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Measurement and modelling of soil hydrological properties for use in the distributed rainfall runoff model – GEOtop

Thesis presented by

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Declaration

I declare that this thesis has not been previously submitted for a degree at the National University of Ireland or any other university and I further declare that the work embodied in it is my own, or else noted. This thesis forms part of a larger Environmental Protection Agency (EPA) project titled *Interactions of soil Hydrology, land use and climate change and their impact on soil quality (SoilH).* As a result of this, sections of chapters 4 to 7 contain contributions from colleagues in the *Soil H* project. The breakdown of my work and that of my colleagues is as follows:

Chapter 4. I was responsible for the field infiltration tests of all 31 mineral sites with assistance from Dr Xianli Xu and Valérie Piaux on a number of sites. I was assisted in the laboratory analysis of bulk density and moisture content by Yvonne Murphy. I was also responsible for the BEST analysis of the infiltration experiments and the mapping of the hydrological classifications of Irish soils. The BEST method was coded in Matlab by Prof. John Albertson of Duke University, N.C., USA.

Chapter 5. I was the lead author on the *Hydrological Processes* peatland hydrology paper (now in press) contained in chapter 5. I was assisted in the field and laboratory work by Tom Fitzpatrick and Debbie Buckley.

Chapter 6. I was responsible for all the hydrological modelling and analysis of peatland afforestation and I am also the lead author in a paper submitted to *Hydrological Processes* on the same topic.

Chapter 7. The incorporation of an erosion module in the hydrological model was carried out by Mr. Tan Zi and Prof. John Albertson of Duke University, N.C., USA. However, I was responsible for hydrological modelling and analysis carried out in chapter 7 and I am the lead author on a paper in preparation arising from this work.

Ciaran Lewis
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Abstract

On site infiltration tests were carried out across a range of different soils in Ireland for the purposes of developing a national soil hydrological classification scheme. The results of the infiltration tests on mineral soils were analysed for soil hydraulic parameters (e.g. saturated hydraulic conductivity (K_{sat})) using the BEST method. Deep well drained mineral soils were found to have a average K_{sat} of 19.2 (max 249; min 0.35) 10^{-6} m s^{-1} with deep poorly drained soils having an average K_{sat} of 0.89 (max 2.4; min 0.25) 10^{-6} m s^{-1}. Investigations into the spatial variation of K_{sat} and bulk density (ρ_b) in a pristine blanket peatland found that the peatland is composed of two distinct zones: one near the margins (i.e. near a stream) and the second at the bog interior. At the near surface (10 to 20 cm depth), K_{sat} was to found to be higher (~10^{-5} m s^{-1}) in the bog interior than the riparian zone (~10^{-6} m s^{-1}) while the converse applied to ρ_b, with lowest density (~ 0.055 g cm^{-3}) at the interior and highest (~ 0.11 g cm^{-3}) at the riparian zone. These results support the theory that areas of lower K_{sat} at the margins control the hydrology of blanket peatlands. A hydrological modelling study using the hydrological model GEOtop into the hydrological response and water balance of upland catchments as a result of water table drawdown and afforestation found that the hydrological response of the peatland catchments changes. It was found that afforestation results in a decrease in streamflow and an increase in evapotranspiration, particularly in summer. However, in winter, following periods of heavy rainfall, peak streamflow increased. The suspended sediment yield (SSY) of an Irish grassland catchment was simulated using GEOtop with an added LISEM erosion module and compared well with measured values. Scenario modelling found that SSY is sensitive to rainfall intensity, with SSY increasing linearly with increasing rainfall intensity.

Chapter 1

Introduction

1.1 Background

Soils are a non-renewable resource that provide a range of agricultural, economic and environmental services. These include: the support of food and fibre production; control of the fate of water in the hydrologic system; affecting the loss, purification, contamination and utilisation of water provision of habitats for organisms; providing the foundation for buildings and road infrastructure; and acting as a store for carbon in the form of organic matter. Fertile soil is essential to food security and human health and must be protected (Boardman and Poesen, 2006). Soils have long endured degradation pressures or threats from natural and human factors. Indeed many societies foundered as a result of unsustainable soil management practices (Montgomery, 2007).

While the importance of air and water quality has long been recognised, the need to ensure soil quality has only recently been appreciated. At EU level there have been a number of initiatives. The Thematic Strategy for Soil Protection (CEC, 2006) identified soil degradation as a serious problem in Europe. It states that this is "*driven or exacerbated by human activity such as inadequate agricultural and forestry practices, industrial activities, tourism, urban and industrial sprawl and construction works*." Such degradation reduces the ability of the soil to perform essential functions with reduced fertility, carbon, biodiversity, water retention capacity, disrupted gas and nutrient cycles and less degradation of contaminants.

In order to adequately protect soils in Ireland, there is a need to understand the soil hydrological processes and assess their variability and to investigate how soil hydrology interacts with land use and climate change. These interactions have implications for soil quality and the vital services that it provides. It is only when the baseline is adequately described (in terms of soil quality) and quantified that future actions to protect soil can be made.

1.2 EPA Proposal

This thesis forms part of a larger Environmental Protection Agency (EPA) project titled *Interactions of soil Hydrology, land use and climate change and their impact on soil quality (SoilH)* and the proposal is summarised as follows.

The SoilH project established a network of benchmark sites throughout Ireland using existing national sites for the measurement of soil hydrological properties (e.g. hydraulic conductivity, porosity and the van Genuchten (1980) parameters α and n) and the establishment of a hydrological classification of Irish soils. A process-based soil hydrological model (GEOtop) (Rigon et al., 2006) was employed and a new module for erosion was developed by Albertson and Zi, colleagues from Duke University, USA as partners in the SoilH project. The model was used to elucidate the interactions between soil hydrology, land use and climate change (with climate projections from the IPCC fourth assessment). These outputs were combined with Irish geo-spatial data to develop a GIS-based risk assessment tool to predict impacts on soil quality based on hydrology, land use and climate change. Other threats to soil including landslides, compaction, loss of organic matter and surface sealing were also investigated as part of the overall EPA project.

This thesis focused on three different aspects of the SoilH project:

A) Measurement of soil hydrological properties in mineral and peat soils through both on-site infiltration tests and laboratory tests for the purposes of establishing a hydrological classification of Irish soils.

B) Use of the hydrological model GEOtop for the purposes of assessing the impact of any future afforestation of peatlands on the rainfall runoff response of blanket peatlands.

C) An assessment of the impacts of increasing rainfall intensity on erosion on a grassland catchment.

Other parts of the project such as the threats to soil quality from loss of organic matter, landslides, surface sealing were assessed by others in the project. Both the GIS-based risk assessment tool and manuals predicting impacts on soil quality and the insertion of the erosion module into the hydrological model code were carried out by collaborating partners.

1.3 Aims

This study investigates the hydrological properties of both mineral and peat soils using field sampling and laboratory analysis of mineral and peat soils along with a distributed hydrological model. For mineral soils, a total of 31 infiltration tests were carried out at field sites thought Ireland, selected from the national soils database (NSD). In addition field tests were also carried out in an Atlantic blanket peatland bog to determine the spatial variation of saturated hydraulic conductivity and bulk density. The findings from the field tests were then used in a modelling effort to assess the impact of possible future afforestation on the rainfall runoff response and erosion rates of a peatland catchment.

This thesis is divided into four projects, each of them with a specific aim;

A) Estimation and analysis of soil hydraulic properties through infiltration experiments for the purposes of developing a national soil hydrological classification.

B) To investigate the spatial variation of saturated hydraulic conductivity and bulk density of an Atlantic blanket peatland.

C) Using the hydrological model (GEOtop) investigate the changing hydrological response from possible future afforestation of blanket peatlands.

D) Analyse the effects of rainfall intensity using GEOtop (and the new erosion module) on the suspended sediment yield of a grassland catchment.

1.4 Layout of Thesis

The thesis contains 9 chapters, a list of references and two Appendices. Chapter 2 is a literature review addressing the current topics relating to soil hydrological properties and erosion as well as a review of peatland hydrology. The methodologies used to perform this work and a general description about the study sites is presented in Chapter 3. Chapter 4 focuses on the infiltration results of the 31 mineral sites. Chapter 5 describes the spatial change in saturated hydraulic conductivity and bulk density along a transect in an Atlantic blanket peatland. Chapter 6 is dedicated to modelling the effects of afforestation on rainfall runoff response of a blanket peatland. Chapter 7 is an assessment of the effects of rainfall intensity on soil erosion from a grassland catchment. Chapter 8 is a general discussion of the thesis and Chapter 9 is some recommendations for future research.

This is followed by the reference list and two Appendices. The first appendix contained in a separate volume, details all the field infiltration tests and results. The second appendix located at the back of this volume contains two papers from the current Ph.D. published in the journals of *Soil use and management* and *Hydrological Processes*.

Chapter 2

Literature Review

2.1 Soils and soil erosion

Land use, farming systems, and agricultural practices may strongly affect water flow over soils and erosion potential. In the light of climate change, where an increase in the frequency and duration of dry periods (droughts) as well as increasing precipitation amounts and extremes events (floods) are expected in many areas of the world (IPCC, 2001), there are increased risks for landslides and soil erosion. Climate change in Ireland from the baseline period of 1961 -2000, is predicted to: have an increase in temperature for all months of between 1.25 and 1.5 °C; a decrease in summer precipitation (of ~ 10%); and an increase in winter precipitation (of ~ 15%) for the 2021-2060 period (McGrath et al., 2005). This increased precipitation trend has already been detected in the west of the country since the mid-1970s (Hoppe and Kiely, 1999; Kiely, 1999). The climate change effects, interacting with land use change could result in increased erosion in both mineral and peat soils.

Soil hydraulic properties are fundamental to quantifying the erosion process. There are however relatively few studies carried out in Ireland where such properties such as hydraulic conductivity, particle size distribution (PSD) and water retention characteristics have been quantified for mineral or peat soils. One of the most comprehensive studies of Irish soils was carried out by Gardiner and Radford (1980) where numerous soil profiles were analysed throughout the country which resulted in 45 different soil classifications of Irish soil. The Gardiner and Radford (1980) study focused on agricultural practices and was not concerned with soil hydraulic properties. While the Gardiner and Radford (1980) study resulted in 45 different soil associations, these associations can be broadly split up into two different soils; mineral and peat soils. Globally peatlands account for 3% of the total land cover (Gorham, 1991; Turunen et al., 2002), but they are one of the most common landscapes in Ireland covering an estimated 13% to 17% of the national land surface (Eaton et al., 2008; Foss et al., 2001). Peat soils are organic in nature and are defined as having a soil organic carbon (SOC) content greater than 15%, with mineral soils having a SOC below 15%. As peat soils are fundamentally different to mineral soils, they are treated separately in this study. The soil hydraulic properties of interest in this study are soil hydraulic conductivity and water retention characteristics. These soil hydraulic properties are important for modelling water and solute transport, managing irrigation and drainage problems, and coupling precipitation and runoff in climate and hydrology models.

2.2 Mineral soils

2.2.1 Measurement and modelling of Soil Hydraulic properties (SHP)

Soil hydraulic conductivities (K_{sat}) have been measured both in the laboratory, and in the field, using different measurement techniques, including the Guelph permeameter, the single-ring pressure infiltrometer, the inverse-auger-hole method with the Porchet solution, the double-ring-type infiltrometer, etc. (Clothier, 1988; White, 1988). Since most methods are time-consuming and costly, many pedotransfer functions (PTF) have been developed and applied widely to translate more readily available soil texture data or soil texture class into soil hydraulic conductivity and water retention characteristics. The Beerkan method proposed by Braud et al. (2005) and Lassabatère et al. (2006), involves combining data from simple infiltration tests in the field and a PTF to estimate the saturated and unsaturated hydraulic conductivity and water retention. To make the Beerkan method more practical in the field, Minasny and McBratney (2007) developed an alternative way of estimating the van Genuchten water retention shape parameter *n* from a soil's sand and clay content using an Artificial Neural Network PTF. Several analytical and empirical methods have been developed to estimate soil hydraulic properties and much effort has been made to evaluate and compare these methods (Christiaens and Feyen, 2001; Islam et al., 2006). Pedotransfer function model uncertainties of soil hydraulic parameters may be large due to soil heterogeneity, and these uncertainties may be propagated into hydrological models (Christiaens and Feyen, 2001). As similar soils can have nuanced differences from country to country, PTF's developed in one country may not adequately describe the behaviour of soils in a different country (Wagner et al., 2001).

2.2.2 Spatial and temporal variation of soil hydraulic properties

Soil hydraulic properties depend on soil structure and texture and therefore tend to vary widely in space. Since the soil hydraulic properties are determined at points in the field (rather than spatially distributed) a large number of determinants are required to assess the magnitude and structure of the variation within the selected area. Distributed hydrological models (e.g. GEOtop) require the spatially distributed input of soil hydraulic properties. Some studies (Herbst et al., 2006a) have revealed that runoff generation (in spatially hydrological modelling) is sensitive to the variation of soil hydraulic conductivity. The question is: how to upscale the point data to regional scale? The use of point measurements and co-variables (e.g., topographical variables) may be optimized for a more accurate spatial prediction (Herbst et al., 2006b; Romano and Palladino, 2002). In addition, Jhorar et al. (2004) pointed out that the vertical variation of soil hydraulic parameters should also not be neglected for successful application of hydrological models. Hydraulic conductivity also shows a temporal variation (Bagarello and

Sgroi, 2007) that depends on different interrelated factors, including soil physical and chemical characteristics affecting aggregate stability, climate, land use, seasonal and dynamics of plant canopy and roots, tillage operations, activity of soil organisms. Further research for this issue is required.

2.2.3 Classification of soil hydrological properties

Clapp and Hornberger (1978) proposed simple power-law descriptors of soil hydraulic properties to maximize parameter identifiability, and for strongly tying parameters to soil texture (i.e. pore size distribution). The work of Clapp and Hornberger (1978) demonstrated this approach for 11 soil textural classes in the US, providing mean and standard deviations for each parameter for each soil class. In a later study by Cosby et al. (1984), 1448 soil samples were examined to formulate predictive relationships describing the hydraulic parameter distributions on the basis of soil structure and particle size distribution. Cosby et al. (1984) demonstrated how discriminant analysis allows for use of the covariation of the hydraulic parameters to construct a classification scheme based on the hydraulic behaviour of soils that is analogous to the textural classification scheme based on the sand, silt, and clay content of soils. There is the potential to add consideration of soil sealing, compaction, and soil organic carbon status to the proposed hydraulic soil classification in this project. Dexter and Czyz (2007) noted that most soil physical properties and behaviour are governed by soil structure as reflected in pore size distribution, which can be determined from the water retention curve. The slope, S, of the water retention characteristic at the inflection point is a measure of the soil structure, and therefore S might be used as an index of soil physical quality and as a variable for the prediction of some soil physical properties and aspects of soil behaviour. They also demonstrated that *S* is related to hydraulic conductivity, friability, tillage (optimum soil water content determination for tillage), compaction, penetrometer resistance, plant-available water, root growth and readily dispersible clay. Dexter (2004) provided descriptive categories of soil physical quality in terms of the corresponding values of *S*.

2.3 Peatlands

2.3.1 Background

Peatlands cover significant areas in northern latitudes and the existence of many peatlands is due to their unique hydrology (Clymo, 2004). Although northern peatlands cover just 3% of the global land surface, they have accumulated between 270 and 450 Pg of carbon, which represents 20 to 30% of the world's estimated soil carbon (Gorham, 1991; Turunen et al., 2002), and their vast stocks of carbon are considered to be particularly vulnerable to climate change (Holden and Burt, 2002b; Oechel et al., 2000; Sottocornola and Kiely, 2010). From a regional perspective, peatlands cover 17% of the land area of the Republic of Ireland (Tomlinson, 2005) and are estimated to contain between 53 and 62% of the national soil carbon stock (Eaton et al., 2008; Tomlinson, 2005). Water table depth has been identified to be of critical importance to the health of peatlands with the possibility of peat undergoing total degradation under dewatering (Bragg and Tallis, 2001). Even a slight drop in water table is expected to affect the chemistry of the bog water, which will likely impact the vegetation composition and distribution (Sottocornola et al., 2009). It is important to maintain these unique landscapes not just because of their large store of carbon but also for biodiversity, as peatlands support a wide range of unique flora and fauna. As such it is important to be able to model the hydrology of peatlands not only for their current status but also under possible future scenarios of climate change which are predicted to result in reduced summer rainfall and increased winter rainfall in Ireland (McGrath and Lynch, 2008).

Knowledge of physical and hydrological properties of soils is a prerequisite for rainfall-runoff modelling and hydrological studies (Albertson and Kiely, 2001; Herbst et al., 2006a). For mineral soils, there is a wealth of information on soil hydrological properties (Brooks and Corey, 1964; Clapp and Hornberger, 1978; Cosby et al., 1984; Montaldo et al., 2001; Nemes et al., 2001; Schaap et al., 2001; van Genuchten, 1980; Wosten et al., 1999). However, the same can not be said for peatlands with limited knowledge of both hydrological properties and of elementary properties (such as bulk density) and particularly their spatial variability (Beckwith et al., 2003a; Egglesmann et al., 1993; Holden and Burt, 2002a; Ingram, 1978; Kiely et al., 2010; Price, 2003; Surridge et al., 2005).

2.3.2 Peatland hydrology

However, despite the importance of blanket peatland hydrology, few studies have carried out detailed hydraulic conductivity measurements (Surridge et al., 2005) including the degree of anisotropy and the spatial variation of vertical and horizontal hydraulic conductivity (Beckwith et al., 2003b). To date efforts at estimating hydraulic conductivity in peatlands using different field and laboratory methods have resulted in a wide range of hydraulic conductivity values (Table 2.1) as low as ~ 10^{-8} m s^{-1} (Hoag and Price, 1995) to as high as ~ 10^{-2} m s^{-1} (Hogan et al., 2006). While some early models of peatland hydrology used in the prediction of the shape of raised bogs (Ingram, 1982), assume that peatland hydraulic conductivity is homogeneous and isotropic, many others since have suggested that peatland hydraulic conductivity is neither homogeneous nor isotropic (Baird et al., 2008; Hoag and Price, 1995; Hogan et al., 2006; Holden and Burt, 2003; Kneale, 1987). Given this reported variable nature of peatland hydraulic conductivity, a wide spread of values within the literature is to be expected. Many of the values reported from the literature are from a single plot or a series of plots scattered through a peatland from the margins to the centre of a bog. Lapen et al. (2005), from a sensitivity analysis of a groundwater model, suggested that areas of lower hydraulic conductivity exist at the margins of peatlands and the margins have a positive impact on bog formation retaining the elevated water table in the bog interior. This hypothesis was tested by Baird et al. (2008) at a raised bog site in Wales and found it to be true but stated that their testing was only confined to one site and may not apply to other sites or other types of peatlands.

The existence of areas of lower hydraulic conductivity at the margins of peatlands which maintain elevated water tables does, however, raise the question of how integral these bogs might be with the stream channels and whether they can provide the flood attenuation function which has often been attributed to wetlands (Evans and Warburton, 2007). However, rainfall runoff from upland blanket peatlands has been observed to be flashy with high flood peaks and short lag times (times of concentration) (Holden and Burt, 2002a). From the catchment perspective and considering the extensive nature of peatlands in the uplands of northern latitudes where many rivers rise, improved knowledge of peatland hydrology is essential for catchment rainfall runoff modelling.

Table 2.1 Hydraulic conductivity (*K*) values reported from a selection of different studies in various peatlands.

Site	Peat type	Depth (m)	Method of analysis	*K* (m s^{-1})	Reference
Canada	Cut over	0.1	Piezometer	1.73 $*10^{-6}$	(Schlotzhauer and Price, 1999)
Ireland	Raised	0.5 – 3	Piezometer	10^{-8} – 10^{-2}	(Kneale, 1987)
Sweden	Raised	0.25 – 2.0	Piezometer	$2*10^{-7}$ – $8*10^{-3}$	(Waddington and Roulet, 1997)
Scotland	Raised	0.1 – 7.0	Piezometer	$1*10^{-6}$ – $1.2*10^{-5}$	(Clymo, 2004)
England	Raised	0 - 0.15	*MCM	$1*10^{-5}$ – $1.2*10^{-3}$	(Beckwith et al., 2003a)
USA	Raised	0.5 – 3.0	Piezometer	$2.5*10^{-6}$ – $2.6*10^{-4}$	(Chason and Siegel, 1986)
Canada	Raised	0.5 – 1.0	Piezometer	$1*10^{-8}$- $5*10^{-6}$	(Fraser et al., 2001)
Poland	Fen		Porous plate	5.5 $*10^{-8}$ – $5*10^{-6}$	(Gnatowski et al., 2010)
Canada	Fen	0 – 2	Piezometer	$1*10^{-6}$ - $9*10^{-3}$	(Hogan et al., 2006)
England	Fen	0 - 1.0	Piezometer and MCM	$1.1*10^{-4}$ – $1.6*10^{-3}$	(Surridge et al., 2005)
Canada	Fen	0 - 0.15	Piezometer	10^{-7} – 10^{-4}	(Kennedy and Price, 2005)
England	Fen	0.1	Piezometer	$6*10^{-7}$ – $6*10^{-6}$	(Baird and Gaffney, 2000)
England	Blanket	0.1 - 0.8	Piezometer	$1*10^{-5}$ – $1*10^{-7}$	(Holden and Burt, 2003)
Canada	Blanket	0.2 – 0.5	Piezometer	$1*10^{-8}$ – $1.6*10^{-2}$	(Hoag and Price, 1995)

* Modified Cube Method

As with hydraulic conductivity there are also limited data available as to the variation of bulk density of different peat types. A number of studies of peatland bulk density have reported bulk density ranges from 0.05 to 0.254 g cc^{-1}, with Tomlinson and Davidson (2000), Kiely et al. (2010) and Wellock et al. (2011) reporting average bulk densities of 0.07, 0.17 and 0.13 g cc^{-1}. Wellock et al. (2011) in a study of 15 afforested peatland sites throughout Ireland found that the bulk density for lowland and high level blanket bogs were similar, but that deeper peats

(>2 m) were found to have a lower density than shallower (<2 m) peats. Wellock et al. (2011) also found that the bulk density of basin peats was higher than blanket peats. There is some conflicting evidence as to the variation of bulk density with depth. Bulk density was shown to generally increase with depth in five peatlands in central and western Europe by Novak et al. (2008) whereas studies by Clymo (2004) and Tomlinson (2005), found no change in bulk density with depth. The studies by Kiely et al. (2010) and Wellock et al. (2011) found that bulk density did not change significantly with depth.

As well as playing an important role in the health and carbon balance of a peatland, hydrological process including surface flow, subsurface flow, buoyancy effects, precipitation timing and intensity, may also play a part in peatland stability. Peatland instability has been documented (Dykes and Kirk, 2006) with hydrological processes being fundamental in determining the spatial and temporal occurrence of peat slides (Warburton et al., 2004) which can be environmentally and geomorphologically significant (Dykes and Warburton, 2008). Failure mechanisms in peat such as shear failure by loading, buoyancy effects, liquefaction and surface rupture all have hydrological controls but a better understanding of peat hydrology is required before realistic models can be developed for predicting future failure (Warburton et al., 2004).

2.4 Peatland Afforestation

Over the centuries much of Ireland's native forestry has been removed, so much so that by the start of the twentieth century forestry only accounted for 1% of the total land cover in the Republic of Ireland (Pilcher and Mac an tSaoir, 1995). However since the 1950s it has been Irish governments policy to increase forest cover and by 2007 the national forest area had risen to 10% (NFI, 2007) with a projected increase to 18% by 2020 (Dept. of Agriculture, 1996).

Much of this recent afforestation has taken place on peatlands which are considered unsuitable for agricultural use. The principle species used in this peatland afforestation is Sitka spruce (*Picea sitchensis* (Bong.) Carr.) and lodge pole pine (*Pinus contorta* Dougl.) (Byrne and Farrell, 2005) with 49% of this afforestation between 1990 and 2000 being carried out on peat soils (Black et al., 2008). Sitka spruce is a non native species but is favoured due to its rapid growth in Irish climatic conditions and ability to withstand difficult site conditions so much so that 57% of the national forest is Sitka spruce (Horgan et al., 2004). While Sitka spruce is able to thrive under the moist Irish weather conditions (Horgan et al., 2004), its root development when planted on peat soils is limited to the upper aerated section of peats (Lees, 1972). Many studies of peatlands, particularly blanket peats have found that the water table remains at or close to the surface for large parts of the year (Bragg, 2002; Hogan et al., 2006; Holden and Burt, 2002a; Iritz et al., 1994; Laine et al., 2007a). This makes many peatlands unsuitable in their natural condition for afforestation from a combination of stunted root development and an associated increase in vulnerability to windthrow. Thus for the purposes for promoting tree growth rate, many peatlands are drained prior to planting (Holden et al., 2004). Drainage is normally carried out using a combination of closely spaced plough furrows and deep (0.5 – 2.0 m) but more widely spaced ditches (Holden et al., 2004).

These drains while beneficial for the development of the forest, have also in the past been linked with higher stream peak flows (Robinson, 1986). Conway and

Millar (1969) found that artificially drained peats produced extremely rapid runoff in the north Pennines (UK) with Ahti (1980) also noting that the flood peak increased after drainage. However Holden et al. (2006) found that surface runoff was greatly reduced following drainage within the Moor house blanket peatlands in the Pennines. Investigations by Iritz et al. (1994) on a selection of forested Scandinavian peatland catchments found that peak flows decreased following drainage works with Prevost et al. (1997) reporting an increase in base flow following drainage in Canadian peatland. While there have been conflicting conclusions drawn from different studies, this is likely due to data limitation and the diversity of ground conditions e.g. as water table depth, which is seen as critical to the amount of storage available and surface runoff production in peatlands (Holden et al., 2006).

While the rainfall runoff response of afforested peatlands is likely to change with drainage, the increased evapotranspiration from the tree canopy must also be considered (Institute of Hydrology, 1991) in any assessment of changing rainfall runoff response. Anderson et al. (2000) attributed a reduction of 7% in runoff to afforestation in a Scottish peatland but also noted that the reduction in runoff was predominately in the spring and summer. Interception losses in a Sitka spruce plantation in Scotland were reported to be greater than 50% of the annual precipitation (Heal et al., 2004). An investigation by the Institute of Hydrology (1991) into the water resources of two upland catchments in Scotland also found significant interception losses of 38% of precipitation with Anderson et al. (1990) also reporting 38% canopy interception and 12% transpiration of gross precipitation. Johnson (1990) found that in a 50 year old Sitka spruce forest in the Scottish highlands the average interception over a 3 year period was 28% with the greatest interception occurring in the summer months and the least in winter. A 25 year old Sitka spruce forest in Northumberland, UK was observed to have an average interception loss of 48% of precipitation (Anderson and Pyatt, 1986). Evapotranspiration from a blanket peatland in southwest Ireland has been observed to vary from 13.5 to 17% of total precipitation representing an average of 394 mm (Sottocornola and Kiely, 2010) with the remaining water leaving the catchment as runoff. Lafleur et al. (2005) found a slightly lower rate of evapotranspiration of 351 mm (5 year average) in the Mer Bleue Bog in Canada. These studies demonstrate that the percentage of precipitation leaving a catchment as evapotranspiration and runoff vary between forests and undisturbed peatlands and any change in land use from natural peatland to forestry is likely to alter the rainfall runoff characteristics of catchment. A review by Hudson et al. (1997a) of studies carried out on a number of catchments including among others the Plynlimon and Lanbrynmair catchments in Wales, concluded that afforestation of uplands had a mainly adverse effect on water quality and water resources. From these studies Hudson et al. (1997a) noted that in the wet windy climate of the British uplands 15 to 20% of rainfall is lost by transpiration in grasslands whereas 30 to 40% is lost from forested areas. Many of the major rivers and their tributaries in Ireland rise in areas of blanket peat and coupled with the likelihood of receiving greater precipitation due to their higher elevation, any change in rainfall runoff response in these areas due to afforestation and associated drainage is likely to impact on the rainfall runoff characteristics of a larger catchment.

2.5 Rainfall-runoff model GEOtop

2.5.1 General

Rainfall runoff models have proved to be a vital tool in hydrology and provide solutions to many practical problems from flood forecasting, assessment of the impacts of effluents on water quality, design of engineered channels and many more. One of the primary drivers for the construction of hydrologic models is the limitation of hydrological measurements as models provide a means of extrapolating known measurements in both space and time to areas where data is not available (Beven, 2001). A review of the literature reveals a wide range of models from simple models such as that based on the unit hydrograph first introduced by Sherman (1932) to complex conceptual distributed catchment models e.g. REW (Reggiani et al., 2000), QUASAR (Whitehead et al., 1997), MIKE SHE (Abbott et al., 1986), TOPMODEL (Beven and Kirkby, 1979), GEOtop (Rigon et al., 2006) and SWAT (Arnold et al., 1998) to name just a few. The choice of model depends on the requirements of the user, as increasing model complexity is generally associated with an increase in cost in terms of data requirements, user input and computational power. It is important at the outset to select the correct model. Models such as SWAT and GEOtop, which simulate the soil-vegetation-atmospheric interactions (SVAT), may be more suited to assessing the impact of land management practices in large complex catchments whereas a model such as TOPMODEL, which bases its distributed predictions on an analysis of topography, might be more suited to assessing the loss of nutrients from a grassland as demonstrated by Scanlon et al. (2004).

2.5.2 GEOtop

The original version of GEOtop includes a rigorous treatment of the core hydrological processes (e.g. unsaturated flow, saturated flow, transport surface energy balances and stream flow generation/routing). The energy process has been extensively tested and validated by Bertoldi et al. (2006). These simulations show that both a more extended channel network and more accentuated slopes result in an increase in the discharge coupled with a decrease in the evapotranspiration. A reduction of the latent heat flux was balanced by an increase in the sensible heat flux. Net radiation also showed a minor sensitivity to topography while the evaporative fraction was shown to be strongly dependent on geomorphic characteristics.

Recently GEOtop has been extended to include treatment of shallow landslides (Simoni et al., 2008). This extension of GEOtop simulated the probability of occurrence of shallow landslides and debris flows. The landslide extension took advantage of the distributed hydrological data such as moisture content, matrix suction, water table depth, etc., all simulated by GEOtop which are fundamental to estimating the stress and/or changes of stress due to precipitation events within the soil matrix at various depths. The resulting model was calibrated in an alpine catchment in the Friuli region of Italy and then used to map future failure probabilities.

2.6 Soil erosion

2.6.1 Background

Soil erosion can result from wind forces or from precipitation induced runoff. However due to the wet climate in Ireland wind erosion is considered negligible (Favis-Mortlock, 2006). Some land use and management practices can lead to precipitation induced soil erosion, which in turn can deteriorate the remaining physical, chemical and biological soil properties and as a consequence reduces soil productivity. The study by Van Oost and Govers (2006) showed that tillage erosion rates can exceed 10 t ha^{-1} yr^{-1}, especially on fields with complex topography. Such rates are at least of the same order of magnitude as average water erosion rates reported for hilly cropland in western Europe. Cerdan et al. (2006) noted that land uses with the highest percentage of bare soil, either spatially (wide inter-row spacing and low leaf cover, e.g. vineyard or maize) or temporally (long inter-crop duration, e.g. maize or spring crop) have the highest soil erosion rates. Evans (1996) estimated that erosion significantly and adversely affects 40% of arable soils in the UK, with these soils losing more than 25% of their agricultural productivity. Grazhdani (2006) noted that poorly built logging roads in forestry operations lose soil by erosion of the road surface and the drainage ditches or the soil exposed by roads cut into hillsides. The study by Xu et al. (2006) demonstrated very high soil erosion rates (108.9 t ha^{-1} yr^{-1}) occurring on road sides slopes in China. Off-site impacts of erosion include sedimentation of rivers and lakes, watercourse pollution and eutrophication, silt build up in rivers with its consequent impact on young aquatic life, and perturbed geomorphological functions of river systems (Owens et al., 2005). While the dominant land use in Ireland is grassland, erosion in the form of particulate matter transports nutrients from the soil to the water courses (Scanlon et al., 2004). The lack of knowledge on soil erosion in the EU has been highlighted by Van-Camp et al. (2004). While studies have provided evidence of past soil erosion in Ireland (e.g. Huang and O'Connell (2000)) there do not appear to be any contemporary experimental field studies that have quantified erosion rates from different land uses in Ireland.

Global climate has changed notably over the past century; this change is expected to continue in the future. Many regions of the world have become drier, in some cases due to decreases in precipitation and in others because of increased evapotranspiration associated with increased temperatures, which also has significant implications for both wind and water induced erosion. In many areas the seasonal distributions of rainfall have changed, with significant implications for patterns of vegetation growth and hence for soil erosion (Nearing et al., 2005). Soil erosion prediction is necessary to fully understand spatial and temporal changes of soil erosion. Many models have been developed and applied for predicting soil erosion and these include empirical models such as RUSLE (Renard et al., 1997) and physically-based models such as LISEM (DeRoo et al., 1996).

Sediment loads in streams have been studied in Northern Ireland but from the perspective of water quality rather than their impact on soil quality (Evans et al., 2006). Related studies by Lewis (2003) measured suspended solids export (suspended sediment yield (SSY)) from a nested set of small grassland catchments in Dripsey, Cork where the SSY export ranged from 0.073 to 0.136 t ha^{-1} for 2002. Recent work (Harrington and Harrington, 2010) measured the SSY from a number of rivers in southern Ireland, as ranging from 0.15 to 0.25 t ha^{-1} yr^{-1}. During the EPA STRIDE Lee Valley Study (1993-1994) project, the total SSY exports were estimated from grassland agricultural land in Dripsey. These were based on continuous stream discharge measurements and an intensive water sampling programme. Measurements were made at three catchment scales (2.28 km^2, 14.91 km^2, 88 km^2) and annual exports of between 0.127 and 0.24 t ha^{-1} yr^{-1} were estimated (Tunney et al., 2000). While SSY is the "instream" deliverable of erosion, SSY may be considered a proxy for erosion, in the absence of field measurements.

Ito (2007) found that the area averaged rate of erosion (in RUSLE) was 9.1 -13 t ha^{-1} yr^{-1}, by comparison with a finding of 10.2 t ha^{-1} yr^{-1} by Lal (2003). Work by Quinton et al. (2006) over a 10 year period in experimental cultivated plots in the UK found the annual average rate of erosion to range from 0.42 to 1.9 t ha^{-1} yr^{-1}. Thus if the global average is estimated at ~10 t ha^{-1} yr^{-1}, then it can be expected that erosion from Irish grasslands are likely to be much smaller and of the order of < 1 t ha^{-1} yr^{-1}. Floods have been found to dominate erosion (Lopez-Tarazon et al., 2009) where, rainfall intensity, soil moisture and infiltration capacity (Romkens et al., 2002; VanDijk and Kwaad, 1996), have been identified as a central process to erosion rates.

2.6.2 Soil erosion modelling

To understand the spatial and temporal changes of soil erosion and to enable mitigation measures, modelling studies of erosion processes and their quantification are required. Many models have been developed and applied for predicting soil erosion and these include the simpler empirical and the more complex physically-based models. By far the most widely used long term annual estimates of erosion (soil loss) are the Universal Soil Loss Equation (USLE), the modified version (MUSLE), the revised version (RUSLE), and the Water Erosion Prediction Program (WEPP), (Kinnell, 2010; Lu et al., 2005). USLE has more recently been also used to predict event erosion (Kinnell, 2010). For modelling the spatial soil erosion risk, these models are often integrated with GIS and Geostatistics techniques (Ozcan et al., 2008). Also, many studies have used

RUSLE (Ito, 2007; Smith et al., 2007) and its related models (e.g. EPIC, see Izaurralde et al., 2007) to simulate the impacts of soil erosion and deposition on the carbon cycle. Each of the above empirical models have strengths and weaknesses and have been applied across a broad range of landscape types (Hancock et al., 2010; Kinnell, 2010).

A further modelling development has been the evolution of catchment scale models which use digital terrain models (DTM's) to detail the topography, soil type, etc. Amongst these, are SIBERIA and CAESAR as described by Hancock et al., (2010). The latter is used to quantify soil erosion rates and processes subject to the action of rainfall and runoff, and can estimate erosion and deposition as well as global erosion or sediment yield. Such models employ spatially variable hydrological and erosion parameters, the spatial distribution of soil type, particle size and predict erosion/deposition at the pixel scale and at the catchment scale. Other erosion models include SHETRAN (Ewen et al., 2000) which is a physically based model suited to the river basin scale. It was specifically designed to model transport of chemicals and sediment thorough various pathways on a continuous basis. SHETRAN also has a component to simulate shallow landslides triggered by groundwater fluctuations. A simpler model that also operates at the catchment scale is the Factorial Scoring Model (FSM) (Verstraeten et al., 2003). It predicts the sediment yield of a catchment, based on a nonlinear equation involving the catchment area, topography, vegetation, gullies, lithology and slope. The rule based STREAM model (Cerdan et al., 2002) is designed for areas that have a clear surface sealing process where a crust is formed on loose soil after tillage by rainfall. The MEFIDIS (Nunes et al., 2005) model was developed to simulate the consequences of climate and land-use changes for surface runoff and erosion patterns during extreme rainfall events. The model relies on physically based runoff and soil detachment equations, dividing the simulation area into spatially homogeneous units and using a dynamic approach for runoff and suspended sediment distribution. The LISEM model (DeRoo et al., 1996) is a physically based model that runs at the event and catchment scale. LISEM runs in a GIS environment and modelled erosion is comprised of splash detachment and flow detachment from over land flow in rills. The transport processes are also simulated with soil transport and deposition carried out on a cell by cell basis. Flow routing is modelled using a four-point finite difference solution of the kinematic wave and Manning’s equation.

Chapter 3

Materials and Methods

3.1 Measurements and estimation of mineral soil hydraulic properties

3.1.1 Selection of mineral site locations

For the purposes of site selection of hydraulic conductivity field test locations, it was decided that the focus of site work would be to collect field samples enabling the determination of the soil hydraulic properties of a range of soils, representing the land uses and geographical spread around Ireland. As texture (percent sand, silt, and clay) is the first measure in understanding soil hydraulic properties, it was decided that soil texture (rather than soil type) should be a key criteria for the site selection. In order to utilise as much as possible of the existing data, the aim of the site selection process was to select as many sites as possible from the National soils Database (NSD) 1310 points (EPA 2007) and the SoilC (Measurement and Modelling of Soil Carbon Stocks and Stock Changes in Irish Soils (Kiely et al., 2010) which are a subset (62 points) of the NSD sites). The soil type associations of Ireland described by Gardiner and Radford (1980) are considered to be the most comprehensive in the country. From these soil associations it was possible to estimate the percentage make up of Irish soils according to the USDA soil textural triangle (Table 3.1). Sites were then selected from the SoilC project to ensure that the sites in this study reflected the make up of Irish soils. As the SoilC project did not include any clay sites, two clay sites from the NDS were identified and chosen to bring the total number of sampling sites to 32. The locations of these sites are shown in Figure 3.1 and details on site land use and soil type are given in Table 3.2.

Table 3.1 Distribution of different texture classes in Ireland.

Texture Classifications	Irish Soils (%) (Gardiner and Radford, 1980)	No of SoilH samples	No of SoilC samples
Clay	3.7	2	
Silty clay	0		
Silty clay loam	3.3	1	1
Sandy clay	0		
Clay loam	17.81	7	7
Medium loam	38.9	11	18
Silty loam	0	1	1
Silt	0		
Sandy clay loam	0.5		1
Sand	0	1	1
Loamy sand	0		
Sandy loam	17.35	8	8
Total Mineral	81.63	32	38
Peat	18.37	1	21
TOTAL	100	32	59+

3.1.2 Mineral soils sampling parameters

While the earlier projects (NSD and SoilC), were focused on the physical make up and carbon and mineral contents, in this study (SoilH) the focus was primarily on the soil hydrological properties (e.g. hydraulic conductivity $K(\theta)$). The moisture retention $\Psi(\theta)$ and the theory and methods of the site and laboratory experiments are described in more detail below. Along with the hydraulic properties, soil samples were also taken for standard soil physical properties such as initial and saturated moisture content, particle size analysis and bulk density.

3.1.3 Hydrological properties sampling methods and theory

From a review of both in field and laboratory methods for determining hydraulic properties, it was decided that the BEST (Beerkan Estimation of Soil Transfer Parameters) method (Lassabatere et al., 2006; Minasny and McBratney, 2007) was most suited (following earlier discussions with Professor Cuenca of Oregon State

University, OR, USA). This method determines both the water retention curve and the hydraulic conductivity curve as defined by their shape and scale parameters.

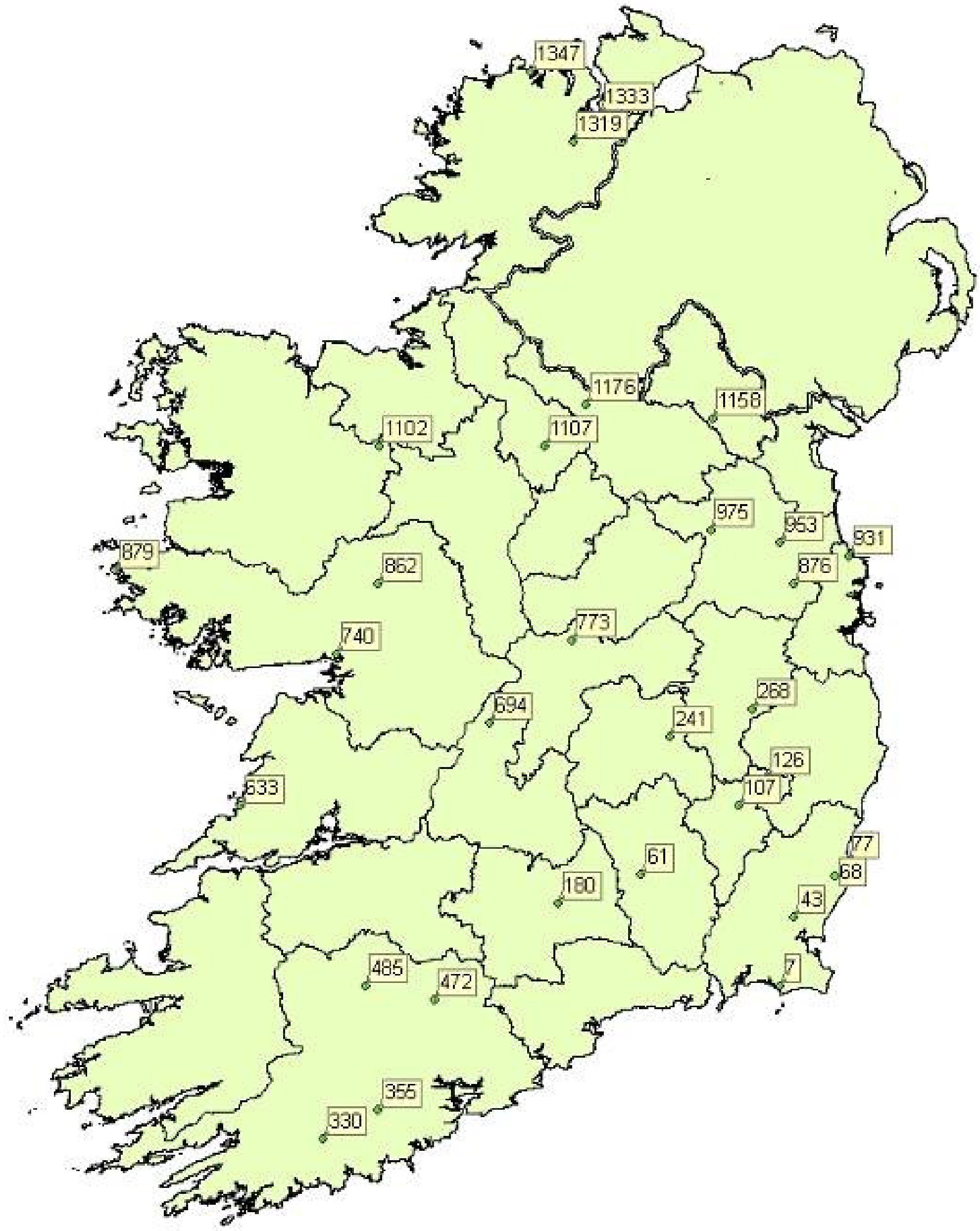

Figure 3.1 Location of all 31 mineral soil sampling sites. The coordinates, soil type and land use of each site is given in Table 3.2.

Table 3.2 Site location, land use and soil type.

Site No	County	northing ING	easting ING	Soil Type	Soil Triangle	Land use
633	Clare	170926	100997	Brown Podzolic	Clay loam	Pasture
330	Cork	50973	131063	Brown Podzolic	Medium loam	Arable
355	Cork	60995	151000	Brown Podzolic	Silty loam	Pasture
472	Cork	101069	171023	Acid Brown Earth	Sandy loam	Arable
485	Cork	106159	146069	Grey Brown Podzolic	Clay loam	Pasture
1176	Cavan	316000	225996	Gley	Sandy loam	Pasture
107	Carlow	170999	281001	Grey Brown Podzolic	Sandy loam	Arable
126	Carlow	181000	291003	Shallow Brown Earth	Sandy loam	Pasture
1319	Donegal	410953	221175	Grey Brown Podzolic	Sandy loam	Pasture
1333	Donegal	420984	231019	Brown Podzolic	Medium loam	Arable
1347	Donegal	436006	205994	Shallow Brown Earth	Medium loam	Pasture
931	Dublin	261008	320978	Gley	Medium loam	Arable
740	Galway	225998	135995	Shallow Brown Earth	Medium loam	Pasture
862	Galway	250998	150996	Grey Brown Podzolic	Medium loam	Pasture
879	Galway	256002	55975	Sand	Sand	Pasture
268	Kildare	206006	286001	Grey Brown Podzolic	Sandy loam	Pasture
61	Kilkenny	146014	245992	Grey Brown Podzolic	Clay loam	Arable
1107	Leitrim	300981	211000	Gley	Clay loam	Rough
241	Laois	195998	256079	Gley	Sandy loam	Forest
876	Meath	251000	301000	Grey Brown Podzolic	Medium loam	Pasture
953	Meath	265998	296002	Grey Brown Podzolic	Clay loam	Forest
975	Meath	270712	270939	Grey Brown Podzolic	Medium loam	Arable
1158	Monaghan	310969	271500	Acid Brown Earth	Medium loam	Pasture
1102	Mayo	300987	151013	Podzol	Sandy loam	Pasture
773	Offaly	230997	220949	Grey Brown Podzolic	Clay loam	Pasture
694	Tipperary	201006	190999	Grey Brown Podzolic	Silty clay loam	Forest

Site No	County	northing ING	easting ING	Soil Type	Soil Triangle	Land use
180	Tipperary	136016	216003	Grey Brown Podzolic	Medium loam	Pasture
43	Wexford	131000	300999	Gley	Medium loam	Arable
68	Wexford	145600	315853	Gley	Clay loam	Pasture
7	Wexford	106269	295929	Clay	Clay	Pasture
77	Wexford	151609	320546	Clay	Clay	Pasture

With BEST, the shape parameters are estimated from particle size distribution analysis and scale parameters determined from infiltration experiments at null pressure head. Saturated moisture content is measured at the end of the infiltration test. Hydraulic conductivity and water pressure scale parameters are calculated from the steady state infiltration rate and prior estimation of sorptivity. The BEST method is based on the van Genuchten relationship for the water retention curve (eqn. 3.1a) (van Genuchten, 1980) with the Burdine condition (eqn 3.1b) (Burdine, 1953).

$$\frac{\theta-\theta_r}{\theta_s-\theta_r}=\left[1+(\alpha h)^n\right]^{-m} \qquad \text{(eqn 3.1a)}$$

$$m=1-\frac{2}{n} \qquad \text{(eqn 3.1b)}$$

$$\eta=\frac{2}{\lambda}+2+p \quad \text{with } \lambda=mn \qquad \text{(eqn 3.2)}$$

where n, m and η are shape parameters and α, θ_r and θ_s, are scale parameters. Usually, θ_r (%) is very low and thus considered to be zero. p (eqn 3.2) is a tortuosity parameter that depends on the chosen capillary model, and a value of 1 is used here following Burdine's condition (Braud et al., 2005; Burdine et al., 1953). n is calculated as proposed by Minasny and McBratney (2007) see eqn 3.3a:

$$n = 2.18 + 0.11\left[48.087 - 44.954S(x_1) - 1.023S(x_2) - 3.896S(x_3)\right] \quad \text{(eqn 3.3a)}$$

where

$$x_1 = 24.547 - 0.238sand - 0.082clay \quad \text{(eqn 3.3b)}$$

$$x_2 = -3.569 + 0.081sand \quad \text{(eqn 3.3c)}$$

$$x_3 = 0.694 - 0.024sand + 0.048clay \quad \text{(eqn 3.3d)}$$

$sand$ and $clay$ refer to sand and clay content (%, w/w)

BEST estimates K_s and parameters from infiltration experiments using a specific algorithm whose main characteristics are briefly described below, and the details can be found in Lassabatère et al. (2006). BEST is referred to as BEST/I or BEST/q according to the choice of time series: I (cumulative infiltration depth) or q (infiltration rate). This study only used the BEST/I method since Lassabatère et al. (2006) showed that BEST/I performed better than BEST/q. If an infiltration experiment with zero pressure on an r_d (mm) internal-radius circular surface above a uniform soil with a uniform initial water content is considered, the three-dimensional cumulative infiltration and infiltration rate can be approached by the explicit transient two-term equation (eqn 3.4a and steady-state expansion (eqn 3.4b).

$$I(t) = S\sqrt{t} + (AS^2 + BK_s)t \quad \text{(eqn 3.4a)}$$

$$q_s = AS^2 + K_s \quad \text{(eqn 3.4b)}$$

where constants A and B can be defined for the specific case of the Brooks and Corey relation (eqn 3.2) and taking into account initial conditions defined by Haverkamp et al. (1994).

$$A = \frac{\gamma}{r_d(\theta_s - \theta_0)} \quad \text{(eqn 3.5a)}$$

$$B = \frac{(2-\beta)}{3}\left[1 - \left(\frac{\theta_0}{\theta_s}\right)^{\eta}\right] + \left(\frac{\theta_0}{\theta_s}\right)^{\eta} \quad \text{(eqn 3.5b)}$$

where $\beta \approx 0.6$ and $\gamma \approx 0.75$, which apply for most soils when $\theta_0 < 0.25\theta_s$ (Haverkamp et al., 1994; Smettem et al., 1994).

BEST first estimates sorptivity by fitting the transient cumulative infiltration to the two-term equations, (eqn 3.4a). The fit is based on the replacement of hydraulic conductivity K_s by its sorptivity function S and the experimental apparent steady-state infiltration rate (q_s) through (eqn 3.4b) and the following conditions: an accurate reproduction of experimental data; a fit for S between zero and a maximum value that corresponds to a null hydraulic conductivity (capillary driven flow) and the use of restricted data subsets to ensure the validity of eqn 3.4a (Xu et al., 2009). Once sorptivity is estimated, the saturated hydraulic conductivity is obtained through eqn 3.4b, assuming that the steady state (apparent steady state) has been reached. MATLAB (MathWorks USA, 7.6.0, R2008a) was the software of choice for the estimation of the hydraulic parameters and the BEST method was coded in MATLAB by Prof. John Albertson of Duke University, NC, USA following the methodology described in Lassabatere et al. (2006) and Minasny and McBratney (2007).

Using the Beerkan field experiment datasets, BEST algorithm did not result in satisfactory hydraulic properties due to the relatively slow rates of infiltration at a number of sites and the high initial moisture content of other sites. Therefore, another algorithm called Wu method (Wu et al., 1999) was used in these cases.

The Wu method is based on the assumption that the following cumulative infiltration curve can be used to describe the infiltration process:

$$I = At + Bt^{0.5} \qquad \text{(eqn 3.6)}$$

where I is cumulative infiltration (mm), t is time (s), and A and B are empirical coefficients. B is equivalent to sorptivity (S) (mm s$^{-0.5}$). This equation is fitted to the (t, I) data pairs measured from the beginning of the single-ring infiltration experiments in order to obtain an estimate of A and B (S). Then K_s (mm s^{-1}) is calculated by the following equation (Wu et al., 1999):

$$k_s = \frac{\Delta\theta\left[\sqrt{(H+G^*)^2 + 4G^*C} - (H+G^*)\right]}{2T_c} \qquad \text{(eqn 3.7)}$$

where $\Delta\theta$ (%) is the difference between the saturated volumetric soil water content θ_s (%), and the initial volumetric soil water content θ_0 (%), H is ponded depth in the ring (mm), and G^*, C and T_c terms have the following expressions, respectively:

$$G^* = d + \frac{r}{2} \qquad \text{(eqn 3.8a)}$$

$$C = \frac{1}{4\Delta\theta}\left(\frac{B}{b}\right)^2 \frac{a}{A} \qquad \text{(eqn 3.8b)}$$

$$T_c = \frac{1}{4}\left(\frac{Ba}{bA}\right)^2 \qquad \text{(eqn 3.8c)}$$

where d (mm) and r (mm) are the insert depth of the ring and the ring radius, respectively and a and b are dimensionless constants where a=0.9084 and b=0.1682 (Wu et al., 1999). An estimate of the α (cm $^{-1}$) parameter can be obtained by the following relationship (Wu et al., 1999):

$$\alpha = \frac{K_s}{\phi_m} \approx \frac{K_s}{\phi_m^{'}} \qquad \text{(eqn 3.9)}$$

where ϕ_m is given by eqn 3.10a

$$\phi_m^{'} = \frac{K_s^2 T_c}{\Delta\theta} \qquad \text{(eqn 3.10a)}$$

3.1.4 Field Infiltration experiments

Field infiltration experiments were carried out at all 31 mineral soil sites at 3 different depths; surface, 15 cm and 30 cm. These field tests were carried out in accordance with recommendations of Prof. Richard Cuenca of Oregon State University OR, USA, who kindly provided us with the field experimental protocol. The field infiltration experiments were done by first carefully cutting the grass leaving the roots intact from a square area approximately 30*30 cm for the surface infiltration test and excavating a trench with minimal disturbance to the soil for the 15 and 30 cm depths (see Figure 3.2). At each depth a heavy duty plastic ring with a bevelled edge and diameter 15 cm was inserted approximately 1-2 cm into the ground with a rubber hammer taking care to minimally disturb the soil. A fixed volume of water (178 ml, corresponding to 10.07 mm of water depth) was then poured into the ring and the time taken for the water to completely infiltrate was recorded. Following this, a second measure of water (same volume as previously) was then poured in and the process was repeated until a steady state of infiltration was achieved. Soil samples were taken for bulk density and moisture content (initial and final) analysis. The initial moisture content of the soil was estimated by taking a soil sample before the infiltration experiment from outside the plastic ring, approximately 30 cm from the infiltration experiment. The final moisture content soil sample was taken from inside the ring after the infiltration experiment was complete and no standing water was remaining on the surface.

Figure 3.2 Infiltration experiment at site 180 with the infiltration experiment at surface in the foregound with the infiltration tests 15 and 30 cm below the surface in the trench behind the surface infiltration test. Replicate trenches were dug 2-3 m apart.

After the infiltration experiment the soil was removed to determine the penetration depth. Once this had been completed at all 3 levels, two more trenches were dug 2 - 3 m away from the first trench and the entire operation was repeated so as to have three replicates for each level. Once the field work (infiltration experiment) had been completed, the cumulative infiltration versus time was then plotted and knowing the pre and post experiment soil moisture, the BEST method was then used to determine K_{sat}. In the cases where the BEST method did not work due to the slow rate of infiltration (i.e. soil close to or at saturation), the Wu method (Wu et al., 1999) was used.

3.1.5 Bulk density, particle size analysis and moisture content.

Bulk density samples were taken at the surface, 15 and 30 cm depths for all three depths using Eijkelkamp ART NR07010253 stainless steel bulking density sampling rings, (volume 10 cc) (Eijkelkamp, Agrisearch Equipment BV, The Netherlands). The bulk density samples were taken before the infiltration experiments commenced at a distance of 20 to 25 cm away from the infiltration test locations so as to avoid disturbing the soil around the infiltration tests. Once the samples were taken they were sealed and transported to the laboratory, where the samples were oven dried at 105°C for 24 hours and sieved to 2 mm. Bulk density (ρ) (g cc^{-1}) was estimated from eqn 3.11.

$$\rho = \frac{M_d}{S_v - CFV} \qquad \text{(eqn 3.11)}$$

where M_d is the dry mass (g) of the sample <2 mm, S_v is the sample volume (cc) and CFV is >2mm coarse fraction volume (cc).

Gravimetric soil moisture (θ) (%) was estimated before and after the infiltration experiments. The samples taken for estimating bulk density were also used to estimate the initial moisture content and a second soil sample from within the infiltration ring was taken after all the infiltration tests were complete to determine the saturated moisture content. Gravimetric soil moisture was estimated from eqn 3.12.

$$\theta = \frac{Ms_{wet} - Ms_{dry}}{Ms_{dry}} \qquad \text{(eqn 3.12)}$$

where Ms_{wet} is the mass of the wet soil (g), and Ms_{dry} is the mass of the dry soil (g) after oven drying at 105°C for 24 hours.

The soil sample that was used to estimate the final moisture content was then analysed for particle size distribution. This analysis was carried out by Brookside Laboratories Inc. New Knoxville, OH, USA, using test S176, see www.blinc.com for further details.

3.2 Peat soils sampling method

3.2.1 Site description

The Glencar catchment is a pristine Atlantic blanket bog near Glencar in County Kerry, southwest Ireland (latitude 51 58′N, longitude 9 54′W) at an elevation of approximately 150 masl (meters above sea level) and is typical of Atlantic blanket bogs in the coastal regions of northwest Europe (Sottocornola et al., 2009). The depth of the bog varies from approximately 1.0 m at the margin (e.g. near the stream or road) of the bog to over 5 m in the bog interior. The water table is at or near the surface of the peat throughout the year (Sottocornola et al., 2009). A meteorological tower has been in existence at this site since 2002 and is run by the Hydromet group in U.C.C.; see section 3.3.2 for further details on the meteorological tower. The range of annual rainfall since 2002 was 2236 to 3365 mm with an estimated eddy covariance estimated evapotranspiration range of 369 to 424 mm and an average of 208 wet days (> 1 mm day^{-1}) per year. The average annual air temperature is 10.5°C. A small stream runs through the centre of the bog and drains approximately 76 ha, 85% of which is relatively intact blanket bog (Figure 3.3).

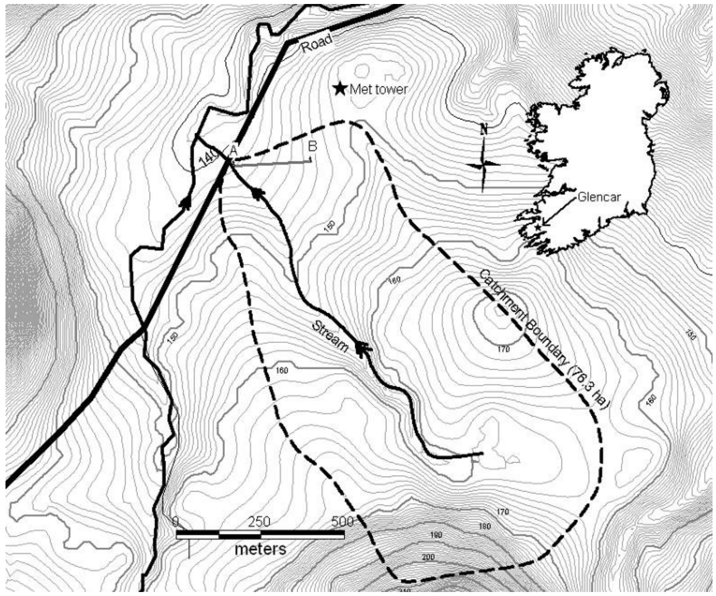

Figure 3.3 A map of the Glencar peatland catchment (76.3 ha) detailing the catchment boundry, the stream draining the catchment and the meteorological tower to the north of the catchment.

3.2.2 Hydraulic conductivity sampling laboratory analysis

Due to the unique nature of peatlands, the mineral soil sampling methods and analysis as previously described were unsuitable for peat soils, a method similar to that described by Beckwith et al. (2003a) was used. This involved extracting an undisturbed sample of peat from the field for laboratory analysis. Field work was carried out between November 2009 and January 2010. A total of 14 locations was chosen in a transect running perpendicular to the surface elevation contours from the stream indicated by section AB on Figure 3.3. A timber peg marked each point and distances between pegs varied from 2.5 m apart adjacent to the stream to 50 m apart at the bog interior at B in Figure 3.4.

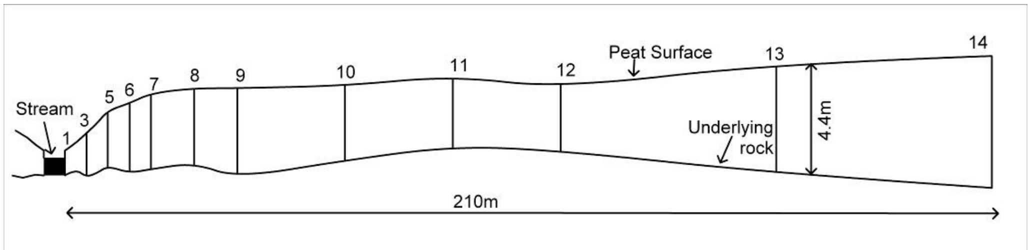

Figure 3.4 Cross section (AB) through bog. Details of peat depths at each peg given in Table 5.3.

Saturated hydraulic conductivity was estimated by employing a peat sampling method, which involved extracting a sample of peat from the field for later laboratory analysis. The peat was removed by hand using a selection of sharp cutting tools. Once the section of peat to be removed for sampling was identified, a narrow trench was dug around the perimeter of the sample. All cuts in the peat were made in long straight lines and care was taken not to damage the sample on extraction from the peat mass, (see Figure 3.5). Once the peat section (approximately 40 cm * 20 cm * 25 cm deep) was cut free from the bog on all sides, a waterproof box was lowered beside the peat and the sample was carefully transferred into the box. This box was then filled with bog water and returned to the laboratory. At each of the 14 locations, two samples were taken: one at a depth 0 to 25 cm and the second at 25 to 40 cm.

In the laboratory the peat samples for hydraulic conductivity were sliced so that three replicate cubes, of sides 10 cm were prepared (see Figure 3.6). The top 10 cm which contained the living plants and mosses was removed. A smooth-bladed knife was used in order to minimise disturbances to the samples by tearing of roots or peat fibres. The surface of each cube was dried with paper towels and then quickly dipped in molten paraffin wax and left to cool. This process was repeated until a thick covering (approximately 1cm) of wax was in place, taking care not to compress the samples at any stage. This method is similar to that of Beckwith et al., (2003a). Once the cubes were covered in wax, the top layer of wax was removed and a wood collar, 5 cm high was placed on the top of the wax surrounding the samples. The samples with collars in place were then sealed again in wax. Once the collar was deemed watertight and stable, the wax covering the bottom face of the cube was removed while it was still hot and easily cut (see Figure 3.7).

Figure 3.5 Excavatation of peat around the 40 * 20 * 25 cm peat sample selected for removal to the laboratory for saturated hydraulic conductivity analysis. Care was taken at all times to minimise the disturbance to the surrounding peat and in particular not to damage the section of peat selected for hydraulic conductivity analysis.

Figure 3.6 Three reciplate peat cubes of sides 10 cm cut from the large 25 *25 *10 cm section taken from the peatland. A 30 cm ruler is in the foreground for scale.

Hydraulic conductivity was determined in the laboratory using a constant head permeability test. First the samples were left to soak overnight in bog water. Filtered peat water was then placed in the reservoir enclosed by the timber collar (at the top of each sample) and the samples were then placed on a 2 mm sieve floor. To eliminate the possibility of water with a different chemical signature affecting the results, water collected from the peatland was used. A constant hydraulic head was maintained at all times by connecting the reservoir on the peat samples to a large reservoir of bog water with plastic tubing (diameter 10 mm) allowing a siphon to form and keeping both levels equal. Peat water then started to percolate through the peat sample and collected in the plastic container at the bottom, see Figure 3.8.

As the test progressed the peat water in the plastic container started to overflow via a section of small plastic tubing (diameter 10 mm) into a glass beaker. The entire apparatus was then left until the discharge became steady which varied from a few hours to days. Once the discharge became steady the flow through the sample (Q) was estimated from eqn 3.13.

$$Q = \frac{v}{t} \qquad \text{(eqn 3.13)}$$

where Q is the discharge (l s^{-1}); v is the volume of water which has percolated though the sample in a given time (m^3) and t is the time taken for the water to percolate (s).

Figure 3.7 Peat samples following waxing with top and bottom layers of wax removed. The samples were dipped in molten wax 15 to 20 times in order to build up a wall of wax approximately 1 cm thick. A timber collar also coated in wax is placed on top of the samples to act as a reservoir. These samples were then ready for saturated hydraulic conductivity tests.

The level of the reservoir was inspected on a regular basis with more peat water being added if any drop in water level was noticed. The air temperature was maintained at constant 18^0C as changes in temperature are known to affect hydraulic conductivity analysis (Surridge et al., 2005). The saturated hydraulic conductivity was calculated using Darcy’s Law; see (eqn 3.14).

$$K_{sat} = \frac{(Q * l)}{(A * \Delta h)} \qquad \text{(eqn 3.14)}$$

where K_{sat} is the saturated hydraulic conductivity (m s^{-1}); Q is the discharge (m^3 s^{-1}); A is the area of the face of the cube (m^2); l is the length of the sample (m); Δh is the difference in head between the water level at the top of the reservoir and the water in the plastic container at the bottom of the sample (m).

Figure 3.8 A peat sample in white paraffin wax set up for an infiltration test for the purposes of determining saturated hydraulic conductivity. Note the large reservoir of peat water (surface area 2400 cm^2) on the right connected to the reservoir of peat water on top of the peat sample (surface area ~ 121 cm^2) via plastic tubing (10 mm diameter) to maintain a constant hydraulic head on the peat samples. The water that has percolated though the peat is collected in the green plastic container which in turn overflows into the glass beaker. The volume of water that has percolated though the sample in a given time is then estimated by pouring the water from the beaker into the graduated cylinder on the left.

The peat samples were first analysed for vertical saturated hydraulic conductivity and then the entire apparatus was dismantled; the cube was rotated on the vertical axis by 90 degrees and the procedure was repeated again to determine the saturated conductivity in the horizontal direction. As the peatland is persistently saturated, only the saturated hydraulic conductivity of the peatland was considered.

3.2.3 Bulk density and moisture content site sampling methods and laboratory analysis

Peat samples for bulk density analysis were also taken. Due to the densely rooted nature of near surface peat, it was not possible to take bulk density samples at or near the surface with conventional bulk density rings. To overcome this problem, bulk density was obtained at the surface using sections of the samples taken for hydraulic conductivity analysis. These samples had a regular shape, which enabled estimates of bulk density. Below this an Eijkelkamp 04.09 peat sampler (Eijkelkamp, Agrisearch Equipment BV, The Netherlands) for bulk density analysis was used. Using this auger which has a semi-circular shape of diameter 5 cm, the full depth of the peat (in some cases as much as 5 m) was sampled in increments of 0.5 m deep (see Figure 3.9). These samples were placed in airtight bags for later laboratory analysis.

The samples for bulk density (below 50 cm) were oven dried for one week at 55 ºC. Samples were then weighed and re-weighed 24 hours later to ensure all moisture had evaporated. All the samples used in the analysis of hydraulic conductivity were also analysed for bulk density. Once the hydraulic conductivity tests had been completed the wax was removed from the samples and the length of each side of the cube of peat was measured so as to determine the volume and the samples were then dried, and bulk density was estimated using eqn 3.15.

Figure 3.9 A 50 cm section of peat removed from below the near surface of the bog using the Eijkelkamp 04.09 peat sampler.

$$\rho_{bd} = \frac{m_d}{V_{or}} = \frac{m_d}{l * h * w} \qquad \text{(eqn 3.15)}$$

where ρ_{bd} is the dry bulk density (g cm^{-3}); m_d is the dry mass of the sample (g); V_{or} is the original (wet) volume of the peat sample (cm^3); l is the length of the sample (cm); h is the height of the sample (cm) and w is the width of the sample (cm). The conventional gravimetric based definition of soil moisture (θ_G) as is used for mineral soils is defined as $\theta_G = M_w/M_s$, where M_w is the mass of water in the soil and M_s is the mass of soil. However, given the large proportion of water in peat and the relatively light mass of peat, the conventional definition of gravimetric soil moisture results in moisture values of order 10^4. Thus, peat moisture content was determined using eqn 3.16 as follows.

$$\theta = \frac{m_{tot} - m_d}{m_{tot}} * 100 \qquad \text{(eqn 3.16)}$$

where θ is the mass ratio based moisture content in %; m_{tot} and m_d are the total wet mass of peat (before drying) and the dry mass of the peat (after drying) respectively. Thus it was possible to estimate bulk density and moisture content for the entire profile of the peat. The Pearson correlation coefficient was used to investigate any significant trends in the variation of both bulk density and moisture content with depth. All statistical analyses as well as the calculations and graphical outputs were determined using MATLAB (Math works USA, 7.6.0, R2008a).

3.3 Rainfall runoff modelling

3.3.1 Catchments

Two catchments were chosen for a rainfall runoff modelling study. The Glencar peatland catchment (see Site description 3.2.1) and the Dripsey grassland catchment. The Dripsey catchment (Figure 3.10) is in southwest Ireland. It is a research catchment of 15 km^2, managed by the U.C.C. Hydromet team and is sited approximately 25 km northeast of Cork city (Latitude 51°59′N, Longitude 8°45′W). A small stream drains the catchment from north to south. It has an elevation range of 60 to 210 masl (meters above sea level). The climate is temperate maritime with mean annual air temperature of 10.2°C and an annual average rainfall of 1470 mm. Rainfall intensity is generally low with the highest rainfall intensity observed over the study period being <15 mm hr^{-1}. The land-cover is almost 100% grassland and is used for beef and dairying agriculture. The soils are gleys and podzols and are described as impeded drainage at the upper elevations to free drainage at the lower elevations.

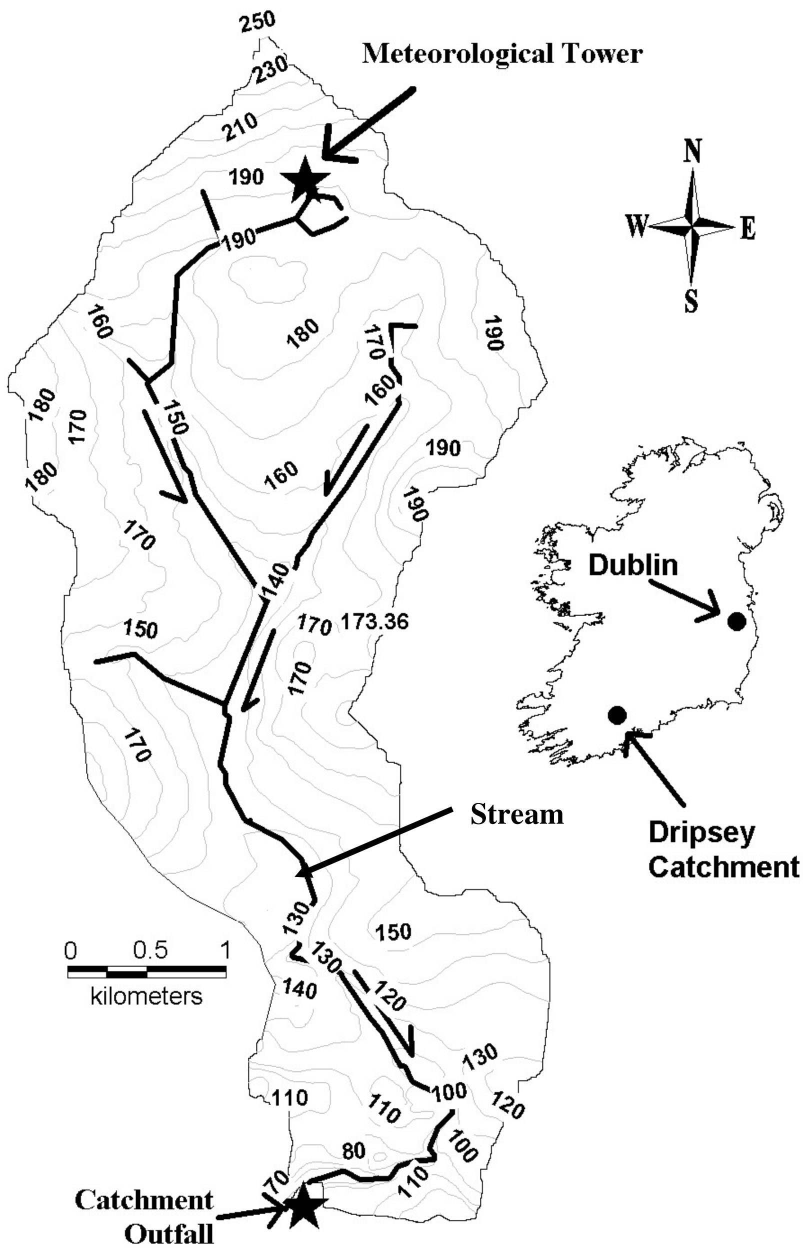

Figure 3.10 A map of the Dripsey catchment detailing catchment boundary, elevation, streams, catchment outfall and the location of the meteorological tower.

3.3.2 Meteorological measurements

In the Dripsey grassland catchment measurements of meteorological variables at a meteorological tower at the top of the catchment (elevation 190 masl) have been ongoing since 2001. They include: air temperature (T_a) and relative humidity (RH) (HMP45A; Vaisala, Helsinki, Finland); net radiation (CNRI net radiometer Kipp

& Zonen, Delft, The Netherlands); 2-dimensional wind speed (W_s) and direction (W_d) (RM Young). Rainfall was measured using a CS-ARG100 rain gauge. At the catchment outfall (elevation 60 masl), the stream height is continuously recorded from which stream flows are determined via a rating curve built up over several years.

In the Glencar blanket peatland catchment (Figure 3.3) a meteorological station was established in 2002 and includes two tipping bucket rain gauges (an ARG100, Environmental Measurements Ltd., UK and an Obsermet OMC-200, Observator BV, The Netherlands) and a WT level recorder which consists of a pressure transducer (PCDR1830, Campbell Scientific, UK) placed inside a metal tube, pierced all along its height. Wind speed was recorded with a 2-D sonic anemometer (WindSonic, Gill, UK). Air temperature (T_{air}) and relative humidity were measured at 2-m height with a shielded probe (HMP45C, Vaisala, Finland), while atmospheric pressure (P) was recorded with a barometer (PTB101B, Vaisala, Finland). An eddy-covariance system for CO_2 fluxes was also located on the same tower. It consisted of a 3-D sonic anemometer (Model 81000, R.M. Young Company, USA) and an open-path infrared gas analyzer for H_2O and CO_2 concentrations (LI-7500, LI-COR, USA) mounted 3 m above the vegetation.

3.4 GEOtop

3.4.1 Background

GEOtop (Rigon et al., 2006) is a distributed hydrological model and simulates the complete hydrological balance in a continuous way during a whole year and is driven by geospatial data (e.g. topography, soil type, vegetation, land cover). It estimates rainfall-runoff, evapotranspiration and provides spatially distributed outputs as well as routing water and sediment flows through stream and river networks (Rigon et al., 2006). GEOtop requires a digital elevation model (DEM), land cover data (including crop height, Leaf Area Index, root depth), soil type data (including vertical and horizontal hydraulic conductivity, depth and shape parameters α and n) in distributed maps for the catchment. Meteorological data such as precipitation, temperature, incoming shortwave radiation, air pressure, relative humidity, wind speed and direction in hourly time steps from one or more points in or near the catchment are also required. GEOtop outputs all major hydrological properties in hourly time steps. Stream flow is provided at the catchment outfall whereas outputs such as temperature, soil moisture, depth of water over soil, evaporation from the soil, transpiration from the canopy, water stored in the canopy, water table and snow depth are all provided in distributed maps suitable for import into a GIS environment such as ArcMap (ESRI® ArcMap™ 9.2. ESRI Inc USA) or GRASS (GRASS 6.4.1 © 1999-2011 GRASS).

GEOtop uses digital elevation models (DEMs) and divides the catchment into cells or pixels. Figure 3.11 illustrates how the Dripsey catchment is split up into cells and how the flow direction of each cell is determined from the digital elevation model. For every cell the model solves both the energy and water balance equations. The GEOtop model is built on an open-source programming framework, which makes it well suited for adaptation and extension.

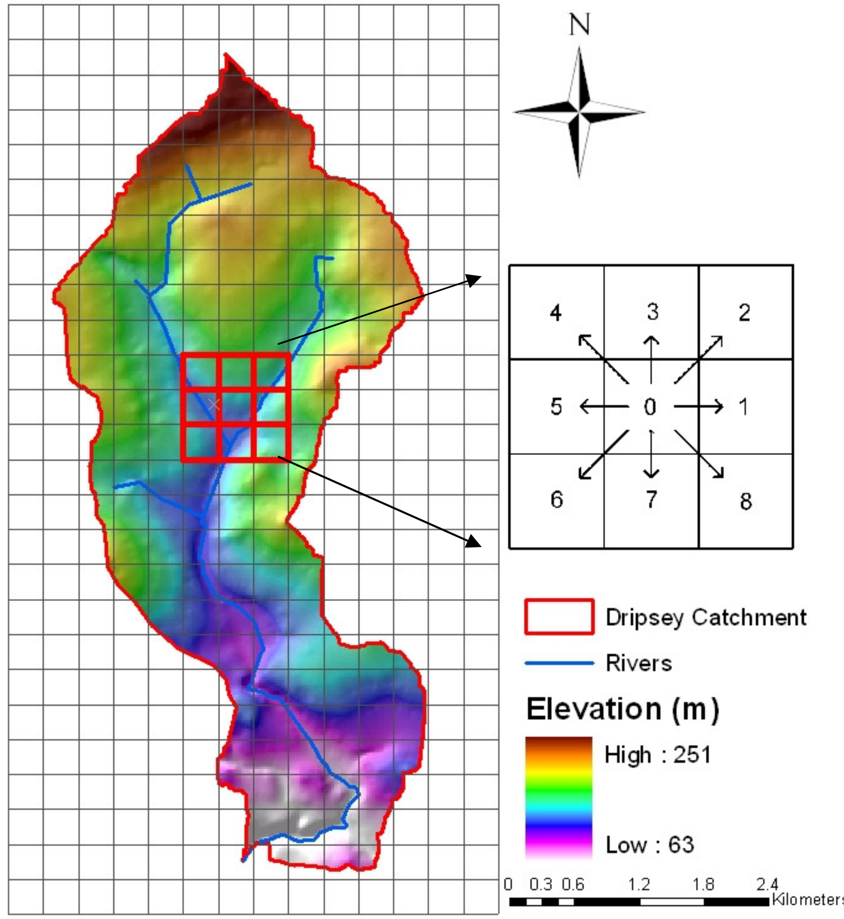

Figure 3.11 Dripsey catchment DEM and cell outline with D8 topology indicating surface flow direction.

3.4.2 GEOtop energy balance

The GEOtop energy balance and water balance (section 3.4.3) is described in detail in Bertoldi et al. (2004) and Rigon et al. (2006). The model calculates in an explicit way the energy balance as a function of the soil temperature and by numerically solving eqn 3.17.

$$G = R_n - H - \lambda ET - \frac{\Delta E}{\Delta t} \quad \text{(eqn 3.17)}$$

Where R_n is the net radiation (W m^{-2}), G the soil heat flux (W m^{-2}) and E the internal energy of the surface layer of vegetation (W m^{-2}) with H the sensible heat flux (W m^{-2}), λ is the water evaporation latent heat (J kg^{-1}) and ET the evapotranspiration defined by eqns 3.21 and 3.23 respectively. R_n is given by eqn 3.18

$$R_n = R_{\downarrow sw}(1-\alpha) + \varepsilon_a R_{\downarrow lw} - \varepsilon_s \sigma T_s^4 \qquad \text{(eqn 3.18)}$$

where $R\downarrow_{sw}$ and $R\downarrow_{lw}$ are measured incoming shortwave and longwave radiation (W m^{-2}) respectively, α is albedo of the surface (-) ε_a is the soil longwave emissivity (-), σ = 5.6704 $*10^{-8}$ (W m^{-2} K^4) is the Stefan-Boltzman constant and T_s is the soil surface temperature (°C). The sensible heat flux (H) is determined by eqn 3.19.

$$H = \rho c_p C_h u (T_s - T_a) \qquad \text{(eqn 3.19)}$$

where T_a is the air temperature at the measurement height (°C), ρ is air density (g cm^3), u is the mean wind speed velocity (m s^{-1}), c_p is air specific heat (J g^{-1} K^{-1}) and C_h the heat transport (J t^{-1}).

Evaporation is estimated as a function of the potential evapotranspiration (E_P) which is calculated as

$$E_p = \lambda(q*(T_s) - q*(T_a)u)/r_a \qquad \text{(eqn 3.20)}$$

where $q{*}T_s$ is the air saturation specific humidity at the surface (%), $q{*}T_a$ is the air saturation specific humidity at the temperature measurement height (%). The specific humidity is the rate between the water vapour mass and the humid air mass. λ is the latent heat of evapotranspiration (J kg^{-1}) and is expressed as a linear function of temperature see eqn 3.21.

$$\lambda = 2501000 + (2406000 - 2501000)\frac{T_a - 273.15}{40} \qquad \text{(eqn 3.21)}$$

The air resistance r_a is expressed as:

$$r_a = \frac{1}{\rho C_E U} \qquad \text{(eqn 3.22)}$$

where the bulk coefficient C_E (J t^{-1}) is equal to the heat coefficient C_H (J t^{-1}).

Evapotranspiration (*ET*) (mm) can then be derived from eqn 3.23.

$$ET = xE_P \qquad \text{(eqn 3.23)}$$

where x is given by eqn 3.24

$$x = \min(1; 0.75/sat) \qquad \text{(eqn.3.24)}$$

where *sat* is the percentage (%) of the entire soil column that is saturated.

Effects on radiation due to topography are also taken into account: shadowing, reduction of the sky view factor and variation in net radiation as a result of aspect and slope. This is achieved by calculating the shadows of the reliefs given by the DEM, by using an algorithm where only convex cells can create shadow on the other cells.

Evaporation from the soil and vegetation is estimated from eqn 3.25a, eqn 3.25b and eqn 3.25c, where E_g (mm) is evaporation from bare soil, E_{tc} (mm) is transpiration of canopy and E_{vc} (mm) is the evaporation from wet vegetation with f_c representing crop fraction (-) of a cell.

$$E_g = (1 - f_c)\frac{r_a}{r_a + r_s}Ep \qquad \text{(eqn 3.25a)}$$

$$E_{tc} = f_c \frac{r_a}{r_a + r_c} Ep\delta_w \qquad \text{(eqn 3.25b)}$$

$$E_{vc} = f_c E_p \delta_w \qquad \text{(eqn 3.25c)}$$

where the soil resistance (r_s) is defined as:

$$r_s = r_a \left(\frac{1.0 - (\theta_1 - \theta_r)/(\theta_s - \theta_r)}{(\theta_1 - \theta_r)/(\theta_s - \theta_r)} \right) \qquad \text{(eqn 3.26a)}$$

where θ_1 is soil moisture content of the top soil layer (%), θ_r is the residual soil content (%), θ_s is the saturated soil content (%) and the wet fraction δ_w is defined as:

$$\delta_w = \min\left(1, \left(\frac{w_r}{w_{r\max}} \right)^{\left(\frac{2}{3}\right)} \right) \qquad \text{(eqn 3.26b)}$$

where $w_{r\ max}$ is the maximum precipitation interception value (4 mm) of a canopy and w_r is the precipitation stored in the canopy.

Canopy resistance (r_c) is defined as:

$$r_c = \min(10^{12}, \frac{60}{f_s f_e f_t f_M}) \qquad \text{(eqn 27a)}$$

where f_s (w m^{-2}) is solar radiation dependence defined as:

$$f_s = \frac{1.25 R_{\downarrow sw}}{R_{\downarrow sw} + 250} \qquad \text{(eqn 27b)}$$

where f_e is vapour pressure deficit dependence defined as:

$$f_e = 1 - \frac{e_s(T_s)x - e_s(T_a)U_r}{40} \qquad \text{(eqn 27c)}$$

with U_r being defined as relative humidity (%)

Temperature dependence (°C) is defined as:

$$f_t = \frac{(T_a - 0)(50 - T_a)}{625} \qquad \text{(eqn 27d)}$$

if T_a is >50°C or < 0°C, $f_t = 10^{-12}$

and $f_M = 0$ if *sat* > 0.5 and 2 if sat < 0.5

3.4.3 GEOtop water balance

The water balance is illustrated by Figure 3.12 where precipitation (*PR*) is partitioned into evapotranspiration (*ET*), sub-surface runoff (Q_{sub}) (m^3 s^{-1}) and overland flow (Q_{sur}) (m^3 s^{-1}). Precipitation data is fed into the model from observed rainfall records from either single or multiple rain gauges or radar observations. If more than one rain gauge is used, GEOtop uses spatial interpolation between the rain gauges to develop the rainfall pattern over the catchment. The soil is split into a number of layers with varying thickness and hydrological properties allowing the subsurface flow rates to be varied not only throughout the catchment but also with depth.

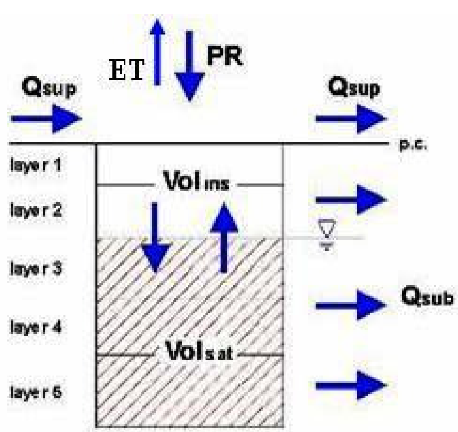

Figure 3.12 The GEOtop portrayal of the water budget in any given cell. (taken from a presentation by JD Albertson at an EPA-Soil H steering group meeting).

The flow of water in the unsaturated zone is estimated by numerically solving Richards equation, while the flow through the saturated zone is estimated from Darcy's law (see Figure 3.13). Surface flow, may be composed of either Horton overland flow or Dunne saturation overland flow (as illustrated in Figure 3.13). Horton overland flow is generated when the rainfall intensity exceeds the infiltration capacity of the surface of a soil whereas Dunne saturation excess flow is generated when the top layer of soil becomes saturated and any further direct precipitation on these soils results in overland flow. This overland flow is routed to the river channel using the 8 cell drainage directions. The flow along the river channels to the catchment outlet is described with a solution of the Saint-Venant parabolic equation.

$$Q(t)=\int_0^t\int_0^L \frac{sW(\tau,s)}{\sqrt{4\pi D(t-\tau)^3}}\exp\left[-\frac{(s-u_c(t-\tau))^2}{4D(t-\tau)}\right]ds.d\tau \qquad \text{(eqn 3.28)}$$

where *Q(t)* is the discharge at the basin's outlet (m^3 s^{-1}), *W(t, s)* (m^3 s^{-1}) is the inflow of the water coming from the hillsides into the channel network at a distance *s* (m) from the outlet and at a time *τ (s)*, *u* is mean celerity (m s^{-1}), *D* a hydrodynamic dispersion coefficient (m^2 s^{-1}) and *L* the maximum distance from the outlet measured along the network (m).

3.4.4 Soil Erosion module

By way of background, we note that there are four essential forms of water erosion: 1) inter-rill erosion - the movement of soil by rain splash and its transport by this surface flow (DeRoo et al., 1996a); 2) rill erosion - erosion by concentrated flow in small rivulets (Boardman and Poesen, 2006); 3) gully erosion - erosion by runoff scouring large channels (e.g. deeper than 30cm) (Poesen et al., 2006) and 4)

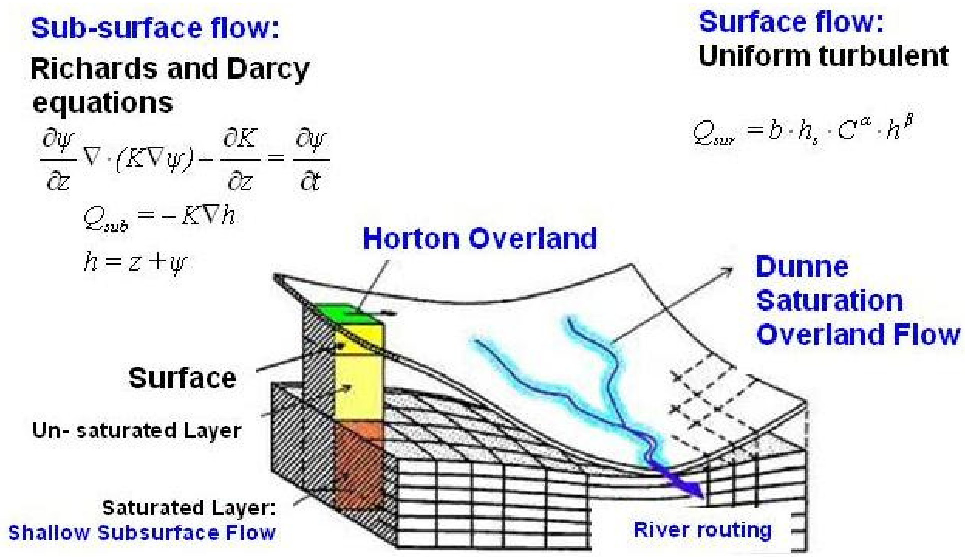

Figure 3.13 Simulation of flow routing through a catchment by GEOtop (taken from a presentation at an EPA-Soil H steering group meeting).

streambank erosion - erosion by rivers or streams cutting into banks (Prosser et al., 2000). Types 1 and 2 tend to occur on normal hillslopes, while types 3 and 4 occur in well developed channels. Since this study focuses on impacts to soil resources (and not necessarily channel integrity) only inter-rill and rill erosion were considered. Therefore only the effects of both rain splash detachment and flow detachment were considered. Splash detachment is the detachment of soil particles due to the impact of rain drops on soil and flow detachment is flow-induced detachment of soil particles from flow forming in small intermittent water gullies or rills over only a few centimetres of depth, see Figure 3.14 (Boardman and Poesen, 2006).

For the development of the erosion module in GEOtop, the LISEM model (DeRoo et al., 1996) was adopted and a new module was developed in GEOtop for the online calculation of distributed erosion, sediment transport, and deposition rates. The LISEM model (DeRoo et al., 1996) is a physical based erosion model that runs at the event and catchment scale. The original LISEM model runs in a GIS environment and modelled erosion is comprised of splash detachment and flow detachment from over land flow in rills. The transport processes are also simulated with soil transport and deposition carried out on a cell by cell basis. Flow routing is modelled using a four-point finite difference solution of the kinematic wave and Manning's equation. The author would like to acknowledge the assistance of Prof. John Albertson and Mr. Tan Zi of Duke University, North Carolina, USA, who developed the code for the new soil erosion module of GEOtop.

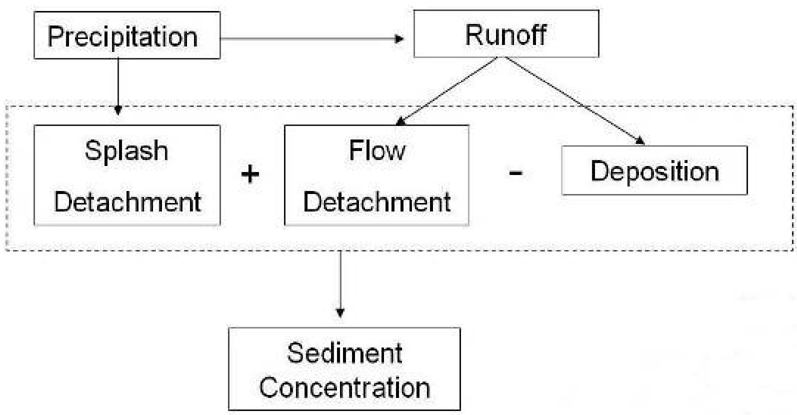

Figure 3.14. Flow chart of erosion (taken from a presentation by JD Albertson at an EPA-Soil H steering group meeting).

As GEOtop operates on a cell by cell basis the erosion module operates in the same way. Splash detachment and flow detachment both happen within the cell. The eroded material is transported from cell to cell by the movement of overland flow downhill. In some cases the eroded material will make its way to the river network and in others the eroded material from one cell may be deposited in cell down slope. Figure 3.15 and eqn 3.29 illustrate the movement of eroded material and the mass balance within a cell.

$$\frac{dM_c}{dt} = Q_{in} * C_{in} + D_s * dx * dy + D_f * dx - D_p * dx - Q_{out} C_{out} \quad \text{(eqn 3.29)}$$

where M_c is the mass of soil in the cell, Q_{in} is flow rate (m^3 s^{-1}) entering the cell, Q_{out} is flow rate (m^3 s^{-1}) leaving the cell, C_{in} is sediment concentration entering the cell (mg l^{-1}), C_{out} is sediment concentration leaving the cell (mg l^{-1}), D_f is flow detachment (kg m^{-3}), D_s is splash detachment (g s^{-1}), D_p is flow deposition (kg m^{-3}), dx is the width (m) of the cell and dy is the length (m) of the cell.

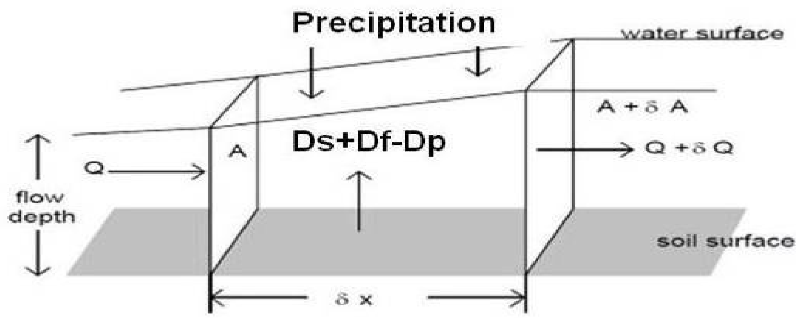

Figure 3.15 Erosion by splash detachment and flow detachment (taken from a presentation by JD Albertson at an EPA-Soil H steering group meeting).

Splash Detachment is simulated as function of soil aggregate stability (*Aggrstab*), rainfall kinetic energy (*Ke*), the depth of the surface water layer and net precipitation, see eqn 3.30.

$$D_s = \left(\frac{2.82}{Aggrstab} Ke * e^{-1.48(h)} + 2.96 \right) P_{net} \left(\frac{dx^2}{dt} \right) \qquad \text{(eqn 3.30)}$$

where D_s is splash detachment (g s^{-1}), h is the depth of the surface water layer (mm), P_{net} is net rainfall which is rainfall less interception (mm), *Aggrstab* is the aggregate stability (median number of drops) taken as 5 for the Dripsey catchment, dx is the size of an element (m); dt is the time increment (s) and the rainfall kinetic energy (J m^2) of the rain drops is given by eqn 3.31

$$Ke = 8.95 + 8.44 \log(P_{net}) \qquad \text{(eqn 3.31)}$$

Flow detachment (D_f) (kg m^{-3}) and *flow deposition* (D_p) (kg m^{-3}) are determined from eqn 3.32 and eqn 3.33.

$$D_f = (T_c - C) y v_s w \qquad \text{(eqn 3.32)}$$

$$D_p = v_s w (T_c - C) \qquad \text{(eqn 3.33)}$$

where w is rill width of flow (m), v_s is the settling velocity of particles (m s^{-1}), *transport capacity* (T_c) is defined by eqn 3.35 (kg m^3) and y, an efficiency coefficient is dependent on grain shear velocity and cohesion of the soil (Morgan et al., 1992; Rauws and Govers, 1988).

$$y = \frac{\mu_{g\min}}{\mu_{gcrit}} = \frac{1}{0.89 + 0.56COH} \qquad \text{(eqn 3.34)}$$

where μ_{gmin} is the minimum value required for critical grain shear velocity (cm s^{-1}); μ_{gcrit} is the critical grain shear velocity for rill initiation (cm^{-1}); and COH is the cohesion of the soil at saturation (kPa). The transport capacity (T_c) is dependent on Cl and Dl which are empirically derived coefficients estimated from D50 (Govers, 1990) see eqn 3.35 and Table 3.3.

$$T_s = c_l (S - 0.4)^{Dl} \rho_s \qquad \text{(eqn 3.35)}$$

here S (m m^{-1}) is the slope gradient (m m^{-1}) and ρ_s is the soil density (kg m^3).

When the GEOtop erosion module is run with GEOtop, GEOtop provides the erosion module with all the necessary hydrological parameters such as precipitation, depth of over land flow, etc. to estimate splash detachment, flow detachment, deposition and transport capacity for each cell. Sediment in the overland flow is then transported from cell to cell with overland flow. If the sediment concentration, C, (on the downslope cell) is less than T_C, then the rill erosion is computed from the flow detachment rate, (and includes effects of vegetation). If, instead, C is greater than T_C, then deposition occurs. Finally, a mass balance equation is integrated on each grid cell in each time state to update the local sediment concentrations and the local soil status. The sediment is routed along the flow and ultimately either deposited in low sloped regions or removed by streamflow.

Table 3.3 Coefficients Cl and Dl derived from D50, taken from (Govers, 1990)

D50	*Cl*	*Dl*
50	0.063	0.56
100	0.038	0.75
120	0.033	0.77
150	0.027	0.82
200	0.022	0.89
250	0.017	0.96

During testing of the erosion module a number of issues were brought to light. Overland flow from either saturation excess or infiltration excess is assumed to spread over the entire cell by GEOtop. Due to computation restraints it was necessary to limit the grid size so as to make runtimes practical which resulted in a cell size of 200 m by 200 m for most simulations. As the overland flow occurred over the full grid this led to very slow shallow flows and resulted in very low erosion rates. This required a modification to the GEOtop code. While overland flow can be generated over the entire cell in GEOtop, the overland flow once generated was automatically transferred to a rill of a width of w (m) to generate a more realistic depth of overland flow (h) (m), (see Figure 3.16) which in turn gave realistic sediment erosion and transport rates.

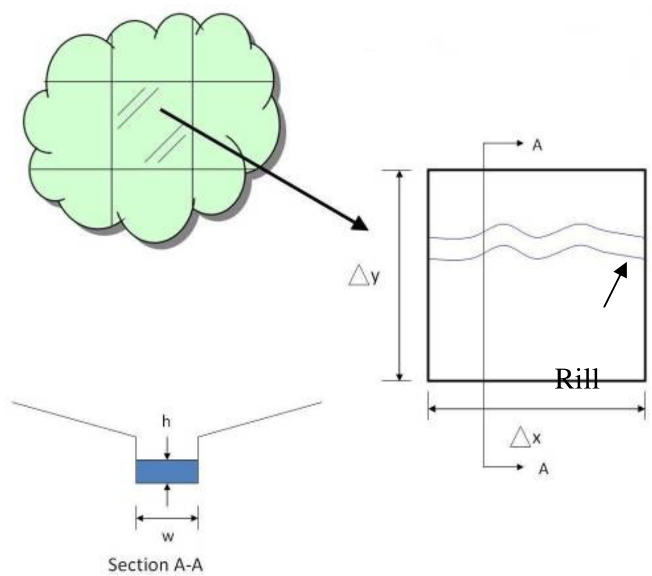

Figure 3.16 Illustration of the conversion of overland flow from an entire cell (Δy*Δx) in GEOtop (in green) to to rill (w*Δx). This results in a larger more realistic depth of flow (h) in the rill (taken from a presentation by JD Albertson at an EPA-Soil H steering group meeting).

Chapter 4

Hydrological properties of mineral soils

4.1 Mineral site sampling

Infiltration tests were carried out on all the mineral sites described in Chapter 3 over the summer months of 2008 and 2009. While the infiltration tests generally took 8 to 10 hours to complete it was necessary to spread the work over two summers. This was due to the difficulty contacting some land owners to obtain permission to carry out the infiltration tests on their land. Furthermore, it was not easy to satisfy the requirement to have an unsaturated soil at the start of the experiment, as these two summers were unusually wet. In order to ensure an unsaturated soil, generally a number of dry days before the planned experiment were needed and due to the nature of the Irish climate this only left a small number of suitable periods to conduct the experiments. A number of sites proved difficult to carry out the experiment in full due to the stony nature of the soil where it was not possible to place the infiltration ring in the ground without leakage. It was not possible to get results from the lower depths of sites 1158 and 241 due to difficult ground conditions. A full description of all the infiltration tests at all the sites, and results can be found in Appendix A.

4.2 Particle size distribution, porosity and bulk density

The results of the particle size distribution analysis and bulk density are presented in Appendix A2, with Figure 4.1 containing a summary of the results from the particle size distribution analysis. From Figure 4.1 we can see that medium loam and sandy loam accounted for a high proportion of the soils analysed. This is to be expected given that between them medium loam and sandy loam soils account for over 56% of Irish soils (Table 3.1) and so confirms the national representivity of our samples.

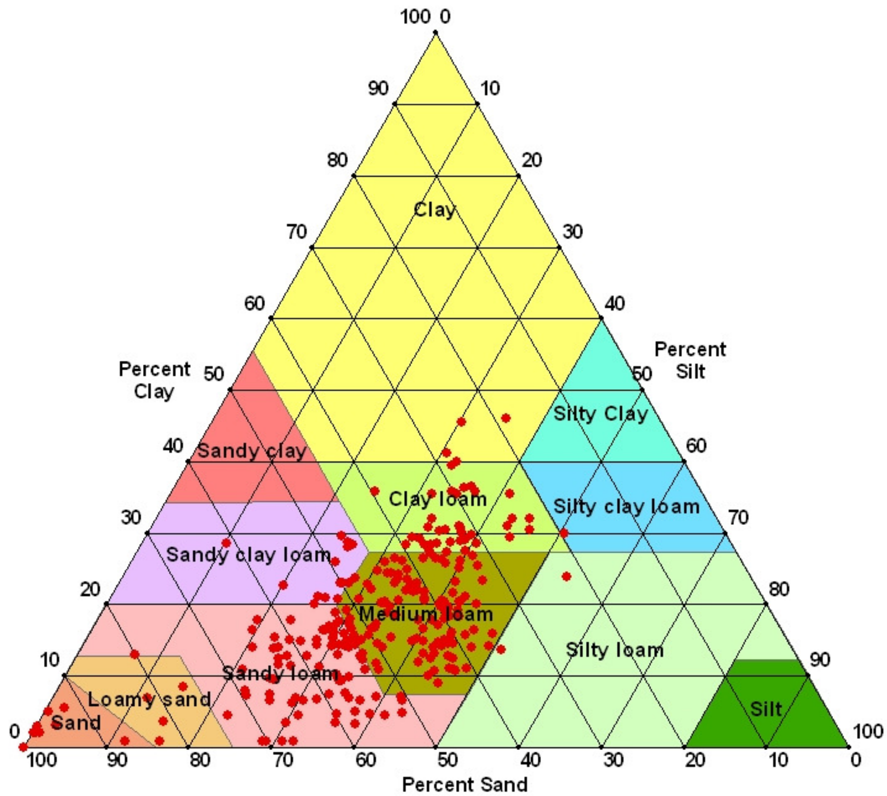

Figure 4.1 Particle size distribution from all 31 mineral sites on the USDA soil triangle. Note the concentration of sites in the medium-sandy loam region. Details of each mineral site are given in Appendix A2.

4.3 Mineral sites infiltration results

Following the infiltration experiments, the hydraulic parameters: K_S, saturated hydraulic conductivity; θ_S, saturated moisture content; and the van Genuchten (1980) parameters α, m and n were estimated for all sites. The van Genuchten (1980) equation (eqn 3.1a) parameters α, m and n are used to establish the water retention curve which is one of the most important features of soil. The water retention curve governs the conditions of plant growth, development and yield as well as the availability and uptake of nutrients and toxic substances by plant root systems, plant water stress, infiltration and drainage (Kern, 1995). The results of each infiltration experiment are given in Appendix A3.

4.4 Hydrological classification of Irish soils

4.4.1 Irish soil surveys and the IFS

One of the most comprehensive surveys of Irish soils resulted from the Strategic Plan for the Development of the Forestry Sector in Ireland (Dept. of Agriculture, 1996). This plan called for the development of a comprehensive inventory and planning system to provide forest resource, geographical and environmental data for management, control and planning purposes. As part of this plan a soil survey was to be conducted to assist in establishing forestry potential for planning and harvesting purposes. This soil survey subsequently became known as the Irish Forestry soils (IFS) database (produced from the project of Soils and Subsoils data generated by Teagasc with co-operation of the Forest Service, EPA and GSI, completed May 2006) (IFS, 2006).

Earlier Irish soil surveys included the General Soil Map of Ireland in 1969 (scale 1:575,000), where the soil association was the unit of mapping and a second edition of this map was published by Gardiner and Radford (Gardiner and Radford, 1980). This second edition of the Generalised Soil Map of Ireland (Gardiner and Radford, 1980) was published at 1:575,000 scale following a ten year period of preliminary reconnaissance, reconnaissance and detailed reconnaissance survey, the map represents a considerable improvement on the first Generalised Soil Map of Ireland, published in 1969 (Meehan, 2003). Detailed reconnaissance survey continued after publication but remains unpublished.

Detailed reconnaissance survey was suspended in 1988 leaving approximately 44% of the Republic of Ireland mapped to this level of detail (Meehan, 2003).

This necessitated the development of soil mapping for the remainder of the country at a similar scale for completion of the IFS. Due to the fact that the mapping project was to be completed within three years a methodology based on remote sensing and GIS from which soil type, productivity and distribution are modelled was developed (Fealy et al., 2006). The soil types being modelled fall into five broad classes; shallow mineral, deep mineral well drained, mineral poorly drained, peat over mineral and peat.

4.4.2 Hydrological classification of Irish soils

In order to build up a national classification of hydraulic properties of Irish soils, the results of the estimates of the hydraulic parameters from the 31 mineral sites were compared to the soil groups of the IFS soil database. The IFS soil database has 7 classes, (Table 4.1 and Figure 4.2).

Table 4.1 IFS soil classes and national coverage.

Soil Class (IFS soil class)	Class Code	Irish soils (%)
Deep well drained mineral	1	31.1
Shallow well drained mineral	2	9.31
Deep poorly drained mineral	3	20.36
Poorly drained mineral soils with peaty topsoil	4	3.3
Alluviums	5	3.55
Peats	6	29.1
Miscellaneous	7	3.28

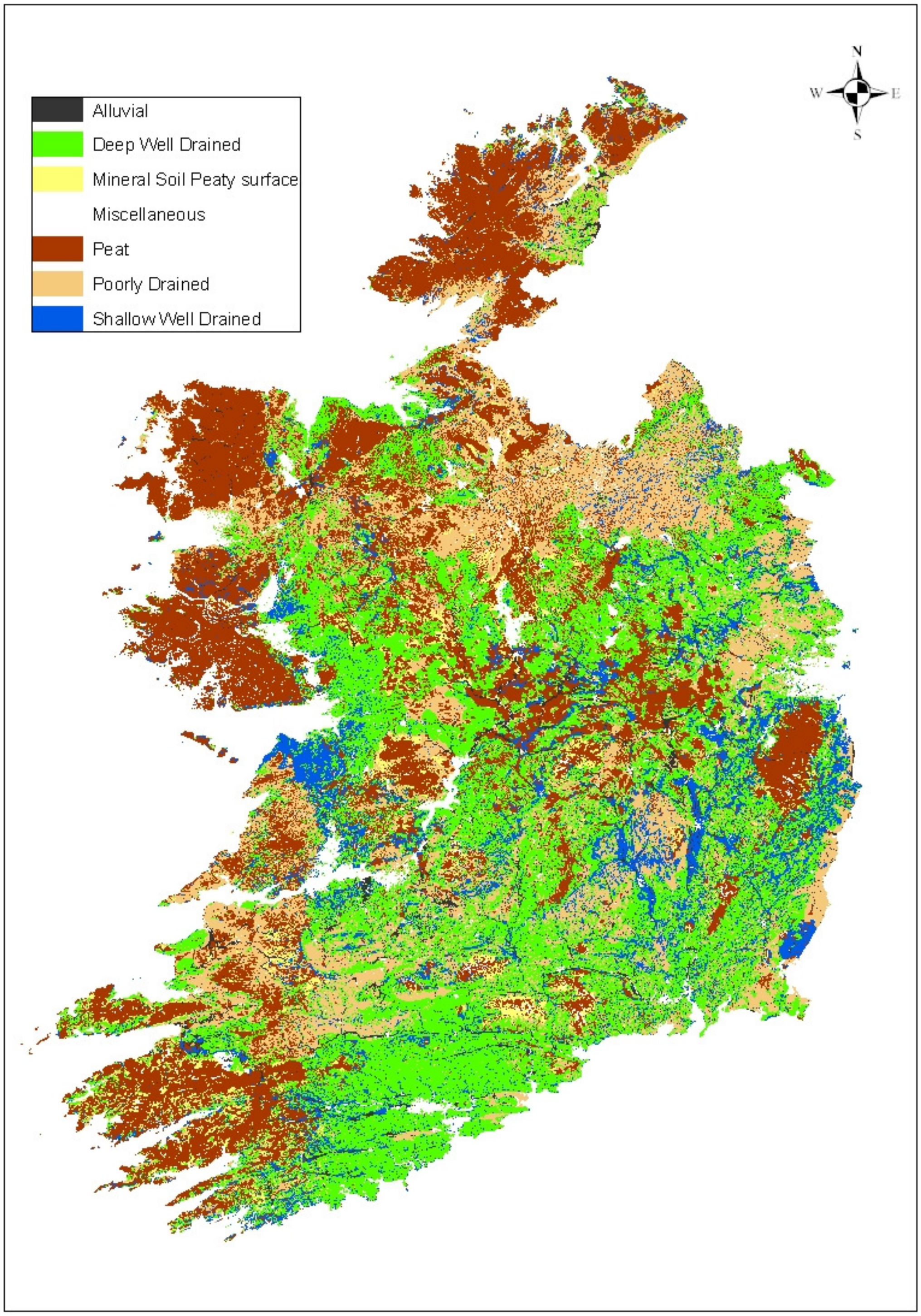

Figure 4.2 Distribution across Ireland of the IFS Soil Classes.

From the 31 mineral sites where we successfully completed the infiltration experiments, 16 sites are distributed across the first category of the IFS soil class (deep well drained mineral) which represents 31.1% of Irish soils. Our selected sites contained only one site in the second category (shallow well drained mineral) representing 9.31% of the country. Twelve sites are in the third category (deep poorly drained mineral) representing 20.36% of Irish soils. The fourth category (poorly drained mineral soils with peaty topsoil) and fifth category (alluviums) each representating just over 3% contain one site each.

We show in Table 4.2 that the IFS soil database captures the difference in the K_s and steady infiltration rate between well drained and poorly drained classes. Deep well drained mineral soils have the highest K_s (average 19.29; max 249; min 0.35; m $s^{-1}*10^{-6}$) with the K_s of poorly drained mineral sites two orders of magnitude lower (average 0.89; max 2.4; min 0.24 m $s^{-1}*10^{-6}$). Excluding the peat soils and alluvium, estimates of θ_S were between 0.36 and 0.46 (l l^{-1}).

The highest values of the van Genuchten (1980) parameter α were observed (0.16 cm^{-1}) in deep well drained mineral soils (0.16 cm^{-1}), with the lowest α in deep poorly drained mineral soils (0.06 cm^{-1}). The van Genuchten (1980) parameter n did not show as much variation with values ranging from 1.99 to 2.28. The author acknowledges the contribution of Dr. Xianli Xu in the construction of Table 4.2.

We also note that due to the high initial water content experienced in many of the sites visited as well as the low infiltration rates, there are relatively few points in the unsteady infiltration rates (sorptivity), see Appendix A1. These limited data points in the sorptive regime (unsaturated) hamper the estimation of sorptivity and as a consequence saturated hydraulic conductivity, as saturated hydraulic conductivity is estimated from sorptivity. A greater number of points in unsaturated infiltration would improve the BEST method. However, due to the nature of the Irish climate there were limited opportunities when the soil conditions would permit a large number of points in the sorptive regime.

Table 4.2 Soil hydrological properties for IFS soil classes.

Soil Class (IFS soil class)	Class Code	No. sites	n	θs (vol vol^{-1})	Ks (m s^{-1} 10^{-6})			α (cm^{-1})
					Mean	Max	Min	
Deep well drained mineral	1	16	2.28 ±0.28	0.46 ±0.17	19.29	249	0.35	0.16 ±0.029
Shallow well drained mineral	2	1	2.25	0.36	2.5	-	-	0.02
Deep poorly drained mineral	3	12	1.99 ±0.65	0.42 ±0.18	0.89	2.4	0.246	0.06 ±0.04
Poorly drained mineral soils with peaty topsoil	4	1	2.16	0.63	0.35	-	-	0.11
Alluviums	5	1	2.16	0.75	1.6	-	-	0.09
Peats	6	0	-	-			-	-
Miscellaneous	7	0	-	-			-	-

Chapter 5

Spatial variability of hydraulic conductivity and bulk density along a blanket peatland hillslope.

Running title: Hydraulic conductivity of a blanket peatland.

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Key words: Peatland, Saturated hydraulic conductivity, Bulk density, rainfall-runoff, Anisotropy

5.1 Abstract

This paper presents the results of a field investigation of saturated hydraulic conductivity K_{sat} and bulk density (ρ_b) in an Atlantic blanket bog in the southwest of Ireland. Starting at a peatland stream and moving along an uphill transect towards the peatland interior, ρ_b and K_{sat} were examined at regular intervals. Saturated horizontal hydraulic conductivity (K_{hsat}) and vertical (K_{vsat}) were estimated at two depths: 10 - 20 cm and 30 - 40 cm below the peat surface while ρb was estimated for the full profile. We consider two separate zones, one a riparian zone extending 10m from the stream and a second zone in the bog interior. We found that the K_{sat} was higher (~10^{-5} m s^{-1}) in the bog interior than the riparian zone (~10^{-6} m s^{-1}) while the converse applied to bulk density, with lowest density (~ 0.055 g cm^{-3}) at the interior and highest (~ 0.11 g cm^{-3}) at the riparian zone. In general, we found K_{hsat} to be approximately twice the K_{vsat}. These results support the idea that the lower K_{sat} at the margins control the hydrology of blanket peatlands. It is therefore important that the spatial variation of these two key properties be accommodated in hydrological models if the correct rainfall-runoff characteristics are to be correctly modelled. Streamflow analysis over 3 years at the peatland catchment outlet showed that the stream runoff was composed of 8% baseflow and 92% flood flow, suggesting that this blanket peatland is a source rather than a sink for floodwaters.

5.2 Introduction

Peatlands cover significant areas in northern latitudes and the existence of many peatlands is due to their unique hydrology (Clymo, 2004). Although northern peatlands cover just 3% of the global land surface, they have accumulated between 270 and 450 Pg of carbon, which represents 20 to 30% of the world's estimated soil carbon (Gorham, 1991; Laine et al., 2007b; Turunen et al., 2002), and their vast stocks of carbon are considered to be particularly vulnerable to climate change (Holden and Burt, 2002b; Laine et al., 2006; Oechel et al., 2000; Sottocornola and Kiely, 2010). From a regional perspective, peatlands cover 17% of the land area of the Republic of Ireland (Tomlinson, 2005) and are estimated to contain between 53 and 62% of the national soil carbon stock (Eaton et al., 2008; Tomlinson, 2005; Xu et al., 2011). Water table depth has been identified to be of critical importance to the health of peatlands with the possibility of peat undergoing total degradation under dewatering (Bragg and Tallis, 2001). Even a slight drop in water table is expected to affect the chemistry of the bog water, which will likely impact the vegetation composition and distribution (Sottocornola et al., 2009). It is important to maintain these unique landscapes not just because of their large store of carbon but also for biodiversity, as peatlands support a wide range of unique flora and fauna. As such it is important to be able to model the hydrology of peatlands not only for their current status but also under possible future scenarios of climate change which are predicted to result in reduced summer rainfall and increased winter rainfall in Ireland (McGrath and Lynch, 2008).

Knowledge of the physical and hydrological properties of soils is a prerequisite for rainfall-runoff modelling and hydrological studies (Albertson and Kiely, 2001; Herbst et al., 2006a). For mineral soils, there is a wealth of information on soil hydrological properties and their spatial variability (Brooks and Corey, 1964; Clapp and Hornberger, 1978; Comegna et al., 2010; Cosby et al., 1984; Montaldo et al., 2001; Nemes et al., 2001; Schaap et al., 2001; Tartakovsky et al., 1999; van Genuchten, 1980; Wosten et al., 1999) However, the same cannot be said for peatlands as there is limited knowledge of both hydrological properties and of elementary properties (such as bulk density) and particularly spatial variability (Beckwith et al., 2003a; Egglesmann et al., 1993; Holden and Burt, 2002a; Ingram, 1978; Kiely et al., 2010; Price, 2003; Surridge et al., 2005).

However, despite the importance of blanket peatland hydrology, few studies have carried out detailed hydraulic conductivity measurements (Holden, 2005a; Surridge et al., 2005) including the degree of anisotropy and the spatial variation of vertical and horizontal hydraulic conductivity (Beckwith et al., 2003b). To date efforts at estimating hydraulic conductivity in peatlands using different field and laboratory methods have resulted in a wide range of hydraulic conductivity values (Table 5.1) as low as ~ 10^{-8} m s^{-1}(Hoag and Price, 1995) to as high as ~ 10^{-2} m s^{-1} (Hogan et al., 2006). While some models of the earlier peatland hydrology used in the prediction of the shape of raised bogs (Ingram, 1982) assume that peatland hydraulic conductivity is homogeneous and isotropic, many others since have suggested that peatland hydraulic conductivity is neither homogeneous nor isotropic (Baird et al., 2008; Hoag and Price, 1995; Hogan et al., 2006; Holden and Burt, 2003; Kneale, 1987). Given this reported variable nature of peatland hydraulic conductivity a wide spread of values within the literature is to be expected. Many of the values reported from the literature are from a single plot or a series of plots scattered through a peatland from the margins to the centre of a bog. Lapen et al. (2005), from a sensitivity analysis of a groundwater model, suggested that areas of lower hydraulic conductivity exist at the margins than at the bog interior and the margins have a positive impact on bog formation retaining the elevated water table in the bog interior. This hypothesis was tested by Baird et al. (2008) at a raised bog site in Wales and found to be true but stated that their testing was confined only to one site and may not apply to other sites or other types of peatlands. Holden (2005a) in transects through blanket peatlands, partitioned the transect into footslope, midslope and topslope, where he found little difference in the bulk density and saturated hydraulic conductivity between the three hillslope parts. However, he noted that the average K was slightly higher in the midslopes, indicative of better and more uniform drainage through the midslope parts of the hillslope.

Table 5.1 Hydraulic conductivity (K) values reported from a selection of different studies in various peatlands.

Site	Peat type	Depth (m)	Method of analysis	K (m s^{-1})	Reference
Canada	Cut over	0.1	Piezometer	$1.73 * 10^{-6}$	(Schlotzhauer and Price, 1999)
Ireland	Raised	0.5 – 3	Piezometer	$10^{-8} – 10^{-2}$	(Kneale, 1987)
Sweden	Raised	0.25 – 2	Piezometer	$2*10^{-7} – 8*10^{-3}$	(Waddington and Roulet, 1997)
Scotland	Raised	0.1 – 7	Piezometer	$1*10^{-6} – 1.2*10^{-5}$	(Clymo, 2004)
England	Raised	0 - 0.15	MCM	$1*10^{-5} – 1.2*10^{-3}$	(Beckwith et al., 2003a)
USA	Raised	0.5 – 3	Piezometer	$2.5*10^{-6} – 2.6*10^{-4}$	(Chason and Siegel, 1986)
Canada	Raised	0.5 – 1	Piezometer	$1*10^{-8}$- $5*10^{-6}$	(Fraser et al., 2001)
Poland	Fen		Porous plate	$5.5 *10^{-8} – 5*10^{-6}$	(Gnatowski et al., 2010)
Canada	Fen	0 – 2	Piezometer	$1*10^{-6}$ - $9*10^{-3}$	(Hogan et al., 2006)
England	Fen	0 - 1.0	Piezometer and MCM	$1.1*10^{-4} – 1.6*10^{-3}$	(Surridge et al., 2005)
Canada	Fen	0 - 0.15	Piezometer	$10^{-7} – 10^{-4}$	(Kennedy and Price, 2005)
England	Fen	0.1	Piezometer	$6*10^{-7} – 6*10^{-6}$	(Baird and Gaffney, 2000)
England	Blanket	0.1 - .8	Piezometer	$1*10^{-5} – 1*10^{-7}$	(Holden and Burt, 2003)
Canada	Blanket	0.2 – .5	Piezometer	$1*10^{-8} – 1.6*10^{-2}$	(Hoag and Price, 1995)

The existence of areas of lower hydraulic conductivity at the margins of peatlands which maintain elevated water tables does, however, raise the question of how integral these bogs might be with the stream channels and whether they can provide the flood attenuation function which has often been attributed to wetlands (Evans and Warburton, 2007). However, rainfall runoff from upland blanket peatlands has been observed to be flashy with high flood peaks and short lag times (times of concentration) (Holden and Burt, 2002a). From the catchment perspective and considering the extensive nature of peatlands in the uplands of northern latitudes, where many rivers rise, improved knowledge of peatland hydrology is essential for catchment rainfall runoff modelling.

As with hydraulic conductivity there are also limited data available as to the variation of bulk density of different peat types. A number of studies of peatland bulk density have reported bulk density ranges from 0.05 to 0.254 g cm^{-3}, with Tomlinson and Davidson (2000), Kiely et al. (2010) and Wellock et al. (2011) reporting average bulk densities of 0.07, 0.17 and 0.13 g cm^{-3}. Wellock et al. (2011) in a study of 15 afforested peatland sites throughout Ireland found that the bulk density for lowland and high level blanket bogs were similar but that deeper peats (>2m) were found to have a lower density than shallower (<2 m) peats. Wellock et al. (2011) also found that the bulk density of basin peats was higher than that of blanket peats. There is some conflicting evidence as to the variation of bulk density with depth. Bulk density was shown to generally increase with depth in five peatlands in central and western Europe by Novak et al. (2008) whereas studies by Clymo (2004) and Tomlinson (2005), found no change in bulk density with depth. The studies by Kiely et al. (2010) and Wellock et al. (2011) found that bulk density did not change significantly with depth. Holden (2005a) in a survey of 160 British blanket peat sites noted bulk density ranges from approximately 0.05 g cm^{-3} near the surface to approximately 0.2 g cm^{-3} at depth.

As well as playing an important role in the health and carbon balance of a peatland, hydrological processes including surface flow, subsurface flow, buoyancy effects, precipitation timing and intensity, may also play a part in peatland stability. Peatland instability has been documented (Dykes and Kirk, 2006) with hydrological processes being fundamental in determining the spatial and temporal occurrence of peat slides (Warburton et al., 2004) which can be environmentally and geomorphologically significant (Dykes and Warburton, 2008). Failure mechanisms in peat, such as shear failure by loading, buoyancy effects, liquefaction and surface rupture all have hydrological controls but a better understanding of peat hydrology is required before realistic models can be developed to predict future failure (Warburton et al., 2004).

With the above motivation and observations in mind, the present study focused on the spatial variation of peatland hydraulic conductivities and bulk density in a pristine Atlantic blanket bog in southwest Ireland. The aims of the study were: (1) to measure the saturated vertical and horizontal hydraulic conductivities, bulk density and peat depth along a hillslope transect in a blanket bog from a stream at the peatland margins to the flat areas at the peatland centre; (2) to examine whether the observed variability (if any) of saturated hydraulic conductivity in blanket bogs matches the pattern of reducing hydraulic conductivity moving from the peatland interior to the exterior, as suggested by Lapen et al. (2005).

5.3 Materials and methods

5.3.1 Methodology

Methods for estimating peat hydraulic conductivity have been described by Gnatowski et al., 2010; Schlotzhauer and Price, 1999; and Surridge et al., 2005. A recent laboratory approach by Beckwith et al. (2003a), referred to as the modified cube (MC) method, has overcome perimeter leakage problems associated with permeameter tests (Surridge et al., 2005) by using paraffin wax to seal the peat samples and has been successfully used by Surridge et al. (2005). While both the piezometer and MC methods give consistent results, it is not possible to assess peat anisotropy using piezometer tests. Because of the need to capture anisotropy and the fact that the water table (WT) in our blanket peatland study site was observed by Sottocornola and Kiely (2010) to be within 17 cm of the surface across the peatland between 2003 and 2007 this study uses the above-referenced MC method. We also note that given the persistently saturated nature of the peatland we only consider saturated hydraulic conductivity.

5.3.2 Site description

The study site is a pristine Atlantic blanket bog near Glencar in County Kerry, southwest Ireland (Latitude: 51°58N, Longitude 9°54W) at an elevation of approximately 150 m (see Figure 5.1). The catchment area monitored for stream flow was 76 ha which was within a larger pristine bog of approximately 121 km^2. The peatland is typical of Atlantic blanket bogs in the coastal regions of northwest Europe in terms of both vegetation and water chemistry (Sottocornola et al., 2009). The water table can be observed at or near the surface of the peat. The depth of the bog varies from approximately 1.0 m at the stream margins to over 5 m in the bog centre. Sottocornola and Kiely (2010) at the same site found that the range of annual rainfall since 2002 was 2236 mm to 3365 mm with an estimated evapotranspiration (using eddy covariance methods) range of 369 mm to 424 mm. Koehler et al. (2009) found at the same site that there was an average of 208 wet days (defined as > 1 mm day^{-1}) per year with an average annual air temperature of 10.5°C. Between May and October the average air temperature was 13.3°C and between November and April it was 7.7°C.

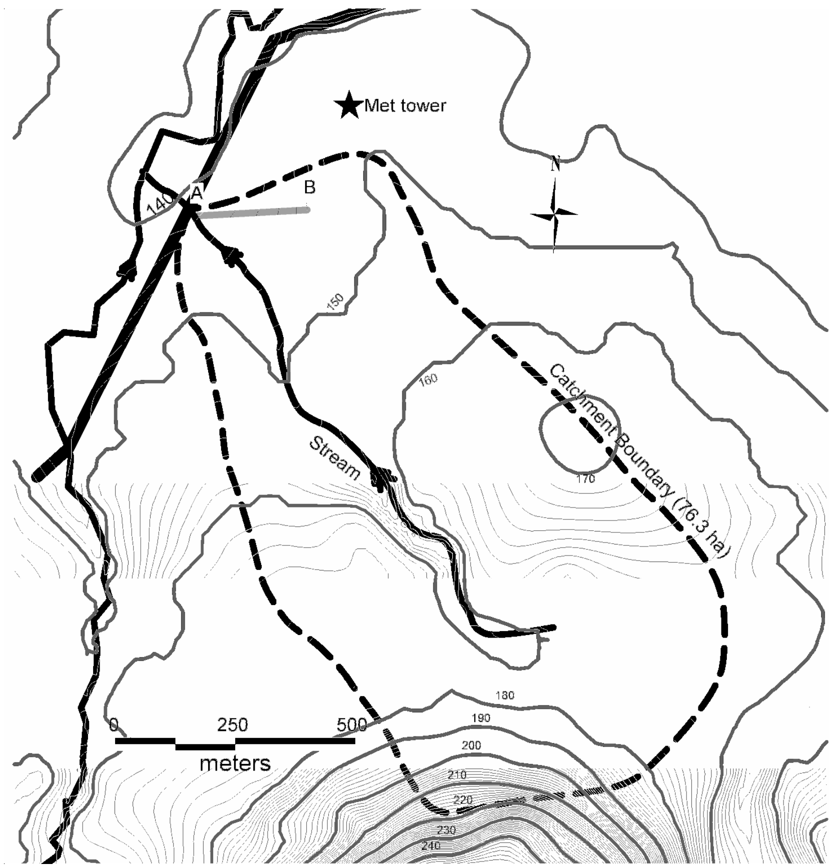

Figure 5.1 Glencar location and peatland site layout.

A small stream runs through the centre of the bog and drains approximately 76 ha, 85% of which is relatively intact blanket bog (see Figure 5.1). The recorded flow ranged between 0.015 and 10.0 l s^{-1} ha^{-1}. Since 2002, WT has been continuously recorded at the centre of the bog. The WT has remained within 17 cm of the surface with the seven year mean ~ 4 cm below the surface. A second temporary

WT recording station was established 50 m from the stream at the start of the field tests and was manually observed every two weeks. The WT at this station was observed to always be within 10 cm of the surface.

5.3.3 Peat samples

Field work was carried out between November 2009 and January 2010. A total of 14 locations were chosen in a transect running perpendicular to the surface elevation contours from the stream indicated by section AB on Figure 5.1. A timber peg marked each point and distances between pegs varied from 2.5 m apart adjacent to the stream to 50 m apart at the bog centre. See Figure 5.2 and Table 5.3 for details of the hillslope transect.

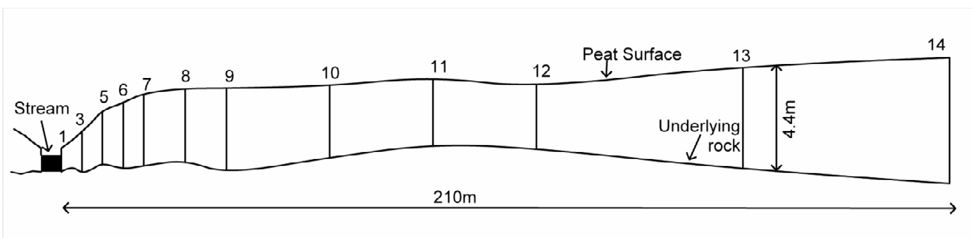

Figure 5.2 Cross section (AB) through bog.

Saturated hydraulic conductivity was estimated by employing a peat sampling method, which involved extracting a sample of peat from the field for later laboratory analysis. The peat was removed by hand using a selection of sharp cutting tools. Once the section of peat to be removed for sampling was identified, a narrow trench was dug around the perimeter of the sample. All cuts in the peat were made in long straight lines and care was taken not to damage the sample on extraction from the peat mass. Once the peat section (approximately 40 cm * 20 cm * 25 cm deep) was cut free from the bog on all sides, a waterproof box was lowered beside the peat and the sample was carefully transferred into the box. This box was then filled with bog water and returned to the laboratory. At each of the 14 locations, two samples were taken: one at a depth 0 to 25 cm and the second at 25 to 40 cm.

Peat samples for bulk density were also taken. Due to the densely rooted nature of near surface peat, it was not possible to take bulk density samples at or near the surface with conventional bulk density rings. To overcome this problem, bulk density was obtained at the surface using sections of the samples taken for hydraulic conductivity analysis. These samples had a regular shape, which enabled estimates of bulk density. Below this and into the catotelm layer, an Eijkelkamp 04.09 peat sampler (Eijkelkamp, Agrisearch Equipment BV, The Netherlands) for bulk density analysis was used. Using this auger which has a semi-circular shape of diameter 5 cm, the full depth of the peat (in some cases as much as 5 m) was sampled in increments of 0.5 m deep. These samples were placed in airtight bags for later laboratory analysis.

5.3.4 Laboratory analysis

In the laboratory the peat samples for hydraulic conductivity were sliced so that three replicate cubes, of sides 10 cm were prepared. The top 10 cm which contained the living plants and mosses was removed from the surface. A smooth-bladed knife was used in order to minimise disturbances to the samples by tearing of roots or peat fibres. The surface of each cube was dried with paper towels and then quickly dipped in molten paraffin wax and left to cool. This process was repeated until a thick covering (approximately 1 cm) of wax was in place, taking care not to compress the samples at any stage. This method is similar to that of Beckwith et al. (2003a). Once the cubes were covered in wax, the top layer of wax was removed and a wood collar was placed on the samples. The samples with collars in place were then sealed again in wax. Once the collar was deemed watertight and stable, the wax covering the bottom face of the cube was removed while it was still hot and easily cut.

In the laboratory, the saturated hydraulic conductivity was determined using the constant head permeability test (Beckwith et al., 2003a). First the samples were left to soak overnight in bog water. Filtered peat water was then placed in the reservoir enclosed by the collar (at the top of each sample) and the samples were then placed on a 2 mm sieve floor. To eliminate the possibility of water with a different chemical signature affecting the results, we used water collected from the peatland. In the constant head method, a head of 50 mm was maintained at all times. This was achieved by connecting the reservoir on the peat samples to a large reservoir with a large surface area of bog water with plastic tubing allowing a siphon to form, keeping both levels equal. The peat samples were placed in a plastic container not much bigger than the peat samples but which did ensure the bottom of the peat samples were under a small depth of bog water at all times. Shortly after the water started to percolate through the sample the plastic container at the bottom overflowed and this overflow was then collected in a graduated glass beaker. The entire apparatus was then left until the discharge became steady which varied from a few hours to days. The water level of the reservoir of bog water was observed on a regular basis and a volume of water equal to the volume percolated through the sample was replaced into the reservoir. This ensured that there was no change in hydraulic gradient across the sample. The room temperature was maintained at a constant 18°C to eliminate any possible affects of changing temperature on the hydraulic conductivity analysis. No particles were observed in the discharge. The saturated hydraulic conductivity was estimated using Darcy's Law: eqn 5.1.

$$K_{xat} = \frac{Q * L}{A * \Delta h} \qquad \text{(eqn 5.1)}$$

where K_{sat} is the saturated hydraulic conductivity (m s^{-1}); Q is the discharge (m^3 s^{-1}); A is the area of the face of the cube (m^2); Δh is the difference in head between the top and bottom of the sample (m); and L is the length of the sample (m).

Following the determination of the vertical saturated hydraulic conductivity (K_{vsat}), the apparatus was dismantled and the peat was allowed to drain. The exposed faces were resealed with wax; the cube was rotated through 90 degrees and two more opposing faces were exposed and the horizontal saturated hydraulic conductivity (K_{hsat}) was then determined using the same procedure. The vertical hydraulic conductivity was determined before the horizontal hydraulic conductivity as a matter of routine. To ensure that determining vertical hydraulic conductivity before horizontal conductivity did not affect the results, the horizontal hydraulic conductivity was determined prior to vertical hydraulic conductivity for a number of cube samples at two locations and in all cases the horizontal hydraulic conductivity was found to be greater than the vertical hydraulic conductivity. Beckwith et al. (2003a) who also investigated whether determining vertical hydraulic conductivity before horizontal hydraulic conductivity influenced the results also found no effect.

The samples for bulk density (below 50 cm) were oven dried for one week at 55 ºC. Samples were then weighed and re-weighed 24 hours later to ensure all moisture had evaporated. All the samples used in the analysis of hydraulic conductivity were also analysed for bulk density. Once the hydraulic conductivity tests had been completed the wax was removed from the samples and the length of each side of the cube was measured so as to determine the volume and the samples were then dried for bulk density analysis using eqn 5.2:

$$\rho_{bd} = \frac{m_d}{V_{or}} = \frac{m_d}{l * h * w} \qquad \text{(eqn 5.2)}$$

where ρ_{bd} is the dry bulk density (g cm^{-3}); m_d is the dry mass of the sample (g); V_{or} is the original (wet) volume of the peat sample (cm^3); l is the length of the sample (cm); h is the height of the sample (cm) and w is the width of the sample (cm). The conventional gravimetric based definition of soil moisture (θ_G) as is used for mineral soils is defined as $\theta_G = M_w/M_S$, where M_w is the mass of water in the soil and M_S is the mass of soil. However given the large proportion of water in peat and the relatively light mass of peat, the conventional definition of gravimetric soil moisture results in values of order 10^4. Thus we determined the peat moisture content using eqn 5.3 as follows.

$$\theta = \frac{m_{tot} - m_d}{m_{tot}} * 100 \qquad \text{(eqn 5.3)}$$

where θ is the mass ratio based moisture content in %; m_{tot} and m_d are the total wet mass of peat (before drying) and the dry mass of the peat (after drying) respectively. Thus it was possible to estimate bulk density and moisture content for the entire profile of the peat.

All the calculations, statistical analysis and graphical outputs were determined using MATLAB (Math works USA, 7.6.0, R2008a).

5.4 Results

5.4.1 Bulk density

Table 5.3, summarises the physical properties of the peat at all 14 trial pit locations. At the stream edge (bog margin) the peat depth was < 1 m and increased to > 5 m at the bog centre. The slope adjacent the bog margin was ~ 1/8 and flattened to almost horizontal at the bog centre. At the time of sampling (winter time), the water table depth at the margin was ~ 10 cm below the surface and was at the surface near the bog interior.

A summary of our measured bulk density (ρ_{bd}) results is presented in Table 5.3 and Figure 5.3, showing a range from 0.038 to 0.165 g cm^{-3}. Near the stream margin the depth averaged bulk density was highest at ~ 0.11 g cm^{-3} while it was lowest near the bog centre at ~0.055 g cm^{-3}. The bulk density values reported here are similar to those of Wellock et al. (2011), Tomlinson and Davidson (2000) and others for blanket peatlands with depths greater than 2 m (see Table 5.2). We found little change in bulk density with depth at either the stream margin or at the bog centre, with the exception of the bottom of the profile. Using the Pearson correlation coefficient, no significant (P<0.05) linear change in bulk density with depth could be found.

5.4.2 Moisture content

The average moisture content from each profile (Table 5.3) ranged from 86.5 to 95.1 %. These values are very high compared to mineral soils. Unlike bulk density, the degree of saturation increases from the bog margin to the interior. As with bulk density, the Pearson correlation coefficient did not show any significant (P<0.05) change of moisture content with depth with the exception of peg 5 which did show a decrease in moisture content (P<0.05) with depth.

Table 5.2 Bulk density values from studies in various peatlands.

Location	Type	Range ($g\ cm^{-3}$)	Average ($g\ cm^{-3}$)	Reference
Minnesota USA	Raised bog	0.06 - 0.14		(Chason and Siegel, 1986)
northern Ireland	Raised bog	0.058 - 0.084	0.069	(Tomlinson and Davidson, 2000)
Ireland	High level blanket bog	0.037 - 0.254	0.123	(Wellock et al., 2011)
Ireland	Low level blanket bog	0.058 – 0.202	0.116	(Wellock et al., 2011)
Ireland	Basin peat	0.095 – 0.264	0.155	(Wellock et al., 2011)
Ireland	Basin, raised & drained peats	0.05 - 0.79	0.17	(Kiely et al., 2010)

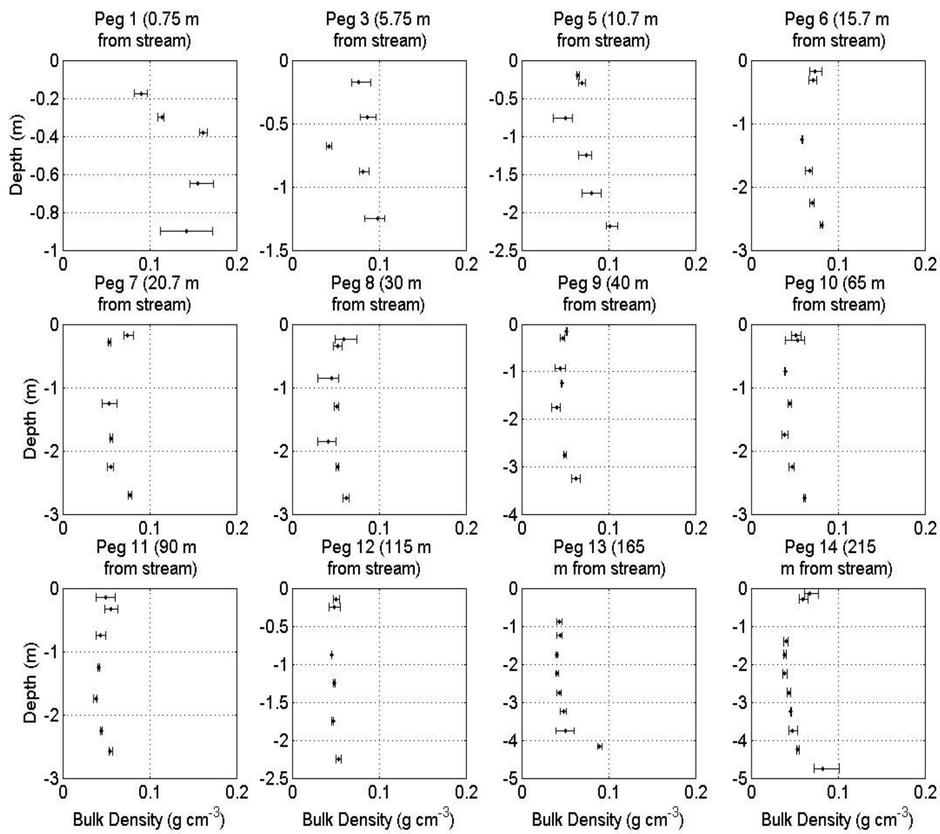

Figure 5.3 Bulk density at 12 locations, for clarity and due to the close proximity to other pegs the plots for pegs 2 and 4 were omitted.

Table 5.3 Summary of results of peat parameters at all 14 cores/pegs along the transect.

Profile Peg Number	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**	**11**	**12**	**13**	**14**
Distance from stream(m)	0.75	3.25	5.75	8.25	10.7	15.7	20.7	30.7	40.7	65.7	90.7	115.7	165.7	215.7
Depth of peat (m)	1.0	1.05	1.77	2.00	2.23	2.80	3.05	3.00	3.6	3.0	2.83	2.5	4.42	>5.3
Elevation (masl)	159.7	160.0	160.4	160.8	161.3	161.6	162.0	162.2	162.2	162.3	162.7	162.2	163.2	163.5
Slope (m m^{-1})	0.126	0.143	0.16	0.189	0.117	0.069	0.035	0.009	0.003	0.011	-0.002	0.007	0.014	0.007
$K_{hsat\ 10\text{-}20cm}$ (10^{-6} m s^{-1})	0.12	-	1.35		3.89	1.70	5.36	10.43	24.13	11.01	10.84	13.36	-	125.43
$K_{vsat\ 10\text{-}20cm}$ (10^{-6} m s^{-1})	1.00	-	0.74	-	4.09	2.81	3.91	4.17	6.57	3.36	9.28	10.28	-	121.08
$K_{hsat\ 30\text{-}40cm}$ (10^{-6} m s^{-1})	0.58	-	0.33	-	0.39	0.17	0.19	18.81	8.56	17.13	9.06	111.07	-	10.42
$K_{vsat\ 30\text{-}40cm}$ (10^{-6} m s^{-1})	0.72	-	2.46	-	1.35	0.48	0.25	2.45	6.63	1.05	5.95	25.75	-	2.23
Moisture content (%)*	86.5	88.25	90.45	90.47	91.39	92.84	93.83	94.5	94.73	94.96	95.04	94.73	94.41	94.48
$\rho_{10\text{-}20}$ cm (g cm^{-3})	0.090	-	0.076	-	0.064	0.074	0.074	0.060	0.052	0.051	0.049	0.050	-	0.068
$\rho_{30\text{-}40}$ cm (g cm^{-3})	0.115	-	0.086	-	0.070	0.071	0.054	0.052	0.047	0.053	0.055	0.049	-	0.059
ρ depth average for profile (g cm^{-3})	0.128	0.111	0.09	0.107	0.068	0.071	0.062	0.052	0.049	0.048	0.047	0.049	0.05	0.052

5.4.3 Moisture content

The average moisture content from each profile (Table 5.3) ranged from 86.5 to 95.1 %. These values are very high compared to mineral soils. Unlike bulk density, the degree of saturation increases from the bog margin to the interior. As with bulk density, the Pearson correlation coefficient did not show any significant (P<0.05) change of moisture content with depth with the exception of peg 5 which did show a decrease in moisture content (P<0.05) with depth.

5.4.4 Saturated hydraulic conductivity

The results for saturated horizontal and vertical hydraulic conductivity at the near-surface (10-20 cm) and sub-surface (30-40 cm) are shown in Figure 5.4 and Figure 5.5 respectively. At each trial pit location on the hillslope transect, the results of three replicates are presented, which show a variation in saturated conductivity of less than one order of magnitude between replicates. To further analyse these results we considered the area between pegs 1 and 5 which is within 10 m of the stream (a riparian zone) and the area between peg 11 and 14 (90 to 215 m from the stream) as the bog interior which is relatively flat by comparison to the slope of ~0.15 m m^{-1} in the riparian zone. Both the K_{hsat} and K_{vsat} at the near-surface and sub-surface showed a significant (P <0.05) increase between the riparian zone and the centre zone when analysed with students t-test. Applying the same principle to bulk density showed that there was a significant decrease (P <0.01) in bulk density between the riparian zone and the centre zone.

K_{hsat} for the near-surface depth (10 – 20 cm) ranged from ~10^{-7} m s^{-1} near the stream to ~10^{-4} m s^{-1} at the bog interior, a difference of three orders of magnitude. K_{vsat} for the near-surface depth ranged from ~10^{-6} m s^{-1} near the stream to ~10^{-4} m

s^{-1} at the bog centre. K_{hsat} for the sub-surface depth (30 – 40 cm) ranged from ~10^{-6} m s^{-1} near the stream to ~10^{-4} m s^{-1} at the bog centre. K_{vsat} for the sub-surface depth ranged from ~10^{-6} near the stream to ~10^{-5} m s^{-1} at the bog centre. From Figure 5.4 and Figure 5.5, it is noted that there is an appreciable range in some replicate samples.

Investigations by Beckwith et al. (2003a) on Thorne Moors, UK, found that anisotropy existed for most of the samples tested, with K_{hsat} generally greater than K_{vsat}. We found that anisotropy does exist with horizontal hydraulic conductivity approximately twice that of the vertical hydraulic conductivity for the near-surface according to (5.4):

$$Kv_{sat} = 0.45Kh_{sat} \qquad \text{(eqn 5.4)}$$

$$R^2 = 0.22, RMSE = 2.81*10^{-6}$$

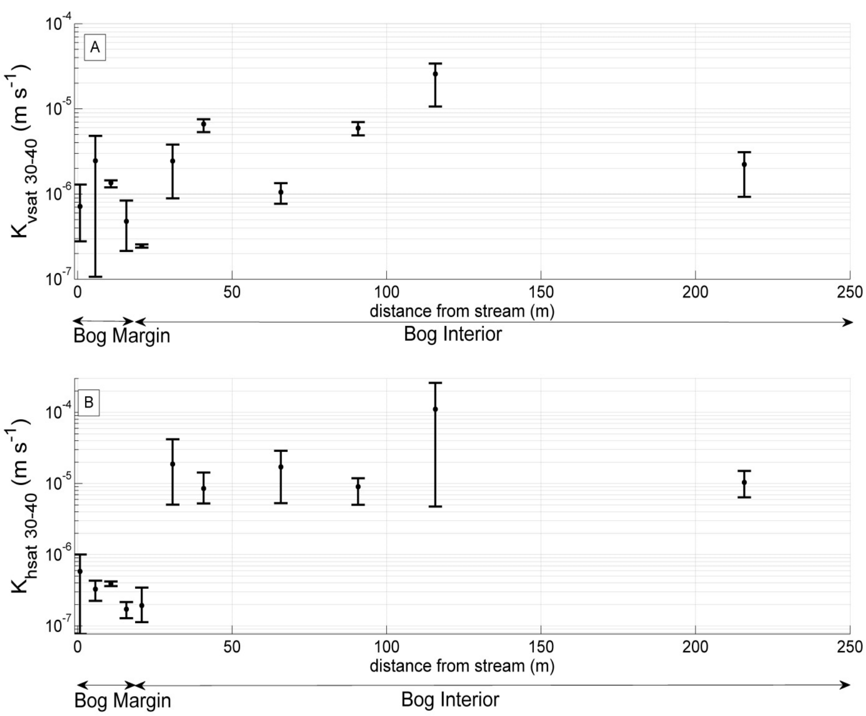

Figure 5.4 Shallow depth vertical (a) and horizontal (b) hydraulic conductivity

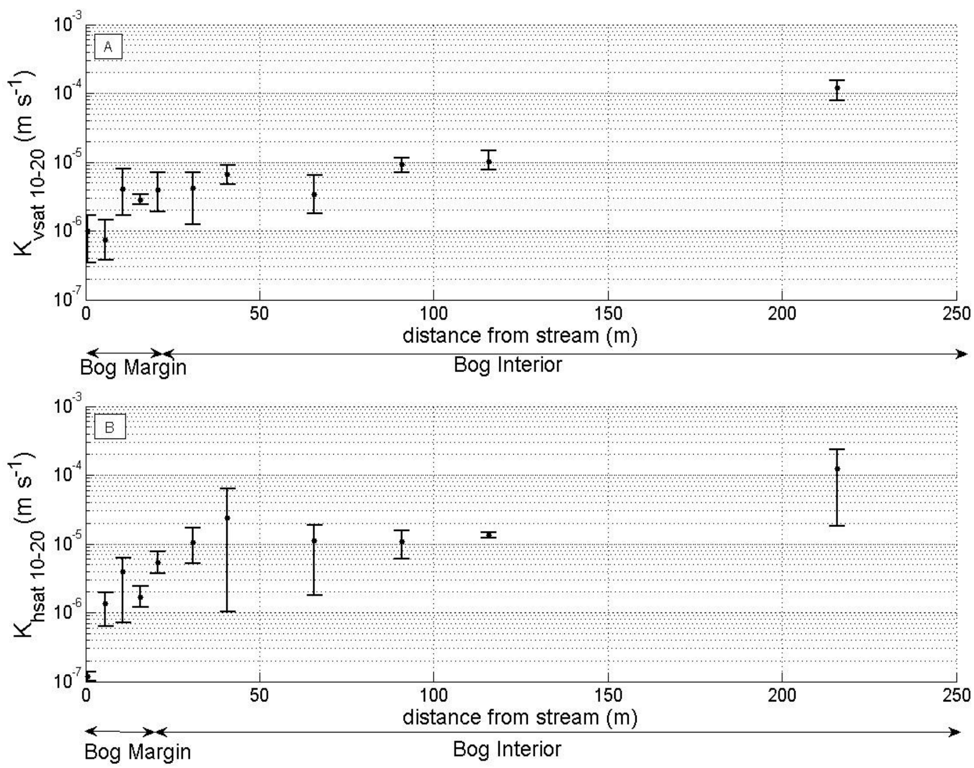

Figure 5.5 Deeper depth vertical (a) and horizontal (b) hydraulic conductivity.

5.5 Discussion

Our values of saturated hydraulic conductivity compare with others (see Table 5.1) including Beckwith et al. (2003a) who used the modified cube method in Thorne Moors, UK (raised bog) and found that the vertical hydraulic conductivity near the surface ranged from 10^{-3} to $1.6*10^{-5}$m s^{-1} and horizontal hydraulic conductivity ranged from $8*10^{-4}$ to $1.6*10^{-5}$ m s^{-1}. Beckwith et al. (2003a) also reported vertical conductivity values at depths of 30 cm that ranged from $3.2*10^{-7}$ to $7.9*10^{-7}$ m s^{-1} and horizontal conductivity values at the same location that ranged from $2.5*10^{-6}$ to 10^{-5} m s^{-1}.

The peat sample size used to conduct the *K* tests in this study was a cube of 10 cm sides. While the literature on sampling peat for *K* is limited, others have used smaller sized cubes or cylinders of smaller cross sectional areas. The cross sectional area of our sample size at 100 cm^2 was larger than: the cylinder of Beckwith et al. (2003a) who used a cross sectional area of 56 cm^2; the cylinder of Chason and Siegel (1986) with a cross sectional area of 18 cm^2 ; the cylinder of Boelter (1965) with a cross sectional area of 75 cm^2; the cylinder of Scholtzhauer and Price (1999) with a cross sectional area of 23cm^2; the cube of Quinton et al. (2008) with a cross sectional area of 36 cm^2; the cube of Kruse et al (2008) with a cross sectional area of 81 cm^2. Recent work (e.g. (Quinton et al., 2008)) suggests that peat hydraulic conductivity (at least for cold regions) is controlled by pore size and pore hydraulic radius. Peat is considered to be made up of large and small pores, the former leading to what is described as “active” porosity and the latter as “inactive” porosity. Many of the small pores are thought to be dead end or closed pores (filled in with the remains of plant cells) and contributing little to the flow. While total porosity of the near surface peat was of the order of 0.80 to 0.95, the active porosity was considered to be ~0.47 to 0.69 with pore size of the order 1 cm (Quinton et al., 2008). Rezanezhad et al. (2009) using 3-D computed tomography suggest an equivalent pore radius of ~2.2 cm. This being the case, then our sample cube size of 1000cm^3 might not be big enough. However from the above citations, our sample is at the larger end of samples examined to date.

We found that saturated hydraulic conductivity is lower in the riparian zone near the stream than the interior of the peatland. This phenomenon was first proposed from a sensitivity analysis of a groundwater model by Lapen et al. (2005) and later found to hold in a raised bog in Wales by Baird et al. (2008). Our results demonstrate the same pattern of spatial variability in the blanket bog in this study. With evidence of similar spatial variability in both raised and blanket bogs, this reinforces the original idea by Lapen et al. (2005) of the importance of the peatland margins in maintaining the elevated water table in the bog interior and thus the overall health and structural integrity of the peatland.

The ecology of the natural border of the bog near the stream was very different to that of the interior of the bog in many aspects. The topography orthogonal to the stream had a gradient of the order of ~0.15 m m^{-1} while the bog interior was mainly flat. The peat depth near the stream was of the order of 1 m while that at the bog interior was as much as 5m. The depth to water table near the stream was about 0.1m while that at the bog interior was close to the surface. The water chemistry of the areas close to the stream had a lower concentration of chloride, a higher water colour, a higher pH and a higher ammonium concentration than the near surface water of the bog interior (Sottocornola et al., 2009). The surface of the bog interior was a mosaic of vegetation communities organized in undulating microforms: hummocks (highest elevation); high lawns; low lawns and hollows or pools triggered by different rates of peat accumulation (Laine et al., 2007a; Sottocornola et al., 2009). However, the area close to the stream was covered almost solely by high lawns, where pools or low hollows cannot structurally exist because of the steep local land gradient. Furthermore, the dominant plant species near the stream were vasculars, which covered less than 30% of the area at the bog interior. The vascular plants contain lignified tissues, that facilitate the conductance of water (enabling transpiration) while the dominance of bryophytes at the bog interior which do not have a root system and as such limits transpiration (Sottocornola and Kiely, 2010).

Furthermore, during the field work, when we sliced down through the bog at the bog interior, to a depth of about 0.3 m, the bog walls left behind collapsed shortly afterwards, suggesting the lack of shear and compressive strength at the bog interior. However, when we sliced down through the bog material near the stream, and removed a cube (>0.3 m by 0.3 m by 0.3 m) no collapse of adjacent peat walls occurred. This suggests the existence of some shear strength of the more compressed bog material near the stream.

While the peat at the riparian zone has a different vegetation cover than the bog interior which may explain some of the different peat properties found there, it may also be possible that an external force has acted on the peat at the margins to compress the peat giving it a higher bulk density and structural strength. It can be noted from Table 5.3 and Figure 5.2 that at a distance of 20 m from the stream the water which lies just below the surface is ~3.3 m higher than the stream bed. Given that for much of the year the flow depth in the stream is of the order 20-30 cm, this represents a 3 m difference in hydraulic head between the stream and peg 7. Thus the water in the centre of the peatland which is made up of peat with a higher hydraulic conductivity and with areas of pools which can be several meters in depth, is likely to exert a significantly hydrostatic pressure on the areas of peat with lower hydraulic conductivity at the margins. We suggest that it is this force that over time has compressed the peat at the margins resulting in its reduced hydraulic conductivity. It is well documented that blanket peatlands in temperate climates accumulate carbon in the soil at a rate of a fraction of a millimetre per year (Koehler et al., 2011) due to the fact that blanket peatlands are a sink for atmospheric carbon dioxide. Thus, over the centuries, this build up of material (at the bog interior in particular) has resulted in additional hydrostatic pressures on the bog margins causing their compaction resulting in increased bulk density, reduced porosity and reduced hydraulic conductivity at the margins which in turn results in elevating the water table in the centre of the bog further promoting peat development and accumulation.

With K_{hsat} about twice that of K_{vsat} for the near-surface, this suggests that at the bog interior, the tendency is for rainfall excess to become (horizontal) flow rather than vertical flow. Studies by Reeve et al. (2000) also suggest that when a peat forms over a low permeability soil, such as exists in Glencar with its clay base, the vertical movement of water through the bog profile is negligible and lateral flow dominates. As the water table at Glencar is close to the surface all the year round (especially at the bog interior), the bog profile is saturated from below and undergoes saturated excess overland flow (SEOF). In this bog environment, with annual rainfall at 2800 ± 500 mm, we found that over a seven year period from 2003 to 2009 the rainfall rate never exceeded 36 mm hr^{-1} which is ~ 10^{-5} m s^{-1} (i.e. lower than the measured saturated horizontal conductivity of 10^{-4} m s^{-1}).

A study of the rainfall hyetographs and the stream flow hydrographs suggest that the time of concentration (ToC), defined as the time it takes runoff water to flow during a rain event, from the most remote point of the catchment to the catchment outlet (where the stream gauge is located) for this small catchment (76 ha) is of the order of 3 hours. The maximum travel distance for rainfall to travel along the surface of the peatland to the stream is approximately 1 km and this equates to an overland flow velocity of ~0.1 m s^{-1} to result in a ToC of 3 hours. In a recent study by Holden et al. (2008), overland flow velocities were found to range from 0.191 m s^{-1} to $1.22*10^{-4}$ m s^{-1} with the higher velocities associated with bare soil and the lowest associated with sphagnum covered ground cover. Our estimate of overland flow velocity is four orders of magnitude faster than K_{hsat} of the near-surface depth

(10-20 cm). As the flood hydrographs of the stream take ~ 3 hours to reach peak flood flow and about 4 hours for the flood recession limb to return to pre flood conditions, this suggests that the flood waves are dominantly composed of overland flow. To further investigate this, an analysis of the baseflow was undertaken at the outfall of the catchment (see Figure 5.1). The base flow index (BFI) which is the relative contribution of baseflow in the stream to total annual flow, was estimated based on the Low Flow Studies Report No 1. from the Institute of Hydrology (1980). The BFI for the catchment was found to be 0.11, 0.04 and 0.08 for the years 2007, 2008 and 2009 respectively with an average of 0.076 for the 3 years. This suggests that approximately 92% of the stream flow occurs from surface runoff. Catchments with low BFI values have low storage capacity thereby resulting in flashy flood peaks. This is due to the water table for most of the bog area remaining near the surface all the year round, which itself is the result of the very low hydraulic conductivities at the bog margins. For other peatlands (such as raised bogs) where the water table can be lower (60 cm below the surface (Lafleur et al., 2005)) there is potential for water storage.

In an effort to compare the stream flow and saturated hydraulic conductivity values obtained from this study, preliminary calculations were made using the stream length and bog depth. From field observations we noted that the bog depth is ~1.0 m at the interface with the stream. The total length of the stream in the bog is ~1500 m (see Figure 5.1). The average K_{hsat} for the peat closest to the stream of ~$0.5*10^{-6}$ m s^{-1} (see Table 5.3). Thus using eqn 5.5 it is possible to estimate the flow through the peat matrix into the stream:

$$Q = 2l * d * kh_{sat} \frac{dh}{dl} \qquad \text{(eqn 5.5)}$$

where Q = flow (L s^{-1}), l = length of stream in the bog (m), d = depth of bog at stream (m), K_{hsat} is the saturated horizontal conductivity (m s^{-1}) and $dh\ dl^{-1}$ is the hydraulic gradient. As the water table depth was observed to remain stable at 10 cm below the surface near the stream for a 12 month period of observations, then we can assume that the hydraulic head can be estimated from the slope (see Table 5.3) of the peat closest to the stream. From eqn 5.5 we estimate the flow through the peat matrix as ~ 1.7 l s^{-1}. As the water table remains close to the surface all year it can be assumed that this flow is constant and represents the groundwater (base flow) discharge into the stream. The average annual stream flow was 2200 mm for the years 2007 to 2009. With base flow at ~ 8% this gives an average base flow of ~ 4 l s^{-1} which compares reasonably with the above estimate of 1.7 l s^{-1}. While estimating the baseflow from hydraulic conductivities in this fashion is not rigorous, it does show that the baseflow estimates using hydraulic conductivity produce flow of the same order of magnitude as the observed flow. This compares well considering the likely variability in depth of peat, hydraulic conductivity and hydraulic head along the length of the stream as well as the possible presence of pipes which can be additional source of stream flow in peats (Holden and Burt, 2002c).

We also suggest that due to the inability of water to resist shear force, peat with high moisture contents will have less structural stability and may be more at risk to peat movement and slides. Creighton (2006) documented such failures in Irish blanket peatlands. The nature of the topography of the peatland in Glencar is such that shallower peat depths occur at lower elevations adjacent to the stream and greater depths at higher elevations were found in the bog interior. This leads us to suggest that it is the peat in the riparian zone which has a lower moisture content and a higher bulk density, that structurally supports the less dense peat of the interior of the bog.

5.6 Conclusions

With regard to saturated hydraulic conductivity and soil physical properties (porosity and bulk density) we found that this pristine blanket peatland is composed of two distinct zones: one near the margins (i.e. near a stream) and the second at the bog interior. At the margins the peat has higher density, lower porosity, lower water content, and lower hydraulic conductivity than the bog interior. The horizontal conductivity was approximately twice that of vertical conductivity. While the findings of Lapen et al. (2005) in a model study, of areas of lower hydraulic conductivity at the peatland margins were supported by Baird et al. (2008) in a raised bog, our results for a blanket bog in the southwest of Ireland also support Lapen (2005). We infer that the areas of lower hydraulic conductivities at the margins play an important role in the overall health of blanket peatlands as they maintain the elevated water table at the interior of the bog. Removal or damage of the peatland at the margins may result in a decrease in the water table height, leading to loss of carbon by decomposition and erosion and to a decrease in the general overall health of the bog. Integrated over the timescale of a year, the stream runoff is composed of 8% baseflow and 92% flood flow, and the latter is from surface runoff rather than subsurface flow. This is because the water table remains so close to the surface for most of the bog area, all year round. It is important in hydrological modelling to take these spatial differences of key properties into account as it appears that it is the hydraulic conductivity at the stream margin controls the runoff from the peatland.

5.7 Acknowledgements

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Chapter 6

How does afforestation affect the hydrology of a blanket peatland? A modelling study

Running title: Hydrological response of afforested peat catchment

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6.1 Abstract

Over the last century, forestry in Ireland has increased from 1% of the land area to 10%, with most plantations on upland blanket peatlands. This land use change is considered to have altered the hydrological response and water balance of upland catchments with implications for water resources and carbon budgets. Due to the difficulty of observing these changes in the field, the aim of this study was to utilise a hydrological model to simulate the rainfall-runoff processes of a drained afforested peatland. The hydrological rainfall model (GEOtop) was calibrated and validated for a small (76 ha) pristine blanket peatland in the southwest of Ireland for the two-year period 2007-2008. The hydrological response of the virgin blanket peatland with regard to stream runoff and water table (WT) levels was captured well in the simulations. Two land-use change scenarios of afforestation were examined with the model; (A) a young 10 year old and (B) a semi mature 15 year old Sitka Spruce forest. Scenario A produced similar rainfall runoff properties to the virgin peatland whereas runoff from Scenario B was 20% lower. On average the WT was drawn down by 16 and 20 cm below the observed WT for Scenarios A and B, respectively. The maximum WT draw down in Scenario B was 61 cm and occurred in the summer months, resulting in a significant decrease in stream runoff. Occasionally in the winter (following rainfall) the WT for Scenario B was just 2 cm lower than the unafforested peatland, which coupled with the drainage networks associated with afforestation led to higher peak stream flows. This study finds that afforestation (following drainage) leads to increased evapotranspiration (from the forest canopy) and decreased the annual stream runoff compared to the non afforested peatland.

6.2 Introduction

Peat is an organic soil composed of partially decomposed plant matter (Hoag and Price, 1995) with depths that range from 30 cm to as much as 1000 cm. Peatlands cover 400 million hectares of the earth's surface and store between 33% and 50% of the world's soil carbon pool (Holden, 2005b) which has been estimated at 25-50% of the current carbon held in the atmosphere (Frolking and Roulet, 2007). This vast store of carbon is considered to be vulnerable to: climate change (Oechel et al., 2000; Sottocornola and Kiely, 2010); artificial drainage (Holden et al., 2004) and land use change (Limpens et al., 2008). More than 80% of these peatlands are located in temperate-cold climates in the northern hemisphere particularly in Canada (Letts et al., 2000), Russia, USA and parts of northern Europe (Limpens et al., 2008). In all regions of these northern peatlands there have been vast areas that have undergone drainage for commercial forestry (Waldron et al., 2009). An estimated 500,000 ha of peatland was afforested between the 1950s and 1980s in the UK (Hargreaves et al., 2003). This land use change and drainage has a profound effect on vegetation and carbon loss of peatlands (Strack et al., 2006).

From a regional perspective, much of Ireland's native forestry had been felled over the centuries so much so that by the beginning of the twentieth century, forestry accounted for only 1% of the total land cover in the Republic of Ireland (Eaton et al., 2008; Pilcher and Mac an tSaoir, 1995). However, since the 1950s it has been the policy of successive Irish governments to increase forest cover and by 2007 the national forest area had risen to 10% (NFI, 2007) with a planned increase to 18% by 2020 (Dept. of Agriculture, 1996).

Much of this afforestation over that past five decades has taken place on peatlands which were traditionally considered unsuitable for agricultural use. An estimated 49% of afforestation between 1990 and 2000 was carried out on peat soils (Black et al., 2008). The principal tree species used in peatland afforestation in Ireland were Sitka spruce (*Picea sitchensis* (Bong.) Carr.) and lodgepole pine (*Pinus contorta* Dougl.) (Byrne and Farrell, 2005). Sitka spruce is non native to Ireland but is favoured due to its rapid growth in the temperate humid Irish climatic conditions and its ability to survive in difficult terrain. An estimated 57% of the national forest stock is Sitka spruce (Horgan et al., 2004). While Sitka spruce is able to thrive under the moist Irish weather conditions (Horgan et al., 2004), its root development when planted on peat soils is limited to the aerated top section of the peat profile (Lees, 1972). Peatlands, and particularly blanket peats, in Ireland are environments with the water table at, or close to, the surface for long periods of the year (Bragg, 2002; Hogan et al., 2006; Holden and Burt, 2002a; Iritz et al., 1994; Laine et al., 2007a). This makes peatlands unsuitable for afforestation in their natural undrained condition which can result in stunted root development and vulnerability to wind throw.

For afforestation purposes, peatlands are typically drained prior to planting (Holden et al., 2004). Drainage is normally carried out using a combination of closely spaced plough furrows and deep (0.5 – 2.0 m) but more widely spaced ditches. Frequently this results in a change in runoff production both in the short term while the drains are active and in the long term when the forest becomes established. These drains, while beneficial for the development of the forest, have also been linked with higher peak stream flows (Robinson, 1986). In an early study by Burke (1975) the runoff: rainfall ratios from an undrained peatland in

Glenamoy, Ireland, were 23.4% compared to 79.2% from a drained catchment. Conway and Millar (1969) found that artificially drained peats produced extremely rapid runoff in the north Pennines (UK). Ahti (1980) found that the flood peak increased after drainage in Scandinavian peatlands. However, Holden et al. (2006) found that surface runoff was greatly reduced following drainage within the Moor House blanket peatlands in the Pennines. Investigations by Iritz et al. (1994) on a selection of forested Scandinavian peatland catchments found that peak flows decreased following drainage. Prevost et al. (1997) reported an increase in stream base flow following drainage in a Canadian peatland. While there have been conflicting conclusions drawn from different international studies, this is likely due to limited data and the diversity of ground conditions (e.g. natural variation of water table depth), which is seen as critical to the amount of storage available and surface runoff production in peatlands (Holden et al., 2006). Furthermore, the type of drainage will impact the degree of change in hydrology. Deep closely spaced drains cause a peat catchment to respond differently to shallow widely spaced drains. Holden et al. (2011) observed that in an intact blanket bog in Oughtershaw Moss in northern England that while the seasonal range of water table depth was 0 to 20 cm, the spatially weighted mean water table depth was only 5.8 cm. In the same bog, they found for a drained section, that while the depth of the seasonal water table ranged from 0 to 40 cm, the spatially weighted mean water table depth was 11.5 cm. The natural water table depth varies from lows of ~10 cm in Irish blanket peatlands to highs of ~60 cm in Canadian and Scandinavian blanket peatlands (Koehler et al., 2011). Thus drainage always precedes blanket peat afforestation in Ireland but this is not necessarily so in Canada and Scandinavia.

The rainfall runoff response of a peatland catchment changes once it becomes afforested (Anderson et al., 2000; Heal et al., 2004; Hudson et al., 1997b). Drainage (water table lowering), peat shrinkage and compression, tree canopy interception and evapotranspiration, all contribute to a changed hydrological (and hydrochemical) regime of a peatland when afforested (Ballard et al., 2011b; Holden et al., 2004; Institute of Hydrology, 1991; Iritz et al., 1994). Anderson et al. (2000) in a study investigating the effects of blanket bog afforestation on the physical properties of the peat soil and on the quantity and timing of runoff, found a reduction of 7% in runoff after afforestation in a Scottish peatland relative to an unforested drained control. They noted that the reduction in runoff was predominantly in the spring and summer, possibly linked with higher evapotranspiration from the forest canopy. Compared with the drained control, peak flows were increased by afforestation while the baseflow component of the total flow was reduced.

Interception losses in a Sitka spruce forested peatland in Scotland were reported to be greater than 50% of the annual precipitation (Heal et al., 2004). An investigation by the Institute of Hydrology (1991) into the water resources of two upland catchments in Scotland found significant interception losses amounting to 38% of precipitation. Anderson et al. (1990) reported 38% canopy interception and 12% transpiration loss in UK afforested peatland. Johnson (1990) found that in a 50-year-old Sitka spruce forest in the Scottish highlands on peat and peaty gley soils, that the average interception over a 3-year period was 28% with the greatest interception occurring in the summer months and the least in winter. A 25-year-old Sitka spruce forest on a peaty gley soil in Northumberland, UK was observed to have an average interception loss of 48% of precipitation (Anderson and Pyatt, 1986). Evapotranspiration from a pristine blanket peatland in south-west Ireland was observed over 5 years to be 15.5% of total precipitation representing an annual average of 394 mm (Sottocornola and Kiely, 2010). These studies demonstrate that evapotranspiration increases and stream runoff decreases after afforestation of peatlands. (Table 6.1). A review by Hudson et al. (1997b) of studies carried out on a number of catchments including the Plynlimon and Lanbrynmair catchments in Wales, concluded that afforested upland catchments (on a mixture of Peaty Gleys, Brown Earths and Podzols soils) had higher evapotranspiration than similar grassland covered catchments. It was further noted by Hudson et al. (1997b) that in the wet windy climate of the British uplands, 15-20% of rainfall is lost by transpiration from grasslands whereas 30-40% is lost from forested areas

Peatlands can serve as important regulators of river flow and hydrochemistry (Koehler et al., 2011), due to their location and precipitation amounts. Many rivers (at least in Ireland, Scotland and Wales) rise in areas of upland blanket peat with high precipitation (due to elevation). In upland blanket peatlands where the water table is perennially close to the surface (as occurs in northwest Europe), flash floods occur as there is very little storage potential in the peat (Baird et al., 2008; Bay, 1969; Bragg, 2002; Holden and Burt, 2002a; Holden and Burt, 2003; Lewis et al., 2011). A recent concept (Lapen et al., 2005) suggests that a layer of lower hydraulic conductivity at the margins of peatlands is responsible for maintaining higher water tables in the centre of the bog. Recent field tests by (Baird et al., 2008) on a raised bog and Lewis et al. (2011) on a blanket bog support this idea.

Table 6.1 A number of WT depths, stream flows and evapotranspiration values from literature.

Ecosystem	Location	Notes	Average annual precipitation (mm)	Annual evapotranspiration (mm)	Average WT depth (cm)	Average stream flow (mm)	Reference
Boreal raised bog	Canada	5 years data	910	351	>0.25 (varies considerably)	560	(Lafleur et al., 2005)
Mire	Sweden	2 years data	683	238	10	445	(Nilsson et al., 2008)
Blanket peatland	Ireland	pristine 2003 to 2007	2597	395	4	~2000	(Sottocornola and Kiely, 2010)
Blanket peatland	Scotland	Drained peatland	943	362	12	581	(Anderson et al., 2000)
Afforested peatland	Scotland	Sitka spruce 5 years old	943	401	19	541	(Anderson et al., 2000)
Afforested peatland	Scotland	Sitka spruce 50 years old	2130	597	-	1533	(Johnson, 1990)
Afforested peatland	Scotland	Sitka spruce 37 years old	2912	1375	-	1726	(Heal et al., 2004)

However to the author's knowledge this spatial variation of hydraulic conductivity has yet to be incorporated into hydrological models.

While there are several hydrological models capable of simulations on mineral soils (Abbott et al., 1986; Arnold et al., 1998; Beven and Kirkby, 1979; Reggiani et al., 2000), the same cannot be said for peat soils as many hydrological models are not well suited to wetlands (Price et al., 2005). A model parameterisation for Canadian peatlands was developed by Letts et al. (2000) for a soil vegetation atmosphere transfer scheme. SHETRAN was employed by Dunn and Mackay (1996) to investigate how peatlands are affected by ditches. Lane et al. (2004) modified TOPMODEL (Beven and Kirkby, 1979) for use with Digital Elevation Models of high resolution, as these high resolution models can lead to many saturated areas that are unconnected to the drainage network. Lane et al. (2004) also noted that TOPMODEL may not accurately represent the lateral movement of water in soils of low hydraulic conductivity into open drains. More recently Ballard et al. (2011a) developed a simplified physics-based model to investigate flow and water table responses to different drainage scenarios. The model was found to perform well under wet conditions capturing the peak flows well, with a poorer performance under drier conditions.

The general objective of this study was to use a two-year hydrological data set at a blanket peatland catchment and a process-based rainfall-runoff model to explore the hydrological response if the peatland were to be drained and afforested. Specifically the aims were: (1) to calibrate and validate the hydrologic rainfall runoff model GEOtop for a small scale blanket peatland catchment in south-west Ireland using two years of observations; and (2) to investigate the catchment hydrological response for (the scenario of) a drained pre-afforested condition with a 50-cm water table lowering and (3) two afforestation scenarios: A, a 10-year-old Sitka spruce afforestation simulation and B, a 15-year-old Sitka spruce afforestation simulation.

6.3 Materials and methods

6.3.1 Site description

The study site is a pristine Atlantic blanket bog near Glencar in County Kerry, southwest Ireland (Latitude: 51°58N, Longitude 9°54W) at an elevation of approximately 150 m (Figure 6.1) and is typical of Atlantic blanket bogs in the coastal regions of northwest Europe in terms of both vegetation and water chemistry (Sottocornola et al., 2009). The study site is part of a larger pristine bog of approximately 121 km^2. A small stream (<1m wide) runs through the centre of the bog and drains approximately 76 ha; 85% of which is intact blanket bog (see Figure 6.1).

The surface of the bog is a mosaic of microforms that differ in relative altitude, plant composition and water table (WT) depth. Four classes were distinguished in relation to their difference in WT level: hummocks, high lawns, low lawns and hollows which cover 6, 62, 21 and 11 % of the study site respectively (Laine et al., 2007a; Sottocornola et al., 2009). Vascular plants cover about 30% of the study site area with the most common species being *Molinia caerulea* (purple moorgrass) and *Calluna vulgaris* (common heather). About 25% of the bog surface is covered by bryophytes with the dominant species being a brown moss, *Racomitrium lanuginosum*. The vegetation of the study site is described in detail in (Sottocornola et al., 2009).

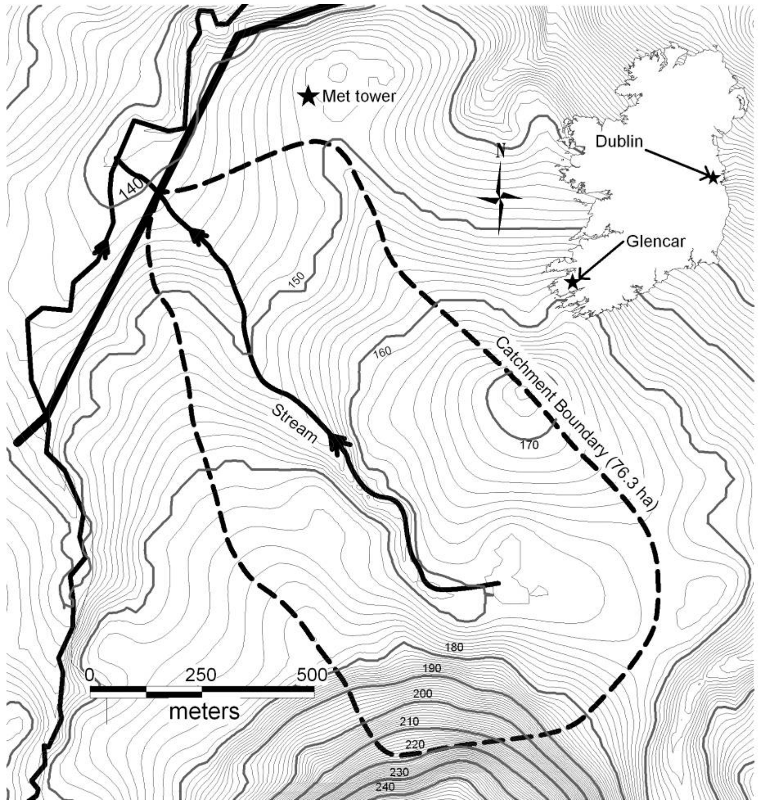

Figure 6.1 Peatland catchment map.

6.4 Site Instrumentation

At the outfall of the 76-ha catchment (Figure 6.1), the stream height was recorded every 30 min using a pressure transducer (1830 Series, Druck Limited, UK). Stream height was converted to discharge using a rating curve (eqn. 6.1) built up from stream height and stream velocity measurements taken over two years.

$$Q = 0.6946 * (h - 0.08071)^{1.441}$$
$$r^2 = 0.9986, RMSE = 0.005 \qquad \text{(eqn 6.1)}$$

where Q is the instantaneous discharge in $m^3\ s^{-1}$ and h is stream height in m. The rating curve was established from manual measurements of instantaneous discharge carried out at a range of stream heights using an OTT current meter (OTT Messtechnik GmbH & Co KG, Germany). It is interesting to note the similarities in the Q-h relationships in the rating curve described in eqn 6.1 and

Manning's equation. An estimate of Q in a wide channel using Manning's equation results in a relationship where Q is proportional to $h^{1.66}$. The rating curve established at the site found a relationship where Q is proportional to $h^{1.44}$. These values of 1.66 and 1.44 are relatively close and the constant value of 0.6946 may offer a physical interpretation of the Manning's equation constant.

The meteorological station (Figure 6.1) was established in 2002 and includes two tipping bucket rain gauges (an ARG100, Environmental Measurements Ltd., UK and an Obsermet OMC-200, Observator BV, The Netherlands) and a WT level recorder which consists of a pressure transducer (PCDR1830, Campbell Scientific, UK) placed inside a metal well pierced all along its height. Wind speed was recorded with a 2-D sonic anemometer (WindSonic, Gill, UK). Air temperature (T_{air}) and relative humidity were measured at 2-m height with a shielded probe (HMP45C, Vaisala, Finland), while atmospheric pressure (P) was recorded with a barometer (PTB101B, Vaisala, Finland). An eddy-covariance system for CO_2 fluxes was also located on the same tower. It consisted of a 3-D sonic anemometer (Model 81000, R.M. Young Company, USA) and an open-path infrared gas analyzer for H_2O and CO_2 concentrations (LI-7500, LI-COR, USA) mounted 3 m above the vegetation.

6.4.1 Climate

Sottocornola and Kiely (2010) at the same site found that the range of annual rainfall (2002 to 2009) was 2236 to 3365 mm with an annual average of 2597 mm. The annual evapotranspiration (ET) (estimated using eddy covariance methods) ranged from 369 to 424 mm with an annual average of 395 mm. From 2002 to

2009 there was an annual average of 208 wet days (> 1 mm day^{-1}) (Koehler et al., 2009). The average annual air temperature was 10.5°C.

The recorded flow at the stream outfall (see Figure 6.1) ranged between 0.015 and 10.0 l s^{-1} ha^{-1} (Koehler et al., 2009) with the 95 percentile flow exceeding 0.037 l s^{-1} ha^{-1}. The flow was observed to be flashy with over 90% of stream flow sourced from surface runoff (Lewis et al., 2011). This is due to a perennially high WT which was observed over the seven years 2002-2009 to be within 17 cm of the land surface (with the 7 year mean WT at ~ 4 cm below the surface). Lewis et al. (2011) investigated the spatial variation of bulk density and saturated hydraulic conductivity at the same site and found that at the surface (top 10 cm) saturated horizontal hydraulic conductivity was lowest at the bog margin near the stream (~10^{-7} m s^{-1}) and increased at the bog interior (~10^{-5} m s^{-1}). The converse was found for bulk density which ranged from ~0.11 g cm^{-3} at the bog margin to ~ 0.055 g cm^{-3} near the bog interior.

6.4.2 Process-based hydrological model - GEOtop

The process-based hydrological model GEOtop (Rigon et al., 2006) was used in this study. It is a distributed hydrological model (operating on a 8 m*8 m grid) and simulates the complete hydrological balance in a continuous way during a whole year (at a temporal increment of 60 minutes). It uses geospatial data e.g. topography, soil type (vertical and horizontal hydraulic conductivity, depth, the Van Genuchten parameters α and n (van Genuchten, 1980)), vegetation cover (including crop height, Leaf Area Index, root depth) and land cover. It provides spatially distributed output fields as well as routing water and sediment flows through stream and river networks.

GEOtop includes a rigorous treatment of the core hydrological processes (e.g. unsaturated and saturated flow and transport, surface energy balances, and streamflow generation/routing). Unsaturated dynamics are treated with a 3-D integration of Richards' equation while surface runoff is routed via a kinematic wave. The authors note that it would have been preferable to use a method of flood routing that would have considered attenuation in the channel. This lack of storage in GEOtop may result in an overestimation in peak flow in some catchments. This may not have been of concern to the developers of GEOtop as it was initially used in Alpine catchments with limited storage. The current version of GEOtop uses the Saint Venant equations for flood routing and a roughness coefficient of 20.0 was used in flood-routing. While the use of the Saint Venant equation in its current form in the GEOtop model does not allow peak attenuation, storage outside of the channel is taken into account as surface water will be stored on the surface of a cell providing attenuation outside of the channel. It must also be noted that there is very limited storage in the Glencar catchment, both inside the channel and across the surface of the peatland. Thus the use of the Saint Venant equation may not have had a large impact on the model performance. Also the parameters that were investigated in this study primarily concerned evapotranspiration, cumulative flow and low flows. The lack of flood attenuation in the model would not have influenced these results.

The space-time fields of radiation that drive the evaporative processes account for terrain effects, such as aspect, slope and shading. The energy processes in GEOtop have been extensively tested and validated (Bertoldi et al., 2006). Using GEOtop in an alpine catchment, Bertoldi et al. (2010) showed that the major factors controlling the land surface temperature in a humid climate are incoming solar radiation and land cover variability. Others have also taken the GEOtop model and added further modules to it including a snow module by Zanotti et al. (2004) and a landside probability function by Simoni et al. (2008).

The hourly meteorological data required (precipitation, atmospheric pressure, temperature, global shortwave radiation, relative humidity, wind speed and direction) by GEOtop were available from the on site meteorological tower. A field measurement study by Lewis et al. (2011) at the same site, found that the saturated horizontal hydraulic conductivity in the riparian zone within 10 m of the stream was ~10^{-7} m s^{-1} which was one to two orders of magnitude less conductive than the bog interior. The saturated horizontal conductivity was found to be approximately twice vertical hydraulic conductivity. To reflect this pattern in the soil matrix in GEOtop, the peat within 10 m of the stream was assigned the hydraulic parameters found by Lewis et al. (2011) in the riparian zone. A second zone was created between 10 and 20 m from the stream and the peat in this zone was assigned a higher hydraulic conductivity value. An incremental process of increasing the hydraulic conductivity was utilised until at the bog interior the hydraulic conductivity assigned was ~10^{-5} m s^{-1}. Vegetation details (LAI, height, root depth) were adopted from Sottocornola et al. (2009) and Laine et al (2007a).

6.4.3 Modelling Scenarios

To reflect the practice of draining peatland prior to afforestation, an artificial drainage network was created for this GEOtop application. This consisted of a series of ditches placed 8 m apart and 350 mm deep. The ditches ran orthogonal to the contours and drained into the stream running through the centre of the bog. Initial trials of this new drainage network however, found that there was little change in the WT as peatland drains have been observed by Stewart and Lance (1991) to only drawdown the WT within a meter of the drain and act mainly to intercept surface runoff. Because of model constraints it was not possible to increase the drainage network density, it was therefore decided to increase the horizontal conductivity of the peatland to represent the reduced travel time for water from the peat matrix to the drainage network that a peatland with a higher density of drains would have.

Two different future land use change scenarios were simulated in this study. The first scenario (A) involved changing the land use from natural peatland to a 10-year-old Sitka spruce forest. Scenario B was similar except the land use is a 15-year-old Sitka spruce forest. The Leaf Area Index (LAI) was adopted from Tobin et al. (2006) in a study on LAI in different ages of Sitka spruce forests in Ireland. Sitka spruce forests at approximately 15 years of age have the highest LAI and therefore deemed to have the largest impact on the rainfall runoff response, mainly from increased interception and transpiration. The changes to the model for Scenario A, were made by increasing the LAI to 4.5, root depth to 40 cm and canopy height to 250 cm. For Scenario B the LAI, root depth and canopy height were increased to 7.5, 55 and 800 cm respectively. Along with the change in land use both these scenarios also included a new drainage network to reflect the practice of draining peatlands prior to afforestation.

6.5 Results

The model was calibrated using observed data for 2007. Figure 6.2 shows the time series of observed and GEOtop modelled flows with in a Nash-Sutcliffe (Nash and Sutcliffe, 1970) efficiency of 0.87. The difference between observed and simulated flow is greatest in times of high flows where the simulated flow is larger than the observed. However, at times of high flow, the observed flows may be underestimated as flow spills overbank. The cumulative rainfall (2229 mm), observed stream flow (1925 mm) and simulated stream flow (2018 mm) are shown in Figure 6.3a. The ET of ~211 mm for 2007 is lower than both the observed value of 304 mm and the nearby eddy covariance estimate of ET of 388 mm for the hydrological year 2006/2007 reported by (Sottocornola and Kiely, 2010).

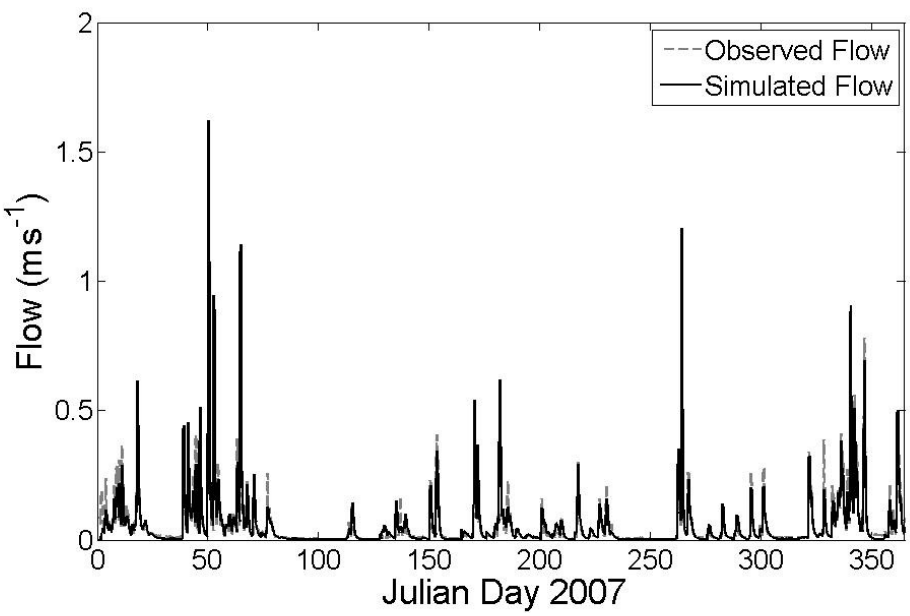

Figure 6.2 Observed and Simulated flows for 2007 at hourly intervals.

Monthly totals of simulated flow also compare well with observed values (Figure 6.3b) as do the instantaneous values of simulated and observed flow (Figure 6.3c) with the exception of the very highest flows. We believe this is due to the overbank flow for which our high observed flows are likely to be underestimated.

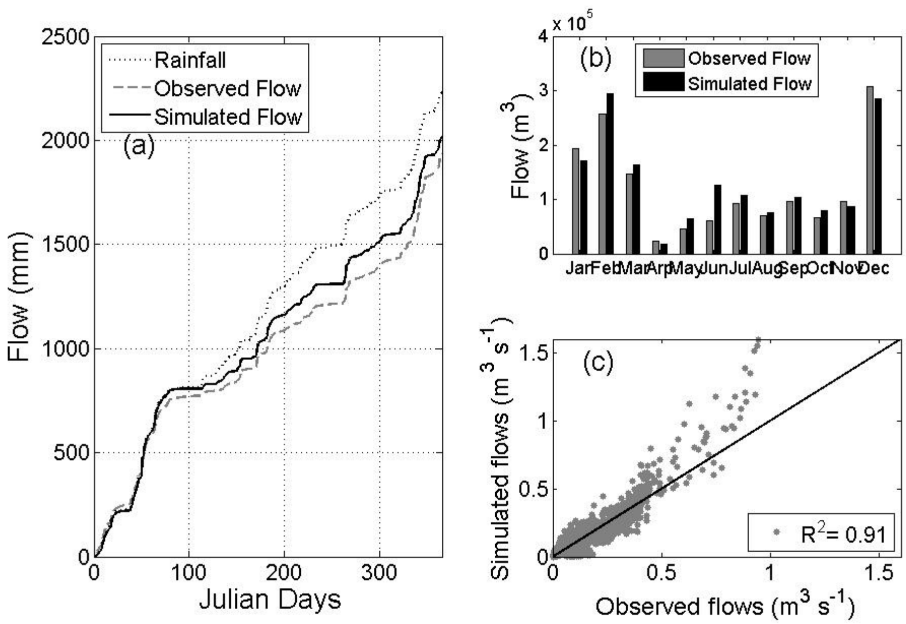

Figure 6.3 For 2007 (a) Cumulative rainfall, observed flow and simulated flow (b) Monthly observed and simulated flows (c) Observed and simulated flows.

The observed and simulated WT depths are shown in Figure 6.4. The (WT) remained close to the surface throughout the year with its lowest value of 17 cm below the ground surface in April 2007. The simulated WT levels show a wider range than the observed. This may be due to the fact that the observed WT levels were at one location in the bog, while the simulated WT is an average for the whole 76-ha bog. From Figure 6.2, Figure 6.3 and Figure 6.4 we suggest that GEOtop simulates well the hydrological process in the peatland for the calibration year.

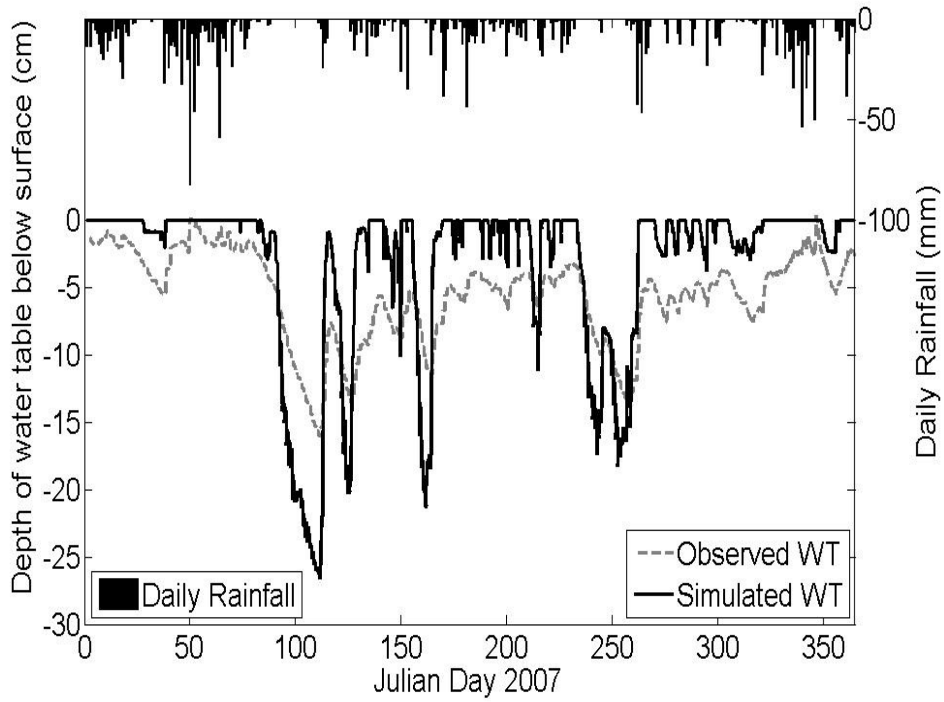

Figure 6.4 For 2007; Daily rainfall (top); observed and simulated water table.

Once the model was calibrated for 2007 it was then validated (no change to parameters) for 2008. Figure 6.5, Figure 6.6 and Figure 6.7 and Table 6.2 show the results of the 2008 simulation. The Nash-Sutcliffe efficiency of 0.89 for 2008 showed a slight improvement on the 2007 value of 0.87 while the simulated flow was slightly greater than the observed flow (Figure 6.6). The simulated WT had an annual mean value of 12 mm below the surface whereas the observed WT had an annual mean depth of 37 mm below the surface for 2008. These figures show that GEOtop provides a similar level of accuracy for the validation year 2008 as it did for the calibration year 2007. From this we suggest that the current configuration of GEOtop simulates well the hydrological processes in the Glencar blanket peatland and that we are able to proceed with a study of land use change scenarios.

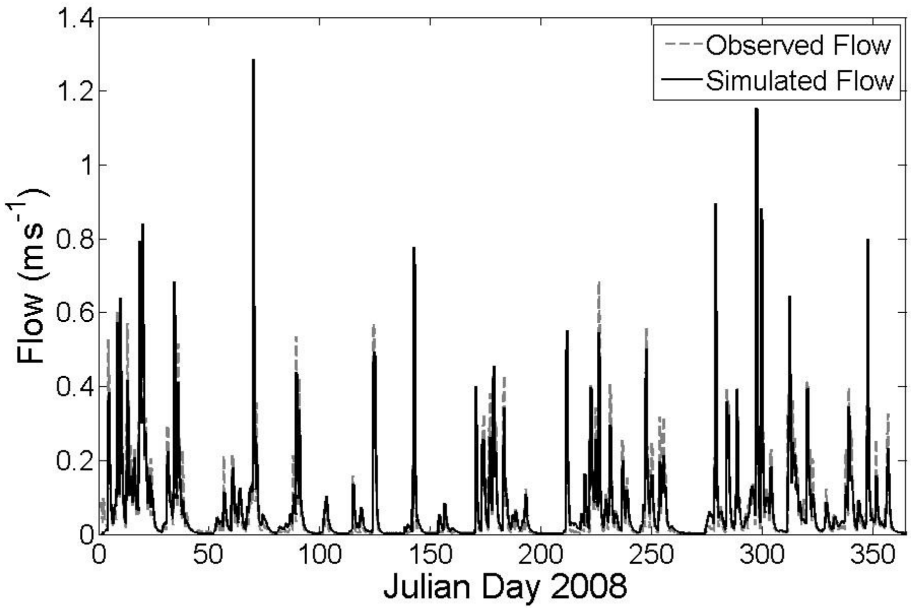

Figure 6.5 Observed and Simulated flows for 2008 at hourly intervals.

While the overall goal of this study is to assess the impact of changing the land use of a pristine peatland from its natural state to forestry on the peatland hydrology, it is not possible to do this by just changing the land use in the model, as in practice, peatlands are drained prior to planting (Holden et al., 2004).

Table 6.2 Observed and modelled scenarios; rainfall, evapotranspiration and streamflow values.

Scenario	Year	Rainfall (mm)	Evapotranspiration (mm)	Streamflow (mm)	Runoff/ rainfall ratio
Observed	2007	2229	304	1925	0.75
Calibrated Model	2007	2229	211	2018	0.82
Observed	2008	2826	421	2405	0.85
Validated Model	2008	2826	330	2496	0.88
Scenario A	2008	2826	501	2325	0.83
Scenario B	2008	2826	913	1913	0.67

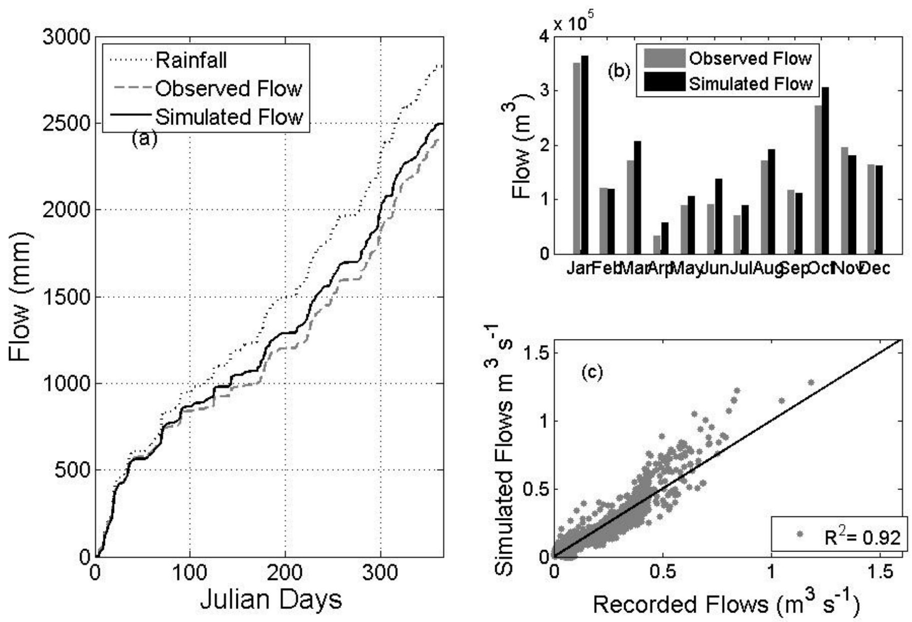

Figure 6.6 For 2008 (a) Cumulative rainfall, observed flow and simulated flow (b) Monthly observed and simulated flows (c) Observed and simulated flows.

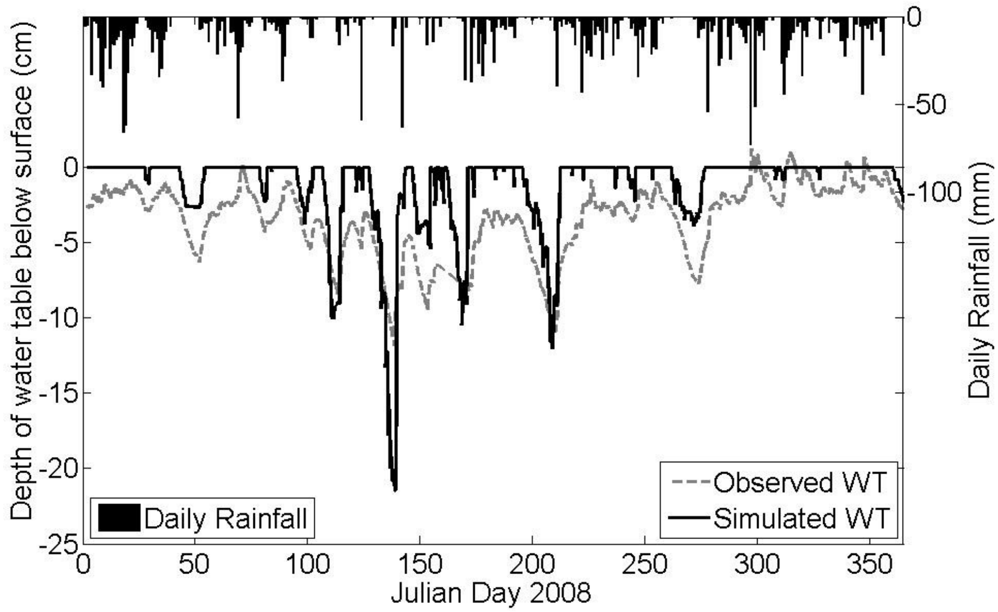

Figure 6.7 For 2008, daily rainfall (top); observed and simulated water table depth.

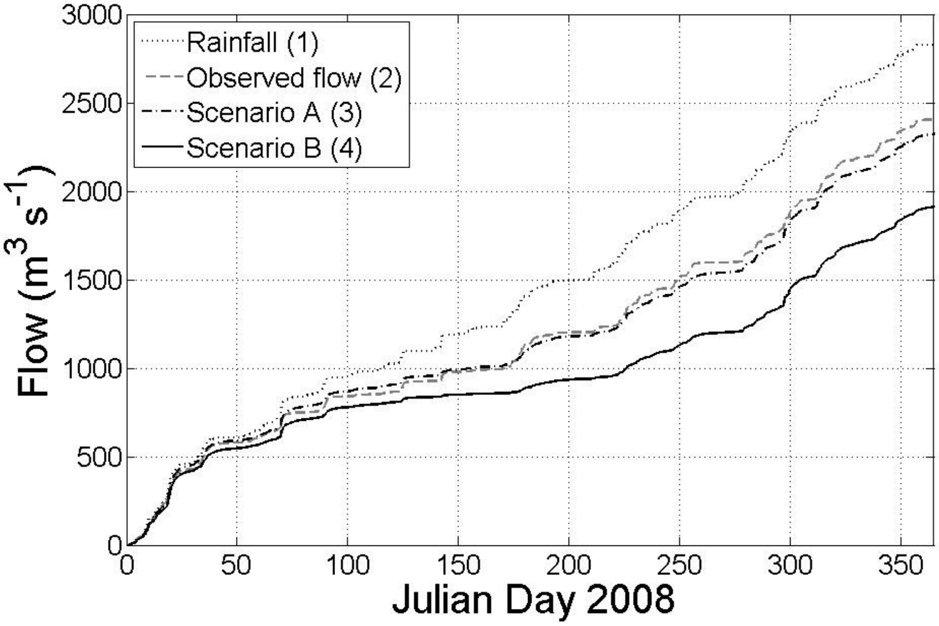

Figure 6.8 Cumulative rainfall, observed flow, simulated flow, Scenario A and Scenario B flow.

The changes in WT due to the drainage and land use change scenarios are shown in Figure 6.9. The simulated drained peatland showed a drop in WT, particularly in the summer months with the mean drained WT 150 mm below the surface and 115 mm below the observed WT. The mean observed WT depth for 2008 was 37 mm below the surface while with mean simulated WT depths for scenarios A and B were 120 and 240 mm, respectively (Figure 6.9).

The cumulative rainfall (2826 mm), observed flow (2405 mm) and simulated flows - 2325 mm for Scenario A and 1913 mm for Scenario B for 2008 are shown in Figure 6.8. From this we estimate ET from the peatland is ~420 mm in 2008 which is similar to the highest value (424 mm) of the eddy covariance estimated ET reported by Sottocornola and Kiely (2010) in the period 2002 to 2007. Scenario A showed very little change in ET from the undisturbed peatland whereas the ET for Scenario B was 925 mm, an increase of 492 mm when compared to the observed ET in 2008.

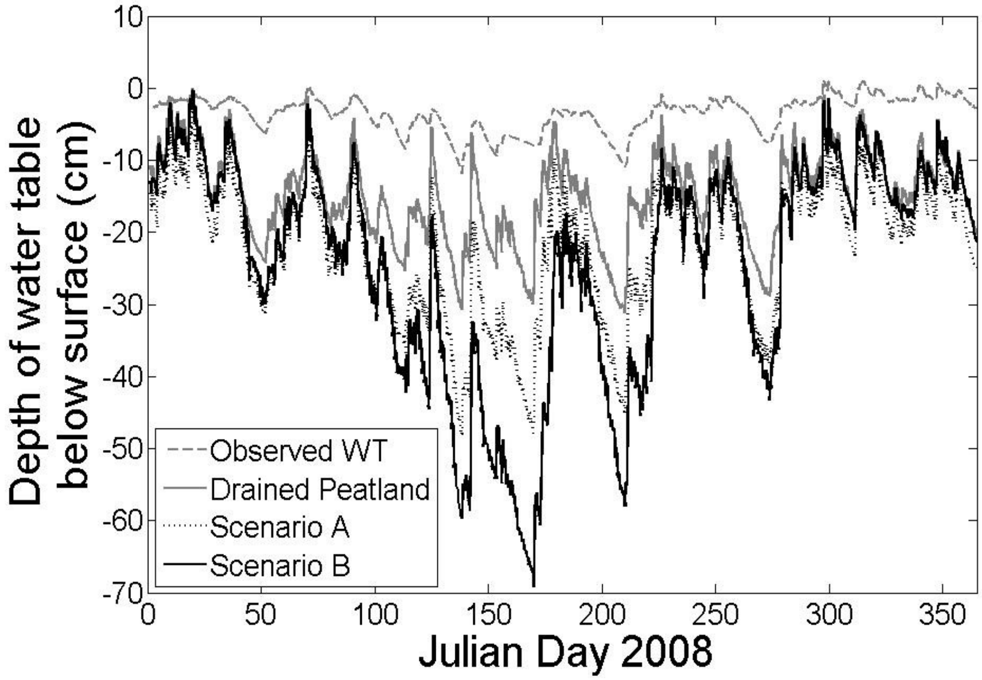

Figure 6.9 For 2008 Water table for observed, drained peat, Scenario A and B.

The total annual runoff for Scenarios A and B were respectively reduced to 96 and 80% of the observed flow. A comparison of the instantaneous flows between the observed flow and Scenario B did not show that Scenario B was consistently reduced by 80%. It showed rather, that Scenario B had on different occasions both higher and lower flows than the observed flow as illustrated in Figure 6.10a and Figure 6.10b. Figure 6.10a shows the observed flow and the flow from Scenario B from Julian day 134 to day 140, 2007. April and May 2007 were unusually dry with April being one of the driest on record at two nearby Met Eireann (Irish weather service) synoptic weather stations and below average rainfall was reported in May (http://www.met.ie/climate/monthly_summarys). This resulted in a large drop in WT. An analysis of the precipitation and stream flow on days 134 to 140 shows that the observed and simulated stream flow from Scenario B produced two very different rainfall runoff responses. The observed stream flow showed a large peak in flow whereas the simulated Scenario B flow showed only a small response to precipitation. Figure 6.10b shows that the converse occurred between days 340 and 344 where the simulated flow from Scenario B was higher than the observed flow. As there is a slight discrepancy between the peak flows and simulated flows for reasons outlined earlier, the simulation of the undisturbed peatland was also placed in Figure 6.10.

6.6 Discussion

The comparisons of observed flows and WT depths with the corresponding simulated values from the GEOtop model in the calibration year and the validation year, show that the current configuration of GEOtop is capable of reliably simulating the hydrological processes.

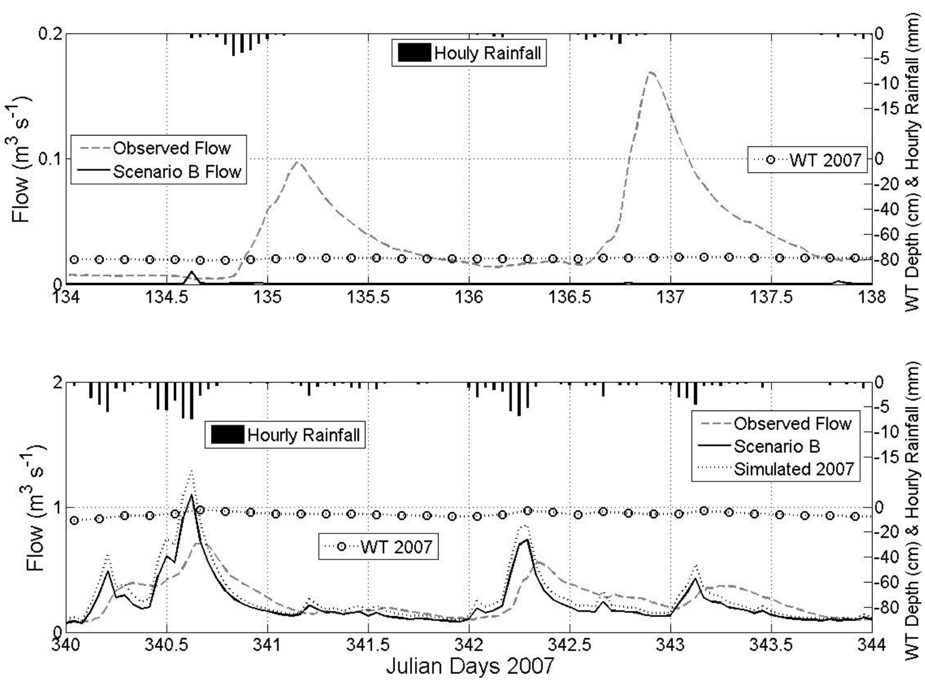

Figure 6.10 (a) for summer 2007: WT Depth, observed and modelled Scenario B flows (b) for winter 2007 WT depth, observed, simulated and Scenario B flows.

Central to this configuration for the peatland is the spatial variation of hydraulic conductivity. Areas of low hydraulic conductivity at the margins near the stream are essential in maintaining the elevated WT in the centre of the bog. For the drainage scenarios, once these areas of lower hydraulic conductivity were modified by the insertion of a drainage network the WT level in the centre of the bog fell. Given that even a slight drop in WT impacts the vegetation distribution and composition (Sottocornola et al., 2009) any disturbance of these relatively small areas of lower hydraulic conductivity will likely affect a much larger area of a bog.

While the practice of drainage prior to afforestation of peatlands will lower the WT, its depth will also be affected by the increase in transpiration and canopy interception with the change of land use from natural peatland to Sitka spruce afforestation. From Figure 6.8 we note that the simulated ET in the mature forest (Scenario B) was 492 mm greater than the virgin peatland. The ET rate of the younger forest (Scenario A) was similar to the undisturbed peatland. A study of ET by Sottocornola and Kiely (2010) at the same site found that while the ET was not water limited, and the observed ET ranged between 369 and 424 mm with an average of 394 mm of ET over a 5-year period. Sottocornola and Kiely (2010) concluded that one of the key limiting factors in ET was the lack of vascular plants. While Figure 6.9 shows that there is a drop in the WT for Scenario A, particularly in the summer months, it must also be noted that at this stage of development of a forest, the tree canopy would not cover the entire peatland. As the WT is lower than in the undisturbed peatland the ET rate from the original peatland vegetation, which has a large proportion of non vascular plants will have decreased. Thus it would appear the increase in ET under the Sitka spruce canopy is offset by the decrease in ET from the original peatland vegetation.

The estimated ET (913 mm) of the second scenario of a more mature forest (Scenario B) was 412 mm higher than Scenario A. It also must be noted that Table 6.2 shows that there was a difference between the observed and simulated (undisturbed peatland) estimates of ET in 2008 where observed and simulated ET represented 14.8% and 11.6% of observed streamflow respectively. However, the difference between the observed and simulated undisturbed peatland is much smaller than the difference between the observed ET and the ET from Scenario B, where the ET from Scenario B represents 32.2% of streamflow.

Such increases in evapotranspiration as a result of changing peatland land use from grass, mosses and bare peat to Sitka spruce have been noted by others. Studies by others in areas such as the Scottish highlands that may be considered similar in soil type and climate to our study site, have found that Sitka spruce may intercept between 28% (596 mm) (Johnson, 1990) and 52% (1514 mm) (Heal et al., 2004) of precipitation. Transpiration of Sitka spruce in Cumbria has been estimated at 12% of precipitation or 172 mm (Anderson et al., 1990). We note that our estimates of evapotranspiration from Scenario B at 32% of rainfall falls between some of the lower and higher values in the literature. Results from a study by the Institute of Hydrology (1991) into the effects of upland afforestation on water resources, suggest that evapotranspiration on a forested catchment receiving 2800 mm precipitation annually was approximately 1190 mm, which is similar to the ET value of 925 mm in our Scenario B. The increased ET in conjunction with the drainage network resulted in a WT draw down in dry periods (Figure 6.8). The

WT in scenarios A and B was on average 165 and 205 mm below the observed WT for the calendar year 2008. This is similar to the draw down noted by Bragg (2002) where the WT in a forested peatland was found to be 100 to 150 mm below that in the adjacent unforrested peatland.

As the runoff from peatlands in their natural state (in western Europe) is known to be flashy due to saturation excess overland flow, a lowering of the WT is expected to reduce the volume of runoff produced from saturation excess. This is noted in the hydrograph for a summer period in Figure 6.10a, which shows the observed flow and the simulated flow from Scenario B, for days 134 to 140. This rain event followed a dry period in April and May 2007 at the end of which the WT was observed to be at its lowest level since records began in 2003. At the start of the precipitation events of day 135, the WT of Scenario B was 698 mm lower than the observed WT of the virgin peatland and 592 mm below the drained simulation. With this lower WT and reduced precipitation reaching the ground due to canopy interception, the stream flow from the simulated forest catchment is greatly reduced. However the converse applies for the stream flow shown in winter as shown in Figure 6.10b. Prior to day 340 there had been frequent rain events which had resulted in a much higher WT both in the simulated and observed cases. A total of 53 mm fell on day 340, resulting in a higher simulated than observed flow. As the observed WT was 19 mm below the surface and Scenario B was just 25 mm below the surface, we consider that both simulations produced saturation excess overland flow. However, the forested simulation had a more extensive drainage network which was able to convey any surface runoff to the catchment outfall more rapidly resulting in a higher peak flow. Such a phenomenon has also been noted by others: with Ahti (1980) finding that increasing density of drainage ditches increased peak flows in a Finish peatland; and model simulations by Iritz et al. (1994) noting that peak flows may be increased by forest drainage when the WT is close to the surface. The studies of Anderson et al. (2000) and (Ballard et al., 2011b) also drew similar conclusions.

This phenomenon of increasing the peak flow following afforestation of peatlands may have implications in larger catchments where afforestation may be considered as a flood mitigation measure. Likewise the reduction of flows from peatlands throughout the summer and in particular after dry periods may be of concern for water resource managers.

6.7 Conclusion

With regard to applying the hydrological GEOtop model to an Atlantic blanket peatland we found that GEOtop was suitable for the purposes of modelling the hydrological processes. Central to this was the input of the spatial variation of the saturated hydraulic conductivity. Peat with a lower hydraulic conductivity at the margins results in elevating the WT depth in the centre of the bog which in turn resulted in saturation excess overland flow during precipitation events. It is also clear from the scenario modelling that afforestation and its associated drainage can change the hydrological response of this pristine peatland catchment. While the evapotranspiration rates from a young Sitka spruce catchment were similar to the existing pristine peatland catchment, a semi-mature Sitka Spruce forest resulted in an increase in evapotranspiration of 492 mm through increased transpiration from the canopy and interception. This increase in evapotranspiration was particularly noticeable in summer and resulted in an increase in depth of the WT and reduction in stream runoff. However, in winter, following periods of heavy rainfall, the WT depth approached that of an unforested drained peatland. This shallow WT depth in combination with a drainage network results in an increase in peak flow in times of heavy rainfall. This suggests that there is limited or no benefit to flood attenuation from peatland afforestation during winter periods when the WT is high while the converse applies to summer flows where the rainfall runoff was reduced in dry periods.

6.8 Acknowledgments

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Chapter 7

A model study of the impact of rainfall rates on erosion from a grassland catchment

Title:

A model study of the impact of rainfall rates on erosion from a grassland catchment.

Running title:

Suspended sediment yield in a grassland catchment

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Key words: suspended sediment yield, rainfall runoff model, rainfall intensity

7.1 Abstract

Soils are a vital non-renewable resource that provide a range of economic and environmental services; however, soil degradation from erosion has been identified as a serious problem globally. The severity and extent of erosion on European soils has increased in the last 50 years following agricultural intensification. We observed a suspended sediment yield (SSY) that ranged from 0.13 and 0.92 t ha^{-1} yr^{-1} for 2002 and 2003 at the outlet of a (15 km^2) Irish grassland catchment. This magnitude of erosion is at the lower end of the global scale. With climate change, precipitation amounts and intensities are expected to change in this region. Our search is then to examine the impact of changing precipitation on erosion yields. We used the hydrological model GEOtop with a new LISEM erosion module to simulate the historical SSY. Individual rain event simulations as well as annual erosion estimates are possible with the new model. The simulations compared well to observed suspended sediment exports, verifying that the new model captures the key erosion processes. A modelling analysis of a number of precipitation events was carried out where rainfall intensities were varied while the total precipitation was kept constant. It was found that SSY is sensitive to rainfall intensity with SSY increasing with increasing rainfall intensity. A simulated precipitation event with a total of 34.4 mm of rain resulted in a SSY of 4.07 and 9.28 kg ha^{-1} for rainfall intensities of 2.5 and 7 mm hr^{-1} respectively. A similar analysis of a 25.2 mm precipitation event resulted in SSY of 8.5 and 14.8 kg ha^{-1} for rainfall intensities of 2 and 7 mm hr^{-1} respectively. Higher precipitation leading up to this event is likely to have resulted in greater overland flow and thus SSY. We found a linear relationship between rainfall intensity and SSY.

7.2 Introduction

Soils are a vital non-renewable resource that provide a range of economic and environmental services. These include: the support of food and fibre production, the control of the fate of water in the hydrologic system; the loss, purification, contamination and utilisation of water; the provision of habitat for organisms; and storage for carbon in the form of organic matter. Fertile soil is essential to food security and human health and therefore, must be protected (CEC, 2006). Soils have long endured degradation pressures or threats as a result of natural and human factors. Indeed, many societies have floundered as a result of unsustainable soil management practices. This recurring phenomenon has recently been charted through history by Montgomery (2007), beginning with the first farmer in the Tigris and Euphrates river basins, through the bronze, iron and industrial ages and up to contemporary industrial farming and smallholder slash and burn practices. In each case agriculture expands on good land, which in turn fuels population growth. This is followed by the expansion of agriculture onto marginal land with a consequent increase in soil erosion and a decline in agricultural production and often societal collapse and emigration. Although advances in crop productivity through biotechnology are possible, Montgomery (2007) argues that soil is a scarce and limited resource and that, on average, we are currently losing soil at least 20 times faster than it is being replaced through natural formation processes.

There are several threats to sustainability of soils including: erosion; loss of organic matter; compaction; landslides; urbanisation or surface sealing; desertification; land use change and climate change and agricultural intensification (Boardman and Poesen, 2006; Claessens et al., 2007; Doyle et al., 2000; Dykes and Warburton, 2008; Fu et al., 2006; Kurz et al., 2006; Lal, 2003; McGrath and

Lynch, 2008). Erosion, either by water or wind, is a global problem and very much dependent on natural and man-made features including: topography and slope; soil type and properties; weather and climate; land use and land-use change and agricultural activities (Ali and De Boer, 2010; de Vente et al., 2008; Lu et al., 2005; Nearing et al., 2004; Nunes et al., 2009; Verstraeten and Poesen, 2001). In this paper our interest is in water initiated soil erosion.

We note that there are four essential forms of water erosion: 1) inter-rill erosion - the movement of soil by rain splash and its transport by this surface flow (DeRoo et al., 1996a); 2) rill erosion - erosion by concentrated flow in small rivulets (Boardman and Poesen, 2006); 3) gully erosion - erosion by runoff scouring large channels (e.g. deeper than 30cm) (Poesen et al., 2006) and 4) streambank erosion - erosion by rivers or streams cutting into banks (Prosser et al., 2000). Types 1 and 2 tend to occur on normal hillslopes, while types 3 and 4 occur in well developed channels. Some land use and management practices can lead to precipitation induced soil erosion, which in turn can deteriorate the remaining physical, chemical and biological soil properties and as a consequence reduces soil productivity. Van Oost and Govers, (2006) showed that tillage erosion rates can exceed 10 t ha^{-1} yr^{-1}, especially on fields with complex and steep topography and these rates are at least of the same order of magnitude as average water erosion rates reported for hilly cropland in western Europe. Cerdan et al. (2006) noted that land uses with the highest percentage of bare soil, either spatially (wide inter-row spacing and low leaf cover, e.g. vineyard or maize) or temporally (long inter-crop duration, e.g. maize or spring crop) have the highest soil erosion rates with reported erosion rates for vineyards and maize being 24.96 and 13.95 t ha^{-1} yr^{-1} respectively. Evans (1996) estimated that erosion significantly and adversely affected 40% of arable soils in the UK, with these soils losing more than 25% of their agricultural productivity. Grazhdani (2006) noted that poorly built logging roads in forestry operations lose soil by erosion of the road surface and the drainage ditches or the soil exposed by roads cut into hillsides. Off-site impacts of erosion include sedimentation of rivers and lakes, watercourse pollution and eutrophication, silt build up in rivers with its consequent impact on young aquatic life, and perturbed geomorphological functions of river systems, (Owens et al., 2005). Floods have been found to dominate erosion (Lopez-Tarazon et al., 2009) where rainfall intensity, soil moisture and infiltration capacity (Romkens et al., 2002; VanDijk and Kwaad, 1996), have been identified as central to erosion rates. Given the number of variables involved in soil erosion, it is no surprise to see a large variation in reported erosion rates in Europe (see Table 7.1). Verstraeten and Poesen (2001) reported erosion rates to vary from <0.5 to >20 t ha^{-1} yr^{-1}, the highest being associated with tillage practices on steep hillslopes with the lowest rates on flat grassland areas.

Global climate has changed notably over the past century and this change is expected to continue in the future (IPCC, 2007; McGrath and Lynch, 2008). In many areas the seasonal distributions of rainfall have changed, with significant implications for patterns of vegetation growth and hence for soil erosion (Nearing et al., 2005a). In Ireland, Kiely (1998) found that annual rainfall has increased by approximately 10% since 1975 by comparison with pre-1975, with the highest monthly increases being in winter (March and October).

Table 7.1 Erosion rates from a number of different land use types in Europe.

Land use	Location	Mean Rainfall (mm)	Erosion (t ha^{-1} yr^{-1})	Reference
Bare soil	European average	674	23.4	(Cerdan et al., 2006)
Vineyeard	European average	629	19.97	(Cerdan et al., 2006)
Cereal	European average	629	2.10	(Cerdan et al., 2006)
Grassland	European average	623	0.29	(Cerdan et al., 2006)
Forest	European average	483	0.1	(Cerdan et al., 2006)
Grassland	UK	580	<0.1	(Fullen, 1998)
Grassland	Swiss Alps	1516	6-37	(Konz et al., 2009)
Grassland	UK	1050	0.84-1.21	(Bilotta et al., 2010)

To understand the spatial and temporal changes of soil erosion and to enable mitigation measures, modelling studies of erosion processes and their quantification are required. Many models have been developed and applied for predicting soil erosion and these range from the simple empirical to the more complex physically-based models. By far the most widely used long term annual estimators of erosion (soil loss) are the Universal Soil Loss Equation (USLE), its modified version (MUSLE), the revised version (RUSLE), and the Water Erosion Prediction Program (WEPP), (Kinnell, 2010; Lu et al., 2005). USLE has more recently been also used to predict event erosion (Kinnell, 2010). For modelling the spatial soil erosion risk, these models are often integrated with GIS and Geostatistics techniques (Ozcan et al., 2008). Also, many studies have used RUSLE (Ito, 2007; Smith et al., 2007) and its related models (e.g. EPIC, see Izaurralde et al., (2007) to simulate the impacts of soil erosion and deposition on the carbon cycle. Each of the above empirical models have strengths and weaknesses and have been applied across a broad range of landscape types (Hancock et al., 2010; Kinnell, 2010).

A further development has been the evolution of catchment scale models which use digital terrain models (DTMs) to detail the topography, soil type, etc. Amongst these, are SIBERIA and CAESAR as described by Hancock et al. (2010). The latter are used to quantify soil erosion rates and processes subject to the action of rainfall and runoff and can estimate erosion and deposition as well as global erosion or sediment yield. Such models employ spatially variable hydrological and erosion parameters, the spatial distribution of soil type and particle size, and predict erosion/deposition at the pixel scale and at the catchment scale. Other erosion models include SHETRAN (Ewen et al., 2000) which is a physically based hydrological-erosion model suited to the river basin scale. It is specifically designed to model transport of chemicals and sediment thorough various pathways on a continuous basis. A simpler model that also operates at the catchment scale is the Factorial Scoring Model (FSM) (Verstraeten et al., 2003). It predicts the sediment yield of a catchment, based on a nonlinear equation involving the catchment area, topography, vegetation, gullies, lithology and slope. The rule based STREAM model (Cerdan et al., 2002) was designed for areas that have a clear surface sealing process where a crust is formed on loose soil after tillage. The MEFIDIS (Nunes et al., 2005) model was developed to simulate the consequences of climate and land-use changes for surface runoff and erosion patterns during extreme rainfall events. The model relies on physically based runoff and soil detachment equations, dividing the simulation area into spatial homogeneous units and using a dynamic approach for runoff and suspended sediment distribution. The LISEM model (DeRoo et al., 1996a) is a physically based erosion model that runs at the event and catchment scale. LISEM runs in a GIS environment and modelled erosion is comprised of splash detachment and flow detachment from over land flow in rills. The transport processes are also simulated with soil transport and deposition carried out on a cell by cell basis. Flow routing is modelled using a four-point finite difference solution of the kinematic wave and Manning's equation.

The lack of knowledge on soil erosion across the EU has been highlighted by Van-Camp et al. (2004), with Bilotta et al. (2010) noting that most UK studies concentrate on lowland arable or upland areas. This also appears to hold true in Ireland with only a small number of field studies at the small catchment scale, despite the importance of agriculture to the Irish economy (Dept. of Agriculture, 2010), where by 2020 the output from agriculture, fisheries and forestry is projected to grow by 33% from the 2007-2009 average. Studies by Lewis (2003), who measured suspended solids export yield (SSY) from a nested set of small grassland catchments in the southwest of Ireland found that the SSY export ranged from 0.073 to 0.136 t ha^{-1} for 2002. During a 1993-1994 EPA investigation (Tunney et al., 2000) the total SSY exports were estimated from a nested set of grassland catchments (2.28 km^2, 14.91 km^2, 88 km^2) in the southwest of Ireland, ranged between 0.127 and 0.24 t ha^{-1} yr^{-1}. A continuous monitoring programme carried out by Harrington and Harrington (2011) found a SSY of 0.256 t ha^{-1} yr^{-1}on the 105 km^2 Owenabue river catchment, in the south of Ireland. Their flux analysis revealed that 85% of the total annual flux was transported over 10% of the year and 69% of the flux over 5% of the year. While SSY is the "instream" deliverable of erosion, SSY may be considered a proxy for erosion, in the absence of plot or field measurements. The values reported for rates of erosion from Irish grasslands are at the lower end of the international scale, (Table 7.1), most likely due to the low rainfall intensity, the relatively flat nature of Irish grasslands and the lack of bare soil. This is aided by the fact that grasslands cover approximately 90% of Irish agricultural lands with tillage accounting for the remainder. While Ireland receives annual rainfall amounts that range from approximately 750 mm in the east to about 1600 mm in the west, the lack of intense short duration convective storms also limits the amount of erosion.

Given the importance of soils for agricultural productivity, and the large growth rate expected in the global populations which is placing increasing demand on agricultural productivity, any loss of soil cannot be taken lightly. As a key driver of soil erosion is precipitation, whose patterns are altering due to climate change, this study aimed to investigate the effect of increasing rainfall intensity on soil erosion from an Irish grassland.

The aim of this study was to: (1) calibrate and validate a distributed hydrological model (GEOtop) in a grassland catchment in the south of Ireland; (2) modify the original hydrological model (GEOtop) to include an erosion module based on the LISEM architecture (DeRoo et al., 1996a); and (3) investigate the effect of increasing rainfall intensities on soil erosion losses.

7.3 Materials and methods

We used the 15 km^2 Dripsey grassland catchment in southwest Ireland (Figure 7.1). It is a research catchment, sited approximately 25 km northeast of Cork city (Latitude 51°59′N, Longitude 8°45′W). A small stream drains the catchment from north to south. It has an elevation range of 60 to 200 masl (meters above sea level). The climate is temperate maritime with mean annual air temperature of 10.2°C and an annual average rainfall of 1470 mm. Rainfall intensity is generally low with the highest rainfall intensity observed over the study period being <15 mm hr^{-1}.

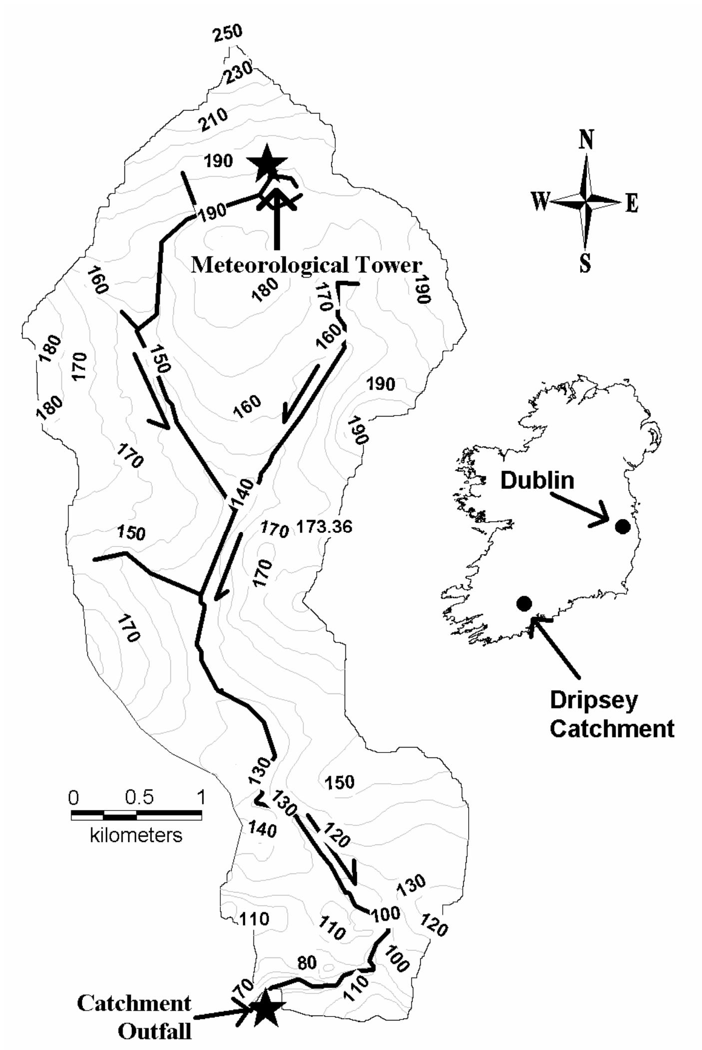

Figure 7.1 Dripsey Catchment (15 km^2) located in the southwest of Ireland. The catchment is drained by the Dripsey river which rises at the top of the catchment just below 200 m and drops over 8.43 km to 60 m at the catchment outfall. The meteorological tower is located in the north of the catchment at an elevation of 192 masl (meters above sea level).

The land-cover is almost 100% grassland and is used for beef and dairying agriculture. The soils are gleys and podzols and are described as impeded drainage at the upper elevations to free drainage at the lower elevations. Measurements of meteorological variables at a meteorological tower at the top of the catchment (elevation 192 masl) have been ongoing since 2001. They include: air temperature (T_a) and relative humidity (RH) (HMP45A; Vaisala, Helsinki, Finland); net radiation (CNRI net radiometer Kipp & Zonen, Delft, The Netherlands); 2-dimensional wind speed (W_s) and direction (W_d) (RM Young). Rainfall was measured using a CS-ARG100 rain gauge. At the catchment outfall (elevation 60 masl), the stream height is continuously recorded from which stream flows are determined via a rating curve built up over several years.

7.3.1 GEOtop hydrological model

The hydrological model used in this study is GEOtop (Rigon et al., 2006). The original version of GEOtop includes a rigorous treatment of the core hydrological processes (e.g. unsaturated flow, saturated flow, transport surface energy balances and stream flow generation/routing). The energy process was extensively tested and validated by Bertoldi et al. (2006). Recently GEOtop has been extended to include treatment of shallow landslides (Simoni et al., 2008). GEOtop is a distributed hydrological model and simulates the complete hydrological balance in a continuous way during a whole year and is driven by geospatial data (e.g. topography, soil type, vegetation and land cover). It estimates rainfall-runoff, evapotranspiration and provides spatially distributed outputs as well as routing water through stream and river networks (Rigon et al., 2006).

The open source nature of GEOtop made it possible to modify the original code to include an erosion module. For the development of the erosion module in GEOtop we have adopted the LISEM model (DeRoo et al., 1996a; DeRoo et al., 1996b) and have developed a module in GEOTOP for the online calculation of distributed erosion, sediment transport, and deposition rates. The LISEM model has been used and tested extensively over the past decade, with over 40 applications of the model published and over 100 papers published that cite the original model presentation (Boer and Puigdefabregas, 2005; Hessel et al., 2003; van Dijk et al., 2005). In the original version which runs in a GIS environment, modelled erosion is comprised of splash detachment and flow detachment from over land flow in rills. The transport processes are also simulated with soil transport and deposition carried out on a cell by cell basis. Since we are focused on impacts on soil resources in this study we do not consider in-channel erosion.

7.3.2 LISEM Erosion model

The LISEM soil erosion model (DeRoo et al., 1996a) is a physically based model that runs at the event and catchment scale. In the original version which runs in a GIS environment, modelled erosion is comprised of splash detachment and flow detachment from over land flow in rills. It is noted that we do not include bank erosion. Splash Detachment (D_s) is simulated as function of soil aggregate stability (*Aggrstab*), rainfall kinetic energy (*Ke*), the depth of the surface water layer and net precipitation, see eqn 7.1.

$$D_s = \left(\frac{2.82}{Aggrstab} Ke * e^{-1.48(h)} + 2.96 \right) P_{net} \left(\frac{dx^2}{dt} \right) \qquad \text{(eqn 7.1)}$$

where D_s is splash detachment (g s^{-1}), h is the depth of the surface water layer (mm), P_{net} is net rainfall which is rainfall less interception (mm), dx is the size of an element (m); dt is the time increment (s) and the rainfall kinetic energy (J m^2) of the rain drops is given by eqn 7.2.

$$Ke = 8.95 + 8.44 \log(P_{net}) \qquad \text{(eqn 7.2)}$$

Flow detachment (D_f) (kg m^{-3}) and deposition (D_p) (kg m^{-3}) are determined from eqn 7.3a and eqn 7.3b.

$$D_f = (T_c - C) y v_s w \qquad \text{(eqn 7.3a)}$$

$$D_p = v_s w (T_c - C) \qquad \text{(eqn 7.3b)}$$

where w is rill width of flow (m), v_s is the settling velocity of particles (m s^{-1}), transport capacity (T_c) is defined by eqn 7.5 (kg m^3) and y, an efficiency coefficient (eqn 7.4) is dependent on grain shear velocity and cohesion of the soil (Morgan et al., 1992; Rauws and Govers, 1988).

$$y = \frac{\mu_{g\,\min}}{\mu_{gcrit}} = \frac{1}{0.89 + 0.56COH} \qquad \text{(eqn 7.4)}$$

where μ_{gmin} is the minimum value required for critical grain shear velocity (cm s^{-1}); μ_{gcrit} is the critical grain shear velocity for rill initiation (cm s^{-1}); and COH is the cohesion of the soil at saturation (kPa). The transport capacity (T_c) is dependent on Cl and Dl which are empirically derived coefficients (Govers, 1990); S (m m^{-1}) the slope gradient, V (m s^{-1}) the mean flow velocity and ρ_s the soil density (kg m^3).

$$T_s = c_l (S - 0.4)^{Dl} \rho_s \qquad \text{(eqn 7.5)}$$

The fundamental parameters driving the LISEM erosion model are COH, T_c and *AGGRSTAB*, see Table 7.2. Cohesion was derived taken from the EUROSEM soil erosion model (Morgan et al., 1998) and is valid for soils at saturation. This is relevant for the GEOtop erosion module, as erosion from overland flow by its nature is associated with saturated soil. The *AGGRSTAB* parameter was calibrated using field experiments and is also associated with saturated soils. The final parameter *Tc* is not connected directly with soils but does also generally occur with saturated soil.

Table 7.2 Erosion parameters of the LISEM model

Form of Erosion	Parameter	Definition	Units	Soil status	Estimated from
Flow Detachment	COH	Cohesion	kPa	Soil at saturation	Measured in the field using a torvane. Taken from EUROSEM (The European soil erosion model) Morgan et al., 1998)
Splash Detachment	AGGRSTAB	Soil aggregate stability	$J\ m^{-2}$	Not mentioned but assumed to be at saturation	Taken from field experiments, no further details given by De Roo et al., (1996)
Transport Capacity	Cl and Dl	Coefficients used in transport capacity	-	NA, overland flow transport capacity	Empirically derived coefficients taken from Grovers (1990)

The LISEM erosion module was coded into GEOtop and is described in detail by Zi et al. (in preparation). The hydrological parameters required to drive LISEM were provided by GEOtop on a cell by cell basis. During testing of the erosion module a number of issues were brought to light. Overland flow from either saturation excess or infiltration excess is assumed to spread over the entire cell by GEOtop. Due to computation restraints it was necessary to limit the grid size so as to make runtimes practical. This resulted in a cell size of 50 by 50 m. As the overland flow occurs over the full grid it leads to very slow shallow flows. A number of modifications to the GEOtop code were necessary to work around this. While overland flow can be generated over the entire cell it is automatically transferred to a single rill of a width *w* (m) to generate a more realistic depth of overland flow *h* (mm).

7.3.3 Suspended sediment data

A study of nutrient export from the Dripsey catchment was carried out over a two year period, 2002 and 2003 by Lewis (2003) which formed part of a larger EPA investigation into Eutrophication from agricultural sources (EPA, 2006). Stream flow was monitored continuously at 30-min intervals over the two years, 2002 - 2003. Composite, flow-weighted water samples at the catchment outlet were collected in flow-actuation mode with an ISCO 6712 auto-sampler with the intake approximately 0.25 m above the streambed. The composite sampling time ranged from about two hours at high flow to two days at low flow. These samples were analysed for suspended sediment (SS) among other water chemistry parameters and covered 42% and 21 % of the years 2002 and 2003 respectively. The SSY from the catchment was estimated from the product of the SS concentrations and stream flow. As the SS sampling regime did not cover the entire year a relationship was built between the measured SS concentrations and streamflow. This relationship enabled an estimation of SS concentration from known flows when there were no SS concentrations available.

7.3.4 Description of model scenario

In an effort to investigate the effect of varying rainfall intensity on erosion, a number of particularly large precipitation events in 2002 of 30.2, 34.4 and 25.2 mm on days 319, 324 and 355 respectively were studied in greater detail. Three different modelling scenarios, A, B and C were created using the precipitation events for days 319, 324 and 355, respectively. Six different simulations were then created for each scenario by varying the rainfall intensity of the precipitation events but keeping the rainfall totals the same. Rainfall intensities of 1.5, 2.5, 4, 5 6 and 7 mm hr^{-1} were chosen for each scenario (see Table 7.3). All other meteorological parameters along with vegetation and soil hydrological properties remained unchanged in the new model scenarios and the model was run for the entire year of 2002 as a spin up to ensure all model parameters would be the same for all model simulations.

7.4 Results

The observed suspended solids concentrations ranged from 0.08 to 218 mg l^{-1} (see Figure 7.2), with the higher concentrations generally observed in periods of higher flows particularly in the winter. When flows of similar magnitudes are binned together as in Figure 7.3, we note that the corresponding SS concentrations show an increase with flow. For low flows (in 2002) under 0.1 m^3 s^{-1} the mean SS concentration was ~ 0.7 mg l^{-1}. For the high flows varying between 4 and 6 m^3 s^{-1} the mean SS concentration was ~ 40 mg l^{-1}. Given that SSY is the product of flow and SS concentrations, Figure 7.3 demonstrates that the SSY of the Dripsey catchment increases rapidly in times of higher flow.

We used the year 2002 to calibrate the hydrological-erosion model, as we had detailed meteorological, streamflow and suspended sediment measurements. During the calibration process parameters such as hydraulic conductivity, leaf area index, root depth and the van Genuchten (1980) parameters, α and n were altered to give the closest fit of simulated flow to observed flow**.** Figure 7.4 shows a comparison of simulated and observed flow for 2002 with a Nash-Sutcliffe (Nash and Sutcliffe, 1970) efficiency of 0.74. It must also be noted that the rating curve at the catchment outlet is less accurate at high flows due to the lack of high flow stream velocity measurements.

Table 7.3 Rainfall patterns for the three modelled scenarios. The total volume remained constant while the intensity was modified.

Time (hr)	Scenario A							Scenario B							Scenario C						
	Original Precip (mm hr-1)	Modified rainfall patterns						Original Precip (mm hr^{-1})	Modified rainfall patterns						Original Precip. (mm hr^{-1})	Modified rainfall patterns					
		1.5	2.5	4	5	6	7		1.5	2.5	4	5	6	7		1.5	2.5	4	5	6	7
	0	0	0	0	0	0	0	1.2	1.2	1.2	1.2	1.2	1.2	1.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
	0.2	0	0	0	0	0	0	2	1.5	2.5	4	5	6	7	0	0	0	0	0	0	0
1	0.2	0	0	0	0	0	0	3	1.5	2.5	4	5	6	7	0	1.2	0	0	0	0	0
2	0.4	0	0	0	0	0	0	3.4	1.5	2.5	4	5	6	7	0	1.5	0	0	0	0	0
3	0.4	0	0	0	0	0	0	3.8	1.5	2.5	4	5	6	7	0	1.5	0	0	0	0	0
4	0.2	0.7	0	0	0	0	0	3.6	1.5	2.5	4	5	6	6.2	0	1.5	0	0	0	0	0
5	0	1.5	0	0	0	0	0	4.4	1.5	2.5	4	5	4.2	0	0.6	1.5	0	0	0	0	0
6	2.2	1.5	0	0	0	0	0	4.4	1.5	2.5	4	4.2	0	0	0.4	1.5	0	0	0	0	0
7	2.4	1.5	0	0	0	0	0	4.2	1.5	2.5	4	0	0	0	0.2	1.5	0.2	0	0	0	0
8	2.4	1.5	0	0	0	0	0	3.6	1.5	2.5	2.2	0	0	0	0.2	1.5	2.5	0	0	0	0
9	2	1.5	0.2	0	0	0	0	1.8	1.5	2.5	0	0	0	0	0.6	1.5	2.5	0	0	0	0
10	0.4	1.5	2.5	0	0	0	0	0	1.5	2.5	0	0	0	0	1.8	1.5	2.5	0	0	0	0
11	0.2	1.5	2.5	0	0	0	0	0	1.5	2.5	0	0	0	0	3.8	1.5	2.5	1.2	0	0	0
12	1.4	1.5	2.5	0.2	0	0	0	0	1.5	2.5	0	0	0	0	4.2	1.5	2.5	4	0.2	0	0
13	1.8	1.5	2.5	4	0.2	0	0	0	1.5	1.7	0	0	0	0	3.4	1.5	2.5	4	5	1.2	0
14	1.4	1.5	2.5	4	5	2.2	0	0	1.5	0	0	0	0	0	2.6	1.5	2.5	4	5	6	4.2
15	2.6	1.5	2.5	4	5	6	6.2	0	1.5	0	0	0	0	0	4.8	1.5	2.5	4	5	6	7
16	1.4	1.5	2.5	4	5	6	7	0	1.5	0	0	0	0	0	1.6	1.5	2.5	4	5	6	7
17	0.6	1.5	2.5	4	5	6	7	0	1.5	0	0	0	0	0	1	1.5	2.5	4	5	6	7
18	0	0	0	0	0	0	0	0.2	1.5	0.2	0.2	0.2	0.2	0.2	0	0	0	0	0	0	0
19	-	-	-	-	-	-	-	0	1.5	0	0	0	0	0	-	-	-	-	-	-	-
20	-	-	-	-	-	-	-	0	1.5	0	0	0	0	0	-	-	-	-	-	-	-
21	-	-	-	-	-	-	-	0	1.5	0	0	0	0	0	-	-	-	-	-	-	-

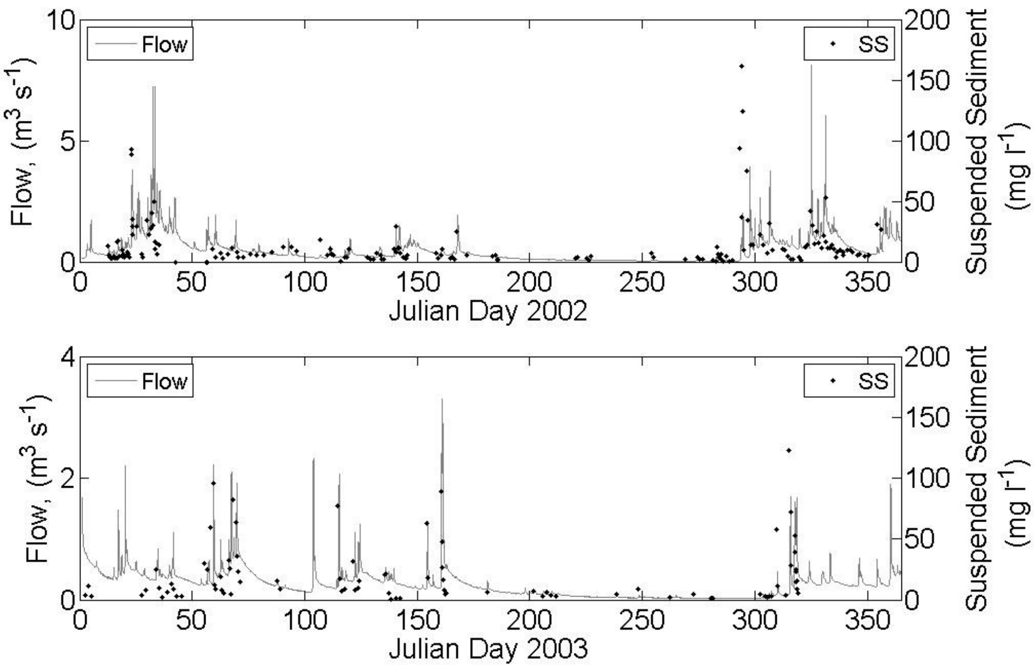

Figure 7.2 Suspended Sediment (SS) and flow for 2002 and 2003 at the outlet of the catchment.

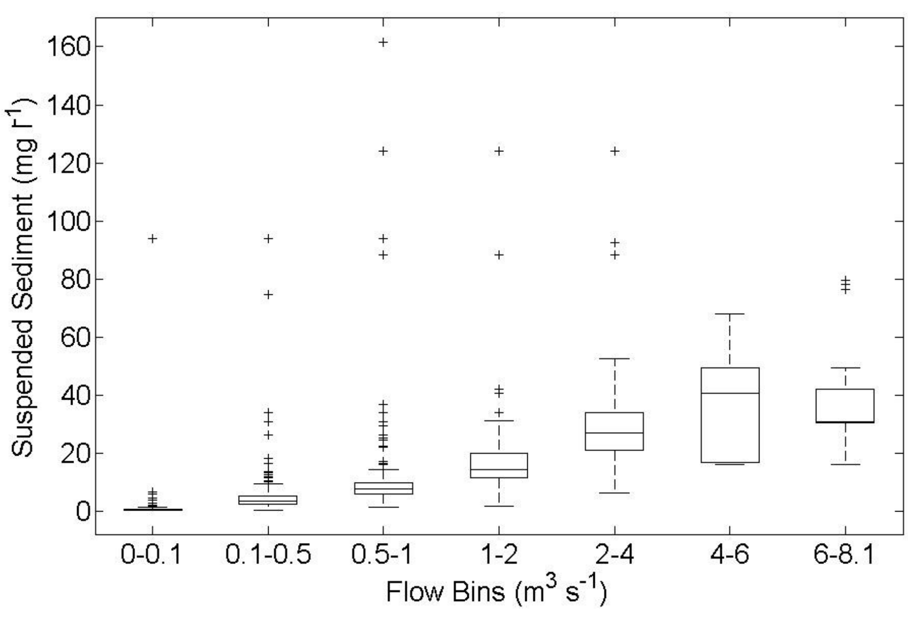

Figure 7.3 Flows for 2002 separated into seven different flow bins with the corresponding observed suspended sediment concentrations demonstrating a trend of increasing suspended sediment concentrations with increasing flow.

The cumulative rainfall, observed stream flow and simulated stream flow are presented in Figure 7.5a. The total observed rainfall for 2002 was 1820 mm which was the highest rainfall recorded at this site since records began in 2001 (mean annual rainfall of 1470 mm, 2000 to 2010). There is good agreement in Figure 7.5b between the annual simulated (1330 mm) and observed annual flow (1268 mm). The residual of ~550 mm is an estimate of annual catchment average evapotranspiration.

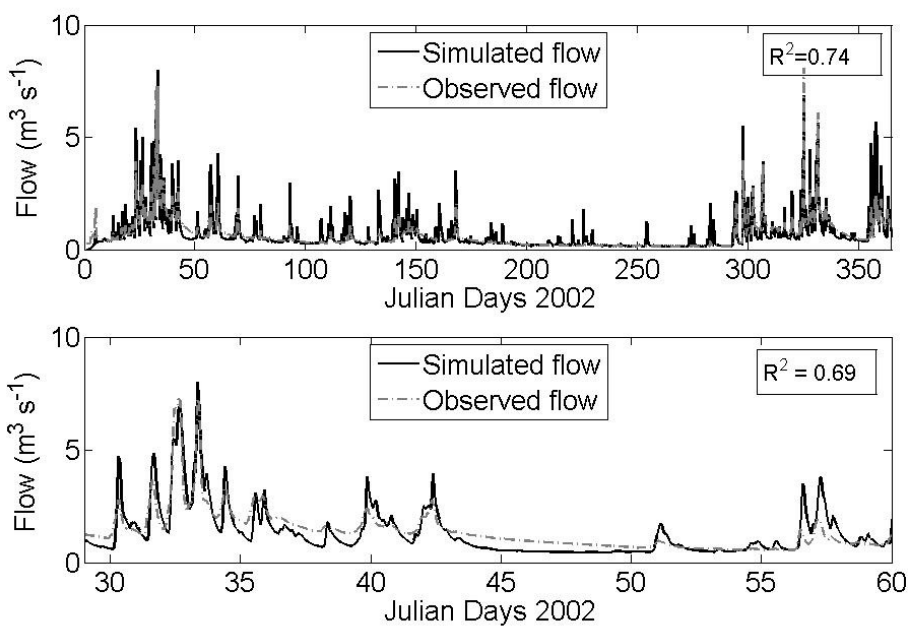

Figure 7.4 (a) Observed and simulated flows for the calibration year 2002 (b) observed and simulated flows from Julian day 29 to 60.

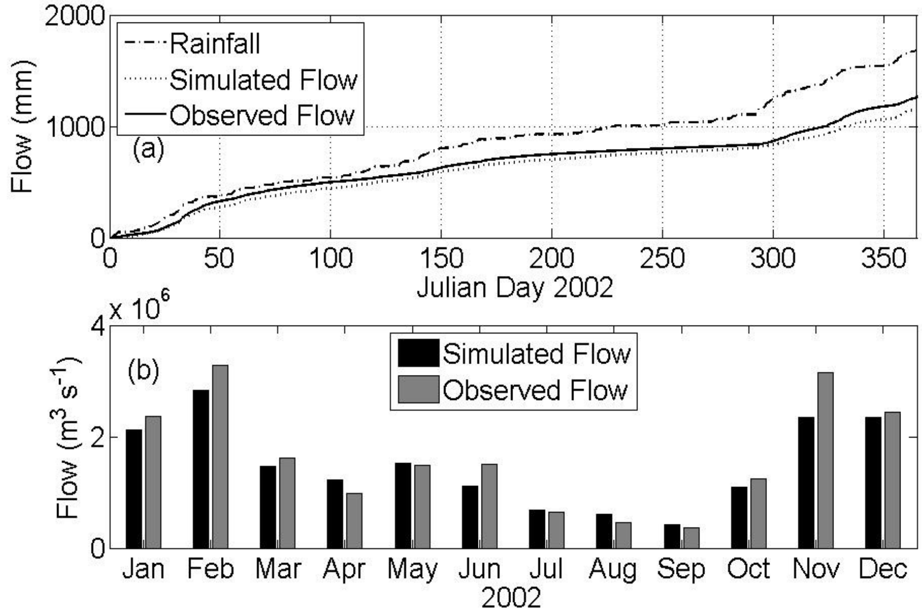

Figure 7.5 (a) Cumulative rainfall, observed and simulated flows; (b) monthly total observed and simulated for the validation year 2002.

GEOtop was then validated using the observed data for 2003 and the optimised parameters set from the calibration exercise. Figure 7.7 shows a comparison of observed and simulated flow for the validation year 2003. The rainfall for 2003 was 1198 mm. The simulated and observed annual flow was 774 mm and 695 mm, respectively. The residual estimate of ET was 503 mm (see Figure 7.8a). The results of the simulated and the observed SSY are presented in Figure 7.9 and Figure 7.10 for the calibration year (2002) and the validation year (2003). For 2002, the simulated SSY was 0.159 t ha^{-1} while the observed SSY was slightly lower at 0.136 t ha^{-1} in 2002. The model estimates of SSY at 0.053 t ha^{-1} in 2003 was lower than the observed SSY of 0.092 t ha^{-1} for 2003. While the water quality observations covered 21% of the year 2003, the remaining 79% of the year was filled with a weak relationship between flow and SS concentrations. Given that the largest difference between observed and simulated SSY occurred in 2003, which had the lowest coverage of water quality observations, it was thought that the reduction in water quality observations may have contributed to this. Figure 7.6a shows a comparison between observed and simulated SSY for 2002. There are two observed SSY estimates, one from the raw data (42% of the year) and the second SSY which is comprised of a combination of the raw data covering 42% of the year and an estimate of SSY from a relationship between stream flow and SSY for the remaining 58% of the year. There are also two simulated SSY estimates, one from the full year and the second a sum of the simulated SSY values from the same 42% of the year where water quality observations were taken. Figure 7.6b shows the corresponding data for 2003. It can be noted from Figures 7.6a and b that both the filled and raw (unfilled) estimates compare well with simulated estimates of SSY in 2002, however the same cannot be said for 2002.

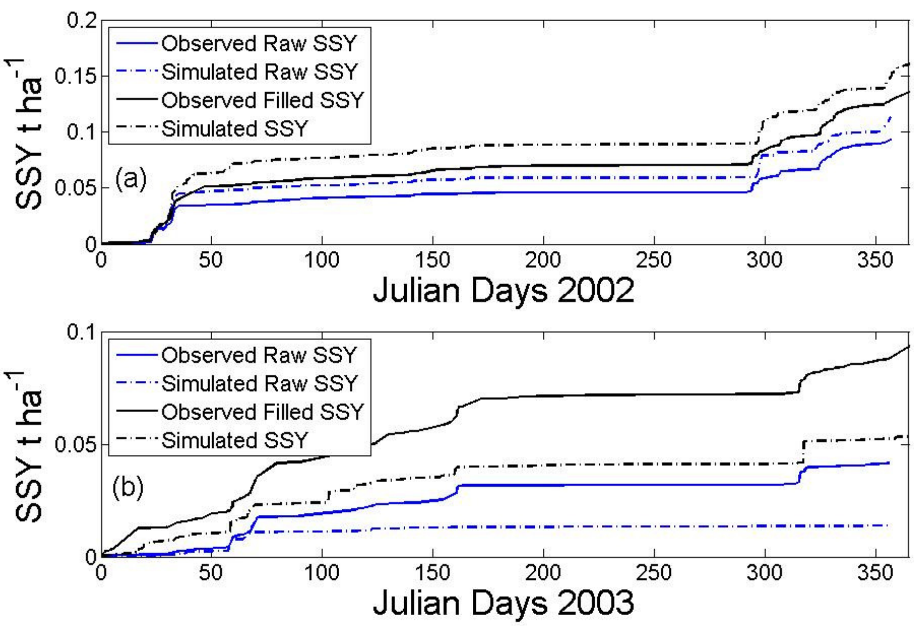

Figure 7.6 for 2002 (a) and 2003 (b), comparison of observed raw (unfilled) SSY and simulated SSY from the corresponding time periods and filled observed SSY and simulated SSY for the entire year. Note the raw SSY estimate does not extend as far as day 365 as there were no water quality observations in the last two weeks of the year.

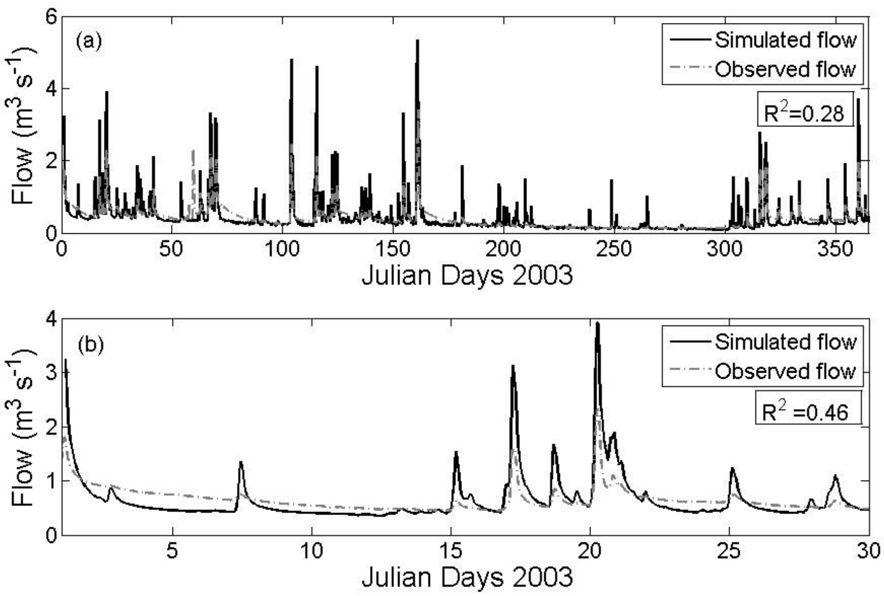

Figure 7.7 (a) Observed and simulated flows for the validation year 2003, (b) observed and simulated flows for the first 30 days of 2003.

For 2002, the low rainfall-streamflow months of July, August and September had a SSY of 0.00052, 0.00047 and 0.00017 t ha^{-1} mo^{-1} respectively. The higher SSY observed in the winter months of January and February were 0.025 and 0.043 t ha^{-1} (see Table 7.4 and Figure 7.9b). A similar pattern was observed in 2003 in Figure 7.10b. The monthly observed flows show a similar pattern, with reduced total monthly flows in July, August and September (Table 7.4). While the monthly flows are reduced, the corresponding SSY is reduced by an even greater extent.

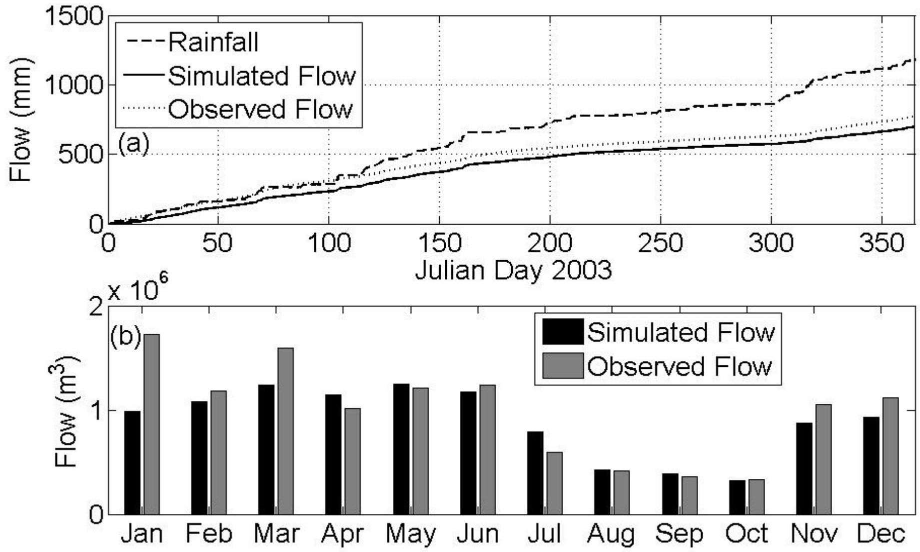

Figure 7.8 (a) Cumulative rainfall, observed and simulated flows and (b) monthly total observed and simulated flows for the validation year 2003.

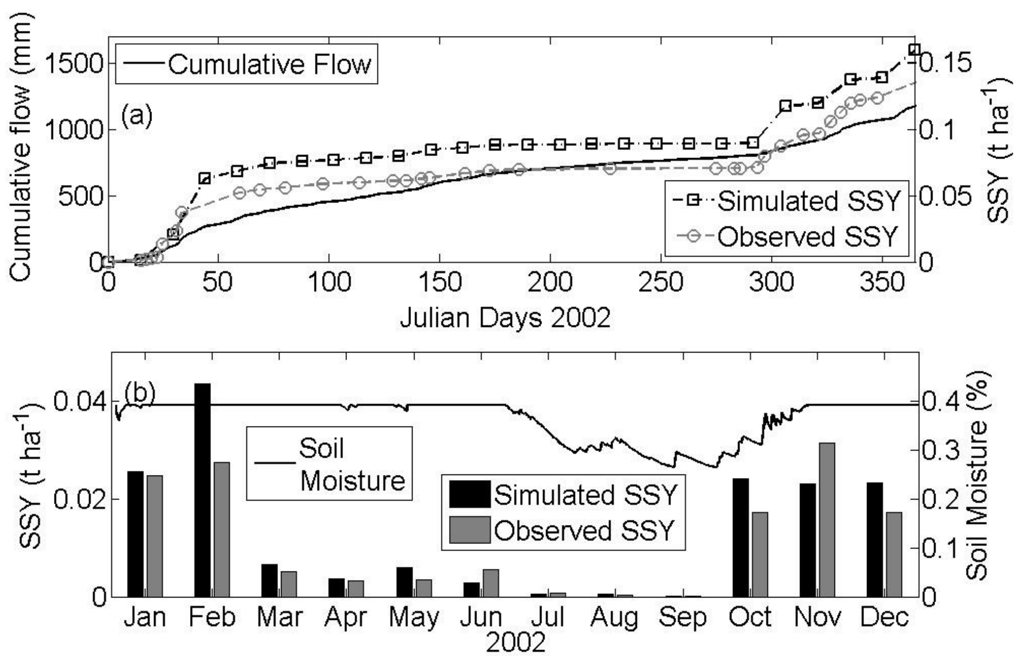

Figure 7.9 (a) Observed and simulated SSY in the Dripsey catchment for the calibration year 2002; (b) monthly totals of simulated SSY, observed SSY and catchment mean simulated soil moisture content.

Figure 7.9b and Figure 7.10b also show the simulated moisture content which ranged from a minimum of 23 % to saturation at 39 %. The minimum moisture content never dropped below the wilting point (21 %) while the soil remained saturated for 57.8 % of 2002 and 38.3 % of 2003. The months of July, August and September tended to have lower soil moisture contents for both 2002 and 2003 while between November and April the moisture content remained close to or at saturation. Liu et al. (2011) in an investigation into spatial variability of soil moisture using remote sensing found a similar seasonal pattern of soil moisture at the same site in 2006. Figure 7.9b and Figure 7.10b also show that the months with higher SSY correspond to the months where the soil moisture remains close to saturation.

The results of the rainfall-erosion scenarios (Table 7.3) are shown in Figure 7.11, Figure 7.12 and Figure 7.13 and are summarised in Table 7.5. The modelling scenarios resulted in SSY ranging from 0.85 kg ha^{-1} to 9.28 kg ha^{-1} (see Table 7.5). The highest SSY resulted from Scenario B (the 34.4 mm precipitation event) with 9.28 kg ha^{-1} resulting for a rainfall intensity of 7 mm hr^{-1} with the SSY decreasing with decreasing rainfall intensity resulting in a SSY of 3.2 kg ha^{-1} from a rainfall intensity of 1.5 mm hr^{-1}. A similar pattern of increasing SSY with increasing rainfall intensity was noticed from the other two scenarios. The lowest SSYs were seen from Scenario A, the precipitation event with the lowest rainfall (20.2 mm) ranging from 0.85 to 1.43 kg ha^{-1} while Scenario C (25.2 mm) had SSY values ranging from 2.1 to 8.1 kg ha^{-1}.

7.5 Discussion

The comparison of observed and simulated streamflows values using the GEOtop model in the calibration year (Figure 7.4 and Figure 7.5) and the validation year (Figure 7.7 and Figure 7.8), confirm that the current configuration of GEOtop

Table 7.4 Monthly totals for rain, observed and simulated SSY and flow for 2002 and 2003.

Month	Rain 2002 mm	Rain 2003 mm	Sim SSY 2002 (kg ha^{-1})	Obs SSY 2002 (kg ha^{-1})	Sim 02 flow (m^3)	Obs 02 flow (m^3)	Sim SSY 2002 (kg ha^{-1})	Obs SSY 2002 (kg ha^{-1})	Sim 02 flow (m^3)	Obs 02 flow (m^3)
1	254.2	94.0	25.61	24.64	2,216,452	2,367,067	7.27	13.11	1,645,420	1,719,742
2	217.8	110.1	43.39	27.38	2,884,129	3,289,127	8.43	11.39	1,332,024	1,181,171
3	81.8	105.8	6.64	5.01	1,448,360	1,612,577	8.11	17.57	1,531,591	1,589,557
4	138.4	143.6	3.67	3.16	1,185,478	974,122	8.27	6.93	1,187,776	1,013,883
5	177.8	130.2	5.92	3.38	1,495,504	1,479,441	3.13	8.02	1,201,128	1,209,751
6	103.0	138.0	2.71	5.60	1,065,387	1,507,850	5.01	13.07	1,138,784	1,239,658
7	46.2	91.2	0.52	0.77	642,349	643,185	0.70	1.36	722,064	597,058
8	70.8	15.4	0.47	0.37	584,991	463,093	0.10	0.40	393,872	417,350
9	44.8	55.6	0.17	0.16	396,151	368,561	0.13	0.15	379,059	358,187
10	244.6	46.0	24.05	17.16	1,127,859	1,236,658	0.0861	0.095	315,789	336,445
11	253.2	191.6	23.00	31.28	2,484,893	3,156,084	10.46	12.06	866,322	1,047,117
12	189.6	102.4	23.31	17.20	2,466,686	2,448,123	1.66	9.13	925,047	1,118,122
total	1822.20	1223.9	159.45	136.1	17,998,241	19,545,888	53.3	93.2	11,638,877	11,828,042

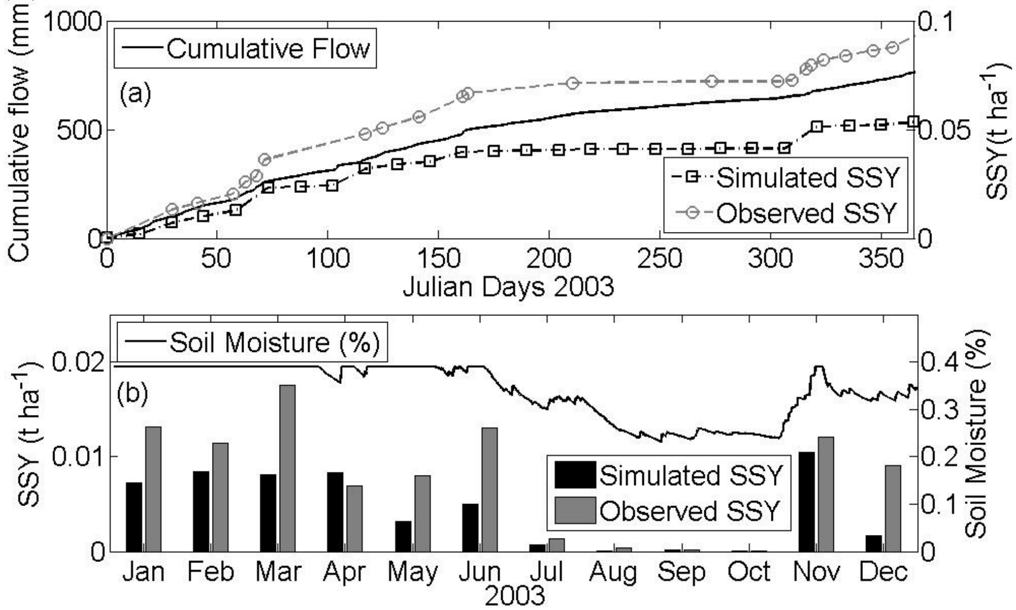

Figure 7.10 (a) Observed and simulated SSY in the Dripsey catchment for the Validation year 2003; (b) monthly totals of simulated SSY, observed SSY and catchment average soil moisture content.

Table 7.5 SSY results from modelled scenarios with varying rainfall intensities.

Scenario A		**Scenario B**		**Scenario C**	
Rainfall intensity (mm hr^{-1})	**Total SSY (kg ha^{-1})**	**Rainfall intensity (mm hr^{-1})**	**Total SSY (kg ha^{-1})**	**Rainfall intensity (mm hr^{-1})**	**Total SSY (kg ha^{-1})**
1.5	0.85	1.5	3.2	1.5	2.1
2.5	0.88	2.5	4.07	2.5	3.7
4	1	4	6.4	4	5.79
5	1.3	5	7.6	5	6.65
6	1.36	6	8.32	6	7.36
7	1.43	7	9.28	7	8.1

captures well the catchment hydrological response of this grassland catchment in the Irish temperate climate.

The observed SSY in this Irish grassland catchment is approximately 0.13 t ha^{-1} yr^{-1}, which is on the low end of the internationally observed erosion values. While there is limited data on erosion in Ireland, a study by Harrington and Harrington

(2010) of SSY in a number of similar catchments in the south of Ireland found SSY to range from 0.15 to 0.25 t ha^{-1} yr^{-1}. This observed rate of erosion is lower than the European wide median soils erosion rate of 0.3 to 1.4 t ha^{-1} yr^{-1} estimated by Verheijen et al. (2009).

The SSY from the GEOtop-LISEM erosion module also compares well with the recorded SSY for the calibration year 2002 (Figure 7.9) but underestimates the SSY in the validation year 2003 (Figure 7.10). This suggests that the model is capturing the main erosion events talking place in the catchment, particularly in 2002. The seasonal pattern of SSY follows the pattern of the cumulative flows with higher exports in winter and times of higher flow. This is to be expected as erosion is primarily rainfall driven in Ireland. The reduction in SSY during the months with lower soil moisture confirms that SSY is dependent on the strength of overland flow. The latter is primarily generated from saturation excess flow in Ireland which results from long duration low intensity rainfall patterns.

The effect on SSY of increasing rainfall intensity is seen in Figure 7.11, Figure 7.12 and Figure 7.13. These figures show the results of a number of GEOtop simulations with varying rainfall intensities, as described earlier (see Table 7.3). All the models were spun up for a number of months prior to the modified precipitation events, thus ensuring all models have the same initial conditions as the unmodified model prior to the start of the precipitation events. Investigations of the rainfall runoff processes of each simulation show that with each increase in rainfall intensity there was an increase in overland flow.

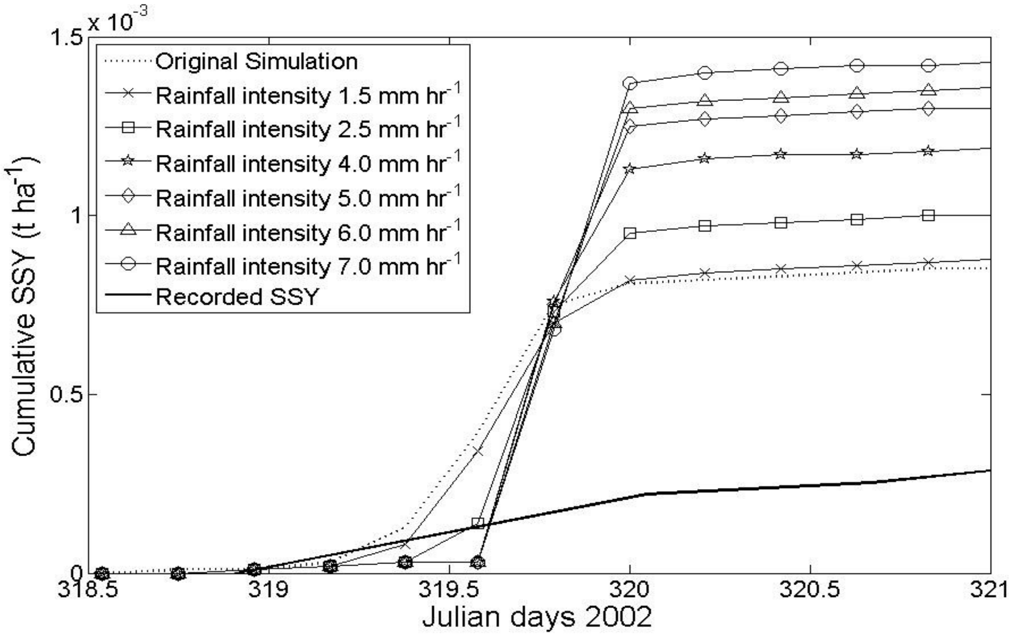

Figure 7.11 SSYs for various rainfall intensities for Scenario A (20.2 mm precipitation event). Note, original and simulated rainfall patterns given in Table 7.3.

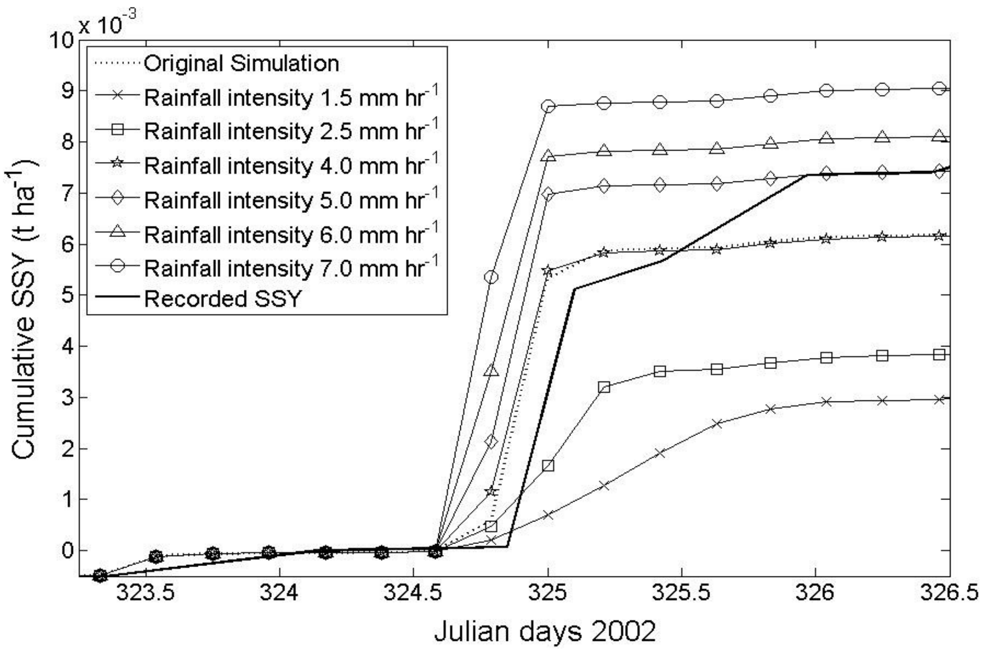

Figure 7.12 SSYs for various rainfall intensities for Scenario B (34.4 mm precipitation event). Note, original and simulated rainfall patterns given in Table 7.3.

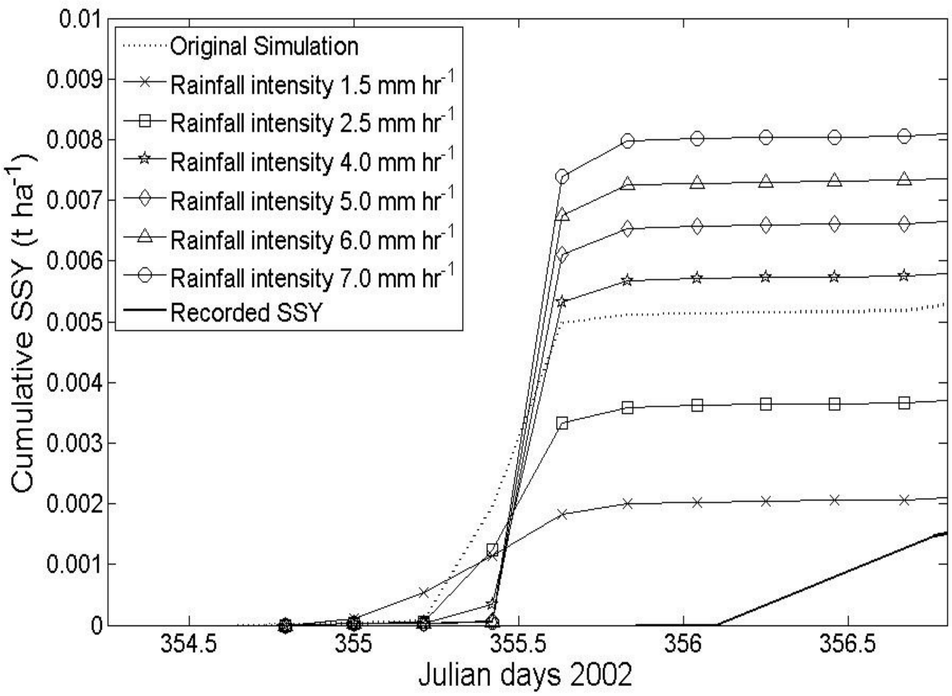

Figure 7.13 SSYs for various rainfall intensities for Scenario C (25.2 mm precipitation event). Note, original and simulated rainfall patterns given in Table 7.3.

We identified a linear relationship between increasing rainfall intensity and increasing SSY (see Figure 7.14). Eqn 7.6, eqn 7.7 and eqn 7.8 show the relationships between rainfall intensity and SSY for scenarios A, B and C respectively.

$$SSY(kg\,ha^{-1}) = 0.118(rain(mm\,hr^{-1})) + 0.6223 \qquad \text{(eqn 7.6)}$$

$$R^2 : 0.92; RMSE : 0.073$$

$$SSY(kg\,ha^{-1}) = 1.438(rain(mm\,hr^{-1})) \qquad \text{(eqn 7.7)}$$

$$R^2 : 0.91; RMSE : 0.7231$$

$$SSY(kg\,ha^{-1}) = 1.083(rain(mm\,hr^{-1})) + 0.925 \qquad \text{(eqn 7.8)}$$

$$R^2 : 0.97; RMSE : 0.4346$$

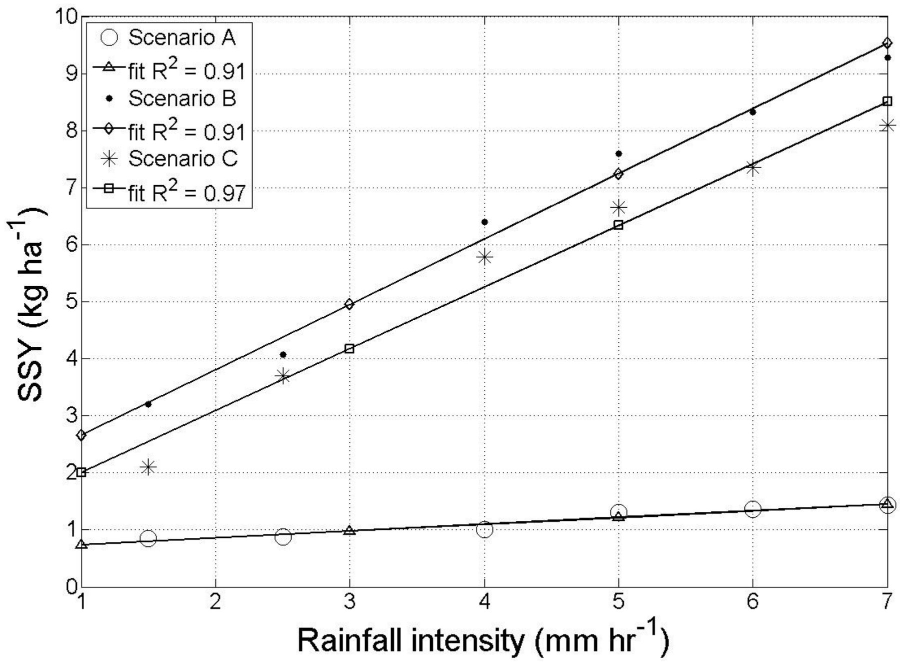

Figure 7.14 Rainfall intensities and SSY from modelled scenarios.

The highest SSY can be seen arising from Scenario B which also had the highest precipitation at 34.4 mm. Investigations into the precipitation 24 hours before these modelling scenarios begun found that 0.6, 9.8 and 12 mm of precipitation had occurred before the A, B and C scenarios, respectively. The higher precipitation before scenarios B and C will have likely resulted in similar soil conditions thereby facilitating the generation of saturation excess overland flow faster than on a drier soil, resulting in a higher SSY. These results indicate that SSY is not only influenced by rainfall intensity but also by prior precipitation and soil moisture status. Similar observations have been noted by others, with Nearing et al. (2005b) noting from modelling work that changes in rainfall amount associated with changes in storm intensity were likely to have a greater impact on erosion than simply changes in rainfall amount alone. Zhang et al. (2011) observed that rainfall intensity had a significant effect on sediment loss from field plots where rainfall intensities of 100 and 200 mm hr^{-1} were respectively observed to produce sediment yields of 2 and 2.5 times the value for 60 mm hr^{-1}. While the rainfall intensity in Zhang et al. (2011) study are much higher that rainfall intensities observed in the Dripsey catchment it is interesting to note that the pattern is the same. It is likely that given this relationship between rainfall intensity and SSY and the predicted increase in magnitude and frequency of severe rainfalls in Ireland identified by Leahy and Kiely (2011) and McGrath and Lynch (2008) will result in an increase in SSY in the future.

7.6 Conclusion

SSY in this Irish grassland catchment was 0.13 and 0.09 t ha^{-1} yr^{-1} for 2002 and 2003 respectively, which is the lower end of the wide range of values reported internationally. An analysis of SS concentrations found that lower flows tend to have lower SS concentrations with the higher flows being associated with higher SS concentrations. Given that SSY is a product of both flow and SS, the SSY of the Dripsey catchment is much higher during periods of higher flows. The hydrological model GEOtop with a new LISEM erosion module simulates the SSY from this grassland catchment and it compares well to the observed SSY. This good comparison between the modelled and measured SSY indicate that the model captures all the main erosion events taking place in the catchment.

From the modelling analysis of a number of rainfall scenarios where rainfall intensities were changed while the total rainfall volume was kept constant it was found that SSY is sensitive to rainfall intensity. SSY was found to increase in a linear fashion with increasing rainfall intensity. A closer examination of the precipitation event of Scenario B on day 324, revealed through scenario modelling, that the same precipitation event with an intensity of 2.5 mm hr^{-1} would have resulted in a SSY of 4.07 kg ha^{-1} whereas the same precipitation event with an intensity of 7 mm ha^{-1} would have resulted in a SSY of 9.28 kg ha^{-1}. Similar patterns were noticed for the other precipitation scenarios modelled. However this precipitation event had a higher precipitation in the 24 hours previous, indicating that initial soil moisture conditions also influence the SSY of a precipitation event. It is likely that any future increase in rainfall intensities due to climate change is likely to result in an increase in erosion from grassland catchments.

7.7 Acknowledgments

This study was funded by the Irish Environmental Protection Agency (EPA) under the Science Technology Research & Innovation for the Environment (STRIVE) Programme 2007–2013 of Ireland (Soil H: Interactions of soil hydrology, land use and climate change and their impact on soil quality; 2007-S-SL-1-S1). We very much appreciate the support provided by the GEOTop team, in particular Prof. Riccardo Rigon and Matteo Dall'Amico. We acknowledge the contribution and discussions with Dr. Xianli Xu. Nelius Foley, Killian Murphy and Adrian Birkby provided essential instrumental and data collection support for this research.

Chapter 8

General Discussion

This thesis formed part of a larger EPA project, titled "*Interactions of soil Hydrology, land use and climate change and their impact on soil quality (SoilH)*". The project proposed to establish a network of benchmark sites throughout Ireland using existing national sites (from previously funded EPA projects such as NSD and SoilC) for the measurement of soil hydrological properties and the establishment of a hydrological classification of Irish soils. The project also proposed to employ a process based soil hydrological model (GEOtop) with an additional erosion module developed from LISEM to investigate erosion and loss of organic matter in Irish catchments. These outputs will be combined (by others in the SoilH projects) with Irish geo-spatial data to develop a GIS-based risk assessment tool to predict impacts on soil quality based on hydrology, land use and climate change.

8.1 Saturated hydraulic conductivity of Irish mineral soils

As part of the Strategic Plan for the Development of the Forestry Sector in Ireland (Dept. of Agriculture, 1996), a soil survey was conducted to assist in establishing forestry potential for planning and harvesting purposes. This soil survey subsequently became known as the Irish Forestry soils (IFS) soil database (produced from the project of Soils and Subsoils data generated by Teagasc with co-operation of the Forest Service, EPA and GSI, completed May 2006) (IFS, 2006). A methodology based on remote sensing and GIS was developed where soil type, productivity and distribution were modelled. The soil types being modelled fall into five broad classes; shallow mineral, deep mineral well drained, mineral poorly drained, peat over mineral soil.

In order to build up a national classification of hydraulic properties of Irish soils, the results of the estimates of the hydraulic parameters from the 31 mineral sites were compared to the soil groups of the IFS soil database. The IFS database was able to capture the difference in saturated hydraulic conductivity between well drained and poorly drained classes. Deep well drained mineral soils have the highest K_s (average 19.29; max 249; min 0.35; m $s^{-1}*10^{-6}$) with the K_s of poorly drained mineral sites two orders of magnitude lower (average 0.89; max 2.4; min 0.24 m $s^{-1}*10^{-6}$). Excluding the peat and alluvium soils, estimates of θ_S were between 0.36 and 0.46 (l l^{-1}). The highest values of the van Genuchten (1980) parameter α were observed (0.16 cm^{-1}) in deep well drained mineral soils, with the lowest α in deep poorly drained mineral soils (0.06 cm^{-1}). The van Genuchten (1980) parameter n did not show as much variation with values ranging from 1.99 to 2.28.

The BEST method (Beerkan Estimation of Soil Transfer parameters though infiltration experiments) was used to estimate hydraulic characteristics such as saturated hydraulic conductivity and water retention in the mineral soil sites. An analysis of the BEST method found that it was a promising method when compared with the DL (Differentiated Linearization (Vandervaere et al., 2000)) method, however some anomalies were encountered when estimating hydraulic properties in some cases where there were two few points in the transient flow state i.e. limited infiltration data in unsaturated soils. This was due either to slow infiltration rates or soils with initial high soil moisture content. This indicates that the BEST method requires a wide range of soil water content from initial to saturated sates so as to include sufficient transient flow. In this study when the

BEST method produced some anomalies, the WU (Wu et al., 1999) method was used to estimate the soil hydraulic properties.

8.2 Spatial variation of blanket peatland saturated hydraulic conductivity and bulk density.

The investigations into the spatial variation of saturated hydraulic conductivity and bulk density in the pristine blanket peatland described in Chapter 5 found that peatland is composed of two distinct horizontally spatial zones: one near the margins (i.e. near a stream) and the second at the bog interior. Saturated hydraulic conductivity was to found to be higher (~10^{-5} m s^{-1}) in the bog interior than the riparian zone (~10^{-6} m s^{-1}) while the converse applied to bulk density, with lowest density (~ 0.055 g cm^{-3}) at the interior and highest (~ 0.11 g cm^{-3}) at the riparian zone. In general, we found saturated horizontal conductivity to be approximately twice the saturated vertical hydraulic conductivity. These results support the idea that areas of lower saturated hydraulic conductivity at the margins control the hydrology of blanket peatlands. We suggest that removal or damage of the peatland at the margins may result in a decrease in the water table height, leading to loss of carbon by decomposition and erosion and to a decrease in the general overall health of the bog. An analysis of stream flow found that stream runoff is composed of 8% baseflow and 92% flood flow, and the latter is from surface runoff rather than subsurface flow and arises as there is limited storage in the peatland due to the perennial near surface water table. These results may also be of particular concern to hydrological modellers, as it is important in hydrological modelling to take these spatial differences of key properties into account. We also conclude that it appears that it is the hydraulic conductivity of the peat at the margin that controls the runoff from the peatland.

8.3 Changing hydrological response of afforested peatland

The investigation into the hydrological response and water balance of upland catchments as a result of afforestation and its associated drainage, described in Chapter 6, suggests that the hydrological response of peatland catchments may be altered as a result of afforestation. Results from the modelling exercise carried out on a blanket peatland using the hydrological model GEOtop where the vegetation has been changed from natural peatland to Sitka Spruce forestry show that afforestation results in a decrease in streamflow and an increase in evapotranspiration. The evapotranspiration rates from a young Sitka spruce (10 years old) forest were similar to the existing pristine peatland catchment. A semi-mature Sitka Spruce forest (15 year old) resulted in an increase in evapotranspiration of 492 mm (121%) through increased transpiration from the canopy and interception.

We noted that this increase in evapotranspiration was particularly noticeable in summer and resulted in an increase in the depth of the water table and reduction in stream runoff. However, in winter, following periods of heavy rainfall, the water table depth approached that of an unforested peatland. This shallow water table depth in combination with a drainage network associated with afforestaation resulted in an increase in peak flow in times of heavy rainfall. From this study we concluded that there is limited or no benefit to flood attenuation from peatland afforestation during winter periods when the water table is high while the converse applies to summer flows where the rainfall runoff was reduced in dry periods. We also note that central to the modelling effort was the input of the spatial variation of the saturated hydraulic conductivity with areas of peat with lower hydraulic conductivity at the margins.

8.4 A model study of the impact of rainfall rates on erosion from a grassland catchment.

An earlier study of the suspended sediment yield (SSY) of an Irish grassland catchment found that SSY was 0.13 and 0.09 t ha^{-1} yr^{-1} for 2002 and 2003 respectively, which is at the lower end of the range of values reported internationally. We analysed the suspended sediment (SS) concentrations from that study found that lower flows tend to have lower SS concentrations with the higher flows being associated with higher SS concentrations. The hydrological model GEOtop with a new LISEM erosion module simulated the SSY from this grassland catchment and it compared well to the observed SSY.

From the results of scenario modelling using the GEOtop/LISEM model of a number of different precipitation events where rainfall intensities were modified while the total rainfall volume was kept constant found that SSY is sensitive to rainfall intensity. The results of this study showed that SSY increase in a linear fashion with increasing rainfall intensity. The analysis of rainfall records in Ireland carried out by Kiely (1999) found at least a 10% increase in annual rainfall since 1975, and also identified that the increased rain came in the form of more frequent wet hours rather than an increase in rainfall intensity. Thus we conclude that it is likely that any future increase in rainfall intensities due to climate change is likely to result in an increase in erosion from grassland catchments.

8.5 Synthesis

The primary goals of the EPA project *Interactions of soil Hydrology, land use and climate change and their impact on soil quality (SoilH)*, was to

A) Establish a network of benchmark sites throughout Ireland using existing national sites for the measurement of soil hydrological properties and the establishment of a hydrological classification of Irish soils.

B) Employ the soil hydrological model GEOtop to investigate the threats to soil quality from erosion, surface sealing, compaction, landslides and loss of organic matter.

C) Investigate the interactions between soil hydrology, land use and climate change.

This thesis details the soil hydrological properties for 31 mineral soil and one peat site and through comparison with the IFS soils database provides a hydrological classification of Irish soils, see Chapters 4 and 5.

A study of the changing hydrological response of a virgin blanket peatland due to land use change (afforestation) was detailed in Chapter 6, with Chapter 7 of this thesis investigating the threat to soil quality from erosion using the hydrological model GEOtop and the additional erosion module. The interaction of climate change and soil hydrology as well as the threats to soil quality from surface sealing, compaction, landslides and loss of organic matter were considered to be outside the scope of this thesis.

Chapter 9

Suggestions for future work

The research conducted as part of the EPA *project Interactions of soil Hydrology, land use and climate change and their impact on soil quality (SoilH)* raises some issues and questions that I would like to recommend for future investigations.

- The hydrological modelling effort conducted in this study highlighted the lack of data on Irish soils in particular the Particle Size Distribution (PSD). A number of projects such as the National Soils Database (NSD) have collected a large number of soil samples but have yet to be analysed for PSD. PSD data from such projects would be invaluable in building up a profile of the PSD of Irish soils.
- This study found that the riparian areas of the Glencar bog had lower saturated hydraulic conductivity than at the bog interior. This, however, is just one blanket peatland and may not be representative of other peatlands. Further investigations should be carried out to see if similar patterns are found in other blanket and raised peatlands.
- The hydrological modelling study of an afforested peatland estimated evapotranspiration rates and stream flow from a semi mature Sitka Spruce. It would be interesting to compare the results of this modelling exercise to hydrological observations such as evapotranspiration, interception and runoff from a number of different afforested peatlands of varying age.
- The afforestation of blanket peatlands alters the hydrology of peatlands. However, it is also likely that other parameters such as peat depth, bulk density and soil organic carbon as well as the carbon balance may also be altered. A study of an afforested peatland and paired pristine blanket peatland may be able to quantify some of these impacts.

- The flashy nature of the hydrological response of virgin blanket peatlands has been well documented and this study suggests that flood mitigation from afforestation of blanket peatlands may be limited. However, raised peatlands forested or otherwise tend to have lower water tables and may provide greater flood attenuation. Further investigations would be required to assess the impacts of afforestation on the hydrological response of raised peatlands.
- The suspended sediment yield of the Dripsey grassland catchment may be at the lower end of values reported internationally, however, many reported values that were higher than the Dripsey catchment were from arable land. A comprehensive sampling regime of suspended sediment concentrations in a number of catchments with different land uses and soil types may show that other areas may have higher suspended sediment yields. It is possible that suspended sediment yields from arable catchments may be higher than the Dripsey catchment and may approach the values reported in the literature where the suspended sediment yield of a catchment is of concern to soil quality.

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Appendix B1

EGU Apr 2009 poster presentation

The Application of GEOtop for catchment scale hydrology in Ireland

Appendix B2

EPA Nov 2010 poster presentation

GEOtop hydraulic rainfall runoff model and its applications

Appendix B3

Xu, X., Kiely, G. and Lewis, C. 2009. Estimation and analysis of soil hydraulic properties through infiltration experiments: comparison of BEST and DL fitting methods. Soil Use and Management, 25: 354-361.

Appendix B4

Lewis, C., Albertson, J., Xu, X. and Kiely, G., 2011. Spatial variability of hydraulic conductivity and bulk density along a blanket peatland hillslope. Hydrological Processes, in press