Runoff production in blanket peat covered catchments

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RUNOFF PRODUCTION IN BLANKET PEAT COVERED CATCHMENTS

Joseph Holden

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To Grandma

Not quite a desert island

ABSTRACT

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Joseph Holden

PhD Thesis, University of Durham, 2000.

Although blanket peat covers many major headwater areas in Britain, runoff production within these upland catchments is poorly understood. This thesis examines runoff production mechanisms within the blanket peat catchments of the Moor House National Nature Reserve, North Pennines, UK. Catchments ranging from 11.4 km^2 down to the hillslope and plot-scale are examined. Runoff from the monitored catchments was flashy. Lag times are short and rainwater is efficiently transported via quickflow-generating mechanisms such that flood peaks are high and low flows poorly maintained. Hillslope and plot-scale runoff measurements show that the flashy catchment response is the result of the dominance of overland flow. Typically 80 % of runoff is produced as overland flow. This occurs both on bare and vegetated surfaces. Most of the remaining runoff is generated from the upper 10 cm of the peat, except where well-connected macropore and pipe networks transfer flow through the lower layers. Below 10 cm depth the blanket peat matrix fails to contribute any significant runoff. Thus most groundwater-based models of peat hydrological process are not readily applicable to these catchments.

Suggestions that infiltration-excess overland flow may be largely responsible for the flashy regime of these upland catchments are not substantiated by the blanket peat infiltration data presented in this thesis. Monitoring of hillslope runoff mechanisms combined with rainfall simulation (at realistic intensities of 3-12 mm hr^{-1}) and tension-infiltrometer experiments shows that saturation-excess mechanisms dominate the response. Infiltration is relatively rapid in the upper peat layers until they become saturated and overland flow begins. High water tables result in rapid saturation of the peat mass such that even at low rainfall intensity runoff production is just as efficient as during high intensity events.

While macropores have largely been ignored in blanket peat, results presented suggest that up to 30 % of runoff may be generated through them. Occasionally these macropore networks develop through the deeper peat such that runoff bypasses the matrix and runs off at depth from small outlets and larger pipe networks. Seasonal variations in runoff-generating processes can be exacerbated by drought which causes structural changes to the near-surface of the peat. This was found to result in enhanced infiltration and macropore flow which may encourage pipe network development.

Flow has been monitored simultaneously from several natural pipes in a 0.4 km^2 headwater catchment of the Tees. This catchment provides one of the few examples of pipeflow monitoring outside the shallow peaty-podzols of mid-Wales. Natural pipes are found throughout the soil profile and can be at depths of up to three metres. Ground penetrating radar was useful in identifying deep subsurface piping and suggestions are made for improvements to the application. The pipe networks were found to be complex and results demonstrate that outlet location and size may bear little relation to the form and depth of the pipe a short distance upslope. Pipes generally contribute less than 10 % to catchment runoff but on the rising and falling hydrograph limbs can contribute over 30 % to streamflow. Pipeflow lag times are short suggesting that both the shallow and deep pipes may be well connected to the surface. Thus while matrix runoff contributions at depth within the peat may be low, macropore flow mechanisms can be significant.

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I confirm that no part of the material presented in this thesis has previously been submitted by me or any other person for a degree in this or any other university. In all cases, where it is relevant, material from the work of others has been acknowledged.

Signed: J. M.........................

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CHAPTER 1

PUTTING BLANKET PEAT HYDROLOGY IN CONTEXT

1.1 Overview

For many centuries peatlands have been a source of fascination to naturalists and scientists. However, most research has concentrated on the ecological and energy resources of these regions. Peatlands are common in cold, wet climates, forming 34 % of the land surface area of Finland and 18 % of Canada (Hobbs, 1986). The humid climate and retentive soils of the British Isles make it suitable for wetland development (Hughes and Heathwaite, 1995). There are now few areas of lowland Britain covered by extensive peat deposits, with the exception of the Somerset Levels and the Cambridgeshire Fens; drainage for agriculture and peat–cutting for fuel and horticulture have reduced their extent. These lowland peats primarily provide sites of inundation for floodwater rather than the source of runoff (Burt, 1995). However, in Britain most of the peat deposits are in the uplands. Here it lies over many of the major watersheds, modifying stream flow and erosion in the rivers originating there. Hence this peat is a major hydrological and geomorphological influence. Nevertheless, there has been a distinct lack of interest in hydrological processes on the blanket peats of the uplands and the details surrounding the spatial and temporal production of runoff within upland peat catchments remain relatively unknown.

Some 7.5 % of the total land surface of the British Isles (Britain and Ireland – 22 500 km^2) is covered by *blanket peat* (Tallis *et al.*, 1998). This represents the largest single contribution of blanket peat (10 – 15 %) to a globally scarce resource – less than 3 % of the world's peatlands are blanket mires. Blanket peats occur on the gentle slopes of upland plateaux, ridges and benches, and are primarily ombrogenous. That is, they are hydrologically disconnected from the underlying mineral layer such that they receive all of their water and nutrients in the form of precipitation. One of the largest areas of peatland in Britain is represented by the blanket peat deposits of the Pennines where the humidity is high enough to allow the spread of the mire over sloping terrain. The blanket peat moorlands of the Pennines are used for many purposes. Reservoirs have been built directly downstream of the moorlands and these upland areas are a major source of water for public supply in the north of England. The moorlands also form an important land resource for agriculture and recreation with tourists, bird-watchers and hill-walkers amongst the list of people to cater for. Peat erosion is common in the

Pennines (Bower, 1961), prominent particularly in the southern Pennines (Conway, 1954; Anderson and Tallis, 1981). Some research seems to suggest that severe erosion onset may have been within the last two hundred and fifty years (Tallis, 1964, 1985; Labadz, 1988). Industrialisation and air pollution resulting in the demise of *Sphagnum* moss carpets is one possible explanation (Tallis, 1965, 1985). Tallis (1997) also suggests initiation of erosion may have occurred in the Medieval warm period (c. 1200 A.D.) due to desiccation. Blanket peat erosion can be very problematic for reservoir management; infilling can be rapid due to high volumetric loads (Labadz, 1988; Labadz *et al.*, 1991).

Excess moisture is critical to the initiation, development, and maintenance of peatlands, and they are very sensitive to changes in the supply of water (Heathwaite, 1995). Change that occurs at the local scale can, in accumulation, have a global impact. Drainage of peat for example can alter carbon sink-source relationships which can increase atmospheric CO_2 (Roulet, 1990). Also the period of time following droughts appears to be strongly correlated with the discolouration of water supplies from peatlands (Mitchell and McDonald, 1992; Butcher *et al.*, 1992); this is thought to be a result of the release of humic and fulvic acids after desiccation, although the actual mechanisms are still not fully understood (Pattinson *et al.*, 1994). With future climate change it may be that increasing problems not just of water quantity but of quality will be presented to water managers in these areas.

In order to predict the consequences and extent of environmental change on blanket peat ecosystems, whether the change is direct, such as drainage or restoration strategies, or inadvertent, such as climate change or chemical deposition, an understanding of the temporal and spatial variability of physical, chemical and biological processes is required. The management of blanket peat catchments clearly requires an understanding of the processes at work so that a holistic integrated management scheme can be adopted which takes into account the variety of uses and changes likely to take place. Clearly an understanding of the hydrology of blanket peat is an essential part of the understanding of the evolution of British upland landscapes, peatland ecology, erosion and water chemistries.

1.2 Scope and organisation of the thesis

This thesis is an attempt to shed light on the exact processes responsible for runoff generation in blanket peat catchments. By doing so, it will also allow improved understanding of some of the interlinked biological and geomorphological processes at work within these important environmental systems. This thesis presents the results of a field investigation conducted in the blanket peat headwaters of the River Tees in the North Pennines. Essentially, there are three interlinked components of runoff generation which merit investigation: flow over the peat surface, flow within the soil matrix and flow within larger macropore networks. The understanding of the roles of these components and the spatial and temporal linkages between them is central to developing an improved understanding of runoff generation in much of the British uplands. This thesis deals with these three elements through use of a mixture of field and laboratory monitoring and experimentation in order to improve our understanding of runoff generation in blanket peat.

The remainder of Chapter 1 will review processes involved in the production of runoff and state the specific aims of the current research whilst the second chapter summarises previous work on peat. Chapter 3 will discuss the study area and methodology for research. In the fourth chapter, results from automated catchment, hillslope and plot scale monitoring are combined with storm based mapping of runoff processes and pathways in blanket peat. These data are used as a context for a series of rainfall simulator experiments discussed in Chapter 5. These experiments allow infiltration and near-surface flow processes to be examined in greater detail. The simulator experiments were performed on field and laboratory peats and include examination of the effects of drought on runoff production in blanket peat. Results from further field experiments using infiltration measurement devices in Chapter 6 allow examination the relative roles of pore size in governing runoff contribution in the infiltration process. These tests also allow the roles of infiltration-excess and saturation-excess runoff generation mechanisms to be assessed (see section 1.3 below). Larger connected pore networks within blanket peat appear to be important runoff production pathways. Chapter 7 assesses the role of piping processes in a small subcatchment of the Tees whilst Chapter 8 explores a new application of ground penetrating radar in order to identify subsurface natural pipe networks in blanket peat. Throughout the thesis the interlinkages between surface and subsurface matrix and macropore processes are examined at different temporal and spatial scales. Chapter 9 brings together many of findings of the thesis and highlights the importance of improved understanding of peat hydrology for management and geomorphology. It is hoped that by adopting a spatially and temporally distributed approach to the basin hydrological system that some of the unknown runoff production mechanisms in blanket peat catchments will begin to be unravelled. Improved modelling and prediction of the effects of climate change on these environments can then follow.

1.3 Runoff production processes

1.3.1 The spectrum of runoff processes

Various runoff pathways attenuate and delay flow to different extents and hence a knowledge of the relevant mechanisms is important (Kirkby, 1985). It is now known that hydrological processes on hillslopes range from infiltration-excess overland flow and the more spatially-localised saturation-excess overland flow, through subsurface flow involving macropores and pipes, to piston or displacement flow and groundwater discharge. The dominance of these processes varies with climate, topography, soil character, vegetation cover and land use, but may vary at one location (e.g. seasonally) with soil antecedent moisture and with storm intensity and duration. The spectrum of runoff processes is related to both the type and the intensity of denudation process and to the resulting mode of hillslope and landform evolution. Therefore it is a necessary pre-requisite to understand hillslope flow pathways before looking into other aspects of hillslope denudation and water quality. On the other hand, water quality and sediment characteristics are often used to identify spatial and temporal sequencing of runoff but as yet the precise manner in which 'old' and 'new' water can be identified from flow mechanisms has still not been clearly identified (Burt, 1989).

1.3.2 The historical lack of a process-based approach

Many drainage basin and wetland studies have adopted a water balance approach (Equation 1.1):

$$P = Q + E + \Delta(I,R,M,G,S) \qquad [1.1]$$

where P is precipitation, Q runoff, E evapotranspiration, I interception, R surface storage, M soil water storage, G groundwater storage and S is channel storage (Ward, 1975). Traditionally the functional school with its use of black-box techniques has required little or no detailed knowledge regarding the relationships between the components and interest is centred on the relationship between the inputs and outputs. The problem with these approaches is that they hide the true complexities within the system. The various processes of runoff act on different spatial and temporal scales such that a ratio of inputs to outputs tells us little about *why* something happens in a drainage basin. Yatsu (1966) noted that "*geomorphologists have been trying to answer the what, where and when of things, but they have seldom tried to ask how. And they have never asked why. It is a great mystery why they have never asked why*". He repeated these feelings 26 years later when he proposed that a dynamism still needed to be injected into a geomorphology which had become more quantitative but with the wrong sort of quantification - statistical relationships rather than physical ones (Yatsu, 1992). So it is within this wider geomorphological context that the history and deficiencies of hydrological research must be placed.

Until the 1970s, hydrologists took the view that the headwaters of a drainage basin were nothing more than source areas for runoff (Burt, 1996). Concern was more with engineering forecasts of floods or water resources and so hydrologists ignored the physical characteristics of the headwaters and the exact processes responsible for generating runoff. Although the focus on water-balance studies and functional black-box methods has allowed an increase in quantitative knowledge of hydrology and geomorphology many of the quantitative relationships that have been derived are statistical and not physically-based. It is for this very reason that Yatsu (1992) accuses Strahler (1952) of 'crying wine and selling vinegar'. Hence, because of a lack of a proper process-based approach, there still remains a fundamental inability to incorporate many of the details of hydrological processes into models of runoff generation.

1.3.3 Dominant storm flow paths: Overland flow and subsurface flow

Horton's (1933, 1945) theory of hillslope hydrology in which he assumed that the sole source of flood runoff was excess water that was unable to infiltrate the soil dominated hillslope hydrology for twenty years (Kirkby, 1985). In this theory infiltration divides rainfall into two parts. One part goes via overland flow (OLF) to the stream channel as surface runoff; the other goes initially into the soil and then through the groundwater flow again to the stream or is lost by evaporation. The dominance of this theory meant that research into subsurface flow mechanisms was neglected. However, despite modifications such as the partial area concept by Betson (1964), who used the idea that infiltration-excess OLF may only be generated from part of the basin, it became apparent that Horton's model was inappropriate in many locations (Burt, 1996). Where permeable soils overlie impermeable bedrock, subsurface stormflow can account for most of the flood runoff. When the soil profile becomes completely saturated, saturation-excess OLF may also occur (Dunne, 1978). Both subsurface stormflow and saturation-excess OLF can occur at much lower rainfall intensities than required to generate infiltration-excess OLF. Furthermore the source areas for subsurface stormflow and saturation-excess OLF will be limited in size and different in location from infiltration-excess OLF. The source areas will also be variable in extent during a storm event and over longer time periods (Hewlett, 1961). The variable source area model has become the dominant concept in hillslope hydrology. Subsurface flow is now regarded as the most important mechanism for generating storm runoff both because of its influence on the generation of 'return flow', which is a component of saturation-excess OLF (Dunne and Black, 1970a), and as an important process in its own right (Anderson and Burt, 1977a, 1978b, 1990a, 1990b; Burt, 1986;). Although work on subsurface flow is now dominant, Burt (1996) suggests that infiltration-excess OLF may not be as rare as proponents of subsurface stormflow have suggested in the past (e.g. see, for example, evidence in Burt *et al.*, 1990). Where infiltration-excess OLF is the dominant producer of storm runoff, overland flow may be generated across large areas of hillside.

1.3.3.1 Macropore flow

Subsurface storm flow may be generated either by Darcian flow through the soil matrix or by non-Darcian flow through macropores or pipes. Most emphasis until recently has been placed on micropore flow, but Beven and Germann (1982) note that macropores have been the object of study for over 130 years and even Horton (1942) stated that runoff may take place "*through a thick matting of grass or mulch cover; through a layer of plant roots close to the soil surface and under forest litter; or even, in some cases, through a network of sun cracks in the soil surface*".

The choice of an effective size to delimit macropores is necessarily arbitrary and is often related more to details of experimental technique than to considerations of flow processes. Previous definitions have ranged from a capillary potential of > -10.0 kPa (equivalent diameter >30 μm) (Marshall, 1959) to Brewer's (1964) coarse macropores with a diameter of 5000 μm. Luxmoore *et al.* (1990) define macropores as having minimum equivalent cylindrical diameter of 0.075 to 1 mm. Edwards *et al.* (1988) found that aerial coverage of directly observed macropores > 0.4 mm diameter averaged

1.4 % varying with tillage and Singh *et al.* (1991) found coverage of macropores > 1.6 mm diameter, as traced on transparent sheets, ranging from 3.5 % at 0.05 m depth to 0.5 % at 0.60 m depth, with no apparent difference due to tillage. However, knowing the number, size and areal porosity of macropores may offer little insight into their role in preferential water flow (McCoy *et al.*, 1994). As Beven and Germann (1982) stress, size alone is not a sufficient criterion for the definition of a macropore. They can promote rapid, preferential transport of water and chemicals through the soil, not only due to their size but also because they are connected and continuous over sufficient distances to transcend agriculturally and environmentally important soil layers.

The presence of macropores close to the surface of the soil may be particularly important in the process of infiltration of rainfall and solutes into the soil. Mosley (1979), Beven and Germann (1982) and Kneale (1986) described the hydrological effects of rapid infiltration down macropores. Kneale and White (1984) described infiltration into a dry cracked clay-loam. Bypassing flow occurred down the cracks once rainfall intensity exceeded the infiltration capacity of the soil. Coles and Trudgill (1985) and Germann (1986) have identified important thresholds governing macropore flow. Antecedent moisture appears to be important. One surprising element is the observation of vertical movement of free water along continuous macropores through an unsaturated soil matrix (e.g. Bouma *et al.*, 1977; Smettem and Trudgill, 1983). This short-circuit bypassing flow invalidates one of the basic tenants of soil water theory which suggests that smaller pores fill before larger pores in any volume of soil. This tenant is operationally defined by the soil water retention curve. Clearly, processes and conditions for macropore flow are complicated.

The question of macropore connectivity in the downslope direction remains unresolved (Burt, 1989, 1996). Studies such as Whipkey (1965), Pilgrim *et al.* (1979) and Imeson *et al.* (1984) demonstrate that macropores can generate rapid downslope runoff as well as aiding infiltration. In other cases their downslope connectivity is not proven (e.g. compare Pearce *et al.*, 1986 with Mosley, 1979).

1.3.3.2 Soil piping

Soil piping is a much neglected process. Selby (1983), Bryan and Jones (1997) and Jones (1997) have indicated the general lack of long-term pipeflow measurements from a variety of catchments and climatic environments. Whilst pipes are not found everywhere they are important in certain localities. Jones (1994) estimates that in Britain three quarters of soil piping occurs in just two major soil groups, podzolic and raw peat. Soil pipes clearly increase the transmissibility of a soil mass (Jones, 1971) with high pipeflow velocities occurring relative to the surrounding matrix (e.g. McCaig, 1979). Average pipeflow velocities recorded in the field exceed overland flow by almost an order of magnitude (Jones, 1987); Wilson (1977) reports velocities of up to 1 m s^{-1}. Piping is commonly associated with soils containing marked reductions in vertical permeability at some point in the soil profile and with horizons above that point which have sufficient lateral permeability and hydraulic gradient to permit significant amounts of throughflow (Jones, 1990; Jones, 1997). Soils likely to crack also encourage pipe networks (Gilman and Newson, 1980; Jones, 1990).

Laboratory and mathematical studies suggest that soil pipes can contribute a significant amount, and in some instances, the majority of subsurface stormflow (e.g. Nieber and Warner, 1991; Sidle *et al.*, 1995). Most field information comes from the British Isles. Those that have involved continuous monitoring of hydrographs have been predominantly in shallow upland peaty soils and podzols and confirm that pipeflow is at least as important a process as early unquantified speculations suggested (Jones, 1971, 1979). At Measnant, mid-Wales, pipes contributed 49 % to stream stormflow and 46 % to baseflow (Jones and Crane, 1984). Most discharge reaches the stream through perennially flowing pipe sections. Average pipeflow contributions give only a partial impression of the role of pipeflow drainage. There are marked variations in pipeflow from storm to storm, depending largely on the antecedent wetness of the basin and the size of the storm. There are also marked spatial and temporal variations across the hillslope both between and within the ephemeral and perennial pipe sections (Jones and Crane, 1984; Jones, 1990). Evidence from the few other British catchments where pipeflow has been examined generally supports the broad responses found at Maesnant, particularly in terms of the sensitivity of pipeflow to soil moisture status and the non-linearity of response (Wilson and Smart, 1984; McCaig, 1983). Uchida *et al.* (1999) also note non-linearity of pipeflow response from an upland Japanese cambisol where sediment movement within pipes caused variations in runoff due to frequent blockages preventing flow.

Both between- and within-storm variations are found to occur in the dynamic contributing area feeding soil pipes (Jones, 1979). At maximum discharges the pipe network can extend the streamflow contributing area to more than 70 % of the Maesnant catchment, linking sources on crests up to 750 m distant from the point of issue near the stream bank (Jones, 1979). This substantially alters the general view of the nature of dynamic contributing areas adjacent to the stream, expanding and contracting in contiguous bands, and recognisable by wetland vegetation and/or concavity in hillslope profiles in plant form (Jones, 1979; 1990). Jones (1994) suggests that pipeflow responses are often similar to the response from other hillslope drainage processes. Bonell *et al.* (1984) showed, for a forested clay soil in Luxembourg, that pipeflow in the saturated upper soil horizon was so rapid it could not be distinguished from saturation-excess OLF.

1.3.4 Linking surface and subsurface flow

Lateral subsurface flow through the soil matrix will occur in any soil in which the hydraulic conductivity declines with depth (Zaslavsky and Sinai, 1981). If both soil and bedrock remain permeable at depth, however, then percolation remains vertical and little lateral flow can occur; infiltrating water will serve only to recharge groundwater storage and to provide baseflow. Subsurface flow can produce storm runoff in several ways; firstly by a large well-connected pipe network as in the Maesnant basin discussed above (Jones and Crane, 1984); secondly if the hydraulic conductivity of the soil is high, infiltration can lead to rapid recharge of the saturated zone at the base of the soil profile; thirdly macropore flow can have the same effect. Soils close to saturation may only require a small amount of infiltration to produce a large rise in the water table. If the hydraulic conductivity is high then large amounts of subsurface runoff will be produced as a result of a rapid rise in the water table. In addition, subsurface storm flow may occur in the form of a delayed hydrograph peaking several days after the rainfall input (Anderson and Burt, 1977a; Burt, 1996).

The variable source area model (Hewlett, 1961) is based on the assumption that water moves downslope through the soil. Hence, the source areas for subsurface stormflow and saturation-excess overland flow are the same (Burt and Arkell, 1986). Indeed it is often difficult to separate the generation of subsurface stormflow from the production of saturation-excess OLF. However, saturation-excess OLF will be a mixture of return flow ('old' soil water) and direct runoff (water unable to infiltrate into saturated soil). Hence solute and sediment loads may be very different from that generated by infiltration-excess OLF alone. As will be discussed in Chapter 2, the relative roles of infiltration-excess OLF and saturation-excess OLF are not well known for blanket peat; hence important information on solutional and particulate denudation is lacking. Maximum soil moisture levels will be reached at the foot of any slope (particularly those concave in profile), in areas of thin soil, and in hillslope hollows where convergence of flow lines favours the accumulation of soil water (Kirkby and Chorley, 1967; Burt, 1986). The extent of the saturated area depends on soil wetness and hence varies seasonally and during a storm. Where surface saturation occurs to any great extent such as on wide valley bottoms, saturation-excess OLF will dominate the flow response with higher peak discharges and lower lag times than are characteristic for subsurface stormflow (Dunne, 1978). Where soils are fairly impermeable through much of their profile, surface saturation may be extensive, with much more rainfall translated into runoff. As can be seen in Chapter 2, this may be the case for blanket peat (Burt and Gardiner, 1984).

1.4 Objective and aims of this research

The overall objective of this thesis is to provide greater understanding of the way the variety of hillslope runoff processes, discussed above, operate within blanket peat catchments. In particular the research aims:

1) To gain an improved knowledge of the spatial and temporal pattern of surface and near-surface flow in blanket peat catchments, including determination of the extent and pattern of OLF generation.
2) To make sufficient measurements of infiltration capacity in blanket peat so that the extent to which infiltration-excess and saturation-excess OLF mechanisms operate can be estimated.
3) To quantify the relative roles of the peat surface, the acrotelm and the catotelm (see Chapter 2) in controlling runoff generation from blanket peat.
4) To gain an understanding of the importance of macropores in infiltration and runoff generation in blanket peat.
5) To investigate the effect of drought on infiltration and runoff processes in blanket peat.
6) To monitor pipeflow processes, their importance, and their interlinkages with other flow processes and to investigate an alternative, non-destructive method of locating subsurface soil pipes in peat.

The following chapter will outline previous work done on peatland hydrology, and highlights why there is a need to know more about the runoff-generating mechanisms within blanket peat catchments.

CHAPTER 2

PEAT HYDROLOGY: A REVIEW OF RESEARCH

2.1 Catchment-scale runoff production

Peatlands cover the headwaters of important river systems in many parts of the world, but despite a history of scientific investigation stretching back fifty years, few studies have reported on the spatial or temporal runoff characteristics of these areas. Their low gradient and near-saturated state make it likely that saturation-excess overland flow (OLF) will be produced. However, Burt *et al.* (1990) noted that where pool-hummock complexes occur, surface runoff may be confined to pools alone, since the hummocks may be well drained. In addition Burt *et al.* (1990) also noted that the infiltration capacity of some peat surfaces may be low enough that infiltration-excess runoff will be produced during high-intensity rainstorms. Thus both infiltration- and saturation-excess models of storm runoff production need to be investigated.

The presence of peat greatly influences the hydrology of a catchment. Until recently, peatlands were considered hydrologically important because they were believed to attenuate floods and sustain baseflow in streams and rivers during periods of low precipitation. Peatlands were seen as 'sponges' because of the high porosity of peat, an idea originally attributed to Humboldt (Ingram, 1983). Most field research has been conducted in small catchments. Of the early studies in blanket peat that of Conway and Millar (1960) is the most notable. They reported results from four small moorland catchments; two had natural drainage channels, and two had artificial networks of moorland grips. The grips are ditches typically 40 cm deep, 45 cm wide and spaced at 15 m contoured intervals. They concluded that artificial drainage of peat moorlands gave an increased sensitivity of runoff response to storm rainfall with peak flows both higher and earlier. Indeed lag times were often within 20 to 30 minutes. They demonstrated that runoff production in peat is extremely rapid especially where hillslopes had a dense gully network, had been burned or were gripped. In contrast relatively uneroded subcatchments exhibited a smoother storm hydrograph with greater lag times and the water balance calculations suggested that uneroded hillslopes could retain significantly more water than drained, eroded or burnt basins. Paradoxically this may have revived the traditional 'aquifer' view of blanket peat catchments. Indeed in an earlier draft of their 1960 paper found in the Moor House Reports, Conway and Millar note:

"During dry summers the peat surfaces exposed in the erosion channels appear dusty and loose, and there may be extremely little or no flow of water in the channels. On the other hand where the peat blanket is intact water flow from its margins continues steadily if not copiously throughout prolonged dry spells"

The results of Conway and Millar (1960) from Moor House National Nature Reserve (NNR) seemed at first 'indisputable' according to Robinson (1985) but soon afterwards a number of overseas studies found that drainage of peat soils reduced the peak flow response (e.g. Baden and Egglesmann, 1970; Burke, 1975a, 1975b, Egglesmann, 1975). The more flashy response shown by Conway and Millar (1960) and Robinson (1985) may be related to the much greater density of channels provided by the open ditches. This yields a large quantity of channel precipitation which is rapidly evacuated from the system (Burt *et al.*, 1990). Huikari (1963) suggested that 30 m was a critical spacing dimension for gripping. During high rainfall, ditches closer than 30 m could receive 4 times as much runoff at one time than ditches spaced at 100 m. Storage changes as a result of drainage present conflicting evidence. Burke (1975a, 1975b), for example, examined a drained and an undrained catchment at Glenamoy, Ireland. Here runoff data for February to December 1968 inclusive were presented. Based on Burke's figures runoff:rainfall ratios from the undrained catchment were only 23.4 % compared to 79.2 % from the drained catchment. This is a remarkable difference and demonstrates the importance of enhanced understanding of the effect of land management practices on the hydrology of peatlands. Moklyak *et al.* (1975) present evidence contrary to that of Burke (1975a, 1975b) to show that drainage can reduce total runoff from peatlands. Out of five catchments investigated, three had reduced annual runoff following drainage, one had an increase and one had no change. There was inconclusive evidence for any explanations for these phenomena. Baden and Egglesman (1974) also present data to show that, at least in the short to medium term, drainage for forestry and agriculture increases temporary water storage capacity in mires and thus enhances their power to regulate stream flow. Recent work on Cuilcagh Mountain, Ireland, has shown that runoff increased by around 11 % following peat extraction upslope (Gunn and Walker, 2000). The extra discharge came from winter low flows and was linked to vegetation destruction. With more OLF on bare peat and a dense drain network storm hydrographs were peakier and higher from the cutover catchment. Ditch blocking reduced the flashy nature of the flow from open ditches and produced a response similar to that of undisturbed bog.

Institute of Hydrology (1972) assess the work of Conway and Millar (1960), Hill Farming Research Organisation (1964) and work by the Agricultural Institute on blanket peat in Ireland where drained and undrained plots were being observed (Anon, 1965). In a remarkable ecologically-unfriendly statement they conclude against Conway and Millar that:

"...in the short term, a drained upland or lowland peat may be a better 'sponge' than an intact mire surface. All long-term planning of peat covered catchments must take into account whether it is better to have bare bedrock or an undrained mire."

One of the four small catchments on which Conway and Millar (1960) based many of their conclusions was partly gripped and partly eroded. The area had also suffered severe burning in 1950 and was subsequently named 'Burnt Hill'. Thus Burnt Hill had three facets of response which were amalgamated into one by the use of a single V-notch weir. The two separate dissected sections of this hillslope have never been gauged and recovery since severe burning never tested. Conway and Millar (1960) state that storage on Burnt Hill had been reduced to very low levels such that flows cease during short summer dry spells whereas intact *Sphagnum* covered basins may store up to 15 cm of rainfall, and flows do not cease from intact catchments. Robinson (1985) in his reassessment of the Conway and Millar (1960) dataset suggests that burning reduced low flows but ditching alone did not.

Although the Conway and Millar (1960) paper may have (accidentally) helped to reinforce the traditional views of intact peatlands as aquifers, other studies have shown that this traditional view is incorrect. Bay (1969; 1970) was one of the first to demonstrate that streamflow from peatlands was very irregular. Runoff was measured on four forested bog watersheds ranging from 24 to 130 acres in size in northern Minnesota for 5 years. Spring runoff from the snow melt period of March to June accounted for 66 % of total annual water yield and streamflow ceased on most of the bogs during each summer. Storm analysis showed that 77 % of lag times were between 1 to 3.5 hours (Bay, 1969). The runoff was directly related to water level in the peat deposits. Indeed Bay (1970) shows that runoff ceased at approximately the same water table elevation each year such that water above this height was available for runoff and below this water was retained in the peat deposit. Like Bay (1969), Chapman (1965) also demonstrated a close relationship between runoff and water table. Once water table fell below 8 cm at Coom Rigg Moss, Northumberland, runoff rates were low. This suggests that the bulk of the water movement in the peat system is fairly rapid flow through the surface layer and that movement in the lower layers is at most very slow.

Studies in Germany have shown sharp runoff peaks and long periods without flow from small, undeveloped raised bogs (Baden and Egglesmann, 1964; Vidal, 1960). Labadz (1988) calculated the water balance for a small stream in the headwaters of Shiny Brook, south Pennines. For the second half of 1984, baseflow accounted for 17.7 % of incident precipitation and quickflow 39.6 %, giving a total of 57.3 % of rainfall input being accounted for by stream runoff. Such storm efficiencies were also seen in 1985 and these compare with usual figures for rural catchments of under 5 % of the precipitation returned as quickflow (Burt *et al.*, 1990). The Shiny Brook flow record (Labadz, 1988) is characterised by long periods with little or no stream flow, punctuated by short events with very high runoff. There is little evidence of any delayed flow, indicating that subsurface flow of water through the peat matrix is unlikely to be a major source of runoff. Crisp (1966) calculated that about 80 % of input rainfall on the 0.83 km^2 Rough Sike catchment on the Moor House NNR emerged as stream runoff. This work also demonstrated the importance of fluvial erosion in the degradation of blanket peats at Moor House. Consideration of the erosional system is therefore strongly linked with the hydrology of the blanket peat (Burt *et al.*, 2000).

Few of the early studies of blanket peat hydrology have given much consideration to the hydrological processes generating storm runoff. Given the lack of sophisticated monitoring equipment, the drainage basin has been viewed as a simple input-output system with little understanding of internal process mechanisms being sought (Burt *et al.*, 1997). Burt and Gardiner (1984) measured flow processes in the Shiny Brook catchment, south Pennines and showed that surface runoff processes dominate. Observations of OLF, infiltration rates and soil moisture status all suggested that surface runoff would be produced even during small storms. Hydrograph analysis suggested that storm rainfall rather than antecedent conditions controlled storm runoff perhaps indicating infiltration-excess OLF development. Furthermore clear spatial differences could be found between eroded and uneroded blanket peat moorland areas with a relatively constant contributing area at the eroded subcatchment comprising peat flush channels and bare peat zones. In the uneroded subcatchment, the source areas for surface runoff appeared to be variable depending on how close the water table was to the soil surface prior to the storm event. Evans *et al.* (1999) studied water table and runoff data from the Trout Beck catchment, a tributary of the Tees at Moor House in the North Pennines. It was found that catchment response was flashy, clearly related to quickflow generating mechanisms dominating within the catchment. Water table analysis suggested that saturation-excess rather than infiltration-excess OLF and near-surface flow was dominant; rates of water table rise were rapid indicative of high infiltration rates. Tomlinson (1980) at Brishie Bog, Scotland also finds that water table rises almost immediately in response to rainfall indicating that infiltration rates are relatively high such that OLF generation may be more likely to be saturation-excess than infiltration-excess dominated. Evans *et al.* (1999) note that further detailed plot-scale measurement of processes are required to elucidate more of the runoff generating mechanisms in blanket peat catchments. Some of the results reported by Evans *et al.* (1999) were collected as part of the research reported here. A more detailed set of results is presented in Chapter 4 of this thesis.

2.2 Peat structure and hydraulic properties

Peat structure is dominated by the state of humification of the decaying vegetation matter, controlling water content and movement. Humification is the decay which occurs by biochemical oxidation of the plant matter. It takes place most rapidly in the upper layers of peat where oxygen availability and relatively high temperatures allow breakdown of cellulose in leaves and stems, but will occur slowly throughout the depth of the peat. The most commonly used assessment of humification in peat is that of von Post (1922) (see also Hobbs, 1986) where a small amount of peat is inspected and manually squeezed to allow grading from H1 (no humification, clear water emerges) to H10 (completely humified, peat and water inseparable). Important for peat hydrology is the clear distinction between the upper, periodically aerated and partly living soil layer, the *acrotelm* with low humification levels, and the lower, anaerobic, more humified, permanently waterlogged lower layer, the *catotelm* (Romanov, 1968). Ingram's (1978; 1983) definition of the acrotelm confines the water table to this soil layer. The acrotelm contains the oscillating water table; possesses high hydraulic conductivity; shows a variable water content; is subject to periodic air entry on de-watering following the lowering of the water table; is rich in peat-forming aerobic bacteria and other micro-organisms; and has a live matrix of growing plant material. Ivanov (1981) noted that all the relatively rapid processes of water and heat exchange are connected with the acrotelm. In contrast, the main thickness of the peat deposit, containing peats at different stage of decay (the catotelm) is inactive. The catotelm has a water content invariable with time; possesses a negligibly small hydraulic conductivity; is not subject to air entry; is devoid of peat-forming aerobic micro-organisms and is poor in microbes in general (Ingram, 1983).

Hobbs (1986) noted that the acrotelm layer would appear to be of hydrological significance as a result of its considerably greater hydraulic conductivity. Living *Sphagnum* and other such unconsolidated material have higher rates of flow than areas of decomposed or herbaceous peat (Boelter, 1964, 1965). Dasberg and Neuman (1977) measured properties of peat in the Hula Basin, Israel. They note in line with Ingram (1978) that the properties changed dramatically when peat becomes partially desaturated and therefore one can distinguish between a permanently saturated layer below the zone of water table fluctuations having relatively uniform characteristics and an overlying partially desaturated layer, the properties of which vary with depth. In general their results for hydraulic conductivites are comparable to those of herbaceous peats found by Boelter (1965). Boelter and Verry (1977) related the runoff in the acrotelm to the position of the water table. As it rises, it enters layers of progressively greater permeability enabling water to runoff sideways at greater and greater rates. For most peat soils the pattern is similar; the greatest contrast in permeability occurs between the upper unhumified acrotelm regions and the catotelm. Whilst in highly porous media hydraulic conductivity tends to be high, peats have low conductivity values despite their porosities being high, typically between 60 and 90 % (Dasberg and Neuman, 1977). In Rycroft's (1971) seepage tube tests in the acrotelm of Dun Moss (Ingram *et al.*, 1974; Rycroft *et al.*, 1975b; Bragg, 1982) the result of 3.1 x 10^{-3} cm s^{-1} obtained for the overgrown ditch was fairly typical of hydraulic conductivites in marginal situations (Ingram, 1983). Romanov (1968), however, reported very much higher values in the acrotelms of bogs in Russia. He quoted hydraulic conductivities in the range 10^{1} to 10^{2} cm s^{-1}, decreasing with depth. Rycroft *et al.* (1975) extensively reviewed reported hydraulic conductivity values from catotelmic peats. Values from blanket peats range from 6 x 10^{-8} cm s^{-1} at 1 m (Ingram, 1967) to 1.1 x 10^{-5} cm s^{-1} at 30 cm depth (Galvin and Hanrahan, 1967). Values in other peats tend to be slightly higher (e.g. Dai and Sparling, 1973; Neuman and Dasberg, 1977). In poorly decomposed fenland peats values as high as 5 x 10^{-3} cm s^{-1} have been reported (Romanov, 1968). There are several factors that appear to affect the hydraulic conductivity including: the botanical composition (Boelter, 1965; Bragg, 1982), degree of humification (Baden and Egglesmann, 1963), bulk density (Boelter, 1969; Paivanen, 1963; 1969), the fibre content (Farnham and Finney, 1965; Boelter, 1970), the porosity (Baden and Egglesmann, 1963), and the surface loading (Wechmann, 1943; Hanrahan, 1954). Further details on the compositional and structural properties of peats are discussed in Clymo (1983), Hobbs (1986), Fuchsman (1986) and Egglesmann *et al.* (1993).

Hydraulic conductivity values often seem to depend on the technique used; laboratory results often provide different results from field tests (e.g. Boelter, 1965; Rycroft, 1971). The difficulties of measuring *in situ* hydraulic conductivities on peat have not yet been resolved. Nevertheless, it is clear that within the main peat mass the hydraulic conductivity is generally very low, yet within the upper (acrotelmic) peat layers it may be very high. Generally the acrotelm is thin as it is rare for the water table to fall as much as 50 cm below the surface in peat-covered catchments (Burt *et al.*, 1990). Hence the thin acrotelm may determine that peat-covered catchments are poor suppliers of baseflow during summer recession. Having only a limited capacity to store precipitation inputs once the water table drops below the most permeable layers, throughflow discharge could fall to a very low level as the hydraulic conductivity of the catotelm is low (Burt, 1995). However, Baird *et al.* (1997) stated that it is dangerous to assume that the catotelm cannot be an important a conveyor of subsurface water; because the catotelm is much thicker, even with low hydraulic conductivities it may provide more water over the longer term. Indeed most models of peatland hydrology are groundwater based. Results presented in Chapters 4, 5 and 6 will show that these models are not particularly well-suited to applications in blanket peat hydrology (see Chapter 9).

The movement of water is a controlling ecological factor (Ingram, 1991). Little work has been done on the spatial nature of subsurface flow in wetland soils, or on the spatial structuring of hydraulic properties; work in this area would prove useful for groundwater flow model development for wetlands (Baird, 1995). Burt and Gardiner (1984) provided flow nets for plots in the Shiny Brook catchment, south Pennines. Generally hydraulic gradients are low except near the channel where the water table is drawn down. Heathwaite (1987) mapped hydraulic potentials around drainage ditches on the Somerset Levels using a network of piezometers. Again the hydraulic gradient was low except near the ditch. Having knowledge of the hydraulic gradient and measured saturated hydraulic conductivity, she was able to estimate discharge of water from the peat into the ditch.

The determination of hydraulic soil properties is well reviewed by Hendrickx (1990). Classically, the empirical relation known as Darcy's Law is used to model the relationship between the specific discharge of water and the hydraulic conductivity:

$$Q = K.i.A \text{ or } Q = K.\frac{d\Phi}{dx}.A \qquad [2.1]$$

Where Q represents discharge, K, hydraulic conductivity, A, cross-sectional area, I, piezometric head divided by unit length, x, and Φ represents piezometric head. Head recovery tests are often performed to obtain values of hydraulic conductivity where either water is added to (slug injection) or removed from (slug withdrawal) piezometers and the recovery to the original water level in the instrument is recorded. In less humified bog peats these tests give results consistent with the behaviour expected from incompressible or rigid soils (Rycroft *et al.*, 1975a). In humified bogs however, numerous workers have reported apparently anomalous test results which seem to show that hydraulic conductivity is dependent on the size of the head difference between the piezometer and the surrounding peat soil. Some workers have attributed these properties to non-Darcian flow processes within the peat (Rycroft *et al.*, 1975a; Waine *et al.*, 1985) while Brown and Ingram (1988), Hemond *et al.* (1984) and Hemond and Goldman (1985) have suggested that apparent non-Darcian water flow in certain peats can be explained by the effects of matrix compression and swelling which cause variable water storage within the peat. Baird and Gaffney (1994) applied the response time theory of Brand and Premchitt (1982) for compressible soils and the rigid soil theory of Hvorslev (1951) to a fenland peat. They showed that compression and swelling of the peat matrix affect the course of head recovery in the piezometers used in the study. Brand and Premcitt's (1982) theory has been applied to blanket peats for the first time as part of the present study and results are presented in Chapter 4.

2.3 Peat saturation and water table

The fact that peat behaves somewhat elastically may result in observed seasonal movements of the ground surface ('Mooratmung' – e.g. Ingram, 1983; 1991). Price and Schlotzhauer (1999) found that water table lowering in a mined peatland caused surface subsidence which was shown to be partly due to shrinkage above the water table and partly due to compression of saturated peat. Hence storage changes can occur in association with peat volume changes. Peat volume changes may increase water limitations to plants when the water table drops below the surface.

Peat is typically 90 % water by mass. In view of this exceptionally high water content peat is an extraordinary material (Hobbs, 1986). The water table is the level at which the water pressure is equal to the atmospheric pressure and hence is the level at which water will stand in a well that is hydraulically connected with the groundwater body (Gilman, 1994). Changes in mire water storage are reflected in changes in the position of the peat water table throughout the year. Ingram (1983; 1991) and Hammond *et al.* (1990) showed that the water table is a crucial control on the vegetational distribution of the mire; both the height and fluctuations are important. In undrained peat the water table is close to the surface most of the time. Tomlinson (1980) provided a description of continuous water table records from Brishie Bog, Scotland. Here water table was within 10 cm of the surface most of the year, although individual transects showed variation depending on their spatial position. Godwin (1931), Heikurainen (1963), Tomlinson (1980) and Gilman (1994) all found that once the water table in a peatland falls below a certain level, then further fall may be dominated by evapotranspiration. Ingram (1983) provided an extensive review of peatland evaporation and stressed the significance of interception losses in winter and a possible stomatal control during periods of declining water table. The dominance of the diurnal pulsing of water table decline sequences once below a critical depth (owing to the effect of evapotranspiration) provides evidence that runoff production is dominant in the upper peat layers as the peat cannot drain freely at depth. The flux of water is greatest during periods of high water table because the upper layers have the greatest hydraulic conductivities (Waddington and Roulet, 1997).

2.4 Infiltration

The process of infiltration involves the passage of water through the surface of the soil into the soil mass. Ingram (1983) in his extensive review of mire hydrology makes almost no reference to the process of infiltration into the peat and little information exists elsewhere. Gardiner (1983) and Labadz (1988) both attempted to measure infiltration in blanket peat but comparison is difficult as they used different methods. Gardiner used a ring infiltrometer on cotton grass and crowberry hummocks whilst Labadz used a simple drip type rainfall simulator on bare peat and cotton grass. Table 2.1 indicates the findings from the two techniques. Gardiner suggests that infiltration-excess overland flow is likely to occur on cotton grass moorland since 13 % of hourly rainfall totals exceed 2 mm hr^{-1}, whereas the infiltration capacities are likely to be exceeded rarely on crowberry areas. Burt and Gardiner (1984) found that crest-stage tubes indicated that OLF is a frequent occurrence on the vegetated hillslopes of an uneroded catchment but that in an eroded and gullied catchment it is only frequent on the peat flush channel and bare peat zones. In addition the lower water table and interception by crowberry vegetation and litter layer in the eroded subcatchment will help to restrict infiltration-excess runoff on the ridges. The implications of the infiltration experiments are therefore confirmed by crest-stage tube data; there is a dominance of surface runoff processes.

Peat infiltration rates are reported to be low and although surface detention storage may be high with some surface pools present for most of the year, these are often highly integrated so that surface runoff is quickly generated. Hence Burt *et al.* (1990) claim that the low infiltration-capacity of saturated peats means that peat catchments are probably one of the few natural types that produce the traditional Hortonian infiltration-excess OLF to any large extent. As blanket peat catchments are generally located in headwater basins their importance may far outweigh their areal extent. Hence it may be that Hortonian OLF is more important in contributing to runoff generation from British upland catchments than hydrological research over the past 30 years may have suggested.

Table 2.1 Mean infiltration capacity of blanket peat; data of Gardiner (1983) and Labadz (1988)

Gardiner (1983) – single-ring constant-head infiltrometer – 20 measurements

10 Cotton Grass	10 Crowberry hummock
2.1 mm hr^{-1}	28.7 mm hr^{-1}

Labadz (1988) - rainfall simulation- intensities from 39 to 96 mm hr^{-1}on a total of 5 0.25 m^2 plots, short duration; 2 runs on each plot.

1 Cotton grass	4 Bare peat
No OLF produced	17.7 mm hr^{-1}

For Labadz (1988) however, higher infiltration capacities were obtained on the one vegetated plots studied, largely as a result of lateral flow through the upper few centimetres of decomposing vegetation. It was clear that very shallow subsurface runoff was occurring profusely. It is therefore difficult to say whether surface flow is more often infiltration or saturation-excess dominated. Indeed the problem may lie with differing interpretation of the peat 'surface'. Both Gardiner (1983) and Labadz (1988) used less than 10 measurements at each site. Tricker (1974) performed around 30 ring infiltrometer measurements on a peaty Derbyshire moorland and found infiltration rates ranging from 0 – 265 mm hr^{-1}. There is thus evidence for wide variability although not all of Tricker's (1974) measurements were made on raw blanket peat. There is clearly a need for a much more detailed assessment of infiltration and surface runoff mechanisms on blanket peat.

2.5 Macropore flow

There has been little work done on macropore flow in peatlands. Most of the research carried out has been done on mined peat stockpiles for power stations in order to determine the most productive water retention and rewetting characteristics. Holden and Ward (1996) found that in some rewetted milled peat stores the water content at depth in the profile was greater than near the surface suggesting a short-circuiting of water flow through them. Some evidence came from 'wet fingers' that were observed in the field (Holden and Ward, 1997). Further evidence came from Holden (1998) who examined milled peat from the surface of a drained and air-dried bog. From core samples outflow was similar to the spray rate and little water accumulated in the peat. Bypassing flow paths appeared to form readily.

Discounting soil pipes, the only work done to the author's knowledge on intact peat to quantify macropore flow was that of Baird (1997) who used a tension infiltrometer to assess the relative roles of matrix and macropore flow at the surface of a fenland peat. Macropore flow appeared to dominate the flux and could play an important role in the infiltration process. Baird (1997) only ran 17 tests on surface peats. No work has been done on macropore flow in blanket peat and no work has been done to quantify subsurface macropore flow in peats (except on pipes).

Macropore flow may be greater following dry weather due to cracking of the peat. Desiccation and associated shrinkage can lead to cracking of the peat and associated changes in the hydraulic properties of the soils. There is a vast literature on the effects of cracking on infiltration and the redistribution of water in non-wetland soils (see Germann, 1990). Little information is available for wetland soils. Cracking has been implicated in soil piping in blanket peats but few attempts have been made to measure and analyse crack flow in peats (Jones, 1981).

2.6 Pipeflow

The importance of surface erosion has long been known, but during the 1960s and 1970s with more observations on the complexity of flow generation in drainage basins and of the impact of subsurface flow on storm hydrographs (eg Whipkey, 1965; Hewlett and Hibbert, 1967; Kirkby and Chorley, 1967; Ragan, 1968; Dunne and Black, 1970a), increasing evidence for subsurface erosion features began to emerge (e.g. Berry, 1964; Ward, 1966; Jones, 1971; Heede, 1971; Temple and Rapp, 1972; Baillie, 1975). Increasing interest, particularly in subsurface flow and erosion processes and their relationship to peat hydrology (eg Bower, 1961; Radley, 1962) and badland geomorphology, resulted in the first extensive reviews of research on piping (Gilman and Newson, 1980; Jones, 1981; Bryan and Yair, 1982).

Jones (1981) and Anderson and Burt (1982) noted the need for increased investigation into soil piping. Since then a limited amount of field research has been pursued yielding enhanced information on the role of piping in hillslope drainage, flood generation and channel development. Nevertheless, as recently as 1997 Jones noted two major problems bedevilling assessment of the role of piping. The first was the lack of measurements; the second, the difficulty of finding and defining the networks. Unlike macropores and small cracks, pipes are not present on all slopes or in all catchments. In peats a change in surface vegetation may often indicate the presence of a pipe (Jones, 1991). Jones and Crane (1984) extensively mapped 4.4 km of pipes in a drainage area of only 0.23 km^2 by dye tracing and ground survey. The peats studied were thin (of the order of only 30-50 cm deep) and no work has been done to monitor pipeflow in deeper blanket peat. Pipe locations were identified mainly by observation of collapse features, of water jets emerging from pipes and the sound of flowing water (Jones, 1982). Often destructive techniques must be used to investigate the pipes (Jones, 1981) and there have been few other attempts to accurately locate and map subsurface piping.

Apart from the Maesnant and Upper Wye studies, other work in Britain has been based on point discharge samples rather than continuous records of storm hydrographs. Work has also been based mainly in shallow peaty soils. Pipeflow may in some locations add significantly to the subsurface contributions to storm flow (Jones and Crane, 1984). Pipes can transfer water rapidly. Little is known about the interlinkages of pipeflow with other flow processes, nor of the form and spatial density of many pipe networks. In peat catchments pipes have been found at the interface between the peat and the underlying mineral layer, and also solely within the peat itself usually where there is rapid change in peat properties such as bulk density (Gilman and Newson, 1980).

The initiation of piping is discussed in detail by Jones (1981; 1990) although process measurement has never been performed. Flow through desiccation, biotic and mass movement cracks may enhance the macropores into pipe networks. Jones (1994) demonstrates that the vast majority of catchments examined with piping in Britain are south-facing suggesting that desiccation cracking may be very important in the formation of piping. Gilman and Newson (1980) and McCaig (1983) find 'thresholds' of rainfall that occur before pipeflow occurs in peatland catchments. It is suggested that this threshold corresponds to a real critical level above which saturated conditions in the soil must rise before lateral drainage occurs in macropores and pipes. Gilman and Newson (1980) suggest that in peat soils this level is related to the depth to which large scale desiccation cracking occurs during dry summer months. Gilman and Newson (1980) observed that vertical cracks in both the roof and floor of peat pipes were a common feature in mid-Wales during the dry summer of 1976, and suggested that this allowed more water to reach deeper levels and created permanent extensions of the pipe networks when re-wetting took place. Bower (1959) observed 'massive cracks' during dry periods in the Pennines which could gape as much as 6 cm at the surface and run perpendicular to the peat face; 'some cracks probably penetrate to a depth of as much as 5 ft' (about 1.5 m). Furthermore the process of 'boiling' (Terzhagi and Peck, 1966) may result from water emerging from a soil surface under pressure causing particles and aggregates to become weightless and 'boil away'. Pop-out failures on the sides of previously developed pipes and macropores may also occur due to this process.

As noted in Chapter 1, most pipe discharge records come from British blanket peat. Anderson and Burt (1982) present data from two types of pipe in the blanket peat of Shiny Brook catchment. One represented a quickflow response closely mirroring the overland flow discharge peak, and the second with a response lagging two hours behind the stream discharge. The first was a near-surface pipe and the second was at depth within the peat mass. Gardiner's (1983) study suggested that pipeflow provided an insignificant amount of storm runoff at Shiny Brook except where very short near-surface pipes (less than 1m depth) linked pools to the main stream channel. Burt *et al.* (1990) noted that it may be that pipeflow is more important in shallow peats whereas in deep blanket peats, pipeflow from the impermeable catotelm will necessarily be restricted. However, this conclusion should be treated with caution as Gardiner (1983) did not monitor the larger pipes in the catchment. This is not to deny that wide variation may exist.

Pipes at Measnant range from 90 mm in diameter on the upper slopes to 240 mm at the stream bank. Three broad categories of pipe flow can be identified in the pipes: perennial, ephemeral and seasonal. Jones (1987) noted that there was a great variety of hydrological response between one pipe and another even within a small area, hence extrapolation is difficult. Piping complicates the descriptions of the nature and pattern of dynamic source areas such that both topology and pedology are critical factors in determining the distribution of quickflow in headwater catchments (McCaig, 1983). Jones (1987) reports that piping doubles the dynamic contributing area in the upper Maesnant, mainly through linking points well beyond the riparian zones of seepage to the stream. Both discharge and sediment transport rates in the major pipes are closely related to the size of shallow surface microtopographic hollows in which they lie, and which themselves are largely created by piping erosion. Thus piping can play a large role in landscape development (Jones, 1990). At Maesnant, pipe discharges are frequently generated by contributing areas larger than the surface depressions and some pipes run counter to the surface topography.

It is likely that there are several sources of pipeflow water. Overland flow can enter collapse features where the pipe is open to the surface and water can also run down cracks and root channels to enter the pipe network. This is seen at some pipe inlets at Maesnant (Jones, 1982). Jones (1982) also claims that main ephemeral pipes also collect both surface and subsurface drainage in shallow bowl-shaped collecting areas of around 5 cm at their heads. For perennial and seasonal pipes clearly the perennially saturated source areas provide a key source. However, there is still much to be done in order to understand the full range of processes operating around piped zones. Pipes receive water more quickly than would be expected from diffuse seepage through the upper layers of peat (Jones, 1978; Gilman and Newson, 1980; Jones and Crane, 1984; Roberge and Plamondon, 1987; Jones *et al.*, 1991). It certainly does not seem that pipeflow is restricted by, as Whipkey and Kirkby (1978) suggested, the long time it takes for rainwater to infiltrate down as far as the level of pipes. Nevertheless Sklash *et al.* (1996) using isotope analysis show that pipes may transmit predominantly 'old' water (water from previous storms) and richer in deuterium during storm flow. This does not prove slow delivery, however. It could be displaced water by the translatory or piston flow mechanisms (Hewlett and Hibbert, 1967). Even so, in an adjacent catchment, Hyett (1990) found strong chemical evidence that stormflow in the pipes is dominated by fresh storm rainfall. Pipeflow can also lower the pH in surface waters by diverting seepage away from mineral layers and reducing residence times (Jones, 1990).

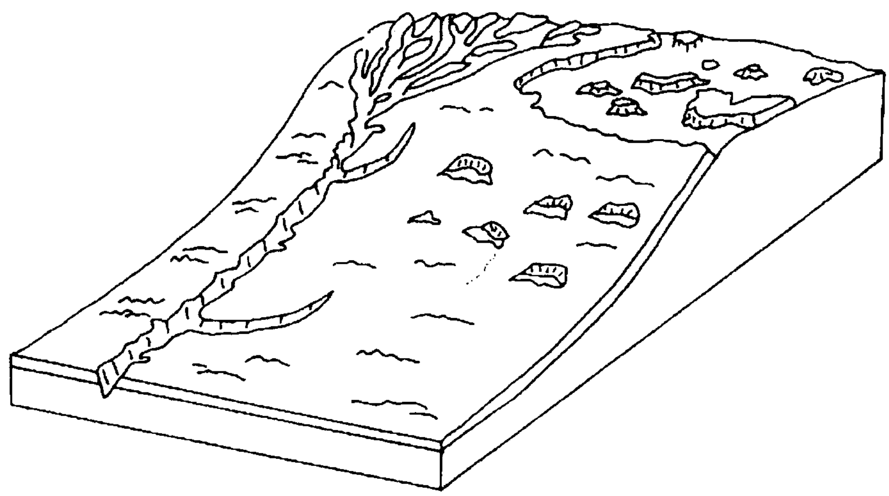
Jones (1990) shows that piping can play an important role in landscape development in some regions. Piping is involved with channel extension through roof collapse often forming gullies (Higgins, 1990). The role of piping in land slips and bank failures in peat is still poorly understood, even by the standards of piping research as a whole (Jones, 1997). Although Jones (1975) describes observations from Northern Ireland which suggest piped bogs are more stable and that piping may stabilise the peat by preventing excessive build up of hydrostatic pressure, he does admit that pipe capacity can be exceeded to create a build up in pressure (Jones, 1981). What is certain is that pipes are not straight, linear passages, which allow water to flow freely. Rather they tend to be tortuous, constantly changing in cross-section (e.g. Gilman and Newson 1980). In many circumstances, water flow is too great to be freely transmitted along these subsurface tunnels and a pressure head builds back up the pipe. In this way subsurface pipes provide routes to the peat base for water which can cause a build up in hydraulic pressure. If a heavy rainstorm occurs, it is possible that restrictions in the dimensions of the pipes acting as bottle-necks will impede flow. Hence water builds back up the pipe and large pressures may build up within the peat mass down toward the substrate. Frequently seen in blanket peat during and after heavy rainfall are numerous springs and even jets of water on the peat surface escaping from the pipe networks as visible evidence of these large hydraulic pressures (e.g. Gilman and Newson,1980; Jones, 1982). Newson (1975) describes 'slide pipes', which are revealed after rapid mass movements of peat during heavy rain. However, it is not known whether these exist prior to the slope failure. Gilman and Newson (1980) link the small crescentic slips which are common in the Pennines and the Plynlimon areas (which Bower (1960) calls 'arcuate tears') to the seepage of water and with pipes. Here the slip is a near a convex break in slope where some concentration of subsurface or surface flow occurs. Investigation of pipes at these locations suggest that it is possible both the pipe and the slip can derive from the same tension crack across the peat hillslope.

2.7 Peat erosion

Erosion of blanket peatlands is widespread in upland Britain (Tallis, 1997). Just as there has been limited detailed research into the runoff characteristics of blanket peat moorlands there has also been little research into peat erosion. Conway (1954) demonstrated the frequency of erosion in the blanket peats of the Pennines. She notes that "few people can see severe peat erosion for the first time without a sense of astonishment". In the decade following Conway's classic paper, a series of investigations were carried out on the morphological features exhibited by eroding peats (e.g. Bower, 1960) and on the possible causes of their erosion (e.g. Bower, 1962; Johnson and Dunham, 1963; Tallis, 1964). These studies emphasised the need for further intensive work on peat build-up and degradation dynamics.

Bower (1960, 1961) divided peat erosional agents into two types: water erosion (through dissection, sheet erosion and marginal face recession) and mass movement (peat slides or bog bursts). It is likely that mass movements in peat are strongly controlled by hydrological processes. There is also evidence to suggest that wind erosion, perhaps in combination with rainsplash, could be an important agent of peat transport (Radley, 1962; Figure 2.1a). Francis (1990) has demonstrated that deflation of dry peat surfaces plays an important role in the lowering of peat surfaces during times of drought. The blanket peat of the Pennines is often heavily dissected. Bower (1960, 1961) divided the dissection systems into two: *Type 1* occurs on flat areas ($< 5^{o}$) of deep peat (> 1.5 m) where a close network of freely and intricately branching gullies exist. *Type 2*, are much more open gully systems with less branching and are more common on sloping ground where gullies form sub-parallel trenches running downslope (Figure 2.1b). Mosley (1972) attempted to quantify Bower's argument. He concluded that it was difficult to identify two distinct types of system and they are probably end-members in a continuum, although plentiful examples of Bower's two types can be found. Bower's dissection types are still widely used today (Burt *et al.*, 2000).

a)



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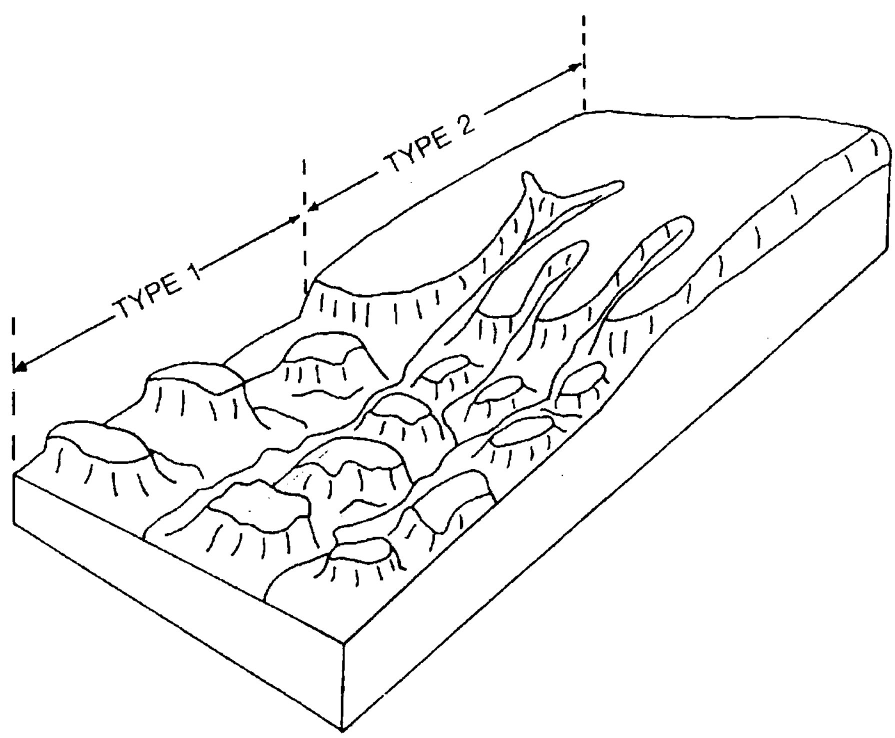


Figure 2.1. Theoretical erosion systems in blanket peat. a) Radley's two types of summit erosion: dissection by water (left) and wind erosion (right) (from Radley, 1962). b) The relationship between Bower's Type 1 and Type 2 dissection systems (from Bower, 1960).

In order for the gully dissection systems to develop, Bower (1961) suggested three mechanisms; headward erosion from the margins, as runnels on the surface, and along lines of weakness within the peat. She also argued for three stages in gully development: early shallow stage; advanced, incised and narrow v-shaped stage, and thirdly a late stage where gullies are wide and separated by small and widely scattered peat islands. It is not clear whether all gullies must necessarily evolve from the early to the advanced stage; evidence presented by Tallis (1994) suggests that a hummock and pool topography can develop within an intact blanket bog, providing drainage without disruption of peat. However, Tallis (1973) suggested that erosion would be inevitable along topographically defined flow-lines once bare peat was exposed. This may occur due to drying of the peat surface during drier climatic periods, by burning, overgrazing, or by air pollution killing sensitive moss species (Tallis, 1965, 1985; Lee *et al.*, 1987; Skeffington *et al.*, 1997). This line of argument goes against Bower's conclusion that peat erosion was the natural and inevitable culmination of peat accumulation and that burning and draining merely encouraged erosion but did not cause it.

Severe peat erosion is problematic for reservoir managers (e.g. White *et al.*, 1996). The low density of peat means there is relatively little in-channel storage until flow velocities are dramatically reduced upon entering a lentic system (Labadz *et al.*, 1991). Thus reservoirs provide good sites for investigation of sediment yields in peat catchments. Calculations of the timing of the onset of erosion based on current sediment yields are complicated by the fact that erosion levels may not have been constant. Furthermore, if erosion onset was earlier than 150 to 200 years ago, information is unlikely to be apparent in reservoir sediments. Radionuclides (e.g. Pb-210) may in future provide evidence of changes in sediment yield. Nevertheless, the 200-year erosion rate from reservoir surveys combined with assessment of contemporary rates presented by Labadz *et al.* (1991) compares favourably with palynological evidence presented by Tallis (1964, 1985, 1994, 1995). Some pools and gully floors show some signs of recent recolonisation, particularly in the North Pennines which shows that peat accumulation is still possible, such that Burt *et al.* (2000) suggest widespread peat erosion today is more likely to be a result of human impact than natural mire decay.

At Moor House NNR investigation has shown that the sources of sediment are not only from OLF on bare peat sections but also from bank and gully collapse (Evans and Burt, 1998) with large blocks of peat moving through the catchment, slowly disintegrating in streams or with rain splash (Warburton and Evans, 1998). Often gully systems are connected to the main channel via small alluvial peat fans such that sediment erosion, transport, and deposition are complex events within the storage–yield system. Crisp (1966) and Crisp and Robson (1979) have shown that individual runoff events account for the bulk of peat transported, but their use of bulk samples collected over a period of hours has allowed comparison only between mean discharge and mean peat transport rates. Re-instrumentation of Crisp's (1966) weir on Rough Sike at Moor House NNR has provided evidence for high sediment concentrations on the rising limb of the hydrograph (Evans and Burt, 1998; Burt *et al.*, 2000, see also Chapter 5). This relationship is indicative of sediment exhaustion whereby the supply of readily mobilised material is quickly depleted (Webb and Walling, 1984). Burt and Gardiner (1984) found that peak suspended sediment concentrations occurred at Shiny Brook in the late summer and early autumn which implies that desiccation of the surface peat in summer combined with intense rainbeat is the main agent of erosion. Because of this, sediment loading may vary with aspect (Bower, 1959; Francis, 1990) and with season (Francis, 1990; Tallis, 1975). Tallis (1975) used peat traps to estimate volumes of peat erosion in the southern Pennines. Large flows caused greater error in the sampling but clear patterns are still discernible. Substantial peat erosion is shown to occur during snowmelt and during heavy rain, when stream flow rates exceed 40-50 $l\ min^{-1}$. Needle ice can often be seen on bare peat, loosening and preparing the peat for removal (Bower, 1959; Burt *et al.*, 2000). Francis (1990) found, however, that peat supply to streams was much greater in the autumn and early winter; the suggestion was that summer desiccation had prepared the peat for removal, but as the winter progressed sediment exhaustion occurred and frost action was of minimal importance. The two opposing results may be related to sediment storage and release mechanisms, and to the nature of the coupling between bare peat areas and streams in the area of study. Furthermore the fieldwork of Francis (1990) during 1983 and 1984 occurred during two atypical dry years (Burt, 1985).

2.8 Water Colour

The discolouration of water supplies from peatland areas is a widespread natural process (Butcher *et al.*, 1995). Discolouration produces problems for management and several measures including transfer network management, allowing highly coloured waters to be diverted from the water supply system, and land use management controls, have been suggested (Pattinson *et al.*, 1994). In the case of Yorkshire Water, some 45 % of the water sent into public supply is obtained from direct supply reservoirs usually draining peat areas (Butcher *et al.*, 1995). Colour in natural waters has very similar properties to known organic fractions such as humic and fulvic acids (Shapiro, 1957) and there is also strong evidence that colour intensity is strongly related to the amount of iron in the water (Hem, 1960). Naden and McDonald (1989) demonstrate the marked seasonality in water colour with peaks in the autumn following warm summer weather. A number of workers have described the rise in levels of water discolouration in supply catchments in the UK since 1976 (McDonald *et al.*, 1988). Colour more than doubled in the 1980s in some Pennine catchments (Naden and McDonald, 1989). This rise has generally been attributed to an increase in the number of drought years and to a lesser extent to changes in the management of upland catchments. Drying out of the peatlands has possibly been a product of both these causes (Butcher *et al.*, 1995). Colour has been shown to rise in the years after a significant drought such as those of 1976 and 1984. Experimentally Mitchell and McDonald (1992) found that particularly high water colours were found not in the autumn immediately following a drought, but in the following autumn flush. It is suggested that this lag time is related to the time taken for the peat to wet back up again. Colour appears to be associated with water table lowering and the aerobic decomposition of the upper organic layers. Thus climate change and peat hydrology are intimately linked to water quality problems in the uplands.

Burt (1979) shows that there can be an important relationship between runoff generation and the solute concentration of soil and stream water such that results can be used in order to explain landscape evolution. Nitrate analysis on stream and soil water samples at Moor House NNR has demonstrated that concentrations found in deeper peats are not reflected in stream water samples whereas there is a clear link between water in near-surface peat (within the upper 10 cm) and the stream water (Adamson *et al.*, 1997). This is in line with the suggestion that catotelmic drainage may be slow, with much flow taking place through the upper soil layers.

2.9 Climate change and peat hydrology

Verry's (1984) 22-year record from a Minnesota mire shows that water tables remained high except during the 1976 severe drought when water table declined to what would have previously been defined as the catotelm. Thus Verry (1984) claimed that Ingram's (1983) definition of acrotelm/catotelm boundaries needed changing to account for drought events. However, the definiton itself may not need changing; the peat properties themselves may change with aeration and drying not normally associated with the lower layers. A drop of water placed on wet peat spreads over the surface of the peat; the angle between the water droplet and the peat tends to zero because the wet peat is hydrophilic. On dry peat, water drops do not spread, but form contact angles between the water and the peat of up to 85^{o} , especially at low pH values. Thus dry peat is water repellent or hydrophobic. The difference in the wetting behaviour of dry peat and wet peat influences the pattern of water movement in the peat and the extent to which precipitation infiltrates the surface (Egglesmann *et al.*, 1993).

Marsh and Turton (1996), Marsh and Sanderson (1997), Burt *et al.* (1998) and Conway (1998) note the recent volatility of the climate in the UK. They suggest that the general stability of the climate with little temporal variability in peatlands may not continue. Flood and drought may recur more frequently and recent summer droughts over the past twenty years have placed great stress on water supplies locally. Conway (1998) presents some climate models which suggest increases in temperature and precipitation for northern England and Scotland. However, summer precipitation may not increase greatly and if it does, then rainfall intensity rather than the number of rain days may increase. According to Church and Woo (1990) this may result in changes in hillslope hydrological processes; evaporation and desiccation increase, shorter snow seasons and more intense rains are likely to occur with future climate change. As we have seen there are important implications for peatland ecology and erosion as well as for hydrology.

It is still difficult to assess what the full effects of increased drought may be for peat hydrological processes. In terms of infiltration, cracking associated with peat surface drying (e.g. Newson and Gilman, 1980; Bower, 1959) may result in increased infiltration rates down the cracks and changes in the runoff generation processes. The bypassing flow work done on milled peat by Holden (1998) suggests that some drying of the peat may result in enhanced macropore flow when the peat is rewetted. The crusting that simultaneously occurs may decrease infiltration rates, however, encouraging further OLF and erosion. As discussed above there is some historical evidence that previous phases of peat erosion may have initiated during, or at the end of, dry periods (Tallis, 1997). Higher temperatures and lower water tables are likely to increase rates of humification in the upper layers of peat and may lead to increasing frequency of severe water colour events exacerbating the management problems for water supply agencies. Carbon flux from peatlands is also closely related to water table conditions (Silvola *et al.*, 1996). If climate change in Britain means warmer drier summers then oxidation of the peat surface is likely to result in increased CO_2 fluxes from British blanket bog. Carbon fluxes may also be enhanced if increased climatic seasonality induces physical erosion of the peat surface; oxidation of detrital peat in the fluvial system will be further enhanced (Evans *et al.,* 1999). There is evidence, however, that CH_4 production from peatlands is reduced as water tables are lowered (Roulet *et al.*, 1992). Hence the effects of climate change on blanket peat are likely to be complex and much more research is required.

2.10 Summary

Successful management of blanket mires requires understanding of how they function hydrologically. Water quality, erosion and carbon flux are all strongly related to the hydrology of the blanket peats in much of the British uplands. The discussion of the available literature on peat hydrology has demonstrated that much remains to be learned. Wetland research has been concerned primarily with botanical and ecological aspects and with the utilisation of peatlands for forestry and agricultural purposes. This has left a gap in our understanding of the process hydrology of wetlands and hence of many headwater catchments. None of the early studies of blanket peat hydrology gave much consideration to the hydrological processes generating storm runoff. With a lack of sophisticated monitoring equipment, the drainage basin was viewed as a simple input-output system with little understanding of internal process mechanisms being sought. The literature on peat hydrology is largely concerned with soil water storage, with most field studies being conducted by botanists rather than by hydrologists. Even Bay (1969), who established that peat was a poor regulator of flow and Conway and Millar (1960) who examined the effects of drainage and burning on the hydrological response of peat hillslopes, seemed somewhat preoccupied with the development and accuracy of suitable techniques for hydrological measurement rather than with discussing the runoff production processes (Burt *et al.*, 1990). Hence the differentiation between catchments was speculatively attributed to the water storage capacity and nature of the ground surface.

While the acrotelm-catotelm model provides a useful starting point for understanding runoff generation, surprisingly little work has been done on the hydrology of the upper peat layer. With high root densities and high a hydraulic conductivity it may be that macropores play an important role in the runoff response of the acrotelm. There have been very few attempts to measure the spatial and temporal variation in runoff production in blanket peat. Given the apparent importance of OLF in peatlands it is crucial to be able to describe its generation as it can produce the highest peak runoff with the shortest response times (Dunne, 1978) and is therefore vital to the understanding of storm hydrographs. Little work exists on measuring the infiltration capacity of blanket peat and so it remains difficult to quantify the relative roles of infiltration and saturation-excess OLF mechanisms. It is also evident that attention needs to be given to the role of subsurface pipes in generating runoff in blanket peat catchments. Pipe networks are often difficult to identify and few continuous hydrograph records from pipes exist.

It is clear that a multifaceted and multi-dimensional approach is required in order to determine the processes that govern runoff from peat catchments. Hewlett and Hibbert (1963) note that

"*we can, of course, record outflow from a watershed, but there is a real need for improved concepts for determining the source and timing of flow...from such knowledge, watersheds can be evaluated as moderators of water flow and future behaviour under specified conditions may be predicted with greater precision*".

Burt *et al.* (2000) argue for a catchment-scale management approach to peatland usage and problems. A 'bottom-up' catchment planning approach with an alliance of interest groups may well be the most efficient and effective means of sustaining a variety of land uses on the blanket peat moorlands of the UK. In order that the management has sufficient science to back its policies (although politics, and economics will be amongst the many other forces in operation) an increase in process-based field studies will be required. However, this may prove difficult without an appreciation of the landscape being a product of an integrated set of processes each of which are operating on smaller and disparate spatial and temporal scales.

This thesis includes investigations at a variety of spatial and temporal scales and will necessarily involve a mixture of field and laboratory experimentation. Research will be confined to the blanket peats of the North Pennines and in particular those of the Moor House NNR. This easily accessible blanket bog is a World Biosphere Reserve. Monitoring of terrestrial and freshwater ecosystems as part of the Environmental

Change Network (ECN) is underpinned by almost 50 years of research at this important field site for environmental science. The area comprises the headwaters of the River Tees and thus Moor House provides an ideal location for study of blanket peat hydrology.

CHAPTER 3

FIELD SITE AND METHODOLOGY

"Many a black wellington and fell boot, armed with microscope, plastic bag and 10 inches of weatherproofing, have emerged mid-Winter from the back door of Moor House in search of the magic species that make up a Ph.D." (Bellamy and Quayle, 1989).

3.1 Selection of study site

This thesis is concerned with the processes responsible for runoff generation in blanket peat. In order to study process, a mixture of monitoring and experimentation was required. The field site therefore needed to be an accessible location in a typical area of blanket peat moorland. The success of the project relied heavily on the collection of storm-based data and on experiments which required suitable weather conditions for their operation. Being able to drive to the study site at short notice, as well as having the opportunity stay at the site over night or for extended periods were important factors in site choice. The ability to transport experimental and monitoring equipment to the site was crucial to the choice of field area. Of practical importance was the need to choose a study site where permission of land access and instrumentation could be gained. Added incentives to site selection would include areas where long term meteorological and hydrological datasets were already available and could provide the context and background information for the thesis.

The field site that satisfied the above criteria with the added bonus of being a base for a large amount of earlier research, primarily in ecology and geology was Moor House National Nature Reserve (NNR) in the North Pennines (Figure 3.1). The region around Moor House contains important headwaters which supply the River Tees, Wear and South Tyne. Research is actively encouraged by the landowners (English Nature) and groups such as the Centre for Ecology and Hydrology. Although much scientific work has been done in the area, studies of the hydrology are sparse. In part, this may reflect the difficulties of working in the uplands and also the previous lack of reliable monitoring equipment that can withstand the harsh climates of these upland areas. Nevertheless, whilst research is sparse, the hydrological work at Moor House is still probably the best and most comprehensive in the UK uplands. Some of the most important early work on peat hydrology and erosion was done at Moor House with the

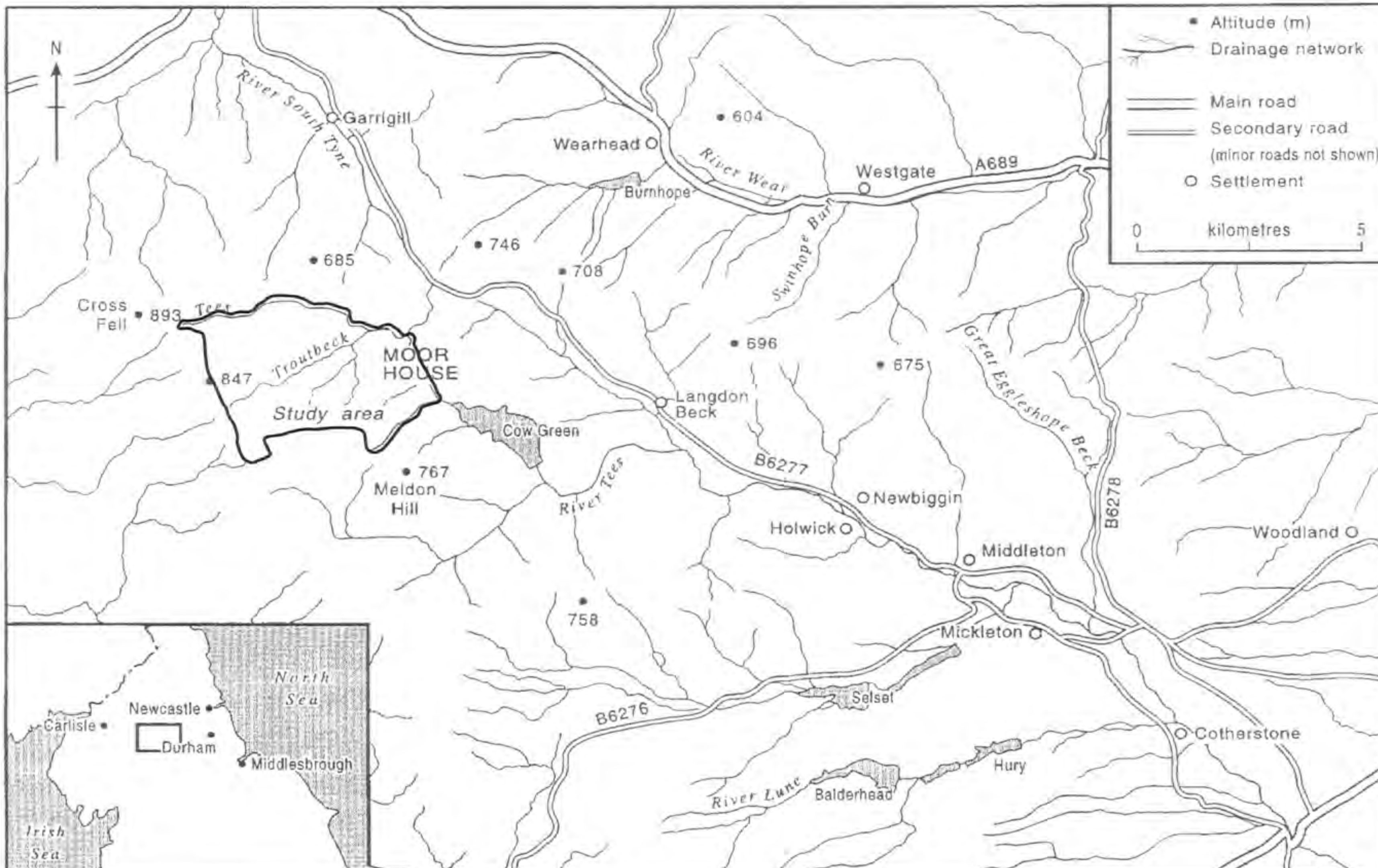


Figure 3.1. Map showing the location of the Moor House NNR in the North Pennines, after Warburton (1998).

work of Bower (1959; 1960), Conway and Millar (1960) and Crisp (1966) being most notable. Results from the revisiting of the Conway and Millar site will form part of this thesis and re-instrumentation of Crisp's original weir on Rough Sike (Figure 3.3) has also been done recently; some early results are reported by Evans and Burt (1998) and Burt *et al.* (2000).

3.2 General description of the study area

The North Pennines are an area of upland moorland at the northern end of the Pennine chain. Most of the area lies above 450 m O.D. with the highest point being the summit of Cross Fell at 893 m O.D. The 'Alston Block' structural unit, which has been active since Devonian times, is bounded to the north by the Tyne Gap (Stublick Fault) and to the south by the Stainmore Trough (Lunedale Fault) (Figure 3.2). To the west the Pennine Fault produces a large sloping escarpment to the edge of the Vale of Eden. To the east the block is more gently tilted under the Durham Coalfield. The area is characterised by an upland landscape of high, open and exposed plateaux and broad ridges which support moorland and montane habitats with few trees (Figure 3.3). Large expanses of blanket peat overlie mineral soils that have developed on glacial, solifluction and alluvial materials (Johnson and Dunham, 1963; Johnson and Hickling, 1970).

3.3 General description of Moor House and Upper Teesdale NNR

The Moor House reserve was acquired by Nature Conservancy Council in 1952 making it the first NNR in England. It is one of the largest areas of blanket bog in Great Britain and is now a World Biosphere Reserve. The site is therefore recognised for its worldwide importance. The reserve occupies approximately 35 km^2 and has an altitudinal range of 290 to 848 m O.D. It extends from the upper edge of the enclosed land in the Eden Valley, over the Dun Fells, to the upper end of Cow Green Reservoir on the River Tees (Figure 3.4). The Tees rises on the edge of the reserve and forms its northern boundary. The geology is Carboniferous in age, with alternating strata of limestone, sandstone and shale into which there are intrusions of the Whin Sill dolerite (Figure 3.3, Johnson and Dunham, 1963). The overlying glacial till has resulted in poor drainage which has led to the development of blanket bog on around 70 % of the reserve (Johnson and Dunham, 1963). The vegetation is dominated by *Eriophorum* sp. (cotton grass), *Calluna vulgaris* (heather) and *Sphagnum* sp. (moss). The land is owned by English Nature and provides free-range common grazing (mainly sheep) for villages in

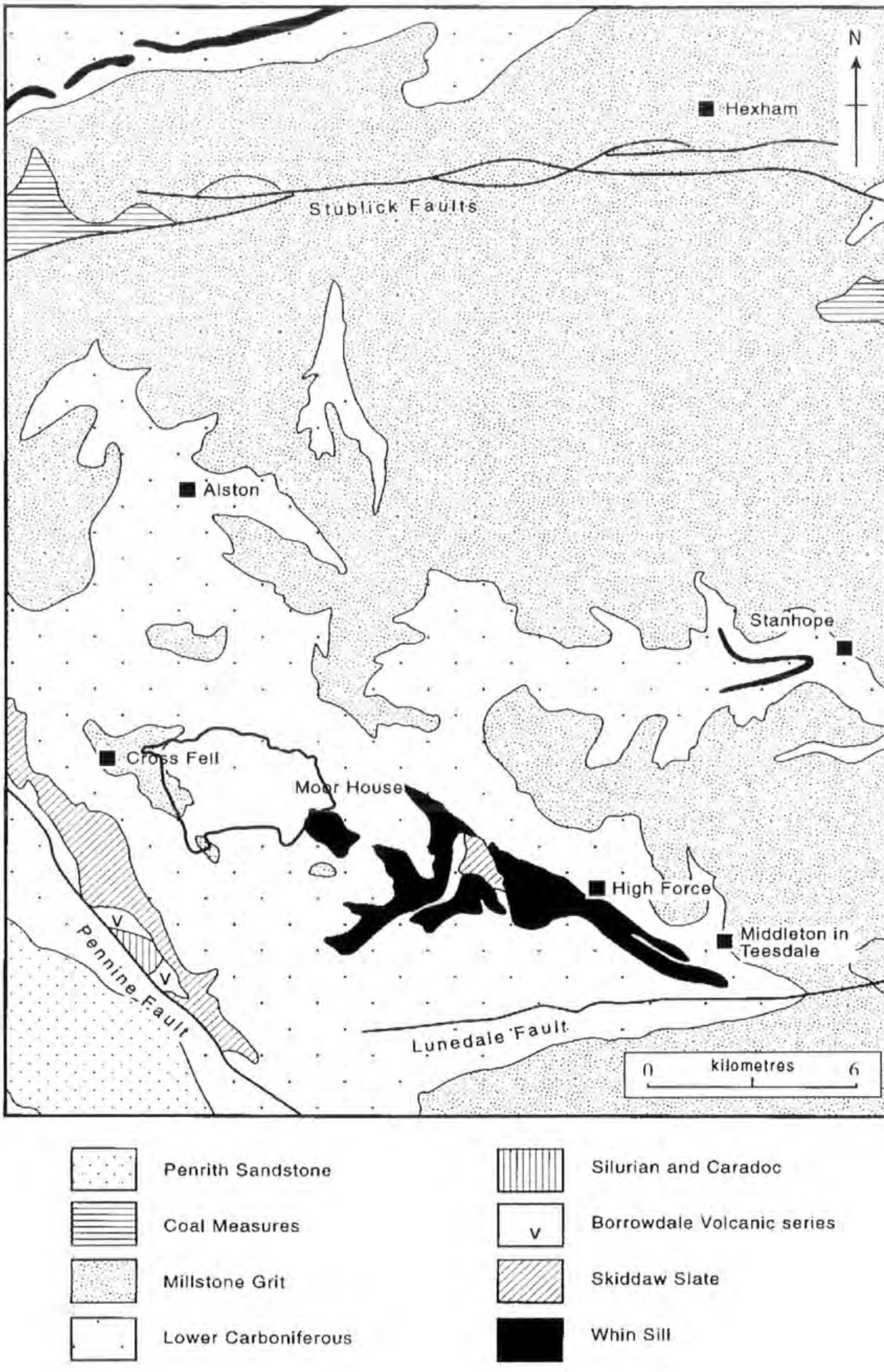


Figure 3.2. Geological map of the central North Pennines (after Johnson and Hickling, 1970).

Figure 3.3. View west across the Moor House NNR towards the peaks of (right to left) Cross Fell, Little Dun Fell and Great Dun Fell on the summit ridge of the Pennines. The gauging station in the foreground is a re-instrumentation of Crisp's (1966) original site on Rough Sike.

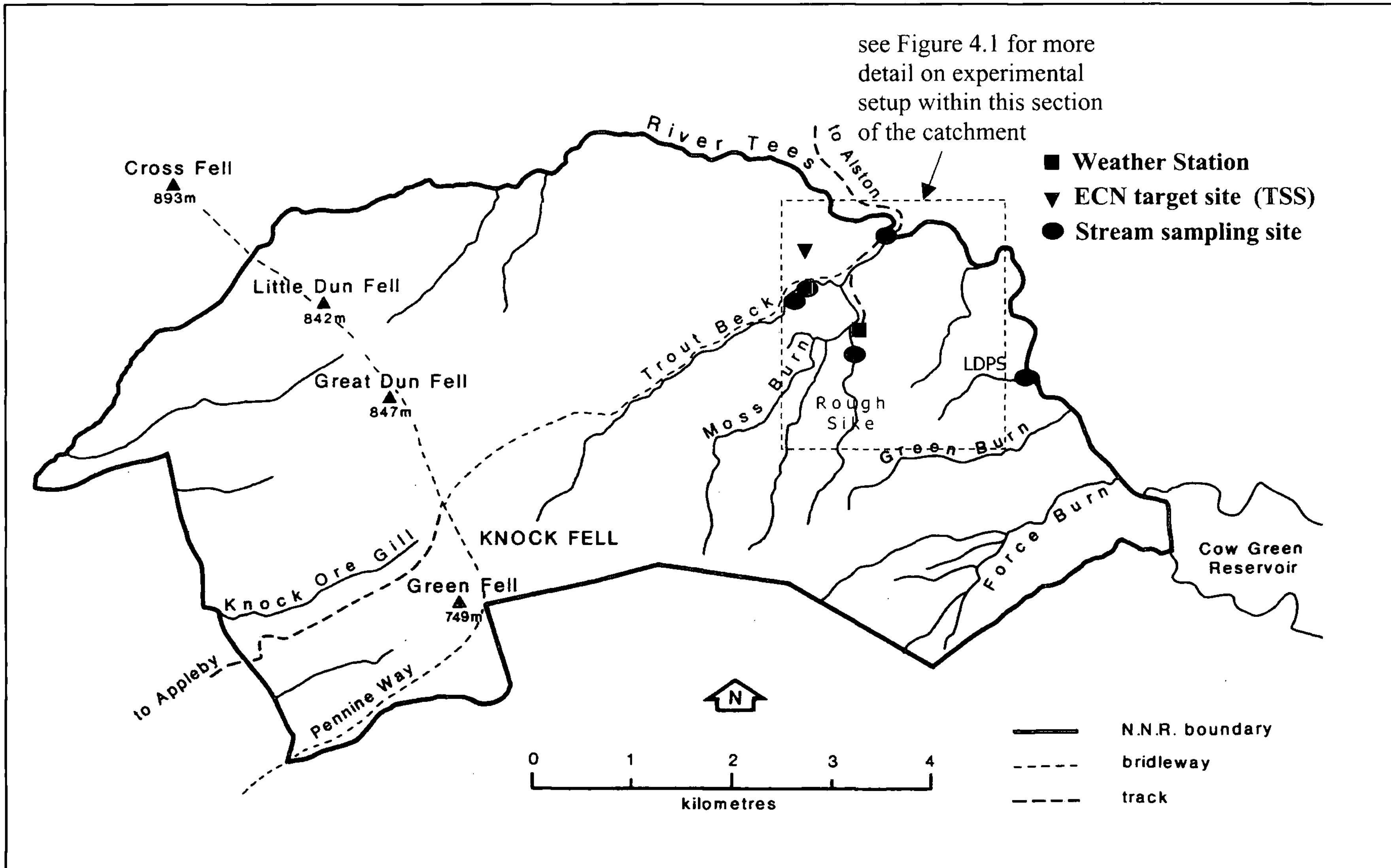


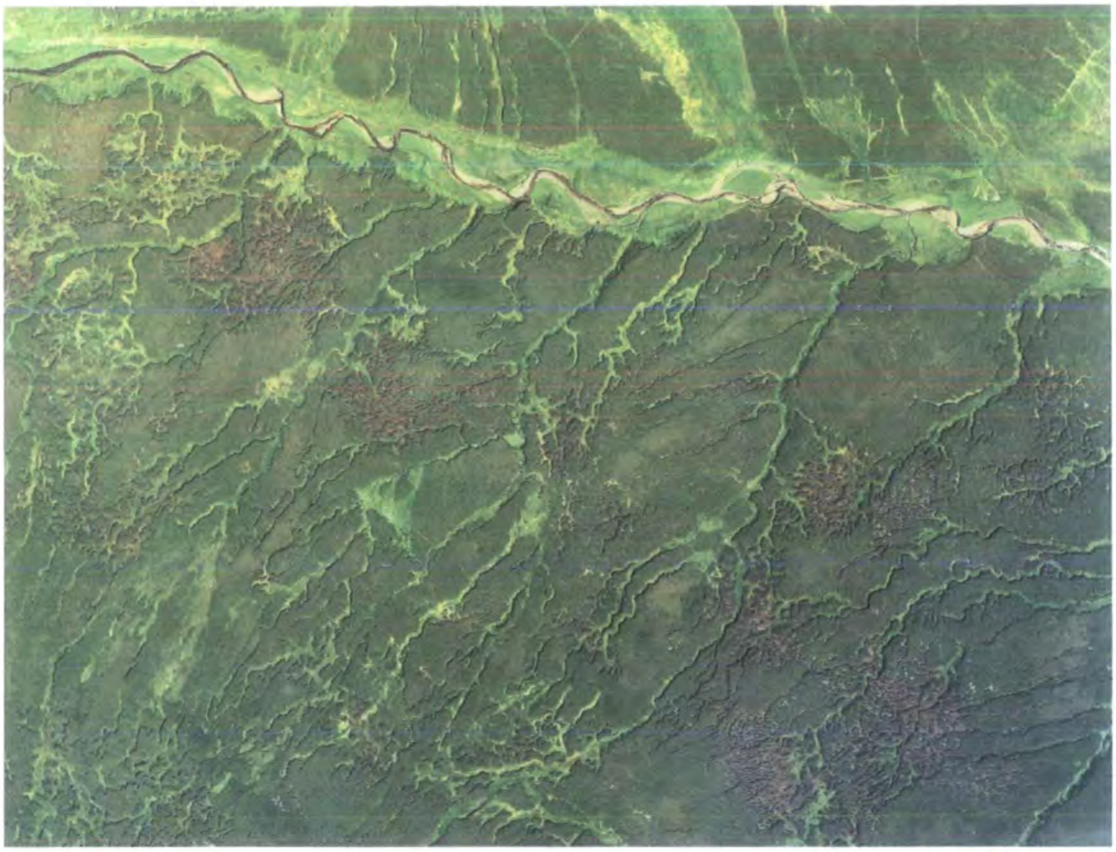
Figure 3.4. Main sampling points in the Trout Beck and Little Dodgen Pot Sike (LDPS) catchments on the Moor House NNR (see also Figure 4.1).

the Eden Valley. Upper Teesdale NNR bounds the south of the Moor House NNR running from Cow Green reservoir to the summit of Mickle Fell and eastward to High Force waterfall on the Tees. Evidence exists for dwelling in the north Pennines since Mesolithic times and flint and chert tools as well as remains of ancient cattle have been found on the Moor House reserve (Johnson and Dunham, 1963). Roman activity has been recorded in the area and mining for iron, lead and other minerals has occurred on the reserve for centuries. Vehicular access to Moor House is via a track from Garrigill, the nearest village, some seven kilometers from the old field station.

3.4 Peat development at Moor House

By 7000 years ago mixed deciduous and pine woodland became established on all but the most exposed summits of the north Pennines (Warburton, 1998). During the early and middle Post-glacial, Mesolithic and Neolithic peoples initiated small-scale clearance of the woodland, first for hunting and then for agriculture. Peat started to replace woodland some 3800 years ago in Upper Teesdale (Pounder, 1989) but in other areas of the uplands this may have started earlier (Taylor *et al.*, 1971). By Roman times the majority of the Pennine uplands would have been cleared of woodland (Atherden, 1992).

Almost all of the Moor House reserve is covered by blanket peat up to around 700 m O.D. which varies in thickness from a few centimetres to four metres on flat ground. Eddy *et al.* (1969) estimated that 18% of the peat cover within the Trout Beck catchment (Figure 3.4) was eroded. Although there are many areas of bare peat, most of the areas of gully erosion have now revegetated with *Sphagnum* and *Eriophorum;* this is in contrast to the Southern Pennines where extensive areas of bare peat remain (Labadz *et al.*, 1991). The aerial photograph shown in Figure 3.5 is typical of the North Pennine blanket peats which are heavily dissected by revegetating gullies. Garnett and Adamson (1997) digitised soil maps based on work by Johnson and Dunham (1963) and Eddy *et al.* (1969) on a 22 km^2 area of the Moor House Reserve. They found that 80 % of this site was covered by blanket peat. Eroding mire occupies 8 % of the area, whilst areas once eroded but now recolonised by vegetation occupy nearly 10 % of the study site. On average they suggest eroded peat is more likely on gentle slopes than steeper slopes. This conforms with Bower's (1960) 'erosion Types' (see Chapter 2). Garnett and Adamson (1997) may overestimate actively eroding sites and underestimate recolonised areas. Revisiting sites where Bower (1959) took photographs of the erosion on the



N

Approximate scale 1 : 12 000

Figure 3.5. An aerial photograph of a northern part of the Moor House NNR showing the densely dissected nature of the blanket peat. The photograph was taken at an altitude of 1524 m (5000 ft), NERC site 94/9 (4), taken 6.8.95, run 5, plate 8887. Reproduced with kind permission of NERC.

reserve suggests that much recolonisation has taken place since the late 1950s (J. Warburton, pers comm, see Chapter 4). This probably continued during and after the work of Johnson and Dunham (1963) and Eddy *et al.* (1969).

Figure 3.6 shows a typical intact blanket peat profile at Moor House. The upper 5 cm consists of poorly humified (H2-H3 on the Von Post (1922) scale) black brown coloured peat with living roots and a crumb structure. Below this to 10 - 15 cm the peat tends to be brown and slightly humified (H3-H4) with occasional bands of light brown *Sphagnum* peat overlying a darker brown *Eriophorum-Calluna-Sphagnum* peat (H4). The soil then very gradually becomes more humified with depth. By 1.5 m into the profile the peat is highly humified with decomposition almost complete (H9). Frequently there are well-preserved remains of birch found at the base of the peat which overlies a light coloured grey clay with sandstone boulders. The clay is often strongly gleyed and waterlogged. This clay then rests on glacial boulder clay. Further information on the peats at sites on the Moor House reserve can be found in Johnson and Dunham (1963) who estimate that peat formation began around the Boreal-Atlantic transition.

3.5 Background data and research at Moor House

Moor House NNR provides an excellent study site not only because of its location but also because of the background of research that has been done on the Reserve since the 1950s and the ongoing monitoring operations that continue today. The house (NGR NY 757 328), after which the area is named, was originally a miners bothy or shop and can be identified on an 1825 mining map (Johnson and Dunham, 1963). The house later became a hunting lodge. In 1952 it was aquired by Nature Conservancy Council and operated as a field station until closure at the end of 1979. The author was one of only four people who witnessed the full demolition of the house on 7th August 1999, the crushed remains of which were used to repair the access track leading to the field site from Garrigill. A small portacabin owned by English Nature is now the fieldworker's only shelter from the severe Moor House climate (see below). It was from the old field station that much work was carried out, mainly on the ecology of this unique upland environment; the flora of upper Teesdale and Moor House are probably more widely known than that of any other area in Britain (e.g. Heal and Perkins, 1978).

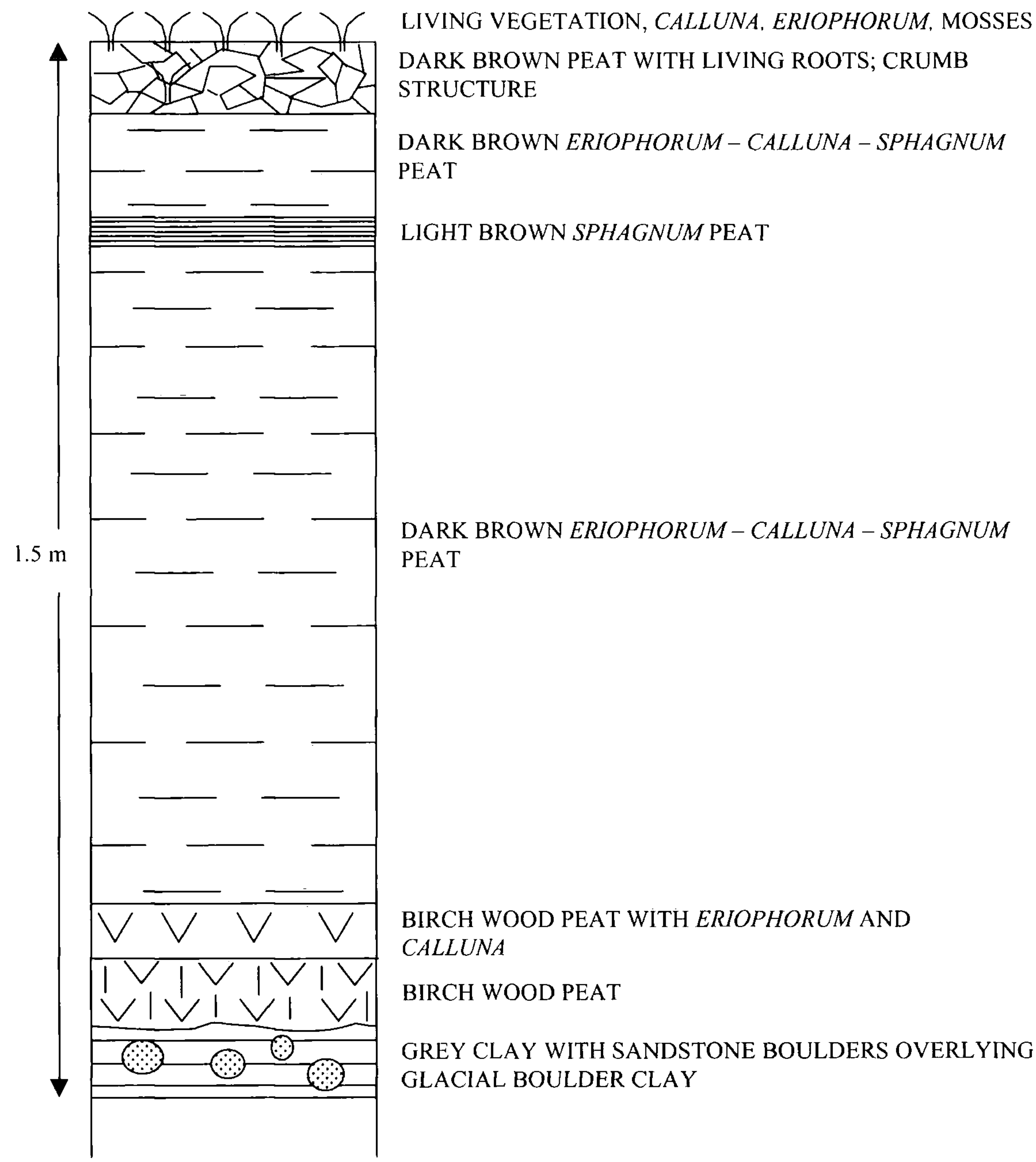


Figure 3.6. A typical profile through intact blanket peat at Moor House.

Meteorological records at Moor House began in the 1930s; at an altitude of 560 m this was then by far the highest point at which such readings were regularly made in the UK, with exception of Ben Nevis (Burt *et al.*, 1998). A record of daily observations exists from 1 May 1952 to 31 December 1979. An AWS was set up on the site in May 1991 which records on a hourly basis. Weekly manual checks at British Meteorological Office specifications are also made.

The Moor House reserve is a UK Environmental Change Network (ECN) monitoring site. The ECN was launched in January 1992. It is a multi-agency long-term research programme to record, analyse and predict environmental changes across the UK. The main objective of ECN is to maintain a network of sites within the UK from which to obtain comparable long-term data sets. This requires measurements at regular intervals of variables identified as being of major environmental importance; protocols for standard measurements can be found in Sykes and Lane (1996). The programme is envisaged to last a minimum of thirty years (Burt, 1994). Recent concerns with climate change, biodiversity and pollution have pushed the importance of long-term observation and monitoring, which for a long time was thought of as low grade science, to the front of agendas for environmental research (Burt, 1994). The ECN collects three types of information: general descriptive information on site characteristics and archiving of past records; routine collection of meteorological, hydrological and hydrochemical data; and monitoring of biological responses. At each ECN site a 0.01 km^2 target sampling site (TSS) is identified within which most of the measurements are made. The location of the Moor House TSS is indicated in Figure 3.4. Data on water table fluctuations and hydrochemistry are among the information collected at the Moor House TSS which will be examined by this thesis. All ECN sites monitor stream discharge: at Moor House, a gauging station opened in October 1957 on Trout Beck, a headwater tributary of the Tees, (see Chapter 4) was reinstated by the Environment Agency in July 1991 after having been closed in June 1980.

3.6 Climate at Moor House

Mean annual temperature based on AWS data (at 556 m O.D.) from 1992 - 2000 was 5.8°C. No comparable data are available for the earlier period of meteorological recording at Moor House because data were based on maximum and minimum thermometers and not automatically logged. Annual mean maximum and minimum temperatures were 8.3°C and 1.9°C respectively between 1953-1978 (Smithson, 1985).

Crudely this suggests a 5.1°C average for the earlier period. Mean monthly temperature data from the AWS are shown in Figure 3.7a. With temperatures hovering around zero for much of the winter the potential for freeze-thaw activity is high. Temperatures can be extreme with values below −15°C recorded in most winters. Air frosts have been recorded in every month of the year and Moor House generally has over 100 days in which the dry bulb air temperature falls below freezing; the climate can be classified as sub-arctic oceanic (Evans *et al.*, 1999).

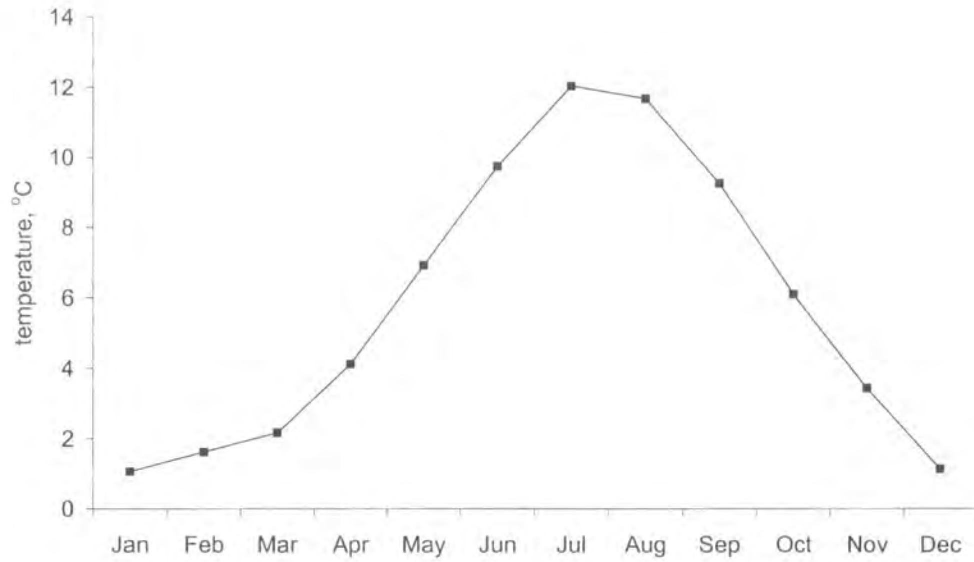
Based on data from 1953 – 2000 mean annual precipitation is 1953 mm with an average of around 240 precipitation days per year (Table 3.1). This is very high but can vary considerably from year to year with 1345 mm recorded in 1971 and 2930 mm in 1979. Burt *et al.* (1998) use the nearby Widdybank Fell observations to infill the missing Moor House records using the regression equation MH = 25 + 0.99WF (R^2 = 61 %) and this has been used to calculate the long-term mean. Between January 1992 and December 1999 mean annual precipitation at Moor House was 1993 mm and from 1953 to 1979 it was 1934 mm. Monthly mean rainfall values are plotted in Figure 3.7b with the recent trend towards enhanced seasonality (Jones and Conway, 1997; Burt *et al.*, 1998) indicated by the lower precipitation totals during summer months and higher precipitation totals during the winter months of the ECN period. The 1995 water year was exceptional in that it combined the fourth wettest winter on record (since 1952) with the second driest summer (Burt *et al.*, 1998).

Table 3.1. Summary of meteorological observations at Moor House (NY 757 328, 556 m O.D.).

Variable	Value, from AWS 1992-2000	Value, based on Smithson (1985)
Mean temperature, °C	5.8	
Days temperature fell below 0°C	107	
Mean wind speed	4.3	
Median wind direction, degrees north	221	
Mean annual precipitation, mm	1994 (1953*)	2010 (1941-1970)
Annual precipitation days	237	247 (1956-1979)
Annual fog days (visibility <1 km at 0900h)		52 (1953-1979)

* calculated from 1953-2000 using Widdybank Fell correlation for 1979-1991

a)



b)

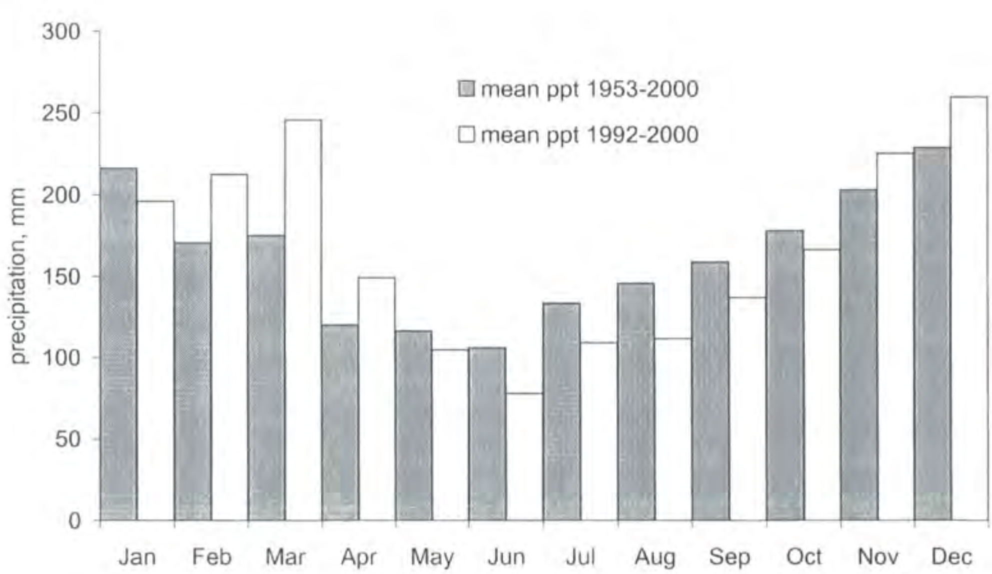
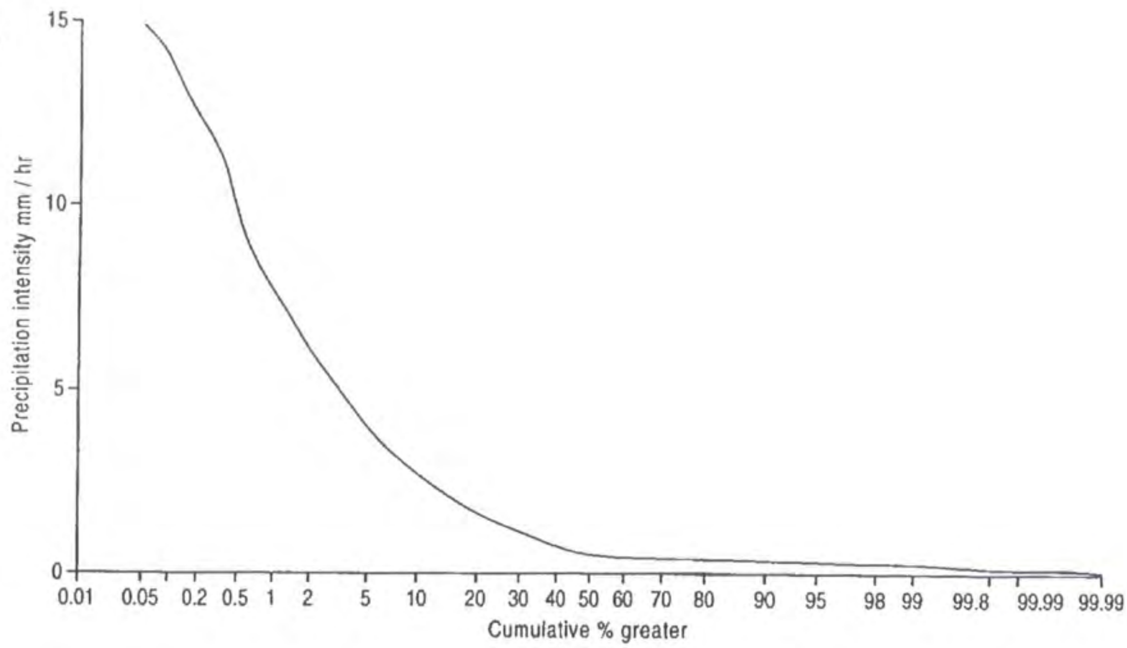


Figure 3.7. a) Mean monthly temperature at Moor House during the ECN period based on hourly averages, b) Mean monthly precipitation at Moor House comparing the ECN period with the long-term average.

c)



d)

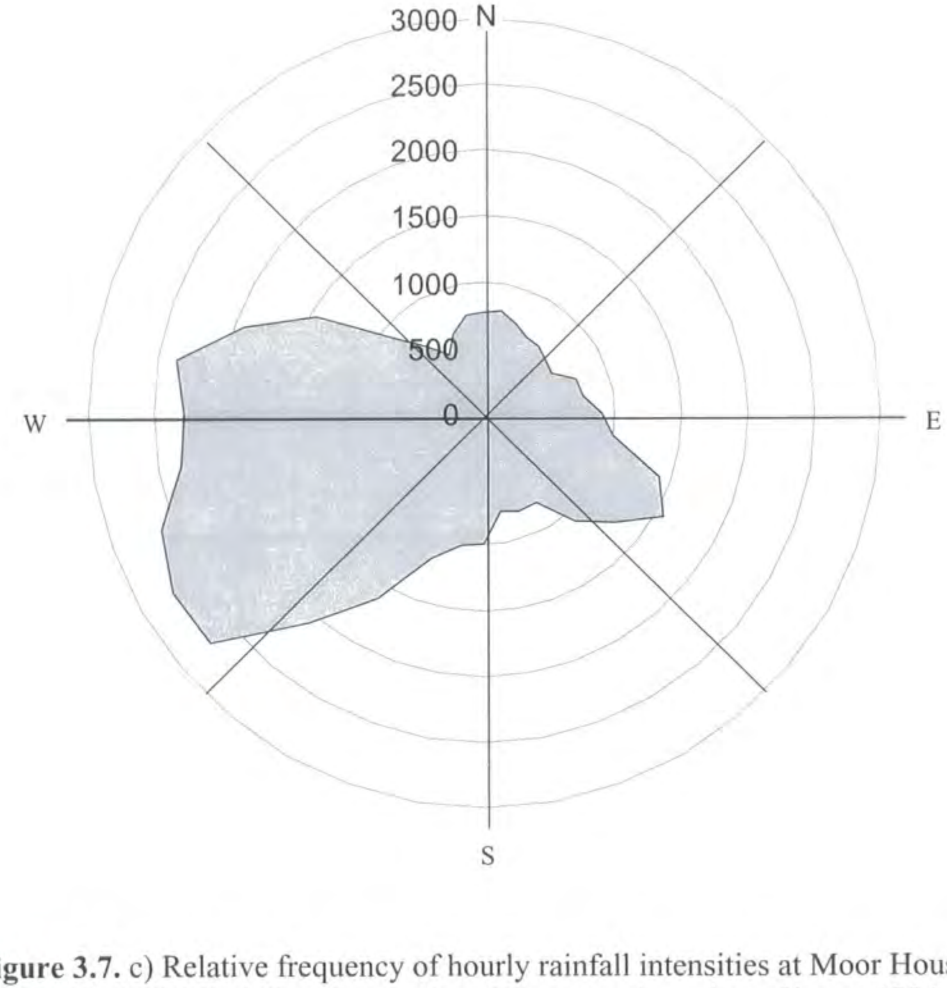


Figure 3.7. c) Relative frequency of hourly rainfall intensities at Moor House during the ECN period, d) Wind direction at Moor House, total number of hours 1994 – 2000, 10° bin width.

The relative frequency of hourly intensities at Moor House as a proportion of all hours with rainfall recorded from the AWS is plotted in Figure 3.7c and shows the dominance of lower-intensity frontal and orographic rainfall at the site. Westerly and south-westerly moist air masses from the North Atlantic dominate the climate as indicated by the hourly wind direction data from 1994-2000 (Figure 3.7d). Between September 1994 and June 2000 the maximum number of consecutive days without precipitation at Moor House was 14 (summer, 1995). Ten days without precipitation was exceeded 8 times during that same monitoring period, with periods of a week or more without precipitation occurring 17 times. Gilman and Newson (1980) note that summer-time desiccation of the peat in the Welsh mountains of the Upper Wye occurs regularly with dry periods of 16 consecutive days having a two-year return period.

Snowfall adds considerable uncertainty to the precipitation measurements. At higher elevations in the North Pennines a significant amount of winter precipitation falls as snow. Average annual snow cover increases from 55 days at 500 m to 100 days on the summits (Archer and Stewart, 1995). Table 3.2 lists the maximum recorded rainfall intensities for each half year of the AWS record from October 1994 – 1999. Very high intensity records in the winter months may be affected by snowmelt. For example the figure of 40.2 mm hr^{-1} in winter 1995 immediately followed a three hour period when the temperature rose from below freezing to 6°C. Relatively high summer rainfall intensities are similar to the 9.9 mm hr^{-1} maximum reported by Carling (1983) for a one-year record at similar elevation in Great Eggleshope Beck further down Teesdale. The peak winter intensities reported by Carling were 4.5 mm hr^{-1} and the discrepancy may again relate to problems with snowfall measurement. The rain gauge at Moor House is unheated. This minimises over-estimation of total rainfall due to drifting snow since the drifted snow is not quickly melted and replaced. However, the unheated gauge does record high hourly intensities during melt periods. The gauge is sited in a Plynlimon pit to optimise performance for wind-blown rain. This, however, increases the possibility of snow drifting over the gauge. Snow cover at the site is synoptically controlled and a typical winter season will see several complete accumulation and melt cycles. The longest continuous period of snow cover at the site during the ECN period to date, as indicated by albedo measurements, was 24 days (19/12/96 – 11/1/97). Melt episodes can have a significant effect on catchment runoff as the stored water is released from the snowpack. The largest floods at Moor House are generally associated with rain-on-snow events (Evans *et al.*, 1999). Whilst snow pillow data is available from nearby

Widdybank Fell (Archer and Stewart, 1995) and elevation-dependent snowmelt models have been derived for the Trout Beck catchment (Bell and Moore, 1999), it is clear that more work is required to improve the quality of Moor House meteorological and hydrological datasets.

Table 3.2. Maximum recorded rainfall intensities at Moor House (summer and winter in mm hr^{-1}) for the 1995 – 1999 water years.

	October-March	April-September
1995	40.2	9.4
1996	15.0	8.0
1997	14.6	6.2
1998	19.4	10.2
1999	24.6	9.0

Another potential problem with hydrological work at Moor House is that the ECN data come from a single station in the lower altitudes of the reserve. Crisp (1966) reports occasions when significant storms on Rough Sike were generated despite there being no recorded rainfall at the meteorological station and attributes this to localised rainfall at higher elevations. The ECN maintains a rain gauge at the top of Great Dun Fell (855 m) but problems with blowing snow are likely to be more severe at this site. Even if a large network of rain gauges was available, determining the true precipitation to the various subcatchments in the area would be made difficult, at least for the winter months. Unpublished data cited by Crisp (1966) collected by Conway and Millar (1960) in the course of catchment hydrology experiments at Moor House suggest that total spatial variability may not be excessive. Four weekly gauges sited within 1.6 km of Moor House showed 5 % deviations from the Moor House gauge over a year. Nevertheless none of these gauges were at a significantly higher elevation than the main Moor House site. Work is currently underway by Andy Joyce (University of Durham) to examine the nature of microclimates at the site through use of an array of automatic weather stations in a variety of settings.

3.7 Research methodology

3.7.1 Location of experimentation and instrumented sites

The main research sites used are shown in Figure 3.4. with more detail given in Figure 4.1. Most of the fieldwork has been done within the Trout Beck catchment where earlier work (e.g. Conway and Millar, 1960; Crisp, 1966) has taken place. The ECN database also provides background information for the Trout Beck catchment. Little Dodgen Pot Sike (LDPS), a much smaller tributary of the Tees, examined in detail in Chapter 7, provides an entirely new site for research. At LDPS streamflow from the catchment has been gauged but soil piping and pipeflow processes are the major components of investigation. Within the Trout Beck catchment the other dominant runoff processes in blanket peat are investigated through monitoring and experimentation. Peat from the Trout Beck catchment is also used in laboratory studies of infiltration and runoff generation before and after drought simulation.

3.7.2 Spatial and temporal scales

To establish the runoff-generating processes which operate in blanket peat catchments it is necessary to observe how the peat behaves hydrologically before, during and after rainfall. This requires intensive field study of sections of peat hillslope small enough for practicable investigation, ideally combined with laboratory experimentation. Fieldwork is an essential function, defining realistic boundary and initial conditions within which to experiment or model. Fluvial and hillslope geomorphology has witnessed a continuing reduction in the time- and space-scales of research, with increasing emphasis on the dynamics of small, site-specific projects (Lane and Richards, 1997). Implicit in this trend is a move away from the mathematical generalisations based on surrogate or 'output' variables to a more detailed study of the physical mechanisms and system state variables associated with particular events. The methodology used in this thesis represents the shift towards the understanding of hillslope hydrology rather than the past concerns that were more with engineering forecasts of floods which resulted in hydrologists ignoring the physical characteristics of the headwaters and the exact processes responsible for generating runoff. Nevertheless, small plot-scale research needs to be placed within the context of catchment-scale hydrology. Research has shown that it is possible to move from small-scale descriptions of fluvial processes to the explanation of larger systems (e.g. Clifford, 1993; Lane and Richards, 1997). The traditional idea that different scales of form and process are causally independent of each other cannot be sustained. Short time-scale and small space-scale processes undoubtedly influence processes over longer time-scales and larger space-scales. Thus, throughout the thesis, results from intensive short-term plot-scale studies and event-based hillslope-scale monitoring and experimentation will be used in parallel with longer-term catchment-scale datasets.

3.7.3 Experimentation and monitoring outline

The main models of peatland hydrology quoted extensively in the literature are groundwater based such as the 'groundwater mound model' of Ingram (1982). This revolves around the nature of water held within the main body of the peat mass and the fluxes that may occur within the peat as related to the shape of a hillslope, or 'mound' of peat. Ingram (1982) stated that to improve and explore the model further ' we need data on the permeability of the deeper catotelm and on groundwater discharge as a water budget item'. Whilst models such as MODFLOW (Harbaugh and McDonald, 1996) and DRAINMOD (Skaggs, 1980) can be used to describe groundwater alone, they inherently lack any adequate representation of the surface water. The blend of field investigation, with experimentation will allow the validity of using these types of models in blanket peat to be tested. These data collected will provide information on surface, near-surface and groundwater flow processes at a variety of temporal and spatial scales. As well as testing the validity of groundwater flow models this should also allow more accurate and useful models of peat hydrology to be developed; not least because the data may aid model parameterisation.

The practical measurement and sampling techniques used in this thesis will be discussed in the relevant chapters and are not dealt with separately here. Figure 3.8 indicates the main experiments and monitoring operations used. The work is heavily field-based but the rainfall simulator experiment did require many laboratory hours. Monitoring work encompasses both automated data collection via dataloggers and manual sampling at regular intervals. Experiments are performed mainly at the plot scale, with monitoring exercises generally covering scales from the plot to the hillslope. Runoff plots, for example, were installed at footslope, midslope and topslope locations on a small hillslope thereby allowing simultaneous process investigation across a hillslope. Streamflow analysis was performed on Trout Beck, LDPS and for two hillslope subcatchments.

The major flow processes under investigation are indicated in Figure 3.8; these are simplified for ease of interpretation such that, for example, OLF covers both infiltration-excess and saturation-excess processes. Nevertheless the diagram acts as a guide to the links between the investigative techniques used for this thesis. The links drawn are specific to this thesis and it is recognised that other links between techniques could have been employed. For example, water table monitoring and crest-stage tube mapping

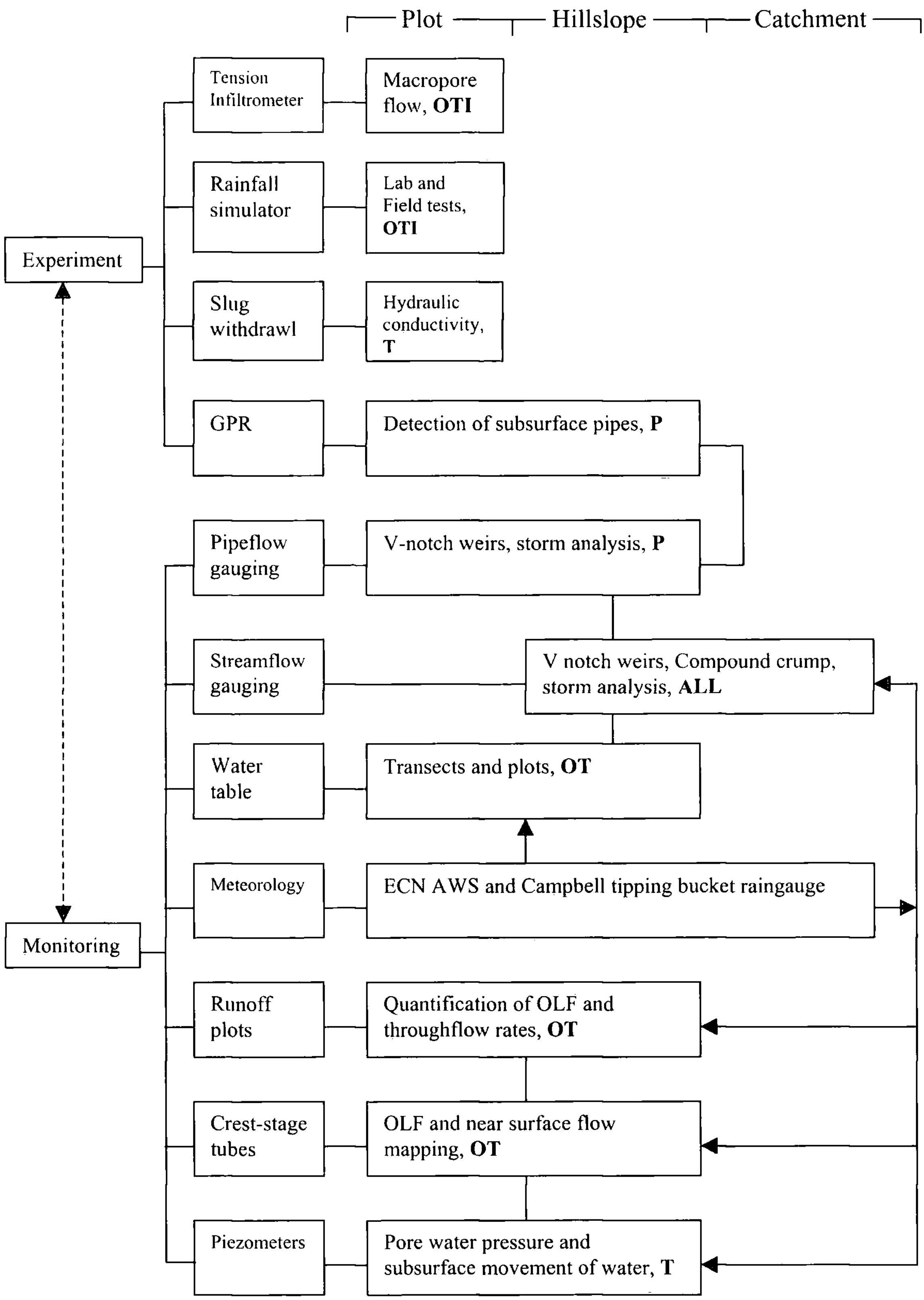


Figure 3.8. Flow chart showing the links between the main experimental and monitoring techniques used in the thesis and the scale on which the work operated. Main processes of investigation are indicated by O = OLF, T = throughflow, I = infiltration, P = pipeflow.

could have been undertaken in conjunction with pipeflow monitoring. However, given time and financial constraints it is was not possible to study all aspects and all interlinkages between the processes involved in blanket peat hydrology; this would indeed require several Ph.D. theses! The role of this thesis is to elucidate more about the operation of hydrological processes within the upland blanket peatlands and the work concentrates on the aims outlined in Chapter 1.

CHAPTER 4

RUNOFF MONITORING IN BLANKET-PEAT: CATCHMENT-, HILLSLOPE- AND PLOT-SCALE PROCESS MEASUREMENT

4.1 Introduction

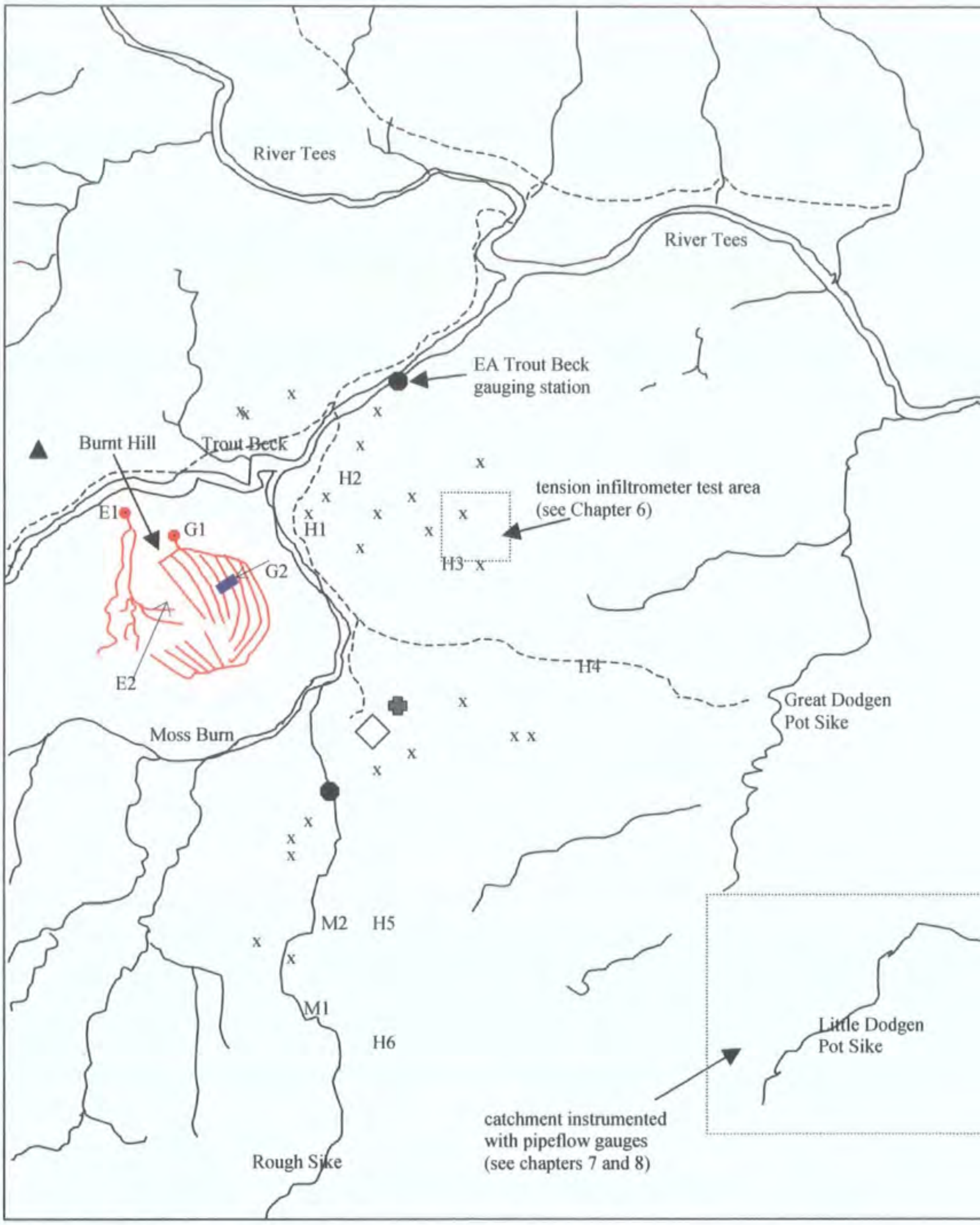
This chapter focuses on the Trout Beck catchment which lies almost entirely within the Moor House NNR. Automatic gauging of this catchment is combined with digitally logged weather station and water table data from the site and with gauging of subcatchments. Tipping-bucket flow recorders allow the runoff processes which produce the catchment and subcatchment scale runoff response to be simultaneously monitored. Mapping and measurement of surface and subsurface flow processes over long time periods (several months) and during storm events on plots and hillslopes will also be presented. These data allow a more detailed understanding of the spatial and temporal generation of runoff in blanket peat.

Trout Beck rises on the slopes of Great Dun Fell at an elevation of about 800 m. The catchment (11.4 km^2) is gauged by a compound Crump weir operated by the Environment Agency (register number 025003) at 535 m, NY 756326 (Figures 3.4, 4.1 and 4.2); stage height is logged every 15 minutes. More detailed characteristics of the catchment and climate are described in Chapter 3 and by Johnson and Dunham (1963) and Burt *et al.* (1998). The ECN target sampling site at Moor House is on a gently sloping fairly homogeneous open hillslope 900 m north-west of the automatic weather station (AWS) at an elevation of 570 m (see Figure 4.1). Here water table is automatically monitored in one dipwell by a pressure transducer. Weekly manual checks are made of water table in the dipwell and at four other dipwells nearby. These data in conjunction with AWS data (NY 757 328, 556 m) collected at the site are used to assess the response of the catchment to precipitation and give some insight into the nature of runoff-generating processes in blanket peat.

4.2 Runoff regime and water table fluctuations in the Trout Beck catchment

4.2.1 Catchment-scale runoff regime

Figure 4.3 plots the runoff from the Trout Beck catchment from October 1994 to December 1999. The flashy nature of the stream response is immediately apparent. Baseflow appears to be of minimal importance. Figure 4.4 is an hourly flow frequency plot during this time period. The pattern is indicative of minimal groundwater flow from



▲ ECN TSS, automatic dipwell

● stream gauging site

H1 coded runoff monitoring plot (see Table 4.4)

x rainfall simulator plot (see Chapter 5)

grip water table transect (see Figure 4.30)

automatic weather station

--- access track

◇ old house site

— main channels on the Burnt Hill (Conway and Millar, 1960) catchment including grip network

— stream network

metres

0 N 5 00

Figure 4.1. Location of main experimental sites on the Moor House reserve. (see also Figures 3.1 and 3.4). Top left corner of map is NGR NY 750 344.

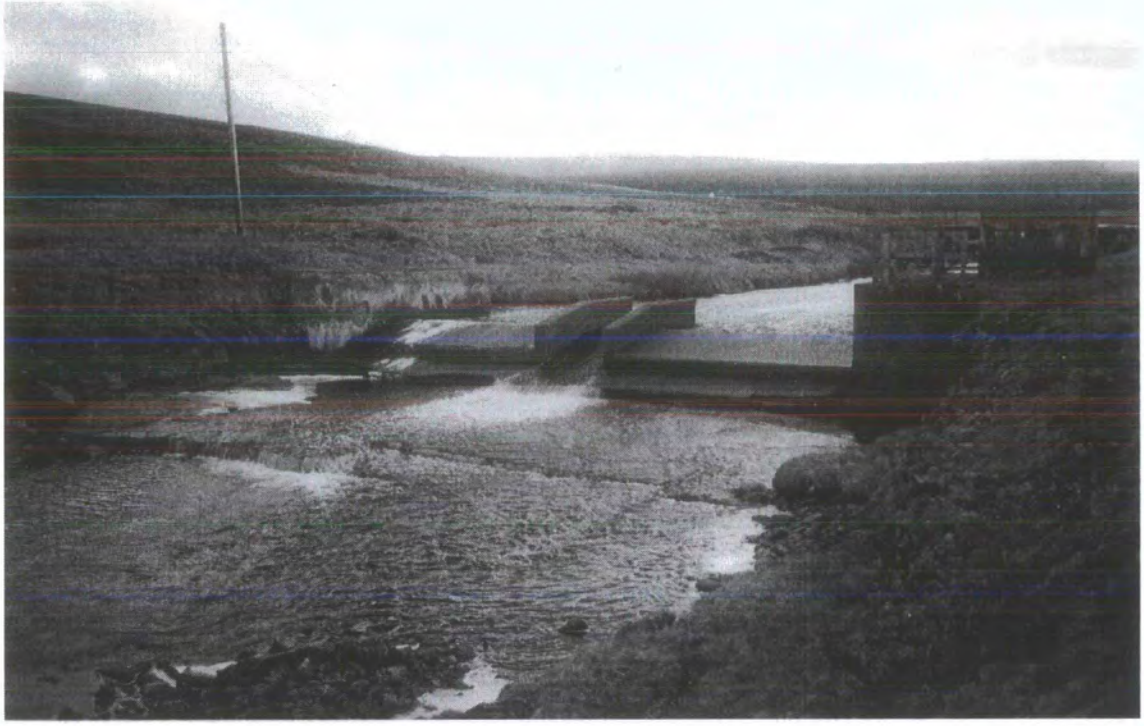


Figure 4.2. Compound Crump gauging station on Trout Beck maintained by the Environment Agency, NY 756326, 535 m O.D., register number 025003.

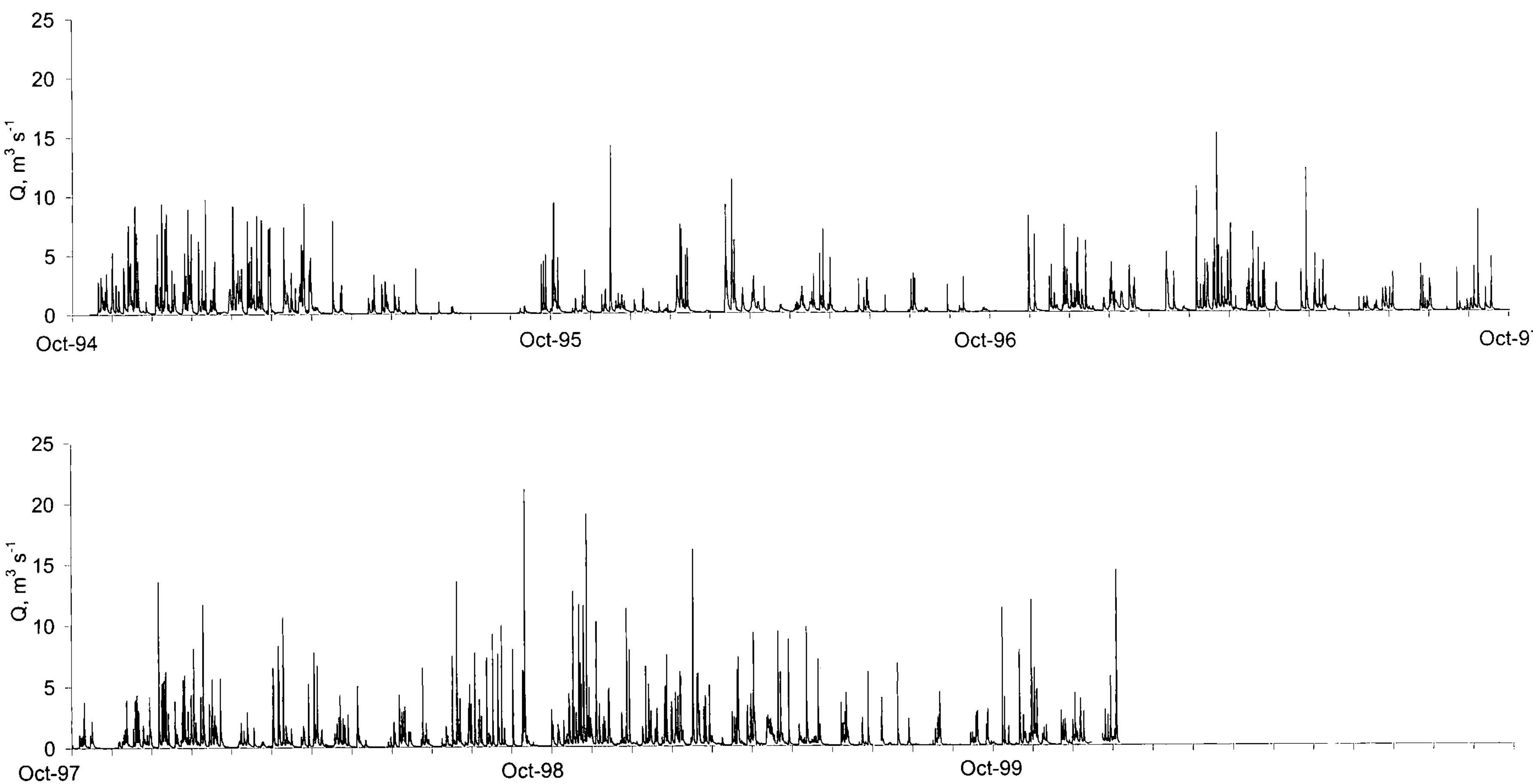


Figure 4.3. Discharge from Trout Beck, October 1994 - December 1999

the peat. Discharges below 0.5 $m^3 s^{-1}$ occurred 75 % of the time but only 21 % of the total discharge volume occurred during this period. Evans *et al.* (1999) reported mean storm rainfall:runoff ratios of 40 % for the Trout Beck catchment. These ratios are very high and are a result of efficient transfer of water to the channel by rapid flow mechanisms. The Trout Beck record shows that, unlike the traditional views of peat hydrology, blanket peat does not behave like a 'sponge'; rather, water is released rapidly following rainfall or snowmelt. During August 1995, discharge from the Trout Beck catchment fell to only 12 l s^{-1}. Monthly precipitation and runoff totals are shown in Figure 4.5. The rainfall:runoff ratio for the entire period of monitoring is 72 %. This is high, and considering the rapidity of the runoff suggests a limited storage capacity of the peat. There is a very close correspondence of rainfall and runoff. The largest disparities occur during February and March 1995, November 1996 and February 1997. This is most likely to be caused by drifting snow blowing into the rain gauge. Further discussion of the problems of precipitation measurement in the catchment can be found in Chapter 3.

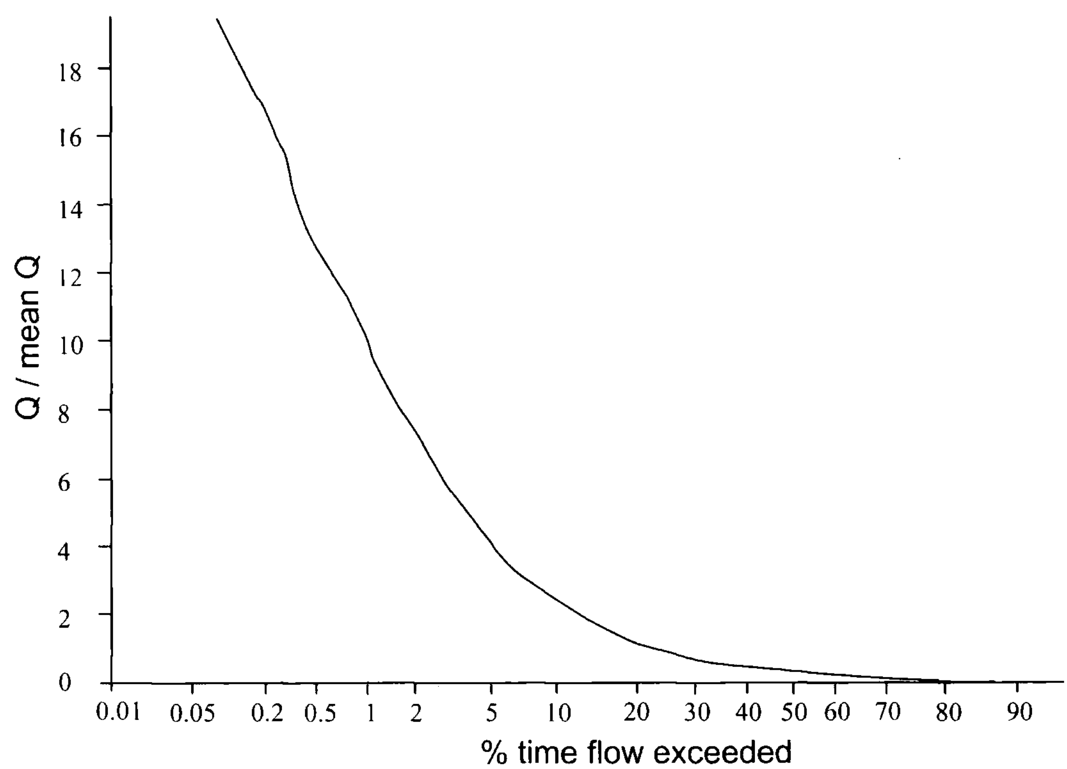


Figure 4.4. Flow duration curve for Trout Beck, October 1994 – December 1999.

Typical hydrographs from Trout Beck can be found later in the chapter (e.g. Figures 4.27, 4.28, 4.29 and 4.35) where they are compared to hydrographs produced by

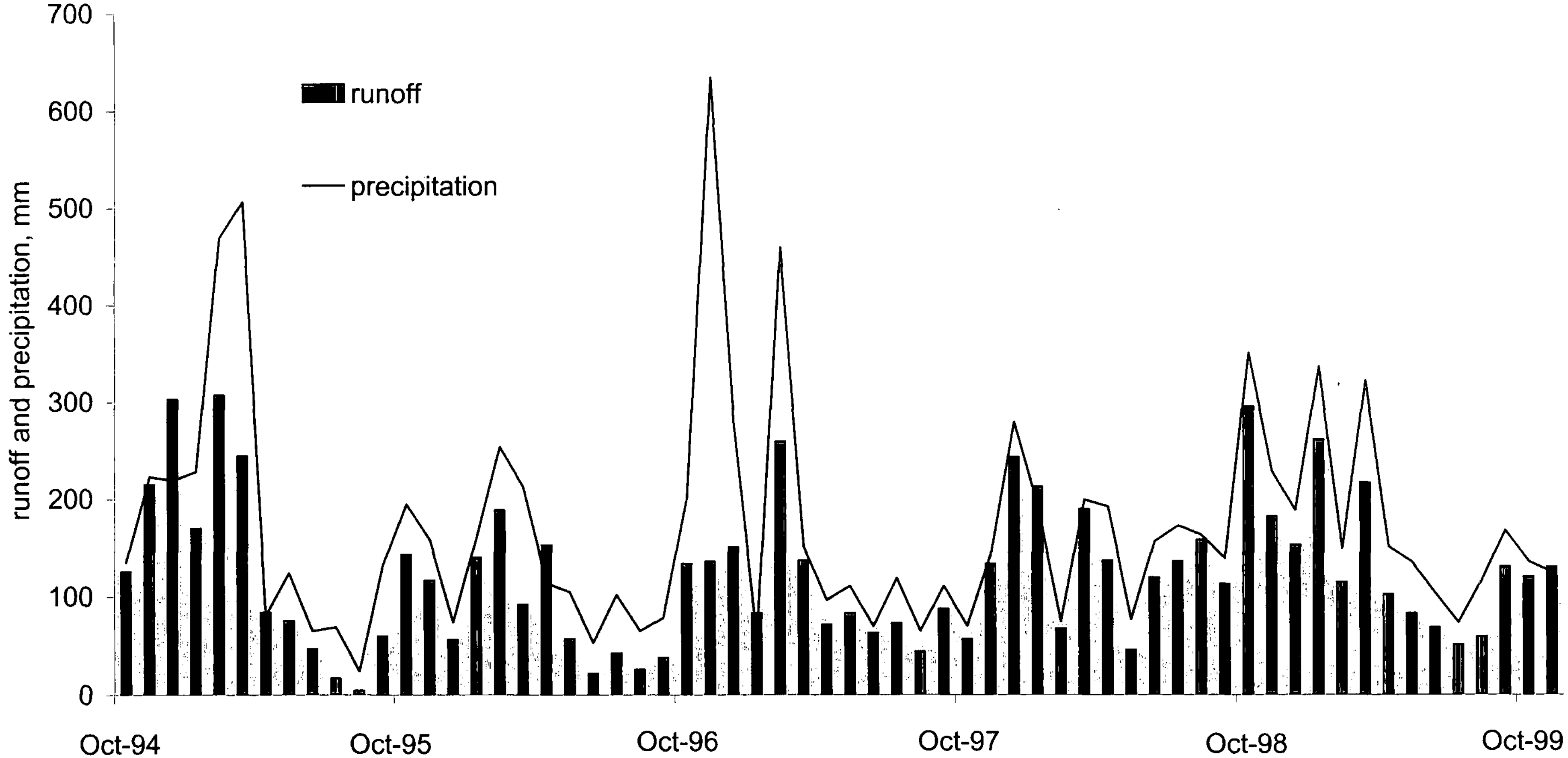


Figure 4.5. Monthly rainfall and runoff totals for Trout Beck

monitoring runoff processes. Hydrograph analysis was performed by Evans *et al.* (1999) on single-peaked storms unaffected by snowmelt spanning the full range of discharge values for Trout Beck for the 1995 to 1997 water years. Additional data from the 1998 and 1999 water years have been added and mean hydrograph characteristics are shown in Table 4.1. Mean peak lag times are only 2.7 hours indicating that the channel is well coupled to the hillslopes that are generating runoff. This also highlights the importance of stormflow in the catchment. Mean peak discharge of the storms analysed was 4.3 m^3 s^{-1} with the maximum recorded discharge during the study period being 21.2 m^3 s^{-1}.

Table 4.1 Mean hydrograph characteristics from Trout Beck

Storm Q	Peak Q	Tpeak	Peak Lag	Intensity
128 200	4.3	6.6	2.7	38.8

n = 72

Storm Q = total storm discharge, m^3
Peak Q = peak discharge m^3 s^{-1}
Tpeak = time from first recorded rainfall to hydrograph peak, hrs
Peak Lag = time from peak rainfall to peak discharge, hrs
Intensity = peak flow divided by total storm discharge x 10^{-6}, (units of s^{-1}). Hydrograph 'intensity' is an indicator of the hydrograph shape. A high intensity storm would indicate a more flashy response with lower intensity hydrographs being much broader in shape.

The Moor House AWS records precipitation on an hourly basis. In order to allow a higher temporal resolution for analysis of the 15-minute Trout Beck streamflow data (and with plot scale work discussed below) a Campbell ARG100 tipping-bucket rain gauge was installed in the catchment near the Trout Beck gauging station. Correlation analysis of the dataset (Table 4.2) reveals strong correlation of flow and rainfall variables so that peak discharge, maximum rainfall intensity (hourly), total rainfall and storm discharge are all significantly positively correlated. Maximum 15-minute rainfall intensity is significantly positively correlated with peak discharge. This is clearly consistent with efficient and rapid transfer of water to the channel. The correlations between Tpeak (the time from rainfall onset to peak flow) and the variables related to storm size are to be expected as a function of hydrograph geometry. Correlations within rainfall characteristics are also to be expected. The stronger association of total precipitation (correlation coefficient = 0.75) rather than maximum 15-minute (0.43) or hourly precipitation intensity (0.60) with peak discharge in such a flashy system may indicate that quick flow is generated from saturated areas and that the role of infiltration-excess OLF may be rather limited. If infiltration-excess OLF was the main runoff-generating mechanism it would be expected that the 15-minute and 60-minute rainfall intensities would be more strongly associated with peak flow than total rainfall. With saturation-excess, the more rainfall that occurs, the larger will be the amount of peat in the catchment that will become saturated and contribute to peak flow. Rather than occurring coincident with the period of heaviest storm rainfall as in the case of infiltration-excess OLF, rapid OLF can be generated by saturation-excess mechanisms even during low rainfall intensities, if the catchment has received a large amount of rain previously. Nevertheless, total rainfall, and maximum 15- and 60-minute rainfall intensity are all significantly positively correlated at $p < 0.01$. This is despite the possibility that local rainfall variations higher up the catchment could influence Trout Beck discharge.

Table 4.2 Inter-correlation of hydrograph variables using 15-minute logged data for Trout Beck, 1994-1999 water years

	Max 60	Max 15	Total ppt	Peak Q	Storm Q	Intensity	T peak	Peak Lag	ROP
Max 60	**1.00**	**0.81**	**0.51**	**0.60**	**0.35**	0.18	**0.44**	-0.33	-0.15
Max 15		**1.00**	0.28	**0.43**	0.17	0.21	-0.17	-0.21	-0.05
Total ppt			**1.00**	**0.75**	**0.88**	-0.07	**0.38**	-0.14	-0.16
Peak Q				**1.00**	**0.81**	0.23	0.05	-0.31	0.09
Storm Q					**1.00**	-0.18	0.33	-0.16	0.25
Intensity						**1.00**	-0.27	-0.24	-0.25
T peak							**1.00**	-0.02	-0.22
Peak Lag								**1.00**	-0.17
ROP									**1.00**

n=72

underlined indicates significant at $p < 0.05$, **bold** type indicates significant at $p < 0.01$

Max 60 = maximum rainfall recorded in one hour

Max 15 = maximum rainfall recorded in any logged 15 minute period (data only available from August 1998, n = 19)

Total ppt = total storm rainfall

Peak Q = peak discharge $m^3 s^{-1}$

Storm Q = total storm discharge, m^3

Intensity = peak flow, $m^3 s^{-1}$ divided by total storm discharge, $m^3 \times 10^{-6}$.

T peak = time from first recorded rainfall to hydrograph peak, hrs

Peak Lag = time from peak rainfall to peak discharge, hrs

ROP = storm rainfall:runoff ratio

4.2.2 Automated water table record

4.2.2.1 Introduction

The water table is the level at which the water pressure is equal to the atmospheric pressure and hence is the level at which water will stand in a well that is hydraulically connected with the groundwater body (Gilman, 1994). Detailed monitoring of the water table at Moor House NNR is carried out as part of ECN (Sykes and Lane, 1996). An array of five dipwells (5 cm diameter) is monitored manually once every week for an area of uneroded gently sloping peat within the Trout Beck catchment. These dipwells fluctuate consistently compared to one another throughout the year and the lowest correlation between any two of the dipwells is 0.96 ($p > 0.01$). The maximum difference between any two of the dipwells averaged over a year is 3.7 cm. A pressure transducer for automatic monitoring is within a dipwell where water table is typically between the minimum and maximum values measured by the array. Water table is monitored every 5 seconds and then averaged for the whole hour.

4.2.2.2 Errors in the automated record

Comparison of the logged dipwell readings with weekly manual readings for the same dipwell show some variability. Figure 4.6 shows what appears to be a seasonal trend in the offset between the two readings. During warmer periods, as indicated by the soil temperature probe, the logger records deeper water tables than those measured manually of the order of 2 to 3 cm. The seasonality in offset suggests either that the pressure transducer is very sensitive to water temperature or that there is some shifting of the transducer. Reported temperature related errors for this sensor are 0.3 % over 360 cm. The error seems to have been reduced during the 1998 and 1999 water years (Table 4.3). During these latter two years of the record there appears to be a shift towards manual readings recording deeper water tables than the sensor. It therefore seems more likely that the transducer is shifting in position perhaps with expansion and contraction of peat with drying and rewetting (Ingram 1983; Ingram, 1991; Price and Schlotzhauer, 1999). The sensor is fixed to the base of the dipwell; however, the dipwell itself may be shifting. Gilman (1994) noted that dipwells for the observation of water table elevation rarely extend to a firm substrate and Hutchinson (1980) discovered a seasonal rise and fall of the ground surface of 50 mm or more in Holme Fen Post, East Anglia. Movements of the ground surface at Crymlyn Bog between 1985 and 1989 varied between 5 % and 12 % of the water table movement (Gilman, 1994) and it is suggested that regular measurements between the top of a firmly fixed datum post and the rim of

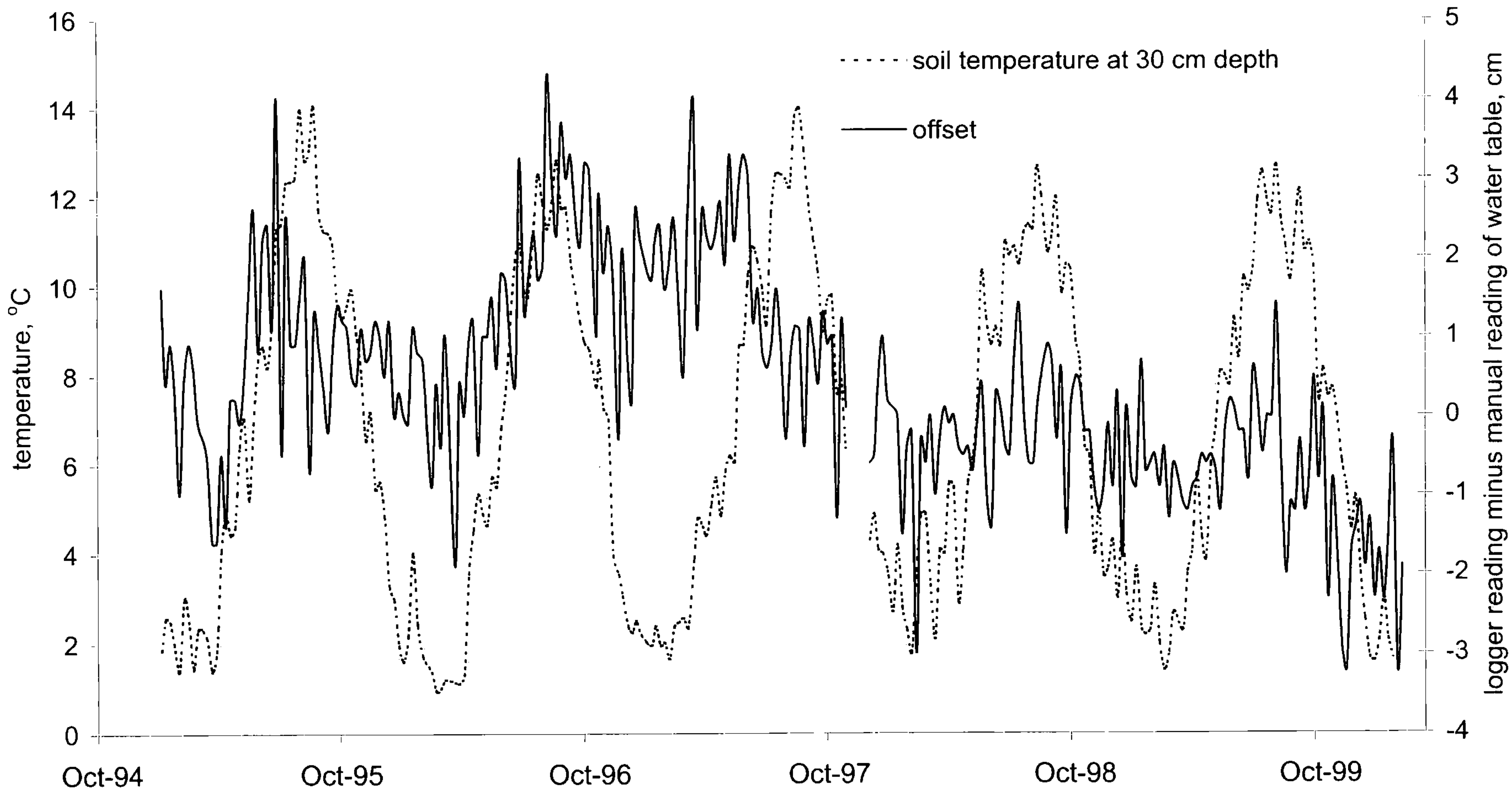


Figure 4.6. Offset of pressure transducer reading of water table depth as checked by weekly manual measurement at the ECN target site.

the dipwell should be made. The pressure transducer may need to be corrected in some way for shifts in the position of the ground surface in relation to its own position; the use of a floating bog shoe device (Roulet *et al.*, 1991) may help with this. The very fact that the ECN are trying to produce standardised long-term records of the environment is important. Given the changing difference between automatic and manual readings, ECN data patterns should be put into the context of possible sensor movement which could result in mistaken identification of environmental change. These results also justify the present practice of regular manual checks of automated ECN instruments. Nevertheless, the variation between logged and manual measurements is not of sufficient magnitude to significantly affect the main trends in water table height and the subsequent analysis is therefore based on the logged dipwell readings. For the purposes of this study the error is minimal, and the water table data have the potential to provide direct insight into the role of antecedent moisture on runoff generation.

Table 4.3 Annual mean and standard deviation (σ) of absolute difference between the logger and the manual measurements of water table, 1995 –1999, based on weekly sampling.

Year	Mean absolute offset, cm	σ in offset, cm
1995	1.0	0.8
1996	1.6	1.1
1997	1.5	0.9
1998	0.6	0.6
1999	0.9	0.7

4.2.2.3 Water table fluctuations

Water table levels from October 1994 to January 2000 are shown in Figure 4.7. It is clear that the water table remains very close to the surface for most of the study period. Fluctuations are swift with recoveries occurring more rapidly than recession. Figure 4.8 plots water table residence times for the entire period. The water table is within 5 cm of the ground surface 82.4 % of the time. Hence there is likely to be a high incidence of saturation-excess overland flow. When the soil profile is completely saturated, excess water added by precipitation from above and/or by subsurface flow from upslope, ponds on the soil surface and may eventually run off downslope. Only in the summer months (May to September) does water table elevation drop further than 5 cm for short periods before a recharge event returns the table to the surface (Figure 4.7). The summers of

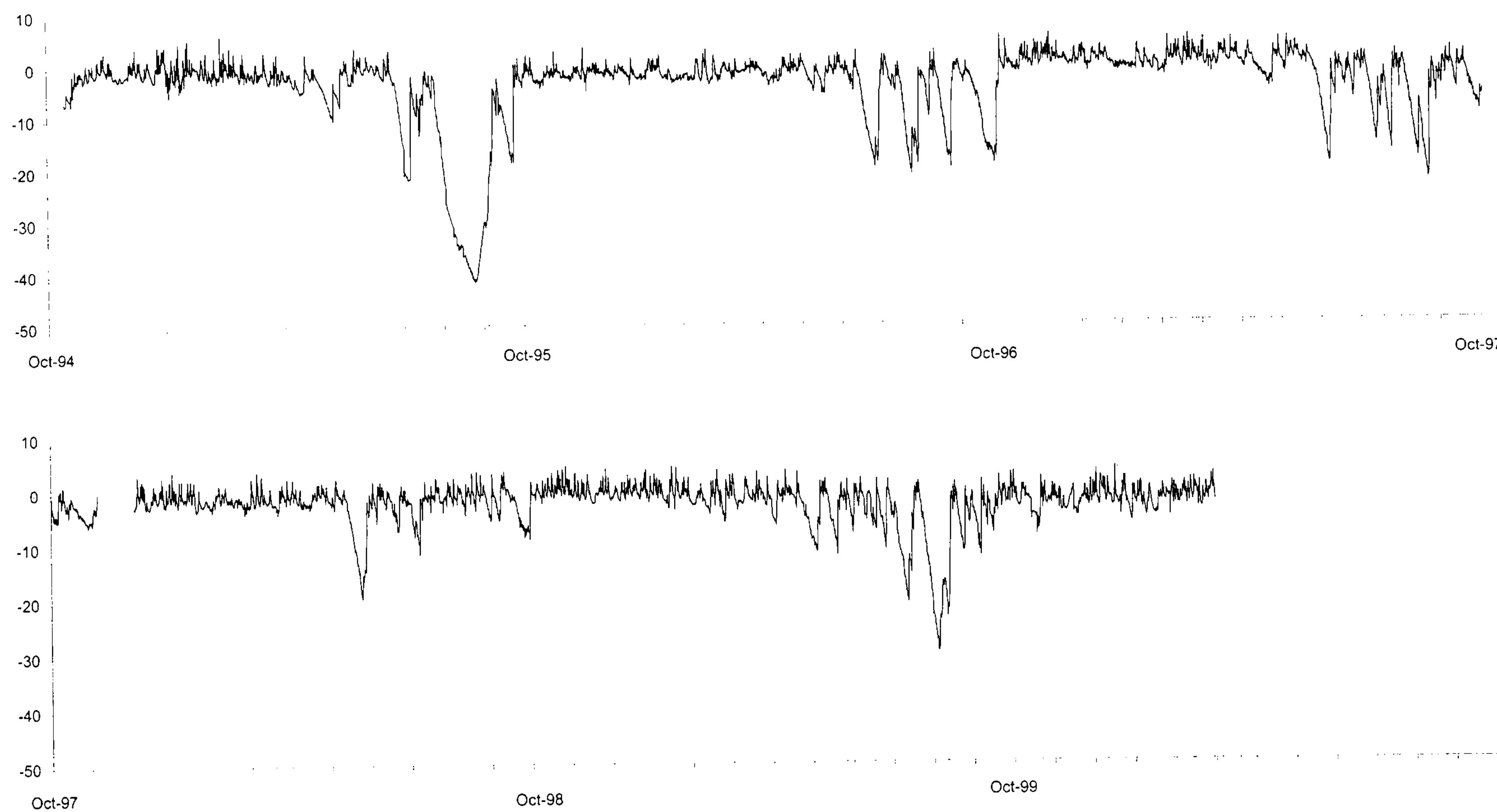


Figure 4.7. Logged hourly water table depth, cm, ECN target dipwell

1996 and 1997 are similar in that there are three or four periods when the water table briefly drops to around 20 cm. During the summer of 1998 this only happens once. There is a warm dry period during July and August 1999 where the water table drops to almost 30 cm. Comparison with the antecedent precipitation index (API) (Figure 4.9) shows that even when there are periods of rainfall deficiency during winter months, such as in the 1996 water year, this has very little effect on water table depth. This suggests an evaporative control on the behaviour of water levels. The only extended period of low water table is during the drought of 1995; even then the minimum levels do not drop below 42 cm. Evans *et al.* (1999) examine the water balance for the Trout Beck catchment for the 1995 to 1997 water years. Significant soil moisture deficits only develop in high summer and changes in storage (calculated using the Penman (1948) formula to assess potential evapotranspiration) are of the order of +/- 100 mm.

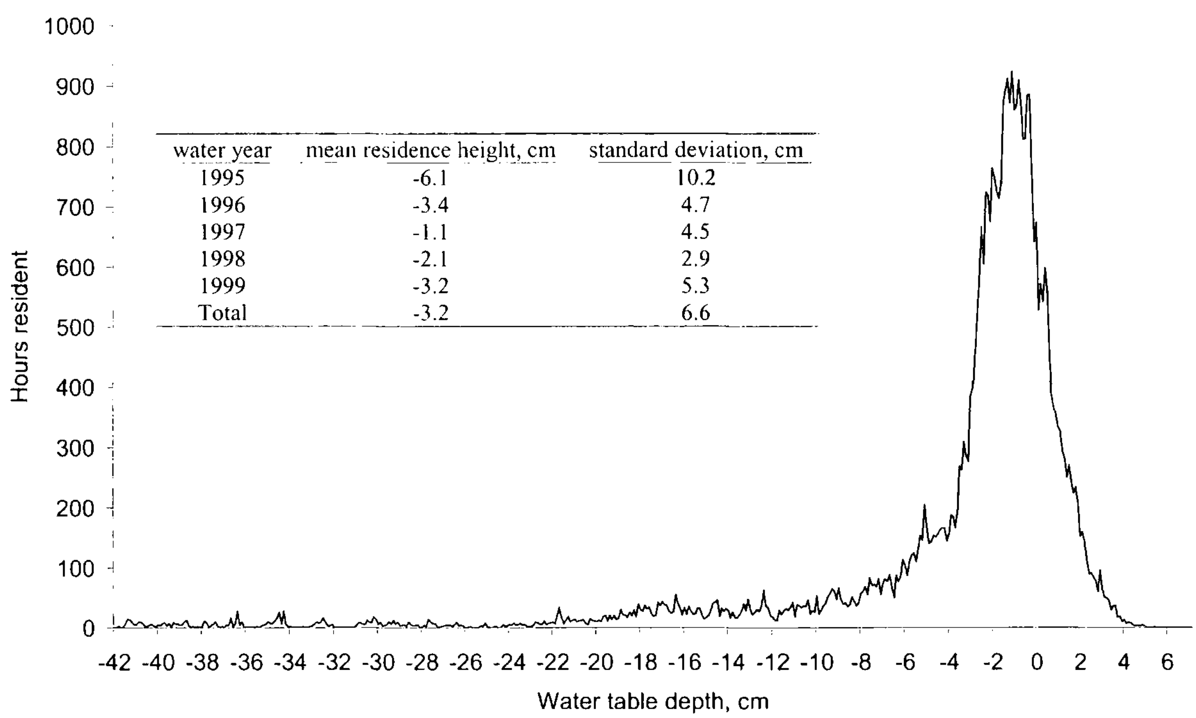


Figure 4.8. Water table residence, ECN target dipwell, 1995 – 1999 water years, bin width = 0.1 cm.

Once the water table falls 5 cm below ground level, the water table recession is controlled by evapotranspiration. Diurnal cycles are evident; there is often no fall in water table during night hours clearly indicating that slow drainage of the peat mass does not occur (e.g. Figure 4.10). Water table recession sequences below 5 cm could

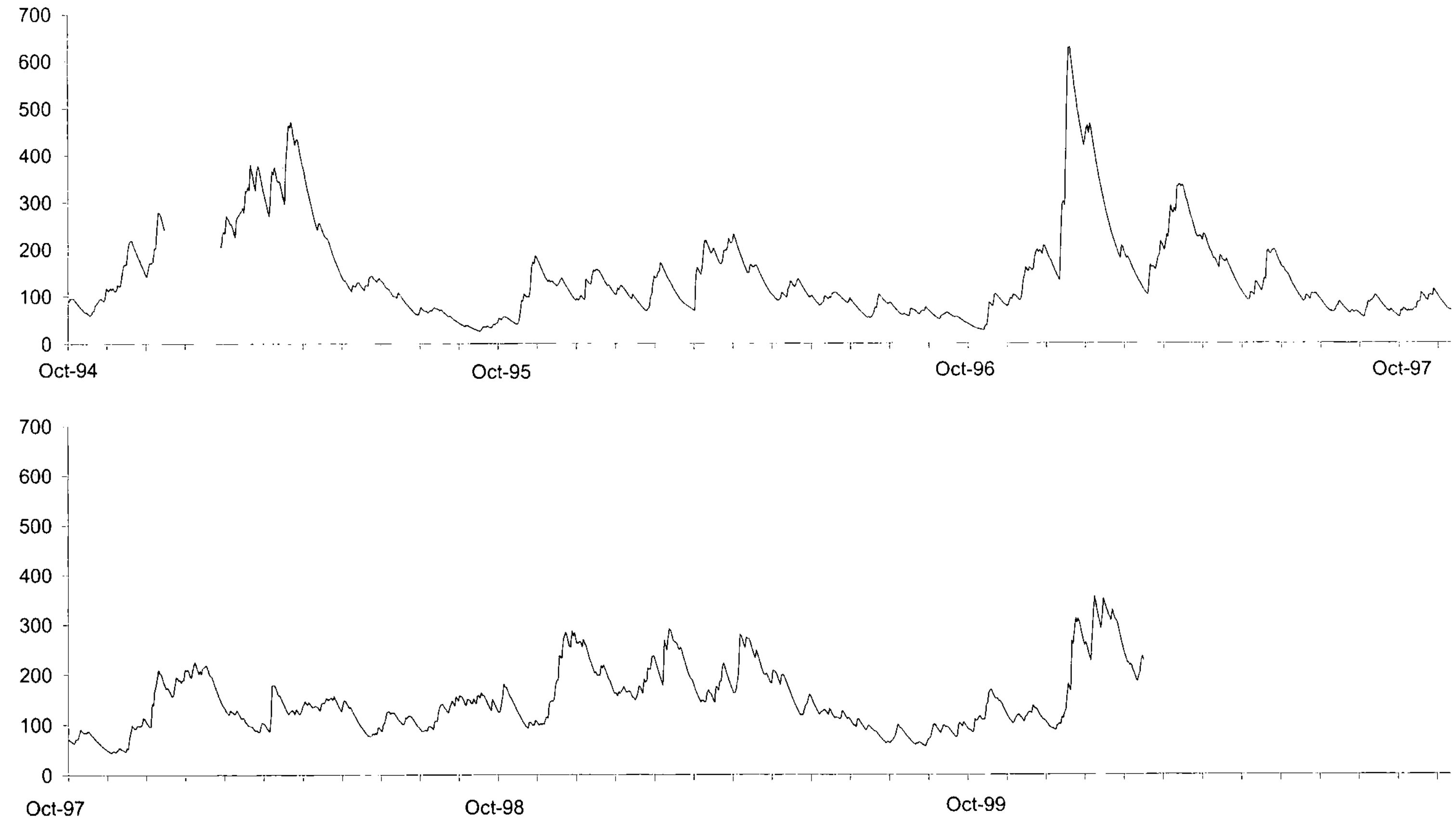


Figure 4.9. Antecedent precipitation index October 1994 - January 2000.

only be identified from May to September each year. Rates of recession varied from 0.55 cm per day in May 1996 to 2.26 cm per day in June 1996, although mean daily fall during the dry periods of the summer months was 1.5 cm. On average 54 % of the daily decline in the water table took place between 12 noon and 6 p.m. (when mean fall rates were 1.3 mm hr^{-1}) and 86 % of the daily falls took place between 9 a.m. and 9 p.m. Heikurainen (1963) and Tomlinson (1980) found comparable diurnal 'steps' which Tomlinson claimed were independent of depth once below the critical level of 5-6 cm below the surface of Brishie Bog. The evidence therefore seems to suggest that when the water table falls to around 5 cm into the peat profile, drainage thereafter is very slow and most runoff production in blanket peat is likely to take place above this level.

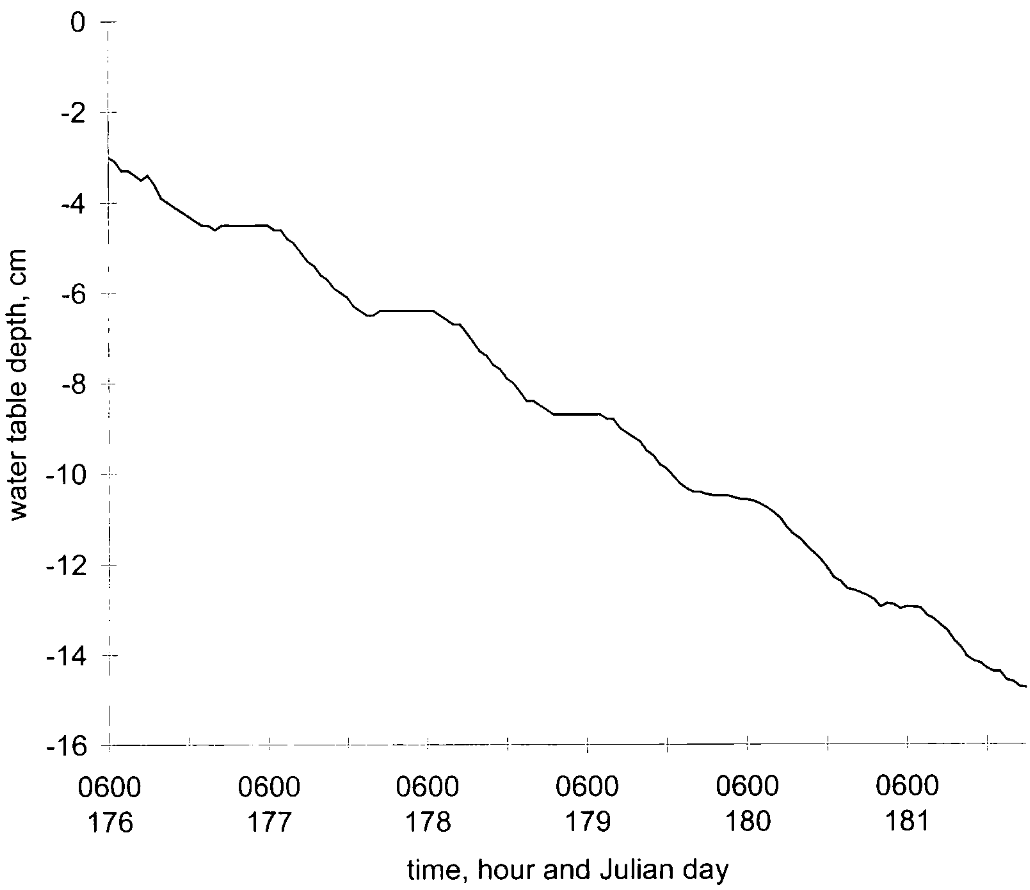


Figure 4.10. Diurnal decline in water table depth, 25th – 30th June 1995.

Recovery rates of the water table during summer recharge events are generally extremely rapid with an average of 5.3 mm h^{-1}; three events produced average recoveries in excess of 20 mm h^{-1}. A multiple regression analysis of rainfall-recharge events shows maximum rainfall intensity as the dominant control upon the rate of rise and the total rise of the water table, with R^2 values of 0.35 and 0.38 respectively.

Total rainfall only contributes 3 % and 2 % to these regressions. Indeed 44 % of peak water table levels are reached whilst it is still raining and, in line with the findings of Godwin (1931) and Heikurainen (1963), the rise in water table usually takes place over a period that is comparable with the duration of the rainfall. It appears that bulk infiltration rates into the upper 20 and 30 cm of the peat could be high allowing rapid recovery of water table during storm events. Given that the peat appears not to drain below about 5 – 10 cm at the ECN dipwell site, presumably related to very low hydraulic conductivities this may seem an unusual observation. The transmission of water to the water table is probably more rapid than could be accounted for by transmission through the soil matrix. It may be that infiltration takes place through macropores (Beven and Germann, 1980) or that water already in the unsaturated zone is displaced (Horton and Hawkins, 1965). It may be that aeration of the peat to depths at which the peat is normally saturated, and surface drying and subsequent shrinkage of the peat (Gilman and Newson, 1980) leads to structural change that temporarily enhances the infiltration, percolation and movement of water below the surface. Further evidence for this comes from experimental work on peat blocks (see Chapter 5).

Regression analysis of the effects of rainfall on water table rise ($n = 70$, $r^2 = 0.54$) predicts that when the water table is at 200 mm depth, 1 mm of rainfall will induce a water table rise of 17 mm. When the water table is within 50 mm of the surface 1 mm rainfall induces a only a 4 mm rise (Figure 4.11). Any change in water table elevation in upper horizons of less decomposed peat therefore represents considerably more water than a corresponding change in deeper, more dense peats (Boelter, 1968). When the water table is close to the surface, a rainfall may raise the level to the surface and any further rain will then be lost by overland flow or by near-surface lateral flow. This will therefore induce error in the regression prediction. Nevertheless these results fall within the range quoted by other authors; Tallis (1973) estimates 1 mm of rain produces a 6 mm rise in the water table at Featherbed Moss, Derbyshire and Chapman (1965) found a rise of 3 mm for 1 mm of rain at Coom Rigg Moss, Northumberland. These relationships give a measure of the air-space volume (in effect, the specific yield fraction of porosity) in the peats which can be filled with water, and are thus also a measure of the degree of compaction of the peats. At Moor House the peat at 5 cm has approximately four times as much air-space volume as peats at 20 cm.

4.2.3 Relation of water table depth to runoff generation

Figure 4.12 plots storm runoff ratio against antecedent water table. Runoff production tends to be less efficient when water tables are lower, presumably linked to water table recharge early in the storm. The greatest variability of runoff production occurs when the water table is within 5 cm of the surface. This suggests that the importance of storage as a control on rapid-flow generation is minimised and that other sources of variability such as the rainfall characteristics of the storm come into play. As the water table is within 5 cm of the surface 83 % of the time, the rainfall characteristics are therefore more closely linked to runoff via quickflow response. At the individual storm level Figure 4.13a illustrates the effect of water table recharge after a summer dry spell on runoff production. Here rainfall in excess of 5 mm hr^{-1} produces minimal hydrograph response while the water table is low. Later in the storm, just under 4 mm in an hour is sufficient to trigger a rapid hydrograph rise. The difference between the two responses is controlled by the level of the water table which was at 24 cm depth before the storm but had risen to within 5 cm of the surface when the hydrograph response was triggered. Response to initial rainfall is much more rapid when the water table is close to the surface as indicated by Figure 4.13b and the result is a greater peak flow.

Hence at the catchment level the data suggest that the generation of rapid flow is not related to infiltration rates but to saturation of the peat mass. Hydrograph response is rapid when the water table is within 5 cm of the surface. This may indicate rapid subsurface flow. However, if the water table is near the surface on the open slope of the ECN water table site it will almost certainly be at the surface in water-collecting sites; saturation-excess overland flow will then occur from specific contributing areas in hollows and adjacent to the channel. Further evidence for this comes from plot and hillslope mapping work detailed below. The relation between runoff and water table for hourly data over the study period is shown in Figure 4.14. Significant storm discharges on Trout Beck are confined to periods when the water table is within 5 cm of the surface. This is further good evidence for the importance of saturation in the production of overland and near-surface flow in the catchment. In the entire period there are no events where high discharges are associated with low water tables (which might otherwise indicate the occurrence of infiltration-excess overland flow in the catchment). Chapman (1965) observed a clear relationship between runoff and water table level at Coom Rigg Moss in Northumberland; once the water table elevation was below 8 cm,

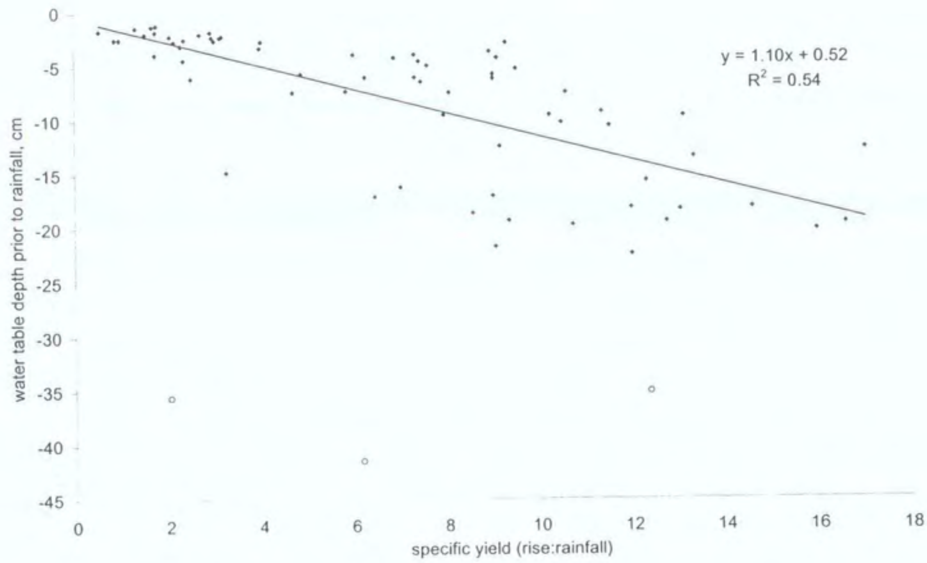


Figure 4.11. Relationship between water table depth and specific yield. Open circles = August 1995.

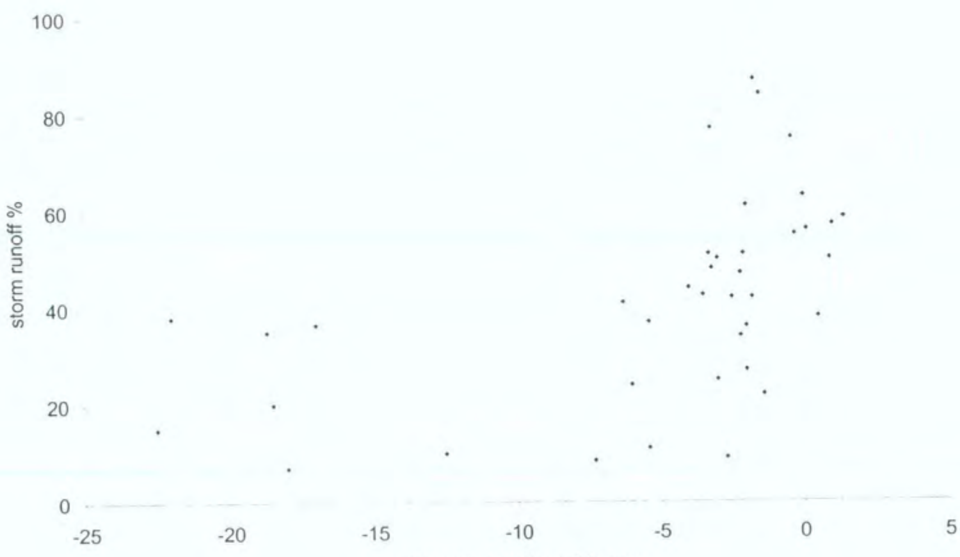
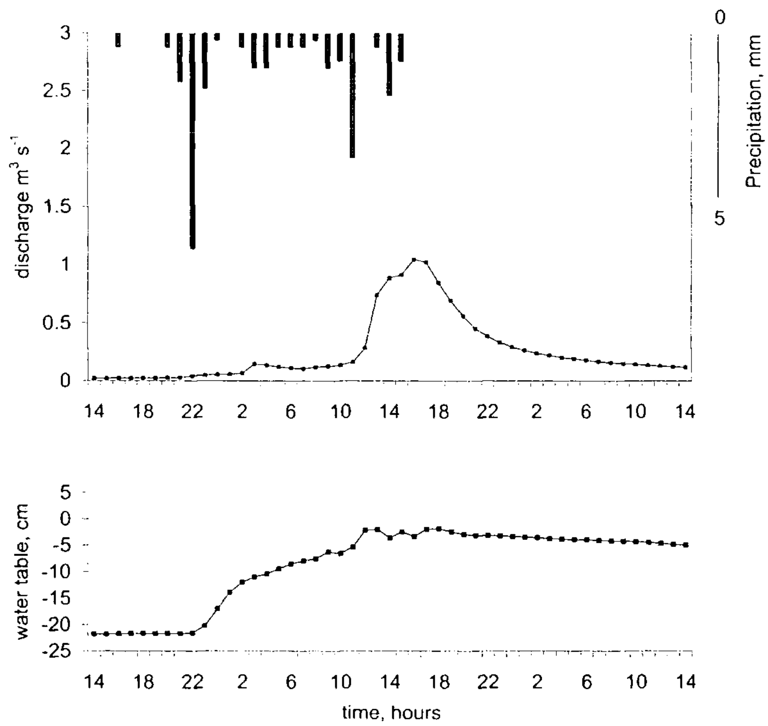


Figure 4.12. Relation of storm runoff, % to antecedent water table

a)



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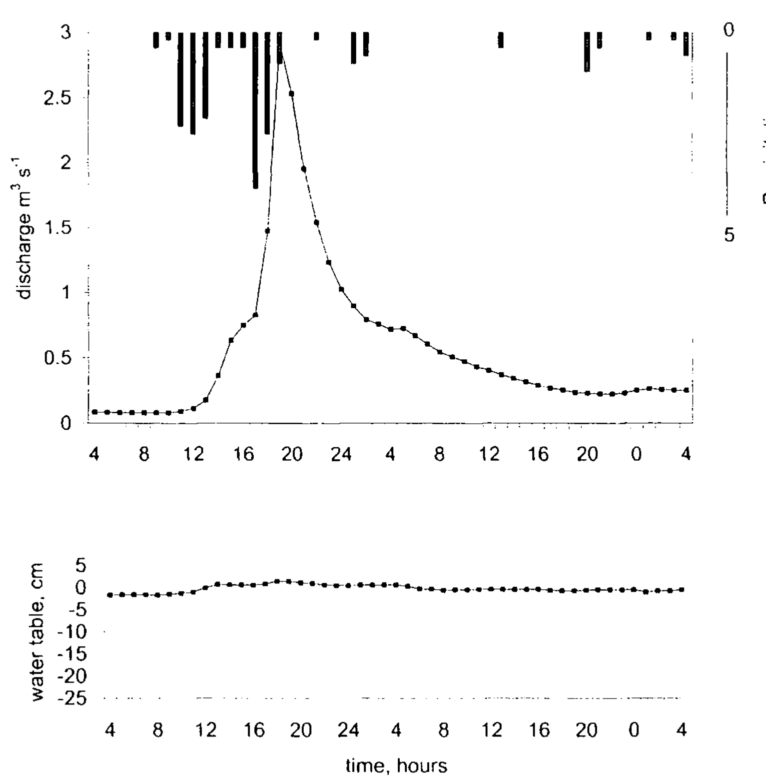


Figure 4.13. Hydrograph and water table data from two storms illustrating the importance of near surface water tables in generating runoff, a) 6/7/95, b) 22/5/96

runoff became negligible implying that the bulk of water movement in the peat was a fairly rapid flow at and near the surface.

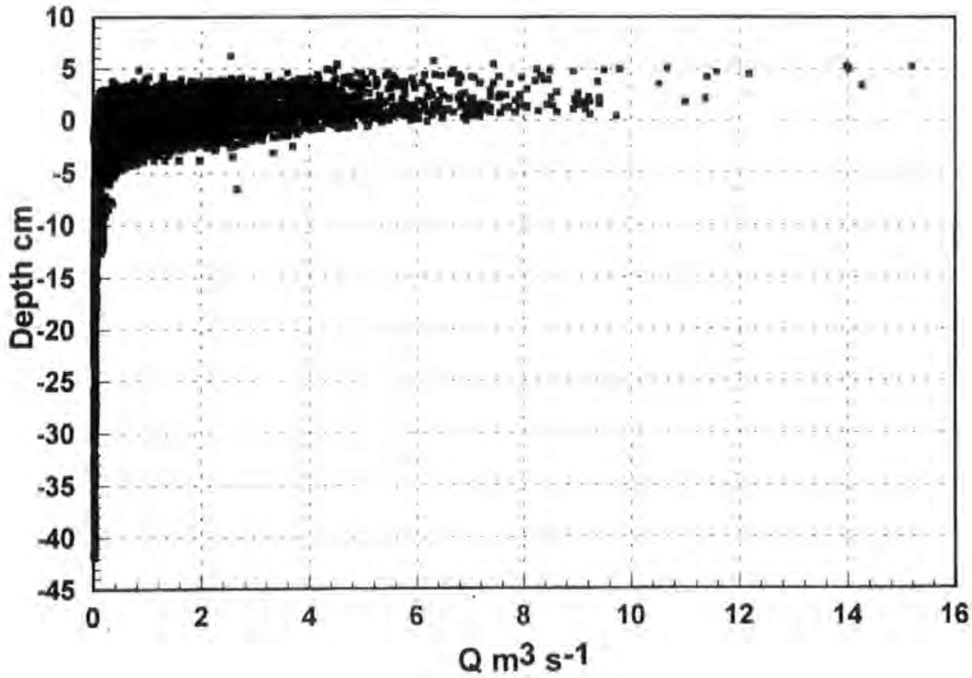


Figure 4.14. Relation of Trout Beck discharge to water table depth, hourly data, October 1994 – 1997.

4.2.4 The 1995 Drought

4.2.4.1 Drought conditions

The meteorological summer of 1995 (June to August inclusive) was the second driest recorded since daily rainfall records began at Moor House in 1952 with only 1976 being drier. The 1995 water year was atypical given that the December-February rainfall was the 4th highest since 1953 (Burt *et al.*, 1998). Nationally July and August 1995 were the warmest in the 335 year Central England Temperature series and the driest in the 230 year England and Wales precipitation series (Hulme, 1998). Evaporative demand for April to August in England and Wales exceeded normal levels by 20 % (Marsh and Turton, 1996). Hulme (1998) suggests that summers as warm as 1995 will in future become 1 in 10 year events rather than 1 in 300 as at present. The rainfall deficit of summer 1995 would remain exceptional but increased evaporative demand would mean that soil moisture deficits of the magnitude recorded in 1995 would be increasingly common. Marsh and Sanderson (1997), Burt *et al.* (1998) and Conway (1998) all note the recent trend in Britain towards enhanced seasonality with warmer wetter winters and hotter, drier summers. Arnell (1996) suggests that with predicted warming in northern Britain accompanied by increased rainfall, particularly in the winter, annual runoff may increase by up to 25 %. However, Burt *et al.* (1998) note that with warmer summer temperatures and only a small increase in summer rainfall, low flows in a peat covered catchment like Trout Beck may well decrease. Therefore knowledge of the impacts of droughts like that in 1995 on blanket peat catchments is important if they are to become more frequent, not least because the headwaters of many UK rivers lie in such areas, and blanket peat moorlands are important source areas for water supply.

4.2.4.2 The effect of drought on streamflow

In such a highly-rainfall dependent system as Trout Beck the effects of prolonged drought on streamflow is dramatic. During August 1995, discharge from the 11.4 km^2 Trout Beck catchment fell to only 12 $l\ s^{-1}$ and for the whole month flows were of the order of 0.01 $m^3\ s^{-1}$. This is less than 5 % of the mean daily flow for the entire study period. Despite a near-record wet winter preceding the drought (Burt *et al.* 1998), and the fact that the water table never fell below 42 cm from the surface, baseflow was virtually non-existent in the Trout Beck system. Thus peat does not act as an aquifer and fails to maintain river flows and reservoir levels during periods of dry weather. The implication is that if climate change does lead to increased seasonality, the impact on the ecology of upland stream systems may be severe, even though mean precipitation levels are greater. Management of water resources will also need careful consideration of the failure of blanket peat to provide sufficient baseflows. Excess winter rainfall will be lost as runoff and will not contribute to maintenance of baseflow in the summer months.

4.2.4.3 The effect of drought on water tables

July and August 1995 are the only two months within the study period where the water table never reached the blanket peat surface at the ECN monitoring site and the water table was below 10 cm for 42 consecutive days and below 20 cm for 35 consecutive days (Figure 4.15). Thus the water table fell to depths which are normally saturated. As noted in Chapter 2, the drying of peat may lead to changes in its hydrological behaviour. In August 1995 there is a disparity in the rates of evapotranspirative decline of water table. Given that the average daily water table depression in June, July or August is 1.85 cm d^{-1} not usually dropping below 1.3 cm d^{-1} it is unusual that the rate of decline for a long spell in August 1995 should be 0.7 cm d^{-1}. Figure 4.15 shows water table

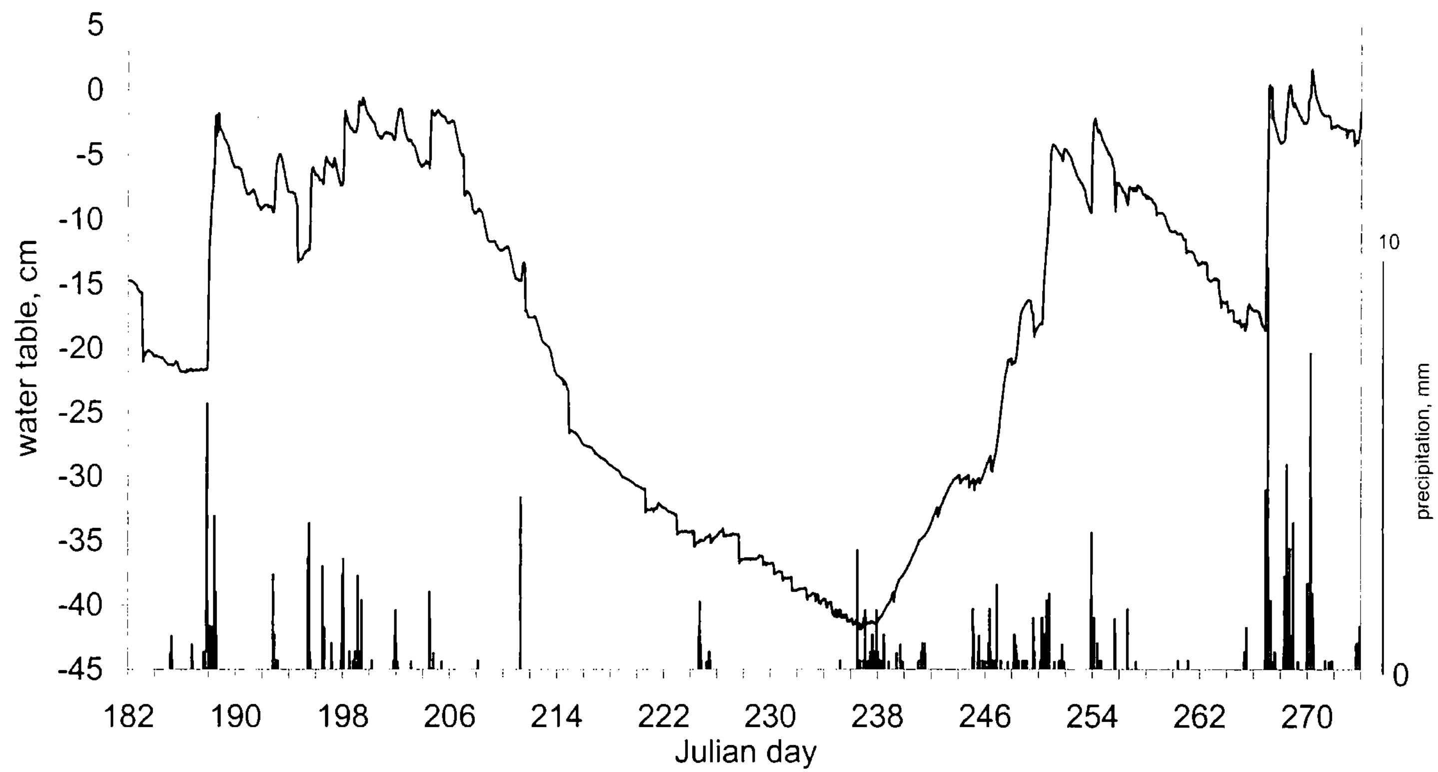


Figure 4.15. Water table and rainfall, July-September 1995.

fluctuations in response to rainfall for July, August and September 1995. There is a more gentle slope to the decline in water table from day 213 onwards. Temperatures were comparable to those in June and July 1995 although daylight hours are slightly shorter. It may be that the water table having dropped to 40 cm below the surface could no longer be tapped by shorter plant roots resulting in a decrease in transpiration rates. Boggie *et al.* (1958) demonstrated that *Calluna* efficiently recovers nutrients from depths of less than 15 cm. *Eriophorum* roots in contrast may withdraw water from depths of up to 50 cm. Diurnal fluctuations in the past have been attributed to transpiration rather than evaporation by comparing water tables under vegetated and cleared areas (White, 1932) and these fluctuations have also been used to provide an estimate of transpiration rates (e.g. Heikurainen, 1963; Gilman, 1994). However, Gilman (1994) found that as the water table declined to the lowest levels in Wicken Fen, the fluctuations just became ripples on a continuous decline which he interprets as a dispersion of the diurnal wave of water demand, the processes of re-distribution of soil water requiring more time to complete as the unsaturated zone became deeper. Essentially increasing depth of the water table limits evapotranspiration (Tomlinson, 1980).

The slowest water table recovery rates during the study period also occur in August 1995 where mean values as low as 0.3 mm hr^{-1} can be observed. This can be seen by the August anomalies in Figure 4.11. The slow recovery rates are also indicated by the much more gently sloping water table rise during days 238-250 compared with the usual very rapid response to rainfall seen for example on day 189, 198, 254 or 268 (Figure 4.15). Some of the slow recharge can be accounted for by low rainfall intensity. Nevertheless, the lag time from rainfall cessation to water table peak is at least ten times greater than seen for any other water table recharge event throughout the three years of study. It may be simply that recharge percolation rates are limited at depths greater than 20 cm. It may also be that temporary structural changes have taken place within the peat itself due to loss of moisture, shrinkage and the aeration of what are usually anaerobic peat layers. Thus with a more volatile climate, the assumption that blanket peat will remain largely saturated with little temporal variability in water table level may not hold in future therefore. There are implications for future peatland hydrology, ecology and erosion given that water tables and their fluctuations are important for vegetational distribution (Ingram, 1983; Hammond *et al.*, 1990), and that continued intrusions of the water table into the usually anaerobic catotelm as seen in the summer of 1995 may change its physical and hydrological properties. Clear evidence for the effect of structural change on peat hydrology is seen in Chapter 5 with drought simulation on peat blocks. The extent to which contemporary studies of peat hydrology at Moor House reflect past or future conditions remains a matter for conjecture. Given the potential for climate change, however, the importance of ongoing ECN monitoring is underlined.

4.2.5 Summary of catchment-scale runoff and ECN water table characteristics

Runoff percentages for Trout Beck are high as rainfall is efficiently and rapidly transferred to the channel producing flashy hydrographs. There is no significant delayed flow and groundwater discharge contributes very little to baseflow. The water table is within 5 cm of the surface for 83 % of the time. During dry spells, in summer, water table falls under evaporative control. Water table recharge is rapid, indicative of relatively high bulk infiltration rates into the catotelm when the upper layers are unsaturated. The rapid generation of surface or near-surface runoff occurs when water tables are close to the surface. This strongly indicates that the dominant runoff pathways are saturation-excess overland flow and/or subsurface storm flow generated by percolation-excess above a saturated catotelm. However, the spatial and temporal details of these mechanisms cannot be determined from catchment-scale datasets. Therefore subcatchment-scale work will now be described to allow some quantification of the relative importance of overland and shallow subsurface flows.

4.3 Subcatchment-scale monitoring

4.3.1 Experimental design

4.3.1.1 The Burnt Hill subcatchments

Of the early studies in blanket peat that of Conway and Millar (1960) is the most notable. They reported results from four small moorland catchments; two had natural drainage channels, and two had artificial networks of moorland grips. They concluded that artificial drainage of peat moorlands gave an increased sensitivity of runoff response to storm rainfall with peak flows both higher and earlier. Some of their study was based in the Trout Beck catchment. They demonstrated that runoff production in peat is extremely rapid especially where hillslopes had a dense gully network, had been burned or were gripped. In contrast, relatively uneroded subcatchments exhibited a smoother storm hydrograph with greater lag times and the water balance calculations suggested that uneroded hillslopes could retain significantly more water than drained, eroded or burnt basins. One of the four small catchments that Conway and Millar (1960) based many of their conclusions on was a hillslope which was partly gripped and partly eroded. The hillslope had also suffered severe burning in 1950 and was subsequently named 'Burnt Hill'. This area was reinstrumented as part of this study. Although Conway and Millar amalgamated runoff from the gripped and eroded sections of Burnt Hill, these were monitored separately here using two 90° V-notch weirs (Figure 4.16). Inaccuracies stem from ice building up in the stilling wells and, occasionally, peat blocking the V-notch. Robinson (1985) notes that one of the problems with Conway and Millar's dataset is that 90° V-notch weirs are insensitive at low flow such that great caution needs to be applied to their interpretation of low flow records. In order to measure low flows more accurately the reinstalled weirs had compound V-notches, with a quarter 90^{0} notch below a 90^{0} notch (Gregory and Walling, 1973; see Figure 4.16) and individual calibrations were made. Figure 4.17 shows the nature of the gripping and gully network on Burnt Hill. Bower (1959) uses Burnt Hill as a classic example of her Type 1 and Type 2 dissection systems with advanced Type 1 stage on the gently sloping hill crest with the pattern of gullying close and complex with deep incision. Drainage from this gullied section feeds into the Type 2 system on the steeper part of the slope with fairly linear gullying. Hydrologically there are two areas of interest: how runoff production is achieved in the artificially drained and naturally dissected sections of Burnt Hill; and to examine the effect of the considerable change in catchment conditions that must have taken place since the late 1950s (one decade after burning). Conway and Millar (1960) suggested that much of the flashy response was related to a peat surface of a 'cheesy, impervious consistency' caused by burning. This is now not immediately obvious with vegetation now recovered substantially since the last fire.

4.3.1.2 Plot and hillslope monitoring strategies

a) Automated OLF and throughflow measurement

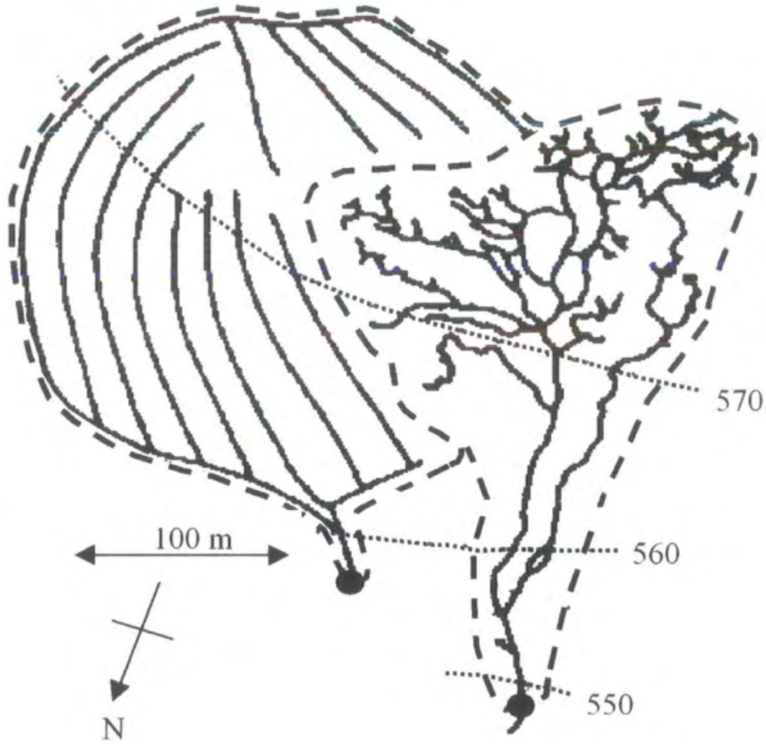
Runoff was collected at a smaller spatial scale from two plots, one on each section of Burnt Hill: upslope of a grip (G2) and at a gully head (E2). Runoff was collected at these two locations and at six other sites on intact slopes (see Figure 4.1 for site locations within the Trout Beck catchment, and Table 4.4 for site characteristics) by use of aluminium throughflow troughs and channelled into tipping-bucket flow recorders (Figure 4.18a and b; Whipkey, 1965; Knapp, 1973; Atkinson, 1978; Reynolds and Stevens, 1987; Khan and Org, 1997). Anderson and Burt (1978a) noted that excavations associated with Knapp's (1973) method of trough insertion may produce errors due to

a)

b)

Figure 4.16. V-notch weirs installed at Burnt Hill. a) On the eroded subcatchment (the site of the original Conway and Millar (1960) station), b) on the gripped subcatchment. Ott float recorders were used to record stage simultaneously at both sites.

a)



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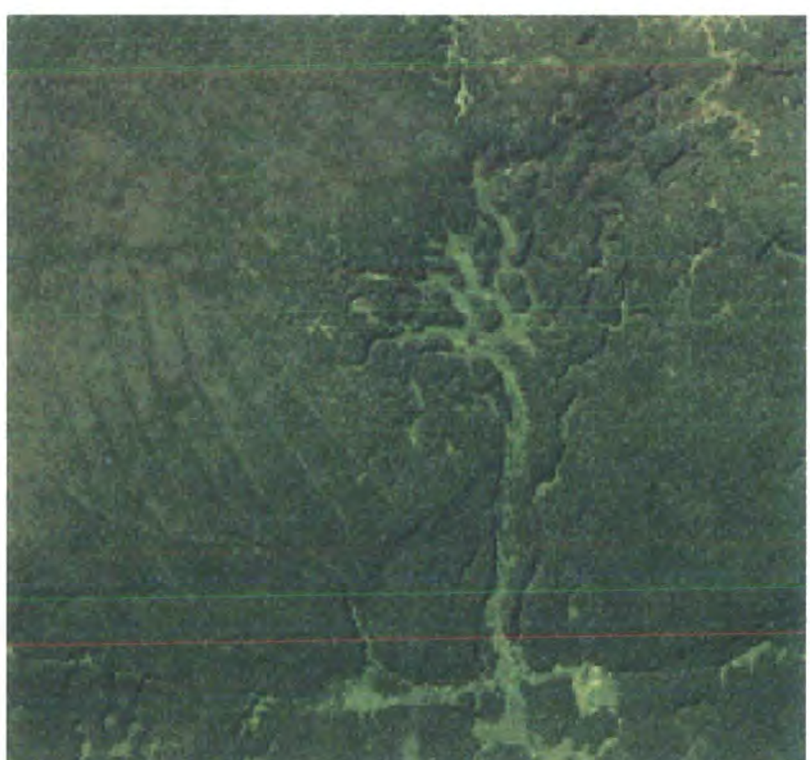
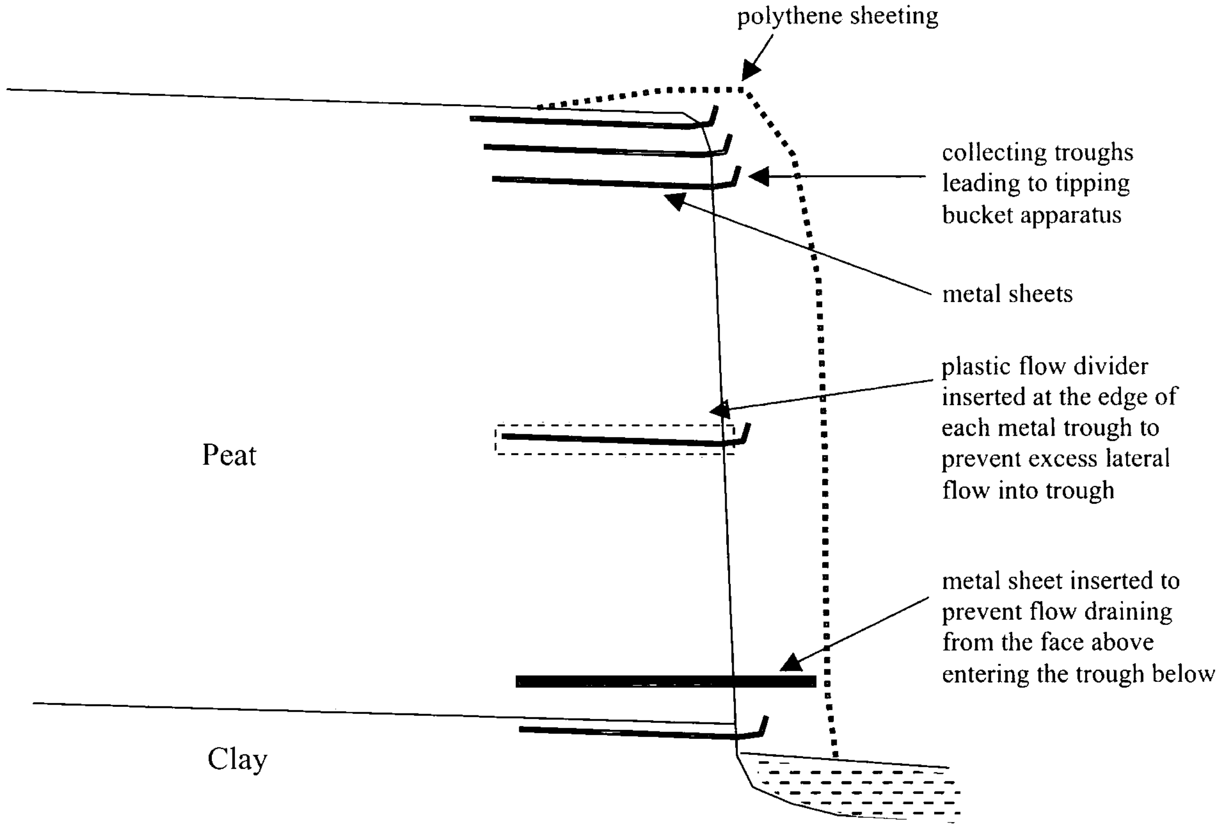


Figure 4.17. The gripped and eroded subcatchments monitored on Burnt Hill. a) A map of the grips and main lines of dissection. b) An aerial photograph showing the nature of the erosion and gripping; reproduced with kind permission from NERC. NERC site 94/9 (4), taken 6.8.95, run 6, plate 8886, scale and direction as indicated in (a).

soil disturbance. This may be especially important in a fibrous and anisotropic soil like peat. Rather than digging away from below and inserting the sheet, much less disturbance is achieved if a rigid sheet is simply slotted into the soil. This is easily achieved in an organic peat soil where there is rarely obstruction of trough insertion by large clasts of soil particles or rocks. A trench was dug in the peat, or at suitable locations a peat face was cleaned off, and 50 cm width aluminium troughs carefully inserted at 1 cm, 5 cm, 10 cm, and 50 cm depth (Figure 4.18a). A trough was also inserted below the interface between the peat and substrate. Flow dividers were inserted flush with the edge of the troughs in order to prevent flow leaking from upper layers into lower troughs and to prevent excess lateral flow distortion. A flow divider was also inserted just above the interface to prevent errors in measurement of flow at the interface. Outflow was measured by tipping-bucket flow recorders connected to a Campbell CR10X datalogger (Figure 4.18b-d). Movement of a magnet attached to the tipping-bucket activated a reed switch such that pulses could be counted by the datalogger. Because of the limitations of the datalogger only five tipping-buckets could be operated at any one time at a given site. Tipping-buckets were calibrated in the laboratory and flow volumes integrated over 15 minutes. Troughs and flow recorders were covered to prevent entry of precipitation and debris. Where the frequency of tips was low due to small volumes of throughflow, tips may not occur in successive 15-minute intervals even though flow was occurring; simply the flow was not great enough to fill a tipping bucket. In these cases a 5-period binomial (Gaussian) filter was added to the dataset to enable hydrographs to be drawn up (N.J. Cox, pers. comm.). This meant that for unfiltered data a tip may have occurred < 15 minutes before it was logged; in the case of the filtered data, at very worst, the flow is displayed 37.5 minutes after it occurred.

The digging of a pit or measurement at a peat face may be problematic as these may be locations where the water table will drop close to the peat face resulting in changes in throughflow and saturation-excess flow pathways and processes. These effects need to be taken into consideration. Monitoring of the water table close to the edge of certain faces did suggest a fall very close to the face (for example see water table transect from G2, Figure 4.30). As the technique is based upon collecting water seeping from a free face the troughs will collect only saturated throughflow (Atkinson, 1978). This is because water at the free face must be at atmospheric pressure in order to leave the pore space of the soil and flow away. This soil at the face must be saturated. Inevitably, if the

a)



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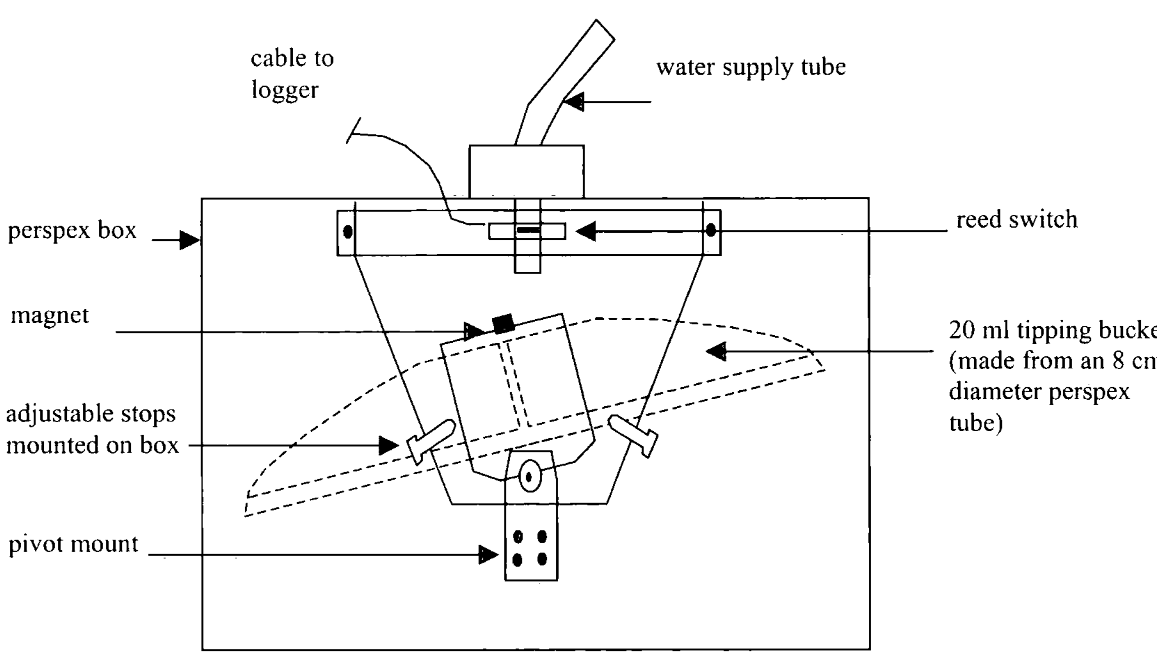


Figure 4.18. Automated recording of OLF and throughflow. a) Trough installation, b) Tipping-bucket design.

c)

d)

Figure 4.18. (continued) Automated recording of OLF and throughflow c) Flow channelled into tipping buckets connected to d) Cambpell CR10X datalogger which could be easily downloaded in the field using a portable PC.

soil at the face itself is saturated, a wedge of saturated soil will extend upslope, perhaps into soil which would not normally be saturated had an artificial free face not been constructed. This process naturally helps counteract the effect of lowering of water table at the face of a soil. Insertion of troughs into the soil face by a considerable distance may also counteract the effect of lowered water table and artificial saturation. Atkinson (1978) states that as a general principle throughflow gutters should ideally be placed on natural faces at stream banks or at the base of slopes as in the study of Weyman (1970, 1973) where distortions of the hydrograph and contributing area were at a minimum. Where possible this scheme was adopted (Table 4.4). Knapp (1973) describes how pits may result in distortion of the net of hydraulic potential such that the troughs receive drainage from areas which are not directly upslope of them. In order to reduce the effect of flow net distortion on throughflow, flow dividers were inserted at the sides of the troughs through the profile. However, because of some distortion and because it was often very difficult to determine the contributing area to the troughs, the results will generally be presented as a guide to the relative roles of runoff processes rather than exact water balances. Nevertheless, the throughflow trough dataset provides essential process-based information on runoff production in blanket peat.

Table 4.4 gives details of the characteristics of each site monitored and of the installation type used. The locations of the sites are given in Figure 4.1. Site 'H1' refers to a small intact blanket peat hillslope where flow was recorded simultaneously at topslope, midslope and footslope sites. The nature of the terrain at these sites on H1 is indicated in Figure 4.19. Because of the limited number of channels available on the datalogger for pulse counting, and in order to allow simultaneous monitoring at the three sites, troughs were only inserted into the upper two monitoring layers of the peat on H1 (i.e. 1 cm and 5 cm).

At some locations discharge could be witnessed from small foci on peat faces and from subsurface pipes. Discharge from 'seepage points' was monitored using plastic guttering inserted below the outlet and flow was channelled to a tipping-bucket flow recorder. Monitoring of pipe networks will be discussed in Chapters 7 and 8. A pipe was found at a gully-head on Burnt Hill acting as an outlet for flow from the monitoring site coded E2. Pipeflow at this site will be discussed in the present chapter as it provides valuable information on runoff production on the eroded Burnt Hill subcatchment. Flow from this pipe was channelled to a tipping bucket flow recorder.

Table 4.4 Site characteristics and instrumentation where runoff was monitored. See Figure 4.1 for site locations.

Plot	Location of flow recording	mean slope, m m^{-1}; length, m; mean (intact) peat depth, m	Veg (Dominated by *Calluna*, *Eriophorum* and *Sphagnum* spp.) in order of prominence	C/P/W*	Face already present or trench dug	Approximate contributing area, m^2	Automated monitoring device	Other details
G1	Burnt Hill, gripped section	0.06, depth 2.0	C-E, Limited S	At G2		24990	90° V-notch weir	
E1	Burnt Hill, eroded section	0.08, uneroded depth 2.3	C-E abundant S in bog pools and C and S on gully floors	At E2		19120	90° V-notch weir	
G2	Burnt Hill, grip face	0.04, 15, 2.0	E-C	C W	F	7.5	Tipping bucket	
E2	Burnt Hill, gully-head	0.05, 30-60, 2.8	S-C-E	C P W	F		Tipping bucket	Pipe flow measured also
H1	Topslope	0.04, 11, 0.8	E-C	C P W	T		Tipping bucket	Flow measured from surface layer only
H1	Midslope	0.09, 38, 0.6	E-C	C P W	T		Tipping bucket	Flow measured from surface and 5 cm only
H1	Footslope	0.11, 56, 0.9	E	C P W	F		Tipping bucket	Flow measured from surface and 5 cm only
H2	Footslope	0.07, 230, 1.2	E-C	C P W	F		Tipping bucket	Runoff also measured manually from 40 troughs
H3	Midslope	0.07, 170, 1.1	S-C		F		Tipping bucket	
H4	Footslope	0.07, 145, 2.3	E-C		F		Tipping bucket	
H5	Midslope	0.05, ?, 1.0	E-C		F		Tipping bucket	
H6	Topslope	0.02, 10, 1.3	E		T		Tipping bucket	
M1	3 macropores at peat face	0.07,145, 1.2	C		F		Tipping bucket	
M2	4 macropores at peat face	0.06, ?, 0.9	C-E-S		F		Tipping bucket	

* C =crest stage tubes, P = piezometers, W = dipwells

a) b) c)

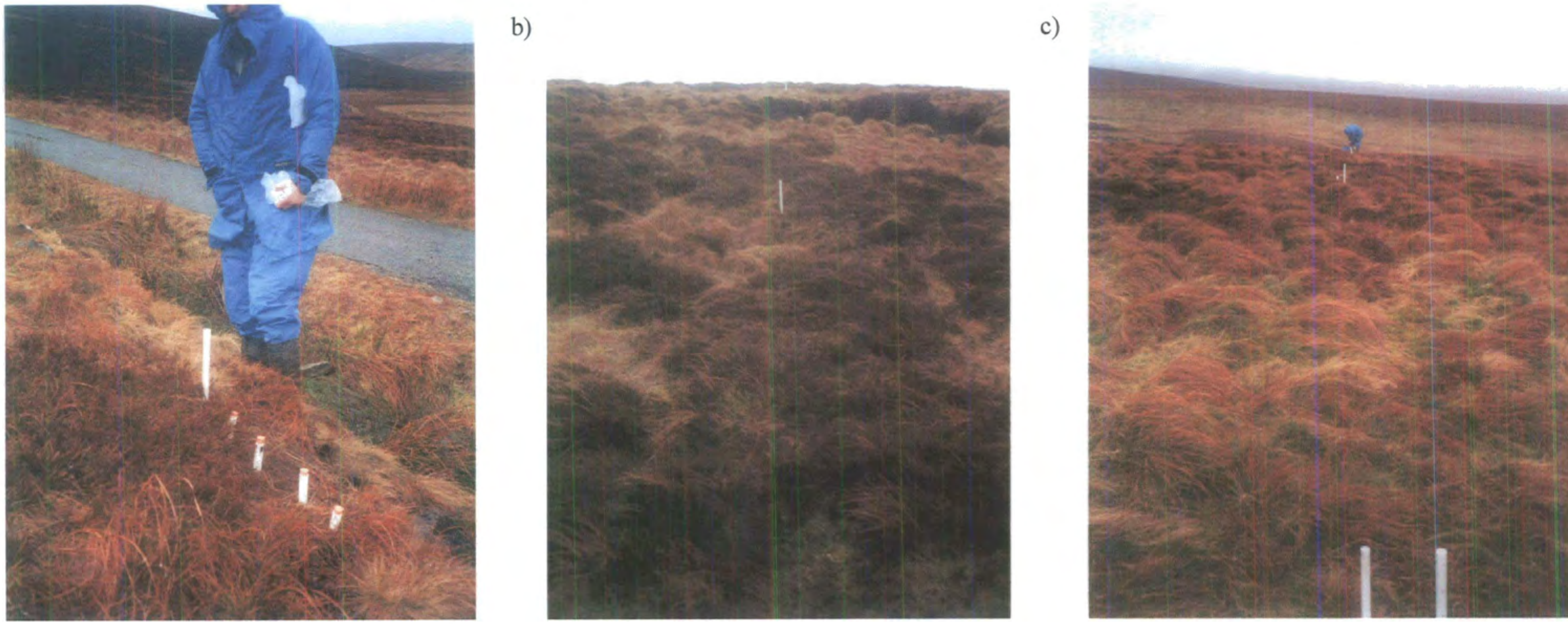


Figure 4.19. Views of study hillslope coded H1. a) Piezometer nest at the foot of H1. b) the midslope of H1 and c) the topslope of H1.

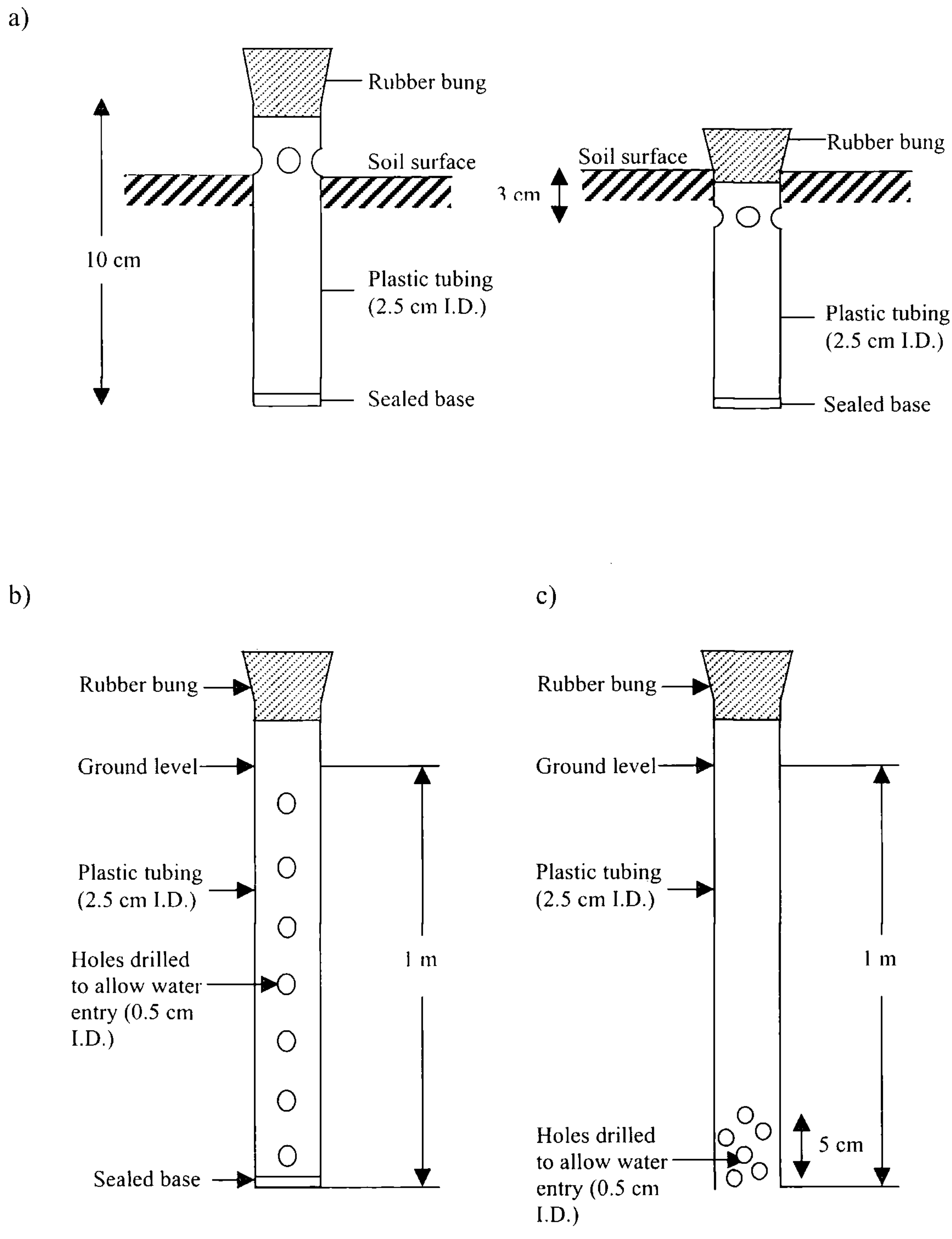


Figure 4.20. Manually operated monitoring devices for examining flow processes on blanket peat hillslopes, a) crest-stage tubes, b) dipwell, c) piezometer.

b) Manual measurement of OLF and throughflow processes

As well as automated flow measurement at sites H1, H2 and at the two Burnt Hill sites (E2 and G2) monitoring of overland and near-surface flow was done using a network of crest-stage tubes. Flow entry points were placed at the surface of the peat and at 3, 6, and 9 cm into the peat mass (Figure 4.20). By burying the tubes to a point where the holes were level with the monitoring height, any flow or ponding to that height resulted in filling of the tube with water. This system provided a means of monitoring the occurrence of OLF and near-surface flow which occurred between the time periods of emptying the tubes. Emptying the tubes was performed with a large syringe such that tube disturbance was kept at a minimum. Networks of thin PVC dipwells and piezometers were also set up at the four sites. A thin borehole was created with a screw auger and the tubes slotted into position. Water depth was measured using an electronic rod inserted into the tube with a sensor at its tip. When the sensor came into contact with the water an LED was activated and the length of rod inserted could be determined. Measurements could be made to the nearest mm. Because of the movement of the ECN transducer as demonstrated earlier (Figure 4.6), the height of the top of PVC tube above the peat surface was also measured coincident with each water level measurement. Slug withdrawal tests were carried out in some of the piezometers at E2, H1 and H2 (see Figure 4.1 and Table 4.4) in order to estimate the hydraulic conductivity of the peats at depth (using compressible soil theory). The piezometers were in position for at least six months before the tests were performed in order to ensure stress-adjustment lags caused by the installation were minimal (Baird and Gaffney, 1994).

4.3.2 Hydraulic conductivity of Moor House peats: the use of compressible soil theory.

The only work estimating the hydraulic conductivity of the peat at Moor House was a basic permeameter experiment performed by Smith (1956). Here cores of around 20 cm length were applied with water and rates from 'severely burned Burnt Hill' ranged from 4.4×10^{-5} to 1.1×10^{-3} cm s^{-1}. For an intact *Sphagnum* peat (probably taken from the surface of the bog) rates were of the order of 6.6×10^{-3} cm s^{-1} to 2.8 x 10-1 cm s^{-1}. These results should be treated with caution however, given the uncertain sampling strategy and method.

Head recovery tests are often performed to obtain values of hydraulic conductivity where either water is added to (slug injection) or removed from (slug withdrawal) the piezometers and the recovery to the original water level in the instrument is recorded.

The degree to which compression and swelling of peat affects head recoveries is not well understood and it is important therefore that both rigid and compressible soil theories are applied to piezometer data from peat soils (see Chapter 2). In the present study head recovery tests (slug withdrawal) were performed on piezometers with 5 cm tips at depths ranging from 80 cm to 10 cm from the surface. Results were obtained using both rigid (Hvoslev, 1951) and compressible soil theories (Brand and Premchitt, 1982) and this allows comparison of hydraulic conductivity values to be made between the two theories. Baird and Gaffney (1994) applied this technique to a fenland peat, but this is the first application of the technique to the author's knowledge in a blanket peat.

Hvorslev's (1951) solution for an incompressible soil uses the basic differential equation that describes saturated flow through a falling head permeameter. The pressure head u, at time t, in a soil of hydraulic conductivity k is related to the initial pressure head u_0 and the equalisation pressure head u_∞ by

$$\frac{u_\infty - u}{u_\infty - u_0} = \exp\left(\frac{-Fkt}{V\gamma_w}\right) \qquad [4.1]$$

where γ_w is the unit weight of water, V is the volume of water required to flow into or out of the piezometer system to equalise a unit pressure difference between the piezometer and the surrounding soil. In a standpipe piezometer V is numerically equal to the cross-sectional area of the piezometer. F is the shape factor (dimensions of length) which describes the flow field geometry around the piezometer (Kirkham, 1945; Hvorslev, 1951; Youngs, 1968; Brown and Hodgson, 1988 and Brand and Premchitt, 1980). For the present study, the shape factor has been determined from the equation of Brand and Premchitt (1980)

$$F = 7\,d + 1.65\,l \qquad [4.2]$$

where d is the diameter of the tip and l the tip length. In developing equation 4.1 Hvorslev (1951) also assumed that the soil around the piezometer tip was isotropic, fully saturated and infinite in extent. When equation 4.1 is solved for k it becomes the familiar solution of Kirkham (1945)

$$k = \frac{V\gamma_w}{-Ft}\log e\left(\frac{u_\infty - u}{u_\infty - u_0}\right) \qquad [4.3]$$

In compressible soils, the compression and swelling of the soil around the piezometer can play a major part in piezometer response and equation 4.1 may not adequately describe the equalisation process. The state of stress in a soil can be described by the effective stress equation

$$\sigma_T = \sigma' + u \qquad [4.4]$$

where σ_T is the total stress and σ' the effective stress. Equation 4.4 can be used to analyse compression and swelling on head recovery in a piezometer. Immediately after slug withdrawal there will be an increase in effective stresses around the piezometer tip as pore water pressure decreases while the total vertical stresses remain the same. As the water level recovers, effective stresses will decline causing more water to enter storage and increase the rate of head recovery. For cylindrical piezometers in compressible soil the rate of pressure head recovery is given by the consolidation equation in axisymmetrical cylidrical coordinates.

$$c\left(\frac{\partial^2 u}{\partial r^2}+\frac{1}{r}\frac{\partial u}{\partial r}+\frac{\partial\ u}{\partial z^2}\right)=\frac{\partial u}{\partial t} \qquad [4.5]$$

where r is the radial distance from piezometer tip mid point, z is the vertical distance from piezometer mid-length, and c is the coefficient of 'consolidation' that accounts for both compression and swelling. Brand and Premchitt (1982) used a numerical solution to equation 4.5 to show that the soil - piezometer system was well represented by a control parameter

$$\lambda = \frac{4\pi a^2 bm}{V} \qquad [4.6]$$

where a is the outside radius, b the half length of the piezometer tip, and m the coefficient of volume compressibility of the soil. The value of λ characterises accurately the shape of head recovery for which there is a unique ratio between t_{90} (time taken for the head to recover to 90 % of initial head difference between piezometer and soil) and t_{50}. Using λ as a control parameter, Brand and Premchitt (1982) derived equalisation monographs based on t_{50} and t_{90} that can be used to calculate the hydraulic conductivity and the coefficient of consolidation.

To illustrate the responses of the piezometers, the head recoveries of two piezometers are shown in Figure 4.21. The piezometers are coded by site (either H1, H2 or E2) and then piezometer nest within that site (A, B, C and so on) followed by the depth of the piezometer tip (in cm). The results from piezometer H2 C60 corresponds quite closely to the response described by Hvorslev (1951). H2 C20 shows pronounced deviation from the curve. All of the recoveries deviated from rigid soil theory, but it is not clear how much error the effects of compressibility and swelling will introduce into hydraulic conductivity calculations. Table 4.5a allows comparison as it shows values of hydraulic conductivity calculated using equation 4.1 and values calculated from the nomograph of Brand and Premchitt (1982).

For those piezometers where insufficient response occurred for t_{50} to be achieved, the hydraulic conductivity was calculated using equation 4.1 for the time of the last reading (t_m). These values are given as a guide in the first column of Table 4.5a. In a rigid soil t_{50}/t_{90} is always 3.322 whereas in a compressible soil the ratio will vary but will always be greater than 3.322 and will increase with the volume of compressibility of soil (Premchitt and Brand, 1981). All of the t_{50}/t_{90} ratios were above 3.322 for the blanket peat piezometers. The ratio, t_{50}/t_{90}, is essentially a measure of the effect of compressibility on the head recovery; as it increases it describes the increasing shallowness of the head recovery sigmoid (seen for example in Figure 4.21). The added bonus of the Brand and Premchitt (1982) method is that a value of the coefficient of consolidation for the peat can be obtained. Values are given in the last column of Table 4.5a. Values of *c* could be important in modelling water flow in peats which are subject to rapid changes in pore water pressures, for example during pump drainage, which would in turn cause changes in the effective stresses in the soil and its storage. Values of *c* determined by Baird and Gaffney (1994) ranged from 0.56 at 2 m to 13.23 at 1.2 m depth for a poorly humified fenland peat. These values fall within the three orders of magnitude variation found at Moor House. The peat at H1 has a higher coefficient of compressibility than at H2 or E2. The peat at H1 also has higher hydraulic conductivities than at the other two sites.

In all but two of the piezometers, values of hydraulic conductivity using t_{50} were greater than the value calculated using t_{90}. This is because equation 4.1 fails to account for variable storage and release of water giving an apparent increase in hydraulic conductivity in early time. The effect is small in those piezometers with a value of t_{90}/t_{50} less than about 4. However, Hvorslev's (1951) theory does appear to be invalid for all the piezometers.

Comparing the hydraulic conductivity values calculated using both theories shows that in all tests where 90 % recovery occurred within the measurement period, the value of k^* was always much lower than k_{90}. Baird and Gaffney (1994) reported that both rigid and compressible soil theories gave values of hydraulic conductivity for each piezometer installation at a fenland site within a factor or two of each other. Clearly this

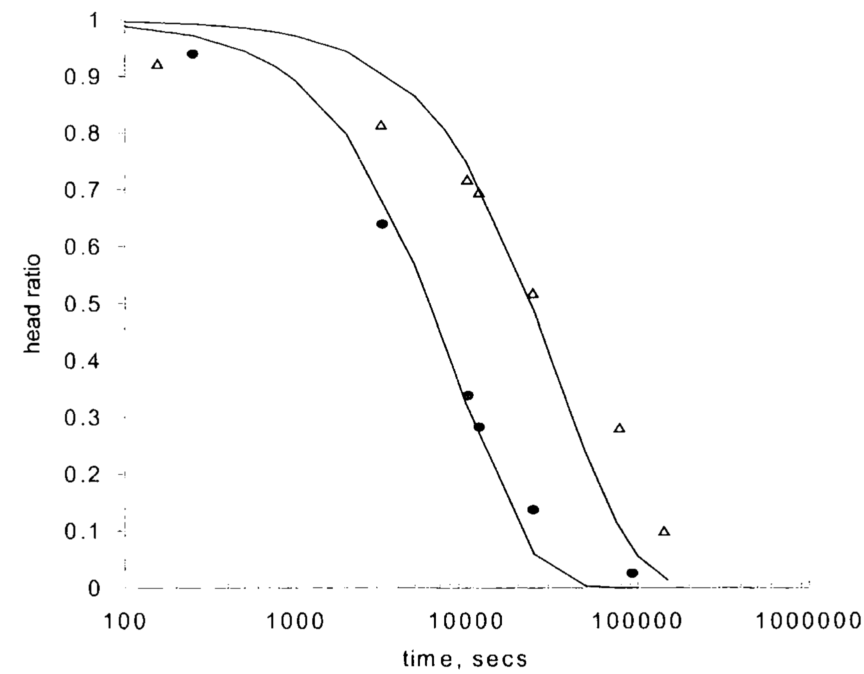


Figure 4.21. Example head recoveries. Closed circles for H2 C60, open triangles for H2 C20. The solid lines are fitted responses (least differences) according to equation 4.1.

is not the case for the Moor House blanket peats where generally the difference is a factor of about five up to an order of magnitude. Penman (1961) compared hydraulic conductivity values from a triaxial cell apparatus set up as a constant head permeameter, and values using piezometer head recovery data applied to equation 4.1. He found that the two values were only in close agreement when $t_{99.99}$ was used in equation 4.1. Hvorslev (1951) similarly suggests that reliable estimates of hydraulic conductivity in compressible soils can only be calculated using equation 4.1 when exchanges to and from storage are nearly complete at the end of the head recovery process. Baird and Gaffney (1994), however, in their fenland peat study found that values of k^* were often closer to k_{50} than k_{90} and concluded that both Hvorslev's (1951) and Brand and Premchitt's (1982) theories give values of hydraulic conductivity that are too high. Results from Table 4.5a do not indicate that Brand and Premchitt's method gives hydraulic conductivity values that are too high. Baird and Gaffney (1994) do recommend Brand and Premchitt's (1982) theory as a standard comparison of hydraulic and storage properties between different peat types.

Although Baird and Gaffney (1994) measured peat from 1.2 to 2 m in depth, the results from 10 cm to 80 cm at Moor House provide hydraulic conductivity values that are generally an order of magnitude lower than those in the fenland bog. Furthermore the

Table 4.5a. Values of hydraulic conductivity calculated using rigid and compressible soil theories for each piezometer. K_{50} and K_{90} at 50 % and 90 % equalisation time. K_m at maximum recorded equalisation time, K* and C calculated using Brand and Premchitt's (1982) response time charts. IT = insufficient response time.

		Hvorslev (1951)			Brand and Premchitt (1982)	
Site/depth	K_m (H ratio) ($\times 10^{-6}$ cm s^{-1})	t_{90}/t_{50}	K_{50} ($\times 10^{-6}$ cm s^{-1})	K_{90} ($\times 10^{-6}$ cm s^{-1})	K* ($\times 10^{-6}$ cm s^{-1})	C ($\times 10^{-3}$ cm^2 s^{-1})
H1						
A80	0.182 (3.9)	IT	IT	IT	-	-
A60		3.478	7.327	6.999	0.885	2.69
A35		IT	13.997	IT	-	-
B80	0.126 (12.0)	IT	IT	IT	-	-
B60		3.815	23.703	20.638	2.474	5.91
B35		3.998	21.903	18.199	3.921	49.24
B20		5.377	121.170	74.856	9.371	18.67
C80	6.494 (5.9)	IT	IT	IT	-	-
C60		3.632	30.689	28.073	4.703	23.80
C35		4.000	34.440	28.602	5.309	30.32
C20		18.014	68.880	12.712	1.240	0.09
C10		3.25	102.421	104.634	14.87	90.25
H2						
A80	6.347 (41.2)	IT	IT	IT	-	-
A60		4.175	9.241	7.352	1.360	6.09
A20	1.343 (24.8)	IT	IT	IT	-	-
A10	1.135 (21.3)	IT	IT	IT	-	-
B80	0.544 (42.2)	IT	IT	IT	-	-
B60		28.924	52.058	5.979	0.697	0.04
B35		3.621	1.492	1.369	0.317	5.23
B20		4.221	1.613	1.269	0.241	0.93
B10		3.516	0.991	0.936	0.201	3.42
C80	5.874 (5.3)					
C60		7.849	10.825	4.581	0.764	0.90
C35		3.504	1.243	1.179	0.327	3.86
C20		3.892	1.876	1.602	0.343	4.11
C10		3.321	0.755	0.755	0.178	2.53
E2						
A80		3.550	0.633	0.592	0.140	2.17
B80		3.963	1.826	1.531	0.249	1.15
B60	0.225 (46.0)	IT	IT	IT	-	-
B35	0.012 (3.4)	IT	IT	IT	-	-
C80		5.923	3.129	1.755	0.221	0.13
D80		4.398	0.916	0.692	0.158	0.09
D60		4.912	0.703	0.476	0.086	0.21
D35		3.635	0.639	0.584	0.107	0.77
E80		13.986	4.817	1.144	0.140	0.03
E60	0.235 (47.6)	IT	IT	IT	-	-
E35		IT	0.502	IT	-	-
E20	0.136 (31.1)	IT	IT	IT	-	-
F80	0.213 (44.2)	IT	IT	IT	-	-
F60		IT	0.504	IT	-	-
F35	0.026 (6.8)	IT	IT	IT	-	-
F20		IT	0.685	IT	-	-
G80		8.002	22.960	9.534	1.034	0.63
G60		3.330	0.356	0.355	0.076	1.26
G35	0.059 (1.1)	IT	IT	IT	-	-
H80		3.940	0.591	0.499	0.097	0.52
H60	0.184 (39.6)	IT	IT	IT	-	-
H35	0.102 (24.4)	IT	IT	IT	-	-
I80		IT	3.461	IT	-	-
I60	0.137 (31.4)	IT	IT	IT	-	-
I35		3.236	0.752	0.823	0.158	90.21
J80	0.170 (37.2)	IT	IT	IT	-	-
J60	0.120 (28.0)	IT	IT	IT	-	-
J35		IT	56.000	IT	-	-

results suggest that the hydraulic conductivity at a particular depth can vary by at least one order of magnitude if not two. Table 4.5b shows the mean values of hydraulic conductivity at the depths tested and indicates the variability found. There is no significant difference ($p < 0.05$) between piezometer results from the different depths measured at Moor House. Even at 10 and 20 cm depth hydraulic conductivity could be found to be as low as 3.43×10^{-7} cm s^{-1} and 1.78×10^{-7} cm s^{-1} respectively. This provides evidence that runoff production measured by the runoff troughs at 5 and 10 cm depths is likely to be a product of percolation-excess resulting in enhanced lateral flow. Infiltration into the surface of the peat may be a rapid process (see Chapters 5 and 6) but at shallow depths (e.g. 10 cm) hydraulic conductivities can be very low resulting in the generation of lateral runoff.

Table 4.5b. Mean hydraulic conductivity for each depth using head recovery piezometer data. Note that the raw data is skewed and that logarithmic transformation is required to compare variability between datasets. Hence the geometric mean is different from the arithmetic mean. K* has been used where available, K_m for the remaining tests.

Depth, cm	N	Mean k, cm $s^{-1} \times 10^{-6}$
10	4	4.096
20	7	1.908
35	12	1.958
60	14	0.889
80	15	1.466

Geometric mean and 95 % confidence interval for each depth

```
-+---------+----+-----+-----+---+------
           (-------------*-------------)
               (---------*----------)
   (-------*-------)
          (-------*------)
          (------*------)

-+---------+----+-----+-----+---+------
0.05      0.3  0.5    1     2   3
```

k, cm $s^{-1} \times 10^{-6}$ (log scale)

4.3.3 Runoff response of the Burnt Hill subcatchments

4.3.3.1 Contemporary runoff response

Total runoff during the study period was greater from the gripped hillslope (G1) than the eroded slope (E1) with G1 showing much higher storm peaks. However, G1 was larger in area (2.49 ha) than E1 (1.91 ha) and so discharge has been converted into runoff per unit area to allow comparison. Figure 4.22 gives discharge data for the 1999 water year for the two hillsides. Response from the slopes appears very similar and is very flashy from both (cf. Trout Beck, Figure 4.3) although E1 produces higher area-weighted peak flows. It is also evident that lower flows are sustained for longer periods

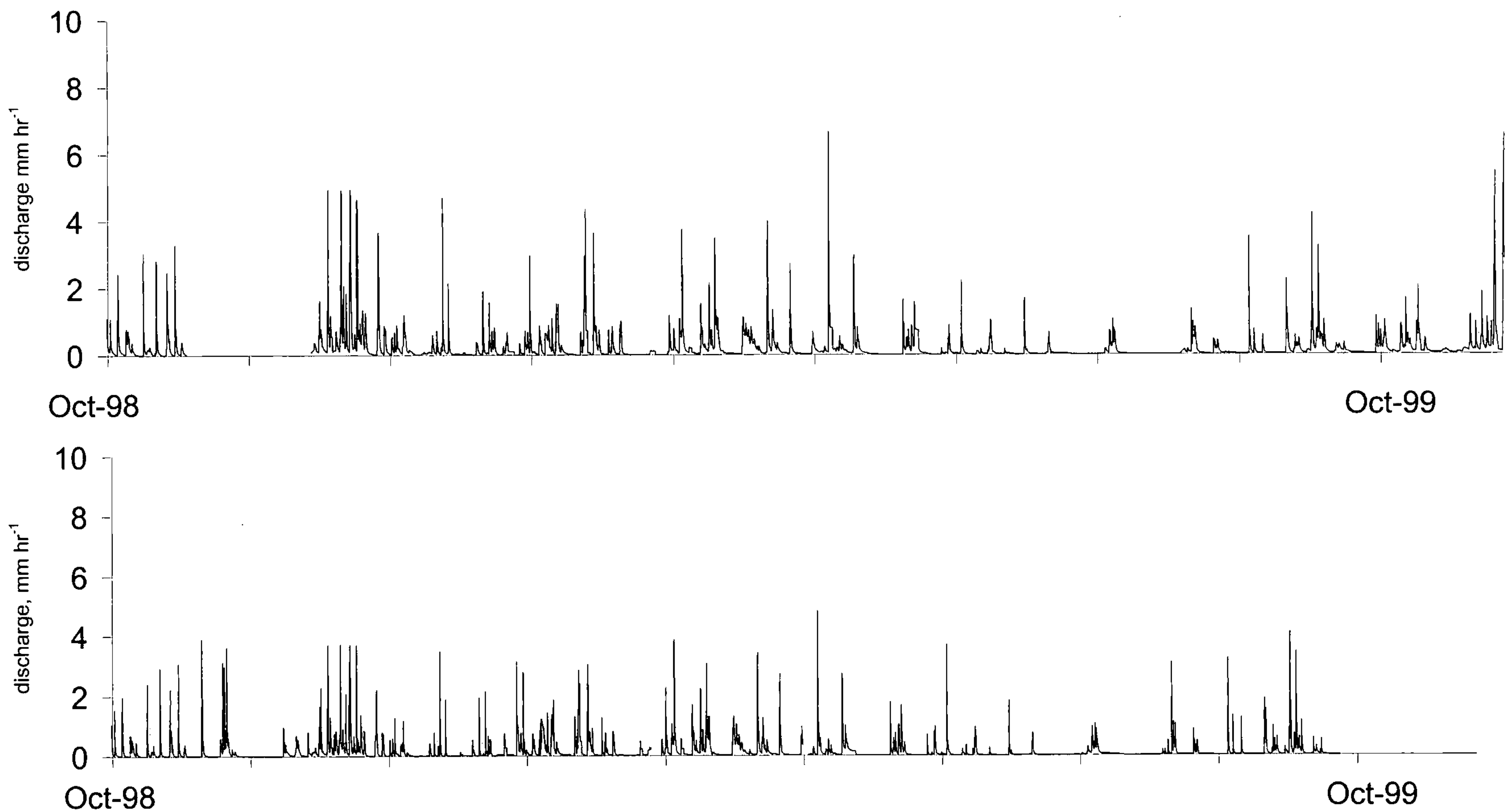


Figure 4.22. Discharge at the Burnt Hill gauging stations, during the 1999 water year, converted to mm hr^{-1}.

from E1 than for G1, with G1 producing narrower, more spiked, hydrographs. Conway and Millar (1960) noted that, during rainless summer periods, of their four monitored catchments, flows ceased earliest from Burnt Hill. Although they did not specifically refer to the effect of artificial drainage on low flows, it does appear from Conway and Millar's (1960) graphs that baseflow from their drained and unburned catchment was lower than from the two undrained catchments. A number of studies (e.g. McDonald, 1973) have interpreted their paper as providing evidence that drainage reduces baseflows (cf. Von Humboldt's 'sponge' theory). Robinson (1985) shows that more flows from Burnt Hill were at the lower range of the flow record than at the other three sites; these sites were very similar to each other in times of low flow. Thus, it is suggested by Robinson (1985) that burning reduced low flows but ditching alone did not. The flow duration curves presented in Figure 4.23 show how the two sections of Burnt Hill compare over the 1999 dataset. It is clear that flows are sustained much more from E1 than G1 over the medium to low flow range. That is not to say low flows are sustained a great deal from either catchment; the data are comparative. Discharge from both weirs almost ceased during some periods of the summer of 1999 and indeed at the very lowest flows runoff from the eroded subcatchment seemed to be more likely cease before runoff from G1. Nevertheless, the ditched section of Burnt Hill, almost 50 years after burning, reduces low to medium flows far more than the naturally dissected section. Essentially the mean recession limb of the hydrographs is much steeper for G1 than E1. Hence measurement of flow from the two subcatchments provides evidence which conflicts with the more normative view of Robinson (1985). Comparison of 1999 data with those from Conway and Millar (1960) will be required, however, to establish whether recovery from burning has effected low flows (see below).

Monthly rainfall and runoff totals from E1 and G1 are shown in Figure 4.24. Although typically the runoff pattern corresponds to the rainfall pattern as discussed for Trout Beck above, it is clear that a much greater proportion of runoff is produced from E1 than G1. For the 1999 water year the rainfall:runoff ratio was 73.8 % for E1 and only 57.6 % for G1 (Table 4.7a). This shows that the original data configuration of Conway and Millar (1960) was inadequate; an important finding given the wide citation that their paper has. Conway and Millar (1960) suggested that Burnt Hill had almost no capacity for storage of rainwater whereas more intact subcatchments could retain up to 15 cm of water. Notwithstanding that change may have occurred since Conway and Millar's study, the evidence presented here suggests that the gripped subcatchment can

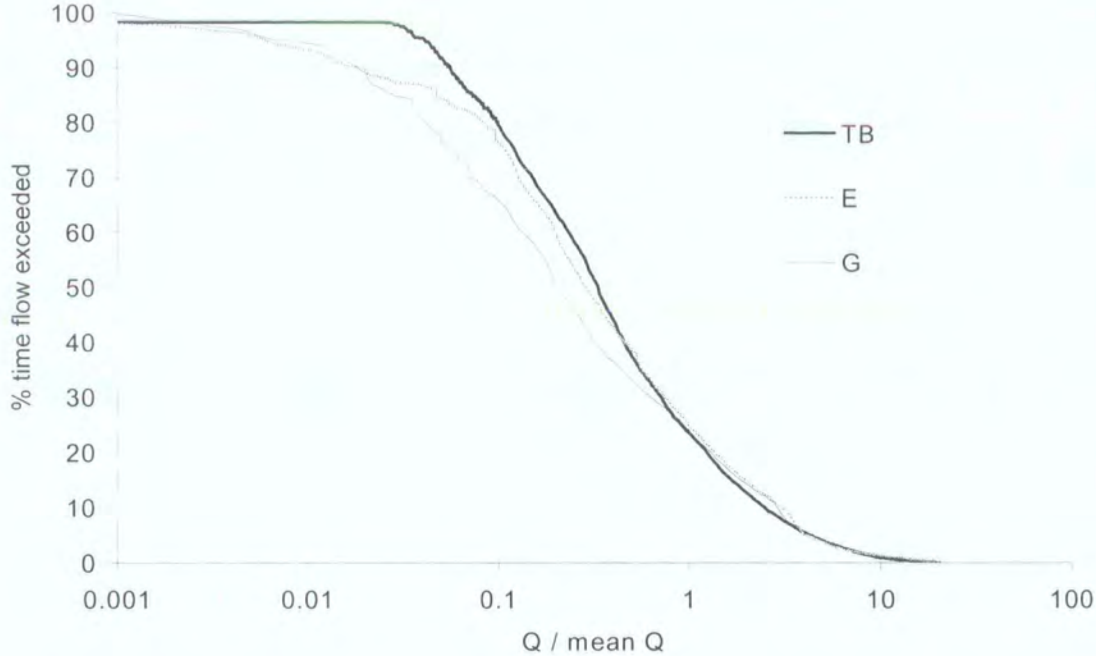


Figure 4.23. Flow duration curves for Trout Beck, E1 and G1 during the 1999 water year. A log scale has been used rather than a probability scale in order to highlight low flows.

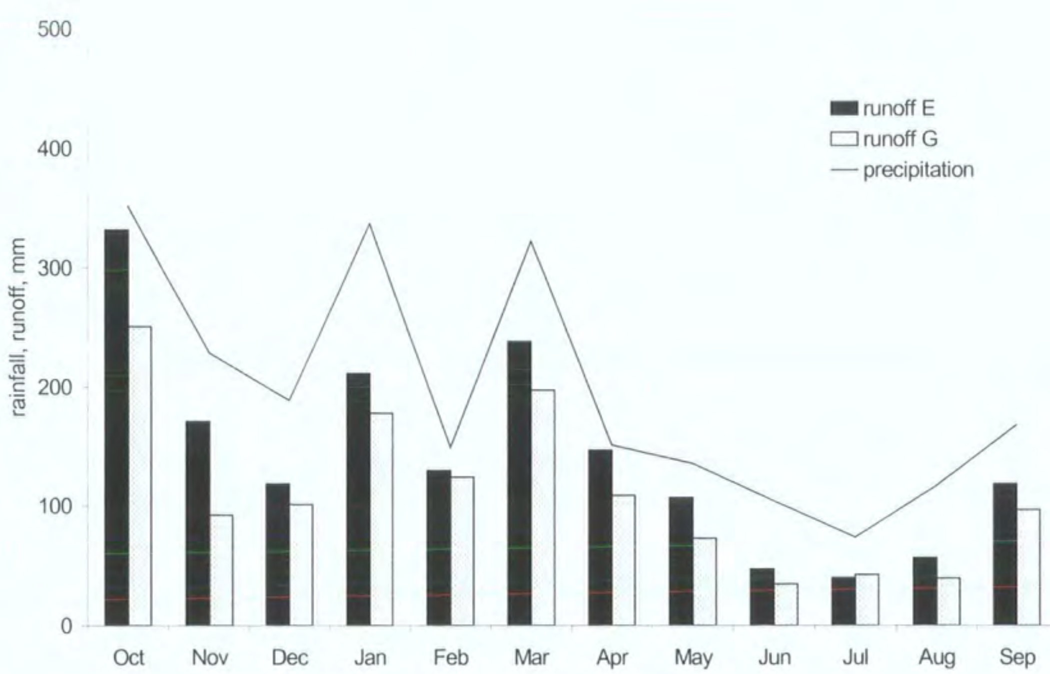


Figure 4.24. Monthly precipitation and runoff totals for the Burnt Hill gauging station during the 1999 water year. Runoff totals are area-weighted for each subcatchment.

store much more water than E1. It would therefore seem that results from the 1999 water year on Burnt Hill bring the effect of ditching on Burnt Hill more into line with results from many other studies as discussed in Chapter 2. There will, or course, have been changes in the capacity of the grips over time related to erosion and vegetation change (see below).

The difference in runoff response from the two subcatchments could be partially explained by comparison of drainage density. The mean drainage density of the Trout Beck catchment is 3.57 km km^{-2}. Such high drainage densities suggest a natural propensity towards quick drainage. Caution is required in assessing drainage densities in peatland catchments since maps in the heavily dissected areas of blanket peat moorland probably underestimate the drainage density (Burt and Oldman, 1986). For Burnt Hill the drainage density was calculated using aerial photographs and maps drawn by Bower (1959) and Conway and Millar (1960) and a GPS survey of the hillslope. The mean drainage density of G1 was 66 km km^{-2} and 77 km km^{-2} for E1. Thus it is to be expected that higher flows would occur from E1. The natural drainage is greater in density than the artificial drainage. The huge difference in drainage density between Trout Beck and the two subcatchments must be related to the measurement scale. The Trout Beck estimate has not included all of the small gully networks and eroded peat flats that contribute to drainage (this measurement would require a great deal of time and benefit from utilising GIS). Nevertheless as drainage densities on E1 and G1 have been estimated at the same scale then the subcatchment data can be used for comparative purposes. These estimates, of course, do not allow an areal estimation as they are based on the length and not the area of ditching or gullying. Gullies occupy approximately 48 – 55 % of the area of E1 with ditches occupying 3 - 4 % of the area of G1. With more water falling directly on to the surface of gullies, there is a natural propensity for increased runoff from E1, and greater peak flows. Furthermore the gullies are often a metre deep whereas the grips are only around 40 cm deep. At the same time, however, runoff ratios from E1 are equivalent to that from the Trout Beck catchment as a whole which is not as heavily eroded.

4.3.3.2 Evidence for change in runoff generation on Burnt Hill since the 1950s.

Mean lag times and hydrograph intensities are given in Table 4.6 for the Burnt Hill sites. The data calculated from Robinson (1985) are also shown. The table suggests that Burnt Hill flows are not as flashy as they were 40 years ago. However, given that

Robinson (1985) only used five storms for analysis this conclusion is rather limited. Conway and Millar (1960) noted that often 'peak lag times' on Burnt Hill could be 0.3-0.5 hours. This is still the case, with 8 of the 44 storms analysed from E1 and 19 of the same storms analysed for G1 having lag times within 0.5 hours. Flows are much more spiked and flashy from Burnt Hill than the Trout Beck catchment. However, the response from a small catchment is likely to be flashier anyway. Peak flows on average are achieved about 0.3 to 0.5 hours earlier from G1 than E1 (although precipitation data were collected at 0.25 hour intervals which affects the resolution of the results). As the catchment area of G1 is greater than E1 it therefore contributes far more to the overall discharge from Burnt Hill.

Table 4.6 Hydrograph characteristics of the Trout Beck and Burnt Hill flows.

	Burnt Hill 1959-1961 from Robinson n=5	Gripped 1999 n=43	Eroded 1999 n=44	Drained C and M subcatchment from Robinson, 1959-1961 n=5	Trout Beck 1994-1999 N= 58
Peak lag, hrs	1.6	1.7	2.1	1.9	2.6
Trec, hrs		31.1	31.9		28.9
Hydrograph Intensity		48.6	44.4		38.8

Peak Lag = time from peak rainfall to peak discharge, hrs
Trec = Time from rain end to return to pre-storm flow
Intensity = peak flow, $m^3 s^{-1}$ divided by total storm discharge, $m^3 \times 10^{-6}$.

The total runoff ratio for Burnt Hill during the 1999 water year appears to be less than during any of Conway and Millar's water years (Table 4.7b) and would suggest that a change has taken place since the late 1950s. Conway and Millar (1960) suggested that after burning the peat lost its ability to allow water to infiltrate with a subsequent increase in surface runoff and reduction in storage. However, although it would appear that some recovery has taken place on the basis of yearly water yields, it is necessary to check the characteristics of the study years for compatibility. For example, 1955 has a runoff ratio within 3.2 % of that in 1999. 1999 was the 4th wettest year in the Moor House record (spanning 1952 onwards) with 2327 mm (mean 1946 mm); 1955 was also wet being the 12th wettest. Following Jones and Conway (1997), if particular attention is paid to the rainfall totals for the three winter months (December – February; DJF) and the three summer months (June – August; JJA) further similarities can be identified between the two years. Importantly, the summers of these two water years were dry

Table 4.7 Rainfall and runoff relationships for Burnt Hill, comparison of present data with data from Conway and Millar (1960).

a) Rainfall:Runoff ratio for the 1999 water year

Weir	Ratio, %
Trout Beck	73.8
Burnt Hill Eroded	73.8
Burnt Hill Gripped	57.6
Burnt Hill Total	64.6

b) Rainfall:Runoff ratio for Burnt Hill, 1955-1958 and 1999 water years, ranking based on the Moor House record since 1952; driest = 1, wettest = 47

Water Year	Runoff Ratio, %	Total rainfall, mm (rank in brackets)	Rainfall DJF	Rainfall JJA	Rank of Ratio DJF:JJA (lowest first)
1955	67.8	2116 (36th)	643.1 (27th)	251.7 (5th)	39
1956	74.5	1802 (15th)	560.4 (19th)	602.3 (47th)	7
1957	74.3	1967 (26th)	743.2 (37th)	522.2 (44th)	23
1958	74.1*	1899 (23rd)	660.2 (29th)	402.8 (29th)	28
1999	64.6	2327 (44th)	674.8 (32nd)	295.0 (11th)	35

*data missing DJFM

c) Rainfall:Runoff ratio not including JJA for Burnt Hill

Year/site	Ratio, %
1955 – Burnt Hill	72.3
1956 – Burnt Hill	77
1957 – Burnt Hill	74.5
1958 – Burnt Hill	*
1999 – Burnt Hill	67.6
1999 – E1	77.4
1999 – G1	60.2
1999 – Trout Beck	75.8

*data missing (DJFM)

with 1999 the 11th driest and 1955 the 5th driest. So, although both years are wet which one would expect to increase the rainfall:runoff ratio, it may be that the enhanced evaporation during dry summer months has reduced the ratio in both years. In contrast, the summers of 1956, 1957 and 1958 are all wet with 1956 being the wettest on record.

The ECN water table record showed that soil moisture deficits only exist during summer at the target site (Evans *et al.*, 1999). Burnt Hill is only 300 m from the target site and has a similar elevation, although aspect may result in some differences in the water balance. Burnt Hill faces a northerly direction whereas the ECN target slope is orientated with a more easterly or south easterly aspect. Soil moisture deficits are therefore less likely to occur on Burnt Hill but if the rainfall and runoff records are examined for all months except those of June, July and August, this makes comparison of the runoff production more valid. Table 4.7c shows that, even though the rainfall during the 1955 and 1999 non-summer periods was very similar, a greater proportion of runoff was produced in 1955 than in 1999. In fact the 1955-58 mean water yield for both subcatchments combined was 72.7 % +/- 3.3 (95%) such that 1999 is statistically different with a ratio of 67.7 %. Hence there is some evidence to suggest that there has been a recovery since the time of burning. This may be due to revegetation of the surface of the peat that has occurred in the decades after severe burning. This would be accompanied by the build up of fresh material on the peat surface above the burnt layer and alteration and penetration of the 'cheesy' layer by vegetation matter. This would allow increased temporary storage of water within the upper peat layers and on the vegetation cover followed by enhanced evapotranspiration. Fires generally only temporarily damage the vegetation of a mire. So long as a root mat is still intact and there is some seed bank, the vegetation re-establishes, mostly in the pattern of the species which resembles the pre-burnt community (Anderson, 1986). If the root mat is killed in an intense fire then erosion can occur. Fires can also produce finely particulate material which settles in the peat pores and reduces permeability; fire can also produce volatile waxes and oils which form an impermeable skin to the peat (Tallis, 1997). The gully floors on Burnt Hill have largely revegetated (Figure 4.25a and b). This revegetation is impressive and contrasts the conditions existing when Bower (1959, 1960, 1962) was describing the extent and nature of the Pennine erosion. Garnett and Adamson (1997) mapped an area of Moor House of which 8 % was eroding blanket peat; areas once eroded but now recolonised by vegetation occupy approximately 10 % of the study site. The revegetated gully floors may be involved with increased

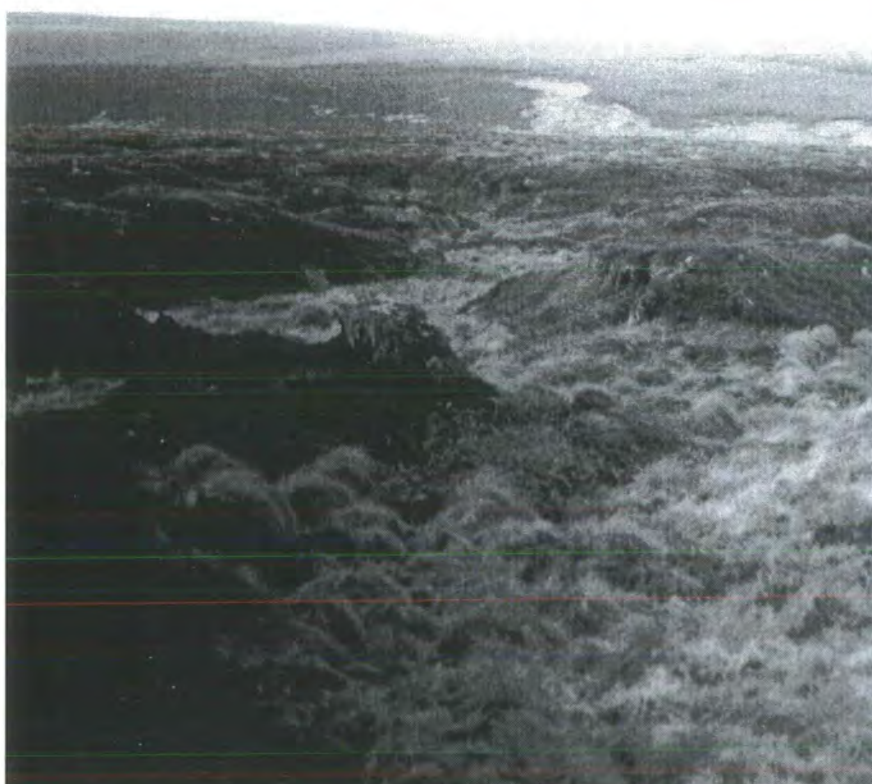


Figure 4.25. Gully erosion on Burnt Hill and the nature of vegetation recovery at the site. Photograph from M. Bower taken in 1958 (upper) compared to one taken in 1998 by J. Warburton from the same point (lower).

temporary water storage, slowing the delivery of water to the catchment outlet and producing the longer hydrograph recession limbs than is the case in the gripped catchment.

Examination of many of the grips on Burnt Hill show that revegetation with *Sphagnum* is taking place on the grip floors and many of them appear to have recovered substantially since cutting. This is occurring particularly on the grips that run across the slope rather than down slope. Mayfield and Pearson (1972) have observed rapid 'healing' of ditches in peatlands where grips are not well maintained. Furthermore many of the grips are not perfectly cut so as to allow smooth and easy passage of water down them. Indeed much standing water lies visible on the surface of the grips trapped in depressions on the grip floor. This may then encourage greater evaporative loss and the greater amount of water storage seen within this catchment. Gunn and Walker (2000) showed that grip blocking by straw bails could reduce flood peaks and bring runoff production more into line with adjacent undisturbed intact slopes. So it seems that revegetation is likely to be the key factor which would result in changes in the hydrology of Burnt Hill since the study of Conway and Millar (1960).

4.3.4. Plot-scale runoff response within the Burnt Hill subcatchments

Robinson (1985) criticises Conway and Millar (1960) because they made little mention of the fact that the catchments they examined differed in other respects than their drainage state. Modelling work by Robinson (1985) shows that topographical differences could not account for the observed differences in flows from the catchments. One important difference between E1 and G1 is that there exists a series of bog pools upslope of the gullies on E1. Indeed Bower (1959) suggests that the end product of these pools is the development of the gully network downstream and its headward extension. Figure 4.26a shows a typical bog pool at the site (plot E2) on the headward side of a gully. At the gully-head nick point, 9 m downslope of the pool shown, a subsurface pipe outlet was found (Figure 4.26b). Subsurface pipes were also found to contribute to some of the other gullies on Burnt Hill. Piping has been assigned as one of the potential mechanisms responsible for gully development in blanket peat (Pearsall, 1950). Tracer examination showed that the pool and the pipe were hydrologically connected such that the pool did not drain yet gully-head flow was maintained. In order to investigate in more detail the runoff processes occurring within the eroded and

a) b)

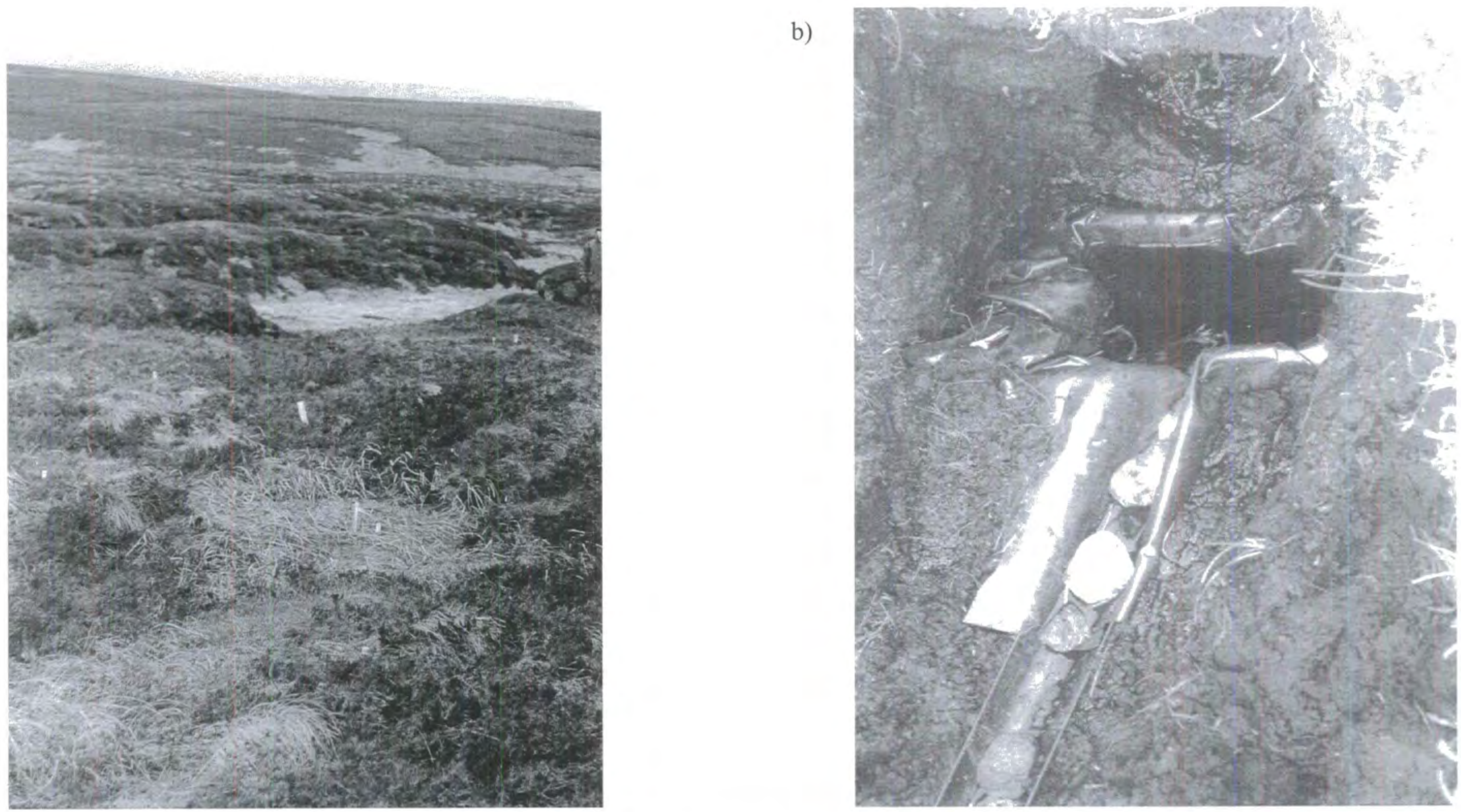


Figure 4.26. a) Pool and gully-head area forming monitored plot E2 on Burnt Hill. b) Pipe outlet and water channelling device at the gully-head which was directed to a large tipping bucket.

gripped Burnt Hill subcatchments, runoff processes were measured at the plot scale at E2 and G2 (see Figure 4.1).

4.3.4.1. Plot runoff processes within the eroded subcatchment

Flow from the pipe monitored at E2 is shown in Figure 4.27e for a 10 day period. Runoff from Trout Beck, E1, and G1 are also shown. For Trout Beck (Figure 4.27b) the steep, almost symmetrical hydrograph with no significant delayed flow indicates dominance of stormflow in catchment runoff (Burt, 1996). Clearly the catchment responds rapidly to rainfall. The eroded hillslope produces higher peak discharges, but has a much smoother and broader hydrograph form than produced by G1 (Figure 4.27c and d). The gripped hydrographs are spikier than the eroded response. Flow from the pool-pipe source at the gully head is perennial and during the monitoring period of 4/8/98 to 16/12/98 only ceased operation during cold periods when freezing affected flow. Peak flows were large given the size of the outlet, often of the order of 3 to 4 l min^{-1}. Nevertheless the hydrograph intensities were low (see Table 4.11) with a very broad shape. It may be that extended pool drainage at gully-heads via overland flow, seepage at depth and pipe networks contribute to the slightly broader shape of the E1 hydrographs when compared to the relatively pool-free gripped catchment. It may be that pool-hummock complexes which are common in blanket peat catchments (Tallis and Livett, 1994; Tallis, 1994) do provide some baseflow to streams, albeit very limited; the data suggests that pipeflow is not dominant in storm response. The pools themselves may simply be a result of the varying growth habits of the different bog plants. This can then be accentuated into the typical hummock and hollow topography (Johnson, 1957; Moore and Bellamy, 1974). The pools may be capable of temporarily storing water and then releasing it at a slow rate through the upper few centimetres of the peat, perhaps connected through macropore networks. If the pools overflow this may produce conditions conducive to pool connection, and eventual gully development. The greater incision on E1 may also allow more drainage whereas the grips favour more evaporation.

Flow from the sides of the gully adjacent to the gully-head at E2 was recorded from a 5 cm deep trough and a 50 cm trough (collecting flow from 30-50 cm depth). No flow was recorded from runoff troughs positioned at 1, 10 or 20 cm depth into the peat face. An example of flow response at the gully bank is shown in Figure 4.28. Flow from the upper layer of blanket peat is flashy, with spiky, ephemeral hydrographs of high

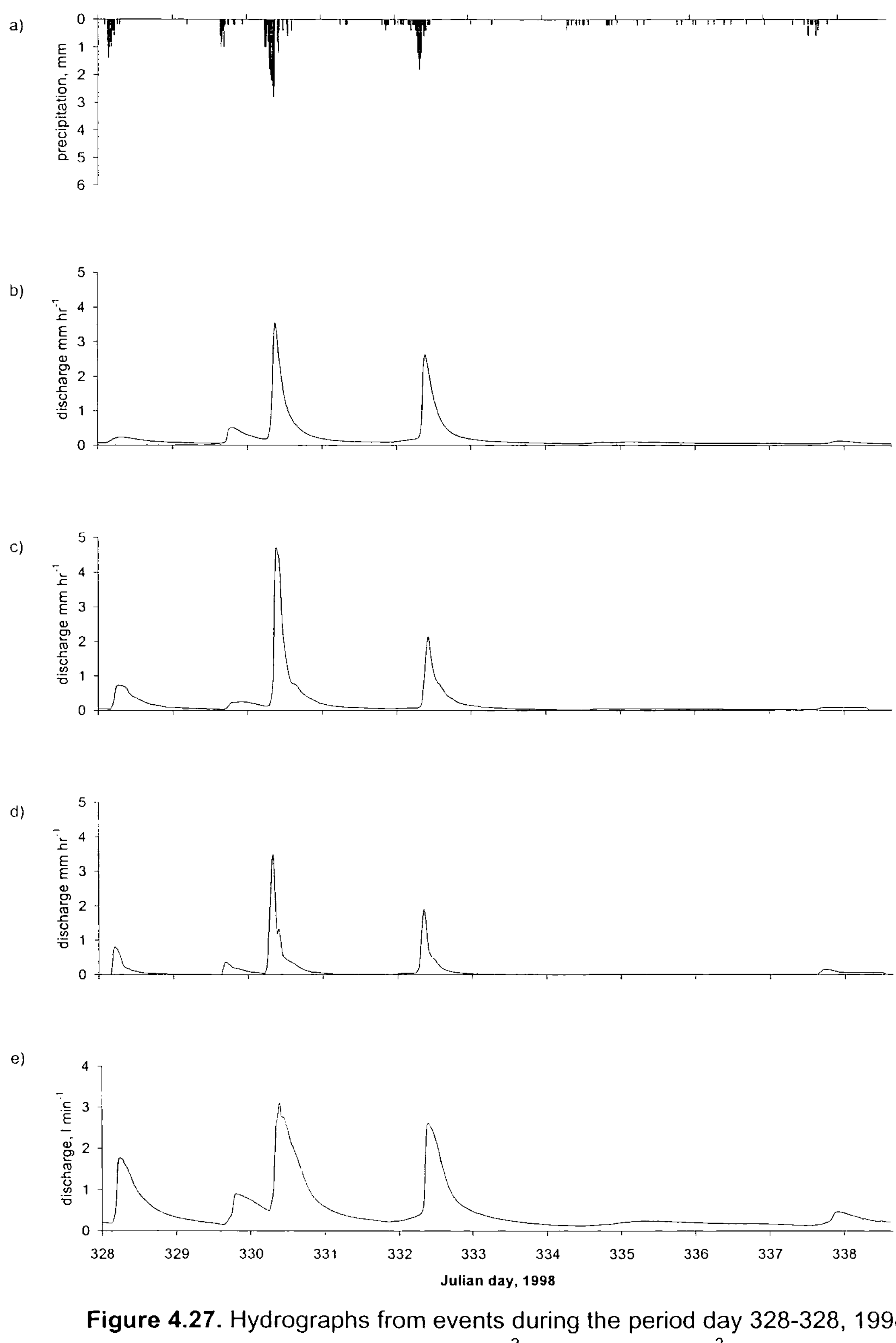


Figure 4.27. Hydrographs from events during the period day 328-328, 1998, a) precipitation, b) Trout Beck (11.4 km^2), c) E1 (0.019 km^2), d) G1 (0.019 km^2), e) pipe at E2 - note the catchment area has not been determined for the pipe such that discharge is measured in l min^{-1} rather than mm hr^{-1}.

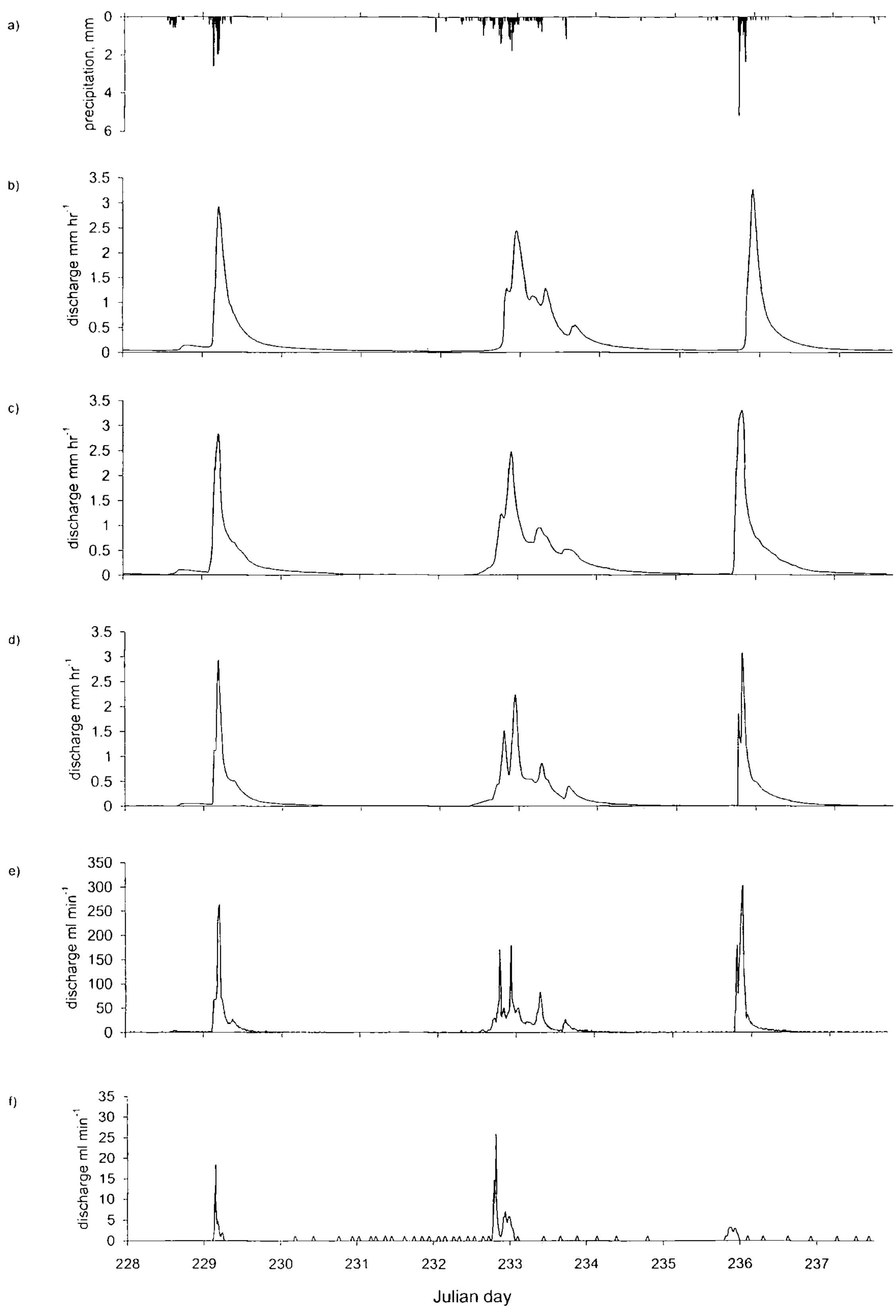


Figure 4.28. Hydrographs from events during period day 228-237, 1998, a) precipitation, b) Trout Beck, c) E1, d) G1, e) 5 cm E2, f) 50 cm E2. b,c and d area weighted, e and f are trough discharge ml min^{-1}

intensity. Runoff appears extremely responsive to rainfall. Discharge volumes are much greater from the 5 cm layer than at 30-50 cm depth. Flow at this gully face was the only monitored location (apart from selected seepage foci and pipes – see below) where flow was recorded deep within the peat. This flow may be linked to the prolonged hydrograph response of the eroded subcatchment over that from the gripped subcatchment where drains are only 50 cm deep. Nevertheless flow at depth from the gully face is ephemeral (Figure 4.28f) such that it is not solely the result of slow drainage of the peat mass. This layer appears to be very well connected to the surface to allow such rapid response to rainfall. It is likely therefore that the peat has different hydrological properties at depth on the gully face to that of uneroded areas (cf. results from other plots below). This is likely to be related to water table drawdown and the desiccation of the peat faces which are exposed to summer baking and winter needle ice formation. This may alter the structure of the peat, which may lead to cracking, macropore network connection with resultant increases in bulk hydraulic conductivity.

4.3.4.2 Plot runoff processes within the gripped subcatchment

Flow response from the gripped plot (G2) is very different in form (Figure 4.29) from that on the gully face. Discharge at G2 only occurred in the top 10 cm of the peat; OLF volumes dominated the response. There was around 10 times more surface flow than flow within the upper 10 cm of the peat. It is striking how similar the shape of the OLF response from G2 is to that of the hydrographs from E1 and G1. Hydrograph form at 5 cm is very much more 'rounded', with lower intensity hydrographs than OLF. Onset of OLF development appears to be rapid. The recession limbs are also steep.

Conway and Millar (1960) and Robinson (1985) suggest that infiltration-excess OLF dominated on Burnt Hill because of the burning that had occurred prior to monitoring. OLF data shown in Figure 4.29 seems to agree with this hypothesis with timing of OLF coinciding closely with the timing of rainfall. However the time period for saturation-excess OLF development is likely to be short on G1 since the drains are only 15 m apart. If saturation-excess OLF occurred across an intact hillslope, it would flow from the topslope down to the footslope. However, on G1, OLF will run into the nearest drain. Thus, the area downslope of a drain will not be supplied with water from from upslope and therefore return flow processes are less likely to occur and will be of a shorter duration. Evidence from longer undisturbed slopes suggests OLF may last for a

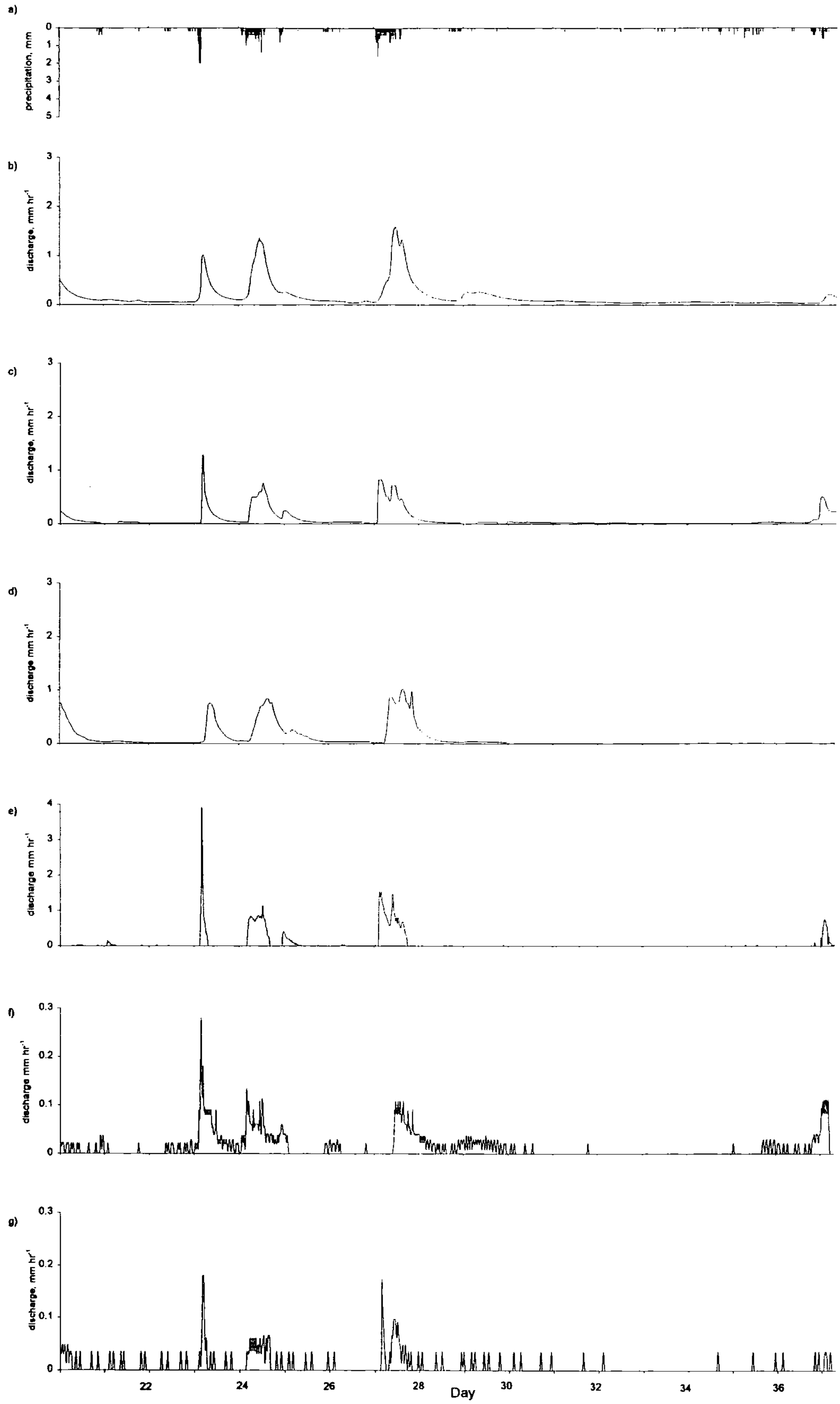


Figure 4.29. Hydrographs from events during the period day 20-39, 1999, a) precipitation, b) Trout Beck, c) G1, d) E1, e) OLF G2, f) 5 cm G2, g) 10 cm G2.

greater length of time because the hillslope takes longer to drain due to saturation-excess mechanisms. These data are discussed below.

4.3.5 Evidence for saturation-excess runoff processes on Burnt Hill

4.3.5.1 Water table and OLF generation at G2

Stewart and Lance (1991) examined the effects of moorland gripping on water tables at Burnt Hill. They found that mean water tables near to drains were lower than at places farther away, but the lowering was slight and confined to a zone a few metres away on either side of the drain. They showed an asymmetrical lowering of the water table around the drains with lowering occurring as far as 2 m downslope but only 1 m upslope. The effects of the drains on vegetation were shown to be confined only to the downslope side. Water table lowering was of the order of a few centimetres even in dipwells immediately adjacent to the drain. Stewart and Lance's measurements were carried out only during 20 days of June and July 1979; nevertheless they did test their results against intact control plots.

A transect at right angles across two adjacent drains on Burnt Hill (see Figure 4.1 for location) was established at the site of the throughflow troughs (G2). Dipwells and crest stage tubes were monitored on an approximately bi-weekly basis (much depended on the weather conditions for accessibility) from June 1998 to May 1999. Monitoring locations are shown on Figure 4.30. The close spacing of these instruments around the grips allows higher resolution data. As Stewart and Lance (1991) found, water tables appear to be only affected close to the drain. One possible error not taken into account by Stewart and Lance (1991) is that when gripping occurs the debris is placed next to the drain and often left. It is clear that on Burnt Hill, the peat dug out of grips was placed downslope and adjacent to the drain. This can be seen on the transect in Figure 4.30 where there are small mounds at sites 7 and 13. This peat is likely to have different properties to that of the intact surrounding peat. This would mean that any measurements of water table close to the drain on the downslope side would be affected by these clumps of peat which are now revegetated. Future investigation of water table drawdown near grips should take into account any deposits from earlier excavation.

Only at sites 7 and 13 did no OLF occur during the study period. This is likely to be related to the combined effect of low supply of water from upslope due to drain interception and the nature of the excavated peat at these two locations. More

site	% of visits with crest stage tube full	site	% of visits with crest stage tube full
1	95	8	13
2	73	9	5
3	25	10	13
4	17	11	50
5	25	12	75
6	32	13	0
7	0	14	25

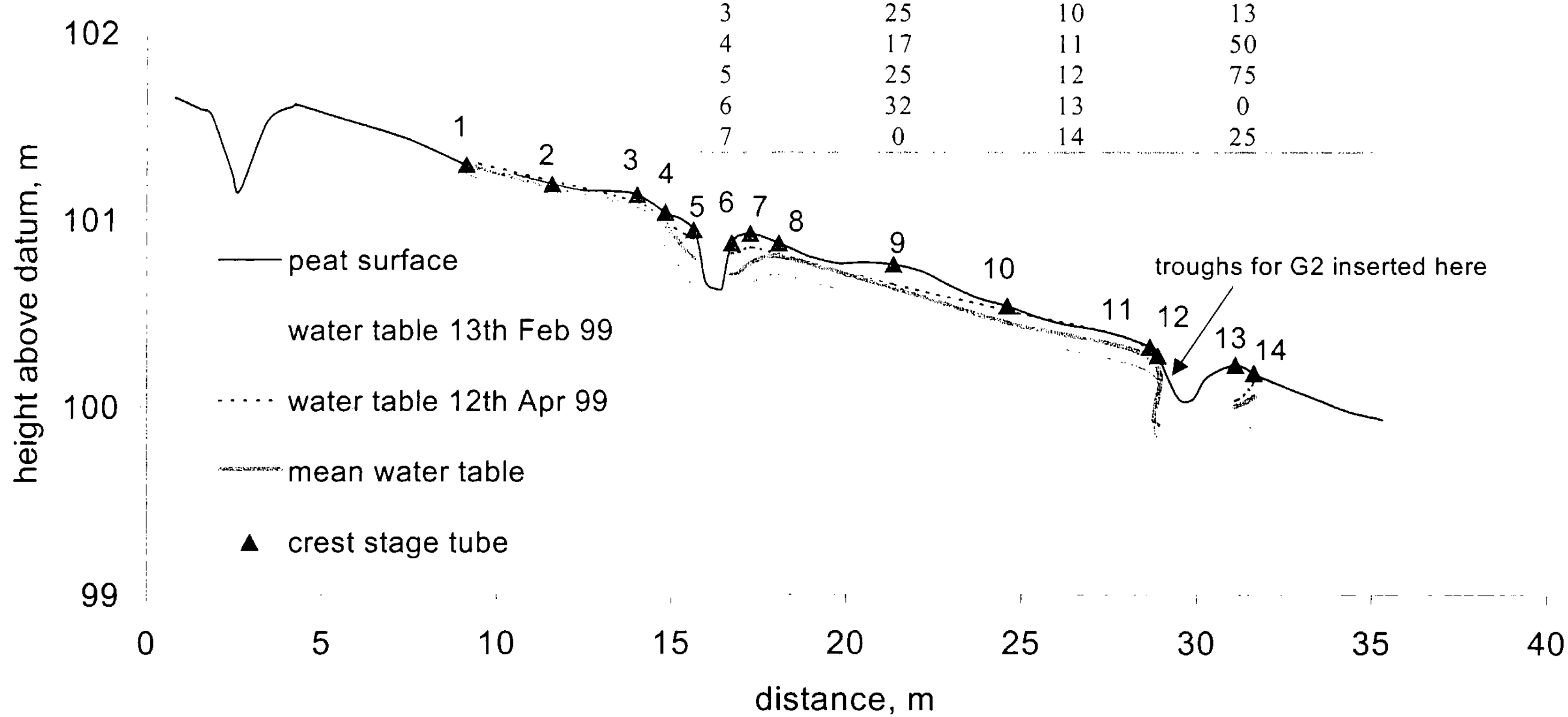


Figure 4.30. Water table and OLF characteristics as monitored bi-weekly June 1998 – May 1999 on Burn Hill gripped transect (see Figure 4.1 for location).

importantly it can be seen that OLF occurrence is spatially much more likely to occur where the water table is found closer to the surface. Indeed at sites 1 and 2 where the mean water table is very close to the surface, OLF was found to have occurred between almost every visit. This provides further evidence that OLF recorded at the monitoring troughs at G2 discussed above was a result of saturation-excess rather than infiltration-excess mechanisms.

Unlike Stewart and Lance (1991) who found that water table variation was greater further from the grip than close to it (during June-July 1979), the results from one year's data show that standard deviations are larger adjacent to the grips. At the trough-recording site the water table clearly drops 10 cm from its edge (site 12). However at site 11, only 20 cm further upslope, there appears to be a limited effect of the drain on the water table, even during drier periods. Hence the insertion of throughflow troughs laterally into the peat face by 30 cm should counteract the effect of the drain on water table decline and hence monitor runoff processes occurring within the peat accurately. No flow occurred from deeper than 10 cm from this site. This suggests that the drains simply act to intercept surface and near-surface runoff, and withdraw water from the peat only alongside the drain edge. Surface runoff is therefore reduced downslope of the drain, and because of the short distances between drains, saturation-excess OLF production is short-lived at the logger site when compared to longer undisturbed slopes (see below).

4.3.5.2 Water table and OLF generation at E2

Mean water table for the monitored plot on E2 demonstrates that water tables may be affected by gullying at a much greater distance than is the case for grips (Figure 4.31a). High water tables surround the pool in the centre of the plot, characteristically with ponding at the surface. Upslope from the pool and gully-head the water table is generally within 5 to 10 cm of the surface. The greatest range in water table occurs nearer steeper sections of the plot, just upslope from the pool and upslope from the steep gully sides (Figure 4.31b). The fact that OLF did not occur in the runoff trough at the edge of E2 on the gully face suggests that a deflated water table resulted in reduced development of saturation-excess OLF and instead water could rapidly infiltrate the surface to runoff below the surface. The proportion of visits between which OLF had occurred over the plot is shown in Figure 4.32. The values tie in well with mean water tables such that areas likely to be more saturated with higher water tables are more likely to allow OLF (or at least a surface ponding) development. Flow appears to be directed into the pool from above and then channelled out of the pool when it overflows into the gully-head nick point. Hence flow here occurs as pipe flow and as OLF. These conditions then are suitable for headward extension. All crest stage tubes on the plot recorded OLF at some stage during the year of monitoring. The only other crest stage tube data from blanket peat, to the author's knowledge, is that presented by Burt and Gardiner (1984); the tubes were found full on 88 % of occasions on an unvegetated peat surface, 88 % on a vegetated flush channel, 47 % on one vegetated hummock and 18 % on another. At three sites on vegetated intact slopes the frequencies were 84 %, 78 % and 78 %. Both Burt and Gardiner's (1984) data and the data from Burnt Hill and from H1 and H2 discussed below, show that OLF generation on blanket peat is spatially distributed, but that OLF does occur over most of the surface at some stage during the year. This is in contrast to the contention of Ingram and Bragg (1984) who suggest that the acrotelm is self maintaining as it does not allow OLF development and is not eroded by the flow.

4.3.6 Matrix throughflow generation on Burnt Hill

Piezometer nests on E2 located at the crest-stage tube sites shown in Figure 4.32 allow pore water pressure and total potential flow nets to be plotted. Figure 4.33 presents these potentials for a transect upslope of the pool (transect indicated on Figure 4.32). Measurements were taken over the upper 80 cm of the peat throughout the year and Figure 4.33 plots the potentials from 24/9/98 as an example. Pore water pressure seems to vary uniformly with depth such that the vertical gradients in total potential are not steep. Slighter steeper vertical gradients can be seen in the upper layers of the monitored zone, indicative of a higher potential vertical flow near the surface. Across the transect the lateral gradient in total potential averages around 0.07. The surface slope averages 0.05.

Pore water pressure varies little throughout the year with a low standard deviation for each piezometer (Table 4.8). Within the peat mass, pore water pressures vary little from wet to dry weather. They do however, change slightly (with mean standard deviations of the order of 3 to 7 cm) which may be important. Heikurainen *et al.* (1964) did present some evidence that water contents below the water table do increase with rising water tables. Indeed this may be related to elasticity or plasticity of the peat and the widely reported phenomenon of the peat mass expanding and contracting such that peat surface

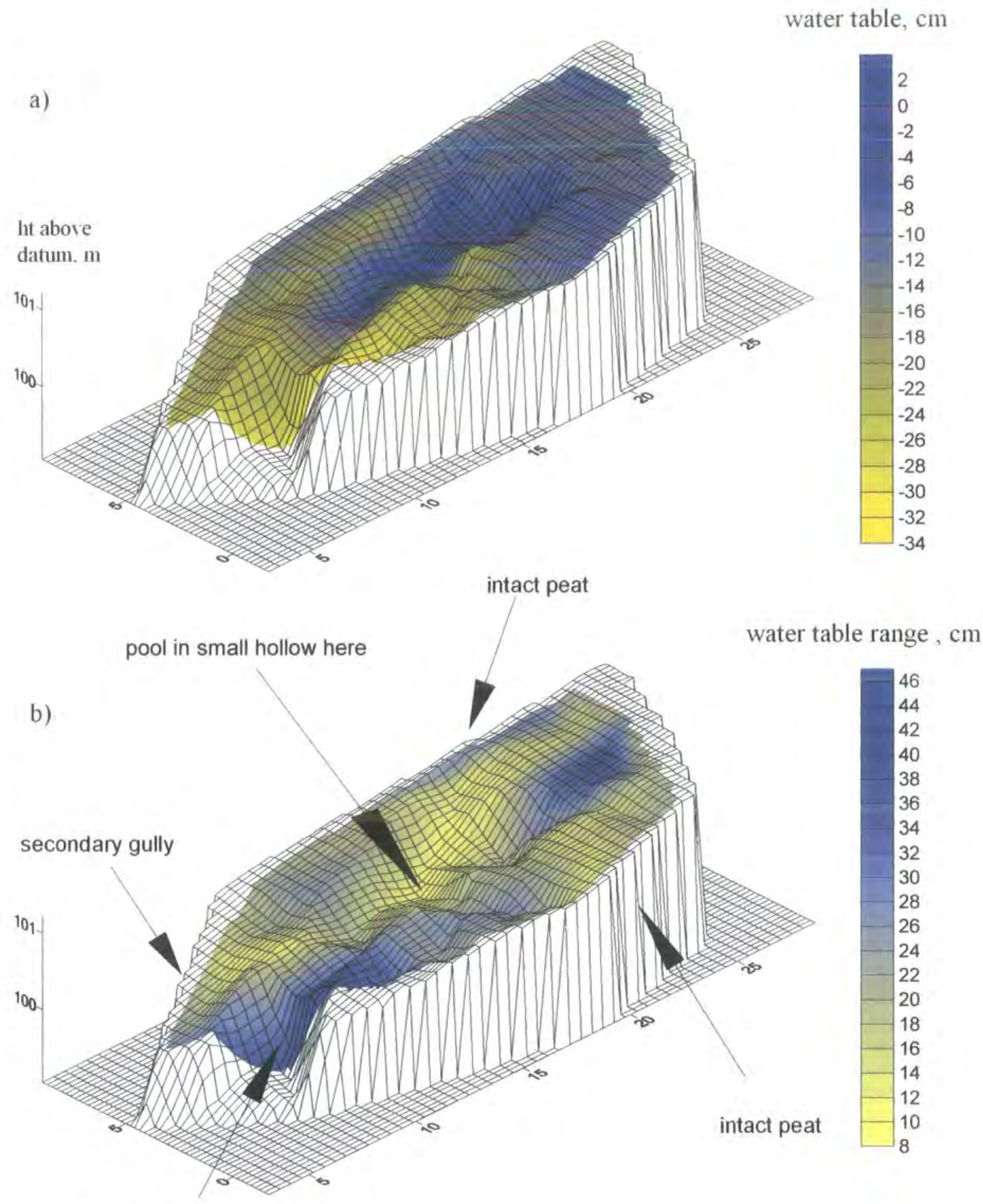


Figure 4.31. Water table at E2 based on bi-weekly sampling, January–November 1998, axis distances in metres a) mean water table depth below surface, b) range in water table depth.

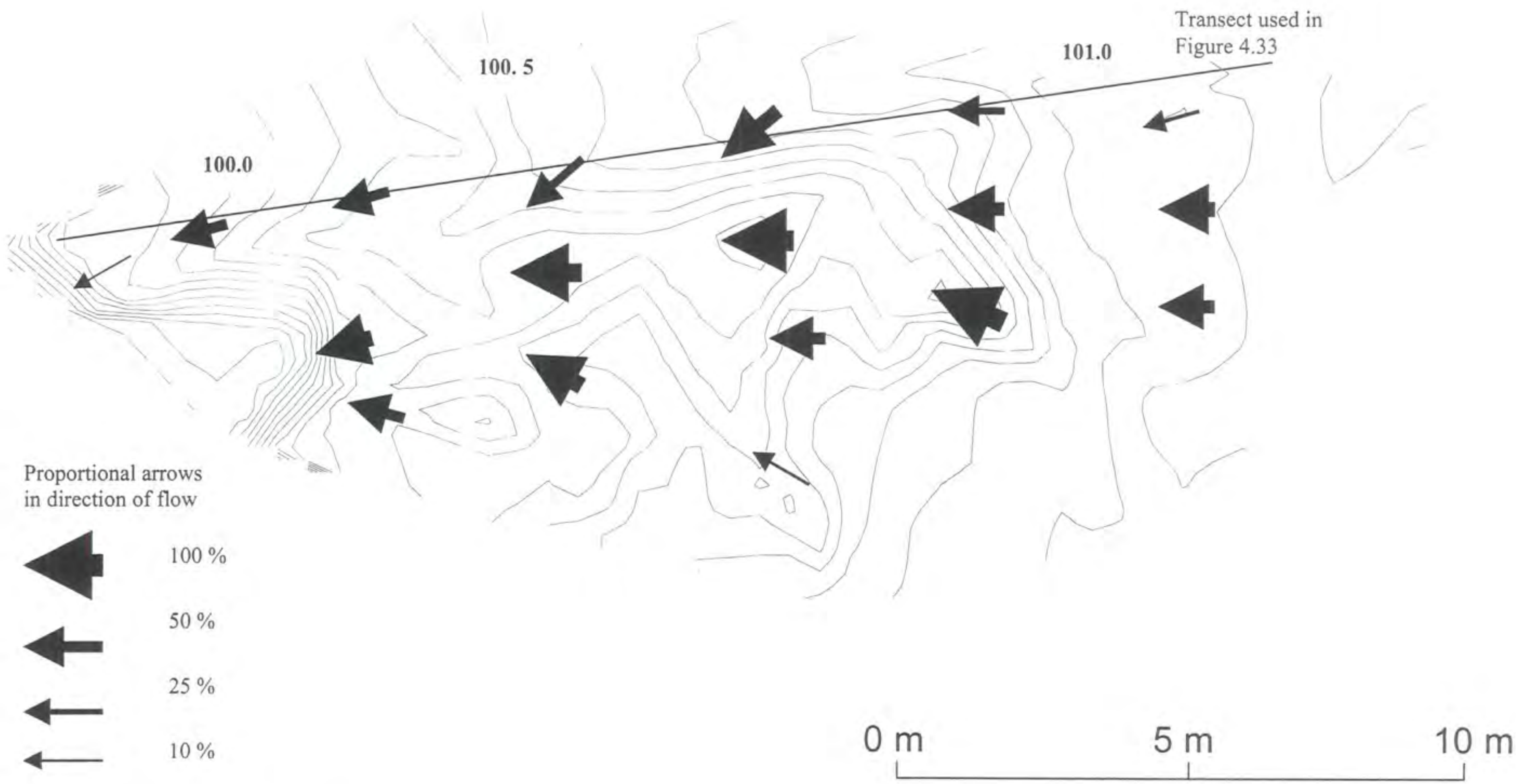


Figure 4.32. Spatial distribution of OLF detected by crest-stage tubes as a proportion of bi-weekly visits when found full on E2, January-November 1998. Snowmelt events ignored.

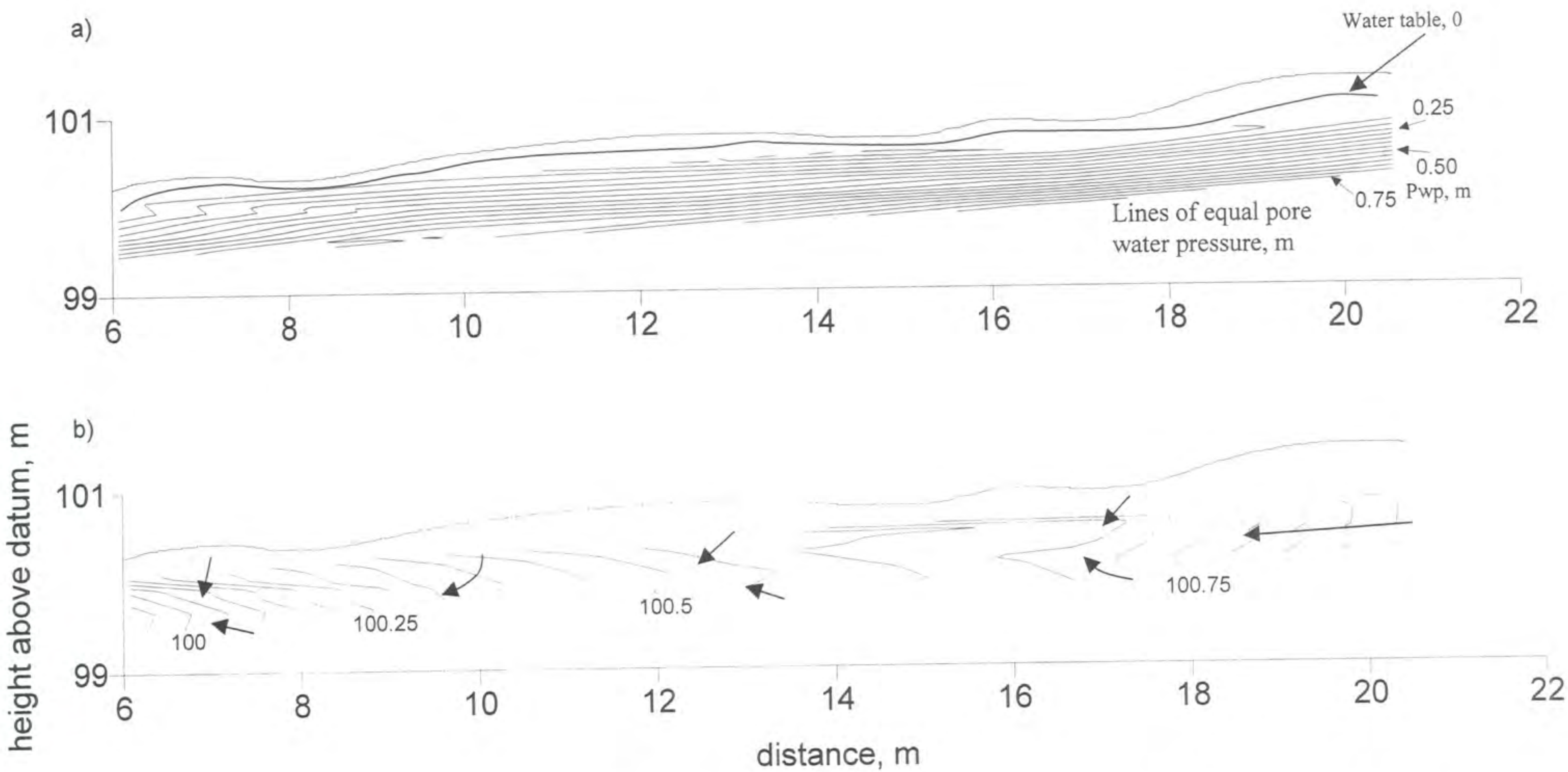


Figure 4.33. Hydraulic conditions across transect 1 on E2 24/9/98, a) equilines of pore water pressure, b) equilines of total potential; flow would be expected to occur perpendicular to the lines of equipotential.

height changes can be clearly measured (the *Mooratmung* discussed earlier). No measurements were taken of peat surface movement at Moor House but it is likely that the effect is negligible given the low variation in pore water pressure throughout the year. Price and Schlotzhauer (1999) provide evidence from a mined peatland that fluctuations in the water table result in water storage changes below the water table due to changes in the overlying weight of water which leads to compressional and hence volumetric change. A reduction of the peat volume by shrinkage or compression entails a decrease in the size of the pores. Consequently, saturation occurs at lower volumetric moisture content. Changes in water storage below the water table is likely to be more important in thicker peat deposits and peat which has been drained or eroding such that water table fluctuations are greater. The compressibility of the peat is also likely to be of great importance (Price and Schlotzhauer, 1999). However, the low variation throughout the year seen on Burnt Hill and at the other monitored sites at Moor House suggests that little extra storage can occur. This may be important for the understanding of peat mass movements (see discussion in Chapter 9).

Table 4.8. Mean and standard deviation of pore water pressure from piezometer nests on E2 January 1998 – November 1998

Depth, m	Mean pwp, m	Mean σ, m*
0.20	0.13	0.03
0.35	0.26	0.04
0.60	0.51	0.05
0.85	0.71	0.04
1.00	0.86	0.07

pwp = pore water pressure
* Calculated values are standard deviations within individual piezometers averaged over all the piezometers at that depth, over the 8 month period of monitoring. This creates a mean standard deviation for each depth category.

3D plots of pore water pressure (Figure 4.34a) indicate greater pressures around the lower-lying surface surrounding the pool. This is presumably a result of ponded water at the peat surface. At 35 cm depth, again the central part of the plot has higher pore water pressures than elsewhere. By 80 cm into the peat there is a more even distribution of pore water pressure in the plot with gully side areas experiencing the lowest pore water pressures. Vertical gradients in potential are very low throughout the plot as there is little change over the four depths shown in Figure 4.34b.

a)

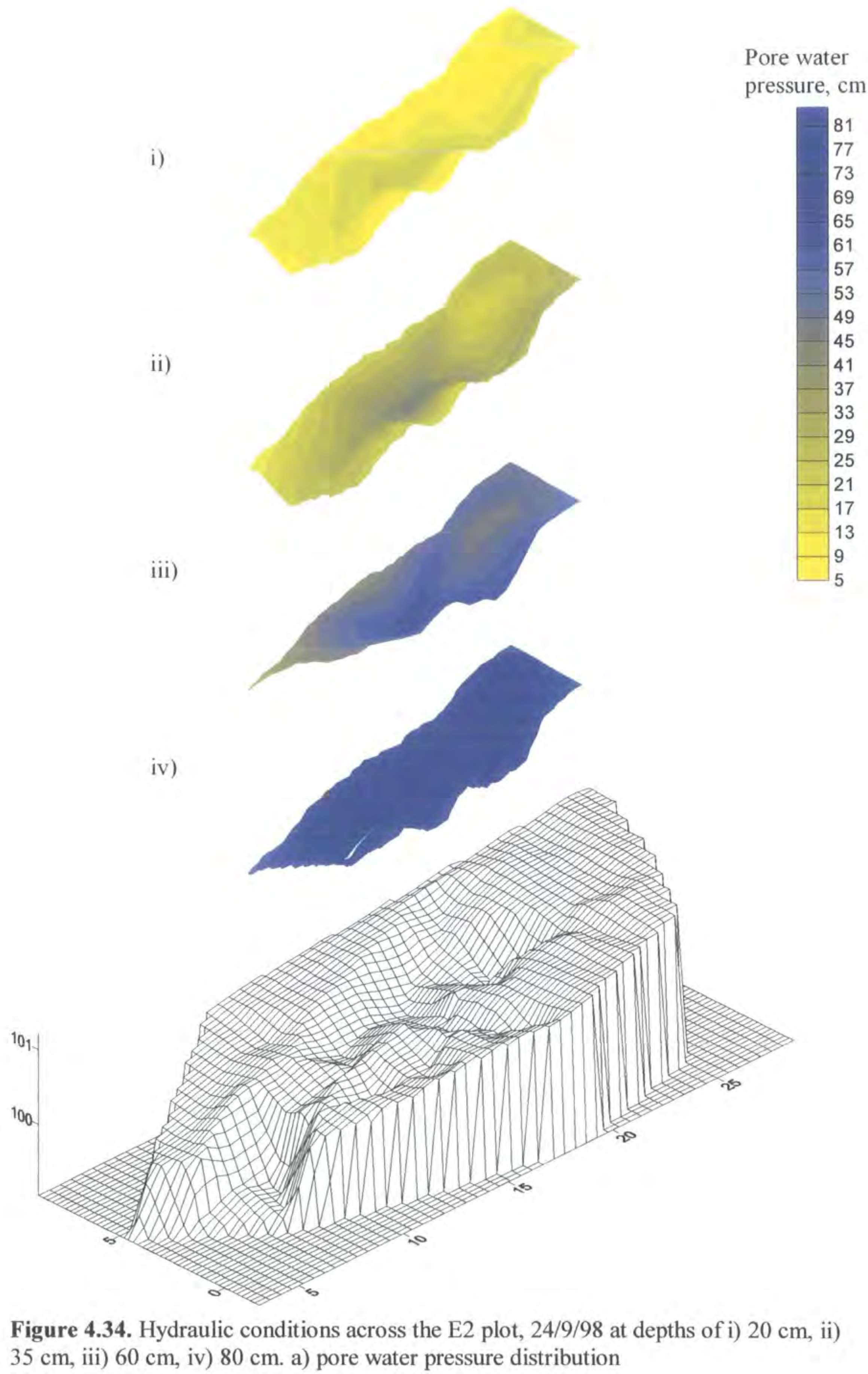


Figure 4.34. Hydraulic conditions across the E2 plot, 24/9/98 at depths of i) 20 cm, ii) 35 cm, iii) 60 cm, iv) 80 cm. a) pore water pressure distribution

b)

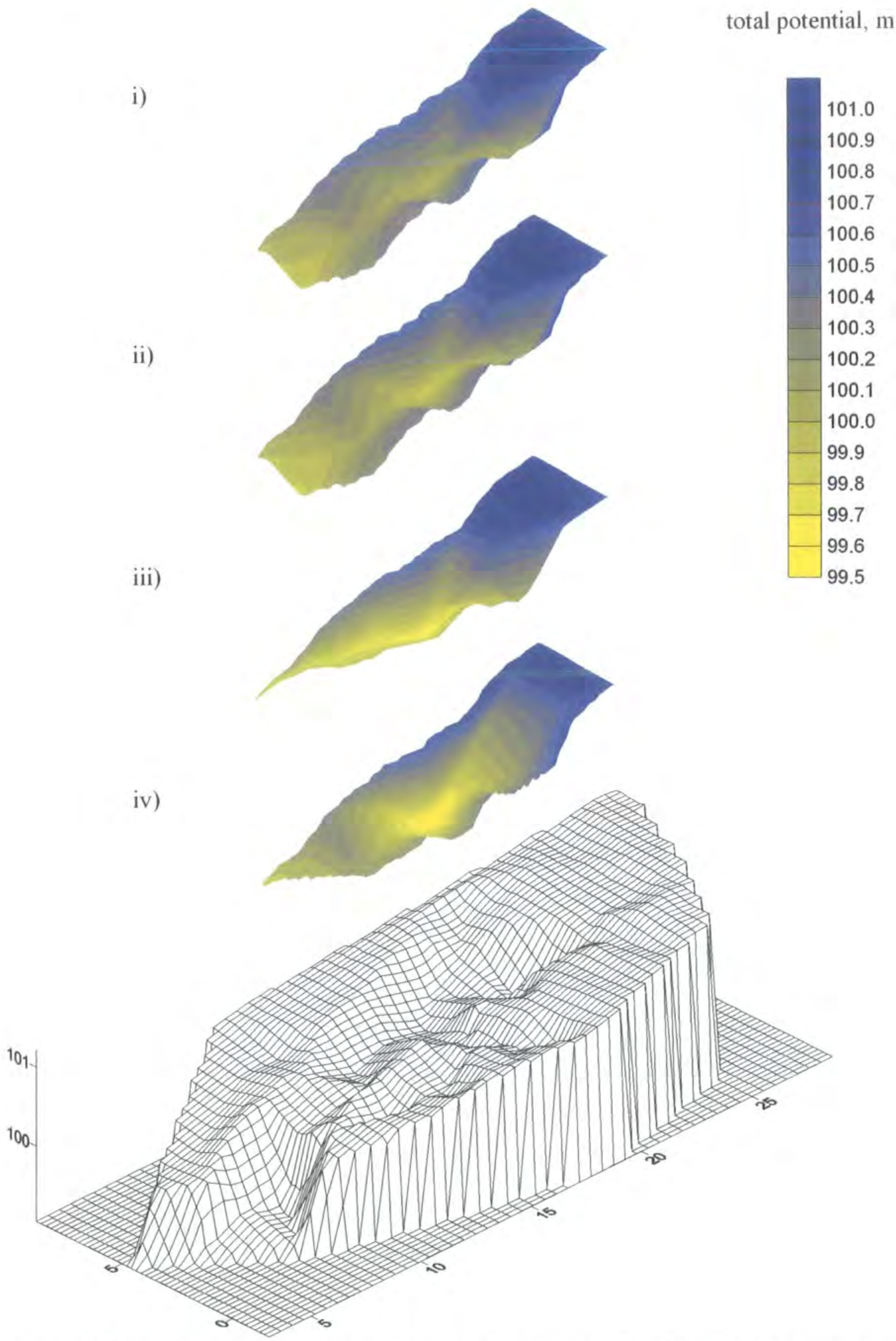


Figure 4.34 continued. Hydraulic conditions across the whole of E2, 24/9/98 at depths of i) 20 cm, ii) 35 cm, iii)60 cm, iv) 80 cm. b) total potential.

4.3.7. OLF and throughflow within the Trout Beck catchment

4.3.7.1 Relative contribution of OLF and throughflow processes

While vertical potential hydraulic gradients are low, lateral gradients do exist. However, flow net analysis does not tell us actually how much flow is occurring in the subsurface peat layers. The flow nets only indicate potential for flow. Table 4.9 gives an indication of the importance of flow process contributions measured by the throughflow troughs from all automated plots installed within the Trout Beck catchment. This provides important quantitative evidence for the relative importance of surface and near-surface flow processes in blanket peat. OLF or at least flow in the upper centimetre of peat, is the most important runoff pathway. Lateral flow at depths greater than 5 or 10 cm is restricted such that runoff contribution from these layers is low. Hence the low hydraulic conductivities found at relatively shallow depths within the blanket peat as determined by slug withdrawal tests (see above) result in minimal flow contributions from most of the peat mass. In this way there is strong evidence against the idea of Baird *et al.* (1997) that because of the thickness of the catotelm it may contribute significantly to flow even if it has a low hydraulic conductivity. The results from plot process measurement suggest that less than 1 % of runoff in blanket peat catchments is generated from the peat below 5 cm depth.

Table 4.9. Mean proportion of flow from 6 automated runoff collecting sites, with depth. H1 and E2 not included due to non-compatible measurement technique.

Source	Mean and (σ) proportion of total discharge collected at the throughflow trough site, %
Surface	81.47 (18.61)
1-5 cm	17.76 (18.64)
5-10 cm	0.74 (1.77)
10-50 cm	0.03 (0.07)

4.3.7.2. OLF and throughflow runoff characteristics

a) General characteristics

A summary of the hydrograph characteristics from monitored runoff processes is shown in Table 4.10. The trends from Burnt Hill have already been discussed. It is clear that lag times are very short in blanket-peat catchments with mean values for Trout Beck of 2.6 hours and most plot scale responses under 2.2 hours. H1 appears to respond very rapidly to rainfall with mean lag times under an hour except at 5 cm depth on the footslope section. Runoff from deeper peat layers (5 and 10 cm), and from footslopes, generally exhibits longer recessions and lower hydrograph intensities.

Table 4.10. Means and standard deviations for storm response characteristics of catchment-, hillslope- and plot-scale runoff measurement.

Source	Mean Peak Lag, hrs	Mean T_{rec}, hrs	Intensity, s^{-1}	n
Trout Beck	2.6	28.9	38.8	58
	3.1	12.6	27.9	
Gripped hillslope	1.7	31.1	48.6	44
	2.2	17.3	27.7	
Eroded hillslope	2.1	31.9	44.4	43
	2.4	15.5	47.0	
Gully-head flow	2.4	33.1	22.4	12
on eroded hillslope	2.6	15.0	7.1	
Plot surface flow	2.0	10.6	85.3	32
	3.5	21.9	73.0	
Plot 5 cm flow	1.3	31.4	76.3	38
	2.4	25.3	62.1	
Plot 10 cm flow	2.2	14.1	54.5	33
	2.4	14.0	60.1	
Seepage foci	1.6	21.7	41.9	31
	2.2	17.0	57.2	
H1 OLF	0.4	18.4	152.7	14
	0.3	10.9	93.1	
H1 midlsope OLF	0.7	7.6	283.0	14
	0.6	7.5	223.8	
H1 midslope 5 cm	0.6	14.2	204.4	14
flow	0.5	7.3	145.0	
H1 footslope OLF	0.9	15.6	149.8	14
	0.7	15.8	100.7	
H1 footslope 5 cm	1.3	57.3	122.7	14
flow	1.2	11.0	82.3	

Peak Lag = Time between peak rainfall and peak discharge, T_{rec} = Time from rainfall cessation to return to pre-storm level, Intensity, s^{-1} = peak flow ($m^3 s^{-1}$) divided by total storm discharge ($m^3 \times 10^{-6}$).

b) The response of H1

The general characteristics of the hydrographs from H1 seen in Table 4.10 are illustrated by the examples in Figure 4.35. OLF on the footslope site is more prolonged than on upslope sites. As the hillslope drains, return flow is produced on gentler slopes producing saturation-excess OLF on the footslopes for a longer period than seen on steeper hillslope sections or at the crest of the hill. When OLF has ceased on the footslope, the flow record from the 5 cm trough indicates that the near-surface layers of the peat continue to drain. Runoff from 5 cm depth tends to be more prolonged than at the surface with much more rounded and less peaky hydrograph forms. This is indicative of a limited flow capacity below the surface of blanket peat and of the dominance of saturation-excess runoff generation. Flow from the surface and 5 cm troughs is clearly ephemeral. Given the minimal contribution of flow from deeper layers in the peat this indicates that peatlands release their gravitationally-free water, rapidly following rainfall.

Examination of cumulative discharge over the example period (Figure 4.36) shows the discharge curve from the eroded Burnt Hill subcatchment overlaps that of Trout Beck at around 0600 day 238 reflecting the slower drainage of bog pools via the revegetated gully floors. For the gripped hillslope, cumulative discharge is similar to that from the topslope and midslope OLF troughs on H1. This again highlights that ditching merely intercepts surface and near-surface runoff and does not result in great increases in water removal from the catchment. Runoff production from the upper layers of H1 is concentrated within a short space of time. Most OLF production on the topslope of H1 occurs in a shorter space of time than for the midslope or the toe of the hillslope. OLF also occurs over a much shorter period of time than flow within the lower layers of peat. Over the short distance of 45 m between the topslope plot and the footslope plot, OLF on the footslope plot ceases 7.5 hours after cessation on the topslope plot.

Figure 4.37 shows some dynamics of runoff production processes on H1 at different stages of the flow recession. Here OLF (or at least surface ponding) was recorded by crest stage tubes over almost the entire hillslope at the peak of the storm at 0300, day 239 (Figure 4.37a). Small-scale microtopographical differences could be found on the hillslope but the measurement network allows the general hillslope runoff production to be displayed. As the hillslope drains following rainfall cessation, the more gently sloping top and footslope regions continue to produce OLF with the steeper slopes

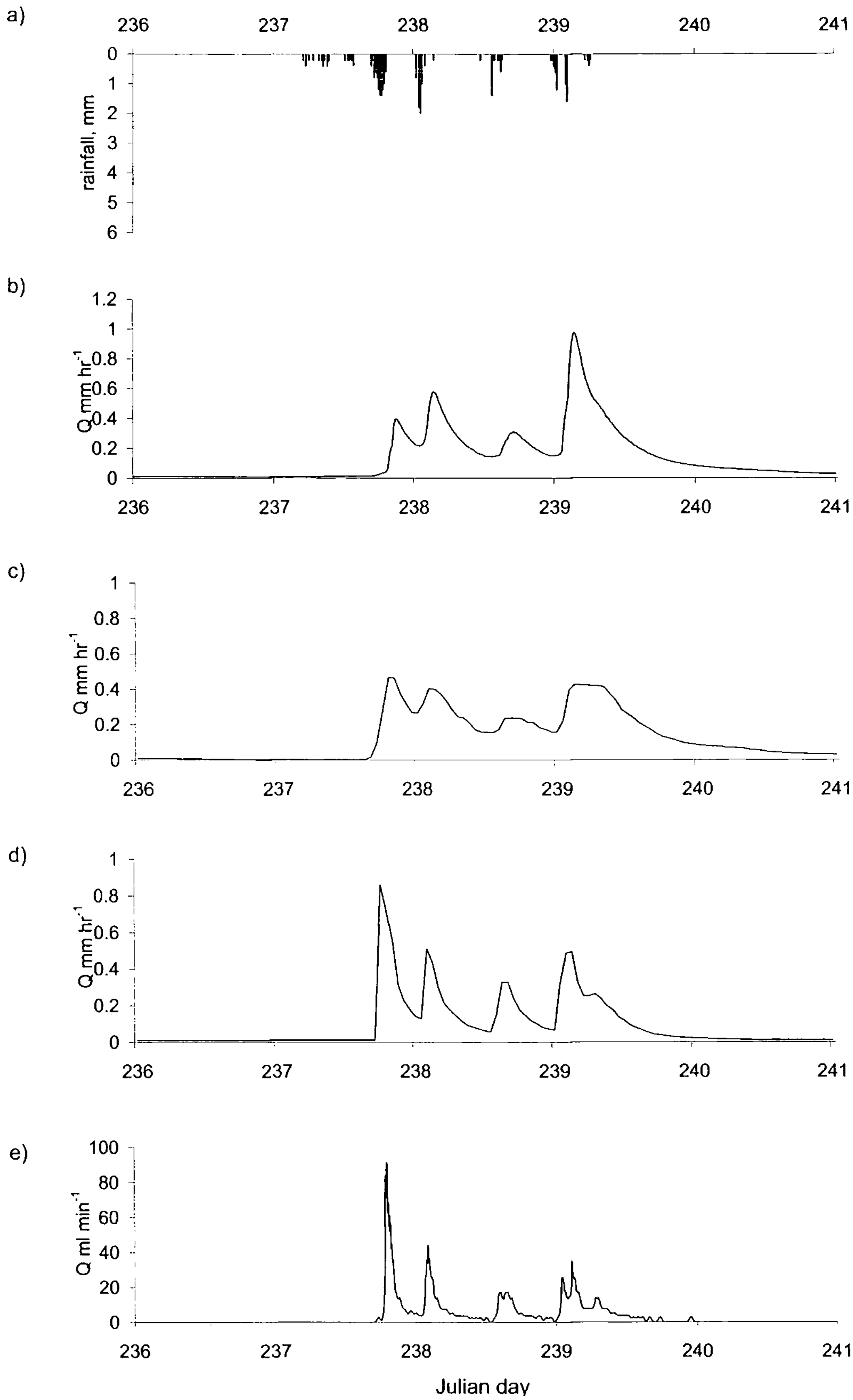


Figure 4.35. Runoff production from monitored sites, days 236-241, 1999, a) precipitation, b) Trout Beck, c) E1, d) G1, e) H1 topslope OLF, f) H1 midslope OLF, g) H1 footslope OLF, h) H1 midslope 5 cm, i) H1 footslope 5 cm.

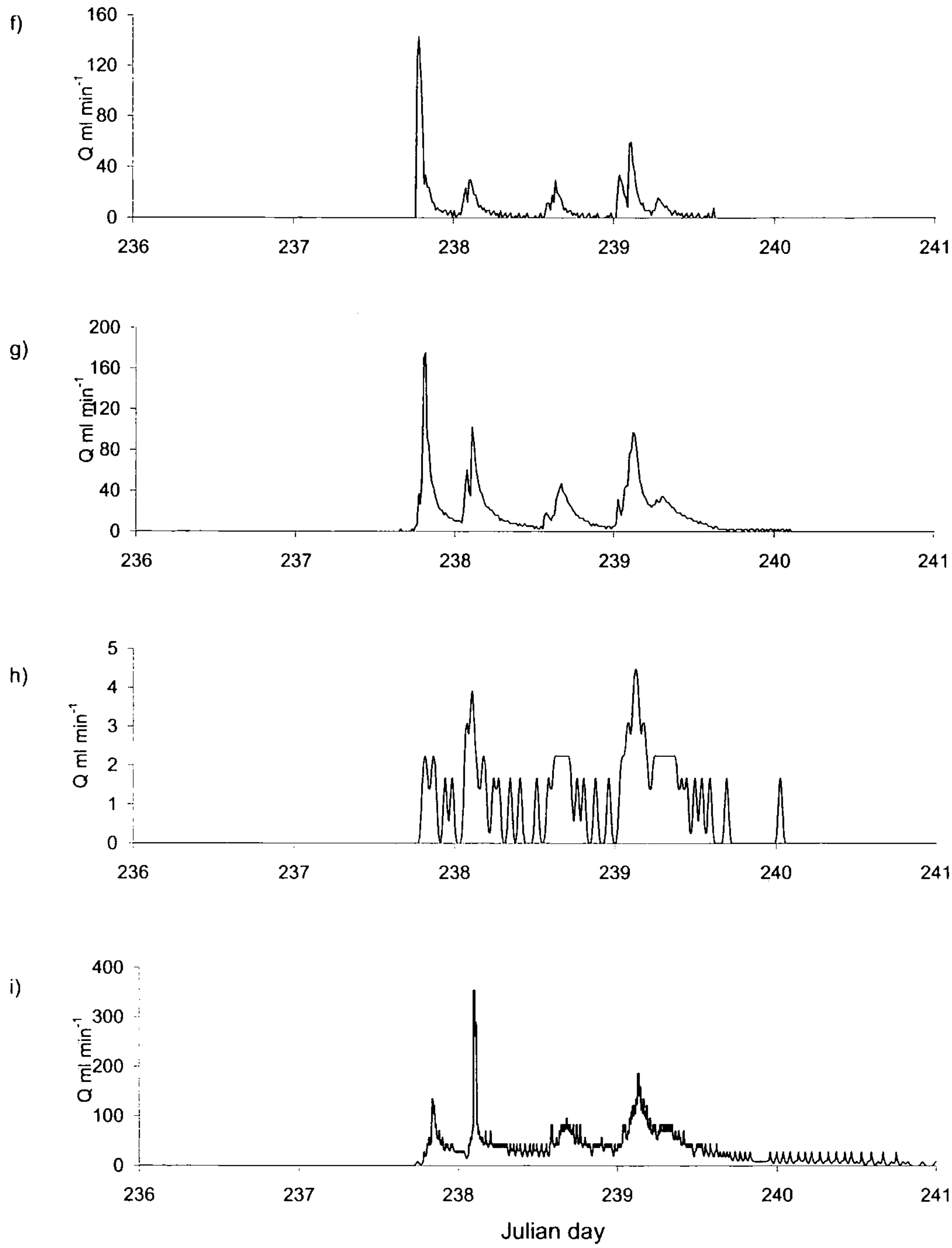


Figure 4.35. continued

producing flow just below the surface at 3 cm (Figure 4.37b). By 1300 (Figure 4.37c) the saturated wedge only exists on the hillslope toe regions whereas steeper areas drain to produce flow down to depths of 6 cm and occasionally 9 cm. After 0900 day 240 (Figure 4.37d) there is only very slow change. Drainage of free water available in the upper soil layers of H1 is rapid such that within 30 hours the hillslope has reached a quasi-equilibrium state with water tables stabilised. Runoff from almost the whole hillslope becomes minimal. The only fully saturated area is on the right flank of the hillslope where monitoring has indicated that the peat is almost permanently waterlogged due to poor drainage. Thus, topography is important for determining dominant runoff process contributions even on low-gradient peat. The steeper midslope sections of H1 produce OLF less frequently than shallower slopes. This suggests that the midslope sections produce more runoff below the surface which collects at the bottom of the slope, and due to impeded drainage manifests itself as return flow.

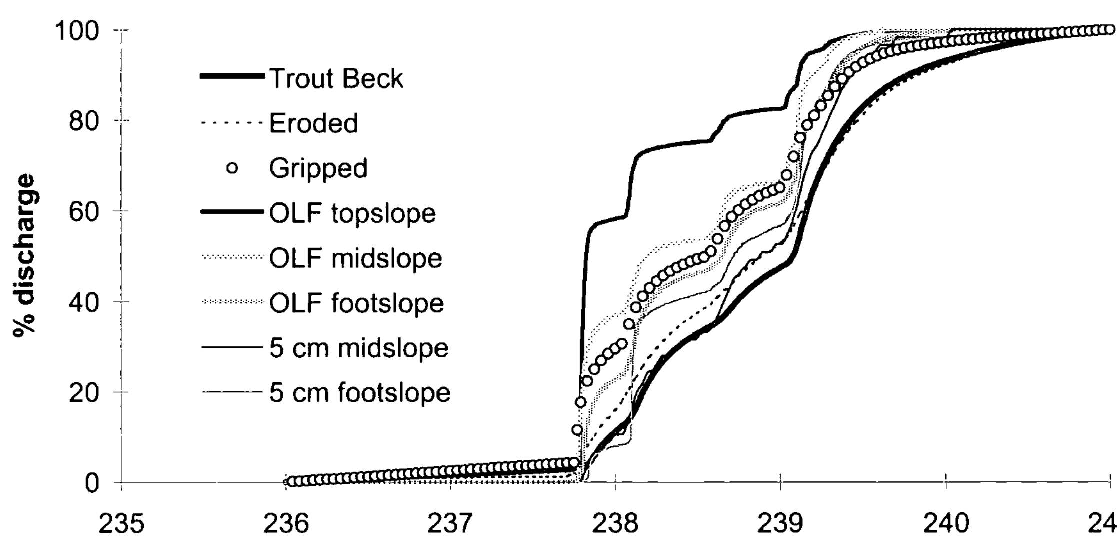


Figure 4.36. Cumulative discharge curves for day 236-241, 1999, for monitored runoff sites.

Mean water table levels are shown for H1 (Figure 4.38a). As on E2 (discussed above), the pattern follows closely the crest stage tube results clearly demonstrating the role of high water tables in producing surface flow. Mean water tables appear to be higher on shallower slopes, notably at the crest of the hillslope and at shallower hill toe locations; the same areas also have the lowest range in water tables (Figure 4.38b). These patterns are clearly important for determining the spatial and temporal production of runoff and for peatland ecology.

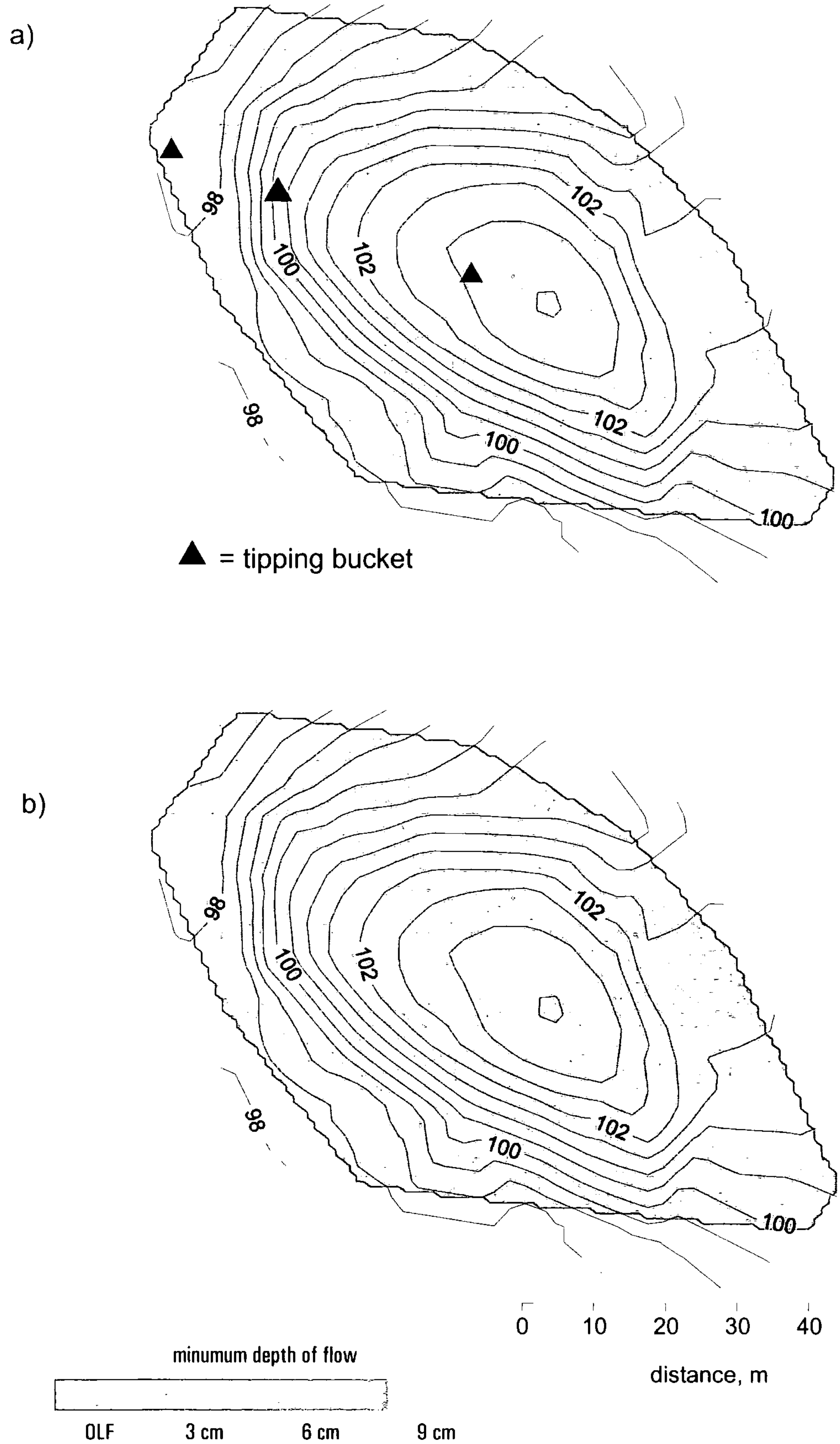


Figure 4.37. Minimum depth of runoff from the peat surface on H1, day 239-240, 1999 as monitored by crest-stage tubes, a) 0300 day 239, b) 0900 day 239, c) 2100 day 239, d) 0900 day 240.

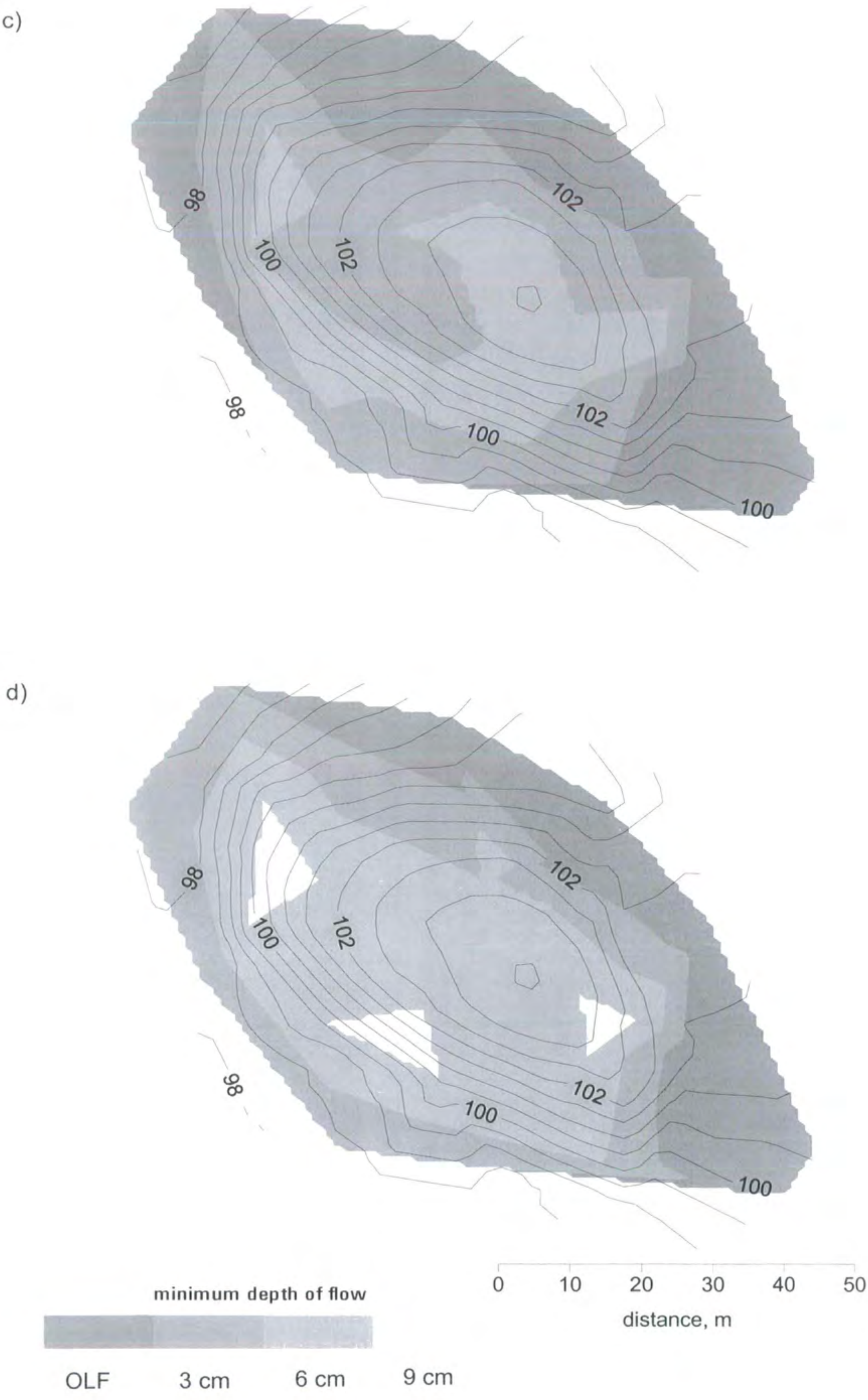
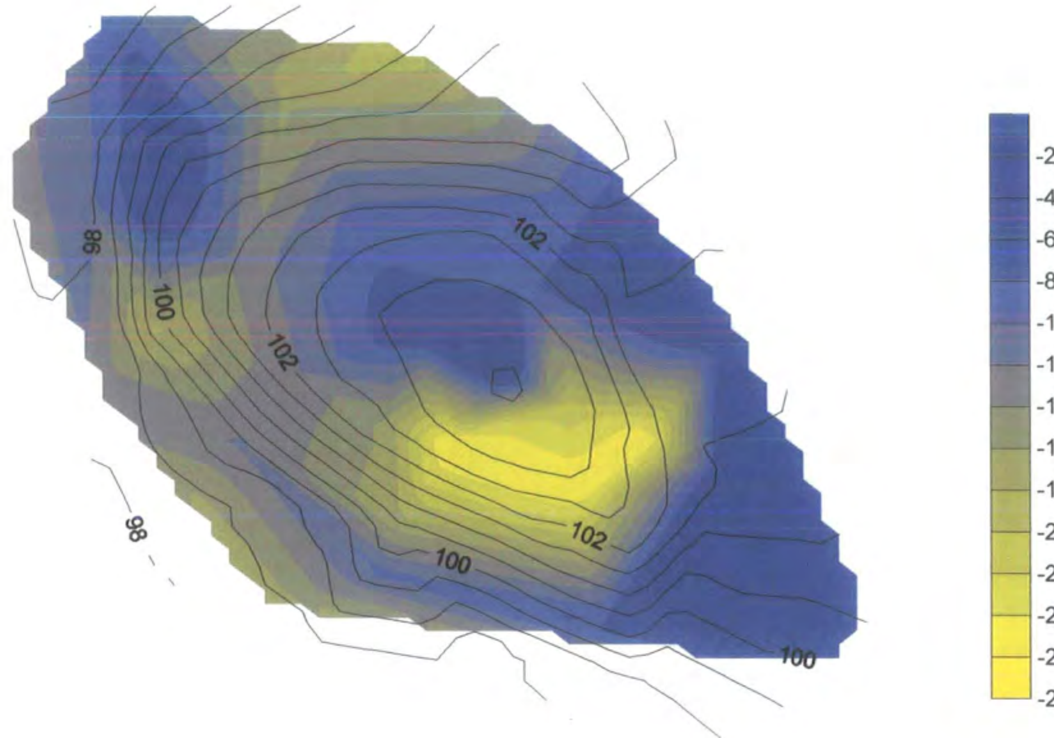


Figure 4.37. continued.

a)



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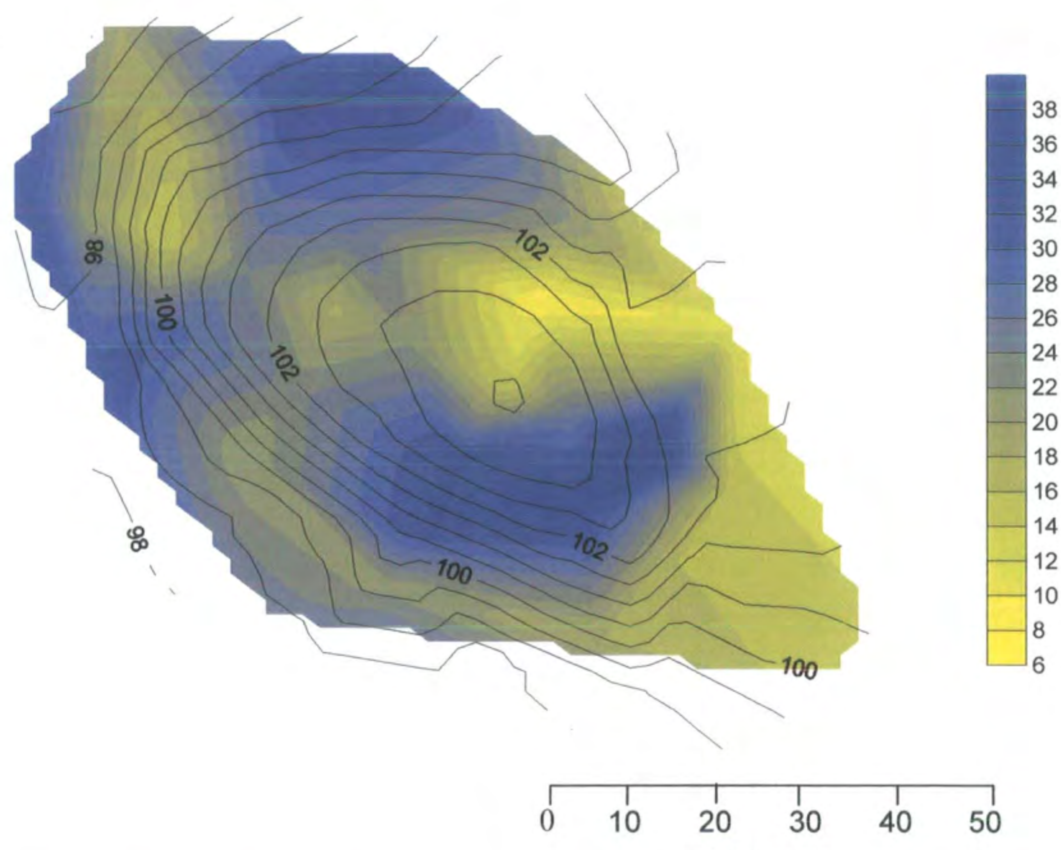


Figure 4.38. Water table characteristics on H1, a) mean water table depth, cm, b) water table range, cm. Bi-weekly sampling, May 1999 – November 1999.

c) The response of H2

Runoff response from the foot of a 230 m slope (H2) is shown in Figure 4.39. Some of the problems with datalogger resolution at low flows are indicated by Figure 4.39e. At very low discharges it takes a long time for enough discharge to fill and trigger the tipping bucket. This therefore causes problems in determining timings of response. However these data still provide valuable information on relative proportions of runoff from the peat mass. It is clear that surface flow is a large proportion of runoff. Small amounts of rainfall produce long-lasting OLF on the toe. This again shows that infiltration-excess is not likely to be the main surface flow-producing mechanism in blanket peat. The flow production seems to be longer than that found on shorter slopes (c.f. G2 response). It is still striking that H2 drains rapidly such that runoff production from the foot of the 230 m hillslope can cease within 48-72 hours of rainfall cessation. The typical acrotelmic response (5 cm) is rounded, with longer recessions than OLF. Simply, hillslope drainage produces prolonged saturation from the foot of the hillslope upwards. As this saturated wedge moves downslope, eventually OLF ceases but the acrotelm continues to drain. Crest stage tube mapping demonstrates how hillslope saturation changes over time during the rainfall event of day 282-283 1998 (Figure 4.40). Much of the monitored section of the slope produces OLF during the main part of the rainfall event. Figure 4.41 shows that water tables are maintained to a much higher level along the entire slope profile than that for the eroding plot seen on Burnt Hill such that water table rise to the surface is more easily achieved. Mean water tables are lower at site 1 (and standard deviations higher) than uplsope presumably because the site is so close to the footslope ditch that bank side drainage is occurring. As the hillslope drains following rainfall Figure 4.40 shows that there are clear areas of the slope which are more likely to be saturated for longer periods of time, consequently these areas become zones where OLF is more likely. Given that surface flow is dominant, then particular areas of a peatland hillslope act as contributing areas for greater volumes of runoff than others. In terms of ecology and hydrochemistry this may be important information. Furthermore, peatland restoration following gripping or milling may benefit from mapping and modelling exercises in order to show where runoff production is more likely to be concentrated so that management schemes can be based on enhanced knowledge of runoff-generating processes. Jones (1982) notes that perennially saturated areas will be important in determining potential source areas for perennially flowing pipes. The particular areas which maintain surface flow according to the crest stage tubes during the rainfall event day 282-283, 1998, are not just temporary features of this

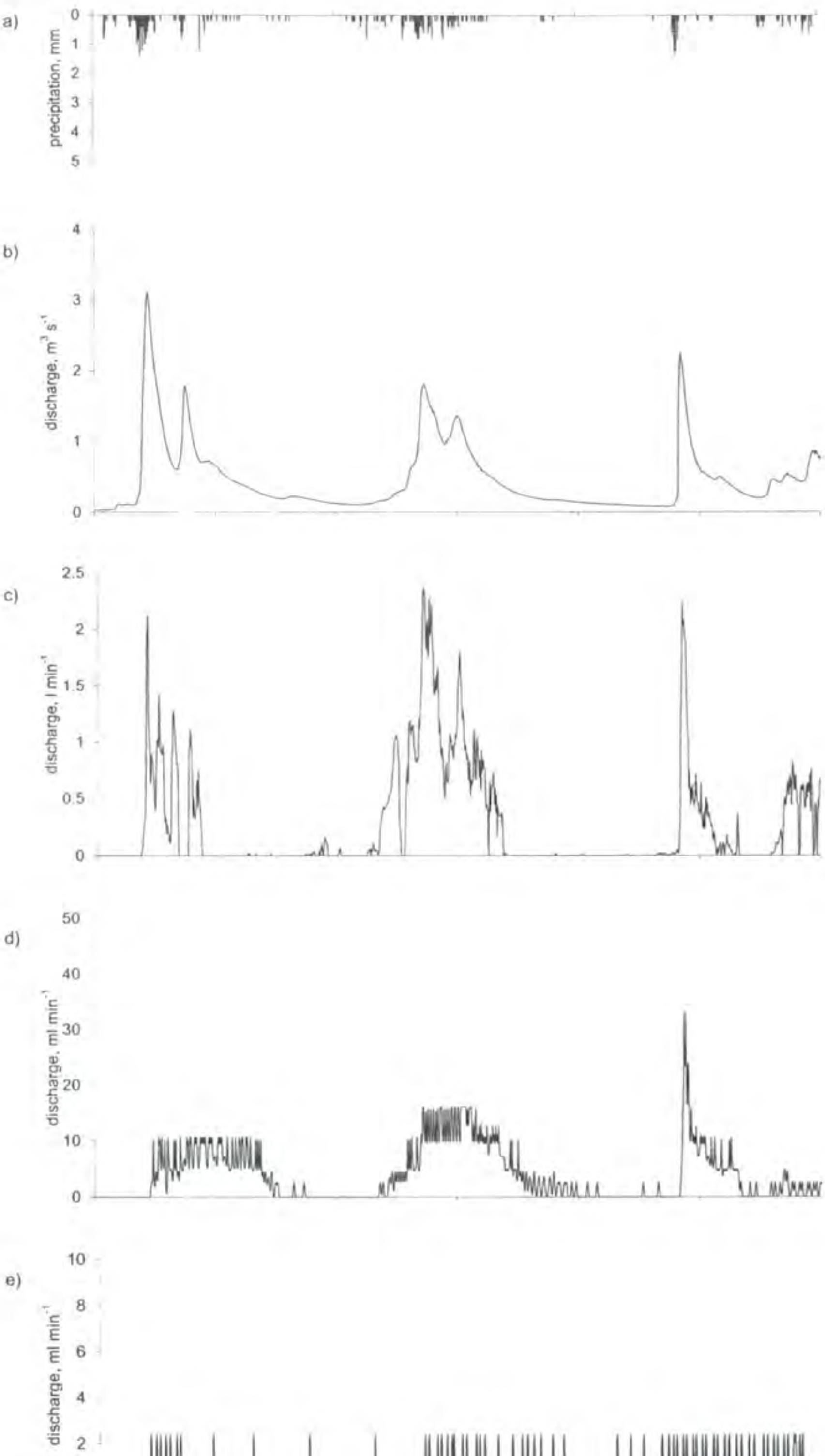


Figure 4.39. Runoff production from Trout Beck and the footslope of H2, day 273-285, 1998, a) precipitation, b) Trout Beck, c) H2 OLF, d) H2 5 cm depth,e) H2 10 cm depth.

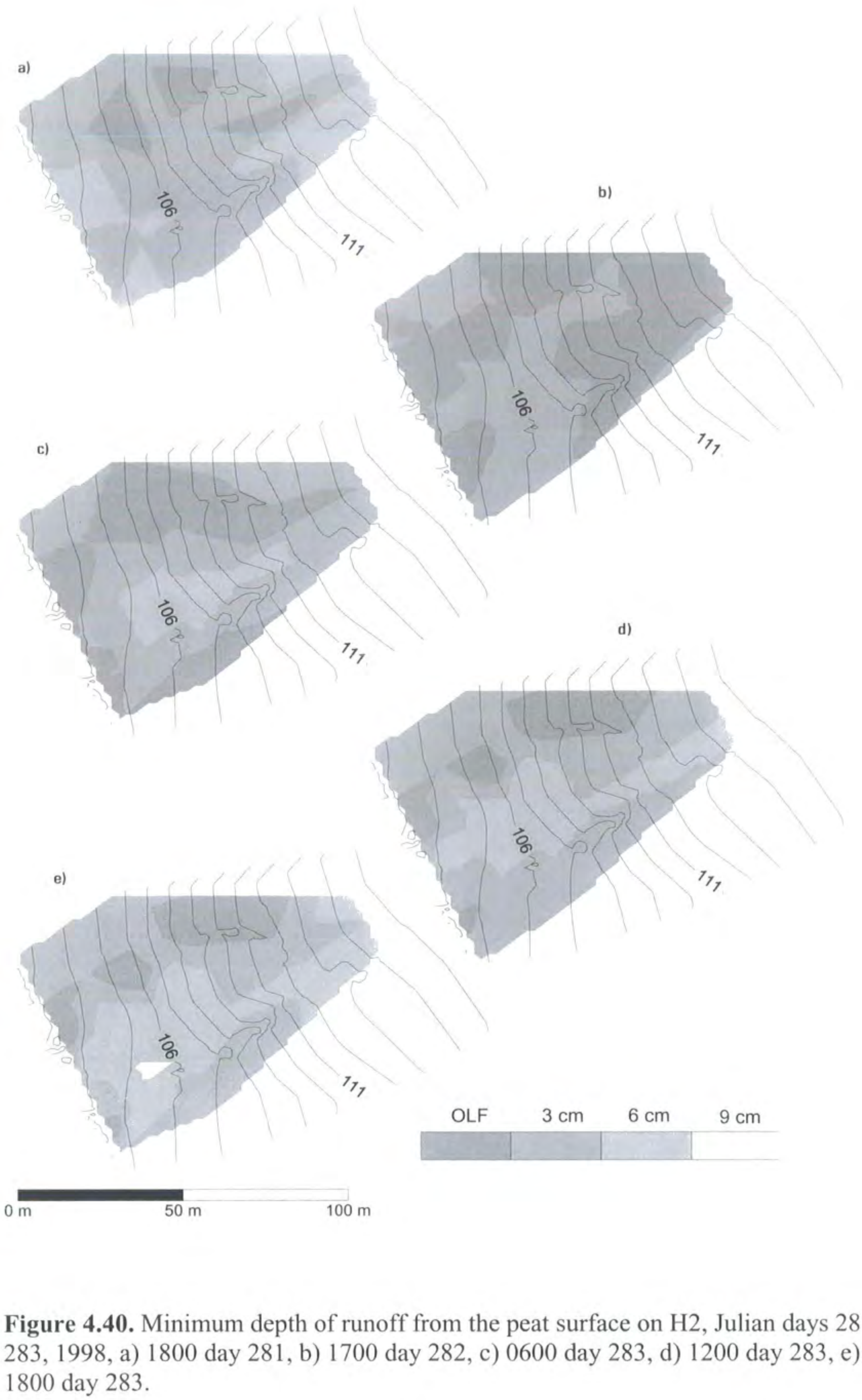
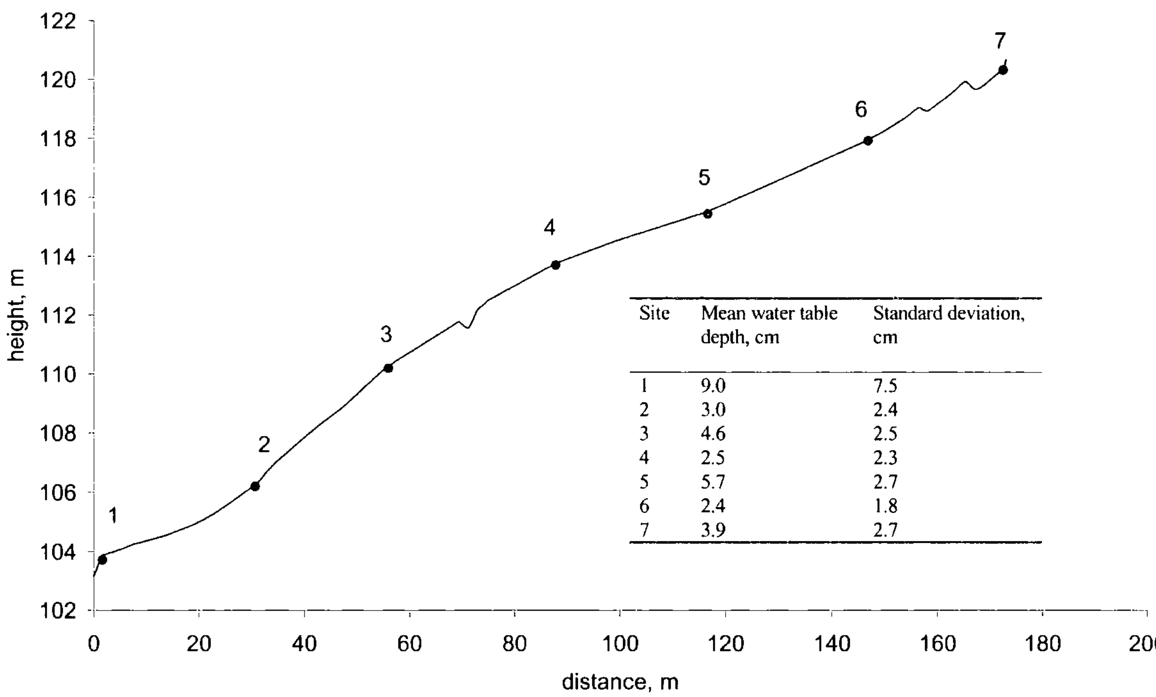


Figure 4.40. Minimum depth of runoff from the peat surface on H2, Julian days 281-283, 1998, a) 1800 day 281, b) 1700 day 282, c) 0600 day 283, d) 1200 day 283, e) 1800 day 283.

event. Bi-weekly sampling of crest-stage tubes over an 8-month period shows that these areas consistently produce OLF more frequently than other zones (Figure 4.42). Hence one side of H2 appears to be a dominant contributing area.



Site	Mean water table depth, cm	Standard deviation, cm
1	9.0	7.5
2	3.0	2.4
3	4.6	2.5
4	2.5	2.3
5	5.7	2.7
6	2.4	1.8
7	3.9	2.7

Figure 4.41. Mean and standard deviation of water table depth over a long transect across H2, bi-weekly sampling, June 1998-September 1999.

Concentrated lines of flow in peat have sometimes been attributed to headstreams which were originally developed in mineral ground, but have become overgrown by peat rather than collapsing later (Tomlinson, 1980). Ingram (1967) also identified 'water tracks' in peats where preferential flow seemed to occur. The evidence from crest-stage tube mapping which may just indicate ponding depth can be shown to be translated into enhanced runoff contribution. The transect of 76 m in length shown on Figure 4.42 is a roadside ditch at the foot of the slope. Here 40 throughflow troughs, mostly made of plastic guttering were inserted 50 cm into the peat at 5 cm depth and at 2 m intervals. The troughs were of four widths, 10 each of 3 cm, 5 cm, 10.5 cm and 50 cm. The last trough (of 5 cm width) on the upstream section of the ditch broke during installation so that there were 39 troughs along a 76 m section of the footslope. Runoff from the troughs was sampled manually and an example is given from three sampling occasions on 27th July 1998 in Figure 4.43a. Trough response is highly variable and Figure 4.43a gives an indication of the spatially distributed nature of runoff contribution from the hillslope. A salt dilution gauging was performed on the ditch using slug and constant rate injection techniques (Burt, 1988) and it can be seen that the large amount of runoff

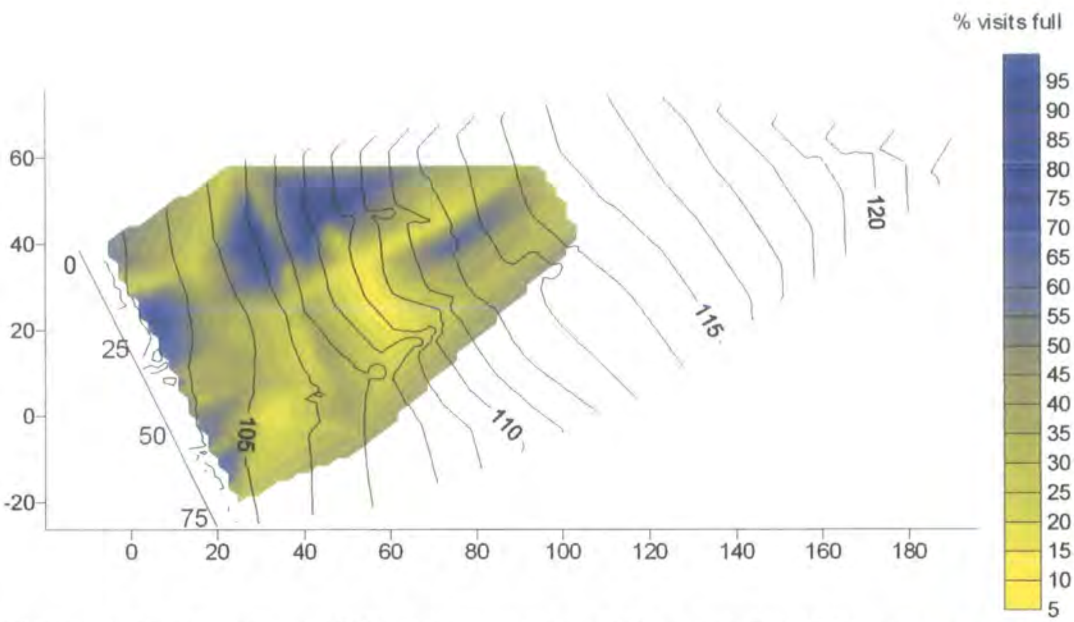
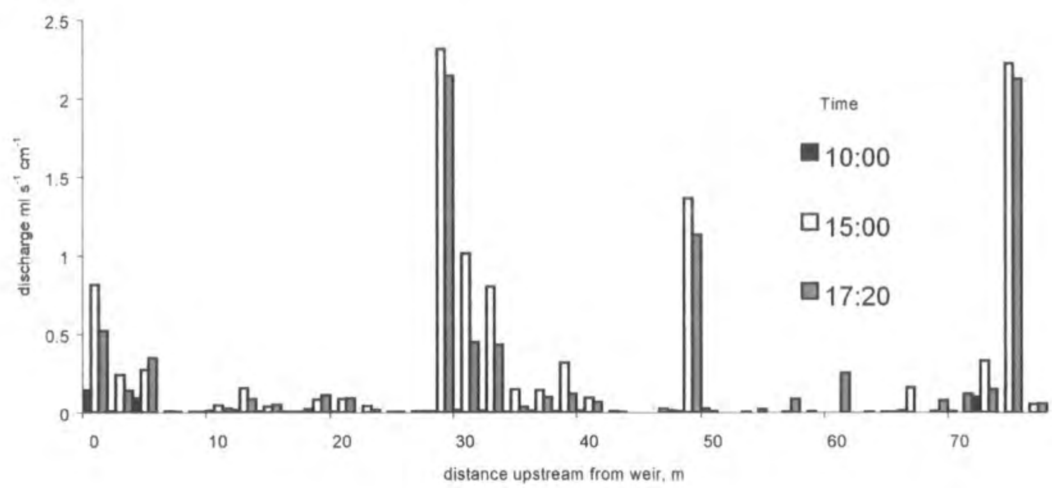


Figure 4.42. Frequency of OLF occurrence on H2 as monitored by crest-stage tubes based on bi-weekly samples, June 1998-September 1999, contours and axis distances in metres.

a)



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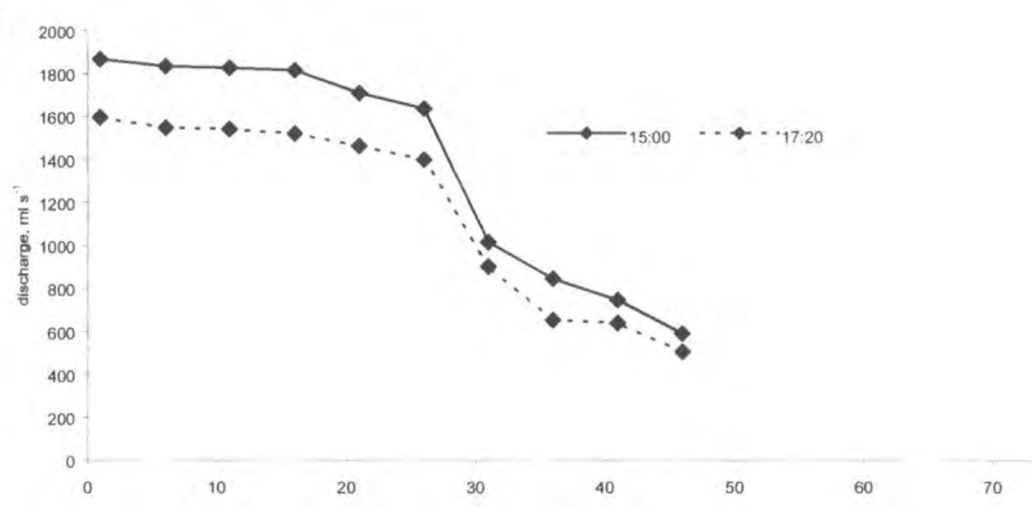


Figure 4.43. Variations in discharge from the footslope of H2, 27/7/98, a) from runoff gutters, b) by salt dilution gauging.

from the troughs at around 28-32 m along the footslope corresponds with where the maximum increase in flow from the slope to the ditch can be found (Figure 4.43b). Comparison of Figure 4.43 with Figure 4.42 shows that this 28-32 m zone along the transect corresponds with the upslope preferential generation of OLF. Thus, it has been possible to map flow production and contributing areas on a peatland hillslope. Clearly the spatial pattern of soil saturation has a dominant influence on runoff production in blanket peat.

d) Changing importance of runoff pathways over time

Throughflow troughs were stacked at 7 sites along the footslope of H2 with 6 depths of manual monitoring at each. The relative proportion of runoff produced from each depth category is shown in Table 4.11. During storm flow most of the runoff is produced from the uppermost layer of peat. However, during lower flow periods most runoff occurs from the acrotelm at 1- 5 cm depth. While much variation in complexity and detail in runoff could be masked by examining the peat from 1 - 5 cm depth as a single unit these data still provide important information on the relative hydrological importance of the surface and near-surface peat. The acrotelm can fill and overflow rapidly because the peat below it is saturated (the water tables are typically very close to the surface); on the recession limb OLF ceases first in most locations such that the upper acrotelm contributes most to flow during lower flow periods.

Table 4.11. Mean flow contributions to total discharge at each site, %, under 'high' and 'low' flow conditions.

Depth, cm	Mean % contribution at high flow	Mean % contribution at low flow
1	84.3	19.3
5	11.6	62.2
10	3.2	13.9
15	0.6	1.6
20	0.4	2.9
50	0.0	0.0

'High' flow was taken to be when flow at the Trout Beck gauging station was greater than 1 $m^3 s^{-1}$, and low flow when discharge was below this level. While 1 $m^3 s^{-1}$ does not appear to be a very high discharge it is exceeded only 13 % of the time, yet 70 % of the discharge volume occurs above this level. 7 sites, 21 low flow and 25 high flow measurements at each site, separated by at least 6 hours; usually by two weeks.

4.3.8 Evidence for bypassing flow in the deeper layers of blanket peat

4.3.8.1 Runoff from macropore outlets

Runoff plot results have indicated that very little flow emerges from peat layers below about 10 cm in depth except from eroded gully sides where water tables have dropped and the peat has suffered desiccation. However, runoff was detected from pipes and small seepage foci. Bower (1959) did note the existence of occasional small seepage points on exposed peat faces and they do represent one of the limited ways in which the lower layers of peat contribute to runoff. An example of one of these outlets is shown in Figure 4.44. Investigation of these seepage points by excavation has shown that they form the outlet of an existing tortuous macropore network within the peat mass. It is clear from mean lag times of just 1.6 hours and fairly rapid hydrograph recession (see Table 4.10) that these macropore networks allow water from the surface layers of blanket peat to reach deeper layers rapidly, bypassing the peat matrix. One potential source of water for the outlets is piston flow with seepage from above resulting in older resident soil water being pushed into the macropore or pipe system. However, the large volume of water emerging from these outlets during storm events suggests that piston flow processes are by no means dominant and it is more likely that new water infiltrating at the surface is bypassing the peat matrix. Runoff from two seepage points is shown in Figure 4.45 with peak discharge measured at over 0.6 litres min^{-1}. Some of the response is clearly diurnal and related to melting ice and snow. Some of the values recorded in the rain gauge may be melting and drifted snow. Analysis of automatic weather station data from Moor House indicates that on Julian day 103 1999 temperatures were below freezing causing M1 to cease flowing. Nevertheless, runoff response from the seepage points generally mirrors that seen in the Trout Beck catchment.

4.3.8.2 Peat-mineral interface flow

Flow at the peat base was monitored from six of the throughflow trough sites; only one produced flow at the peat-mineral interface (Figure 4.46). Discharge here is clearly low with a maximum recorded level of 14.5 ml min^{-1} per metre of contour width. Flow is ephemeral and linked very strongly to rainfall events. Discharge at the peat-clay interface is therefore not a result of continuous slow seepage from the peat mass. No flow was recorded from a depth of 10 cm to the base of the peat (100 cm) at this plot (H5). It is probable that a macropore network exists to connect the surface or near-surface of the blanket peat to the base at this point, bypassing the soil matrix. This may be important for the stability of blanket peat slopes (see discussion in Chapter 9).

Figure 4.44. Small seepage point located on a peat face indicative of flow occurring through the lower layers of the peat.

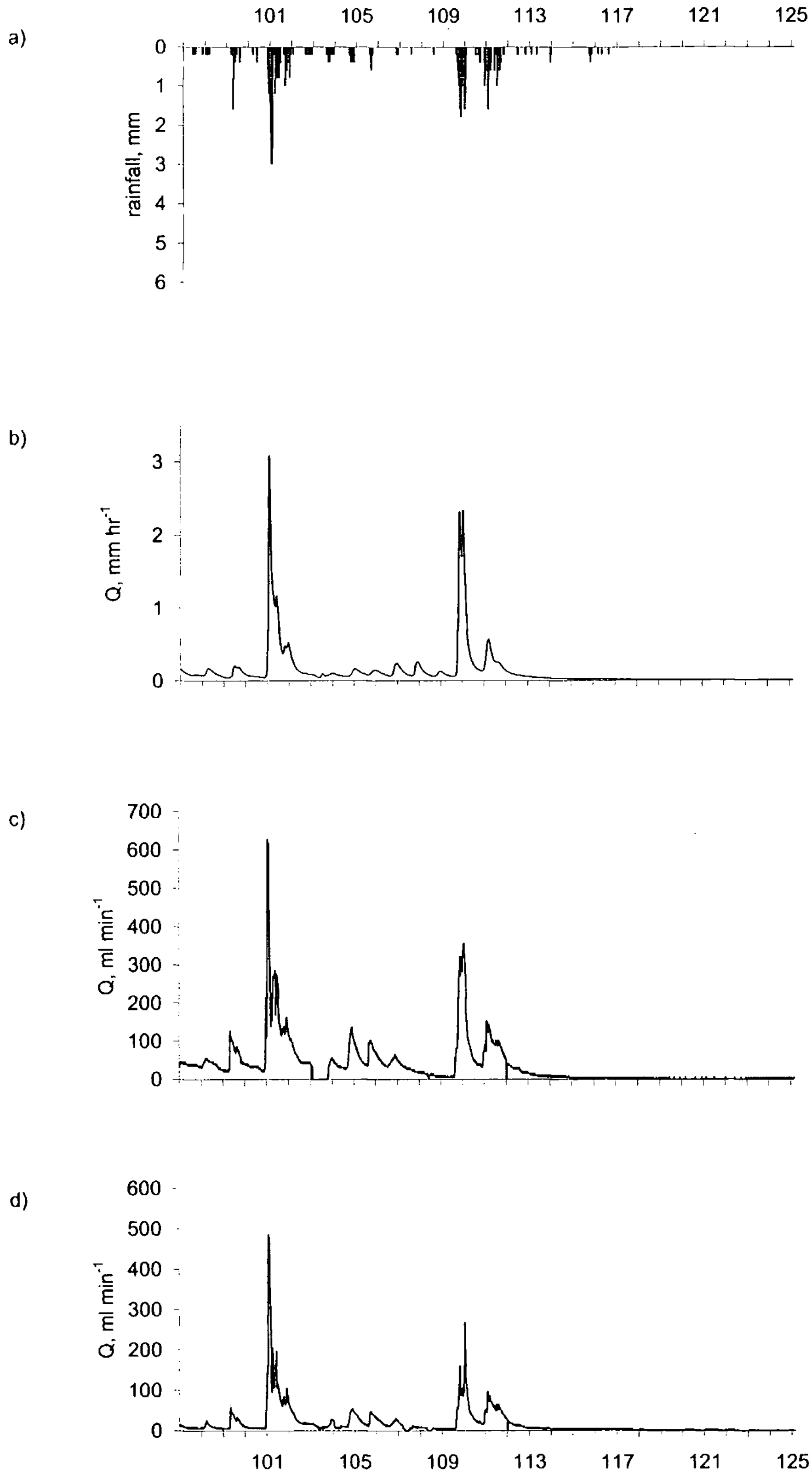


Figure 4.45. Discharge from two seepage points during day 97-125, 1999, a) precipitation, b) Trout Beck, c) seepage outlet at 45 cm depth, d) seepage outlet at 60 cm depth.

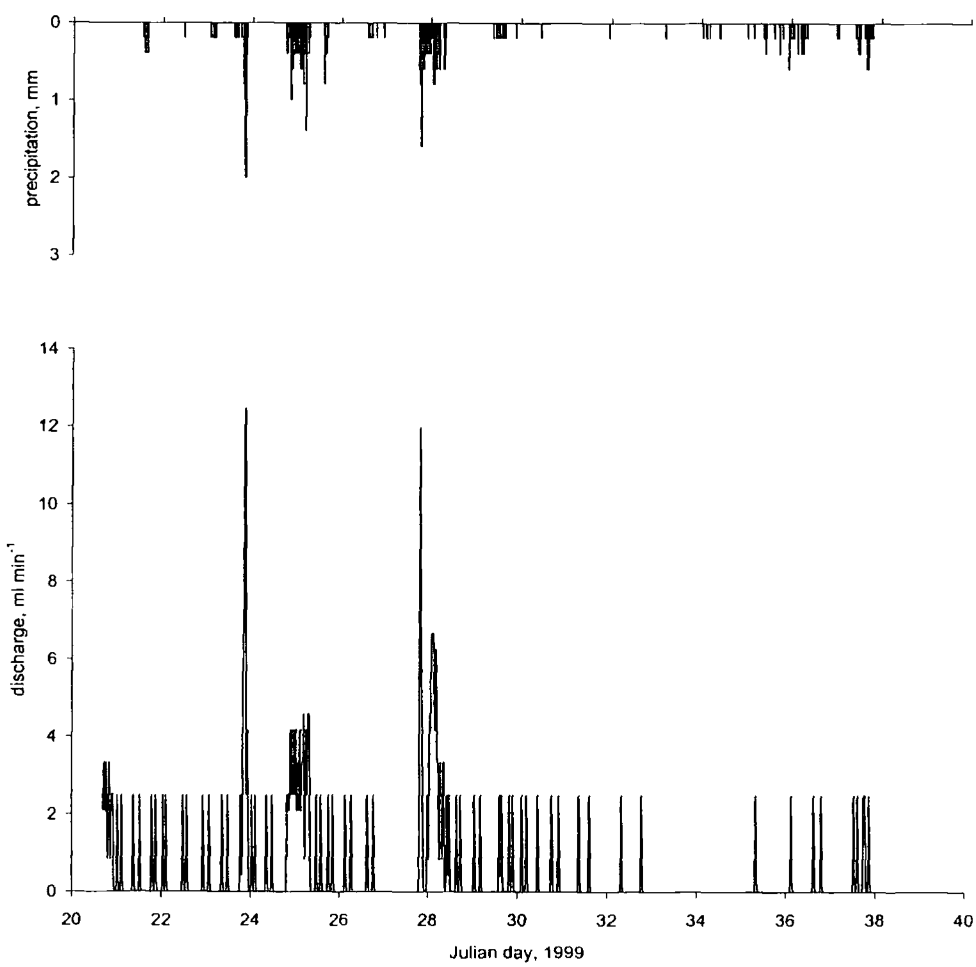


Figure 4.46. Runoff production at the peat-clay interface.

4.4 Conclusions

Blanket peat catchments exhibit flashy regimes. Runoff percentages for the Trout Beck catchment are high as rainfall is rapidly transmitted to the channel producing a flashy hydrograph response. During warm dry weather water tables are controlled by evapotranspiration. This is demonstrated by the diurnal cycles in the ECN water table record during summer dry periods with very little movement in the water table during night hours. Water table recharge is rapid. This suggests that infiltration rates are high when the lower acrotelm and upper catotelm are unsaturated. However, further testing of this assumption is required and Chapters 5 and 6 examine infiltration processes in more detail. The rapid generation of runoff from Trout Beck occurs when the water table is within 5 cm of the surface at the ECN target site. This water table elevation was achieved during 83 % of the 5-year study period. This indicates that runoff pathways are likely to be saturation-excess dominated. Plot-scale monitoring allows an insight into the detailed operation of the important infiltration-, percolation- and saturation-excess surface and near-surface runoff processes. Flow within the upper 5 cm of blanket peat significantly dominates plot-scale runoff response. During high flow, OLF is dominant; during low flow, flow between 1-10 cm into the acrotelm is dominant. Overall there is a dominance of saturation-excess OLF, particularly on more gentle slopes with impeded drainage, and on footslopes where OLF occurs most frequently. On steeper slopes, more flow seems to occur within the near-surface layers of blanket peat, rather than at the surface. Storm mapping of flow processes can elucidate the nature of variable source areas for runoff production in blanket peat. Topography and preferential flow tracks are important controls on areal contribution to runoff. Gripping will simply divert OLF and near-surface flow away from a hillslope such that upslope supply of water to the midslope near-surface flow pathway and as return flow on the footslope will be limited. Therefore, unlike intact slopes, in a gripped catchment footslopes will not be more important runoff generators than topslopes or midslopes. This is simply because the runoff supply from upslope is reduced since flow bypasses footslopes via the grip network.

No significant discharge emerges from the lower layers of peat except from eroded gully sides where water tables are suppressed and the peat desiccated, and from small seepage points which are the outlet for macropore networks within the peat mass. These networks often appear to be well connected to the surface and near-surface such that flow response to rainfall from outlets is rapid. Importantly, spatially localised ephemeral rainfall induced flow has been detected at the peat-mineral interface. The spatial distribution of this flow process may have implications for the stability of peat masses and the formation of subsurface pipes (see Chapter 9).

Re-examination of one of Conway and Millar's (1960) four sites has shown the inadequate nature of their monitoring strategy. This is of particular importance given the wide citation of their work. The eroded subcatchment produces higher peak flows than the drained subcatchment. Drainage densities and areal coverage of drainage is greater in the eroded catchment such that this would be expected. Moreover, the eroded subcatchment sustains low flows slightly better than the gripped catchment and greater runoff production occurs from the former. The artificially drained hillslope acts as a better store for water but is still a poor regulator of flow. Increased water table drawdown near gully sides or the occurrence of bog pools on the eroded site may be partly responsible for runoff patterns. There is some evidence that recovery has taken place since the last burning of Burnt Hill, such that lag times may have increased and runoff production from Burnt Hill has been reduced. Although heavily dissected, total (area-weighted) runoff production from the eroded catchment is very similar to that from Trout Beck which is only 18 % eroded (Eddy *et al.*, 1969). Production from the gripped subcatchment is much lower yet more 'intense'.

The data spanning the 1995 drought provided the opportunity to assess the response of the Trout Beck catchment to conditions which may become more common in the future. Flows from the Trout Beck catchment were reduced to extremely low levels during the summer of 1995 and the catchment was unable to sustain baseflow despite near-record rainfall during the preceding winter. Thus, if droughts become more common with enhanced seasonality, there will be severe impacts on the ecology of these upland blanket peat catchments. Reduced rates of water table recovery at the end of the 1995 summer may indicate physical changes in the peat due to oxidation and/or desiccation. These changes would significantly impact the timing and quality of runoff production and may also have the ability to trigger phases of erosion and alter the carbon flux from the peatland system (Roulet *et al.*, 1992; Silvola *et al.*, 1996). It is unclear to what extent drought affects the infiltration and runoff generation processes in blanket peat; these issues will be addressed in the following chapter. Some control over the rainfall intensity is desirable in order to elucidate more of the detailed infiltration and near-surface flow responses to rainfall.

CHAPTER 5
RUNOFF PRODUCTION FROM BLANKET PEAT DURING SIMULATED RAINFALL EVENTS

5.1 Introduction

This chapter presents results from rainfall simulation experiments performed on plots of blanket peat. These experiments allow determination of infiltration and runoff generation processes within the upper peat layers. The first part of the chapter will describe the equipment used. The results and discussion section is then split into two parts. The first (5.3) is based on field experiments performed during spring 1999 and also examines some repeated tests during the dry summer of 1999. The second (5.4) is based on laboratory experiments. Here pilot study work is performed to examine the effects of drought on runoff generation in blanket peat and also to examine the role of bypassing flow within the acrotelm.

5.1.1. The use of rainfall simulators to examine hydrological process

Rainfall simulators have been extensively used in hydrological, pedological and geomorphological problems (eg. Bergkamp, 1998; Bork and Rohdenburg, 1981; Cerda, 1998; De Ploey *et al.*, 1976; Imeson and Kwaad, 1980; Imeson, 1983; Lusby, 1977; Morgan, 1995; Pilgrim and Huff, 1983, Foster *et al.*, 2000). Simulators allow much greater control over the rainfall variable. The amount, intensity and duration of rainfall can be controlled along with other parameters such as drop-size distribution and water chemistry to varying degrees, depending on the system of application. Unlike a ring infiltrometer, the surface does not pond immediately, but will only do so at some later stage if the input is great enough. The time to ponding will depend on the application rate as well as on the hydraulic properties of the soil. Runoff can be collected to determine infiltration rates (by subtracting runoff rates from application rates) and erosion either as a whole sample for smaller plots or in a sub-sampling strategy for larger plots.

Almost all field rainfall simulator experiments have collected runoff solely from the surface layer and have disregarded any lateral throughflow in deeper soil layers. Tsuboyama *et al.* (1994) successfully collected matrix and macropore flow from a steep forested hillslope during application of tracer solutions, but such process measurement is rare. Field throughflow measurements of processes resulting from natural precipitation events are more widespread (e.g. Tsukamota, 1961; Whipkey, 1965; Dunne and Black, 1971; Knapp, 1970 and Weyman, 1971). The lack of throughflow investigation beneath rainfall simulators may be related to the large depth to which water can infiltrate in permeable soils. Data collection may also be subject to problems of pit throughflow convergence (Atkinson, 1978) as discussed in Chapter 4. Most reported rainfall simulator runoff plots operate some sort of pit collection system at their lower end such that infiltration and throughflow properties may be altered to some extent. Results in Chapter 4 suggest that most of the infiltrating water in blanket peat experiments will runoff laterally fairly near the surface. As a main subject of interest is elucidation of the exact runoff generating processes within the upper peat layers it seems sensible to measure subsurface runoff processes.

5.1.2. Types of rainfall simulator and choice of appropriate equipment

Broadly, there are three rainfall simulator systems: sprays, rotating sprays and drip-screens. Spray systems generally supply rainfall in pulses to the ground (e.g. Costin and Gilmour, 1970), and rotating sprays deliver rainfall over a large surface area although intensity usually decreases with distance from the rotating nozzle. The spray-type systems often provide rainfall at terminal velocities which approach that of natural rainfall. Drip systems use hypodermic syringes (Romkens *et al.*, 1975) or other drop formers over a fixed grid to produce rainfall over relatively small surface areas. The drop formers are often not raised high enough to allow representative terminal velocities, but they generally allow a constant rainfall rate with drop sizes more easily controlled than the in spray systems (Bowyer-Bower and Burt, 1989).

Various factors including money and time available, and the purpose of the experiment, influence the choice of simulator. Table 5.1 lists the factors of importance for blanket peat experiments and the most suitable type of simulator for each factor. The drip-type simulator is the most appropriate for this work. The logistics of use in the field in a remote area such as Moor House NNR is one of the main criteria for system choice. The amount of water used to provide rainfall at any given intensity is generally much less for a drip-screen design than for spray-type simulators (Bowyer-Bower and Burt, 1989). An intensity of 6 mm hr^{-1} requires only 12 litres of water hr^{-1} for the simulator described by Bowyer-Bower and Burt (1989). Hence an experiment can be run for over four hours before the water supply bottles (50 litre total capacity) need to be refilled. The ability to move the system easily between plots is also crucial and there is a need to control the wind variable (often problematic on the exposed moors) in order that accurate water supply to plots can take place. This is achieved more easily for a drip-type simulator, by simply surrounding the system with wind proofing (plastic sheeting). The use of a drip-type simulator means, however, that smaller plot areas will be used for analysis. Of course there is a trade-off between various factors of importance and it is recognised that small plots may not necessarily be fully representative of the blanket peat catchments. Nevertheless, several plots can be used and the use of a small plot means that all of the surface and near surface runoff can be collected and this avoids problems of inter-plot sub-sampling of the runoff processes.

Table 5.1 Desirable characteristics of a rainfall simulator for use on the blanket peat moorland of the North Pennines

Desired attribute	Best suited simulator
Easily transportable across blanket peat moorland	Drip-type
Efficient use of water	Drip-type
Protection from the wind to increase accuracy of rainfall application volumes into selected plot	Drip-type
Reproduction of low intensity rainfall to reflect expected natural conditions	Drip-type
Long duration rainfall	Drip-type
Achievement of 'natural' terminal velocity	Spray-type
Control over drop size distribution	Drip-type (but not very flexible)
Attainment of desired drop-size relative to rainfall intensity (i.e. flexibility)	Spray-type
Accurate replication of rainfall parameters	Drip-type
Uniformity of rainfall over the whole plot area	Drip-type
Adaptability to difficult terrain and vegetation	Spray-type

5.1.3. Previous work using rainfall simulators in blanket peat

The only previous rainfall simulation work performed on blanket peat to the author's knowledge was a 'pilot study' of Labadz (1988) who used a drip-type simulator on undisturbed peat samples. Only five 0.25 m^2 field plots were used with rainfall intensities ranging from 39 to 92 mm hr^{-1} which often fluctuated during the experiment and the longest duration of any of the 12 runs was 36 minutes. The duration of simulated rainfall experiments is usually shorter than events occurring naturally in temperate climates; test runs shorter than one hour are reported by most rainfall simulator workers, although Onstad *et al.* (1981) extended their experiments for two hours to allow runoff and soil loss to reach equilibrium rates. Labadz (1988) notes that the achievement of rainfall intensities comparable with those occurring naturally is important if results are to be compared with 'real' data, but in temperate climates this has proved difficult.

As discussed in Chapter 3 rainfall intensities in the Pennines are typically low (see Figure 3.7c). There is a dominance of low-intensity frontal and orographic rainfall at Moor House. Disregarding possible snowmelt occasions, 10 mm of rainfall in one hour was exceeded only five times in the four water years studied with a maximum of 11.6 mm hr^{-1}. Data from a raingauge at the study site were logged every 15 minutes between August 1998 to December 1999. Disregarding possible snowmelt occasions, results from this gauge indicate that 10 mm hr^{-1} was exceeded 18 times and 12 mm hr^{-1} six times. On only one occasion rainfall intensity occurred greater than 14 mm hr^{-1} when 5 mm fell in 15 minutes (20 mm hr^{-1}). Furthermore, the flashy response of river regimes in blanket peat areas appears to occur for all rainfall intensities above a threshold of 1-2 mm hr^{-1} and not just for those at the higher end of the precipitation range (Evans *et al.*, 1999). Hence it was felt that intensities of rainfall below 14 mm should be used for the simulation runs. The experiments of Labadz (1988) discussed above are probably unrepresentative of the conditions generally experienced in the blanket peat of the Pennines.

5.2 Characteristics of the rainfall simulator used in the study

5.2.1. Simulator design

A drip-type rainfall simulator as described by Bowyer-Bower and Burt (1989) was used to provide the rainfall. The principal components are shown in Figure 5.1. This design was developed by the Laboratory of Physical Geography and Soil Science, at the University of Amsterdam and by the Agricultural Research Station Zaidin in Grenada and has been used by many workers (e.g. Imeson and Verstraten, 1986; Gerits, 1988; Bowyer-Bower, 1993; Foster *et al.*, 2000). Drops were formed by controlling flow

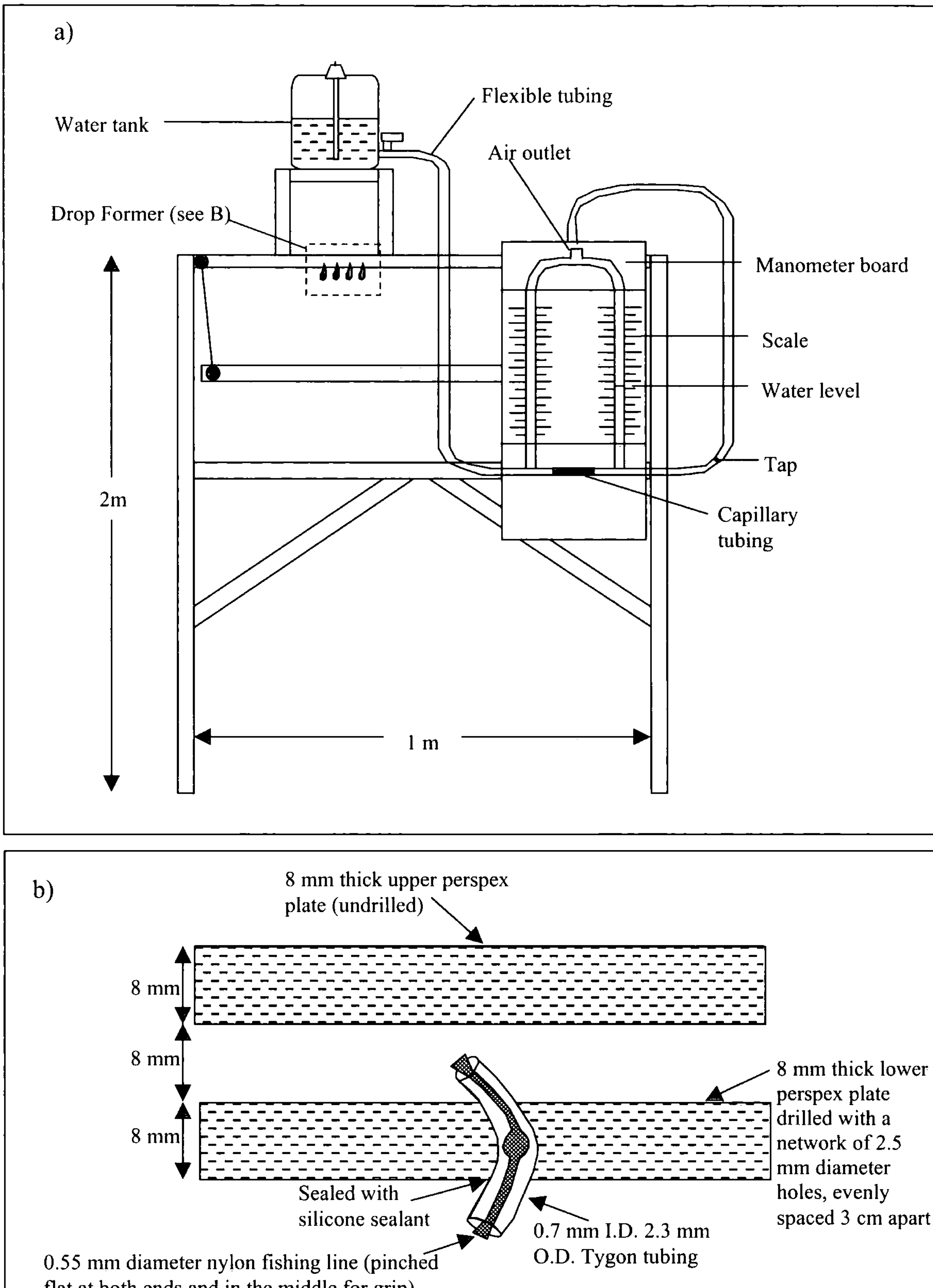


Figure 5.1. Design of the rainfall simulator used in the study.

through Tygon tubing of 2.3 mm outside diameter (OD) and 0.7 mm inside diameter (ID) through which was threaded 25 mm long, 0.6 mm OD fishing line. The upper perspex plate contained 627 drop formers in 19 rows of 33. A constant head system of two 25 litre water tanks mounted above the perspex drip-screen was used. A manometer board controlled the rainfall intensity and careful calibration through accurate measurement of rainfall production (by collecting the discharge in a tray and measuring the volume produced) allowed a head difference and rainfall intensity relationship to be accurately determined. Repetition of the calibration procedure showed that as long as the simulator was kept level accurate simulations of rainfall intensity could be reproduced ($r^2 = 0.98$). Intensities below 3 mm hr^{-1} could not be reproduced by this system. The rainfall simulator was supported by a metal frame with adjustable legs for levelling and adjusting the apparatus to the required distance form the ground.

5.2.2. Drop size distribution

Positioned 200 mm below the perspex plate hung a wire mesh which was used to scatter, break up and coalesce water drops into a distribution of drop sizes closer to that of natural rainfall. The dimensions of the mesh used provides a strong control on the distribution of drop sizes produced. Foster *et al.* (2000) used a 4 mm by 4 mm spacing although Bowyer-Bower and Burt (1989) suggest that a mesh of 3 mm x 3 mm spacing and 1 mm diameter wire is more suitable. The estimated drop size distribution of Foster *et al.* (2000) for a 20 mm hr^{-1} intensity run suggests a modal diameter of 3 mm using the filter paper method, although drops equal to or less than 12 mm diameter were recorded. Half of the of the drops were 3 - 5 mm in diameter. From the results presented by Foster *et al.*, it appears that only 68 droplets were measured, yet Salles *et al.* (1999) demonstrate that a minimum sample size of 10 000 drops is required in order to estimate the drop-size distribution with an accuracy of 3 % or less from rainfall simulators. For the present study the flour pellet method was used as it was cheap and easy and although time consuming, is one of the most common means of measuring drop size (e.g. Laws and Parsons, 1943; Costin and Gilmour, 1970; Cerda, 1998; Erpul, *et al.*, 1998). Figure 5.2 presents results from a 12 mm hr^{-1} intensity rainfall with a 3 mm x 3 mm mesh. The modal drop size was ≤ 0.5 mm, with a D_{50} (the drop diameter at which half the sample by volume is composed of larger drops and half of smaller drops) of 1.5 mm. This compares more favourably with natural drop size distributions (Best, 1950). Low- intensity rainfall is composed mostly of small drops (Laws and Parsons, 1943). Hudson (1971) discusses the properties of natural rainfall including drop-size distribution and terminal velocity. D_{50} increases from around 1.8 mm at 12.7 mm hr^{-1} to 2.5 mm at intensities greater than 65 mm hr^{-1}.

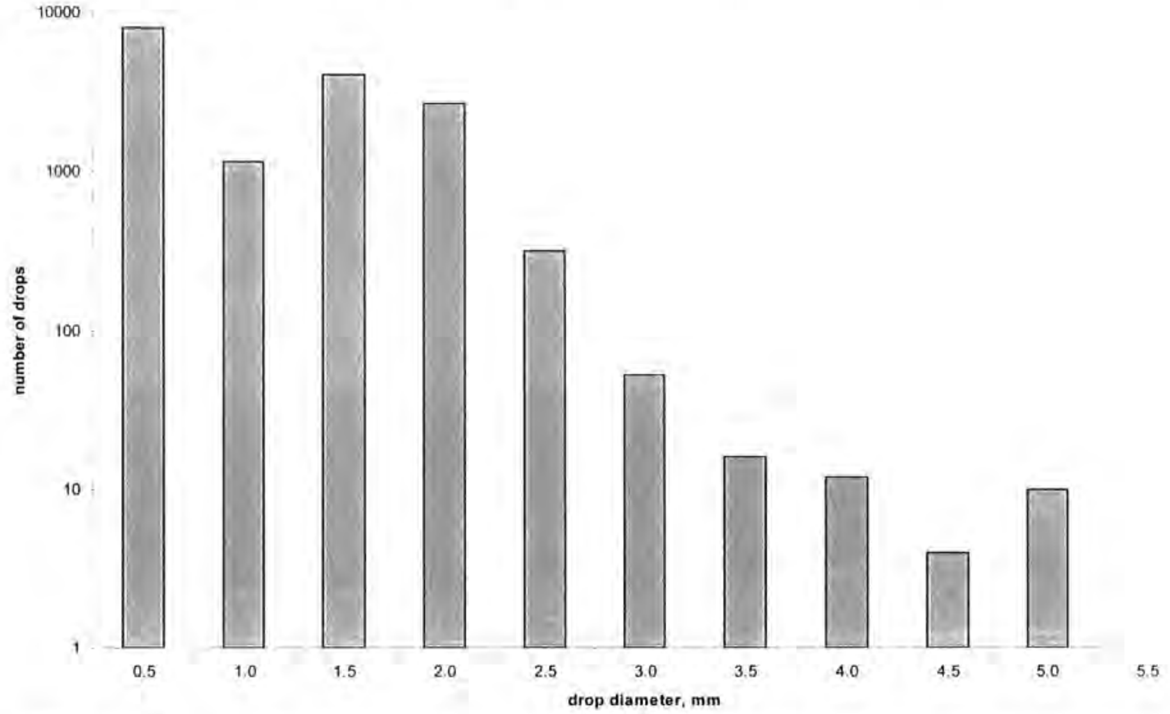


Figure 5.2. Size distribution of raindrops produced by the rainfall simulator at 12 mm hr^{-1}.

5.2.3. Energy characteristics

Although most of the present study examines the infiltration and runoff processes on blanket peat, some erosion work was carried out for bare peat plots. Here the kinetic energy (KE) of the raindrops is crucial for soil detachment and transport. Laws (1941) performed experiments on the fall velocity of water droplets using photographic techniques. He found drops of 1 mm diameter reached 95% of their terminal velocity after falling only 2.2 m whereas drops of 2 mm diameter required a height of 5.0 m and drops of 3 mm a height of 7.2 m. The drip screen simulator described above tends to produce simulated rainfall with lower KE because of the difficulty in raising the droplet formers to sufficient heights and because drops fall as a response to gravity and not due to an applied pressure from a pump or mains supply. For a drip screen at 1.8 m, and with the drop-size distribution described in the present study above, the range of terminal velocities were between 60-90% except in a few cases, with the D_{50} at around 80%. The mean KE of the rainfall produced by the simulator at 12 mm hr^{-1} based on the drop size distribution data was 0.069 J m^{-2} s^{-1}. Parsons (pers comm) calculated a KE of 0.089 J m^{-2} s^{-1} using a simulator of the same design at 25 mm hr^{-1}.

5.3. Rainfall simulation on field plots of blanket peat

5.3.1. Field methods

The rainfall simulator described above was used to provide the rainfall. The legs were inserted into the peat, with horizontal bars preventing the simulator from sinking further. The drip screen was adjusted so that in each case it was 1.8 m above ground level ensuring constant velocity for drops of any given size. On the blanket peat moorlands of the Pennines there tends to be significant air movement even on fine days so a protective polythene sheet was used to prevent deflection of rainfall outside the test area. Intensities could be easily varied between 3 mm to 140 mm hr^{-1}. Given the low natural rainfall intensities, 3, 6, 9 and 12 mm hr^{-1} were used.

Unlike Labadz (1988) who used mains water to refill the Mariotte reservoirs, it was decided wherever possible to use natural rainwater to refill the cannisters. Differences in the chemistry of tap water may affect soil erosivity (Barton, 1994) perhaps through ionic exchange capacity and hence the strength and stability of the peat (Hobbs, 1986). Furthermore peat erosivity may affect infiltration rates for example through the blocking or opening of hydrologically functioning macropores for water transfer. Rainwater was collected on the Moor House NNR in a large barrel. Occasionally when this source ran out, or when the study plot was too far from the barrel, stream water was used, being typically low in solute concentrations. This water was passed through a 63 μm filter before being used in the rainfall simulator.

Testing of the apparatus indicated an approximately uniform distribution of rainfall over the 1 m x 0.5 m area of rainfall delivery. Hence it was decided that the peat plots investigated should also be of this size. The plot was delimited on three sides by an aluminium plot boundary, inserted to a depth of 20 cm using a sharp cutting edge and protruding 10 cm above the surface (Figure 5.3). At the lower edge a small pit was excavated and three runoff troughs constructed of aluminium inserted against the clean front edge of the plot, being slightly inclined to ensure flow into the collecting vessel. The troughs were inserted at 1 cm, 5 cm and 10 cm below the surface to collect flow from the layer directly above the trough. The positioning of the troughs had to be taken

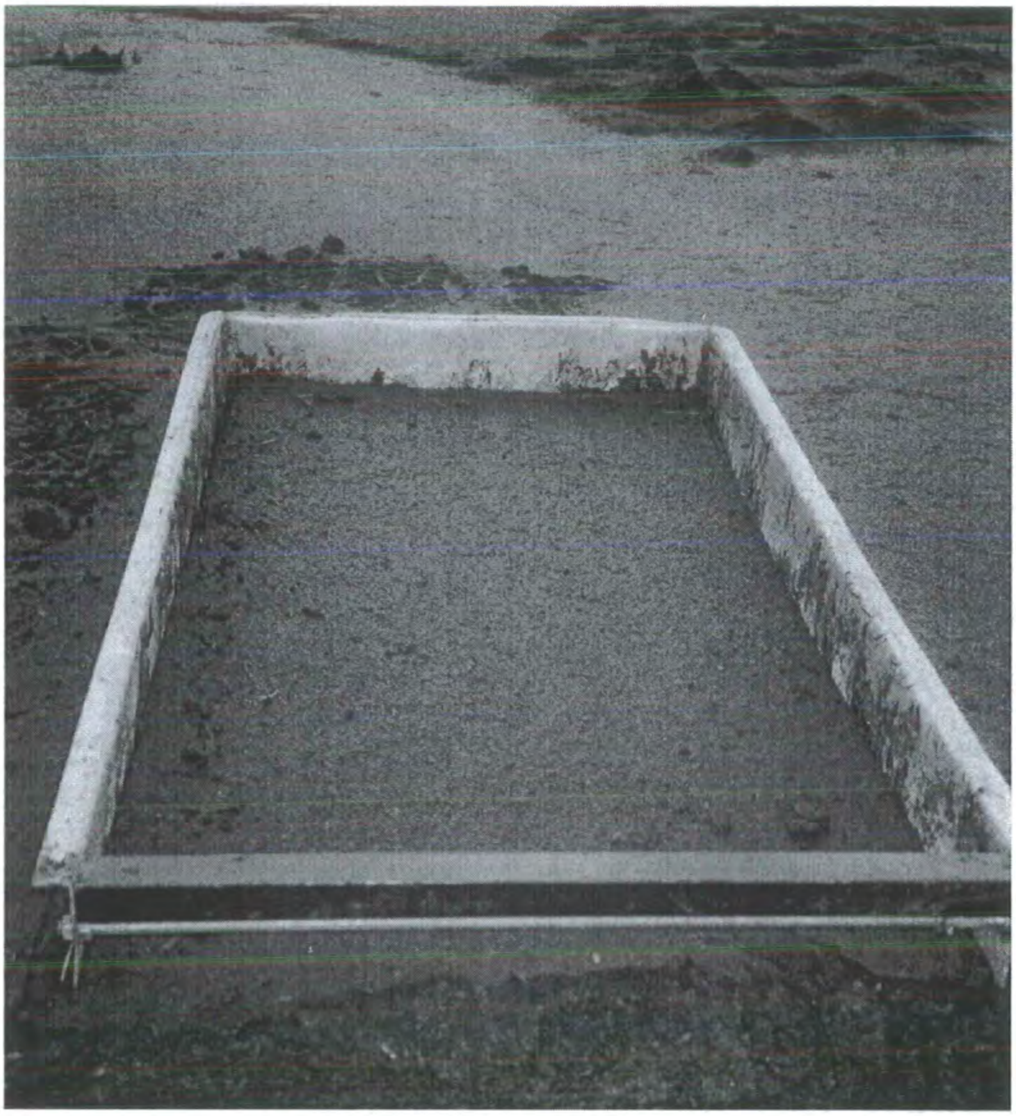


Figure 5.3. Aluminium plot boundary for rainfall simulator runs set up in the field, 1 m x 0.5 m area.

with great care in order to ensure that no water emerging from upper collection layers could leak down the face to lower troughs to give a false runoff record and to minimise disturbance of the peat. Bowyer-Bower and Burt (1989) note that a careful and complete seal of the trough lip to the soil surface is critical for accurately routing the runoff from the plot to the trough for collection and measurement. The upper trough was inserted at 1 cm below the surface because it was found to be very difficult to create suitable contact to collect surface runoff above this depth. Hence infiltration rates are indicative of infiltration to depths greater than 1 cm and any lateral flow within 1 cm of the surface contributes to 'surface runoff'. Runoff was measured manually every five minutes from each layer and measured in the field using volumetric measuring cylinders. From bare peat plots, the surface runoff was collected and poured into bottles for storage. The suspended sediment load was then measured in the laboratory by filtration and oven drying at 105°C within four days of collection.

Six plots for each of the four main surface cover types at the study site were selected (bare peat, *Calluna, Eriophorum,* and *Sphagnum*). Each plot contained at least 90 % of the selected cover and had a slope of between 2° to 3°. At each site, rainfall was simulated at four intensities (3, 6, 9 and 12 mm hr^{-1}). Rainfall intensity was not increased from 3 to 6 to 9 to 12 mm hr^{-1} on each plot. Instead the order of the runs was chosen at random so as to reduce the overall effect of any antecedence which might bias mean values. If for example, the starting intensity was 3 mm hr^{-1} for each plot then dry antecedence may influence mean infiltration and runoff values for this intensity. Once rainfall at a particular intensity produced steady-state runoff from the three runoff troughs (often this took 1-2 hours, and was longer at 6 and 3 mm hr^{-1}), the rainfall supply was stopped and the plot allowed to drain. The plot was left for at least 2 hours before the next run began. Runoff from the three layers generally fell to an extremely low volume within 30 minutes of rainfall stopping. The simulations all took place during April and May 1999. Two of the plots for each vegetation type were then revisited in August 1999 during a dry spell to see if surface desiccation and water table drawdown had any impact on infiltration and runoff processes from the acrotelm. In total therefore, 24 sites were examined, with 8 revisited during dry, warm weather. Hence a total of 128 rainfall simulation runs were conducted in the field. This represents a considerable increase in data on the infiltration properties of blanket peat and the runoff production processes within the acrotelm compared to previous studies.

5.3.2. Calculation of infiltration and runoff rates

Runoff rates were measured as volume of runoff from the plot over five minutes, and then converted to a rate in mm hr^{-1} since the exact volume of water applied was known. For 6 mm hr^{-1}, 12 litres of water were discharged in one hour, such that in 5 minutes 1 litre would be discharged. Hence if 1 litre was collected in a trough over 5 minutes, this would be equivalent to 6 mm hr^{-1} of runoff. Infiltration rates were calculated by subtracting surface runoff rates from rainfall intensity. This neglects any possible influence of evaporation and more importantly the effects of surface depression storage and storage on vegetation surfaces. As Slattery (1994) notes there are several difficulties in applying the theoretical infiltration curve of Philip (1957). Here it was found more appropriate to assess final steady-state runoff and infiltration rates as a mean value of the readings taken over a time period when the runoff was considered as 'steady'. An example is shown in Figure 5.4 where the initial trend associated with rainfall is the typical fall in infiltration rate up to about 40 minutes. After 40 minutes there does not appear to be any long-term rise or fall in infiltration rate and oscillation around a mean value appears to occur. Here the infiltration rate can be considered as quasi-stable. Observation of the surface runoff processes suggests that the oscillation occurs as a response to waves of water movement linked to surface ponding and the episodic cut and fill of micro-topographical features, and the burst out of water followed by a period of pool refill and micro-channel change. In a sense then, these fluctuations are not a direct response of changes in infiltration rate at the peat surface, but an error associated with the method of data collection. The Philip curve is necessarily fitted through the first and last data point collected if the simultaneous equation method is used. As the last data point may be a function of the episodic pool burst and fill effect, rather than actual infiltration rates, it was therefore felt more appropriate to use the quasi-steady state average.

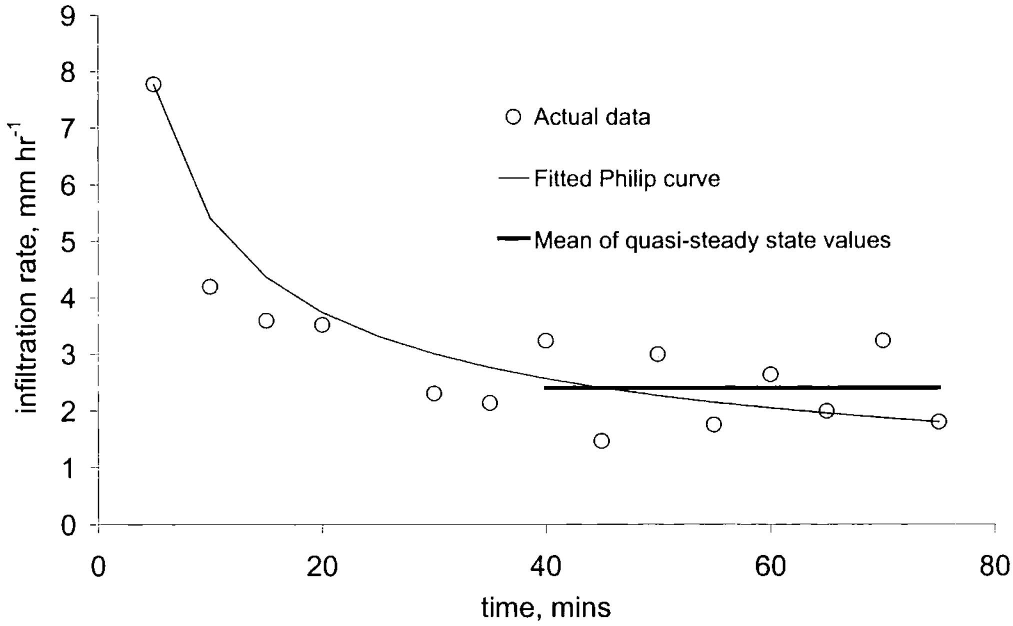


Figure 5.4. Comparison of infiltration rate determination over time with actual values, fitted Philip curve, and mean quasi-steady state value.

5.3.3 Results from tests in spring 1999.

<u>5.3.3.1. Infiltration rates</u>

Mean steady-state infiltration rates for each intensity of rainfall are shown in Table 5.2. The slight decrease in the proportion of applied rainfall infiltrating at steady-state is also indicated. This would be expected if the infiltration capacity of the peat was independent of rainfall intensity. However, mean final infiltration rate increases with rainfall intensity. Some runs produced no surface runoff such that infiltration rate could not be determined; clearly here the infiltration rate was greater than the application rate for these tests. Nevertheless, for most runs OLF was produced even at low rainfall intensities. Thus, as found in Chapter 4, OLF can develop at low rainfall intensities even on vegetated blanket peat. Labadz (1988) found infiltration rates ranging from 52.5 to 3.0 mm hr^{-1} with a mean of 17.7 mm hr^{-1} on bare peat. Clearly results from the present study are lower than the results of Labadz. Given the tendency shown here for infiltration rates to increase with rainfall intensity, comparison of results must therefore be related to rainfall intensity. It may be for this reason that ring infiltrometer tests do not provide adequate information on typical infiltration rates in blanket peat. For Labdaz (1988), rainfall intensities in her tests ranged from 32 to 96 mm hr^{-1}. Furthermore, the results of Labadz are based on 4 bare peat plots and one *Eriophorum* plot, compared with six of each of the four surface covers examined in the present study. Figure 5.5 shows how mean steady-state infiltration rates vary with intensity for each vegetation type. For lower intensities, mean rates of infiltration into bare peat are slightly greater than for pristine mire. At the 9 and 12 mm hr^{-1} intensities, the infiltration capacity of uncovered peat is slightly less than for *Eriophorum* or *Calluna* surfaces. However, there is little overall difference in rates between a *Eriophorum* or *Calluna* covered surface and bare surfaces. Mean infiltration rates into a peat surface below that of a *Sphagnum* vegetation appears to be lower than for other vegetation types.

Table 5.2. Mean and standard deviations (in brackets) for infiltration rate, mm hr^{-1} and % of applied rainfall infiltrating for field rainfall simulation experiments, Spring 1999.

Intensity mm hr^{-1}	3	6	9	12
Mean infiltration rate at steady-state, mm hr^{-1}	2.04 (0.73)	4.02 (1.21)	5.51 (1.78)	7.02 (2.42)
Minimum infiltration rate at steady-state, mm hr^{-1}	0.89	1.31	2.40	3.52
% of applied rainfall infiltrating at steady-state	67.94 (24.35)	67.02 (20.19)	61.22 (19.83)	58.51 (20.21)
Number of runs with no overland flow	3	3	2	1

n = 24 for each intensity

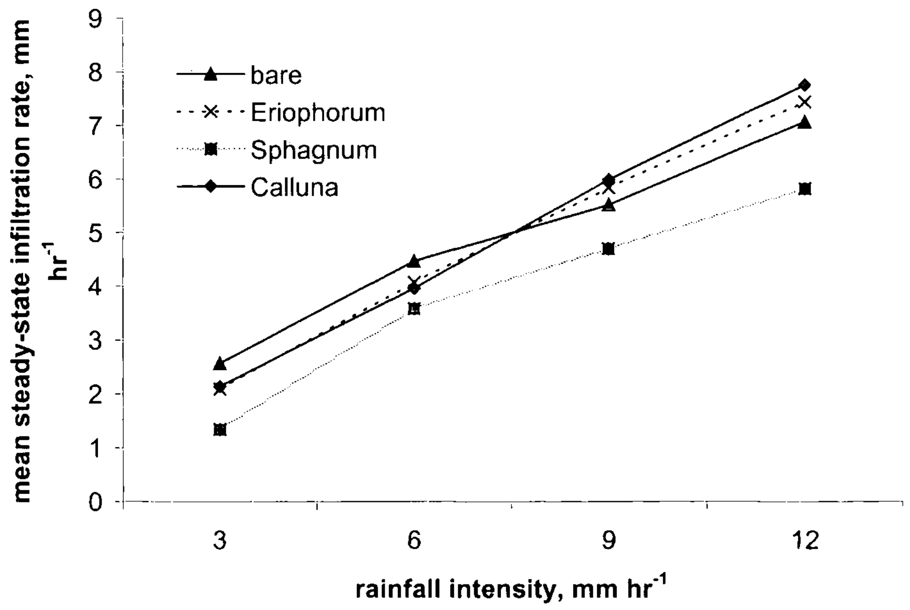


Figure 5.5. Mean infiltration rate against rainfall intensity for surface cover types on field plots exposed to rainfall simulation. Standard deviations can be calculated from Table 5.4.

5.3.3.2. Runoff production

a) Significance of factors

Runoff was collected from the surface layer and at 5 cm and 10 cm depth. Data for runoff rates from all layers at different intensities is highly variable and positively skewed, ranging from 0 to 9.99 mm hr^{-1} with a mean of 1.77 mm hr^{-1} and a skewness of 1.49. Given this, it is natural that the variability within groups of values defined by depth, vegetation and rainfall intensity categories is not even roughly constant, as required for application of ANOVA. A square root transformation of the data was found to work well. The square roots of runoff rates are less skewed with a range from 0 to 3.16 $\sqrt{(mm\ hr^{-1})}$, a mean of 1.05 $\sqrt{(mm\ hr^{-1})}$ and a skewness of 0.35. More important, the variability with depth, vegetation and intensity categories is now more nearly constant.

The ANOVA results (Table 5.3) show that the depth and intensity controls are overwhelmingly significant as the calculated significance levels are less than 0.00005. Hence both controls can be regarded as genuinely influencing runoff rates. As suspected during examination of infiltration rates above, surface cover is of some importance to the model but its influence is not as strong as depth and rainfall intensity controls.

Table 5.3. Analysis of variance of steady-state runoff rates from bounded rainfall simulator plots, square root data.

Source	df	F	Prob > F
Model	8	21.84	0.0000
Depth	2	54.40	0.0000
Intensity	3	19.44	0.0000
Vegetation cover	3	2.49	0.0600

$R^2 = 0.39$

Detailed cross-tabulation of the means (Table 5.4) demonstrates that runoff decreases with depth and increases with intensity. Runoff from all three soil layers increases with rainfall intensity.

A similar data transformation to that above was required for comparison of runoff efficiency data. Here raw data ranges from 0 to 85.3 % with a mean of 23.2 % and a skewness of 0.78. After transformation, skewness was –0.01 with a mean of 3.90 $\sqrt{\%}$,

Table 5.4 Means and standard deviations for runoff rates for vegetation and depth groups by intensity, square root data, $\surd(\text{mm hr}^{-1})$.

Vegetation	Depth	Rainfall intensity				
		3	6	9	12	Total
B	0	0.55	1.20	1.82	2.20	1.44
		0.39	0.35	0.43	0.31	0.73
	5	0.53	0.97	1.20	1.29	1.00
		0.48	0.61	0.77	0.68	0.67
	10	0.95	1.03	1.12	1.25	1.09
		0.47	0.57	0.59	0.70	0.56
	Total	0.68	1.07	1.38	1.58	1.18
		0.47	0.50	0.66	0.72	0.68
C	0	0.75	1.15	1.41	1.74	1.27
		0.59	0.92	1.10	1.20	0.99
	5	0.68	1.04	1.32	1.56	1.15
		0.55	0.79	0.97	1.14	0.90
	10	0.33	0.39	0.46	0.55	0.43
		0.39	0.47	0.58	0.64	0.50
	Total	0.59	0.86	1.06	1.28	0.95
		0.52	0.79	0.96	1.11	0.89
E	0	0.88	1.30	1.69	2.03	1.48
		0.39	0.52	0.60	0.72	0.69
	5	0.68	0.96	1.44	1.71	1.22
		0.55	0.92	1.21	1.40	1.08
	10	0.08	0.11	0.11	0.18	0.12
		0.19	0.27	0.27	0.43	0.29
	Total	0.57	0.79	1.08	1.31	0.94
		0.57	0.79	1.03	1.22	0.96
S	0	1.25	1.53	2.06	2.47	1.83
		0.34	0.24	0.23	0.29	0.55
	5	0.70	0.99	1.23	1.48	1.10
		0.67	0.33	0.31	0.52	0.45
	10	0.40	0.36	0.48	0.71	0.45
		0.47	0.45	0.45	0.44	0.43
	Total	0.64	0.92	1.26	1.55	1.13
		0.51	0.67	0.74	0.84	0.74
Total	0	0.86	1.30	1.75	2.11	1.50
		0.48	0.56	0.67	0.73	0.77
	5	0.70	0.99	1.30	1.51	1.12
		0.28	0.65	0.83	0.94	0.80
	10	0.24	0.47	0.54	0.67	0.52
		0.30	0.55	0.59	0.66	0.57
	Total	0.64	0.92	1.20	1.43	1.05
		0.51	0.67	0.86	0.98	0.83

and values ranging from 0 to 9.24 √ %. Again, variability with depth, vegetation and intensity categories is now more nearly constant. ANOVA indicates that for runoff efficiency, of the three controls, only depth can be accepted as a genuine control (Table 5.5). Rainfall intensity can be disregarded as a control of efficiency and vegetation cover is of limited importance, although it can again be argued that some influence can be identified.

Table 5.5. Analysis of variance of percent runoff at steady state as a proportion of incident rainfall from bounded rainfall simulator plots, square root data.

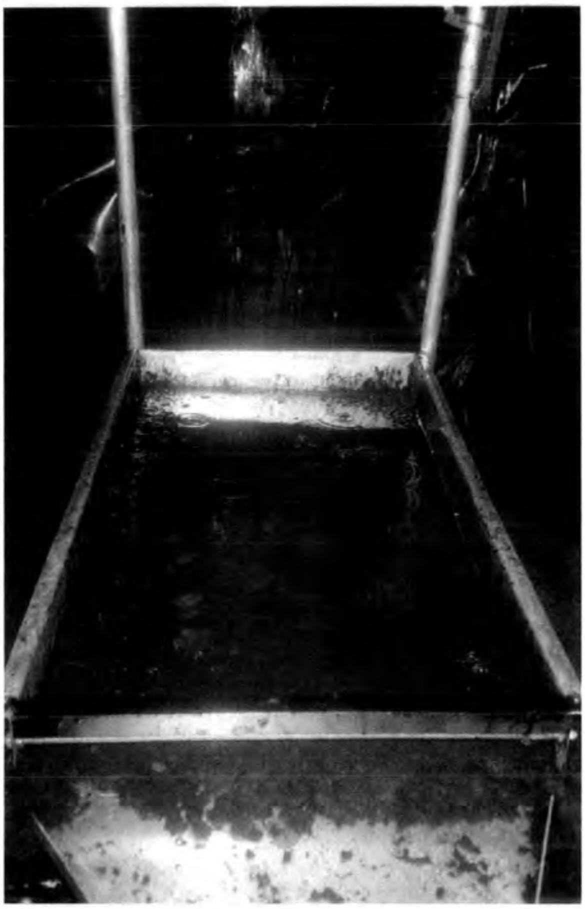
Source	d.f.	F	Prob > F
Model	8	14.15	0.0000
Depth	2	52.10	0.0000
Intensity	3	0.48	0.6960
Vegetation cover	3	2.52	0.0590

$R^2 = 0.29$

b) Rainfall intensity control

Table 5.5 shows that rainfall intensity exerts very little influence on runoff production efficiency. This is probably a reflection of saturation-excess runoff development rather than infiltration-excess. As noted in Chapter 1 saturation-excess OLF can occur at much lower rainfall intensities than is required for infiltration-excess OLF. Thus if the peat becomes saturated to the surface even under low intensity rainfall then OLF is likely to be produced no matter what the rainfall intensity is, as long as there is enough water supply to keep the peat saturated. The rainfall simulator results suggest that blanket peat runoff production is just as efficient for low-intensity as high-intensity storms. This evidence backs up the catchment-scale findings of Evans *et al.* (1999) who suggest that at Moor House rainfall events with intensities greater than about 1-2 mm hr^{-1} produce steep hydrograph responses as a result of rapid and efficient overland flow. Given that infiltration rates also increase with intensity, this suggests that a mechanism operates by which a similar proportion of rainfall can infiltrate into the peat, independent of intensity (over a 2 mm hr^{-1} threshold). This may be related to ponding development on the surface of the peat, with higher rainfall intensities inducing greater depths of ponding and hence a greater head of water and a resultant increase in percolation rates and subsurface runoff. Figure 5.6a shows the development of a ponded surface on the same bare peat plot as shown earlier in Figure 5.3. In other cases ponding was not

a)



b)

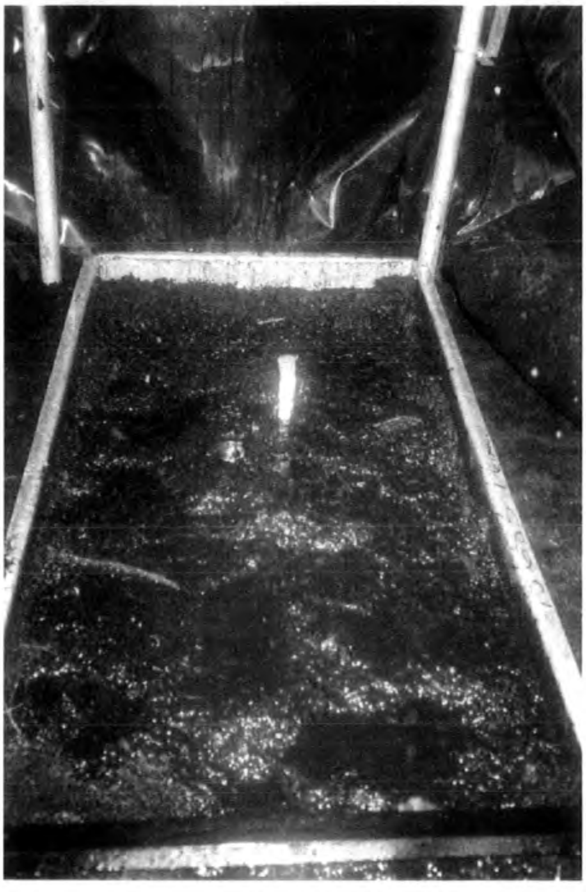


Figure 5.6.

c)

Figure 5.6. Observation of ponding development on runoff plots during rainfall simulation, a) ponding over almost the whole plot surface on bare peat, b) ponding in depressions on bare peat, c) bounded plot with runoff troughs on a densely vegetated peat surface illustrating the difficulty of witnessing ponding development on vegetated surfaces.

always as uniform across the surface of the peat and often only occurred in depressions dependent upon the microtopography of the plot (Figure 5.6b). Ponding development across the surface of a vegetated peat was more difficult to observe, as one can appreciate from the density of the cover in Figure 5.6c. However, some limited observations were made of possible ponding development within the vegetated plots. For example, for *Sphagnum*-covered plots, the level of water could often be seen to rise at the peat face within the living *Sphagnum* carpet at the throughflow trough junction.

Schiff (1953) showed that the infiltration rate below a depth of ponded water on a loam (with infiltration rates typically below 2 mm hr^{-1}) increased proportionally with ponding depth. Philip (1958) suggested theoretically that there should be a relationship between ponding depth and infiltration rate and tested this numerically for a light clay soil. It was suggested that infiltration rates would be affected by 2 % for every cm of ponded water. This is relatively small but importantly the effect was predicted to be greater in wetter and non-homogenous soils. Schmid (1989) incorporated OLF of depth 20 mm into infiltration models on sandy loams and demonstrated up to 11 % error in the 'no OLF' model. No work of this type has been done on peats and because the infiltration models are based on infiltration into unsaturated homogenous soils it is difficult to establish how important ponding depth may be on blanket peat without field experimentation.

Since most rainfall simulation studies report only one intensity of rainfall there are few reports of relationships between rainfall intensity and infiltration. Bowyer-Bower (1993) did find that rainfall simulation at higher rainfall intensities resulted in greater infiltration rates in a semi-arid soil. Here, the increase was attributed to the greater energy of high-intensity rainfall disrupting the soil crust in response to wetting. For example, finer material produced by slaking and dispersion is kept in suspension instead of blocking pores and thus decreasing infiltration rates. Similar processes can be envisaged on blanket peat.

One further point stems from the non-uniform nature of a soil over a $0.5m^2$ plot. The surface of part of the plot may well have a higher infiltration capacity than the rest of the plot. Therefore as Hawkins (1982) demonstrates numerically, mean infiltration rate over a plot will increase with rainfall intensity simply because a greater flux of water is occurring through the parts of the plot surface that have the higher relative infiltration capacities. These factors are often ignored by workers using rainfall simulators. Not only are the findings presented here of importance for our understanding of blanket peat hydrology, but these results have implications for the way in which we measure infiltration rates themselves. It is clear that the infiltration process into blanket-peat is more time and space bound than first thought: it is linked to the temporal distribution of rainfall intensities as well as durations. Comparison of the infiltration properties of blanket-peat must therefore be contextualised within rainfall intensity data.

c) Depth control

Mean steady-state runoff rates from all three runoff-collecting troughs are shown in Figure 5.7. Standard deviations are indicated (although bars are not used as the chart becomes too difficult to interpret) and indicate the wide variability within the dataset. There is a large amount of overlap but on average the greatest amount of the applied rainfall runs off as overland flow (be it saturation-excess or infiltration-excess). Lateral flow between 5-10 cm depth accounts for a mean of only 7.2 % to 13.0 % of incident rainfall volume, compared with 31.6 % to 40.8 % at the surface and 21.7 % to 25.5 % from the peat layers between 1-5 cm depth. Mean runoff increases in all layers with increasing rainfall intensity, a result of increased infiltration followed by enhanced lateral flow, irrespective of whether the mechanism involves increased head through ponding or other intensity-dependent processes. For the lowest layer of monitored peat, the gradient of the rise in runoff with intensity is far less than for the overlying layers. This is obviously linked with the larger proportion of overland flow occurring, but may also be to some extent a reflection of a limited capacity for lateral flow within this layer restricted by a lower hydraulic conductivity and reduced percolation rates. On average 77% of the input rainfall is collected from the three monitored layers. The rest of the rainfall may be infiltrating deeper into the peat, some of which may be occurring through leakage down the sides of the aluminium plot boundary.

d) Surface cover control

ANOVA demonstrated that there was a minor vegetational control on runoff generation. It is not necessarily surface cover that is the control, however, rather the surface cover is representative of characteristic properties of the peat below that cover. It is well known that particular vegetation types prefer different water table conditions, height and fluctuations being important (Heikurainen, 1968) for example. Furthermore, the vegetation may interact with the peat structure by rooting, litter deposition and building up of the peat deposit. Thus 'surface cover' is used as a simple classificatory approach.

More detail is added to the broad trend of a decline in runoff with depth by Figure 5.8 which examines surface cover controls. Here error bars have not been added for ease of interpretation of the diagram but standard deviations are included in an inserted table. For *Eriophorum*-covered peat, the mean runoff between 1-5 cm is just as great as that at the surface, but between 5-10 cm only 1.2 % of input rainfall is collected as throughflow from this layer. So peat below cotton grass clearly allows profuse flow within the top 5 cm of the peat but below this layer very little lateral flow occurs at all. Labadz (1988) set up a drip-type simulator above one *Eriophorum* plot on blanket peat in the Southern Pennines. Here, rainfall intensities of up to 90 mm hr^{-1} failed to produce 'Hortonian overland flow' (p255). Nevertheless flow into the soil pit created to install the only runoff collection trough (at the surface) was found to be profuse as the pit filled with water and had to be emptied regularly, something which was not found at the four bare peat plots of Labadz (1988). Hence it was suggested that rapid flow was occurring in the uppermost layers of decaying vegetation and that erosion of these areas is unlikely unless the vegetation is removed. Unlike the results from Labadz (1988) however, the results presented in the present study indicate that vegetated peats are very capable of producing overland flow. However, *Calluna* plot 5 provided results of a similar nature to that of Labadz (1988) with no surface runoff nor any runoff from the 10 cm trough. Instead profuse flow occurs between 1 and 5 cm into the peat mass.

There may be some difficulties in comparing data sets due to different definitions of the peat surface. For the present data the surface is defined as the first centimetre of intact peat and any very loose leaf litter layer is not really considered as the peat 'surface', although often the distinction is very difficult due to the partially living nature of the upper peat profile. Ingram and Bragg (1984) suggest that the acrotelm itself possesses the essential characteristics of a layer which suppresses sheet flow. At the same time, however, we have already seen evidence from crest-stage tubes, storm mapping and runoff plots (Chapter 4) that widespread overland flow does occur on vegetated peat hillslopes often to depths of more than 1 cm. Not only will definitions of the surface vary but it is likely that acrotelms of different natures and hence different surface properties exist, spatially distributed throughout the areas of study in the literature and indeed throughout small catchments.

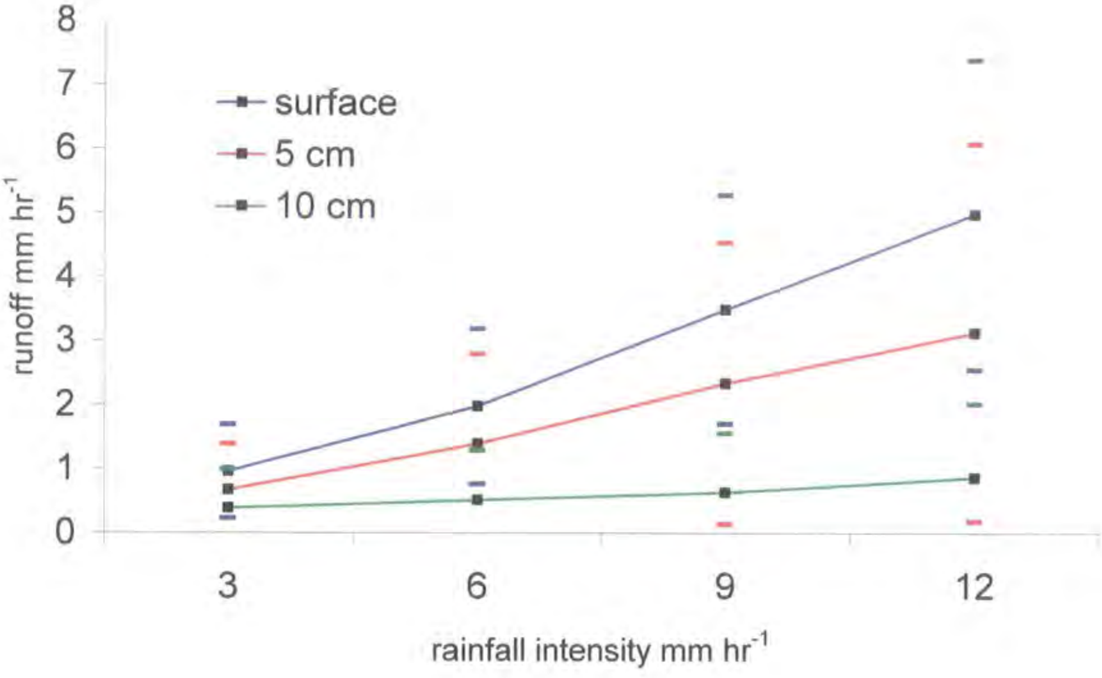
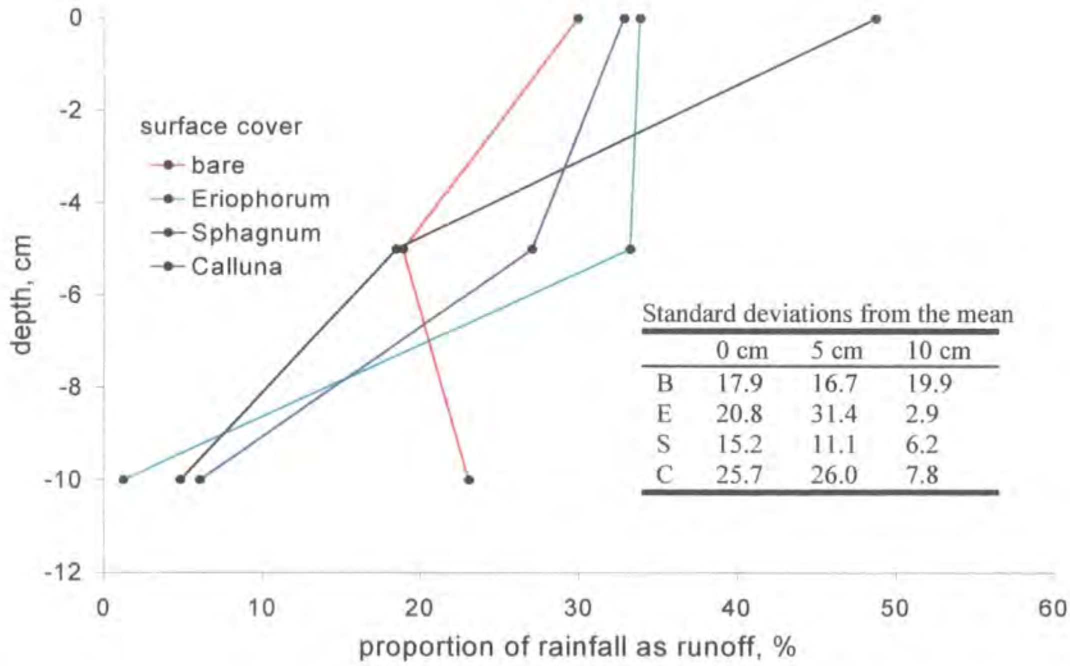


Figure 5.7. Mean steady-state runoff from field plots by depth against rainfall intensity. Plus and minus one standard deviation from the mean indicated by horizontal coloured bars.



Standard deviations from the mean

	0 cm	5 cm	10 cm
B	17.9	16.7	19.9
E	20.8	31.4	2.9
S	15.2	11.1	6.2
C	25.7	26.0	7.8

Figure 5.8. Proportion of input rainfall produced as runoff by depth for surface cover type.

For vegetated surface types runoff decreases with depth, but for bare peat the mean proportion of runoff between 5-10 cm is 8.0 % greater than that between 1-5 cm and only 2.4 % less than surface runoff. This may be related to some form of peat desiccation. It may be that weathering on an unprotected peat surface leads to the development of a more permeable upper peat mass; any weathering effect is lessened with penetration depth and percolation-excess once again occurs at 10 cm such that lateral flow occurs more readily at this level. Ingram and Bragg (1984) note that on a bare peat surface with the downwasting and removal of the acrotelmic layer, that the result is a mire with restricted infiltration leading to enhancement of sheet flow on the surface. Results presented here, however, indicate that bare peat has equivalent infiltration rates to that of a vegetated peat. In this way a dynamic feedback mechanism may operate because the peat itself changes its hydraulic properties as the emerging bare surface becomes susceptible to surface drying or frost heave, and to aeration. Hence the bare peat surface degrades and allows infiltration to take place, such that the near-surface peat that was once the catotelm now itself becomes a thin acrotelm. Ingram and Bragg (1984) ignore this mechanism. An indication that the surface properties of the bare peat are very different to that of the peat below comes from analysis of the dry bulk density (DBD) of bare peat with depth (Figure 5.9). The top 10 cm of bare peat has a much lower DBD than the peat below, with a sharp transition after about 10 cm to a much denser peat. As well as desiccation of the surface, erosion may lead to reworking of the surface peat, probably through a mixture of water and wind-driven mechanisms, such that the top layer of peat may in certain locations contain a depth of unconsolidated deposited peat. In this case it is likely that bulk densities are decreased and this will allow increased infiltration to a shallow depth just below the reworked layer where lateral runoff can take place. For the *Eriophorum*-covered peat core, DBD increases gradually with depth, although the bare peat is more compact, probably representative of greater humification and age.

Runoff rates from all layers and plots are shown in Figure 5.10. Only 1 of the 24 plots produced no surface runoff during a 12 mm hr^{-1} rainfall. This suggests that on bare peat and below a vegetation cover, surface flow is likely to be a widespread phenomenon if rainfall is prolonged. It may be that overland flow is a product of a mixture of infiltration and saturation-excess mechanisms. The variability in runoff with depth between plots indicates that water movement in the acrotelm is highly variable. It is probable that runoff pathways to deeper layers are spatially localised, perhaps related to macropore connectivity, or to the connectivity and spatial distribution of more permeable matrix.

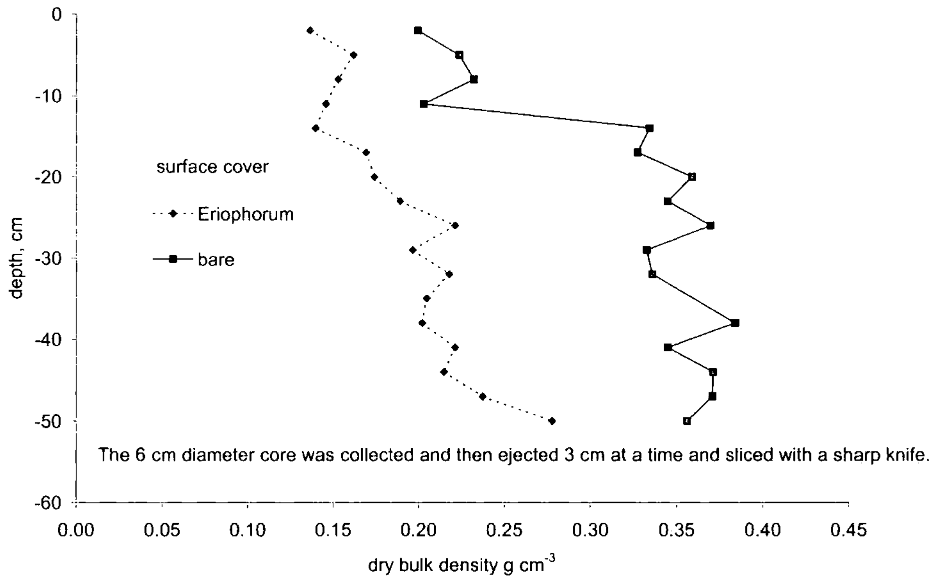


Figure 5.9. Variation in dry bulk density with depth below an *Eriophorum* cover and a bare surface.

5.3.3.3. Examination of rainfall-runoff processes from individual plots

Although there are too many simulator runs to present for individual analysis here, it is still useful to examine a small selection in order to elucidate some of the findings from closer examination of plot-scale rainfall-runoff response.

The runoff and suspended sediment response from bare plot 2 for a rainfall intensity of 12 mm hr^{-1} is shown in Figure 5.11. Observations of operational processes leading to surface runoff oscillations were mentioned earlier. Oscillations are also recorded for runoff at depth. Walsh and Voigt (1977) found similar oscillations in percolation rate from a rainfall simulation experiment on leaf litter. Percolation rates were seen to oscillate such that they could frequently be in excess of rainfall intensity. Here the development of unsteady layers in response to alternating relatively impermeable and permeable layers could lead to an alternating build-up and release of water within the soil, and hence fluctuations in the output measured by the throughflow trough collector. Other factors of importance may be temporary blockages developing within the upper litter and soil layers and changes in the characteristics of the peat with progressive wetting.

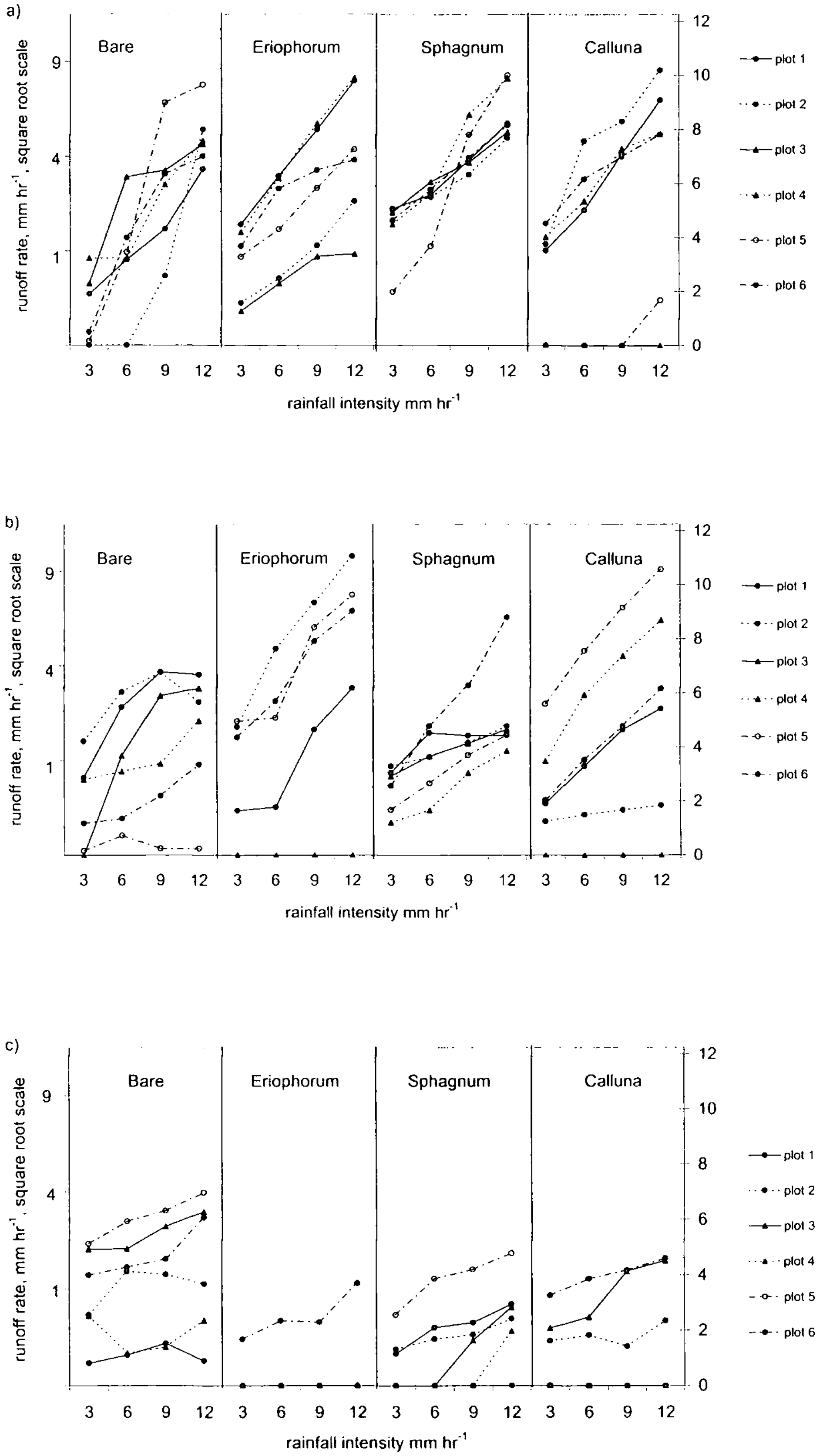


Figure 5.10. Steady-state runoff rates by vegetation type with rainfall intensity for each field plot, a) surface runoff, b) runoff at 5 cm, c) runoff at 10 cm.

At the plot-scale, some of the catchment-scale characteristics seen earlier (see Chapter 4) can be observed. For example, Figure 5.11 shows there is a very rapid response to rainfall from all three layers from plot B2. Within about 20 minutes of the onset of rain, runoff response from all three layers rises rapidly. Runoff is greater nearer the surface and declines with depth. Steady-state rates of runoff are achieved within 80-100 minutes for all three layers. Equally the typically fast recessional responses of blanket peat catchments can be seen at the plot scale. The recession appears much more rapid than the time to peak for surface runoff for this run, although the hydrographs are fairly symmetrical in appearance. Surface runoff recession is faster and more dramatic than those of lower layers with more time being taken for subsurface drainage to occur. This is a result of continued percolation of water into the peat mass from the surface layer after rainfall input has ceased. The drainage of the lower layers slows as the excess water drains more quickly from above. This suggests that surface flow in this case is more likely to be a result of saturation-excess mechanisms rather than infiltration-excess. Sediment loading appears to be supply limited with concentrations decreasing through time after an initially high peak although other mechanisms operating may produce the same trend (see below).

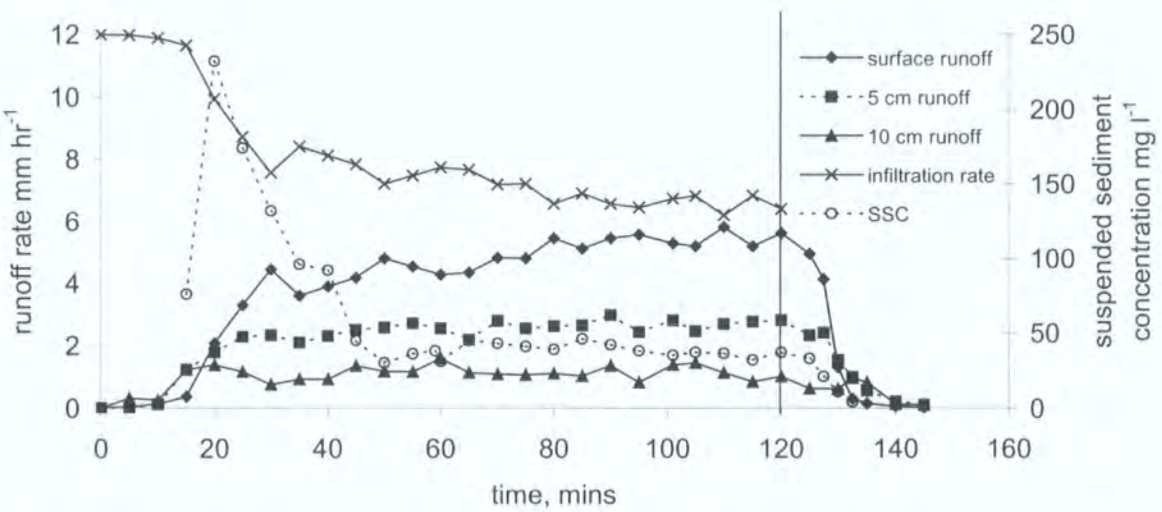


Figure 5.11. Runoff production, infiltration and suspended sediment concentration for field plot B2 during a 12 mm hr^{-1} event. No suspended sediment concentration data over first 15 minutes as insufficient runoff from the surface produced for analysis.

Runoff response from a *Sphagnum*-covered plot is shown in Figure 5.12 for 12, 9, 6 and 3 mm hr^{-1}. Again runoff is greater nearer the surface. Steady-state overland flow is greater for higher intensities, but runoff draining into the 10 cm and 5 cm troughs is about the same for each rainfall intensity except at the very lowest application rate. This provides some evidence for a limited capacity for flow in these layers for this plot. It may be that the effects of surface ponding are not transmitted via a pressure head mechanism so readily in this plot and that flow is restricted to spatially distributed flow pathways via macropores or more permeable sections of matrix, such that a bypass flow may operate (Beven and Germann, 1990).

An example of runoff flowpaths in the acrotelm resulting in bypassing of a layer of soil is shown in Figure 5.13 for a 9 mm hr^{-1} run on bare plot 5. Here no flow occurred between 1- 5 cm depth from the plot. Instead, more of the infiltrating water was found to runoff laterally between 5 and 10 cm such that once infiltration had occurred, vertical percolation allowed water to be channelled down to the layers below 5 cm. Clearly at a depth between 5 and 10 cm almost all of the flow was then diverted laterally out of the plot as nearly all of the infiltrated water was collected from this layer at steady-state. Figure 5.13 shows that the rise in flux from the 10 cm layer was greater than the surface flux such that for the first 30 minutes of the run lateral runoff at 10 cm exceeded surface runoff. Here it may be that macropores within the near surface layers of the bare peat (perhaps produced as a product of weathering) result in rapid percolation of water down to the deeper layer. As these macropores, cracks or more permeable matrix sections have a limited flow capacity, once this flow level is exceeded the bypassing flowpaths fill up and infiltration rate falls such that surface flow can occur.

On average, for all runs, steady state runoff is reached in 59 minutes, and this time is reduced for higher intensities (Table 5.6). The time to steady state is about the same for all soil layers which suggests that there is a close connection between the lateral flow production and surface flow production. This may be related to the development of saturation-excess OLF and a head of water at the surface and hence the time to maximum ponding (and surface flow) coincides with the time to steady flow from layers at depth.

Recession limb characteristics were only measured for 8 plots, and here the mean time from the end of rainfall to runoff falling below 0.1 mm hr^{-1} was achieved in 25 minutes on average at the surface, 29 minutes at 5 cm and 33 minutes from the 10 cm layer. Given that runoff volumes were lower to start with at depth, then the recession gradients were even greater for surface flow.

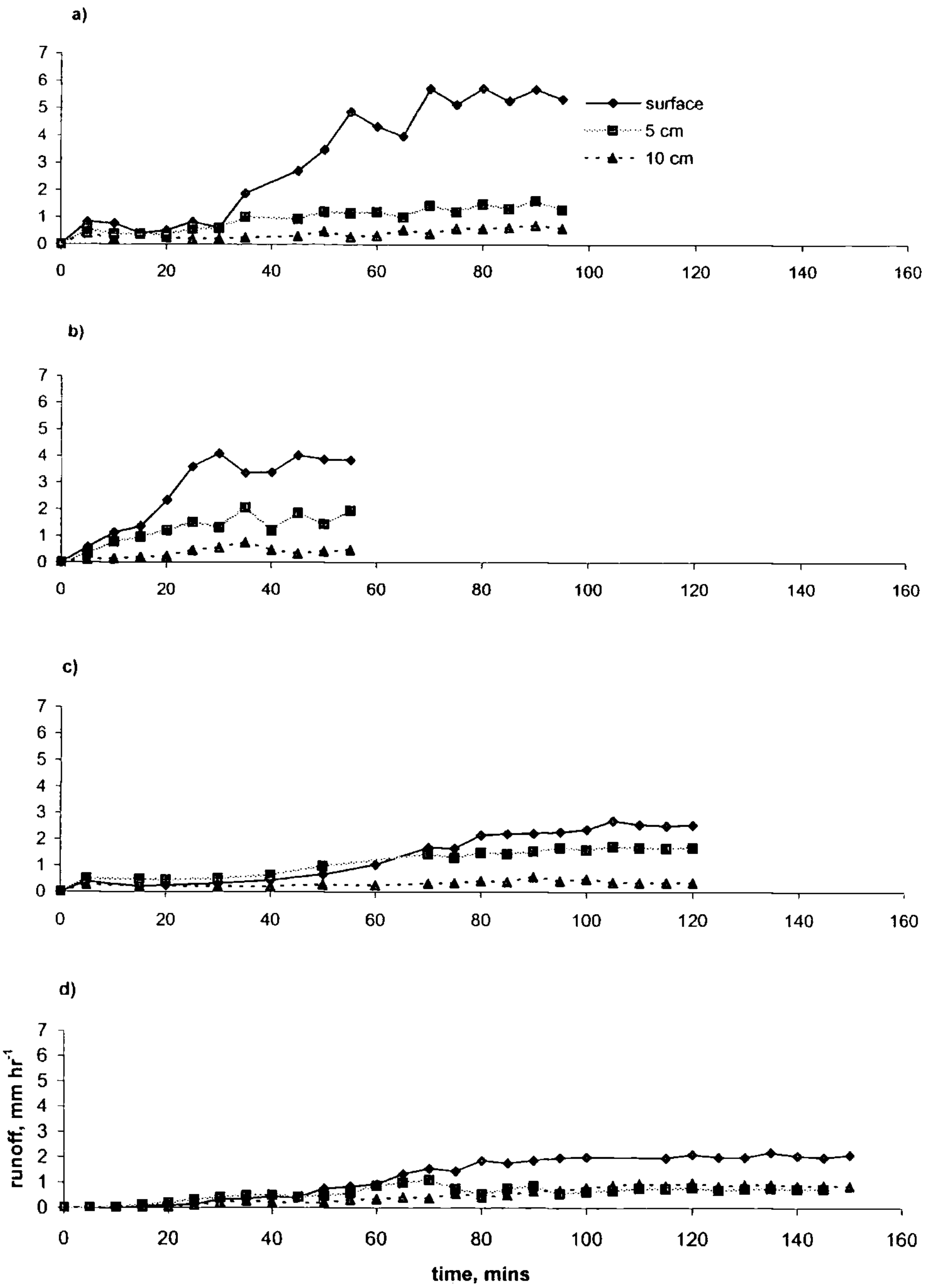


Figure 5.12. Runoff production from field plot S1 for four simulated rainfall events, a) 12 mm hr^{-1}, b) 9 mm hr^{-1}, c) 6 mm hr^{-1}, d) 3 mm hr^{-1}

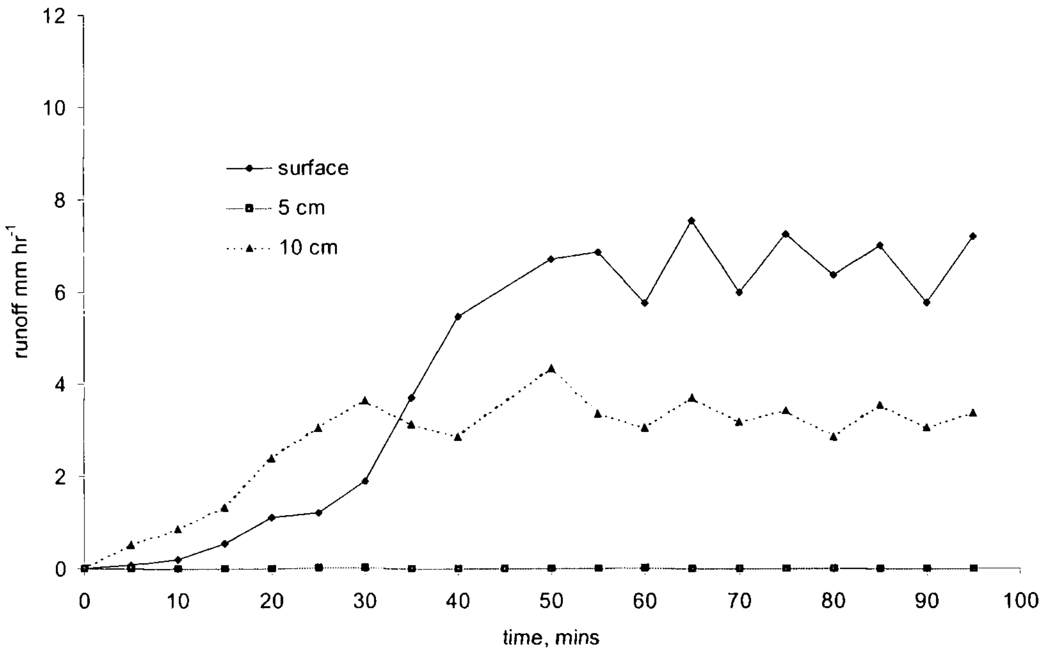


Figure 5.13. Runoff production from field plot B5 during a 9 mm hr^{-1} rainfall simulation event.

Table 5.6 Means, standard deviations and frequencies (top to bottom of each row) of time to steady-state runoff production, minutes

Depth, cm	Rainfall intensity, mm hr^{-1}				
	3	6	9	12	Total
0	70	76	50	45	60
	24	19	20	40	34
	21	22	22	23	88
5	67	74	49	40	58
	29	18	17	26	31
	21	21	21	21	84
10	57	81	45	50	58
	30	26	21	48	38
	13	13	14	15	55
Total	69	77	49	45	59
	24	25	34	30	34
	55	56	57	59	227

As mentioned above, each plot experienced an initial rainfall event, after which it was left to drain for at least 2 hours before a rainfall event was simulated at a different intensity on the same plot. Moisture conditions may have changed within the peat such that the first event affects time to peak for subsequent events. An examination of the antecedence effect on time to steady-state shows that previously 'dry' plots take between 30-25 minutes longer on average to reach steady-state runoff, than those plots which have already had a simulated rainfall event on them (Table 5.7). Unlike Bowyer-Bower (1993) who found that initial moisture status could affect steady-state runoff values in semi-arid soils, no antecedence effect could be determined for runoff rates in the blanket peat investigated during spring 1999. However, there may be an important seasonality (see below) to the processes.

Table 5.7 Antecedence effect on time to steady-state runoff, mean for all layers, minutes

Antecedence event	Depth		
	0 cm	5 cm	10 cm
'dry' events	82	75	78
'wet' events	53	50	55
Total	60	58	58

Core samples were taken from four plots immediately after rainfall. Cores were also taken immediately adjacent to the plots to provide a comparison for the peat with no applied rain. Although there may be small-scale spatial variation, some information may be gleaned from analysis. Each core was segmented with a sharp knife into 3 cm sections in the field and placed into air sealed bags. This prevented any chance of changes in moisture within the core occurring due to drainage and leakage along the walls of the coring tube. Cores from two plots immediately after rainfall indicate that peat blocks may undergo a change in moisture content in their upper layers such that some of the incoming rainwater is taken up by the peat mass (Figure 5.14). Moisture within the upper 10 cm of peat increases with rainfall for B1 and E5 such that some of the infiltrating rainwater is absorbed into the acrotelm and was not recorded as runoff. For B1 there are moisture changes down to 14 cm, which indicates that percolation has taken place to this depth (if the exterior core is fully representative of the interior core). There is 4.8 mm of unaccounted flow for this plot at steady-state (Table 5.8). This suggests that percolation has not just occurred to increase the peat moisture content, raise the water table and lead to saturated conditions whereupon percolation to this depth ceases; moreover, percolation is occurring to a deeper level than the runoff troughs at steady-state. For E5 at steady-state only 0.1 mm hr^{-1} is unaccounted for. The moisture curve indicates that there is no change in the moisture content of the peat below about 10 cm. Here initial rainfall has been absorbed into the acrotelm but as percolation below 10 cm is limited, almost all of the runoff is collected in the throughflow troughs. For E1 and B3 there is very little discernible difference in moisture between the cores before and after rainfall application.

Table 5.8 Steady-state runoff at 12 mm hr^{-1} rainfall intensity for four field plots

Depth	B1	B3	E1	E5
Surface	3.5	4.6	7.8	4.3
5 cm	3.6	3.1	3.1	7.6
10 cm	0.1	3.2	0.0	0.0
Total collected	7.2	10.9	10.9	11.9

5.3.3.4. Sediment movement

Vegetated peat produced very little sediment, such that measurement from test runs frequently produced no measurable sediment. It was thus decided that resources would be better spent on examining sediment removal from bare plots which proved to be easier to measure. Rain splash was an important agent of disturbance and entrainment as often particles were splashed up to the top of the plot boundary boards (10 cm in height) and up to 15 cm against the rainfall simulator legs and on the internal sides of the wind proofing around the plot. Suspended sediment concentration was measured during the 24 runs on bare peat. The sudden decline in sediment concentration seen coinciding with rainfall cessation provides evidence for the strong erosional role of rain splash. This was seen earlier for example in Figure 5.11. As runoff decreases rapidly after rainfall cessation however, the effect is combined with transport limiting flow reduction. Surface wash was observed to be the main agent of transport with individual particles and fibres of peat easily observed as moving in micro-rills and the sediment supply to the runoff troughs was oscillatory in nature related to micro-pool and micro-rill, cut and fill processes. The supply of available sediment was in most cases found to be limited as concentrations decreased during a simulation run, very quickly at first, and more slowly later, typically producing clockwise hysterisis loops (e.g. Figure 5.15). This effect may also be related to the development of a pool of surface water which attenuates the erosive power of rain (Klove, 1998). The trends in sediment

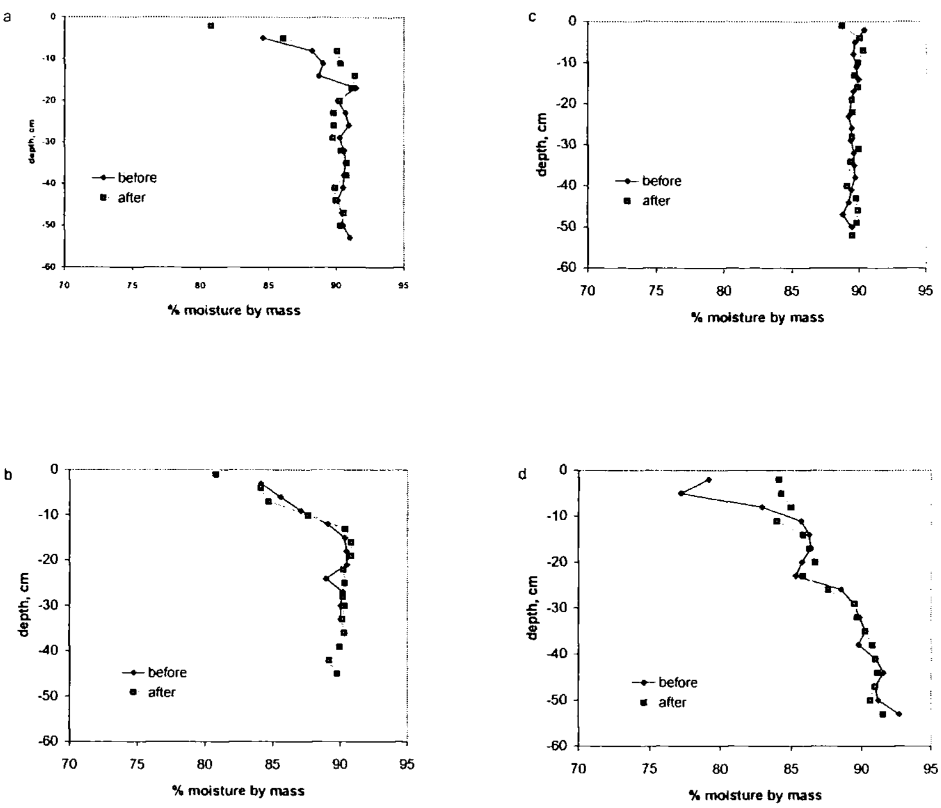


Figure 5.14. Moisture status of peat cores taken from within rainfall simulator plots after rainfall ('after') and just next to rainfall simulator plots as a surrogate for pre-event status ('before'), a) B1, b) B3, c) E1, d) E5

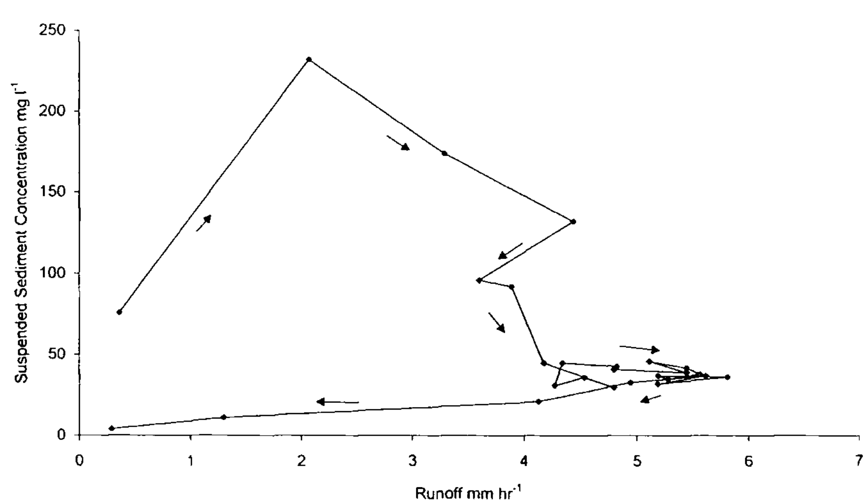


Figure 5.15. Hysteresis plot of suspended sediment concentration against surface runoff for field plot B2, during a 12 mm hr^{-1} rainfall event (runoff production from the plot during this run is shown in Figure 5.11).

concentration during the runs are similar to those reported by Klove (1998) who examined a degraded mined peat surface in Finland.

Peak concentrations were generally recorded during the rising limb of surface hydrographs and are illustrated in Table 5.9 with mean concentrations over the length of the run also shown. Values of peak concentration of sediment were found to be between 33 to 3852 mg l^{-1}, generally increasing with intensity on a particular plot. These values are somewhat lower than that found by Labadz (1988) who found peak values ranging from 947 to 9110 mg l^{-1}. However, the runs of Labadz were at very high intensities (39 to 92 mm hr^{-1}) and hence a greater total raindrop impact energy would have been supplied to the peat surface allowing increased detachment and entrainment. Bare peat areas are frequently surrounded by vegetation, such that sediment may become trapped, so that sediment yield measured at a particular outlet is therefore a result of differential production, storage and deposition within the catchment (Walling, 1983). For the Rough Sike catchment, within which many of the rainfall simulator tests were performed, Crisp (1966) estimated an annual sediment yield of 93 tonnes with an estimate for about 10-20% of the catchment as eroding. In reality though this eroding peat is concentrated in gullies corresponding to Bower's (1961) late stage of development. Re-instrumentation of Crisp's weir has provided evidence for a strong sediment relationship on the rising limb of the hydrographs (Burt *et al.*, 2000) with most of the peak concentrations occurring during this time. The relationship is indicative of sediment exhaustion whereby the supply of readily mobilised material is quickly depleted (Webb and Walling, 1984). The rainfall plot studies reflect these catchment-scale data. Peak suspended sediment concentrations from the catchment outlet were found to be around 50-60 mg l^{-1} (Evans and Burt, 1998) with peat fans at the end of gully networks playing a large role in mediating the connection between the main channel and the eroding source areas. The rainfall simulator tests were, of course, performed on isolated disconnected plots. On a hillslope, sediment will be transported, deposited and stored several times and supply of sediment from upslope will be an important feature. The mean concentrations of suspended sediment produced on bare peat plots indicate that loads are spatially highly variable. Over the 3-12 mm hr^{-1} range the mean suspended sediment concentration was 224 mg l^{-1}. Although this value is not huge, the very low density of the peat (0.1 g cm^{-1} compared to 1.2 g cm^{-1} for mineral soil peds and aggregates and 2.6 g cm^{-1} for typical quartz grains) means that this represents a significant volumetric load (Burt *et al.*, 1997).

Burt and Gardiner (1984) note the importance of desiccation in creating a peat surface that can provide high sediment loading. Because of this sediment loading may vary with aspect (Bower, 1959; Francis, 1990) and with season (Tallis, 1975: Francis, 1990). It may be that the value quoted above is at the lower end of the scale when winter frost activity has been reduced and before summer desiccation has occurred. Future rainfall simulation experiments could be performed at different times of the year in order to elucidate further the effects of surface drying and needle-ice desiccation on both infiltration processes and erosion rates. Experiments could also be performed which control both wind and rainfall variables in order to investigate the erosional roles of each mechanism.

Table 5.9 Mean and peak suspended sediment concentrations from bare peat

Plot	Intensity	Infiltration rate	Time of run	Mean conc, mg l^{-1}	Peak conc, mg l^{-1}
1	3	2.71	55	61.13	98.5
	6	5.18	50	72.75	111.4
	9	7.48	50	39.25	149.2
	12*	8.51	100	73.58	210.5
2	3*	2.03	150	23.34	50.1
	6	3.94	90	43.21	77.5
	9	5.88	90	36.67	90.1
	12	6.76	90	47.65	100.9
3	3	2.57	90	21.43	33.4
	6*	2.82	125	32.13	70.7
	9	5.56	90	36.43	100.2
	12	7.45	140	53.28	232.0
4	3	2.15	50	204.62	635.1
	6	5.14	95	226.82	1028.6
	9*	6.08	55	1220.24	2096.1
	12	7.31	75	2377.48	3852.5
5	3	3.00	180		
	6	5.02	250	84.34	156.3
	9	2.40	200	99.85	301.2
	12*	4.40	100	149.65	555.4
6	3	2.98	180	45.12	67.2
	6*	4.70	85	72.12	111.2
	9	5.70	75	56.43	145.2
	12	7.98	120	71.22	189.4

* = first run on the plot

5.3.4. Effect of summer desiccation on runoff processes

5.3.4.1. Summer desiccation processes

Some pilot study work looking at the effect of warm summer dry periods on blanket peat was performed in the field during the summer of 1999. Out of the 24 plots examined in the spring study, 8 were revisited during a warm dry period of the summer in 1999, with two from each surface cover type. It is unusual in the blanket peat catchments of the North Pennines to experience periods of more than 10 days without rain which coincide with warm weather. Figure 5.16 illustrates the dry bulb air temperature and rainfall characteristics for 1st July 1999 to 31st August 1999. Only 19.2 mm of precipitation occurred between day 203 and 225, with only 0.2 mm of rain between day 186 and 195 and none between 206 and 216. Total precipitation for June, July and August 1999 was 294.8 mm, which compares with the mean for this time of year of 384.7 mm (mean at Moor House since 1953). Air temperatures fluctuated daily, with temperatures frequently above 20 °C.

With the dry warm weather, cracking on the surface of bare peat was observed (Figure 5.17). Measurements of crack dimensions were performed on two 1 m^2 gully plots at various stages during the summer. Maximum crack width was 28 mm and maximum depth of penetration was 145 mm. Clearly this will have important implications for the depth of penetration of supplied water and the infiltrational and runoff generating processes in operation. Crack length values could not be determined accurately because cracks tended to be continuous and often polygonal in nature (Bower, 1959; Francis, 1990), although one almost straight surface crack was measured at 702 mm in length. Mean values of crack sizes for the two plots are shown in Figure 5.18a. There is a peak in crack dimensions around day 214 which is after 10 days without rain and at a time when summer temperatures reached a maximum (see Figure 5.16). The surface took many months to recover, with cracks still evident in late summer. The cracks usually filled with re-deposited material rather than by complete re-swelling of peat. Rainfall simulation plots were revisited from Julian day 209-216 during the height of the summer surface desiccation.

5.3.4.2.Comparison of spring and summer runoff production

T-test analysis of square root data (to produce more normally distributed datasets) demonstrates that there is a significant difference between the runoff production in the plots during the original runs of the spring and the runs during the warm dry summer

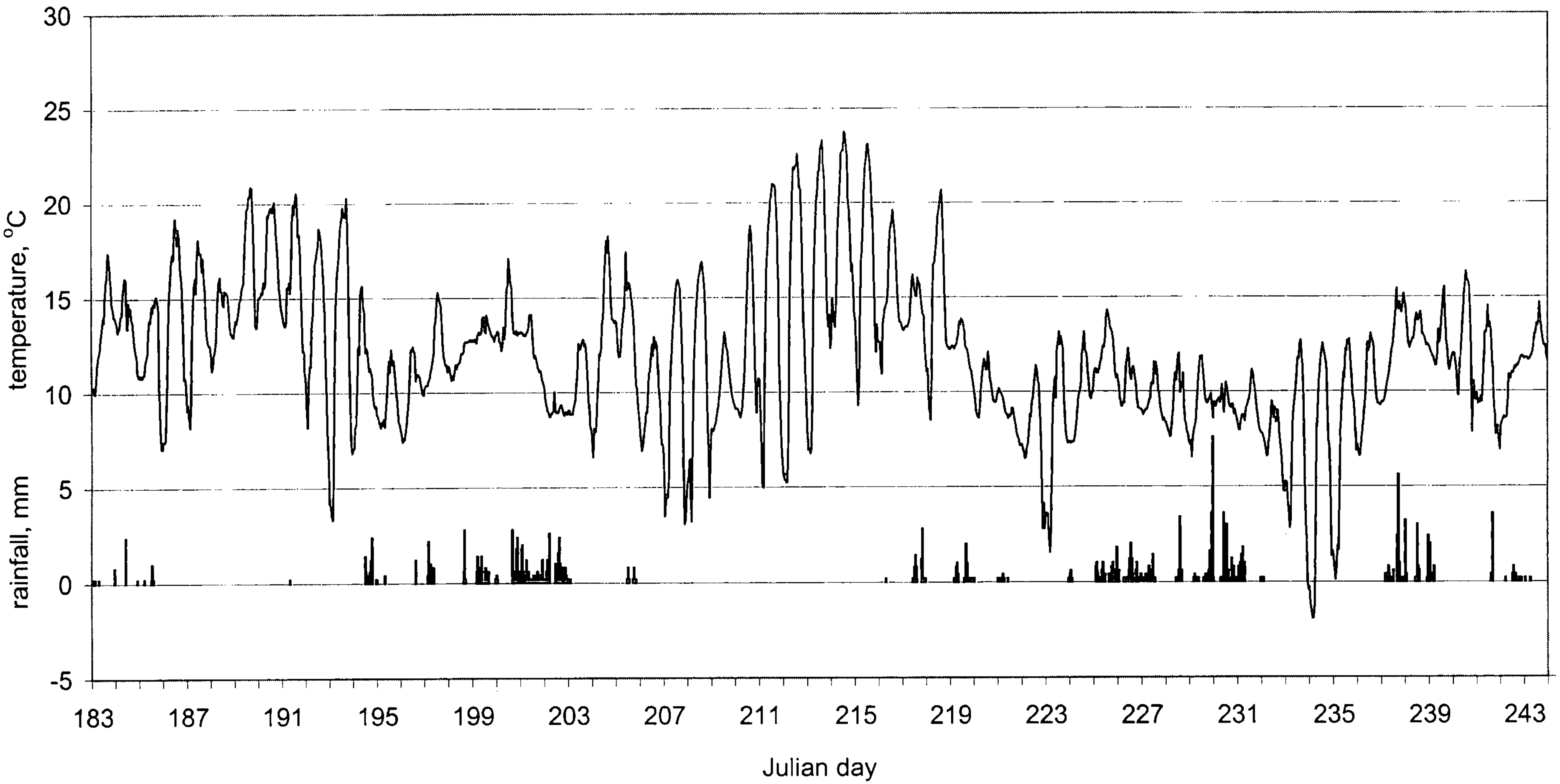
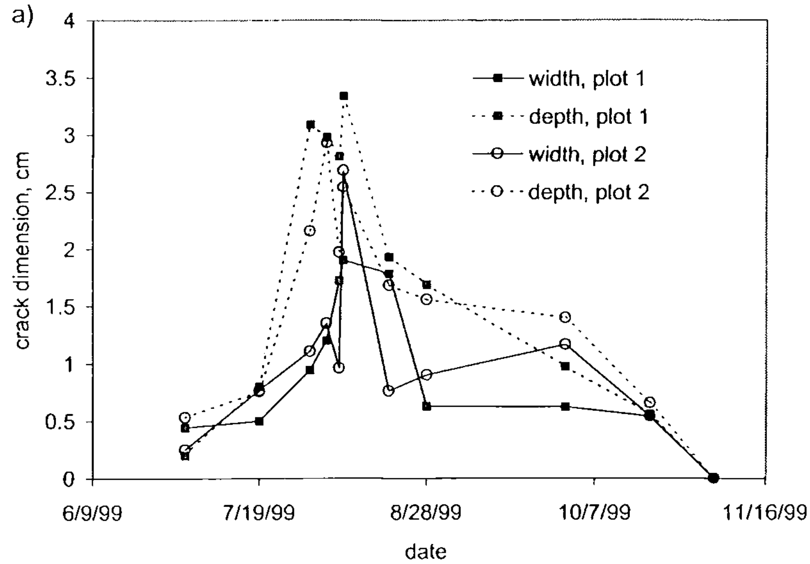


Figure 5.16. Air temperature and precipitation, Moor House, July and August, 1999.

Figure 5.17. Cracking on the surface of bare peat, Moor House, August 1999.

a)



b)

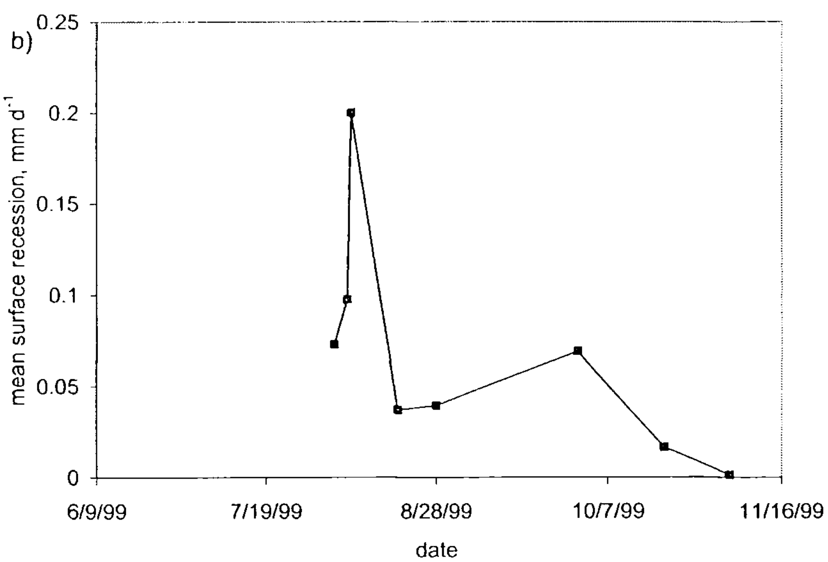


Figure 5.18. Peat desiccation on two bare 1 m^2 plots within a gully at Moor House, summer 1999, a) mean crack widths and depths, b) mean 'apparent' surface lowering based on data from 20 erosion pins.

period ($t= 2.03$, $p < 0.03$). Comparison of steady-state runoff production is shown for all layers and plots for pre- and post-drought tests in Figure 5.19. For *Sphagnum*-covered plots there is no change in the runoff production rates. The *Sphagnum* cover may protect the surface from damage. Furthermore, because *Sphagnum* species tend to grow in wetter areas which have higher water tables, such as in topographical hollows which are very poorly drained, these areas may be less likely to become very dry, unless the drought is very severe.

For one of the *Calluna* plots (C5) there is no change with most of the runoff recorded from the 5 cm layer during both the spring and summer. For the other *Calluna* plot (C2) where surface runoff is recorded, this has decreased, with a resultant increase in the 1-5 cm layer. Clearly therefore there has been an increase in the infiltration capacity of the surface layer. Similarly for both the *Eriophorum* plots, infiltration rates have increased and flow from the 5 cm trough has also increased, with no runoff recorded from the lower layer. For the bare peat plots, total steady-state runoff collected has decreased. This suggests that water is infiltrating to deeper levels than the 10 cm throughflow trough such that increased vertical percolation results in a decrease in lateral flow in the upper peat layers. For plot B2, runoff from 5 and 10 cm has decreased, with a slight increase in runoff at the surface for 3, 6, and 9 mm hr^{-1}, but at 12 mm hr^{-1} intensity infiltration rates are greater during the summer than in the spring. The variation with intensity may be related to micro-topographical crusting and crack flow dynamics. For B5, infiltration rates are greatly increased for all intensities such that surface runoff is reduced in the summer test.

Whilst there may be significant changes in interception storage between sampling periods this is unlikely to affect the results of these comparative tests. This is because rainfall was simulated until steady-state runoff had been occurring for some time. The data being compared from spring and summer are runoff values at steady-state. Hence, although the interception component may store more water at the beginning of the tests in the summer experiments than in spring, once this store is at capacity then flux through the system should not be altered. Therefore at the time when steady-state runoff values are measured a few hours into the test almost all of the rainfall will be reaching the peat surface in both spring and summer tests.

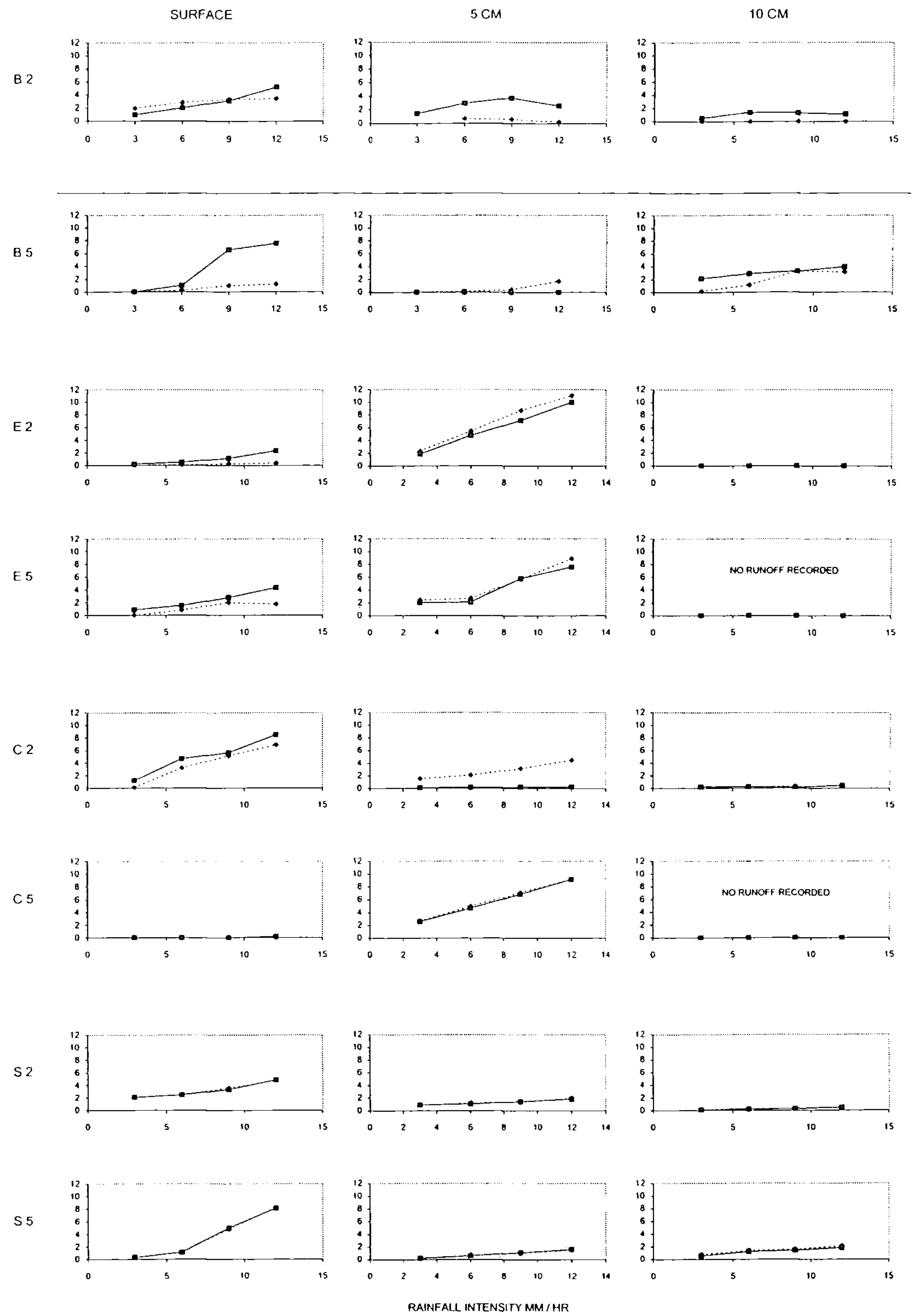


Figure 5.19. Steady-state runoff by rainfall intensity for field plots revisited during summer 1999, full trendline = spring 1999, dotted trendline = summer 1999.

Where any change has occurred, rainfall can more readily infiltrate into the peat during the summer test, than during the spring. This suggests that less overland flow may be expected after warm dry spells. Not only is this related to a lower water table and water table recharge in the first instance, but as these tests ran to steady-state such that the water table had time to rise, the evidence suggests that alteration in hydrological properties of the peat at and near the surface has occurred. It may be that matrix and macropore flow can be increased within part of the acrotelm as evidenced by the increasing fluxes measured from the 5 cm layer during the summer tests. On bare peat where crack flow was visible, some absorption into the peat was noticed before any runoff occurred. This was followed by a period of ponding in depressions on surface crusted sections. The ponds then overflowed into cracks to be channelled away. Eventually, the cracks themselves filled with water as flow capacities were exceeded and as some blockage by sediment and perhaps some re-swelling occurred. Only at this stage could overland flow be collected from the runoff trough.

5.3.4.3. Sediment movement

Sediment loading from the two bare plots was lower during the summer runs (Table 5.10). Partly this is a response to a reduction in surface runoff. Because of this and as mean and peak concentrations are lower in the summer runs, this provides more data on the seasonality of peat erosion. Figure 5.18b displays mean surface lowering from the two 1 m^2 crack measurement sites based on twenty erosion pins. This indicates that peat wastage, rather than runoff and rain splash, coincident with hot, dry weather may be an important recessional mechanism (Francis, 1990). Wind erosion could also have removed material from the dry peat surface and may in fact be many times more important than sheetwash erosion and rainsplash. Often peat groughs on the exposed moors are orientated in the direction of the prevailing wind indicative of the potentially important role of wind in erosion of blanket peat. More work needs to be done to examine these erosional processes in the uplands. However, the data shown in Figure 5.18b may be misleading since the apparent surface lowering may just be a result of expansion and contraction of the peat itself.

Observations of weir pools on the Moor House reserve, indicate that maximum supply of eroded material to streams occurs in early spring due to frost action; the weir pools had to be cleared of sediment frequently during the spring but very infrequently at other times of the year. Tallis (1975) showed that substantial peat erosion occurs during snowmelt and during heavy rain, when stream flow rates are high. Francis (1990) however found that peat supply to streams was much greater in the autumn and early winter; the suggestion was that summer desiccation had prepared the peat for removal, but as the winter progressed sediment exhaustion occurred and frost action was of minimal importance. The two opposing results may be related to sediment storage and release mechanisms, and to the nature of the coupling between bare peat areas and streams in the area of study.

Table 5.10 Comparison of suspended sediment concentration (SSC) in surface runoff from spring and summer runs on bare peat plots

Plot/ season	Rainfall intensity, mm hr^{-1}	Steady-state infiltration rate, mm hr^{-1}	Mean SSC, mg l^{-1}	Peak SSC, mg l^{-1}
2 spring	3	2.03	23.3	50.1
	6	3.94	43.2	77.5
	9	5.79	36.7	90.1
	12	5.02	47.7	100.9
2 summer	3	1.02		
	6	3.13	17.2	26.3
	9	5.64	24.2	45.1
	12	8.58	32.4	67.2
5 spring	3	3.00		
	6	5.02	84.3	156.3
	9	2.40	99.9	301.2
	12	4.40	149.7	555.4
5 summer	3	2.93		
	6	5.79	8.5	14.3
	9	8.03	15.6	23.4
	12	10.80	17.3	23.5

The 8 runs on two plots of dry peat clearly do not provide significant evidence for catchment-wide predictions and the tests are merely presented as a pilot study. It may be that the summer crusting of the surface prevents removal of sediment in the first instance, but as wetting up occurs and significant ponding and runoff begin to disturb the crust, then more sediment erosion can occur. For the dry peat runs, peak sediment concentrations were found coincident with peak runoff (Figure 5.20). Here the sediment supply appears to be transport limited. The suggestion is that summer desiccation occurs through peat wastage (a combination of biochemical oxidation, shrinkage, consolidation and compaction: Stevens and Stewart, 1976) when there is no rainfall and surface runoff. Once some wetting up of the peat has occurred, there is a time lag before sediment removal reaches a peak as supply is transport limited. This may explain the results of Francis (1990) for late autumn and early winter of 1983 and 1984 (two atypical dry years; Burt, 1985). Much more work is required on linking process mechanisms and rates of erosion in blanket peat areas.

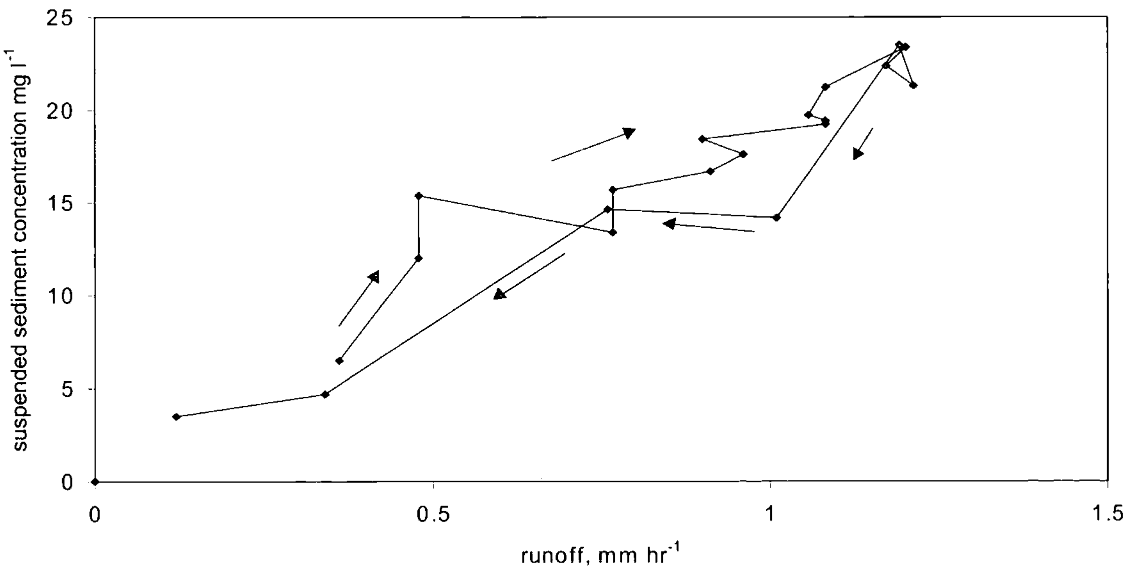


Figure 5.20. Hysteresis plot of suspended sediment concentration against runoff for a 12 mm hr^{-1} rainfall event on field plot B1, summer 1999.

5.4. Rainfall simulation on laboratory blocks of blanket peat

5.4.1. Context

In order to study the effects of drought on runoff production in blanket-peat in more detail than could be seen in the summer of 1999, a laboratory experiment was undertaken. As discussed in Chapter 4 the summer drought of 1995 was an extreme event. Comparing July and August 1995 with the predicted climate conditions for 2021 – 2050 Hulme (1998) suggests that summers as warm as 1995 will become 1 in 10 year events rather than 1 in 300 as at present. As temperatures rise and rainfall appears more variable, it is possible that flood and drought may recur more frequently in the future. For these reasons it is important to establish the possible effects of drought on blanket peat hydrology.

5.4.2. Laboratory Methods

In order to examine the effects of drought on the acrotelmic runoff generating processes in more detail than could be seen from the dry spell of summer 1999, 16 intact blocks of peat were carefully removed from the field and taken to the laboratory. The majority of laboratory rainfall simulation studies have examined reconstituted or remoulded soils, a technique that is particularly unsuitable on peat due to its fibrous and anistropic structure. 1 m x 0.5 m blocks were found too difficult to remove and transport so blocks of 0.32 m x 0.42 m (0.13 m^2) and 35 cm depth were used. Block boundaries were smeared with petroleum gelatine prior to emplacement. Four blocks of each surface cover type were used. Runoff was collected from 1 cm, 5 cm and 10 cm depth from the blocks. Rainfall was simulated at 3, 6, 9 and 12 mm on each block. This was to test whether these blocks were representative of the plots sampled in the field tests, and in order to examine in more detail infiltration and runoff pathways within the acrotelm through the use of tracers. Two of the blocks from each surface cover category were then left for four weeks in the laboratory without rainfall simulation. Laboratory temperature fluctuations in both magnitude and duration were found to be very similar in nature to those found at Moor House during extended spells of the summer of 1995. Laboratory temperatures fluctuated from 25 °C at 2 p.m. to 10 °C at 1 a.m. during December 1998 to February 1999. Environmental conditions were obviously not identical to field conditions as humidity, albedo, direct sunlight, wind effects and so on were different. Nevertheless, the temperature similarities were very convenient.

The rainwater supplied by the simulator was made up in solution to replicate mean rainwater characteristics at Moor House as determined by bi-weekly sampling of precipitation chemistry maintained by ECN. It was decided that a direct mains water supply should not be used due to the differing hydrochemistry and because mains supply to the simulator can often result in poor attainment of desired intensity (Foster *et al.*, 2000).

For the two peat blocks of each surface cover type that were not to be subject to drought conditions, two tracers were added to the Mariotte supply at very low concentrations (0.2 mg l^{-1}); Lissamine FF, and potassium bromide. The use of fluorescent dyes for identification of water flow routes in soil is well documented (Aubertin, 1971; Omoti and Wild, 1979). Trudgill (1987) recommends (cautiously) the use of Lissamine FF as an indicator of travel times and as a useful fluorescent dye for column work, particularly for soils with a high organic content and it survives in soils with low pH. Bromide has also been successfully used in travel time analysis (e.g. Nachabe *et al.*, 1999; Hoag and Price, 1995). Runoff from the blocks was collected and analysed to determine timings of breakthrough of newly applied rain. For Lissamine FF, the samples were tested using a Perkin-Elmer LS-3B fluorescence spectrometer with excitation and emission maxima set as specified by Smart and Laidlaw (1977). A linear relationship was found between concentration and fluorescence ($r^2 = 0.99$). Bromide concentrations were determined using a Dionex™ ion chromatograph.

Once runs had been completed, the peat blocks were carefully divided into five horizontal sections, 5 cm apart. Nine microsamples (5 g) and one bulk sample (500 g) were taken from the lower face of each section. The microsampling strategy is shown in Figure 5.21. Soil was extracted using saturated calcium sulphate solution (Omoti and Wild, 1979) at 1:3 soil-to-solution ratio. Blanks containing soil and deionized water, and dye solutions without soil, were also tested to correct for background fluorescence and dye loss onto polypropylene containers used in the experiments. Dye adsorption onto the containers was negligible and no correction procedure was necessary. A maximum background level equivalent to 3 $\mu g\ l^{-1}$ was recorded and subtracted from the experimental data. Microsamples were classified into two categories on the basis of optically visible structural features (Smettem and Trudgill, 1983):

Class 1: Samples containing fissures, 'macropores' and roots with a diameter >1 mm were considered as possessing a bypass capability.

Class 2: Samples with non-visible voids, generally of a uniform and smooth nature, were classified as 'matrix' and assumed to possess no bypass capability.

Some degree of operator error was unavoidable when adopting a classificatory approach but is not critical for exposition of gross differences (Smettem and Trudgill, 1983).

For the peat blocks subject to drought conditions, the same procedure was adopted with tracer input occurring in the final post drought rainfall application. These blocks were subject to post-drought rainfall at 6 and 12 mm hr^{-1}. Rainfall application was repeated on each block every day for the 6 days following initial re-wetting in order to investigate whether there is a post-drought recovery of infiltration and acrotelm runoff on blanket-peat towards pre-drought levels.

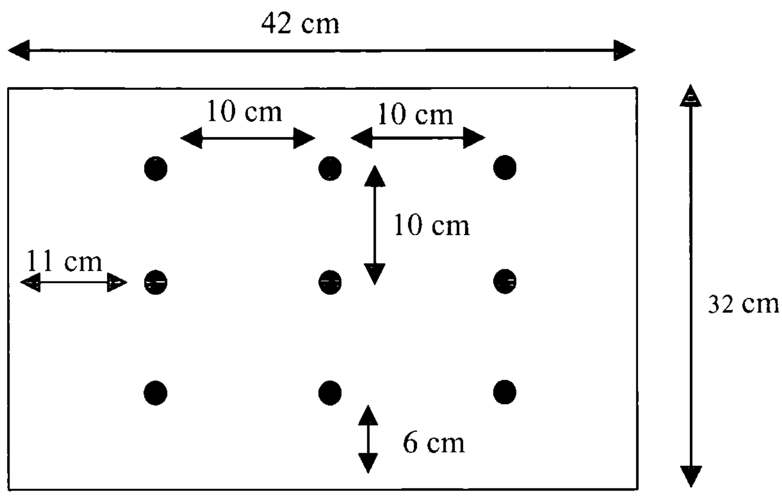


Figure 5.21. Microsampling scheme; this scheme is used at 5 cm, 10 cm, 15 cm and 20 cm depth

5.4.3. Results of laboratory analysis

5.4.3.1. Infiltration and runoff production in blocks before drought simulation

Table 5.11 presents mean results from the laboratory tests before drought simulation. Comparison of these data with those from the field (Table 5.2) shows that mean steady-state infiltration values are around 10 to 15 % lower than those measured in the field. This may be due to some form of surface compaction taking place during transit related to acrotelm distortion and blocking of macropores. Statistically, however, there is no significant difference between the two datasets ($p<0.01$). As in the field tests, infiltration rate increases with applied intensity, and the proportion of applied rain infiltrating remained around 50 %.

Table 5.11 Mean and standard deviations (top and bottom of each row) for infiltration rate, mm hr $^{-1}$ and % of applied rainfall infiltrating

	Rainfall intensity mm hr^{-1}			
	3	6	9	12
Infiltration rate mm hr^{-1} at steady-state	1.71	2.91	4.18	6.01
	0.85	1.86	2.88	3.78
% of applied rainfall infiltrating at steady-state	56.97	48.59	46.42	50.04
	28.46	31.01	32.03	31.50

For runoff data, a square root transformation was found necessary before applying ANOVA. Table 5.12 shows how depth, intensity and vegetation cover are all overwhelming controls as the calculated significance levels are less than 0.00005. The model is clearly much stronger than that for the field data with an overall R^2 of 63 % compared to the 39 % found in field tests. The depth variable is the most important control on runoff production. For runoff efficiency data, ANOVA again demonstrates similar findings to that of the field simulations such that rainfall intensity can be discarded as an influence with depth and surface cover as genuine controls (Table 5.13).

Detailed cross-tabulation of the means (Table 5.14) shows that runoff declines with depth. Less runoff is collected from the *Sphagnum* blocks than for peat with other surface cover types which suggests more water is infiltrating to depths below 10 cm. Field-based *Sphagnum* tests produced the highest amount of overland flow, yet in the laboratory tests, *Sphagnum* covered plots had the lowest mean overland flow rates. The laboratory results agree with field results from the tension infiltrometer experiments discussed in Chapter 6 (see discussion in Chapter 9).

Table 5.12. Analysis of variance of steady state runoff rates from laboratory rainfall simulator blocks, square root data.

Source	d.f.	F	Prob > F
Model	8	38.61	0.0000
Depth	2	121.5	0.0000
Intensity	3	10.66	0.0000
Vegetation cover	3	11.10	0.0000

$R^2 = 0.63$

Table 5.13. Analysis of variance of percent runoff at steady state as a proportion of incident rainfall from laboratory rainfall simulator blocks, square root data.

Source	d.f.	F	Prob > F
Model	8	37.77	0.0000
Depth	2	133.05	0.0000
Intensity	3	0.26	0.8530
Vegetation cover	3	11.38	0.0000

$R^2 = 0.62$

All steady-state runoff values for each block and layer are shown in Figure 5.22. No runoff is collected from the lowest layer of any of the *Calluna* blocks, and almost all of the input rainfall emerges at steady state from the upper two peat layers. A greater proportion of runoff is collected from the 10 cm layer beneath an *Eriophorum* cover than from below other cover types. Nevertheless, bare, *Eriophorum* and *Calluna* surfaced block responses are similar to each other and the *Sphagnum* contrast is again clear.

Figure 5.23 shows a typical runoff response from rainfall simulation on an *Eriophorum*-covered peat block. The majority of runoff occurs from the uppermost centimetre of peat, with lessening amounts with depth. A steep rise in surface runoff occurs at around 15 minutes into the experiment suggesting that either initial infiltration and absorption into the peat was rapid and declining, or that surface depression storage has now overflowed or peaked such that overland flow can now occur much more rapidly. Runoff response from lower layers is slower to react to rainfall onset in this case and only after 25 minutes does flow begin to occur. This suggests that infiltration-excess overland flow is occurring at the surface and that any infiltrating water is first absorbed into the peat before any percolation-excess lateral flow occurs. Recessional response is steep in the upper layers and shallow and subdued in the lower layers, and even after 30-40 minutes, water still drains from all layers. Mean recession times, as calculated by the time taken for flow to be reduced below 0.1 mm hr^{-1} since rain cessation, ranged from 40 minutes for the surface layer to 86 minutes for the 10 cm layer.

5.4.3.2. 'Old' and 'new' water production

a) Specific conductivity

The chemographs of specific conductivity shown in Figure 5.23. indicate that some chemical separation of runoff is possible, with surface flow having a much lower conductivity than at depth. Mean specific conductivity of the runoff produced from each layer in the laboratory tests is given in Table 5.15. For the 10 cm layer, standard deviations are low, such that conductivity varies little throughout an event. This is indicative of the slow movement of water through the lower layers of peat and the resulting longer residence times before emergence. Higher deviations are found for the upper layers, representative of larger intra-storm changes, and the importance of new water during storm events.

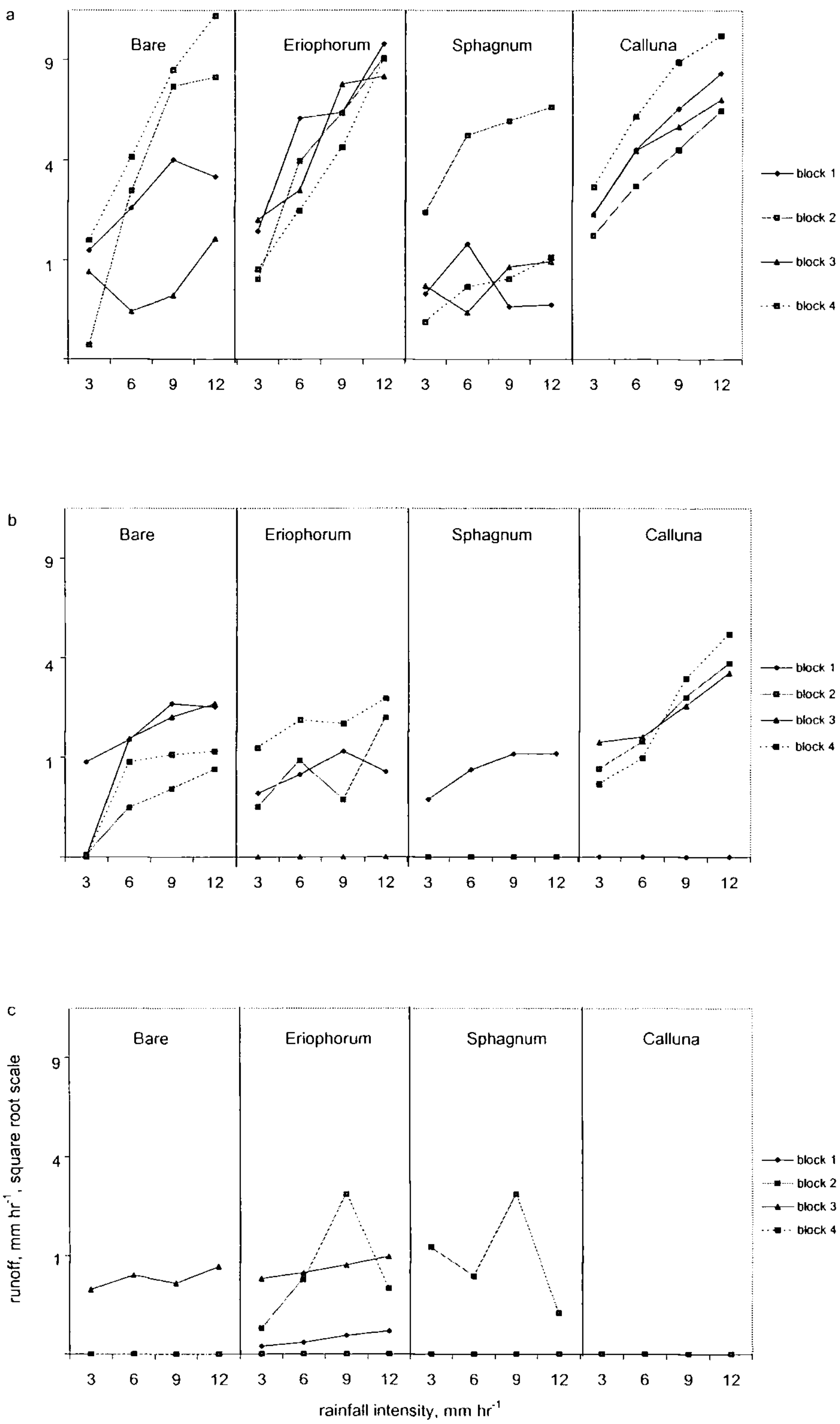


Figure 5.22. Steady-state runoff rates by vegetation type with rainfall intensity for each laboratory peat block, a) surface runoff, b) runoff at 5 cm, c) runoff at 10 cm.

Table 5.14 Mean and standard deviations (top and bottom of each row) for runoff rates for vegetation and depth groups by intensity, square root data, $\sqrt{(mm\ hr^{-1})}$ laboratory peat blocks

Vegetation	Depth	Rainfall intensity				
		3	6	9	12	Total
B	0	0.83	1.43	2.07	2.12	1.61
		0.48	0.67	1.03	0.94	0.91
	5	0.24	0.95	1.16	1.24	0.90
		0.47	0.32	0.39	0.33	0.53
	10	0.16	0.20	0.18	0.17	0.18
		0.33	0.40	0.36	0.40	0.34
	Total	0.41	0.86	1.14	1.10	0.88
		0.50	0.69	1.01	1.01	0.86
C	0	1.47	2.09	2.48	2.80	2.21
		0.20	0.28	0.37	0.33	0.58
	5	0.69	0.84	1.23	1.51	1.07
		0.49	0.57	0.83	1.01	0.75
	10	0.00	0.00	0.00	0.00	0.00
		0.00	0.00	0.00	0.00	0.00
	Total	0.72	0.98	1.24	1.44	1.09
		0.69	0.96	1.16	1.32	1.06
E	0	1.10	1.90	2.45	3.01	2.16
		0.29	0.40	0.26	0.13	0.77
	5	0.58	0.79	0.74	0.96	0.76
		0.45	0.58	0.59	0.71	0.55
	10	0.28	0.43	0.39	0.63	0.42
		0.34	0.43	0.39	0.38	0.37
	Total	0.64	1.04	1.20	1.61	1.11
		0.49	0.78	1.02	1.19	0.94
S	0	0.81	1.15	1.17	1.20	1.08
		0.47	0.78	0.83	0.72	0.66
	5	0.14	0.22	0.26	0.26	0.22
		0.29	0.44	0.52	0.52	0.40
	10	0.27	0.20	0.41	0.10	0.24
		0.54	0.40	0.81	0.21	0.49
	Total	0.41	0.52	0.61	1.16	0.52
		0.51	0.69	0.78	1.12	0.66
Total	0	1.05	1.64	2.04	2.28	1.76
		0.44	0.64	0.83	0.92	0.85
	5	0.41	0.70	0.84	0.99	0.74
		0.45	0.52	0.70	0.78	0.64
	10	0.18	0.21	0.24	0.20	0.21
		0.34	0.35	0.47	0.35	0.37
	Total	0.55	0.85	1.04	1.16	0.90
		0.55	0.79	1.00	1.12	0.92

Table 5.15 Specific conductivity of runoff waters from peat blocks subject to simulated rainfall, μ S cm^{-1}, input rainfall at 31.1μ S cm^{-1}

Layer	Mean conductivity	Mean standard deviation	Mean maximum conductivity	Mean minimum conductivity
Surface	56.5	16.4	103.9	42.6
5 cm	100.2	10.2	106.8	94.5
10 cm	103.1	7.4	112.1	100.4

The low overall conductivity values are typical of the values found in blanket peat catchments and is representative of low soil residence times and the nutrient deficient status of blanket peat (Moore and Bellamy, 1974). Goreham (1956) sampled waters from the Moor House reserve during dry weather in May 1954. Here conductivity ranged from 40 to 89 μS cm^{-1} in pools and drains. During wet weather in August 1954, conductivity ranged from 18 to 20 μS cm^{-1}, with Trout Beck at 25 μS cm^{-1}. Cryer (1980) reported values of specific conductivity for a peat-covered area in mid-Wales, with mean conductivities of 80 μS cm^{-1} for peat matrix flow and 35 μS cm^{-1} for stream water and pipeflow. Stream sampling indicates that lower levels of conductivity are coincident with high discharge (Burt and Gardiner, 1984) and rainfall simulator induced runoff follows this trend with a clear clockwise hysterisis in the chemographs (e.g. Figure 5.24). This, of course, is unusual and indicates and exhaustion effect within the peat. Indeed it is likely that much of the OLF produced on the plot at first is return flow produced at the lower end of the plot as the peat becomes saturated. Then, as further ponding develops, the OLF is diluted with rainwater.

The evidence from the simulator runs suggests that old water in the acrotelm is pushed out during a rainfall event such that when the rainfall commences conductivity is higher due to the longer residence time of the water and as rainfall progresses, the conductivity decreases as new water is provided. When supply is switched off, conductivity in all layers begins to rise again, with water draining that has had greater contact time with the peat mass such that time for diffusion into the mobile water is increased.

b) The use of tracers

The variation in Lissamine FF and bromide concentration recorded from runoff samples from a 12 mm hr^{-1} run on S1 are shown in Figure 5.25. For surface flow, bromide and Lissamine FF concentration levels and variations are very similar. However, for the

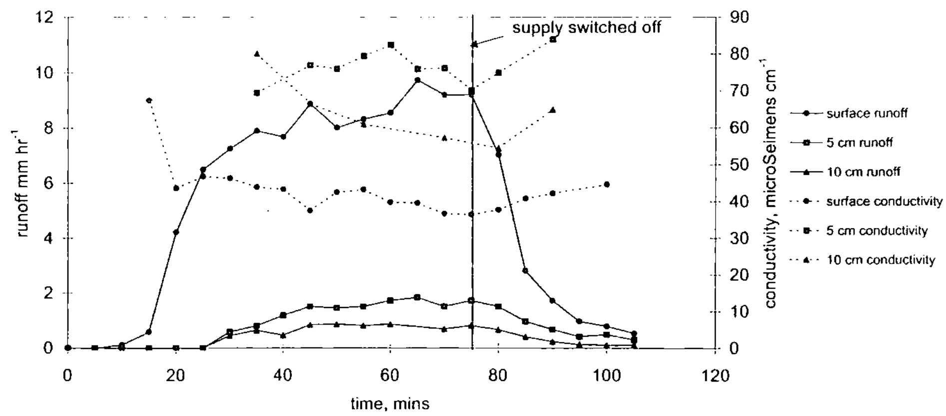


Figure 5.23. Runoff response and specific conductivity for laboratory black E2, rainfall at 12 mm hr^{-1}, conductivity of rainwater 31.1 micro Seimens cm^{-1}

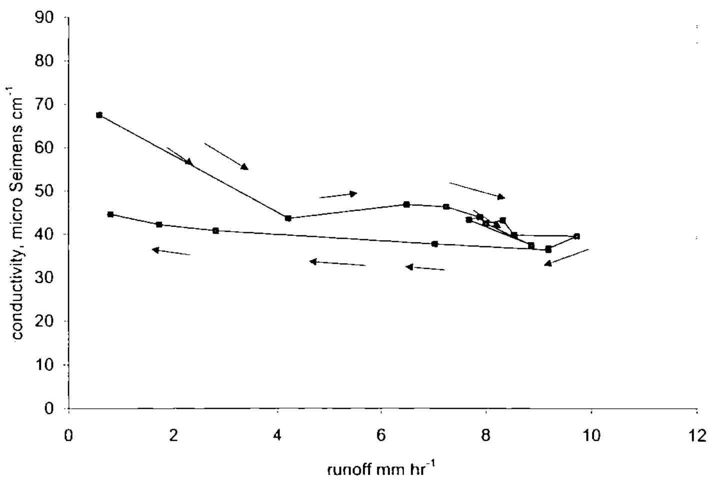


Figure 5.24. Hysteresis plot of specific conductivity against discharge for surface layer E2 during a 12 mm hr^{-1} rainfall event

deeper layer, Lissamine FF concentration is subdued compared to the breakthrough of bromide to near input value (0.2 mg l^{-1}), after two hours. This indicates that Lissamine FF is more suitable for identification of flowpaths within the soil as it is adsorbed more readily than bromide. Hence extraction of soil should allow preferential flow paths to be assessed. For estimating the timing of arrival of the newly applied rainfall to runoff and the amount of mobile water pushed out of the blocks by incoming new water however, bromide acts as a more suitable tracer simply because it is less readily adsorbed. If there is some adsorption then the figures given for old water contributions to runoff will be overestimates. Importantly soil sampling indicated low levels of bromide, with no significant difference from background levels detected.

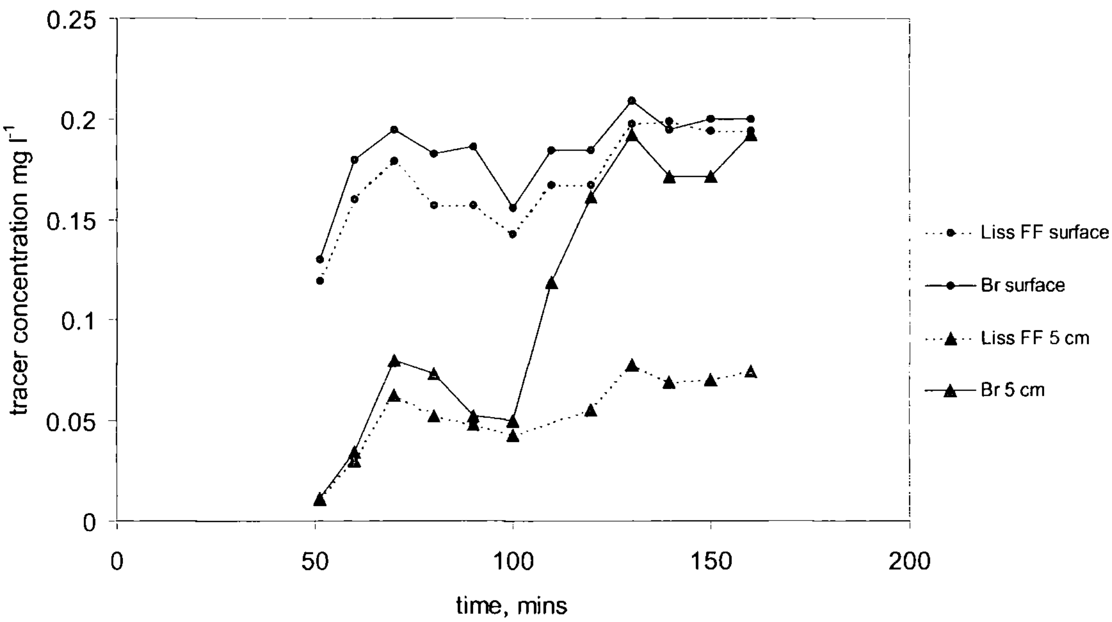


Figure 5.25. Breakthrough of Lissamine FF and Bromide in runoff from laboratory peat block S1, with rainfall at 12 mm hr^{-1}

Following Pilgrim *et al.* (1979) who used specific conductance to perform a hydrograph separation, since the input bromide concentration is known, and if no adsorption is assumed, then old water contribution (Q_o) to total discharge (Q_t) can be calculated from equation 5.1:

$$Q_o = Q_t (C_t - C_n) / (C_o - C_n) \qquad [5.1]$$

where C_t is the concentration of bromide at a given time, C_n the applied concentration of bromide in the rainfall and C_o is the background concentration.

For the 8 blocks tested the estimated mean depth of old water produced as OLF was 1.03 mm (standard deviation = 0.70), and from 1-5 cm was 0.78 mm (standard deviation = 0.75). No data were available for the lower layer; where flow occurred bromide concentrations never reached a steady C_n over the given time of runoff collection. These data are based on the volume of peat sampled by a runoff trough. However, the data indicate that OLF production is a result of saturation-excess return flow combined with fresh rainfall once the surface becomes saturated. There may also be some contribution from old water in surface pools and on vegetation stems and leaves. Less old water is flushed out of the 5 cm layer per cm per unit volume of peat, although overall flux of old water is greater.

A 12 mm hr^{-1} rainfall simulation is shown on block E4 in Figure 5.26. Here no flow was recorded from the 10 cm runoff trough. Steady-state was achieved in around 105 minutes from the other two monitored layers. Steady-state runoff was 9.1 mm for the surface layer with 2.5 mm for the 5 cm trough. After around the same amount of time bromide levels approximately reached C_n, although slightly lower concentrations are recorded from the 5 cm trough than the surface trough, suggesting a minor amount of loss. The estimated amount of old water available for runoff mobilisation can be seen to decrease over time, presumably as old water is pushed out of the block and is replaced with new water (Figure 5.26b). Over the first centimetre 2.3 mm of old water was mobilised, and for 1 - 5 cm, 0.8 mm was released.

Response from a 9 mm hr^{-1} simulation on block E2 is indicated in Figure 5.27. Here surface runoff reaches steady-state after around 45 minutes (6.1 mm hr-1), with old water contributions (1.1 mm) ending after the first 40 minutes for this layer. Initially, runoff is a mixture of old and new water, but at 20 minutes almost all of the runoff from the first cm of peat is old water. After this, the remaining old water is flushed out rapidly. Results from the lower layer indicate that runoff production may be a complex process involving mixing and interaction of flows at depth. Initially, bromide levels rise rapidly indicating a quick channelling of new water out of the 10 cm trough. Later, after around 50 minutes, bromide concentrations fall, and old water is now being flushed through the system in greater quantities. This trend may be indicative of some form of preferential flow with a strong link to the surface, dominating runoff from the lower layer over the first 50 minutes of the run. Later, as the slower response of matrix flow begins to dominate, this allows a larger volume of the peat mass to act as a source area

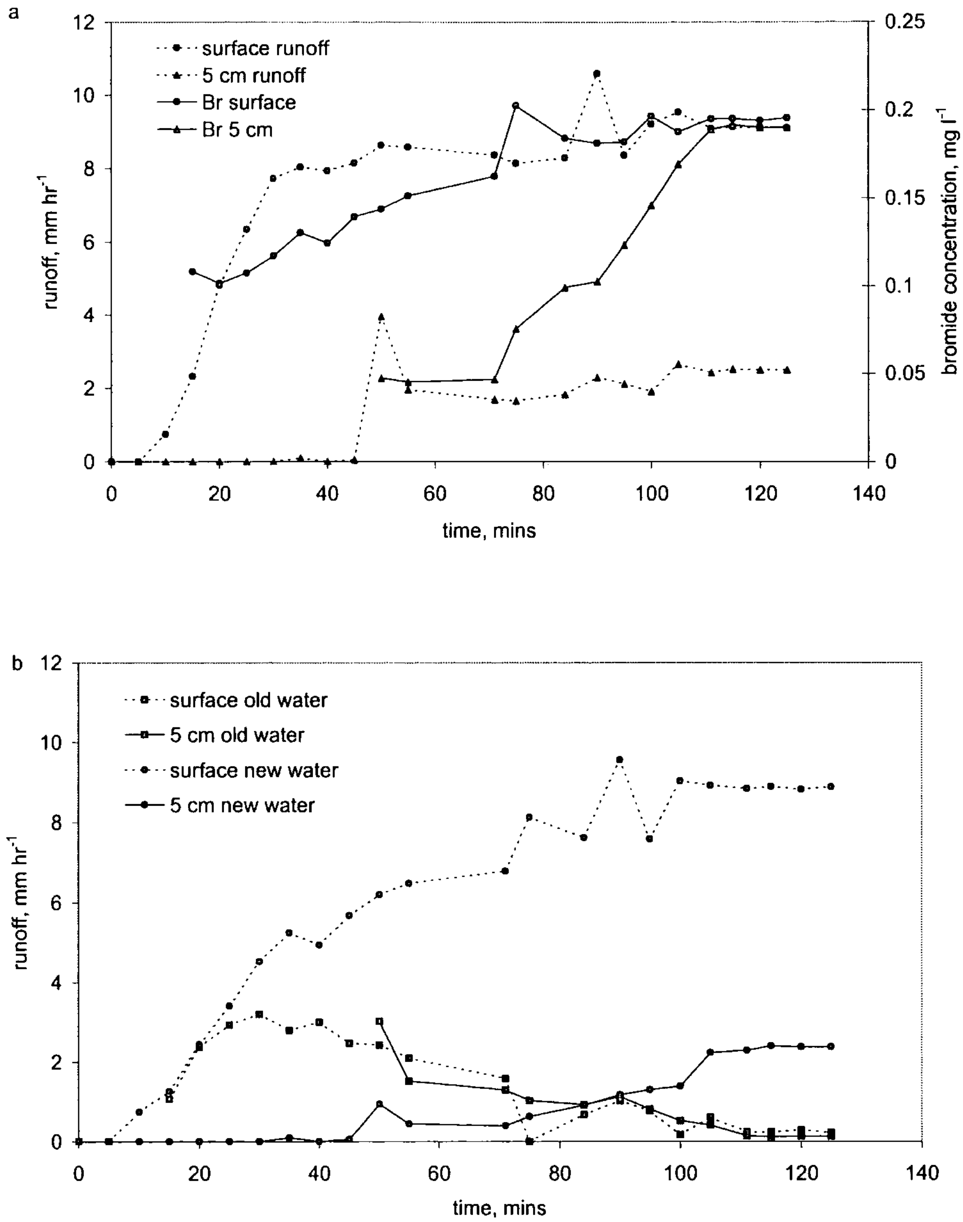
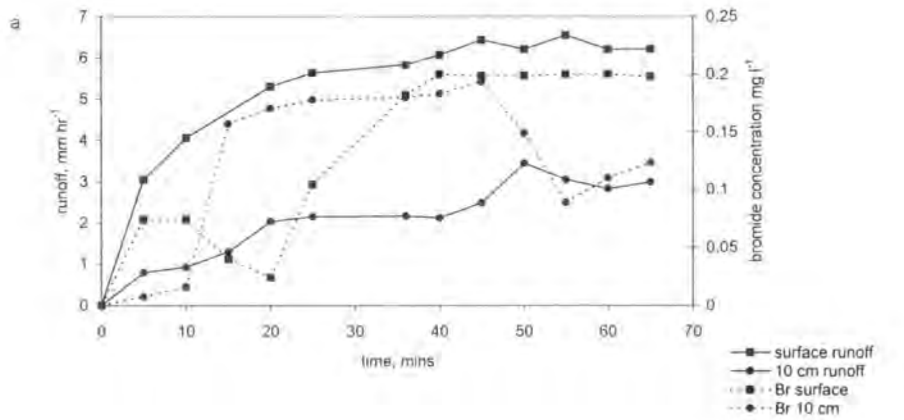


Figure 5.26. Runoff production from laboratory block E4 during a 12 mm hr^{-1} rainfall simulation, a) recorded runoff and bromide concentration, b) calculated old and new water contributions to flow



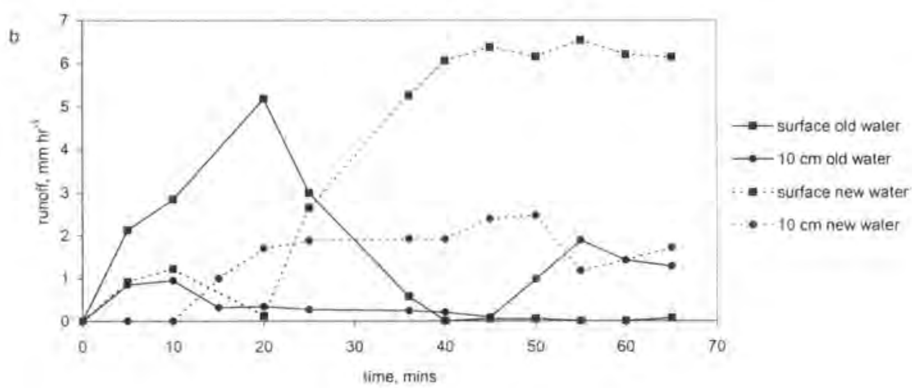


Figure 5.27. Runoff production from laboratory block E2 during a 9 mm hr^{-1} rainfall simulation, a) recorded runoff and bromide concentration, b) calculated old and new water contributions to flow

for runoff production. More evidence for preferential flow in the peat blocks comes from Lissamine sampling of soil (see below), and from tension-infiltrometer experiments on blanket-peat (Chapter 6).

5.4.3.3. The effects of drought on runoff generation and infiltration

After 4 weeks without rain, the 8 blocks subject to drought were re-wetted at 6 mm and 12 mm hr^{-1} intensities. These two runs were repeated each day for 6 more days. Steady-state runoff from the peat blocks subject to drought conditions is indicated in Figure 5.28 with pre-drought values indicated for comparison. Runoff from the initial rewetting on the first day of post-drought testing and values from the last post-drought test on the seventh day since re-applying rainfall are shown.

On post-drought day 1 all blocks display a decrease in surface runoff over pre-drought values, except for S3. This is an important finding and adds weight to the pilot study performed during the field in summer 1999. Rainfall is added to the blocks until steady-state is reached. Therefore increased infiltration rates are not simply a result of the initial dry antecedence but time is given for the blocks to wet up as much as they are capable of doing so. Immediately following a drought, infiltration is likely to be increased and more subsurface flow will result. This may take place through the increased number of connected shrinkage cracks within the peat which encourages lateral flow in the upper few centimetres of peat. By day 7 OLF is greater than on day 1, and values are generally closer to pre-drought levels demonstrating that some recovery of the peat blocks has occurred. At 5 cm depth, there has been a decrease in runoff from the two bare peat blocks, with smaller changes in the *Calluna* and *Eriophorum* blocks. At 10 cm only the *Eriophorum* blocks and one *Sphagnum* block recorded runoff. For E1 a large increase in runoff at this depth can be seen. Here increased infiltration and percolation has resulted in increased lateral flow to the 10 cm trough. This percolation appears to be reduced by the seventh day of re-wetting as runoff from the 10 cm layer has reduced and runoff from the upper two layers increased. The amount of rainfall collected as runoff from the upper 10 cm of the peat is lower after the drought for 6 of the 8 blocks, even after 7 days of re-wetting (Table 5.16).

Hence, it would appear that much more of the applied rainfall is infiltrating deeper into the peat than before the drought. It is difficult to say whether this would be a permanent post-drought feature of the blocks, but the 7 day post-drought runoff trend (Figure 5.29)

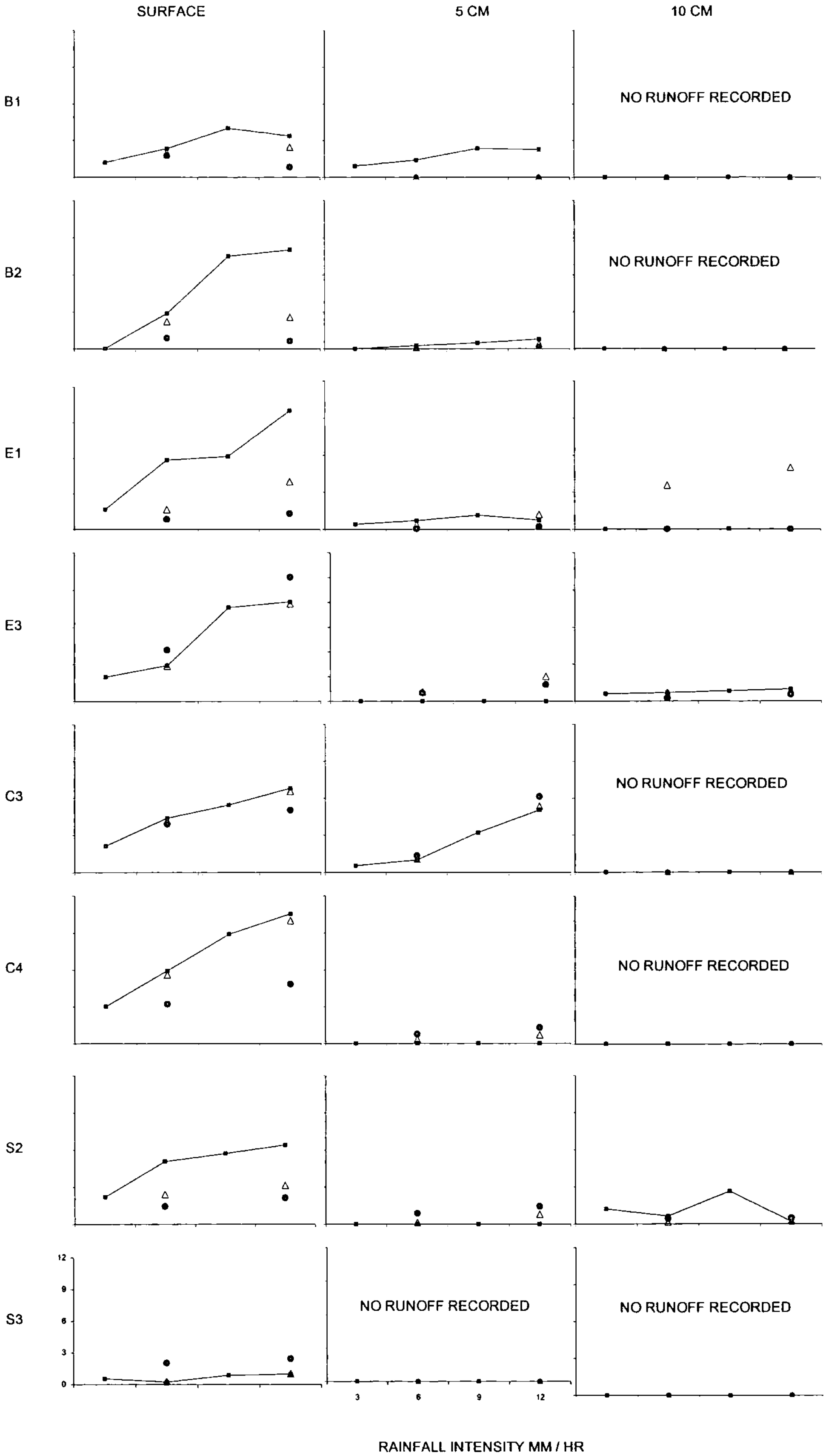


Figure 5.28. Steady-state runoff production from peat blocks for pre and post drought simulation, squares = pre drought runoff, circles post drought day 1, triangles post drought day 7

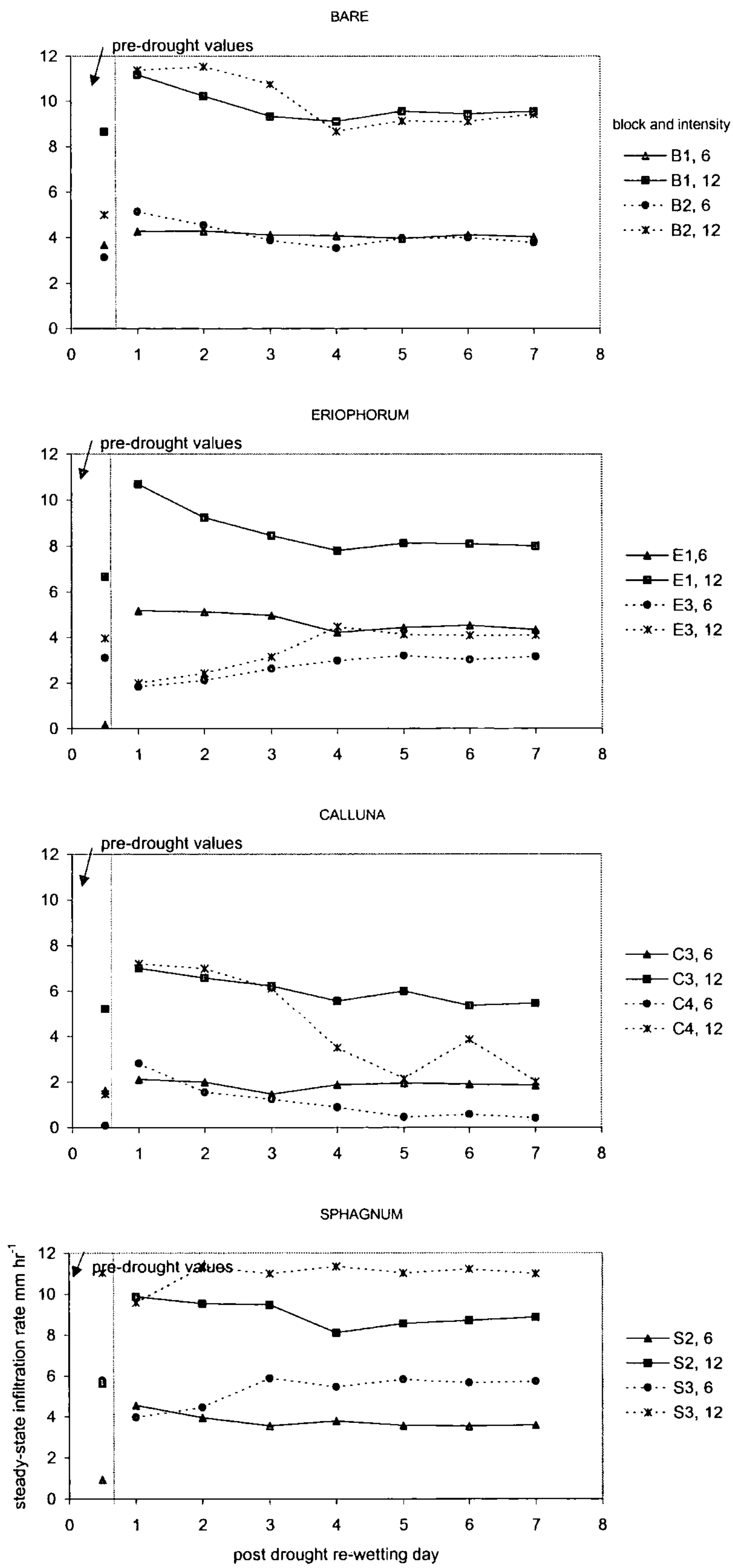


Figure 5.29. Changes in steady-state infiltration rates over 7 post-drought rewetting days

does indicate some semi-permanent change at least. Generally there is some 'recovery' in infiltration rates back towards original values over the 7 days, although in many cases the recovery is not total such that infiltration is still higher after 7 days of rain than before the drought. In many cases, recovery is more pronounced during the first 3 - 4 days, after which there appears to be a stabilisation. This trend would suggest some permanence surrounding the changes in infiltrational and runoff properties of the surface layers of blanket-peat. If this is the case then recurrent droughts in blanket peat catchments may have a major impact on the runoff-generating processes. This may in turn effect the hydrochemical and erosive processes operating within these upland catchments and more work will be required to establish the direct linkages involved.

Table 5.16 Percent of input rainfall not collected as runoff at steady-state, laboratory peat blocks

Non-drought		Drought	
Block	% uncollected	Block	% uncollected
E2	4.1	E1	15.1
E4	3.4	E3	9.7
B3	65.3	B1	79.6
B4	0	B2	75.5
S1	88.7	S2	65.7
S4	91.2	S3	91.5
C1	0	C3	11.2
C2	19.8	C4	1.0

n = 8 in both cases

5.4.3.4. The role of macropores in runoff generation before and after drought: implications of a qualitative assessment

The microsampling and bulk sampling strategy for flowpath separation discussed in section 5.4.2 was used on all 16 blocks of peat (8 non-drought and 8 post-drought blocks). Depth of penetration of applied rainwater as indicated by recovery from bulk samples is shown in Figure 5.30. Only 3 of the 8 non-drought blocks showed any signs that applied rainfall infiltrated deeper than 10 cm. These were the two *Sphagnum* blocks and B3. As Table 5.16 indicates these are the three blocks for which total runoff collected from the upper 10 cm was found to be low. It is clear that for these three blocks infiltrating water reaches a depth of at least 20 cm into the peat. For the other five blocks however, all runoff production appears to take place within the top 10 cm of the peat mass. For the peat blocks which had been subject to drought the depth of penetration is greater than in blocks not subject to drought conditions. Hence, a larger proportion of runoff can originate from deeper within the peat mass than before. This is not simply an effect of newly applied rainfall penetrating deeply after a drought because there had been 6 days of re-wetting events before the tracer was applied. The effect appears to be more long-term than a week.

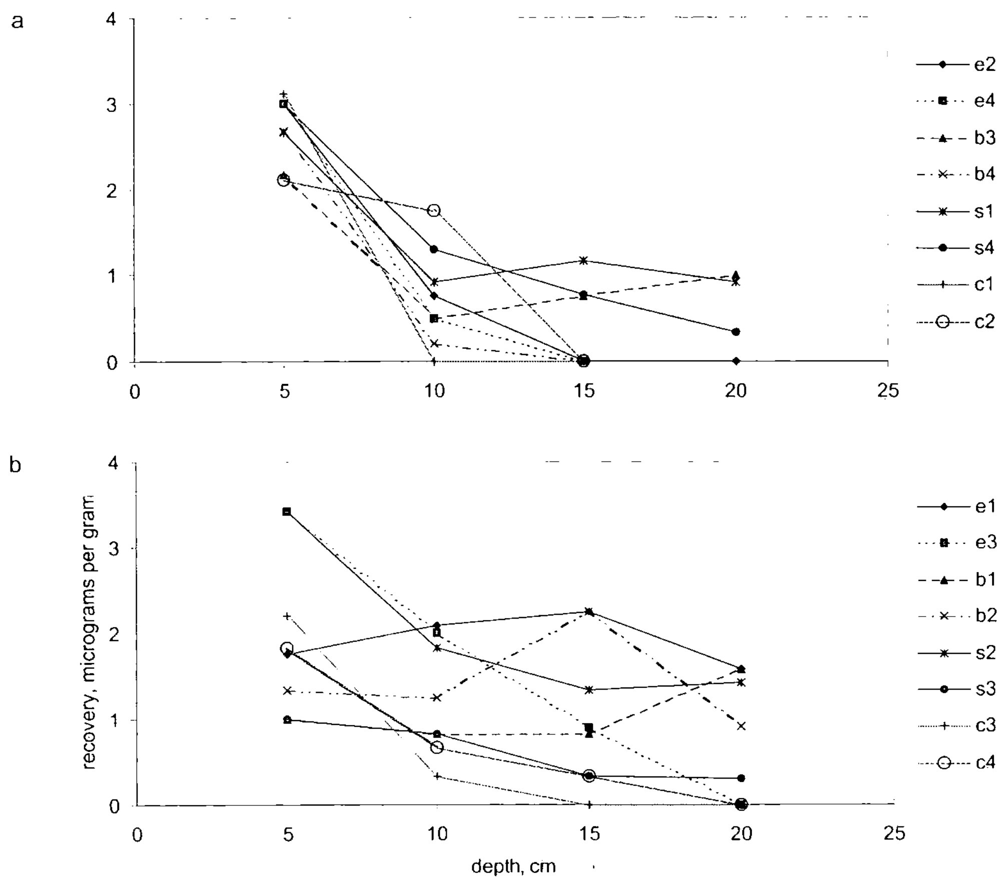


Figure 5.30. Lissamine FF distribution with depth, bulk sample recovery, a) non-drought blocks, b) drought blocks.

To facilitate comparison between depths, decreases in the mean concentration of Lissamine FF recovery with depth were allowed for by using the formula:

$$R = C_m/C_b \qquad [5.2]$$

where R is the recovery ratio (Smettem and Trudgill, 1983), C_m is the microsample dye concentration ($\mu g\ g^{-1}$), and C_b is the bulk sample dye concentration ($\mu g\ g^{-1}$). R values calculated from this equation are classified into two groups (>1 and ≤ 1) and interest is directed to the number of occurrences falling into the two structural classifications as described above. The hypothesis (H_1) states that the structural categories will show a differing frequency of R value occurrence, with R values > 1 occurring most frequently in class 1 and R values $\leq$ 1 most frequently in class 2. The χ^2 test with one-tailed significance levels yields the significance level for the difference between an observed number of occurrences in each category and expected number based on the null hypothesis (Ho).

Microsample R values given in Table 5.17a for blocks not subject to drought indicate that dye penetration is well correlated with visible structural features at the 5 cm layer. The majority of R values > 1 correspond to class 1 and R $\leq$ 1 to category 2. The χ^2 test (Siegel, 1956) accepts the difference (H_1) statistically at the 99.9 per cent level for the uppermost sampled layer and for the overall trend. For layers below 5 cm Ho must be accepted. However, given that dye penetration was only greater than 10 cm in 3 cases, this may be related to the low number of cases. The number of cases in class 1 is greater at 10 cm than for other layers. Structurally then, more macropores were found at 10 cm, but it may be that these were not sufficiently connected to upper layers to allow flux.

For blocks subject to drought conditions, the number of samples in class 1 is much greater than in blocks not subject to drought (Table 5.17b). Even at 15 cm the number of class 1 samples seems to be greater than in non-drought blocks. It may be that the drought has induced greater macroporosity in the blanket-peat. Furthermore, H1 can be accepted at 10 cm depth after the drought, such that macropore flow is important at deeper levels within the peat than before. Clearly then drought conditions encourage the development of functional macropores within the blanket peat deposit. The data presented are an important development in our understanding of the way in which runoff generation can be effected by droughts in blanket peat catchments. The increased infiltration and lateral subsurface flow that follows a warm dry spell on the moorlands may be partly explained therefore by structural changes within the peat caused by lowering of the water table into the normally anaerobic peat layers and shrinkage of the upper peat.

Further evidence for a more permanent physical change within the peat blocks comes from examination of moisture content of the peat layers (Table 5.18). In all four sampled layers down to 20 cm, the moisture content of the peat has fallen; even after substantial subsequent rainfall. The peat has not regained its original moisture content

Table 5.17. Chi-squared evaluation of R (C_m/C_b) values in relation to structural categories, a) blocks not subject to drought conditions, b) blocks subject to drought conditions

a)

	Number of events					H_1 acceptance level (one-tailed) (1d.f.)
	Class 1		Class 2			
Depth	R > 1	R ≤ 1	R > 1	R ≤ 1	χ^2	
5 cm						
Observed	15	6	2	49	37.6	99.9
Expected	5.0	16.0	12.0	39.0		
10 cm						
Observed	30	21	8	4	0.25	Not accepted
Expected	30.8	20.2	7.2	4.8		
15 cm						
Observed	5	18	3	10	0.01	Not accepted
Expected	5.1	17.9	2.8	10.1		
20 cm						
Observed	1	5	15	15	2.25	Not accepted
Expected	2.7	3.3	13.3	16.7		
Total						
Observed	51	50	28	78	12.7	99.9
Expected	38.5	62.5	40.5	65.5		

b)

	Number of events					H_1 acceptance level (one-tailed) (1d.f.)
	Class 1		Class 2			
Depth	R > 1	R ≤ 1	R > 1	R ≤ 1	χ^2	
5 cm						
Observed	28	8	9	27	20.1	99.9
Expected	18.5	17.5	18.5	17.5		
10 cm						
Observed	32	7	17	16	7.67	99.0
Expected	26.5	12.5	22.5	10.5		
15 cm						
Observed	19	5	25	14	1.60	Not accepted
Expected	16.8	7.2	27.2	11.8		
20 cm						
Observed	2	3	13	27	0.11	Not accepted
Expected	1.7	3.3	13.3	26.7		
Total						
Observed	81	23	64	84	30.0	99.9
Expected	59.8	44.2	85.2	62.8		

even after a week of wet weather, and although 6 of the blocks had a vegetation cover. It is well known that a drop of water placed on wet peat spreads over the surface of the peat; the angle between the water droplet and the peat tends to zero because the wet peat is hydrophilic. On dry peat, water drops do not spread, but form contact angles between the water and the peat of up to 85°, especially at low pH values. Thus dry peat is water repellent or hydrophobic. The difference in the wetting behaviour of dry peat and wet peat may influence the pattern of water movement in the peat and the extent to which precipitation infiltrates the surface (Egglesmann *et al.*, 1993). Clearly a four-week drought under conditions similar to that of the summer 1995 has caused structural changes to the peat blocks down to 20 cm which allows greater macropore flow and increased infiltration and percolation. It is unclear whether the surface moisture deficiency and structural changes are permanent; the rainfall simulation experiments do suggest that some recovery occurs, although this is less likely (or at least much slower) after the first three days of recovery. The fact that dried peat rarely returns to its original status upon re-wetting (Egglesmann *et al.*, 1993) also tends to suggest that changes may be more permanent.

Table 5.18 Mean moisture content, % by mass, of peat block layers in peats not subject to drought and peats subject to drought followed by a 7 day re-wetting procedure, n = 8 in all cases.

Layer	Non-drought	After drought and re-wetting
0 - 5 cm	88.9	79.4
5 - 10 cm	93.2	83.9
10 – 15 cm	90.4	84.6
15 – 20 cm	90.5	87.7

5.5. Conclusions

The rainfall simulation experiments on blanket peat have added great detail to our knowledge of the infiltration and runoff-generating processes within this soil type. Runoff collection has suggested that infiltration rate increases with rainfall intensity. This has important implications for inferences drawn from infiltrometers and simulators alike. Comparison of infiltration rates must be contextualised within intensity, and hence timebound datasets. The use of low-intensity rainfall has allowed a more realistic evaluation of infiltration rates and flow processes than previous studies. Overland flow appears likely to occur on both vegetated and bare surfaces although surface cover does exert some control. This, in combination with variability between plots, suggests that flow development will be widely distributed in space. Runoff decreases rapidly with depth, with the largest proportion of flow occurring within the top centimetre of the peat at steady-state. Not much vertical percolation takes place to depths greater than 10 cm such that most of the runoff production is within the upper layers of blanket-peat.

From bare peat, sediment loading tends to be supply limited. Seasonality may affect this relationship such that after a warm dry spell, surface desiccation allows sediment supply to become transport limited. Rainfall-runoff response may also vary with season.

Rainfall following warm dry weather tends to infiltrate more readily into blanket-peat, not just initially but to the extent that steady-state runoff rates are altered. Surface runoff is reduced and more flow takes place within the acrotelm. Peat subject to drought conditions displays this trend in a more marked way. Klove (1998) used high-intensity rainfall simulation (35 – 260 mm hr^{-1}) using a spray nozzle on large 100 m^2 plots of heavily disturbed (mined) peat in Finland. Work concentrated on sediment erosion processes but OLF was often found not to occur below rainfall intensities of 30 mm hr^{-1}. This peat surface would have been heavily disturbed and this result probably represents the important effect of changing environmental conditions on runoff generation in peatlands. The effect of vegetation removal and of exposing peat more readily to the processes of surface desiccation is to increase infiltration rates and promote lateral subsurface flow, often via preferential flow paths.

It is clear that structural changes take place within the near surface of the peat following a drought which allows enhanced infiltration and runoff at depth. Even after a week of wet weather, infiltration is still greater than before the drought and the peat does not regain its initial moisture status. Some reversion back towards pre-drought infiltration and runoff trends does occur in the first three or four days following re-wetting in many cases. After this changes appear less obvious. A mixture of qualitative and quantitative assessment of flow pathways suggests that macropore flow is an important pathway for water movement within the upper layers of blanket peat. The role of macropores in the runoff generation processes appears to increase after dry weather. A more quantitative-based experimental study examining the role of macropores in runoff-generating processes in blanket peat catchments is presented in Chapter 6.

CHAPTER 6

MACROPOROSITY AND INFILTRATION IN BLANKET PEAT: THE IMPLICATIONS OF TENSION DISC INFILTROMETER MEASUREMENTS

6.1. Introduction

It is well known that macropores such as worm holes, root channels and shrinkage cracks can exert a significant influence on water and solute movement (Beven and Germann, 1982). The relationship between unsaturated hydraulic conductivity (K) and unsaturated pressure head (ψ) is important for describing macropore functioning (Messing and Jarvis, 1993). Tension infiltrometers are a standard tool for *in situ* determination of saturated and near-saturated soil hydraulic properties (Jarvis *et al.*, 1987; Zhang, 1998; Zhang *et al.*, 1999). In order to assess the role of matrix and macropore flow a tension infiltrometer allows infiltration of water into the matrix, while not allowing flow into larger pores that may otherwise dominate the infiltration process. The infiltrometer provides a source of water at a small negative pore water pressure at the surface. The negative pressure prevents the larger pores that fill at greater pore water pressures from wetting up and short-circuiting the flow. Hence, by subtraction, the hydrological role of larger pores during the infiltration process can be evaluated. Most studies using tension infiltrometers have been conducted at the soil surface, although Azevedo *et al.* (1998) looked at infiltration properties of an Iowa loamy soil at surface and 0.15 m depth, and Logsdon *et al.* (1993) and Messing and Jarvis (1993) conducted measurements at different depths for different agricultural tillages. Only one study using tension infiltrometers has examined infiltration into peat, and this was performed on a thin fen peat where 17 *surface* runs were conducted with no vegetative discrimination (Baird, 1997).

6.2. Methods

6.2.1. Data collection

A porous disk infiltrometer similar to that designed by Ankeny *et al.* (1988) which controls the supply pressure head with a Mariotte bottle was used in the study. The set-up is shown in Figure 6.1. Infiltration rates were measured manually, although pressure transducers can be used in conjunction with a data logger for continuous measurement (Ankeny *et al.*, 1988). A 100 m x 100 m area containing the four most common surface vegetation types found at the field site (*Calluna, Eriophorum, Sphagnum* and bare peat) was used for sampling. For each vegetation type eight random measurement locations were determined. Each location selected had to consist of at least 90 % of the specified vegetation cover. The bare peat plots were all located in peat that was around 50 cm lower than the intact surrounding peat. At each location vegetation was cut back to the peat surface and a fine layer of moist fine sand of the same diameter as the circular base of the infiltrometer (26.5 cm) was applied. This smoothed out any irregularities at the soil surface and improved contact between the disk and soil surface. Moist sand is essential as air-dry sand may readily fall down into surface-vented macropores, forming 'wicks' (Messing and Jarvis, 1993). The infiltrometer was then placed on the sand. The weight of the infiltrometer may have resulted in some compression of the peat producing a slight restriction in flux. A *Sphagnum* cover would be compressed by the instrument but with the vegetation cover removed, field observations suggested that the effects of compression were minor, even for surface peats. The supply reservoir was narrow such that total water volume held in the infiltrometer was low, not only resulting in reduced weight but also aiding accurate measurements of discharge.

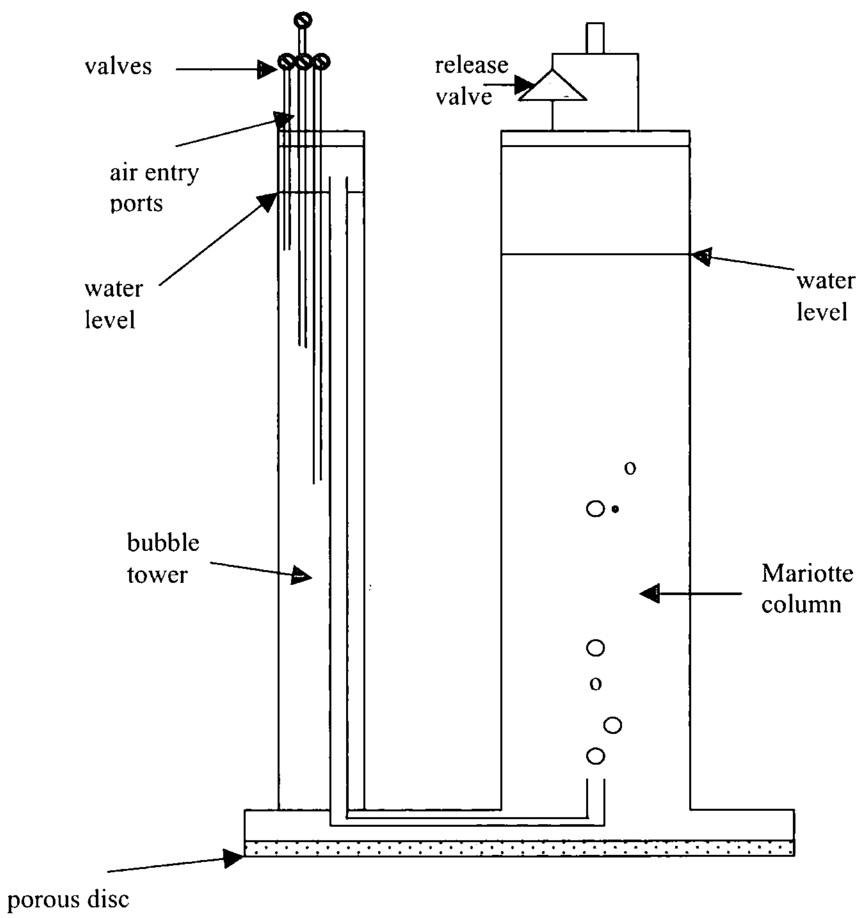


Figure 6.1. Schematic diagram of the tension infiltrometer

Infiltration measurements were performed with supply heads of –12 cm, -6 cm, -3 cm and 0 cm. Tests were conducted with the lowest supply head first (-12 cm), as reversal of this may lead to hysteresis where drainage occurs close to the disk while wetting continues near and at the infiltration front (Reynolds and Elrick, 1991). Infiltration measurements continued until a steady state was achieved. Frequently this took over 30 minutes to achieve but often the main problem was the low infiltration rates, which meant that experiments could last over several hours before satisfactory volumetric measurements could be attained. The problems of sunlight heating the supply reservoir were reduced by shading (Baird, 1997). After the surface test was completed at each measurement site it was allowed to drain for at least 96 hours (while running experiments elsewhere) before excavating 5 cm and repeating the test. The excavation was performed with great care to prevent macropores from becoming blocked during this process. The excavation was repeated at 10 cm and 20 cm depth. In total, therefore, there were 32 sample sites and runs were attempted at four depths at each site. It was decided not to test infiltration at depths greater than 20 cm as previous site tests had shown that the water table was never much lower than 20 cm and that the effective hydraulic conductivity would be so low that experiments would take too long to complete. Nevertheless a dry summer 1999 at Moor House NNR meant that 121 runs were successful with the water table dropping to 30 cm during an extended warm period (see below).

6.2.2. Data analysis

$K(\psi)$ values, including field saturated hydraulic conductivity, (K_{fs} – the hydraulic conductivity of a field soil when it is saturated) were obtained from the steady-state infiltrometer data using the method outlined by Reynolds and Elrick (1991). Here Wooding's solution for infiltration from a shallow pond (Wooding, 1968) is combined with Gardner's (1958) unsaturated hydraulic conductivity function. It is hence assumed that both are suitable for the instrument and soil in question (see discussion of data limitations below).

For about half the runs, plots of ln Q_s against ψ_0 were found to be linear and about half were not (Figure 6.2). In non-linear cases Reynolds and Elrick (1991) suggest that a reasonable approximation to $K(\psi)$ is to assume that ln Q_s versus ψ_0 is piecewise linear. Hence Reynolds and Elrick's piecewise method was used for the non-linear runs.

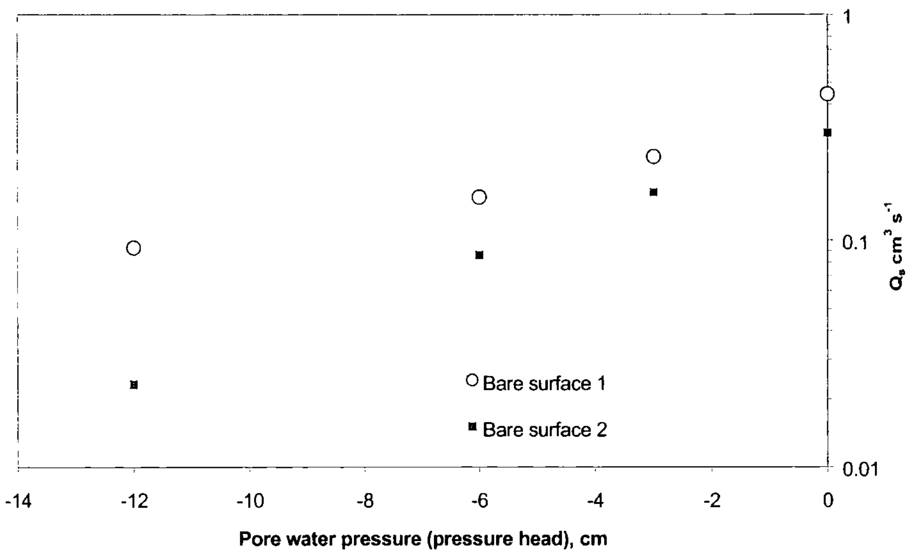


Figure 6.2. Examples of steady infiltration rates (Q_s) against pore water pressure.

Definitions of macropores vary widely and the choice of an effective size to delimit macropores is necessarily arbitrary. Luxmoore (1981), Watson and Luxmoore (1986) and Baird (1997) use the value of –3 cm pressure head to distinguish between macropores and smaller pores. According to capillary theory this indicates that macropores are larger than 0.1 cm in diameter. The proportion of K_{fs} governed by macropores for each run was therefore calculated by subtracting K at a pressure of -3 cm from K_{fs} (Baird, 1997).

6.3. Results and discussion

6.3.1. Macropores

Infiltration rates at K_{fs} were higher than at the other measured or calculated negative supply heads. Hence under saturated flow conditions, macropore flow is a major component of the infiltration process. Macropore contribution to K_{fs} has a mean value of 35.9 % and is approximately symmetrically distributed around that mean with a range from 1.0 % to 79.9 % and a standard deviation of 19.7 %. It is therefore suitable in raw form as the response variable for an analysis of variance (ANOVA) in relation to depth and vegetation type as controlling variables. In this analysis of variance depth is treated as a categorical variable, not a numerical variable, that is, just as defining four distinct categories, not as a series of numerical values from 0 cm to 20 cm. The ANOVA results (Table 6.1) show that each control is overwhelmingly significant as the calculated significance levels are less than 0.00005. Hence both controls can be regarded as genuinely influencing percent macropore contribution with an overall R^2 value of 35 %.

Table 6.1. Analysis of variance of macropore contribution to K_{fs}, %

Source	df	F	Prob > F
Model	6	10.08	0.0000
Depth	3	8.89	0.0000
Vegetation cover	3	10.91	0.0000

$R^2 = 0.35$

The detailed cross-tabulation of the means (Table 6.2) shows that *Sphagnum*-covered peats are associated with a greater role for macropores in the infiltration process. In non-*Sphagnum* covered peats the depth control is also evident with a maximum macropore flux control at 5 cm and a minimum at 20 cm depth. Results for all observations are shown in Figure 6.3. Here depth is taken literally, and essentially the figure represents a histogram that has been rotated and reflected. Relative to each mean, *Sphagnum* plot values tend to be higher and *Eriophorum* plot values tend to be lower in many cases. The surface macropore flux contribution ranges from 21 % to 68 %. These values are lower than found by Baird (1997) in a fen peat where the proportion of surface K_{fs} due to macropore flow was between 51 and 78 %. Now a wetland meadow nature reserve, this fenland site was drained, ploughed and used for arable farming between the late 1960s and 1990 which would probably have altered the surface properties of the peat (although Baird (1997) maintains that the soil profile was very similar to undisturbed peats nearby). In the blanket peat studied here, there has been no draining, ploughing or arable farming; before 1952 there may have been some burning related to grouse shooting (see below).

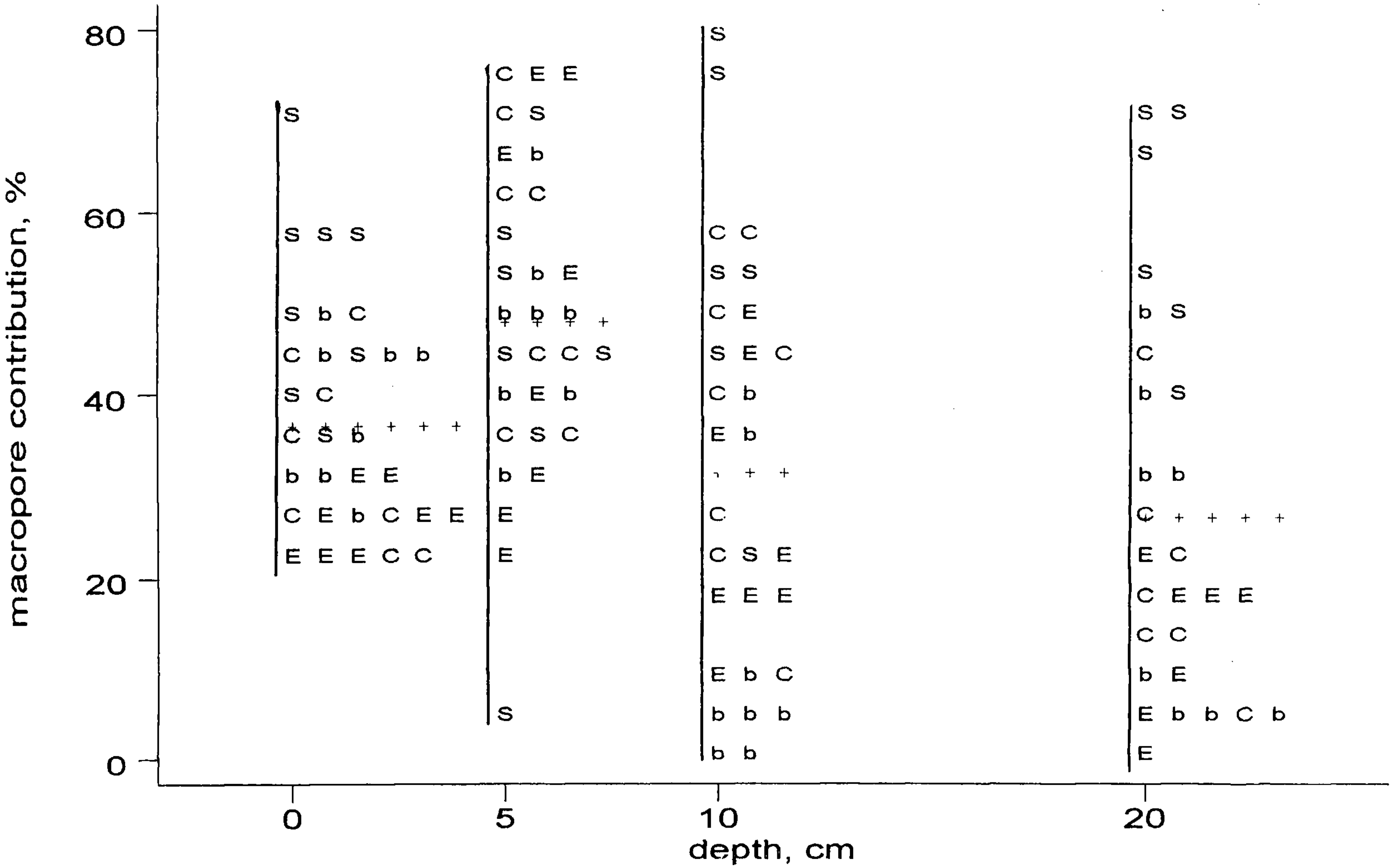


Figure 6.3. Macropore contribution to saturated hydraulic conductivity, %, for each depth by vegetation type. b = bare, E = *Eriophorum*, C = *Calluna*, S = *Sphagnum*, +++ = mean value. The figure displays a histogram by vegetation for each depth value.

Table 6.2. Means, Standard Deviations and Frequencies (from top to bottom of each row) of macropore contribution to K_{fs}, %

Depth, cm	vegetation type				Total
	Bare	*Calluna*	*Eriophorum*	*Sphagnum*	
0	37.7	32.7	25.7	50.5	36.7
	8.6	10.1	2.8	11.5	12.5
	8	8	8	8	32
5	46.6	52.9	47.6	43.5	47.8
	10.3	15.3	21.6	20.5	16.8
	8	8	8	7	31
10	12.9	38.6	26.1	54.0	31.5
	15.6	17.8	14.4	21.2	22.0
	8	8	8	6	30
20	22.0	19.2	12.9	57.4	26.6
	17.6	12.4	7.7	12.1	20.9
	8	7	7	6	28
Total	29.8	36.4	28.6	51.0	35.9
	18.6	18.2	18.1	16.5	19.7
	32	31	31	27	121

Given that bare peat has been eroded away to a depth of more than 50 cm in many places, it is striking that the proportion of macropore flow at the surface and near-surface (5 cm) appears to be much higher than in deeper layers and is also roughly equivalent to that found at the surface of vegetated peats. This is likely to be related to the cracking and desiccation of bare peat during dry periods. There is a vast literature on the effects of cracking on infiltration and the redistribution of water in non-wetland soils (see Germann, 1990), but little information is available for wetland soils. Cracking on the peat surface has been reported many times (e.g. Bower, 1959; Gardiner, 1983; Gilman and Newson, 1980, see also Chapter 5 and Figure 5.17), although no reliable estimates have been made of its hydrological role. Cracking has been implicated in soil piping in blanket peats (Jones, 1981). Desiccation cracking is a complex process, depending both on the shrinkage potential of the soil and its strength, and on rates of desiccation (Gilman and Newson, 1980). Peat is generally 90% water by weight and when the surface dries, significant crusting and shrinkage can occur. It is certain that the surface tension cracks act as hydrologically active macropores, channelling water both laterally and vertically from the peat surface crust. This was clearly visible from observations of water flow when sprinkled on to the surface at the study site. Furthermore, when water enters a desiccation crack in soil, it may in certain circumstances erode a tubular channel along the base of the crack. This process of macropore formation has been described by many authors including Henkel *et al.*

(1938); Hughes (1972) and Rathjens (1973). Needle-ice formation on the peat surface, seen very often in the winter months in the North Pennines, may also play a role in loosening the near-surface peat in order to help create macro-channels for water flow. Whatever the mechanism, the removal of vegetation and downwasting of peat to a depth where macropore functioning appears less likely to occur has resulted in the emergent bare peat surface being transformed to contain functioning macropores of the same order as in the surface of vegetated peats. The role of desiccation in altering the surface infiltration and runoff processes within blanket peat is an area for further research.

The proportion of flow occurring through macropores at 5 cm depth is greater than in any other measured layer. The same is true for peat below each vegetation type, except in the case of *Sphagnum*. It may be that old root channels act as lines of weaknesses in the peat at around 5 cm depth, which are then enlarged by macropore erosion processes. For bare peat between 5-10 cm the role of macropores appears to drop off rapidly suggesting that desiccation at the surface has only reached down to the upper few centimetres of peat. This finding is backed up the dry bulk density cores discussed in Chapter 5 (see Figure 5.9). *Eriophorum* and *Calluna* covered peats appear to behave in a very similar way to each other. Here we would expect more active macropores near the surface where root channels and loose woody and leafy material are located with minimal decomposition and compaction. Before 1952 the study site was a grouse moor and may have suffered periodic burning. Given that rates of peat growth at Moor House have been estimated between 0.6 mm to 1 mm yr^{-1} (Turner *et al.*, 1972) this would result in about 3-5 cm of growth since the time of burning. Hence it may be that this macroporous layer at 5 cm represents a result of a period of burning. No evidence of burning could be found within the soil profile at the tension-infiltrometer test sites, however.

Clearly *Sphagnum* peat provides a more macroporous (and generally more permeable - see below) route for water transfer up to 20 cm depth than peats with other surface covers. The *Sphagnum* cushion comprises a dense and finely porous 'roof' of side branches, supported on a much less dense layer of vertical columns interspersed by much larger spaces and obtaining some lateral bracing from the occasional divergent side branches (Ingram, 1983). At the same time *Sphagnum* does not produce significant root systems (unlike *Calluna* and *Eriophorum*.) thereby reducing the mechanisms available for the creation of macropores at depth.

6.3.2. Saturated Hydraulic conductivity

Figure 6.4 gives some examples of changes in K_{fs} with depth for 5 sites. Clearly the saturated hydraulic conductivity at these sites drops rapidly over a very short distance. Over just 20 cm K_{fs} is often reduced by up to 2 orders of magnitude. There are some non-conforming data runs such as *Calluna* site 1 (C1) which witnesses a dramatic increase in K_{fs} at 20 cm depth over that at 10 cm. One might expect this to be related to an increase in functioning macropores at this depth, but C1 in fact has the lowest proportion of macropore flow of all the runs at 10 and 20 cm depth. C1 is a clear example therefore of the inherent variability of matrix flow within blanket peat.

K_{fs} is highly variable, and positively skewed, ranging from 0.013 x 10^{-6} cm s^{-1} to 545.6 x 10^{-6} cm s^{-1} with a mean of 133.5 x 10^{-6} cm s^{-1} and a skewness of 0.91. Given this, it is natural that the variability within groups of values defined by depth and vegetation type categories is not even roughly constant, as required for application of ANOVA (despite its name, ANOVA is all about comparing mean values on the assumption that within-group variation is approximately constant). A square root transformation of the data was found to work well. The square roots of K_{fs} are less skewed with a range from 0.114 x 10^{-3} $\sqrt{(cm\ s^{-1})}$ to 23.36 x 10^{-3} $\sqrt{(cm\ s^{-1})}$, a mean of 9.27 $\sqrt{(cm\ s^{-1})}$ and a skewness of 0.32. More important, the variability with depth and vegetation categories is now more nearly constant.

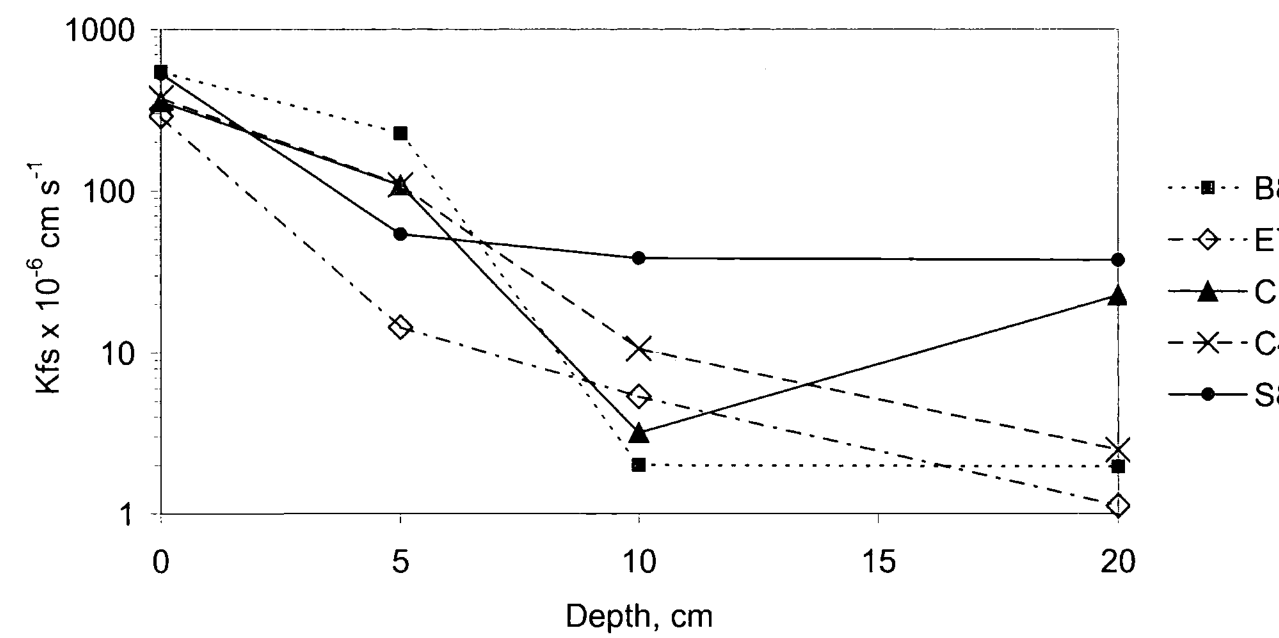


Figure 6.4. Examples of individual site changes of K_{fs} with depth.

The ANOVA shows that both depth and vegetation type can be accepted as genuine controls, but depth is very much the dominating factor (Table 6.3). Table 6.4 presents cross-tabulation of the means which vary from 20.0 x10^{-3} √(cm s^{-1}) at the surface of bare peat to 1.1 x 10^{-3} √(cm s^{-1}) at 20 cm depth below *Eriophorum*. These values are equivalent to 4.0 x 10^{-4} cm s^{-1} to 1.5 x 10^{-6} cm s^{-1}. Summary data on hydraulic conductivities in peat are given by Rycroft *et al.* (1975) from values of 6 x 10^{-8} cm s^{-1} for highly humified blanket peats up to 5 x 10^{-3} cm s^{-1} for slightly humified fen peats. Romanov (1968) quoted results for *Carex-Hypnum* peats from Belorussia showing progressive decline in hydraulic conductivity from 3.1 x 10^{-3} cm s^{-1} at 0 to 50 cm depth to 6 x 10^{-5}cm s^{-1} in the 100 to 150 cm layer. Clearly the latter peat is far more permeable than that of the blanket peat studied here. Nevertheless, data presented here adds to the existing knowledge base on the hydraulic conductivity of peat.

Results for all values are shown in Figure 6.5. The strong control of depth on K_{fs} is clear from the rapid decline in mean flux as depth increases (Figure 6.5a) and from the fact that all surface K_{fs} values (0 cm) are above the mean value for each vegetation category and all 20 cm K_{fs} values are below the mean (Figure. 6.5b). For bare peat, values of K_{fs} are largely above the mean at 0 cm and 5 cm depth and below the mean in deeper layers. Cross-tabulation of the means (Table 6.4) demonstrates how bare peat has the highest mean surface infiltration rate. It would be logical to relate the greater flux at the surface of bare peat to the desiccation of the surface and perhaps to the role of cracking in creating macropores. However, given that the role of macropores at the surface was only slightly greater in bare peat than *Calluna* covered peat, these data demonstrate that the propensity for matrix flow in the surface layers of unvegetated peat is probably greater.

Table 6.3. Analysis of variance of saturated hydraulic conductivity, square root data

Source	df	F	Prob > F
Model	6	71.00	0.0000
Depth	3	136.51	0.0000
Vegetation cover	3	3.84	0.0116

$R^2 = 0.79$

a)

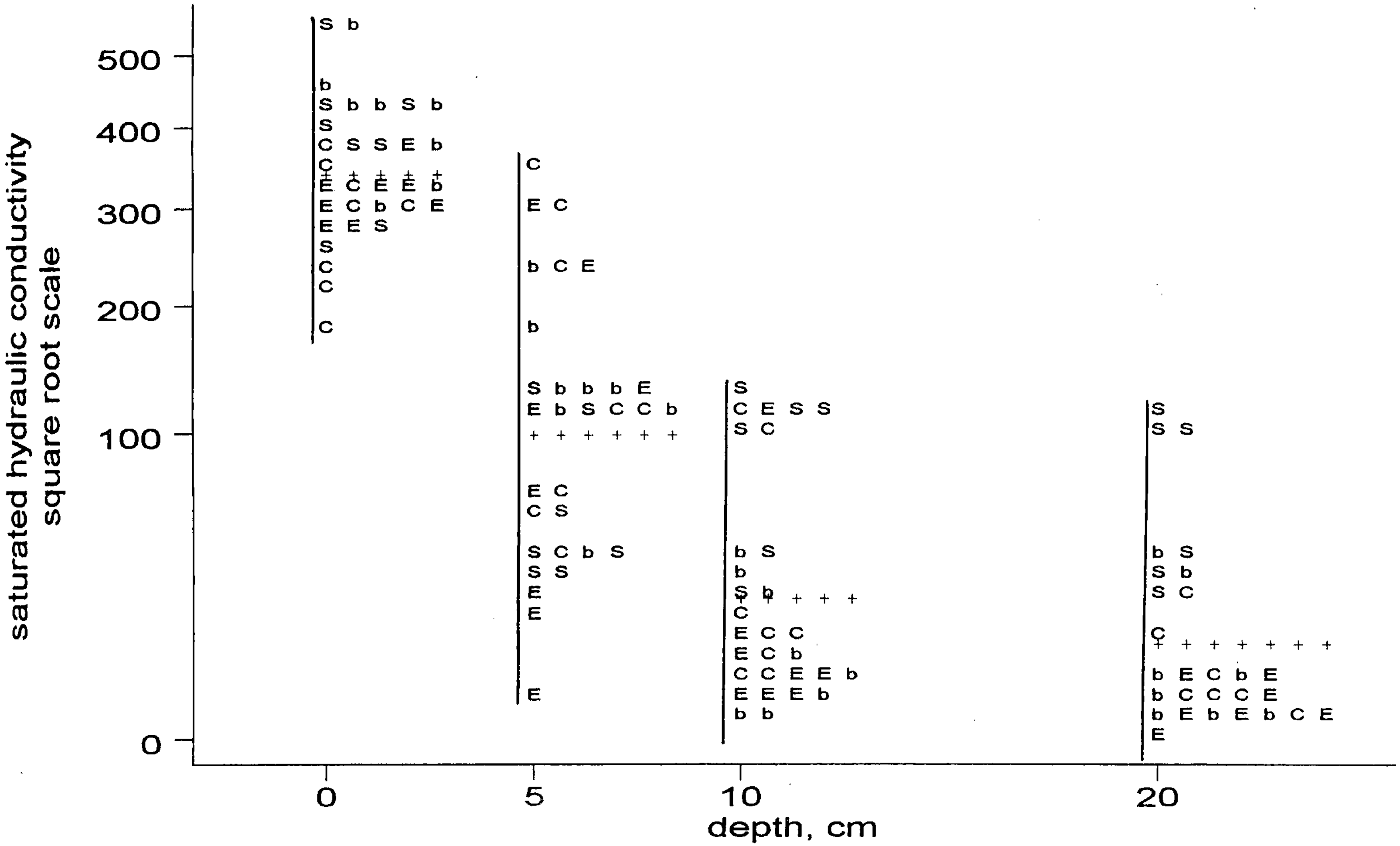


Figure 6.5. Saturated hydraulic conductivity of blanket peat, a) for each depth by vegetation; b) for each vegetation by depth. b = bare, E = *Eriophorum*, C = *Calluna*, S = *Sphagnum*, +++ = mean value.

b)

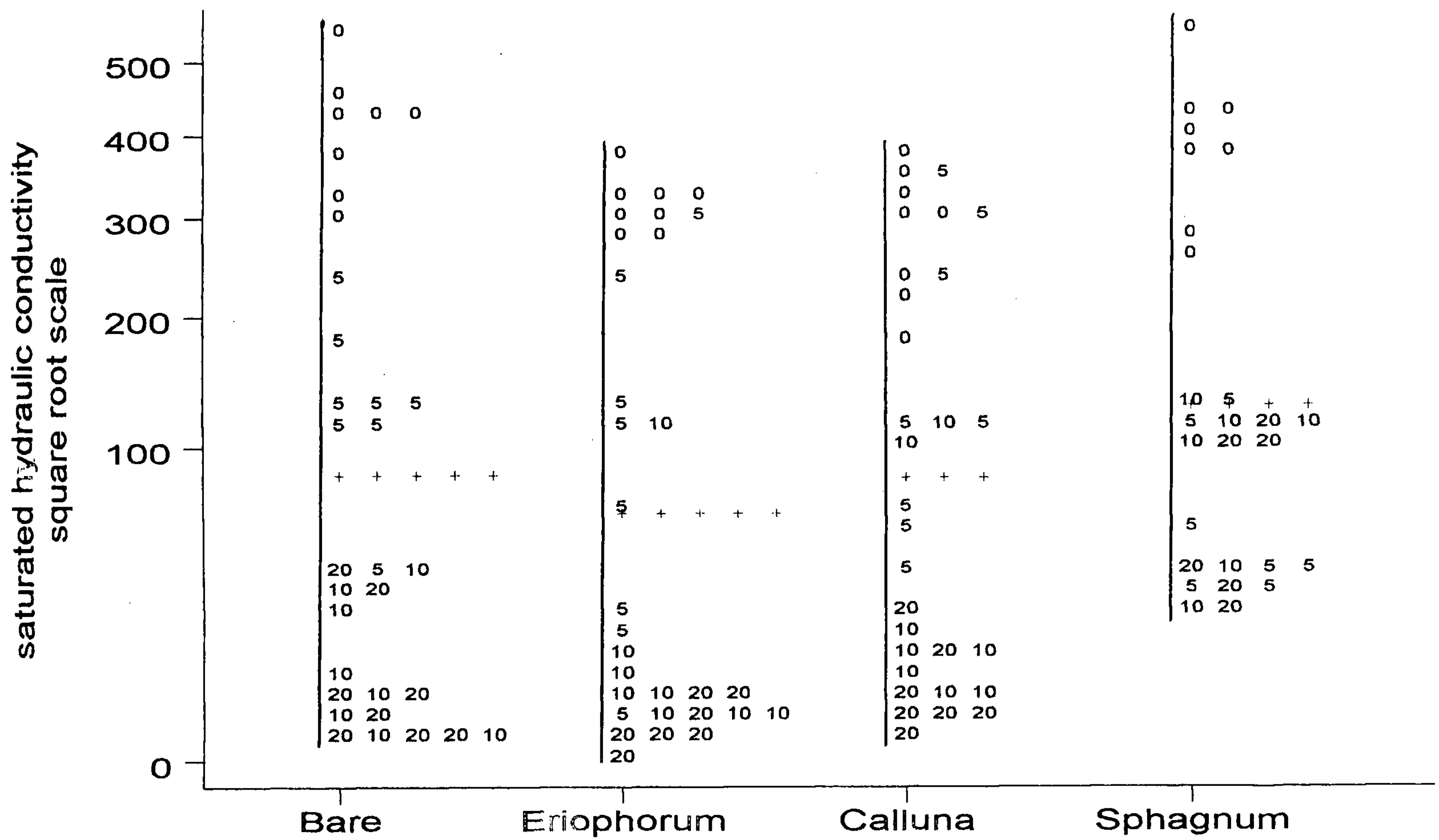


Figure 6.5 continued

Table 6.4. Means, Standard Deviations and Frequencies (from top to bottom of each row) of square root values of saturated hydraulic conductivity data.

Depth, cm	vegetation type				Total
	Bare	*Calluna*	*Eriophorum*	*Sphagnum*	
0	20.0	16.8	17.5	19.3	18.4
	2.0	2.1	0.8	2.2	2.2
	8	8	8	8	32
5	11.2	11.7	9.1	7.4	9.9
	2.7	4.7	5.6	2.5	4.3
	8	8	8	7	31
10	2.9	4.7	3.0	8.8	4.6
	2.2	3.5	3.0	2.8	3.6
	8	8	8	6	30
20	2.4	2.0	1.1	7.7	3.1
	2.2	1.4	0.7	2.7	3.1
	8	7	7	6	28
Total	9.1	7.9	9.0	11.3	9.3
	7.6	7.2	6.6	5.8	6.9
	32	31	31	27	121

Mean values of $\sqrt{K_{fs}}$ in units, $x10^{-3}\ \sqrt{(cm\ s^{-1})}$

For bare peat, and for *Calluna* and *Eriophorum*-covered peat, mean K_{fs} declines with depth. *Eriophorum*-covered peat has a slightly lower mean K_{fs} below the surface layer than *Calluna*. This may be related to rooting structures found below these two vegetation types or to the greater earthworm density found associated with *Calluna* (Svendsen, 1957). Earthworm densities at Moor House are typically low ranging from one per 100 m^2 on *Eriophorum* dominated areas to 13 per 100 m^2 on *Calluna* peats (Svendsen, 1957). At 10 and 20 cm depth, peat with a *Sphagnum* cover tends to have a greater K_{fs} than peats with other covers. This is indicated not only by comparison of mean values but also by the dominance of individual *Sphagnum* plots with K_{fs} greater than in peats 10 to 20 cm below a bare, *Calluna* or *Eriophorum* cover (Figure 6.5a). For peats beneath *Sphagnum* it appears that K_{fs} declines very rapidly over the first 5 cm and then stabilises somewhat. Clymo (1983) found that below a *Sphagnum* peat profile, the bulk density in the top 1 cm was relatively high, but that immediately below the tightly packed capitula of the mosses it falls to about 0.02 g cm^{-3}. Below this, density increases gradually until the weakened mosses can no longer support the load above (commonly 10-30 cm deep) and density increases rapidly to 0.1 g cm^{-3}. However, given that the proportion of functional macroporosity does not change greatly with depth below *Sphagnum* from the surface down to 20 cm, it would appear that both matrix flow and macropore flow at the peat surface is much greater than at depth. Over the entire depth range of sampling, there is a greater macropore and water flux associated with peat beneath growing *Sphagnum* than in peat beneath *Calluna, Eriophorum* or a bare surface. It is well known that *Sphagnum* communities prefer higher water tables and a much wetter environment than that of the woody or grassy species found at the study site, yet the peat below *Sphagnum* is more permeable, and indeed macropores contribute more to this permeability than under other peats. Peat surface hollows can develop in which water can pond up, encouraging the development of *Sphagnum* mats and thereby increasing local peat permeability and functioning macroporosity. The more permeable peat is still confined by less permeable peat, however, and water continues to pond beneath the *Sphagnum*. This sort of 'hummock-pool' complex is commonly found in the Pennine blanket peats (Tallis 1994; Tallis and Livett 1994).

Burt *et al.* (1990) argued that areas of blanket peat may be one of the few locations where infiltration-excess overland flow can occur frequently. The typically low rainfall intensities received in the blanket peat moorlands of Britain was discussed in Chapter 3. The relative frequency of hourly rainfall intensities at the study site as a proportion of all hours with rainfall is shown in Figure 3.7c. Mean surface infiltration rates determined from the tension infiltrometer data, as indicated in Table 6.5, are exceeded only occasionally. Disregarding possible snowmelt occasions, 10 mm of rainfall in one hour was exceeded only five times in the four water years studied, with a maximum recorded intensity of 11.6 mm hr^{-1}. Fifteen-minute data are available from a raingauge at the study site between August 1998 and December 1999. Results from this gauge indicate that intensities equivalent to 10 mm hr^{-1} were exceeded 18 times and 12 mm hr^{-1} six times. On only one occasion did rainfall greater than 14 mm hr^{-1} occur when 5 mm fell in 15 minutes.

Furthermore, the flashy response of river regimes in blanket peat areas appears to occur for all rainfall intensities above a threshold of 1 to 2 mm hr^{-1} and not just for those at the high end of the precipitation range (Evans *et al.*, 1999). The substantial fall in K_{fs} with depth clearly contributes to this. Hydraulic gradients within the main peat mass at the study site are low, often below 0.1, with flow nets indicating mainly vertical hydraulic gradients near the surface (Chapter 4). Measurement of near surface hydraulic gradients (within the upper 5 cm) during rainfall has shown that these gradients can approach 1, thus indicating that the near surface layers of peat can readily transfer water away from the surface. It is therefore seems logical to conclude that infiltration-excess overland flow is only generated on rare occasions and that where it occurs more frequently, it will be spatially very localised because of the high variability in surface infiltration rates. It appears more likely that a saturation-excess mechanism, combined with percolation-excess above a much less permeable layer at 10 - 20 cm depth, dominates the runoff response. Evans *et al.* (1999) note that it is possible that subsurface storm flow in the acrotelm is limited by absorption of water into unsaturated peat at a fixed rate, and that subsurface runoff is generated when precipitation inputs exceed that rate. The tension infiltrometer results suggest that low-intensity rainfall events of the order of 0.2-0.8 mm hr^{-1} can be absorbed at up to 20 cm in depth but above this intensity percolation-excess occurs. This adds weight to the threshold hypothesis of Evans *et al.* (1999).

Table 6.5. Mean field saturated steady infiltration rates, standard deviations and frequencies (top to bottom of each row) for each peat surface type and depth, mm hr^{-1}

Depth, cm	vegetation type				Total
	Bare	*Calluna*	*Eriophorum*	*Sphagnum*	
0	14.60	10.30	11.11	13.64	12.41
	2.95	2.53	0.99	3.04	2.99
	8	8	8	8	32
5	4.72	5.63	3.97	2.31	4.28
	2.64	4.25	3.89	1.57	3.29
	8	8	8	7	31
10	0.45	1.20	0.61	2.8	1.11
	0.52	1.62	1.31	1.65	1.50
	8	8	8	6	30
20	0.41	0.21	0.06	2.05	0.58
	0.57	0.29	0.06	1.43	0.99
	8	7	7	6	28
Total	5.20	4.47	4.06	6.13	4.90
	6.17	4.78	4.93	5.80	5.41
	32	31	31	27	121

6.4. Data limitations

There are two important problems associated with the data collection. Firstly the analysis is based on an assumption that the unsaturated hydraulic conductivity before the test ($K(\psi_b)$) is much less than the hydraulic conductivity under the imposed tension infiltration condition ($K(\psi_0)$) (Reynolds and Elrick, 1991). Because of this it is likely that surface tests will have produced more reliable results than the measurements taken at depth. Indeed, if the water table is close to the surface then for tests at depths greater than 5 cm and for applied pore water pressures of –12 cm and –6 cm it may be sufficiently close to the tension infiltrometer for the assumption of $K(\psi_b) << K(\psi_0)$ to fail. Reynolds and Elrick (1991) however show that this theoretical limitation may only produce minor errors. Although for most of the year the water table was within 5 cm of the surface, water tables were up to 30 cm below the surface during the testing period of summer 1999 (see Figure 4.7). The low water table was a pre-requisite for the tests and gives more credence to the methods used. Figure 6.6 examines the errors associated with the assumption of $K(\psi_b) \ll K(\psi_0)$ if $K(\psi_b)$ is high. If $K(\psi_b)$ is less than 40 % of $K(\psi_0)$, which is a reasonable assumption at depths of 20 cm when the water table is at 30 cm, then errors are generally less than an order of magnitude. For $K(\psi_b)$ values less than 10 % of $K(\psi_0)$, the potential error produced in K_{fs} estimation is on average a factor less than 2 and at most less than 3. For the surface measurements presented in this paper the mean range in calculated K_{fs}, for a cover type category was less than a factor of two; at depth the average range in K_{fs} for within depth-cover category data was a factor of 68. Given this natural variability, and the range of 4 orders of magnitude in the dataset presented, the errors produced by the theoretical limitation of $K(\psi_b) \ll K(\psi_0)$ within the context of this study are therefore likely to be minor.

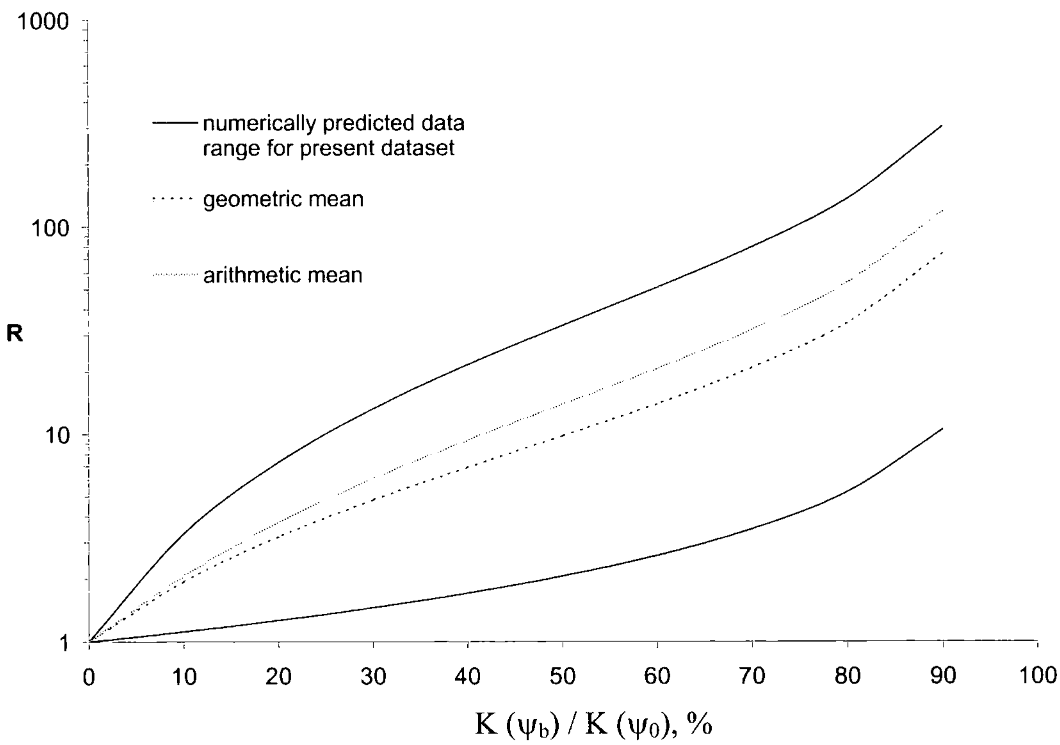


Figure 6.6. Magnitude of potential errors produced in K_{fs} estimation when the assumption of $K(\psi_b) \ll K(\psi_0)$ is tested for values of $K(\psi_b)$ close to $K(\psi_0)$. R = the ratio of K_{fs} estimated when $K(\psi_b) \ll K(\psi_0)$ to K_{fs} estimated when $K(\psi_b)$ is close to $K(\psi_0)$.

The second problem with the dataset is that Wooding's (1968) solution for infiltration from a shallow pond assumes that the soil below the tension disc is homogenous, isotropic and uniformly unsaturated. For the vast majority of tension infiltrometer measurements reported in the literature, these assumptions are only ever approximately met. The soil profiles at the present study site indicate increasing DBD and increasing humification over the first 20 cm; the increase in DBD is most pronounced between 10 and 15 cm into the bare peat. Furthermore, the evidence suggests that hydrologically the peat is anisotropic. The key question is whether the peat properties change sufficiently within the zone of influence of the tension disc to render the dataset meaningless. The maximum volume of water discharged during any of the infiltrometer tests, was 173 cm^3 over two hours. Due to the large area of the tension disc compared with the cylindrical area of the supply reservoir, this meant that apparently large changes in water depth in the supply reservoir actually occurred at low infiltration rates. This aided accurate measurement and allowed establishment of steady-state infiltration in circumstances of low flux. Hence 173 cm^3 equated to 3.1 mm water depth of infiltration over the area of the disc. This water depth does not tell us how deep the front of the wetting bulb actually reached because firstly it is unlikely to extend to a uniform depth below the tension disc (and will also extend laterally[1]). Secondly a 3.1 mm water depth is less than the depth of infiltration because the soil has a limited functional porosity. Peat has a very high porosity. Mean porosity measured in the upper layers at the study site were around 55 % from 0 to 10 cm and 35 % from 10 to 25 cm. This change with depth generally means that any change in water table elevation in upper horizons of less decomposed peat represents considerably more water than a corresponding change in deeper, more dense peats (Boelter, 1968). As shown in Chapter 4 (Figure 4.11) automatic logging of water table and rainfall at Moor House by ECN allows regression analysis of the effects of rainfall on water table ($n = 70$, $R^2 = 0.54$). The regression predicts that when the water table is at 20 cm, 1 mm of rainfall will induce a water table rise of 17 mm. When the water table is within 5 cm of the surface 1 mm rainfall induces a only a 4 mm rise. These water table elevation values suggest that 3.1 mm of water supply may not result in a deep wetting bulb and because infiltration rates were lower at depth, much smaller volumes of water were used in those tests. This along with the low

[1] The potential for lateral spread of the wetting bulb could be important and would be reduced by the use of a dual (or concentric) disc tension infiltrometer. This equipment has only recently been developed and tested (Zhang *et al.*, 1999) and was found to have the additional time-saving benefit of not having to rely on infiltration reaching steady-state. However, the concentric disc infiltrometer would still be subject to the problems associated with the two assumptions outlined above when applied to blanket peat. The concentric disc tension infiltrometer is also more expensive to construct.

water tables during the study period provides further evidence to suggest that the wetting bulb is unlikely to have extended far below the infiltrometer disc.

Thus although use of Wooding's (1968) solution does mean that hydraulic properties of the soil that affect a reading at 5 cm for example, may come from soil at depths below 5 cm, it is unlikely that hydraulic properties of the soil that affect readings at 5 cm will affect readings at 10 cm and deeper. This therefore allows the independence necessary for use of ANOVA on the depth categories. These data also suggest that although the peat structure changes fairly rapidly with depth in the upper peat (see Chapter 5), the zone of influence of the wetting bulb is likely to be low such that inhomogeneity of the peat may not affect the results substantially. The results from 10 cm under a bare surface are most likely to be affected by this error because as discussed in Chapter 5 the soil profiles indicate the most dramatic density changes at this depth (e.g. Figure 5.9). Although the data from the surface layer is least problematic and that there may be up to an order of magnitude error in the data from the tests at 20 cm, it is suggested that the below surface dataset still adds significantly to the sparse literature on peat soils. This is especially the case given that other methods of estimating K_{fs} at depth in blanket peat result in similar errors (Rycroft *et al.*, 1975). There is still a need to find more accurate methods for determining *in situ* K_{fs} in blanket peat.

6.5. Conclusions

Results of this study showed that infiltration rates at 0 mm were significantly larger than infiltration rates at the three other tensions for all plots. Therefore, under saturated flow conditions, macropore flow is a significant pathway for water in the upper layers of blanket peat. Macropores appear to be an important component of upland hydrology, a component which has so far largely been ignored. Given this information it is likely that macropores also play an important role in upland hydrochemistry and further work is required in this field. Peat depth and surface vegetation cover appear to be linked to macroporosity and to values of field-saturated hydraulic conductivity. It would appear that macropores provide an important runoff-generating pathway in blanket peat-covered catchments. Potential for water movement through the peat matrix at the surface is also high and the tension infiltrometer results suggest that infiltration-excess overland flow is not a frequent occurrence, although it may happen locally a few times each year. Shallow subsurface stormflow occurs (percolation-excess runoff) because K_{fs} decreases rapidly with depth, often by 2 orders of magnitude over 20 cm. Because of this, water that infiltrates relatively quickly into the upper peat (~5 cm) becomes restricted as the matrix K_{fs} level decreases and the functional macroporosity falls. Hence, ponding at a relatively shallow depth will occur. There is a higher level of functioning macroporosity at 5 cm than at any of the other measured depths within the peat. Because K_{fs} is generally lower at this depth than at the surface, it would appear that at this level matrix flux is reduced much more than macropore flux. *Sphagnum*-covered peat appears to have a greater macroporosity and permeability up to 20 cm depth than other surface types, but this is probably of little overall consequence to runoff regimes, because *Eriophorum-Calluna* mixes dominate the blanket peat at the study site. The original hydrological properties of the bare peat have now been altered by desiccation producing both greater macropore flow and greater matrix flow. With an increasing number of hot and dry summers (Marsh and Sanderson, 1997), peat desiccation may increase, as may the areal extent of bare peat cover. This might allow surface infiltration rates to increase, but the evidence indicates that the overall proportion of functioning macropores will not change greatly as bare peat surfaces become altered to act similarly to those with a vegetation cover. It may be, however, that with hotter, drier summers, cracking and desiccation will affect deeper levels within the peat.

CHAPTER 7

THE HYDROLOGY OF PIPES IN THE LITTLE DODGEN POT SIKE CATCHMENT

7.1 Introduction

There has been a lack of continuous pipeflow measurements (Bryan and Jones, 1997). The limited data available from Europe and North America suggest that pipeflow can be an important contributor to streamflow especially during storms. Unfortunately most evidence is limited to a few catchments in Wales (see Chapter 2) and there is a dearth of data outside the Welsh peaty podzol catchments of the Upper Wye and the Maesnant.

In the Upper Wye the pipes monitored were in shallow peats (*circa* 27 cm) and were all close to the surface (Morgan, 1977; Gilman and Newson, 1980). Measurements of pipe cross-sections were taken at the stream bank and by digging pits and were found to average 9.2 cm in diameter (standard deviation = 2.6 cm) within the Cerrig yr Wyn subcatchment (Morgan, 1977). In terms of monitoring, Gilman and Newson (1980) were only really concerned with the response of ephemeral pipes in three storms. The piped Measnant catchment is the only one to have been monitored in more than a handful of storms (Jones, 1994). The pipes at Maesnant (in peat and peaty gleyed podzol soils which are typically of 1 m depth) range from 9 - 30 cm in diameter but again are only shallow and were found 15 - 80 cm from the surface (Jones, 1982). Thus, although most continuous monitoring of piping comes from the peaty catchments of the Upper Wye and Maesnant, these areas only contain examples of relatively shallow piping with small cross-sectional areas. Piping has been observed at much greater depths in blanket peats, with Pearsall (1950) and Bower (1960) both observing deep-seated pipeflow. Anderson and Burt (1982) report pipe diameters up to 50 cm in Shiny Brook, South Pennines and the existence of deep and shallow pipes. Evidence from Gardiner (1983) suggests a minor contribution from pipeflow in the deeper blanket peat (Burt *et al.*, 1990) although he did not monitor the larger pipes in the catchment. Gunn (2000) notes that pipes on Cuilcagh Mountain, Ireland, range from a few centimetres in diameter to those that are large enough to crawl into. Apart from undergraduate dissertations, no work has been done on these pipes.

Detection of soil pipes has often proved problematic (Bryan and Jones, 1997). Ground Penetrating Radar (GPR) has been applied to some of the Northern Pennine pipes in Chapter 8. Here it has been found that GPR often fails to detect pipes that are smaller than 10 cm in diameter and the set-up discussed in Chapter 8 also fails to detect near-surface pipes. Thus, for many of the Welsh pipes (which dominate the literature) the technique may be limited given the typical range of pipe sizes and depths found there. Most of the pipes in the Cerrig yr Wyn subcatchment would not be detected as they are usually less than 10 cm in diameter but in the Maesnant the range of pipe sizes suggests that they could be detectable by GPR (see above). The shallow depths of many of the Measnant pipes, however, means that there may be difficulties in some of them using GPR. However, because the pipes are closer to the surface in the mid-Wales catchments they are visibly easier to detect. Many of the pipes in the deeper North Pennine blanket peats are at a much greater depth in the soil profile (see below). Thus GPR application may be more successful in the Pennines than in the shallow peaty podzols of mid-Wales. However, use of a different range of antennae for the shallower soils may allow greater transference of the technique (see Chapter 8).

The Little Dodgen Pot Sike (LDPS) catchment on the Moor House Reserve (see Figure 3.4) contains soil pipes which often have large cross-sectional areas (see below) and are frequently found throughout the soil profile, with some pipe outlets located entirely within the clay substrate. The raw blanket peat in the LDPS catchment is often deeper than 2 m and is therefore very different in nature from the Plynlimon catchments. Thus the LDPS catchment provides an alternative location to monitor pipeflow than in the peaty podzols of mid-Wales. This chapter will examine results from mapping and continuous hydrological monitoring of the pipes in the catchment.

7.2 The Little Dodgen Pot Sike (LDPS) catchment

The location of the LDPS catchment on the Moor House Reserve is shown in Figure 3.4. Delimitation of catchment boundaries is often difficult in blanket peat because of the nature of the gently sloping terrain, and the subsurface pipe networks (Burt and Gardiner, 1982; Burt and Oldman, 1986). In addition, the head of LDPS emerges from two limestone risings. Sinkholes were found upslope of the outlets. It is notable that these sinkholes were on the other side of the visible watershed such that estimation of catchment area based on contour maps would not have been sufficient. No detailed work has been done on the limestone drainage systems of this area. By using salt tracing techniques it was possible to identify which sinks were feeding LDPS and which were feeding other catchments. Thus it was possible to more accurately define the catchment area which was larger than the surface topography would have suggested.

The LDPS catchment covers an area of 0.44 km^2 (+/- 0.04 km^2) falling from 570 m to 515 m where it enters the Tees around 2 km upstream of Cow Green Reservoir. The Whin Sill outcrops just to the south of Little Dodgen Pot Sike towards Cow Green Reservoir (See Figure 3.2). Most of the LDPS catchment is underlain by bands of middle Carboniferous limestone (Johnson and Dunham, 1963). Over almost the entire catchment a layer of glacial clay till forms the base for blanket peat. This clay layer is usually around 30 cm deep, although it can contain many coarse clasts resulting in a clayey diamict. Most of the peat in the catchment is intact, with only three gullies in the main part of the catchment (Figure 7.1 and 7.5). There is an eroded peat hagg-island system at the head of the catchment which feeds one of the limestone sinkholes (see Figure 7.1). Examination of the aerial photograph combined with ground survey indicates that less than 5 % of the catchment is eroded with gully floors well vegetated. Peat flush zones are common and can be identified by the wide areas of lighter-coloured more grassy vegetation on the aerial photograph (Figure 7.1). The blanket peat cover is typically 1.5 – 2.5 m in depth although it is up to 3.2 m in places. The peat in the headwater zone tends to be shallower (circa 80 cm) where there are steeper slopes (averaging 5^{o}). The slopes below the confluence of the two headwater sections (originating from the two risings discussed above) are gentler, often around 1 – 2^{o}. The stream long profile is shown in Figure 7.2. The stream is slightly steeper in its upper and lower reaches, but slopes more gently along its mid section. Along much of the upper sections of the stream channel, the watercourse is enclosed by a roof of clay and vegetation (Figure 7.3a) such that the stream itself is effectively contained within a pipe. The enclosed sections are indicated on the long profile (Figure 7.2). Occasionally these enclosed stream sections slump as shown on Figure 7.3b causing a change in stream course and episodically adding sediment to the system.

Most of the LDPS basin faces northeast, although the lower third of the river course runs eastwards. Jones (1994) showed that most piped catchments that have been examined in Britain face south such that piping has been associated with cracking of the peat surface during the summer months. Whilst summer desiccation is common at Moor House (see Chapter 3) it does not appear to be as common as in the Welsh uplands (Gilman and Newson, 1980), the numerous examples from which may skew Jones' (1994) results. There is one grip running across from the catchment divide to the stream channel.

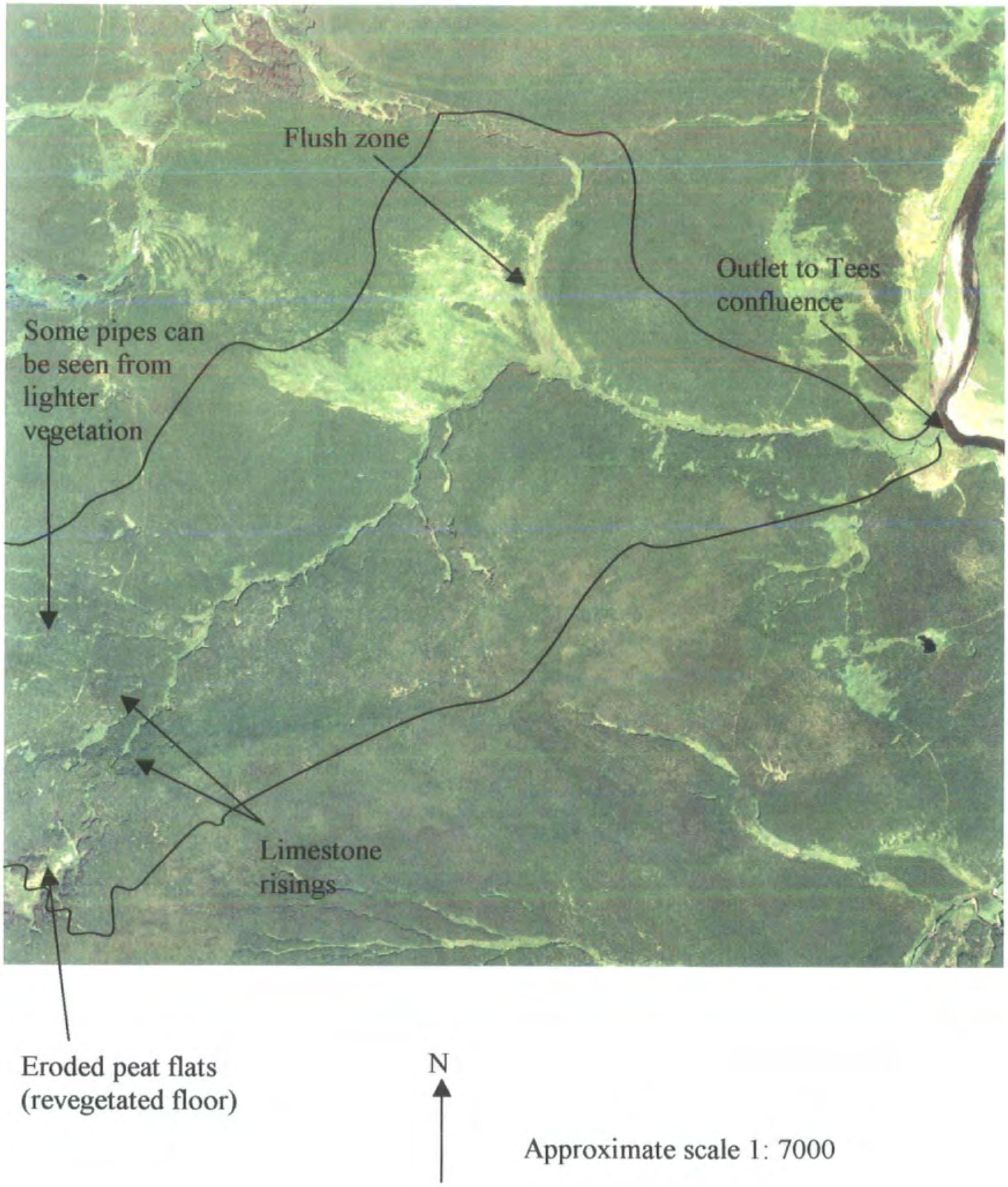


Figure 7.1. An annotated aerial photograph of the LDPS catchment. Reproduced with kind permission from NERC, site 94/9(4), taken 6.8.95, run 7, plate 8856.

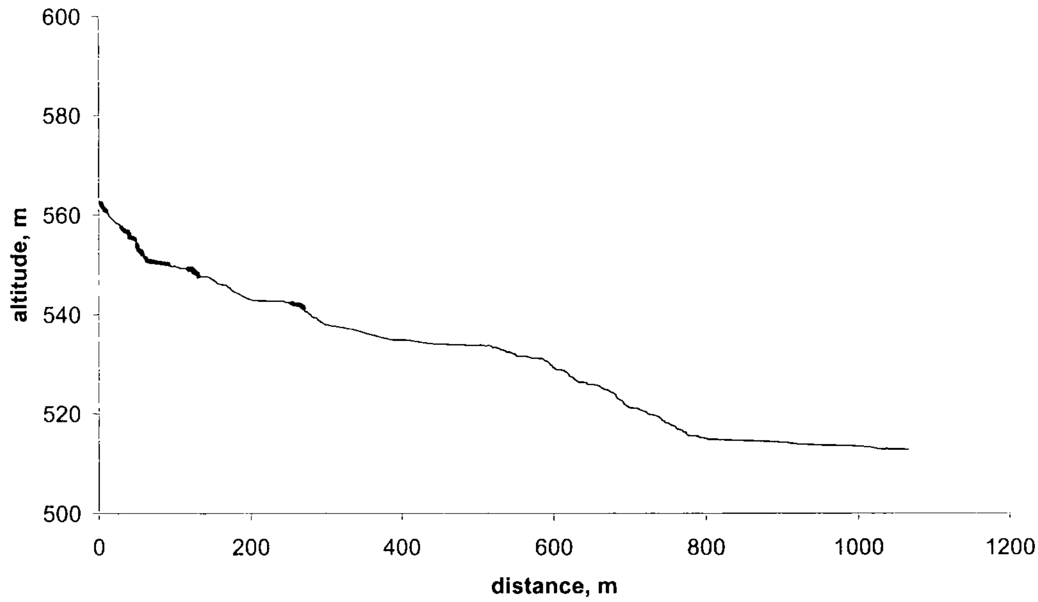


Figure 7.2. The long profile of LDPS. Bold sections indicate where the stream is enclosed by a peat or clay deposit.

7.3 Stream Discharge at LDPS

Stream discharge was gauged by an Ott R16 stage recorder installed in June 1999 on a straightened section 60 m upstream from the outlet to the Tees. A rating curve for the straightened section was derived from repeated flow measurements using an electromagnetic velocity probe. Median discharge for the study period 0.009 m^3 s^{-1} (0.07 mm hr^{-1}) and runoff to rainfall ratio for the catchment was 83 % which is higher than for Trout Beck (Table 7.1) and indicates the limited storage capacity of this blanket peat catchment. Mean data for the comparative period over which the same storms have been analysed for both Trout Beck and LDPS is indicated in Table 7.1. The LDPS catchment displays similar lag times to Trout Beck but has a shorter recession period, as one would expect for a smaller catchment. Hydrograph intensities are greater for Trout Beck, though, suggesting a flashier response. Nevertheless, the response of the two catchments is very similar. Area-weighted discharge peaks are usually lower in LDPS than in Trout Beck, although the maximum peak discharge (mm hr^{-1}) during the comparative period was slightly greater in LDPS as occasionally storm peaks are greater than in Trout Beck (e.g. day 263 1999, Figure 7.4). Figure 7.4 suggests that LDPS displays slightly smoother hydrograph peaks than the more spiked Trout Beck response.

a)

b)

Figure 7.3. Example areas where LDPS runs below the surface. a) Large outlet within clay deposit. Note the smaller outlet above and to the left of the main outlet which operates during high flow. b) Crescentic slumping of river terrace around a subsurface section of LDPS.

Table 7.1 Mean hydrograph characteristics from LDPS compared with Trout Beck for the period Jul 99 to Dec 99 (in bold). Brackets indicate Trout Beck data for Oct 94 to Dec 99 (second row) and LDPS data for Jul 99 to Jun 00 (fourth row).

	Runoff ratio	Peak Q	Median flow	Time to peak	Recess	Peak Lag	Intensity
Trout	**77.8**	**4.79**	**0.05**	**6.9**	**27.3**	**3.5**	**34.4**
Beck	(72.0)	(6.28)	(0.05)	(6.6)	(28.9)	(2.7)	(38.8)
LDPS	**80.4**	**4.97**	**0.04**	**7.2**	**25.1**	**3.3**	**31.5**
	(83.0)	(5.25)	(0.07)	(7.9)	(24.8)	(3.2)	(32.3)

Runoff ratio = Total rainfall divided by total runoff, %
Storm Q = Total storm discharge, mm
Peak Q = peak discharge mm hr^{-1}
Time to peak = time from first recorded rainfall to hydrograph peak, hrs
Peak Lag = time from peak rainfall to peak discharge, hrs
Intensity = peak flow/10^6, m^3 s^{-1} divided by total storm discharge, m^3 (s^{-1}).

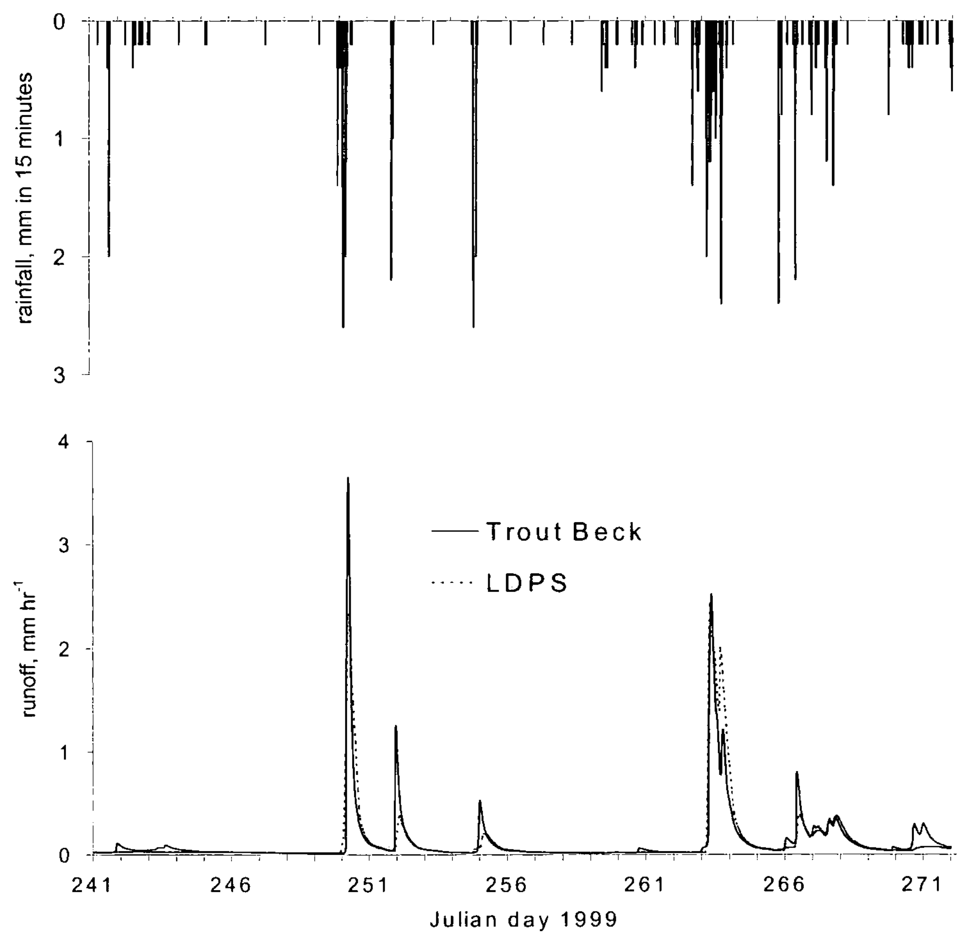


Figure 7.4. Discharge from Trout Beck and LDPS during days 241-272, 1999.

7.4 Pipe-form characteristics in the LDPS catchment

Figure 7.5 maps the main soil pipes discovered in the LDPS catchment. These pipes were originally identified by walking along the river channel and observing pipe outlets. The outlets were then traced back upslope where possible by following slight depressions in the surface and watching for occasional collapsed peat sections which allowed the pipe to become visible. Often the pipes were easier to identify during storm events. This is because jets of water emerging from surface outlets could be observed where the pipes were full and a back pressure was operating (e.g. Figure 7.6). Similar jets were observed by Gilman and Newson (1980). The gurgling of pipeflow water could also be heard beneath the peat during some (non-windy) storm events. Nevertheless it was very difficult to accurately map pipe direction, length and continuity. Those pipes that could be identified were mapped using a differential Global Positioning System (Higgitt and Warburton, 1999) as were other hydrological features such as gullies, seepage/flush zones and areas of bog pools.

The four areas where a concentration of bog pools can be found are associated with piping. Pipes 11, 13, 16 and 18 (pipe identification numbers are given in Figure 7.5) run downslope from the bog pool areas; pipe 13 then spills on to a *Sphagnum* flush or seepage zone area. Several other pipes in the catchment also feed these flush areas such as pipes 2 and 25. Flush zones can also feed pipes, as in the case of pipe 1; McCaig (1979; 1984) observed similar features in the Southern Pennines and termed these flushes which feed pipes 'secondary source areas'. Many of the pipes discharged onto the surface causing overland flow which then ran downslope often back into the pipe system via sinkholes. Both gullies 1 and 2 have pipes entering at their heads. Taylor and Tucker (1970) were among the first to suggest that piping in peat could be lead to dissection. Burt *et al.* (1990) note that the role of pipes for gully extension and stream channel initiation is uncertain. On the deep blanket peat of the southern Pennines, pipe collapse seems only to be important at a few sites. At LDPS gullies 1 and 2 run directly into the stream channel and can be seen in Figure 7.9. Gully 3 runs over the catchment divide from the neighbouring Great Dodgen Pot Sike catchment, and then feeds a flush zone which also has a pipe beneath it. Pipe 8 could be identified from collapsed peat features and varied widely in cross-section. It was impossible to identify from the surface the exact direction and depth of the piping down to the river channel but the GPR was used on this pipe and the results will be discussed in Chapter 8.

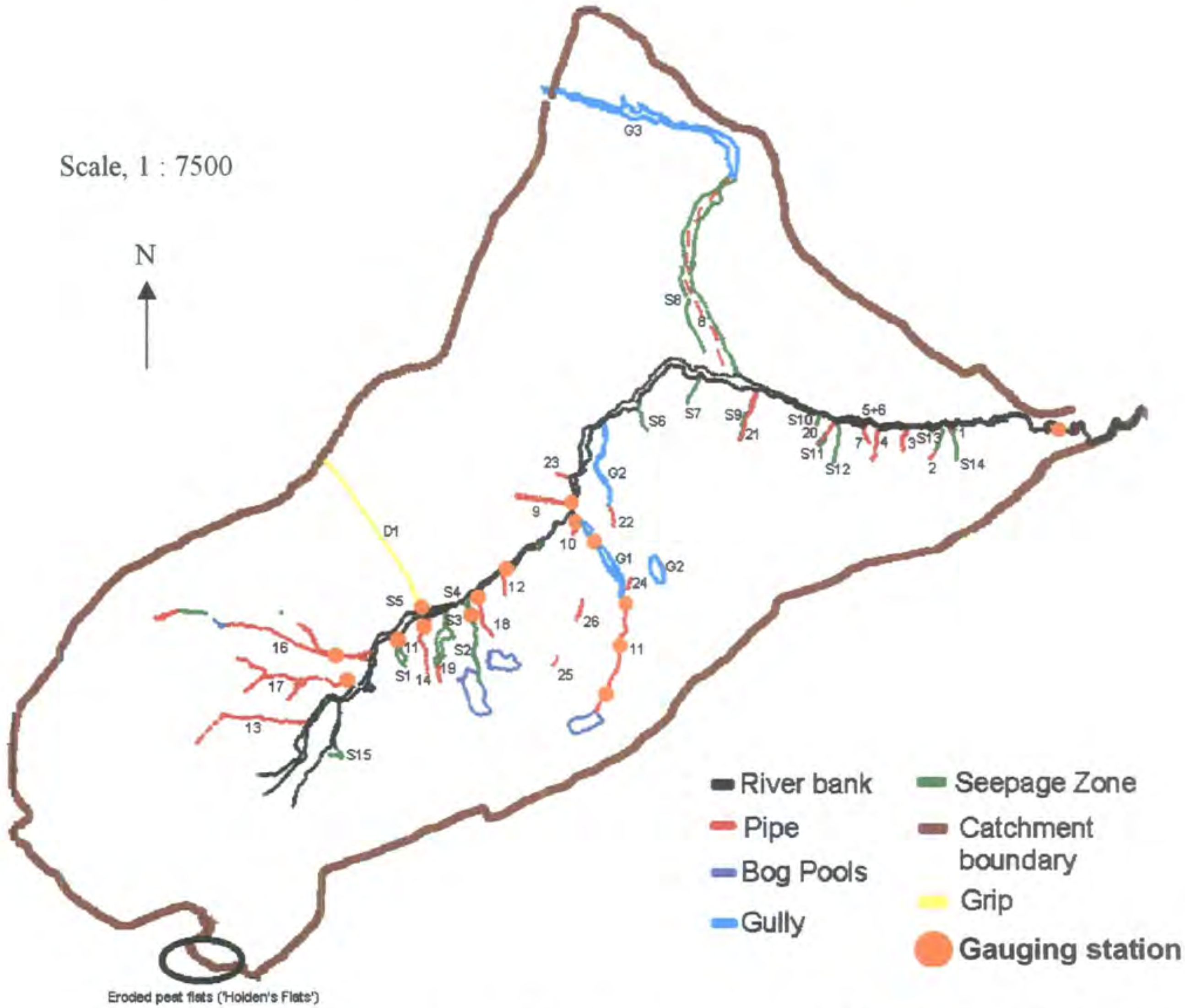


Figure 7.5. Main features and pipes of the LDPS catchment. Mapping was done using a Magellan GPS. Table 7.2 gives further details on the features within the catchment. Pipes 9, 10, 11, 12, 14, 16, 17 and 18 are monitored at locations indicated by the orange circles. The grip (D1), two seepage zones (S1, S2) and gully 1 are also gauged.

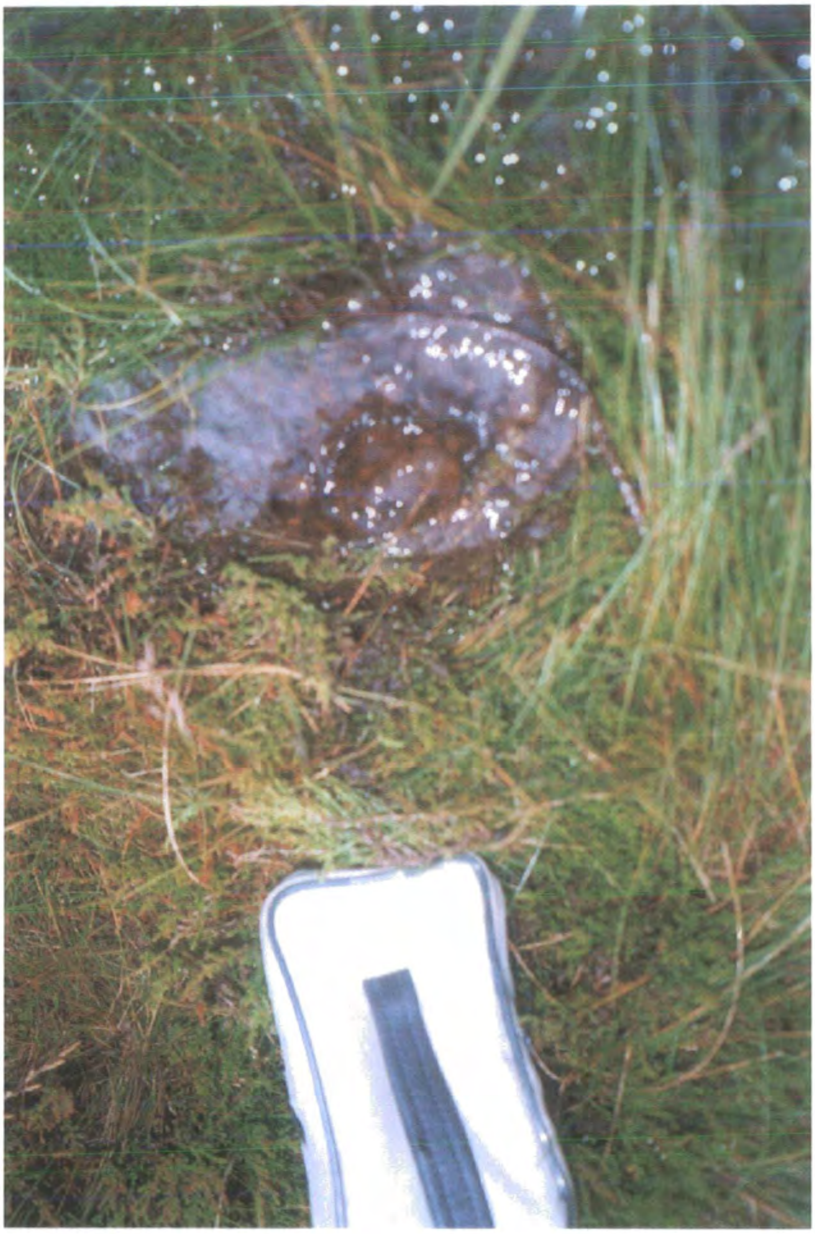


Figure 7.6. Spring of water emerging upwards at the surface from a subsurface pipe. As pipe flow capacity is reached it fills and a head of water develops from upslope.

Several of the pipes in the headwater area are associated with vegetation patterns; grasses dominate some piped areas and can be identified on the aerial photograph (Figure 7.1). Jones *et al.* (1991) and Jones (1994) describe similar associations of piping and grass 'lanes' in the Maesnant basin. The location of pipes 13, 16 and 17 can be identified on Figure 7.7 by the lighter-coloured vegetation. Notably these pipes are located in areas of fairly shallow peat (Table 7.2), although deeper than on Plynlimon. In most cases, however, no vegetation changes were associated with the soil pipe. Figure 7.8 gives an example where no vegetation change is associated with the pipe but a slight topographic depression can be observed.

The longest flowing pipes extend over 150 m across the 1-3 degree river terrace slopes and have mean diameters ranging from 3 cm to 70 cm. The pipe characteristics are shown in Table 7.2. These are the characteristics at the outlet of each pipe. Nine of the 26 pipes were ephemeral with flows in pipes 19 and 25 also reaching very low levels (*circa* 1 litre hr^{-1}). The pipes vary from being shallow within the peat layer, deep within the peat, at the peat-substrate interface, or entirely within the substrate. Half of the pipes were at a depth of over 1 m with some being at almost 2 m. Thus, the LDPS catchment is the first blanket peat catchment study with continuous pipeflow monitoring of both deep and shallow soil piping. The ephemeral pipes at LDPS are not like those reported at Nant Gerig (Gilman and Newson, 1980), Maesnant (Jones, 1981, 1987; Jones and Crane 1982, 1984), or Shiny Brook (Gardiner, 1983; Anderson and Burt, 1982) because they are not simply the shallowest of pipes in the peat. Instead both ephemeral pipes and perennial pipes can be found at shallow and deep locations in the soil profile (Table 7.2). Thus Jones' (1982) theory that ephemeral pipes found at around 15 cm depth on Maesnant were fed by raising of the phreatic surface would be difficult to support at LDPS. This is because the phreatic surface would already be well above the height of many of the ephemeral pipe outlets at LDPS. Water tables are typically within a few centimetres of the surface for most of the year.

The distinction between ephemerally and perennially flowing pipes is often difficult to determine for the pipes in LDPS. This is because during dry periods all but one of the pipes almost completely cease flowing, perhaps with just a very slow dribble of less than 1 ml min^{-1} flowing from the pipe outlet. This is well below the threshold of most monitoring devices. In fact only one of the pipes in the catchment continued to produce significant (continuously measurable) flow during rainless periods (pipe 10). Perhaps

Figure 7.7. Vegetation changes along the tracks of pipes on the slopes of the upper reaches of LDPS. Left to right pipes 13, 17 and 16.

Figure 7.8. Surface depressions following the route of pipe 11.

Figure 7.9. Gully 1 and 2 within the LDPS catchment. A flat bog pool area can be seen on the upper slopes (see Figure 7.14d) which appears to be the source for some runoff into pipe 11. Pipe 11 supplies runoff to the head of gully 1 (the gully on the right).

this was the only truly perennial pipe. For the purposes of this study, ephemeral pipes are those which completely cease producing runoff. For example pipe 12 only produces runoff during high stream flows (see below). This is likely to be because this pipe is connected to another pipe which at high discharge overflows into the secondary channel. Thus the difference between ephemeral and perennial pipes is again not clear; ephemeral pipes may simply be extensions of the perennial channel network. The distinction between the two pipe types widely quoted in the literature may not be very useful in these upland peat catchments because it is not often clear when a very low discharge should be considered as 'zero discharge'. In terms of pipe size dimensions there are no significant differences between the two types of pipe.

Figure 7.10 indicates that all but one of the 'ephemeral' pipes are located entirely within the peat; the other, pipe 12, is within the substrate. All of the pipes found at the peat-mineral interface are 'perennial'. The largest six of the pipes in diameter are perennially flowing. The pipes are generally of a much larger diameter than in the case of the Upper Wye pipes (Gilman and Newson, 1980). This is probably related to the fact that the shallow nature of the peat in the Upper Wye restricts the dimensions of the pipes. In deeper peat it seems that pipes can erode to greater diameters. At LDPS eight of the pipes were based on the interface between the peat and the underlying substrate. Piping is typically found in soils associated with marked reductions in vertical permeability (Jones, 1990) and are often at the interface between organic and mineral horizons (Jones, 1981). Four of the pipes were found to be entirely within the substrate.

Figure 7.10 demonstrates that the pipes at the interface tended to be elongated along the horizontal whereas pipes entirely within the peat are more rounded or tend to be elongated in the vertical; Figure 7.11 gives examples of these tendencies. This may be related to the difference in the ability of the peat to degrade in comparison to that of the clay and till beneath it. Jones (1981) suggests that there is some evidence to indicate that small rounded pipes evolve to larger flat bedded or rectangular pipes and suggests that 'horizontally lenticular' pipes are typical of shallow peats in Britain (Weyman, 1971; Jones, 1975; Morgan, 1977; Atkinson, 1978). Jones (1975) found that 37 % of pipes at the streambank of Burbage Brook were flat bedded and horizontally lenticular compared with 12.5 % in Afon Cerist. This is the type of geometry generally expected in open channels which would therefore suggest non-capacity flow control on the geometry (Jones, 1981). Gilman and Newson (1980) observed smooth beds and rough

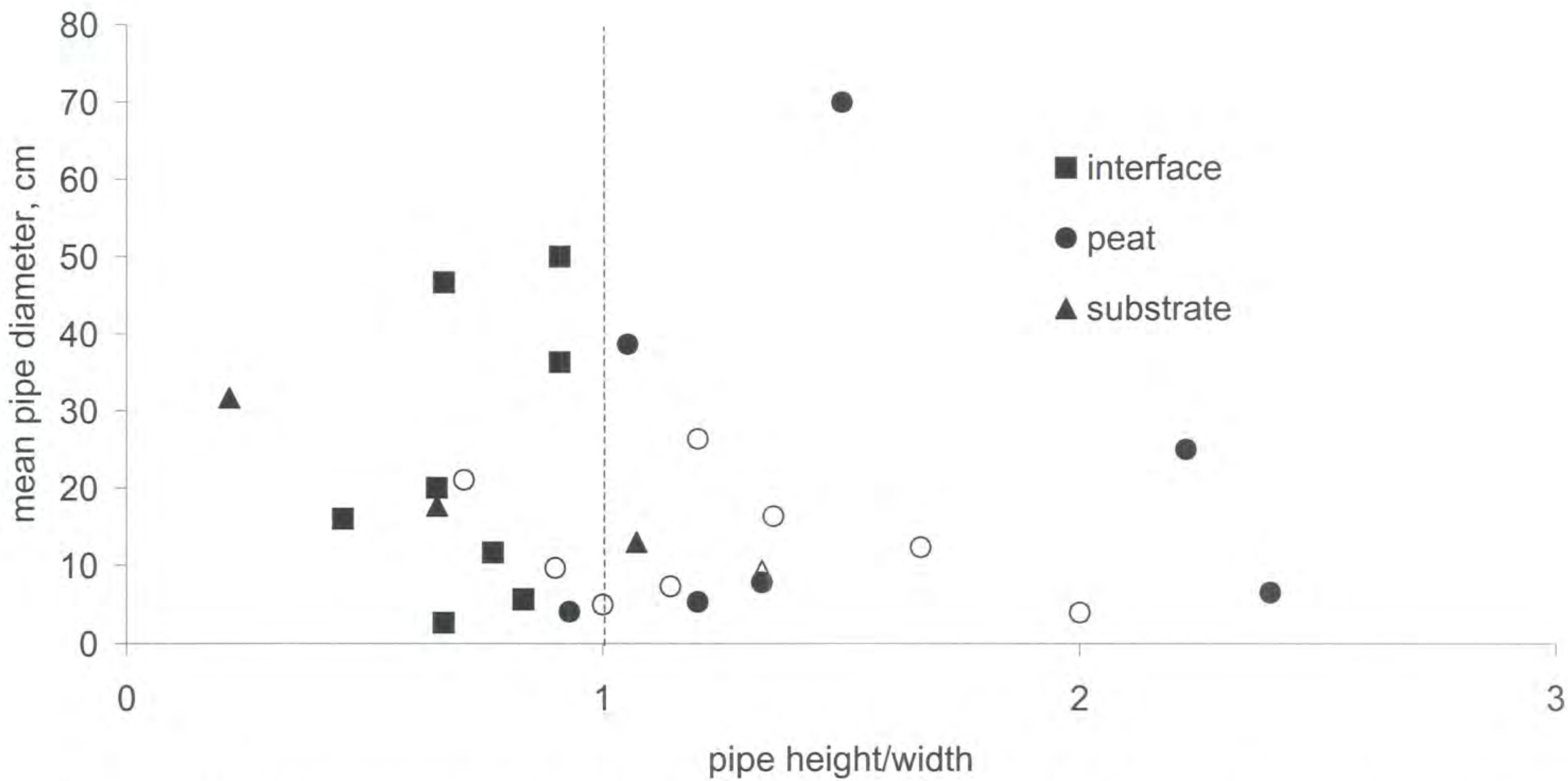


Figure 7.10. Pipe diameter, shape, and location in the soil profile at LDPS. closed = perennial, open = ephemeral.

pipe roofs in Cerrig yr Wyn, Plynlimon. However, the fact that the peat-mineral interface seems to affect pipe geometry at LDPS (generally being associated with horizontal elongation) suggests that erodibility of floor material may be an important factor. Concurrent with the findings of Jones (1981), there is no relationship between pipe length and pipe cross-sectional area. There is also no relation between pipe cross-sectional area and depth or location within the soil profile at LDPS; unlike the findings of Jones (1981), vertically elongated pipes at LDPS were not usually larger in diameter.

The average cross-sectional area of pipe outlets per kilometer length of streambank is taken as the best measure of intensity of piping activity along the streambank (Jones *et al.*, 1997). Table 7.3 shows that LDPS has a relatively low intensity of piping along the streambank compared to other catchments studied and a slightly lower frequency of piping along the streambank than at Maesnant. Jones *et al.* (1997) suggest that soil piping in Britain tends to occur on catchments with steeper stream slopes than average (7.7^{o} compared to 5.9^{o} national average. The volume and density of piping that has so far been identified on LDPS is much lower than at the other sites tabulated and mean stream slope and valley side slope are much gentler at LDPS. However, there may be a much greater density of pipes than indicated by this preliminary mapping exercise with the pipes being more difficult to find than at Maesnant and other sites because they are often deeper.

7.5 Pipeflow measurement

7.5.1 Choice of gauging sites

Runoff was monitored at 15 pipe and flush zone sites. It was impossible to monitor discharge from all pipes and all seepage areas due to limitations on expense, disturbance, and equipment availability. Ten piped sections, one grip (D1), one gully (G1) and two flush zones (S1 and S2) were monitored as well as the main stream gauge just upstream of the Tees outlet. The monitoring sites are shown in Figure 7.5. It was decided not to monitor the pipes and seepage zones downstream of pipe 23. This is because field observation and the measurement of runoff with stopwatch and measuring cylinder showed that many of these pipes were not major sources of runoff; manual measurement also indicated that runoff response was similar to those in the upper part of the catchment, although fairly diverse. It was hoped that the pipes that were monitored provided a good cross-section of the response types found over the entire catchment and were the major pipeflow inputs to the stream.

Table 7.2 Pipe characteristics in the LDPS catchment. Note that pipe dimensions, cm, are measured at the outlet.

Pipe	Peat depth	Depth of pipe roof	Depth of pipe base	Mean diam	Length	Flow type: eph/per	Monitored
1	170	168	170	3		P	
2	150	133	130	4	10	E	
3	105	60	65	5	30	E	
4	160	115	165	47	40	P	
5	160	90	110	16		E	
6	160	60	77	21		E	
7	75	73	78	4	20	P	
8	175				225	P	
9	180	20	25	12	60	E	Y
10	110	115	130	13		P	Y
12	130	135	147	10		E	Y
13	80	90	75	20	80	P	
14	110	85	110	19	55	P	Y
15	95	25	37	6	5	P	Y
16	110	105	115	32	150	P	Y
17	60	20	40	47	125	P	Y
18	75	5	10	7	60	E	Y
19	125	30	24	5	17	P/E	
20	120	115	125	16	20	P	
21	135	150	135	18	60	P	
22	220	30	34	4	20	P	
23	180	183	190	12	15	P	
24	250	150	180	27	10	E	
25	200	30	20	20	10	P/E	
26	220	175	184	10		E	
D1	100		50		180		Y
S1	95				25	P	Y
S2	90				55	P/E	Y
11a	260	15	35	25	10	P	Y
11b	225	5	50	40	70	P	Y
11c	245	0	90	70	115	P	Y
11d	55					P	Y

Table 7.3 Identified intensity of piping in LDPS compared to other piped sites

Catchment	Cross-sectional area of pipes m^2 km^{-1} streambank	Pipe frequency km^{-1} stream bank	Mean diameter of pipes, cm	Pipe volume in main area of piping, m^3 km^{-2}	Pipe density in main area of piping, km km^{-2}	Mean annual ppt, mm	Mean altitude, m	Mean main stream slope, degrees	Mean valley side slope, degrees
LDPS	**0.026**	**9.5**	**19**	**22**	**4**	**2000**	**540**	**2.2**	**3.0**
Maesnant, Cambria (Jones and Crane, 1984)	0.656	14.5	10	2099	98	2200	541	8.1	9.5
Afon Cerist Snowdonia (Jones, 1975)	0.567	80	10			2000	150	1.7	7.5
Burbadge Brook, Peak District (Jones, 1975)	0.544	89	9			1000	357	2.0	10.2
Cerrig yr Wyn, Cambria (Gilman and Newson, 1980)			5	353	180	2200	472		9.0
Nant Gerig, Cambria (Gilman and Newson, 1980)			10	55.3	44	2200	495		9.0
East Twins, Blackdown Hills (Stagg, 1974)			4	156.8	142	1100	244		2.3
Wansfell, Lake District (Jones *et al.*, 1997)			3	124	175	2000	360		11.0

a)

b)

Figure 7.11. Example pipe outlets. a) Vertically elongated pipe outlet entirely within the peat. Note the mineral sediment on the floor of the outlet indicating upslope pipe contact with the mineral substrate. b) Horizontally elongated pipe at the peat-clay interface.

One of the main sources of runoff came from the hillslopes draining into seepage zone and pipe 8. Approximately 15 % of the catchment fed this zone (shown in Figure 8.11). However drainage was generally too diffuse to monitor and the pipe was too awkwardly located within the peat with no clear outlet to the river to monitor flow. Pipe 13 was not monitored because its base was on a clayey diamict with loose gravel base and it became too difficult to ensure that all runoff was not leaking around any measurement device.

7.5.2 Discharge measurement

Pipe discharge was monitored either by insertion of a weir plate into a pipe or, where this was too difficult, water from a pipe outlet was channeled via plastic sheeting and tubing into a plastic box with a V-notch at the front end. The weirs were gauged by the use of a water level sensor consisting of a one-turn potentiometer; this is turned by a float attached to a pulley wheel and counterbalance by 70 kg strain braided fishing wire (Figure 7.12). The design details are given in full by Jones *et al.* (1984) except to note that the tape recording data logging system described by Jones *et al.* (1984) has been superceded. The potentiometer was connected to an available channel on a Campbell CR10X datalogger. The setup allowed stage to be recorded with a resolution of +/- 1 mm thus allowing high flow discharges to be recorded to the nearest 50 ml s^{-1}, and low flows to the nearest 50 ml min^{-1}, averaged over 15 minutes. Flows lower than around 100 ml min^{-1} (1.6×10^{-6} m^3 s^{-1}) could not be gauged and tipping buckets would prove more accurate under these conditions.

Jones *et al.* (1984) note that, although the British Standard (BS 3680 Part 4A) for thin plate weirs should be followed as far as possible, there is no standard to cover small weirs suitable for many applications in hillslope hydrology. For most of the pipes, the sharp-crested weir plates were set directly into the peat where possible and a good length of plate kept either side and below the cut-out portion to limit seepage and erosion around the edge of the plate. V-notches were usually 45° although where higher flows were likely 90° V-notches were cut. Suitable floats were constructed from plastic cistern ball-floats, or rounded plastic jars part filled with water and antifreeze to float at the maximum diameter when counterbalanced by a metal weight of 120 g. Stage was recorded at 15-minute intervals and converted to discharge using a calibrated rating curve produced separately for each weir. Examples of the low-cost pipeflow weirs with float and pulley system in operation are shown in Figure 7.13.

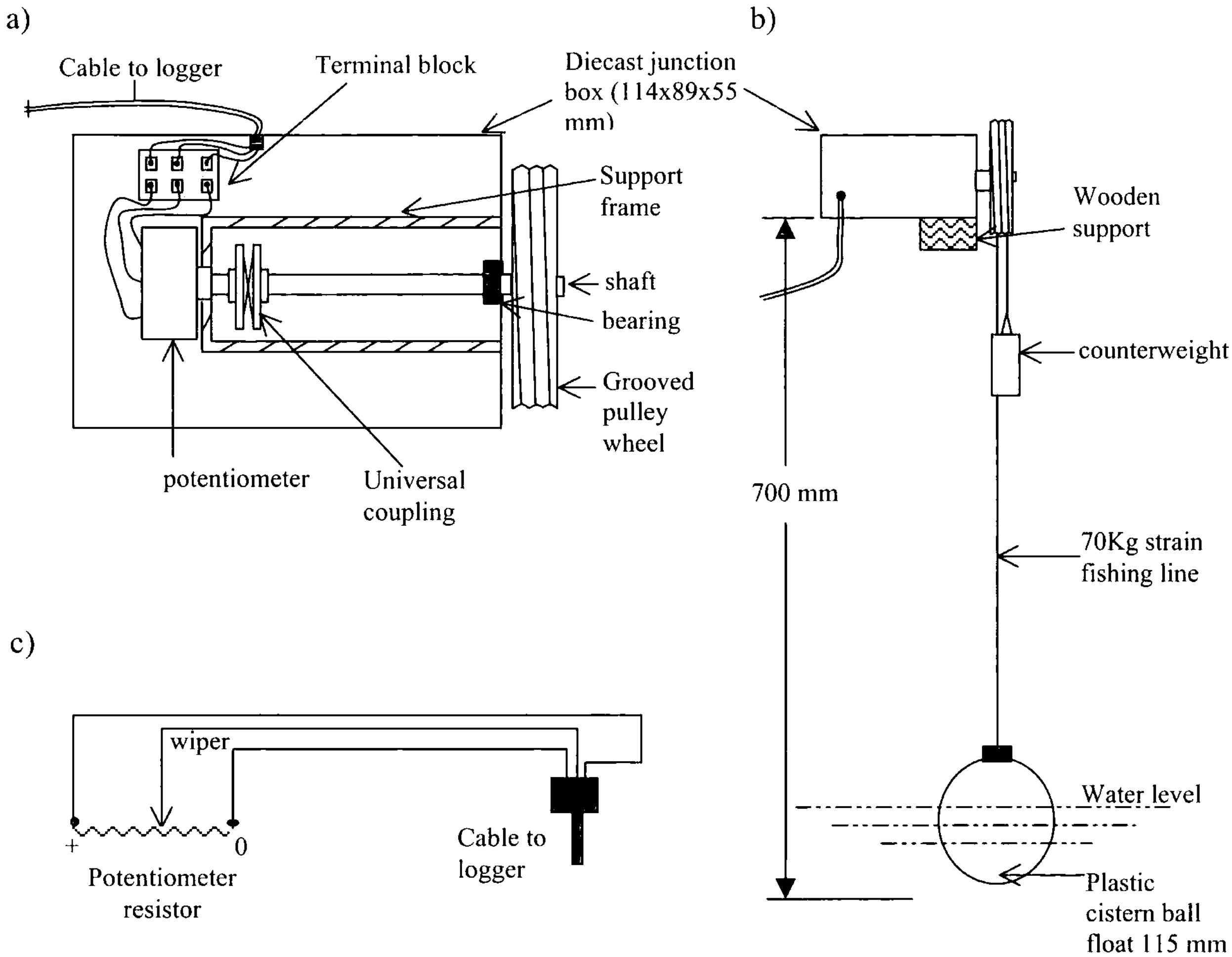


Figure 7.12. Design of water level sensors used to measure discharge from the pipes in the LDPS catchment. a) potentiometer housing b) float assembly c) potentiometer connectors.

a)

b)

Figure 7.13. Examples of the pipeflow gauging structures constructed in the LDPS catchment. Flow is measured by the calibrated float-stage recorder through either a) insertion of V-notch weir or b) channelling flow into a stilling box with v-notch cut into the front end.

Stage recorders at pipes 14 to 18 and S1, S2 and D1 were all fixed to one Campbell logger whilst pipes 9 to 12, including gully flow from G1 and three gauged sections of pipe 11 were fixed to the other logger. Thus 15 weir stage records could be simultaneously produced from July to December 1999. This period was when both loggers became available for use at the same time. Logger locations had to be carefully chosen in advance in order to make the most efficient use of cabling. Nevertheless around 900 m of three-core cabling was used in the monitoring configuration.

7.6 Pipeflow response

7.6.1. Pipe blockages

Two days after logger installation, flow at pipe 18 ceased and has not restarted. Zhu (1997) found that in the loess soils of China pipes were frequently blocked by collapses which could be re-opened in subsequent events. Therefore the piping had erratic discharges whereby instability of piping was a key factor in determining hydrological response. Although peat is not quite as readily mobilised and erodible as the loess soils of China it is likely that collapse within pipe 18 has blocked the flow of water to the original outlet. Uchida *et al.* (1999) also found pipes that periodically blocked with sediment producing erratic hydrological response in a Japanese Cambisol. The case at LDPS demonstrates the dynamic nature of piping within blanket peat. As no storms were recorded from pipe 18, it will be ignored from the subsequent hydrograph analysis.

7.6.2. Ephemeral pipe response

Discharge from the LDPS monitoring sites for a 30-day period is given in Figure 7.14. It is immediately apparent that, although all of the sites display flashy regimes, there is a marked difference between sites in runoff response. Pipes 9 and 12, which are ephemeral, show different responses with pipe 12 only responding to the larger events. The outlet for pipe 12 is entirely within the clay and the data could suggest that this pipe is connected to another pipe such that it only operates for short periods at the height of the storm when another pipe (as yet undiscovered) overflows. This is not to say that there may be other threshold mechanisms operating within the LDPS catchment. Pipe 9 behaves very differently from pipe 12 with much slower recessions and broader peaks. Pipe 9 is within the peat layer at around 20 cm from the surface although its source may be deeper as water seems to be rising upwards at the outlet due to backpressure.

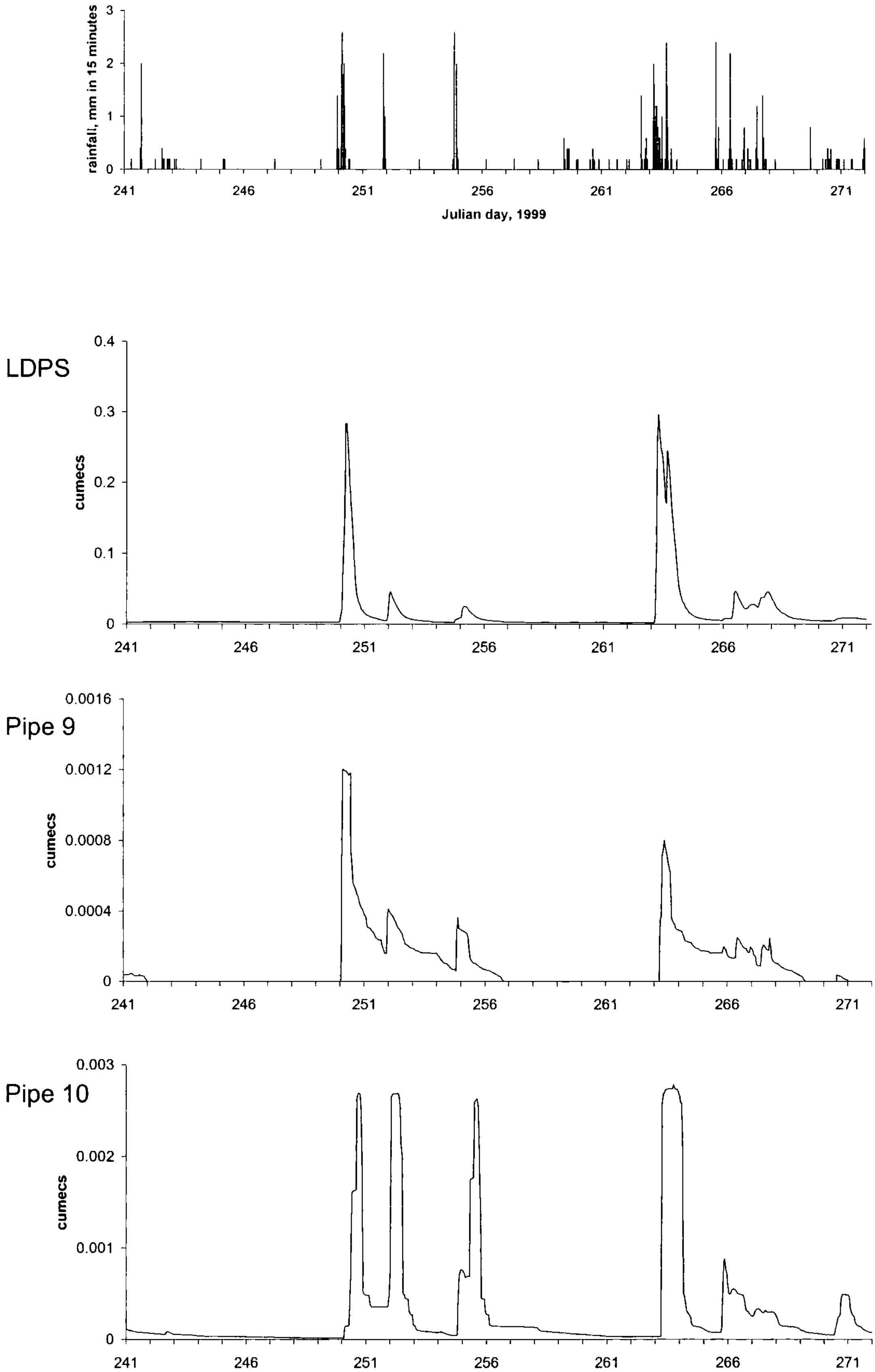


Figure 7.14. Discharge from the LDPS monitoring stations during days 241-272, 1999.

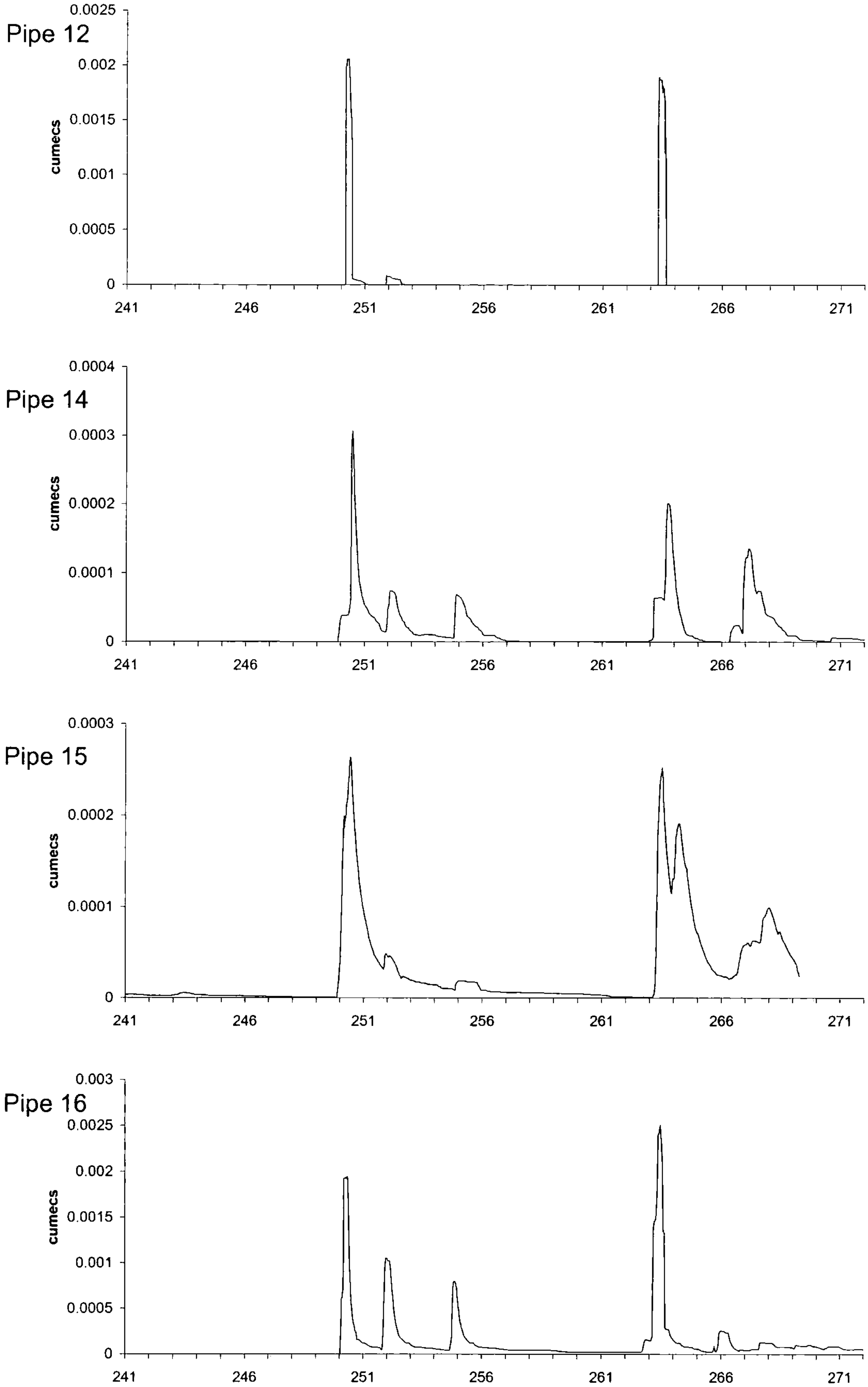


Figure 7.14. continued

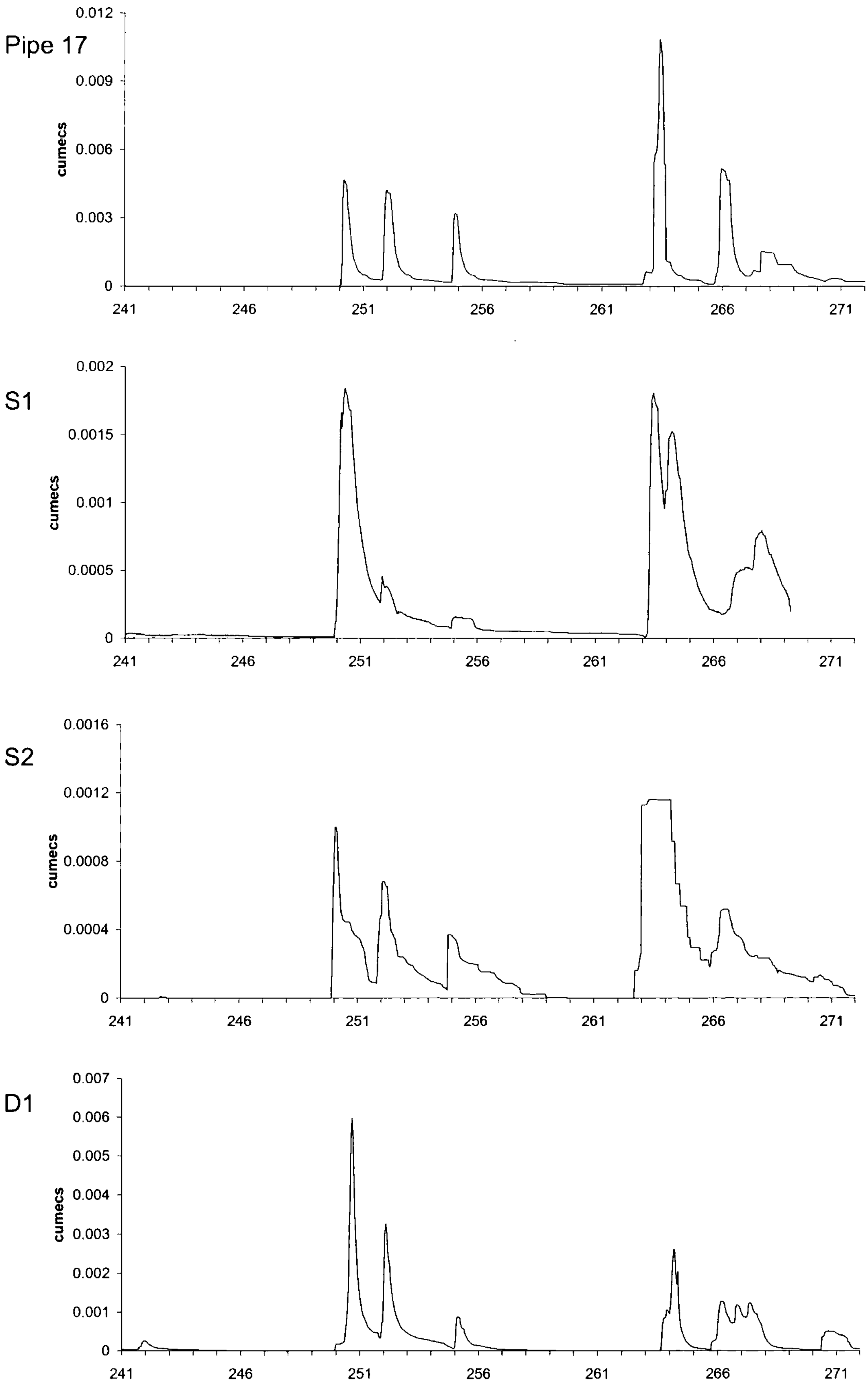


Figure 7.14. continued

7.6.3. Perennial pipe response

Pipe 10 behaves as if there is a limited capacity to the pipe such that most storms produce approximately the same peak flows. S1 and S2 and pipe 15 display similar discharge characteristics to each other with much broader hydrographs than the other sites. Pipe 15 is immediately adjacent to S1 and the close similarity of the hydrograph form suggests that pipe 15 is linked directly to S1. Pipes 10, 14, 16, 17 and D1 all have narrower storm hydrographs such that response to each rainfall event is much more distinct than from the other sources.

The pipes at LDPS all generally respond to low rainfall intensity and low rainfall total events, even after a dry antecedent period. There is no evidence to suggest that there is a 10 to 50 mm rainfall threshold which is required before pipeflow will respond as found in the mainly ephemeral systems that have been monitored by Gilman and Newson (1980) and McCaig (1983). It seems clear that the pipes in the LDPS catchment receive drainage far more quickly and in greater volumes than would be expected simply from diffuse seepage through the overburden. Nevertheless flow from the Sike itself is more flashy than any of the other monitored sites (except pipe 12) as indicated by the hydrograph intensity index (Table 7.4). Thus runoff production other than pipeflow probably dominates the catchment response and as demonstrated in Chapters 4, 5 and 6 OLF and acrotelm flow processes are important quickflow mechanisms in blanket peat catchments.

7.6.4. Pool- pipe-gully linkages – pipe 11.

Newson (1976) notes that the hydrological significance of rapid pipe drainage may be reduced because pipes do not always discharge directly into the surface streams of the catchment. Pipe 11 was monitored at source (Figure 7.15d), 70 m downslope and where it fed the head of gully 1 (Figure 7.15c). There was a further monitoring site near the mouth of the gully (site 11d). The gully is shown in both wet and dry conditions in Figure 7.15 a and b. OLF can clearly be seen running across the vegetated floor of the gully. Figure 7.15d shows the flat bog pool area which feeds the head of pipe 11.

a)

b)

Figure 7.15. a) Gully 1 during low flow conditions, b) Gully 1 during high flow conditions. Flow over the revegetated surface can clearly be seen.

c)

d)

c) The head of gully 1 where pipe 11 enters the gully. The wire which connects the flow recorder (situated 1 m upslope from the gully head) to the datalogger can be seen to the left of the photograph. d) Flat bog-pool area from which pipe 11 seems to emerge.

Table 7.4 Results from hydrograph analysis of 14 storms between July to December 1999

Location	Mean storm Q, m^3	Peak Q, $m^3\ s^{-1}$	Start Lag, hrs	Peak Lag, hrs	T_{rec}, hrs	Mean Intensity, s^{-1}
LDPS	10150	4.97	1.7	3.3	25.1	32.3
9	48.2	0.00080	2.4	1.8	39.2	15.1
10	157.2	0.00269	3.3	7.8	58.8	13.9
11a	56.9	0.00296	1.5	4.8	25.4	25.2
11b	31.7	0.00198	2.1	4.9	26.7	21.9
11c	103.0	0.00461	1.3	4.8	34.7	26.0
11d	335.0	0.01310	2.2	2.4	40.5	19.7
12*	32.9	0.00202	3.7	1.7	21.3	36.5
14	7.7	0.00021	3.2	3.9	20.6	20.0
15	12.1	0.00025	1.2	2.8	17.5	13.0
16	35.7	0.00251	0.2	2.6	29.9	26.6
17	35.7	0.00128	5.8	8.5	12	26.0
S1	93.2	0.00181	1.1	2.7	19.5	14.1
S2	78.4	0.00116	4.6	3.4	45.5	10.3
D1	266.8	0.01250	3.2	5.9	25.7	22.6

*pipe 12 responded to 10 of the 14 storms analysed

Storm Q = Total storm discharge, mm
Peak Q = peak discharge mm hr^{-1}
Start Lag = time from first recorded rainfall to hydrograph rise, hrs
Peak Lag = time from peak rainfall to peak discharge, hrs
Intensity = peak flow/10^6, $m^3\ s^{-1}$ divided by total storm discharge, m^3 (s^{-1}).

Runoff response from the four sites is shown in Figure 7.16. Response is broadly similar at all sites, although sites b and c most closely match each other. Discharge at site b is actually lower than upslope at site a. Newson and Harrison (1978) reported significant losses of pipeflow during experiments using artificially pumped water in natural ephemeral pipes and surmised that this situation was normal. However, Jones (1982) suggests that at Maesnant the pipes gain more in the form of effluent seepage than they lose by influent seepage so that discharge continues to increase downslope. It is unlikely that pipes lose much water through seepage on their floors and sides in LDPS particularly as the peat matrix at the depths of these pipes has a very low hydraulic conductivity. The loss of water via blocked sections of pipe spilling out to the

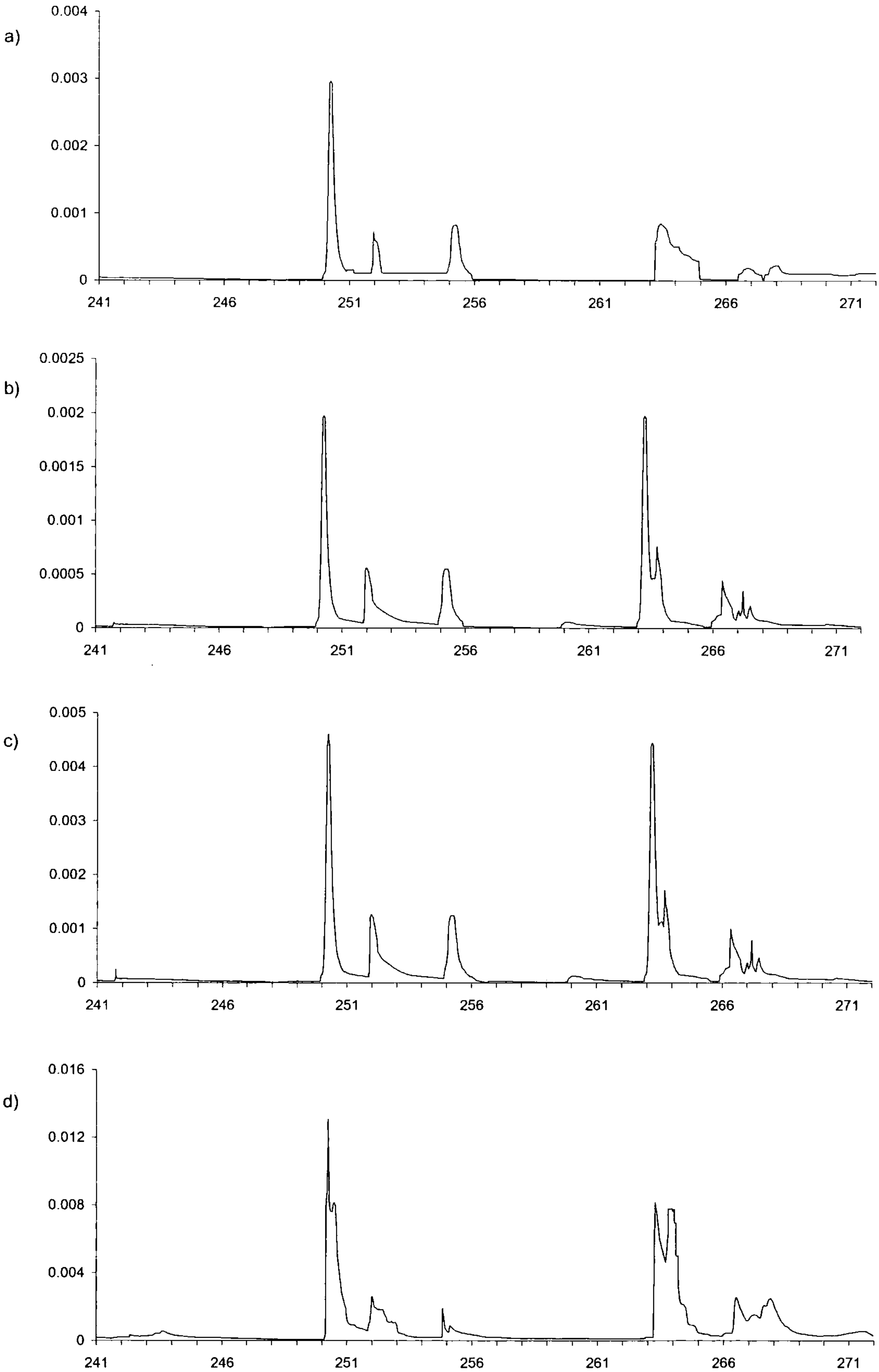


Figure 7.16. Discharge, cumecs, from pipe 11 and gully 1 monitoring stations during days 241-272, 1999.

surface or overflowing to other connected pipes seems more feasible. Observations of pipe 11 in the field showed that the pipe upstream of site b seemed to lose some water along its length. It was later found that about 10 m upstream of site b another pipe was discharging vertically upwards into pipe 11. There was thus a much more complex network of piping than originally thought. It may be that water was being leaked upslope to this secondary pipe, some of which was re-entering just upslope of 11b and some of which was being lost to an unobserved secondary pipe system. Discharge increases along the pipe from 11b to 11c where the pipe heads into gully 1. Runoff at the gully monitoring station is generally around three times that at the head. OLF and shallow subsurface flow running downslope into the gully will affect the overall hydrograph response. Thus Figures 7.16c and d appear different with secondary peaks on storm responses during days 250 and 255 present at site 11d but not at 11c.

Storm analysis for the monitored sites shows that peak lag times are shorter and recession times longer for the gully than at the three stations on pipe 11 upstream (Table 7.4). Mean start lag time (time from first rainfall to hydrograph rise) is slightly greater in the gully than at 11b but about 1 hour slower than at 11a or 11c. At Maesnant, Jones and Crane (1984) found that the storm hydrograph was recognisably established at the head of perennial pipes four hours before it reaches the outlets at the stream edge. There is no evidence from LDPS to suggest that this is the case here. Mean peak lag times are shortest from the gully such that the well vegetated gully floor does not seem to slow the storm wave down, although the vegetation may intercept and store some of the initial moisture such that start lags are greater.

7.6.5. 'Flashiness' of response

Hydrograph intensity is greatest from ephemeral pipe 12 which only operated during high flows (Table 7.4, Figure 7.17). The broadest (and hence least flashy) runoff response came from seepage zone 2 with a hydrograph intensity of 10.3 whilst S1 also had a smoother and less peaky hydrograph response. Pipes 10 and 15 similarly have less peaky responses than other pipes yet pipe 10 is at the mineral interface and pipe 15 is near the peat surface (see Table 7.2). Burt *et al.* (1990) suggested that pipeflow may be more important on shallow peat soils whereas on deeper blanket peats pipeflow from the impermeable catotelm will necessarily be restricted. However, the evidence presented from LDPS suggests that pipe depth has little to do with the nature of runoff response.

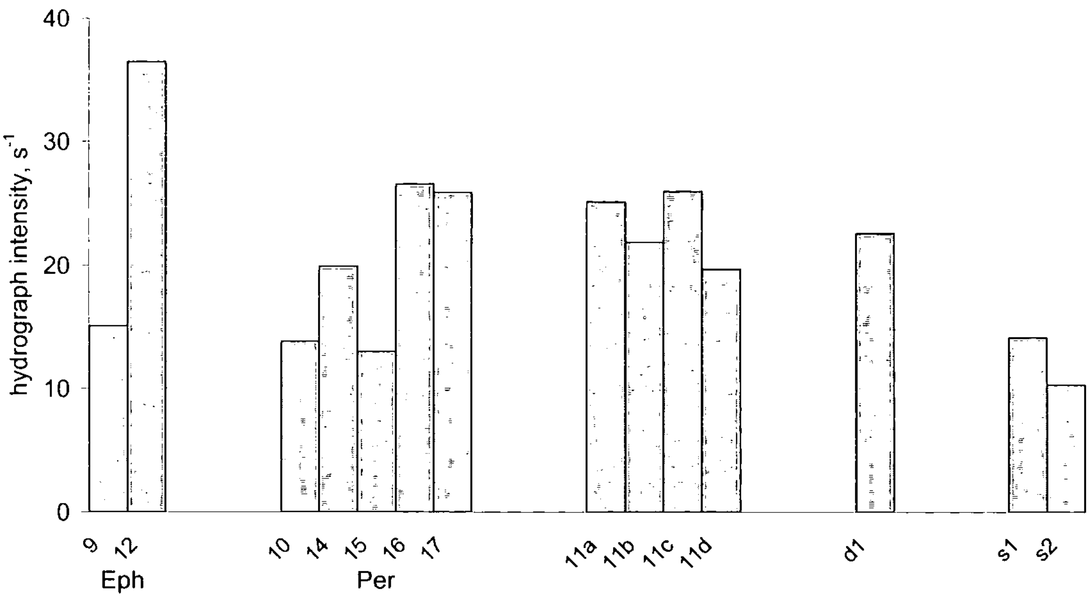


Figure 7.17. Mean hydrograph intensity at each of the LDPS monitoring stations, for 14 storms between June to December 1999.

7.6.6. Lag times

The shortest peak lag times (time from rainfall peak to discharge peak) for any of the pipes is the response from ephemeral pipe 9 with mean a peak lag of 1.8 hours, followed one hour later by pipe 15. Six out of eight of the pipes have peak lag times under 5 hours. Pipeflow lag times are similar to those at Maesnant and several other reported sites (Jones and Crane, 1984; Table 7.5). The initial speed of response from the LDPS pipes and seepage zones (0.2 to 5.8 hours) is much quicker than at Maesnant where start lag times (from rainfall onset to initial rise in hydrograph) ranged from 8.6 to 13.2 hours (Jones and Crane, 1984). The low hydraulic conductivity of the peat below 5 or 10 cm depth (e.g. see Chapters 4 and 6) means that it is unlikely pipeflow in the LDPS catchment is derived from diffuse seepage through the peat matrix. It seems much more likely that OLF and near-surface flow enters pipes where they are open to the surface at sinkholes or where a layer of *Sphagnum* provides the pipe roof. Macropores may also provide a bypass route for water to enter the pipe system.

Pipes that are deep in the peat at one point along their course may not necessarily be so deep at another. Pipe morphology appears to be very variable such that you can cut a peat face back a short distance to reveal a completely different set of dimensions. In this way pipe outlets dimensions can be misleading. Pipes are not simple linear channels for the passage of water; rather they are tortuous and constantly changing in cross section. Frequently the pipes run uphill such that back-pressures are required to transport the water upwards through those sections. In some instances a pipe can become a runnel where for a few metres there is no roof to the pipe. An example can be seen in Figure 7.18 just upstream from station 11b. It is notable that Gilman and Newson (1980) still called these open topped features pipes. Anderson and Burt (1982) suggested that routing of water could occur between cotton grass mounds along runnels. Subsequent growth of the peat could then roof-in the channels. Since many of the pipe systems seem to originate around areas of bog pools or flush zones it is likely that pipes tap surface and near-surface excess water from such collecting areas as the water filters through the surface living *Sphagnum* cover. Hence the extended flow found in longer recession times for many of the pipes is probably derived from a larger catchment area with very wet flush or pool features. Jones and Crane (1984) noted that much of the late recession drainage in the Measnant stream seemed to be coming from 'pools and bogs in the headwaters'.

For 12 of the 14 monitoring stations discharge starts to rise within +/- 2 hours with respect to streamflow rise (Figure 7.19a). Flow at three of the eight pipes rises, on average, before streamflow. This would suggest that peak pipeflow contribution may be on the rising limb of the hydrograph. Distribution of peak lag times is slightly positively skewed with a mean of 4.75 being higher than the mode of 4.05 hours. On average the time between streamflow peak discharge and pipeflow peak discharge is only 0.02 hours with a modal value of 0.71 hours. Hence if flow from all of the pipeflow stations is added together one may expect that maximum cumulative pipeflow discharge would be likely to occur within a few minutes of streamflow discharge peak.

The ephemeral pipes both have peak lag times around 2 hours shorter than that of streamflow (Figure 7.19b) and yet start lag times are longer than streamflow (Figure 7.19a). Flow in pipe 12 falls back to zero on average around 4 hours before the end of stream stormflow (Figure 7.19c). Figure 7.19c, however, shows that stormflow in pipe 9 lasts around 14 hours longer than in ephemeral pipe 12. There is also a diversity of response between the perennial pipes and the seepage zones with stormflow ceasing in some pipes up to 13 hours before stream stormflow whilst in others it may continue for a further 30 to 40 hours after stream stormflow has receded.

Table 7.5 Selected pipeflow characteristics recorded in the literature
*Ground surface angles at outlet †Ephemeral/Perennial

Source	Soil Type/Location	Peak discharge, l s^{-1}	Flow Type†	Diam, cm	Slope m m^{-1}*	Start lag hrs	Peak lag hrs
Present Thesis	**LDPS**	**4.6**	**E/P**	**3-70**		**0.1-3.7**	**1.6-8.5**
Weyman (1971)	Upper East Twins Basin, peaty podzol	1		2.5-5			
Stagg (1974)	Upper East Twins Basin, peaty podzol	0.75		2.5-5			
Jones (1975; 1978)	Bourne Brook, Cambridge	0.3		9			
Finlayson (1977)	Lower East Twins Basin, brown earth	0.11					
Waylen (1976)	Lower East Twins Basin, brown earth	0.12					
Knapp (1970: 1974)	Upper Wye, Plynlimon, peat	0.67-0.83		100			
Wilson (1977)	Nant Cwmllwxh, Brecon Beacons	1.5		60			
Jones (1987)	Maesnant, peat and peaty podzol	59.3				8.6-13.2	
Roberge and Plamondon(1987)	Lac Laflamme, nr Quebec, sandy till	1.11					
Gilman and Newson(1980)	Upper Wye, shallow peat	2.0	E	5-24			
Uchida *et al.* (1999)	Japan	0.18	E	5	0.71	12.1	3.7
Muscutt *et al.* (1990)	Afon Cyff	1.5	E	5-10	0.25	5	6
Zeimer and Albright (1987)	Casper Creek, USA	8.5	E	15-45	0.3-0.7		
Tsukamoto and Ohta (1988)	Hakyuchi, Japan	0.5	P	5	0.52	9	5
Koyama (1994)	Hiruzen, Japan	1.85	E	50	0.47	34	28
Woo and diCenzo (1988)	James Bay Coast, Canada	0.7	E	6-7	0.0005	0	1
Elsenbeer and Lack (1996)	La Cuenca, Peru	0.22	E	8	0.51	0	0
Uchida *et al.* (1999)	Kyoto, Japan, Forest Cambisol	0.18	E	5	0.5	11-12	1.6-3.7

Figure 7.18. Pipe section which became a runnel along a short reach just uplsope of gauging site 11b.

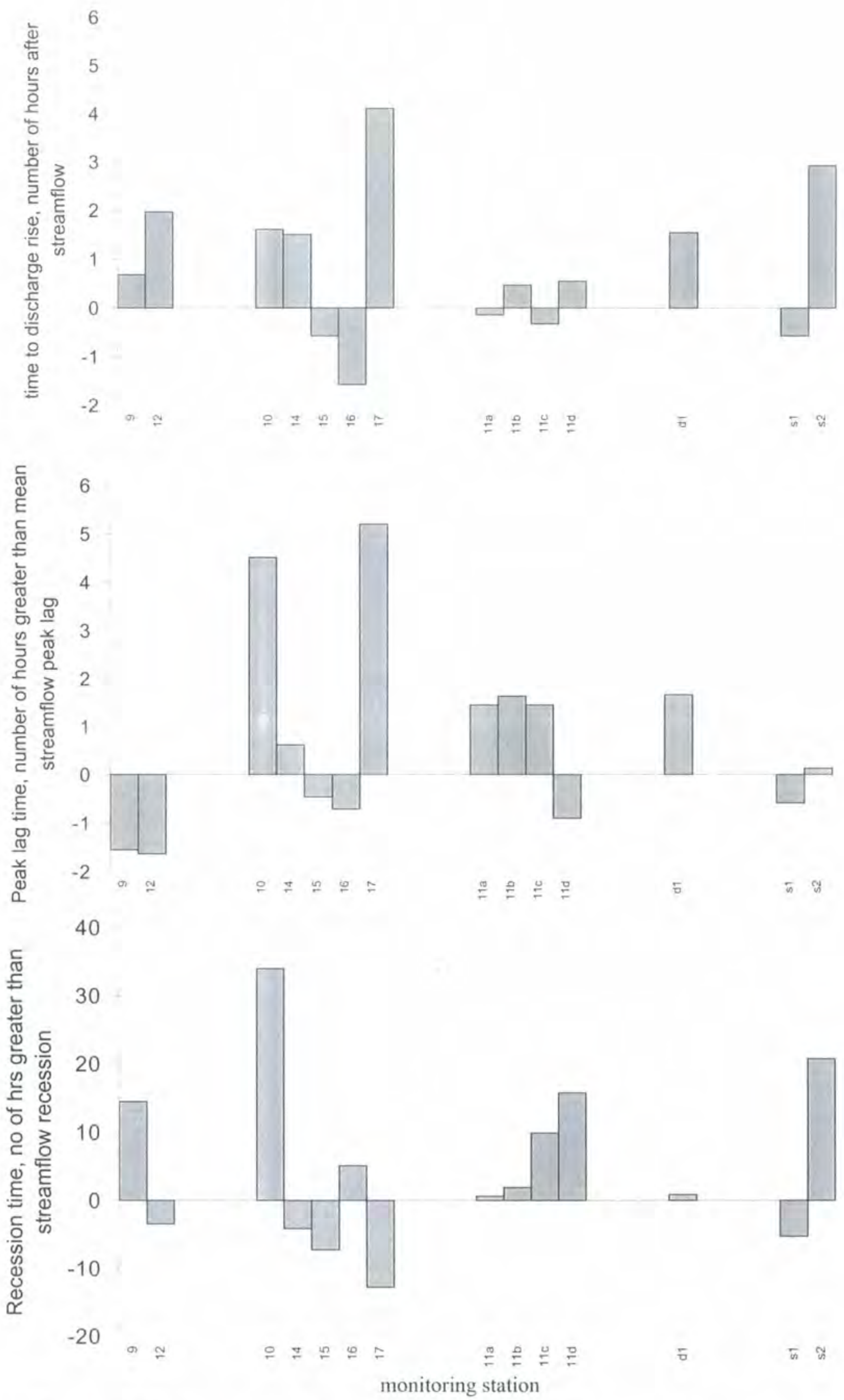


Figure 7.19. Lag time characteristics of the LDPS monitoring stations compared to streamflow lag times, a) Start lag, number of hours greater than streamflow begins to rise, b) Peak lag, number of hours greater than streamflow peak lag, c) Fall lag, number of hours greater than mean streamflow recession (time from rain end to flow back to pre-storm level).

Generally stormflow in d1 ceases around the same time as stream stormflow. The ditch is fed by OLF and near-surface runoff (as discussed in Chapter 4). For pipe 11 there is clear evidence of downslope drainage through the system. The upslope site (11a) drains first followed by the sites in order of distance downslope. This is more likely to be related to the downslope drainage of OLF and near-surface flow (as discussed in Chapter 4) than to slow drainage of the pipe. Mean flow velocities of the order of 8 cm s^{-1} were recorded in pipe 11. For the 115 m length of pipe this would give a mean travel time of 24 mins from top of pipe to bottom. This is far too short to account for mean recession times of 15 hours longer at 11c than at 11a. Notably pipe 11 is a shallow pipe often having its roof within 5 cm of the surface (see Table 7.2). As discussed in Chapter 4 the source areas producing OLF will move downslope after rainfall, as the saturated gentle slopes drain from the topslopes down. As the hillslope drains, runoff from the near-surface layers becomes minimal upslope and stormflow in the pipe head area ceases. Where OLF and near surface flow is being produced further downslope runoff can enter the pipe system via highly permeable *Sphagnum* pipe roofs where the pipes are shallow or even where there is no roof. Runoff may also enter the pipe system via macropores which bypass the peat matrix. Some of these were identified and monitored in Chapter 4. As was seen in Chapter 4 runoff produced from these macropore outlets on peat faces, which could be deep within the profile, was flashy and generally coincident with OLF and near-surface flow generation.

7.6.7. Pipe contributing areas

Mean storm discharge divided by approximate pipe length is greatest at site 11a, 15, and S1 (Figure 7.20a). These are all pipes fed by pools or wet flush areas, and have a larger catchment area. Calculating catchment areas for the pipes was difficult as it was often impossible to tell what areas were feeding the pipes, particularly on the gentler slopes and since occasionally pipes run counter to the surface topography. Comparisons have been made between pipeflow and other hillslope drainage processes in terms of velocity (Jones, 1987) and estimates of the total contributions to stream runoff from various sources in a basin (Jones and Crane, 1984). However, these comparisons lack a clear relationship with basin area that would allow wider generalisations about the relative efficiency and importance of pipeflow.

Dunne (1978) provided a valuable basis for making such generalisations for OLF and throughflow with collations of American and British data. These data have been plotted and extended by Kirkby (1985), Anderson and Burt (1990a) and Burt (1996). In order to map pipeflow data onto the graphs of Kirkby (1985) and Anderson and Burt (1990a) Jones (1997) advocates the estimation of surrogate pipe basin areas. This requires estimating the micro-catchment area feeding the pipes. Jones (1987) demonstrated that surface depressions are poor indicators of pipeflow contributing areas, probably because piping can develop routes that are at variance with the surface topography. Dye tracing can be used to test links between pipes but is impractical for delimiting catchment areas (Jones, 1997). Thus Jones (1997) advocates calculating surrogate 'basin area' through use of storm discharge and rainfall information. The largest contributing areas for each pipe were selected. This was done by calculating the dynamic contributing area (DCA) for each storm as given by equation 7.1 for perennial pipes and 7.2 for ephemeral pipes:

$$\text{DCA (per)} = \text{Total storm discharge in pipe} / \text{Total storm rainfall} \qquad [7.1]$$

$$\text{DCA (eph)} = \text{Total storm discharge} / \text{Total storm rainfall before end of pipeflow} \qquad [7.2]$$

After the areas had been calculated for each storm, the largest area was taken for each pipe to be a surrogate for basin area. The first of these formulae (7.1) was advanced by Dickinson and Whitely (1970) and used by Calver *et al.* (1972). It is purely an arithmetic estimate of the area of a catchment with a runoff coefficient of 1.0 in a storm. Essentially it is a minimum contributing area for the pipe during a storm event. Equation 7.2 was adapted by Jones (1997) for where pipeflow may end before rainfall stops. The limitation of applying the LDPS dataset to this technique comes from the fact that only 14 storms were analysed. Thus the largest contributing areas calculated for each pipe are likely to be underestimated. Nevertheless, the data from LDPS probably contain the greatest quantity of continuous pipeflow data outside of Maesnant. Three of the larger storm events during the monitoring period had precipitation totals of 25 mm, 36 mm and 43 mm respectively with 9.4 mm and 7.6 mm and 7.2 mm occurring in one hour. These are near the higher end of typical rainfall events in the North Pennines (e.g. see Figure 3.7c).

The distribution of peak discharge recorded from the pipes over the monitoring period closely matched the mean storm discharge patterns, except in ephemeral pipe 12 where storms were peakiest (Figure 7.20a and c). The peak flows found in the seepage zones were lower than expected when compared to the distribution of storm discharges and these sites were where storms were least peaked (cf. Figure 7.17). The maximum

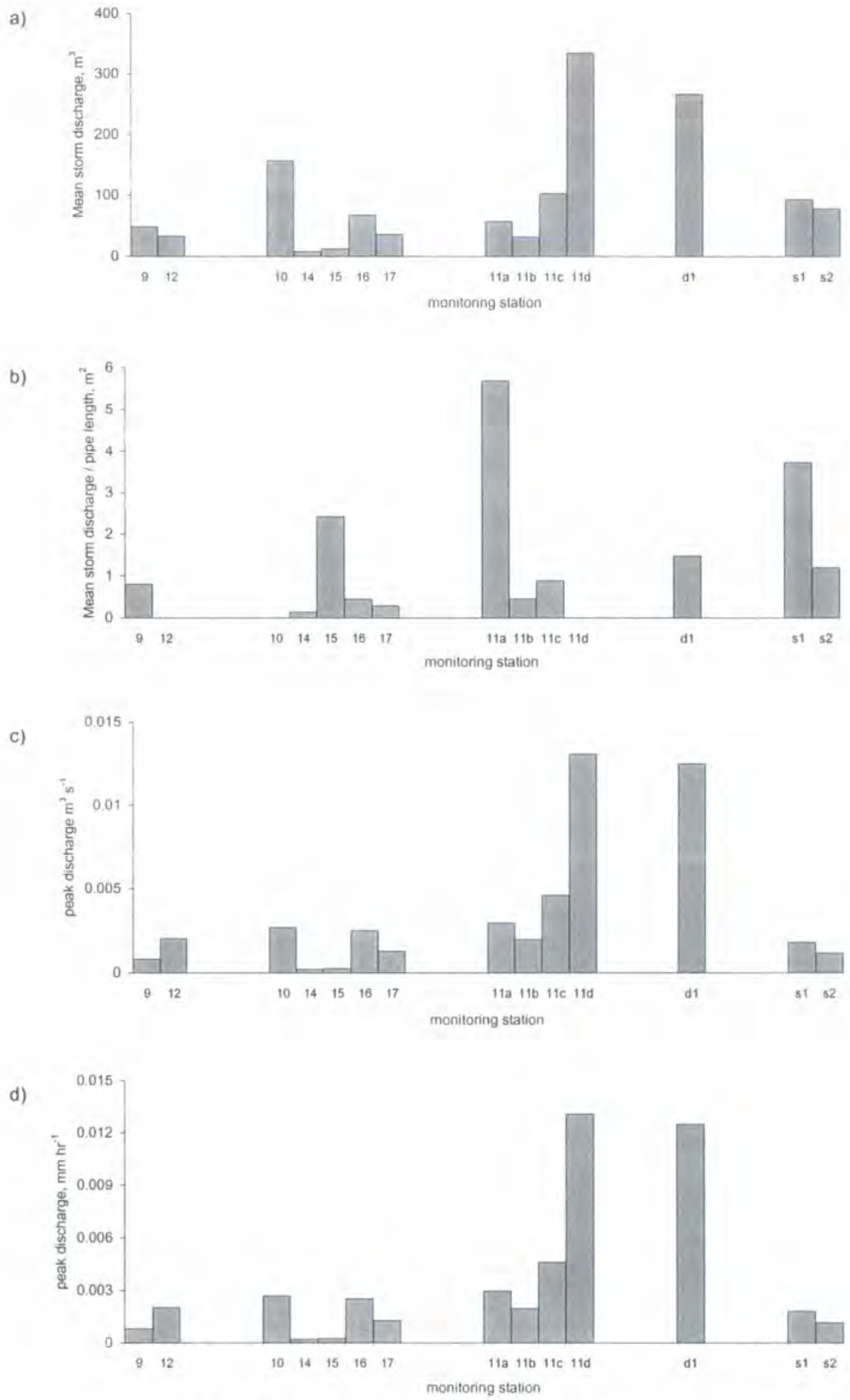


Figure 7.20. Storm discharge charactersitics at the LDPS monitoring stations, a) mean storm discharge, b) mean storm discharge divided by estimated pipe length (no data for pipes 10 and 12 as length undetermined), c) peak discharge recorded during study period, cumecs, d) area-weighted peak discharge during study period, mm hr^{-1}, as determined from calculation of surrogate basin area. Monitoring stations coded as given in Table 7.2 and Figure 7.1. d1 = ditch, s1, s2 = seepage zones, 11a-d = monitoring sites along pipe 11.

recorded pipe discharge was at 11c, with 4.6 l s^{-1} and 2.7 l s^{-1} at pipe 10 (Figure 7.20c). This is not as great as reported for Maesnant but greater than for most other reported pipeflows (Table 7.4). Peak discharge recorded from the grip was 12.5 l s^{-1}.

In terms of area-weighted peak discharges (mm hr^{-1}) calculated from the surrogate area technique outlined above, the situation is different (cf Figure 7.20c with 7.20d). Ephemeral pipe 12 has the greatest peak flows with the gully (11d) having the lowest. The flow at the head of pipe 11 recorded a higher area-weighted discharge peak than further down the pipe. Peak area-weighted flows from the perennial pipes 10, 14, 15, 16 and 17 ranged from 6.4 to 15.9 mm hr^{-1}.

Figure 7.21 plots peak runoff rates and lag times with catchment area calculated using Jones' (1997) 'surrogate basin area' technique. The pipeflow data from LDPS can be compared to the diagrams prepared by Jones (1997) which are based on Kirkby (1985) and Anderson and Burt (1990a) and the Maesnant pipeflow data. Uchida *et al.* (1999) fitted their pipeflow response to these diagrams and found that their monitored headwater ephemeral pipes in a forest Cambisol fitted into the Maesnant envelope. Of course, this type of diagram does not take into account typical rainfall intensities found in different environments nor the variety of soil parameters. Nevertheless they are useful indicators of typical responses.

The catchment areas of the piped sections estimated at LDPS are clearly smaller than found by Jones (1997). Whilst fewer storms have been analysed at LDPS than at Maesnant the maximum contributing areas feeding the pipes at LDPS are unlikely to be much greater than the current estimate. They are also very unlikely to be anything like as high as at Maesnant (Jones, 1987). At Maesnant much greater total discharges are measured issuing from the monitored pipe systems. So, as the catchment areas are smaller at LDPS and peak runoff rates are higher, this pushes the main envelope of the LDPS pipeflow dataset to the left of Jones' (1997) pipeflow envelopes (Figure 7.21a).

Importantly the peak flow response of LDPS and Trout Beck at the catchment level fit into the saturation-excess OLF envelope on the Anderson and Burt (1990a) diagram. This is interesting given the findings of previous chapters indicating the dominance of saturation-excess OLF in blanket peat catchments. Similarly, in terms of peak lag times Trout Beck fits into the saturation-excess OLF data envelope (Figure 7.21b). The effect

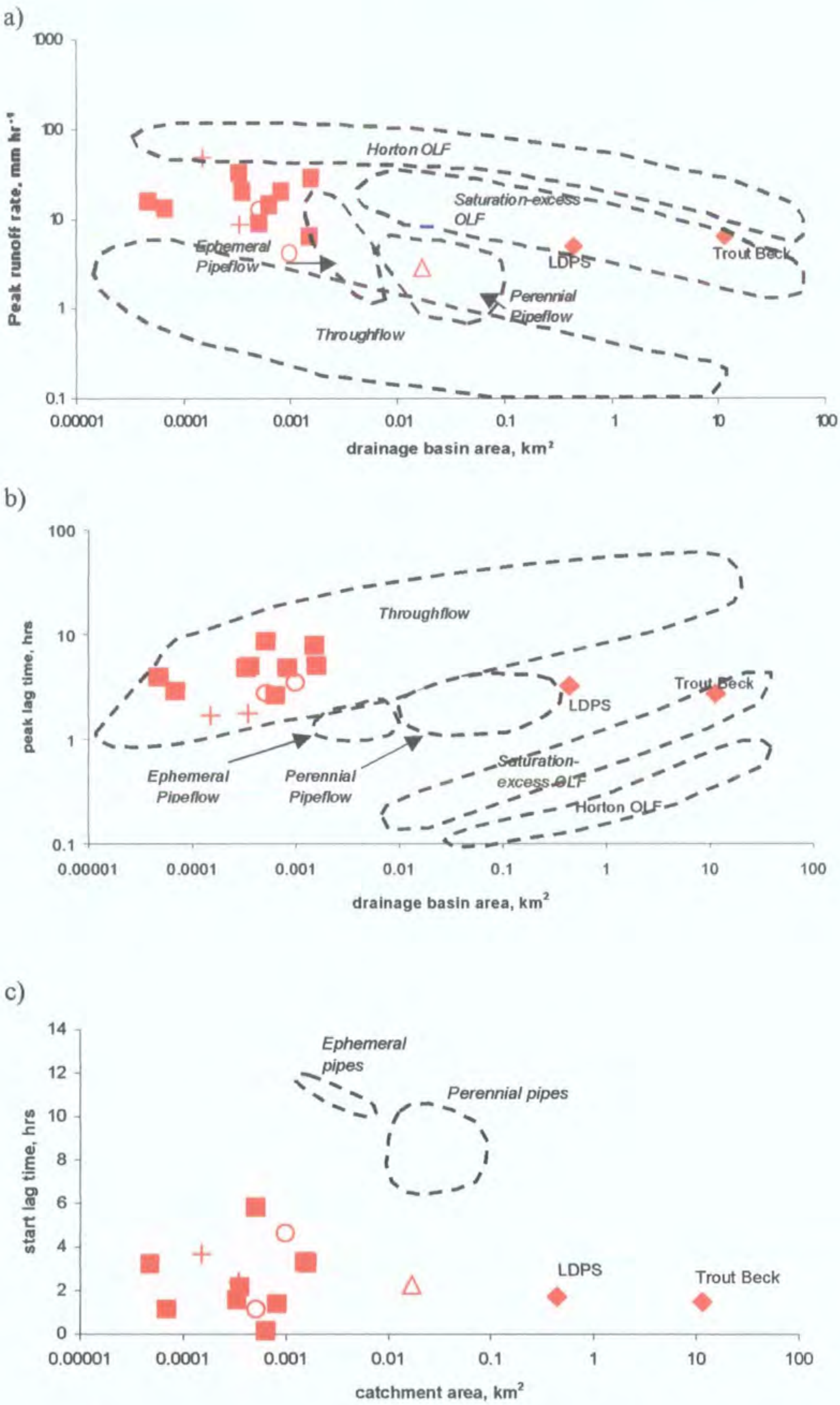


Figure 7.21. Peak runoff rates (a), peak lag times (b) and start lag times (c) for hillslope processes. A comparison of the LDPS data with that from Maesnant and the collations of Dunne (1978), Kirkby (1985) and Anderson and Burt (1990). Red squares = perennial pipes, red open circles = seepage zones, red crosses = ephemeral pipes, red triangle = gully 1.

of piping in the LDPS catchment may push the LDPS mean lag time response away from the saturation-excess OLF envelope and toward the perennial pipeflow envelope as found on Plynlimon. As the peak lag times for pipeflow are around the same in LDPS as found by Jones (1997) but the maximum contributing area smaller, this pushes the LDPS pipeflow responses away from that found by Jones (Figure 7.20b). The pipeflows at LDPS are thus within the throughflow envelope on the Anderson and Burt (1990a) diagram. Data are not available on start lag times for the hillslope drainage processes compiled by Dunne (1978). However, the pipeflow data are plotted in Figure 7.21c and compares LDPS with results from Maesnant (Jones, 1997). Given the shallow nature of the Maesnant and Upper Wye pipes on Plynlimon, which dominate the literature on storm pipeflow response, one may expect start lag times to be shorter than for the frequently deeper pipes found at LDPS. However, the much more rapid response of the LDPS pipes to rainfall than on Plynlimon means that the LDPS pipeflow response fails to fit the rough limits of the earlier data (Figure 7.21c). Given that catchment area is larger at Maesnant, one would expect to find longer lags there. Jones (1997) argues that the trend from the Maesnant and Upper Wye data which is dominated by the high lag times experienced in the ephemeral pipes would not be expected for OLF, in which smaller basins should have shorter lag times. Hence he suggests that this is one instance in which the trends within hillslope processes actually diverge. The inclusion of the data from LDPS to the diagram of Jones (1997) suggests otherwise. Furthermore, in the case of LDPS, the similarity of response of ephemeral and perennial pipes to rainfall events fails to allow separation of the two types in terms of lag times or peak flows.

7.7 Pipeflow contribution to streamflow

During the monitoring period of July to December 1999 the eight monitored pipes contributed 9.5 % of the total streamflow recorded. The two monitored seepage zones contributed 2.5 %, the ditch 1.9 % and the gully 5.1 %. Manual sampling of the other pipes which were not automatically monitored (apart from pipe 8 which was too difficult to assess) during high and low flows suggested that these pipes may contribute a further 2 - 4 % of total discharge.

Total monitored pipeflow contributions to runoff during the 30-day period which was examined earlier in the chapter are shown in Figure 7.22. It is clear that pipeflow is more important for smaller events such as on days 252, 255 and 266, whereas for larger events like that on days 250 and 263 it is probable that saturation of a greater extent of

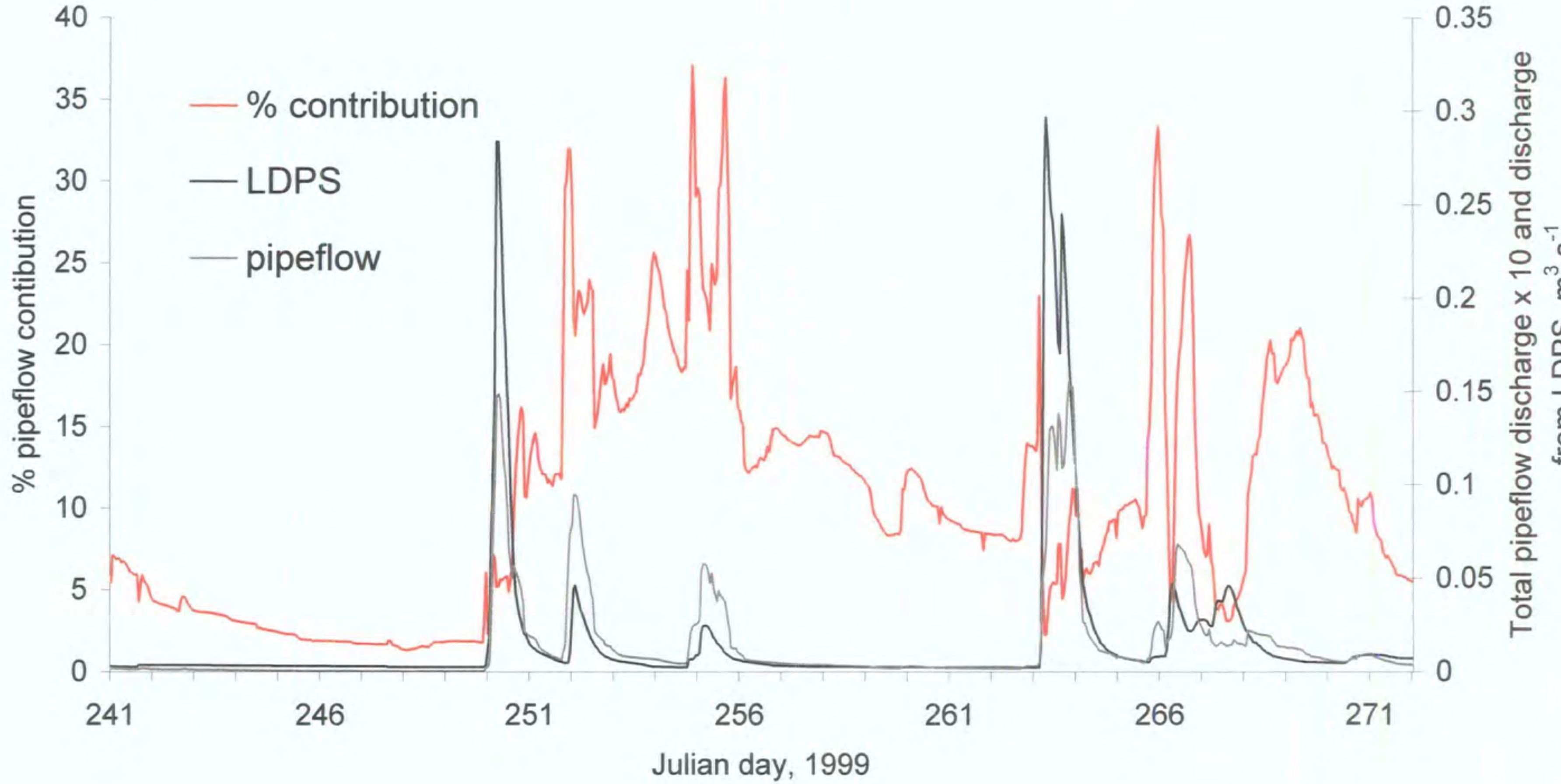


Figure 7.22. Pipeflow contribution to streamflow in the LDPS catchment during days 241-272, 1999.

the hillslopes means that OLF and near-surface drainage become more important relative to pipeflow. Peak contributions to streamflow from piping generally occur on the rising limb of stream hydrographs, with a minimum coincident with the streamflow peak. There is then a rise in the proportion of pipeflow contributing to runoff as stream flow recedes. Often there are two or three peaks in the proportion of pipeflow to the streamflow falling limb. This is probably related to the timings of individual pipe recessions relative to that of streamflow. It is clear from Figure 7.22 that during intermediate streamflow pipes contribute a larger proportion of runoff. The 8 monitored pipes can contribute at times over one third of streamflow. During both high and low flows however, pipeflow contributions can fall to 2 or 3 % such that over the 30 day period 9.2 % of streamflow moved through the piped system, 2.5 % through seepage zones 1 and 2, 1.9 % from the grip and 5.0 % moved through gully 1.

McCaig (1983) estimated pipeflow in Slitherough Clough, Yorkshire. He suggested that as runoff increased, the proportion of runoff from piped areas also increased. However, these results were based on estimations using a mixing model and McCaig (1983) did not actually measure the pipeflow. Jones (1978) and Jones and Crane (1982) presented evidence for the Maesnant to suggest that pipeflow contributions were of reduced significance under very wet antecedent conditions and in the heavier rainstorms. There was also some additional evidence for another fall-off in percentage contribution in drier antecedent conditions and in the lighter storms (Jones and Crane, 1984). The density distribution shown in Figure 7.23 shows how for LDPS both high and low flows are accompanied by reduced relative pipeflow contributions. The highest contributions are recorded during medium flows of around 0.07 - 0.10 mm hr^{-1}. The highest densities on the plot occur when streamflows are low and therefore pipeflow contributions are low such that most of the time pipeflow contributes less than 15 % to streamflow.

A further important point arises from the very particular nature of the hydrology of blanket peat which has been discerned in this and previous chapters. In most soil types (other than peat) where active piping occurs, pipeflow would be expected to increase the rate of runoff from a catchment. However, in blanket peat the low hydraulic conductivity of the matrix means that OLF and very near-surface flow dominates the catchment response. Given the lag times and pipeflow response in comparison to streamflow it therefore seems that soil piping in blanket peat does not increase the rate of runoff production in these catchments. It may in fact be that piping actually supplies more of the recessional and 'baseflow' components which would otherwise be almost non-existent. Yet at the same time the pipes respond rapidly to rainfall in that almost all of them rise within two or three hours of rainfall onset. The difference between peat piping and other soil piping is that although pipeflow is flashy in most cases, in comparison to other dominant flow processes occurring within the catchments, pipeflow is no more flashy than the other sources in peat. In other soil types, however, pipeflow is comparatively more flashy and will therefore contribute higher proportions of discharges on the rising limbs and peaks of the streamflow hydrograph. In catchments where OLF is absent, pipeflow will tend to provide a runoff peak before throughflow, as noted in the East Twin catchment (Weyman, 1970).

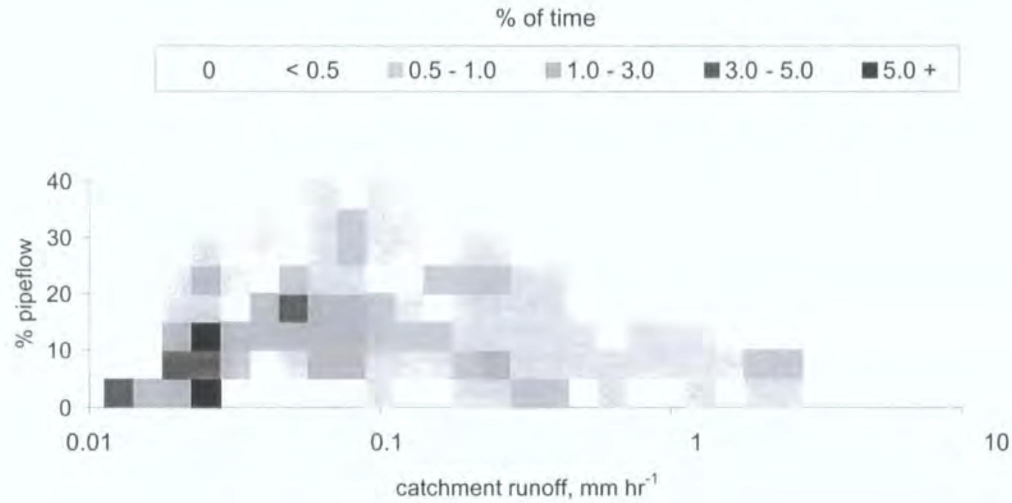


Figure 7.23. Density distribution of the proportion of time pipeflow contributes a given percentage to streamflow in the LDPS catchment, July to December 1999. A darker cell indicates that there are a greater number of occasions when pipes contribute a given percentage to catchment runoff than for a lighter cell. For example, when catchment runoff is 0.1 mm hr^{-1}, pipes contribute between 5 and 10 % of the streamflow volume during less than 0.5 % of the total monitoring time.

Storm hydrographs and ratio of discharge at each monitoring station to for the storm on day 250, 1999, is given in Figure 7.24. The diversity of response is clear. Three types of response seem evident; rapid rise and fall of hydrograph with almost symmetrical form such as seen down the length of pipe 11 (a to c), from the grip (d1), and from pipes 16 and 17; these are generally long pipes. The shorter pipes such as the ephemerals 9 and 12 and perennial 10 produce squarer hydrograph responses with flattened peaks and are slightly broader in appearance. The third type of response to this storm event comes from the much broader hydrograph form of S1, S2 and pipe 15 (which is probably linked to S1). In terms of the contributions to stream stormflow there are two main

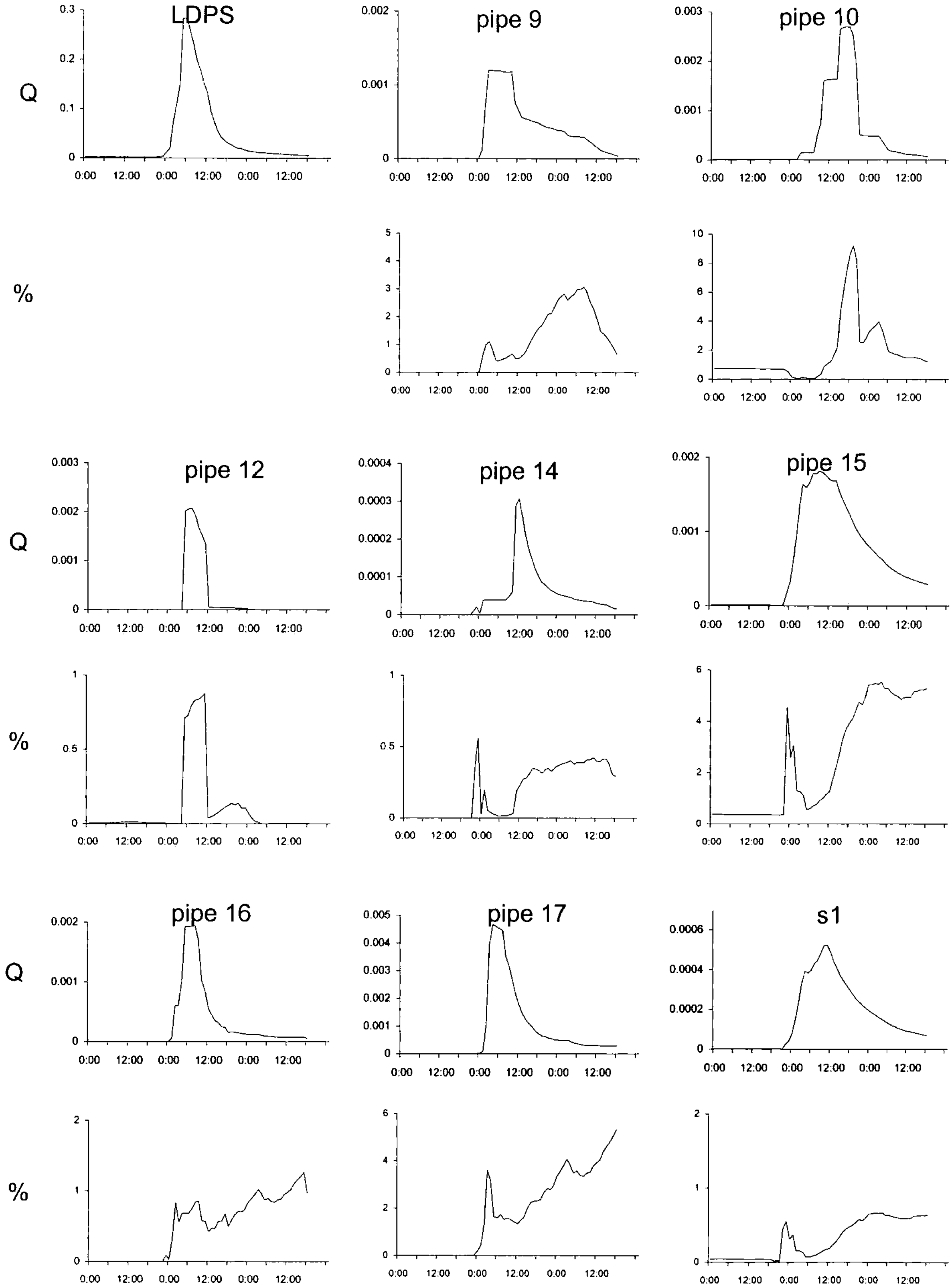


Figure 7.24. Discharge, cumecs, and percent of streamflow discharge for storm on 6-8th September, 1999, for the LDPS gauging stations.

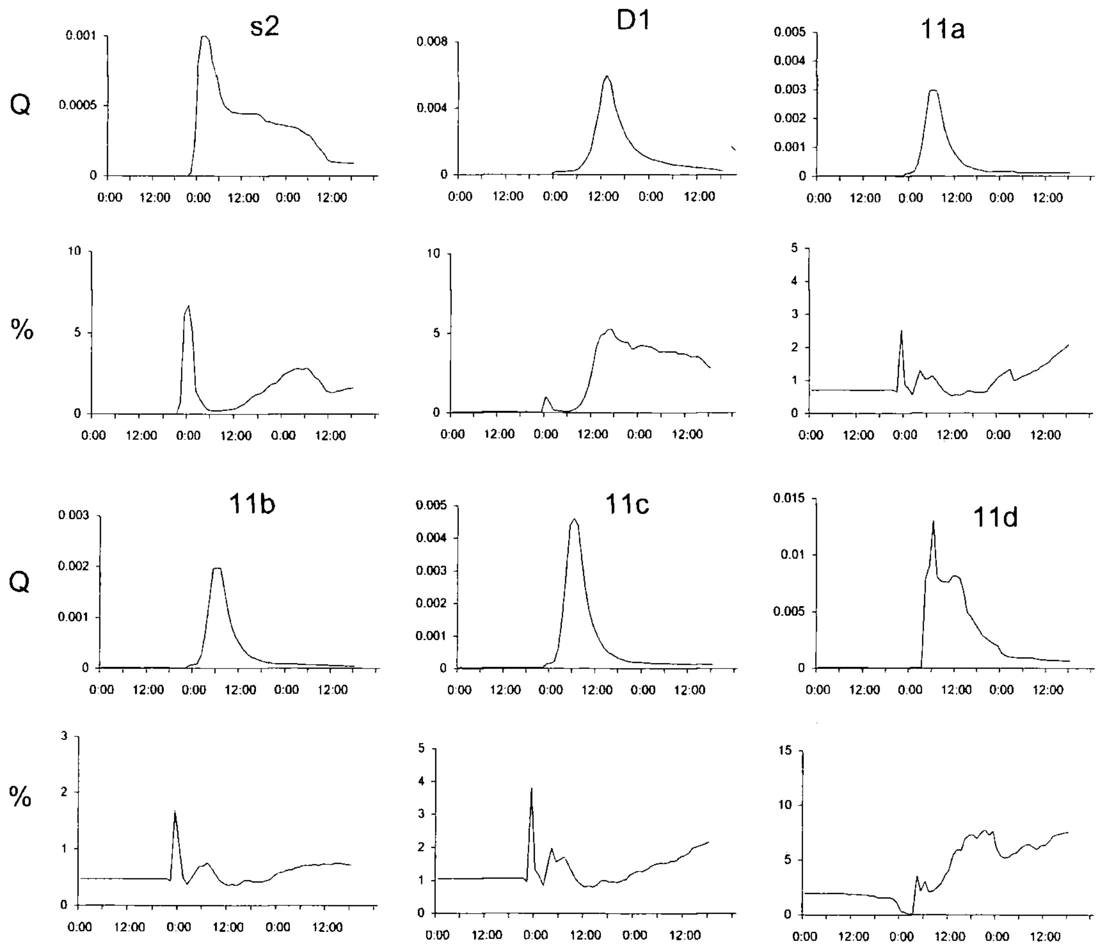


Figure 7.24. continued

patterns. The first is where there is a sharp peak in contribution on the streamflow rising limb due to the rapid response to rainfall of many of the pipes. This sharp peak is followed by a fall in contribution followed by a more extended broad rise and fall in contribution along the streamflow falling limb. The secondary peak is not as high as the initial peak but the rise in contribution is more long lasting. Pipes 11 (a to c) 14 and S2 can be included in this category. The second pattern under which the remaining sites can be included (except for pipes 10 and 12) is that where there is an initial sharp peak in the proportion of pipeflow on the stream rising limb but then the secondary rise on the stream falling limb is greater than this initial peak. A good example of this is the ratio of flow from pipe 9 (Figure 7.24).

7.8. Summary

The monitoring work performed in the LDPS catchment has allowed simultaneous and continuous flow records from a wide range of pipes within a raw blanket peat catchment. Whilst the record is only 5 months long and only 14 storms were analysed, 15 gauging sites were continuously monitored during the study period including 8 separate pipes. This is easily the most extensive continuous record of soil pipeflow outside of the Maesnant on Plynlimon. The pipeflow response from LDPS was found to be different from that on Plynlimon. This is important given the wide citation of the Plynlimon work. Both perennial and ephemeral pipes were found in the LDPS catchment throughout the soil profile. Importantly, the distinction between the two pipe types is often not clear and may not be useful within these upland blanket peat catchments. Pipe outlet depth had little relationship with the flow regime of the pipe in LDPS, although pipe outlet shape appeared to be affected by proximity to the peat-substrate interface. Pipe outlet dimensions are the most common reported feature of soil pipes in the literature. As will be shown in Chapter 8 pipe outlet characteristics are misleading because the pipe shape, size and depth may be very different a short distance upslope.

Weyman (1975) distinguished between streambank and hillslope piping. He claimed that the small pipes underlying extensive areas of hillslope in the Mendips seem to be connected to the surface by open roots and responded rapidly to rainfall. The other pipes seen in streambanks represent the concentration of streamflow from the lower part of the slope and were fed directly from the soil matrix. The LDPS data show that this is not always the case as many pipes issuing into the streambanks can react quickly to rainfall and produce large volumes of discharge. These pipes can also extend up the hillslope for a considerable distance; some clearly fed in part by surface inlets. Direct capture of OLF through pipe inlets may be a major source of storm runoff resulting in 'start lag' times of 2 hours or less. Cryer's (1980) water quality analysis of piping at Maesnant led him to agree with Jones (1978) that both soil cracks and seepage supplied the pipes with water.

Calculation of 'surrogate basin area' allowed the plotting of the Moor House data onto the generalised graphs of Anderson and Burt (1990a) and Jones (1997). This allowed a simple comparison to be made (Figure 7.21). The plots suggest that at the catchment-scale, saturation-excess OLF is the dominant runoff-generating mechanism in blanket peat catchments. This agrees with the work presented in Chapters 4-6. The pipes within the LDPS catchment behave differently from those on Plynlimon (Jones, 1997). Whilst 'start lag' times for pipeflow in the LDPS catchment are shorter than on Plynlimon 'peak lag' times are approximately the same. Peak runoff rate, peak lag time and start lag time data from LDPS all plot outside the Plynlimon data envelopes.

An important aspect of pipe hydrology in these blanket peat catchments is that medium flows are sustained for a longer period of time than would otherwise be the case. Unlike the effect of soil piping in most soils (where other subsurface flow processes would dominate) which would be to increase the speed of runoff production within a catchment, soil pipes in blanket peat catchments appear to provide a greater proportion of flow to the falling limb of the stream hydrograph. Pipeflow in LDPS, despite accounting for only around 10 % of streamflow in total can nevertheless be a very important contributor to flow, particularly on the rising or falling limb of the stream hydrograph when pipeflow contributions can be in excess of 30 %. Thus, although OLF and near-surface flow processes are more important than pipeflow within LDPS, the dominance of the various processes changes through time and space during a storm event. Hence, in line with Jones (1979), the source areas for runoff within the LDPS catchment may be more dynamic than the classical variable source area model of Hewlett (1961) may suggest.

The sediment and solute loading of the catchment response may be affected by such dynamism. Pipe waters in LDPS were found to contain at least three times as much sediment as streamwater during most sampling runs for example. It was noted that one of the pipes ceased flowing two days after weir installation and it was probable that sediment blocked this pipe. Furthermore the stilling pool (of volume around 0.3 m^3) at pipe 17 filled entirely with sediment during an active period from December 1999 to March 2000. It may be that these pipes play a much more important role in sediment and solute budgets in the uplands than work has hitherto suggested.

It is likely that the pipe network contributing to streamflow in LDPS is larger than indicated by surface mapping. Bryan and Jones (1997) note that new techniques are urgently needed for surveying the piped networks and measuring subsurface catchments. These may help identification of source areas and mechanisms of pipe supply. A pilot study was carried out to assess the suitability of ground penetrating radar to locate subsurface pipes in blanket peat catchments and the results are discussed in the following chapter.

CHAPTER 8

THE APPLICATION OF GROUND PENETRATING RADAR TO THE INDENTIFICTION OF SUBSURFACE PIPING IN BLANKET PEAT

8.1 Introduction to Ground Penetrating Radar (GPR)

Traditional point-measurement techniques, such as soil coring or pit excavation are destructive and provide an incomplete characterisation of the subsurface; GPR, originally developed for military applications, provides an alternative. GPR has been used in fields as diverse as architecture, engineering, environmental management and mineral prospecting (Mellet, 1995; Reynolds, 1997). GPR is frequently used to study contaminants in groundwater (Benson, 1995; Daniels *et al.*, 1995), the nature of subsurface faulting (Benson, 1995) and the location and size of plastic, metal pipes (e.g. Peters *et al.*, 1994) and other objects, particularly in archaeology (Conyers and Goodman, 1997). GPR has been used successfully to map peat deposits (e.g. Warner *et al.*, 1990; Hanninen, 1992; Theimer *et al.*, 1994; Lapen *et al.*, 1996), soil and rock stratigraphy (e.g. Olson and Doolittle, 1985; Davis and Annan, 1989; Dominic *et al.*, 1995), bedrock topography (e.g. Olson and Doolittle, 1985) and the water table (e.g. Lapen *et al.*, 1996). It has been used to construct continuous bottom profiles through peatlands (Bjelm, 1980; Hanninen, 1992) and is capable of recording peat depth and differentiating internal irregularities due to peat composition, water content, and bulk density (Warner *et al.*, 1990). GPR has been shown to produce better near-surface resolution in the upper few metres of soil and bedrock than seismic refraction (Olson and Doolittle, 1985). With GPR, where there is a sharp variation in water content, there will be a strong reflection. Hence cavities and soil pipes may be detectable within the blanket peat. To the author's knowledge GPR has never been used before to identify soil piping in peat. This chapter will present results from an exploratory pilot investigation done in order to assess the application of GPR in identifying subsurface pipes within blanket peat.

8.2 Basic Principles of the GPR

Much of the basic physics of GPR is described in detail by Davis and Annan (1989). Short pulses of high frequency (10 -1000 MHz) electromagnetic energy are transmitted by an antenna through the ground surface and reflected from boundaries between zones or from internal irregularities which have differences in electrical properties. The reflection is detected on the surface, and the time between transmission and detection is proportional to depth. Moving the transmitter and receiver antennae across the test area builds up a complete cross section of the site. The depth of penetration depends on the ground conditions at each site. Increased depth can be obtained with lower frequency radio waves, but this reduces the resolution of the radar reflections. Therefore the frequency used is a compromise. In the case of detecting relatively small features such as soil pipes a sufficiently high frequency must be selected so that the radar wavelength is short, allowing detection and suitable resolution.

Although the length of time for the electromagnetic waves to reflect to the receiver of the GPR is proportional to depth, the velocity of the wave through the medium must be calculated in order to determine the depth. The dielectric constant (γ_r) (otherwise known as the relative permittivity), is a direct measure of the velocity (V) of an electromagnetic wave through a material. This and electrical conductivity (EC) govern radar propagation velocities through a medium. When EC is small it can be ignored as γ_r overwhelmingly controls V. Many soils, because of their high EC, are essentially 'radar opaque'. Gain factors are generally applied to take into account the attenuation arising within the soil and from simple geometric spreading of the GPR signal. Soil moisture tends to increase radar attenuation. Indeed this fact has been used in order to allow detection of soil water content. For example, Chanzy *et al.* (1996) demonstrate a strong correlation between GPR data and soil water content with an error of only $0.03 m^3 m^{-3}$. Hubbard *et al.* (1997) use GPR data to estimate γ_r which is then used to estimate intrinsic permeability and saturation values.

The most critical parts of the GPR system are the antennae (Peters *et al.*, 1994). The most common version is a simple pair of parallel antennae, one to transmit and one to receive. The radar system causes the transmitter antenna to generate a wavetrain of radiowaves which propagates away in a broad beam. As radio waves travel at high speeds (in air 300 000 km s^{-1}) the travel time of a radiowave from instant of transmission through to its subsequent return to the receiving antenna is of the order of a few tens to several thousand nanoseconds (10^{-9} seconds). This requires very accurate instrumentation to measure the transmission instant precisely enough to allow accurate time and therefore accurate depth calculations.

8.3 Application of GPR to peatlands

As early as 1980 Ulriksen found that GPR could provide peat thickness data in Swedish bogs to a good degree of precision (Hanninen, 1992). Much of the work on peatlands utilising GPR technology has been performed by the Geological Survey of Finland who in 1983 noted a desire to improve the rapidity and accuracy of fieldwork which was heavily based on traditional drilling methods (Hanninen, 1992). 2000 km of GPR measurements were carried out in 104 peatlands across Finland resulting in detailed datasets on peat thickness, stratigraphy and underlying topography. Interestingly, Bjelm (1980) noted that knowledge of the base topography and hydrogeological condition of Swedish peatlands through use of GPR would be of use not only for utilisation of the peatlands for energy but for peatland restoration. Hanninen (1992) notes that the depth data obtained from GPR in peatlands are markedly more accurate and detailed than those obtained by traditional means. Predictions of mineral basement depth have been reported to an accuracy of 10 cm, which is at least comparable to the likely accuracy of reference coring (Theimer *et al.*, 1994). Warner *et al.* (1990) applied GPR to the mapping of peat stratigraphy and thickness in a large bog in south-western Ontario. Their survey was undertaken in conjunction with a conventional coring survey and measurement of peat physical properties. The results indicated that GPR responds to peat moisture content and bulk density, which vary with stratigraphic changes. In particular, the acrotelm-catotelm boundary and the basal clay were GPR reflectors. Lapen *et al.* (1996) used GPR to delineate subsurface features along a wetland catena in south-eastern Newfoundland. Signals were less attenuated when soils were dry but due to the high water content of the peat average propagation velocities were found to be relatively slow (0.04 m ns^{-1}). Nevertheless, the total peat depth could clearly be assessed on a continuous survey and the bedrock was a prominent reflector.

The low EC of the soil pore water in peatlands results in non-dispersive signal propagation and allows velocity profiles of the organic and mineral soil to be estimated (Theimer *et al.*, 1994). This fact means that although the signals are attenuated by the high moisture content in peat they are not attenuated by the EC of the pore water. The depth of penetration required in the North Pennines is generally lower than 3 m whilst 8-10 m depths have successfully been investigated in peatlands (e.g. Chernetsov, *et al.*, 1988, Theimer *et al.* 1994). However, care is needed; Warner *et al.* (1990) undertook a survey when the bog was frozen to allow ready access and a solid working surface for the portable GPR instruments, but unfortunately the frozen layer acted to overwhelm other received signals and attempts to remove these signals were difficult. Hanninen (1992) noted similar problems with winter peatland measurements in Finland.

The frequencies used in peatland GPR surveys have ranged from 50 MHz to 600 MHz, but surveys using frequencies above 300 MHz have been less successful. Nevertheless, a 500 MHz antenna was found to lend itself well to examination of structural layers in peat by the Geological Survey of Finland (Hanninen, 1992) particularly to the surface layer which could not be well examined using 80, 100, 120 or 300 MHz. However, the 500 MHz antenna had limited probing depths. Therefore for the present study, and due to limited time available to use the equipment, it was decided to use the more typically successful 100 and 200 MHz antenna.

8.4 Field equipment and methodology used at Moor House

The GPR used was a Ramac GPR from Mala Geoscience; Figure 8.1 is a schematic diagram of the system when connected. The equipment was loaned from the Geology Faculty at the University of Barcelona and used in conjunction with Dr. Maria Vilas who provided training in the interpretation of the radargrams. In its simplest form the system consists of a computer, control unit, transmitter and receiver. The control unit is connected to the transmitter and receiver with optical fibres and to the computer with a communications cable. The control unit organises procedures and controls the transmitter and receiver. It also keeps track of current position and time. Each component in the system, except for the computer, is powered by a specially designed battery pack. These packs give up to 8 hours field usage, although it is best to have spare in case one of the batteries is weak. Given the poor life of many internal laptop battery cells (often less than one hour), it is usual to have an external long-life battery source for the laptop computer too. This can be strapped around the user's shoulder in the field. The contact points on the system, particularly to battery packs, are susceptible to fail during even the lightest of rain such that it is advisable to cover the points with a small amount of plastic sheeting to keep the sensitive areas dry (see Figure 8.2 for example).

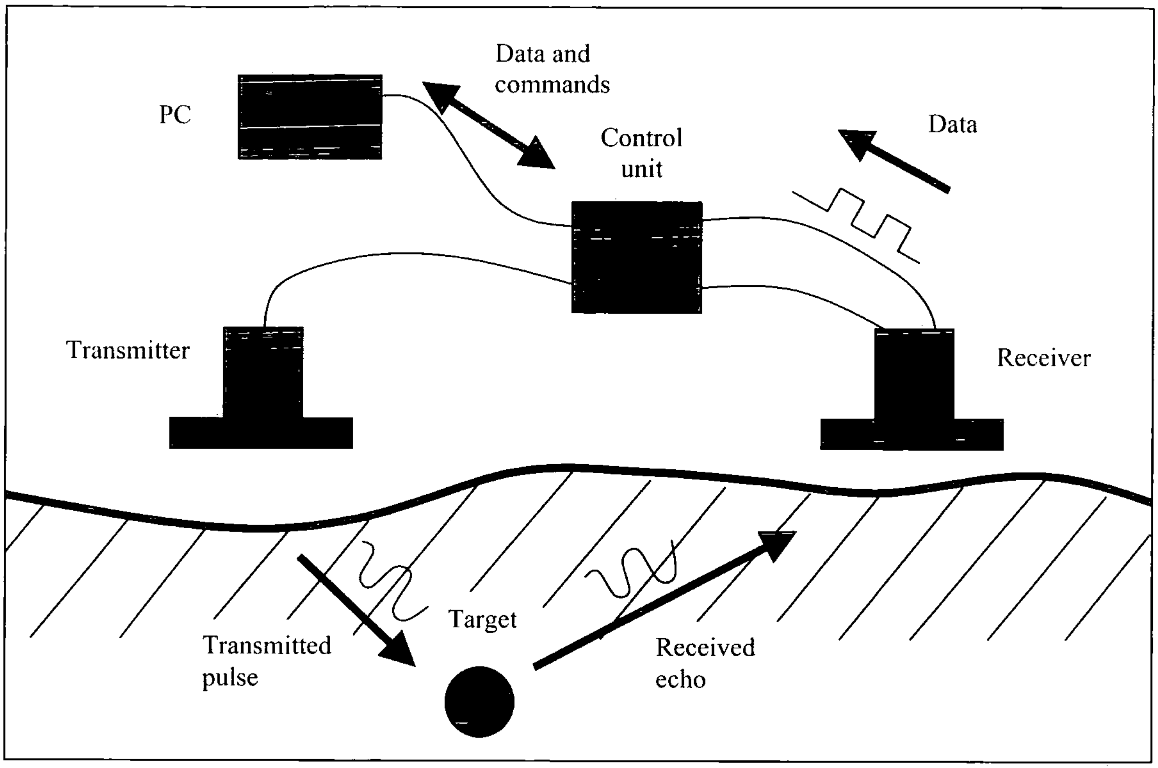


Figure 8.1. Schematic diagram of the GPR equipment

When collecting a sample[1], the control unit sends a control signal to the transmitter and receiver separately. After the transmitter has received the signal, it generates a pulse through the antenna element. The pulse reflects on any objects or structures beneath the ground surface and is echoed back to the receiver. Once the receiver has detected the control signal, it collects a sample and passes it to the control unit. By repeating this process at very finely controlled intervals, the control unit can collect all the samples in a trace[2]. The control unit places each incoming sample in its correct position in the current trace. When the trace is complete, it is sent to the computer where it is saved on the hard disk and displayed on the screen. There are two opto-connectors on the unit, one of which is used to transfer the control signals from the control unit and the other to send the collected data to the control unit. There is also a connector for a battery pack. The transmitter unit generates electromagnetic energy and transmits it to the surrounding area, especially into the medium that is being studied. The energy is in the form of a pulse at high amplitude that is fed to the antenna element. A pulse is transmitted every time a control signal is received through the optical fibre. The unit has one opto-connector for the control unit and a connector for a battery pack.

During data collection the whole system is transported along the transect to be scanned (Figure 8.2), and collects data at given points (or time intervals) as determined by the user. The system is fairly lightweight and portable, with the control unit carried on the fieldworker's back, the laptop computer being a standard item and the GPR antennae easily lifted and moved to the next sample point along a survey transect. GPRs can be dragged across the terrain on a trolley or even moved using a powerful vehicle with a winch (e.g. Welsby, 1988). For difficult ground such as peat with its soft nature, eroded steep-sided hags and awkward vegetation it was better simply to carry the antennae to each sample point. It is possible for one person to take measurements with the Ramac but with the control unit braced on the field worker's back and the need to carry the antenna to each sample point, it is better to have another person present. The second person can operate the laptop computer which is used to control when a sample is taken (simply by pressing a key on the keyboard), enter profile parameters and store data. The best set up was found to be that where one person carries the computer and control unit

[1] In a completely digital system, the incoming signal is measured a certain number of occasions per unit of time. The result of every such measurement is a numeral, a sample. [2]At each point of measurement along the profile, a specific number of samples are collected. Together, these samples make up a trace.

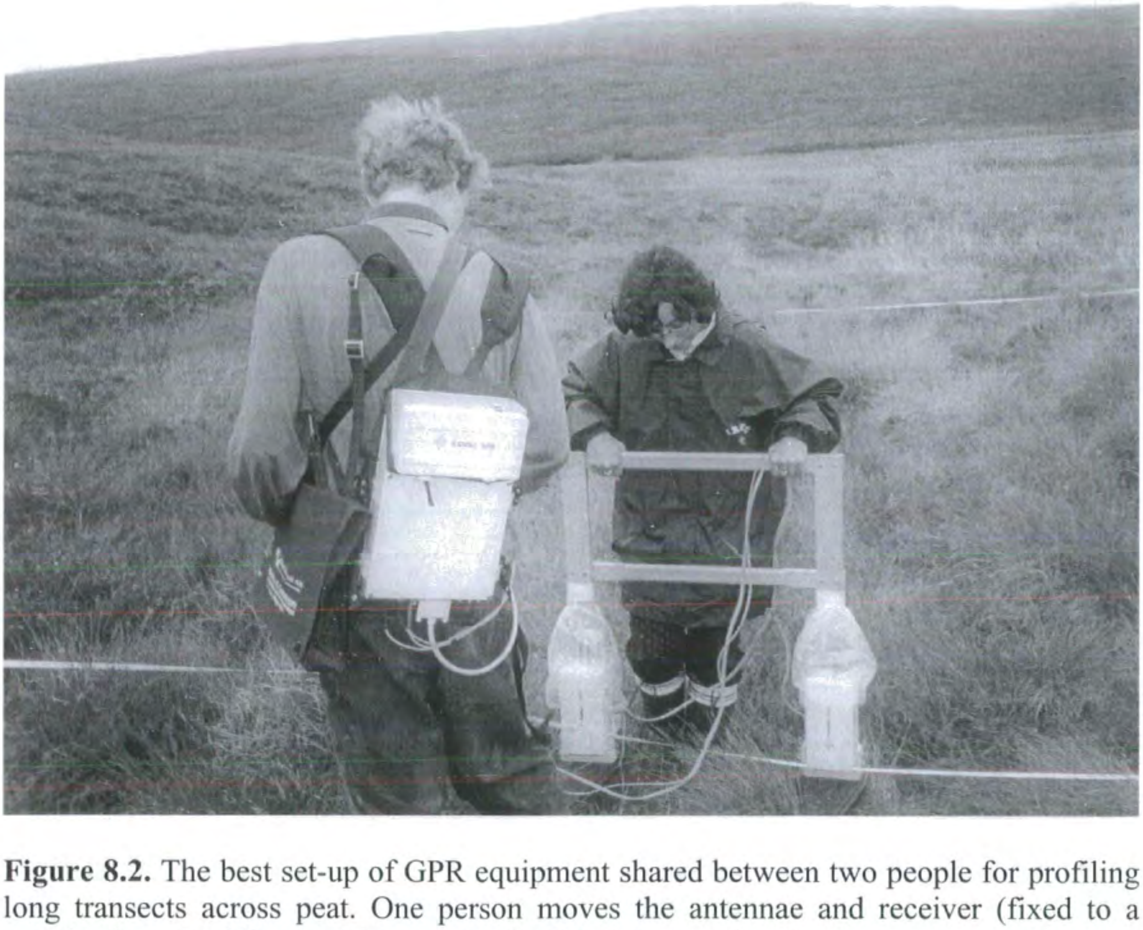


Figure 8.2. The best set-up of GPR equipment shared between two people for profiling long transects across peat. One person moves the antennae and receiver (fixed to a wooden brace) whilst the other carries the PC and control unit.

whilst the other carefully places the antennae at the correct location on the transect (seen in Figure 8.2 for example).

A 50 m tape was placed taut across the surface of the transect and used as a guide to the nearest two centimetres as to where to place the antennae (which were spaced at a 0.5 m interval using a wooden brace – see Figure 8.2). Locations were later accurately surveyed (see below). Traces were taken every 10 cm along shorter transects and every 50 cm or 1 m on longer transects (over 20 m). It is better if the scan is conducted so that traces are collected equidistantly as this simplifies location of subsurface objects.

Ramac software was used on the laptop computer for data collection, and field display of results from the operating transect. Post-processing was done using the GRADIX software from INTERPEX Inc. which filters the data to allow particular features to be identified more easily. It also allows data to be presented in a variety of ways. A colour display of the reflection data from the profiles was found to provide the clearest results and these have been used in data analysis. Frequent measurements of peat depth done by reference coring allowed checking of depth conversions. Ground-surface truthing was performed using an EDM for short transects or a differential GPS for longer transects (Higgitt and Warburton, 1999). Although not very accurate for altitudinal measurements (+/- 20 m) the Magellan GPS was found to provide fairly accurate relative height data during one continuous session (+/- 5 cm) when compared to EDM surveys of the same transect. However, terrain surfaces are notoriously uneven in peat. The quality of the GPS return signal will therefore depend on the nature of the contact with the peat surface under foot and the nature of the movement of the individual holding the mobile GPS (e.g. what phase of footstep they are in when a signal is returned).

8.5 The identification of subsurface features in blanket peat

8.5.1 Subsurface topography

In the process of collecting information on soil piping using the GPR, data are also collected on the nature of subsurface topography. A good example of the ability of the GPR to help quickly and easily generate digital elevation data on peat depth comes from a survey of the 100 m^2 ECN target sampling site at Moor House. Here five transects were taken running north-south spaced at 25 m intervals with similar transects running east-west. GPR traces were collected at 50 cm intervals. A north-south profile is shown in Figure 8.3a. Here an area of deeper peat can be identified around 60 – 90 m along the profile. Reflections can also be identified from subsurface layers of peat. More stratigraphic reflections are seen in the deeper section of the peat. This suggests that peat growth may have started in the hollow before spreading over the rest of the hillslope, the wetter location allowing peat to build up locally in advance of more widespread peat formation. Whilst not the focus of the present chapter it is worth noting that with careful coring it may be possible to correlate specific peat layers such as those which are *Sphagnum*-rich or rich in birch wood remains with the reflections indicated in the GPR profiles. This may then allow the development of improved models of hillslope blanket peat development.

Figure 8.3b shows that surface topography generally slopes in the direction of the substrate topography on the ECN target site. However, the surface topography is usually much gentler in slope than that of the substrate, the topography having been dampened by the greater build up of blanket peat in the hollows with their faster peat accumulation rates. A contour plot of surface slope and peat depth across the ECN target site is shown in Figure 8.3c. This demonstrates how easy it is to map subsurface topography and to identify hollows beneath the peat mass using the GPR technique. Depth probing using a soil auger or rod would have been very time consuming across this ECN plot and would have required 2000 separate measurements to achieve the same spatial density of measurements as the GPR. This would have taken several hours, maybe even days. Whilst for the GPR the same number of measurements were taken, it only took one hour to perform the survey.

It would be meaningless, of course, to plot how accurately the GPR measures peat depth. This is because 'real' peat depths, as determined by soil augering, were used to convert the time taken for waves to reflect from the substrate and return to the receiver antenna to depth measurements. Nevertheless, it is possible to test the application of the GPR to detect the depths of objects within the peat, such as pipes, simply by measuring the depths of several pipes by hand and comparing with the GPR estimates (see below).

8.5.2 Detection of soil pipes

The ability of the GPR to detect soil piping in blanket peat can be seen in Figure 8.4a. Here a short transect was tested across a known pipe location in the LDPS catchment. The pipe was just upslope of gauging site 11a (see Chapter 7) and the pipe is clearly

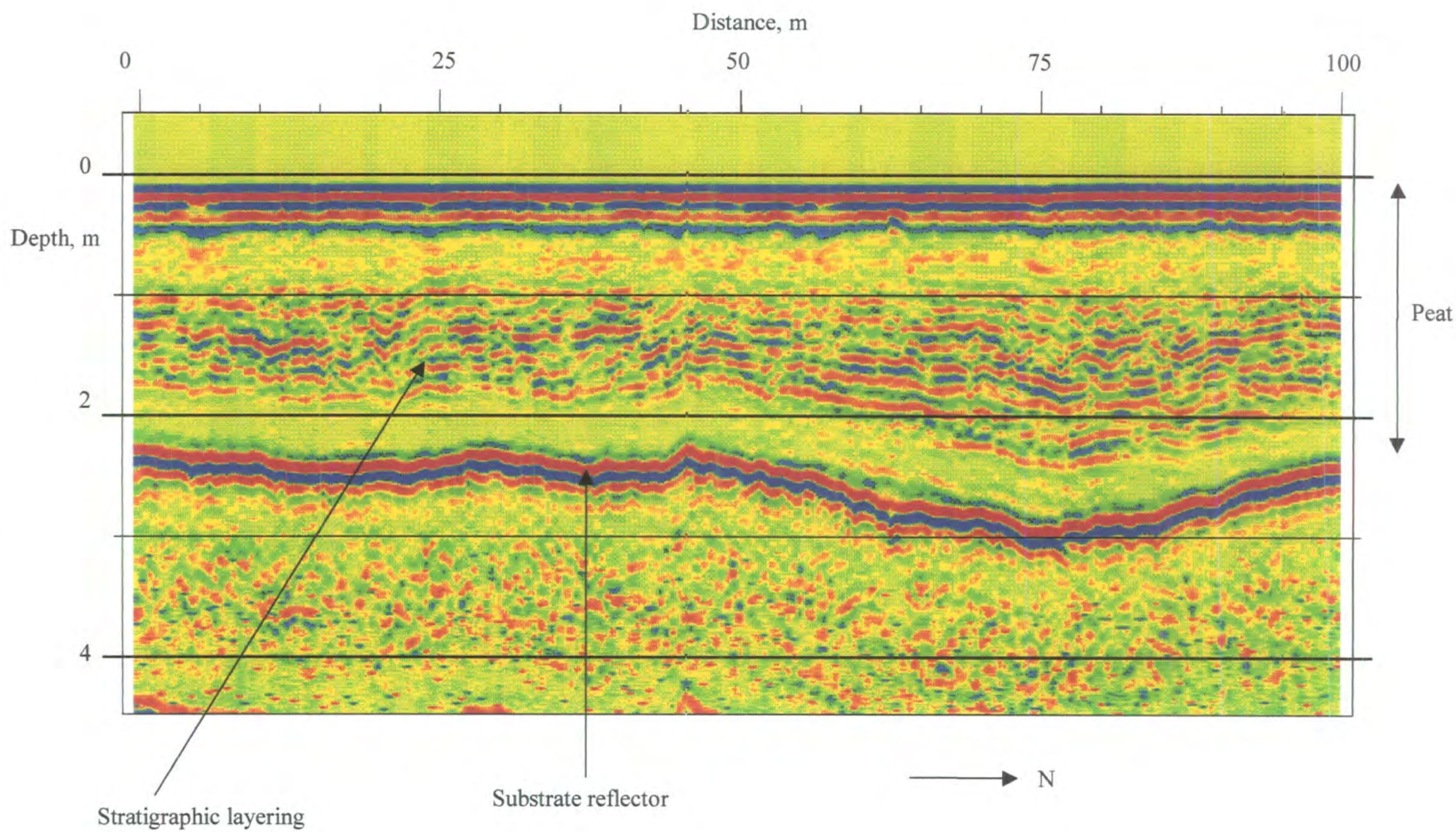


Figure 8.3a. Radargram across a transect on the ECN target site. Traces were taken every 50 cm. The subsurface topography is a clear reflector. There are many other thin stratigraphic layers that are acting as reflectors within the peat mass itself.

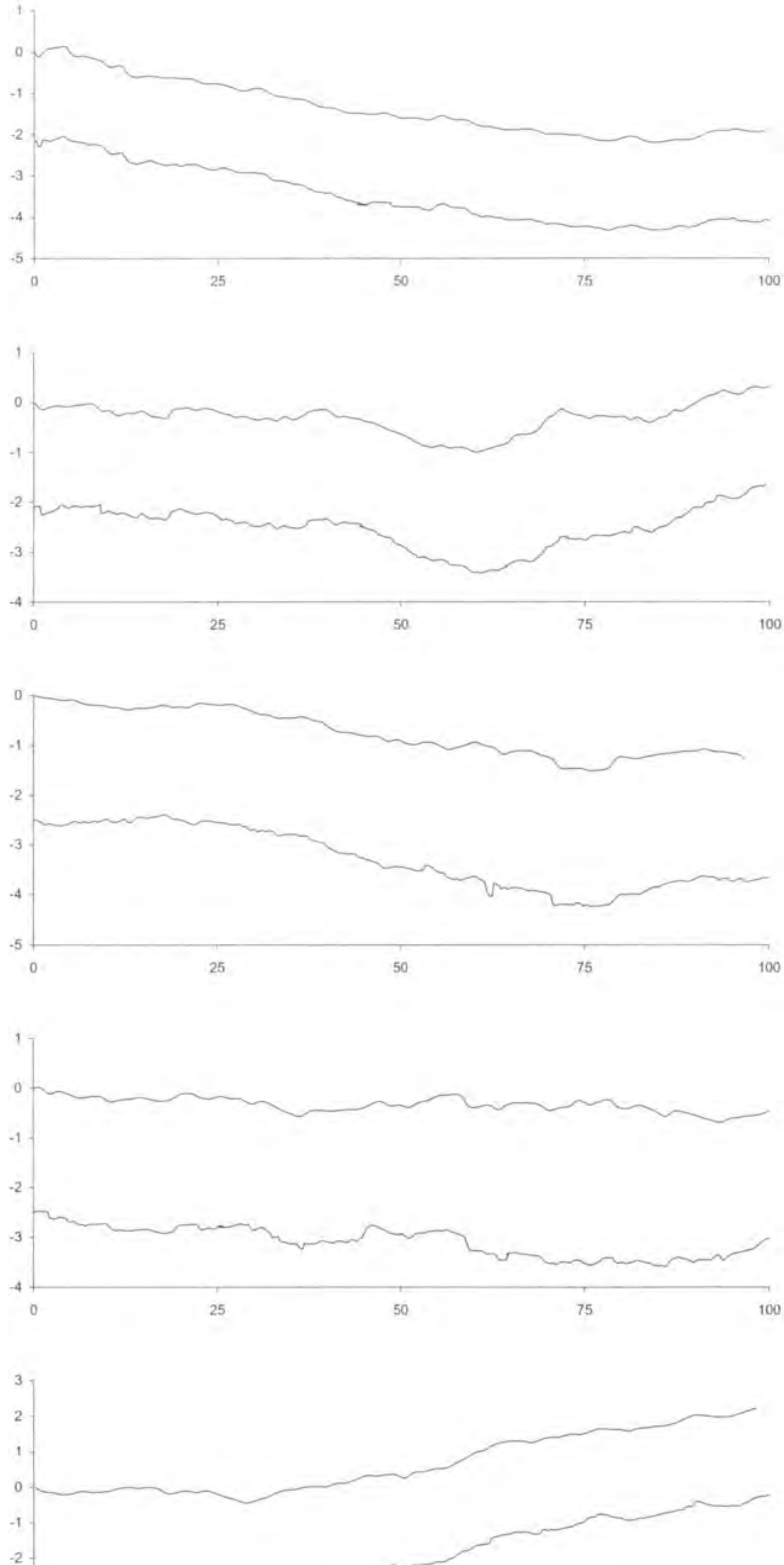


Figure 8.3b. North-south transects across the ECN plot showing surface slope and substrate topography. Axis depths and distances in metres.

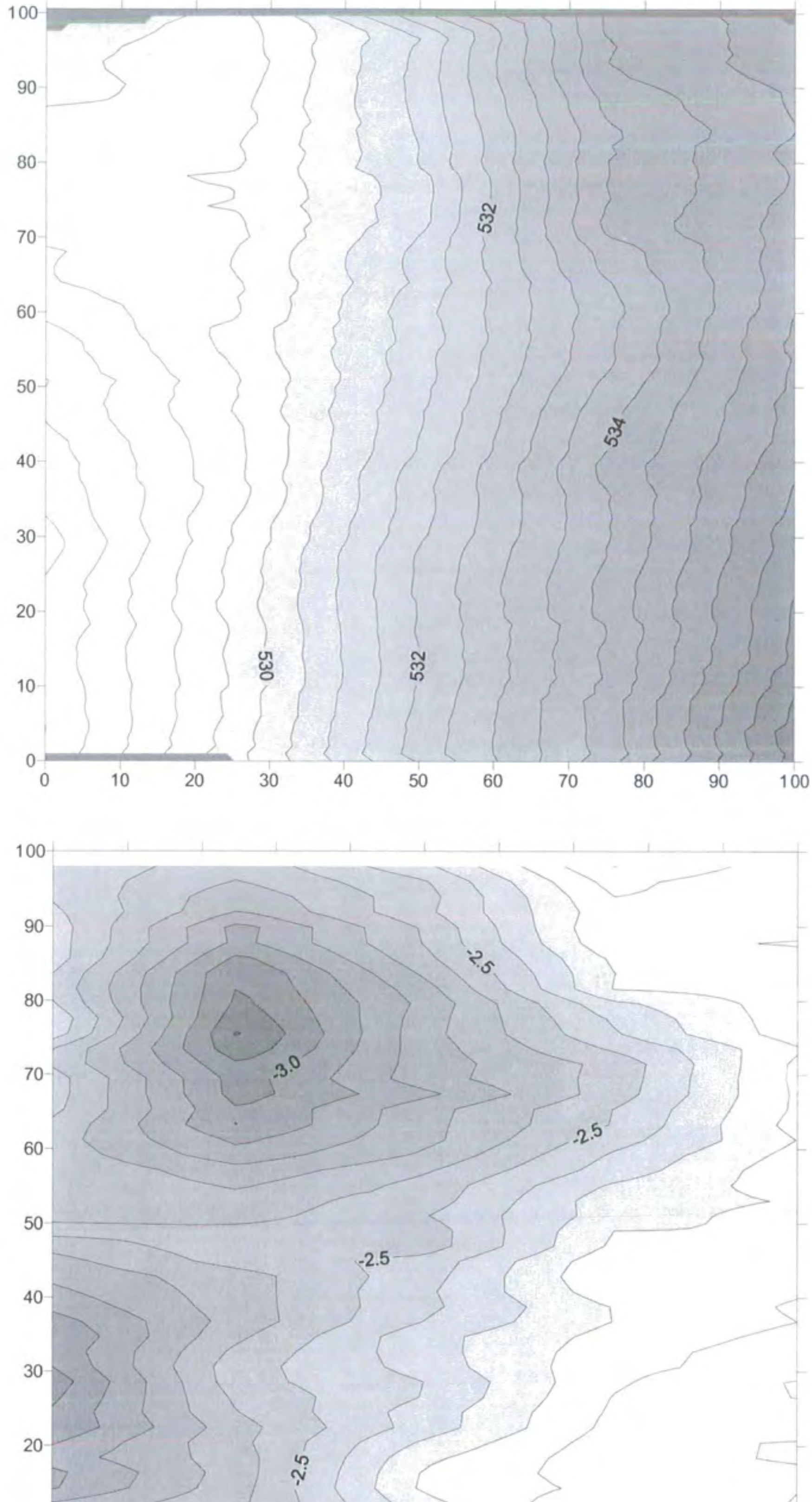


Figure 8.3c. Surface topography (upper) and peat depth (lower) across the ECN target site. North is vertically up the page, axis distances in metres.

detected using the 200 MHz antenna. The pipe was found at the interface between the peat and the substrate and results were verified by field observation. Upslope of this point it was unclear where the pipe ran below the surface. Therefore five more short transects were traversed across the slope. The GPR profiles created for transects 2 and 6 are shown in Figures 8.4b and c as examples. Again the pipe was located at the interface. It became evident that pipe 11 ran much further upslope than original field mapping had suggested; discharge measurements at site 11a (Chapter 7) had indicated that this was likely.

8.5.2.1 Pipe depth, size and interpretation of radargrams

Accurate measurement of the dimensions of the subsurface pipes using the GPR technique is difficult. This is because of multiple reflections from the pipe roof, floor and sides and also because the cross-sectional area of a pipe displayed by the GPR will depend upon what angle the profile cuts across the pipe. A perfectly cylindrical pipe cut across at right angles will produce a circular pipe form on a profile. However, if the profile cuts through the pipe at any other angle, the result will be an ellipse. Theoretically, if the GPR transect followed the lateral direction of the pipe perfectly then a pipe of infinite width would be displayed. It is probable that the very elongated form of some of the pipes displayed is due to this factor. Indeed the insertion of a long staff into the outlet of the pipe studied in Figure 8.6 showed that GPR transect 1 did not run at right angles to pipe direction thereby producing an elongated pipe form on the radargram. Images seen in radargrams are often not at all like images from X- rays in medical technology. Reflection profiles printed in two dimensions can look significantly different from the buried structures being searched for. The GPR antennae transmit energy through the ground in a wide beam; the antenna is therefore not only looking straight down but also in front, back and to the sides. For example when the antenna is in front of a soil pipe the travel time for a wave to leave the antenna is longer than when the antenna is directly over the pipe. Thus, the net effect is a hyperbolic-type reflection (Conyers and Goodman, 1997) of the pipe as the GPR moves over it, with the apex of the hyperbola denoting the top of the pipe. This hyperbolic reflection can be seen in Figures 8.4 a-c. Clearly interpretation of the radargrams is subjective, but with a little experience it is possible to confidently identify subsurface features from the plots.

In order to test the application of the GPR to detecting the depth of soil pipes, profiles were taken across several known pipes. The depth of the pipe roofs was determined by

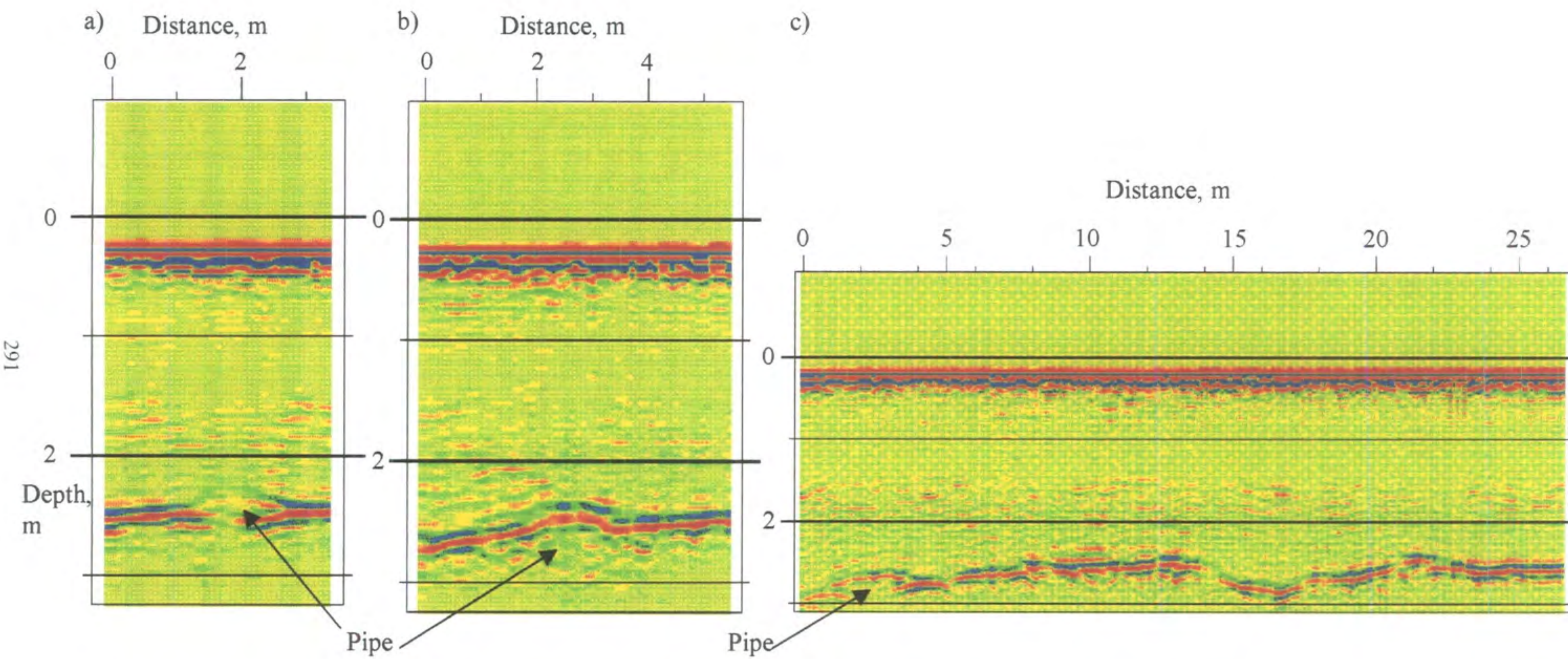


Figure 8.4. Example radargrams which allow identification of subsurface piping at LDPS monitoring station 11a, a) transect 1, b) transect 2, c) transect 6.

excavation or where outlets were visible on streambanks or gully sides. Figure 8.5 plots the relationship between actual pipe depth and GPR detected pipe depth. There is a close correspondence between the two measurements. Thus, although no pipes were observed by eye at depths greater than 2 m (due to obvious difficulties with digging trenches to observe them) it was decided to accept the GPR detection of pipe depths below 2 m as correct detection with an accuracy of +/- 30 cm. The plot in Figure 8.5 does not indicate, however, that pipes smaller than around 10 cm in diameter could not be identified, and four pipes at depths of 5, 5, 15 and 20 cm depth could not be observed on the GPR radargrams. The mean diameter of pipes in the LDPS catchment is 19 cm (standard deviation = 16 cm). Thus application of the GPR may be limited hydrologically as many of the pipes are below 10 cm in diameter. However, 70 % of the pipes measured manually were larger than 10 cm in diameter (e.g. see Table 7.2).

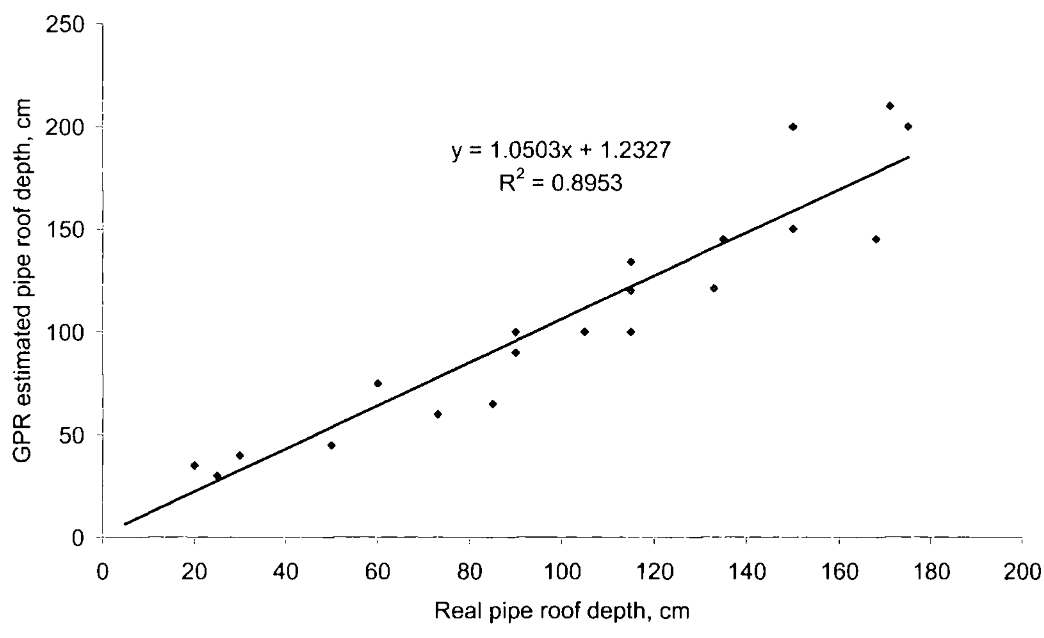


Figure 8.5. Pipe roof depth against GPR measured pipe roof depth as determined by the depth of the apex of each pipe on the radargrams.

The GPR profiles shown in Figures 8.3 and 8.4 indicate that use of the 200 MHz antenna produces multiple reflections from the peat surface. The thick red and blue lines on the figures within the top 30 cm are merely multiple reflections, or echoes, of the peat surface. Thus, use of the 200 MHz antenna means that objects (including pipes) within the upper 30 cm of the peat would be difficult to identify; neither the water table nor the acrotelm/catotelm boundary can be observed.

8.5.2.2 Comparison of antenna frequency

Data produced using the 200 MHz antenna compared with those from the 100 MHz antenna are shown in Figure 8.6a and b for a transect in the LDPS catchment. Clearly the resolution is poorer using 100 MHz although it does seem to pick up extra information on layering within the peat, whilst not picking out the substrate as clearly. A greater depth of the upper layer of peat is covered by surface echoes when using the 100 MHz antenna, although at the same time more information appears to be provided on laminations within the peat over the upper 60 - 70 cm. However, it is clear that in line with the findings of Hanninen (1992) higher frequencies are required to allow more detailed analysis of the near-surface layers. With higher frequencies, the depth of probing will decrease. Given the nature of soil pipes at Moor House which are often found at the base of the peat layer, it was decided to use the 200 MHz antenna as this gave better resolution for identification of pipes down to around 10 cm in diameter and was found to give penetration depths up to 5 m. Furthermore it allows more near surface pipes to be identified than when using the 100 MHz antenna whilst allowing the full depth of peat to be examined. Ideally it would have been best if the 200 MHz antenna was used in conjunction with an antenna of much higher frequency (e.g. 500 MHz). This would have allowed both deep and shallow pipes to be identified. This is of importance given the need outlined in Chapter 7 for a technique to aid identification of source areas and mechanisms of pipe supply.

8.5.2.3 Spatial distribution of pipes

Figure 8.7 provides data on the spatial distribution of the GPR-located pipes using the 200 MHz antenna uplsope of the gauging station 11a in the LDPS catchment. Figure 8.7a shows that soil pipes were detected along 5 of the 6 transects. With the peat almost 3 m deep in places and the pipes frequently located near the base of the peat, the benefits of being able to identify pipes using this remote sensing technique are clear. The digging of trenches is destructive and time consuming. However, it is not clear from the map in Figure 8.7a whether the pipes are connected, nor was it possible to locate the sources of the pipe network. Three pipes were found along transect 5; Figure 8.7b indicates that two of them were close together just below the peat mass, with one pipe within the peat. It is difficult to say from the spatial distribution of transects how these three pipe sections are linked but it may be that occasionally a pipe which is well connected to the surface feeds deeper pipes. Gilman and Newson (1980) described a shallow anastomising pipe system in the Welsh uplands. The pipe system at LDPS may

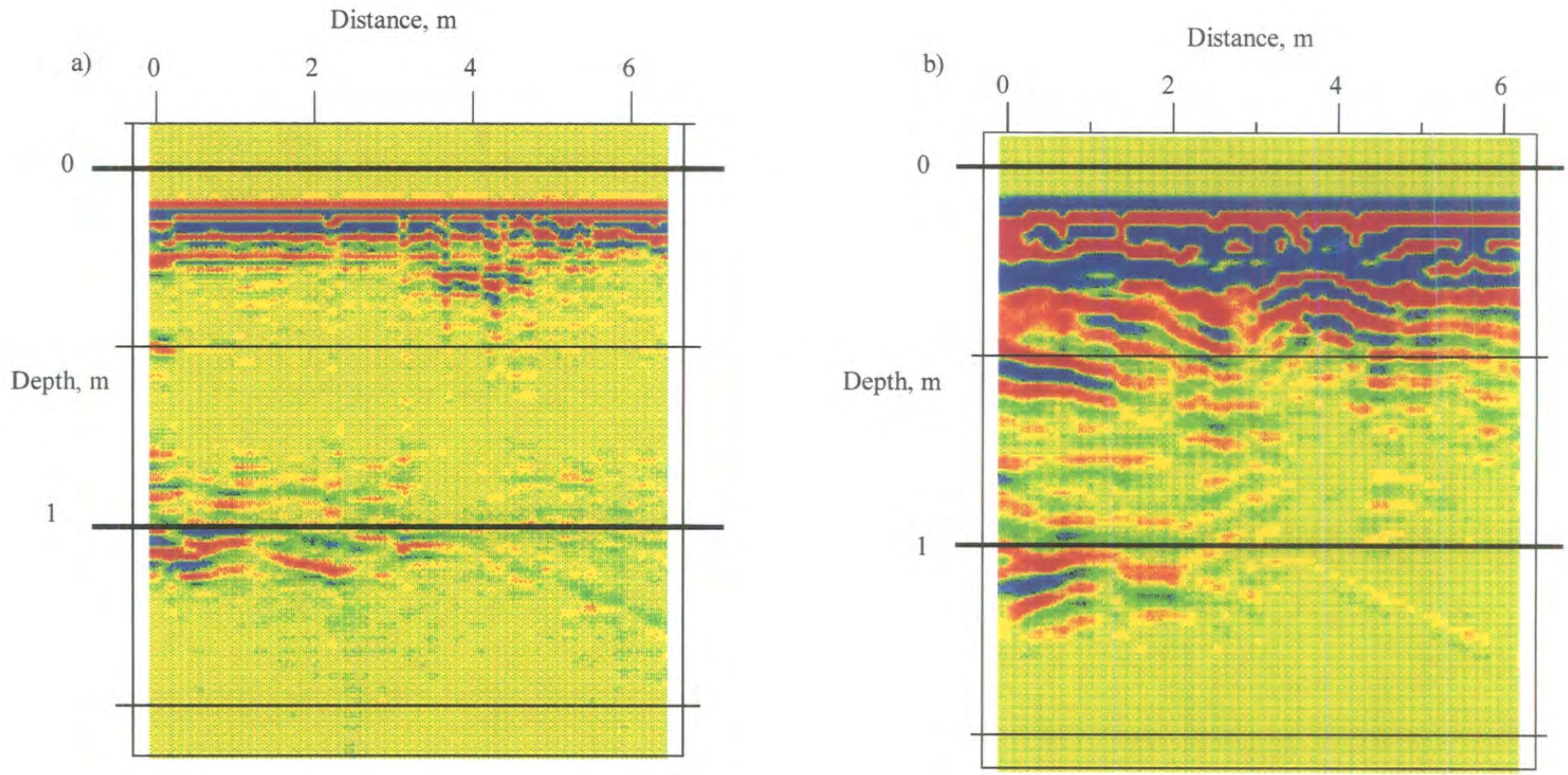


Figure 8.6. Comparison of two radargrams on the same transect. a) using 200 MHz, b) using 100 MHz.

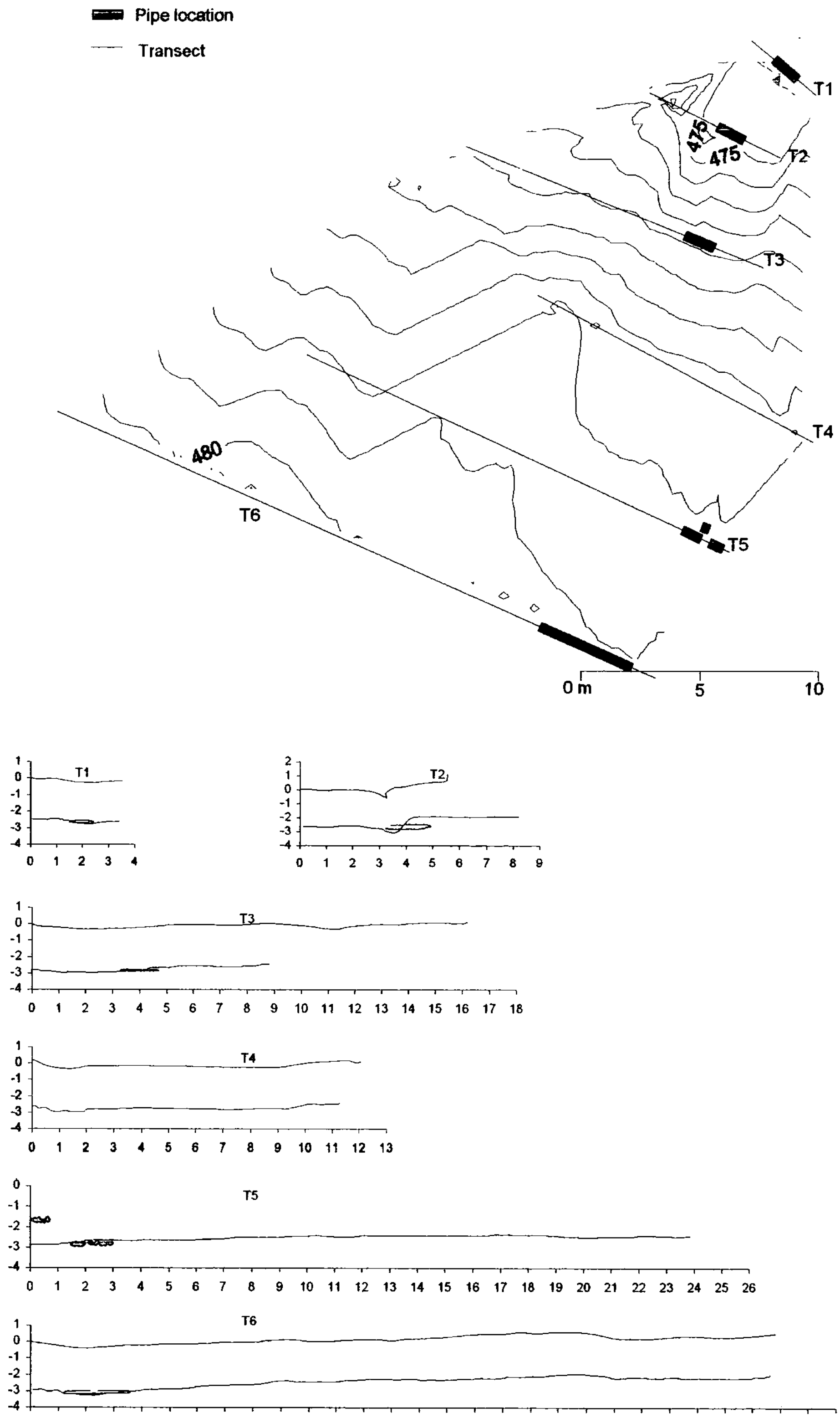


Figure 8.7. Pipe identification using GPR upslope of the 'visible source' of pipe 11. a) Spatial distribution of piping identified along the transects, b) depth of piping determined from the radargrams. Pipe sizes and shapes, indicated by red, are very approximate due to limitations outlined in the text. Axis distance is in metres.

be of a similar nature; the observations in Chapter 7 which indicated that pipeflow could decrease and increase along the course of a pipe because of other pipes draining the main pipe upstream and then feeding back into the pipe downstream, provide further evidence that this may be the case. It is clear that to accurately map the deep LDPS pipe network a very high density of transects would be required. Nevertheless this preliminary experimental work with the GPR gives some indication of the complex nature of deep-seated piping in blanket peat.

Figure 8.8 shows an example of where the GPR appeared to detect a pipe which could not be seen from surface observation. The location of the pipe detected varies from being at the interface to entirely within the peat mass. This backs up the field evidence discussed in Chapter 7 and shown in Figure 7.11a where pipes with an outlet entirely within the peat mass can produce sediment which is minerogenic in nature. Thus, the feeding pipe network upslope must make contact with the substrate at some point. Figure 8.8 also demonstrates that there is more than one pipe upslope of the single streambank outlet (transect 3). Thus, it is very likely that the subsurface drainage density of pipes is much greater than mapped in Chapter 7 on the basis of surface observation. This would help explain the high proportion of streamflow that passes through the pipe outlets monitored in LDPS.

8.5.3 The link between pools and pipes

The gully-head plot on Burnt Hill, which was examined in Chapter 4, was also traversed using the GPR. The transects were spaced at 2.5 m intervals (Figure 8.9a, b.). The pipe from which flow was measured on Burnt Hill can be identified on 5 of the seven transects and is associated with wet flush or ponded areas on three of the transects. The pipe seems to run below the pool on the plot. This situation has also been observed in the Flow country, Scotland (A. Baird, pers comm.). The pipe broadly follows the surface depressions, which in turn seem to follow subsurface depressions (Figure 8.6b and c). There is therefore some suggestion that initiation of gully erosion may in some cases follow pre-formed substrate drainage lines. On the other hand erosion of the substrate by the pipeflow coupled with pipe roof collapse or slight slumping due to the removal of material from the walls of the pipe by pipeflow erosional processes may account for the form of these features.

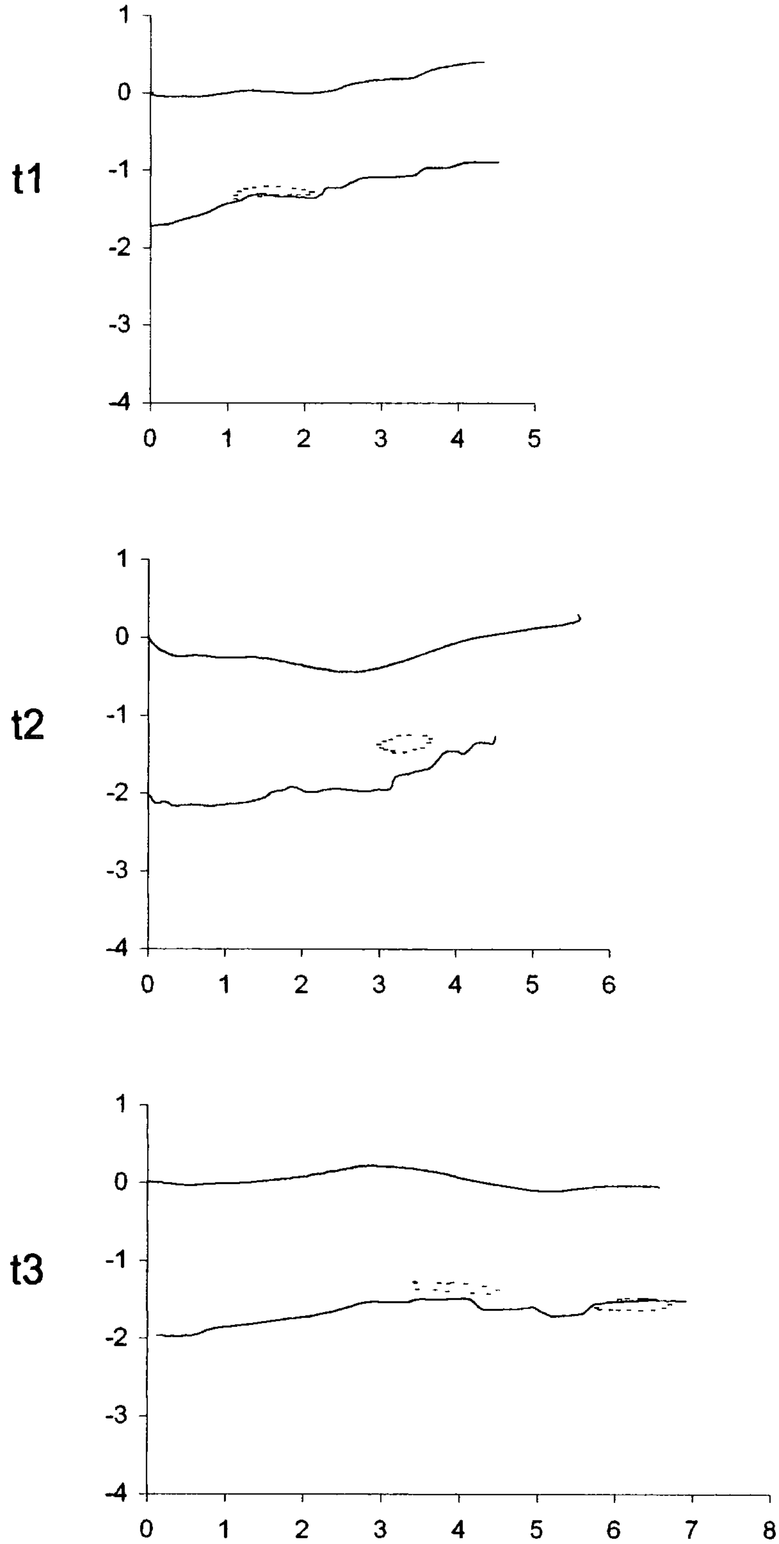
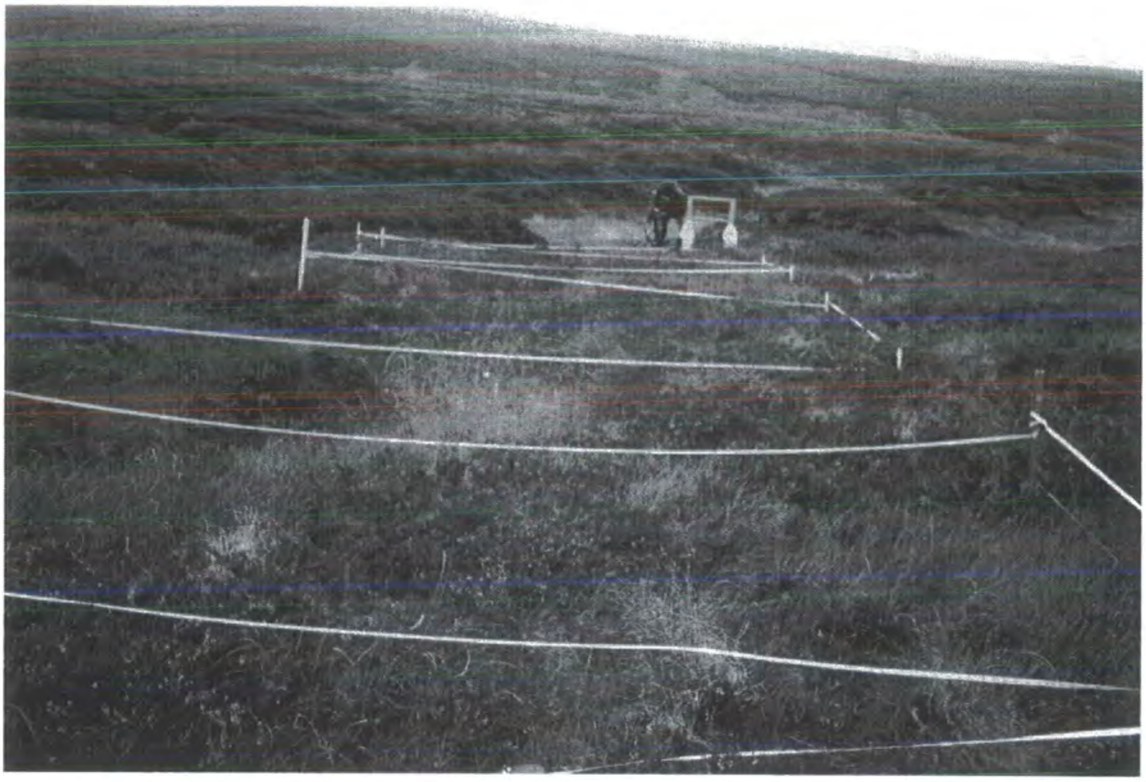


Figure 8.8. Profile identification of a pipe which was not visible from surface observation. The pipe indicated by the red zone appears to branch upslope. Axis distances in metres, x-axis = horizontal distance, y-axis = vertical distance (depth).

a)



b)

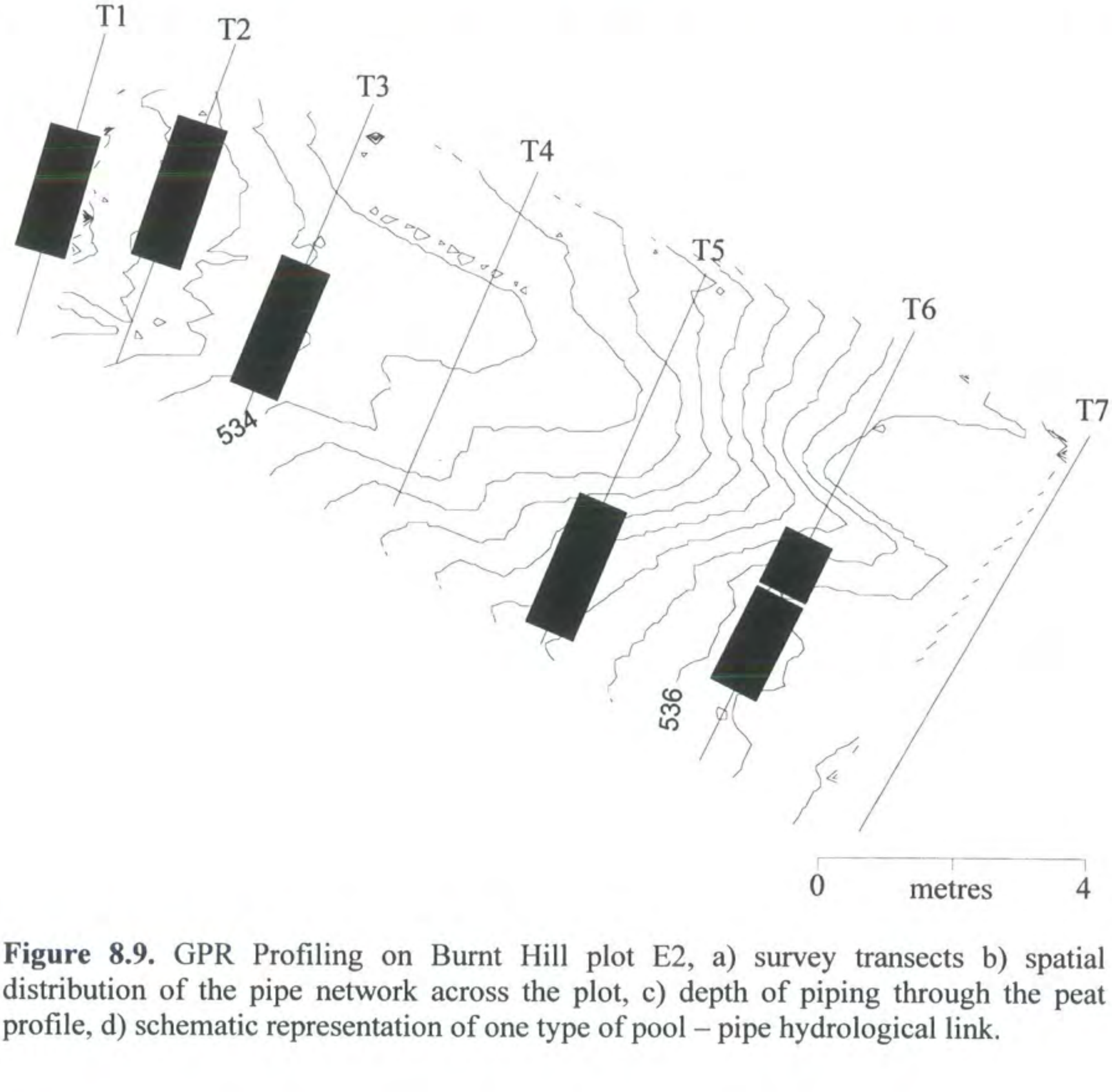


Figure 8.9. GPR Profiling on Burnt Hill plot E2, a) survey transects b) spatial distribution of the pipe network across the plot, c) depth of piping through the peat profile, d) schematic representation of one type of pool – pipe hydrological link.

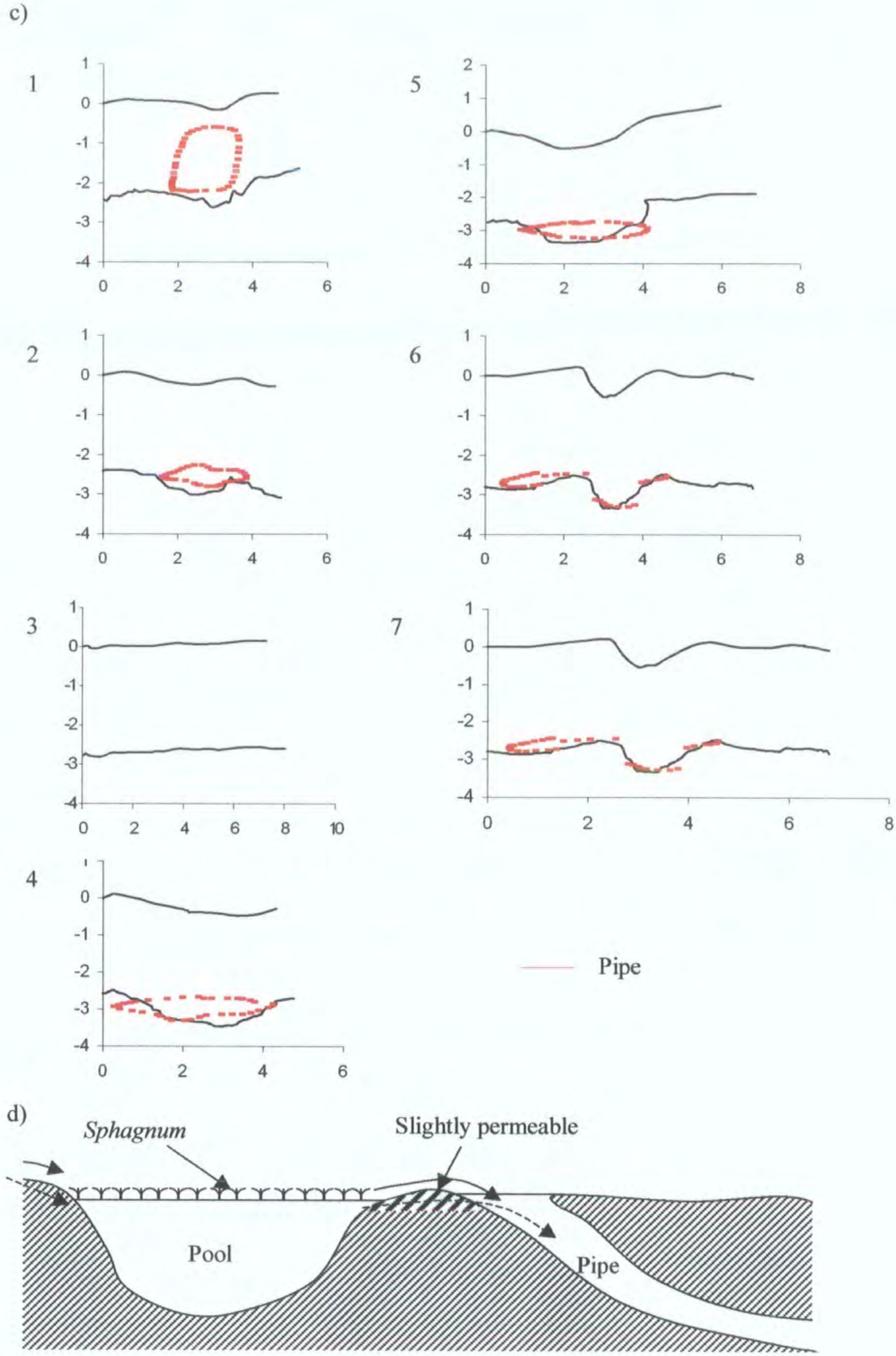


Figure 8.9. c) depth of piping through the peat profile, d) schematic representation of one type of pool – pipe hydrological link.

Along transect 6 apparent piped sections were in fact ponded areas which contained very little peat. They were hollows, or old blocked pipes, filled with water with a covering of *Sphagnum* and *Eriophorum*. The roof of these 'apparent' pipes on the GPR profiles could not be identified, indicative of the fact that these features were watery hollows. Hence the GPR was able to detect areas where peat was not present in the profile and thus both full and empty cavities could be identified. The water-filled sections identified by the GPR differ from pipes in that water cannot readily flow through them due to entrapment by surrounding peat. That is to say, they cannot readily transmit water through the peat mass. In the case of the Burnt Hill plot it seems that the wet ponded areas which collect water from the surrounding hillslope in turn feed the pipe system (dye tracing confirms this to be the case). During high flow OLF and near-surface flow runs into the depressions causing the pools to overflow and generate further runoff downslope. This water then enters the pipe system downslope of the blockage. Figure 8.9d schematically represents such a situation. As rainfall ceases, OLF and near-surface flow continue to flow into the ponded hollows until the hillslope has drained to the base of the acrotelm. Because the pool receives drainage for an extended period it continues to overflow producing a more prolonged recession limb from the pipe, as seen in Chapter 4. Due to the nature of the ponding and growth of peat around the edges of the pools, the downslope lip of these zones may have a deeper, or more permeable acrotelm. Simple laboratory permeameter tests of the peat at Moor House do suggest that the edges of pools are more permeable down to a depth of 20 cm than non-pool connected peat (Table 8.1). Thus a slow shallow drainage of the pool can occur. Whilst recognising that peat does not behave in a Darcian way (see Chapter 4) and indeed narrow hysterisis loops were found on the rising and falling head permeameter tests, the experiment was used as a quick and easy rough guide to hydraulic conductivities for comparative purposes. The pools can drain to a shallow level but once the water level is around 20 cm below the surrounding surface, the pools seem to only lose water by evapotranspiration.

8.5.4 Multiple interface reflections and the effect of water and air-filled cavities on signal attenuation

Given that EC governs wave propagation speed, one may expect that a large pipe filled with air or a deep pool of water may alter the signal such that a much different estimate of substrate depth is given. This would happen because the time for the signal to reach the substrate and reflect back to the receiver would be different from the time taken if there were just a layer of peat between the GPR and the substrate. The radargrams indicate, however, that where areas of peat are very wet or ponded, such as on hummock-pool terrain, there is little adverse effect on identification of the substrate (e.g. Figure 8.10a). Ponding does seem to have an adverse effect on the interpretation of the upper peat layers, however, with the surface echoes being pronounced around 20-30 cm deeper than usual. This may simply reflect the fact that the peat itself is 20-30 cm deeper with a pool of water on the surface filling the upper depth. Of geomorphological importance is the observation shown in Figure 8.10a that some of the peat gullies have formed over pre-existing drainage lines in the substrate.

Table 8.1 Hydraulic conductivity values from near surface peats on pool edges and peats from non-pool areas as determined by simple laboratory permeameter tests

Location	Sample depth, cm	Mean K cm s^{-1}	Mean DBD g cm^{-1}	Mean % water by mass	N° of samples with no flow
Pool edges	0 – 5	> 1 *	0.08	95.6	0
	5 – 10	4.35×10^{-2}	0.10	91.5	0
	10 – 15	6.59×10^{-2}	0.10	91.8	1
	15 – 20	3.48×10^{-4}	0.12	90.3	1
Non-pool peat	0 – 5	7.76×10^{-3}	0.12	88.4	0
	5 – 10	2.60×10^{-5}	0.15	91.0	1
	10 – 15	2.93×10^{-7}	0.16	89.7	2
	15 – 20	$< 10^{-8}$ *	0.18	88.5	5

* discharge rate was too difficult to accurately measure due to high / low flow

ANOVA for K values was performed after logarithmic transformation: $F = 15.82$ (Prob>F = 0.004)

GPR transects which ran across gripped hillslopes demonstrate that signal movement through the air gap created by a grip does not distort the estimate of the depth of the substrate to any great extent. Figure 8.10b shows that grips of 50 - 80 cm depth result in an apparent rise in substrate depth of approximately 10 cm. Thus, when peat is deep, the effect of GPR moving over a pipe or cavity will not cause a large error in substrate depth estimate, although it may cause a slight apparent rise in the substrate. Figures 8.10a and b both indicate a secondary layer at around 4 m in depth. This is not a real geological layer but an echo from the original reflection of the bedrock surface at around 2 m. Most of the radar energy that is reflected at the subsurface interface is transmitted directly back to the surface and recorded at the receiving antenna. Some of the reflected energy has, however, been re-reflected back into the subsurface at the ground-air interface, and then re-reflected back again to the surface from the same

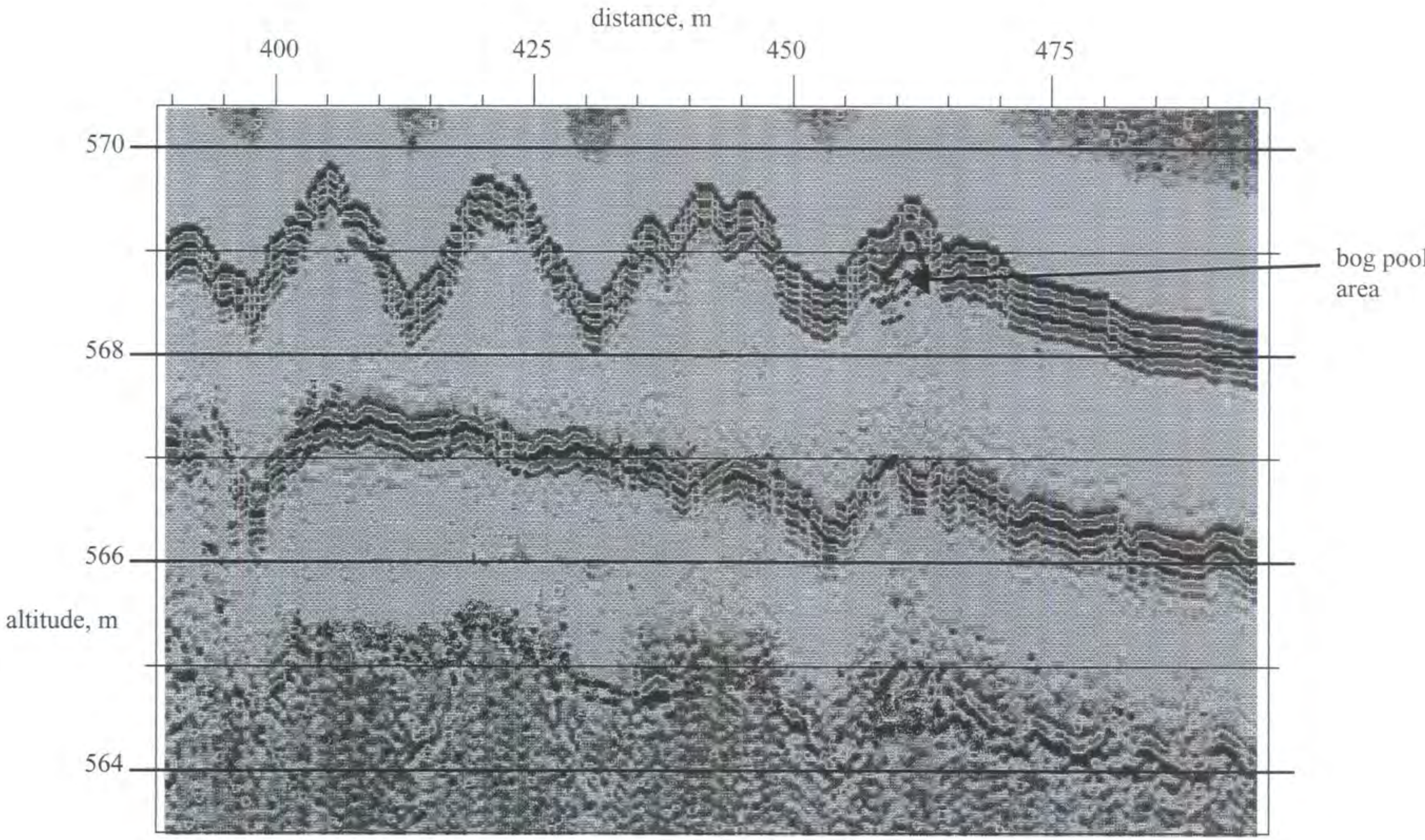


Figure 8.10a. An example of greyscale radar profiling across the eroded Burnt Hill subcatchment (E1). Note that two of the four gullies formed in the peat appear to have formed on pre-existing depressions in the subsurface topography.

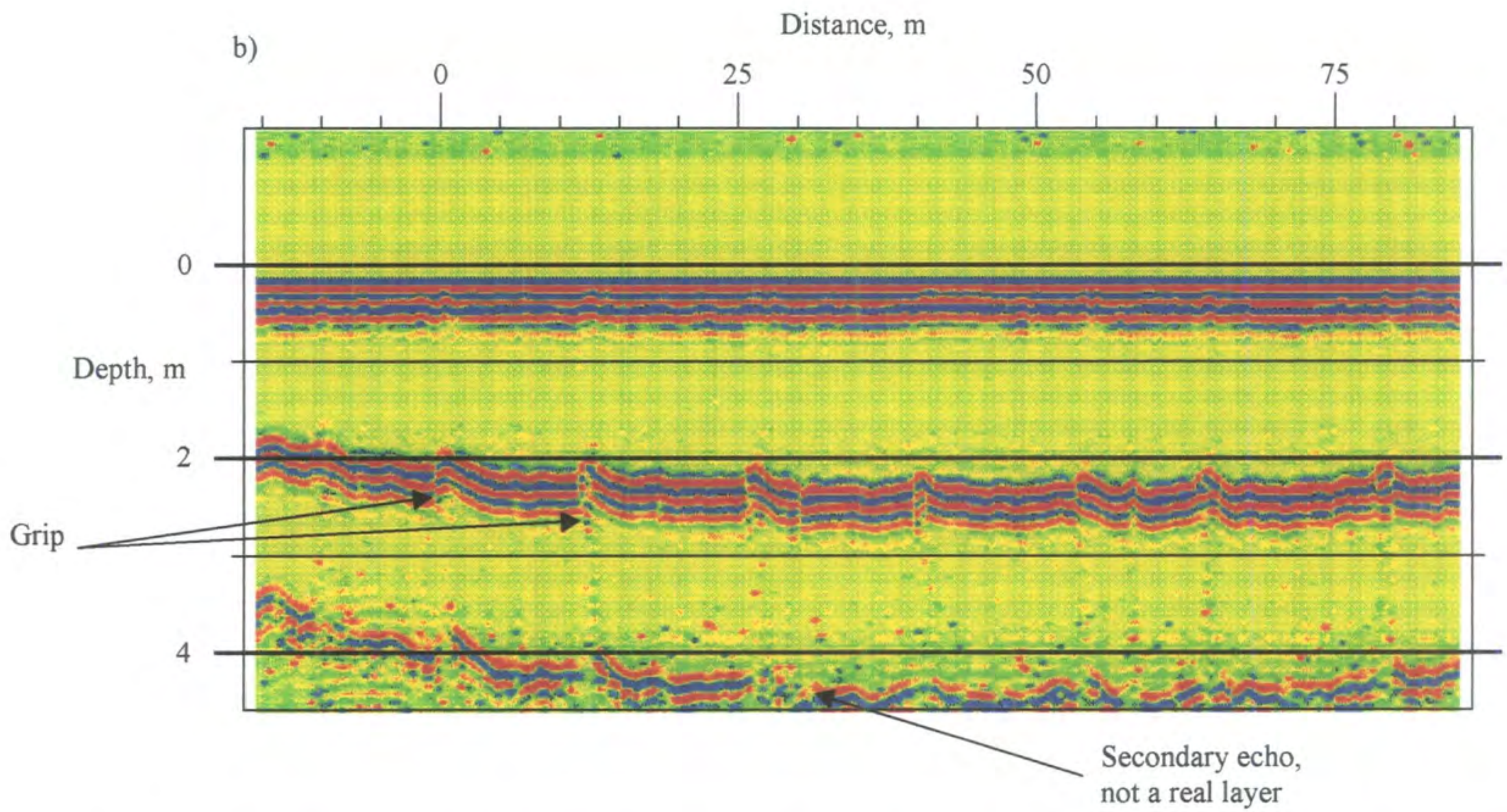


Figure 8.10b. An example of GPR profiling across the gripped Burnt Hill subcatchment and the effect of the grips on substrate depth measurement.

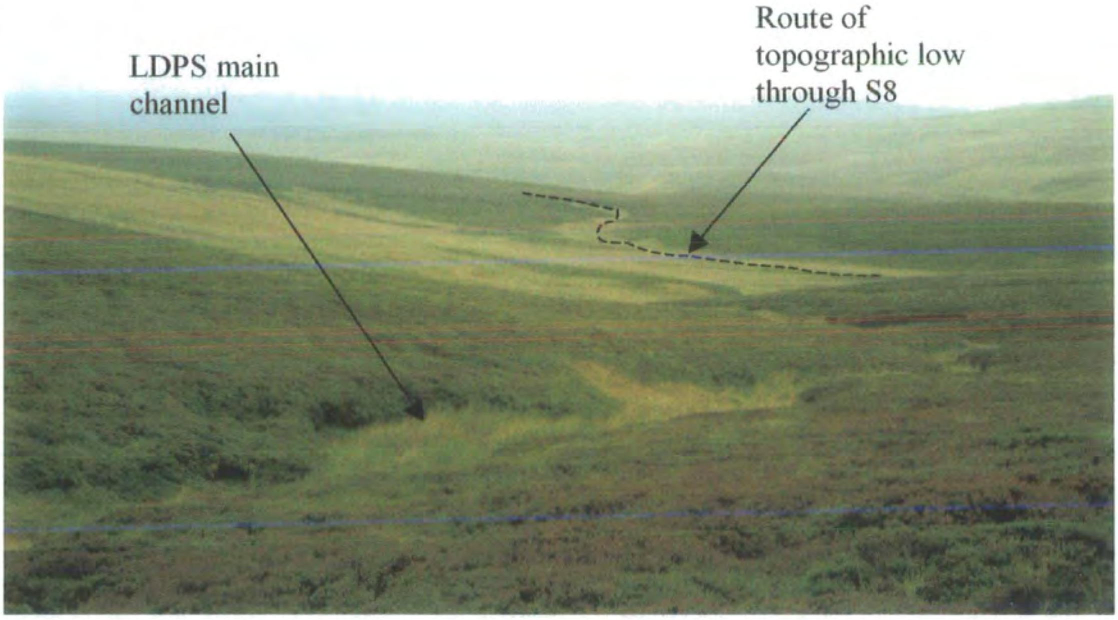
subsurface interface. Thus, the secondary 'apparent' bedrock layer appears at approximately twice the distance as the real bedrock interface. The multiple reflection, is usually much lower in amplitude due to geometric spreading during its travel, energy attenuation, and additional reflection from numerous other interfaces along its path. The colour palette used for data analysis can display the relative amplitudes of reflections. Nevertheless, GPR profile analysis is often subjective and requires some experience in handling data to allow correct interpretation of reflections.

8.5.5. A search for piping along a valley floor.

LDPS pipe 8 was discussed briefly in Chapter 7. Its origin was known to be a small plunge pool where water running from the floor of a gully upslope entered an opening in the peat which led to a coarse gravel layer at the peat-substrate interface. A significant volume of water was seen to enter the pipe system at this location. The gully ran into a valley-bottom area (which can be seen in Figure 8.11a) which then connected to the main stream channel. It was difficult to map the subsurface drainage system down through the valley-bottom from surface observation. Occasionally pipe roof collapse allowed the presence of a pipe to be identified and also allowed some estimation of flow volumes and pipe size. Figure 8.11b shows one example. At high flow these collapse features can fill with water such that they overflow on to the surface. Water also flows over the surface and near-surface to enter these holes and so access the pipe system.

GPR transects were taken across the valley bottom at right angles to the slope at 10 m intervals from the gully mouth at the top of the valley to the stream at the bottom. Figures 8.12a and b plot the transect locations and the spatial distribution of GPR identified piping within the valley. Piping can be traced down toward the stream over a 160 m section. Along some transects more than one pipe can be identified. The piping cannot be identified on some transects, however, and thus cannot be confirmed as continuous. It is possible that the pipes deviate away from the transect locations and moved upslope or that in some places were not identifiable by the GPR. Examination of slope data suggests that the areas with the greatest number of pipes (perhaps where the pipe system anastomises the most) are on the gentlest slopes (Figure 8.12a). Twelve of the eighteen pipe profiles found using the GPR were within the peat layer whereas the other six were at the interface between the peat and underlying substrate (Figure 8.12b). As in the examples discussed above, it seems likely that feeder pipes located close to the surface connect down to the deeper pipe networks. In terms of the development of the

a)



b)

c)

Figure 8.11. Seepage zone and pipe 8. a) Aspect of the valley, b) sinkhole with c) pipe

a)

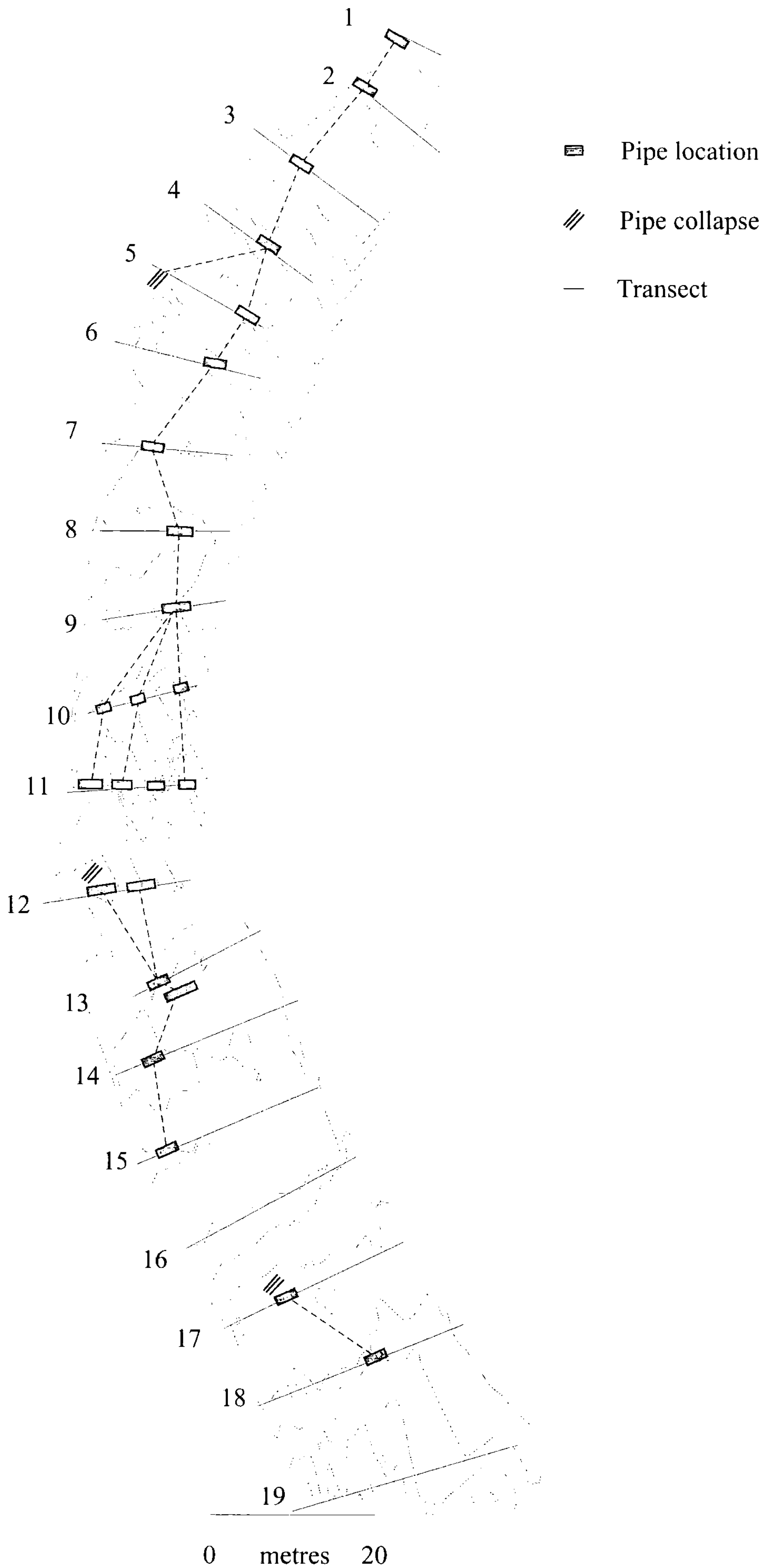


Figure 8.12. Location of pipe 8 as determined by GPR survey.

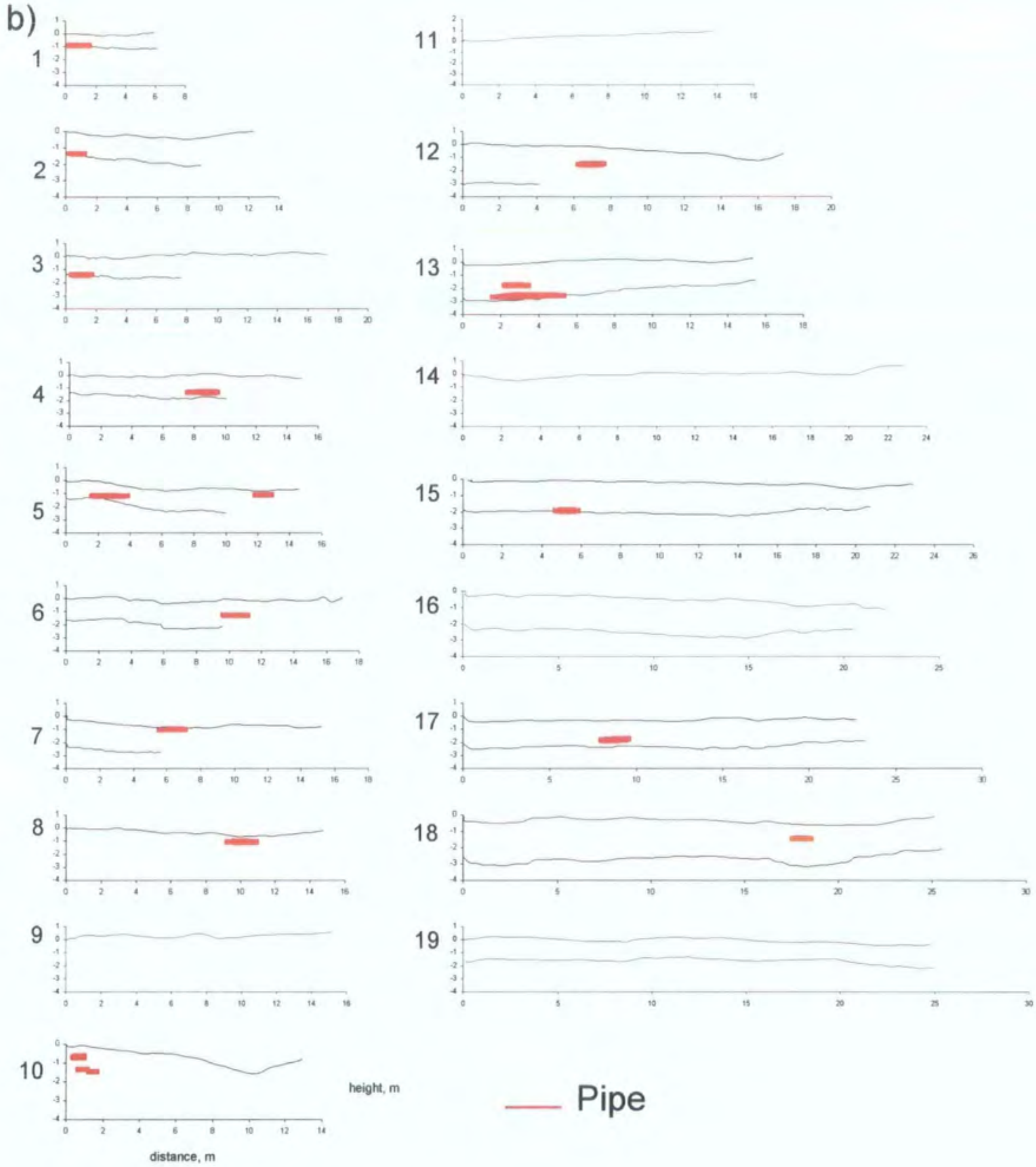


Figure 8.12 continued

pipe system over time, it may be that a pre-existing drainage channel which ran along the valley floor became overgrown by peat deposits and was eventually roofed in. The fact that the pipes can be identified along some transects well above the peat-mineral interface may be related to blockages within the original system causing backpressures and resulting in vertical as well as horizontal erosion of fresh pipe tracks through the peat mass.

8.6 Conclusions

8.6.1. Summary

GPR has been used for the first time, to the author's knowledge, to remotely sense soil pipes in blanket peat. The pilot study done at Moor House has shown that the application of GPR to subsurface pipe detection is successful within these upland catchments. The following findings were made regarding use of GPR in peatlands to remotely sense soil pipes:

1) The GPR can identify pipes in blanket peat catchments.
2) Comparison of data on pipes identified by the GPR and data verified by manual measurement suggest that pipe depth can be located in the soil profile with an accuracy of 20 to 30 cm.
3) In agreement with ground survey, as discussed in Chapter 7, soil pipes were identified throughout the soil profile (except near the surface – see below).
4) Pipes very close to the surface of the peat could not be identified using the 100 or 200 MHz antennae due to multiple surface reflections.
5) The smallest verified pipe identified by the GPR was 9 cm in diameter.
6) The GPR work did strongly suggest that pipe densities were much greater than could be recognised from the surface.
7) It was not possible to confirm the connectivity of pipe networks between transects.
8) Substrate topography and peat stratigraphy appear to be easily identifiable on radargrams. Thus deeper peat-filled hollows have been identified on small hillslope plots at Moor House.

8.6.2. Suggestions for improvements to the technique and further applications

The suggestion of a more anastomising pipe system on flatter areas of peat seen in Figure 8.12a is reminiscent of Bower's (1960) Type I and Type II gully erosion systems (see Chapter 2). Bower argued that gullies are more branched and anastomised on flatter slopes, feeding into straighter, unbranching gullies on steeper slopes. Further work is required to test the possibility that gentler peat slopes contain more anastomising pipe networks and to test linkages between piping and gully erosion, and those between gully erosion and substrate topography.

Higher frequency antenna are required to increase the data resolution in order to identify pipes smaller than around 10 cm in diameter (as long as the reduced probing depth of the higher frequencies still provides adequate coverage). In order to map the full range of pipes in deep peat from the substrate to the surface it may be necessary to develop multiple frequency antennae, the results from which should be digitally combined to produce full-depth profiles. For blanket peats around 3 m deep a 200 MHz antenna should be combined with that of a much higher frequency (500+ MHz). This would allow both deep and shallow piping to be identified at the same time and to examine whether deep pipes are frequently connected to the near-surface to be fed by the shallow runoff that dominates blanket peat catchments. The use of a higher frequency antenna would aid identification of the very shallow water tables and acrotelm/catotelm boundaries that are found in blanket peat. Repeated GPR surveys during storm events may then yield more continuous data on saturation-excess hillslope runoff processes through water table monitoring.

Improvements in using the GPR for soil pipe work would come from applying recently developed GPR techniques in archaeology. In archaeology synthetic computer GPR profiling and computer-generated three-dimensional maps have been developed which allow the subsurface features to be visually plotted and viewed from all angles on-screen (Conyers and Goodman, 1997). Used on blanket peat this would aid our understanding of pipe form, connectivity and sources within upland blanket peat. If GPR technology can advance over the next few years such that large areas of hillslope can be surveyed much more quickly, perhaps by the use of an array of probes scattered over a hillslope, using the data to produce a three-dimensional model of the soil profile, then hillslope hydrologists will make significant gains in understanding an important component of subsurface catchment runoff.

As it stands GPR, with the appropriate antennae, can remotely detect the location of soil pipes in blanket peat catchments if they occur below the areas along which the equipment is taken. Given the difficulty in detecting pipes from surface observations this is an important geomorphological and hydrological tool. The application is limited, however, in that GPR demonstrates the presence of pipes but does not establish their hydrological importance, or connectivity.

CHAPTER 9

CONCLUSIONS

9.1 Review of research objectives

Studies of the hydrology of blanket peat are rare and there have been limited measurements of the processes responsible for runoff generation within these important upland catchments. Therefore our understanding of the links between hydrology, ecology, erosion, hydrochemistry and climate change in blanket peat catchments is, at present, still very incomplete. The overall objective of the research presented in this thesis was to provide greater understanding of the processes responsible for runoff generation in blanket peat catchments. The thesis has presented results from monitoring and experimental work on runoff production in the blanket peat catchments of the Moor House NNR, North Pennines. The work was focussed on improving our knowledge of the patterns of OLF and near- surface flow generation, infiltration processes, the role of macropores in runoff generation and the role of subsurface soil pipes in deep peat catchments. This chapter aims to draw together the research findings from the individual investigations.

9.2 Major findings

9.2.1 Quickflow production

Runoff from the blanket peat catchments monitored in this study was flashy. Lag times were short and rainwater and snowmelt were efficiently transported out of the catchments via quickflow-generating mechanisms such that flood peaks were high and low flows poorly maintained. There are implications for catchment management, particularly in terms of winter rain-on-snow events and the increased occurrence of summer drought conditions. The ECN water table data analysed in conjunction with Trout Beck flow data suggested that peat saturation rather than low infiltration rates may be responsible for rapid runoff from the catchments. There was a strong relationship between river flow and water table height such that high flows only occurred when the water table was very close to or at the surface. On no occasions was a low water table at the ECN site accompanied by high stream flow. This suggests that infiltration-excess OLF may not be a common occurrence in blanket peat catchments. For 84 % of the time the water table at the ECN target site on gently sloping peat was within 5 cm of the surface; thus surface saturation can be quickly achieved.

In Chapter 7 peak flows and lag times from the LDPS and Trout Beck catchments were plotted on a summary diagram from a variety of soil types collated by a number of authors (Figure 7.21). The data correspond well with the theoretical saturation-excess OLF data envelopes. Judging from the catchment-scale data, saturation-excess OLF appears to be the primary runoff-generating process in blanket peat catchments. Much of the monitoring and experimental plot- and hillslope-scale work presented in this thesis adds weight to this simple comparison.

9.2.2 Importance of the near-surface and surface peat for runoff generation

Runoff plot work suggests that around 82 % of flow in non-piped areas of peat is generated across the surface, with 17 % from the top 5 cm of the peat deposit. Baird *et al.* (1997) suggested that, even though the lower layers of blanket peat have a low hydraulic conductivity and little water seems to emerge from them, it may be that because these layers are thicker they may generate comparatively more runoff than the upper layer. This line of argument does not stand up to testing. The lower layers of peat matrix produce very little runoff. During low flow conditions when the quantities of runoff were small, around 63 % of the runoff was generated from the top 5 cm of peat whereas runoff below 10 cm depth was rare. At high flow, the contrast is much greater still.

As well as quantifying the importance of runoff generation from the upper and lower peat layers, these data are also of immense importance to the development of useful hydrological models of blanket peat catchments. The main models of peatland hydrology quoted extensively in the literature are groundwater-based such as the 'groundwater mound model' of Ingram (1982). Whilst this is a raised bog model, such models are often applied to other mires. The model revolves around the nature of water held within the main body of the peat mass and the fluxes that may occur within the peat as related to the shape of a hillslope, or 'mound' of peat. Ingram (1982) stated that to improve and explore the model further 'we need data on the permeability of the deeper catotelm and on groundwater discharge as a water budget item'. Results presented in this thesis show that the lower layers of the peat matrix are unimportant contributors to flow in blanket peat catchments. Indeed the groundwater mound model of Ingram (1982) and Ingram and Bragg (1984) appears to be somewhat irrelevant to modelling requirements in blanket peat. Whilst models such as MODFLOW (Harbaugh and McDonald, 1996) and DRAINMOD (Skaggs, 1980) can be used to describe groundwater alone, they inherently lack any adequate representation of the near-surface conditions. The data presented in this thesis, which provides information on surface and near-surface flow processes at a variety of temporal and spatial scales, should allow more accurate and useful models of peat hydrology to be developed; not least because the data may aid model parameterisation. The inclusion of macropore flow through the upper and lower peat layers along with pipeflow processes would add further realism to such models. With surface cover type (perhaps representing the nature of the underlying peat) playing an important role in the rainfall simulator and tension infiltrometer tests, it seems that links to ecology should be incorporated into flow models in peatlands. Miss Charlotte MacAlister, University of Newcastle, is doing some initial work in this field (e.g. MacAlister and Parkin, 1999; MacAlister, 2000).

9.2.3 Spatial distribution of overland flow production

Ingram and Bragg (1984) suggested that OLF may be generally confined to bare peat areas where the acrotelm has been removed. They suggest that in vegetated areas the acrotelm may be self-sustaining by not enabling OLF generation and thereby protecting itself from sheetwash erosion. However, it is clear from the findings of this study that OLF is not restricted to bare peat. Whilst OLF on vegetated surfaces may be more difficult to see because of the plant canopy, rainfall simulator and tension-infiltrometer experiments along with plot-scale monitoring all confirm that OLF is generated under moorland vegetation. Indeed OLF can be generated on most peat surface types even during the lightest of rainfall events. This is in agreement with the work of Burt and Gardiner (1984) although they suggested that infiltration-excess may be most important for OLF generation.

Rainfall simulator results indicated OLF development on vegetated and bare surfaces over the 3–12 mm hr^{-1} rainfall intensity test range. This could suggest that blanket-peat infiltration rates are low. However, the work has demonstrated that low infiltration rates are not due to low surface permeability but more to do with low percolation rates below the surface and ponding at depth resulting in saturation of the near-surface layers. Thus, it is because saturation-excess OLF is being generated that results in a low infiltration rate. More evidence for this comes from the fact that even when the water table was low at the ECN site, (i.e. deeper than 5 - 10 cm), rainfall resulted in rapid recharge of the water table back to the surface within a few hours. Therefore infiltration rates appeared to be high enough when the peat was unsaturated that infiltration-excess OLF was unlikely to be the main cause of flashy runoff generation, particularly given the low rainfall intensities common in the Pennines.

Hillslope monitoring of water table and OLF occurrence by use of crest-stage tubes (described in Chapter 4) allowed some of the spatial characteristics of surface and near-surface flow generation to be assessed. OLF was more likely to occur in areas where the water table was regularly maintained very close to the surface. This was readily achieved on gentle slopes, such as on hill crests and footslopes. Steeper midslopes had a less frequent occurrence of OLF and a resultant greater proportion of subsurface flow. Footslopes are more important contributors to runoff, particularly in the generation of saturation-excess return flow. Hillslopes tend to drain their gravitationally available water quickly following a storm event such that runoff levels drop to very low levels within 24- 48 hours. Hence catchment-scale runoff is flashy. Gripping simply intercepts and transfers OLF and near-surface flow out of the catchment at a quicker rate. It thus reduces saturation-excess OLF development on the downslope side of the grip because upslope supply of OLF and near-surface flow is cut off.

9.2.4 Conway and Millar revisited

Conway and Millar (1960) stated that an intact blanket peat catchment could store much more water than a drained (eroded or grip-dissected) catchment. A dissected catchment would provide earlier and higher peak flows. This conclusion was then misinterpreted in the literature (see Chapters 2 and 4) as suggesting that undamaged blanket peat catchments were good regulators of flow. In fact their data never suggested that this was the case; runoff production from intact catchments was also very flashy. However, comparison of process operation on Burnt Hill has yielded important information on how differences in runoff generation were missed in the original Conway and Millar (1960) study of the site. This is important given the wide citation of their paper. Whilst the eroded subcatchment is very heavily dissected, water yield was equivalent to that of the Trout Beck catchment which is much less dissected. The water yield was also similar to that of LDPS where there is very little gully dissection. Flows from the gripped subcatchment were peakier with shorter lag times than from the eroded subcatchment but water yields were around 15 % lower. While peak flows from the eroded subcatchment are generally higher than the gripped section, the bulk of the extra water yield from the eroded subcatchment appears to come from the increased maintenance of low flows. That is not to say that significant volumes of baseflow are generated, far from it: flows drop to extremely low levels, becoming almost unrecordable, during dry summer spells. In comparison to the gripped subcatchment, however, the recession limb is extended. This appeared to be due to differences in catchment characteristics resulting in different runoff processes operating. However, the results should be treated with some caution. This is because they are not likely to simply represent a comparison of the effect of gripping on runoff with the effect of gully incision on runoff because the catchments examined are very different. The eroded subcatchment has many bog pools on its summit for example; these are lacking on the gripped hillslope. Bog-pools appear to supply more diffuse drainage via *Sphagnum* mats and pipes at gully-heads which helps maintain flow after runoff from the gripped slope had receded. The deep gully dissection results in water table lowering, weathering of the gully sides and an increased amount of runoff emerging from the lower layers of the peat.

9.2.5. Flow through the lower peat layers

This thesis has demonstrated that macropores are important generators of runoff within the upper layers of blanket peat. For the lower peat layers macropores may well be far more important than the matrix in runoff production. The observation of ephemeral flow occurring at the base of the peat in some locations suggests that a bypassing flow mechanism operates to connect surface waters to the peat base. Occasionally seepage points, as discussed in Chapter 4, can be seen on peat faces at depth. Monitoring of runoff from these outlets indicates that they are well connected to the surface peat layers. It is difficult to assess the importance of these well connected deep macropores to runoff-generation in peat catchments, particularly as it is difficult to measure how many of them have an unimpeded route to the stream channel. It may be that many macropore networks exist but, as they are not directly connected to a peat face, they merely fill during a storm event and runoff production from the networks is limited. Nevertheless the outlets of these networks which can be found occasionally on peat faces do represent one of the few ways the lower layers of peat contribute to catchment runoff.

The peat-substrate interface flow may also have implications for the stability of peat masses. Peat slides are a fairly common occurrence in the North Pennines (Warburton and Higgitt, 1998) and the failure plane is often considered to be close to the peat base. Further work is required to establish how common flow at the interface actually is and exactly how this runoff-generating mechanism operates. Field observation suggests that it is spatially localised but fairly frequent. Gardiner (1983) recorded substrate flow from some peat slopes in the southern Pennines. It may be that this flow can be linked to field observations of hollows within the peat mass where water can pond at the interface. Often, when cleaning up a peat face for investigation with a spade, water can suddenly gush from the lower peat layers as a reservoir is broken through. The variability in peat characteristics near the interface and the spatial variability in pore water pressure changes near this interface may be important for peat mass stability. The low variation in pore water pressure at piezometer nests in the study (Chapter 4) suggests that peat mass failure is unlikely to occur due to overburden pressure following a heavy rainstorm; other hydrological mechanisms such as the build up of pressures within subsurface pipes, or a reduction in frictional strength at the peat base seem more likely candidates. With warmer summers and wetter winters in Northern England in the future (Conway, 1998), peat slides may become a more frequent phenomenon. This may occur as more desiccation cracks can form in peat during the warmer summers, and with subsequent wetter winters a greater amount of flow can take place through the developing crack / macropore network to lubricate the peat base. This may also be linked to the further development of subsurface piping within blanket peat. However, much more work needs to be carried out in order to develop our understanding of these potential mechanisms.

9.2.6. The effects of drought on runoff production

Droughts result in very low flows from blanket peat catchments. This has implications for water management and ecology. The results from the pilot study of the effect of seasonal weather conditions on infiltration and runoff production during the spring and summer of 1999 was backed up by data from laboratory drought simulation. Steady-state infiltration rate, that is the infiltration rate after a long spell of constant rainfall when infiltration rate is no longer changing, was found to be greater during the summer than in spring. It is likely that this is related to the summer desiccation of the peat surface. Imeson and Kwaad (1990) and Burt and Slattery (1996) also identified time-dependent changes in infiltration rates, soil properties and runoff. Their work was done on agricultural soil however, and was related to agricultural practices, the effects of which were tied to seasonal controls. On blanket peat the effect was more pronounced on bare peat surfaces than on *Sphagnum*-dominated (high water table) zones. Drying and shrinking of the peat surface appears to cause both macropore flow (e.g. cracking) and matrix flow to increase. This effect is a more important control of surface runoff than the surface crusting of the peat. Following drought conditions OLF is reduced, macroporosity increased and subsurface flow becomes more dominant. The peat block experiments indicate that a structural change takes place within the peat that is not fully reversed once re-wetting takes place.

9.2.7. Piping

Anderson and Burt (1984) suggested that deeper pipes in peat may not supply much runoff. This is because throughflow through the deeper peat layers is restricted. However, the deep pipes studied in the LDPS catchment provide important contributions to flow. The monitoring work done at LDPS was the first to combine monitoring of deep and shallow pipes in a raw peat soil (c.f. the shallow peaty podzols on Plynlimon). The pipes were found at depths ranging from the surface to within the mineral substrate. The pipeflow response at LDPS was found to be different from that on Plynlimon. The Welsh study catchments dominate the literature and so it is important that pipe processes are presented from elsewhere. Unlike on Plynlimon (e.g. Jones, 1981; Jones and Crane, 1984) the ephemeral pipes at Moor House were not necessarily the shallowest and were found at all depths. In fact there was no significant difference between the dimensions or locations of ephemeral or perennial pipes in the LDPS catchment. The distinction between the two types of pipe is often not appropriate. Evidence from the LDPS catchment, including the GPR work, indicates that the pipes move regularly from deep within the peat to near the surface along their profile such that supply of water is not restricted to deep water from the peat. It may be that well-connected macropores from the surface and near-surface of the peat extending down into the profile provide one source of water to these pipes, but it seems that most of the water comes from OLF and near-surface flow directly entering the pipes where they are connected to the surface. Thus start lag times are short and hydrographs flashy. Jets and springs of water emerging from the pipes on the peat surface during storm events provide important evidence for the connectivity of surface flows and pipeflow.

At the same time, whilst runoff from the pipes is highly variable, generally the pipes exhibit longer recession flows than streamflow. It is this maintenance of low flows after streamflow has receded that results in pipeflows contributing over 20 % to streamflow in the catchment on the falling limb. During peak flows, when saturation-excess OLF dominates, pipes may only contribute 2-3 % to streamflow despite having high flow peaks. The 'perennial' nature of many of the pipes may stem from linkages between the pipes and hillslope hollows or flat bog pool areas which are poorly drained. It may be that pipes are connected via the acrotelm and the peat surface to these pool areas and flow is maintained, albeit at a very slow rate. Most of the pipes may be perennial, but flows are often extremely low.

Pipe networks in blanket peat are very complex. Pipes can lose and gain water along their length depending on how other pipes feed and drain the study pipe. Flows can also stop emerging from pipe outlets for long periods of time because of blockages upslope. Development of GPR technology is required in order to enhance its ability to locate subsurface pipes in blanket peat. Currently the GPR can easily detect pipe existence across a survey transect, but a dense network of transects is required to map the anastomising pipe networks. More intensive and dense surveying of hillslope areas, coupled with improved software development and other techniques such as dye tracing will be required in order to build more accurate models of the subsurface pipe network and its links with the surface layers.

9.3. Reconciling conflicting data; effects of measurement technique used

The rainfall simulator experiments in the field suggested that *Sphagnum* plots had lower steady-state infiltration rates than other plots. At the same time the tension-infiltrometer experiments suggested that *Sphagnum*-covered peat had a higher hydraulic conductivity and a greater proportion of functioning macropores. This was in agreement with laboratory rainfall simulator tests where *Sphagnum* peat had higher infiltration rates than peat below other cover types. Whilst this seems to be conflicting evidence, it may in fact be further evidence of saturation-excess OLF development. The control of surface cover type on infiltration rates and functional macroporosity may not necessarily be related to surface cover type alone as the vegetation often reflects near-surface peat conditions, particularly in terms of water table fluctuations. In the field *Sphagnum* tends to grow in areas with high water tables, often in slight depressions or hollows where drainage is impeded. Thus, saturation will be achieved more readily achieved and more OLF will be generated. At the same time the peat below the *Sphagnum* may have a higher permeability than the surrounding peat, with a greater proportion of macropores. Water is not quickly transferred out of the peat through the macropores, however, probably because the macropores are not well connected and also because the more permeable *Sphagnum*-covered peat is trapped by surrounding peat with a lower hydraulic conductivity. In the laboratory, the higher infiltration and subsurface runoff rates occurring in blocks of *Sphagnum*-covered peat may be a response to the fact that the block is no longer in a tightly closed naturally waterlogged system. Instead water is able to flow from the sides and front of the block via guttering into flow collectors. In the field, water was only able to flow out of the front of the bounded rainfall plot area to be measured. Thus comparison of laboratory rainfall simulation with field results should be treated with some caution. Nevertheless the laboratory tests do provide at least some comparative information on drought conditions and on runoff processes occurring within the peat blocks which would not be too dissimilar to that occurring in the field. The laboratory results on untreated peat blocks were, after all, of the same order and within the same data envelope as the field results.

A lower proportion of runoff occurred as OLF during rainfall simulator tests than was collected by the runoff troughs during natural rainfall events. This may reflect the scale and nature of approach. The rainfall simulator experiments used bounded plots in order to be able to estimate infiltration rates more accurately. The natural storm event monitoring did not use bounded plots as interest was in the timings and nature of flow processes at different locations and how flows were linked both upslope and downslope. Bounded plots of 0.5 m^2 are not equivalent to large hillslopes which can produce vast amounts of return flow from upslope drainage. Small bounded plots can only respond to the rainfall that hits their surface and runs off. Large hillslopes on the other hand, have to cope not only with incident rainfall but the flow that drains from upslope, often over hundreds of metres. Thus OLF generation on hillslopes is likely to be greater simply because of the head and supply of water from upslope. On small plots the effects of upslope drainage are discounted and thus the relative proportions of flow are not as they may be under natural circumstances. For example, if a hillslope were 100 m long and 0.5 m wide, an area of 50 m would supply runoff to the footslope. If rainfall fell uniformly across the slope the volume of water passing through the footslope would be 100 times that of a bounded plot of 0.5 m^2. Thus the dominance of saturation-excess OLF development would be less clear in the bounded plots. As with all experimental and modelling work, boundary conditions are changed by the way the process is measured. Whilst it is acknowledged that by bounding plots they become disconnected units from the hillslope, rainfall simulation experiments allowed control of the rainfall variable and allowed some examination of the way in which infiltration and near-surface runoff processes operate in blanket peat. It is clear, however, that both scale and location matter and at least by adopting a spatially and temporally distributed approach the data provided by this thesis allow plot- hillslope and catchment-scale comparison.

9.5 Concluding remarks

The data presented in this thesis are clearly applicable to North Pennine blanket peats which have high water tables and poorly developed acrotelms. This is not to say that the results are not applicable to other blanket peat deposits. Much will depend on the nature of saturation of the upper peat layers and on how deep the acrotelm has developed. Heavily eroded peat deposits may behave very differently with low water tables, desiccated gully-sides and heavily drained bog-pool systems. Climatic variations may also cause runoff production to vary between regions (for example, the North York Moors are much drier than the North Peninnes).

The next stage in the study of blanket peat catchments should be to tie the process-based hydrology presented in this thesis to geomorphological, ecological and water quality studies. Several key themes are worthy of further investigation:

1. Process-based measurement of the role of hydrological (and other) agents in the erosional development of the blanket peat uplands. Links should be made to the revegetation and recovery that is occurring on much of the North Pennine moors in contrast to the continuing degradation of the South Pennines.
2. The role of piping in sediment and solute delivery.
3. The exact nature of the sources of pipeflow in blanket peat, particularly the deep-seated perennial pipes.
4. The further development of subsurface mapping techniques so soil pipe networks can more effectively mapped. The improvement of appropriate GPR processing software would be of great benefit.
5. The effects of enhanced macropore flow and other structural changes following drought on water quality.
6. The role of hydrological processes in peat mass movements.

An improved understanding of the issues outlined above will allow enhanced management tools to be implemented. For example, it may be that removal of vegetation cover on peat by burning, coupled with warmer drier summers, results in cracking of the bare peat surface and enhanced macropore development. This may lead to a greater amount of bypass flow reaching the base of the peat resulting in a reduction of shear stress at the peat-substrate interface. In order that management is based on process-based science, it is critical that the full spectrum of linkages between process operation at a variety of scales is studied. It is hoped that this thesis provides a contribution to the knowledge of the nature of hydrological processes operating in blanket peat catchments.

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