Chapter 1

THE PROBLEM IN CONTEXT

1.1 Introduction

Soil is an important resource for the survival of the human race and is a central component of environmental systems, together with water, air and radiation from the sun. Undoubtedly, soil is one of the essentials for life on Earth. Scientists have investigated various soil properties to ensure good yields of crops, fibre and fuel (Cresser *et al.*, 1993). However, soils do not necessarily always provide ideal conditions for plant growth. Many soil processes can constrain plant growth: soil hydrology is fundamental for most of these (Hudson, 1971; Evans, 1980; Kirkby, 1980; Morgan, 1995). Among the most serious is soil erosion by water.

Globally, soil erosion by water is a serious present-day environmental problem and its consequence is subject to extensive investigations (Kirkby, 1980; Morgan, 1995). Previously published simulation studies of the effects of future climate change upon erosion indicate that, under land usages that leave the soil unprotected, even minor increases in rainfall amounts are likely to result in disproportionately large increases in erosion (Kirkby, 1980; Favis-Mortlock and Boardman, 1995).

Soil erosion rates may be expected to change in response to changes in climate for a variety of reasons (Pruski and Nearing, 2002*a*), the most direct of which is the change in the erosive power of rainfall (Favis-Mortlock and Savabi, 1996; Williams *et al.*, 1996; Favis-Mortlock and Guerra, 1999; Nearing, 2001; Pruski and Nearing, 2002*a*). Existing studies however almost invariably make the simplifying assumption that distributions of future rainfall intensities remain unchanged from the present (Favis-Mortlock, 1995; Favis-Mortlock and Boardman, 1995; Favis-Mortlock and Guerra, 1999; Pruski and Nearing, 2002*a*,*b*; O'Neal *et al.*, 2005). This is unlikely to be the case. Intensities may change and/or the frequency of occurrence of high-intensity events may change (Karl *et al.*, 1995; Houghton *et al.*, 1996; Watson *et al.*, 1998; Karl and Knight, 1998; Osborn and Hulme, 1998, 2002). Any increases in the occurrence of high-intensity rainfall—even without any associated increases in rainfall amounts—may well increase runoff, and hence erosion rates (Kirkby, 1980; Morgan, 1995; Parsons and Gadian, 2000). Future climate change will certainly affect rainfall intensities but our ability to forecast future intensities is limited by the shortcomings of General Circulation Models (GCMs) (Favis-Mortlock and Boardman, 1995).

Few studies have attempted to quantify changes in future rainfall intensity (Karl *et al.*, 1995; Houghton *et al.*, 1996; Watson *et al.*, 1998; Karl and Knight, 1998; Osborn and Hulme, 1998, 2002). Results from these studies suggest that more (or similar) rainfall than at present will occur on fewer raindays—implying an increase in the frequency of heavy rainfall. If these predictions are correct, the implication for future erosion rates are clear. For these reasons, there is a urgent need for greater understanding of future rainfall intensity changes in order to improve the ability of soil erosion prediction.

This thesis, therefore, aims to address some of the issues involved in predicting future soil erosion as a response to future changes of rainfall intensity. More detailed aims and issues associated with achieving those aims are presented at the end of Chapter 2. Prior to Chapter 2, topics that are related to soil erosion processes, rainfall intensity, climate changes and erosion prediction models are discussed in the following sections.

1.2 Soil Erosion Processes

1.2.1 Introduction

> "Erosion by water is the redistribution and removal of the upper layers of the soil, both by the action of falling rain, and by water flowing over the soil during and after rain or following snowmelt." (Favis-Mortlock, 2002)

The erosion of soil by water and wind is a naturally occurring process, which is commonly accelerated by human activity. However, when soil erosion occurs at a greater rate than the rate of soil formation, soil erosion is considered as an environmental problem. Soil erosion is a ubiquitous problem that threatens an important and non-renewable resource such as the agricultural land that is suitable for cultivation (on-site impact) (Boardman, 2003). In addition to removing a valuable resource, soil erosion leads to increased sediment input to nearby watercourses, resulting in, for example, the silting-up of dams and contamination of drinking water (off-site impact) (Mejia-Navarro *et al.*, 1994; Kitchen *et al.*, 1998).

Soil erosion problems can be viewed in three different ways (Kirkby and Morgan, 1980). Firstly, in the broadest view, soil erosion can be compared with other processes of landscape denudation. When and where it is the most rapid process, soil erosion should be recognised as the dominant problem. This view leads to the question of what erosion rates can be tolerated in the long-term. Secondly, a narrower overview examines soil erosion with its immediate climatic and vegetational controls. This then leads to the question as to how well the processes involved in raindrop impact, flow generation, and sediment resistance are understood. Thirdly, soil erosion can be considered in relation to its broad patterns in time and space. Yet, the reasons for the temporal and spatial distributions of soil erosion are only partially understood (Quine and Zhang, 2002; Gómez *et al.*, 2005; Wakiyama *et al.*, 2010).

Soil erosion by water is most active where rainfall cannot infiltrate the soil, but flows over the surface. As the flow travels down a hillslope, it is able to carry soil materials away mainly by shear stress although other sub-processes also contribute (Kinnell, 2000, 2005*a*). In some cases, only an hour or two of contact time with the surface soil is needed to carry away an appreciable amount of material. Thus, where overland flow is dominant, soil erosion by water is likely to be the main process of landscape denudation. When a large depth of water flows rapidly over the surface with correspondingly large hydraulic forces, soil erosion acts catastrophically. These conditions are most commonly found in semi-arid areas, but fields cleared for agricultural purposes are also subjected to erosion in almost any climate, which can on occasion be severe (Boardman, 2001, 2003).

Semi-arid areas are very sensitive to small natural changes in climate and in such areas it is difficult to separate natural from man-induced changes in erosion rates. However, even in temperate-humid areas increased erosion resulting from farming can be sensitively dependent on the extent of the change in vegetation cover, the total rainfall at periods of low cover, and the intensity of the rains.

Therefore, two distinct types of area appear to be at great risk of soil erosion. The first are semi-arid areas, and the second are locations in temperate areas that have been stripped of vegetation for crop cultivation. A soil erosion rate can reach at its maximum where intense rainfall occurs during the period of lowest vegetation cover. This is normally the case in semi-arid climates or in temperate areas which have been left bare at the time of the heaviest annual rainfall. In such cases, when the rainfall increases, soil loss increases, so that the erosional peak tends to be synchronised with the rainfall peak and this relationship becomes more distinctive when the soil becomes more unprotected.

When soil erosion problem is to considered, it also is worthwhile to take long-term effects of soil erosion into consideration. For example, when soil erosion occurs at a rate of one millimetre per year, it might not have apparent effect in a human lifetime. However, over a longer-term, the effect can be considerable. To put this into perspective, topsoil of 15 cm thickness in general would be completely removed after 150 years if erosion rates stay as high as 1 mm/year in average with no additional soil formation in the area. Topsoil contains a high proportion of soil organic matter and the finer mineral fractions, which provide water and nutrient supplies for plant growth. This may look as an oversimplification of the erosion process and soil formation, but it gives us an idea of the long-term effect of soil erosion.

1.2.2 Rainfall

1.2.2.1 Raindrop Splash

The process of erosion by water is a two-phase process: detachment and transport (Morgan, 1995). Individual soil particles are detached from the soil mass by the impact of raindrops. The erosive power of raindrops weakens and loosens the soil surface, and flowing water transports the soil particles (Kinnell, 2000). When sufficient transporting energy is no longer available, a third phase, deposition, can occur.

Raindrop splash distributes soil particles radially away from the site of detachment. The raindrop detachment-splash transport (RD-ST in Figure 1.1) process is effective where rainfall intensities are high, for example, as a result of convective rainstorms. However, splash transport (ST) is a generally inefficient transport mechanism. If the soil has virtually no slope, soil particles splashed away from the point of impact are

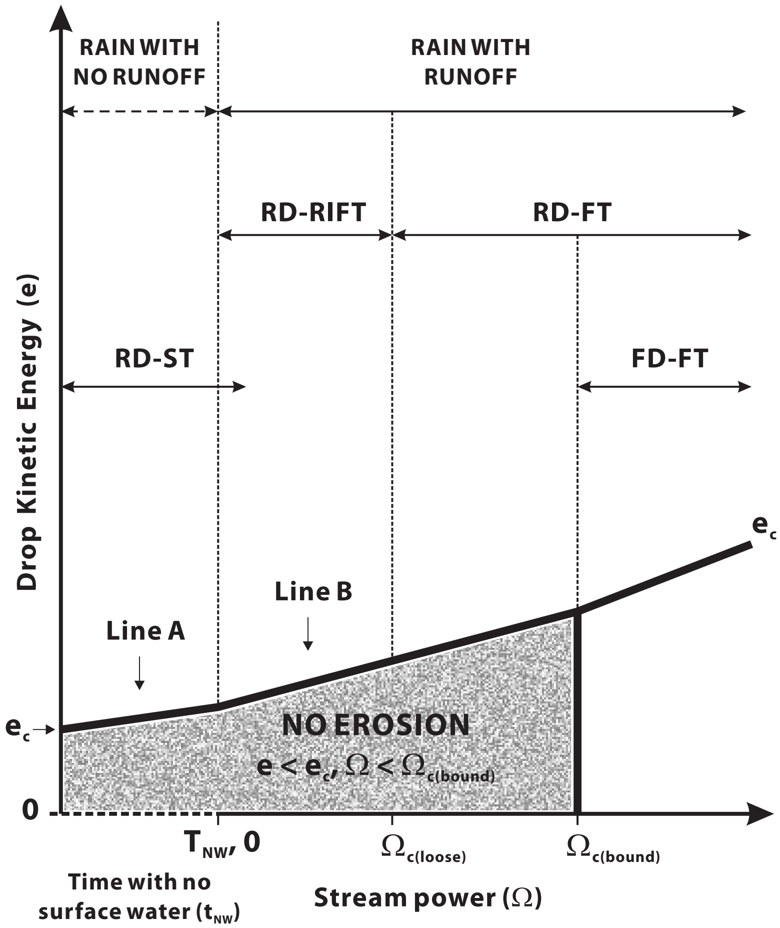

Figure 1.1 Detachment and transport processes associated with variations in raindrop and flow energies. T_{NW}: total time when rain falls and there is no surface water. e_c: critical raindrop energy to cause detachment; raindrop-induced erosion occurs when drop energy is equal or greater than e_c. Line A: e_c when raindrops are detaching soil particles from the soil surface prior to flow developing. The slope on this line is used to indicate increasing resistance to detachment caused by, for example, crust development. Line B: e_c when raindrops are detaching soil particles from the soil surface when flow has developed. The slope on this line is used to indicate increasing utilization of raindrop energy in penetrating the flow when flow depth increases as flow power increases. $\Omega_{c(\text{loose})}$: critical stream power required to transport loose (pre-detached) soil particles. $\Omega_{c(\text{bound})}$: critical stream power required to detach particles bound within the soil surface (held by cohesion and interparticle friction). RD-ST: raindrop detachment and splash transport. RD-RIFT: raindrop detachment and raindrop-induced flow transport. RD-FT: raindrop detachment and flow transport. FD-FT: flow detachment and flow transport(From Kinnell, 2005*a*).

replaced by soil particles detached by other raindrops in the surrounding area (Kinnell, 2000; Zartl *et al.*, 2001). Even if the soil surface has a slope, net downslope transport by raindrop splash alone is generally small (Kinnell, 2001).

When water flows start to build up, the soil surface becomes protected from direct raindrop impact, and another transport mechanism begins to active. Raindrops with sufficient kinetic energy to penetrate through the flow may detach and lift soil particles into the flow, which then carries them downstream until it looses sufficient transporting energy to carry the particles. Soil particles transported by the flow then fall back to the soil surface of lower grounds. This transport process is termed Raindrop-Induced Flow Transport (RIFT in Figure 1.1) (Kinnell, 1990). While RIFT is more efficient than ST, it still requires numerous raindrop impacts to move soil particles downstream.

As rain continues, thin surface water flows become capable of moving loose soil material on the top of the surface, but might not be capable of detaching soil material from the soil mass. In many cases, soil particles are detached by the help of raindrop impacts, and carried away downstream without the need for raindrops to be involved in the transport process. This raindrop detachment-flow transport (RD-FT) process is more efficient than RD-RIFT. In a typical field, both RD-RIFT and RD-FT occur simultaneously in the same flows.

When the critical stream power ($\Omega_{c(\mathrm{bound})}$) for flow to detach soil particles from soil mass exceeds stream power (Ω), flow detachment (FD) occurs (Kinnell, 2000). Once soil materials are detached and transported by flow (FD-FT), erosional channels are generated (Figure 1.1). As these channels develop and increase in size to become large rills and possibly even gullies, processes such as gravitational collapse of channel walls and heads become important (Boardman *et al.*, 2003).

1.2.2.2 Rainfall Intensity

The erosive power of rainfall has long been appreciated by studies on soil erosion (Musgrave, 1947; Wischmeier and Smith, 1958). Nevertheless, obtaining information on rainfall intensity for soil erosion is very much problematic. One way of measuring rainfall intensity would be measuring size, distribution and velocity of raindrops, so that the kinetic energy of the rainfall can be calculated (Cerda, 1997; Lascelles *et al.*, 2000). This can be seen as a 'bottom-up' approach. Another method would be simply to measure rainfall amount and duration, so that intensity can be obtained by dividing rainfall amount by duration (i.e. rainfall amount per unit time) (Osborn and Hulme, 1998). This is a 'top-down' approach. However, both approaches have their own shortcomings (Parsons and Gadian, 2000; Schuur *et al.*, 2001; Garcia-Bartual and Schneider, 2001).

Although rainfall intensity plays a very important role for soil erosion, it is important to recognise that the vital variable for soil erosion by splash is not rainfall intensity itself, but rainfall energy. This rainfall energy varies in association with rainfall intensity. As raindrops increase in size, their terminal velocity increases. This increases the kinetic energy of raindrops. The total kinetic energy of rainfall also increases with increasing number of raindrops during a given time. The total kinetic energy of rainfall may be estimated from the distribution of raindrop size and number of raindrops during a storm. The accuracy of this estimation is, however, limited by natural variations in rainfall characteristics (van Dijk *et al.*, 2002). Yet, in natural rainfall events, the relationship between rainfall intensity and energy is neither so clear, nor simple. Despite this, simple assumptions about the rainfall intensity-energy relationship are often made in studies on soil erosion, in particular, modelling studies, as rainfall intensity is the only easily modifiable control on rainfall energy in such studies (Laflen *et al.*, 1997; Morgan *et al.*, 1998).

Parsons and Stone (2006) ran a laboratory-based rainfall simulation experiment to determine the implications of temporal variation of rainfall intensity for rates of soil loss. He found that erosion is least for the constant-intensity storms. This is highly significant because soil-erosion models are typically calibrated using data obtained from constant-intensity experiments. Moreover, storm pattern does not appear to affect the volume of runoff, but it does affect the quantity of eroded sediment. In particular, the constant-intensity storm patterns are associated with low erosion rates. Storm pattern also affects the size-distribution of the eroded sediment. Parsons and Stone (2006) therefore concludes that the relationship between rainfall energy and interrill erosion is more complex than is currently assumed in process-based models of soil erosion.

Other studies also note that there are complex interactions between raindrop size, velocity and the duration of rain, which control the erosive power of rainfall (Kinnell, 1981; Brandt, 1990; Salles *et al.*, 1999; van Dijk *et al.*, 2002).

1.2.3 Soil Type

Soil erodibility is an estimate of the resistance of the soil to erosion, based on the physical characteristics of each soil (Morgan, 1995). Although erodibility varies with soil texture, aggregate stability, shear strength, infiltration capacity and organic and chemical contents, soils with high infiltration rates, higher levels of organic matter and improved soil structure have a greater resistance to erosion, in general (Morgan, 1995).

Erodible soils have restricted clay content (Bryan, 2000). Soils with more than 30–35% clay are generally coherent and form stable soil aggregates, which are resistant to raindrop impact and splash erosion (Evans, 1980). Clays often have rough surfaces to store much water, and are resistant to sheet and rill erosion. Sands and coarse loamy sands, on the other hand, have high infiltration rates and resistant to erosion, and even if this is exceeded, sands (more than 0.3 mm diameter) are not easily eroded by flowing water or by raindrop impact (Evans, 1980; Marshall *et al.*, 1996).

Sandy soils are however more erodible than clayey soils because the aggregates of these sandy soils slake more readily and seal the soil surface (LeBissonnais, 1996). Loamy soils are also particularly at risk of sealing (Ramos *et al.*, 2000).

After cultivation, the soil surface becomes rough. The amount of water, which can be stored on the surface before runoff takes place, is thus large at this time. Surface roughness is least after drilling and rolling of the seedbed, and differences between soil types are smallest (Robinson and Naghizadeh, 1992). For similar soil types, the timing of cultivation can affect the storage volume, for example, a clay surface prepared in winter can have more than twice as much storage volume as a surface prepared in spring (Evans, 1980).

Moreover, stony soils are generally less vulnerable to erosion as the surface stones not only protect the soil, but also increase infiltration by providing larger pores between stones (Agassi and Levy, 1991; Poesen and Lavee, 1994; de Figueiredo and Poesen, 1998). However, when rock fragments are well-embedded in a surface seal, a positive relation for runoff and sediment yield is found (Poesen and Ingelmo-Sanchez, 1992). A negative relation occurs either where rock fragments are partly embedded in a top layer with structural porosity or where the rock fragments rest on the surface of a soil having either textural or structural pore spaces (Poesen and Ingelmo-Sanchez, 1992).

1.2.4 Topography

There are two aspects of topography that affect erosion: slope angle and length. Normally, erosion would be expected to increase as the slope steepness increases (Liu *et al.*, 1994). Soil erosion by water also increases as the slope length increases because of increases in velocity and volume of runoff (Liu *et al.*, 2000). Water depth increases with downslope distance so that interrill soil erosion is affected by slope length (Gilley *et al.*, 1985*b*). Water depth then affects soil detachment and overland flow sediment transport capacity (Gilley *et al.*, 1985*a*). Slope angle is also closely related to the effectiveness of splash erosion (Kinnell, 2000; van Dijk *et al.*, 2003).

The location of downslope is an important factor that determines the development of rills on a hillslope. However, there is another factor that is closely related to the dynamics of initiation and growth of rills. The minute variations of soil surface topography, also known as microtopography, can play an important roll on this "rill competition".

Microtopography is not temporally static because erosional processes will continuously modify the surface of soil during a rainfall event. As a result, runoff during the latter part of the event will flow over a soil surface that has been modified and different from the surface earlier in the rainfall. Thus, erosive modification of microtopography constitutes a feedback loop which might be expected to operate in a positive sense. The most 'successful' rills (i.e. those conveying the most runoff) will modify the local microtopography to the greatest extent, and so will most effectively increase their chances of capturing and conveying subsequent runoff.

Favis-Mortlock *et al.* (2000) previously recognised the importance of microtopography in the initiation and the development of rills, and developed a erosion model, RillGrow, using a self-organising dynamic systems approach. More about RillGrow is included in Section 1.5.5.

1.2.5 Land Use

Soil erosion potential is highest where the soil has no or very little vegetative cover. Vegetation cover protects the soil from direct raindrop impact and splash, and tends to slow down surface runoff. On a field with complete vegetation cover, runoff and erosion are comparatively small, often less than 5% of runoff and 1% of erosion from bare soil, respectively (Braskerud, 2001; Rey, 2003). One reason is because the infiltration rates of the vegetated field are relatively higher than those on bare soils as the field often has a better soil structure and more stable aggregates (Robinson and Phillips, 2001). When runoff does take place, the leaves and roots of plants inhibit the flow by reducing the velocity of the flow (Braskerud, 2001; Rey, 2003). On soils with less than 70% vegetation cover, runoff and erosion increase rapidly when rainfall occurs (Favis-Mortlock and Savabi, 1996). Under less than 20–30% vegetation cover, runoff and erosion are related to the amount of bare ground, increasing as the proportion of bare ground increases (Favis-Mortlock and Savabi, 1996).

The effectiveness of any crop management system against soil erosion by water also depends on how much protection is available at various periods during the year, relative to the rainfall amount that falls during these periods. In this respect, crops which cover for a major portion of the year (e.g., alfalfa or winter cover crops) can reduce erosion much more than can crops (e.g., row crops) which leave the soil bare for a longer period of time and particularly during periods of intense rainfall (Zhang *et al.*, 1995*a*,*b*).

1.3 Soil Erosion and Rainfall Intensity

This section provides examples of some notable erosion events which are documented in selected publications.

South Downs, East Sussex, UK, October 1987 (Boardman, 1988) Heavy rainfall on 7 October 1987 and subsequent storms resulted in soil losses over 50 m^3/ha on several fields and over 200 m^3/ha on one field in the eastern South Downs (Boardman, 1988). Monthly rainfall totals at Southover, Lewes, were 54.3 mm for September and 270.9 mm for October 1987. Rainfall recorded at Southover, Lewes, on 7 October 1987 was 50.2 mm with a maximum short period intensity of 6.7 mm/h for 5.5 hours including 40 mm/h for 15 minutes.

Substantial rills or gullies were formed by the rainfall event on 7 October 1987. As a result of this, following rainfalls as low as 7 mm caused runoff and erosion (Boardman, 1988). Although there are no event-by-event records available for soil losses, it is evident that the rainfall on 7 October 1987 played an important role, by contributing to rill or gully generation, on soil erosion in the area. However, the main factors responsible for the severe erosion were land use and farming practices.

Vicrello, Tuscany, Italy, May 1994 (Torri *et al.*, 1999) A rainfall depth of 77.8 mm fell on a field plot with a bare soil in Vicrello, Tuscany, Italy (Torri *et al.*, 1999). The storm lasted for over 28 hours and caused a soil loss of 126.2 t/ha. Maximum intensity averaged over 10 minute was 120 mm/h.

Hadspen, Somerset, UK, May 1998 (Clark, 2000) Total rainfall amount of 47.6 mm fell in Hadspen, Somerset, UK on 13 May 1998 (Clark, 2000). Most of rain fell between 2115 GMT to 2130 GMT reaching rainfall intensity of >100 mm/h. In Nettlecombe Hill and Higher Hadspen, ploughed fields on slopes with 2–11° eroded at the rates of 1.412 tonnes/m^3 and 1.312 tonnes/m^3, respectively. Total soil loss from two area was 72.1 tonnes.

Ashow, Warwickshire, UK, August 1999 (Harrison and Foster, 1999) On 20 August 1996 in Ashow, Warwickshire, the storm commenced at 1930 BST. Rainfall intensity was low until 2030 BST when 24.5 mm of rain fell in 30 min and a total of 33.5 mm fell before midnight.

One of two fields in the catchment was planted with oilseed rape eight days before the storm. The field was ploughed and power-harrowed, and then seed drilled with a low ground pressure buggy. It was subsequently rolled by a tractor with low ground pressure tyres. The other field was harvested of wheat and barley, and then rough ploughed, the soil clods being broken up using rotating discs.

Extensive erosion of top soil occurred, followed by the development of gullies and rills by overland flow during the storm. Approximately 790 t of sediment was eroded from the two fields excluding the sediment that reached nearby river (River Avon, UK). Average sediment yields was 49.7 t/ha which is equivalent to the average ground lowering of 3.8 mm.

Northern Ethiopia Highlands, 1998-2000 (Nyssen *et al.*, 2005) Rainfall intensity in Northern Ethiopia Highlands was monitored using a tipping bucket rain gauge during 1998-2000 (Nyssen *et al.*, 2005). Overall rain intensity in the area is low. 88% of total rain volume falls with an intensity <30 mm/h. Most storms have a low intensity with a brief high intensity part. This high intensity can be observed at the beginning, in the middle or at the end of the storm. Although area-averaged intensity was low in this area, it was found that maximum rain intensity at individual locations exceeded by far the threshold values for excessive rain (see Table 5, Nyssen *et al.*, 2005). Rainfall intensities beyond these thresholds were known to cause >50% of total soil losses (Krauer, 1988). Large rain erosivity in the area is due to larger median volume drop diameters (D_{50}) than those reported for other regions of the world, rather than due to high intensity.

South Downs, East Sussex, UK, October 2000 (Boardman, 2001) Exceptional rainfall in October and November 2000, especially a 24-hour fall of about 100 mm, led to extensive erosion and property damage (Boardman, 2001). The rainfall was typical of frontal, low-intensity events that usually occur in British winters but it lasted for a longer period than usual period. In a 24-hour period prior to 09:00 on 12 October (i.e. 11 October rainday), a total rainfall of 89.9 mm was recorded (Boardman, 2001). In a 10-hour period of continuous rainfall (23:00–09:00) 63.8 mm fell with a maximum intensity of 11.4 mm/h and a maximum short-period intensity of 3.6 mm/min (i.e. 216 mm/hr) (Boardman, 2001).

Rainfall of 100 mm in 24 hours has a return period of well over 100 years and a intensity of 11 mm/h is to be expected every year (Boardman, 2001). This means that the rainfall on 11 October 2000 has a rainfall intensity that is commonly observed in the area, but the total amount and duration are very unlikely in the area. It is noted that high intensity rainfall within prolonged low intensity rainfall at the time of year when the agricultural land is most vulnerable may result in extensive erosion events.

1.4 Rainfall Intensity and Climate Change

Many studies using GCMs predict an increase in global average precipitation in response to global warming induced by greenhouse gases (Houghton *et al.*, 1996; Jones and Reid, 2001; IPCC, 2001*b*,*a*, 2007*a*,*b*). This increase in global average precipitation has been based on the assumption that an increasing global-mean temperature will intensify the hydrological cycle (Nearing *et al.*, 2005). The IPCC reported that there has been a very likely increase in precipitation during the 20th century in the mid-to-high latitudes of the Northern Hemisphere (IPCC, 2001*b*, 2007*b*). Climate models are also predicting a continued increase in intense precipitation events during the 21st century (IPCC, 2001*a*, 2007*a*).

In addition, there has been a number of investigations using observed data that provided some evidences for a significant increase in extreme precipitation (Karl *et al.*, 1995; Karl and Knight, 1998; Osborn *et al.*, 2000; Osborn and Hulme, 2002). Karl *et al.* (1995) and Karl and Knight (1998) observed increases in extreme precipitation (greater than 50 mm per day) in the United States using historical data over the period 1910–1996.

Osborn *et al.* (2000) and Osborn and Hulme (2002) also observed an increasing trend in intense daily precipitation over the period 1961–2000 in the United Kingdom. They found that, on average, precipitations were becoming more intense in winter and less intense in summer.

The findings by Osborn *et al.* (2000) and Osborn and Hulme (2002) are generally consistent with the results from the GCM simulations (Jones *et al.*, 1997; Jones and Reid, 2001). However, IPCC (2001*a*,*b*) indicated that potential changes in intense rainfall frequency are difficult to infer from global climate models, largely because of coarse spatial resolution. The ability of GCM integrations and operational analyses to simulate realistic precipitation patterns, spatially and seasonally, is also generally not as good as the ability to predict temperature (McGuffie *et al.*, 1999). The likelihood of finding real trends in the frequency of extreme events becomes lower the more extreme the event (Frei and Schär, 2001). The same authors demonstrate this by applying known trends in the scale parameter to synthetic data series, and then attempt to identify statistically significant trends in the frequency of various extreme events.

There are various physical reasons (see Trenberth, 2000) why a large increase in the magnitude of heavy precipitation may occur with only a correspondingly small increase in mean precipitation. It is even possible that heavy precipitation occurrence could increase when mean precipitation decreases, if there is a more radical change in the precipitation distribution (Osborn and Hulme, 2002).

A study by Nearing (2001) estimated potential changes in rainfall erosivity in the United States during the 21st century under climate change scenarios. He concluded that, across the United States over an 80 year period, the magnitude of average changes in rainfall erosivity was 16–58%. This variability in the magnitude was due to the method (two GCM models and two scenarios) that he used to predict the changes in rainfall erosivity. Regardless of which method was used, he suggested that changes in erosivity will be critical at certain locations.

In order to run a soil erosion model such as WEPP (Water Erosion Prediction Project, See Section 1.5.3 for more details), for example, various weather parameters for each day of the simulation period are required (Flanagan and Nearing, 1995). These weather variables (e.g. rainfall depth and duration, peak storm intensity and time to peak, minimum and maximum temperatures, dew point temperature, solar radiation, wind speed and direction) can either be generated by CLIGEN (CLImate GENerator, See Section 1.5.3.1 for more details) or compiled manually from observed climate data.

Generating climate data for studies on future soil erosion is not a simple task, even with today's climate data, as a starting point, since all erosion predictions must involve modelling extreme weather events. Extreme weather events (e.g., heavy showers, gusts and tornadoes) are rare and occur on the synoptic and even smaller temporal and spatial scales (Schubert and Henderson-Sellers, 1997; Katz, 1999; Coppus and Imeson, 2002). Long integrations of very high-resolution models are required to simulate those extreme events and even then, there is little prospect that sub-synoptic scale events can be successfully resolved in GCMs. GCM grid sizes are too large to properly capture convective elements in the atmosphere, so that precipitation within a short period (e.g., one day) is poorly reproduced by GCMs (Schubert and Henderson-Sellers, 1997).

There are a few ways for resolving this scale issues with GCM data. One way is by using climate data generated directly by Regional Climate Models (RCMs) that are capable of generating climate data with a sub-daily resolution (i.e. 20-min). Another can be achieved by downscaling. There are several approaches for downscaling GCM data into regional scale. Wilby and Wigley (1997) divided downscaling into four categories: regression methods, weather pattern (circulation)-based approaches, stochastic weather generators and limited-area climate models. Among these approaches, circulation-based downscaling methods perform well in simulating present observed and model-generated daily precipitation characteristics, but regression methods are preferred because of its ease of implementation and low computation requirements. RCM data and downscaled data allow predictions to be made at a finer scale than GCMs. All of these methods are widely accepted methods that were often chosen to generate climate data with a sub-daily resolution.

Lastly, IPCC (2001*a*) reported generalised results from the analysis of five regional climate change simulations. Although scenarios for precipitation produced by these experiments varied widely among models and from region to region, the results provide very important working envelopes for this research. The results related to precipitation are summarised as follows:

1. Regional precipitation error spanned a wide range, with values as extreme as approximately −90% or +200%.
2. Simulated precipitation sensitivity to doubled CO_2 was mostly in the range of −20% to +20% of the control value.
3. Overall, the precipitation errors were greater than the simulated changes. It can be expected that, due to relatively high temporal and spatial variability in precipitation, temperature changes are more likely to be statistically significant than precipitation changes.

1.5 Soil Erosion Prediction Models

1.5.1 Background and Categories of Erosion Models

To assess the risk of soil erosion, estimates of soil loss rates may be compared with what is considered to be acceptable for conservation purposes; the effects of different conservation strategies may then be determined. Consequently, a technique is required to compare possible soil losses under a wide range of conditions. One way of doing this is using computerised models for soil erosion, which are (like all models) simplified representations of reality. Types of erosion models are categorised by their structures in Table 1.1. It needs to be noted that the categories in Table 1.1 tend to be mixed, nowadays.

(Favis-Mortlock *et al.*, 2001)

Another categorisation scheme is based on objectives and levels of performance. In this scheme, there are two basic types of models in addition to the categories in Table 1.1. One is a screening model, which is relatively simple and designed to identify problem areas. This type of model only requires predictions of the right order of magnitude. The other type is an assessment model, which requires better, more robust, and accurate predictions because it is mainly designed for evaluating the severity of erosion, for example, under different soil management systems. Thus, depending on the purpose to which a model is put, the appropriate level of complexity/simplicity of the model should be established. A clear statement of the purpose of the study is essential; this will serve as a starting point for all modelling procedures.

Our current understanding of erosion processes is greatest over short time periods, seconds to minutes. It is thus problematic when applying this understanding to longer periods, e.g. months to years or even longer, as is necessary for real-world conservation tasks. It may just be feasible for slightly longer periods such as hours or days, but continuous extrapolation is not appropriate (Kirkby *et al.*, 1992; Morgan, 1995). Therefore, longer-term prediction can only be achieved by summing the predictions for individual events, or developing models empirically, using data collected on a long-term basis, or improving our understanding of processes to be able to build physically-based models.

In addition to these temporal extrapolation issues, the spatial extrapolation issues must also be considered. For instance, the detailed requirements for modelling erosion over a large drainage basin (Hooke, 2000) may differ from those demanded by models of soil loss from a short length of hillslope (Goff *et al.*, 1993), or even at the point of impact of a single raindrop (Sharma *et al.*, 1991). Until recently, integrating researches in different scales (i.e. plot, field and catchment-scale) has been neglected because it is a difficult task (Boardman, 1996).

Therefore, prior to using a soil erosion model, where the model is to be used and why the specific model is appropriate should be considered carefully.

For error and uncertainty involved in modelling approaches, Oberkampf *et al.* (2002) suggest two definitions of uncertainty: aleatory and epistemic. Aleatory uncertainty (also can be termed as 'unknown unknowns') refers to irreducible uncertainty, inherent uncertainty, variability and stochastic uncertainty. A probability or frequency distribution is generally used to quantify aleatory uncertainty, when sufficient information is available. Epistemic uncertainty (also can be termed as 'known unknowns') refers to reducible uncertainty, subjective uncertainty and cognitive uncertainty. This is a source of non-deterministic behaviour that comes from lack of knowledge of the system or environment. This uncertainty can also be viewed as a potential inaccuracy in any phase or activity of the modelling process that is due to lack of knowledge.

Thus, even if we eliminate epistemic uncertainty (or known unknowns) by studying

Table 1.1 Types of erosion models (from Morgan, 1995)

Type	Description
Physical	Scaled-down hardware models usually built in the laboratory; need to assume dynamic similitude between model and real world.
Analogue	Use of mechanical or electrical systems analogous to system under investigation, e.g. flow of electricity used to simulate flow of water.
Digital	Based on use of digital computers to process vast quantities of data.
Physically-based	Based on mathematical equations to describe the processes involved in the model, taking account of the laws of conservation of mass and energy.
Stochastic	Based on generating synthetic sequences of data from the statistical characteristics of existing sample data; useful for generating input sequences to physically-based and empirical models where data only available for short period of observation.
Empirical	Based on identifying statistically significant relationships between assumed important variables where a reasonable database exists. Three types of analysis are recognised: - Black-box: where only main inputs and outputs are studied; - Grey-box: where some detail of how the system works is known; - White-box: where all details of how the system operates are known.

and obtain absolute knowledge, we will still not be able to predict future weather perfectly because of the aleatory uncertainty (unknown unknowns).

Oberkampf *et al.* (2002) defines error as a recognizable inaccuracy in any phase or activity of modelling and simulation that is not due to lack of knowledge. Using this approach, there are two types of errors: errors that are acknowledged and errors that are unacknowledged. Acknowledged errors are those inaccuracies that are recognised by the analysts. Unacknowledged errors are those inaccuracies that are not recognized by the analysts, but they are recognizable. The CLIGEN errors found by Yu (2000) can be seen as an example of unacknowledged errors (See Section 1.5.3.1).

Oberkampf *et al.* (2002) further suggest a comprehensive and new view of the general phases of modelling and simulation, consisting of six phases (Figure 1.2):

1. conceptual modelling of the physical system
2. mathematical modelling of the conceptual model
3. discretization and algorithm selection for the mathematical model
4. computer programming of the discrete model
5. numerical solution of the computer program model
6. representation of the numerical solution

This framework is a synthesis of the reviewed literature, with three substantial additions compared to a more conventional viewpoint. First, it makes a more precise distinction between the system and the environment. Second, it places more emphasis on the distinction between aleatory and epistemic uncertainty in the analysis. Third, it includes a dominant element in the simulation of complex physical processes; the numerical solution of non-linear Partial Differential Equations (PDFs).

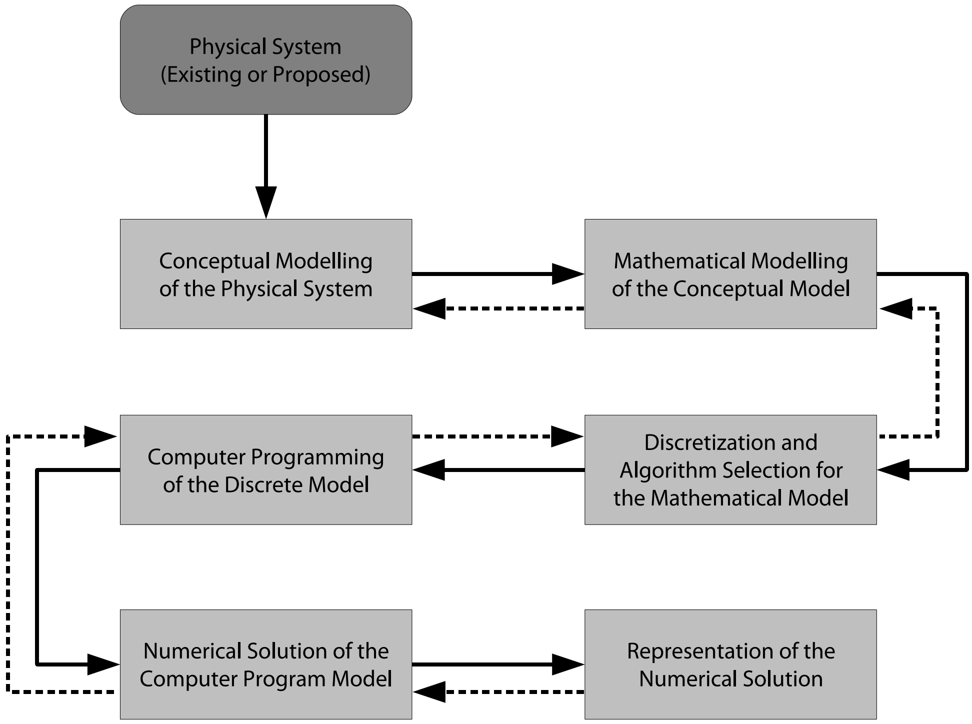

Figure 1.2 Proposed phases for computational modelling and simulation. (From Oberkampf *et al.*, 2002)

Conceptual modelling of the physical system Conceptual issues about the physical system are considered by determining all possible factors that might affect the system.

Mathematical modelling of the conceptual model The primary activity of this phase is to develop detailed and precise mathematical models. The complexity of the models depends on the physical complexity of each phenomenon being considered, the number of physical phenomena considered, and the level of coupling of difficult types of physics. Emphasis on comprehensiveness in the mathematical model should not be interpreted as an emphasis on complexity of the model. The predictive power of a model depends on its ability to correctly identify the dominant controlling factors and their influences, not upon its completeness or complexity. A model of limited, but known, applicability is often more useful than a more complete model. Any mathematical model, regardless of its physical level of detail, is by definition a simplification of reality.

Discretization and algorithm selection for the mathematical model Converting the mathematical models into a form that can be addressed through computational analysis. Conversion of the continuum mathematics form of the mathematical model into a discrete, or numerical, model. Specifying the methodology that dictates which computer runs will be performed in a later phase of the analysis to accommodate the non-deterministic aspects of the problem.

Computer programming of the discrete model—modular approach Algorithms and solution procedures defined in the previous phase are converted into a computer code.

Numerical solution of the computer program model The individual numerical solutions are actually computed.

Representation of the numerical solution The representation and interpretation of both the individual and collective computational solutions. Basically this phase concerns how to present results for a group of specific audiences.

The erosion models used in the present research are reviewed in the next section. However, an additional model, the Universal Soil Loss Equation (USLE), is first discussed since this model embodies the basic concepts underpinning many more recent models such as the WEPP (Water Erosion Prediction Project) model, which is also reviewed in this chapter.

1.5.2 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. 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 account of the protection-effectiveness of different crops, a conservation factor and a soil erodibility factor, consecutively. All these factors were then incorporated together, modified and up-dated to the Universal Soil Loss Equation (USLE) (Wischmeier and Smith, 1978).

The USLE consists of six factors, which are simply multiplied together to estimate soil loss although there is substantial interdependence between the variables (Wischmeier and Smith, 1978):

$$A = R \times K \times LS \times C \times P \tag{1.1}$$

where A (tonnes·ha^{-1} yr^{-1}) is average annual soil loss, R (MJ·mm·hr^{-1} ha^{-1} yr^{-1}) is rainfall erosivity, K (t·hr·MJ^{-1} mm^{-1}) is soil erodibility, L (dimensionless ratio) is the slope-length factor, S (dimensionless ratio) is the slope-steepness factor, C (dimensionless ratio) is the cropping factor, and P (dimensionless ratio) is the conservation practice factor.

The rainfall erosivity factor (R) is related to the raindrop impact effect. R factor provides relative information on the amount and rate of runoff associated with the rain. The soil erodibility factor (K) is used to represent the differences of natural resistances of soils to erosion. The slope length (L) and steepness(S) factors provide the topographic information that can affect the rate of energy dissipation. The cropping factor (C) is the ratio of soil loss from cropped field under specific conditions to the corresponding loss from tilled, continuous fallow conditions. The conservation practice factor (P) is the ratio of soil loss with a specific conservation practice to the corresponding loss with conventional slope tillage.

The USLE uses the empirical results of erosion studies conducted at many locations over nearly a half-century of research, including rainfall erosivity, soil erodibility, slope length, slope steepness, cropping and management techniques, and supporting conservation practices of more than 10,000 plot-years of data from about 50 locations in 24 states in the US (Wischmeier and Smith, 1978). The results were statistically analysed and the relationships between the factors incorporated into equation 1.1.

Both the strength and weakness of the USLE lie in its estimation of erosion as the product of a series of terms for rainfall, slope gradient, slope length, soil, and cropping factors. However, it does not account for any non-linear interactions between the factors (Wischmeier and Smith, 1978; Meyer, 1984).

Nicks (1998) suggests that USLE may be used to estimate soil loss on a storm by storm basis where incremental rainfall is available. Rainfall erosivity index (EI) for a rainfall event is calculated by

$$EI = R_{0.5} \sum (210 + 89\, log_{10} I) \tag{1.2}$$

where I is the incremental rainfall intensity and $R_{0.5}$ is the maximum storm 30 minutes rainfall. Individual storm erosion amounts may then be calculated with the USLE using this EI value to replace the R factor in equation 1.1, summed to give a yearly soil loss, and then averaged to produce a mean annual erosion estimate.

In contrast, Kinnell (2005*b*) points out important problems of predicting event erosion using the USLE. One of the main problem described is that, in the USLE, there is no direct consideration of runoff even though erosion depends on sediment being discharged with flow, which varies with runoff and sediment concentration. Kinnell (2005*b*) concludes that the failure to consider runoff as a primary factor in the USLE is the factor that causing the USLE to produce the erroneous prediction of event erosion, which in turn leads to systematic errors in predicting average annual soil loss.

Since the introduction of the USLE to estimate soil loss, it has become the conservationists' primary tool for planning purposes (Diaz-Fierros *et al.*, 1987; Centeri, 2002). The USLE provides an ease of use and relatively reliable results, and requires only readily obtainable information in order to estimate average annual soil loss. However, Wischmeier (1976) warned about the problem of the misuse of the USLE.

The database for the original USLE is restricted to the US east of the Rocky Mountains (Wischmeier and Smith, 1978). The base is further restricted to slope where cultivation is permissible, normally 0 to 7°, and to soils with a low content of montmorillonite; it is also deficient in information on the erodibility of sandy soils (Wischmeier and Smith, 1978). It is important to note that, because the USLE was designed to estimate average annual soil loss from any specific field over an extended period, soil loss estimates for a specific year may substantially differ from the long-term average predicted by the equation (Wischmeier, 1976). Extrapolating the relationship beyond the database for the original USLE, therefore, should be conducted with care.

The basic concepts of the USLE were subsequently used and developed by some continuous simulation models. Some of these models are CREAMS (Chemicals, Runoff, Erosion, and Agricultural Management Systems) (Knisel, 1980), EPIC (Erosion-Productivity Impact Calculator) (Williams *et al.*, 1984), SWRRB (Simulator for Water Resources in Rural Basins) (Williams *et al.*, 1985), WEPP (Water Erosion Prediction Project) (Nearing *et al.*, 1989; Flanagan and Nearing, 1995).

1.5.3 Water Erosion Prediction Project (WEPP)

WEPP (Water Erosion Prediction Project) is a process-based model that describes the processes, such as infiltration and runoff, soil detachment, transport, deposition, plant growth, senescence, and residue decomposition, that lead to erosion (Flanagan and Nearing, 1995). The model takes four input files, climate, soil characteristic, slope, and crop management.

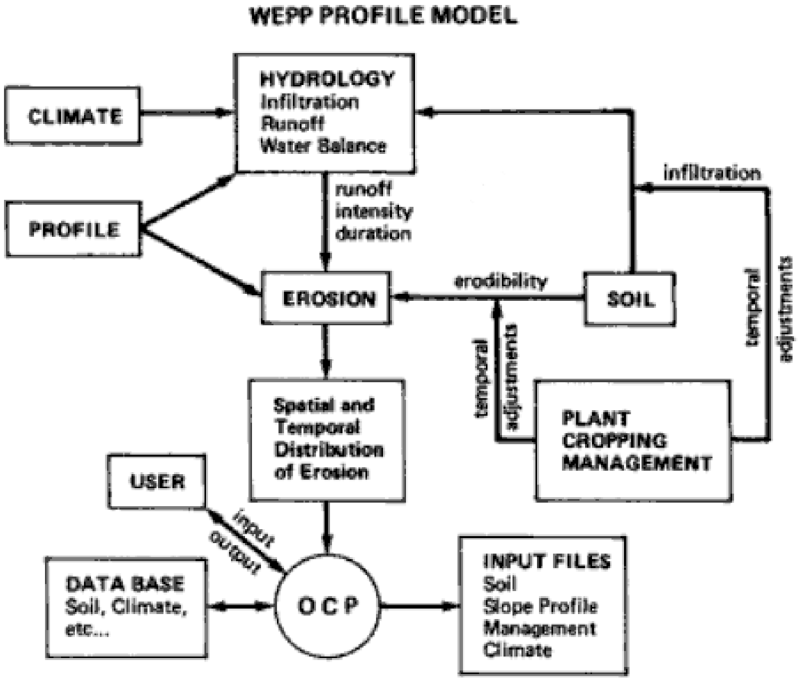

Figure 1.3 Flow chart illustrating the components of the hillslope profile of the WEPP model (from Lane and Nearing, 1989)

WEPP was developed by the USDA-ARS (United States Department of Agriculture-Agricultural Research Service) as a new-generation water erosion prediction technology for the routine assessment of soil erosion for soil and water conservation and environmental planning and assessment (Flanagan *et al.*, 2007). The development of WEPP was initialized with an intention to replace the 'long-used' USLE (See Section 1.5.2) (Nearing *et al.*, 1989; Flanagan and Nearing, 1995). WEPP no longer relies on factor values from the USLE, instead uses separate erodibility parameters for interrill (K_i) and rill erosion (K_r) (Flanagan and Nearing, 1995). In WEPP, rills are assumed to have a uniform rectangular cross-section with a uniform spacing of 1 metre. All rills are assumed to be equally hydrologically efficient (Flanagan and Nearing, 1995).

The steady state erosion component of WEPP is based on:

$$\frac{dG}{dx} = D_f + D_i \tag{1.3}$$

where G represents sediment load, x is the distant downslope, D_f is the rill erosion rate, and D_i is the interrill erosion rate. D_f and D_i are calculated on a per rill area basis. Rill erosion, D_f, is positive for detachment and negative for deposition, and calculated by:

$$D_f = D_c(1 - \frac{G}{T_c}) \tag{1.4}$$

where T_c is the transport capacity of flow in the rill, and D_c is detachment capacity of the rill flow and:

$$D_c = K_r(\tau_f - \tau_c) \tag{1.5}$$

where K_r is rill erodibility parameter, τ_f is flow shear stress acting on soil particles, and τ_c is the critical shear stress or rill detachment threshold parameter of the soil. Interrill erosion is given by:

$$D_i = K_i I_e \sigma_{ir} SDR_{RR} F_{\text{nozzle}} \frac{R_s}{\omega} \tag{1.6}$$

where K_i is the interrill erodibility, I_e is the effective rainfall intensity, σ_{ir} is the interrill runoff rate, SDR_{RR} is the sediment delivery ratio, F_{nozzle} is an adjustment factor to account for sprinkler irrigation nozzle impact energy variation, R_s is the rill spacing, and ω is the width. Interrill erosion is also expressed with baseline interrill erodibility as (Nicks, 1998):

$$D_i = K_{ib} I_e^2 C_C C_G \frac{R_s}{\omega} \tag{1.7}$$

where K_{ib} is baseline interrill erodibility, C_C is the effect of canopy cover on interrill erosion, and C_G is the effect of ground cover on interrill erosion. The hydrologic variables that drive the WEPP are the effective rainfall intensity and duration, the peak runoff rate, and the effective runoff duration.

The USLE erosion database could not be used directly for the WEPP parametrisation. Three field experiments on cropland, rangeland and forestland were conducted to determine parameters for K_r and K_{ib} given in equations 1.6 and 1.7. A total of 77 sets of plot data were collected (Nicks, 1998). Fixed rainfall intensity was applied to the plots using a rainfall simulator. A comprehensive model description is available in Flanagan and Nearing (1995).

Rainfall is represented in the WEPP with the double exponential function. A storm is described with four parameters, storm amount, average intensity, ratio of peak intensity to average intensity, and time to peak intensity. A stochastic weather generator, CLIGEN (CLImate GENerator; Nicks *et al.*, 1995) is used to generate these storm precipitation inputs. More about CLIGEN is covered later in Section 1.5.3.1. WEPP then disaggregates these storm inputs into a single peak storm intensity pattern (time-rainfall intensity format) for use by the infiltration and runoff components of the model. (Flanagan and Nearing, 1995).

The WEPP model can be used for hillslope erosion processes (sheet and rill erosion), as well as simulation of the hydrologic and erosion processes on small watersheds. The hillslope mode predicts soil erosion from a single hillslope profile of any length. It can be applied to areas up to about 260 hectares in size. The watershed mode links hillslope elements of specified widths together with channel and impoundment elements. WEPP is designed to run on a continuous simulation but can also be operated for a single storm. A modified version of the hillslope WEPP has been developed for research purposes (Favis-Mortlock and Guerra, 1999; Favis-Mortlock and Savabi, 1996). This is designed to account for the effects of atmospheric CO_2 concentration changes on plant growth.

In this research, only hillslope mode of the WEPP (v2004.7) was used for continuous or single-event simulation, depending on the purpose. The current version of the WEPP model is 2010.1[1].

[1]http://www.ars.usda.gov/Research/docs.htm?docid=10621, Accessed in February 2012

1.5.3.1 CLIGEN

CLIGEN is a stochastic weather generator, which generates daily time series estimates of precipitation, temperature, dew point, wind, and solar radiation for a single geographical point, based on average monthly measurements for the period of climatic record (Nicks *et al.*, 1995). The estimates for each parameter are generated independently of the others (Nicks *et al.*, 1995).

In comparison to other climate generators, CLIGEN is better at preserving the low-order statistics of rainfall, temperature, and solar radiation on a daily, monthly, and annual basis (Nicks *et al.*, 1995). Unique to CLIGEN is the capacity to simulate the three additional weather variables to characterize the storm pattern, namely storm duration, time to peak, and peak intensity, which are specifically developed for the WEPP simulation (Flanagan and Nearing, 1995).

CLIGEN stochastically generates four precipitation-related variables for each wet day, which are precipitation amount (mm), R, storm duration (hour), D, time to peak as a fraction of the storm duration, t_p, and the ratio of peak intensity over average intensity, i_p. Average intensity is defined as R/D. Although it is possible to calculate individual variables manually for each rainfall event, it is a labour intensive task to calculate the variables for multiple events.

CLIGEN requires observed precipitation statistics in order to generate these four precipitation-related variables (Table 1.2).

Precipitation data required to derive the CLIGEN input parameters in Table 1.2 are time series of daily precipitation data and sub-daily precipitation data with a time intervals no greater than 30 minutes (Nicks *et al.*, 1995; Yu, 2000). In principle, there is no need to distinguish these two types of precipitation data because sub-daily data can be accumulated to produce daily values. In practice, however, these two types of data

Table 1.2 Precipitation parameters required by CLIGEN to generate WEPP precipitation inputs (from Nicks *et al.*, 1995)

Parameter	Description
meanP	Average precipitation (inches) on wet days for each month
sdP	Standard deviation of daily precipitation (inches) for each month
skP	Coefficient of skewness of daily precipitation (inches) for each month
P(W/W)	Probability of a wet day following a wet day for each month
P(W/D)	Probability of a wet day following a dry day for each month
MX.5P	Average maximum 30-min peak intensity (in/hr) for each month
TimePk	Cumulative distribution of time to peak as a fraction of the storm duration

usually come from two different sources. The coverage of the daily data, both in space and time, is much more extensive in comparison to sub-daily data at short time intervals. In addition, the two types of data are normally stored in different formats. It is therefore useful to treat the two types of precipitation data separately.

CLIGEN (version 4.2) was previously released with WEPP version 2001.3. However, this version of CLIGEN had a major coding error and was modified substantially (Yu, 2000). CLIGEN (version 4.2) computed a ratio $\omega = R_{0.5}/R$, where both $R_{0.5}$ and R were rainfall depth (originally in inches). $R_{0.5}$ had been converted from inches into millimetres (mm), while R was not (Yu, 2000). The CLIGEN code was thus changed to correct this error. This however led to extensively increased storm durations.

To accommodate the correction of unit conversion error, it was necessary to incorporate two important modifications in the CLIGEN codes. First, a new algorithm to determine the monthly means of the maximum 30-min rainfall depth was implemented. Secondly, the parameter values for storm duration and the coefficient of variation for the ratio of the maximum 30-min rainfall depth to daily rainfall required in CLIGEN were estimated using the break-point rainfall data (Yu, 2000).

This error in the CLIGEN code has certainly affected the results from the earlier studies, which employed the previous versions of WEPP and CLIGEN to estimate soil loss (Truman and Bradford, 1993; Zhang *et al.*, 1995*a*,*b*, 1996; Baffaut *et al.*, 1996; Laflen *et al.*, 1997; Baffaut *et al.*, 1998; Favis-Mortlock and Guerra, 1999).

The version of CLIGEN at the time of writing is version 5.22564[2]. This is the version used in this research. However, this version is not the current version any longer. The current version of CLIGEN is 5.3[3].

1.5.4 European Soil Erosion Model (EUROSEM)

EUROSEM is a dynamic distributed event-based model for simulating erosion, transport and deposition of sediment over the land surface by interrill and rill processes (Morgan *et al.*, 1998). The model has explicit simulation of interrill and rill flow; plant cover effects on interception and rainfall energy; rock fragment effects on infiltration, flow velocity and splash erosion; and changes in the shape and size of rill channels as a result of erosion and deposition (Morgan *et al.*, 1998). It can be applied to a small field and up to a small catchment.

EUROSEM requires a one-minute resolution breakpoint rainfall data for the storm. The model then computes, using the breakpoint rainfall data, the interception of the rain by the plant cover, the generation of runoff as infiltration excess, soil detachment by raindrop impact, soil detachment by runoff, transport capacity of the runoff and deposition of sediment. The model has a modular structure that aims to make further improvements of the model easier. The model considers the followings:

- the interception of rainfall by the plant cover
- the volume and kinetic energy of the rainfall reaching the ground surface as direct throughfall and leaf drainage
- the volume of stemflow

[2]http://horizon.nserl.purdue.edu/Cligen/, April 2006
[3]http://www.ars.usda.gov/Research/docs.htm?docid=18094, Accessed in February 2012

- the volume of surface depression storage
- the detachment of soil particles by raindrop impact and by runoff
- sediment deposition
- the transport capacity of the runoff
- frozen soils and stoniness

The flow chart for EUROSEM is shown in Figure 1.4.

Runoff generator and the water and sediment routing routines of EUROSEM are from another model called KINEROS (Woolhiser *et al.*, 1990). The volume of sediment passing a given point on the land surface at a given time is calculated by a mass balance equation:

$$\frac{\partial(AC)}{\partial t} + \frac{\partial(QC)}{\partial x} - e(x,t) = q_s(x,t) \tag{1.8}$$

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

$$e = DR + DF \tag{1.9}$$

where DR is the rate of soil particle detachment by raindrop impact ($m^3\ s^{-1}\ m^{-1}$), and DF is the balance between the rate of soil particle detachment by the flow and the particle deposition rate ($m^3\ s^{-1}\ m^{-1}$).

The EUROSEM simulates erosion and deposition by calculating three main processes, soil particle detachment by raindrop impact, soil particle detachment by runoff, and transport capacity of the flow.

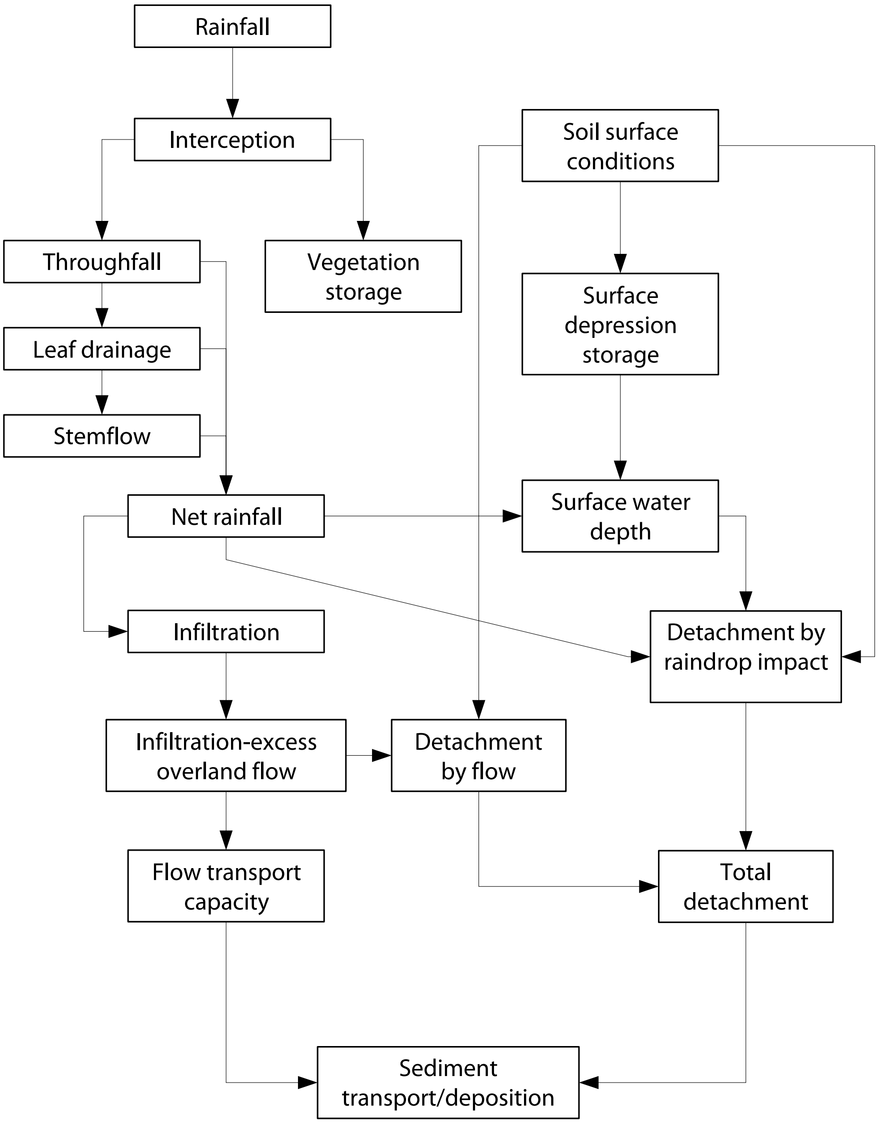

Figure 1.4 Flow chart of EUROSEM (from Morgan *et al.*, 1998)

Soil particle detachment by raindrop impact Soil detachment by raindrop impact (DR) for time step (t_s) is expressed as a function of the kinetic energy of the rainfall at the ground surface, the detachability of the soil and the surface water depth:

$$DR = \frac{k}{\rho_s} KE e^{-zh} \tag{1.10}$$

where k is an index of the detachability of the soil (m^3/J), ρ_s is the sediment particle density (=2.65 Mg/m^3), KE is the total kinetic energy of the net rainfall at the ground surface (J/m^2), z is an exponent taken as equal to 2.0 which varies between 0.9 and 3.1 (Torri *et al.*, 1987), and h is the mean depth of the surface water layer (m). The kinetic energy of the rainfall is the combined energy from direct throughfall and leaf drainage. The energy of the direct throughfall is computed using raindrop size distribution found by Marshall and Palmer (1948). The energy of the leaf drainage is based on a study by Brandt (1990).

Soil particle detachment by runoff Soil particle detachment by runoff is based on a theory proposed by Smith *et al.* (1995), and is given as:

$$DF = \beta w v_s (TC - C) \tag{1.11}$$

where DF is the rate of detachment of soil particles by the flow, β is a flow detachment efficiency coefficient ($\beta = 1$ when deposition is taking place and $\beta < 1$ for cohesive soils when DF is positive), w is the width of the flow (m), v_s is the setting velocity of the particles in the flow (m/s), TC is the sediment concentration in the flow at transport capacity, and C is the actual sediment concentration in the flow.

Transport capacity of flow EUROSEM uses two separate transport capacity relationships for rill and interrill flows. Rill and interrill transport capacities are based on Govers

(1990) and Everaert (1991), respectively. The equation for rill transport capacity (TC_r) is expressed as:

$$TC_r = c(\omega - \omega_c)^{\eta} \tag{1.12}$$

where ω is unit stream power (cm/s) which is defined as $\omega = 10vs$ (v = mean flow velocity (m/s) and s = slope (%)), ω_c is a critical value of unit stream power (0.4 cm/s), and c and η are experimentally derived coefficients related to the median particle size of the soil. Interrill transport capacity (TC_{ir}) is modelled as:

$$TC_{ir} = \frac{b}{\rho_s q}\left[(\Omega - \Omega_c)^{\frac{0.7}{n}} - 1\right]^{\kappa} \tag{1.13}$$

where b is a function of particle size, ρ_s is the sediment density (Kg/m^3), Ω is Bagnold's modified stream power, Ω_c is a critical value of Bagnold's modified stream power, n is Manning's n, and $\kappa = 5$. Sediment delivery to the rills is simulated depending on the transport capacity of the interrill flow.

Since EUROSEM uses a dynamic rather than steady-state approach used by WEPP, it may be assumed to better represent the spatial and temporal variation of runoff and erosion. However, the result of the model simulation may become considerably uncertain due to its process-based nature that requires detailed model parametrization (Quinton and Morgan, 1998). Particularly, EUROSEM requires high resolution rainfall data (ideally, 1-min breakpoint data), soil hydrological information, detailed surface geometry, and soil mechanical and vegetation characteristics. Because of the detailed requirements of the model, application of the model is greatly restricted to where such data are available.

Parsons and Wainwright (2000) found that because EUROSEM ignores small-scale heterogeneities in the infiltration characteristics of soil, the model generates delayed initiation times for runoff, so that predicted hydrographs showed the commencement of runoff later than observed. Such variabilities in the infiltration characteristics may be responsible for the comparatively rapid initiation of runoff on the plot. They also found that the subsequent soil detachment by runoff in interrill areas is overestimated by the model even though, according to the model document, detachment by flow should be negligible in interrill areas.

After personal communication (16 Jun 2004) with Anthony J. Parsons , it is noted that EUROSEM may have a unit conversion error. The model document states that the flow depth (h) in equation 1.11 is in metres. However, a study by Torri *et al.* (1987) on which this equation is based indicates that the height is in millimetres (see Figure 2 in Torri *et al.* (1987)). Anthony J. Parsons suggests that the height is in centimetres rather than either metres or millimetres. If confirmed, this error would have major effects on EUROSEM's ability to estimate runoff and erosion. However, this was not investigated further in this thesis.

The version of EUROSEM used in this research is 3.9 (14/12/1998), which is the current version of the model. It seems that the model development has been ceased for some time.

1.5.5 RillGrow

While the later two erosion models (i.e. WEPP and EUROSEM) reviewed previously are capable of realistically simulating rates of soil erosion, they (in common with all other present-day erosion models) have a number of conceptual shortcomings. For example, rills are considered to be equally spaced, with regular cross-sections, and to be of similar hydrological efficiency. In reality, rills are not necessarily spaced regularly and often have irregular cross-sections. Adjacent rills may also vary greatly in their ability to transport runoff and sediment (Favis-Mortlock, 1996). Additionally, while such models separately describe the different processes responsible for erosion in rill and interrill areas, they largely fail to acknowledge the physical link that exists between the processes operating in the two zones (Favis-Mortlock *et al.*, 2000).

The development of the RillGrow model (Favis-Mortlock, 1996, 1998*b*; Favis-Mortlock *et al.*, 2000; Favis-Mortlock and de Boer, 2003; Favis-Mortlock, 2004) started with a consideration of these shortcomings, and a question '*Is the initiation and development of hillslope rill systems driven by relatively simple rules acting on a much smaller scale?*'. To test this hypothesis, a self-organising dynamic systems approach was used to simulate the initiation and development of a rill network on the bare soil of a small (e.g. plot-sized) hillslope area.

RillGrow is a single-event model, which generates realistic rill networks by simulating, on a grid of microtopographic elevations, the combined erosive action of overland flow moving between the cells of the grid. Surface water arrives, a single raindrop at a time, on random cells of the grid. Runoff moves over the grid following the steepest microtopographic gradient; as it moves, it erodes the soil surface by lowering the elevation of the soil's surface. Each change in elevation affects the routing of subsequent runoff: the result is a feedback loop in which flow patterns over the grid at any time depend on earlier flow patterns (and hence erosion), these flow patterns then condition subsequent flow patterns, and so on (Figure 1.5).

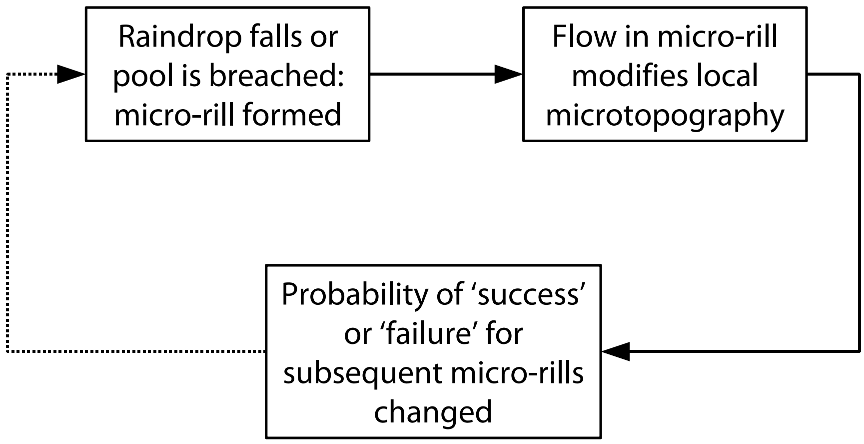

Figure 1.5 The conceptual feedback loop of RillGrow (From Favis-Mortlock, 1998*b*)

The first version of the RillGrow model was able to realistically simulate rill initiation and development, reproducing several features of observed rill systems (Favis-Mortlock, 1996, 1998*b*):

1. a narrowing of rill spacing with increased slope angle
2. an increased contribution of rill erosion with downslope distance
3. a non-linear increase of total erosion with slope steepness
4. an increase in rill depth below confluences and micro-rill piracy

However, RillGrow 1 has some important limitations. It ignores important process descriptions (e.g. infiltration, deposition), and the hydraulics of rill initiation are oversimplified. It also does not operate in a true time domain, since at any moment during the simulation, only a single 'packet' of overland flow is moving over the topographic grid. Infiltration is ignored, thus all water on the grid is assumed to be rainfall excess, and transport capacity is assumed to be infinite because no deposition can occur (Favis-Mortlock *et al.*, 2000). All these limitations meant that it was not possible to validate the model in a deterministic way, e.g. by comparing simulated and laboratory-produced rill networks.

A second version of the model ('RillGrow 2') was developed to overcome some of these limitations, and allow more rigorous validation (Favis-Mortlock *et al.*, 2000). In RillGrow 2, overland flow in effect moves concurrently between cells of the microtopographic grid. Such concurrency is not easy to achieve on a serial-processing computer, since only one instruction can be carried out at a time. Concurrent processing is simulated in RillGrow 2 in the following way: during a given timestep, outflow from each 'eligible' wet cell is routed in a random sequence which differs for each timestep. This variation in sequence is necessary to prevent any artefacts of flow pattern which might result from a fixed routing sequence. Over a sufficient number of timesteps, the model in effect operates in 'parallel', i.e. concurrently, in a true time domain. RillGrow is now at version 6, parallelised for multiple computer processors (personal communication with David Favis-Mortlock, 2012).

Table 1.3 The main routing algorithm used in RillGrow 2 (From Favis-Mortlock *et al.*, 2000). Note that, while this set of rules does not change during the model simulation, the result of applying the rules to a given cell depends on the past history of elevation change both for that cell, and for adjacent ones.

For each iteration:

Drop raindrops on random cells on the spatial grid. Each drop makes (or adds to, if the cell is already 'wet') a store of surface water for that cell.

Go through all 'wet' cells in a random sequence (which is different each iteration). For each 'wet' cell, check whether sufficient time has elapsed for flow to have traversed the cell. If it has, then do the following:

- Find the adjacent cell which has the steepest downhill (i.e. outflow) energy gradient. Note that this adjacent cell may or may not be already 'wet'.
- Attempt to level the energy gradient between these cells by outflow of an appropriate volume, adding to (or creating) the store of surface water on the adjacent cell.
- Erode this cell (i.e. lower its soil-surface elevation) by an amount which depends on the outflow volume and velocity.
- If there are other adjacent cells with downhill energy gradients, process these as above.

RillGrow 2 also uses a refined set of basic rules for the routing algorithm (Table 1.3). A probabilistic expression by Nearing (1991), based on the random occurrence of turbulent bursts, is used to represent flow detachment. Sediment load is compared with transport capacity, which is calculated using stream power in an s-curve expression developed by Nearing *et al.* (1997). Deposition is estimated using the approach of Lei *et al.* (1998): this assumes deposition to be a linear function of the difference between sediment load and transport capacity.

Infiltration is now simulated by RillGrow 2, however the approach used is still very simple (Favis-Mortlock, personal communication 2006). Splash redistribution is also represented, using the diffusion-based approach of Planchon *et al.* (2000), modified to include attenuation due to surface water depth (Favis-Mortlock, personal communication 2006). Additionally, a simple linkage is made between overall volumes of sediment detached or deposited by splash in any timestep, and volumes of sediment currently being transported by overland flow (Favis-Mortlock, personal communication 2006). The distribution of raindrop volumes is assumed to be normally distributed (Favis-Mortlock, personal communication 2006).

In comparison to models such as WEPP and EUROSEM, RillGrow makes no explicit separation between rill and interrill areas. Soil erosion amount is calculated as the result of summation of soil losses from all areas. Flow velocities are so low on interrill areas that little detachment occurs as a result. On such areas, splash redistribution dominates.

Favis-Mortlock *et al.* (2000) observes that RillGrow 2 replicates the responses of RillGrow 1 including realistic rill spacing with respect to slope gradient. In addition, the simulated spatial patterns of erosion compare well with laboratory-based observations, as do total erosion, variation in rill depth, and time-series of runoff and soil loss. Braided or dendritic flow patterns can be made to emerge by varying rainfall intensity. At low rainfall intensities, however, the definition and stability of rill patterns is less well defined.

As originally constructed, RillGrow 2 assumed each storm to have a constant rainfall intensity. For the research described in this thesis, RillGrow 2 has been modified to use breakpoint rainfall data. This allows the model to simulate rill initiation and development and total soil loss, as affected by rainfall intensity changes (Favis-Mortlock, personal communication 2006).