channel flow. Another distinction in equations is that in bedload equations and total load equations. Neither type specifically includes wash load. Some authors claim (Borges et al, 1995, Smart & Jaeggi, 1983, Rickenmann, 1991) that their formula is applicable to steep slopes, but this usually means up to slopes of only about 20%. Furthermore, almost all formulae for channel flow are bedload formula. Some equations for channel flow and rill flow will be discussed in chapter 7.

For overland flow less equations are available. Huang (1995) studied soil loss from 1.2-m soil pans with slopes ranging from 4-30%. He found that concentration was best predicted using stream-power based polynomials of the form:

\[
C = D_1 \cdot q^2 \cdot S^2 + D_2 \cdot q \cdot S + D_3
\]  

(2.13)

Where the \(q\) is discharge, \(S\) is slope and \(D_1\) to \(D_3\) are coefficients depending on soil type.

Everaert (1991) performed measurement in a very small flume (0.05 by 0.3 m) to test transport capacity for interrill conditions with laminar flow regime. Slopes ranging from 1 to 10 degrees were used, while discharge was between 0.2 and 2.5 cm²/s. He used several grain sizes, the smallest of those was 33 µm. The results were best predicted with effective stream power (depth corrected stream power), but good results could also be obtained with shear velocity, unit stream power and a combination of \(q\) and \(S\). His unit stream power equation for a grainsize of 33 µm is:

\[
V = \log(q_s) = -1.31 + 1.51 \cdot \log(S \cdot V)
\]  

(2.14)

Where \(V\) is given in cm/s and \(q_s\) is predicted in g cm⁻¹ s⁻¹.

Neither Huang (1995) nor Everaert (1991) discussed whether or not the small plot size used by them is large enough to reach transport rates that equal transport capacity.

As Beschta (1987) noted each equation has usually been developed for a limited range of conditions and when used in field application the estimated transport rates for the different equations may vary over several orders of magnitude. There is no such thing as
a universally applicable transport equation. Many studies have evaluated the use of different transport equations under different circumstances (e.g. channel flow: Van den Berg & Van Gelder, 1993, Van Rijn, 1993, Hossain & Rahman, 1998; overland flow: Alonso et al., 1981, Govers, 1992a, Guy et al., 1992; flume data: Low, 1989, Lu et al., 1989). Almost all studies reached different conclusions about the suitability of certain equations. Some of the studies mentioned here will be discussed in more detail in chapter 7. In this section, some attention will be given to the more theoretical evaluation of transport equations by Julien & Simons (1985) and Prosser & Rustomji (2000).

Prosser and Rustomji (2000) reason that discharge ($q$) and slope ($S$) are the basic controlling factors in sediment transport and that other parameters such as shear stress and stream power are derived from these two basic parameters. Therefore, they expressed a large number of equations in terms of $q$ and $S$ to make comparison possible:

$$TC = A \cdot q^a \cdot S^b$$

(2.15)

To express the different equations in this way it is necessary to neglect the threshold values in the equations. The $a$ and $b$ coefficients were calculated with different methods ($q$&$S$, shear stress, stream power) and compared. The results showed that the resulting $a$ and $b$ coefficients depended on the method used, but that $a$ and $b$ were comparable for all types of experiment (lab-plot, plot, river). Only flume-studies gave slightly different results. The median of both $a$ and $b$ was found to be 1.4, and both ranged between about 1.0 and 1.8. Any equation with coefficients within this range would be valid if the choice for that particular equation can be justified for the specific conditions to which it is applied.

Julien & Simons (1985) reviewed a number of bedload equations for their applicability to overland flow. They rewrote the equations to a form similar to that of equation 2.15 and compared the exponents of slope ($b$) and discharge ($a$) with equations developed for laminar flow. They assumed that overland flow is laminar and found that the discharge exponent ($a$) varies with type of flow, but that the slope exponent ($b$) is fairly constant. The discharge exponent of streamflow equations (turbulent flow) is generally lower than that of overland flow equations (laminar flow). Only the equations by Engelund-Hansen and Barekyan were found to be relevant to overland flow.

A number of transport equations will be evaluated for the Danangou catchment in chapter 7. An equation that could not be tested, but that is nevertheless interesting was developed by Abrahams et al. (2001). In recognition of the fact that the use of transport equations is often hampered by the limited range of conditions for which they were developed, Abrahams et al. (2001) used a very large data set obtained from flume experiments to develop a total load transport equation for interrill flow. Experiments were conducted in a 5.2 metre long flume and were performed under a wide range of conditions with respect to: flow depth and velocity, Reynolds number, Froude number, slope, sediment size, sediment concentration, roughness concentration and diameter, flow density and viscosity. Based on a dimensional analysis the following transport equation was obtained:
\[ q_s = a \cdot D \cdot U_\ast \cdot Y \left( 1 - \frac{Y_c}{Y} \right)^b \left( \frac{V}{U_\ast} \right)^c \left( \frac{w_i}{U_\ast} \right)^d \]  

(2.16)

\[ w_i = \left( \frac{g \cdot (\rho_f - \rho_s) \cdot D}{\rho_f} \right)^{0.5} \]  

(2.17)

Where:
- \( q_s \) = sediment transport rate (m²/s)
- \( D \) = median grain size (m)
- \( U_\ast \) = shear velocity (m/s, see equation 7.4)
- \( Y \) = Shields parameter (see equation 7.3)
- \( Y_c \) = critical value of Shields parameter
- \( V \) = flow velocity (m/s)
- \( w_i \) = inertial fall velocity (m/s)
- \( g \) = gravitational acceleration (m/s²)
- \( \rho_f \) = fluid density (kg/m³)
- \( \rho_s \) = sediment density (kg/m³)
- \( a-d \) = coefficients

This equation is interesting since it was developed using data obtained under a range of conditions. For application on the Loess Plateau especially maximum volumetric concentration (0.3), minimum grain size (98 mu) and maximum slope (10 degrees) are relevant. These values compare favourably with those of some other transport equations (see chapter 7), but grain size is still too large and slope angle too low for Loess Plateau conditions.

### 2.4 High concentrations

The effects of high concentrations on streamflow and sediment transport have traditionally been studied more in the context of hydraulics than of hydrology. Flows with extreme concentrations such as debris flows have received much attention and have been modelled with different degrees of success. Flow of debris flows is very different from channel flow. Soil erosion models, however, have not paid any specific attention to high sediment concentrations. For most regions, concentrations in runoff will not be very high, so that no special attention is needed. For the Loess Plateau, however, very high concentrations have been reported regularly. These kinds of flow occupy intermediate positions between clear water flow and debris flow and could have properties that differ significantly from clear water flow. More specifically high sediment concentrations could change density, viscosity, resistance to flow, velocity profile and transport capacity. This subject, therefore, requires special attention in the case of the Loess Plateau, and will be discussed in chapter 5.
2.5 Gullies

2.5.1 Introduction

Nordström (1988) summarised several definitions of gullies. She mentioned the following characteristics of gullies: a steep incised channel, often with a headcut, no permanent water, evidence of present or past rapid extension and the incision is mainly formed in unconsolidated materials. Gully erosion can have major effects, both on-site (due to soil loss) and off-site (due to sediment). On-site agricultural land can be lost, while off-site the major consequences are flooding and silting up of reservoirs.

2.5.2 Channel head

Dietrich & Dunne (1993) defined the channel head as the upstream boundary of concentrated flow between definable banks. The upper gully head will therefore often (though not always) coincide with the channel head. The position of the channel head in the landscape has been a topic of investigation for a long time in theoretical geomorphology. Channel initiation by overland flow can be viewed in two ways (Dietrich & Dunne, 1993, Montgomery & Dietrich, 1994, Kirkby, 1994, Prosser & Dietrich, 1995, Bull & Kirkby, 1997): as a balance between erosion and infilling and as a threshold phenomenon.

The first approach is called the instability view by Kirkby (1994) and Prosser & Dietrich (1995). It was developed by Smith & Bretherton (1972). This approach assumes that at every point of a slope incision processes and diffusion processes are operating. On the higher parts of the slopes diffusion processes will dominate because incision is limited by lack of water. Further downslope incision dominates. Channel initiation is then assumed to occur at the point where incision starts to dominate over dissipation. This will usually be around the inflexion point. Valleys tend to be without a sharp edge.

The second approach (the threshold view) assumes that for channel initiation to occur a threshold must be exceeded. In other words, erosivity of overland flow must surpass resistance of the soil. This threshold is usually expressed in term of shear stress or stream power. The critical shear stress (or critical stream power) will depend on the properties of the soil. Shear stress itself is related to discharge and slope. In the case of Hortonian overland flow, discharge should be proportional to drainage area. Valleys tend to have sharp edges and headcuts. This concept was proposed by Horton (1945). The threshold approach can also be applied to other processes such as sapping and mass movements (Montgomery & Dietrich, 1994). Gerits et al. (1987) used a threshold approach to explain piping on badland slopes. The link between threshold and process might not always be straightforward as reaction time and relaxation time can also play a role, so that the system could still be reacting to some threshold exceedance in the past.

Several authors (Montgomery & Dietrich, 1994; Kirkby, 1994; Rauws, 1987) mentioned the compatibility of these two views. No incision will occur until the threshold is exceeded; afterwards incision and diffusive processes will both operate. Kirkby (1994) stated that threshold behaviour is likely to dominate over instability behaviour in semi-arid areas,
while the reverse is true for humid areas. Prosser & Dietrich (1995) reasoned that the
threshold approach is most appropriate for materials with cohesion, while the instability
approach should be used in the case of cohesionless materials.

2.5.3  Gully processes

The main difference between rills and gullies is their size but processes operating in gullies
can also differ from processes in rills. Imeson and Kwaad (1980) stated that gullies
resemble river valleys, while rills resemble river channels in their behaviour. Kalman
(1976) stated that rills are in principle self-stabilising, while gullies are not. Rills are most
likely formed by overland flow erosion, but gullies can develop in several ways.

Role of overland flow

Overland flow is likely to play a role in gully formation in semi-arid regions since
infiltration excess overland flow (or hortonian overland flow) is likely to be important in
semi-arid regions because of high rainfall intensities. The assumption of hortonian overland
flow can however also be used in other regions, for example on soils with low permeability.
It has the advantage that drainage area can be used instead of discharge (at least in steady
state conditions), which is much harder to measure. Vandaele et al. (1996) used upslope
drainage area and slope gradient to predict the initiation of rills and gullies. They found that
larger drainage areas were necessary for rill initiation when there was more stone cover or
vegetation. Antecedent moisture conditions proved also to be important. Overland flow is
concentrated because of small accidental variations in topography (Bryan, 1987). Because
of concentration shear stress will increase rapidly. Overland flow shear stress (or stream
power) must exceed a threshold value, the value of which is determined by surface material
properties (such as cohesion, texture and aggregate stability) and by vegetation. Rauws
(1987), for example, found the following thresholds for rill initiation: slopes of more than 2
degrees and flow velocities of more than 3 - 3.5 cm/s. According to Prosser & Dietrich
(1995) and Prosser (1996) vegetation can increase the threshold shear stress several times.
It is often assumed that this threshold will be exceeded when flow becomes turbulent (Loch
& Thomas, 1987), while it is also often stated that the flow must be able to transport all
particle sizes (the flow is non-selective) for a rill to form (Bryan, 1987, Torri et al, 1987,
Rauws, 1987, Crouch & Novruzi, 1989). When a rill is formed transport capacity increases
greatly. The rill can grow into a gully when it is not removed by hillslope processes
(diffusive) or management by man. Because of headcut migration the gully can retreat into
un-rilled areas, thereby obliterating the rills that preceded its formation. Gullies formed by
overland flow tend to be v-shaped. If a gully is u-shaped, seepage flow probably plays a
role, even though overland flow might still be dominant (Imeson & Kwaad, 1980).

Role of piping

Piping can be caused by two mechanisms. Firstly, a pipe can be formed by animal activity
or plant root decay. Alternatively, it can be formed because of seepage. Seepage erosion
(also called sapping) sensu stricto can only occur in cohesionless materials, because the
process involves liquefaction (Montgomery & Dietrich, 1994). Pipes can, however, also be
formed in cohesive material. Chemistry often plays an important role, as it controls
swelling/shrinking properties of materials. Swelling/shrinking and dispersion are mainly
controlled by the type and amount of clay. Bocco (1990) mentioned the following factors to be helpful in piping: high soil dispersion, soil cracking, steep hydraulic gradients and a convex slope profile. Most of these factors are also mentioned by Nordström (1988). Pipes can collapse to form lines of sinkholes and ultimately gullies. Pipe directions are mainly controlled by subsoil properties, such as geology and the direction of joints and faults, and can therefore differ from surface water directions.

**Role of mass movements**

Mass movements can be important in gully development because gully side slopes can be quite steep. Undercutting and seepage can both be important. Collison (1996) found that most material from gullies is produced by head and wall instability. He assumes that instability of the gully head is enhanced by overland flow infiltrating into cracks near the edge of the gully. Govers (1987) assumed that rill widening is caused by mass movements and rill deepening by overland flow incision. Rill incision thus creates the opportunity for mass movements to occur on the rill walls, but the actual mass movements on the rill/gully walls need not occur at the time of the storm. Govers (1987) viewed mass movements as delayed reactions to gully incision by water flow. Water distribution in the soil is important for stability, and this process takes time.

The above-mentioned processes are not all equally effective in producing gullies. Sapping, for example, is most effective when a free face is present. As a result sapping is more efficient in headcut migration than in actually forming a new gully on an undisturbed slope.

**2.5.4 Gully development**

To be able to predict the future behaviour of gullies it is necessary to understand the development of gullies over time. Gully heads tend to migrate because of e.g. plunge-pools, seepage, mass wasting and headcut erosion by flowing water. This decreases the upstream area of the gully head, which should eventually result in stabilisation of the gully head because less water is available at the headcut. Side slope processes will however continue and will dominate over incision. The upstream area of the gully outlet will remain constant, because as flow over the headscarp is decreased flow over the sidewalls is increased (Burkard and Kostaschuk, 1997). All material that is delivered to the gully must be transported by the water flowing into the gully. If sediment transport is decoupled from sediment production stabilisation can occur (Harvey, 1994). Imeson and Kwaad (1980) also stressed this balance between sediment production in gullies and sediment removal from gullies. Coupling is likely to decrease with growing gully size. Gully erosion should therefore become a transport-limited process, and stabilisation could occur. The sequence of events can also be important in gully development. Moderate rainfall events might for example be capable of enlarging existing gullies, without being able to form new ones. If loose material is produced between rainfall events, the first storm after a dry period might produce much sediment, while later storms of equal magnitude produce less sediment. It also means that a series of small events might cause no overland flow erosion, but could result in for example mass movements. This makes it harder to relate rainfall amount or rainfall intensity to the observed amount of erosion. Ultimately, gully development will,
however, be controlled by base level. If base level remains constant, sidewall stabilisation will occur eventually.

2.5.5 Gully erosion and soil conservation

From the foregoing discussion, it should be clear that soil conservation methods should be adapted to the soil erosion process that is operating. If, for example, incision by overland flow is the dominant process methods should be aimed at increasing infiltration. But this will only aggravate the problem if piping is the dominant process (Bocco, 1990). Another example of the complexities of conservation measures is described by Nir & Klein (1974), who suggested that introducing contour ploughing increased gully erosion. Because of the contour ploughing field erosion was decreased and sediment content of water entering the gully system was lower than before, which made the water more erosive. Kalman (1976) performed field experiments on rill erosion in Morocco and found that rills were self-stabilising. He applied a certain discharge and observed progressively lower sediment concentrations that finally approached zero. Erosion after that only occurred when a higher discharge was applied. Kalman did not discuss the causes of this but the implication of his result is that erosion rates in his case will be higher when rills are removed than when they are left alone. This shows that a thorough knowledge of local erosion processes is necessary. Gullies did not show this self-stabilising behaviour. These examples show that erosion-restricting measures should be well thought out.

2.5.6 Gully erosion in loess areas

Loess is an erodible material that is nevertheless able to form vertical walls that can be 10 metres or more in height. This is mainly due to bonds between the silt particles and to tension (Derbyshire, 1989). When loess becomes wet, however, it looses much of its strength because the bonds that exist between the primary silt particles are destroyed. Loess is therefore said to be a collapsible (or metastable) material (e.g. Handy, 1973). Erosion of loess can therefore be a major problem and gullies commonly occur (Leger, 1990).

In the loess area of central Belgium a distinction is commonly made between ephemeral gullies and bank gullies (e.g. Poesen, 1993; Poesen et al., 1998). Ephemeral gullies occur were overland flow concentrates, such as in thalwegs or along linear landscape elements. Bank gullies occur were concentrated flow encounters an earth bank. Bank gullies depend more on local conditions than ephemeral gullies and less on overland flow intensity. The distinction in ephemeral gullies and bank gullies is useful for gently sloping agricultural loess areas (such as in Belgium), but looses its relevance for other loess areas where gullies are much larger.

Large loess gullies, with depths of over 50 metres, have been reported from areas with a continental climate, such as Ukraine (Leger, 1990), China and Nebraska, USA (Pye & Sherwin, 1999). These very large gullies are often more like small valleys, and in fact, there are no clear criteria to distinguish between large gullies and small valleys. Sidewalls are often almost vertical near the top and headcuts can be several tens of metres in height.
Leger (1990) noted the role of desiccation cracks in causing the detachment of blocks of material in such large gullies. Derbyshire et al. (2000) found that the Chinese loess is prone to slab development that is due to tensional stresses in the loess. They also confirmed the findings of Lohnes & Handy (1968) that stability analysis can be used to show that loess slopes would have near vertical slopes at the top (70-85°) and slopes of 51-59° lower down. These gullies can cause severe dissection of previously low relief areas by headcut retreat. A well-known example of this is the retreat into the tablelands of the Chinese Loess Plateau.

Loess is not only susceptible to gully erosion, but also to sheet and rill erosion. Not only is it an erodible material, it is also prone to sealing and crusting. Cerdan et al. (2002), for example, found that sealing was important on loessic soils in Normandy, France. They measured maximum concentrations in runoff of about 100 g/l. These concentrations are much lower than those found elsewhere (e.g. on the Chinese Loess Plateau), but the slope angles in Normandy were gentle (1-10%).

2.6 Erosion modelling

2.6.1 Introduction

Morgan & Quinton (2001) and Doe & Harmon (2001) distinguished two broad model categories: predictive models and research models. Predictive models are used in practical applications, e.g. to assist in making land management decisions, while research models primarily aim to increase process understanding. In practice, this distinction is not so clear-cut, but it clearly shows that different models should be used for different aims. Therefore, the starting point of modelling should be a clear statement of objective (Morgan & Quinton, 2001). In this thesis, the focus will be on process understanding and research models.

Many models have been developed for simulating soil erosion. These models range from simple empirical models to very complicated process based models. Process based models apply as much process-knowledge as is available and use general laws or principles such as conservation of mass (continuity), Newton’s second law of motion (momentum) and the first law of thermodynamics (energy) (Doe & Harmon, 2001). Process based models may, in principle, be used for conditions outside those tested since the laws on which they are based must be obeyed in all circumstances. Empirical models, on the other hand, do not necessarily model the right process and can only be used for the range of conditions for which they were developed. Which type of model is most suited to any particular situation depends on what the purpose of modelling is as well as on time and funds that are available. Complex process based models do not always give better predictions than simple models for several reasons: 1) Not all processes of soil erosion are sufficiently understood, so that the model structure might be flawed and empirical components are still used. 2) The more complex the model the more input data is required. Since input parameters are often difficult to determine accurately, uncertainty regarding the model input will increase with model complexity (Brazier et al., 2000,
Jetten et al., in press). Favis-Mortlock et al. (2001) showed that for more complex models error propagation becomes more important too. 3) Often, there are no unique input parameter sets, so that different input parameter sets can result in equally good predictions. This is called equifinality. On the other hand, process based models are the only type of model that can help us to better understand the processes that are operating. Besides, if physical processes are correctly incorporated in the model this would increase the reliability of predictions regarding the effects of management and landscape change (Brazier et al., 2000). Finally, though process based models might not give better predictions than empirical models they do give additional information, e.g. regarding the spatial and temporal distribution of erosion (Morgan & Quinton, 2001).

Erosion models can also be subdivided in those that simulate erosion for single storms and those that simulate erosion for longer time periods (continuous models). The first type depends on accurate data about model parameters at the start of an event (the initial conditions), while the latter models these parameters in between events. Thus, what happens in between events is not relevant for storm based models, as long as the initial conditions for the next event are correctly specified, but should be modelled by models operating on longer time scales. Processes that should be dealt with in continuous models include plant growth, evapotranspiration, percolation and the seasonal change of soil properties. Therefore, continuous models need much more data than event based models.

A third subdivision of soil erosion models is that in lumped and distributed models. Lumped models use only a few spatial elements for any application, while in distributed models the number of spatial elements can be in the tens of thousands. Obviously, lumped models can only predict the amount of soil loss leaving the study area, while distributed models should be able to give spatial predictions of erosion and deposition, but at the cost of having larger data need and computing time.

Scale issues play a role in soil erosion modelling in several ways. First, it is possible that at different spatial scales different processes are operating, or that they are operating in different ways. For any particular study, a model should therefore be chosen that is relevant to the scale of the problem being studied. Generally, a finer time and space resolution requires more detailed process knowledge, so that modelling at catchment scale is inevitably a compromise between detailed process understanding and reasonable computation times (Kirkby, 1998). At the same time, however, additional processes might appear at larger scales. Gully erosion, for example, does not operate at plot scale, but only on hillslope or catchment scale. Secondly, models usually require the use of effective parameter values that are representative values for the spatial units of the model. In most cases, the values of effective parameters are scale dependent. Another problem is that measurement scale of these parameters is often different from model scale. Most measurements of soil characteristics, for example, are point based (King et al, 1998), and might therefore not be representative for the larger model units. Finally, the way in which the actual landscape is represented in the model can also influence the model results. For grid-based models, for example, the results depend on grid size. This issue will be further discussed in chapter 9.
This section will briefly discuss some of the present day soil erosion models. LISEM will be discussed in section 2.7. Some models that have been developed with the specific purpose of simulating gully erosion will also be discussed.

2.6.2 Soil erosion models

**USLE**
The Universal Soil Loss Equation (Wischmeier & Smith, 1978) is an empirical equation developed for the United States. It is based on a large number of Wischmeier plot measurements and predicts average annual erosion for plots or fields. Deposition is neglected. The basic equation is a simple multiplication of a number of factors. As shown by Haan et al. (1994), the individual factors can, however, be calculated in complex ways, especially in the revised version (RUSLE, Renard et al., 1997). The USLE has later been adapted to other areas as well. Since the plot scale is not the appropriate scale to study gully erosion the model is not capable of simulating gully erosion.

**WEPP**
The Water Erosion Prediction Project (Flanagan et al., 2001) is a process-based erosion model that simulates erosion for hillslope profiles. By combining several profiles with channel segments and impoundments small catchments can also be modelled. The model structure is described by NSERL (1995). WEPP is a continuous simulation model that can also be used for single storms. Since it is a continuous model it needs many parameters that are not needed in event-based models. For example, one needs to take into account soil management and changes of soil properties over the season. The minimum number of input parameters required to run the model is about 100 (Brazier et al., 2000).

Infiltration is simulated with the Green-Ampt Mein-Larson equation. Overland flow is divided into rill and interrill flow. Friction factors are calculated as the sum of partial friction factors, such as surface residue friction, bare soil friction and so on. Rill density must be specified beforehand and it is further assumed that all rills have equal discharge. Sediment transport is calculated with a modified Yalin equation. Water routing is in principle done with the kinematic wave, but to limit computation time approximations are used instead. For erosion calculation the steady-state runoff is used, which in practice means that the peak runoff is calculated and used in the erosion calculations. The runoff duration is adapted to a so-called effective duration to ensure that the total runoff volume remains constant. WEPP uses 5 sediment classes: clay, silt, sand, small aggregates and large aggregates. Transport and deposition are calculated separately for these classes. Gullies cannot be modelled.

**EUROSEM**
EUROSEM (Morgan et al., 1998a,b) is a process based soil erosion model developed to operate on event basis. The model was developed to overcome some of the problems and limitations of the USLE. The main problem of the USLE is that it is a model that gives only mean annual soil loss, while for the European circumstances it is more important to model within single storms (Morgan et al., 1998a,b). Some newer American models, such
as CREAMS (Knisel, 1991) and WEPP can be run for individual storms, but still they
only model total storm soil loss. To model individual events a dynamic model is needed.
EUROSEM therefore uses a 1-minute time step. Catchments are divided in a series of
planes and channels that are supposed to be internally homogeneous. The main problem
with a distributed, storm-based model is that many data about initial conditions are
required.

In EUROSEM, all flow is assumed to be turbulent, which makes the use of the Manning
equation possible. Morgan et al. (1998a,b) discussed the value of Manning's n for
overland flow and channel flow. Theoretically, Manning's n should be higher for
overland flow, but Morgan and co-workers argued that surges in velocity and presence of
sediment could counterbalance this. They therefore use the same Manning's n for both
overland flow and channel flow. Sediment is detached by both rainsplash and flow
erosion and transported by flow. Rill erosion is modelled explicitly, but rill position has
to be specified in advance. The sediment concentration in the flow is modelled as a
balance between erosion and deposition (which both operate continuously). Transport
capacity for rills is modelled with the use of transport equations developed by Govers
(1990) and for interrill flow with those of Everaert (1991). Both equations use stream
power to calculate transport capacity for different median grainsizes.

EUROSEM cannot be considered to be a fully distributed model as the possible amount
of planes and channel segments is limited. Rill position has to be specified in advance.
Gully erosion is at present not included explicitly and can only be modelled if the gullies
can be either seen as large rills or as small channels. If different processes are operating
in gullies these cannot be modelled. In addition, the location of gullies has to be specified
in advance.

2.6.3 Gully erosion models

RILLGROW
Favis-Mortlock et al. (1998) developed the RILLGROW model to simulate the initiation
of rill systems and their subsequent development. The only input is detailed micro
topography data (which are not always available). The micro topography is adapted
during the simulation run according to the calculated erosion. Erosion is calculated with a
stream power based equation. This method appears promising for very small areas, but it
is unlikely that it can be used for catchment scale or even field scale as the data
requirement and the computational demands will be too large for that.

EGEM
The Ephemeral Gully Erosion Model (EGEM), as described by Woodward (1999), was
especially developed to simulate ephemeral gully erosion, in particular in the USA.
EGEM consists of a hydrology and an erosion component. Erosion is driven by peak
discharge and runoff volume, where peak discharge is assumed to occur as long as there
is runoff. Gully depth is assumed to be constant along the length of the gully. It is
assumed that the gully will erode vertically downwards until a less erodible layer is
reached. The maximum allowed depth is 46 cm (18 inches) because deeper gullies are considered to be ‘classical’ gullies instead of ephemeral gullies. When the maximum depth is reached the gully will widen. Nachtergaele et al. (2001) tested the model for mediterranean conditions in southern Portugal and southern Spain. They found that the ephemeral gully volume was predicted well, but this proved to be caused by the fact that gully length is also in input parameter of EGEM. In fact, the relationship between gully length and gully volume had a higher $R^2$ than the EGEM-predicted volumes with measured volumes. EGEM assigns values for a number of soil parameters (like shear stress, grainsize) based on soil type and tillage method. This method might not be accurate for the stony soils in the mediterranean study areas (Nachtergaele et al., 2001). The theory underlying EGEM was also not based on mediterranean conditions and might therefore not apply in this case.

GULTEM

Sidorchuk and Sidorchuk (1998) developed the three-dimensional hydraulic GULTEM model to simulate the first stage of gully development. This first stage is the phase in which there is incision of the gully floor (gullies are assumed to be rectangular) with mass movements after the runoff events (resulting in a trapezoidal gully cross-section). The system develops rapidly during this stage. Flow width and depth are calculated from discharge using empirical equations developed for the Yamal Peninsula in arctic Russia. Particle detachment is calculated using the product of bed shear stress and average velocity. The output of the model is gully depth, width and volume. The main limitation of the model seems to be that the final gully length has to be specified beforehand. Headcut retreat is not modelled. Later Sidorchuk (1999) also presented a model for the second, so called stable, stage of gully development. During this stage, there is negligible erosion and deposition along the gully bed and the gully’s bottom and walls are morphologically stable. Crucial in this model is how to specify the channel-forming discharge, since the stable condition implies that the threshold for transport is exceeded, but the threshold for erosion not reached. These thresholds depend on discharge. Sidorchuk reasoned that each discharge produces channel transformation and that therefore each discharge should be used as well as the return period for the particular discharges.

2.6.4 Discussion

In the previous sections only a few of the large number of available soil erosion models (see e.g. Jetten et al, 1999; Morgan & Quinton, 2001) have been discussed. It is clear that the models presented here do not take into account the effects of very steep slopes or high concentrations.

Gully erosion has received more attention, but up until now no erosion model can model gully erosion accurately. The more recent soil erosion models do, however, perform some sort of rill erosion calculation. It is, however, generally necessary to specify beforehand where the rills or gullies will develop. Even in a model such as the Ephemeral Gully Erosion Model (EGEM), which was especially designed to model ephemeral gullies, one has to specify the estimated depth and final length before simulation (Woodward, 1999). It
seems likely that the trend towards incorporating gully erosion into erosion modelling will continue as field measurements have shown how important gully erosion can be. At the moment none of these models can model for example mass movements on gully walls or headcut retreat (Poesen et al., 1998). Haan et al. (1994), however, report the development of models that take at least one of these processes into account.

Another trend is to use the grid based Digital Elevation Model (DEM) to estimate the amount of water at a certain point and to use a threshold concept to model concentrated flow erosion. From the DEM a slope map and a map of drainage direction can be calculated. Several authors (e.g. Ludwig et al., 1996; Takken et al., 1999 and Van Dijck, 2000) have recognised that tillage direction can change the runoff direction. Ludwig et al. create a drainage direction map by combining slope, tillage direction and linear features such as field boundaries. They assume that water is more likely to follow the slope on steeper slopes and more likely to follow the tillage direction on gentle slopes. Thresholds of upstream area and slope length are subsequently used to predict where concentrated flow erosion will occur.

The way in which to model gully erosion depends on what exactly one wants to model. Do we want a geomorphological model or an erosion model? In the first case we are trying to model the shape of the gully, in the second case only the amount of eroded material. So, geomorphological modelling would require adaptation of the DEM after each time step. In the second case the DEM maybe does not need to change and it might be possible to model the process of gully erosion by just indicating if a gully is present in any particular model unit.

Gully development is also a process that operates on several timescales. We can for example assume that loose material will be produced by mass movements between storms. Initial conditions and throughflow are also better modelled on a daily basis than on a storm basis. Produced material will then be removed during a storm. Incorporating these 'slow' processes into storm-based erosion models does not seem logical, since this would take too much computing time. It would be preferable to link storm-based erosion models to another model that simulates gully processes operating on longer timescales. In the case of the Chinese Loess Plateau this means modelling on a daily basis with a soil water model and modelling with a storm-based erosion model during storms.

2.7 LISEM

Water erosion on the Chinese Loess Plateau almost exclusively occurs during the few heavy storms that occur in a year. This makes the choice for a storm-based model logical. LISEM (Limburg Soil Erosion Model) was used in this study because it is not only a storm-based model, but it is also raster based and can therefore simulate detailed spatial patterns of erosion. Many soil erosion models do not pay much attention to simulating erosion patterns. The principles of LISEM have been described in several papers (De Roo et al. 1994; 1996a,b; Jetten and De Roo, 2001). LISEM is also a process based model, so that any new process knowledge can be incorporated where necessary. A practical
advantage of LISEM is that it has been integrated with a Geographical Information System (GIS). It reads GIS-maps as input and produces GIS-maps as output. As GIS PCRaster is used (Wesseling et al., 1996). PCRaster is a grid-based GIS that has been specifically designed at Utrecht University to perform spatial calculations. Both LISEM and PCRaster are written in the C++ programming language.

### 2.7.1 Model structure

Figure 2.1 shows a simplified flow chart of the LISEM model. LISEM can be divided into two parts: a water part and an erosion part. Rainfall is the basic input of the water part. Interception is subtracted from the rainfall. The remaining rainfall reaches the soil surface, where it can infiltrate or form a surface storage. Since LISEM is a storm-based model the infiltrated water is essentially a loss of water in the sense that infiltrated water cannot resurface. Infiltration can be simulated using one of several available equations. Partly empirical equations such as the Green&Ampt and Holtan equations can be chosen as well as the physically based Richards equation (using the SWATRE sub model, Belmans et al., 1983). Surface storage will result in surface runoff once a certain threshold is exceeded. Flow velocity is calculated with the Manning equation and surface runoff is routed over the landscape with the kinematic wave equation. The effects of steep slopes on flow velocity are not specifically considered. The user can specify separate wheeltrack and channel networks. Overland flow and flow from wheeltracks can flow into the channel and is then routed to the catchment outlet as channel flow.

![Figure 2.1 Simplified flow chart of the LISEM soil erosion model](image-url)
Overland flow and channel flow are both routed with the kinematic wave, which is solved by a 4-point finite difference solution. An implicit method is used, so that the solution is usually stable and the choice of time step length and grid size is more important for accuracy than for stability. To solve the equation a linear scheme is used to obtain a first estimate of discharge, while iteration proceeds thereafter using a non-linear scheme and the Newton method (Ven Te Chow et al., 1988). The present version of LISEM (version 1.63) does not perform a kinematic wave for sediment. Instead, the sediment is redistributed according to the redistribution of water as determined with the kinematic wave. This procedure seems to introduce some error and a mass balance correction is used for the sediment to overcome this problem. In this correction the mass balance error of sediment is redistributed over the network according to the sediment output of the different pixels. Future versions of LISEM might include a separate kinematic wave for sediment in which discharge is routed first and the results from the discharge routing are used to route the sediment.

LISEM simulates erosion by rainfall and erosion by overland flow and channel flow. Rain splash erosion is calculated as a function of rainfall kinetic energy and depth of the surface water layer. Sediment transport only occurs by overland flow and channel flow. For both overland flow and channel flow, LISEM uses the transport equation developed by Govers (1990) for slopes of up to 12 degrees. The equation is based on a stream power approach:

\[ TC = c(SV - SV_{cr})^d \cdot \rho_s \]  \hspace{1cm} (2.18)

Where:
- \( TC \) = transport capacity (g/l)
- \( S \) = slope (m/m)
- \( V \) = mean velocity (cm/s)
- \( SV_{cr} \) = critical unit stream power (cm/s)
- \( \rho_s \) = density of solids (kg/m³)
- \( c, d \) are coefficients

According to Govers the critical unit stream power is 0.4 cm/s. The coefficients \( c \) and \( d \) depend on grainsize and can be calculated with equations 7.17 and 7.18. Figure 2.2 shows \( c \) and \( d \) as function of median grainsize (D50), while figure 2.3 shows transport capacities calculated with equation 2.17. Figure 2.3 shows that transport capacities increase with decreasing grainsize. Net flow detachment and deposition are calculated with an equation derived from the EUROSEM model (Morgan et al, 1998a,b), but reformulated for use with pixels:

\[ D_f = y \cdot (TC - C) \cdot \omega \cdot w \cdot DX \cdot dt \]  \hspace{1cm} (2.19)

Where:
- \( D_f \) = net detachment/deposition flow (kg)
- \( y \) = efficiency coefficient
- \( TC \) = transport capacity (g/l)
- \( C \) = concentration (g/l)
\( \omega \) = settling velocity (m/s)

\( w \) = flow width (m)

\( DX \) = pixel length (m)

\( dt \) = time step length (s)

The efficiency coefficient \( \gamma \) is 1 for deposition and smaller than 1 for detachment. In case of detachment the value of \( \gamma \) depends on cohesion. After detachment and deposition are calculated with equation 2.19 a check is performed to ensure that no physically impossible situations occur. This is necessary because using equation 2.19 detachment and deposition might become larger than \( TC-C \). Therefore, detachment is limited to that amount of detachment that would just fill all available transport capacity, while deposition is restricted to the total amount of sediment present in the flow.

![Graph showing c and d coefficients as a function of median grainsize (D50)](image)

Figure 2.2. c and d coefficients as a function of median grainsize (D50)

Like the other erosion models LISEM does not take the effect of high concentrations into account. Fluid density is assumed to be constant, and predicted water discharge is a clear water discharge. Neither is gully erosion explicitly modelled by LISEM.

### 2.7.2 LISEM input and output

As a process based model LISEM requires a significant amount of input. The input differs from one infiltration equation to the other. Table 2.1 shows the input that is required by LISEM if the Richards equation is used for infiltration, and when the wheeltrack option is not used. All names with extension .map are PCRaster maps. Provided that field data are available almost all maps can be derived from the 3 basic maps: DEM, land use map and soil map. Table 2.1 also shows how these data can be obtained.
As can be seen from table 2.1 there is a number of parameters that should be measured in the field, either once or repeatedly. How these parameters were determined will be discussed in chapter 4. Table 2.1 only shows which data are needed to actually run the model. To test the model result additional field data are needed. Those data are at least runoff and sediment concentration at the outlet of the catchment, and preferably also data about the actual distribution of erosion in the catchment. Those data will also be discussed in chapter 4.

The standard output of LISEM consists of several ASCII files that can be used to prepare hydrographs and sedigraphs for a maximum of 3 different points in a catchment as well as several maps. Apart from an erosion map (tonnes/ha) and a deposition map (tonnes/ha) a number of runoff maps (l/s) is created for user-specified times. One of the ASCII-files gives a summary for the outlet of the catchment.

In this thesis Windows version 1.63 of LISEM is used. This version will be referred to as LISEM 163. In the course of the thesis 2 adaptations of LISEM 163 were developed. For sake of clarity these versions will be called LISEM LP (Loess Plateau) and LISEM TC (Transport Capacity).

2.7.3 LISEM calibration and validation

LISEM was developed for the province of Limburg, The Netherlands, and has been previously calibrated and validated for several catchments in the Loess region of northwestern Europe. De Roo et al. (1996b) performed a sensitivity analyses and found that the most sensitive variable in the prediction of runoff is saturated conductivity. For
Table 2.1 Input data for LISEM version 1.63, with the use of the SWATRE infiltration sub model but without the use of a wheeltrack network.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Name b</th>
<th>Method a</th>
<th>Unit</th>
</tr>
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<td></td>
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<td>Drainage direction</td>
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<td>derive from DEM</td>
<td>-</td>
</tr>
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<td>derive from DEM</td>
<td>tangent</td>
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<td>Catchment outlet</td>
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<td>-</td>
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<td>mm/h</td>
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<td>-</td>
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<td>measure once</td>
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</tr>
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<td>Splash delivery ratio</td>
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<td>Soil profile map</td>
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<td>profgras.map</td>
<td>profile description</td>
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</table>

*a* Once and repeatedly refer to a frequency in time, not space. Measurements should be performed in a sufficient number of places to be able to create a map.

*b* When no name is specified the value of the parameter applies to the entire catchment and should be specified in the LISEM interface.
the prediction of soil erosion LISEM was most sensitive to changes in Manning’s n and transport capacity. De Roo et al. (1996b) found that LISEM gave reasonable results for 60% of the storms that were modelled. They attributed the discrepancy for the other 40% to spatial and temporal variability in saturated conductivity and initial moisture content and to differences between summer storms and winter storms. De Roo & Jetten (1999) identified saturated conductivity, suction at the wetting front and initial moisture content as sensitive parameters. Takken et al. (1999) showed that LISEM predicted catchment runoff reasonably well, but failed to reproduce observed erosion and deposition patterns inside a catchment.

2.7.4 LISEM and the Loess Plateau

Some of the main algorithms of LISEM are only valid for a given range of circumstances. Since the circumstances on the Loess Plateau are very different from the circumstances for which LISEM was developed it is possible that application limits will be encountered when applying LISEM to the Loess Plateau. Gullies play a much more dominant role than in northwestern Europe and they may function both as sinks and source of sediment. Likewise, the combination of steep slopes with small-scale land use, might provide an environment that LISEM is not able to simulate. Some important characteristics of LISEM that might affect its application on the Loess Plateau will be discussed briefly here.

The first characteristic is the way in which transport capacity is calculated. LISEM calculates flow velocity with Manning’s equation and subsequently calculates stream power and transport capacity. Since stream power is the product of the velocity and the (energy) slope, the slope angle influences the stream power considerably. According to Govers (1990) the transport capacity equation derived from stream power is valid for slope angles up to 20%, and this will pose one of the largest potential problems in the application of LISEM to this area. The erosion and deposition are modelled as transport deficits and surpluses and are therefore also strongly determined by the flow conditions.

Another important characteristic of LISEM is that it is a grid-based model. Flow circumstances are determined locally in each grid cell and inertia of water is not taken into account. Thus while in reality sediment remains in suspension when the slope angle changes abruptly, LISEM may simulate sudden deposition because it assumes that the transport capacity decreases radically. It is therefore possible that the presence of gullies with extreme changes in slope pose problems for the kinematic wave that is used in LISEM to route the water and sediment. Such problems are likely to be more pronounced for steeper areas. On the other hand, models that do not use a grid based approach cannot model spatial distribution of erosion at all.

A final feature of LISEM that may be important is that concentrated flow in ditches and ephemeral streambeds can be simulated as a separate process, by defining these features as ‘channels’. The channels are assumed impermeable and the channel dimensions and characteristics influence the hydrograph shape considerably.
This study must be seen as a performance test of a process based distributed model under extreme circumstances.

2.8 Conclusions

The specific characteristics of the Chinese Loess Plateau that have been identified have so far not been studied in the context of soil erosion modelling.

Both the velocity equations and the sediment transport equations that are commonly used in erosion models have not been developed for such steep slopes as those of the Chinese Loess Plateau. The effect of slope angle on flow velocity and sediment transport has not been evaluated for such models either. Likewise, the effect of sediment concentration of flow properties has not received any particular attention in erosion modelling.

Gully erosion is a topic that has received more attention in erosion modelling. Some of the present day erosion models do simulate some sort of gully erosion, but even models that have been specifically developed to model gully erosion require that some properties of the gullies be set in advance.

The LISEM model is an up to date erosion model that is in principle suitable for simulating erosion on the Loess Plateau for several reasons. LISEM is storm-based, so that it should be able to handle the storm-dominated water erosion of the Plateau. LISEM is also a distributed model, so that spatial predictions of erosion inside a catchment are possible. Finally, LISEM is a process based model. This means that process descriptions in the model can be adapted if the specific characteristics of the Loess Plateau require this. At present LISEM does not specifically take into account the effects of steep slopes, high sediment concentrations and the presence of gullies.