AN INVESTIGATION OF SOIL WATER MOVEMENT ON DRAINED AND UNDRAINED CLAY GRASSLAND IN SOUTH WEST ENGLAND

by

PAULA JANE ADDISON

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ABSTRACT.

The Rowden Moor experimental site (A.F.R.C. I.G.E.R., North Wyke) provided an opportunity to characterise discharge regimes, elucidate runoff generation mechanisms and to consider implications for solute movement under natural and drained conditions. Research was conducted on a heavy clay grassland soil in an area of high rainfall (1053 mm a^{-1}) in South West England. A combined hydrometric and tensiometric study was undertaken within a nested experimental design (1 m^2 to 1 ha) on one undrained and one drained site throughout a drainage season (October to March).

Results at the hectare scale demonstrated that drainage did not substantially alter the volume of field runoff (~ 400 mm) but did change the dominant flowpaths. Drainage diverted water from surface/near surface routes to depth so that drain storm runoff was lagged by some 30 minutes over undrained site discharge. The drained site also exhibited a more peaky regime, with a maximum daily discharge of 45 mm being almost twice that for the undrained field.

At the field and plot scale, the significance of macropore flow was noted. To investigate this in more detail, a tracer experiment was performed on an isolated soil block which had been mole drained and so had enhanced macroporosity. Macropore flow was generated under unsaturated conditions (little matric potential response and no water table was identified). Stable oxygen concentrations were $\delta^{18}O$ +3.5 and -5.8 in tracer and background water respectively. Drainflow indicated that there was rapid interaction between applied tracer and soil water (peak flow $\delta^{18}O$ -1.1). Thus, the matrix-macropore interface was not a boundary between two separate domains of old and new water, high and low conductivity but a site of rapid interchange and mixing. Temporal variability of soil status and matric water composition, also indicated that limited areas of the matrix were capable of transmitting rapid flow. It became clear that even in a heavy clay soil such as that found at Rowden, where macropore flow was promoted by drainage operations, the matrix still had an important role to play. On the basis of potential, soil moisture and observation of tracers, it is proposed that discrete (finger-like) volumes of the matrix are capable of rapid water transmission. Although it was frequently impossible to relate moisture content and soil water potential because instrumentation monitored different volumes of soil, hysteretic soil moisture behaviour over the drainage season was evident in both data sets.

This study confirmed the importance of rapid subsurface runoff generation mechanisms on drained soils, but noted that discontinuous translatory flow in the matrix and macropore flow occurred and that the two 'domains' were inextricably linked. Further work should be undertaken at the detailed scale to elucidate the soil characteristics which promote rapid runoff mechanisms and the consequences for water quality, especially where the soil subsurface represents a major reservoir (e.g. nitrates).

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AUTHOR'S DECLARATION

At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award.

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Addison, P.J.; Williams, A.G.; Scholefield, D.; Dowd, J.F.; Wenner, D. (1992)
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External Contacts:

Department of Geology, University of Athens, Georgia, U.S.A..
A.D.A.S. Soil and Water Research Centre, Ansley Hall, Trumpington, Cambridge, U.K..

Signed ....

Date ....

CHAPTER 1: INTRODUCTION.

1.1 General introduction.

Substantial areas of lowland Britain have been underdrained to increase agricultural productivity and intensity of land use. Hydrological effects of drainage extend well beyond field boundaries (Hill, 1976). For example, there is currently concern over nitrates, pesticides and other agricultural pollutants which are transported through underdrained soils to river courses (Burt *et al.*, 1992; Haycock *et al.*, 1993). Drainage directly and indirectly affects soil properties, such that water pathways and runoff processes are altered. Impacts at a wider scale can only be understood and effectively managed through an appreciation of drained soil hydrology (Robinson and Beven, 1983; Reid and Parkinson, 1984; Randall, 1990; Scholefield *et al.*, 1993).

An experiment was formulated to monitor the hydrology of a drained pasture field (1 ha) and to compare characteristics with those of an undrained control site. Alterations in field hydrology such as volumes and timing of runoff could be determined. Within each field, a plot (10 m^2) was instrumented to observe spatial and temporal variation of soil water status. The importance of different flow routes was investigated and changes in runoff processes identified. At a more detailed scale, conservative tracers were used to characterise soil hydrology. Studies regarding water movement and solute interaction in soils must provide a theoretical basis for environmental management. Information from these should underpin the formulation, evaluation and implementation of schemes which aim to maintain or improve soil and water environmental quality. For example, Burt *et al.* (1992) propose riparian buffer zones to lessen problems of agricultural pollution such as nitrate loss and disposal of slurry.

1.2 Field drainage and its impact.

Since the nineteenth century much heavy, impermeable land has been subject to extensive field drainage. Ideas advocated by agricultural improvers could only be put into effect on many areas of land after drainage had been carried out. Drainage has been conducted over large areas of arable and pastoral production. Permanent pasture, the focus of this study, occupies about 50 % of enclosed land in England and Wales and much of this grassland is located on soils with naturally imperfect drainage (Forbes *et al.*, 1980).

Two innovations were of particular importance for the large scale development of drainage. The first was the invention of a process for producing clay tiles and pipes cheaply (Green, 1975). This led to deep land drainage, using permanent piped drains installed in trenches at up to 80 cm depth. Second, the development of steam powered ploughs quickly opened the way to the introduction of mole drains: unlined, round channels constructed without excavation (Raadsma, 1974).

Moles are installed at relatively shallow depths (0.45-0.55 m) and are closely spaced (2 m). They usually convey water to deeper, more widely spaced tile drains via permeable gravel, and are installed using a mole plough. The mole, a pointed cylindrical steel core of 5 to 8 cm length and 7.5 cm diameter, is attached to a horizontal beam by a leg (Figure 1.1). As the plough is pulled almost horizontally through the soil, the mole and leg create a number of cracks and fissures. An expander, of approximately 10 cm diameter, is attached to the mole by a short chain. This clears the channel and induces further fissuring (Godwin *et al.*, 1981). The efficiency of the drainage system is determined by these crack systems which provide an important and direct route for transmission of water to the channel.

The degree of success and longevity of mole drainage is largely dependent on soil type, especially clay content and mineralogy, and soil moisture content at the time of installation (Leeds-Harrison, 1982; Spoor *et al.*, 1982). The poor hydraulic conductivity of many soils necessitates close drain spacing but the low economic return on drainage in grassland prohibits installation of pipe drains at the close optimum density required (Findlay *et al.*, 1984). Therefore, mole drainage presents a relatively cheap technique and has proved to be the most effective treatment on many clayey soils (Trafford, 1976).

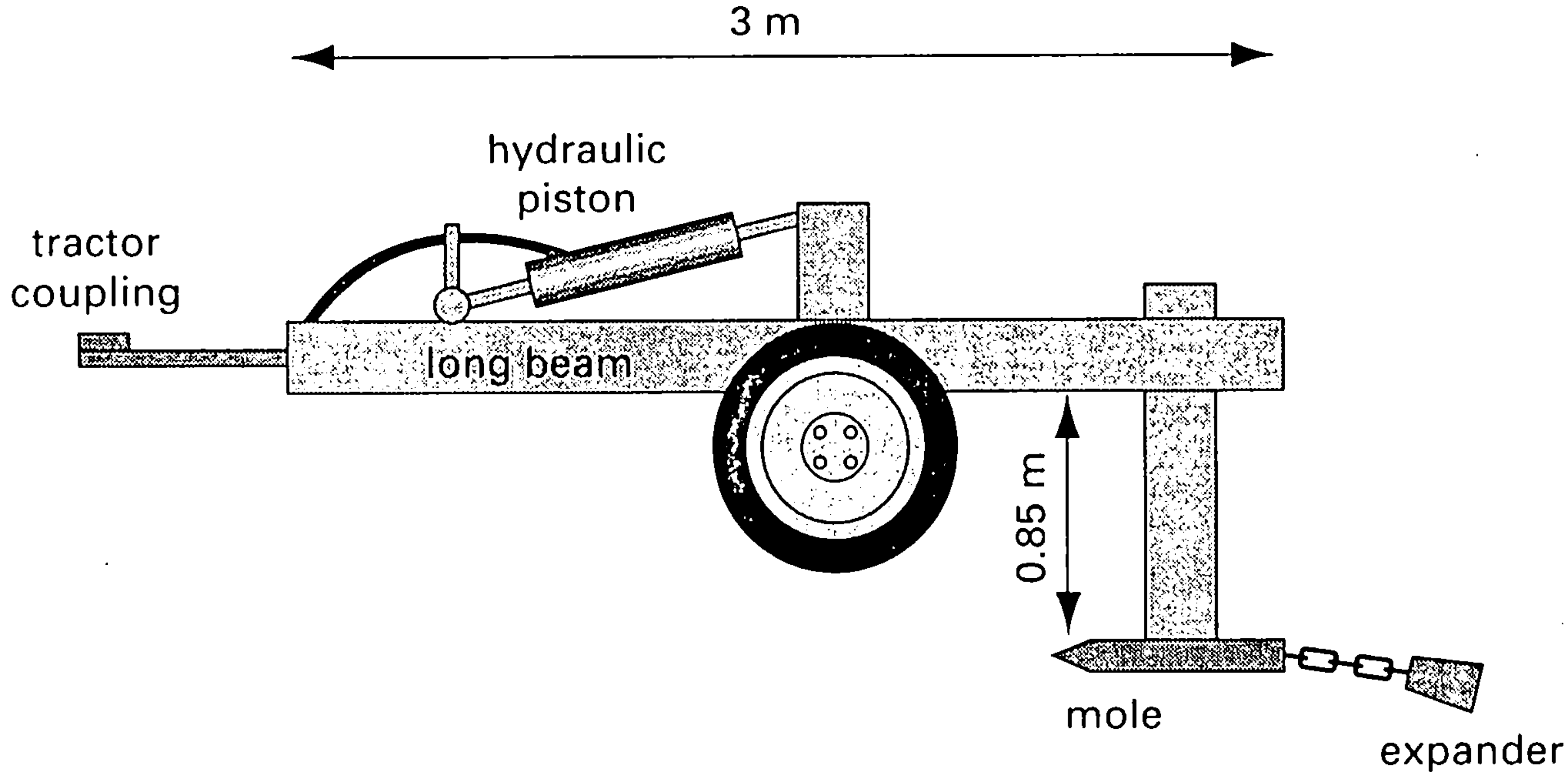


Figure 1.1 Mole plough.

By the end of the nineteenth century 5 million hectares, roughly one third of England and Wales, was drained by tile and mole drains. Drainage activity declined during the agricultural depression (12,000 ha a^{-1} in the early 1940s) and then rose to a maximum of 100,000 ha a^{-1} in the 1970s. It is difficult to determine the extent of field drainage in England and Wales today, because there has been inadequate documentation and much installation has involved the improvement or replacement of existing schemes. However, Green (1980) notes that the promotion of drainage was so successful that the United Kingdom is now one of the most extensively drained countries in Europe.

Drainage has four main types of impact: agricultural, hydrological, pedological and chemical. Agriculturally, field drainage in grassland areas lowers the water table, leading to drier soils during critical periods of the year (Armstrong and Garwood, 1991). Among the benefits are greater trafficability, an extended grazing season, increased rooting depth and a diminished risk of damage to crops due to lack of oxygen. Historically, agricultural engineering research has considered the efficiency of different drainage designs (Trafford, 1975; Le Grice, 1980), the impact of drainage on farm operations and the costs and benefits associated with field drainage (Armstrong *et al.*, 1988; Clark *et al.*, 1988; Tyson *et al.*, 1992). These areas of applied research are now less valid than before because of agricultural surpluses and the imposition of quotas.

Second, soil hydrological response is affected by field drainage, but evidence is complex and often conflicting. Research such as that quoted in Section 1.3 has led to two fundamentally different conclusions. One purports that drainage leads to an increase in the speed of runoff generation and higher peak discharges. Conversely, other workers report that efficient drainage leads to increased soil water storage capacity so that flood peaks are reduced by delaying runoff. The starting point for this thesis is an investigation of the hydrological nature of an undrained and a drained field or small catchment.

Third, pedological changes resulting from field drainage will lead to hydrological modification. Alterations in soil properties may be caused by the installation of mole drains. Soil structure is of crucial importance as it affects soil water pathways influencing both relative amounts and rates of overland flow and subsurface flow. The presence of macropores and fissures following drain installation can substantially alter routes of water movement. Similarly, increased biotic activity in the soil profile results in structural amelioration (Gilbey, 1986). The combined effect is of greater pore space and increased hydraulic conductivity, also changing the pattern of water movement.

Finally, spatial and temporal changes in soil hydrological characteristics influence the transport of chemicals and pollutants and ultimately affects stream water quality. Rapid transport of nitrate and farm waste from field to stream is causing grave concern (Randall, 1990; Addiscott *et al.*, 1993; Burt *et al.*, 1992; Scholefield *et al.*, 1993).

1.3 Field drainage and runoff processes.

Despite the extent of drainage in the United Kingdom, understanding of drainage affected runoff processes is limited. Impacts are complex and research results have been inconclusive. The lack of consensus from field drainage experiments reflects historical emphases on either technological issues (such as drainage design and efficiency) or on catchment hydrological questions (such as the influence of drainage density, slope or land use). Catchment studies of drainage effects have been inconclusive because they have been site specific: results have been greatly influenced by the type of drainage, soil type, agricultural practice and topography, thereby presenting conflicting conclusions; field drainage may enhance or reduce stormflow.

Decreased stormflows were reported by Egglesman (1972). He noted that peak storm runoff from an undrained catchment was three times higher than that generated by a drained catchment. This was attributed to reduced occurrence of flows in soil surface layers, lower water tables and greater subsurface storage. In other experiments, increased flood peaks have been recorded. Conway and Millar (1960) and Robinson *et al.* (1985) reported that drainage of peats increased the magnitude of peak flows.

Bailey and Bree (1980) observed that mole drainage modified the subsoil enabling significant water movement to depth. Rapid hydrograph rise was also due to increased drainage density. Thus, following mole drain installation, water was transported more rapidly from the soil surfsace, through the profile, the many mole channels and discharged into river channels.

Hallard (1988) and Armstrong and Garwood (1991) reported that mole drainage changed active pathways on grassland by lowering the water table such that throughflow (as opposed to surface flow) dominated contributions to hydrological outputs. This change in runoff generation was accompanied by a delay to the outflow hydrograph and reduced peak drain discharge. Armstrong and Garwood (1991) noted that there was no change in peakflows, the total proportion of water leaving drained sites or the rapid response and multiple peaks typical of clay soils. However, they calculated that peakflows from drained soils were delayed by 30 minutes.

The nature of soil water movement is a function of soil structure. Prior to drain installation water movement is dependent on the natural soil porosity and conductivity. As most soils suitable for moling are relatively impermeable and clayey, the contribution of crackflow is significant (Findlay *et al.*, 1984). During installation, systems of fissures are created around the mole drain. On wetter soils, the introduction of fissures enhances macroporosity and allows rapid movement of subsurface water and drainflow. In a study of the hydrological response of a clay soil to mole drainage, Robinson and Beven (1983), reported that the presence of macropores and fissures was critically important in determining the volume and timing of discharges. Similarly, Harris *et al.* (1984), indicated the significance of the mole slit and cracks to the removal of water from drained sites.

Soil structure, in turn, is influenced by agricultural practices. For example, permanent pasture characteristically has a well structured soil with a network of large pores surrounding the fine blocky aggregates in the rooting zone, resulting in strong grass growth and active water and air movement. Since grass roots are undisturbed for long periods, they very effectively ramify to depth, enhancing water infiltration and percolation. Good drainage encourages structural amelioration, while poor drainage leads to structural deterioration. For example, stocking when the soil is too wet can result in compaction and puddling (poaching) of the surface and lead to reduced infiltration capacity and greater occurrence of overland flow (Davies, 1983).

The fundamental importance of soil structure for water movement, as influenced by agricultural practices, was conclusively demonstrated by Hallard (1988). He showed that the hydrological response of reseeded mole drained fields and undrained permanent pasture control sites were dissimilar even though the mole drains were installed many years after reseeding. Ploughing and reseeding substantially altered soil structure, destroying the continuity of natural macropores. The bulk of the soil lacked a network of connected flow routes to contribute to new cracks and fissures. Mole drain macropores were hydrologically isolated from the surrounding soil so that drains were ineffective. On arable land Johnson *et al.* (1993) noted that saturation of the ploughed layer (which lacked continuous macropores) initiated macropore flow in an unsaturated subsoil.

Therefore, with all these factors affecting the hydrological response and water pathways of drained sites intensive and comprehensive investigations are required. The value of gaining deeper insights into the mechanisms controlling discharge should be substantial. Better comprehension will help in the explanation of impacts of drainage at the larger scale, and to address questions of great relevance to the environmental movement such as the passage of pollutants through soil. Perception of this issue has been heightened in recent years as the consequences of farming practices have been more apparent.

Nitrate pollution provides a good case study. Scholefield *et al.* (1993) noted that 'grassland agriculture had been considered environmentally benign' in terms of nitrate pollution. This was in spite of rising applications of nitrogen fertilisers (application of 250 to 400 kg N ha a^{-1} on dairy pasture would not be uncommon). Nitrogen from slurry variability of soil and hydrological processes. For example, Scholefield and Stone (1992) found that most nitrate fertiliser is fixed immediately upon application, but noted that mechanisms by which the nutrient is seized are unknown.

One problem has been the extrapolation of conclusions from experiments under different land management For example, nitrate leaching is not substantial under cut grass systems (Barraclough *et al.*, 1983). However, in grazing systems, ingested nitrogen is rapidly cycled and returned to the soil in excreta, which is concentrated in an available form in patches across fields (Whitehead, 1970). Further, with drainage, mineralization of organic matter is increased so that grazed land can be a significant source of water pollution (Ball and Ryden, 1984; Scholefield *et al.*, 1993). Therefore, the mechanism and speed of pollutant transport through grassland soils are pressing issues (Haigh and White, 1986).

Garwood, *et al.* (1986), found that nitrate losses from drained grassland fields at Rowden were two to three times those from undrained sites (Table 1.1). Following drain installation, water table drawdown (Bentley *et al.*, 1989) resulted in lower mean water table levels leading to less favourable conditions for denitrification (Table 1.2) and greater availabillity of nitrogen for leaching. Garwood *et al.* (1986), attributed nitrate leaching disparities to differences in soil water regime and water pathways.

Table 1.1 Loss of NO_3-N in drain and non drainflows (kg N ha^{-1}).

Treatment	1983-84	1984-85
Undrained (200 kg N)	16	28
Drained (200 kg N)	57	56
Undrained (400 kg N)	36	18
Drained (400 kg N)	79	72

Source: Garwood *et al.* (1986).

Table 1.2 Loss of N by denitrification (kg N ha^{-1}).

Treatment	March 1984 - March 1985
Undrained (400 kg N)	64
Drained (400 kg N)	46

Source: Garwood *et al.* (1986).

There are strong relationships between land management practices, the abundance of macropores and the infiltration rates of soils (Edwards *et al.*, 1984). Many workers have found that macropores represent a major pathway for solute movement (see for example, Shuford *et al.*, 1977; Tyler and Thomas, 1977; Shaffer *et al.*, 1979). Hallard (1988) argued that the role of macropores in the leaching process largely depended on the source of nitrate. If nitrate was easily contacted by rainfall and infiltrating water before macropore flow became operational then macropores would transmit both water and solute more rapidly to depth than uniform displacement approaches would predict (Smettem *et al.*, 1983). Barraclough *et al.* (1983) and Hallard (1988) believed that if nitrate was relocated within soil peds then macropore flow would produce inefficient leaching. Both the contact with the nitrate source and residence time would be insufficient for runoff to equilibrate with nitrate in the soil profile.

1.4 Research aims and conceptual approach.

The foregoing discussion has suggested that there is a need to investigate the hydrological impact of mole drainage on permanent pasture since this represents typical conditions of substantial tracts of lowland Britain and because the effects of field drainage extend well beyond immediate field boundaries (Hill, 1976) but are still poorly understood.

Specifically, the first aim was to compare the hydrology of an undrained and a drained field. Runoff regimes were characterised, and flow duration curves were computed for each treatment. Volumes of storm and total runoff were measured, and hydrograph characteristics were examined for selected storms. From these data, water pathways were deduced and an evaluation of their relative contributions was attempted.

The second aim was to study subsurface flow pathways in more detail. More precisely, the objective was to determine the nature and relative importance of matric flow routes compared with faster bypass routes in a well structured, mole drained, grassland soil. This was attempted through observation of soil hydraulic properties by instrumentation centred on mole drains. From soil water status measurements the gradient or water movement driving force can be determined and the soil water flux calculated from Darcy's Law. However, as explained in Section 1.6, Darcy's Law assumes that water moves through the soil matrix, so it may be inappropriate. For this reason, one- and two-domain modelling approaches were evaluated. An indication of the importance of matrix and macropore flow can be gained by comparing the predicted discharge volumes per unit time with those monitored at the subsurface monitoring weirs. Soil water status was also characterised for the control field to explain the patterns of water infiltration, percolation and movement in undrained soils. Furthermore, the temporal change in water pathways are examined by considering fluxes at different times of the year and relating this to the abundance of cracks and fissures in the soil and antecedent soil moisture conditions.

An investigation of the interaction between water moving rapidly through macropores and that stored in the aggregates themselves is the third aim. The hydrological behaviour of water moving down cracks and channels in a mole drained soil can be studied at a detailed scale using conservative tracers. These can be applied at the soil surface, and water samples collected at different depths as they pass through a block of soil. The speed of movement to and along the mole channel can be determined. Of further interest is the extent of mixing between new and existing water. These findings have implications for understanding how subsurface water and particularly groundwaters may be contaminated. Rapid flow may lead to decreased adsorption of pesticides, fertilisers and other pollutants so that they are transported rapidly below the root zone.

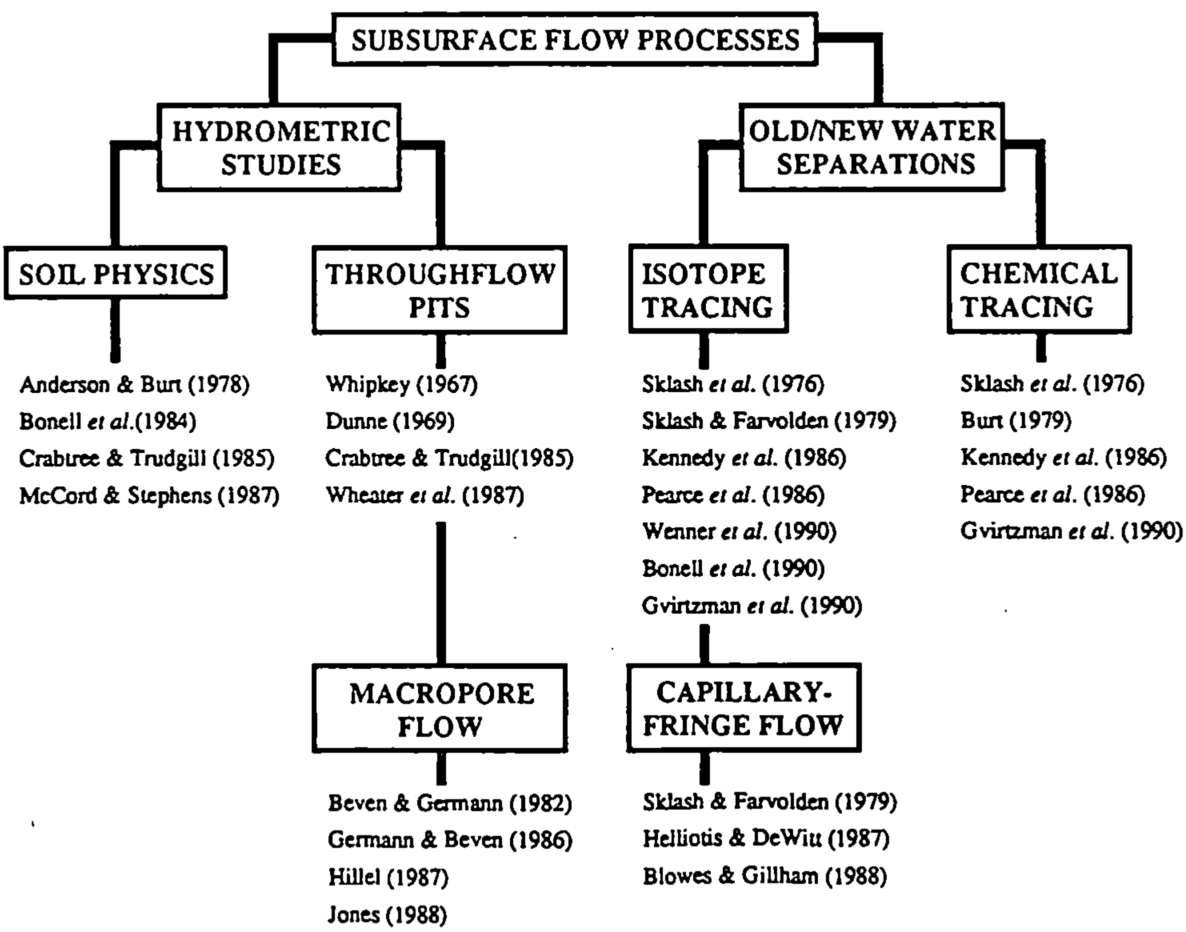
1.5 Experimental approach.

The experimental design is described in Chapter 3. The isolation of a small relatively homogeneous drained fields, with undrained fields acting as controls, provides a useful framework to study the hydrological effects of drainage. This mini-catchment method, in which the site boundaries are known, facilitates the calculation of water balances, and removes variations in climate, relief, soil type and management. The water budget for an area can be calculated from the continuity equation in which input equals output plus or minus change in storage. In this instance, the rainfall volume equals drainage discharge and evapotranspiration losses together with any change in soil moisture storage. The usual problem with determining a water balance is that catchment boundaries cannot be clearly identified. While this aspect did not pose a difficulty in this experiment, there are errors in the balance due to limitations of the way that evapotranspiration losses are calculated. This in turn affects the determination of deep drainage losses from the fields. Other, possibly more serious questions relate to how typical the small fields and their management are, and to what extent the four lateral drains represent a typical scheme. However, despite these problems, the isolated field can provide a useful framework within which to characterise water balances under various treatments, and to assess gains, losses and changes in storage.

The use of small replicated experimental fields is especially valuable to the detailed study of water pathways. Many drainage studies have neglected the surface flow component but Armstrong *et al.* (1984) has shown that significant quantities of water may leave the areas of heavy clay in this manner. Surface flows can be measured individually from the undrained and drained fields using V-notch weirs, installed outside the field boundaries so that equipment does not affect runoff generation. An additional weir on the drained field monitored subsurface discharge. With appropriate research design, the hydrological response in terms of volumes and timings of flow for the various surface and subsurface pathways can be compared. Explanation of differences, in terms of water routes, requires a rigorous interpretation based on the factors influencing the water pathways. Straightforward descriptive accounts may be misleading since several different combinations of flow routes could be proposed which would give the same results.

Central to an explanation of how water flows in the mole drained fields is an understanding of subsurface flow processes. Unfortunately, unlike surface flows, subsurface flows cannot be monitored directly. Many of the intellectual problems considered in this thesis revolve around the twin difficulties of conceptualising how subsurface water moves and in trying to monitor it. Such is the complexity that a wide variety of approaches is possible as Figure 1.2 reveals. Some of these, such as the use of throughflow troughs, can be shown to be of limited value since these installations totally alter the local hydrology by providing a sink for saturated flow to such an extent that they generally show that macropore flow is the most important runoff producing mechanism. Conversely, tensiometers have much to offer the investigation of both saturated and unsaturated subsurface flows. The soil matric potential at a point in the soil can be measured relatively easily but there are problems associated with attempting to mathematically solve the magnitude and direction of subsurface flow using Darcy's Law.

Figure 1.2 Summary of investigations into subsurface flow processes, highlighting processes and rapid response mechanisms (adapted from McDonnell, 1988).



In classic terms, saturated and unsaturated subsurface movement are a function of the energy gradient, known as soil potential, and the ease with which water can move through the soil matrix in response to the gradient. This relationship is formally expressed by Darcy's Law as explained in more detail below. Recently, a number of studies have shown that soil hydrological properties are highly variable even over a small area (Nielson and Biggar, 1977). However, studies of subsurface flow dynamics have demonstrated that, under field conditions, the soil matrix is often bypassed by water flowing rapidly through macropores (Bouma *et al.*, 1981; Beven and Germann, 1982). Darcy's Law is not applicable in these conditions.

One way to evaluate macropore flow is to trace water flow using naturally occurring tracers (Section 1.7). This study reports on an experiment conducted on a small block during which stable oxygen isotope and chloride tracers were used. The block was isolated from the surrounding soil to enable calculation of water balances. Where the composition of tracer and soil water are known, a simple mixing model can then be used to determine the flux of new and old water. The importance of macropore flow during a storm event can be determined as the relative amount of new water in the drainage water (Johnson *et al.*, 1993). Such a technique is invaluable for demonstrating whether water applied at the surface reaches the subsoil very rapidly or whether new water is mixed with old and merely pushes out the existing water with a piston action. Mixing of the new water with old water stored in the soil aggregates is most likely to occur if the passage of the water is retarded, for example, as it passes through the aggregates (Sklash, 1990). However, there are suggestions that mixing occurs almost instantaneously (Dowd *et al.*, 1991), and therefore the tracer experiment was particularly important because it may shed light on how pollutants are held or transported through soil.

The remainder of this chapter considers Darcy's Law and how this can be solved for transient conditions by Richards' Equation. The literature on macropore flow is then examined to demonstrate our limited knowledge in this area. Finally, the potential contribution of soil water tracer experiments are discussed.

1.6 Darcy's Law and Richards' Equation.

The movement of subsurface water is governed by the hydraulic conductivity or ease of water movement through a soil and the gradient of gravity and matric forces (matric potential). Hydraulic conductivity is a function of soil structure and soil moisture content. In formal terms, matric pore flow is the movement of water in small pores within and between soil aggregates in response to the total energy potential. Darcy (1856), derived the first quantitative description of water flow through a porous medium (Eqn. 1.1).

$$\frac{\delta \theta}{t} = \frac{\delta}{\delta z} \left[K(\theta) \frac{H}{\delta z} \right] - U(z,t) \qquad \text{[Eqn. 1.1]}$$

where:
- θ = volumetric water content ($m^3\ m^{-3}$).
- H = hydraulic head (mm).
- K = hydraulic conductivity (mm d^{-1}).
- t = time (d).
- z = depth (mm) - positive downward.
- U = sink (water loss per unit time, d^{-1}).

Darcy's Law makes a number of assumptions (Marshall and Holmes, 1988):

(i) average flow occurs through a homogeneous and isotropic medium. This implicitly assumes that the soil volume under consideration is sufficiently large to represent heterogeneities over the entire sample;

(ii) flow is laminar. Darcy's Law no longer applies once turbulent flow occurs, that is, when mean flow is greater than mean pore velocity (Luxmoore, 1981).

Much of the physics of soil water movement in the saturated phase for steady-state conditions is summarised by Darcy's Law. However, Equation 1.1 cannot be solved for the transient conditions found in unsaturated soils because hydraulic conductivity is a function of soil moisture content. As the soil drains, the remaining water is held in progressively smaller pores at lower potentials and so the conductive volume decreases and hydraulic conductivity declines. A reduction in soil moisture content of a few per cent can cause a reduction in conductivity of three to five orders of magnitude (Wellings and Bell, 1980). Therefore, since conductivity is a function of moisture content which depends on flux, there are two unknowns in Equation 1.1 so it cannot be solved.

Richards (1931) recognised that by combining Darcy's Law and the equation of continuity, also known as the principal of conservation of mass, the soil water equation for transient vertical flow could be solved. For steady-state conditions, continuity requires that the amount of water flowing into a volume of soil is equal to the amount flowing out. Assuming that there are no sources or sinks, the equation usually referred to as the Richards' Equation states:

Defining the differential water capacity, C (θ) as:

$$C(\theta) = \frac{\delta \theta}{\delta h} \qquad \text{[Eqn. 1.2]}$$

where: h = the soil water pressure head (cm H_2O).

This allows Equation 1.1 to be transformed into an equation in which pressure potential is the only dependent variable:

$$\frac{\delta h}{\delta t} (C \theta) = \frac{\delta}{\delta z} \left[K(\theta) \frac{H}{\delta z} \right] - U(z, t) \qquad \text{[Eqn. 1.3]}$$

To simulate flow and the redistribution of water in a soil profile, the first and second order partial differential equations must be integrated between given limits, usually referred to as boundary conditions.

The complexity of the equations is such that they are usually solved by computer models such as LEACHM (Hutson and Wagenet, 1992) or SWIM (Ross, 1993). They are both finite difference models which solve the Richards' Equation at a number of nodes within a soil profile to predict water movement. Soil layering and saturation at the surface can be taken into account. However, the Richards Equation is applied assuming that the soils are homogeneous horizontally and do not have a strong vertical structure so that macropore flow is unimportant (Hutson and Wagenet, 1992).

Problems with applying Darcy's Law get worse when soil heterogeneity is considered. From an experimental point of view, the temporal and spatial variation in soil hydrological properties must also be taken into account. Even small scale heterogeneities can cause large changes in soil water flux. For example, Nielson *et al.* (1973) demonstrated that saturated hydraulic conductivity could change by several orders of magnitude within several metres in a well mixed agricultural soil. Furthermore, temporal variations can also be significant because the seasonal drying and wetting of soils results in the formation of desiccation cracks which are subsequently sealed. Although there are many problems due to soil spatial heterogeneity and variations through time, soil water physics approaches offer the only method for calculating soil water fluxes and for determining the resultant soil water pathways.

1.7 Macropores and their role in runoff generation.

Pores are the voids within a soil. They can be defined by their origin, size, morphology or function in terms of water movement (Yeh and Luxmoore, 1982). Through a discussion of Darcy's Law (Section 1.6) flow through the smaller pores of the matrix has been considered. This section focuses on water flow through macropores. Commonly, macropore diameters vary from 1 to 3 mm (Table 1.3). Broad categories of macropores include biologically created channels, such as decayed root and faunal tunnels, as well as the cracks and fissures along ped faces. The functions of pores are determined by their size: Skopp *et al.* (1981) and Chen and Wagenet (1992) argued that macropores should transmit water at a sufficiently slow rate to facilitate the extensive mixing of pore water. This definition requires the macropores to be relatively small. Other researchers such as Luxmoore (1981) and McDonnell (1988) claim that the pores must be capable of conducting water independently of the soil matrix, and therefore they define macropores as greater than 1 and 3 mm, respectively.

Table 1.3 shows that workers have offered various definitions of macropores and the flow of water in these 'voids'. Their different studies have resulted in apparently contradictory conclusions about the nature and consequences of macropore flow. However, further consideration of such studies reveals very different methodologies and study strategies (e.g. *in situ* dye tracing on field soils was used by Hallard and Armstrong (1992) (dyes) and Johnson *et al.* (1993) (isotopic and chemical), while McDonnell (1988) looked at the hillslope and stream and Jarvis *et al.* (1991) used chemical tracers on a soil core).

Table 1.3 Definitions of macropores and characteristics of macropore flow.

Authors.	**Observations.**
Bouma *et al.* (1983)	Macropore flow under unsaturated conditions gave no mixing
Hallard and Armstrong (1992)	Water travelled to mole channel as a discrete pulse, no mixing
Jarvis *et al.* (1991 b) Chen and Wagenet (1992)	Macropore flow more important under grass ley than under continous barley, but in both cases resulted in 'significant' mixing of new and old water Macropore flow occurred slowly enough to allow 'extensive' mixing of water
Luxmoore (1981)	Macropores defined by energy required to retain water (ψ < -0.3 bar)
Beven & Germann (1981) Sklash *et al.* (1986) Johnson *et al.* (1993)	} matrix must be at/near to } saturation Macropore flow was initiated only after saturation of Ap horizon
McDonnell (1988)	Macropores 1 to 3 mm diameter. Flow could occur independent of matrix

Several authors sought to define pore size by capillary potential, that is the energy required to retain water within the pores (Luxmoore, 1981; Luxmoore *et al.*, 1990):

- *micropores:* water is conducted through micropores by laminar flow (Darcy's Law, Eqn. 1.1) and is subject to capillary forces $\psi < -30$ kPa

- *mesopores:* water is subject to capillary forces of $-30 < \psi < -0.3$ kPa

- *macropores:* soil pores not subject to forces $\psi < -0.3$, which conduct water at greater rates than the mean pore velocity, i.e:

$$\langle q_{pore} \rangle = \frac{\langle q_{macroscopic\ flux} \rangle}{\langle \eta \rangle - dps} \qquad \text{[Eqn. 1.4]}$$

where:
- q = specific flux or velocity (dim.LT^{-1}).
- $\langle\,\rangle$ = mean.
- η = porosity of the control volume ($cm^3\ cm^{-3}$).
- dps = proportion of dead or inactive pore-space;

where: η-dps = η_e effective.

These definitions are rather arbitrary since many other factors influence the potential of a pore to promote macropore flow. Such factors include continuity, orientation, tortuosity, surface characteristics, connection to the surface and stability, as discussed below.

Macropore flow, sometimes referred to as bypass flow, can be defined as rapid subsurface flow through macropores (Table 1.3). According to Beven and Germann (1981) and Sklash *et al.* (1986) macropore flow is significant only under saturated or near-saturated conditions as pore walls or ped faces must be at, or close to, saturation to prevent the water from being retained by smaller pores. Flow into the larger pores is only initiated when the infiltration capacity of the surrounding matrix is exceeded or ponding occurs because pressure in the pore must be atmospheric. The macropores behave as pipes or tubes and are capable of transmitting water rapidly. Ideally, a relatively large amount of water, such as supplied by stemflow or that which collects in small topographic depressions, is needed to supply sufficient water to the macropores in order to equal the throughput. The effectiveness of macropores in water transmission is therefore a function of surface conditions, stable soil structure, initial moisture content and rate of water supply. Pore structure (connectivity, direction and tortuosity) is also important: unless macropores are continuous throughout the profile, that is they are linked in a network rather than dead-end, rapid water movement from the surface to depth is impossible.

A number of studies have shown that water flow in these large pores is partially independent of the hydraulic conditions in the matrix (McDonnell, 1988; Dowd *et al.*, 1991). The term 'preferential flow' is sometimes used to describe the same phenomenon in unsaturated soils because water movement will always occur in preference in the centre of the larger voids. As well as the macropore flow described for saturated soils, macropore-like behaviour has been noted in unsaturated soils in which no ponding was observed (Gaskin, 1988; McDonnell, 1988). A preferential flow or fingering mechanism has been suggested by researchers such as Tamai *et al.* (1987), Dowd *et al.* (1991) and Yeh (1972). When the soil is close to saturation water is retained by matric potential but many of the large pores are filled with water. These are capable of conducting flow many times greater than smaller pores. The importance of soil moisture is rather ambiguous because soils which are initially dry encourage preferential flow due to hydrophobic conditions and desiccation cracks (Edwards *et al.*, 1989). More research is required to describe soil physical characteristics which control the occurrence of preferential flow.

The traditional Darcian concept of water flow is not applicable to bypass flow for several reasons. Firstly, as Beven and Germann (1981) observed, the hydraulic gradient in a soil with non-capillary macroporosity is different from the hydraulic gradient in surrounding smaller pores. Secondly, turbulent rather than laminar flow occurs in macropores leading to the rejection of the traditional Darcian concept of water flow in soil.

Currently, there are several different approaches which are used with variable success. Macropores are similar to small pipes and therefore equations for describing flow in open channels and pipes can be applied (Atkinson, 1978). The flow through a hollow open cylinder is generally determined using the Hagen-Poiseuille Equation applied to gravity induced flow in a vertical channel (Childs, 1969; Atkinson, 1978):

$$q = (^{*}pgr^4)/8v \qquad \text{[Eqn 1.5]}$$

where:
- q = the flow rate.
- p = fluid density.
- g = acceleration due to gravity.
- r = the radius of the tube.
- v = the dynamic viscosity.

While this approach holds for one macropore the problem is how to deal with macropore flow at the field scale. Jury *et al.* (1982) and Jury and Roth (1990) developed a series of stochastic models claiming that the populations of macropores can be represented by their frequency distribution. They assume that processes in soil water systems and their outcome are uncertain and therefore only definable in statistical terms. These models include descriptions of spatially and temporally variable processes but have not been developed sufficiently for general application.

An alternative approach tries to average the behaviour of the macro and micropores and has led to the definition of the representative elementary volume (R.E.V.). This is the volume of soil, usually incorporating both macropores and micropores, at which if the soil is texturally uniform and homogeneous, the microscopic effects will average out. The R.E.V. is usually an area of 1 to 10 m^2, multiplied by depth determined by variability of macropore distribution (Beven and Germann, 1981). Water flux calculations based on Darcy's Law can now be computed as the concept of R.E.V.s takes account of both macropore (bypass) and micropore (matrix) flow. However, this assumption that soils are homogeneous, in which mean hydraulic conductivity can be defined, is highly questionable: if the soil exhibits heterogeneity or discontinuity in the soil physical properties at the larger scale, then the simple R.E.V. approach is no longer sufficient. Drained fields, such as the one used in this experiment, offer a practical and convenient way to observe subsurface flow response over a large area (Richard and Steenhuis, 1988). Andreini and Steenhuis (1990) comment that the spatial variability of macropore flow, which might be more pronounced at a smaller plot scale, will be integrated to give a spatially averaged response.

Despite the difficulties associated with soil heterogeneities and complexities associated with bypass and matrix flowpaths, soil water physics approaches based on Darcy's Law are the only method for quantifying moisture relations in the soil and resultant pathways. The soil water physics approach in this experiment is used at the field scale and results compared with a simple bypass model (Hutson and Wagenet, 1992). When these approaches are used in combination with chemical and isotopic tracing techniques, we have a powerful investigative tool for analysing soil water pathways.

1.7 Tracer Studies.

Many hydrological studies have recognised the value of using tracers for determining the direction and velocity of water movement through the soil (Reeves and Beven, 1990). At the catchment scale, tracer studies have been used to elucidate major hillslope hydrological processes and their role in influencing the behaviour of a storm runoff (Sklash *et al.*, 1986; Sklash, 1990). Within the context of this field drainage experiment, tracers may be used to advantage to identify and characterise active flowpaths. Other workers who have used tracers have reported the significance of of macropores (Table 1.4). Both isotopic and chemical tracers have shown the importance of macropore flow in soil cores (White *et al.*, 1984) while fluorescent dyes have been used in the laboratory to observe preferential flow in laboratory sand profiles (Omoti and Wild, 1979 a) and fissure flow in the field (Omoti and Wild, 1979 b; Smettem *et al.*, 1983; Hallard, 1988).

Table 1.4 Tracer studies which have demonstrated macropore flow.

Reference	Observation
Blake *et al.* (1973)	- Found rapid movement of solute through structured soil under natural conditions.
Coles and Trudgill (1985)	- Same observation as above on a weakly structured soil.
Gvirtzman and Margaritz (1986)	- Observed the bypassing of soil water by percolating rainfall; the immobile water representing 40-55% of the water content.
Shufort *et al.* (1977)	- Under continuously flooded field conditions, about 32% of the water was immobile.
Bowman and Rice (1986)	- Significant preferential flow occurred even in soils with little structure. Rate of downward flow was more than 60% greater than predicted in the absence of preferential flow.
Rice *et al.* (1986)	- Immobile water fraction estimated to be 80% of the soil water content.

Reeves and Beven (1990) identified seven requirements for an effective soil water tracer.

Tracers should:

(i) not be sorbed by the soil unless required to physically delimit water pathways;

(ii) be present in insignificant quantities in the water being studied;

(iii) be conservative, i.e. not significantly degraded chemically or biologically;

(iv) have low toxicity;

(v) be inexpensive;

(vi) be easily detected and quantified;

(vii) not alter the natural direction of movement of the water body under investigation.

Chemicals such as chloride, bromide, iodide and nitrate have been widely used for soil and ground water tracing applications. Many anion species are unsuitable for soil water tracing as they increase the surface activity and solid/water ratio (they are subject to increased sorption, Bowman, 1984; Reeves and Beven, 1990). Chloride and bromide, however, are particularly useful chemical tracers of water pathways in the soil because of low susceptibility to adsorption.

The most powerful technique for tracing soil water movement is the use of naturally occurring stable isotopes (McDonnell *et al.*, 1990). They have three advantages as tracers:

(i) stable isotopes are the constituent parts of natural water molecules and therefore they travel at the same speed;

(ii) natural isotopes are chemically conservative at ambient temperatures and, therefore, do not sorb strongly like most active tracers;

(iii) changes in concentration only occur by mixing, diffusion, dispersion or fractionation and therefore any alteration is due to a physical i.e. hydrological process.

There are conflicting views about whether incoming water retains its isotopic signature (McDonnell, 1988; Sklash, 1990) or whether mixing occurs so that identities of the various components of moving and stored water are lost. Therefore, studies of the changes in the isotopic signature should provide a valuable insight into soil water movement and solute transport processes.

1.8 Thesis Structure.

It has been demonstrated that there is a need to investigate the hydrological impact of mole drainage and to identify the importance of the various overland and subsurface flow pathways. As such, this investigation addresses key issues about water and solute transport in permanent grassland, which are applicable over a large area of Britain. Specifically, the study aims have been identified as:

(i) to develop a more complete understanding of how water moves in a grassland soil and to compare the hydrology of a drained and an undrained heavy clay field;

(ii) to determine the importance of macropore subsurface flow routes;

(iii) to investigate the movement of water in cracks and fissures using conservative tracers.

The introduction (Chapter 1) sets the aims of the thesis within the context of other drainage investigations and reviews some of the literature relating to subsurface water movement. Various types of drainage impact such as agricultural, hydrological, pedological and chemical are described. Several approaches to studying matrix and bypass flow are outlined and some of the problems of monitoring subsurface flow are considered.

In Chapter 2, information on the experimental site at North Wyke, Devon is presented. The grassland site is ideal for studying the impact of mole drainage as it incorporates drained fields and undrained control sites. The factors affecting soil hydrology are set into the wider regional context and the local climate, geology and soils are described.

Details of the equipment used and information acquired are presented in Chapter 3. The experimental set up in the field consisted of weirs equipped with continuously recording stage recorders to monitor the surface and subsurface flows, as appropriate. Measurement of the discharges provide much of the data on which the study is based. Instrumentation employed to elucidate flow pathways at the detailed plot scale are described. Pressure transducer-datalogging techniques were used to monitor soil water status continuously and a Time Domain Reflectometer determined volumetric soil moisture. Techniques for measuring the soil hydraulic properties and the soil water retention characteristics are also discussed.

Soil properties exert a fundamental influence on water movement. Chapter 4 presents data on the following properties: particle size, organic matter content and bulk density. Soil water retention and transmission properties were also measured. Hydraulic conductivities in the undrained and drained fields were determined to identify possible differences in hydrological behaviour between the sites.

The hydrological results for the three scales of investigation, namely field, plot and soil block are presented in Chapters 5, 6 and 7. The annual water balance is described in Chapter 5 to place the more detailed discharge results and hydrological responses in the context of the yearly budget. Hydrographs from the undrained and drained sites are analysed to identify the influence of the mole drain system. The hydrograph parameters and the statistical procedures used in this procedure are outlined in this chapter.

The hydrology of the two plots is discussed in Chapter 6. The soil water potentials in the plots are described for two periods: one in autumn during the wetting up of the soil; and the other in late winter. In this way seasonal trends are investigated. Elements of soil water physics, as presented in Chapter 1, are used to determine the importance of macropore flow: comparison of measured subsurface flow discharges and those predicted by Richards-Darcy's Equation, using LEACHM (Hutson and Wagenet, 1992) is presented. Parameters for the water flux model use data for the soil hydraulic properties as set out in Chapter 4.

Chapter 7 presents a detailed study of subsurface flow routes on the soil block. The movement of water in the soil and generation of mole drain runoff is traced using stable oxygen and chloride tracers. The soil moisture status was characterised using the transducer and Time Domain Reflectometer technology, and the soil water samples were collected for tracer analysis, using the techniques described in Chapter 3. Two sets of calculations are described which were performed to determine first, the water velocity in the block, and thus the relative importance of bypass flow, and second, how much mixing was occurring in the soil peds between new and existing water. The final chapter presents a synthesis of the main findings of the investigation together with a discussion of the wider implications.

CHAPTER 2 RESEARCH SITE.

2.1 Introduction.

The research site, Rowden Moor, is located at the Agriculture and Food Research Council (A.F.R.C.) Institute of Grassland and Environmental Research (I.G.E.R.)[1], North Wyke Research Station. North Wyke is in central Devon, 7 km north of Dartmoor at grid reference SX 660984 (Figure 2.1). This is an area of dissected plateau at an altitude of 130 to 158 m O.D., with slopes ranging between 2 and 5°. The Okement and Taw Rivers drain the area, flowing north from Dartmoor.

The experimental site was established in 1982 (Section 2.6) and offers a valuable resource for the investigation of the hydrological characteristics of heavy agricultural soils (Section 2.5) in a high rainfall area (Section 2.2).

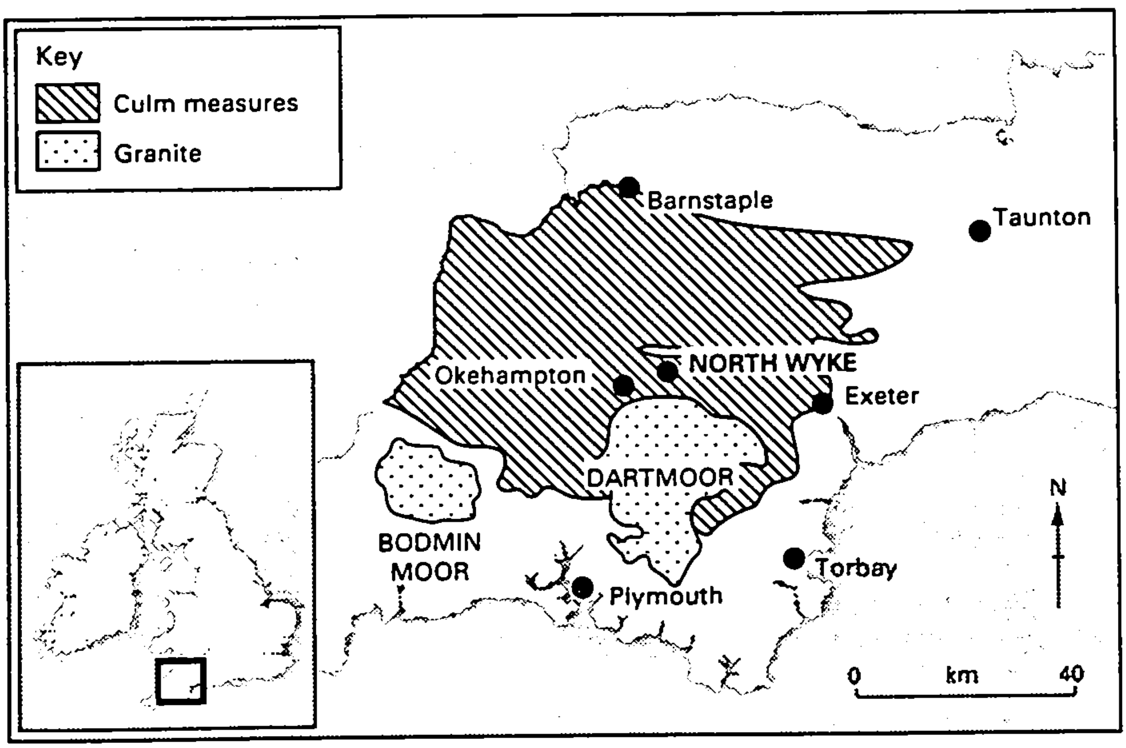


Figure 2.1 Site location and geology.

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[1] This was the Permanent Grassland Division of the Grassland Research Institute (G.R.I.). In 1986, it became the Permanent Grassland Division of the Animal and Grassland Research Insitute (A.G.R.I), which was then succeeded, first by the Institute of Grassland and Animal Production (I.G.A.P.), and finally by the Institute of Grassland and Environmental Research (I.G.E.R.).

2.2 Climate.

The region has an oceanic climate. Temperatures are mild, monthly averages are 15.3° C in the summer and only fall to 4.3° C in the winter months (Table 2.1). Yearly rainfall totals have a wide range, with values of 781.6 mm and 1453.2 mm recorded at North Wyke in 1964 and 1960, respectively (North Wyke Meteorological Records, Unpubd.) but mean annual rainfall is high at 1034.7 mm, occurring over an average of 201 rain days (Table 2.1). The dominant rainfall characteristic is the marked October to February maximum (Figure 2.2). The prevailing south-westerly winds and associated depressional activity from the Atlantic are largely responsible for the nature of the climate.

Table 2.1 Meteorological means, North Wyke (1961 to 1990) (Tyson, 1992)

Month	Rainfall (mm)*1	Rain days	P.Et. (mm)*2	Mean air temp (°C)*3	Mean sunshine (hrs d^{-1})
Jan	129.1	22	0	4.5	1.6
Feb	97.2	16	8	4.3	2.3
Mar	88.8	18	32	5.7	3.3
Apr	62.1	15	49	7.7	5.1
May	61.1	15	75	10.5	6.2
June	59.8	13	84	13.5	6.4
July	52.9	13	83	15.3	6.2
Aug	67.8	15	70	15.2	5.4
Sept	75.6	15	42	13.3	4.3
Oct	100.8	18	20	10.8	3.0
Nov	113.6	20	4	7.2	2.2
Dec	125.9	21	0	5.6	1.6
Total	1034.7	201	467	9.5	4.0

*1 calculated from a 30-year period (1961-1990; February to September 1981 missing) for North Wyke.

*2 M.A.F.F. 30-year average estimates of potential transpiration for the area including North Wyke (M.A.F.F., 1976).

*3 calculated from a 30-year period (1961-1990) for North Wyke.

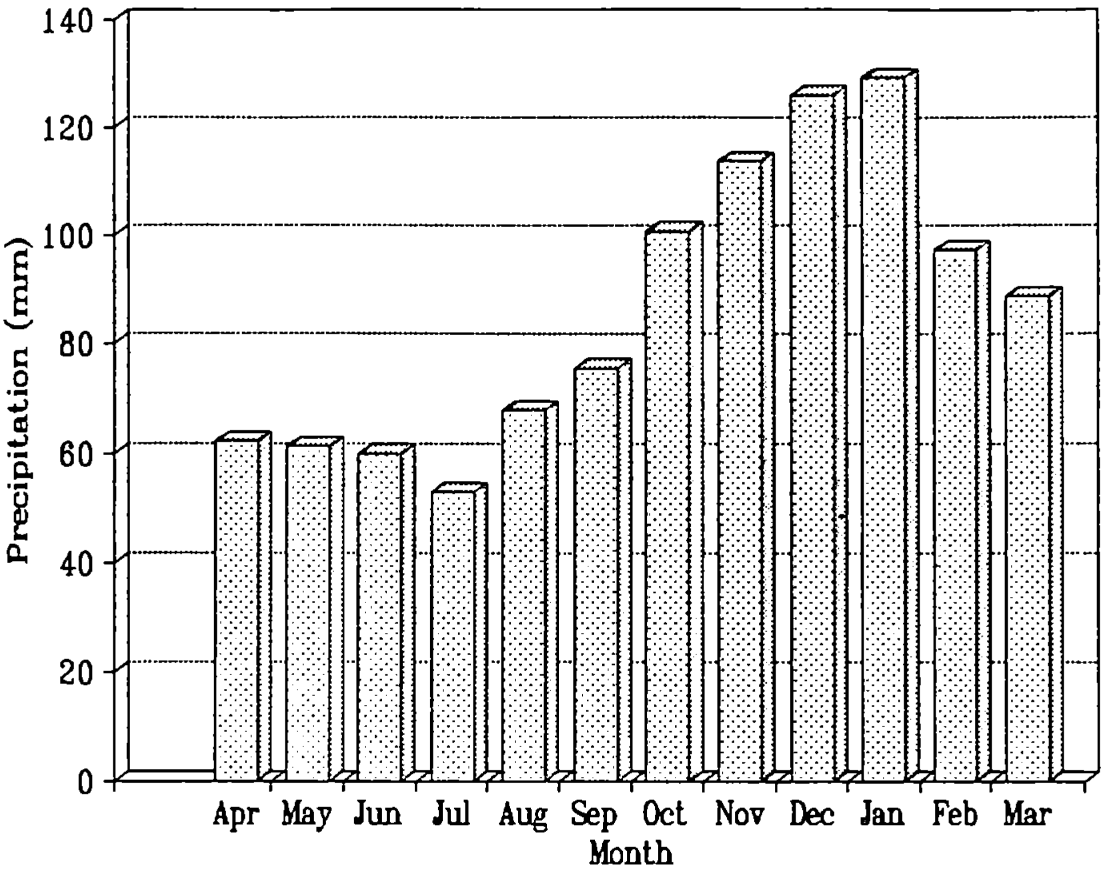

Figure 2.2 Mean precipitation at North Wyke, Devon (1961 - 1990).

Potential transpiration was estimated for Agro Met area 43 N, which includes North Wyke, by Smith (1976) using the procedure outlined by the Ministry of Agriculture (M.A.F.F., 1976 b). Regional results highlight low winter potential transpiration rates, with obvious implications for the water balance. The evapotranspiration estimates are substantiated by mean daily sunshine figures, and mean monthly air temperatures for each month at North Wyke (Table 2.1).

2.3 Relief.

The experimental site at Rowden Moor is situated on the eastern flank of a ridge which is centred on Beacon Cross. From the summit of the ridge, at just over 180 m O.D., the land slopes away at angles of between 2 and 5° (Figure 2.3).

The highest point of the site at 158 m is situated on the western boundary. The main slope from east to west displays a notable convexity in form as it drops to its lowest point on the eastern tip of the field site at 130 m. However, the angle and direction of slope vary across the site with probable implications for slope hydrology. Parkinson and Reid (1985; 1986), have shown that drainflow response characteristics can be significantly influenced by topographic factors.

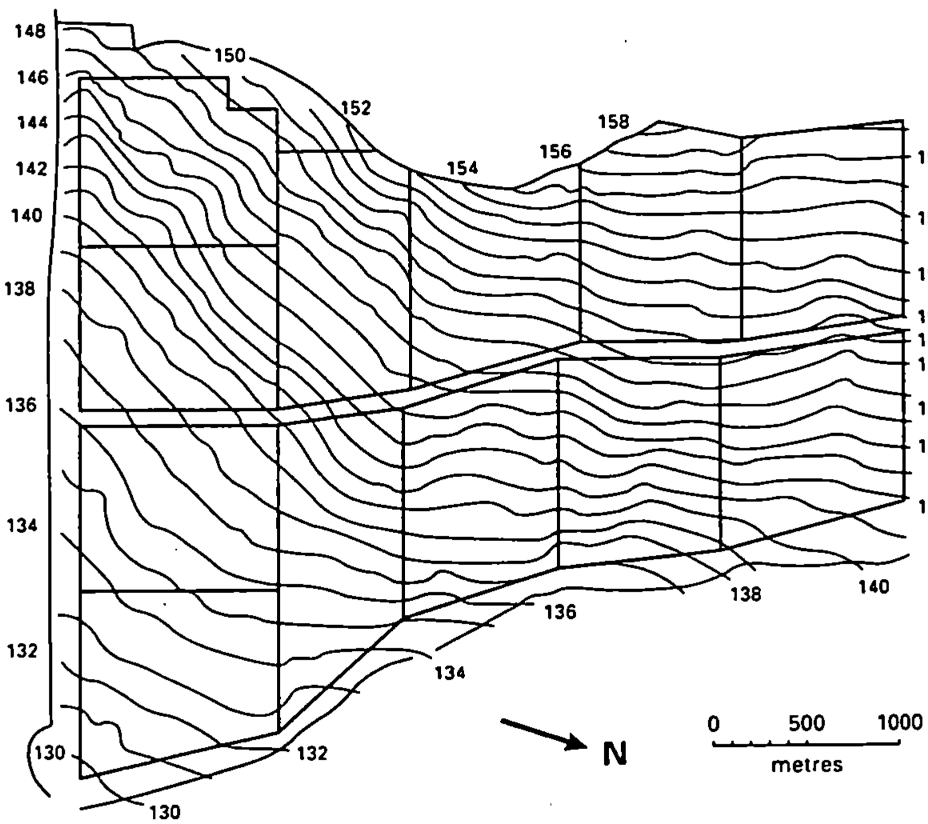

Figure 2.3 Topographic survey of the field. Contours at 1 m intervals. (Data source: Armstrong *et al.* 1984).

2.4 Geology.

The Carboniferous Crackington Formation, also known as the Culm Measures, underlies the area. The Culm Measures extend under 100 000 ha of Devon and Cornwall (Figure 2.1). The formation is composed of clay shales and thin sandstone beds. Under waterlogged conditions the grey black shales break down to buff illitic clays. Although the sandstone beds are rarely greater than 0.3 to 0.4 m in depth, they constitute one quarter of the sequence.

Frost working and solifluction led to the formation and deposition of locally derived rubbly drift or head material in the Pleistocene. Such deposits, of between 2.0 and 4.0 m depth, are found over the *in situ* Carboniferous rocks, and can lead to stony subsoils (Harrod, 1981).

The fractured shale layer found at depth in this area is hydrologically significant as it is known to carry ground water under pressure. Isolated, deep drains are often installed to remove this water when it approaches the surface. However, this does little to alleviate the problem of high water contents caused by high rainfall.

2.5 Soils.

2.5.1 Introduction.

The Soil Survey of England and Wales conducted an investigation of the soils at North Wyke and Rowden. Soils have been derived from clay shales which were weathered in head of variable thickness (Section 2.4). They are classified as belonging to the Hallsworth series, and were described in detail by Harrod (1981). Soils in this series are very similar, being distinguished largely by the depth at which the shaly head is encountered. Soil development is strongly controlled by local hydrology (Findlay *et al.*, 1984). Clayden (1964) explained the distribution pattern of soils over the Carboniferous shale as a function of faster weathering to clay on relatively level, less well drained ground. The Tedburn soils, situated on gentler slopes, are the wettest in a hydrologically and/or physiographically related sequence. Here, release of clay from Carboniferous shales is more intense, resulting in coarsely structured impermeable subsoil (Findlay *et al.*, 1984).

The Hallsworth series, a clayey pelostagnogley, is characterised by a shallow Ag horizon (0.24 m) with a high organic matter content (9.6 %). This is a grey-brown clay loam, silty clay loam or clay which is strongly mottled. Manganese accumulation below 0.2 m lies above a blue and ochreous mottled clayey Bg horizon (0.41 to 0.73 m). This horizon has a high bulk density and a low hydraulic conductivity (Section 4.3). A soil profile description produced by F.D.E.U.[2] (Table 2.2) shows the clay subsoil to a nominal depth of 1.2 m.

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2 The A.D.A.S. Field Drainage Experimental Unit became the Soil Water Research Centre (S.W.R.C.) in 1992.

Table 2.2 Soil profile description - Rowden Moor experimental site (adapted from Harrod, 1981).

Depth range	Horizon description
0-0.24 m	**Ag.** Brown (10YR 4/5), clay loam with few fine ochreous mottles in root channels; well developed fine/medium angular, blocky structure becoming coarser below 0.15 m; common fine pores; common fine and medium fissures; highly permeable; abundant fine, fibrous roots; common small stones forming a pan-like layer at 0.2 m.
	Distinct boundary to:
0.24-0.66 m	**Bg.** Light grey (2.5Y 7/2) clay, with abundant small and distinct yellowish red (5YR 5/8) mottles; well developed medium/large prismatic structures; common fine pores; few large fissures; slowly permeable; common fine fibrous roots found only in fissures; few small and medium stones.
	Merging boundary to:
0.66-1.2 m	**Cg.** Light grey (5YR 7/1) clay, with abundant small and medium reddish yellow (5YR 6/6) mottles; very poorly developed coarse prismatic structures; few small and medium fissures; no roots; common small and medium stones.

2.5.2 Soil structure and texture.

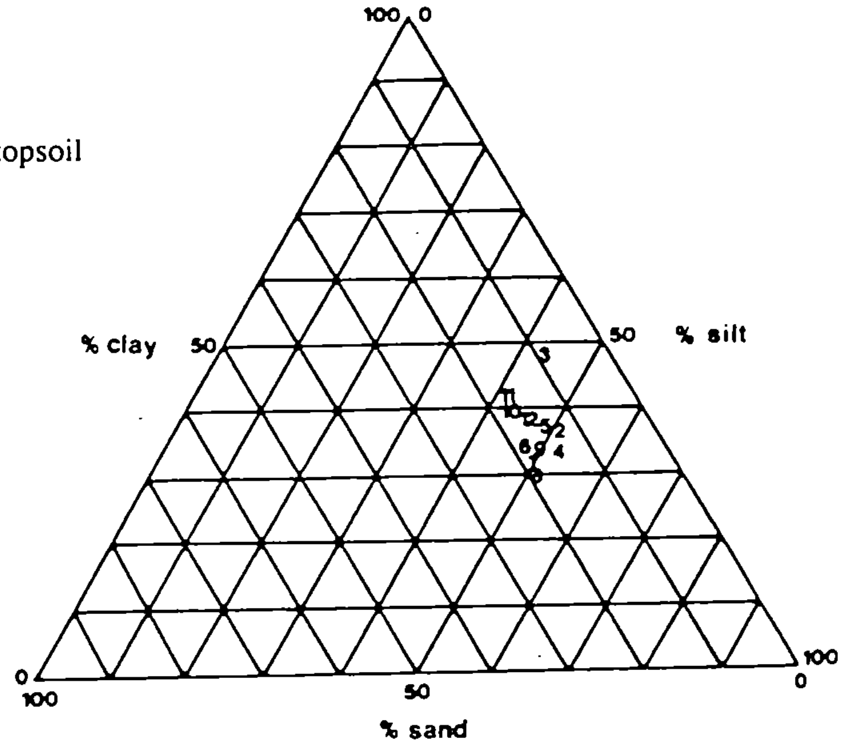
The prismatic structure, typical of the B horizon, deteriorates below 0.7 m with decreasing clay content and increasing stoniness. In the summer, prismatic structure and the presence of smectite ensure that the soil cracks as moisture content decreases. Under natural drainage the soil has limited shrink-swell capacity. However, artificial drainage induces seasonal drying, shrinkage and cracking. By using observation pits, Hallard (1988) noted changes in structural cracks throughout the wetting-up period. The faces of coarse structural units provide a pathway for aeration, water transmission and roots are found concentrated along them.

Soil characteristics over the site are very uniform but differences in erosional history, hydraulic characteristics and waterlogging result in variable topsoil depths. The relative homogeneity of the soil is illustrated by the clustering of data points in the particle size diagrams produced for topsoil and subsoil samples by S.W.R.C. (Figure 2.4). The classification is based on the British Standards Institution particle size grades. It can be seen that topsoil particle size characteristics are more uniform. The high clay content, of 48 %, is probably partially responsible for surface wetness problems and high moisture retention characteristics indicative of this soil type. Despite the higher degree of variability in the subsoil, all samples are in the silty clay/clay class.

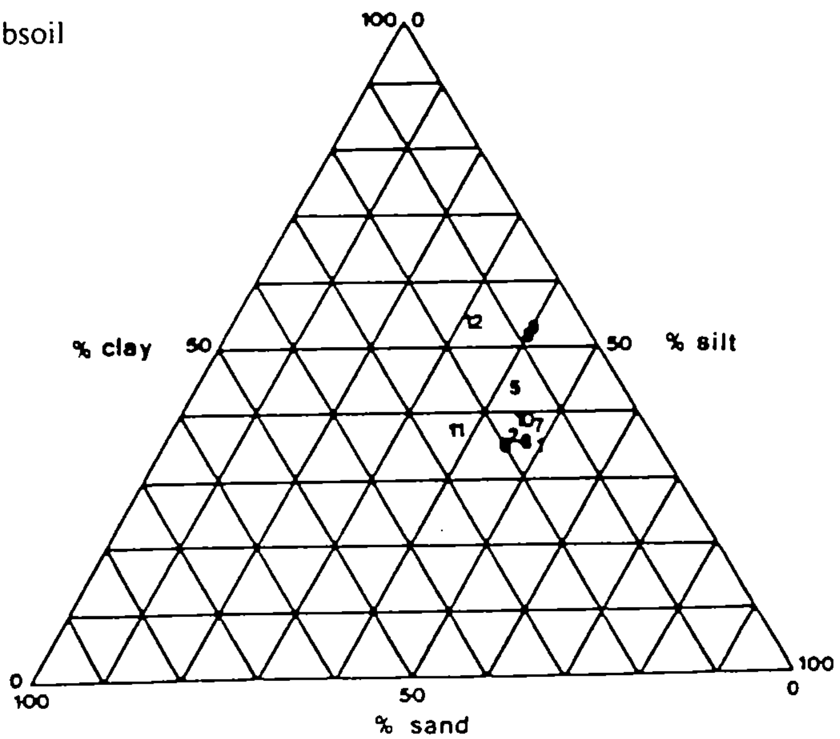
Figure 2.4 Particle size characteristics for Rowden Moor

(F.D.E.U. cited by Hallard, 1988).

a) topsoil



b) subsoil



2.5.3 Soil hydrology.

The hydrology of soils in the area is significantly affected by the presence of a fractured shale layer at depth. This layer, locally known as shillet, carries groundwater under pressure. As this may lead to groundwater in upper soil horizons, a widespread local drainage practice is to install deep, isolated drains. However, such drains do little or nothing to alleviate surface water problems which are caused by high rainfall and hydraulic characteristics of upper soil horizons. At Rowden a pre-drainage survey identified a number of 'wet spots' which were then the focus of appropriate drainage (Hallard, 1988). None of these wet spots were on the plots considered by this study (Section 3.1.2). Further, on this site the shale layer does not approach the main drain depth at 0.9 m.

The dense, highly impermeable subsoils with slow hydraulic conductivity (Section 2.5.2) are waterlogged for much of the year. Drains normally function from October until early April discharging a minimum of 225 mm and sometimes > 600 mm of runoff (Tyson, 1990).

During certain years the insides of subsoil peds never become completely wetted. However, the retained water capacity (water at field capacity) is well above the surface soils plastic limit such that even with effective drainage, a period of drying is still required before stocking or trafficking to avoid structural damage (or poaching) to upper horizons. Findlay *et al.* (1984) calculated profile available water based on the technique developed by Hodgson (1974). They found that much of the available water is held at relatively high tensions largely due to soil texture (Section 2.5.2).

2.6 Land management.

The research is part of a larger experiment concerned with agricultural and environmental management in such areas with high rainfall and heavy clay soils. The site was established in 1982 by the Grassland Research Institute, now I.G.E.R., and S.W.R.C. Its position within the farm layout is illustrated in Figure 2.5. Fourteen 1 ha fields were hydrologically bounded on permanent grassland (Figure 2.6). Seven fields were drained using mole drains at 2 m spacing and 0.55 m depth together with lateral collector drains 40 m apart and at 0.85 m depth (Figure 2.7). Both the undrained and drained fields have been instrumented with continuously recording stage recorders to monitor surface and subsurface flow, as appropriate (Section 3.2.4).

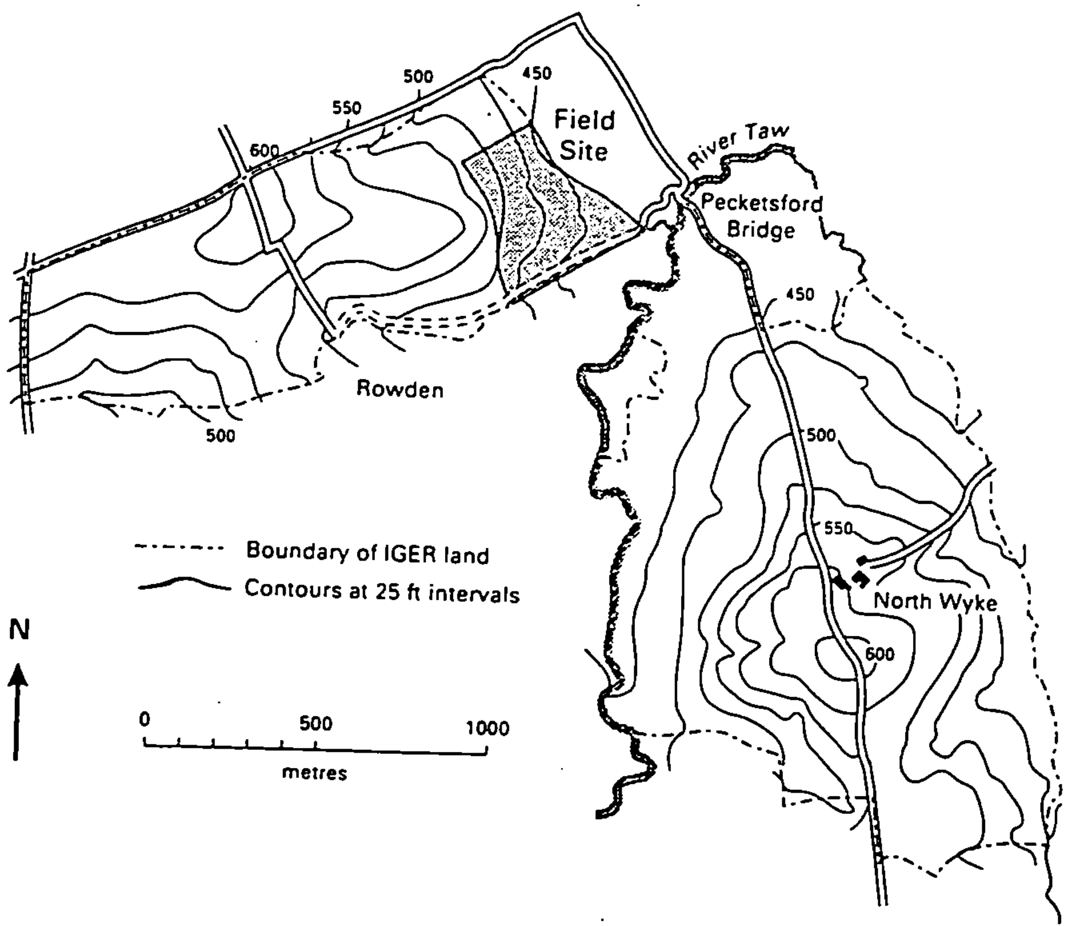


Figure 2.5 Farm plan - location of field site.

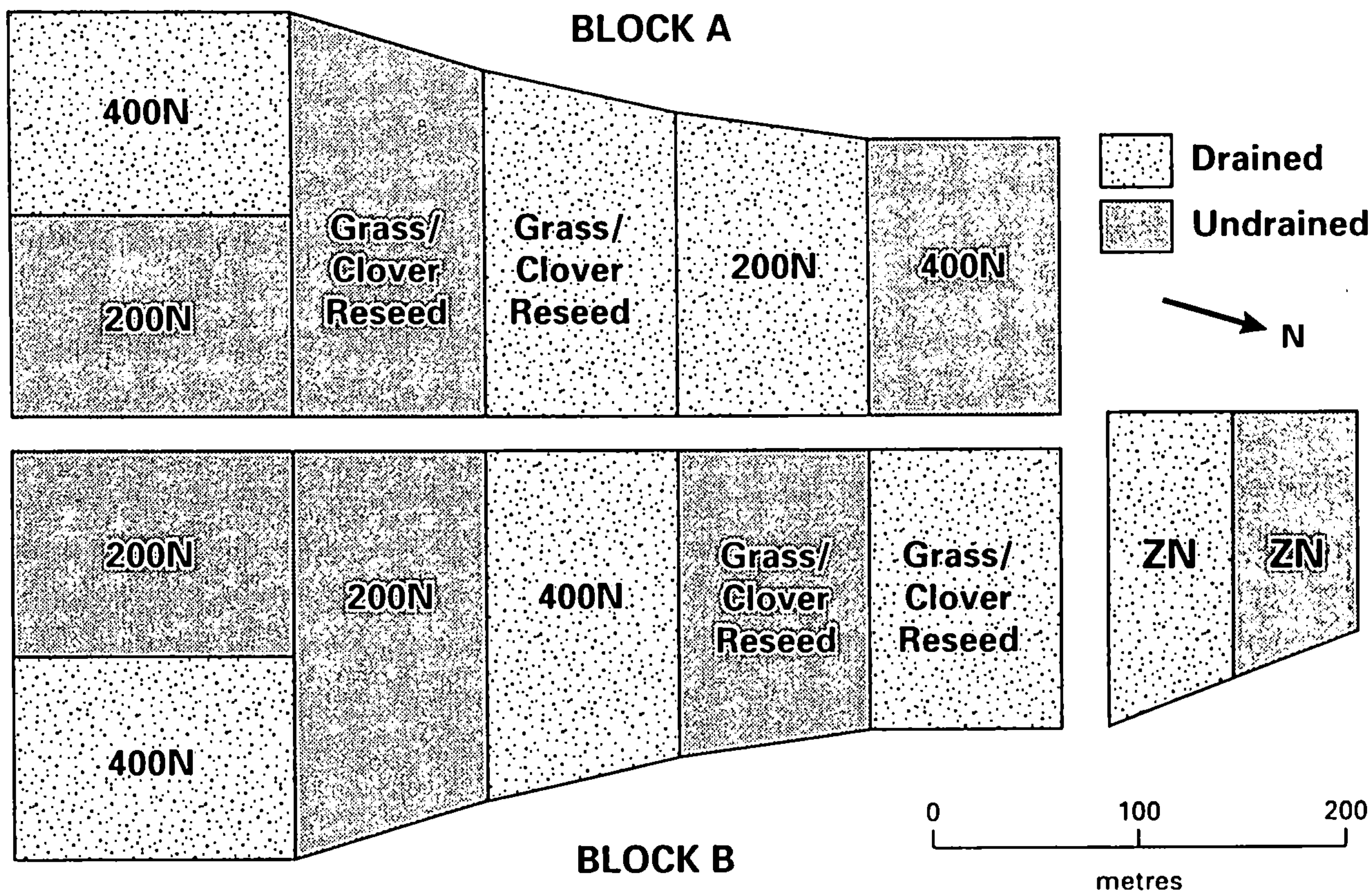


Figure 2.6 Drainage economics experiment layout. (200N, 400N = kg of N added per year, ZN = no fertilizer added)

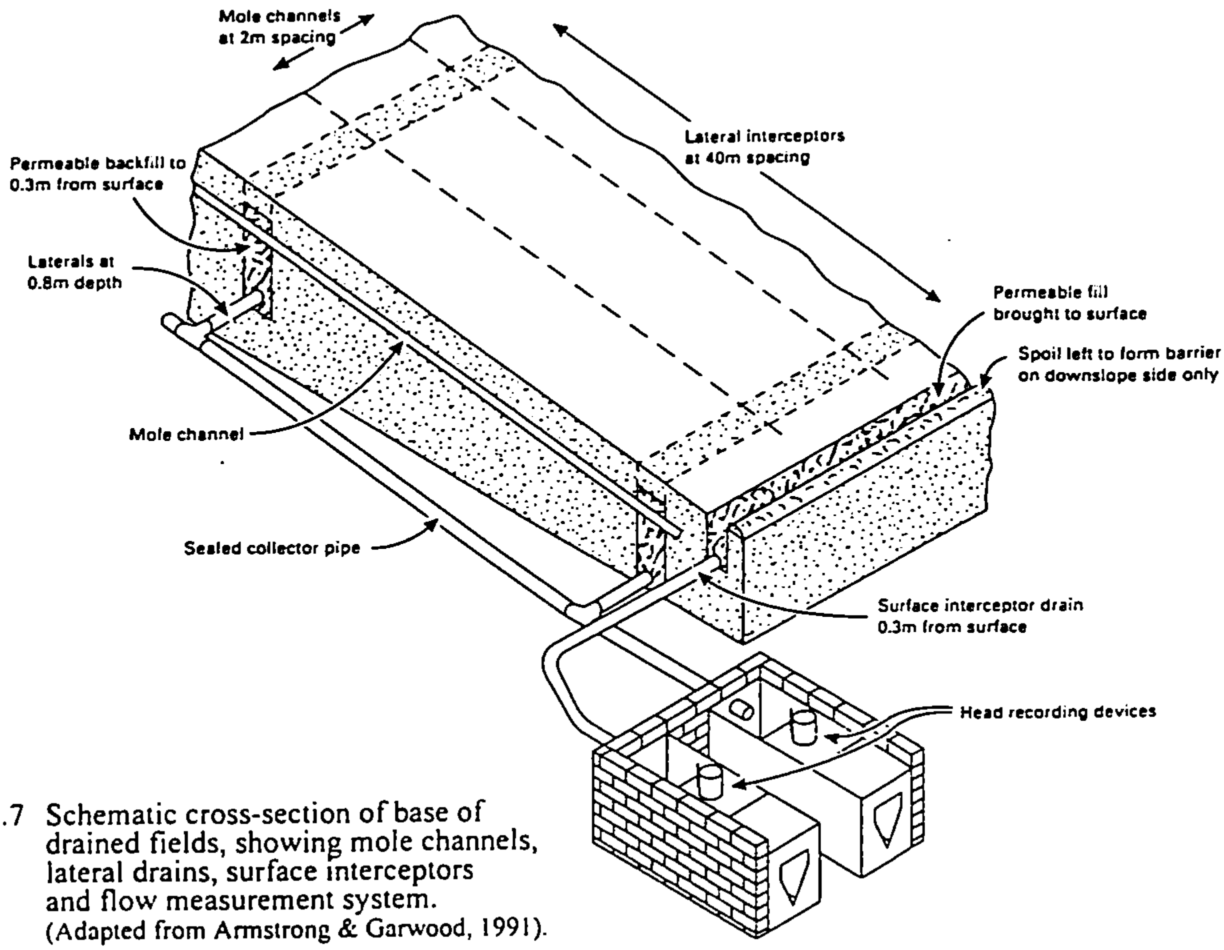


Figure 2.7 Schematic cross-section of base of drained fields, showing mole channels, lateral drains, surface interceptors and flow measurement system.
(Adapted from Armstrong & Garwood, 1991).

Much of the land use on this soil is permanent grassland, frequently under a rushy sward. Rowden Moor has been under permanent pasture since before 1942. Sward composition at the commencement of the experiment is described by (Wilkins, 1982). Vegetation was dominated by the following species: *Lolium perenne* (17 %), *Agrostis spp.* (40 %), *Holcus lanatus* (17 %), *Juncus effusus* (9 %) and *Trifolium repens* (5 %).

This soil can produce large yields of grass. However, Findlay *et al.* (1984), state that without drainage, these soils are usually waterlogged for long periods in the growing season because of their slowly permeable subsoil. With field capacity ranging between 175 and 250 days management, grazing and conservation are confined to the summer months due to poaching. Drainage can achieve some improvement according to Trafford (1971) and has been observed to lower the water table and reduce surface wetness, thereby lengthening the grazing season on this site (Armstrong and Garwood, 1991). The Langabeare and East Worlington experimental sites have been subject to the same treatment on similar soils (Trafford, 1971) with similar results in terms of better soil surface conditions (Harrod, 1981).

CHAPTER 3: METHODS AND INSTRUMENTATION.

3.1 Introduction.

3.1.1 Study strategy.

The primary aim of this research was to investigate the hydrological impact of drainage at a range of scales (Section 1.4). A major objective was to characterise the hydrologies of undrained and drained fields (Section 3.1.2) by monitoring field discharge (Section 3.2). The second aim was to study subsurface flow pathways on instrumented plots to aid in the explanation of response at the gross scale (Section 3.1.3). Field observations were made over one drainage season, from October 1990 to April 1991. To examine mechanisms of subsurface flow in more detail a lysimeter was established on which a tracer experiment was conducted (Section 3.1.4) to provide information on the relative importance of macropores and the matrix in conveying soil water.

In this thesis, field, plot and lysimeter are defined with reference to scale as follows:

- field: a one hectare hydrologically bounded area (Section 2.6);
- plot: a smaller instrumented area (4.0 x 2.5 m) within a field;
- lysimeter: an isolated plot (4.0 x 2.4 x 0.8 m), established for a tracer experiment.

3.1.2 Field study.

The following sections will introduce the rationale for the selection of fields. In addition, the instrumentation required to calculate hydrological budgets and to characterise field discharge is outlined. Field and laboratory techniques used to determine soil hydraulic properties are also described.

The fields were selected on the basis of several criteria:

(i) they had not been subject to reseeding, so possible hydrological impacts associated with ploughing are avoided;

(ii) they had not been subject to high fertiliser applications (> 200 kg N ha^{-1} a^{-1}) as this was viewed as increasingly unrepresentative in the economic and political climate of deintensification of agricultural practices (which is important if conclusions are considered alongside leaching data);

(iii) they permitted an investigation of the impacts of agricultural drainage for field hydrology.

These fields, under permanent grassland, had been grazed throughout the drainage economics experiment (Section 2.6). Both had been subject to annual applications of lime, phosphate and 200 kg of N. Their location is shown in Figure 3.1. A quantitative approach to the study of water movement in undrained and drained fields requires measurement of both timing and amounts of hydrological inputs and outputs. This enables the calculation of runoff coefficients, determination of water flow characteristics, such as flow duration curves, as well as quantification of water balances. Such data are required to set the results of this limited field catchment area in a wider spatial and temporal context. Budgets were calculated from precipitation inputs (Section 3.2.2), evapotranspiration losses (Section 3.2.3) and outflow determined as stated in Section 3.2.4. Hydrological characteristics and budgets are presented in Chapter 5.

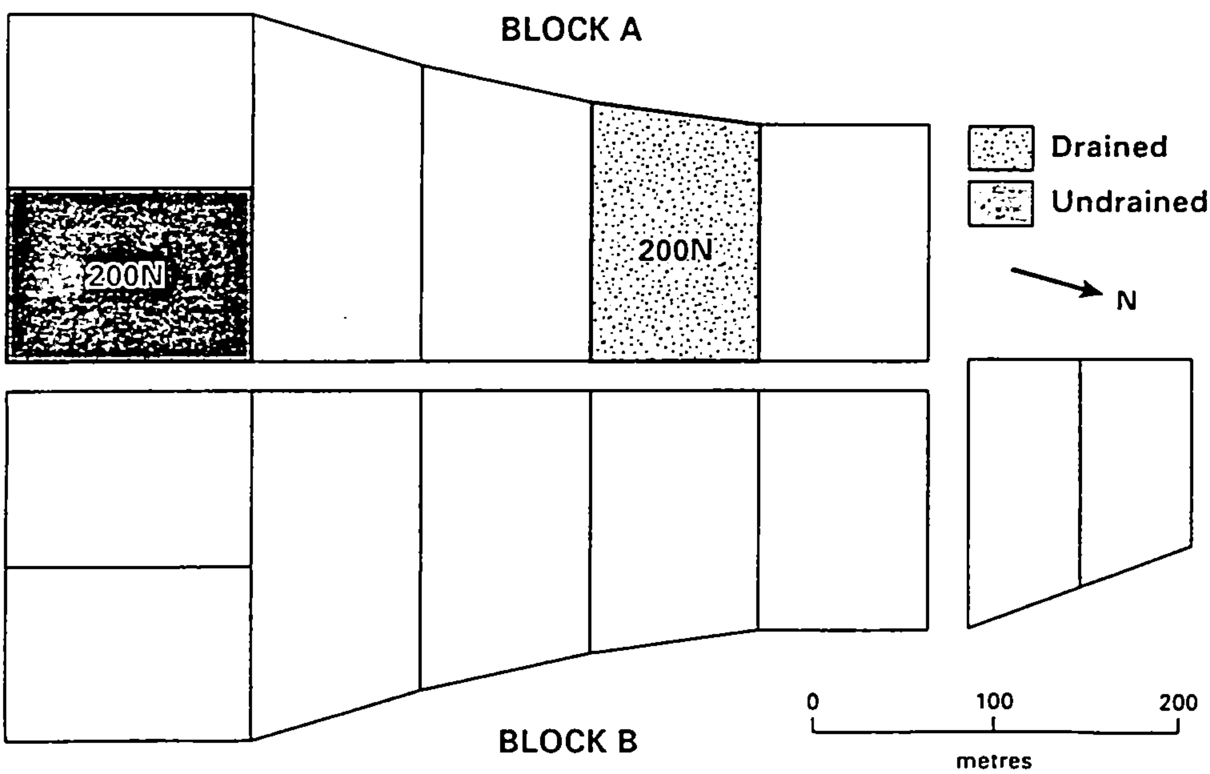


Figure 3.1 Fields selected for instrumentation.

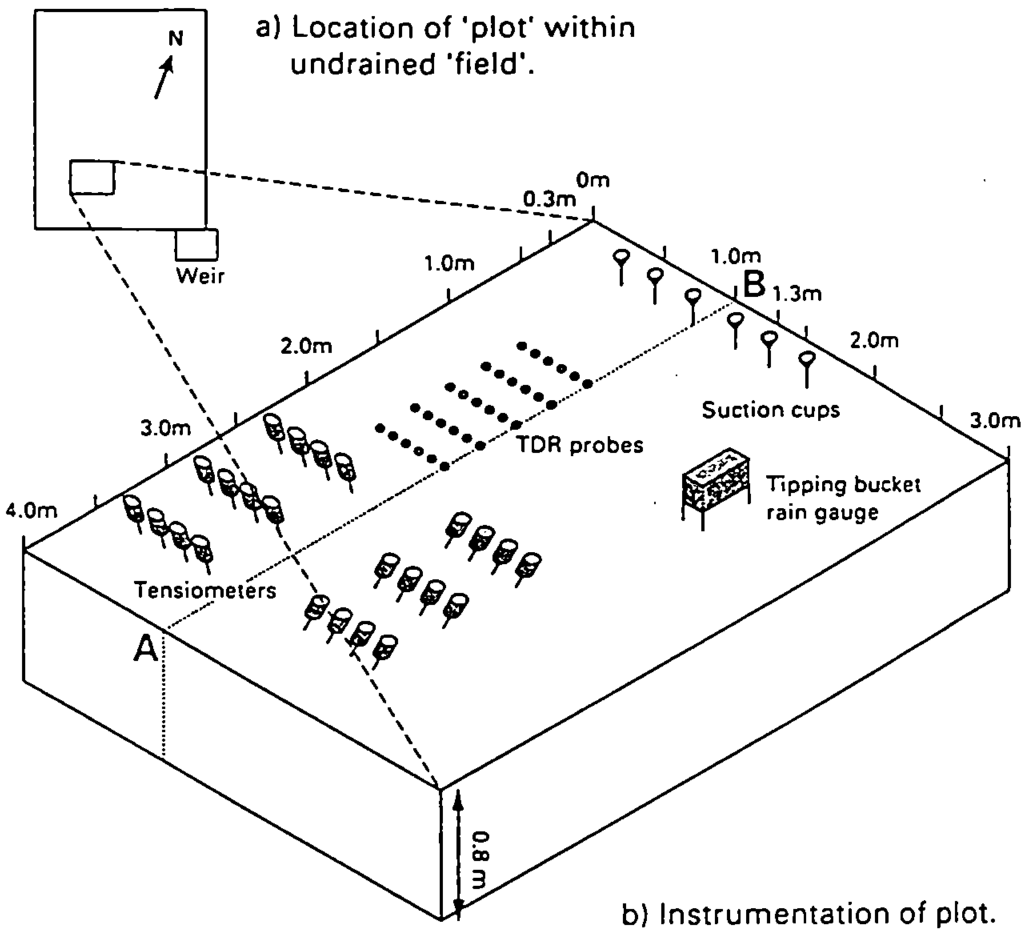
Soil hydraulic properties are fundamental to the control of soil water movement (Bathurst, 1986 b). Much attention must be given to their description to facilitate more detailed analysis of flow pathways. Their quantification is required for solving the equations which describe soil water flow. The hydraulic properties of the soil, and physical properties related to them, were characterised for both fields and the lysimeter (Section 3.3). Physical soil properties which have an effect upon the movement of water include soil bulk density, porosity, soil moisture content, potential gradient, water retention capacity and hydraulic conductivity. The soil characteristics and techniques adopted are summarised in Table 3.1 and results obtained are presented in Chapter 4.

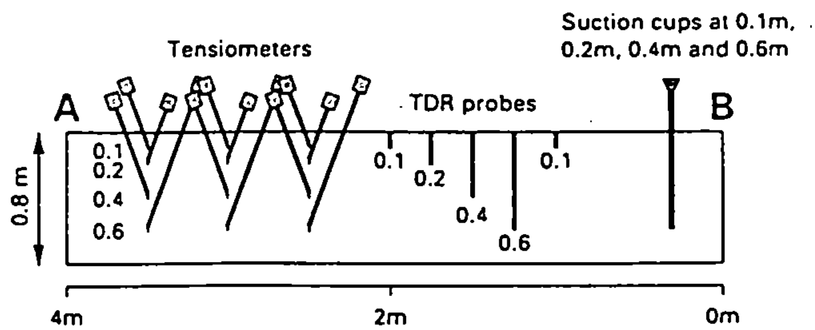
Table 3.1 Hydraulic and related physical soil properties investigated.

Characteristic	Technique
Soil water characteristic curve (Section 3.3.2)	sand tension table, pressure plate
Bulk density (Section 3.3.2)	undisturbed core
Structure (Section 3.3.2)	water release curves bulk density and porosity
Saturated hydraulic conductivity (K_{sat}) (Section 3.3.3)	ring permeameter
Particle size distribution (Section 3.3.4)	wet sieving and sedimentograph
Soil strength (Section 3.3.5)	Bush penetrometer
Organic matter content (Section 3.3.6)	loss on ignition
Matric potential (Section 3.3.7)	tensiometer-transducers
Soil moisture content (Section 3.3.8)	T.D.R. and volumetric cores

3.1.3 Plot Study.

To examine variability of hydrological properties within fields, two small (4 x 2.5 m) intensively instrumented plots were established, one on the undrained field and another on the drained field (Figures 3.2 and 3.3). Soil moisture status (Section 3.3.7) and content (Section 3.3.8) were monitored to determine suction and moisture gradients to infer pathways and processes of water movement.





c) Cross-section of Instrumentation.

Figure 3.2 Undrained plot.

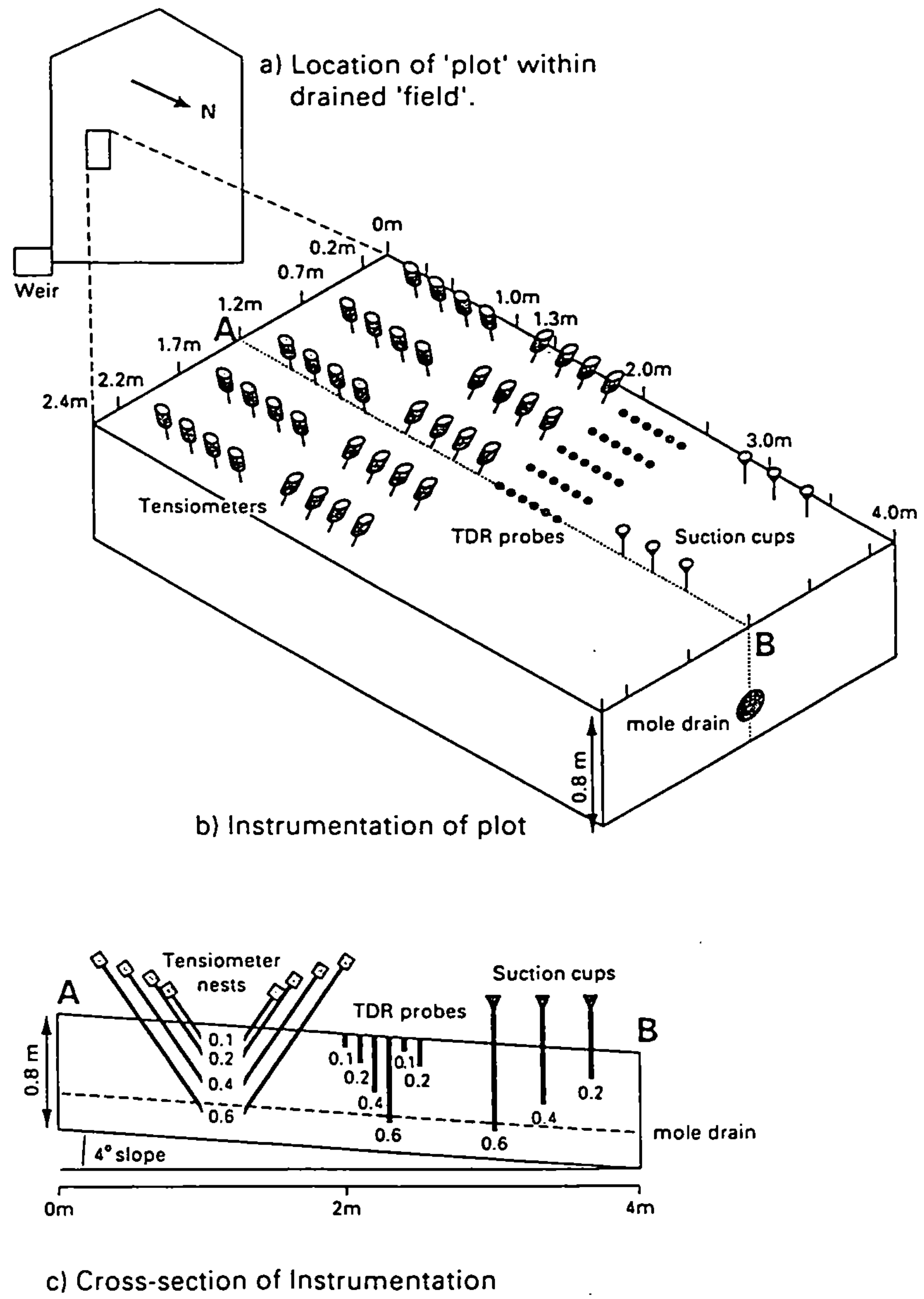


Figure 3.3 Drained plot.

Instrumentation on the drained plot was located along transects, centred on a mole drain and the associated crack system (Figure 3.3). Such transects allowed observation of spatial variation in hydrological characteristics relative to the mole drain (mid mole to mid mole). This pattern of instrumentation was replicated on the undrained plot (Figure 3.2).

3.1.4 Lysimeter Study.

To study flow pathways in more detail, a lysimeter experiment was established: application and monitoring of tracer movement was undertaken. This experiment comprised a 10 m^2 soil block located on the periphery of the field site (Figure 3.4). A mole drain was installed by drawing a mole plough (Figure 1.1) by a tractor at 0.55 m below the soil surface. The drain was established in May, 1991, during a period when soil moisture content was deemed suitable for such operations, conforming to the requirements of appropriate soil plasticity and soil strength, as identified by Spoor *et al.* (1982) and Leeds-Harrison *et al.* (1982). A mechanical digger was then employed to isolate the block by trenching around an area 2.4 m wide, 4.0 m long and 0.8 m deep, centred on the newly installed mole drain and also including a 5 year-old mole drain. The trenches, walls and floor were covered with marine plywood which had been treated with preservative, and were supported by vertical and horizontal struts. The top of the boarding was above the soil surface, allowing a clay seal to be placed around the upper edges of the block. The bottom of the plywood boarding was hammered into the clayey subsoil and a clay seal was placed at the base. Corner seals were further reinforced using silicon sealant. Two holes were made in the board on the lower side of the block to provide an outlet for mole drainage. A 0.1 m length of guttering was inserted into each drain in order to direct the drainflow out of the block. A drain was laid from the lower, outer trench wall to the weir of the adjacent field in order to dispose of drainflow from the block. Compaction by machinery and personnel was avoided by careful planning of the order of work and use of duckboards.

The lysimeter was intensively instrumented, in a similar manner to the plots enabling control of hydrological inputs, application of tracer and accurate observation of drainflow generation (Figure 3.5).

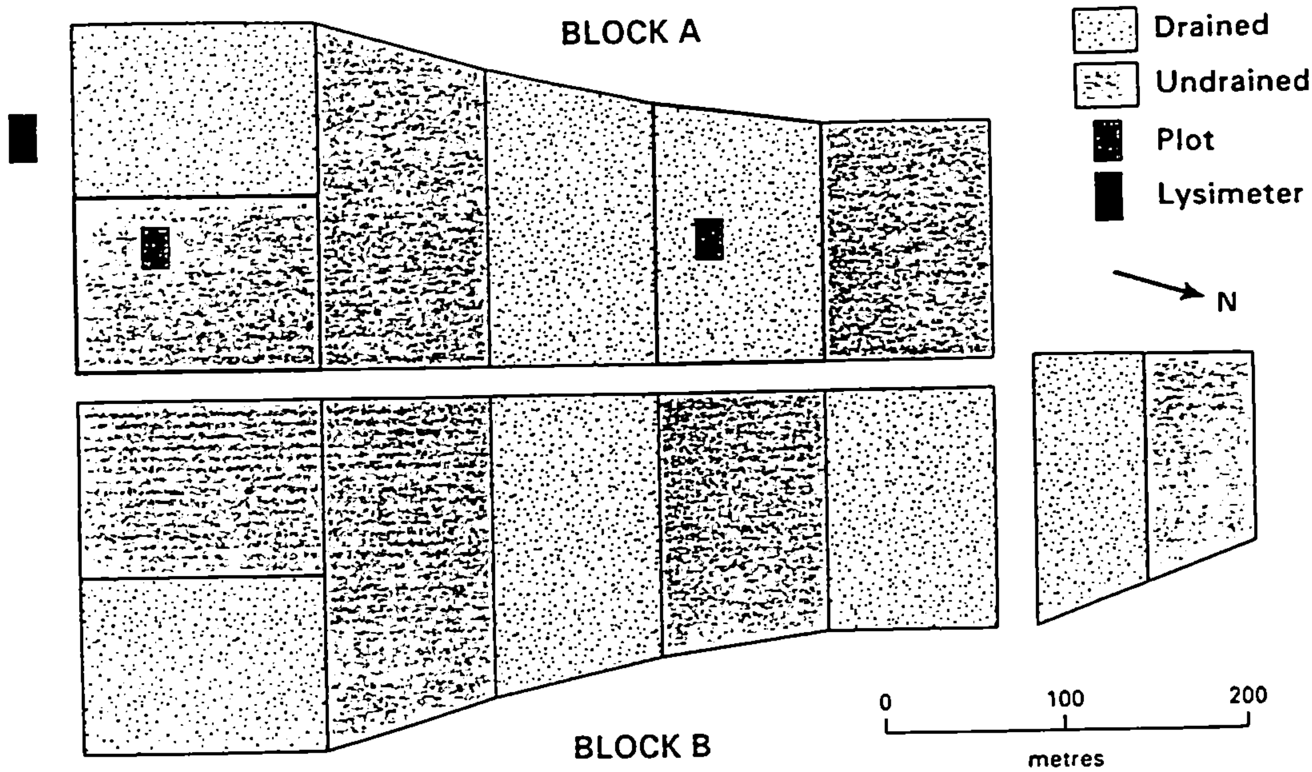


Figure 3.4 Location of plots and lysimeter.

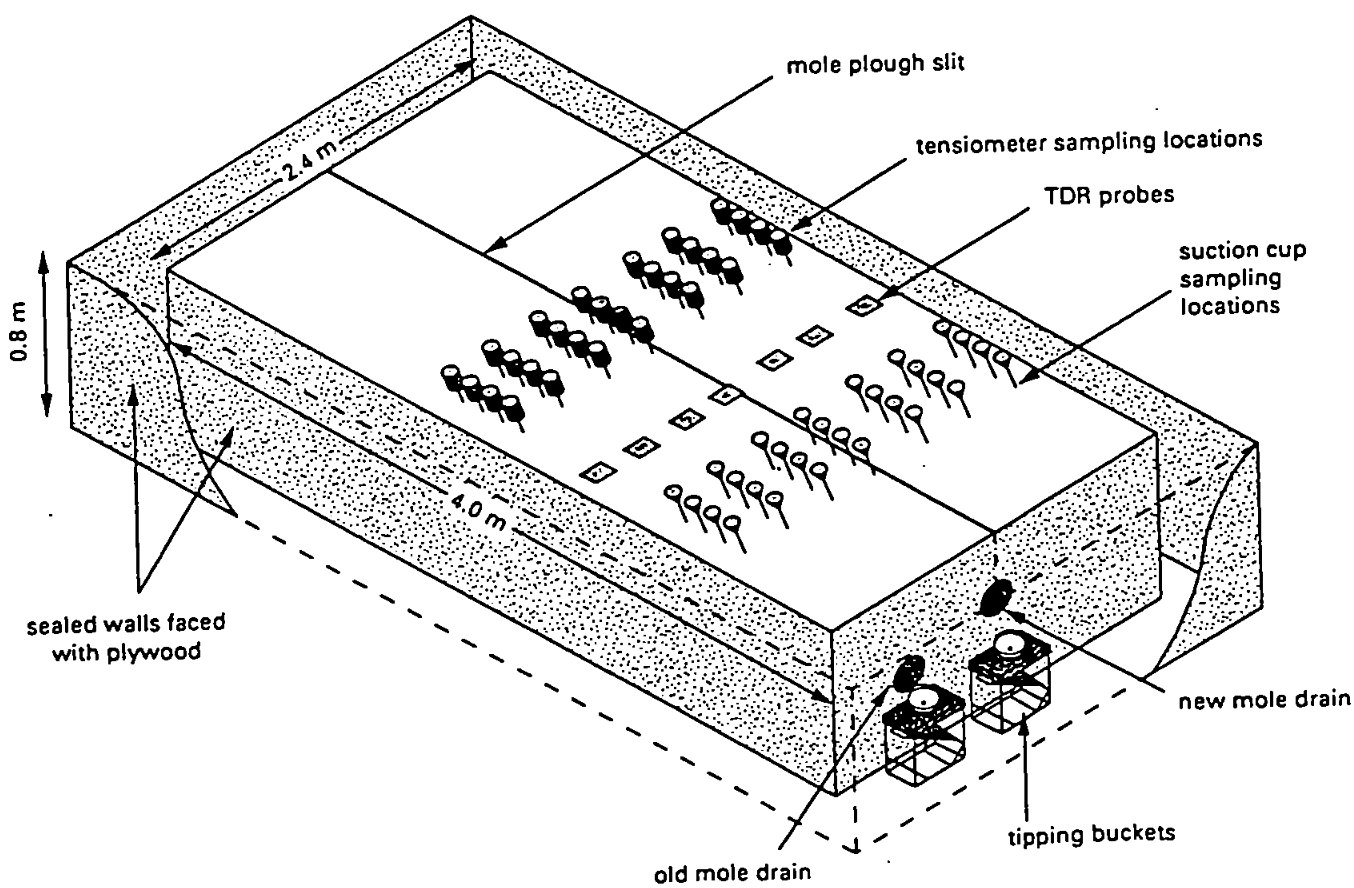


Figure 3.5 General design of the lysimeter.

Water supply was diverted from a mains source and a rotary piston water meter (BS 5728 Class D) was employed to control and monitor the volume of water applied. The smallest unit registered was 0.1 *l*. With an accuracy of ±2 % at discharges of 11.5 *l* h^{-1}, which increased with the rate of discharge, this provided a good indication of the amount of irrigation. The water was applied to the soil surface using a 10 m length of porous horticultural irrigation (or leaky) pipe which was spread over the block and moved at regular intervals. The pressure in the pipe was sufficient to ensure water sprayed over an average of about 0.5 m around the pipe.

Output from the lysimeter was determined using tipping-buckets, similar to those described by Knapp (1973). Two triangular buckets were connected side by side and centrally pivoted. Water fed into one bucket causes it to tip when sufficiently heavy, thus presenting the other bucket to collect the water. A magnet attached at the junction of the buckets, passes a dry reed switch every time a bucket empties, thereby operating an electromagnetic counter (Figure 3.6). The main disadvantages of this equipment occur during extreme flows. When low flows are prevalent evapotranspirational losses may represent a significant error. High flows may mean that discharge is underestimated during tips. A further problem is presented by the measurement of water left in the bucket after the last tip.

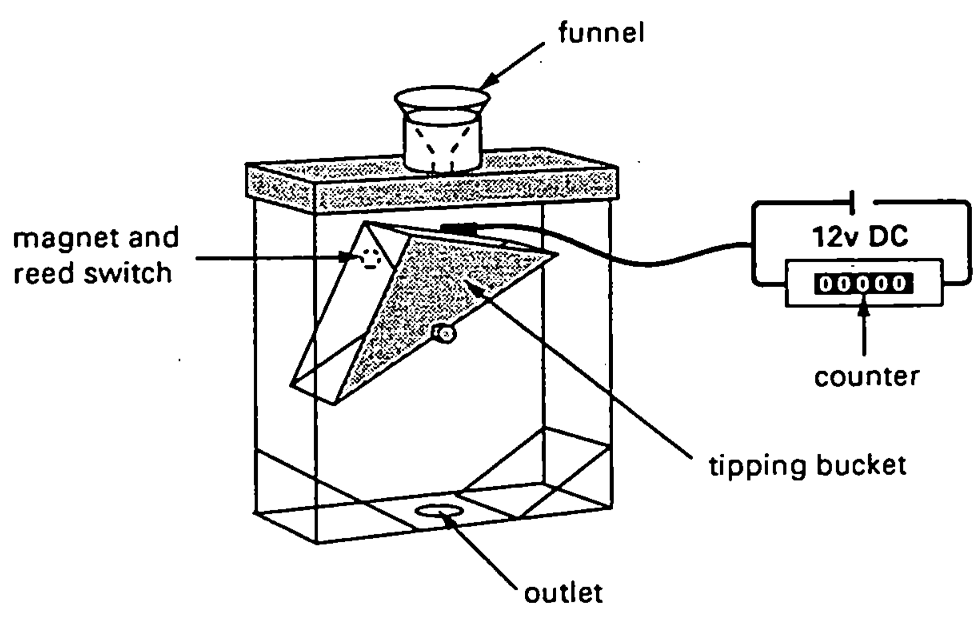


Figure 3.6 Tipping-bucket assembly

3.2 Determination of hydrological inputs and outputs.

3.2.1 Introduction.

Monitoring hydrological inputs and outputs allows the determination of rainfall-runoff response and the water balance of a field (Chapter 5). Measurement of precipitation is described in Section 3.2.2. Runoff encompasses drainflow as well as overland flow/shallow subsurface flow, to be referred to as non drainflow. Quantification is outlined in Section 3.2.4. These data allow the relative significance of different exit routes (evapotranspiration, drainflow and non drainflow) to be calculated. The timing and volume of hydrological outputs can be determined and compared spatially (between fields) or temporally (between events, months or drainage seasons). Impacts of varying antecedent moisture conditions and different precipitation volumes and intensities upon weirflow generation on the two fields may be assessed. In this way, it is possible to identify seasonal variation of field response. Furthermore, an examination of the consequences of agricultural drainage for drainflow generation in terms of timing, reliability and quantity is possible.

3.2.2 Precipitation.

Precipitation was monitored at the North Wyke Meteorological Station (8836), located 1.5 km south of the experimental site at Rowden (Figure 3.7). Daily readings of total rainfall were recorded at 0900 h. A tipping-bucket raingauge was established at the site to determine whether rainfall data from the North Wyke Meteorological Station were representative: no significant difference between the volumes recorded was observed (Table 3.2). Rainfall data were supplemented by information on timing and intensity from an automatic weather station located 1.5 km west of Rowden Moor.

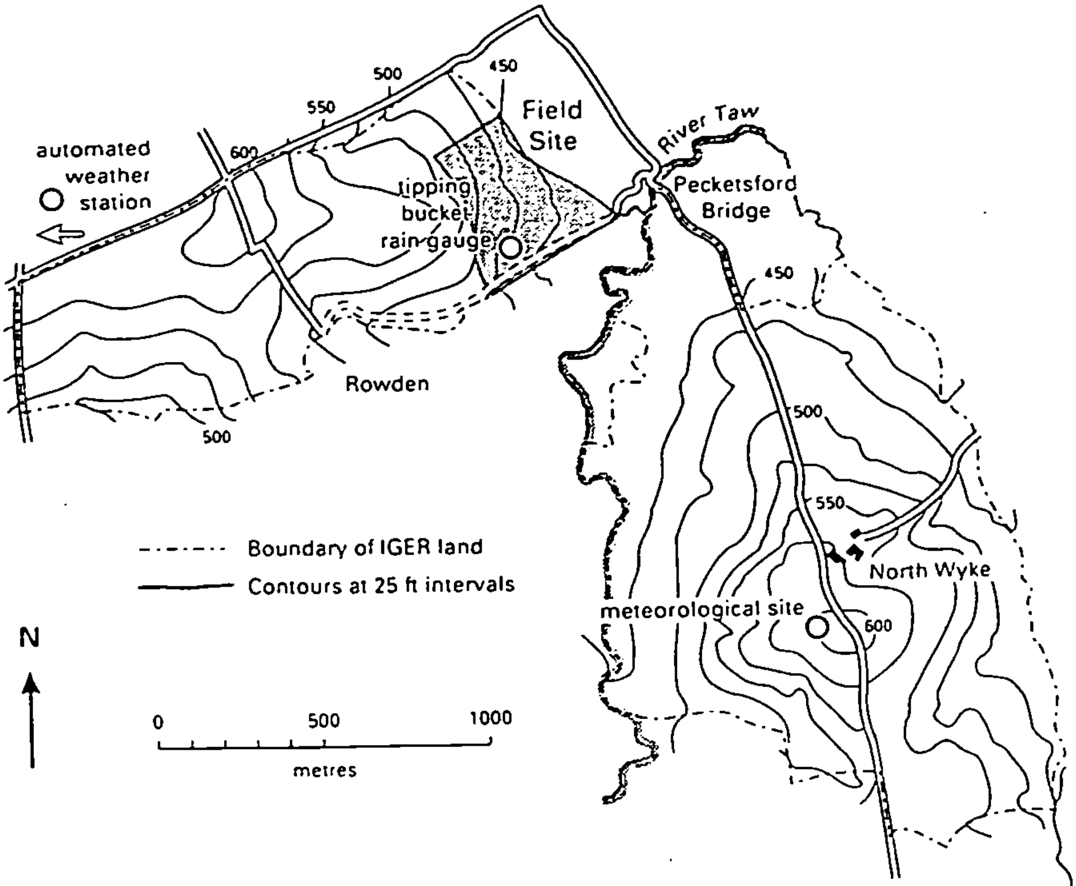


Figure 3.7 Locations of Meteorological Observation

Table 3.2 Rainfall recorded at Rowden and at Meteorological Station (mm).

Date	Rowden	Met. site
22.12.90	2	2.6
23.12.90	1	0.9
24.12.90	5	5.2
25.12.90	17	17.5
26.12.90	9	8.0
27.12.90	11	13.0
28.12.90	2	1.7
29.12.90	12	10.5
30.12.90	4	4.6
31.12.90	6	3.1
01.01.91	2	5.2
01.02.91	34	33.9
Total	105	106.2

3.2.3 Determination of evapotranspiration outputs.

An open pan lysimeter, situated at the North Wyke Meteorological site, was employed to determine the rate of actual evaporation from a body of standing water. Data were collected daily, from March to September, and less frequently throughout the remainder of the year. These values were then corrected according to Penman (1952) to calculate the rate of evapotranspiration from turf. Penman (1952) proposed that evapotranspiration from an area under a full crop cover amounted to 80 % of the volume of evaporation from a standing body of water, for May to August; 70 % for March, April, September and October; and 60 % for the remainder of the year.

Net or effective precipitation, that is incoming precipitation minus evapotranspirational losses, was calculated using evapotranspiration figures. The net precipitation is the total amount of water available for soil water storage and drainage.

3.2.4 Measurement of runoff.

To study soil water movement at the gross scale, one undrained and one drained field were selected (Section 3.1.2). Undrained fields have one weir outlet to monitor runoff outputs (surface flow and shallow subsurface flow) intercepted at the field boundaries by shallow drains (at 0.3 m), covered with gravel to the surface. Non drainflow is intercepted in the same manner on drained fields, and a second weir monitors drain discharge (Section 2.6, Figure 2.7). Therefore, it has been assumed that all, or nearly all, natural drainage on the undrained fields, occurs within the A/Ag horizon. This assertion is based on the low permeability of the soil and especially of the subsoil (Section 2.4.2) and the near surface location of the water table throughout much of the winter period (Hallard, 1988). All weirs were instrumented with continuously recording stage recorders to monitor surface and subsurface flow, as appropriate.

The weirs, based on an S.W.R.C. design (Talman, 1980), have a British Standard ½ 90° V notch. Whilst flows of up to 30 $l\ s^{-1}$ can be measured, accuracy is also maintained at low flows. The influence of turbulence is minimised by a 2 m long fibreglass weirbox. In addition, the head recording device is situated behind a perforated zinc plate on the opposite side of the water inlet. Fluctuating water levels are recorded by dataloggers and on a 30-day drum chart. Individual calibration of weirs was conducted by S.W.R.C. to give direct readings in $l\ s^{-1}$, with an accuracy of ± 0.83 mm over a range of 300 mm (Talman, 1983).

3.3 Soil hydraulic properties.

3.3.1 Introduction.

Soil hydraulic properties were characterised in the field and laboratory (Table 3.1). These characteristics influence the rate and direction of water movement, as explained in Section 1.3. Soil moisture retention curves were produced for soil horizons on both fields to examine the influence of artificial drainage on porosity (Section 3.3.2). Soil moisture retention curves illustrate the soil moisture status-soil moisture content relationship. This indicates the pore size distribution as the volume of water released at various suctions can be related to pore diameters (Hall *et al.*, 1977; Avery and Bascomb, 1982). In addition, saturated hydraulic conductivity (K_{sat}), the rate at which water is transmitted by a soil under saturated conditions, was determined. Despite the problems associated with this technique (Section 3.3.3), it provides a reference or standard which partially overcomes problems due to differing antecedent moisture conditions.

Particle size distribution was determined for each K_{sat} core (Section 3.3.4) as this property exerts a strong influence on the porosity and structure of soil, thereby influencing soil hydrology. Byers and Stephens (1983), Bresler *et al.* (1984) and Warrick and Nielsen (1980) showed that hydraulic conductivity can be empirically related to particle size distribution, which can be quickly and easily measured.

3.3.2 Structure (soil water retention characteristics, bulk density and porosity).

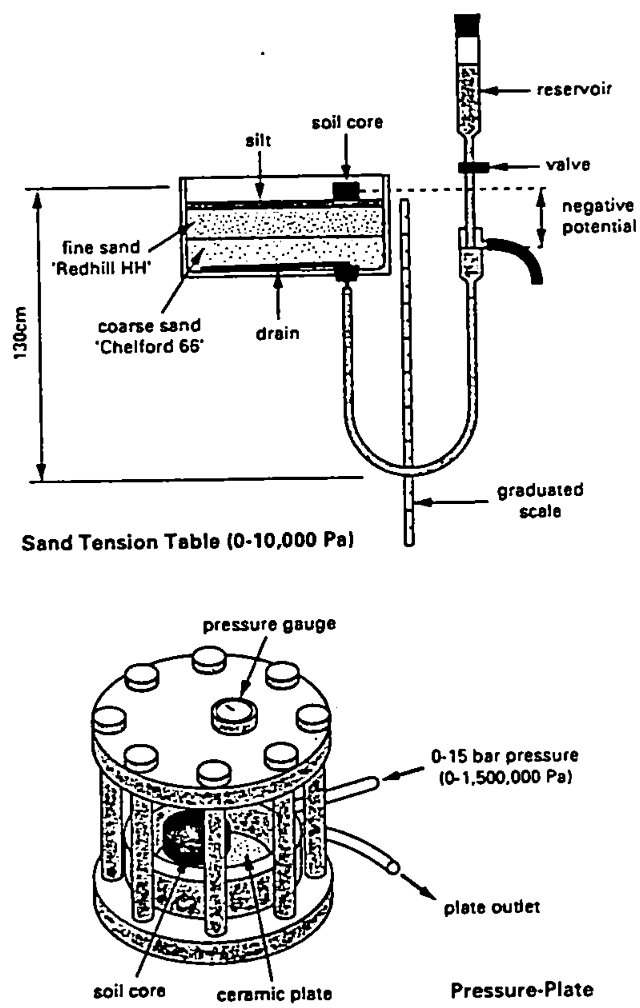
Thomasson (1978) has defined soil structure as the assemblage of aggregates (peds) and voids (pore space), including the voids within and between aggregates. Pore spaces between peds are important, as their size and frequency control the drainage characteristics and water retention properties of soils. Changes in structure due to management practices have implications for pore size distribution. The significance of increasing the proportion of large pores, by mole drain installation, for example, would be improved root development and drainage and reduced water retention capacity (Section 2.5.2).

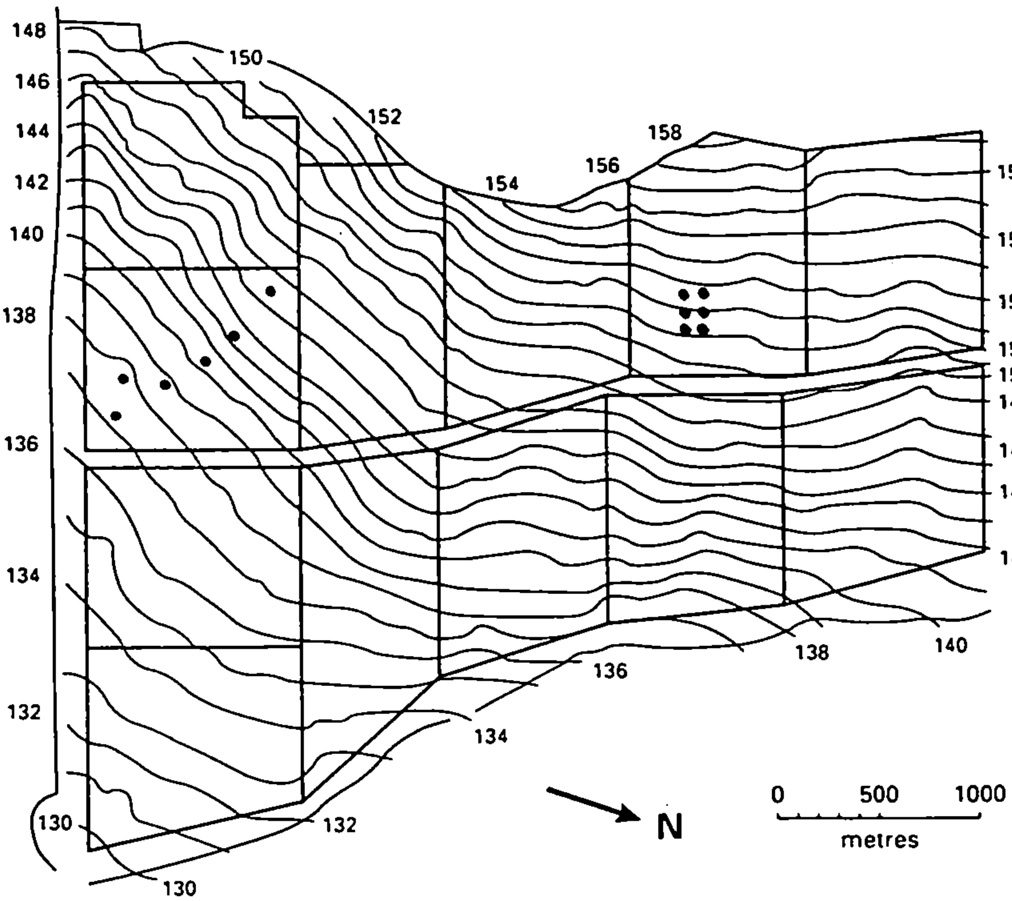
Structure may be assessed subjectively in the field on the basis of the degree of ped development and packing density (Hodgson, 1976). However, there is no universally agreed quantitative definition (Williams *et al.*, 1987), although Thomasson (1978), proposed a structural index for surface soils, based on the volumes of two categories of pore size: air capacity and available water. The air capacity of a soil is measured by the percentage of pores greater than 50 μm diameter; those which drain under gravity (Hall *et al.*, 1977). Where air capacity is less than 5 %, permeability is low and very little water drains under gravity. Air capacities of between 5 and 10 % represent moderately low permeabilities. Thomasson (1978) states that the air capacity of a soil reflects its ability to accept and transmit rainfall. The proportion of pores between 0.5 and 50 μm determines the available water capacity of soils. It is these pores which retain water against gravity.

Thomasson (1978) did not consider soils in which stones constituted more than 10 % of the soil mass, or where organic matter content was greater than 4 %, so his survey had a bias towards arable soils. In addition, Thomasson's classification ignores the structural water held in smaller pores at higher suctions which can form a significant proportion of the total water content of a soil (e.g. 30 % for clay soils; 5 % for sands). He only considers pores which can be drained at suctions of less than 15000 cm H_20, those which hold soil water which is active in terms of moisture and solute fluxes. This approach must, therefore, be employed with caution.

In this study, structural information is based on the water release curve, bulk density and porosity as well as field observation. The moisture release characteristic of soil is defined as the relationship between water content and tension. Standard laboratory techniques using sand tables and pressure plate apparatus (Klute, 1986; Figure 3.8) are very well established: they provide a high degree of accuracy of measurement, although they use a relatively small volume of soil (*c.* 50 cm^3) and are time consuming (Hall *et al.*, 1977). According to Ragab and Cooper (1990), they can be regarded as a reference method against which others can be compared. Mercury intrusion provides an alternative method (Danielson and Sutherland, 1986). However, the effectiveness of this technique is impaired by the small sample size (approximately 0.2 cm^3), which means that all samples are disturbed. The equivalent standard for the field is a direct comparison of water content and potential (e.g. tensiometry and Time Domain Reflectometry, Sections 3.3.7 and 3.3.8, respectively) in this case accuracy is limited by the instrumentation and spatial variability. Determination of the soil water characteristic is possible through virtually the entire range from saturated to very dry soil (Jury *et al.*, 1991). However, it is subject to hysteretic behaviour. Wetting and drying curves form an envelope that gives the extreme ranges of water contents that can be associated with any particular matric potential (Hanks and Ashcroft, 1980). The sampling procedure adopted in this study and the laboratory techniques are outlined below.

Figure 3.8 Sand table and pressure plate.





• Sampling site.

Figure 3.9 Sampling locations for water retention.

A Pitman corer was used to extract thirty one soil cores of 5 cm diameter and 3 cm height. A stack of three metal sleeves is supported in a heavy stainless steel tube to which a cutting ring is attached. The whole coring device is hammered carefully into a flat horizontal surface avoiding compaction of the sample (Hall *et al.*, 1977). Following excavation, the middle section is disposed of and, after trimming, the upper and lower rings provide the samples on which analyses may be performed. Seventeen soil cores were obtained from locations along a diagonal, downslope transect across the undrained field (Figure 3.9). This transect involved sampling along all of the topographical components of the plot (convex, concave and linear slope segments). Fourteen cores were removed from mole and mid mole positions, downslope from the plot, on the drained field (Figure 3.9). The bases of trimmed cores were wrapped with a thin gauze mesh to prevent any loss of sample during analysis.

Samples were weighed at field moisture then placed in a tray, in which water stood slightly shallower than the ring depth. They absorbed water until no further weight gain could be detected. Wetted samples were then equilibrated at 0 cm, 5 cm, 10 cm, 20 cm, 30 cm, 40 cm and 50 cm suction on a sand tension table and reweighed. A Soil Moisture Equipment Corporation pressure plate was employed to equilibrate samples to 300 cm, 1000 cm, 3000 cm, 5000 cm and 15000 cm. The analysis was conducted at a standard temperature of 20°C. Tension table and pressure plate equipment were constructed according to Soil Survey of England and Wales procedures (Hall *et al.*, 1977). Therefore, water content-tension relationships discussed are for a drying curve.

Bulk density: this is defined as mass per unit volume of dry soil (White, 1979). It is influenced by soil texture and porosity and hence related to percentage air capacity and retained water capacity (Reeve *et al.*, 1973). Typical bulk densities for undisturbed soils range between 1.0 g cm^{-3} and 1.3 g cm^{-3} depending on texture and condition (Brady, 1990). Sandy soils without structure and well aggregated clays are found at the upper and lower ends of the range, respectively. Trafficking or working of soil when excessively wet results in failure of structures and a squeezing out of soil pores (Williams *et al.*, 1987). The resulting soil compaction is demonstrated by an increase in bulk density (Soane, 1970).

Following water release to 15000 cm H_2O, all cores were oven dried at 80°C for 24 hours and reweighed. The soil cores were removed from the rings and the weight of each ring was determined. Bulk density was then calculated.

Porosity: defined as the ratio of the volume of pore space to the bulk volume of soil, this was calculated from the water release data by dividing the mass of water (at saturation) by the mass of dry soil.

3.3.3 Saturated hydraulic conductivity (K_{sat}).

Accurate determination of K is of paramount importance as ti reflects the cumulative impact of soil characteristics on the route and rate of water movement. Further, it is a key parameter for modelling most aspects of water and solute movement (Hutson and Wagenet, 1992; Ragab and Cooper, 1990). There are a number of field and laboratory techniques which can be employed to measure K_{sat}, but all are based on transformations of Darcy's original equation (Section 1.6; Eqn. 1.1). K_{sat} is calculated from Darcy's Law after measuring or imposing a soil water flux.

Laboratory measurement of K_{sat} is likely to be unreliable due to problems associated with the use of smaller cores and increasing disturbance experienced by the sample (Talsma, 1969; Berryman, 1974; Gilmour *et al.*, 1980; Ragab and Cooper, 1990). Saturated hydraulic conductivity was determined in the field using ring permeametry (Bonell *et al.*, 1983; Figure 3.10). The ring permeameter primarily measures vertical hydraulic conductivity of an excavated soil core. Measurement in the field has the advantage of minimising disturbance of the soil sample, thereby maintaining the continuity of macropore systems. In addition, larger volumes of soil can be handled relatively quickly (Chappell, 1990).

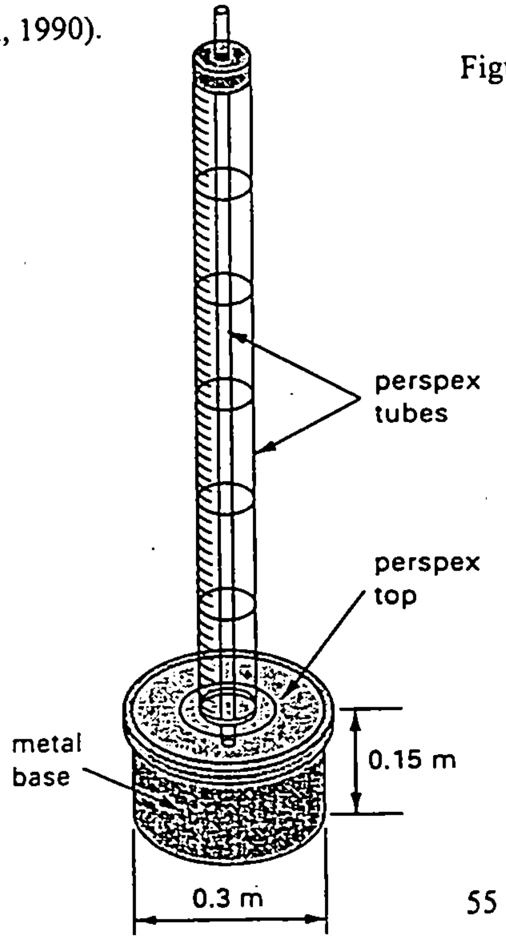


Figure 3.10 Ring permeameter.
(After Talsma, 1969)

Insertion of the ring may disturb the structure of the soil contained in the core and if the sides of the ring are not adequately sealed water may bypass the sample. However, careful application of this technique limits such errors. It is also a widely used technique, which allows isolation of different soil horizons to determine the significance of each in the overall profile or for hillslope hydrology.

The ring employed was 0.15 m deep (inserted to 0.1 m) and 0.3 m in diameter, which Youngs (1983) suggested would provide a representative sampling volume. Smaller cores were used by Chappell (1990) but he noted considerable variation due to limited sample size. Cores were extracted at depths of 0 - 0.1 m and 0.1 - 0.2 m. Stones made insertion of the ring problematic, particularly in the subsurface horizon of the drained field. Random sampling was employed on the undrained field. The cores were retrieved from above mole and mid mole positions on the drained field. Following wetting of cores for a minimum of 20 minutes, a constant head of 4 cm was applied (Figure 3.10) and the volume of water required to maintain this head noted, until a constant rate was established. This field analysis was conducted in a dry period over two weeks in August, 1991.

Saturated hydraulic conductivity may rarely, if ever, be achieved in many soils as considerable quantities of air may be entrapped during rapid wetting (McDonnell, 1988). This is an important consideration when interpreting data obtained from techniques such as the ring permeameter which will generate very different results compared to an approach such as gypsum blocks (Armstrong and Garwood, 1991).

Possible problems associated with antecedent soil moisture conditions were anticipated. Therefore, three undrained and three drained cores were retained and progressively wetted over fourteen days. During this period K_{sat} was determined on five more occasions. K_{sat} is supposed to be a constant for a particular soil, as soil moisture content is controlled. However, it could take months for desiccation cracks in clay soils to close (Emerson, 1959) and Chappell (1990) found that the measured permeability of surface soils on a grass hillslope increased during summer months. A longer wetting period may decrease

K_{sat} due to swelling of clays. On the other hand, exploitation of preferential flow paths and reduction of hydrophobic effects may increase conductivity.

The rate of fall of water in the permeameter is recorded and K_{sat} determined using Darcy's Law:

$$K = Q \quad L/do \qquad \text{[Eqn. 3.1]}$$

where:	K	=	hydraulic head ($cm^3\ s^{-1}$)
	L	=	sample length (cm)
	do	=	total head (cm)
	Q	=	flow rate through the sample ($cm^3\ s^{-1}$)

where: $Q = \dfrac{\text{rate of fall x scale factor}}{\text{area of sample}}$ [Eqn. 3.2]

where: rate of fall of water in permeameter and scale factor is the volume of water per unit fall of water.

3.3.4 Particle size analysis.

Particle size analysis was conducted for a subsample of each core extracted for K_{sat} determination. Subsampling of dried fine earth fraction in the laboratory was conducted in accordance with the recommendations of Mullins and Hutchison (1982) and Smith and Pratt (1984), with the aid of a riffle splitter in preference to coning or quartering to ensure maximum representativity. The subsample was dispersed in 'Calgon' solution before wet sieving conducted according to BS1377 test was employed. This procedure determined the particle size distribution of samples down to, but excluding, silt and clay particles.

A Fritsch Scanning Photo-Sedimentograph was used for the analysis of the silt and clay fractions. Preparation included removal of fine organic particles and the dispersion of silt and clay particles by using an ultrasound bath. The sedimentograph measures the change in optical density of a sample over time (this being related to settling time and therefore the size of particles). This differs from the standard British Soil Survey technique which involves the removal of aliquots and calculation of particle size distribution based on weight. However, the relationship between particle size determined by the two techniques is very consistent (r = 1.0, Hartley, pers. comm.).

Laurens *et al.* (1988) found a relationship between K_{sat} and macrovoid area, percentage silt and clay and soil texture can be seen to be related to other soil physical and hydraulic properties (Chapter 4).

3.3.5 Soil strength.

The ease with which a horizon can be penetrated provides an indication of the compaction of the horizon and enables an assessment of the soil conditions. This is a reflection of soil texture and structure. The Bush penetrometer was used to measure resistance to a depth of 0.6 m in fifteen stages. The penetrometer is pushed into the ground using even, manually applied pressure until the maximum depth is reached or the maximum load is exceeded. Resistance was determined across mole - mid mole transects on the drained field and also at K_{sat} sampling sites (Section 3.3). At each K_{sat} sampling site ten readings were made around the core to obtain an average reading. The presence of stones proved to be a problem, causing readings to be aborted, especially on the drained site.

3.3.6 Organic matter content.

In mineral soils, organic matter content is important for soil structural development and moisture retention. It was determined for subsamples of K_{sat} cores by placing oven dried soil in a muffle furnace at 350°C for 24 hours. Loss on ignition was expressed as a percentage of the oven dried weight.

3.3.7 Matric Potential.

The difference in matric or soil moisture potential between two points within a soil provides an energy gradient for the movement of soil water (Darcy, 1856; Hubbert, 1940). Matric potential is most commonly expressed in terms of pressure (Bear, 1972). This pressure can be measured by a variety of instruments (Burt, 1978; Curtis and Trudgill, 1974) the most widely used of which are tensiometers and piezometers. Piezometers are open to capillary pressure at their base and atmospheric pressure at their top. The depth of water in the piezometer represents the head or positive pressure within the soil. Tensiometers were adopted in this study, and comprise a porous ceramic cup, with an

ABS plastic body (Figure 3.11). The tensiometer cup is buried in the soil, the cup and body hold de-aired water which is drawn through the cup until water tension in the soil and the cup are at equilibrium. Water is constantly interchanged between the unit and the soil, so that pressure within the unit is determined by that of the soil. Increasing or decreasing soil moisture contents, causing variations in tension, result in water movement in or out of the cup until equilibrium is attained once more. Klute (1962) gave the minimum detectable pressure change as 2 to 3 mm of water. Summer potentials are below the range of measurement of tensiometers (850 cm of water; Watson, 1965). However, Long (1984) stated that at this suction 90 % of available water in sandy soils would be removed. At the other extreme of the measurement range, tensiometers can indicate positive pressures. In this way, the water table can be located. Gradients of hydraulic potential indicate directions and magnitudes of water flux in the field.

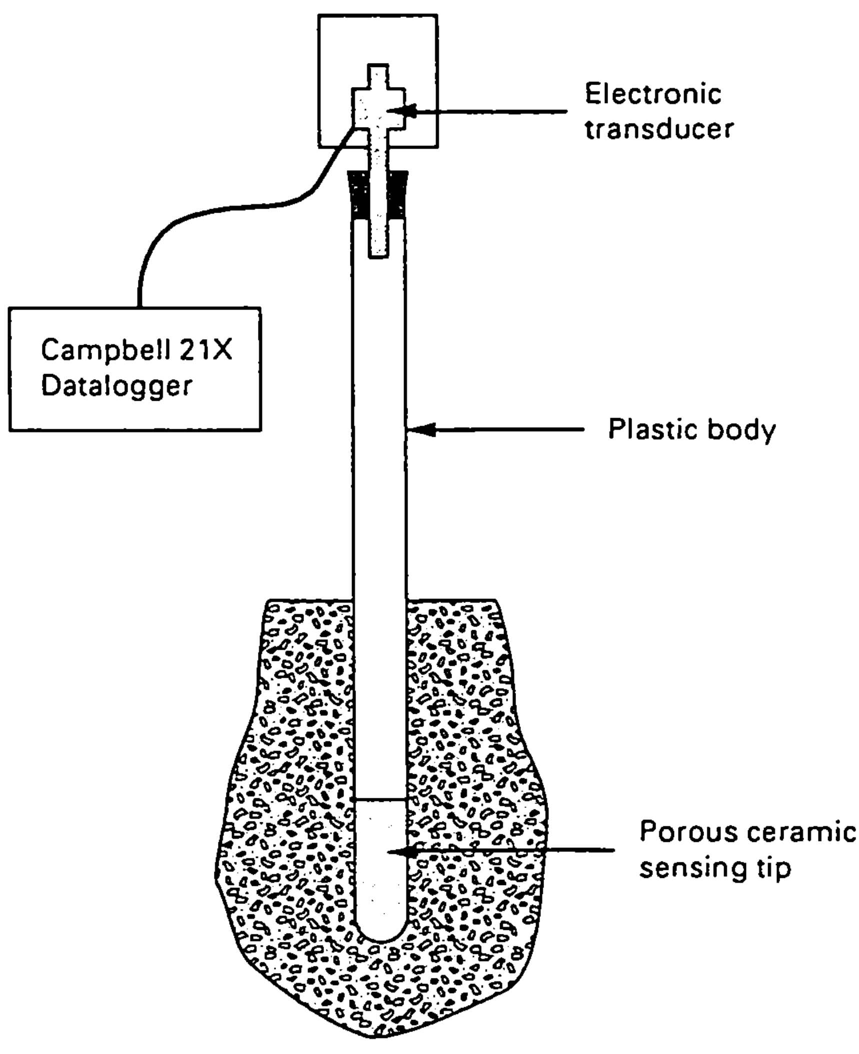


Figure 3.11 Tensiometer design.

Two techniques are commonly used to monitor soil water suction in tensiometers: mercury manometers and pressure transducers. The former involves manual reading which is time consuming, and is often limited to recording at daily or weekly intervals (Lowery *et al.*, 1986). Infrequent measurement means small changes in soil water potential may be neglected (Long, 1984). This inaccuracy is compounded by the low sensitivity of vacuum gauges. Concerns regarding safety in grazed fields provide a further disadvantage of mercury manometers, which can be overcome by the use of transducers. Pressure transducers also have the advantage of rapid response and greater accuracy (Long, 1982). Dowd and Williams (1989) indicated that soil water potential can be measured to within ±5 cm H_2O whilst conventional mercury manometers are only accurate to within 20 cm H_2O. Further, pressure transducer data can be datalogged, facilitating the monitoring of rapid changes in soil water suction as well as computer processing of results.

Calculation of total potentials: pressure potential measured by tensiometers encompasses all effects on water other than gravity. A temporary bench mark was established and levelling was referenced to this point to establish the instrumentation positions and the perimeters of the fields. A theodolite was used to set out these locations. In this way, the elevation and, therefore, gravitational potential of each instrument could be determined. Total potential is calculated by the combination of gravitational potential and matric potential.

Instrumentation: an automated tensiometer-datalogger system was used to monitor soil water suction on the undrained and drained plots and the lysimeter. Tensiometers were constructed using 2 cm diameter Soil Moisture Corporation 1 bar ceramic cups, affixed to A.B.S. tubing of the same outer diameter by epoxy resin. Nests of tensiometers were installed at depths of 0.1 m, 0.2 m, 0.4 m, and 0.6 m, set at an angle of 45°. Duckboards were employed during installation and monitoring to avoid compaction of the soil surface. Installation at 45° reduces the likelihood of cracks (induced by coring and tensiometer insertion) coinciding with the cup or body. A 2 cm diameter screw auger and a guiding jig were employed to excavate the tensiometer holes. A small amount of a clay and sand mixture was deposited in the bottom of the hole to ensure good soil-cup contact. This installation procedure was adopted in preference to the standard technique of excavating a larger hole and backfilling (MacDuff, 1986) to minimise the risk of failure due to poor soil-cup contact and to reduce disturbance of the soil. Twenty four tensiometers were installed on the undrained plot (Figure 3.2) and a further twenty seven were established across a mole drain (Figure 3.3). The lysimeter was also instrumented with nested tensiometers (Figure 3.4) but to minimise disturbance, 1 cm diameter cups were used. Such instrumentation allowed the study of variations in matric potential within and between soil horizons (Section 2.4.1) at each site and a comparison of soil hydrological response under different treaments.

The automated tensiometer system monitored fluctuations in soil moisture tension at a 30 minute interval. Honeywell pressure transducers were employed (Dowd and Williams, 1989). Mains direct current was stepped down to 12V through a power supply, the input to transducers was regulated to 10V by adjustable voltage regulators. Pressure transducers were at a maximum distance of 2.5 m from the power supply, AM32 and logger. All transducers were connected by cables of equal length to minimise problems associated with voltage drop. Transducers were placed in water-tight containers with silica gel.

Electrical output from transducers was recorded on a Campbell 21X datalogger via an AM32 multiplexer. Data were downloaded using a Husky Hunter field computer, before transferring to a mainframe. Conversion of data was achieved using individual transducer calibrations derived in the laboratory using a hanging column. Regression of head against electrical output was conducted to derive an X coefficient and a constant to be used in the conversion of the field electrical output (Table 3.2. and Figure 3.12).

Figure 3.12 Transducer calibration
(Numbers 11, 12 and 13).

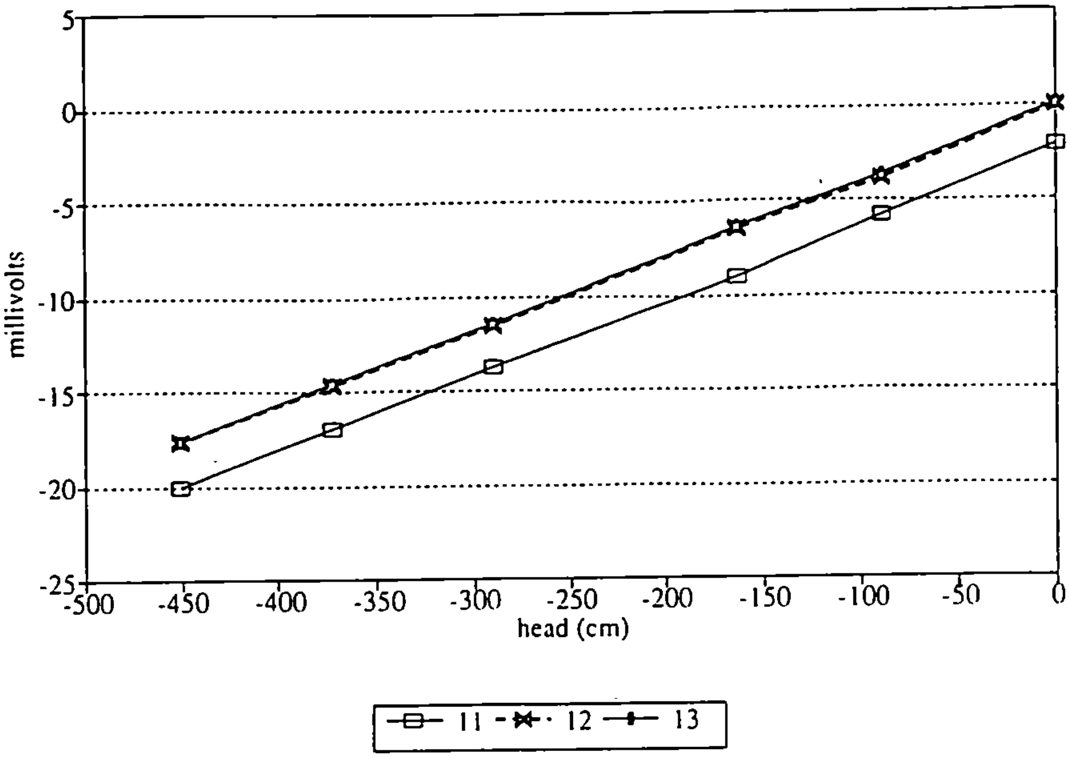


Table 3.2 Calibration of Honeywell transducers - examples.
Output from transducers and potentials calculated using regression output.

Calibrat heads cm	Left column: transducer output; right column: calibrated potential. transducer 1 mv	cm	transducer 2 mv	cm	transducer 3 mv	cm	transducer 4 mv	cm	transducer 5 mv	cm
0	-5.43	1.688	-4.01	0.032	-0.62	2.128	-0.46	1.443	-0.65	1.63
-118	-9.93	-122	-8.31	-119	-5.06	-123	-4.97	-122	-5.09	-122
-275	-15.4	-273	-13.8	-273	-10.4	-274	-10.5	-273	-10.5	-273
-330	-17.5	-330	-15.9	-330	-12.4	-330	-12.5	-329	-12.5	-330
-387	-19.5	-385	-17.9	-386	-14.4	-383	-14.5	-384	-14.5	-384
-471	-22.7	-473	-21	-472	-17.5	-473	-17.8	-473	-17.7	-473
regress	X	K	X	K	X	K	X	K	X	K
	27.51	151.2	27.75	111.2	28.08	19.61	27.38	14	27.89	19.81

3.3.8 Soil Moisture Content.

Soil moisture content can be analysed by a variety of techniques including gravimetric determination, the neutron probe, capacitance probe and Time Domain Reflectometry (T.D.R.). T.D.R. provides a rapid and *in situ* means for determining soil moisture content. Unlike gravimetric techniques, it is non-destructive and offers greater precision. Spatial resolution is higher than that for neutron probe moderation (Jury *et al.*, 1991). Electromagnetic pulses are transmitted along a set of parallel metallic rods installed in the soil. The properties of a wavelength which is reflected back to the T.D.R. provide an indication of volumetric soil moisture content (Heimovaara and Water, 1993). This is possible because of the variation in the dielectric constants of air (k = 1), soil grains (k = 3 to 5) and water (k = 81). The dielectric constant of soil is, therefore, highly sensitive to moisture content and relatively independent of soil type. The relationship between dielectric constant and volumetric water content has been established by many authors (Hoekstra and Delaney, 1974; Topp *et al.*, 1980; Ansoult *et al.*, 1985). It is complex, for example as expressed by Topp *et al.* (1980) it is not linear due to the influence of bound water (Ansoult *et al.*, 1985):

$$\theta_v = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} K - 5.5 \times 10^{-4} K^2 + 4 \times 10^{-6} K^3 \qquad \text{[Eqn. 3.4]}$$

where: θ_v is volumetric soil moisture content; K is dielectric constant.

A Tectronix 1502B reflectometer was used, in conjunction with 3 mm diameter stainless steel transmission lines. The lines were not impedance balanced (Stein and Kane, 1983; Zegelin *et al.*, 1989).

On the plots, instrumentation consisted of pairs of parallel stainless steel probes installed in September, 1990. They were located in grids (0.5 x 1.0 m) covering a transect from mole to mid mole on the drained site and a similar area on the undrained site (Figures 3.3 and 3.2). Rods were embedded to depths of 0.1 m, 0.2 m, 0.4 m and 0.6 m where possible but stones resulted in some pairs being shortened. Installation of rods was conducted using a mallet and guides, produced by drilling parallel holes in wooden blocks. Various workers have found that T.D.R. observations are least reliable when short probes are employed (Dowd pers. comm.; Goss pers. comm.). Therefore, probes in the upper 0.1 m of the soil profile were installed at 45° to the surface, giving an actual length of 0.14 m.

On the lysimeter, probes were located to enable estimation of volumetric soil moisture content along a transect which crossed the block and intercepted the mole drains. Four pairs per depth were located at eight points along the transect (Figure 3.4).

3.4 Hydrochemical properties.

3.4.1 Introduction to field and plot study.

Indirect methods, such as determination of K_{sat} (Section 3.3.3) and drain discharge (Section 3.2.4), are normally adopted to study subsurface hydrology. However, the use of tracers is the only direct way in which complex flowpaths can be determined, all others being indirect or destructive (Reeves and Beven, 1990). As discussed in Section 1.7, tracer studies have been used to elucidate the major hydrological processes occurring within hillslopes (Figure 1.2). At the plot scale, McDonnell (1990) has shown that tracer techniques are powerful means of determining the importance of macropore flow.

The chloride content of precipitation, overland flow, subsurface or drainflow and matric water samples was analysed. Precipitation samples were bulked for each event. Undrained and drained field runoff was collected manually from the three weir outflows. The mole drain tipping-bucket outflows provided the sampling point for lysimeter runoff. Overland flow was only observed on the undrained plot and was sampled manually (the number and location of sampling sites was determined by duration and extent of flow). Matric water was sampled by suction samplers located within the plots and the lysimeter (Figures 3.2 to 3.4; Section 3.4.2).

In addition to chloride content, stable oxygen concentrations of ten samples were analysed to determine their origins and their residence time. Fifty samples were analysed for nitrate and ammonia content to provide an indication of variability associated with the various hydrological regimes. The analytical techniques used to determine chloride and stable oxygen contents are described in Section 3.4.3.

3.4.2 Suction cup samplers.

Suction samplers were constructed of the same porous ceramic cups and plastic bodies as tensiometers (Section 3.3.7, Figure 3.11). They were embedded in the soil in a similar manner, ensuring a good contact between cup and soil. A vacuum was established in the body of the instrument. If the matric potential of the soil was lower than that in the sampler, water was drawn across the cup wall, until the potentials of the cup and the soil were in equilibrium (Talsma *et al.*, 1979; Stevens, 1981).

A total of six suction cups at 0.2 m, 0.4 m, and 0.6 m were installed on the undrained plot (Figure 3.2). On the drained plot they were situated over the mole drain and at mid mole, downslope from the tensiometers (Figure 3.3). These depths were sampled in order to provide data for the two horizons above the mole drain (Section 2.4.1). Four cm diameter cups were used to create a large contact area with the soil. On the lysimeter, twenty four suction cups were established downslope from the tensiometers (Figures 3.5, 3.13). Due to the high density of installation, 2 cm diameter cups were used. Suctions of 0.4 to 0.7 bars were applied for 12 to 48 hours using a manual vacuum pump.

3.4.3 Tracer techniques.

A tracer experiment was conducted on a 2.4 m by 4 m isolated soil block, centred on a newly installed mole drain and its related system of macropores, and also included a 5-year old mole drain. The plot was intensively instrumented to enable soil tension and moisture content determination (Figures 3.4 and 3.13). Tipping-buckets indicated the volume and timing of drainflow (Section 3.1.4). After a wetting period of one week, a KCl and stable oxygen labelled tracer was applied to the soil surface in 20 mm of water, over a one hour period. Nine subsequent additions of unlabelled water were made. Over two hundred samples of soil matric water and drainflow were analysed for Cl and $^{18}O_{SMOW}$ at the Department of Geology, University of Georgia.

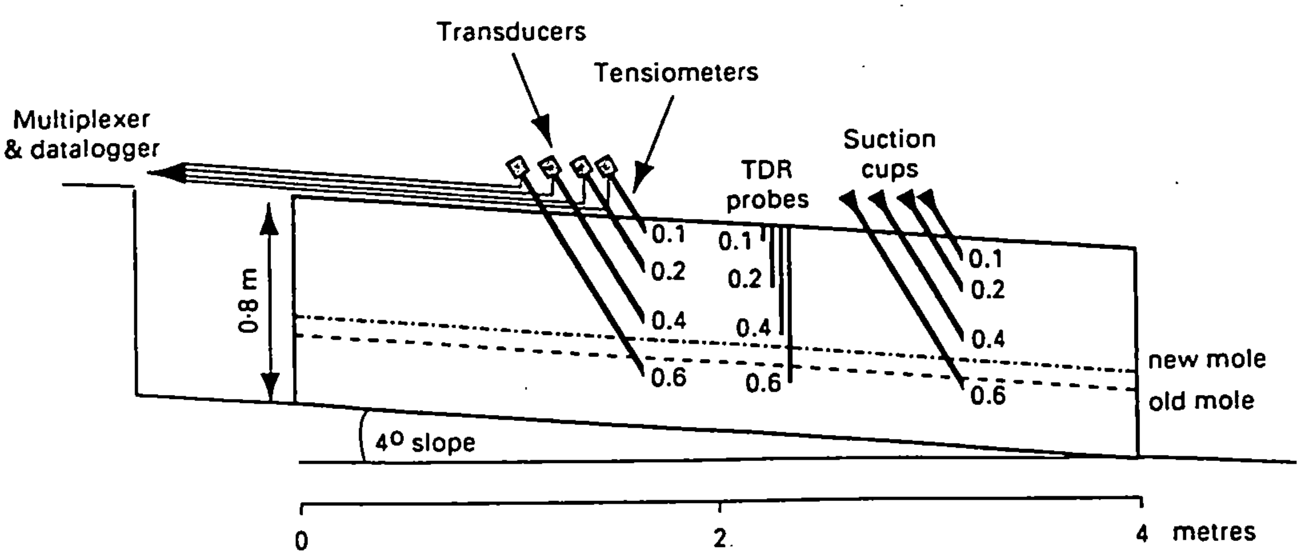


Figure 3.13 Lysimeter cross section

Chloride concentrations were determined using a Technicon AA II GTPC Autoanalyser. A colour reagent which contained 15 % v/v mercuric thiocyanate and 15 % v/v ferric nitrate reagent was employed. On mixing with the sample, the thiocyanate ion is released from mercuric thiocyanate through replacement of mercury by the chloride ion to form un-ionised mercuric chloride. In the presence of ferric ions, liberated thiocyanate ions form highly coloured ferric thiocyanate in concentrations proportional to the original chloride concentration. The concentration can be quantified by determination of absorbance using a spectrophotometer set to a wavelength of 480 µm. The calibration, derived by employing a blank and a range of standards, is linear for the range 0 to 40 mg l^{-1}, samples lying outside this range were diluted (Solman pers.comm.).

Stable oxygen concentrations were determined by the carbon dioxide (CO_2) gas equilibration technique and mass spectrometry, both are outlined here. The CO_2 equilibration procedure relies on the isotopic exchange between oxygen in the unknown water and that in gaseous carbon dioxide. Liquid nitrogen cooled ethanol freezes the CO_2, impurities are drawn off, and pure CO_2 is added to vials containing water samples. The vials are agitated and equilibration occurs over 24 hours. The equilibrated CO_2, is drawn into a vacuum, isolated by liquification using liquid nitrogen and extracted through a capillary system. It is transferred to sample bottles, which include two controls to validate the replicability of each run. The international standard V-SMOW is employed. This has an isotopic composition relative to standard mean ocean water of $^{18}O = \pm 0.04$ %.

Following the gas equilibration procedure, CO_2 in the water samples was analysed using the Finnigan MAT, Model Delta E Isotope Ratio mass spectrometer. Mass spectrometry can provide precise $\delta^{18}O$ values ($\pm < 0.2$ %; Bowen, 1988). The mass spectrometer separates charged atoms and molecules by their masses, based on their motions in magnetic and/or electrical fields. There are four major parts: the inlet system; the ion source; the mass analyser and the ion detector. The gas is drawn into a chamber through an aperture, the different weights of isotopes, and therefore, the molecules of which they are a part, determines the trajectory of each molecule within the chamber. The exit point of molecules provides an indication of their weight and so their isotopic values. Thus, the ratio of ^{16}O:^{17}O:^{18}O can be calculated. In the inlet system, an instability of ions and mass separation is produced by the application of a high vacuum. The flow velocity of lighter components is greatest, so heavier isotopes become enriched in the reservoir from which the gas flows to the mass spectrometer. The next part of the spectrometer, the ion source, forms and accelerates ions, focusing them in a narrow beam, which is collimated by a weak magnetic field. In the mass analyser, ion beams are separated according to the M/e ratio. Passing the ion beams through a magnetic field deflects them into circular paths. Ions are collected in the ion detector, where they generate an electrical impulse which is then fed into an amplifier. The simultaneous collection of two beams provides a more accurate determination of ratios.

This approach, comparing chloride and stable oxygen, allowed an analysis of the transmission, retention and the degree of mixing of water in the soil profile, as discussed in Chapter 7.

CHAPTER 4: SOIL HYDRAULIC PROPERTIES.

4.1 Introduction.

This chapter characterises soil properties which influence soil water movement: they are listed and their significance is outlined in Table 4.1. Techniques used to obtain these results were described in Chapter 3 and summarised in Table 3.1.

4.2 Soil properties.

4.2.1 Introduction.

To a large extent, soil properties such as structure and density as well as the others listed in Table 4.1 are interdependent. Organic matter and texture, for example, influence many other properties (White, 1979). However, structural characteristics are of paramount hydrological importance, integrating physical, chemical and biological factors and, therefore, they are given due prominence in this section.

Table 4.1 Hydraulic and related soil properties investigated.

Characteristic	Significance
Soil water characteristic curve (Section 4.2.2)	relationship between soil water content and soil water potential
Bulk density (Section 4.2.3)	reflects porosity
Structure and porosity (Section 4.2.4)	determines K_{sat}, soil moisture content
Saturated hydraulic conductivity (K_{sat}) (Section 4.2.5)	ability of soil to transmit water
Particle size distribution (Section 4.2.6)	related to structure, porosity, K_{sat}
Soil strength (Section 4.2.3)	indication of compaction
Organic matter content (Section 4.2.7)	improves structure, water retention
Soil moisture content (Section 4.2.8)	dictated by porosity, texture and retentivity, a control of K_{sat}

4.2.2 Soil water retention characteristics.

Soil water retention data for the undrained field are presented in Table 4.2, summarised in Table 4.3 and shown graphically in Figures 4.1 to 4.3. The volumetric soil moisture content at saturation ranged from approximately 40 to 50 %, averaged 44 % and showed minimal variation with depth (Tables 4.2 and 4.3). At 0.1 m depth, water contents were about average, at 0.2 m they were lower and at 0.4 m they were higher than average.

When 50 cm H_2O suction was applied, volumetric soil moisture content ranged from 37 to 47 %, averaging 42 %. This suction is significant as it represents the lowest negative pressure at which water can be transmitted under the influence of gravity. At this suction, drainage from pores greater than 60 µm in diameter occurs. The water held between 50 cm and 15000 cm H_2O is known as the available water capacity (Gradwell, 1978; Danielson and Sutherland, 1986). At 15000 cm H_2O suction, water content ranged from 26 to 36 %, averaging 30 %. The boundary between stored water and structural water is defined as occurring at 15000 cm H_2O (Hanks and Ashcroft, 1980). On the undrained soil, the greatest amount of structural water was observed at 0.4 m (33.2 %).

These data indicate that the amount of transmission, i.e. the number of large pores, is very limited, accounting for only 2.6 % of the soil volume on the undrained site. Pores of less than 60 µm in diameter contribute 11.6 % of the soil volume. Hogan (1978) working on the same soil, measured air capacities of 12.0 and 4.8 % for the A and B horizons, respectively (Table 4.4). At saturation, Hogan found that total pore space was 63.3 and 50.4 % by volume in the A and B horizons. Soil moisture content decreased to 51.3 and 29.2 % in the A horizon at 50 cm and 15000 cm H_2O suction and 45.7 and 29.1 % in the B horizon. The higher values recorded by Hogan for the A horizon are partially explained by lower bulk density values (Section 4.2.3).

Table 4.2 Output from water retention analysis.

Site	Initial weight	5 cm	10 cm	20 cm	30 cm	40 cm	50 cm	0.3 ba	1 bar	3 bars	5 bars	15 bar
		SUCTION APPLIED										
T 21	36.7	46.05	46.06	45.91	46.05	46.18	45.97	43.71	40.01	35.57	33.35	28.06
B 21	35.06	44.8	44.74	44.42	44.46	44.58	44.45	40.52	37.63	33.61	31.99	28.6
T 22	26.94	40.41	40.23	39.88	39.91	40	39.78	36.3	34.12	30.97	29.49	27.27
B 22	28.9	42.63	42.38	42.15	42.01	42.1	41.47	38.54	36.64	34.18	32.52	30.25
T 24	23.34	45.96	46.02	45.72	45.81	45.83	45.83	41.27	38.16	34.1	32.05	29.05
B 24	23	44.43	44.42	44.09	44.13	44.3	44.15	39.55	36.72	33 14	30.99	29.29
T 26	26.48	41.32	41.06	40.7	40.74	40.68	40.5	37.17	34.74	32.29	30.47	27.98
B 26	29.84	43.57	43.38	42.95	42.94	42.88	42.42	38.82	36.07	32.59	30.93	28.96
T 27	32.92	48.47	48.23	47.78	47.81	47.84	47.44	44.34	41.18	38.31	37.14	34.09
B 27	34.92	47.59	47.43	47.05	47.03	47.11	46.75	44.18	40.85	36.83	35.44	33.15
T 11	33.45	44.06	43.98	43.66	43.78	44.06	43.93	39.81	36.86	32.98	31.53	28.05
B 11	30.22	41.8	41.65	41.2	41.29	41.35	40.92	37.36	34.65	30.05	28.27	26.12
T 12	34.56	46.45	46.32	45.96	46.08	46.26	46.13	42.15	38.31	33.62	31.59	29.36
B 12	31	42.62	42.36	41.91	41.98	42.09	41.91	38.25	35.15	31 55	28.81	27.01
T 13	29.93	38.93	38.66	38.23	38.26	38.22	37.81	35.18	33.35	30.61	28.74	27.5
B 13	25.51	35.69	35.4	34.91	34.97	34.92	34.62	31.61	29.73	27.53	26.06	24.97
T 18	33.8	41.51	41.3	40.91	40.68	40.5	38.52	37.17	36.54	34.09	32.59	31.35
B 18	28.61	40.02	39.75	39.32	39.38	39.41	39.02	36.3	34.83	32.33	30.91	29.78
T 19	31.96	45.87	45.54	45.14	45.17	45.19	44.48	42.2	39.91	37.01	35 81	34.29
B 19	35.25	48.24	48.02	47.64	47.64	47.64	47.26	45.17	42.23	38.61	37.52	36.15

Site: Letter prefix indicates position in stack (T = top; B = bottom)
First number indicates field site (1 = undrained; 2 = drained)
Second number indicates sampling point.

Table 4.3 Summary of water retention data.

		Volumetric soil moisture				Loss of soil moisture			Overall loss
		Sat'n	5 cm	50 cm	15 bar	sat'n - 50 cm	5cm - 50 cm	50 cm - 15 bar	
Drained field	T 21	50.93	46.05	45.97	28.06	4.9538	0.0755	17.91216	22.866
	B 21	49.24	44.8	44.45	28.6	4.7942	0.3478	15.84973	20.644
	T 22	42	40.41	39.78	27.27	2.2232	0.6352	12.50737	14.731
	B 22	43.48	42.63	41.47	30.25	2.0115	1.1645	11.22185	13.233
	T 23	47.91	47.26	45.9	32.23	2.0115	1.3611	13.6719	15.683
	B 23	51.8	50.76	50.04	36.95	1.7544	0.7108	13.0972	14.852
	T 24	47.99	45.96	45.83	29.05	2.1627	0.1361	16.77228	18.935
	B 24	47.61	44.43	44.15	29.29	3.4633	0.2874	14.85156	18.315
	T 25	53.6	45.52	44.69	27.77	8.9079	0.8318	16.92352	25.831
	B 25	56.34	47.52	46.41	29.49	9.9212	1.104	16.92352	26.845
	T 26	44.71	41.32	40.5	27.98	4.2044	0.8167	12.5225	16.727
	B 26	48.65	43.57	42.42	28.96	6.231	1.1494	13.46017	19.691
	T 27	48.28	48.47	47.44	34.09	0.8318	1.0284	13.3543	14.186
	B 27	50.03	47.59	46.75	33.15	3.2819	0.8469	13.59629	16.878
	Averages	48.75	45.45	44.7	30.22	4.0538	0.7497	14.47603	18.53
Undrained field	T 11	48.53	44.06	43.93	28.05	4.5976	0.121	15.87998	20.478
	B 11	43.75	41.8	40.92	26.12	2.8281	0.8772	14.80619	17.634
	T 12	49.64	46.45	46.13	29.36	3.5087	0.3176	16.77228	20.281
	B 12	46.63	42.62	41.91	27.01	4.7186	0.7108	14.89693	19.616
	T 13	42.86	38.93	37.81	27.5	5.0513	1.1192	10.31442	15.366
	B 13	39.47	35.69	34.62	24.97	4.8547	1.0738	9.648977	14.504
	M 14	47.1	45.13	44.12	32.15	2.9794	1.0133	11.96292	14.942
	T 15	41	39.94	38.64	26.54	2.3593	1.3006	12.09903	14.458
	B 15	39.5	39.03	38.07	26.42	1.4368	0.9679	11.64532	13.082
	T 16	46.38	45.42	44.89	33.91	1.4973	0.5293	10.97987	12.477
	B 16	45.13	44.72	43.96	32.2	1.1645	0.7562	11.76631	12.931
	T 17	50.85	47.32	46.31	30.08	4.5371	1.0133	16.22782	20.765
	B 17	51.36	47.05	45.72	30.55	5.6412	1.3309	15.16916	20.81
	T 18	41.76	41.51	38.52	31.35	3.2365	2.9945	7.168676	10.405
	B 18	37.93	40.02	39.02	29.78	-1.089	0.9982	9.240635	8.1517
	T 19	44.96	45.87	44.48	34.29	0.484	1.3914	10.19343	10.677
	B 19	47.14	48.24	47.26	36.15	-0.121	0.983	11.11598	10.995
	Averages	44.43	42.96	41.81	30.21	2.6251	1.1559	11.60211	14.227

Figure 4.1 Undrained field water retention (0.1 m depth).

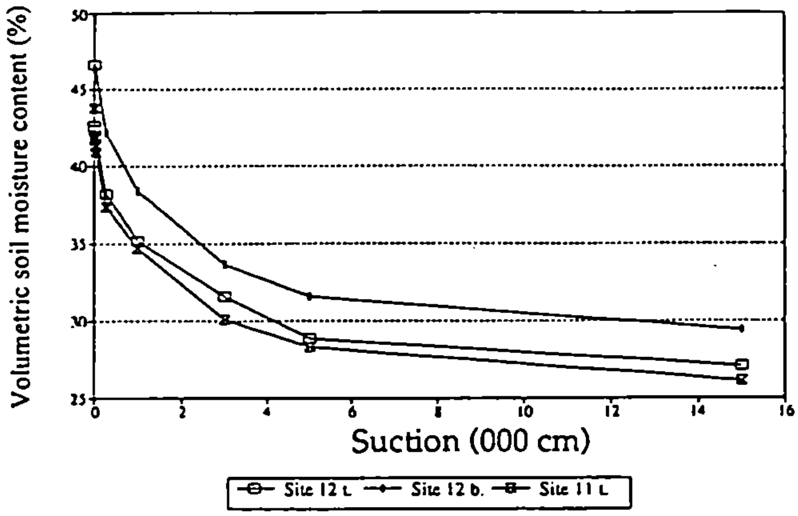


Figure 4.2 Undrained field water retention (0.2 m depth).

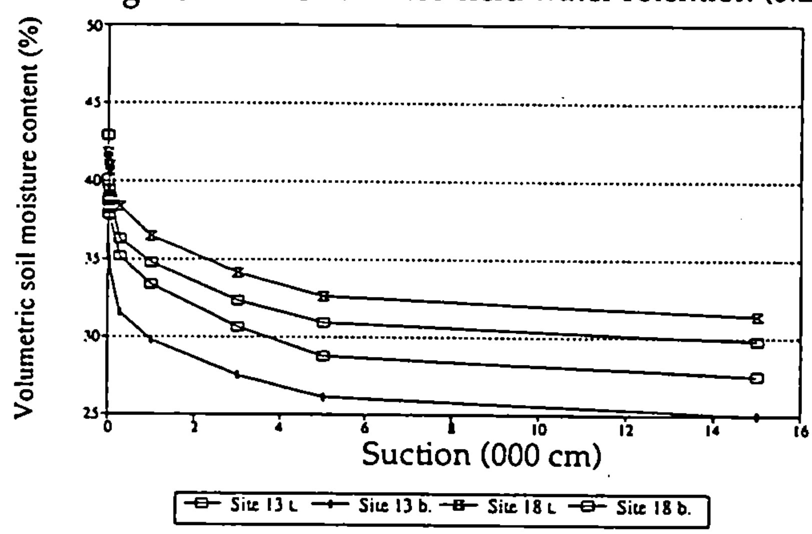
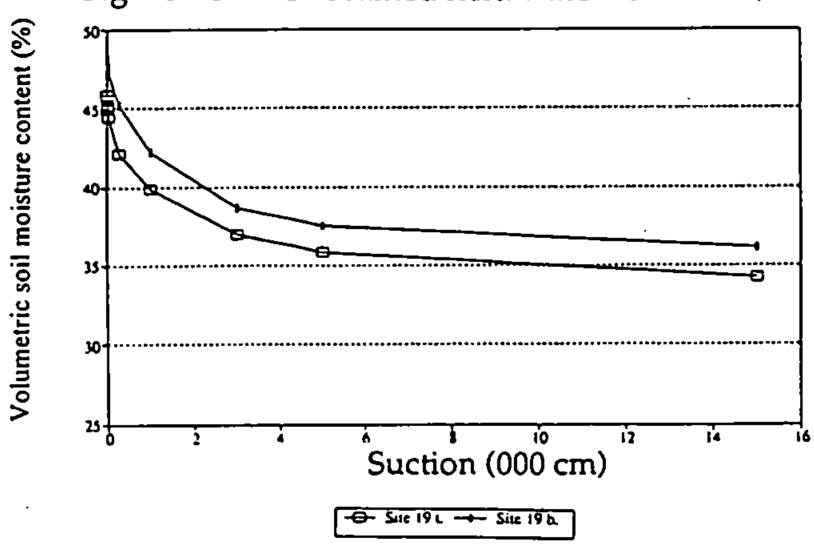


Figure 4.3 Undrained field water retention (0.4 m depth).



Water release data for the drained field are presented alongside those for the undrained field in Tables 4.2 and 4.3. Results are illustrated in Figures 4.4. to 4.6. Volumetric soil moisture content at saturation ranged between 42.0 and 56.3 %, averaging 48.8 %. At 50 cm H_2O mean soil moisture content was 44.7 % (cf undrained soil 41.9%), thus the air capacity was greater on the drained field (4.1 %). The available water capacity.was 14.5 % as soil moisture content decreased to 30.2 % at 15000 cm H_2O (Table 4.3).

Figures 4.4 and 4.6 show that drained field cores from 0.1 m and 0.4 m had high soil moisture contents (largely between 45 and 50 %) at saturation, while the cores from 0.2 m had lower values (less than 45 %). At 0.1 m, the slope of the water release curve was steep (Figure 4.4) as water content fell to 28.6 % at 15000 cm H_2O.

These results relate to the soil water characteristics observed at the Pitman core scale (Section 3.3.2). The data confirm that this soil is like others in the Hallsworth series, in that it retains a high volume of soil moisture at low suctions in all horizons (41.8 to 44.7 % at 50 cm H_2O). Mann Whitney analysis showed that samples from undrained and drained fields were significantly different (99.9 % significance). Volumetric soil moisture content at saturation, as well as air capacity and available water capacity were all higher on the drained field (Table 4.4). However, there were similarities between the soil water characteristics of the two fields. First, most available water drained at low suctions (2.5 to 5 cm H_2O). Second, averaged soil moisture contents at 15000 cm H_2O were the same for both fields (30.2 %).

Results for both the undrained and drained soils concur with those of Hogan (1978), indicating the small proportion of drainable pores (Table 4.4). The retained water capacity at 50 cm H_2O suction was greater than the plastic limit (Hogan, 1978) and so the need for an effective drainage system is demonstrated. Following winter rainfall, even with efficient drainage, a period of drying is required before stocking can occur without poaching damage (Hallard, 1988). The analyses indicate that mole drainage increases the water retention capacity of the topsoil even at the Pitman core scale.

Figure 4.4 Drained field water retention (0.1 m depth).

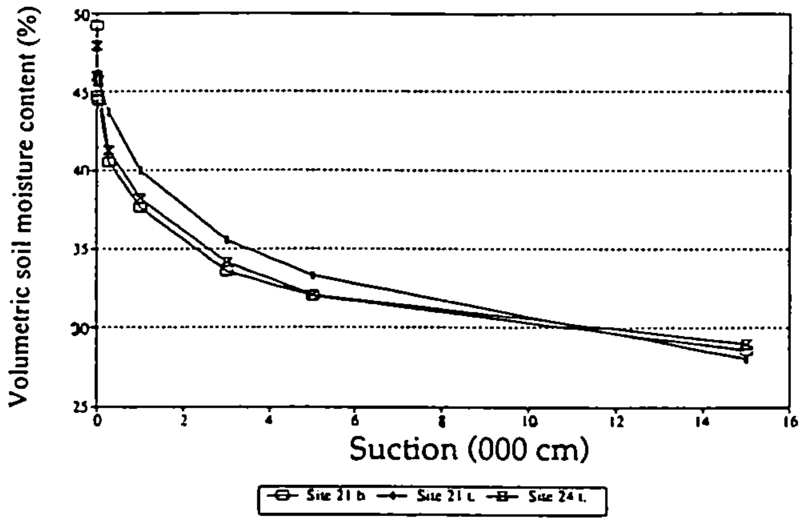


Figure 4.5 Drained field water retention (0.2 m depth).

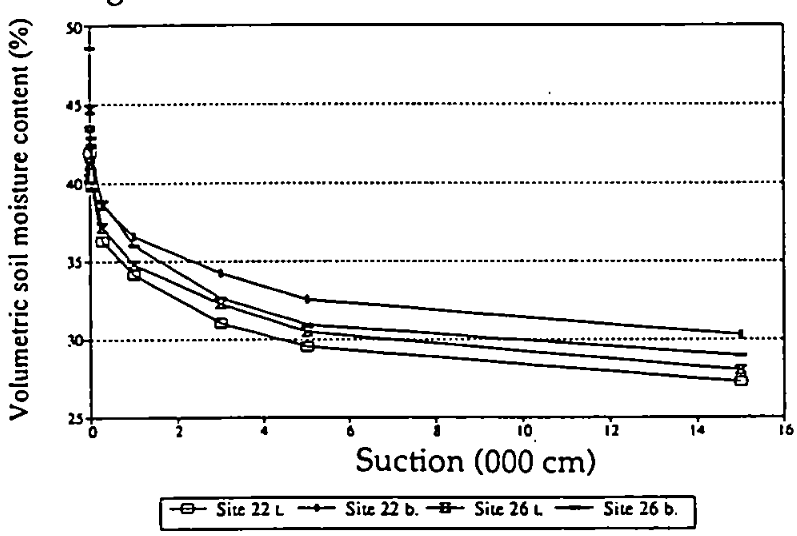
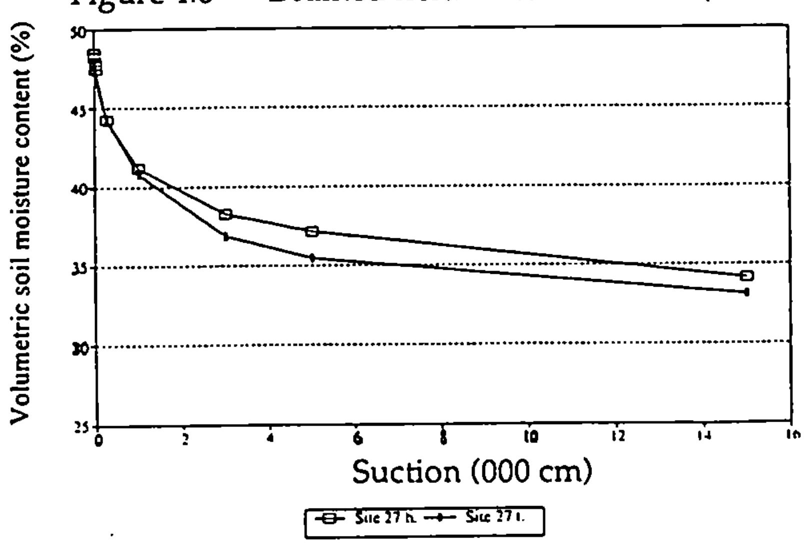


Figure 4.6 Drained field water retention (0.4 m depth).



This may reflect an improvement of topsoil structure as a result of increased aeration, faunal activity and organic matter content in the topsoil (Sections 1.2, 4.2.4 and 4.2.7; Gilbey, 1986), all essentially increasing porosity in the upper range of pore sizes. A consideration of water transmission at grosser scales is undertaken in Section 4.2.5, where the effect of fissuring by mole drainage is incorporated.

Table 4.4 Pore volumes at critical suctions (%).

Water category	Undrained	Drained	Hogan (0 - 27 cm)	Hogan (27 - 66 cm)
θ at saturation	44.4	48.8	63.3	50.4
θ at 50 cm H_2O	41.8	44.7	51.3	45.7
θ at 15000 cm H_2O	30.2	30.2	29.2	29.1
Air capacity	2.6	4.1	12.0	4.8
Available water	11.6	14.5	22.1	16.6

Figures 4.2 and 4.5 shows that results for 0.2 m depth on both fields were similar. This may reflect the influence of the boundary of the Ag and B horizons and the associated hydrological conditions (i.e. seasonal waterlogging indicated by manganese accumulations; Table 2.2).

4.2.3 Bulk density and soil strength.

In this experiment, bulk density was calculated for all the cores used to determine soil water retention curves (Section 3.3.2), results are detailed in Table 4.5.

Table 4.5 Mean bulk density data (g cm^{-3})

Depth (m)	Undrained field	Drained field	Hogan (1978)
0.1	1.19	1.10	—
0.2	1.49	1.30	0.99
0.4	1.45	1.47	1.31
Profile mean	1.38	1.29	1.20

On the undrained field, mean bulk density was 1.19 and 1.45 g cm^{-3} for the surface and subsoil, respectively. These values compared with 1.10 and 1.47 g cm^{-3} on the drained field. Results are consistent with those recorded by Hogan (1978; Table 4.5) and Armstrong and Garwood (1991) at Rowden Moor. The higher bulk densities of the undrained field (Table 4.5) confirm the soil water data above. On both fields, lower bulk densities for the surface are due to their improved structure and higher organic matter contents (Section 4.2.7). Figure 4.7 illustrates increasing bulk density with depth and the consequence of increasing density at 0.2 m is demonstrated by the decrease in saturated soil moisture content data.

Figure 4.7 Bulk density and soil moisture content.

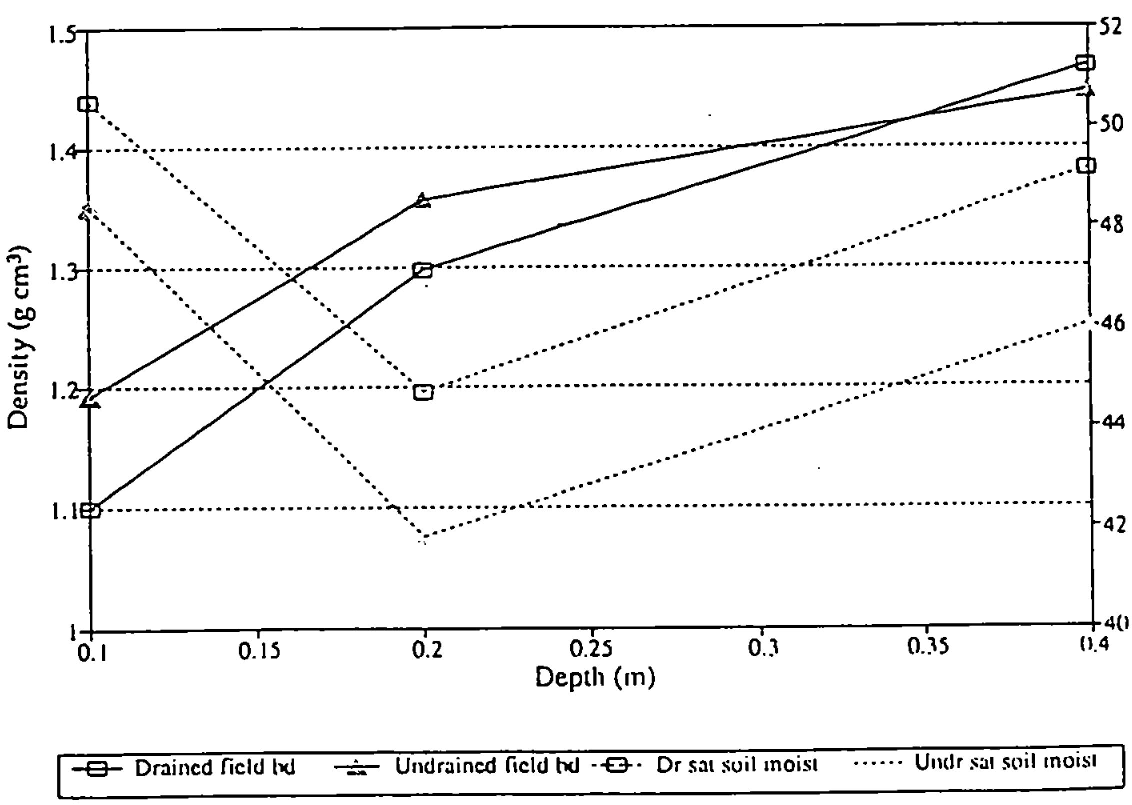
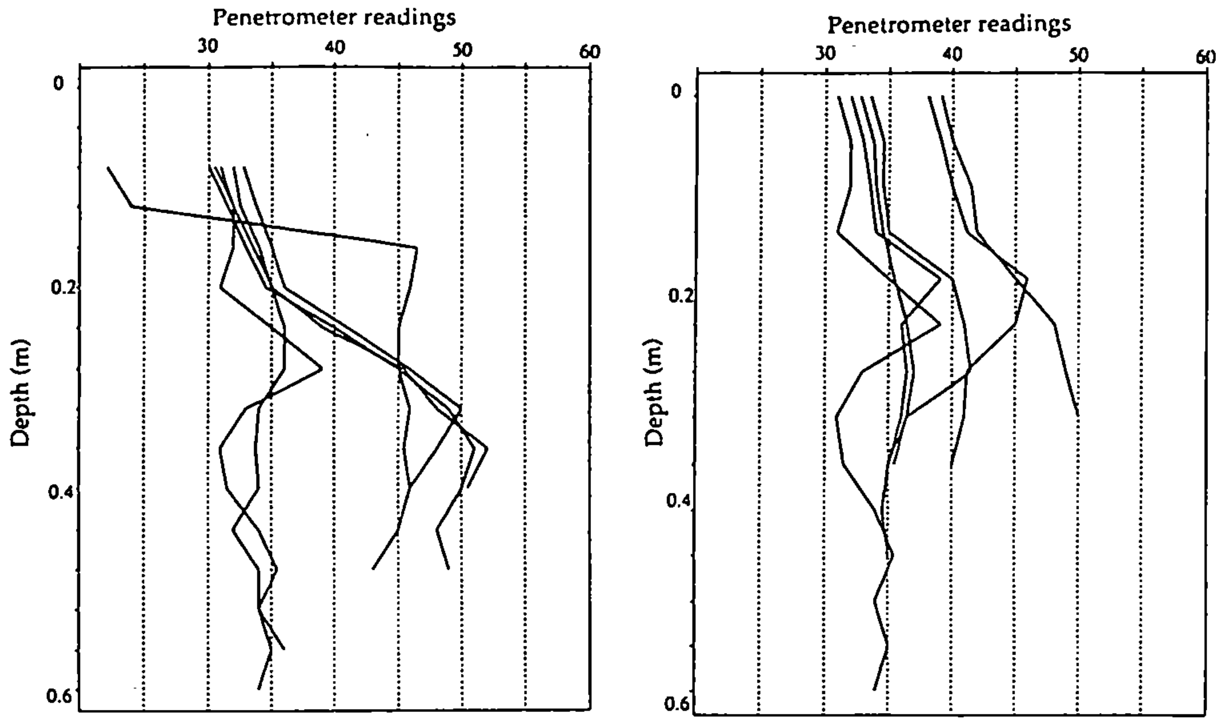


Figure 4.8 shows soil strength results for both fields. On the undrained field, resistance increased with depth. A distinct increase in values at 0.2 to 0.24 m depth is consistently indicated. This coincides with the boundary between the A and B horizons (Table 2.2), also identified in bulk density and saturated soil moisture content profiles (Figure 4.9). Resistance was relatively uniform until 0.2 m depth when soil strength became more variable.

Figure 4.8 Soil strength profiles.



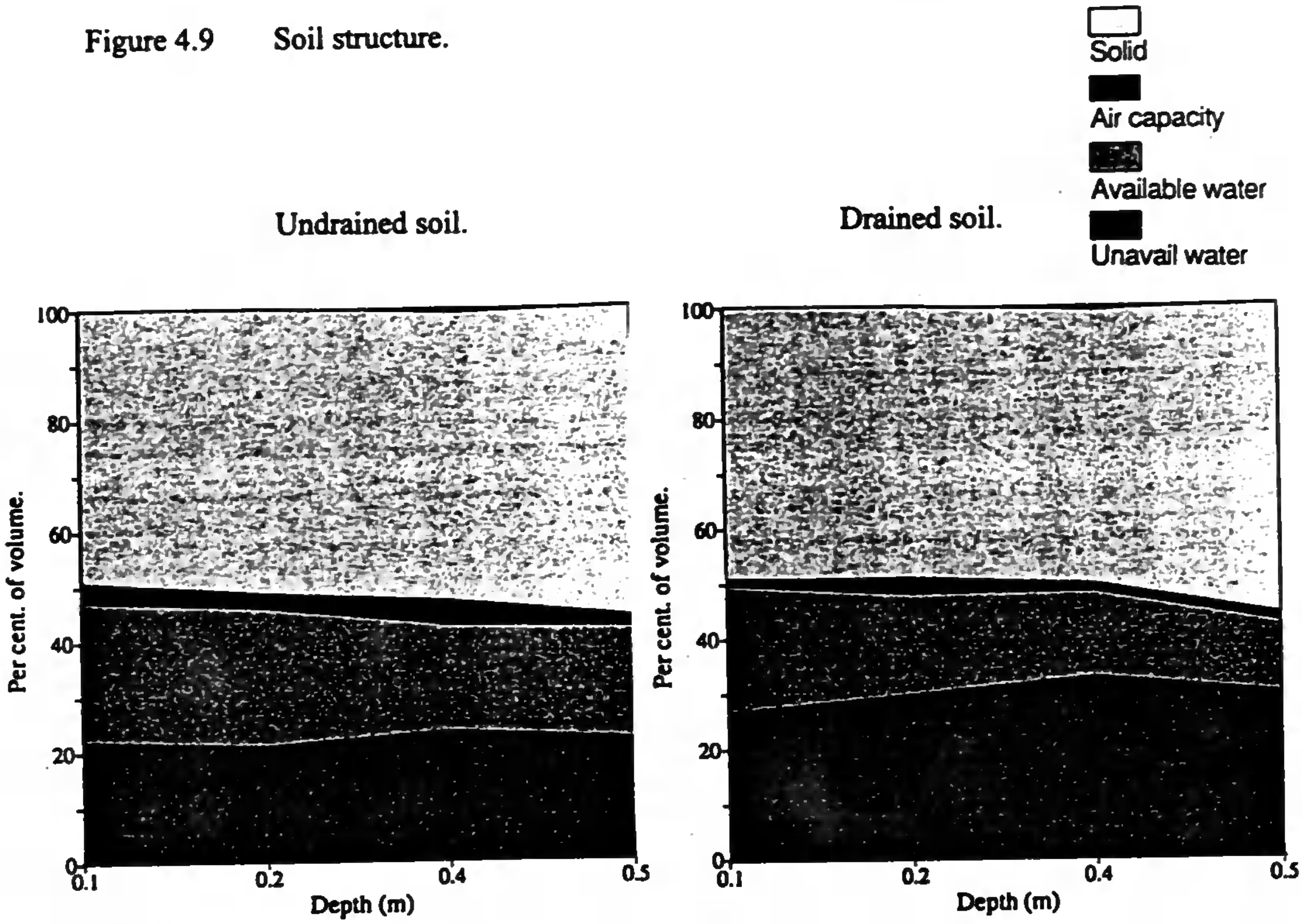
a) Undrained field. b)Drained field.

A similar pattern is exhibited by the drained field results. However, the increase in resistance around 0.2 m was less marked. Readings taken on this field were more frequently aborted due to contact with stones resulting in pressure exceeding the maximum load. Therefore, results were more difficult to obtain with depth. So, it is possible to see that soil strength was less variable in the upper 0.2 m on the undrained field, but it is not possible to compare values for lower depths. Penetration resistance was slightly higher on the drained field in the upper soil. This could be caused by soil structural changes resulting from drainage i.e. fissuring or lower antecedent soil moisture content creating more resistance to penetration, or it could be due to more stones.

4.2.4 Soil structure and porosity.

Soil structure is categorised in Figure 4.9 according to Thomasson's classification. Despite variability between individual soil water release curves, averaged results revealed that the soils on both fields had poor structures.

Figure 4.9 Soil structure.



4.2.5 Saturated hydraulic conductivity (K_{sat}).

This section outlines and analyses the K_{sat} results, first for the field experiment, then for the laboratory experiment.

In the field, the ring permeameter technique was employed, using cores of 0.3 m diameter (Talsma *et al.* 1979; Section 3.3.3). Saturated hydraulic conductivity results for undrained and drained cores are presented in Table 4.6. The K_{sat} value for the undrained field averaged 79.1 mm hr^{-1} in the surface and 42.0 mm hr^{-1} in the subsoil (Table 4.6). On the drained field, K_{sat} was about three times higher (223.0 and 149.4 mm hr^{-1} for the upper and lower depths, respectively).

As K_{sat} data tend to be skewed, the $\log_{10}$ mean value is more appropriate for the comparison of data than the arithmetic mean (Baker, 1978; Bonell *et al.*, 1983; Ternan *et al.*, 1987). Two drained plot results exercised a great influence over the $\log_{10}$ means, so samples 5 and 10 have been excluded from K_{sat} analysis. However, the $\log_{10}$ means varied little from arithmetic means (Table 4.7). Looking at the two measures, it can be seen that on both fields K_{sat} decreases with depth. This is expected as clay content (Table 2.2) and bulk density (Section 4.2.3) increase with depth and soil moisture decreases with depth (Section 4.2.8).

Table 4.6 Saturated hydraulic conductivity of undrained and drained soils.

Depth (m)	Undrained K_{sat} mm hr^{-1}	Drained K_{sat} mm hr^{-1}
0-0.1	93.79	502.70
	140.96	178.81
	71.33	154.86
	41.68	374.10
	34.69	54.55
	122.73	112.10
	95.23	344.44
	51.58	101.60
	65.04	99.85
	74.15	86.90
	—	138.33
0.1-0.2	55.82	173.50
	35.52	181.53
	39.75	93.09
	37.08	—

Table 4.7 Averaged hydraulic conductivity results mm hr^{-1} (m day^{-1})

Depth (m)	n	Arithmetic mean	Log_{10} mean	Range
UD 0-0.1	10	79.10 (1.90)	72.28 (1.73)	34.70-141.00 (0.83-3.38)
UD 0.1-0.2	4	42.04 (1.01)	41.35 (0.99)	35.52-55.82 (0.85-1.34)
D 0-0.1	9	223.00 (5.35)	187.15 (4.49)	99.80-502.70 (2.39-12.06)
D 0.1-0.2	3	149.37 (3.58)	143.13 (3.43)	93.09-181.53 (2.23-4.36)

where: 'UD' is undrained field and 'D' is drained field.

Table 4.7, shows that at the lower end of the ranges, K_{sat} was *c.* 35 and 96 mm hr^{-1} on the undrained field and drained fields. K_{sat} is most variable, within and between fields, in the upper 0.1 m. The ranges of data illustrate the variable nature of K on the site, and agree with the findings of Nielsen *et al.* (1973) and McDonnell (1988). These workers measured conductivity values which varied by two orders of magnitude within small catchments (*c.* 1 ha). Chappell (1990), using the ring permeameter technique, found that K_{sat} at 0 - 0.1 m varied by one order of magnitude, i.e. it ranged between 37 and 390 mm hr^{-1} for a grassland soil. Further, as in this study, Chappell's results were skewed towards the upper end of the range. Armstrong and Garwood (1991) used the encased gypsum block technique (Baker, 1978) to characterise conductivity to 0.3 m and the single auger hole technique (Bouma and Dekker, 1981) for lower depths. Permeability was very low on the undrained field, leading to waterlogging. Armstrong and Garwood (1991) did not provide K_{sat} values for drained fields but noted that they were higher. Armstrong (1986 b) argued that a low proportion of drainable pores, high clay content and impermeable subsoil result in low saturated hydraulic conductivity of 10 mm day^{-1}, based on water balance calculations.

The relationship between K_{sat} and soil properties such as bulk density and organic matter was investigated (Table 4.8). This analysis showed that little of the variance in K_{sat} could be explained by these soil properties.

Table 4.8 Relationship between K_{sat} and soil properties.

Soil property	Undrained field R^2	Drained field R^2
Bulk density	1.2 %	3.3 %
Soil moisture	4.5 %	11.4 %
Organic matter	1.8 %	30.5 %
% > 2 mm	21.9 %	0.5 %

Multiple regression was carried out to assess how much combined influence soil properties had on K_{sat}. R^2 values for the upper 0.1 m were:

Undrained $R^2 = 44.5$ % Drained $R^2 = 50.1$ %

The multiple regression values show that the soil properties measured accounted for 50 % or less of the variation in K_{sat}. Despite an inability to tie K_{sat} to individual soil properties, multiple regression analysis suggests that the characteristics act together to determine K. Other researchers have suggested that conductivity is related to bulk density, organic matter and activity of soil fauna. For example, Bonell *et al.* (1983) found that bulk densities in South Creek catchment, Queensland increased with decreasing permeability. However, this parameter did not totally explain the observed variations in hydraulic conductivity. For Rowden, organic matter content, when regressed with hydraulic conductivity ($\log_{10}K$), explained about 30 % of the variation in K_{sat} values on the drained field. Drainage increases the rooting depth, distribution of organic matter and the presence and activity of soil fauna on the drained field (Gilbey, 1986). Baker (1978) found that as saturated conditions are approached, soil macropores, including structural cracks and worm channels tend to dominate moisture flow. The presence of earthworm channels was noted on both sites but it was difficult to determine if there was any significant difference in frequency.

McDonnell (1988), indicated that results for the Guelph permeameter technique, may not be indicative of true saturated hydraulic conductivity, but provide an illustration of field saturation. The same is true for the ring permeameter procedure, thus a laboratory experiment was devised to determine changes in K_{sat} in six cores, following wetting over fourteen days (Section 3.3.3). On average, 20 litres of water was applied to each core per day. After the fourth test (eleven days) cores were placed in a bath of water for three days, after which a final permeability test was conducted. Soil samples were extracted from each core to determine particle size distribution, soil moisture content, bulk density and organic matter content.

Figures 4.10 and 4.11 demonstrate the K_{sat} results. Conductivities of cores extracted from the undrained field remained lower than those from the drained field. However, even between cores from the same field there were large differences, especially for the drained field. Variability was greatest at the outset of the experiment, as wetting-up proceeded there was a general trend of decreasing K_{sat} as clays swelled and cracks closed. This behaviour was observed for all but two of the drained cores. The slight increase in K_{sat} experienced by these two cores may have been due to disturbance during handling.

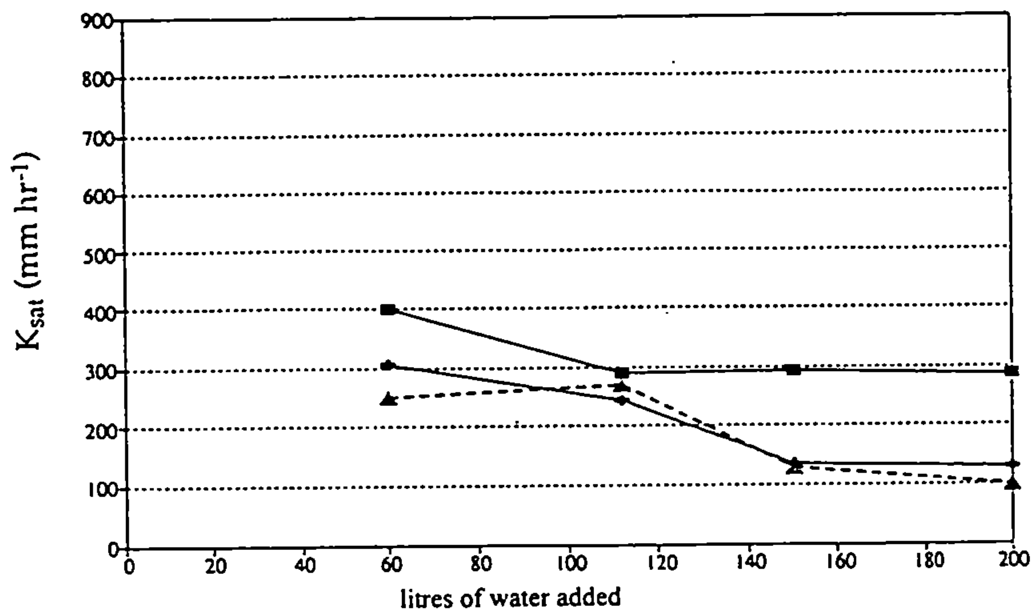


Figure 4.10 Laboratory determination of K_{sat}, undrained field.

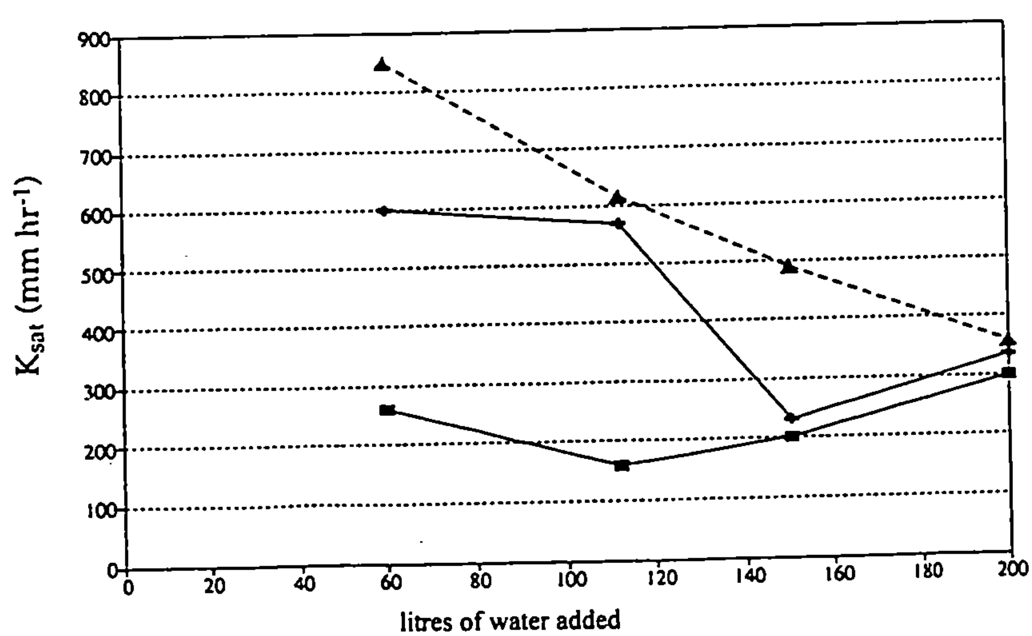


Figure 4.11 Laboratory determination of K_{sat}, drained field.

Section 3.3.3 discusses problems associated with this technique, but since these problems were common to all cores results remain useful for comparative purposes. Conductivity was measured only for vertical flow, whilst in most situations, water flow in the unsaturated zone is predominantly vertical (Ragab and Cooper, 1990). It is, therefore, commonly assumed that horizontal components of unsaturated water flow can be ignored, enabling considerable simplification of flow equations and computation. In the saturated zone, however, flow is often predominantly horizontal. The flow properties of soil are often anisotropic (i.e. the vertical hydraulic conductivity is different from that in horizontal directions).

When examining these data the effect of scale must be considered. The Pitman core samples were relatively small, and sampling may have avoided many of the structural cracks and mole drain related fissures which may have been present in these larger cores. The differences in the initial K_{sat} values obtained from the drained field cores can be attributed to the degree of fissuring and cracking caused by mole drainage and, more extensively, improved soil structure and texture. The decrease in variability of hydraulic conductivity with progressive wetting for both fields may reflect closure of cracks and high soil moisture contents. However, Emerson (1959) noted that it could take months for cracks to close completely.

K_{sat} values will vary through time, for example, drying of soil during the summer results in soil shrinkage and increased conductivity. The drained soil was drier, with more cracking. This would explain the higher K_{sat} values obtained. The experiment, conducted at this time of the year, demonstrated the effectiveness of macropores in increasing the permeability of the upper horizon especially on the drained field. In addition, it illustrated the substantial volumes of water and periods of time required to wet even a small volume, of what is the most permeable horizon of the soil.

Measurement of permeability showed that drainage had a positive effect on saturated hydraulic conductivity. Fissures and cracks introduced by mole drainage increased subsurface flow routes and connectivity and provided a sink for subsurface water.

4.2.6 Particle size distribution.

Soil texture is described in Section 2.5.2 and Figure 2.4 where it is classified as lying in the silty clay or clay textural categories. The survey conducted by Harrod (1981) illustrated the homogeneity of soil across the site. Within the profile, the topsoil is the most uniform portion, and has an unusually high clay content of 48 %. This has obvious implications for surface wetness and moisture retention values (Section 4.2.2). Soil samples were analysed from the upper 0.1 m and 0.1 to 0.2 m depth for both fields and the lysimeter (Table 4.19). Results were less clustered (Figure 4.12) but indicated that the soil was generally a clay (except at 0.2 m on the undrained plot: silty clay). The Analysette laboratory technique was used and this gives good replication of the Andreasan pipette technique (Appendix 1).

Figure 4.12 Soil textural classification.

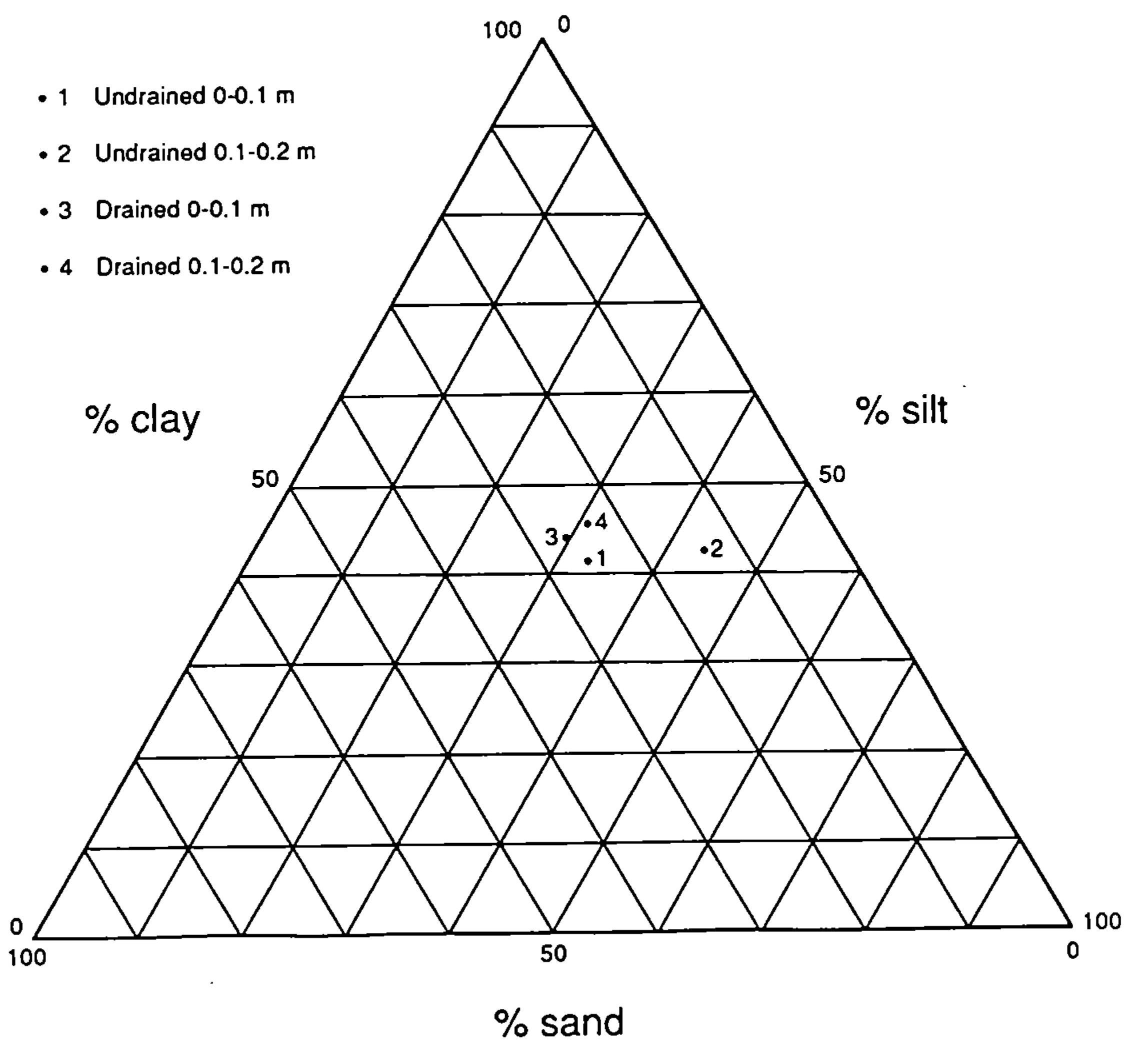


Table 4.9 Mean particle size data.

Classification	Undrained 0-0.1 m	Undrained 0.1-0.2 m	Drained 0-0.1 m	Drained 0.1-0.2 m
Coarse sand >2 mm (%)	0.9	1.7	2.9	12.3
Sand 2 mm - 63 μm (%)	25.5	14.3	26.7	23.3
Silt 63 μm - 2 μm (%)	33.9	43.5	29.5	31.5
Clay <2 μm (%)	40.6	42.2	43.8	45.2

The trends on each site were of decreasing sand content accompanied by increasing proportions of the smaller fractions, with depth. Comparison between sites showed that the drained soil contained a lower proportion of silt in the fine earths fraction (91 % significance level). Further statistical analysis (Table 4.10) indicated that the proportion of particles greater than 2 mm was significantly higher on the drained field (95 % significance).

Table 4.10 Mann Whitney U Test analysis of particle size data.

Particle Class	Hypothesis	Significance Level
% > 2 mm	D > UD	0.0004
% Sand	D > UD	0.3693
% Silt	D < UD	0.0941
% Clay	D > UD	0.1820

where: 'UD' is undrained field and 'D' is drained field.

4.2.7 Organic matter content.

Decreasing organic matter content was associated with increasing depth on both fields (Table 4.11). The decrease was slightly greater on the undrained field, which may be explained by restricted rooting depth and structural variation due to waterlogging. However, the difference in organic matter content between sites was not significant. Correlation demonstrated a negative relationship of -0.519 between organic matter and bulk density (Figure 4.13) significant to 95 %. Soil organic matter content is strongly related to aggregate stability and structure (Tisdall and Oades, 1982; Chaney and Swift,

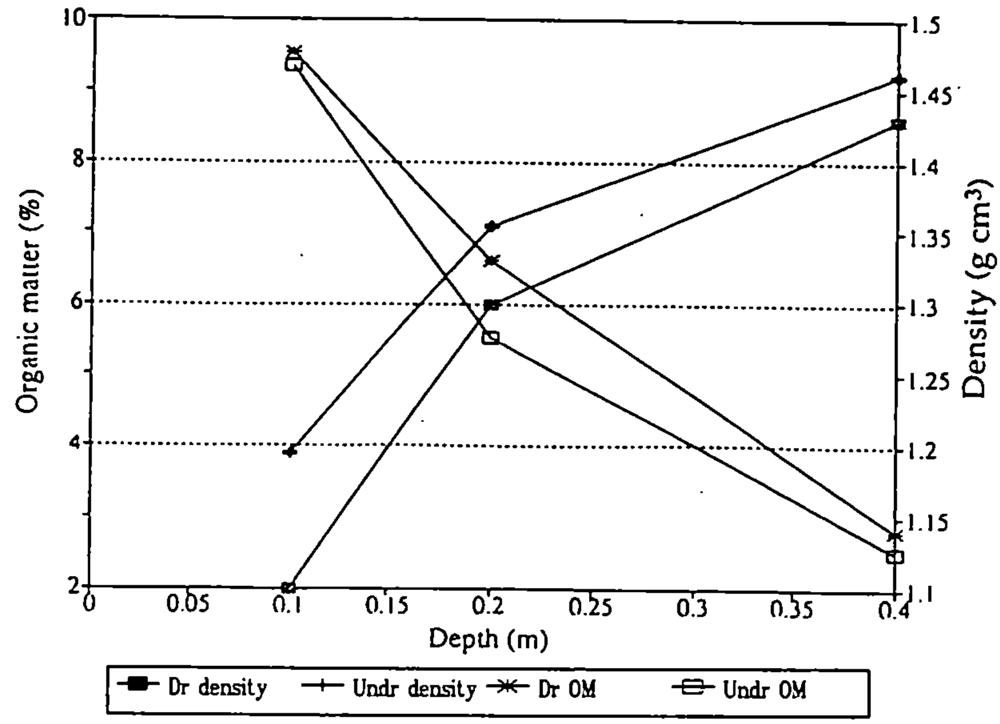


Figure 4.13 Organic matter and bulk density.

Table 4.11 Mean organic matter content (%)

Soil property	Undrained 0 - 0.1 m	Undrained 0.1 - 0.2 m	Drained 0 - 0.1 m	Drained 0.1 - 0.2 m
Organic matter	9.34	5.52	9.54	6.61

4.2.8 Soil moisture content at saturation.

The soil moisture contents presented in Table 4.12, were those prior to water release analysis, i.e. under saturated conditions. Therefore, this variable was a function of the pore volume of the soil. Mean values for the upper 0.1 m indicated that the drained field had a greater porosity.

Table 4.12 Mean volumetric soil moisture content (%) at 0 cm H_2O (total of air capacity and available water).

Depth (m)	Undrained field	Drained field
0.1	44.46	50.56
0.2	41.82	44.70
0.4	46.05	49.16
Profile mean	44.11	48.11

On both fields, soil moisture content decreased from 0.1 to 0.2 m depth, but increased at 0.4 m. This suggests that pore space decreased at the A/Bg boundary and is consistent with the increased bulk density values with depth (Armstrong and Garwood, 1991) and soil descriptions (Harrod, 1981). Spearman's Rank Correlation analysis showed a negative correlation between saturated soil moisture and bulk density on both fields, significant to 95 % (undrained field $r = -0.394$; drained $r = -0.508$).

4.3 Summary.

The soil on Rowden Moor is clay rich and poorly structured. The distinct horizon boundary at approximately 0.2 m depth was consistently evident in all profiles. Analyses showed that bulk density increased with depth in all profiles and was inversely proportional to organic matter and soil moisture contents. Further, bulk density was lower on the drained field, where higher porosity was observed throughout the range of pore sizes. This was reflected in increased saturated hydraulic conductivity on the drained site.

Results were generally in agreement with others for this site (Hogan, 1978; Harrod, 1981; Hallard, 1988; Armstrong, 1986 b). However, the significance of the scale and method adopted was consistently noted (e.g. for soil water retention, K_{sat} and soil moisture). The implications of the results for water movement at the hectare scale are explored in Chapter 5.

CHAPTER 5: FIELD SCALE RESULTS: ANALYSIS OF RAINFALL-RUNOFF CHARACTERISTICS.

5.1 Introduction.

Analysis of hydrology at the field scale is required to infer the significance of surface and subsurface water pathways, for example, the relative contribution of overland flow versus subsurface flow in generation of field discharge. This will be achieved using a qualitative and quantitative approach to flow data analysis. The water balance, annual and monthly runoff coefficients, seasonal drainage trends and flow duration characteristics will be described to characterise the field hydrology. In addition, the different flow characteristics of the undrained and drained fields will be compared prior to the more detailed study of soil water in Chapter 6.

The water balance is an expression of the relationships between various components of the hydrological cycle including precipitation, evapotranspiration and runoff, shown in Figure 5.1. This budget approach provides the structure for the following discussion. The field results for components are examined in turn, before drawing these together in the calculation of the water balances in Section 5.5.

Chapters 5 and 6 focus on the hydrological behaviour of the fields and plots between 1.10.90 and 31.3.91 (Section 5.3.1) hereafter referred to as the study period or drainage season. This chapter provides hydrological information required to compare the 1990-91 results with those for other drainage seasons at Rowden (Armstrong *et al.*, 1984; Hallard, 1988; Armstrong and Garwood, 1991), as discussed in Section 5.5.

The water balance components of undrained and drained fields are represented by equations 5.1 and 5.2 respectively:

$$R - Et = (Of + If) \pm dP \pm S \qquad \text{[Eqn. 5.1]}$$

$$R - Et = (Of + If) + Df \pm dP \pm S \qquad \text{[Eqn. 5.2]}$$

where:	R	=	precipitation inputs
	Et	=	evapotranspiration losses
	Of	=	overland flow losses
	If	=	shallow subsurface (interflow) losses
	Df	=	drainflow losses
	dP	=	deep percolation-water entry from depth
	S	=	change in soil water storage

These water balance components are illustrated in Figure 5.1. Results for the meteorological inputs and outputs are described in Section 5.2. Drainage design dictated that overland flow and shallow subsurface flow were measured collectively as outlined in Section 3.2.4, and labelled 'surface flow', for the undrained field and 'non drainflow', for the drained field (Section 5.3).

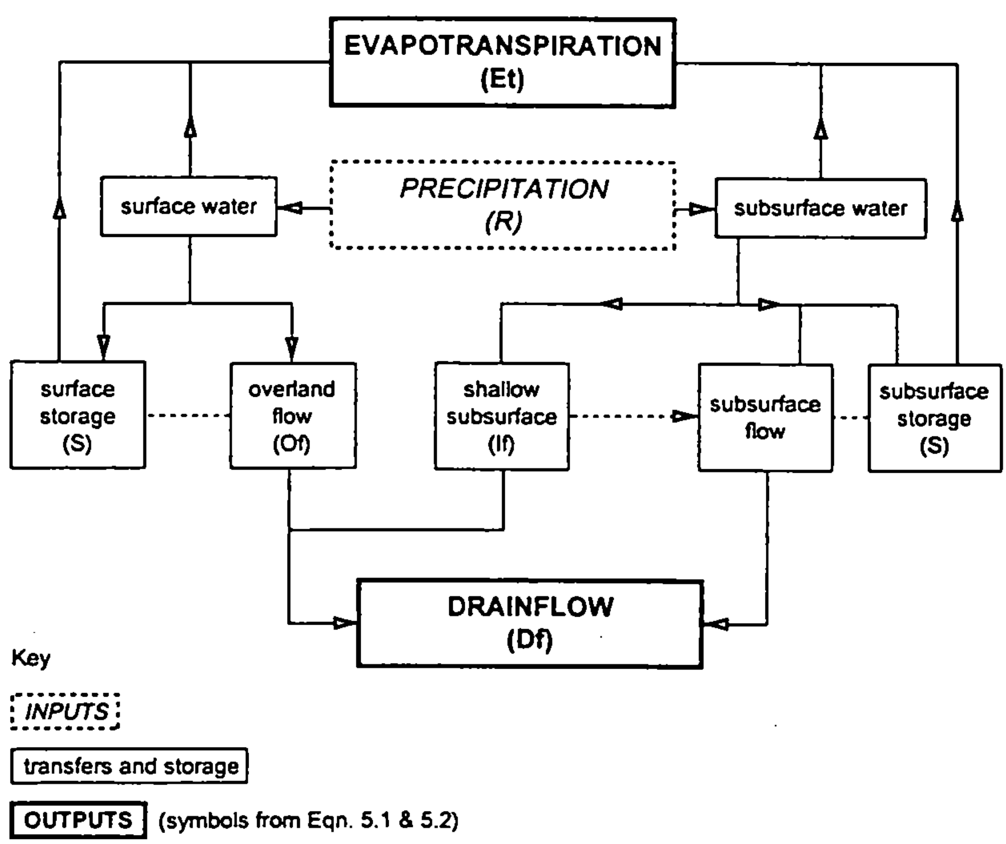


Figure 5.1 Water balance components.

The amount of water stored in the soil (Section 5.4) is an important parameter in the water balance equation. However, it is difficult to assess for a clay soil where measurement is time consuming (Section 3.3.8) and subject to large problems of spatial variability. Temporally, variation in soil water storage is most problematic during storm events. Spatially, a perched water table may exist, or drainage can complicate soil water distribution (Hillel, 1980; Armstrong, 1983). However, the T.D.R. grids which were established in the two fields (Section 3.3.8) made it possible to monitor soil moisture contents to a fine spatial resolution and on occasions to a fine temporal resolution (as discussed in Chapter 6).

When calculating water balances it is preferable if there is no leakage from the site as this is difficult to quantify. Piezometric data collected at the site (Armstrong *et al.*, 1984; Hallard, 1988) indicated that deep percolation was unlikely. Indeed, Armstrong *et al.* (1984) suggested that the possibility of water entering some of the fields (Section 2.4) was more likely, although this was not the case for the fields under consideration here. Monthly runoff coefficients (Sections 5.3.2; 5.3.7) and those described by Hallard (1988) suggested that the amount of water entering the fields in this manner was small and unlikely to have a direct impact on runoff quantities. This may, however, have been significant over limited areas as higher antecedent soil moisture content could result in different hydrological regimes. Section 5.4 discusses this in more detail.

5.2 Meteorological inputs and outputs.

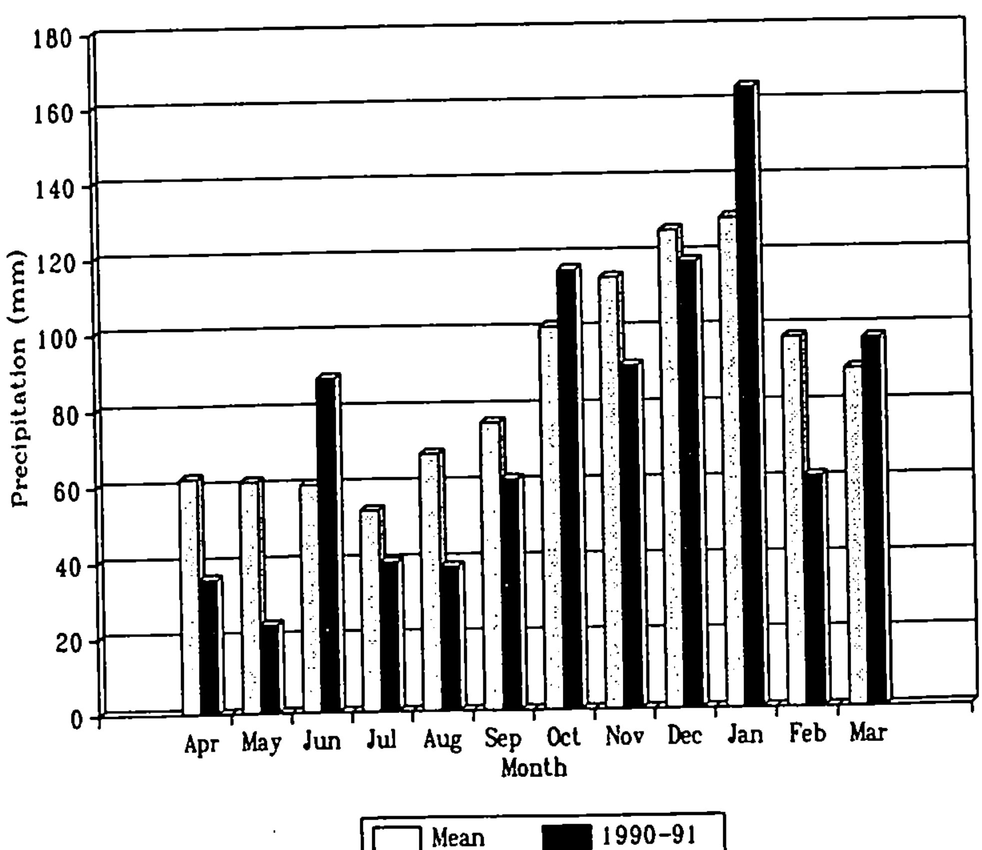
5.2.1 Precipitation inputs (R).

From April 1990 to March 1991, the precipitation total was 929.3 mm which was 89.8 % of the 30 year average (Figure 5.2 and Table 5.1). Summer rainfall was slightly less than average and winter rainfall higher than usual. There was a low volume of precipitation between April and September 1990 (inclusive) when 284.4 mm of precipitation fell compared with a 30 year average of 379.3 mm (i.e. 75.0 %). October rainfall was higher than average (115 %). During the period of drainflow (November to March) rain was close to the 30 year mean (95.4 %). However, this general similarity in rainfall volumes during the drainage season hides large monthly variations (Figure 5.2 and Table 5.1).

Table 5.1 Precipitation (mm) during study period and 30 year mean (Section 2.2, Table 2.1)

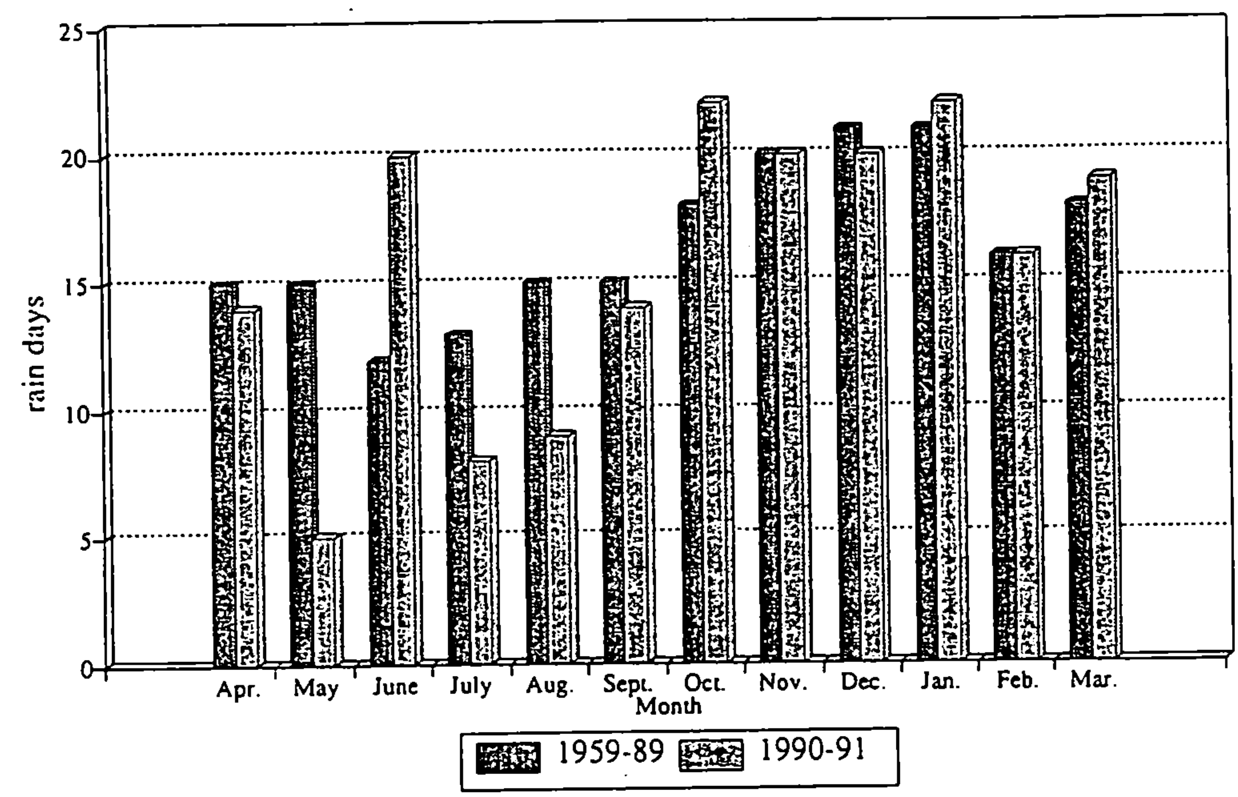
Month	1990-91	30 year mean	study period rain % of 30 yr average)
April	35.5	62.1	57.2
May	23.1	61.1	37.8
June	87.8	59.8	146.8
July	39.3	52.9	74.3
August	37.9	67.8	55.9
September	60.8	75.6	80.4
October	115.9	100.8	115.0
November	90.2	113.6	79.4
December	117.6	125.9	93.4
January	163.6	129.1	126.7
February	60.7	97.2	62.4
March	96.9	88.8	109.1
Totals	929.3	1034.7	89.8

Figure 5.2 Precipitation: 30 year mean and study period. North Wyke, Devon.



There were 119 rain days during the study period, a value close to the 30 year average of 201 (Table 2.1). A lack of correlation between the number of rain days and total precipitation was notable in 1990-91. For example, October and January both had 22 rain days, but while October was relatively dry, January was particularly wet (Figures 5.2 and 5.3; Table 5.1). The number of rain days in January does not have a direct link to the total precipitation, rather, the high precipitation total was related to the greater magnitude of storms.

Figure 5.3 Rain days: 30 year mean and study period, North Wyke, Devon.



A 'storm' can be arbitrarily defined as occuring when more than 10 mm of precipitation falls within 24 hours. Based on this definition, there were twenty five storms between 1.10.90 and 31.3.91 with most occurring from October to January (Figures 5.4 and 5.5). Four storms occurred between 25.10.90 and 29.10.90, four between 19.11.90 and 24.11.90, nine between 20.12.90 and 9.1.91 and three occurred between 4.3.91 and 8.3.91 (Figure 5.5). These four periods of heavy rainfall account for twenty three of the twenty five storms, more than one third of the rainfall (367.8 mm) but covered only thirty seven days, or one fifth of the drainage season. Substantial runoff was generated from both fields at these times (Sections 5.3.7 and 5.3.8).

Figure 5.4 Summary of distribution of storm events.

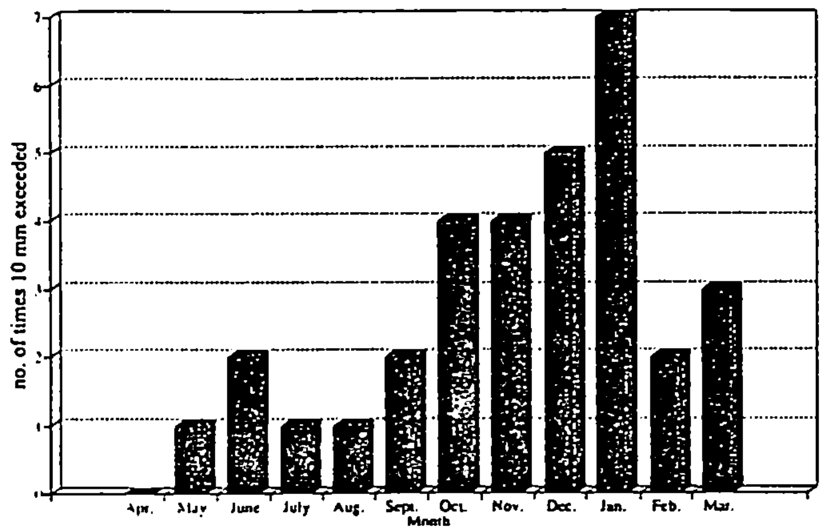
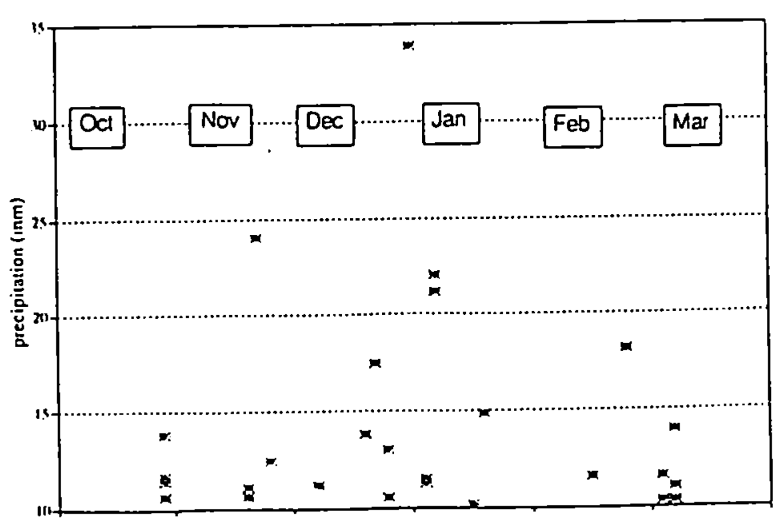


Figure 5.5 Magnitude and distribution of storm events.



5.2.2 Evapotranspiration losses (Et).

Total evapotranspiration losses for April 1990 to March 1991 were 379.2 mm (Table 5.2) and for the drainage season (October 1990 to March 1991) were 97.2 mm. Such losses were considerably higher than average values (1959-85) and those for 1984-85 (Table 5.2). The variation between results is partially explained by the different methods of calculation employed. The 1990-91 results were calculated using an evaporation pan and by applying a correction factor for 'land under a full crop cover' to results (Penman, 1952; Section 3.2.3). Hallard (1988) used the M.A.F.F. estimates of potential evapotranspirational losses for Agro-Met Area 43N (M.A.F.F., 1976) which are not site specific (being relevant for a large area).

Table 5.2 Monthly evapotranspiration losses (mm) for study period compared with 1984-85* and 1959-85*.

Month	(1959-85)	(1984-85)	(1990-91)
Sept	42	47	58.4
Oct	20	23	31.2
Nov	4	0	10.9
Dec	0	0	5.8
Jan	0	0	5.9
Feb	8	0	10.4
Mar	32	0	33.0
Apr	49	49	46.7
Total	155	119	202.3

*source: Hallard (1988).

5.2.3 Net precipitation (R-Et).

Net precipitation was determined by deducting evapotranspiration from gross precipitation and represents the amount of water draining through the soil. Water which infiltrates the soil surface and is temporarily stored in the upper soil horizons before being abstracted by plants or evaporation has wider hydrological significance. It is part of antecedent soil moisture content which is an important factor in determining infiltration rates, soil hydraulic conductivity and operative water pathways.

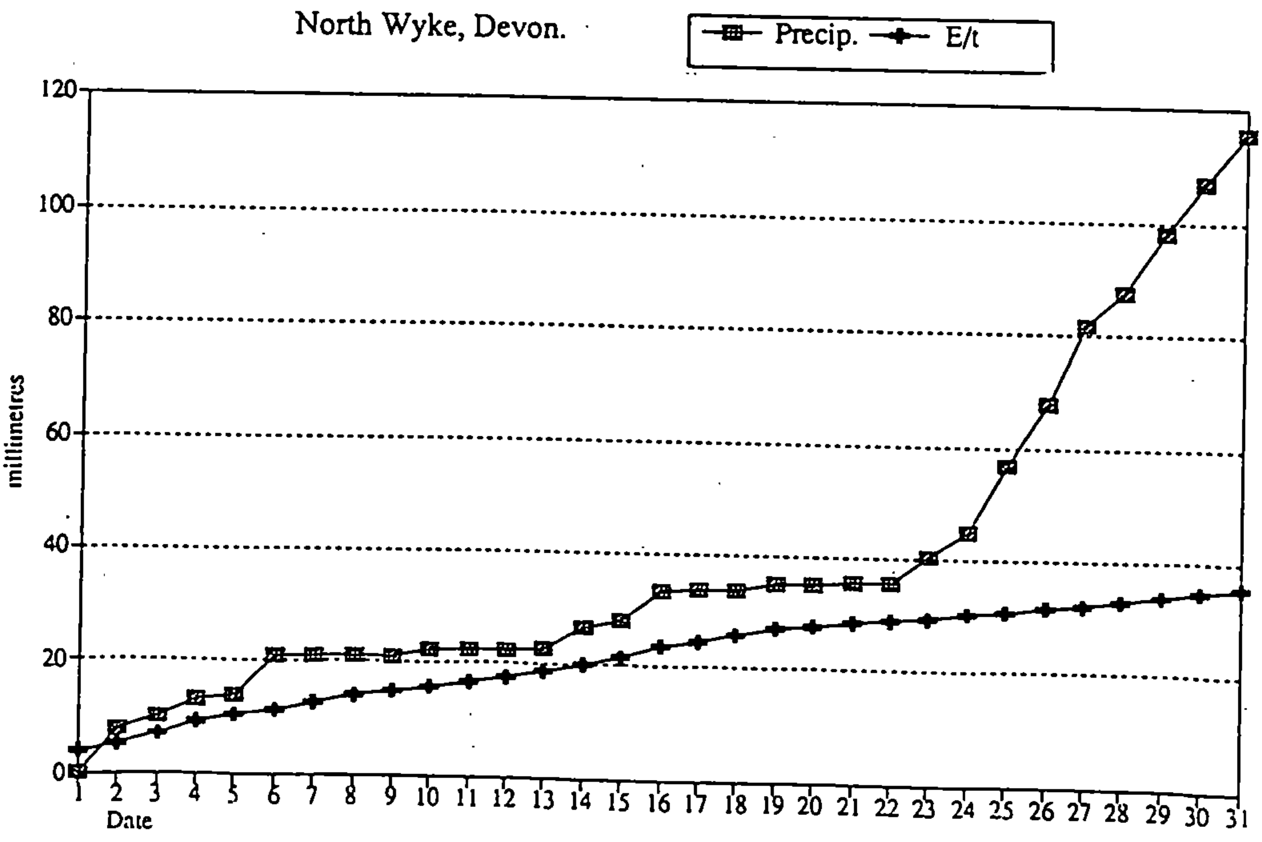
Net precipitation followed a similar, but more exaggerated trend to that of gross precipitation. Due to the seasonal pattern of evapotranspirational losses, high winter precipitation values were only slightly amended by low evapotranspiration rates (Table 5.3).

Table 5.3 Net precipitation (mm) 1990-91.

month	net ppn	gross ppn	evapo-transpiration
Sept	2.4	60.8	58.4
Oct	84.7	115.9	31.2
Nov	79.3	90.2	10.9
Dec	111.8	117.6	5.8
Jan	157.7	163.6	5.9
Feb	50.3	60.7	10.4
Mar	63.9	96.9	33.0
Total	550.1	705.7	155.6

Pre-drainage season evapotranspiration was higher than average, heightening the soil moisture deficit already exaggerated by lower than average precipitation (Section 5.2.1). As evapotranspiration remained relatively high (Figure 5.6), the effectiveness of precipitation in reducing this deficit, and in realising shrink-swell capacity was checked (Section 5.2.1).

Figure 5.6 Cumulative rainfall and evapotranspiration, October 1990. North Wyke, Devon.



5.2.4 Extreme weather conditions.

Extreme weather conditions adversely affected monitoring. A period of cold weather with severe night frosts occurred between 24.1.91 to 14.2.91. Snow lay on the ground from 30.1.91 until 14.2.91. February had a minimum temperature of -9.5° C and a monthly mean of 2° C. On the three coldest days (7.2.91 to 9.2.91) temperatures averaged -5.7° C. Minimum temperatures rose above freezing point on 19.2.91. The frost and snow held water at the soil surface, preventing infiltration and surface runoff. Water in the weir boxes also froze, preventing passage of water and effective gauging. On 15.2.91, a thaw occurred, coinciding with 11.6 mm of precipitation at the site.

5.3 Discharge from undrained and drained fields (Of, If and Df).

5.3.1 Introduction.

Discussion of field discharge in Chapter 5 is arranged in the following order: the runoff characteristics of the undrained field and non drainflow are considered first as these represent the 'natural condition' and are followed by a consideration of drainflow losses. Discussion of each aspect of runoff, on each treatment, is juxtaposed to reflect the major objective of assessment of drainage impact (Section 1.4).

Table 5.4 Structure of Section 5.3

Runoff duration and pattern	Undrained	- flow duration - overland flow and shallow - subsurface losses (Of + If) - runoff coefficients
	Drained field	- flow duration - overland flow and shallow subsurface losses (Of + If) - drainflow losses (Df) - runoff coefficients

*Abbreviations are those used in Equations 5.1 and 5.2.

Weirflow data were collected by S.W.R.C. as described in Section 3.2.4. The results provide a record of flow in millimetres per hour. Field discharge of 1 mm hr^{-1} is equivalent to 10000 l hr^{-1}. All fourteen undrained and drained fields on Rowden Moor had weirs operating by 27.11.90. Drainage continued until 8.5.91 from all fields, with undrained field discharge being particularly 'persistent' (Tyson pers. comm.). Discharge from the undrained and drained fields under investigation began on 18.11.90 and 25.10.90, respectively. From 26.3.91, flow became intermittent on the two fields and comments are, therefore, limited to the period between 1.10.90 and 31.3.91, hereafter referred to as the 'drainage season'. Measurement of weirflow was hindered by frost and snow which persisted for two weeks in February (Section 5.2.4).

Runoff coefficients were used to investigate varying aspects of field hydrology at different times in the drainage season. At the beginning of the season, this was a measure of the balance between satisfaction of the soil moisture deficit and the relative importance of bypass flow. Later, when the soil was wetter, it was regarded as a measure of the efficiency of the drainage system in removing excess water, thereby lowering the water table and reducing surface wetness.

While illustrating basic field discharge, runoff coefficients gave little indication of the variability within the discharge record. Such variability was demonstrated by flow duration curves which allowed the comparison of overall field performances, and the identification of discharge thresholds marking changes in the relative importance or contribution of different magnitudes of flow (Sections 5.3.2 and 5.3.3). Flow duration curves were constructed for the two fields. A logarithmic scale was used to more effectively illustrate detail concerning low discharge values. A normal probability scale was employed for the duration ordinate so that the curve approximated to a straight line.

5.3.2 Undrained field: flow duration analysis.

Figure 5.7 illustrates the flow duration curve for overland flow (Of) and shallow subsurface flow (If) from the undrained field. The dominance of low flows for the majority of the drainage season is apparent, one possible explanation for this result is shallow interflow (investigated in Chapter 6). High flows generated by overland flow (up to 23 mm d^{-1}) were coincident with storm periods (Section 5.2.1) and occurred for a low proportion of the drainage season.

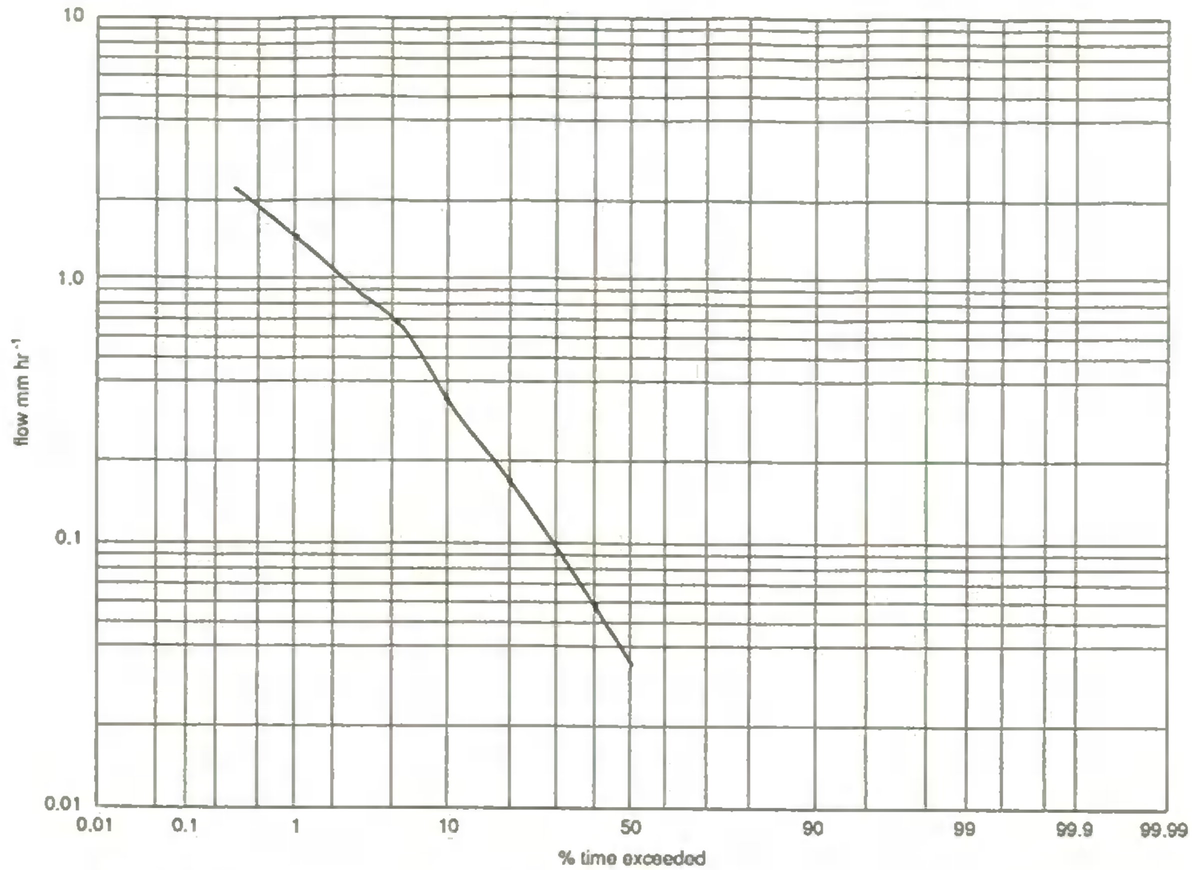


Figure 5.7 Undrained field flow duration curve.

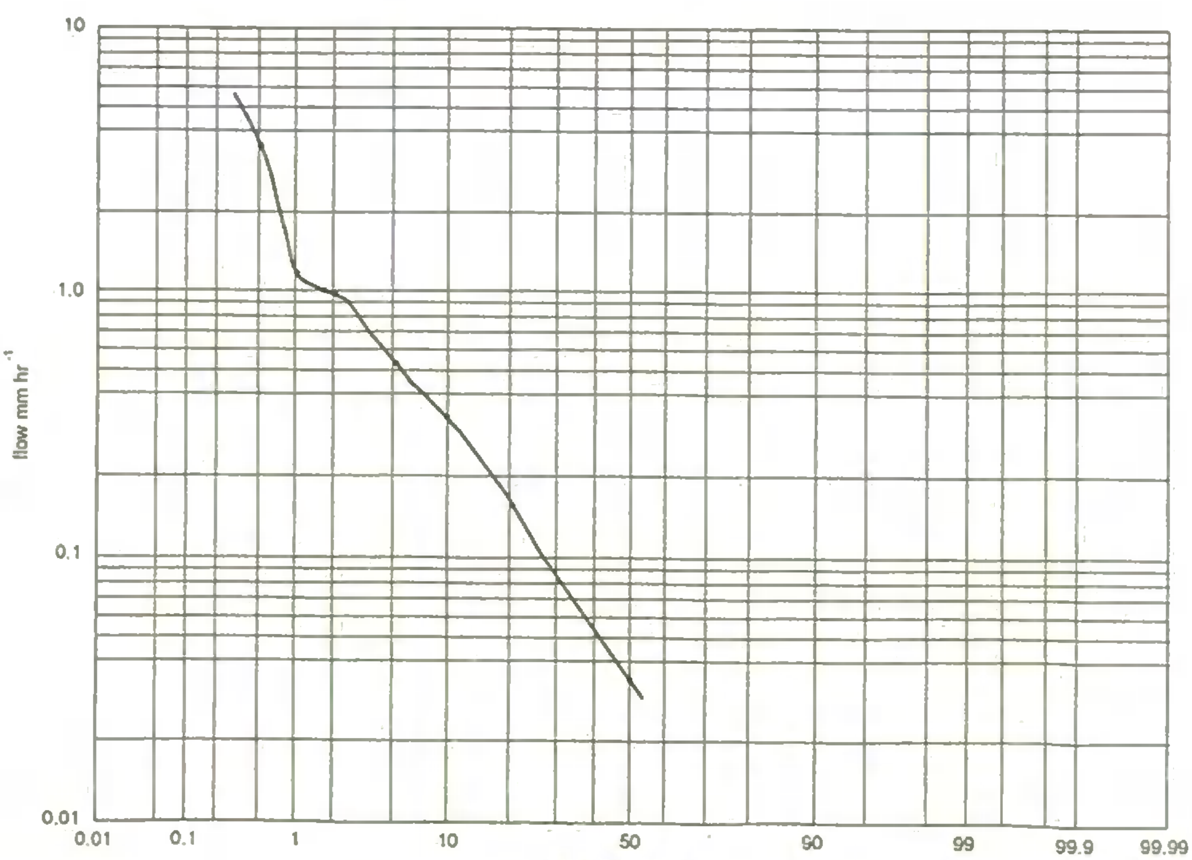


Figure 5.8 Drained field flow duration curve.

5.3.3 Drained field: flow duration analysis.

Drained field data were separated to produce a drainflow (Df) duration curve (Figure 5.8). There were some periods of zero flow which are not incorporated in the flow duration curve. Figure 5.8 illustrates that drainflow rates were less than 0.05 mm hr^{-1} for almost 50 % of the time. A rate of 1 mm hr^{-1} or greater was attained for only 1.6 % of the flow record compared with 2.5% on the undrained field. The dominance of low discharge rates concurs with the findings of Hallard (1988). The extended period of flow coupled with the relatively low discharge suggests that a large proportion of the water draining through the field moves via subsurface pathways. Peaky response indicates an important role for macropores, while the sustained baseflow is due to water draining from the matrix.

5.3.4 Undrained field: seasonal pattern of flow (If + Of).

To summarise, the flow record of the undrained field was shorter than the drained field mole discharge in 1990-91 (Figures 5.9 and 5.11). One weir was used to determine the output of water both by overland flow and by shallow lateral subsurface flow of water, intercepted by 0.3 m deep, backfilled ditches at field boundaries (Section 3.2.4). There were only three runoff events, each lasting a matter of hours, between 27.10.90 and 18.11.90. Discharge continued until 24.3.91, with only short periods of zero output (Table 5.5). Periods of no flow amounted to nearly one third of the drainage season.

If the flow record for the start of the drainage season is examined (Table 5.6) it is apparent that discharge followed a similar pattern to that observed in 1983 by Hallard (1988). Field capacity was not achieved until late in October and discharge response was not related to precipitation. In October 1990, weirflow amounted to 0.1 % of the output from the field (Figure 5.9). In both years, dry summers were followed by drainage seasons of only 5 months. In 1984, by contrast, early wetting followed a wet summer so that October discharge did reflect rainfall distribution, and the drainage season was extended (7 months). Persistent discharge only began on 18.11.90. The weir discharge roughly approximated the distribution and magnitude of precipitation events with the exception of snowmelt (Section 5.2.4).

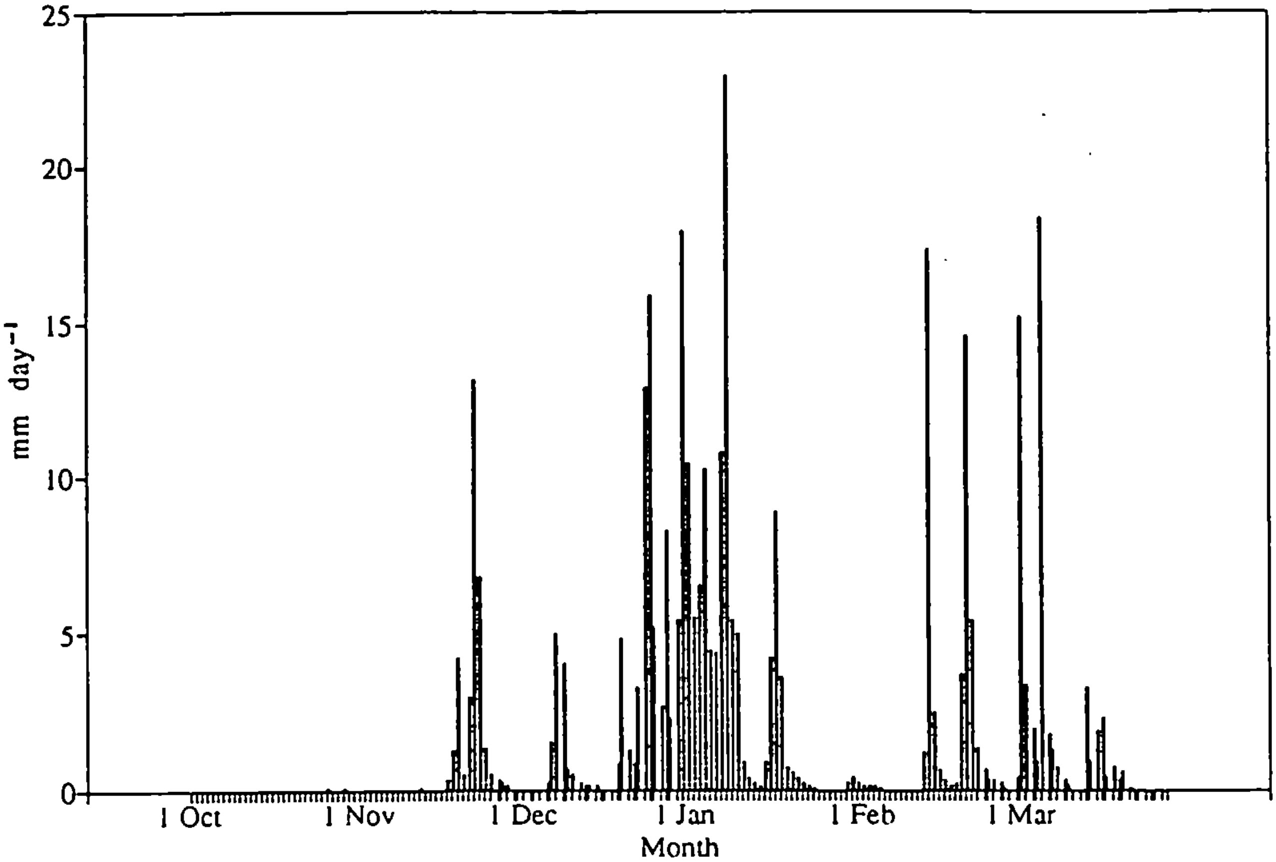


Figure 5.9 Undrained field weirflow (1990-91).

Table 5.5 Undrained field weir discharges and periods of no flow (1990-91).

Month	Monthly discharge	Total days with no discharge
October	0.04	29
November	31.38	1
December	93.33	8
January	107.44	5
February	48.94	7
March	54.32	9
Total	335.45	59

Table 5.6 Monthly rainfall and runoff totals (mm) - undrained field.

Month	1983-84		1984-85		1990-91	
	ppn	runoff	ppn	runoff	ppn	runoff
Sept	—	—	106.8	0.3	60.8	0 0
Oct	—	—	113.4	20.3	115.9	0.04
Nov	50.4	8.5	211.6	142.1	90.2	31.38
Dec	173.5	144.8	126.6	86.3	117.6	93.33
Jan	223.5	216.8	88.4	59.9	163.6	107.44
Feb	77.6	70.6	42.8	24.4	60.7	48.94
Mar	74.9	30.3	105.8	51.4	96.9	54.32
Apr	12.2	10.5	93.8	39.3	88.2	27.2
Total	612.1	481.6	889.2	424.0	793.9	362.65
Runoff coefficient						
(Nov - Apr)		78.7 %		60.2 %		58.8 %
(Sept-Apr)		—		47.7 %		45.7 %

The monthly discharge for December (93.33 mm) was approximately three times that for November. This represented three quarters of the cumulative runoff in the first three months of the drainage season (124.74 mm). The magnitude of average daily flows more than doubled to 2.43 mm day^{-1}. A monthly maximum of 17.9 mm day^{-1} occurred on 31.12.90. The impact of a reduction in soil moisture deficit (Section 5.4.2) continued to be evident in January when maximum recorded daily runoff occurred (22.9 mm; 8.1.91). In February, high discharge was related to a thaw releasing snow and ice from surface storage. The thaw also reinstated overland flow pathways to the field boundary and therefore to the weir. Persistent discharge ceased on 24.3.91. The response of field runoff to the storm periods highlighted in Section 5.2.1 is illustrated in Table 5.7. Discharge during these periods amounted to 59.6 % of the drainage season total.

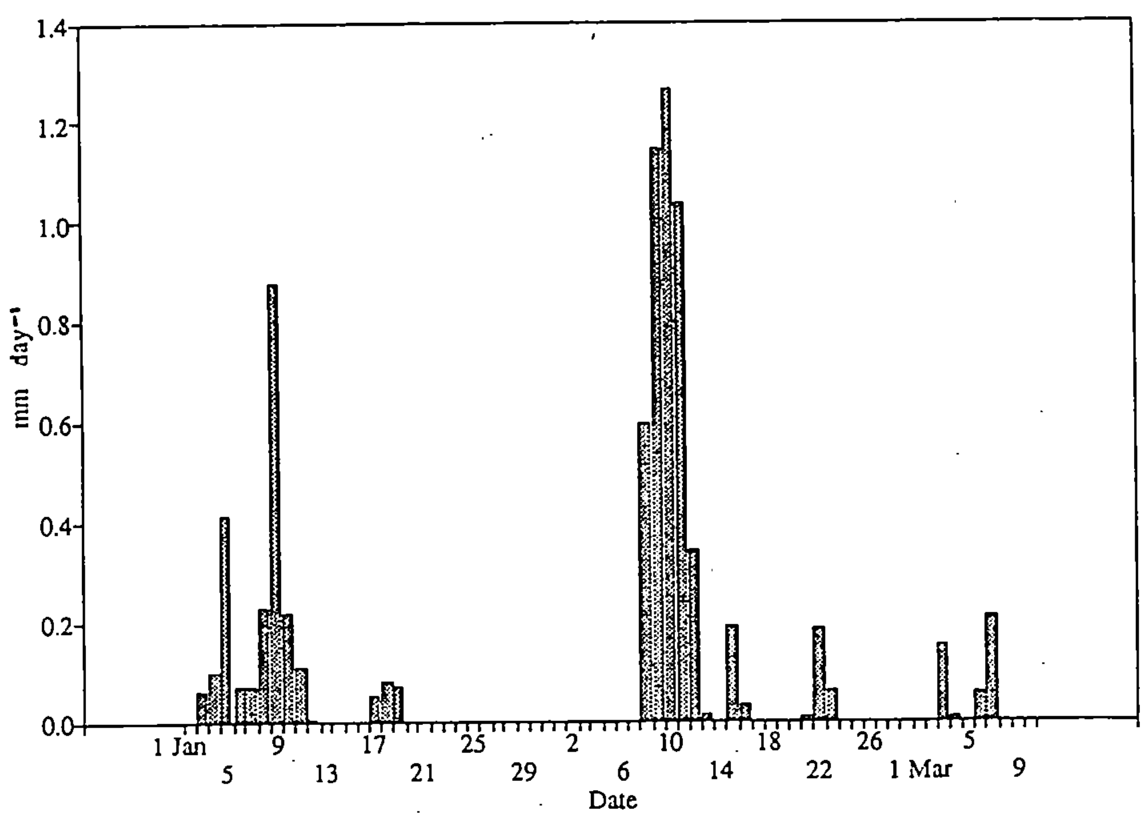
Table 5.7 Undrained field: storm runoff, 1990-91 (mm)

Date	Precipitation	Discharge
25.10 - 29.10	53.1	0.03
19.11 - 24.11	58.6	28.8
20.12 - 09.01	206.6	161.0
04.03 - 08.03	49.5	26.3
37 days	367.8	216.1

5.3.5 Drained field: seasonal pattern of non drainflow (If + Of).

The non drainflow discharge from the drained field (Figure 5.10) had an intermittent and short flow record (only 28 days between 3.1.91 to 8.3.91). Initiation of non drainflow in January was synchronous with runoff coefficients for the drain (Df) near to, or equal to 100 % (Section 5.4). Mean discharge, calculated for those days on which non drainflow did occur was 0.27 mm d^{-1}, compared with 9 mm d^{-1} for the drains on the same days. The highest daily surface weirflow of 1.266 mm, occurred on 10.2.91, when snow thawed and an additional 11.6 mm of precipitation fell. Runoff generated over 11 days in February, accounted for almost 53 % of the total. The weir discharge record suggests that the soil was above field capacity throughout the period, however, sustained rainfall over at least three days was required to generate non drainflow rather than high daily precipitation amounts.

Figure 5.10 Drained field: non drainflow discharge.



Note: includes all periods of non drainflow on drained field.

5.3.6 Drained field: seasonal pattern of drainflow (Df).

Unfortunately, a failure of the stage-recorder, between 4.1.91 and 14.2.91, resulted in a period with no available data (Section 5.2.4). A seasonal runoff coefficient was applied to calculate discharge for the drain weir for this period. However, using this technique for days on which weirflow data were available gave an over-prediction of 142 %. Therefore, a runoff coefficient was calculated for a restricted period: 16 days before, and 16 days after the failure of the stage recorder. This was to ensure that the runoff coefficient related to a similar hydrological period. Regression analysis showed that discharge (Q) was closely related to precipitation (R) ($r^2 = 67.57$ %).

$$Q = 0.72\,R + 1.58 \qquad (\mathrm{mm\ d^{-1}}) \qquad \text{[Eqn. 5.3]}$$

Application of regression results gave mole drain discharge of 153.78 mm for 4.1.91 to 13.2.91, inclusive. For January and February, this led to the prediction of 152.55 mm and 61.46 mm, respectively. All subsequent discussion of drainflow discharge, in this thesis, considers data which include these extrapolated results for January and February.

Mole drain weir discharge started on 13.10.90 and was persistent from 26.10.90 (Figure 5.11). Thereafter, the pattern of discharge was more uniform than that from the undrained field (Section 5.3.4), showing clear exponential recessions in the days following hydrograph peaks. Persistent discharge continued into April, 1991. However, from 25.3.91 the rate of discharge had decreased to less than 0.15 mm d^{-1} (less than 1500 l d^{-1}). Drainage season data are shown in Table 5.8 and illustrate that most of the discharge from the drained field was conducted by the moles and the deep lateral drains. Total discharge from the field was 441.7 mm, of which 433.8 mm (98 %) was drainflow (Df) and 8 mm (2 %) was non drainflow (If + Of).

Figure 5.11 Drainage season mole drain discharge.

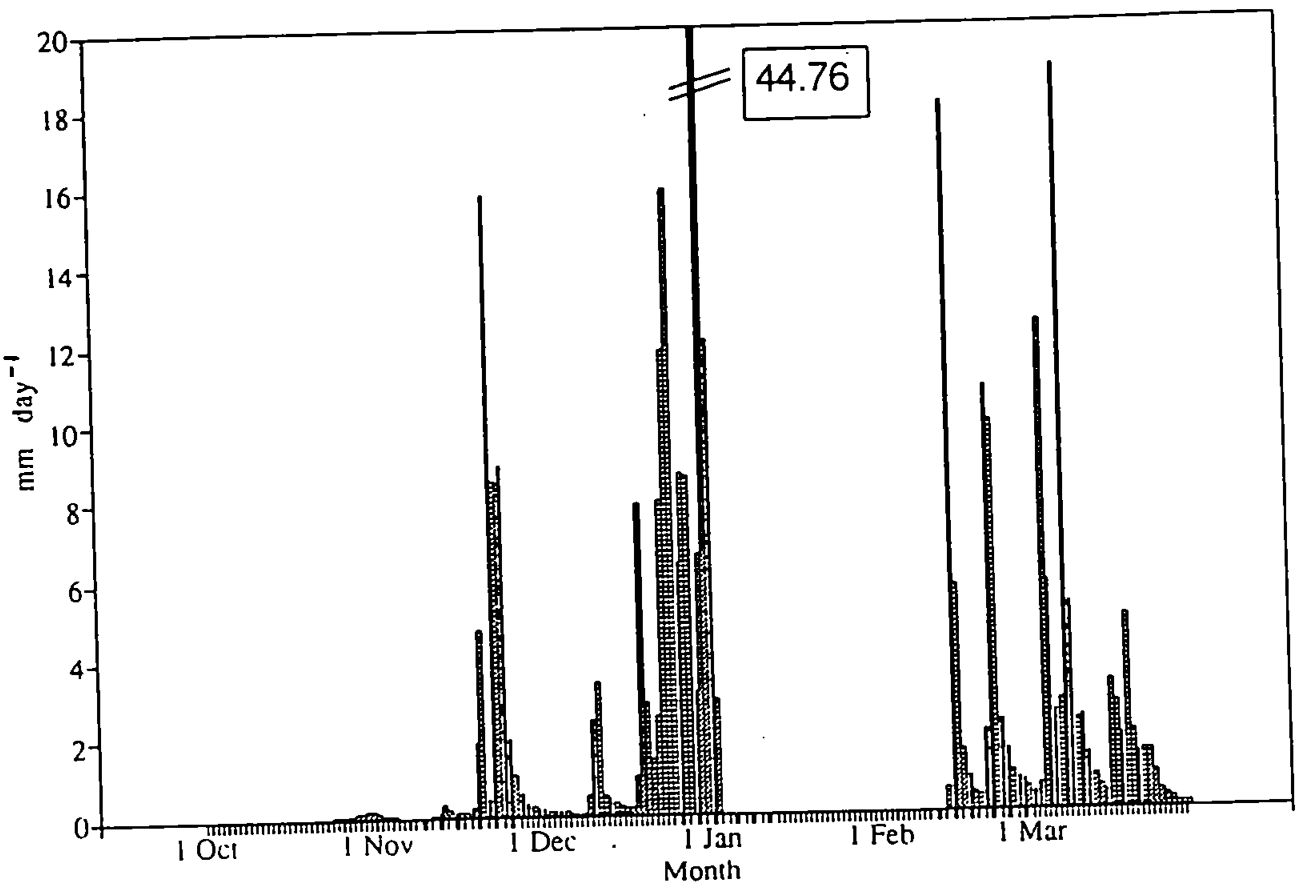


Table 5.8 Drained field: summary of drainage season data (1.10.90 -31.3.91).

All units mm.

Component	Seasonal total	Daily maximum	Daily average
Precipitation (R)	644.9	33.90	3.55
Mole drain (Df)	433.8 *	29.73	1.77
Non drain (If + Of)	7.7	1.27	0.04
Potential Et	97.2	—	—

*see Table 5.9

To set the monthly discharge from the drained field into context, data from Hallard (1988) are presented in Table 5.9. The data confirm the dominance of water movement to the drains through all of these drainage seasons.

Table 5.9 Drained field: monthly rainfall and runoff totals - discharges separated into drain (Qsub) and non drainflow (Qsurf). All units in mm.

	1983-84*1			1984-85*1			1990-91		
Month	ppn	Qsub	Qsurf	ppn	Qsub	Qsurf	ppn	Qsub	Qsurf
Sept	—	—	—	106.8	0.1	0.0	60.8	0.00	0.00
Oct	—	—	—	113.4	11.5	0.2	115.9	0.47	0.00
Nov	50.4	7.8	0.0	211.6	148.8	1.6	90.2	46.10	0.00
Dec	173.5	99.7	3.0	126.6	85.4	0.7	117.6	93.94	0.00
Jan	223.5	181.8	3.7	88.4	64.6	0.5	163.6	152.55*2	2.36
Feb	77.6	78.8	0.8	42.8	29.1	0.2	60.7	61.46*2	4.87
Mar	74.9	31.6	0.2	105.8	55.5	0.5	96.9	79.31	0.42
Apr	12.2	14.5	2.9	93.8	41.8	0.9	88.2	33.65	0.00
Total	612.1	414.4	7.8	889.2	422.2	4.6	793.9	467.48	7.65

*1 Hallard, 1988.

*2 The measured discharge was 47.03 mm, giving a cumulative drainage season discharge of 328.31 mm. When regression was incorporated to predict January discharge, drained field runoff was 152.55 mm. February rainfall was also altered to 61.46 mm.

In 1990-91, monthly drainflow volumes were greatest in December and January, as in 1983-84 and 1984-85 (the exception to this was November, 1984, where exceptionally high rainfall generated higher runoff). In February, runoff was low (66.33 mm), reflecting low rainfall volumes (60.7 mm) (Section 5.3.4).

Daily drainflow peaked at 29.73 mm on 1.1.91 when 33.9 mm of precipitation was recorded (Table 5.10). There were nine other high discharge events (> 10 mm d^{-1}). Like rainfall (Section 5.2.1), these were clustered according to frontal weather patterns (Section 2.2), as illustrated in Table 5.11. Days with higher runoff than precipitation, are explained by the combination of quickflow and soil drainage following storage of water from consecutive days of rainfall. Such events illustrate the impact of drainage in diverting water to subsurface water pathways, thereby delaying weirflow response, and are discussed in more detail in Chapter 6.

Table 5.10 Drain discharge for high runoff events (mm).

Date	Precipitation	Drainflow
23.10.90	24.0	15.77
25.12.90	8.0	11.77
26.12.90	17.0	15.87
01.01.91	33.9	29.73
02.01.91	7.1	12.09
15.02.91	11.6	17.98
22.02.91	18.2	10.75
03.03.91	13.9	12.38
07.03.91	16.5	18.85

Table 5.11 Drained field storm runoff in 1990-91 (mm).

Date	Precipitation	Drainflow	Non drainflow
25.10 - 29.10	53.1	0.2	0
19.11 - 24.11	58.6	31.6	0
20.12 - 09.01	206.6	128.3	1.822
04.03 - 08.03	49.5	42.3	0.424
Total 37 days	367.8	202.4	1.246

5.3.7 Undrained field runoff coefficients.

From October to March, the proportion of precipitation which exited the site via weir discharge was almost 50 %. The monthly runoff coefficients are detailed in Table 5.12. The rise in the proportion of runoff to rainfall was consistent from October. When weir discharge started (27.10.90) cumulative discharge represented only 0.03 % of rainfall since 1.10.90. Before 18.11.90 there were only two more occasions on which weir discharge was recorded. The highest discharge of the month (13.13 mm) occurred on 14.11.90.

Although higher and more consistent discharges were observed throughout November, the runoff coefficient for cumulative monthly discharge remained relatively low at 34.78 %. Highest runoff coefficients occurred in December and January. The magnitude and frequency of storms experienced (Figure 5.3) led to high runoff volumes (Section 5.3.6,

Table 5.11) from the already wetted soil (Section 5.4) causing the runoff coefficient to increase to 83.48 %. In the first 25 days of January 106.9 mm of rain fell and resultant weirflow was 107.0 mm. Highest flows and runoff coefficients were related to large rainfall inputs (33.0 mm on 1.1.91; 22.9 mm on 8.1.91). Runoff coefficients, calculated from cumulative data, remained between 65 and 70 % for the rest of the month. Drainage ceased during a period of cold weather when snow and ice lay on the ground for two weeks (Section 5.2.4).

Table 5.12 Undrained field: rainfall and runoff totals (mm) - for months with > 10 mm weir discharge.

Month	1983-84*[1]		1984-85*[1]		1990-91	
	ppn	runoff	ppn	runoff	ppn	runoff
Oct	-	-	113.4	20.3	-	-
Nov	-	-	211.6	142.1	90.2	31.38
Dec	173.5	144.8	126.6	86.3	117.6	93.33
Jan	223.5	216.8	88.4	59.9	163.6	107.44
Feb	77.6	70.6	42.8	24.4	60.7	48.94
Mar	74.9	30.3	105.8	51.4	96.9	54.32
Apr	12.2	10.5	93.8	39.3	88.2	27.93
Total	561.7	473.0	782.4	423.7	617.2	363.34
Runoff coefficient	84.2 %		54.2 %		58.6 %	
Dec - April	84.2 %		57.1 %		63.0 %	

*[1] Data from Hallard (1988)

Runoff coefficients calculated for three drainage seasons illustrated the potential for variable field hydrological response to different meteorological conditions (Table 5.12). The runoff coefficients were lowest in 1984-85. In this year, precipitation was more evenly distributed. Runoff coefficients were highest in 1983-84, unfortunately the data set only includes drainage from December so there is no indication of the extent of wetting that had occurred but evapotranspiration was low. Two consecutive months of high rainfall then generated high runoff volumes. Therefore, despite the shorter duration of significant drainage and lower precipitation more drainage occurred than in 1990-91 or 1984-85. When runoff coefficients for December to April were calculated for 1984-85 and 1990-91, no change in this pattern was observed.

5.3.8 Drained field: runoff coefficients.

Table 5.13 details the runoff coefficients for the drained field and compares them with those observed by Hallard (1988). Over the drainage season, the most striking trait was of high runoff coefficients in the winter, when evapotranspiration was negligible, with less runoff in the early autumn and spring. However, even at the monthly scale, more detailed trends were apparent. Drainage began before the soil reached field capacity (Section 5.4). As the soil moisture deficit was satisfied and evapotranspirational losses decreased, the proportion of runoff to rainfall increased. A soil moisture threshold may have been reached, when evapotranspiration was negligible and runoff discharges accounted for 100 % of gross precipitation.

The highest runoff coefficients for the undrained field occurred in December and January. However, it was assumed that the relationship on the drained field was more complex as water table response and water transmission were complicated by fissure networks and mole drains. Drainage management increased the depth of drainage and therefore potential temporary storage, providing alternative water pathways. When the runoff coefficients were calculated on a daily basis this became more apparent.

On a daily basis, Figure 5.10 shows that the highest non drain discharge (If + Of) of 0.192 mm occurred at the same time as a high mole drain output (18 mm d^{-1}) and a period of high soil moisture content on 15.2.91 (Section 5.4.3). Examination of days when there was rain and surface runoff reveals that the mean coefficient was 55.82 % for non drainflow and over 100 % for the drainflow weir. Therefore, on these days, drainflow included old water conveyed by the matrix as well as new water transmitted by macropores (Section 1.6). Mechanisms responsible for generating high discharges are discussed in Section 7.2.5.

Daily drained field runoff coefficients were highly variable. Initial autumn values were very low because a soil moisture deficit existed (Section 5.4.3). Before drainflow commenced on 25.10.90, the proportion of precipitation which left the fields by this route drainflow ranged between 21 and 39 %. Between 25.10.90 and 27.10.90 rainfall was high (39.3 mm) and evapotranspiration remained constant so drainage increased (Figure 5.12). However, precipitation ceased and by the end of the month cumulative output fell to 29.3 % of input, most being accounted for by evapotranspiration (Figure 5.13). Net precipitation was diverted to soil moisture storage. By 18.11.90, the cumulative runoff coefficient was still only 0.25 %. However, there was a rapid rise in the value so that by 20.12.90, subsurface and surface discharge (Df + If + Of) represented 20 % of rain (since 1.10.90). Between 3.1.91 to 31.3.91 the proportion of runoff ranged from 40 to 50 %.

On 15.2.91 the runoff coefficient was 155.2 % with mole drain discharge (18 mm) being considerably higher than incoming precipitation (11.6 mm) due to release of water by snow and ice thaw (Section 5.2.4).

Table 5.13 Drained field: monthly rainfall and runoff totals (mm) and runoff coefficients (%) *1.

Date	1983-84			1984-85			1990-91		
	ppn	runoff	%	ppn	runoff	%	ppn	runoff	%
Sept	—	—	—	106.8	0.1	0.1	60.8	0.0	0
Oct	—	—	—	113.4	11.7	10.3	115.9	0.5	0.4
Nov	50.4	7.8	15.5	211.6	150.4	71.1	90.2	46.1	51.1
Dec	173.5	102.8	59.3	126.6	86.1	68.0	117.6	93.9	79.9
Jan	223.5	185.6	83.0	88.4	65.0	73.5	163.6	152.6*2	93.2
Feb	77.6	79.6	102.6	42.8	29.3	68.5	60.7	61.5*2	101.3
Mar	74.9	31.8	42.5	105.8	56.1	53.0	96.9	79.3	81.9
Apr	12.2	14.6	120.7	93.8	42.7	45.5	88.2	33.7	38.2
Totals									
(Nov-Apr)	612.1	422.2	68.9	669.0	429.6	64.2	617.2	425.1	68.9
(Sept-Apr)				889.2	441.4	49.6	793.9	467.5	58.9

*1 Calculated from monthly cumulative data (runoff/rainfall * 100)

*2 Measured runoff was 47 mm (runoff coefficient 28.8 %) for January - however, this included a period with no data (Section 5.3.2). February was also altered.

Figure 5.12 Drained field: precipitation and evapotranspiration (mm). October 1990.

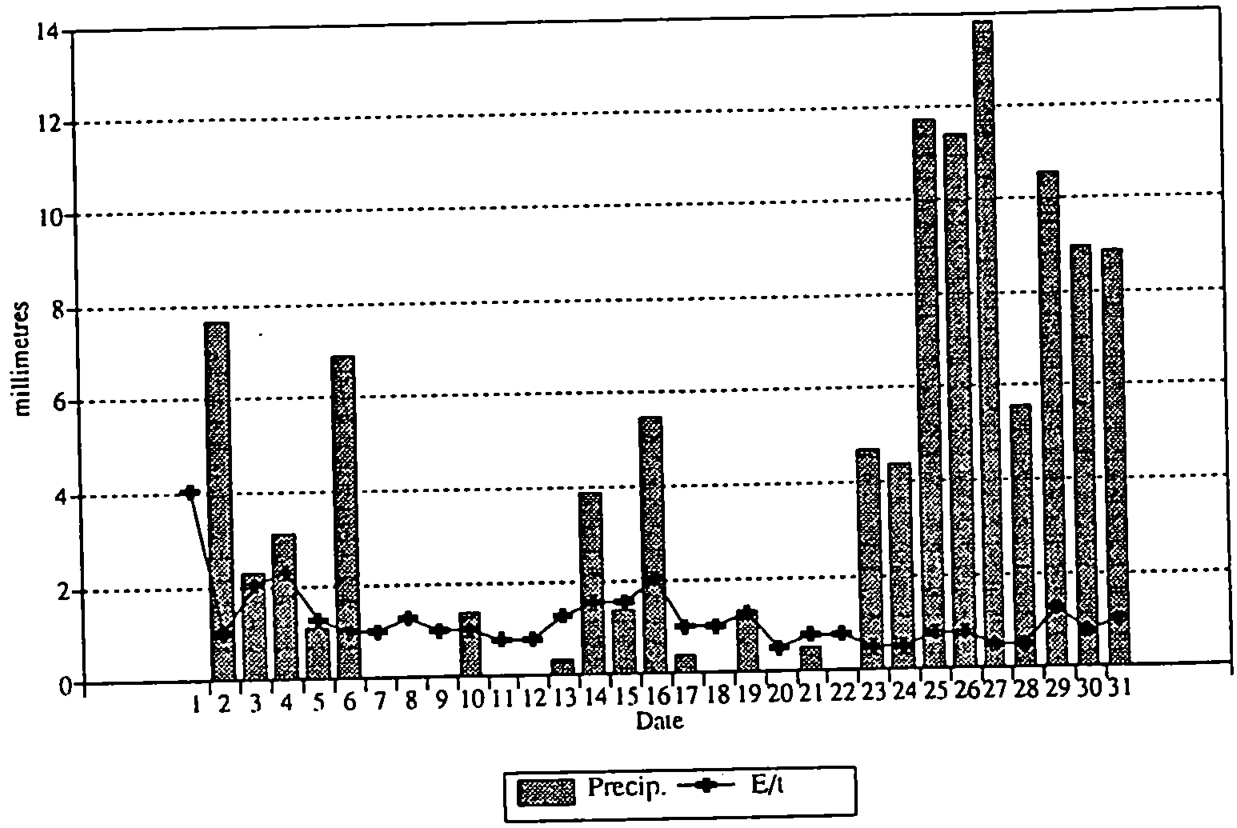
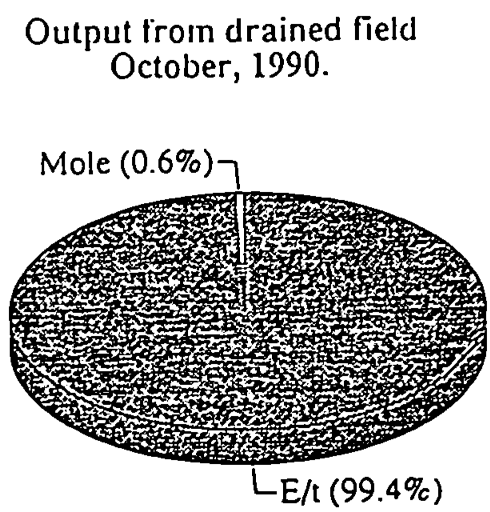


Figure 5.13 Drained field: relative importance of water balance outputs. October 1990.



At the end of the 1990-91 drainage season, high runoff coefficients fell quickly to low or zero values. This was coincident with a decrease in soil moisture content to less than the winter mean and the cessation of persistent drainflow but did not correspond to a significantly lower input of precipitation (Table 5.13). These results are similar to those of Parkinson (1984) and Hallard (1988).

5.3.9 Summary: comparison of discharge regimes on undrained and drained fields.

Table 5.14 and Figure 5.14 summarise the relative contributions of the input and various output components of the water balance on a monthly basis (Section 5.1; Eqn. 5.1; Figure 5.1). These demonstrate the dominance of evapotranspiration on both fields early in the drainage season. By November, mole drain discharge accounted for nearly 80 % of output compared with 73 % weirflow output on the undrained field. In December and January, weir runoff on the two fields accounted for a similar proportion of the output from both fields, ranging between 90 and 95 %. The notable development in January was the contribution of 5 % of drained field output by non drainflow, and therefore, the initiation of a further water pathway. As soil moisture content increased (Section 5.4.3) this mechanism became more significant, contributing nearly 7 % of output in February. Evapotranspiration controls the seasonal deficit, while rainfall is the major influence on pathways and hence the magnitude of discharge (Table 5.14).

Table 5.14 Monthly rainfall and runoff totals (mm).

Month	Precipitation (R)	Evapo-transpiration (Et)	Undrained field (Of + If)	Drainflow (Df)	Drained field (Of + If)
September	60.8	58.4	0	0	0
October	115.9	31.2	0.04	0.48	0
November	90.2	10.9	31.38	46.10	0
December	117.6	5.8	93.33	93.94	0
January	163.6	5.9	107.44	152.55	2.36
February	60.7	10.4	48.94	61.46	4.87
March	96.9	33.0	54.32	79.31	0.42

Figure 5.14 a Monthly outputs from the undrained field.

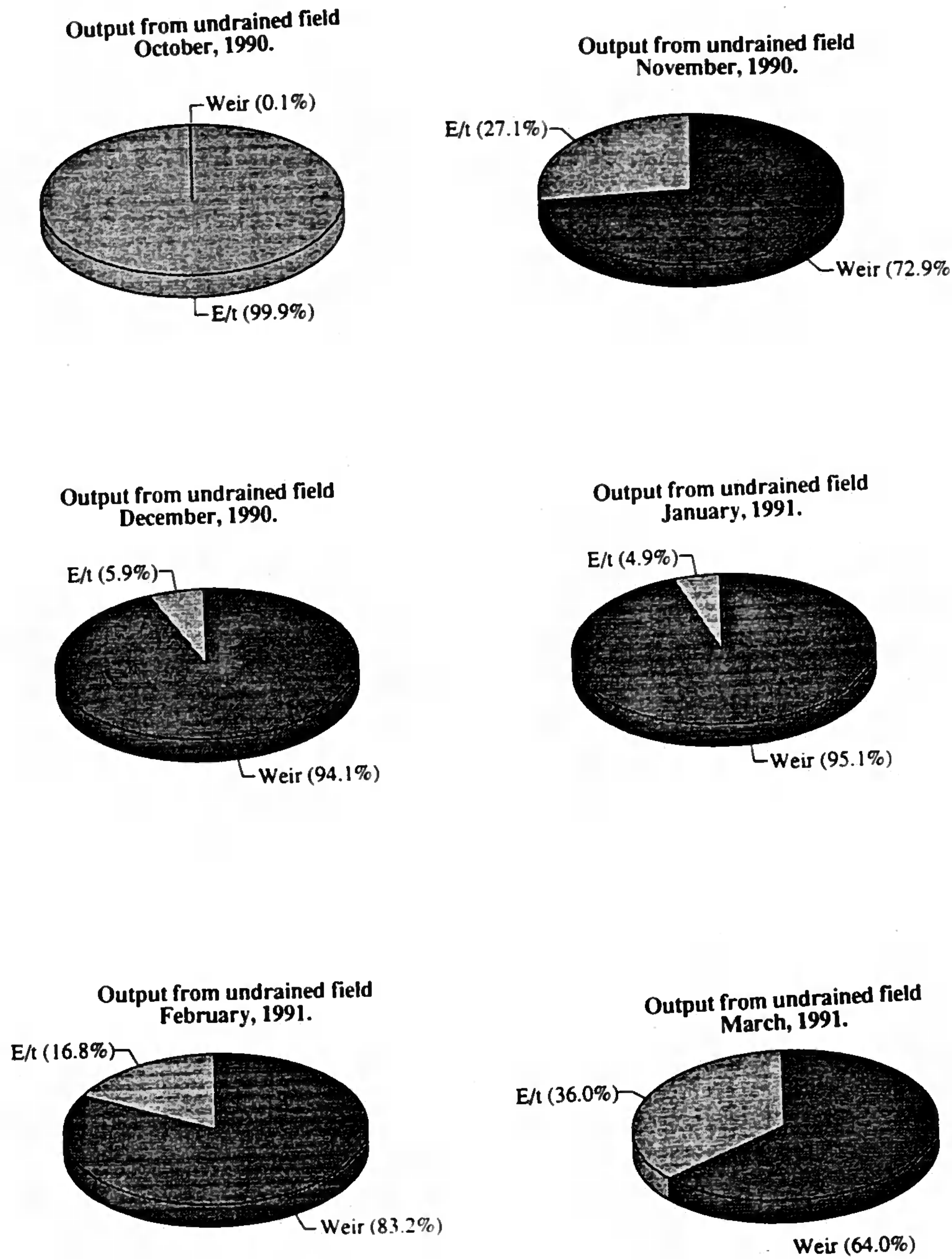
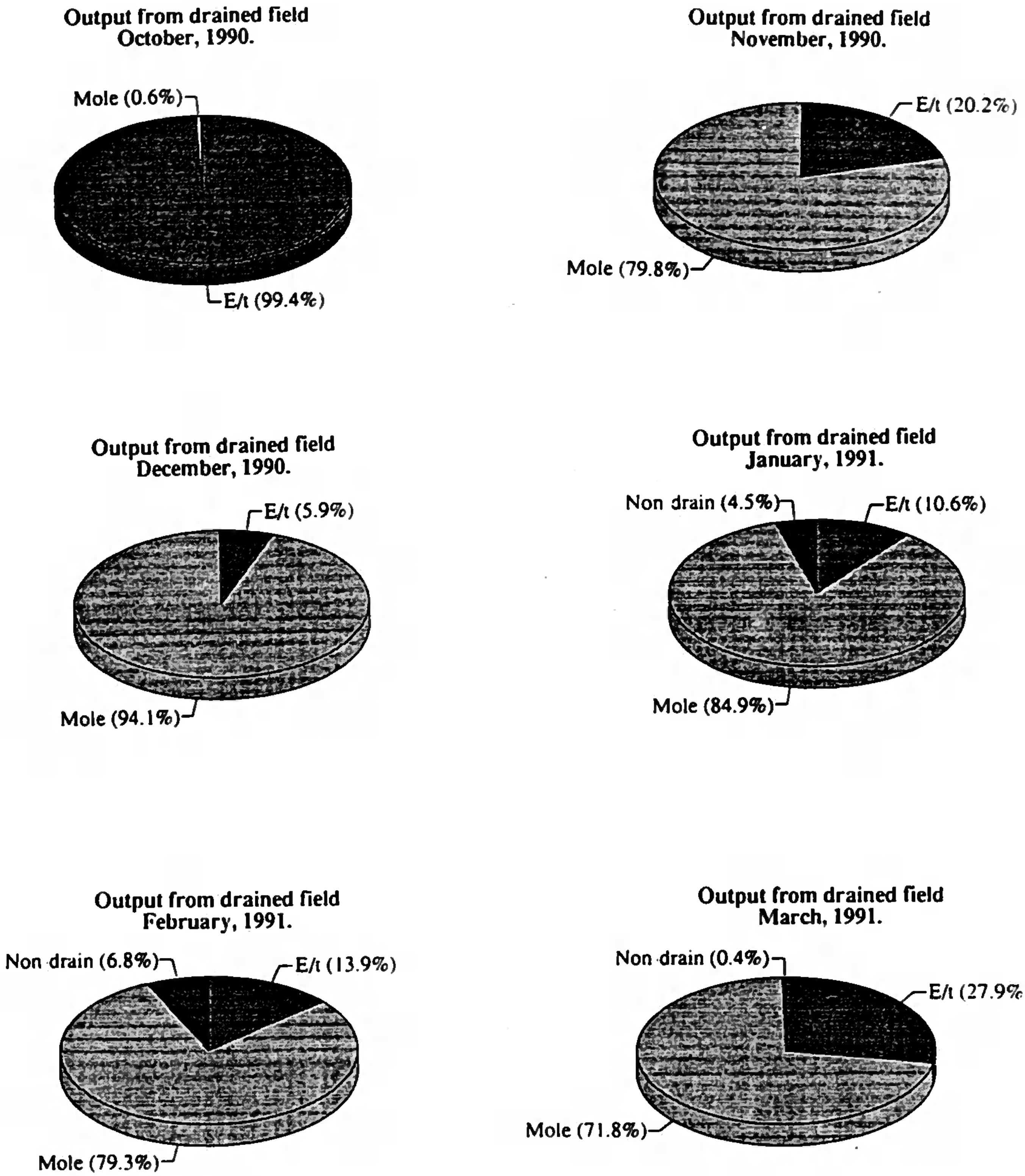


Figure 5.14 b Monthly outputs from the drained field.



Drained field runoff occurred over a considerably longer period than undrained runoff (Table 5.15) and within the 1990-91 drainage season a number of distinct hydrological phases were identified. On the undrained field, there was little or no subsurface sink for drainage (Hallard, 1988), therefore, Of and If pathways were initiated. It was assumed that this was coincident with satisfaction of the soil moisture deficit On the drained field, drains began to operate when the soil immediately adjacent to the mole drain was saturated, due to rapid water flow through a network of cracks. These hypotheses are investigated by more detailed observations in Chapter 6.

Total field discharge was higher from the drained site and can be attributed to the ability of the mole drain to conduct water from the subsurface. Maximum discharge generated from the drained field (Df) was twice as high as that from the undrained field (Of + If) both on an hourly and a daily basis. High rates of runoff from the drained field could be explained by fissure flow/mole drainage. The peaky regime is attributed to these subsurface pathways quickly switching on and off.

Consequently, mean daily flow was lower for the drained field due to the large proportion of quickflow and because baseflow was maintained for long periods. Baseflow was contributed by the slowly draining matrix. This situation is similar to that described by Whipkey (1967) and Hewlett and Hibbert (1963). On the drained field, mean Of + If discharge was minimal. Infiltration capacity was not exceeded until January and the soil could only have been saturated in very locally limited areas close to field boundaries due to the low volumes of runoff generated.

5.4 Soil water (S) at the field scale.

5.4.1 Introduction.

Volumetric soil moisture content was determined on three occasions by T.D.R. (Section 3.3.5) at thirty six sites on each field. The sites were spaced at 10 m intervals across four transects to investigate the spatial variability of soil moisture content, and to relate variation to the more detailed spatial and temporal measurements taken within the plots (Section 6.4). Observations were made in January, March and April. More intensive spatial and temporal monitoring of volumetric soil moisture was undertaken on the plots (Section 6.4). Comparison of field and plot data, using a Mann Whitney U test, identified no significant difference between the data sets. Therefore, it is assumed that plot data are representative of the wider scale and can be used in the calculation of the soil moisture budget (Section 5.5).

5.4.2 Undrained field: volumetric soil moisture content.

Mean volumetric soil moisture content was 48 % from 0 to 0.2 m depth, 45 % from 0 to 0.4 m and 45 % from 0 to 0.6 m; and there was little variation about the mean within the three surveys. The presence of large pores at the surface explains the higher soil moisture content from 0 to 0.2 m, as confirmed by evidence from the soil moisture characteristic curves (Section 4.2.2). Results for the soil moisture characteristic curve showed that the saturated volumetric soil moisture content was 44 % (Table 4.3). The upper 0.2 m of the undrained profile was saturated from mid November onwards. On 16.12.90, soil moisture throughout the 0.6 m profile was greater than 44 % and the soil was more consistently at saturation from January to March.

5.4.3 Drained field: volumetric soil moisture content.

Volumetric soil moisture content was more spatially and temporally variable on the drained field. Moisture content values were at a maximum in January (48.5 % from 0 to 0.2 m depth, 47 % from 0 to 0.4 m and 46 % from 0 to 0.6 m) but declined by March, and had fallen quite markedly by April (42.1 % from 0 to 0.2 m depth, 42.9 % from 0 to 0.4 m and 44 % from 0 to 0.6 m). Soil moisture was least variable in January, and as the soil dried, variability rose. Results for the soil moisture characteristic curve show that the soil was saturated from 0.2 to 0.6 m depth in January, but not in March and April.

5.4.4 Soil water storage (S).

On the undrained field, volumetric soil moisture content was 29 % on 1.10.90 and briefly reached saturation (45 %) in the upper 0.3 m on 21.10.90. Water content of the upper 0.6 m increased from 174 mm on 1.10.90, to 257 mm on 30.10.90, and to 276 mm by December a total rise of 102 mm for the period. By mid March, volumetric soil moisture averaged 49 % over the 0.6 m profile, an increase of 120 mm of stored water over the drainage season.

On the drained field, the difference between the soil moisture content in October (33.2 %) and the maximum on 14.2.91 (49.4 %) represented 97.2 mm of water. The high soil moisture value was coincidental with high runoff coefficients for all weir sources (drain and non drain). By 30.3.91, drainage reduced the seasonal increase of total soil water storage in the 0.6 m profile to 98.4 mm.

In October, volumetric soil moisture contents rose from *c.* 30 to 44 % as the soil moisture deficit was quickly reduced on both fields. Between 1.10.90 and 29.10.90, the undrained and drained soils imbibed 78 mm and 34 mm, respectively. While weir discharge was absent or negligible (Section 5.3.4) the soil soaked up precipitation. The continued wetting process resulted in an increase in soil water of 102 mm, raising average volumetric soil moisture content to around 45 % by December.

5.5 Calculation of water balances.

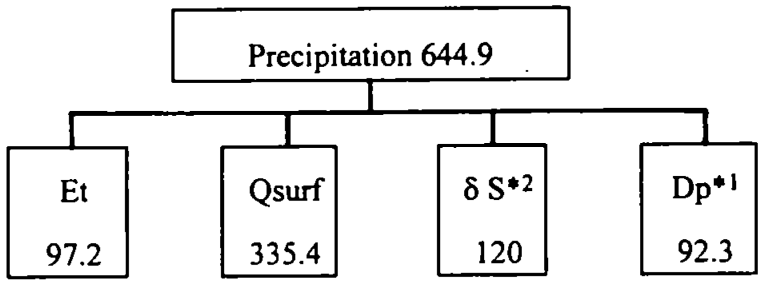
The components of the water balance were calculated on a daily basis. Results were summed and analysed for monthly and six monthly periods. The drainage season water balance (1.10.90 to 31.3.91) is the more meaningful, and is discussed to set the context for the more detailed discussions of Chapter 6.

Figure 5.15 a shows that on the undrained field, 86 % of rainfall could be accounted for by measured and estimated outputs (weirflow and evapotranspiration) and increased soil water storage. The water balance data for the drained field (Figure 5.15 b) accounted for 95 % of rainfall. The remaining 14 and 5 % may be accounted for by various fates. It is probable that the depth of the soil profile which responds to seasonal wetting and drying extends below 0.6 m. If the increased soil water content observed over 0.6 m was continued to greater depth this would largely account for the 'surplus' water (Figure 5.15 footnotes). Deep percolation or leakage from field boundaries is a further possible explanation and the problems experienced with the weirs in January and February (Section 5.2.4) should not be forgotten.

Figure 5.15 Diagrammatic representation of drainage season water balance (Oct to Mar) Rowden Moor (to drain depth 0.6 m). All units in mm.

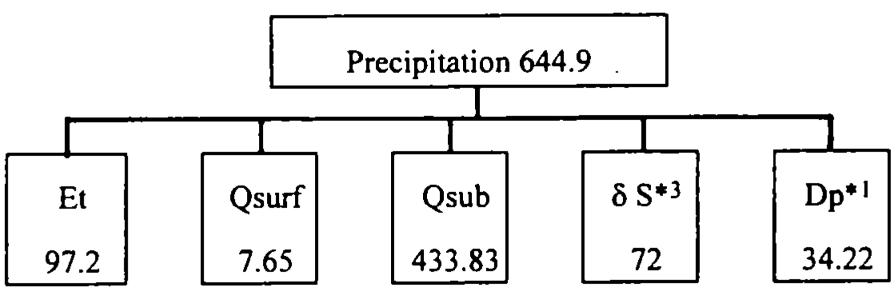
A) Undrained field. $R = Et + (Of + If) \pm S \pm Dp$

$$644.9 = 97.2 + (335.4) + 120 + 92.3$$



B) Drained field.. $R = Et + (Of+If) + Df \pm S$

$$644.9 = 97.2 + (7.65) + 433.83 + 138$$



*1 deep percolation or other 'leaks'

*2 over 0.8 m = 160 mm (to account for Dp soil moisture would have to increase to 1.06 m depth).

*3 over 0.8 m = 96 mm (to account for Dp soil moisture would have to increase to 0.86 m depth).

Once more, the dominance of surface pathways on the undrained field (50 % of drainage season rainfall, 61 % of monitored output) and subsurface pathways on the drained field (67 % of drainage season rainfall) is demonstrated. Increased soil water (11 %) is less important on the drained than on the undrained field (19%). The improved structure and higher organic matter content of the drained soil (Sections 4.2.2 and 4.2.7) resulted in higher moisture contents in early October and the soil dried more quickly in March (Section 5.4.4).

5.6 Summary.

Using a water balance approach, field runoff and soil moisture patterns have been described under natural and drained conditions at the field scale. The main conclusions, based on these observations, are:

(i) based on T.D.R. data the soil on the undrained field was at saturation from November. However, this was not so consistently demonstrated by tensiometry;

(ii) the major pathway on the undrained field was overland and shallow subsurface flow (Of + If);

(iii) volumetric soil moisture content on the drained field was usually below saturation;

(iv) regardless of (iii), drain discharge from the drained field began earlier as connectivity of macropores and wetted areas had been augmented;

(v) on the drained field, *c.* 60 % of net precipitation left the field via the drains;

(vi) water budget calculations for the undrained field showed that the volumetric soil moisture content reached saturation on 1.11.90, and during October the net precipitation of 78 mm was exactly balanced by increased soil moisture storage.

CHAPTER 6 PLOT HYDROLOGY.

6.1 Introduction.

In Chapter 5, hydrology at the field scale through the drainage season was considered. Analysis of hydrographs, flow duration data and runoff coefficients indicated that the relative importance of surface and subsurface water pathways varied through the drainage season and according to management regime (Section 5.3). To further investigate the role of soil properties and impact of drainage on soil water movement mechanisms, this chapter will examine hydrology at the plot scale, as defined in Section 3.1.1.

Soil matric potential and moisture content were monitored on each plot (Section 3.3.7 and 3.3.8) vertically through profiles, and horizontally along transects. The spatial scales under consideration were, therefore, 50 cm^3 sampling volumes (tensiometers) found within 1 m horizontal transects, and 0.6 m soil profiles (T.D.R.). The temporal scale was reduced to half hourly intervals, at a minimum and daily intervals within a monthly context, at a maximum.

Examination of soil matric potential data (Section 6.2) with precipitation, soil moisture storage and field discharge patterns, enabled inferences to be made regarding the nature of subsurface flow. Such hydrometric observation aimed to identify subsurface water fluxes and pathways. The water flow subroutine of LEACHM was employed as a further aid to interpretation of field data (Section 6.3).

Data were collected throughout the drainage season (October to March; Section 5.3.1). Periods discussed in detail occurred first, during autumn wetting when drainflow was initiated (26.10.90 to 27.11.90) and second, during the recession of surface and subsurface discharge (27.2.91 to 30.3.91). These two periods were chosen to investigate the relative importance of various pathways of field runoff, based on differences identified at the gross scale in Chapter 5. Summary data give some indication of seasonal patterns of rainfall and runoff relationships (Table 6.1), but they mask variations within these time periods which are discussed in more detail in Sections 6.2 and 6.3. Drained plot matric potential results were automatically logged at half hourly intervals (Section 3.3.7). Operational difficulties with the undrained plot datalogger led to manual recording throughout the study period. Therefore, matric potential data on the undrained field were collected less frequently, but still allowed comparison of soil hydrological response under different management regimes.

For each plot, discussion of matric potential in November and March are juxtaposed (Section 6.2). Observations are then compared with a simulation of one- and two-domain flow (Section 6.3). Soil hydrological response on each plot is then considered in terms of soil moisture contents (Section 6.4). In Section 6.5, work reported in this chapter is summarised and discussed.

Table 6.1 Rainfall and runoff: selected data for Rowden Moor (including % runoff coefficients)

Dates	Rainfall	Undrained field discharge	Drained field drain discharge	Drained field surface discharge
26.10 - 30.11.90	90.2 mm	31.38 mm 34.8 %	46.1 mm 56.1 %	0.0 mm 0 %
27.02 - 31.03.91	96.9 mm	54.32 mm 56.1 %	78.9 mm 81.9 %	0.4 mm 0.5 %

6.2 Soil water potential.

6.2.1 Introduction.

The principles governing soil water movement are discussed in Section 1.6, where the Darcian approach and the subsequent Richards' Equation are outlined. Total potential within the soil is the driving force behind soil water movement and is determined by the combination of the gravitational and matric potentials (Section 1.6).

Values quoted here for matric potential are **positive** when referring to **suction** and **negative** when referring to **head**. Hence, a wetting soil has declining matric potential values, whilst values for a drying soil increase. Matric potential is discussed first for the undrained and then the drained site. Temporally, some seasonal trends are outlined, highlighting the dynamics of wetted areas, before discussing the two periods chosen to investigate soil water transmission at different stages of the drainage season. Spatially, discussion focuses on variation of matric potential in two directions: vertically (in profiles) and horizontally (along transects).

6.2.2 Undrained plot: matric potential - November 1990.

On the undrained plot, November rainfall (Figure 6.1) resulted in a highly variable matric potential response. Matric potential data for 1.11.90 to 30.11.90 are presented to illustrate the soil status prior to, and during drainflow initiation (Figure 6.2).

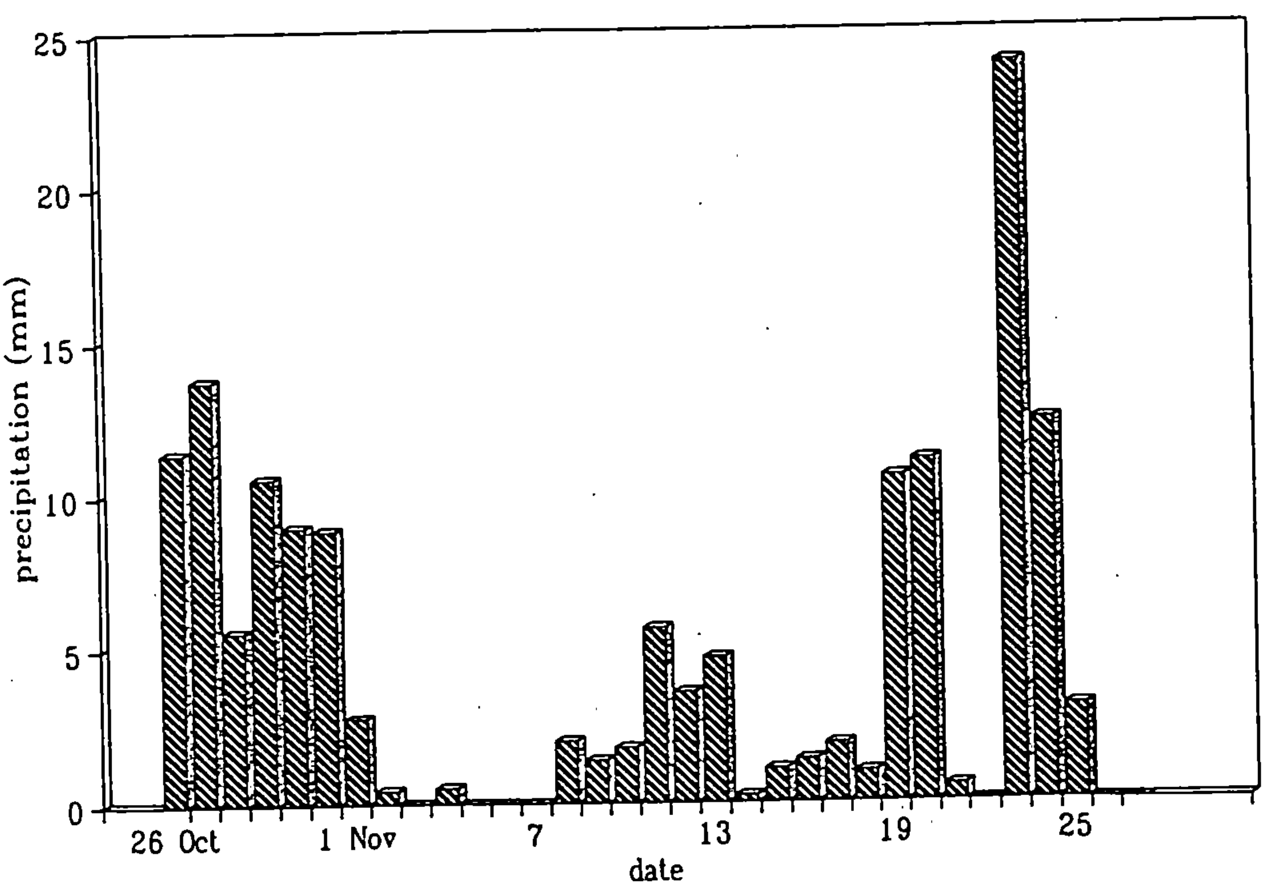
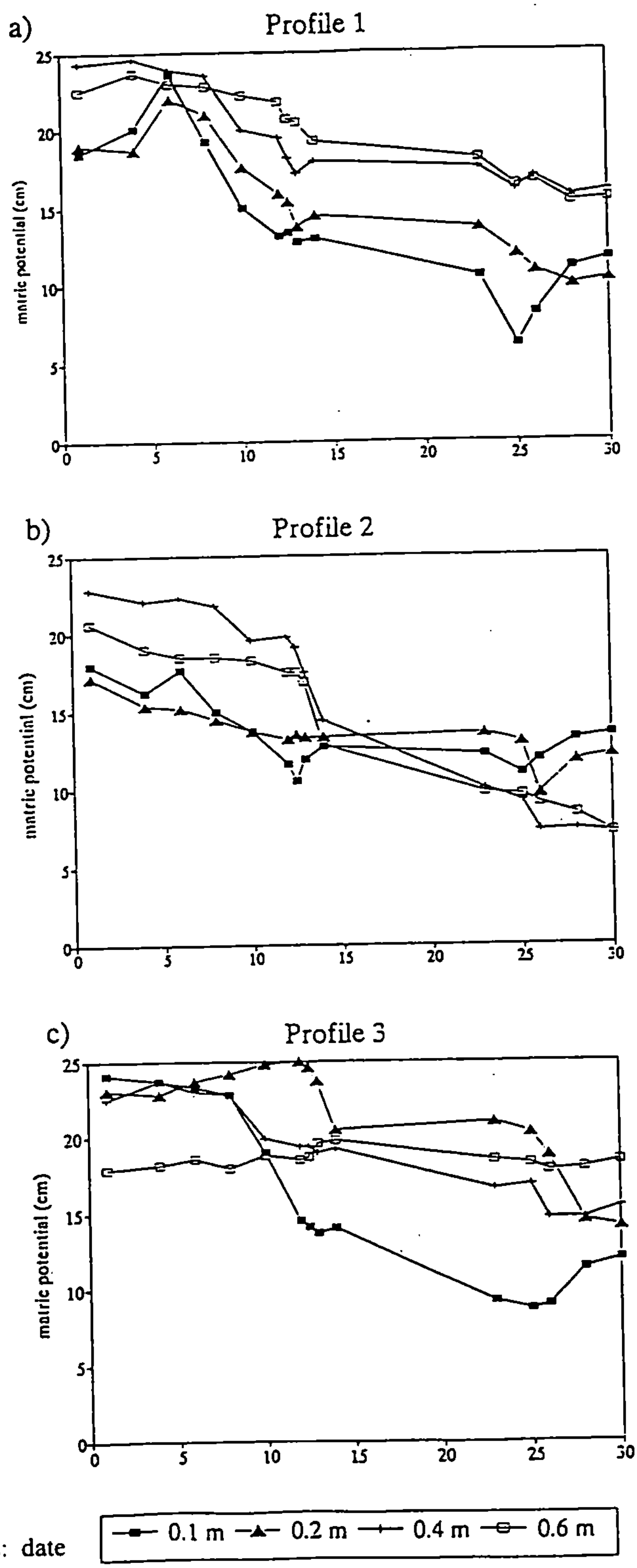


Figure 6.1 Rainfall at Rowden (26.10.90 - 30.11.90).

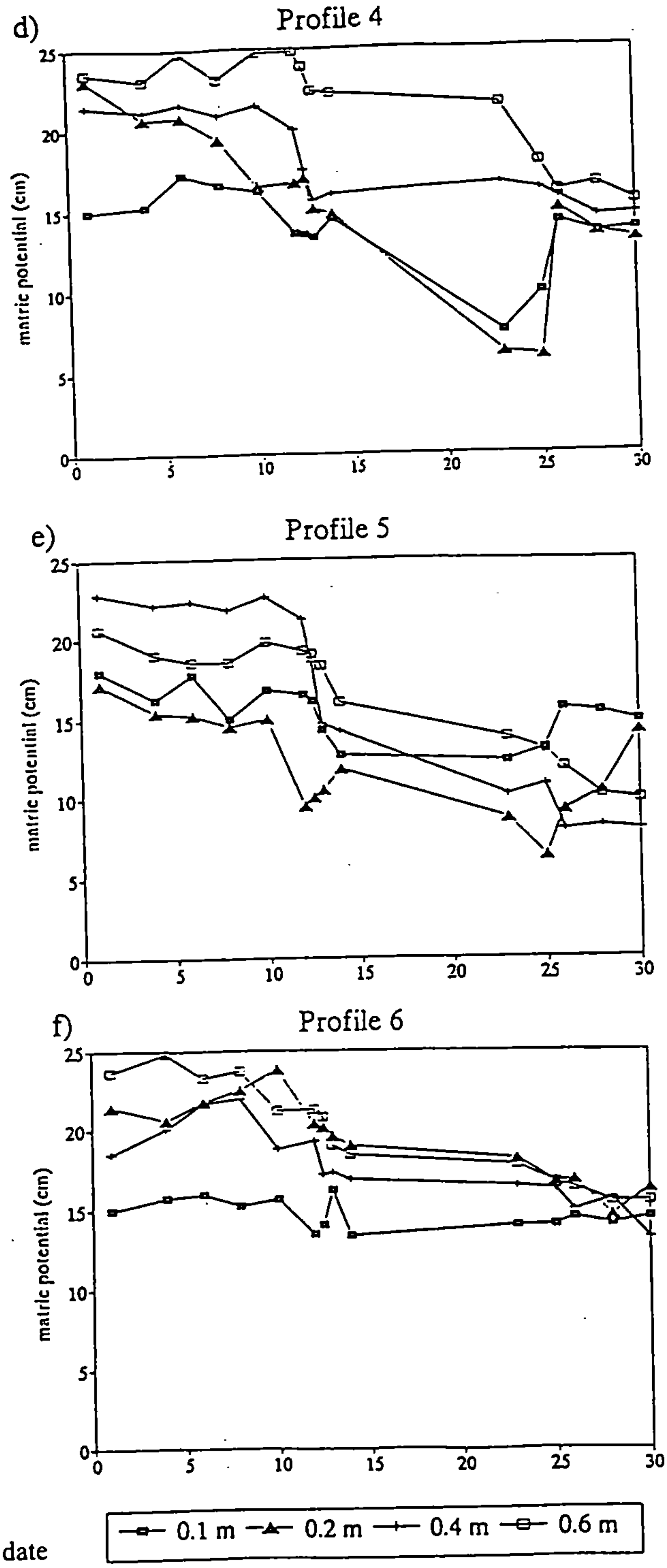
Figure 6.2 Undrained plot: matric potential (November, 1990).



x axis: date

Figure 6.2 continued

Undrained plot: matric potential (November, 1990).



x axis: date

Table 6.2 presents matric potential data for the four depths of each profile on 12.11.90. This date was chosen from a visual inspection of Figure 6.2 to give representative results. The data show that the soil was wet throughout the profile, being 13 cm H_2O (suction) at the surface and 18 cm H_2O at the base. This is equivalent to a volumetric soil moisture content of about 43 % at the surface as compared to 42 and 44.7 % at field capacity and saturation, respectively (derived from the soil moisture curve, Section 4.2.2). With the exception of profile 5, the upper 0.1 m was the wettest portion of each profile. Soil at 0.6 m was driest in three of the six profiles. Table 6.3 includes data for 14.11.90, when there had been a further 6 mm of rain. Despite wetting at most sites, this shows that the relative status of different sites was unchanged. Additional information is given in Table 6.4 which details matric potential data for 24.11.90, when a further 12.4 mm of rain had fallen, one week after field runoff was initiated (Figure 6.3). Between 14.11.90 and 24.11.90, matric potential values in the upper 0.1 m and at the base decreased by *c.* 3 cm H_2O, however, response was more variable at the surface. These data show that the surface and subsurface are connected.

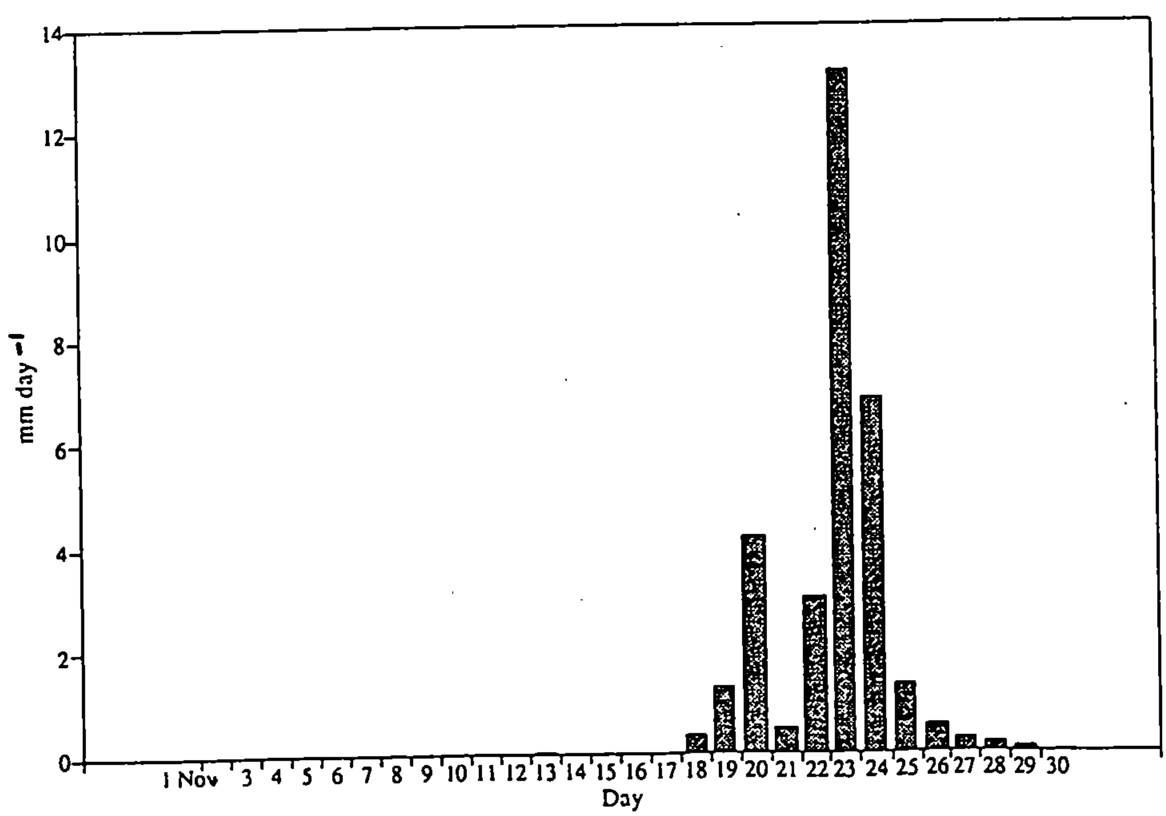


Figure 6.3 Undrained field: field discharge (November 1990).

Table 6.2 Undrained plot: matric potential values (12.11.90) (cm H_2O)

Profile	0.1 m	0.2 m	0.4 m	0.6 m
1	13.4	15.4	18.2	20.7
2	10.5	13.5	19.1	17.5
3	14.1	24.6	19.4	18.7
4	13.5	17.0	17.5	23.9
5	16.1	10.0	18.5	19.0
6	14.0	20.2	17.2	20.8

Table 6.3 Undrained plot: matric potential values (14.11.90) (cm H_2O)

Profile	0.1 m	0.2 m	0.4 m	0.6 m
1	13.0	14.5	18.0	19.3
2	12.7	13.3	14.4	12.8
3	14.0	20.5	19.2	19.8
4	14.5	14.8	16.0	22.2
5	12.7	11.8	14.2	16.0
6	13.3	19.0	16.9	18.4

Table 6.4 Undrained plot: matric potential values (24.11.90) (cm H_2O)

Profile	0.1 m	0.2 m	0.4 m	0.6 m
1	6.2	12.0	16.1	16.4
2	11.0	13.0	9.2	9.6
3	8.7	20.3	16.9	18.3
4	9.9	6.0	16.2	17.9
5	13.1	6.3	10.8	13.0
6	14.0	16.8	16.3	16.7

Potentials monitored in profile 1 were between 18.5 and 24.3 cm H_2O on 1.11.90 and declined to 10.4 to 16.0 cm H_2O by the end of the month. Inspection of Figure 6.2 shows that matric potentials declined rapidly from 6.11.90 to 14.11.90 when 20 mm of rain fell. The matric potential declined more slowly throughout the rest of the month, despite 67 mm of further rainfall.

Potentials at 0.1 and 0.2 m were approximately 5 cm H_2O lower than those at 0.6 m, however, the wetting curves were parallel. As rainfall ceased on 25.11.90, the upper soil drained causing potentials to rise, while the lower soil continued to wet and thus the matric potential curves deviated from their earlier pattern (Figure 6.2 a).

Profile 2 exhibited a more complicated temporal pattern than profile 1 (Figure 6.2 b). Matric potentials in the surface horizons remained relatively constant at *c.* 14 cm H_2O, while the lower horizons declined from 18 to 9 cm H_2O. Therefore, the soil at depth became wetter, while the surface did not respond to incident rainfall. So, by exploiting desiccation cracks, water may have bypassed the upper matrix. Soil segments in profile 3 also exhibited two different patterns of response. Values of matric potential at the surface declined from 24 cm to 8.7 H_2O throughout the period (Figure 6.2 c). In contrast, the matric potential at 0.6 m depth remained constant throughout, at 17 cm H_2O.

There was a shallow decline in matric potential overall, throughout profile 4 (Figure 6.2 d). When the pattern for each depth was examined in detail, it was found that the surface horizons were more responsive to heavy rainfall between 19 and 24.11.90 (14.6 to 6.8 cm H_2O) but they also drained quickly: potentials rose to 14.2 cm H_2O once more on 24.11.90. In contrast, the lower horizon showed no response at 0.4 m, while at 0.6 m the response was delayed (21.5 cm H_2O on 23.11.90 to 16.1 cm H_2O on 26.11.90). The surface horizons (0.1 and 0.2 m) in profiles 4 and 5 responded in a similar manner. Figure 6.2 e (profile 5) shows the greatest decline in potential occurred at 0.4 m (21.2 cm H_2O on 12.11.90, to 14.2 cm H_2O on 14.11.90). This coincides with continued rainfall (9 mm) and increasing matric potential (drainage) at the surface.

Profile 6 had the highest matric potentials, ranging from 15 to 23 cm H_2O on 1.11.90. The surface horizon which was relatively wet (15 cm H_2O) and changed little (Figure 6.2 f). The lower three depths dried initially but responded to rainfall from 10.11.90, this pattern being similar to that described for profiles 4 and 5. By the end of the month, the potentials of the lower segments had declined from between 18.5 and 23.6 cm H_2O to 15 cm H_2O.

Examination of the individual segments for profiles 1 to 6 showed that the suction at 0.1 and 0.2 m was, generally, in the lower range for each profile. Potentials decreased in early November and were relatively uniform by 12.11.90. From 25.11.90, potentials increased in response to percolation and evapotranspiration of soil water. At 0.4 m, potentials were often high before the period of regular rainfall. Response to rainfall was sometimes lagged (profiles 4 and 5). The response at 0.6 m varied from no observable change (profile 3) to gradual or stepped wetting (profiles 1 and 4, respectively). This illustrates the slow progression of wetting on the undrained plot, the impact of the hydraulic discontinuity at approximately 0.2 m (Chapter 4) and a lack of observed pathways which transmitted water to depth.

In the last six days of October, 59.3 mm of rain fell, an additional 24 mm in the first half of November lowered the mean potential from 20.5 to 15.9 cm H_2O. At this stage, field runoff was initiated, explaining the fate of additional rainfall. Despite a further 66.8 mm of rainfall, mean potential was still 8.8 cm H_2O at the end of the month. This result suggests that the soil could not imbibe the water rapidly enough to affect the matric potential, so infiltration excess or shallow interflow was generated. Alternatively, rainfall may have been absorbed and rapidly transmitted as throughflow throughout the profile, hence matric potential and soil storage was not affected. However, the low hydraulic conductivity values recorded (Section 4.2.5) preclude the second explanation.

Results for the wetting undrained plot can be summarised as follows:

1. there are three possible types of response:
   a. rapid response, e.g. profile 4, 0.2 m (Figure 6.2 d) cracked soil, highly conductive;
   b. nil response, e.g. profile 3, 0.6 m (Figure 6.2 c) massive structure;
   c. gradual/stepped/lagged response, e.g. profile 3, 0.2 m (Figure 6.2 c) 0.6 m profiles 1 and 4 (Figure 6.2 a and d) slow matric percolation;
2. soil was wettest at the surface;
3. a. soil wetted up quickly in early November (7 cm H_2O decline in 14 days);
   b. soil wetted up very slowly later (on 30.11.90 values ranged from 7 to 18 cm H_2O and the range declined to 6 to 15 cm H_2O by 29.12.90).

6.2.3 Undrained plot: matric potential - March 1991.

Rainfall for March 1991 (Figure 6.4) and matric potentials for the various depths in each profile (Table 6.5 and Figure 6.5) are presented. By March 1991, four transducers had broken, so data for their locations do not appear. Further, in profile 6, maintenance of a transducer resulted in loss of some data for 0.6 m.

Figure 6.4 Rainfall at Rowden (1.3.91 - 30.3.91).

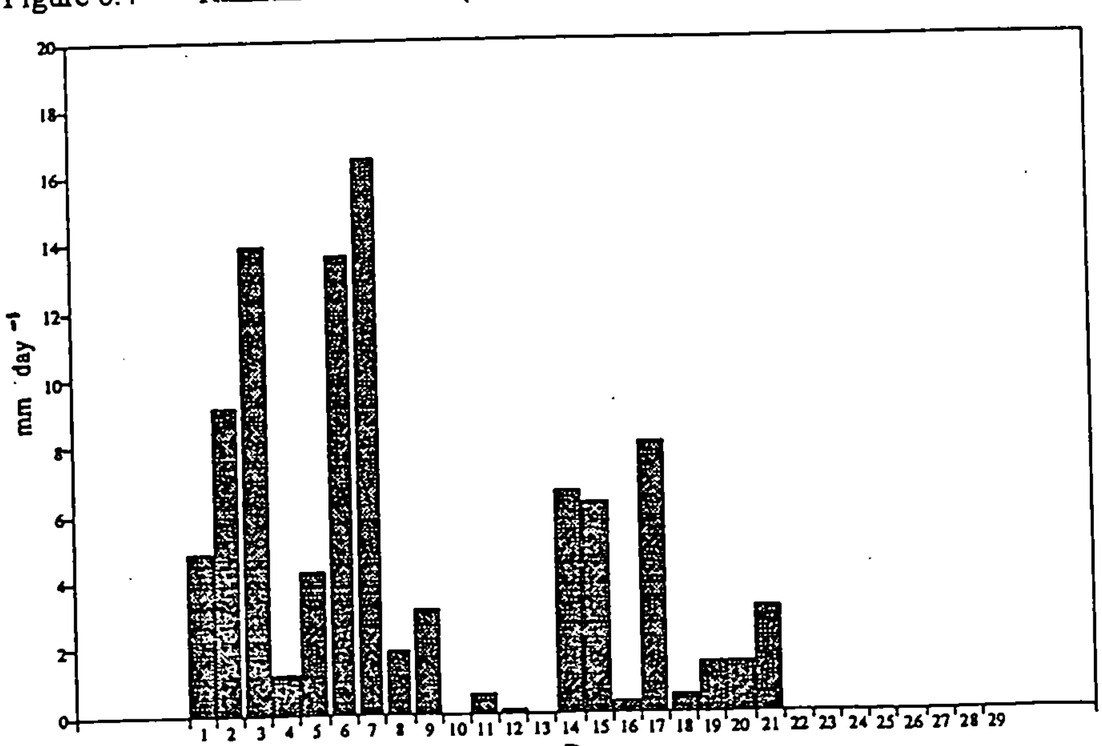


Table 6.5 Undrained plot: matric potentials (15.3.91) (cm H_2O)

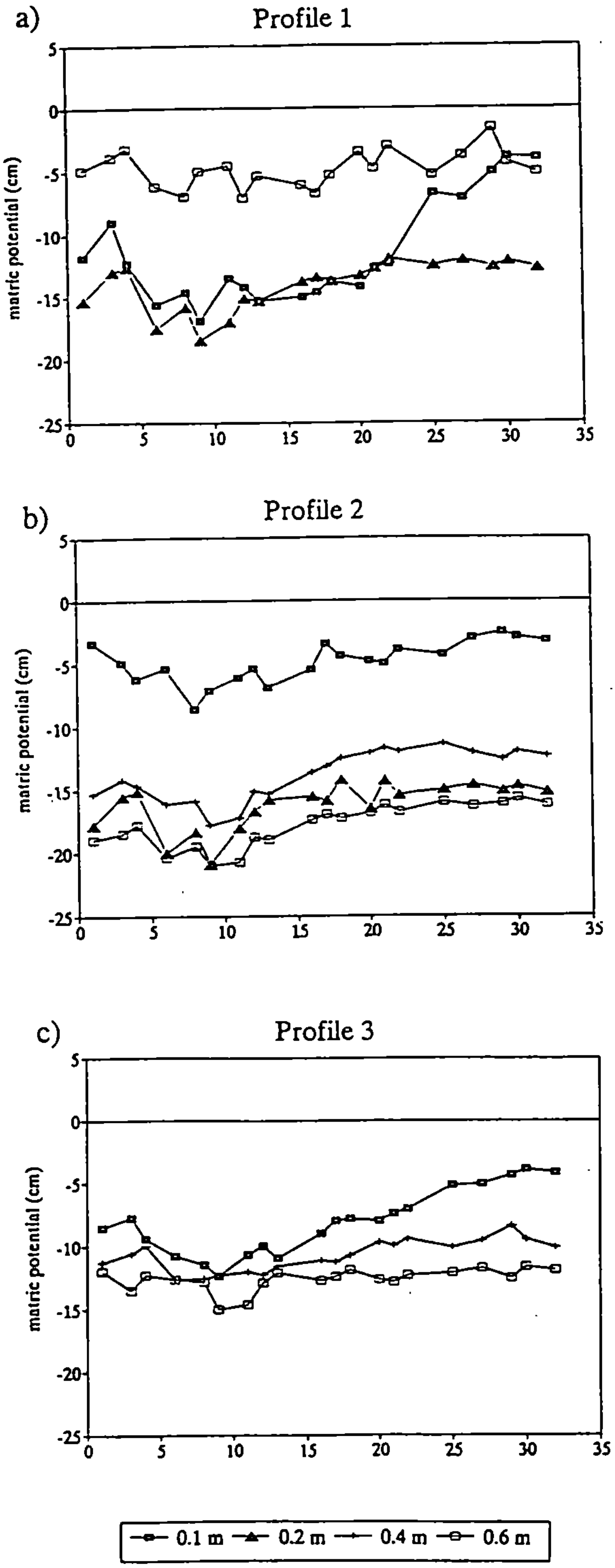
Profile	0.1 m	0.2 m	0.4 m	0.6 m
1	-14.6	-13.4	—	-6.8
2	-3.4	-15.9	-13.2	-16.9
3	-8.0	—	-11.2	-12.4
4	—	—	-10.3	-15.3
5	-7.0	-6.0	-13.0	-20.0
6	-6.0	-11.0	-13.0	<-17.0

Table 6.5 shows that the soil was saturated in most cases. The matric potential ranged from -20 cm H_2O at depth to -3.4 cm H_2O at the surface, i.e. pressure head was +20 to +3.4 cm H_2O. With the exception of profile 1, the base of the profile had the greatest head, averaging -16 cm H_2O. The soil in the upper 0.1 m had the lowest head within the profile (*c.* -8 cm H_2O). Despite matric potential values which indicated saturation at every depth in every profile, the absolute depth of saturation recorded by tensiometers was not uniform and cannot, therefore, be attributed to the presence of a water table. This agrees with the observations of Holden for the nearby Crediton series (Holden pers. comm.).

Potentials monitored in profile 1, ranged between -15.3 and -4.9 cm H_2O at the end of February. During the first week of March, all depths registered a decrease in matric potential (Figure 6.5 a) but by 14.3.91, all depths were drying, this being most apparent at 0.1 m. At the end of March, values ranged from -12.7 to -4 cm H_2O. The marked increase in potentials at 0.1 m depth (-12.4 to -6.8 cm H_2O from 20.3.91 to 23.3.91) coincided with the cessation of rainfall in March (Figure 6.5 a). Matric potential response at 0.2 m followed the trend at the surface. At 0.6 m, the potential fluctuated around -5 cm H_2O, with some evidence of lagged responses compared with the upper horizon.

In profile 2, matric potential ranged between -19.0 and -3.3 cm H_2O on 27.2.91. All depths responded to rainfall from 1.3.91 to 9.3.91 before draining slowly until the end of the month. On 27.2.91, potentials in profile 3 were similar (-12.0 to -8.5 cm H_2O)

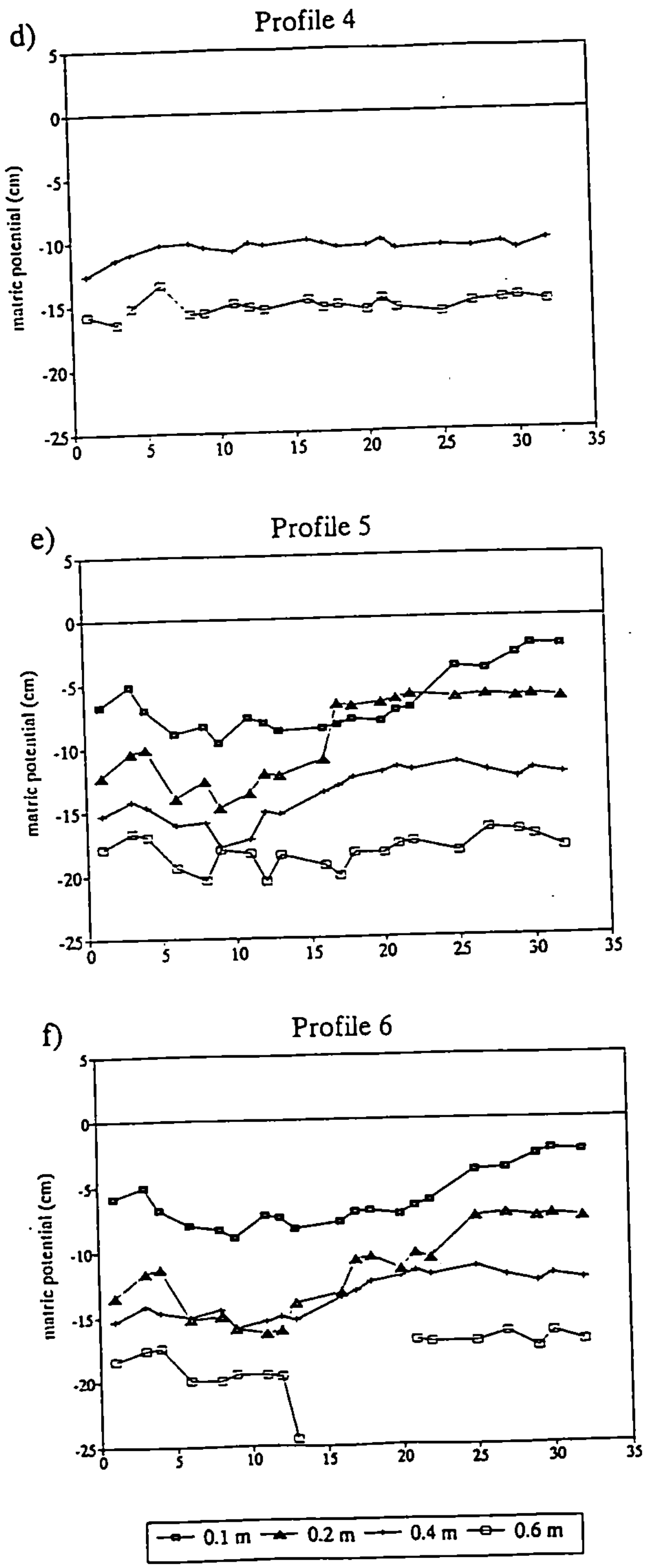
Figure 6.5 Undrained plot: matric potential (27.2.91 - 30.3.91).



x axis: date

Figure 6.5 continued

Undrained plot: matric potential (27.2.91 - 30.3.91).



x axis: date

and converged on 7.3.91 (*c.* -12 cm H_2O). Values diverged as the surface soil dried through the rest of the month, while potentials at the base of the profile were constant. Similarly, there was little response at either 0.4 or 0.6 m in profile 4 (Figure 6.5).

In profiles 5 and 6 the surface soil wetted following rainfall in early March and remained fairly constant until 19.3.91 when the soil began to dry (Figures 6.5 e and f). Response at 0.2 m was more marked, potentials ranging between -16 cm H_2O on 10.3.91 to -8 cm H_2O on 23.3.91 in profile 6, for example. At 0.4 m depth matric potentials in both profiles were almost identical. Drying was marked from 7.3.91 until 19.3.91 after which potentials levelled out (*c.* 13 cm H_2O). Values at depth were relatively constant, ranging between -20 and -17 cm H_2O.

Figure 6.6 illustrates the erratic nature of the undrained field runoff. The soil was already wet at the beginning of March 1991 (e.g. -9 cm H_2O) but matric potential decreased by about 5 cm H_2O in response to high rainfall (60 mm) in the first 8 days of March. During this period, 24.3 mm of field runoff were generated. Subsequently, drainflow ceased although matric potential was maintained at *c.* -8 cm H_2O. From 15 to 23.3.91, a further 34 mm of more evenly distributed rainfall resulted in minimal matric potential response and minimal field discharge (drainage stopping on 21.3.91).

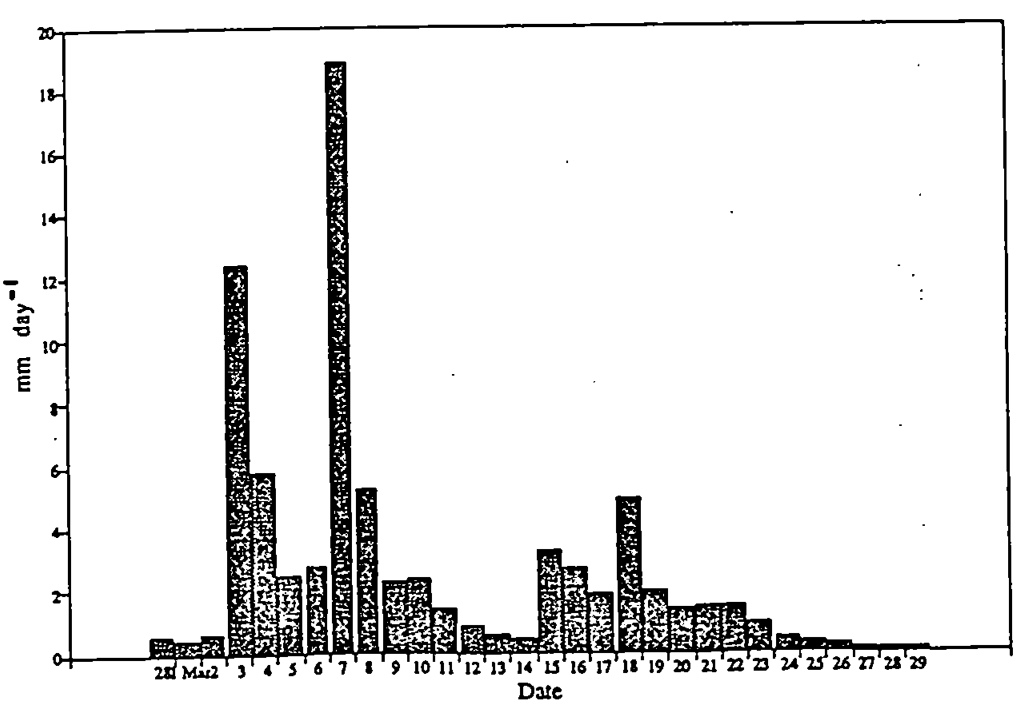


Figure 6.6 Undrained field : field discharge (March 1991).

Results for the undrained plot in March can be summarised as follows:

1. the soil was largely saturated throughout the month (negative matric potential). Rainfall resulted in decreasing potentials; on cessation of rainfall, potentials increased minimally. Hydraulic conductivity was low, and there was a lack of exit routes/connectivity of macropores. Therefore, water remained in storage;
2. no water table was observed;
3. 66 mm of rain in the first 8 days of March resulted in 34 mm of field discharge. However, low intensity rain in the third week of March produced negligible runoff;
4. head was lowest at the surface and highest at the base;
5. unlike the wetting period (Section 6.2.2), it was difficult to delineate types of response;

6.2.4 Undrained plot: matric potential - summary.

Through the autumn and spring there were some tensiometers which responded in a similar manner within the plot and a few of these were adjacent to one another. However, it was not clear whether adjacent tensiometers (in horizontal, vertical or diagonal directions) always sampled continuous soil volumes or discrete pockets. During the wetting period, matric potential was most variable. In terms of its hydrology, the soil was anisotropic (Figure 6.2 a to f). In March, the soil was saturated and heads were less variable.

Evidence of response at 0.6 m in November particularly, but also in March, indicated that water was moving vertically through the soil in saturated/near saturated conditions. With the low conductivity of the soil (Section 4.2.5) it is difficult to attribute this to transport through the matrix. However, fissure flow could convey water to depth, especially in November when the soil cracks remained open.

The observed lagged responses highlighted above can be attributed to tortuous pathways as well as the operation of a number transmission mechanisms, i.e. fissure flow and throughflow. Such an argument is substantiated by the patterns of response in profiles 3 and 6 in November, i.e. the matric potential at 0.4 m decreased while that at 0.2 m did not. In contrast, there were no such examples in March.

Wetted areas were observed on the undrained plot as well as on the drained plot (Section 6.2.7). These areas were considered to be sites with notably higher moisture contents (> 44.7 %, Section 4.2.2) and lower matric potential values than the rest of the soil. The size and development of these zones varied in four ways: between fields, between horizons, along the transect and through the drainage season. These zones could be expected to be important in terms of rapid runoff generation.

On the undrained plot, wetted areas were extensive and persistent. For significant periods, entire profiles were wetted (for example, profiles 2 and 5, from December through to March). More frequently, however, wetted areas were not continuous through the profile, being interrupted by drier soil, especially at 0.4 m (Section 6.2.4).

6.2.5 Drained plot: matric potential - November 1990.

Table 6.6 presents matric potential for four depths at mole, mid and quarter mole profiles (for the full set of tensiometers), results will be discussed in this order, noting similarities and disparities along transects and following the structure adopted for undrained plot results above.

Matric response was considerably more variable than that observed on the undrained plot, therefore, it is not possible to generalise about whether the soil was wet or dry, so, each mole drain location will be dealt with separately. For example, at the mole, potentials in the lower three depths ranged from *c.* 10 cm H_2O (equivalent to 46 % volumetric soil moisture content) to more than 300 cm H_2O (equivalent to 35 % volumetric soil moisture content) over a period of five days. Figures 6.7 a, b and c show the matric response of individual tensiometers, exemplifying the behaviour at each location.

Figure 6.7 a Drained plot: matric potentials above the mole drain (November 1990).

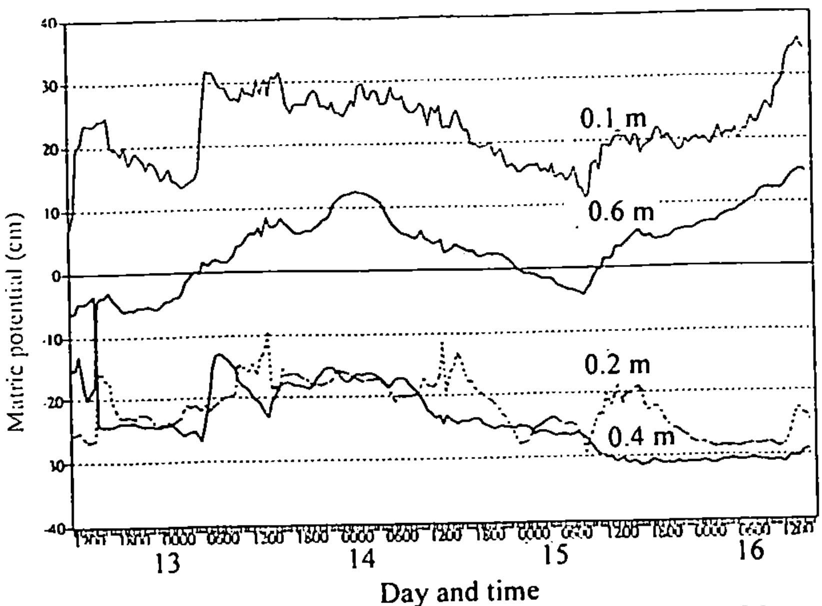


Figure 6.7 b Drained plot: matric potentials at mid mole location (November 1990).

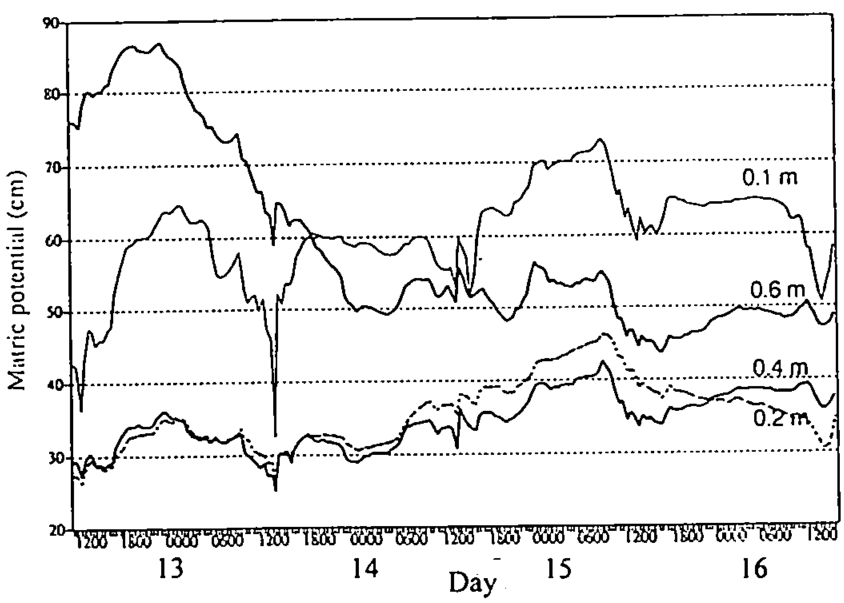


Figure 6.7 c Drained plot: matric potentials at quarter mole location (November 1990).

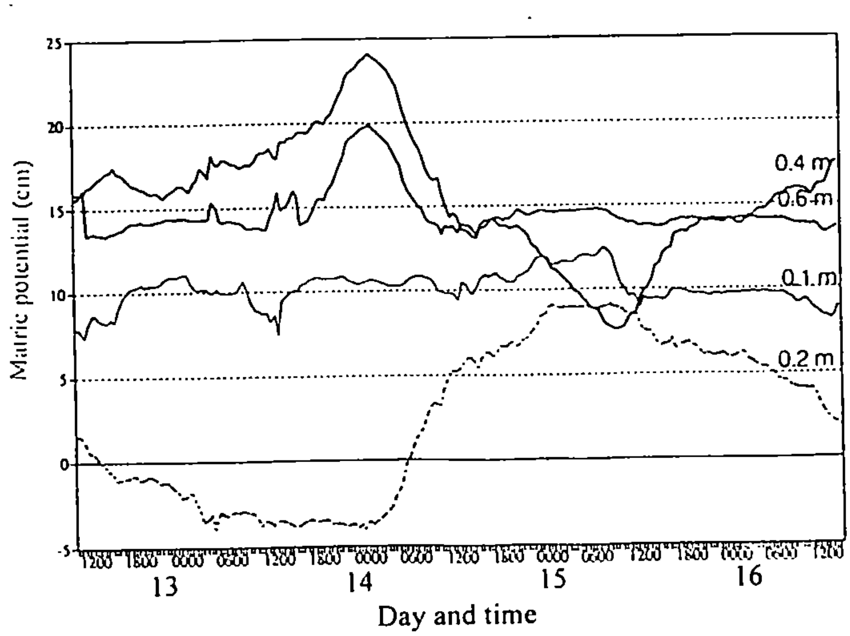


Table 6.6 Drained plot: observed ranges of matric potential, with depth and across transect (12.11.90 - 16.11.90). Where: n = 250; all units in cm H_2O.

Depth (m)	Mole drain			Quarter mole			Mid mole		
	min	max	mean	min	max	mean	min	max	mean
0.1	6.0	34.1	22.3	4.7	13.0	10.5	32.3	73.0	60.1
0.2	-28.0	349	47.7	-1.0	11.9	5.3	24.9	48.2	35.5
0.4	-31.8	339	45.4	1.2	23.9	9.4	3.2	42.3	10.3
0.6	-4.0	300	119.6	4.8	20.1	9.8	16.5	320.2	125

Above the mole drain (Figure 6.7 a), matric potentials indicated that soil was saturated from 0.2 m down. Matric potential between 12 and 16.11.90 was complex, showing a sigmoidal response at the surface and 0.4 m depth. At 1200 h on 12.11.90, values for matric potential were 23 cm H_2O (0.1 m), -27 cm H_2O (0.2 m), -20 cm H_2O (0.4 m), and -5 cm H_2O (0.6 m). Examination of the rainfall event beginning 2200 on 12.11.90 (4.7 mm), shows that values at the surface declined to 15 cm H_2O by midnight, remained almost constant until 0430 h on 13.11.90 (+32, -22, +1 and -20 cm H_2O). They then declined once more from 1200 h on 14.11.90.

The main trends were as follows:

1. at 0.1 m, rapid response to rain followed by rapid drainage;
2. at 0.2 and 0.6 m, little variation;
3. at 0.4 m, rapid response to rain followed by gradual drainage and subsequent wetting, due to percolation.

At mid mole (Figure 6.7 b), soils were unsaturated, i.e., much drier than above the mole. Matric potentials in the upper horizon (0.1 m depth) fluctuated around 55 cm H_2O, the middle horizons (0.2 and 0.4 m) rose from 30 - 37 cm H_2O, while at depth (0.6 m) the tension dropped from 80 to 47 cm H_2O. At 1200 h on 12.11.90, profile matric potentials were 46 cm H_2O (0.1 m depth), 29 cm H_2O (0.2 m), 29 cm H_2O (0.4 m) and 80 cm H_2O (0.6 m), they then rose. However, following the rainfall event beginning 2200 (4.7 mm) matric potentials dropped to 32, 25, 25 and 59 cm H_2O at 1200 h on 13.11.90. After this minimum suction, the surface and middle horizons dried rapidly (54, 30 and 31 cm H_2O at 1400 h) while at 0.6 m depth the soil dried only for a short time (64 cm H_2O) before potentials declined once more falling to 50 cm H_2O at 2200 h on 13.11.90. As Figure 6.7 b illustrates, the rainfall resulted in matric potentials falling until 1200 h on 13.11.90. The soil drained quickly at first, before establishing a more gradual, but fluctuating, increase in potentials which continued until 0800 h on 15.11.90. These rising traces for 0.1, 0.2 and 0.4 m were approximately parallel, while that for 0.6 m was more shallow. Matric potential then levelled out throughout the profile, only increasing slightly before 1000 h on 15.11.90 when 1.4 mm of rainfall caused a further drop. By 0000 h on 16.11.90, potential at the same locations had equilibrated: (65, 36, 39 and 49 cm H_2O). The two stage change in soil status during drainage of water substantiates the conclusions drawn from determination of the water retention characteristic (Section 4.2.2), that there is a bimodal pore size distribution.

Rainfall on subsequent days: 0.2 mm (13.11.90); 1.1 mm (14.11.90) and 1.4 mm (15.11.90) resulted in the troughs shown in Figure 6.7 b. The main trends are that the 0.1 to 0.4 m showed evidence of wetting due to rainfall of a few hours duration. The 11 mm of rain during this period had no overall impact on matric potential at these depths, showing that the water was bypassing the soil peds. This is substantiated by results at 0.6 m which showed a pattern of wetting up.

The range of values at mid mole is summarised in Table 6.6, demonstrating that this profile was far less variable than the mole profile. There was limited evidence of delays in response between the various depths. One explanation may be that lags were not identified due to the temporal resolution of monitoring (30 minutes). At mid mole, response was very peaky at 0.1 m, improved structure with lots of large pores (Section 4.2.4) gave highly variable matric potential. At 0.2 and 0.4 m, the suctions were relatively low and had correspondingly low ranges. These parts of the profile were probably hydrologically connected, as they responded together. The pattern at 0.6 m was the most consistent. Matric potentials continued to decrease throughout the period due to continued percolation.

On 12.11.90, soil was only saturated at 0.2 m in the two quarter mole profiles, where the hydraulic discontinuity resulted in saturation for 41 hours (Figure 6.7 c). Potentials ranged between zero and +16 cm H_2O at1200 h on 12.11.90. In the first profile, the already wetted soil at 0.1 m continued to be well drained, so that matric response was very damped. Drying at 0.1 and 0.2 m at mid and quarter mole locations was coincident. Potentials were similar at 0.4 and 0.6 m, a rise in values was evident from 1800 h on 13.11.90, and they peaked simultaneously at 0000 h on 14.11.90. The initial steep rise can again be attributed to drainage of gravitational water in large cracks (Section 4.2.2). Unlike the other two profiles on this plot, there was only limited evidence of decreased matric potential at 0.2 and 0.4 m, prior to drainflow increase (Table 6.7).

Table 6.7 Drained field precipitation and discharge (mm)

Date	Precipitation	Drainflow
12.11.90	3.6	0.097
13.11.90	4.7	0.334
14.11.90	0.2	0.118
15.11.90	1.1	0.096
16.11.90	1.4	0.096

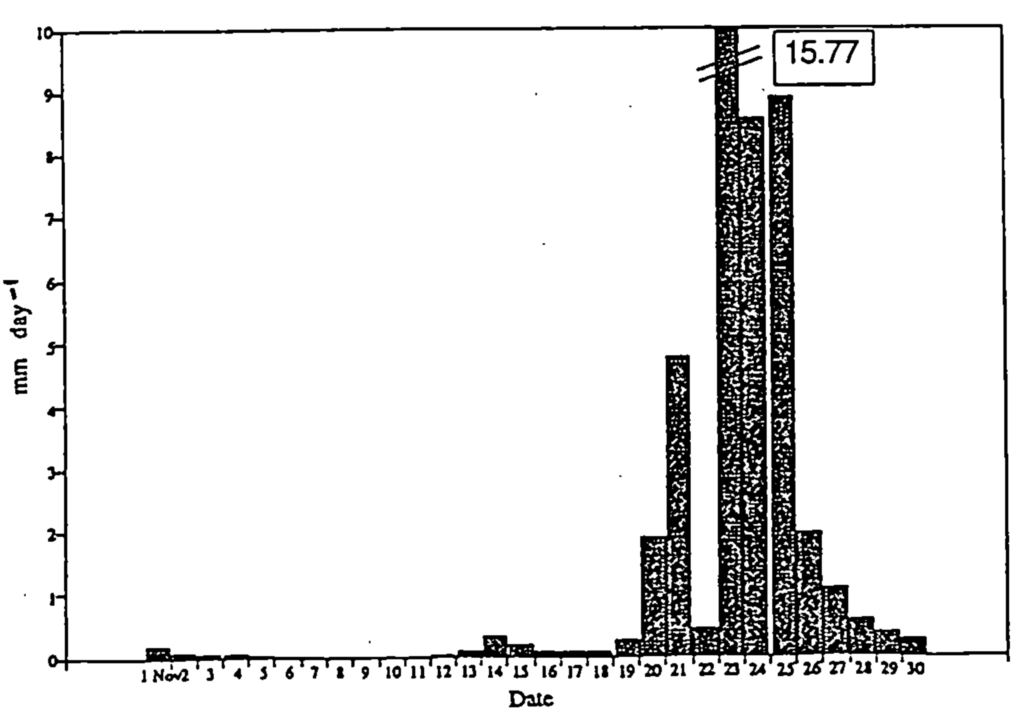
Note: no non drainflow was recorded during this period.

Across the transect, low potential values at 0.2 m (means from 5.3 to 47.7 cm H_2O; Table 6.6), demonstrate the hydrological significance of the Ag/B horizon boundary. Here, changes in texture and structure, cause a decrease in conductivity resulting in temporary waterlogging. As well as inhibiting percolation, this boundary may also represent a divide in terms of capillary water movement. Water below the boundary is held at higher matric potentials, in smaller pores. Fissuring and increased soil porosity caused by mole drainage (Section 1.2) may reduce the discontinuity between the hydraulic conductivity of different horizons. This is substantiated by the similar mean matric potential values for 0.2 and 0.4 m over the mole, compared with the much lower values at 0.2 m at quarter and mid mole positions (Table 6.6). This would, in turn, reduce potential gradients.

Matric potential response and field discharge: persistent field drainage began on 12.11.90 (Figure 6.8), and during the period 12 to 16.11.90 total discharge was 0.659 mm. The soil wetted through 11 and 12.11.90, matric potential declined over the mole and discharge increased to 0.014 mm hr^{-1} at 0330 h on 13.11.90. A plateau in matric potential was maintained with higher discharges as drainflow decreased from 2230 h on 13.11.90 to 0030 h on 15.11.90. Matric potentials declined slowly thereafter.

Drainflow was initiated after potentials at all depths and across the transect had declined for approximately 9 hours. This period of wetting was necessary to produce sufficient head and connectivity of wetted areas to facilitate drainflow, i.e. an area of soil which provided a pathway to the drain had to be wet before flow could begin. At mid mole, matric potentials declined to their minimums around midday on 13.11.90 which was soon followed by the maximum drain discharge. As the soil profile dried, receding discharge was observed.

Figure 6.8 Drained field: drain discharge (November 1990)



6.2.6 Drained plot: matric potential - March 1991.

Matric response from 27.2.91 to 22.3.91 was examined to study soil response at the end of the drainage season. At 0.1 m, a diurnal pattern of matric potential was evident. However, this did not obscure the overall trend of increased matric potential. When daily potential was calculated as the mean of troughs and peaks, the overall increase in matric potential values was 30 cm H_2O in March.

Above the mole, suctions at the lower two depths were low and similar in late February (Figure 6.10). However, throughout early March the soil at 0.4 and 0.6 m appeared more responsive to weather conditions. For example, while the soil was drying on 1.3.91 the matric potential at 0.4 and 0.6 m depths rose by 6 and 3 cm H_2O respectively. The narrow range of potentials was maintained through March. Rainfall on 15.3.91 and 16.3.91 (Figure 6.10) caused declining potentials in the upper three depths. This occurred first at 0.1 m, and shortly afterwards at 0.2 and 0.4 m. Potentials at 0.2 m depth stabilised, whilst those at 0.1 and 0.4 m continued their descent. Further light rainfall on 17.3.91 sustained the decline of potentials at 0.1 m, while that at 0.4 m can be attributed to continued percolation. Following 8 mm of rainfall, potentials at 0.1 and 0.4 m depths crashed on 17.3.91 and response of a lower magnitude was observed at 0.2 and 0.6 m. At the lower depth response was lagged by 14 hours. Rainfall in the next four days temporarily halted the recovery of potentials at 0.1 and 0.4 m

In late February, the potentials at mid mole were more variable than those above the mole. At 0.1 and 0.4 m depths, the soil was relatively dry, resulting in high suctions (*c.* 14 cm H_2O). Potentials at 0.2 m and especially at 0.6 m were markedly lower (*c.* 2 cm H_2O). Throughout the month there was little of a discernible pattern in the potentials at 0.1 and 0.4 m depth, as the soil behaved erratically. However, at 0.2 and 0.6 m there was evidence of drying in the form of rising potentials resulting in concave curves (Figure 6.9 b). The response at 0.2 m depth was markedly shallower. Rainfall in the second week of March resulted in obvious decreasing potentials at 0.1 m depth. As potentials increased overall, the occurrence of rainfall did not always result in the same response at lower depths. For example, on 12.3.91 the soil responded at all depths but on 13.3.91, despite a

Figure 6.9 a Drained plot: matric potentials above the mole drain (March 1991).

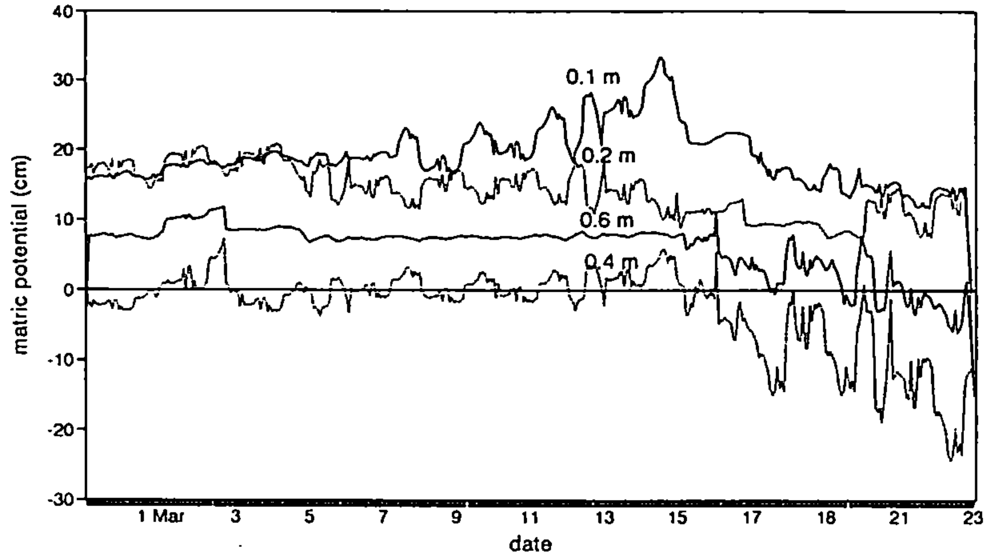


Figure 6.9 b Drained plot: matric potentials at mid mole location (March 1991).

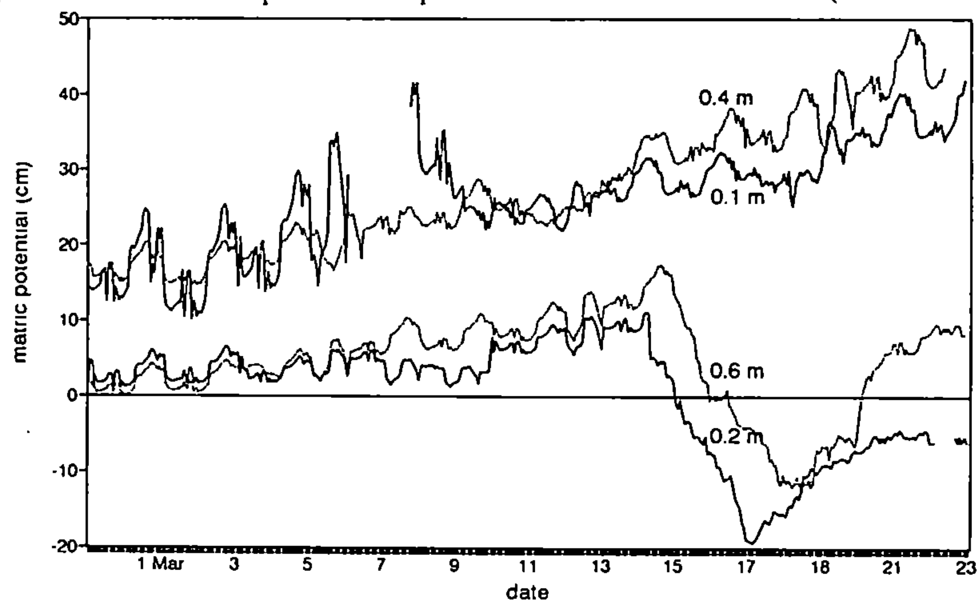
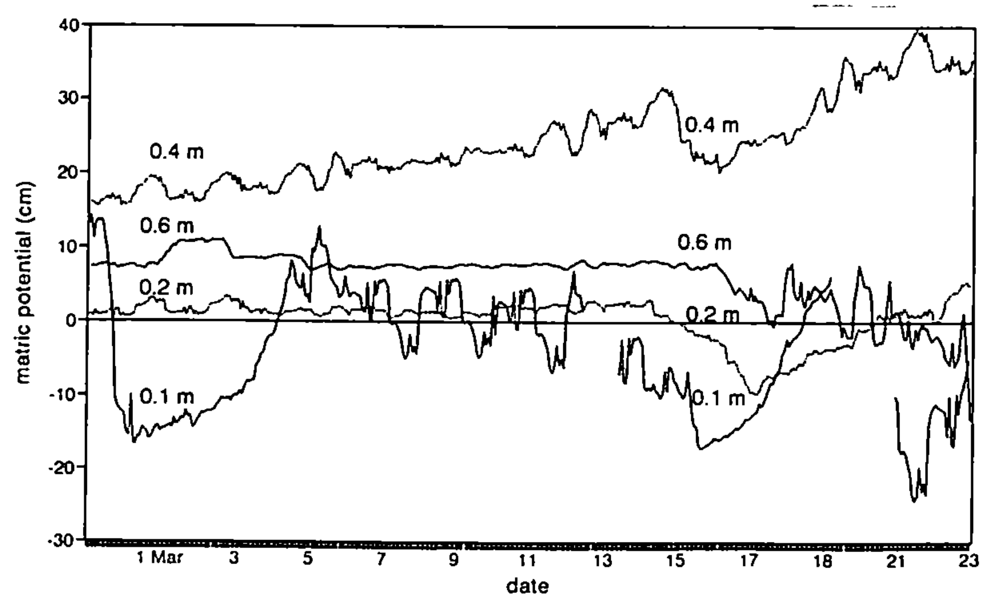


Figure 6.9 c Drained plot: matric potentials at quarter mole location (March 1991).



greater volume of rainfall, the tensiometers at 0.2 and 0.4 m did not respond. Following 23 mm of rainfall between 15 and 19.3.91, potentials at 0.2 and 0.6 m fell once more and the soil was saturated (*c.* -20 cm H_2O). However, this response was lagged by about 14 hours (Figure 6.11). Despite continued low intensity precipitation between 19 and 22.3.91 (6.6 mm), potentials rose at 0.2 and 0.6 m depth so that by 22.3.91, the potentials through the profile were similar once more (-5 to +8 cm H_2O).

Potentials in the quarter mole profile were in a similar range as those of other profiles at the end of February and through March (Figure 6.9 c). The quarter mole location had the lowest potentials for 0.1 m depth across the transect. Here response to rainfall was quickest perhaps because the high soil moisture content resulted in an increased hydraulic conductivity. However, based on the evidence of potentials, drainage at this depth was more gradual than elsewhere in the profile. At 0.2 m depth, potentials fluctuated around 0 cm H_2O, showing little response to rainfall and only a small and gradual increase overall. This behaviour at 0.1 and 0.2 m might be explained by the hydraulic discontinuity noted around 0.2 m (Section 4.2.5). Matric potentials at 0.4 m were high and response was smoothed. A slow increase in values until midday on 15.3.91 was followed by a relatively gradual decline and then recovery following rainfall. At 0.6 m depth, there continued to be little variation in potentials, although values did fall from mid March, following rainfall.

A notable characteristic of the drained soil matric potential behaviour was the uniformity of results at field capacity/on the tail recession. This indicates that following drainage of gravitational water, there was relatively little change in potentials. This concurs with the water retention curves (Section 4.2.2).

On the drained plot, wetted areas occurred at 0.2 m depth, so they were coincident with manganese accumulation at the A/Bg boundary (Section 2.5) but were less extensive than on the undrained plot (Section 6.2.3). The tendency towards saturation at the horizon break indicated that there was a marked discontinuity between the moisture flux of the two horizons, the upper horizon having a much higher flux (Section 4.2.5). Persistent and substantial wetted areas throughout the soil profile, on the drained plot, developed later in the drainage season. This was attributed to the greater capacity for water transmission, and therefore, lower antecedent moisture conditions throughout the season (Section 5.4).

In the first half of March, rainfall and runoff were relatively high, with notable peak mole drain discharges on 4 and 8.3.91 (Figure 6.10). Throughout this period matric potential in all profiles were little changed. For example, mole drain discharge increased by four times between the relatively similar profile potentials shown in Figure 6.11. In the latter half of March, less intense rainfall continued and as evapotranspiration increased (*c.* 25 % of field hydrological output), runoff was maintained at > 2 mm d^{-1} until 23.3.91 before declining (1 mm or less). Potentials increased most markedly in the lower soil, so that suction was between 2 and 9 cm H_2O throughout the profile by 30.3.91.

6.2.7 Drained plot: matric potential - summary.

In summary, mole, mid mole and quarter mole matric potential patterns discerned were as follows:

1. in November, the soil above the mole drain was saturated from 0.2 m to depth and initially only minimal drainflow was generated;
2. soil elsewhere, with the exception of 0.2 m at mid mole was unsaturated;
3. at the mole, rapid wetting, was followed by release of water to the mole drain, yet drainflow was sustained even though matric potential increased (i.e. the soil dried). Mid mole and quarter mole profiles exhibited a more standard relationship between wetting soils and increased discharge;
4. towards the end of the drainage season, most of the soil drained rapidly, but mole drain discharge was maintained by a relatively small proportion of the soil volume which remained saturated;

6.2.8 Comparison of undrained and drained conditions.

This section discusses the effect of drainage on matric response to rainfall at the onset of the drainage season. This can be summarised in eight points:

1. on the drained plot matric potentials were higher in October, except at the mole drain;
2. throughout November 1990, the magnitude of matric response was greater on the drained plot. Field drainage and consequent structural amelioration of the soil (Sections 1.2 and 4.2.4) increased hydraulic conductivity allowing water to be conveyed to depth more quickly (Section 4.2.5). Furthermore, following storms, gravitational drainage was more effective and resulted in rising potentials, as a hydrological sink was present within the drained plot;
3. variability on the undrained plot was anisotropic but on the drained plot fissuring resulted in greater connectivity between the surface and subsurface, leading to isotropic heterogeneity;
4. at the end of the drainage season, potentials were higher once more on the drained plot. Most of the soil remained saturated on the undrained plot and yet drainage ceased;
5. during wetting on the undrained plot, runoff was generated before the soil was saturated. This was due to interped spaces being filled by water while intraped spaces remained unsaturated. Shallow interflow plus saturation overland flow (Hewlett and Hibbert, 1967; Section 1.3) were dominant on this plot during rainfall inputs;
6. during drying on the undrained plot, storage pores continued to slowly release matric water, sustaining field runoff;
7. during wetting on the drained plot, the soil was unsaturated with the exception of the soil above the mole drain. However, there, was rapid matric potential response throughout the profiles, thus bypass flow generated field runoff. Connectivity of wetted macropores and fissures was a precursor to field discharge generation;
8. during drying on the drained plot, gravitational drainage of matric water sustained runoff.

6.3 Simulation of matric potential and soil water flux using LEACHW.

6.3.1 Introduction.

The Leaching Estimation And CHemical Model is a one dimensional simulation model which describes the water regime and chemical and solute transport in unsaturated and partially saturated soil to a maximum depth of 2 m. The water flow subroutine (LEACHW) is based on the work of Hutson (1983). It includes options to describe non homogeneous profiles. However, the model is not intended to use unequal depth increments or to predict runoff quantity or quality, although subsurface discharge in the form of leachate is predicted.

Gradual development of this model has been undertaken by Wagenet and Hutson (1987 and 1989) and Hutson and Wagenet (1992). The authors recognise that natural soils display a complexity impossible to reflect adequately in a simple one dimensional model, but represent many important processes which influence the fate of chemicals in the root zone in a semi-quantitative manner. Each process included in the model is identifiable by a mathematical equation. The use of such models may improve insight into, and appreciation of, the relative importance of various water pathways in the soil, thereby facilitating evaluation and clarification of field data which is often confusing and variable (Ewen, 1990). The model is employed here as a tool to aid the interpretation of field results. Mathematical modelling is an accepted scientific practice which can comprehensively integrate basic processes and describe the system beyond simple subjective judgement. Objective employment of models may contribute to the development/planning of future research as well as contemporary management.

Information on soil properties and the initial conditions of the soil segment are required, in addition to soil surface boundary conditions (Table 6.8).

The estimation of water retention parameters is achieved by fitting the retention function to a measured or predicted retention curve. Regression models derived by Thomasson and Carter (1989) for British topsoil and subsoils were adopted for this study. The independent variables are particle size distribution, organic carbon and bulk density.

Table 6.8 Input data required for LEACHW.

Soil properties	Soil surface boundary
Moisture content / matric potential	Rainfall amount and rate
Hydrology (retentivity, K, particle size)	Mean temperature and amplitude
Chemistry	Evapotranspiration

Caution should be exercised when using the regression equations to predict K-θ-h. The regression is calculated using regional data sets, therefore it incorporates a large amount of generalisation. Thus, R^2 varies between 0.5 and 0.9 when K-θ-h is derived, representing considerable potential error (Hutson and Wagenet, 1992). In this context, predictions should be examined critically, and knowledge of local soil conditions should override the approximate estimations of the regression prediction. Leaching represents the predicted subsurface output, there is no option to incorporate a sink such as a mole drain. However, the conductivity value employed in the simulations discussed here is that determined in the field for the drained site. This is the only consideration taken of the impact of drainage.

Output is in the form of conductivity and soil moisture content for each layer at potentials ranging from zero to 15000 cm H_2O kPa. In addition, cumulative totals and mass balances of water and solutes are calculated. An example of output is included in Table 6.9. LEACHW can generate large amounts of data and so Table 6.9 only indicates values for ten times per day.

The profile water regime is simulated by using a finite difference form of Richards' Equation to calculate water content, flux and potential. Therefore, to solve the equation, source and sink terms, soil hydrological characteristics and boundary conditions are required (Table 6.8). Richards' Equation (Eqn. 1.3) is a soil water flow equation for transient vertical flow derived from Darcy and the equation of continuity as described in Section 1.5.

Table 6.9 Output from LEACHW.

Leachw simulation - summary of results
26.10.90 - 30.11.90 (Richard's Eqn.)

Date	Cumul Q	Daily Q	Cumul Ppn	Daily Ppn
26 Oct	0	0	11.4	11.4
27	0	0	25.2	13.8
28	0.01	0.01	30.8	5.6
29	0.04	0.03	41.4	10.6
30	0.27	0.23	50.4	9
31	1.37	1.1	59.3	8.9
1 Nov	3.43	2.06	62.1	2.8
2	5.25	1.82	62.5	0.4
3	6.65	1.4	62.5	0
4	7.74	1.09	63	0.5
5	8.64	0.9	63	0
6	9.39	0.75	63	0
7	10.02	0.63	63	0
8	10.56	0.54	65	2
9	11.05	0.49	66.4	1.4
10	11.52	0.47	68.2	1.3
11	12.02	0.5	73.9	5.7
12	12.77	0.75	77.5	3.6
13	14	1.23	82.2	4.7
14	15.94	1.94	82.4	0.2
15	17.72	1.78	83.5	1.1
15	19.2	1.48	85.4	1.9
17	20.53	1.33	86.4	1
18	21.8	1.27	97	10.6
19	23.04	1.24	108.1	11.1
20	25.35	2.31	108.6	0.5
21	29.78	4.43	108.6	0
22	32.85	3.07	132.6	24
23	35.04	2.19	132.6	0
24	40.01	4.97	145	12.4
25	45.96	5.95	148.1	3.1
26	50.23	4.27	148.1	0
27	52.86	2.53	148.1	0

Decimal day	Potential (cm H2O) 0.1 m	0.2 m	0.4 m	0.6	leach-ate
0.00	306.60	306.60	306.60	306.60	0.00
0.10	307.60	306.27	306.47	306.60	0.00
0.20	308.47	306.06	306.33	306.60	0.00
0.30	309.24	305.94	306.19	306.60	0.00
0.40	319.61	306.24	306.04	306.60	0.00
0.50	345.99	308.24	305.89	306.60	0.00
0.60	52.45	80.06	305.37	306.60	0.00
0.70	74.45	86.96	300.80	306.60	0.00
0.80	86.05	92.80	290.41	306.59	0.00
0.90	92.76	96.61	277.00	306.59	0.00
1.00	98.02	99.55	263.10	306.59	0.00
5.00	38.84	29.45	16.08	277.19	0.33
5.10	39.81	30.34	16.27	270.81	0.40
5.20	40.64	31.11	16.52	263.52	0.49
5.30	41.36	31.79	16.79	255.55	0.59
5.40	42.79	32.61	17.06	247.12	0.70
5.50	44.95	33.88	17.38	238.47	0.81
5.60	27.95	20.46	15.87	229.68	0.94
5.70	33.52	24.08	13.19	220.96	1.06

Darcy and Richards provide popular approaches to the study of soil water movement. However, they have their drawbacks as outlined in Section 1.5. In addition, data, especially hydraulic conductivity, are not always obtainable and numerical solution procedures are time consuming (Hutson and Wagenet, 1992). Simulations were conducted using one- and two-domain approaches. The first algorithm applied Richards' Equation (Section 1.6; Eqn. 1.3), predicting water flux within a horizontally homogeneous soil. It assumes no strong structure, preferential flowpaths (cracks) or fingering, so that all water flux occurs within the matrix. The second method was based on the model developed by Addiscott (1977) and Addiscott *et al.* (1986). Two processes are incorporated: matric flow and bypass flow. Both types of simulation were conducted to aid in the interpretation of the field data (Section 6.4.1).

6.3.2 Richards' model - prediction of soil hydrology (November 1990).

The following section presents simulation results for November and March (in seasonal terms wetting and recession periods). Predictions for each month are made using Richards' Equation and then Addiscott's algorithm. Each section discusses matric potential and then considers leaching. Input parameters for K_{sat} and bulk density are those for the drained field. The shape of the soil moisture characteristic curve was analysed using K_{find} (Dowd, pers. comm.) to predict the Millington Querk a and b coefficients. By using these coefficients the relationship between soil moisture content and K_{sat} can be derived.

The matric potential profiles for the simulation are depicted in Figure 6.11. A diurnal swing in potential values was evident. This was caused by temperature amplitude generating diurnal evapotranspiration variation. Potentials were high at the start of the simulation, exceeding 300 cm H_2O in all segments. The highest potential occurred at 0.1 m depth (346 cm H_2O) and this increased at first due to evapotranspiration. Inputs of 11.4 mm of precipitation on 26.10.90 and 13.8 mm on 27.10.90 led to a sharp decline in potentials between the soil surface and 0.4 m. The falls in potential were steepest at 0.1 and 0.2 m depth. At 0.4 m, the descent was slightly shallower, and this depth still had the lowest potential (12.8 cm H_2O) on 1.11.90.

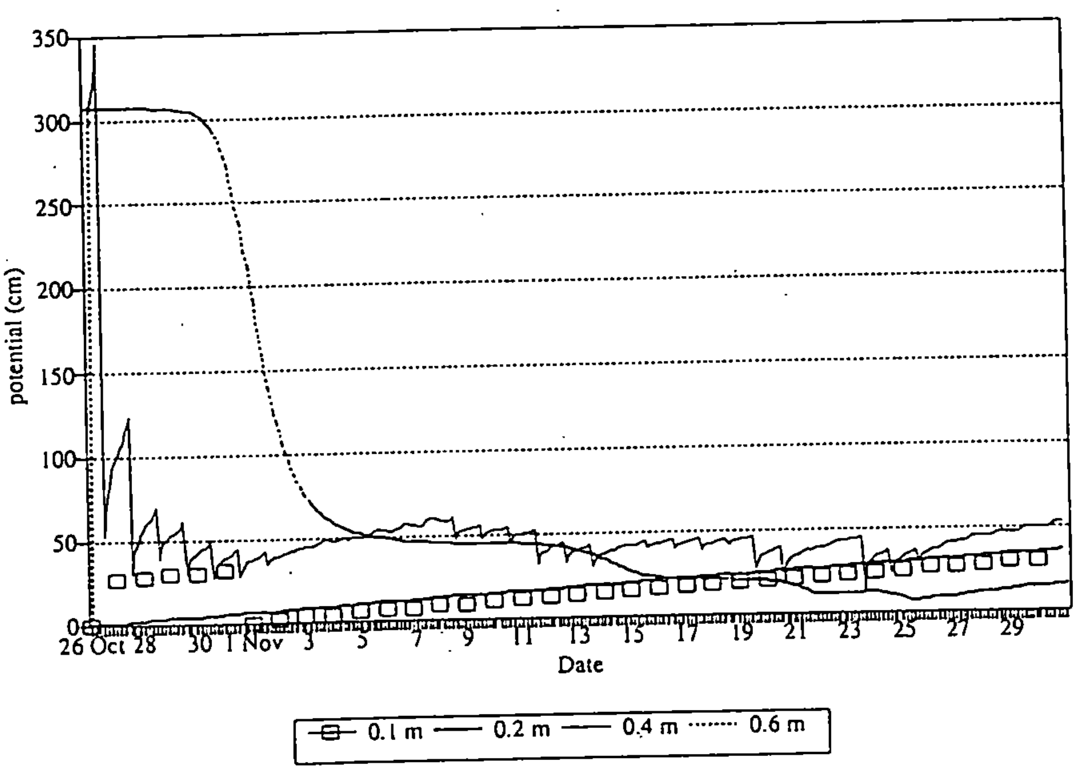


Figure 6.11 LEACHW Richards' simulation of matric potential (November)

Water from distinct rainfall events dominated in different soil segments. This was illustrated on 14.11.90. The upper soil segment experienced recharge, while water from previous events was still draining from lower segments. At 0.1 and 0.2 m, a diurnal swing was evident as the simulated potentials progressed to minimums of 11.5 and 10.2 cm H_2O, respectively, on 23.11.90 which coincides with a decrease in the recharge rate (Figure 6.11). The minimum values for each segment occur at different times during the simulation (Table 6.10 a).

Table 6.10 a Minimum potentials predicted by Richards' Equation (26.10.90 to 30.11.90). Potential in cm H_2O.

Depth (m)	date	time	potential
0.1	23.11.90	2136 h	11.49
0.2	23.11.90	2136 h	10.24
0.4	24.11.90	1648 h	3.09
0.6	25.11.90	0448 h	6.67

Table 6.10 b Relationship between recharge and drainage in Richards' simulation (26.10.90 to 30.11.90).

Size of event	Date	Input (mm)	Actual Q (mm)	Richards Q (mm)	Richards runoff coefficient
Medium	01.11.90	2.8	0	2.06	73.6 %
	13.11.90	4.7	0	1.23	26.2 %
	16.11.90	1.9	0	1.48	77.9 %
High	24.11.90	12.4	8.4	4.97	40.1 %
Total	26.10 - 30.11.90	148.1	46.1	56.72	38.3 %

Table 6.10 a illustrates that within the simulation, the matrix at 0.4 m experienced the lowest predicted potentials. Those at 0.1 and 0.2 m depth were three times and two times higher, respectively. At 0.6 m, the lowest potential was an order of magnitude greater than that at 0.4 m. In addition, LEACHW predicted a lag, controlled by hydraulic conductivity, which therefore reflected depth from the surface. Consequently, the lowest potential at 0.6 m occurred 33 hours and 36 minutes later than those at 0.1 and 0.2 m depth.

While the range of values converged in the upper three soil segments, potential at 0.6 m remained significantly higher until 5.11.90. From 1.11.90, rises in matric potential at 0.1 and 0.2 m were synchronous with relatively constant potentials at 0.4 and 0.6 m (i.e. flat curve) and steps on the cumulative rainfall curve. After 1.11.90, the potential at 0.4 m remained fairly constant (5 to 30 cm H_2O). However, at 0.6 m, potential continued to decrease until 25.11.90 as drainage continued through the soil matrix in response to matric potential gradients. By 25.11.90, potential at all depths began to behave more homogeneously, though a lag between each soil segment was still evident. This indicated that water movement occurred throughout the profile, i.e. there was flux between all segments. At the end of the simulation (30.11.90) matric potential from the upper to lower segments was predicted to be 49.9, 39.9, 20.4 and 14.1 cm H_2O.

The simulated leaching using Richards' Equation was characterised by a smoothing of the recharge rate (Figure 6.12). The magnitude of response to precipitation events was little changed throughout the period. Recharge, discharge and runoff coefficients are detailed in Table 6.10 b.

Daily runoff bore little resemblance to daily precipitation (Figure 6.4), as illustrated in Table 6.10 b. Despite only 0.9 mm of rainfall, leaching was maintained between 2.11.90 and 7.11.90, when output of 6.59 mm was generated. This compares with no observed discharge on the undrained field and 0.251 mm on the drained field. Drainflow throughout this period experienced a steady decline, ceasing on 6.11.90. The predicted and actual outputs are compared in Figure 6.13. Differences are explained by the assumption in LEACHW that all infiltrating water was conducted through the soil matrix. In reality, bypass flow occurred on the drained field so that water content of the matrix was lower than that predicted by LEACHW, and insufficient to sustain drainflow.

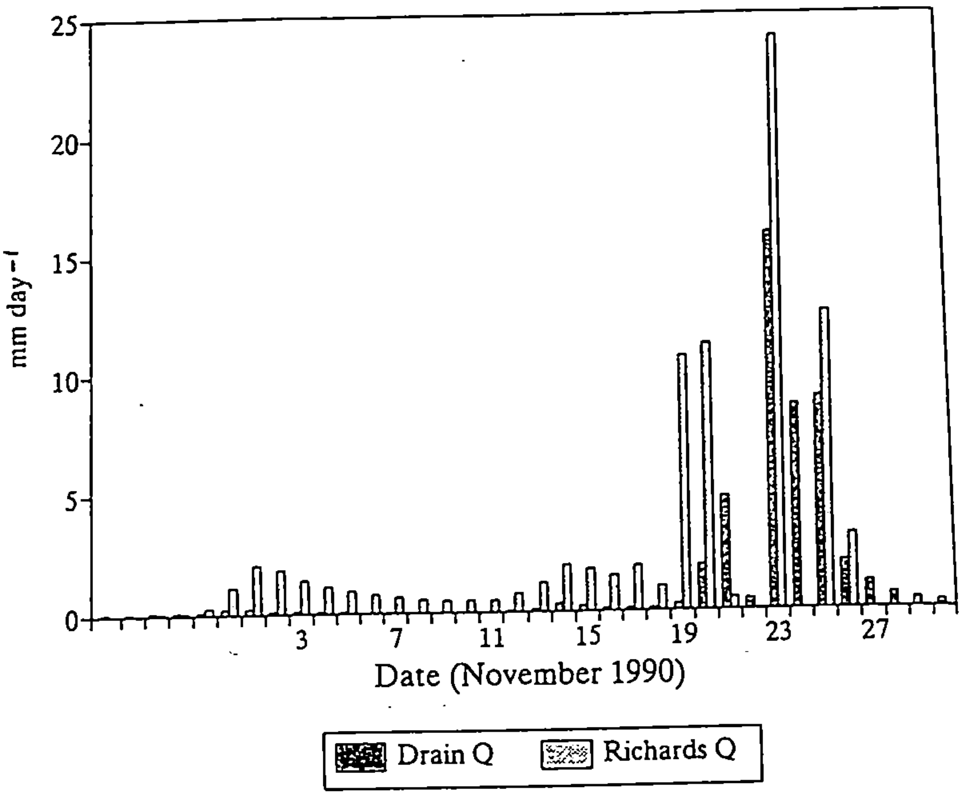


Figure 6.12 LEACHW Richards' simulation of leachate (November).

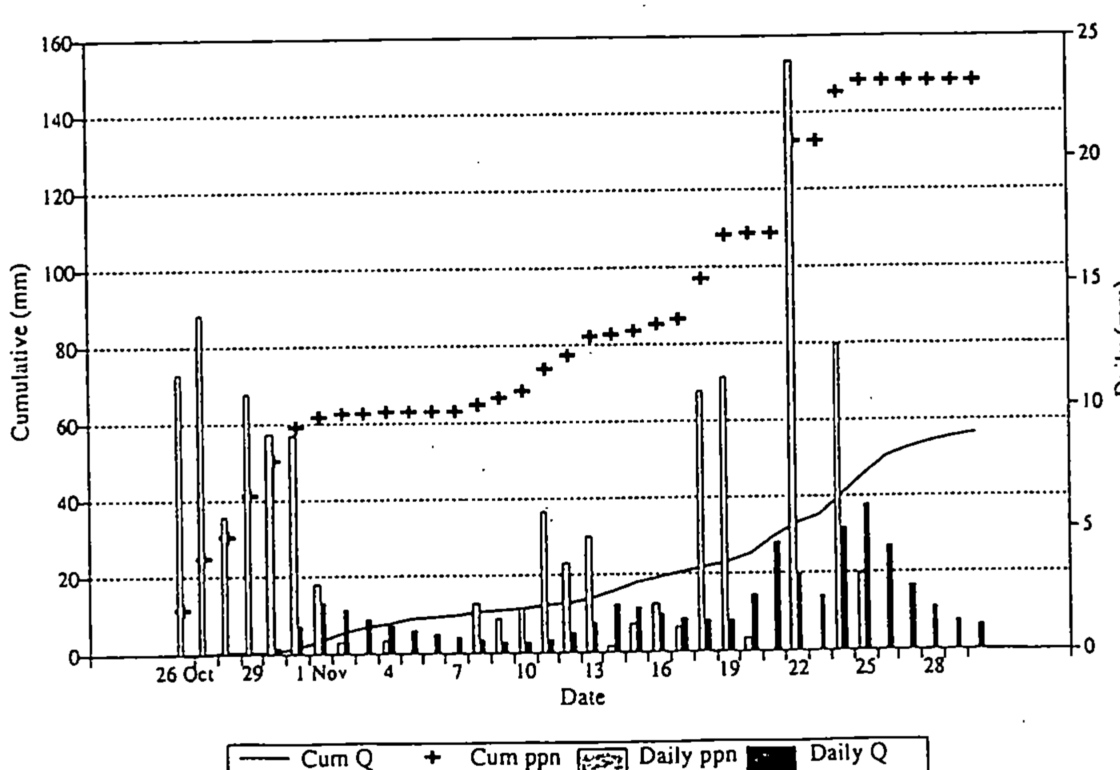


Figure 6.13 Actual and Richards' simulation of discharge (November).

Therefore, the cumulative discharge record exhibited a gradual rise. This was due to the assumption that water was conducted through the soil matrix. A lag time in matric potential was imposed as movement of water was slowed according to the K_{sat} of the soil segment. The predicted hydrograph based on daily discharge was characterised by a smoothed shape, with well-defined recession curves. This is also shown by the maximum predicted daily leaching value which was 5.95 mm (25.11.90) compared with drainflow volumes of 15.77 and 13.1 mm for the drained and undrained fields, respectively (23.11.90). The highest predicted peakflow within a 2 hour 24 minute time period occurred on 25.11.90 with a discharge of 0.73 mm (0.31 mm hr^{-1}) while the drained field peak discharge occurred at 1800 h on 23.11.90. On this day weirflow had increased from 0.032 mm hr^{-1} (0030 h) to a maximum of 2.518 mm hr^{-1}. The simulated cumulative drainage curve was relatively shallow in the first half of the period, as soil storage capacity was satisfied. However, after 16.11.90 the curve was much steeper, due to higher antecedent moisture conditions and four days with more than 10 mm of precipitation which occurred in close proximity. The final cumulative predicted discharge of 56.72 mm (38.3 % of rainfall, Table 6.5) was higher than the observed weirflow.

6.3.3 Addiscott's model: prediction of soil hydrology (November 1990).

The two-domain model showed that the upper soil was near saturation at the start of the simulation (Figure 6.14). All other soil segments had relatively high soil matric potentials, ranging between 220 and 310 cm H_2O. The models' response to precipitation events of 11.4 and 13.8 mm on 26.10.90 and 27.10.90 (Figure 6.1) was rapid falls in potentials at lower depths. Values at 0.2 and 0.4 m converged with those in the top segment on 27.10.90 and 30.10.90, respectively. The recessions at 0.4 and 0.6 m exhibited a tailing off which can be explained by the start of leaching output on 1.11.90, which provided a destination other than soil moisture storage for recharge water. A dry period then led to increased matric potential in all segments but 0.4 m.

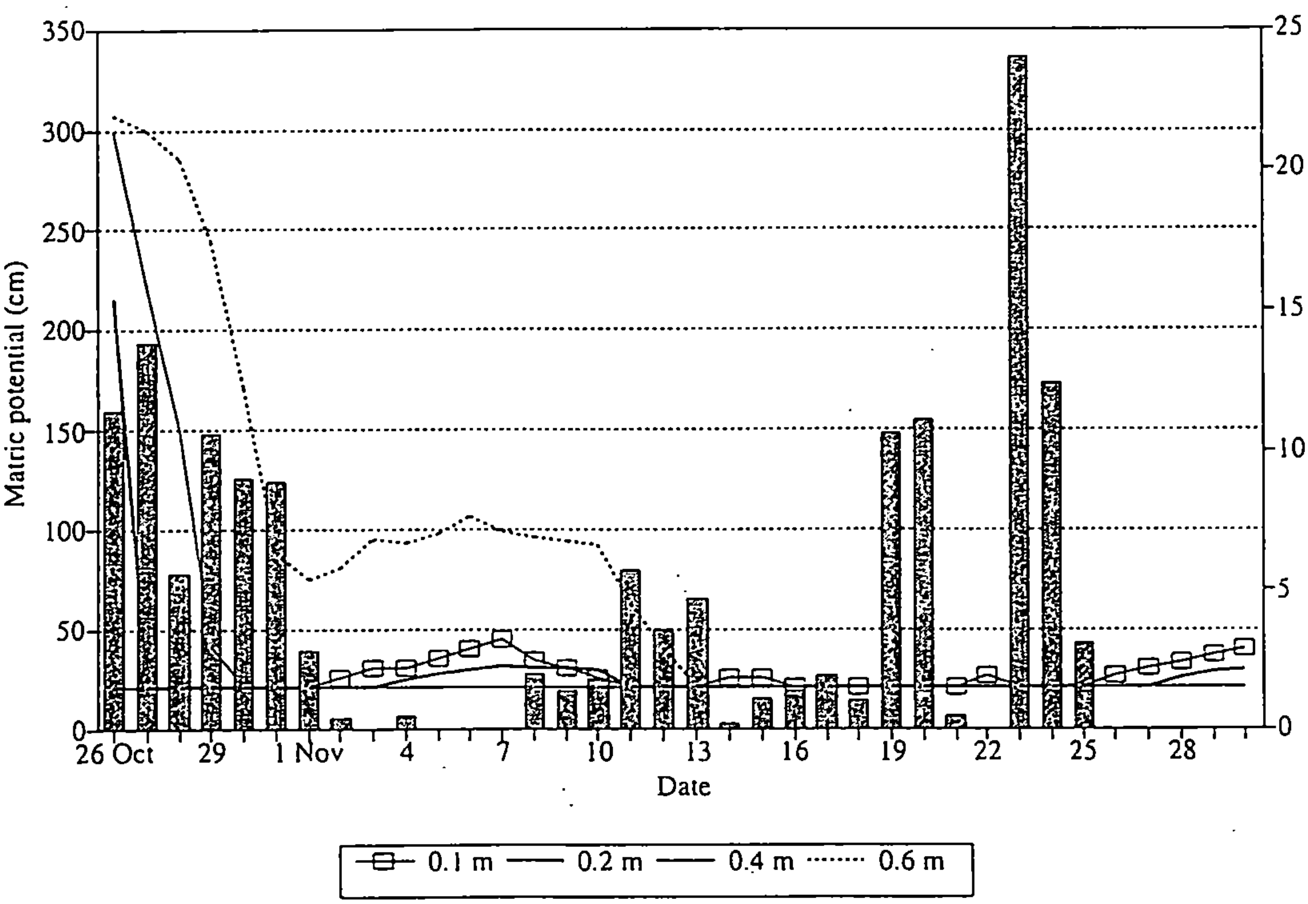


Figure 6.14 LEACHW Addiscott's simulation of matric potentials (November).

More precipitation meant that matric potentials became more uniform throughout the profile. By 13.11.90, the soil moisture deficit to 0.6 m was also satisfied, so that matric potential was equal at all simulated depths. Further periods without precipitation resulted in greatest and most rapid drying response at 0.1 m and uniform potentials elsewhere until 27.11.90 when values at 0.2 m rose once more.

Between the matric potential recession which ended on 1.11.90 and the following group of consecutive days of recharge starting on 8.11.90, the order of matric potential values through the profile altered. Matric potential was initially greatest at 0.6 m with values decreasing from 0.1 m through to 0.2 and 0.4 m. By 13.11.90, potential at 0.1 m was equal to, or greater than, that at all other depths, reflecting the greater effect of evapotranspiration loss, and drainage of water from this segment.

The clustering of six days of relatively high precipitation at the start of the simulation was not reflected in the leaching output from the soil. This recharge was largely employed in fulfilling the soil moisture deficit throughout the profile as illustrated in Figure 6.15. On the drained field, response, in terms of soil moisture content and weir discharge, to recharge is illustrated by Figure 6.16. Precipitation initially satisfied some of the soil moisture deficit. However, after 26.10.90 wetting of the profile was more gradual and subsequent recharge topped up the existing soil moisture until drain discharge began. Neither of the other two weirs operated. By 13.11.90, the field was at field capacity (or recharge rate > permeability) so discharge more closely mirrored the pattern and magnitude of rainfall. In this way, the Addiscott two-domain simulation more closely reflected the situation in the field (Table 6.11). The cumulative discharge curve was, therefore, shallow in the first half of the simulation and steeper, but stepped in the second half (Figure 6.15). Regression of rainfall and leaching (Table 6.12), also shows that the Addiscott model gave a more accurate prediction of early season drainage. While matric potential values were still high, discharge was highly dependent on input and the operation of bypass flow so regression values are higher. As soil moisture deficit began to be satisfied, flow was maintained in periods with no rainfall and the two-domain prediction moved away from the observed field situation.

Table 6.11 Runoff coefficients for fields and simulations (26.10.90-30.11.90).

Dates	Rain mm	Field discharge: Undrained*1 mm %	Field discharge: Drained*1*2 mm %	Simulated discharge: Richards' Eqn. mm %	Simulated discharge: Addiscott mm %
26.10 - 12.11.90	77.5	0.04 0.05	0.9 1.2	12.8 16.5	18.4 24.0
13 - 30.11.90	72.0	31.4 43.6	45.6 63.4	44.0 61.0	68.3 94.8
26.10 - 30.11.90	149.5	31.4 43.6	46.6 31.1	56.7 38.3	86.6 58.0

*1 runoff or leaching (mm) and runoff coefficients (%)

*2 there was no recorded non drainflow

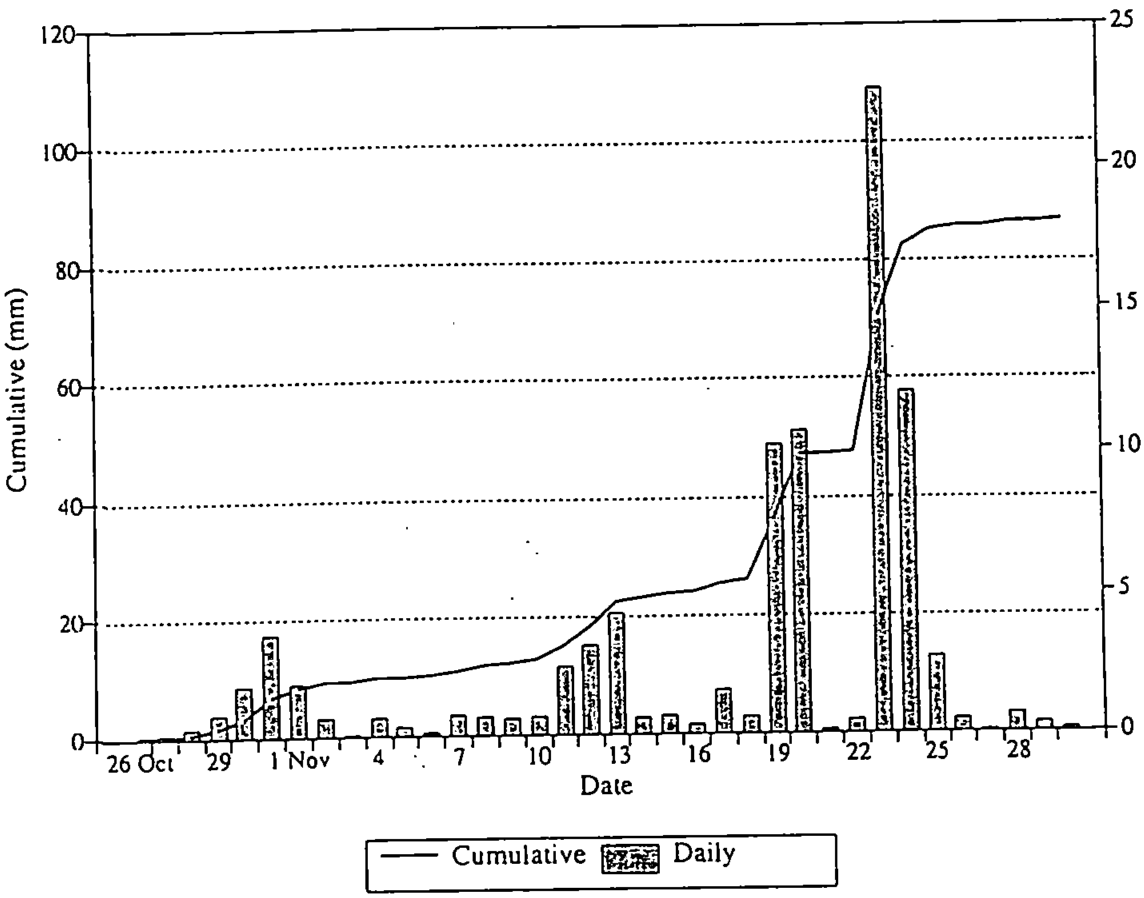


Figure 6.15 LEACHW Addiscott's simulation of discharge (November 1990)

Comparison of Richards' and Addiscott's simulations and field drainage data illustrates that Richards' algorithm diverted too much water to baseflow and imposed too much of a lag throughout this period. Best fits are for cumulative and not daily data, by predicting slow percolation through the matrix, the Richards' Equation gave the best simulation for early November. Regression of cumulative field data gave good results; these were best where drier and wetter periods were assessed separately. Again, Addiscott's algorithm, produced the better simulation but predicted too much runoff. Evaluation of the runoff coefficients for the first and second half of the period further illustrate differences between the two simulations and reality (Table 6.11).

Table 6.12 illustrates the complexity of the fate of rain throughout late October and early November. On the drained field, there was a poor relationship between daily rainfall and runoff totals and application of Addiscott's algorithm gave a similar result. As the soil wetted, the relationship between daily rainfall and runoff became stronger on the drained field (56 %). However, Addiscott related discharge too closely to rainfall distribution.

Table 6.12 Actual and simulated rainfall runoff relationships (November).

Data source	Dates	R^2 daily data	R^2 cumulative data	D.f.
Drained field observation	26.10 - 12.11.90	3.1	84.6	16
	13.11 - 30.11.90	55.9	93.3	16
	26.10 - 30.11.90	30.8	75.9	34
Richards	26.10 - 12.11.90	34.4	66.7	16
	13.11 - 30.11.90	1.8	90.3	16
	26.10 - 30.11.90	0.2	49.9	34
Addiscott	26.10 - 12.11.90	3.0	84.4	16
	13.11 - 30.11.90	99.5	99.9	16
	26.10 - 30.11.90	59.8	94.5	34

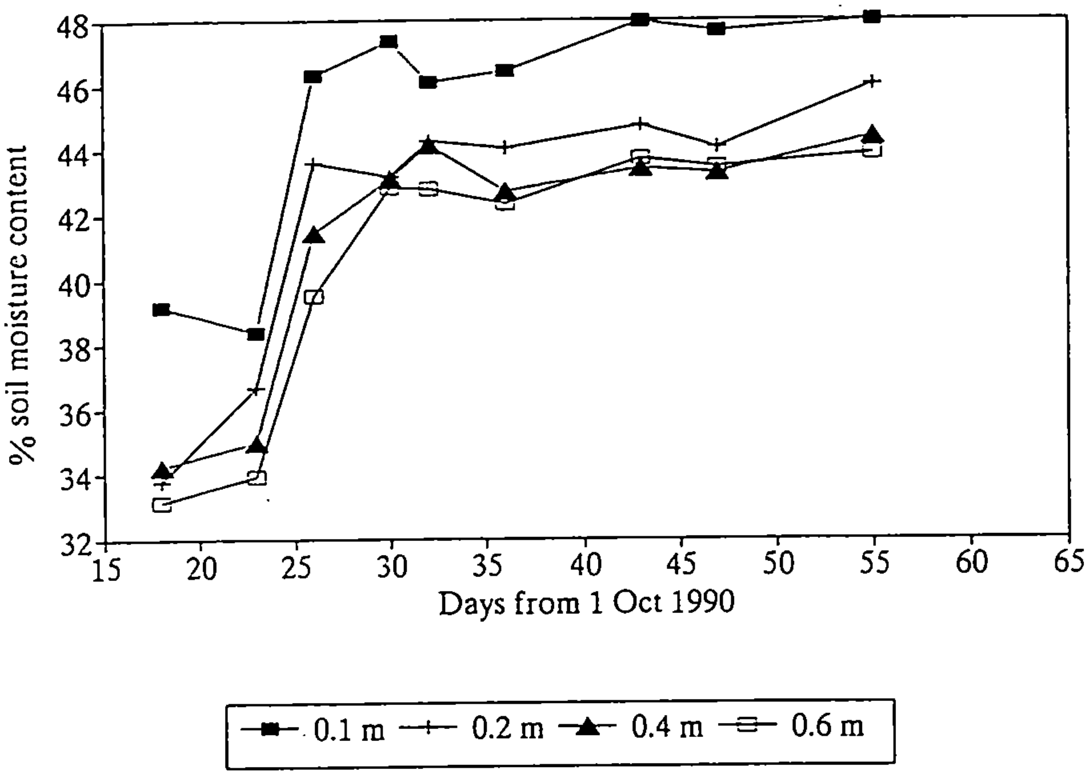


Figure 6.16 a Drained plot: soil moisture content (November 1990).

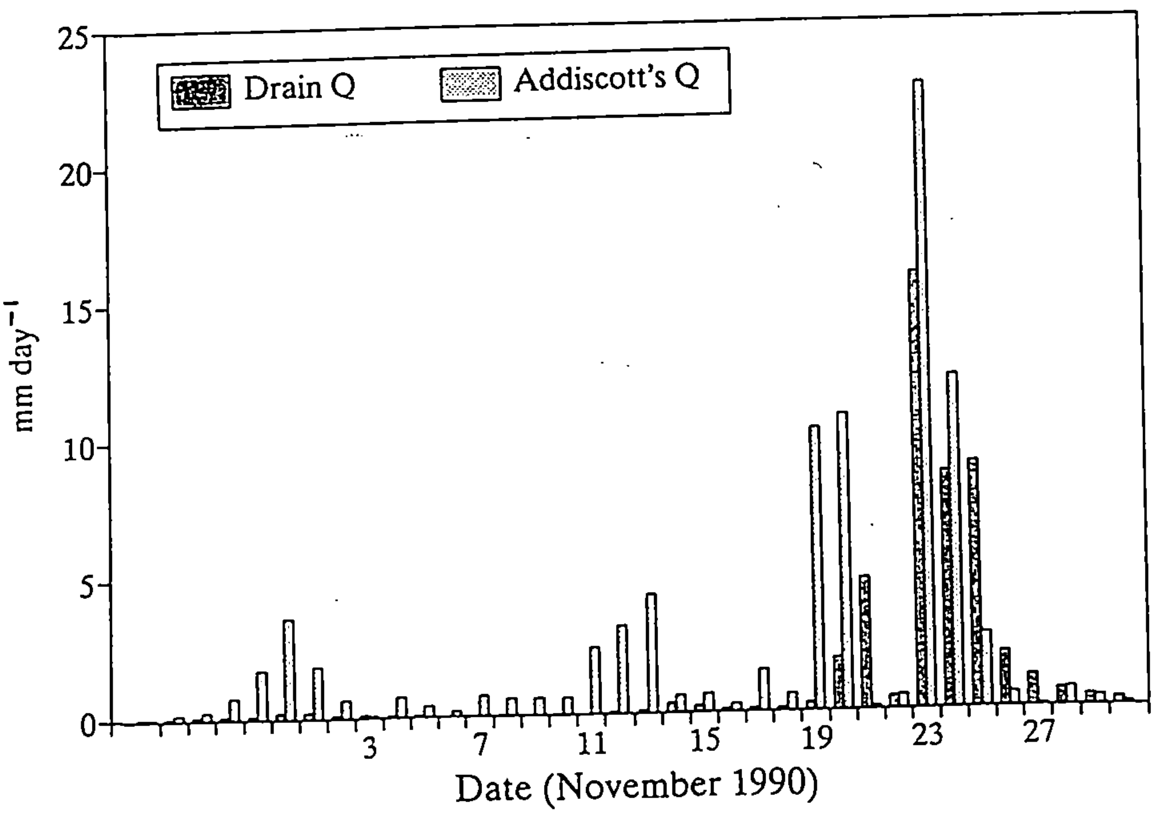


Figure 6.16 b Drained plot: actual and Addiscott's discharge (November 1990).

6.3.4 Richards' and Addiscott's models: prediction of soil hydrology (March 1991).

Much of the discussion of Sections 6.3.2 and 6.3.3 is pertinent to the LEACHW simulations for March. However, this month covers a period of drainflow recessions and soil drying, so its inclusion here expands the discussion which has gone before. Figures 6.17 to 6.19 illustrate the actual and predicted discharge for the drained field for 27.2.91 to 31.3.91. Total discharge of about 80 mm over the period, compares with predicted output of less than 60 mm. A total of 0.424 mm of overland flow occurred on the drained field (0.21 mm on 7.3.91) but the model does not incorporate a surface runoff component, so all water moves to subsurface pathways.

A notable characteristic of drained soil matric potential behaviour was the uniformity of results at field capacity and on tail recessions (Figure 6.20). This indicates that following drainage of gravitational water there was relatively little change in potentials. It concurs with water retention curves (Section 4.2.1) derived using sand table and pressure plate techniques (Section 3.3.2) but differs from predicted behaviour using one- and two-domain simulations (Figures 6.21 and 6.22). Using LEACHW, matric potentials on tail recessions were seen to continue to rise between precipitation events. The difference between actual and predicted potentials can be explained by the inability of the model to simulate the two stage conductivity of the soil as it represents the porosity as evenly distributed pore sizes.

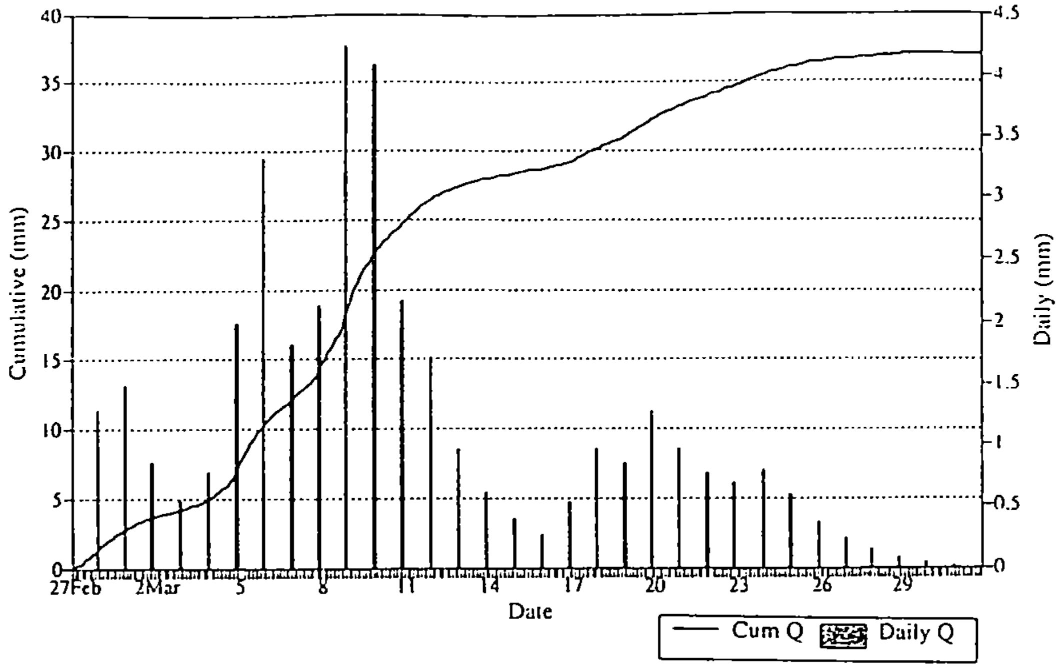


Figure 6.17 LEACHW Richards' simulation of discharge (March 1991).

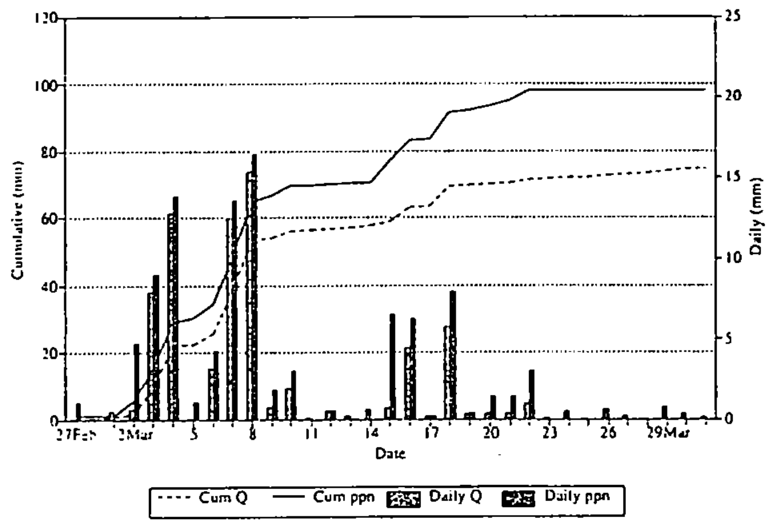


Figure 6.18 LEACHW Addiscott's simulation of discharge (March 1991).

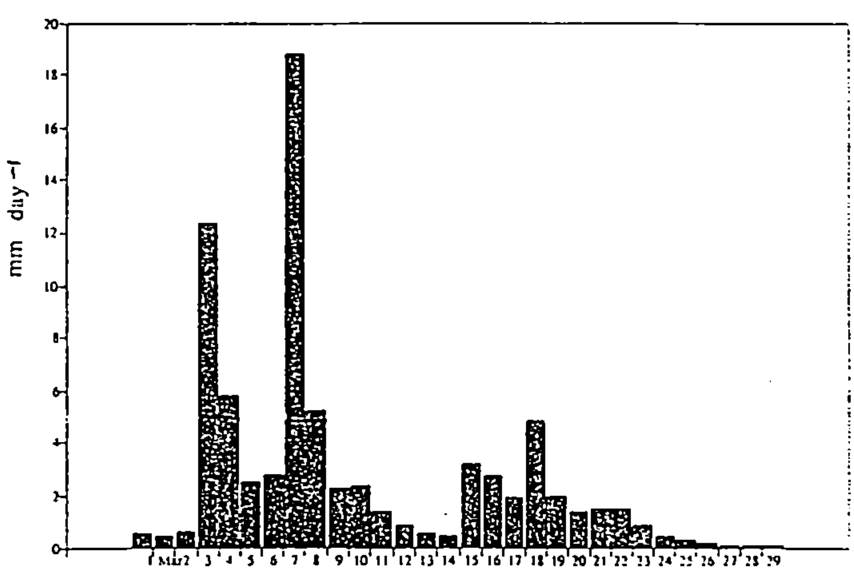


Figure 6.19 Drained field: drainflow (March 1991).

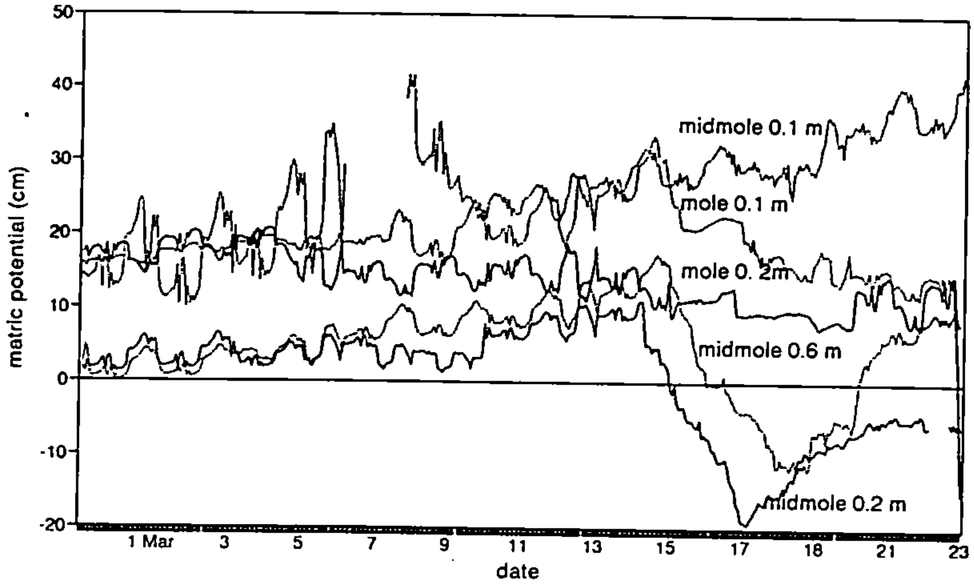


Figure 6.20 Drained plot: matric potential (March 1991).

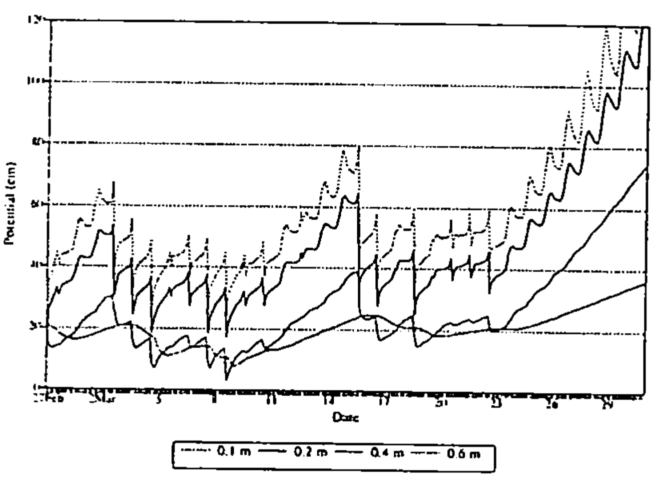


Figure 6.21 LEACHW Richards' simulation of matric potential (March 1991).

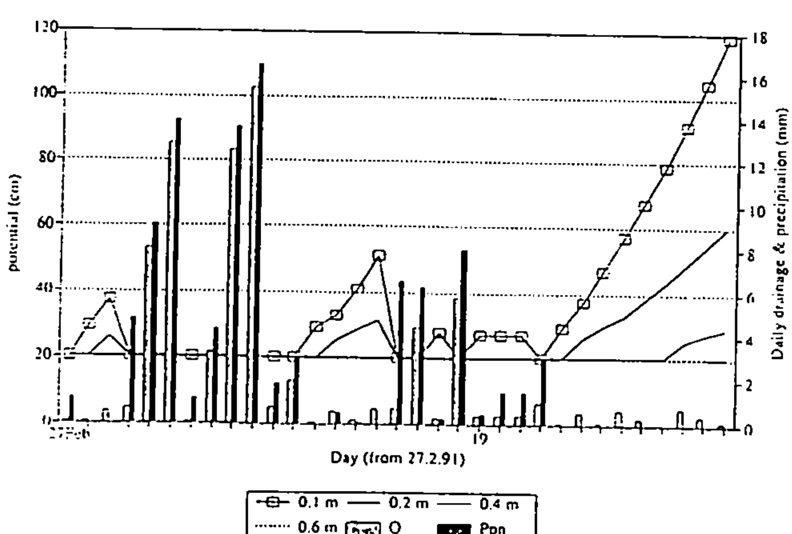


Figure 6.22 LEACHW Addiscott's simulation of matric potential (March 1991).

Observed field discharge was more peaky than that predicted by the Richards' Equation simulation. For example, maximum mole drain weirflow on 9.3.91 was 19.0 mm compared with a predicted output of 6.35 mm. In addition, in the field, daily discharge exceeded 10 and 5 mm on two dates and five dates, respectively. LEACHW, however, predicted zero days with more than 10 mm discharge from the profile and only two days with more than 5 mm. The peaky nature had obvious implications for the cumulative discharge curve. In the field, there were two notable increases in cumulative discharge (Figure 6.23). Consecutive days of relatively high rainfall (23 mm and 30 mm) gave daily runoff of 12.38 mm (4.3.91) and increasing to 18.36 mm (7.3.91) from the drained field.

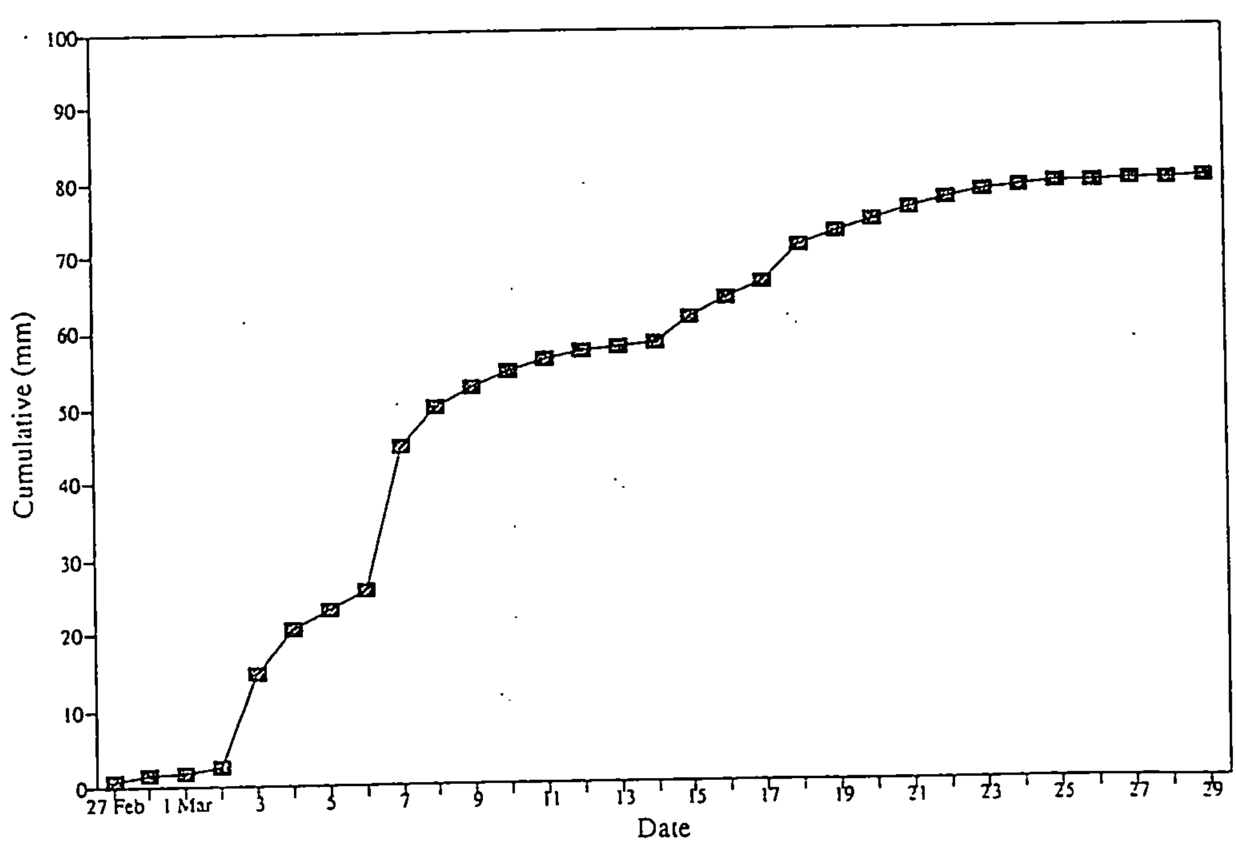


Figure 6.23 Drained field: cumulative drainflow (March 1991).

Table 6.13 Runoff coefficients for fields and simulations

(27.2.91 to 31.3.91)

		Field discharge						Simulated discharge			
Dates	Rain	Undrained		Mole drain		Drained non drainflow		Richards'		Addiscott	
	(mm)	mm	%	mm	%	mm	%	mm	%	mm	%
27.2 - 11.3.91	69.6	44.2	63.5	55.5	79.7	0.4	0.6	37.0	53.1	56.3	80.6
27.2 - 22.3.91	98.1	54.5	55.6	76.5	78.0	0.4	0.4	51.9	52.9	71.4	72.8
27.2 - 31.3.91	98.1	54.5	55.6	80.4	82.0	0.4	0.4	56.9	58.0	74.5	75.9
peak day^{-1}	16.5	18.4		18.8		0.2		6.4		15.4	

Table 6.14 Summary of undrained field runoff data.

Dates	Total (mm)	Maximum (mm)	Mean (mm)	Runoff coefficient (%)
27.2 - 11.3.91	44.2	18.4	3.4	63.5
27.2 - 22.3.91	54.5	18.4	2.3	55.6
27.2 - 31.3.91	54.5	18.4	1.8	55.6

Evaluation of LEACHW.

Application of the Richards' and Addiscott's approaches has shown that neither provides a reasonable approximation of soil hydrology at Rowden. There may be a need to define soil hydraulic characteristics more precisely (true K_{sat} for a larger R.E.V. to characterise the combined conductivity of the matrix and macropores) and recognise the spatially variable relationship between volumetric soil moisture content and potential, i.e. retentivity. A fundamental failing is that water pathways cannot be seen as separate entities, there is a continuum of pore sizes and associated transport mechanisms, thus runoff generation simply cannot be considered as occurring in one- or two-domains.

6.4 Volumetric soil moisture content.

6.4.1 Introduction.

Volumetric soil moisture content (θ_v) was determined on thirty dates between 14.10.90 and 3.4.91 on both plots, using T.D.R. (Section 3.3.8). Soil moisture content was calculated on both plots for four soil segments: from the soil surface to 0.1, 0.2, 0.4 and 0.6 m depth. In addition, there were sufficient sets of probes for the determination of moisture content at 0.3 m on the undrained plot. The drainage season results for the undrained and drained fields are presented in Sections 5.4.2 and 5.4.3, respectively. Discussion focuses on the same periods as those considered in Section 6.2 and 6.3.

6.4.2 Undrained plot: volumetric soil moisture content (November).

Assessment of whether the mole drain exerts any control over spatial variation of soil moisture content and the significance of soil moisture content for generation of field runoff is reported here. As with matric potential (Section 6.2.2) soil moisture contents were highly variable in November. Soil moisture data for 26.10.90 to 30.11.90 are presented to illustrate the soil status prior to and during drainflow initiation (Figure 6.24).

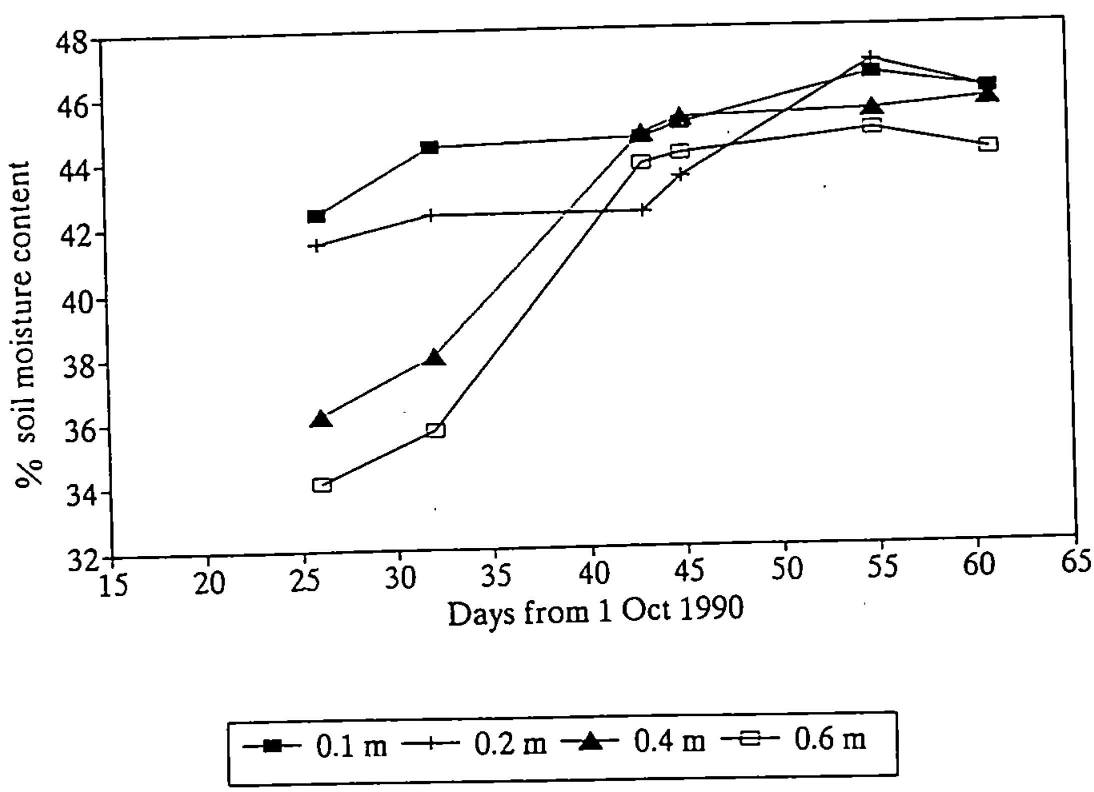


Figure 6.24 Undrained plot: volumetric soil moisture content (%) November 1990.

Table 6.15 Undrained plot: mean volumetric soil moisture content (%) November 1990.

Date	Measure	0 - 0.1 m	0 - 0.2 m	0 - 0.4 m	0 - 0.6 m
26.10.90	mean	42.4	41.5	36.2	34.1
	st dev	3.2	2.6	1.9	2.4
01.11.90	mean	44.4	42.3	38.0	35.7
	st dev	3.0	2.1	2.5	2.9
12.11.90	mean	44.6	42.3	44.7	43.8
	st dev	3.0	1.9	2.3	1.4
14.11.90	mean	45.0	43.4	45.2	44.1
	st dev	3.3	2.1	1.8	0.8
24.11.90	mean	46.5	46.9	45.4	44.8
	st dev	3.2	2.1	2.0	0.9
30.11.90	mean	46.0	46.0	45.7	44.1
	st dev	2.7	1.1	1.6	1.0

where: $n = 8$ per depth per day.

Great fluctuations of potentials through time (Section 6.2.2) were accompanied by small variations of volumetric soil moisture content (Table 6.15). This was especially true for the deeper profiles. In the clayey B horizon, recharge or drainage of small volumes of water resulted in relatively large changes of potential. Wetting of the upper 0.2 m was fairly steady, but Table 6.15 shows that there was relatively little change in the lower soil segments (most of the increase in θ_V recorded over 0 to 0.4 m and 0 to 0.6 m being accounted for by increased storage in the upper 0.2 m). In late October, decreasing variability in the upper two segments was coincident with increasing variability in the 0 to 0.4 and 0 to 0.6 m profiles. Throughout the remainder of the period, variability increased in the upper portions of the soil and decreased in the lower segments (Table 6.15). Field runoff was initiated when profile soil moisture content was between 45 and 47 %.

Volumetric soil moisture content was monitored by probes installed in a grid (Section 3.3.8) to investigate spatial change. Visual inspection of results revealed that there was much variation between adjacent pairs of probes (between horizons, across and up-down slope). There was one persistently dry zone observed by five adjacent pairs of probes, and two spatially continuous wetted zones as monitored by four and five pairs of probes.

6.4.3 Undrained plot: volumetric soil moisture content (March).

Soil moisture content was between 45 and 52 % on 1.3.91. Figure 6.25 illustrates that undrained plot soil moisture content generally decreased with depth, although 0 to 0.2 m was 3 to 4 % wetter than 0 to 0.4 m. This relationship was maintained throughout March. The decrease with depth was due to increasing bulk density and an associated smaller pore size distribution.

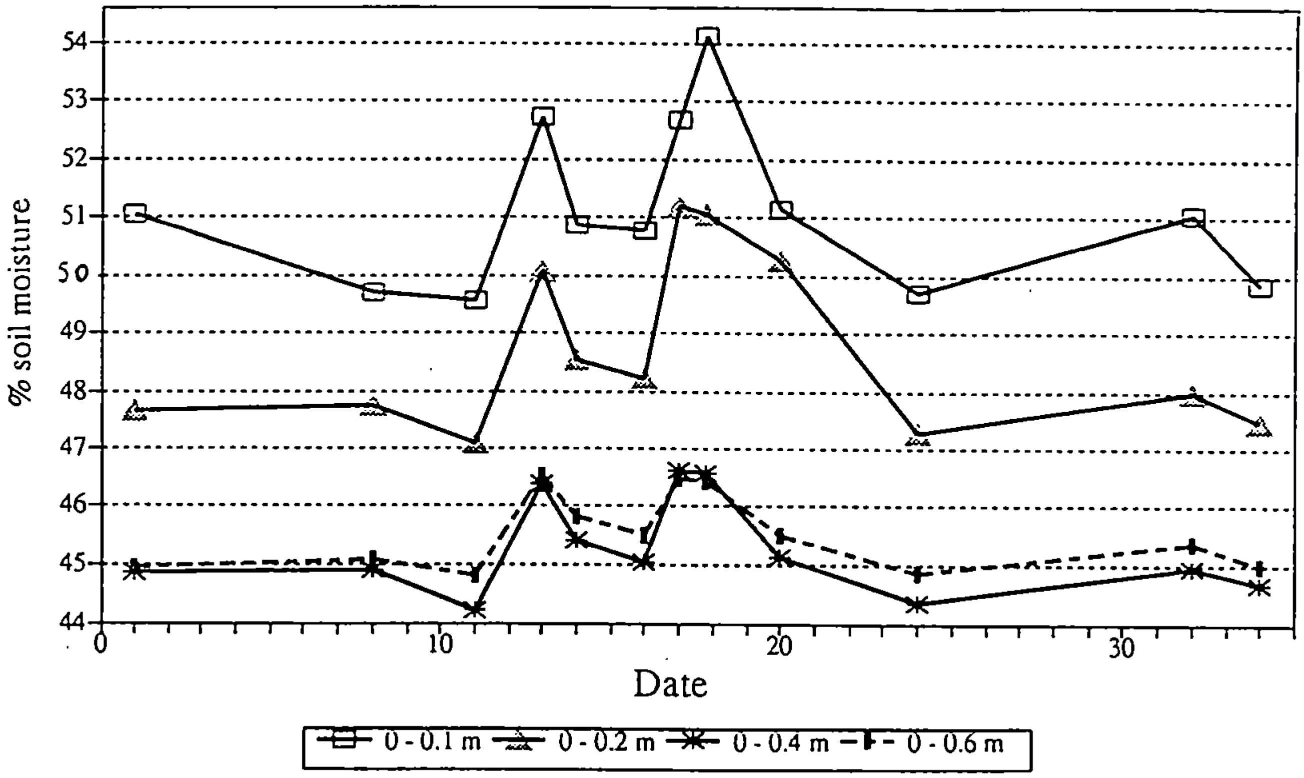


Figure 6.25 Undrained plot: volumetric soil moisture content (March 1991).

Moisture content in the upper 0.1 m of the soil was most variable. Values at 0.2 and 0.4 m depth were least variable. Highest contents were recorded on 8.3.91 after 25.5 mm of rain (59.5, 52.5, 48 and 52 % with decreasing depth). All soil segments then dried until the end of the month, however the variability between profiles remained low (5 %; Figure 6.23) and moisture content on cessation of field runoff was between 45 and 50 %.

6.4.4 Drained plot: volumetric soil moisture content (November).

Table 6.16 shows that in mid-October, volumetric soil moisture content for the 0.6 m profile averaged 33 %. By the beginning of November this had risen to almost 43 % and remained little changed throughout the rest of the month. Soil moisture contents at all depths were between 43 and 50 % by 24.11.90 (Figure 6.16 a).

Averaged water contents on the drained plot were about 5 % above those for the undrained plot in October and early November (Tables 6.15 and 6.16). This indicates that the drained soil had greater storage capacity and its improved structure aided infiltration and percolation. However, variability of soil moisture contents was higher for the drained soil. Here, as on the undrained plot (Section 6.4.2), as the soil wetted a number of temporal trends were discernible. Between the surface and 0.1 and 0.2 m, variability decreased until 1.11.90, whilst in the deeper profiles there was no observable pattern. From 1.11.90, volumetric soil moisture content to 0.1 m became more variable whilst the other three segments exhibited a slight but consistent decrease in volumetric soil moisture content from 5.11.90 (Table 6.16). Soil moisture contents had rapidly increased in the last week of November. However, Figure 6.26 shows that drainflow initiation occurred when mean moisture content for the 0.6 m profile was still below 40 %, but when some parts of the soil were considerably wetter (e.g. 0.1 m averaged 46.1%).

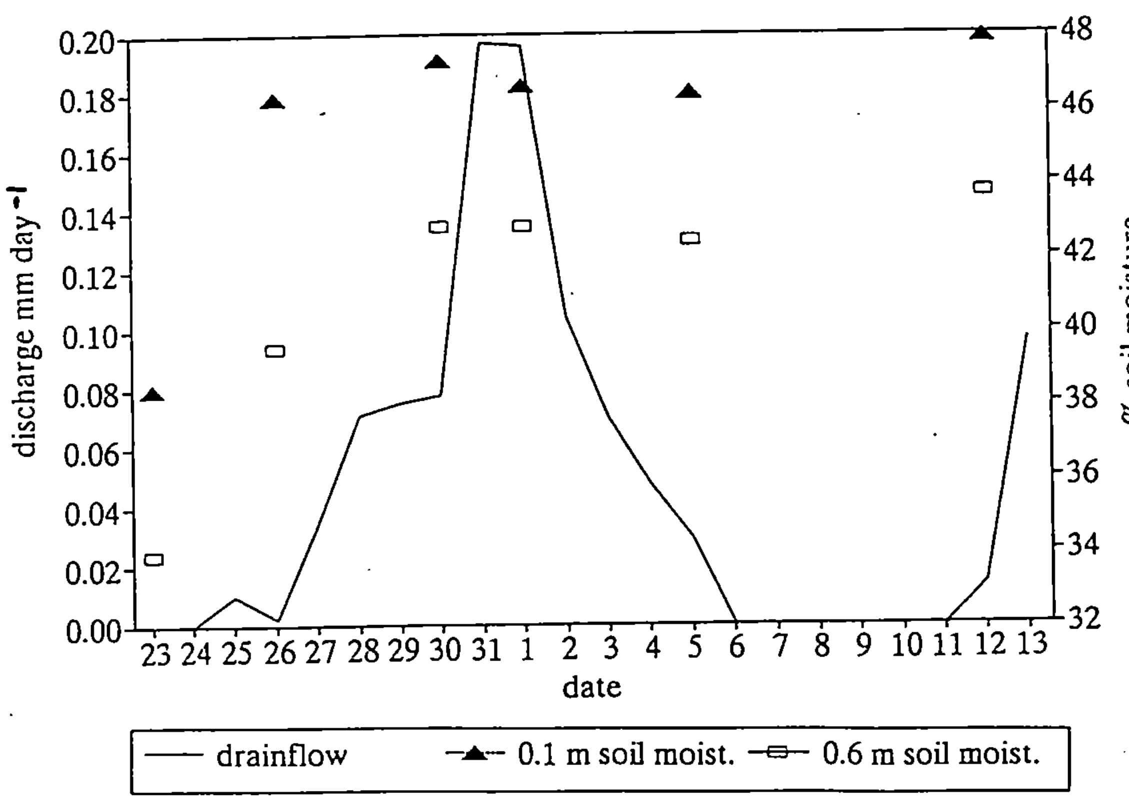


Figure 6.26 Drained plot: mean soil moisture contents during drainflow initiation.

Table 6.16 Drained plot: mean volumetric soil moisture content (%) November 1990.

Date	Measure	0 - 0.1 m	0 - 0.2 m	0 - 0.4 m	0 - 0.6 m
18.10.90	mean	39.2	34.5	34.2	33.2
	st dev	3.5	2.6	2.5	2.7
23.10.90	mean	38.4	36.4	35.3	33.9
	st dev	3.6	2.7	2.2	1.0
*26.10.90	mean	46.1	43.4	41.5	39.5
	st dev	3.4	2.5	1.9	2.6
30.10.90	mean	45.5	43.9	43.1	42.9
	st dev	3.2	2.5	2.6	3.1
*1.11.90	mean	46.1	44.3	44.1	42.8
	st dev	2.9	2.1	2.5	3.0
5.11.90	mean	46.4	44.0	42.9	42.4
	st dev	3.2	2.4	2.7	2.8
*12.11.90	mean	48.0	44.6	43.5	43.8
	st dev	3.0	2.2	2.4	1.7
16.11.90	mean	47.7	44.0	43.3	43.5
	st dev	3.3	1.7	2.2	1.5
*24.11.90	mean	50.0	45.2	43.4	43.4
	st dev	3.5	1.7	2.2	1.3

where: n = 10 per depth per date

* dates for which undrained plot soil moisture contents are listed (Table 6.15)

Table 6.17 illustrates that the mole drain had no discernible impact on soil moisture in October and November. The strongest relationship between distance from drain and volumetric soil moisture content was observed in the 0.4 m profile, in which approximately half of the variation of soil moisture contents in November could be explained by position relative to the drain.

6.4.5 Drained plot: volumetric soil moisture content (March).

The mean soil moisture contents for March were 49.3 % for the upper 0.2 m and about 45.8 % for the 0.6 m profiles (Table 6.18). Spatially, there were two distinct patterns of soil moisture for 2.3.91. First, moisture content decreased with depth, until 0.4 m when results were generally slightly lower than those for 0.6 m. The exception was around the mole where values were closer. This may have been due to the drain creating greater potential and causing drainage from larger pores or to increased porosity around the drain

Table 6.17 Correlation coefficient (r) of volumetric soil moisture content and distance from drain (autumn 1990 and spring 1991).

Date	0 - 0.1 m	0 - 0.2 m	0 - 0.4 m	0 - 0.6 m
18.10.90	0.10	0.23	0.24	0.10
23.10.90	0.30	0.21	0.10	0.18
26.10.90	0.29	0.15	0.21	0.15
30.10.90	0.00	0.48	0.54	0.11
05.11.90	0.00	0.30	0.63	0.82
12.11.90	0.12	0.04	0.51	0.24
16.11.90	0.12	0.05	0.46	0.06
24.11.90	0.12	0.05	0.46	0.06
04.03.91	0.25	0.58	0.66	0.86
05.03.91	0.52	0.70	0.81	0.74
07.03.91	0.41	78.2	38.1	84.7
11.03.91	60.3	0.71	0.59	0.95
15.03.91	0.35	0.52	0.70	0.85
23.03.91	0.81	0.80	0.16	0.73
25.03.91	0.22	0.49	0.30	0.74

where: n = 16, significance level = 95 %.

(Section 4.2.4). Second, soil moisture content increased throughout each profile with distance from the mole drain (Figure 6.27). By 4.3.91, soil moisture contents at all positions had increased (> 2.5 % or 6 mm in the upper 0.4 m; 1.7 % or 6.8 mm over 0.6 m). The smallest variation, over 0.6 m depth was due to the more massive soil characteristics at depth, i.e. higher bulk density, less fissuring, poorer connectivity with the soil surface. The influence of the mole drain on volumetric soil moisture content is investigated in Table 6.17 and was most marked in the three deepest segments, in the first half of March. The relationship was stronger and more prolonged for the full profile (0 to 0.6 m).

Figure 6.27 Drained plot: volumetric soil moisture content (%) with distance from drain (m). March 1991

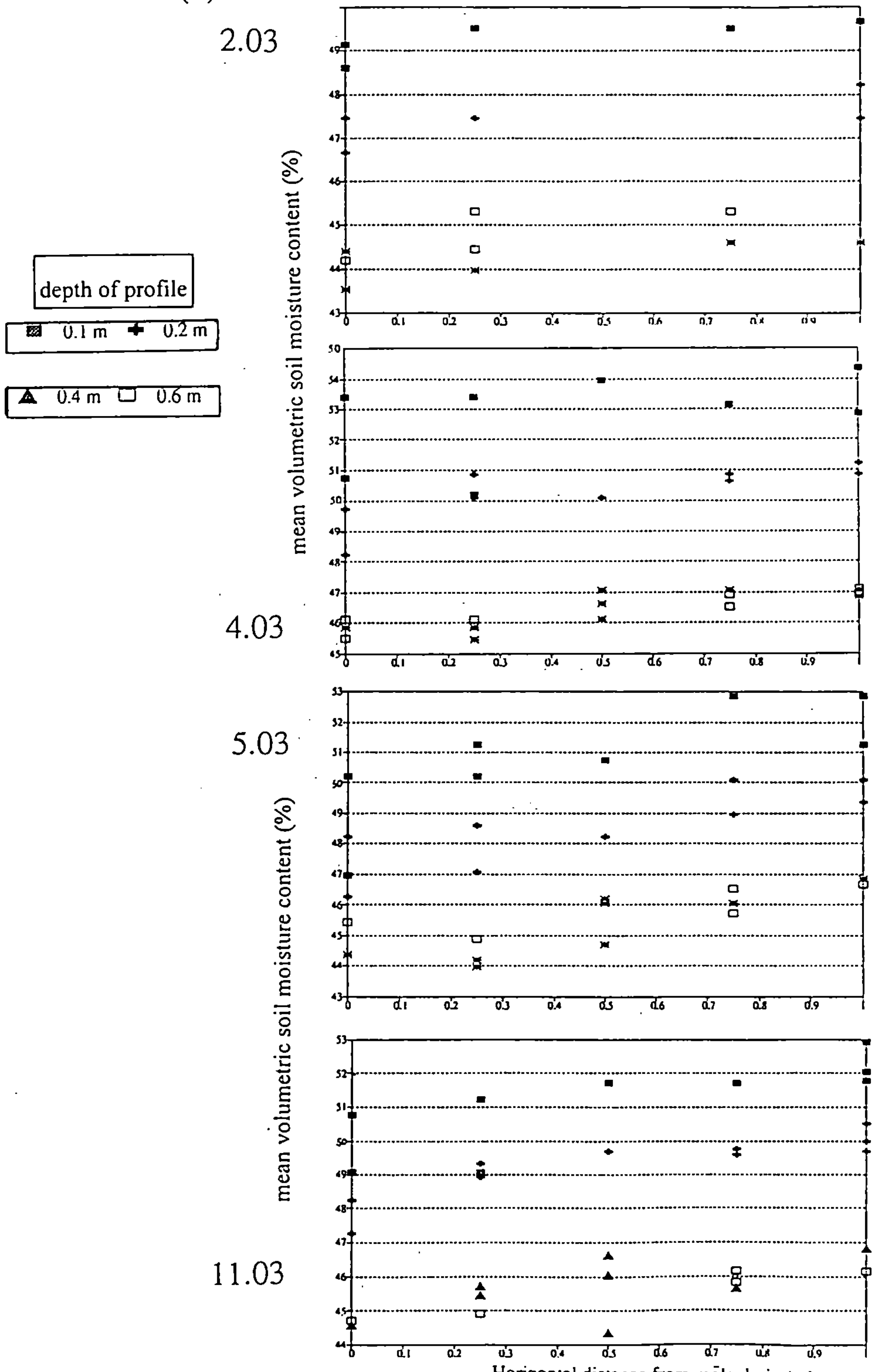


Table 6.18 Drained plot: mean volumetric soil moisture content (%) March.

Date	0-0.1 m	0-0.2 m	0-0.4 m	0-0.6 m
March mean	—	49.3	45.9	45.0
27.2.91	49.5	47.6	44.4	45.1
2.3.91	49.6	47.1	44.2	44.8
4.3.91	52.8	50.1	46.7	46.5
5.3.91	50.8	48.5	45.9	45.8
7.3.91	50.8	48.4	45.3	45.5
8.3.91 (i)	52.7	51.2	47.2	46.5
(ii)	54.1	51.1	46.6	46.4
11.3.91	51.2	50.2	45.4	45.5
15.3.91	49.7	47.3	44.3	44.8
18.3.91	49.9	47.8	44.6	45.2
23.3.91	51.1	48.0	44.9	45.3
25.3.91	49.8	47.3	44.7	45.0

Soil moisture was examined temporally: after precipitation stopped at 1900 h on 4.3.91 increasing moisture content with distance from the drain was heightened (5.3.91) as the soil drained. Volumetric soil moisture contents were determined twice on 8.3.91. The first readings were taken at 1200 h, prior to which there had been 18 mm of rainfall in 10 hours, with almost 6 mm since 0600 h. After 1200 h, intensity decreased to less than 1 mm hr^{-1}. By 1800 h, a slight drop in soil water storage was observed between readings at 0.2 to 0.6 m, while a slight rise was recorded at 0.1 m. Results on 11.3.91 were very similar to those from 5.3.91. They were taken after a similar interval without recharge.

Summary of drained field T.D.R. results.

Mean soil moisture contents for the drained plot were different at the start and end of drainflow (Table 6.19). In October, simultaneous observation of dry soil and drainflow generation may be explained by the occurrence of macropore flow. This is to be expected for such a heavy and impermeable soil (Harrod, 1981; Hallard, 1988) and concurs with matric potential observations (Section 6.3). It is not uniform wetting, a rising water table or a capillary fringe which initiate drain discharge, but exploitation of macropores and development of wetted zones which can convey preferential flow.

Table 6.18 Drained field: mean volumetric soil moisture content (%) at start and end of drainage season.

Depth (m)	26.10.90	02.04.91
0-0.2	43.4	47.6
0-0.4	41.5	44.8
0-0.6	39.5	45.2

Table 6.19 Drained field: range of soil moisture content (%) at start and end of drainage season.

Depth	26.10.90			2.4.91		
(m)	max	min	range	max	min	range
0.2	45.85	38.55	7.30	50.17	45.69	4.48
0.4	43.82	38.59	5.23	45.98	43.30	2.68
0.6	40.85	38.72	2.13	45.85	44.10	1.75

Mean volumetric soil moisture contents were relatively constant throughout the drainage season but variability decreased (Table 6.19). By April, soil moisture contents were more uniform and averaged 45 %. The T.D.R. effectively averages over the monitored profile or segment of soil, so that immobile water is included in any result. This water may form a substantial proportion of soil moisture (Section 4.2.2) and is important in terms of antecedent soil moisture content. This means that relatively small changes in soil moisture content may represent large differences in terms of soil status.

6.5 Discussion.

Three directions of variable soil hydrological response were recognised: through the drainage season; between plots and finally within plots. Results showed that runoff from both fields was initiated, and was then sustained, well before the soil matrixes were entirely wetted. Over the season, variability of soil status on both plots reduced as water slowly diffused through most of the matrix. Thus, when field discharge ceased, profile soil moisture contents and potentials were higher than those recorded during drainflow initiation. A few tensiometers indicated that some parts of the matrixes did not wet at all.

Macropore flow appeared to be important on both plots. Unlike the findings of Johnson *et al.* (1993) for a mole drained arable site, macropore flow at Rowden was not dependent on a significantly wetted matrix. On the contrary, low hydraulic conductivities of the dry matrix were believed to promote fissure flow. As the continuity of macropores is a major control of soil water movement (Section 1.7), the drained plot has more effective macroporosity (moling introduces fissures which are connected to a major sink). However, the consequences of drainage are complex. Whilst increased macroporosity enables subsurface water to bypass the matrix, the same increase in porosity diverts water from surface to subsurface pathways. In effect, macropores 'open up' the soil thereby encouraging infiltration and movement of water to depth. A more detailed characterisation of the soil volumes in which fissure flow occurs and of those which exclude water (thereby encouraging macropore flow) would both be a legitimate focus for future work. Further, it may be possible to identify a range of thresholds (rainfall intensity, antecedent soil moisture content) which determine the generation of macropore flow at different points in the drainage season.

Hydrologically, soil status of the drained plot was more spatially and temporally variable than that of the undrained site. In addition to augmenting macroporisity, by reducing the duration of saturated conditions, drainage allowed increased biological activity and improved topsoil structure (Gilbey, 1986; Blantern, pers comm.). The resultant enhanced porosity was recorded in the laboratory (Section 4.2.4) and at a range of scales in the field (from the K_{sat} cores up to the 1 ha field scale). This concurs with the findings of other drainage impact studies (Hallard, 1988; Armstrong and Garwood 1991). Thus, with an improved infiltration capacity and porosity the drained soil profile responded more quickly to rainfall and soil moisture status was more variable.

In terms of intraplot variability, the undrained soil hydrology was anisotropically heterogeneous. However, the mole channel produced consistent patterns of soil moisture content (increasing with horizontal distance from mole channel) on the drained plot.

While this was apparent in T.D.R. data (which characterised relatively large volumes of soil including macropores and the matrix) spatial trends in tensiometric data were difficult to define. Matric potential data was valuable in that it indicated the range of hydrological behaviour (rapid response ⇒ no response) of soil volumes in close proximity. Relationships between soil moisture content and matric potential observed in homogeneous soils in laboratory conditions are more difficult to discern in the field, not least because tensiometers and T.D.R. monitor different volumes of soil. They are complicated by anisotropic heterogeneity due to inherent soil variability and are also related to independent hysteretic behaviour of soil water. Discrete wet zones (identified by soil moisture contents and potentials) were identified in the matrices of both plots. The nature of these wetted pockets and their role in runoff generation are considered in Chapter 7.

LEACHW was employed to aid in the interpretation of field data (Section 6.3). Structural conceptualisation of the soil usually treats the matrix and macropores as different flow media (Chen and Wagenet, 1992). The matrix is assumed to behave as a homogeneous soil without macropores and water fluxes are described by equations derived from Darcy's Law (Section 1.5). Macropores are regarded as a separate domain. If macropores acted independently of the matrix, the predominant discharge characteristics would be consistently high runoff-rainfall values and low lag times. As neither of these are proved it is assumed that the matrix and macropores do not behave independently; rather they are inextricably linked in terms of water movement. This means that the one- and two-domain models utilised in the LEACHW programme (Section 6.3) are too crude.

Where 'classic' macropore or bypass flow (as represented by Addiscott's simulation; Section 6.3) occurs the implications for solute/pollutant transport are complex and largely depends on the source of the solute under consideration. If bypass flow is the dominant runoff generation mechanism, the opportunity for buffering percolating soil water is reduced, thus acidifying chemical species from atmospheric pollution (Reynolds *et al.* 1986; Mason, 1992) or surface applied fertilisers for example, will be prone to rapid transport to water courses (Barraclough *et al.*, 1983; Hallard 1988). Conversely, compounds held in the soil matrix (e.g. nitrates or pesticides; Johnson *et al.*, 1993) will not be available for leaching. To investigate the validity of these assumptions (i.e. is macropore flow at Rowden 'unbuffered'?) an isotopic tracer experiment was conducted in which macropore flow was induced (Chapter 7).

Thus, plot scale observation characterised soil hydrology which was highly spatially and temporally. Drainage imposed some spatial order on the otherwise unpredictable behaviour of the soil, but increased temporal varaibality of soil status, by altering soil structure and water pathways. Within the nested experimental design, this intermediate level of study raised more questions than it answered, however it informed the design of the lysimeter experiment described in Chapter 7.

CHAPTER 7: TRACER EXPERIMENT.

7.1 Introduction.

The relative importance of macropore flow in runoff generation was demonstrated in Chapter 6. In this chapter, these rapid flowpaths are studied in more detail using an intensively monitored lysimeter. The aim of the experiment was to trace the movement of labelled water through and out of a soil block in which a mole drain and associated fissure system had been installed. Macropore flow was induced by irrigation. This represents an extension of previous work reported in this thesis. Rather than inferring the nature of water movement by observation of soil status, the fate of water labelled with a conservative tracer could be observed. The tracer experiment allowed the potential pathways suggested in Section 6.5 to be examined. More particularly, the relative importance of advection and dispersion can be considered, contributing to theoretical debates and methodological concerns. For example, Wenner *et al.* (1990) argued that mixing between recently arrived and existing water occurred, whilst Johnson *et al.* (1993) and McDonnell (1988) assumed that soil water retains a discrete chemical/isotopic signature so that water from various precipitation events remains distinct and traceable. The theoretical debate has obvious repercussions for various methodologies such as contaminant transport predictions (Johnson *et al.*, 1993) and hydrograph separation (McDonnell *et al.*, 1990; Beven, 1991).

The work reported in this chapter aimed to identify and characterise a number of features: the mechanisms which were responsible for the infiltration of water into the block; the extent of soil water movement; and whether new or old water generated discharge. Comparison of hydrographs, matric suction and soil moisture contents with those from the plots should facilitate an understanding of plot and field hydrology.

The lysimeter, installed on a site adjacent to the undrained field (Figure 3.4), included a five year-old mole drain and a newly installed mole channel (Section 3.1). From 5.6.91 to 11.6.91, 1098 *l* of water was applied to the soil block (Table 7.1).

Table 7.1 Prewetting, tracer application and storm simulations (June 1991).

Date and start time h	Volume (l)	Rate (mm hr^{-1})	Cumulative volume (l)
4	152	20	152
5	98	10	250
6	200	20	450
7	173		623
8	180		803
9	183		986
10	112	20	1098
11 (1500)	200	20	1298
12 (1500)	195	20	1493
13 (1355)	116	17	1609
13 (1805)	76	11	1685
14 (1500)	85		1770
15 (1300)	70		1840
17 (1000)	150		1990
18 (1000)	152		2142
19 (1000)	150		2292
20 (1000)	143		2435

Mean soil moisture contents ranged between 41.1 and 42.5 % (0.6 and 0.4 m; Table 7.2). However, in places the soil was at or close to saturation (46 %): maximum values were 44.8 to 51.5 %, but no sustained drainflow was recorded.

Table 7.2 Volumetric soil moisture contents (%) prior to tracer irrigation.

Depth (m)	Minimum	Maximum	Mean
0 - 0.1	35.5	51.5	41.9
0 - 0.2	34.2	47.0	42.0
0 - 0.4	40.2	44.8	42.5
0 - 0.6	38.5	45.0	41.1

An irrigation experiment, incorporating a chloride and stable oxygen tracer investigation, was conducted over ten days starting on 11.6.91 (Section 3.1). Following surface application of the tracers, eight storms were simulated over nine days. The volume and rate of water application was monitored (Section 3.4, Table 7.1, Figure 7.1). For the first five storms, hydrographic data were recorded by tipping buckets (Section 3.4.2) and drainflow samples were collected throughout the duration of discharge. Matric potentials (Section 3.3.7) were monitored at 30 minute intervals between 11.6.91 and 30.6.91. Volumetric soil moisture contents were determined at a maximum of a daily interval using the T.D.R. (Section 3.3.8). In addition, suction was applied to soil solution samplers immediately before storm simulations. Cups were evacuated 24 hours later, a field pilot test indicated that such a sampling regime would extract sufficient volumes of water for Cl and ^{18}O analyses. Volumes and rates of drain discharge were recorded for the final four simulations and automatic logging of matric potential continued.

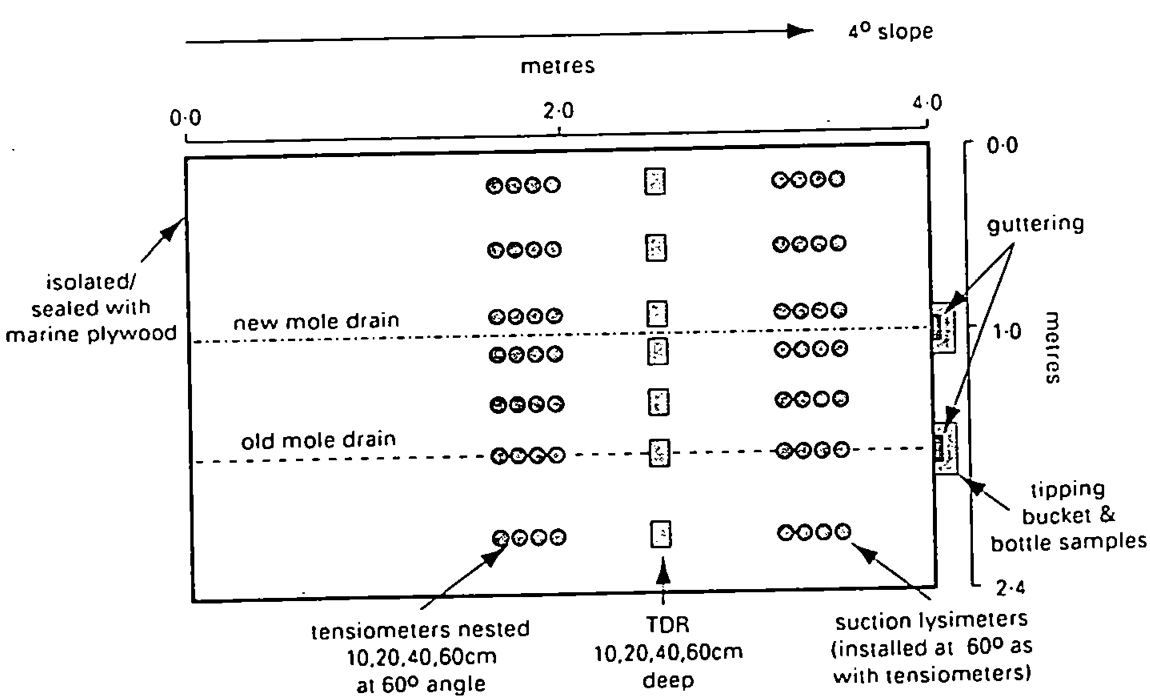


Figure 7.1 Plan elevation of isolated soil block.

Water balance.

Approximately 30 % of the irrigation water was output through the mole drains, 23 % being recovered from the new mole. Matric potential results indicated that water was retained by the soil but the T.D.R. showed the block only held 14.4 mm of irrigation water (2.4 % increase over 0.6 m) by 15.6.91. On the fields and the lysimeter not all of the macropores are connected to the mole drains, water is able to move to depth and as the block was not completely sealed water was observed leaking from the base. Deeks (1995) reported a similar loss during a carefully controlled mole drain laboratory experiment. She found that only 10 % of applied water left a 1 m^3 soil block via a mole drain.

Chapter 5 showed the value of hydrograph analysis for flowpath description, but observations were limited to generalisations at the field scale. Chapter 6 reported on the detailed plot scale, relating soil hydrology to field scale discharge and identified a need to more directly relate the two. To attempt this, simulations on 11.6.91 (1500 h) and 13.6.91 (1355 and 1805 h) are described in detail in the following sections. The first simulation allows consideration of the immediate fate of labelled irrigation water, which will be the carrier throughout the ensuing period. More specifically, infiltration of water into the block and runoff generation mechanisms were focussed on. The block was irrigated once more on 12.6.91 with unlabelled water. Analysis of data for the simulations on 13.6.91 enabled consideration of mechanisms which were responsible for water movement over longer periods.

Soil hydrology, runoff and tracer movement results for 11.6.91 and 13.6.91 are discussed separately (Sections 7.2 and 7.3). They are arranged in the following order:

- drain discharge;
- soil matric potential;
- LEACHW simulation (for 11.6.91);
- soil moisture content;
- tracer experiment - stable oxygen fate;
- discussion of hydrology and tracer movement.

It was assumed that stable oxygen would provide the best tracer (Section 3.4.3) however, chloride was employed as a precaution against analytical mishaps. Therefore, discussion of water composition focuses on stable oxygen (the behaviour and performance of chloride on both dates is discussed in Section 7.4).

7.2 Analysis of simulation data (11.6.91).

7.2.1 Hydrograph description.

The soil block responded rapidly to application of the labelled irrigation water. Irrigation at 20 mm hr^{-1} commenced at 1500 h (before which there was no drainflow) and water emerged from the new mole drain after 18 minutes (Figure 7.2). The hydrograph illustrates a rapid rise, with peakflow of 0.62 l min^{-1} (equivalent to 3.72 mm hr^{-1} rain) attained 30 minutes after application began. Old mole drainflow commenced after 30 minutes, however, a more gradual rising limb meant that peakflow was not attained until 55 minutes after the start, this was maintained for 10 minutes (Figure 7.2)

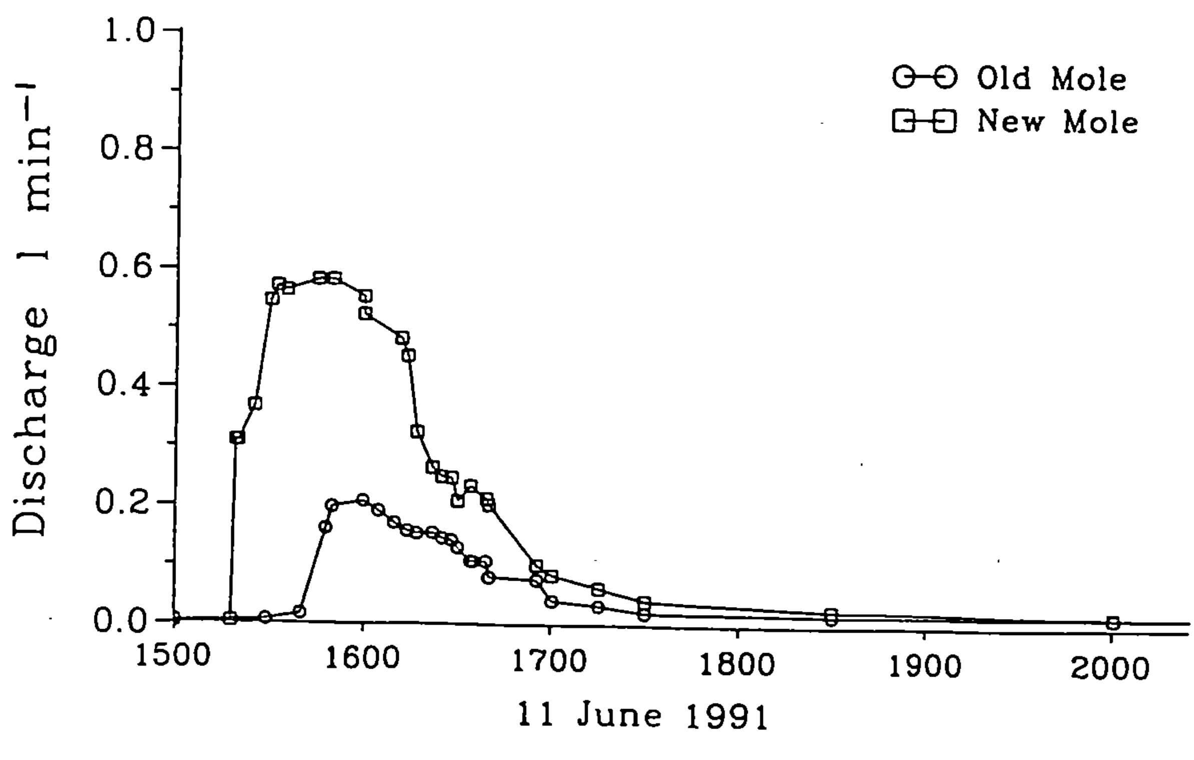


Figure 7.2 New and old mole drain hydrograph (11.6.91).

A steady flow was maintained from both drains until irrigation ceased at 1600 h. At this point, discharge from the new mole decreased dramatically as evidenced by the steep recession curve and runoff from the old mole exhibited a more gradual decline. Surface ponding, which occurred for a matter of minutes on the lower edge of the soil block, is assumed to have caused a step on the recession curves of both hydrographs. Discharge occurred for 9 hours in total. However, flow rates after 1700 h were less than 0.1 l min^{-1}.

The steep rising limb indicated that once a threshold had been passed, flowpaths were established which conveyed water rapidly to the drain. The nature of this threshold will be considered along with matric potential (Section 7.2.2) and soil moisture content (Section 7.2.4).

The minimal lag and peaky hydrograph indicated that flow was generated by one of three possible mechanisms: macropore flow, translatory flow, or a perched watertable. The absence of baseflow at either drain means that the last two explanations can be discounted given the nature of the soil porosity and the low hydraulic conductivity of the soil (Sections 4.2.2 and 4.2.5).

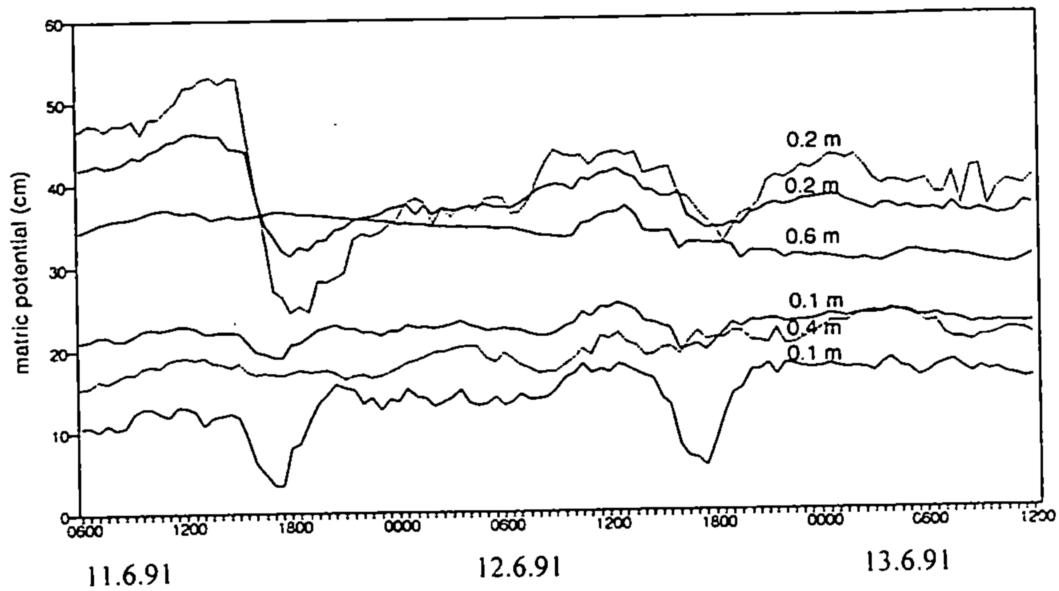
7.2.2 Matric potential.

Matric potential was monitored at 0.1, 0.2, 0.4 and 0.6 m to investigate soil hydrological conditions during irrigation and to relate potentials to drainflow. Chapter 6 indicated two distinctly different hydrological regimes: on the drained plot, incident rainfall generated stormflow from unsaturated soil in November (Section 6.3); in March, positive heads were observed on the undrained plot whilst stormflow but no baseflow occurred.

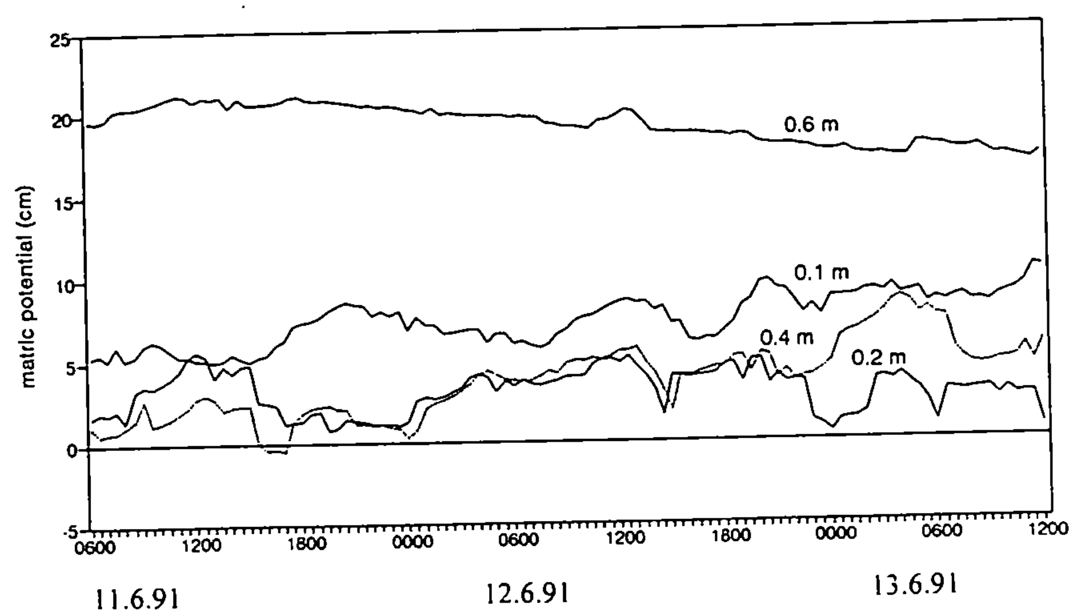
Figure 7.3 shows that on 11.6.91 matric potential was variable throughout the block, ranging from 62 cm H_2O (negative pressure) in profile 1 at the edge of the block, to near saturation at several points above the old and new mole drains (2 cm H_2O). Response to irrigation was also variable. Wetting did not appear to be related to depth from surface, instead the extent and speed of wetting were related to pre-event soil moisture status so that soil water heterogeneity was reinforced. Antecedently wetter sites (e.g. 0.2 and 0.4 m

Figure7.3 Matric potential (11 to 13.6.91).

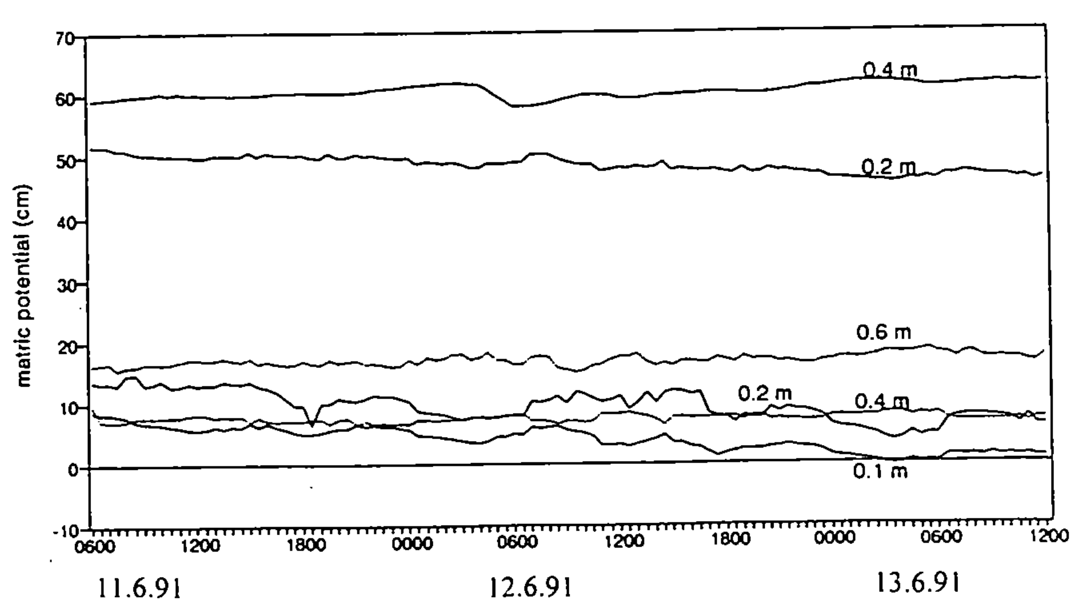
Profiles 4 and 5 (over new mole)



Profile 3 (midmole)



Profiles 1 and 7 (edge of lysimeter)



x axis: hours

in profile 3) received new water within 30 minutes. Tensiometers located above the mole channels, particularly at the surface, were most responive. For example, matric potential at 0.2 m above the new mole declined from 53 to 27 cm H_2O between 1400 and 1600 h on 11.6.91. The more rapid response illustrates that fissures provided an effective route for water to move into the matrix. Other tensiometers monitored slow wetting (0.6 m in profile 3 and 4) and some drier sites showed no wetting at all (0.1 m and 0.4 m in profile 5). Sites with lower matric potential have higher conductivities either because of their hydrological status or due to their physical character.

Therefore, in most instances, no relationship was observed between change in matric potential and generation of runoff. Water was conveyed through a relatively small volume of the soil. This concurs with findings for the plots in November, 1990 (Section 6.2). Section 7.2.3 compares field matric potentials with those predicted by LEACHW. Flow rates were not calculated for reasons given in Chapter 8.

In summary:

1. high matric potential was maintained throughout much of the soil block, even during irrigation. Drain discharge occurred while the bulk of the soil was unsaturated. Water bypassed most of the soil;
2. flowpaths were relatively complicated (profiles and horizons did not respond uniformly);
3. event/post event soil moisture status was dependent on antecedent soil moisture status, which strongly influenced infiltration and percolation of event water;
4. soil conductivity was at a maximum at the wettest sites.

in profile 3) received new water within 30 minutes. Tensiometers located above the mole channels, particularly at the surface, were most responsive. For example, matric potential at 0.2 m above the new mole declined from 53 to 27 cm H_2O between 1500 and 1700 h on 11.6.91. The more rapid response illustrates that fissures provided an effective route for water to move into the matrix. Other tensiometers monitored slow wetting (0.6 m in profile 3 and 4) and some drier sites showed no wetting at all (0.1 m and 0.4 m in profile 5). Sites with lower matric potential have higher conductivities either because of their hydrological status or due to their physical character.

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2. flowpaths were relatively complicated (profiles and horizons did not respond uniformly);
3. event/post event soil moisture status was dependent on antecedent soil moisture status, which strongly influenced infiltration and percolation of event water;
4. soil conductivity was at a maximum at the wettest sites.

7.2.3 LEACHW.

LEACHW (Section 6.4) was employed to aid the interpretation of field results. The discharge data in Figure 7.4 show that the lag predicted by Richards' Equation was too great and cumulative leachate exceeded that which was discharged by the two mole drains. Potentials predicted by the one-domain model for the start of the experiment were higher than those in the field (Figures 7.3 and 7.4). On irrigation, LEACHW values fell sharply at all depths. The drop in potentials was steepest and occurred first at 0.1 m.

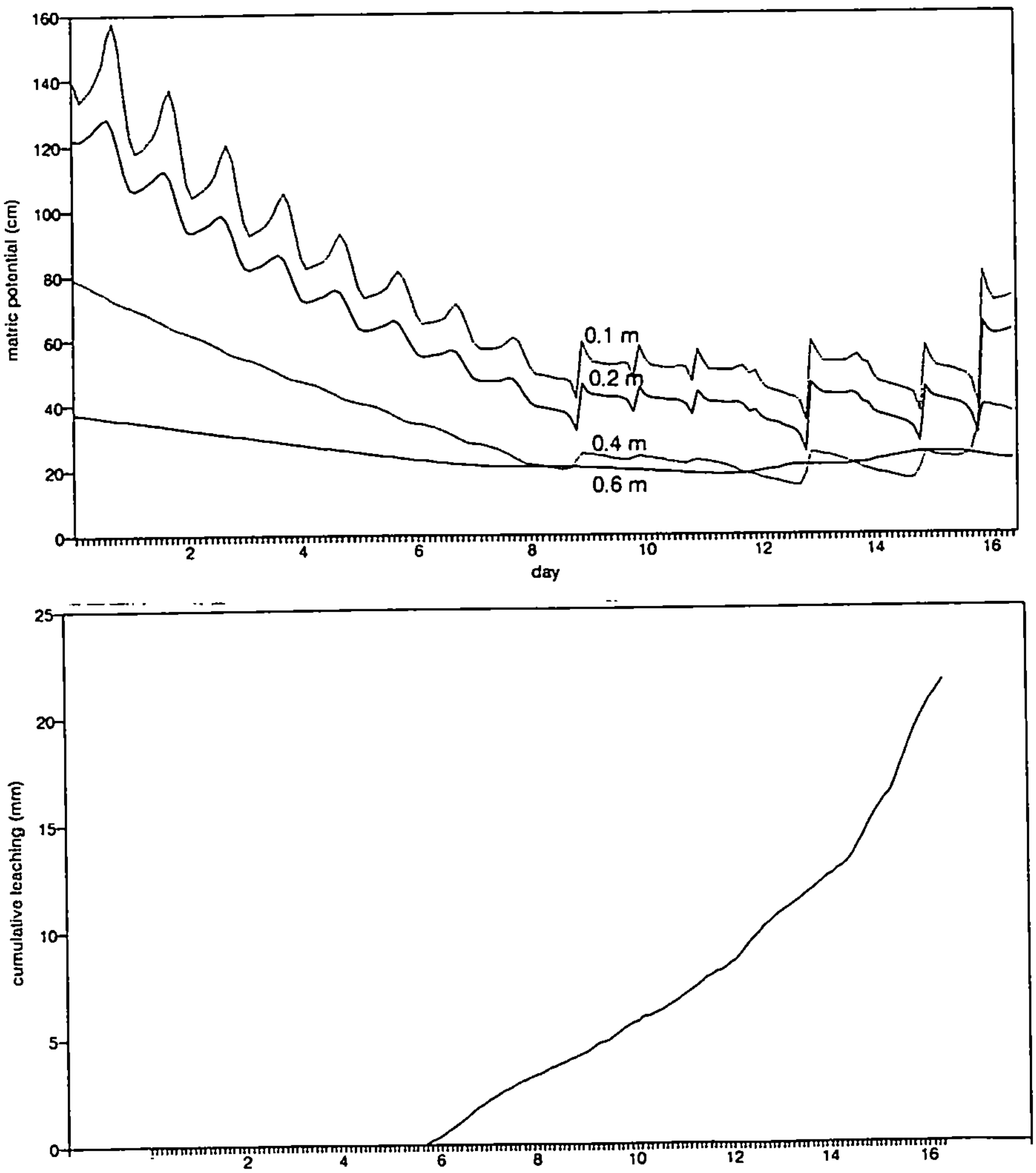

Figure 7.4 Richards' simulation of matric potential and discharge (1 . - 16.6.91).

The two-domain model produced a better simulation of timing of discharge from the block with a lower lag, shorter duration and higher total leachate (Figure 7.5). Potentials remained higher for longer when Addiscott's algorithm was applied. They followed a stepped pattern, as evaporation was deemed to have a significant effect.

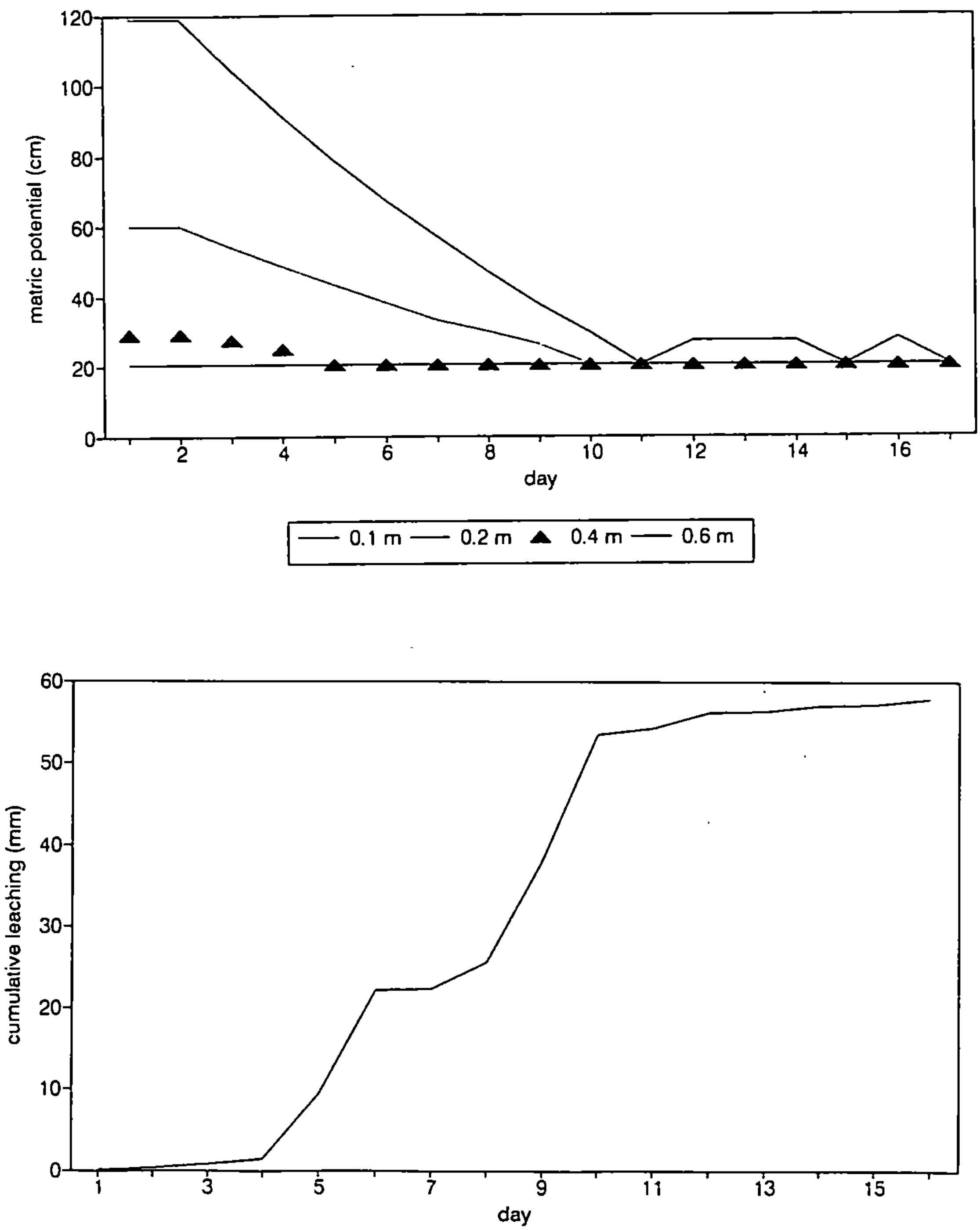

Figure 7.5 Addiscott's simulation of matric potential and discharge (10 - 16.6.91).

7.2.4 Soil moisture content.

Volumetric moisture content of the lysimeter soil was observed to:

- characterise detailed soil hydrology;
- determine the fate of irrigation water and attempt to identify the water pathways which generate drainage water;
- calculate the water balance.

Section 6.4 illustrated that water did not move uniformly through the Rowden Moor soil and Section 7.2.2 indicated that the soil block behaved in a similar way. An intensive transect of T.D.R. probes was installed across the lysimeter to provide additional information on soil hydrological conditions (especially changes in soil water storage) and monitored water content over larger volumes of soil than tensiometers. This was an attempt to determine whether water flowed through preferential zones or fingers (Steenhuis and Parlange, 1988), or through more discrete macropores and fissures.

Prior to irrigation, at 1400 h on 11.6.91, the range of soil moisture content in the block was 35.5 to 51.5 % (0.1 m at sites 4 and 8, respectively). Over the 0.6 m profile, moisture content was less variable (38.5 to 45 %). This small range was equal to 38.9 mm of water in the 0.6 m profile. The pattern seen in Figure 7.6 for each site, was of decreasing soil moisture content with depth, from 0.2 to 0.6 m, while the upper 0.1 m was either the driest or wettest portion of the soil.

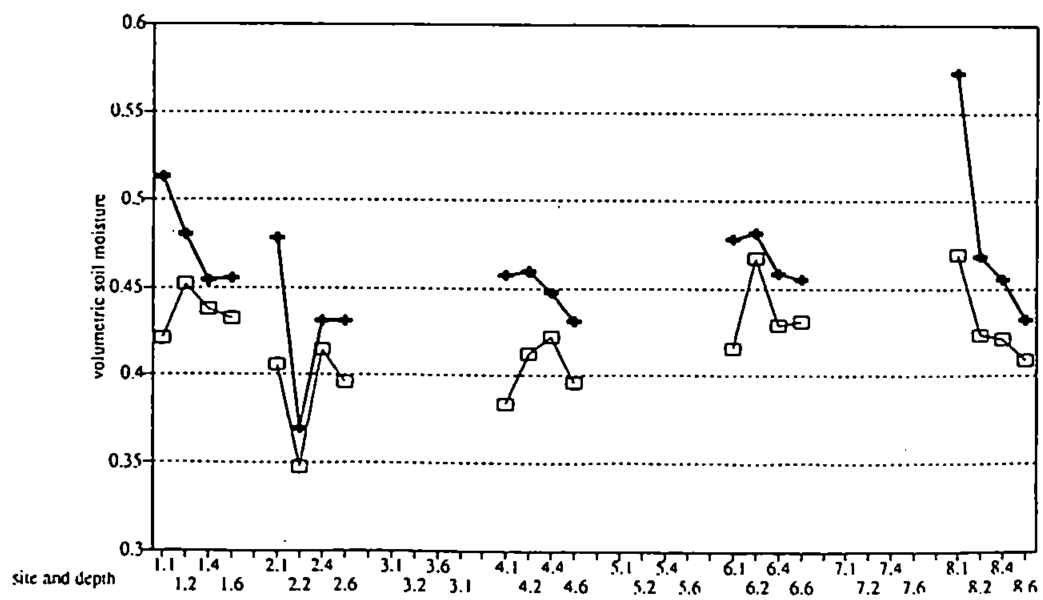

Figure 7.6 Lysimeter soil moisture content 11.6.91

From 1500 h, increased volumetric soil moisture content was registered at all sites (Figure 7.6). Soil moisture increased from 41.9 to 48.5 % at the surface and from 41.1 to 44.5 % at depth. The average increase in water storage was 5.2 %, however, there was a relatively large variation between depths and sites (Table 7.3). The smallest change was observed at 0.2 m depth at site 6. Here, an increase of 0.1 % (0.2 mm) was noted. The maximum difference recorded at 0.2 m at site 2 was considerably higher (15.2 % or 30.3 mm).

Table 7.3 Summary of lysimeter soil moisture results (%).

Date	measure	0-0.1 m	0-0.2 m	0-0.4 m	0-0.6 m
11.6.91	mean	41.9	42.0	42.5	41.1
(1400 h)	st dev	5.75	4.32	1.14	1.90
11.6.91	mean	48.5	47.4	45.3	44.5
(1600 h)	st dev	3.25	1.40	0.69	1.32
12.6.91	mean	43.6	44.7	44.5	42.6
	st dev	3.45	2.09	2.14	1.7
13.6.91	mean	43.8	44.0	43.8	43.5
	st dev	2.85	1.94	0.99	1.32
14.6.91	mean	43.6	46.0	44.5	43.5
	st dev	2.20	1.33	1.61	1.64

The change in moisture content will be discussed over the four depths monitored. By 1600 h, irrigation had resulted in an altered soil moisture content gradient. Highest values occurred in the upper 0.1 m and at this depth, variability of soil moisture reduced from 16 % (16 mm) at 1400 h to 8.4 % (8.4 mm) at 1600 h. The increase in mean volumetric soil moisture content from 41.9 % to 48.5 % was equivalent to the uptake of 6.6 mm of water. All other depths also experienced decreases in the range of values. Increased average moisture content between the surface and 0.2 m was only 4.2 mm greater than that to 0.1 m, at 10.8 mm and over 0.4 m it was only 11.2 mm. However, over 0.6 m an average increase of 20.4 mm of stored water was observed. Increased soil moisture content values for various depths, regardless of their situation on the transect, are detailed in Table 7.3. The wetting noted in 0.2 and 0.4 m profiles was largely contributed by storage in the upper soil. The overall increase through the profile indicated greater wetting between 0.4 and 0.6 m than between 0.2 and 0.4 m. This suggests that water bypasses the upper soil and is conveyed to depth by macropores. Field observations showed that macropores induced by drain installation extend to this lower soil horizon. Although mole drainage initiates fissures and improves porosity, connectivity to drains is not guaranteed (Section 7.1).

The hydrological behaviour of profiles is now considered. Table 7.4 details the mean increase in soil moisture content for three sites following application of tracers. A pattern which might be interpreted as a form of preferential flow was observed. The greatest increase in soil moisture content occurred in profile 2, which had the highest antecedent water moisture content. Conversely, profile 6 the driest site, took up least water. Thus, wet soil provided pathways for water to enter the matrix first and captured event water. In terms of soil profiles, the biggest increase in water storage occurred in the upper horizon.

Table 7.4 Lysimeter: % change in volumetric soil moisture content (1400 h and 1600 h, 11.6.91).

Site	0-0.1 m	0-0.2 m	0-0.4 m	0-0.6 m
2	10.7	13.3	4.0	6.5
4	6.6	4.0	2.5	3.3
6	7.9	1.4	2.9	2.7

Hydrological monitoring highlighted the heterogeneous nature of the soil at all stages of the experiment. Rapid generation of runoff in the absence of a uniform wetting front or a water table points to the occurrence of bypass flow once detention along fissures is satisfied. Tensiometer and T.D.R. arrays both identified wet and dry parts of the profile and recorded some soil hydrological response to irrigation as the wettest parts of the matrix provided more conductive routes for water movement. However, the bulk of the matrix remained unsaturated: the typically low permeability of the clay soil prevented uniform wetting.

7.2.5 Analysis of water composition.

Stable oxygen and chloride tracers were employed in an effort to elucidate flowpaths operating in this fissured soil. Hydrological results and modelling suggested that macropore flow and matric wetting and drying were important. However, their relative contributions and interaction were not quantified.

The following theories were investigated:

- water moves through cracks and fissures without mixing; incoming rain would therefore retain its chemical makeup (Wenner *et al.*, 1990; Johnson *et al.*, 1993);
- water moves slowly through the matrix and mixes, displacing old water (Weyman, 1970).

The stable oxygen concentration of irrigation water was $\delta^{18}O$ +3.5 while that of matric and drainage water was $\delta^{18}O$ -5.8. Composition of drain water and then soil water is discussed. The isotopic tracer was immediately evident in discharge from both moles

The first water from the new drain, emerged after 18 minutes and had a value intermediate between those of old and new water at $\delta^{18}O$ -3.6. By 1522 h, the value had risen to $\delta^{18}O$ -1.1 and declined to $\delta^{18}O$ -3.5 by 1526 h. The highest value for new mole samples at 1522 h is related to the high discharge at that time (Figure 7.7). The isotopic signature rose gradually to another peak of $\delta^{18}O$ -1.3, 55 minutes after the experiment began. On cessation of tracer application (1600 h), the stable oxygen concentration declined rapidly to $\delta^{18}O$ -3.1 at 1630 h and to $\delta^{18}O$ -4.1 at 1650 h. A new equilibrium of $\delta^{18}O$ -4.9 was reached 90 minutes after irrigation stopped.

A similar pattern was observed for the old mole drain which had a double peak and a long recession limb (Figure 7.8). The first sample of water arrived after 40 minutes and had a stable oxygen value of $\delta^{18}O$ -4.0. One minute later the value increased to $\delta^{18}O$ -3.7 before falling to $\delta^{18}O$ -4.7, 4 minutes after this (1445 h). The concentration rose to a peak of $\delta^{18}O$ -1.5 by 1510 h and declined to $\delta^{18}O$ -3.8 by 1630 h.

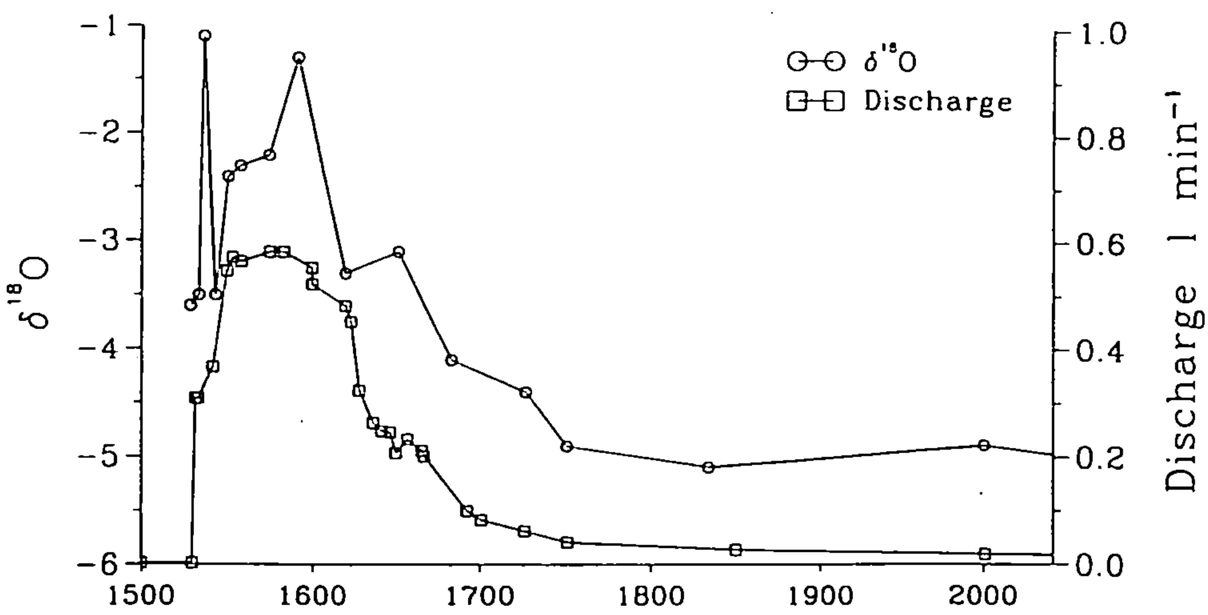

Figure 7.7 Tracer concentrations in new mole drain discharge (11.6.91)

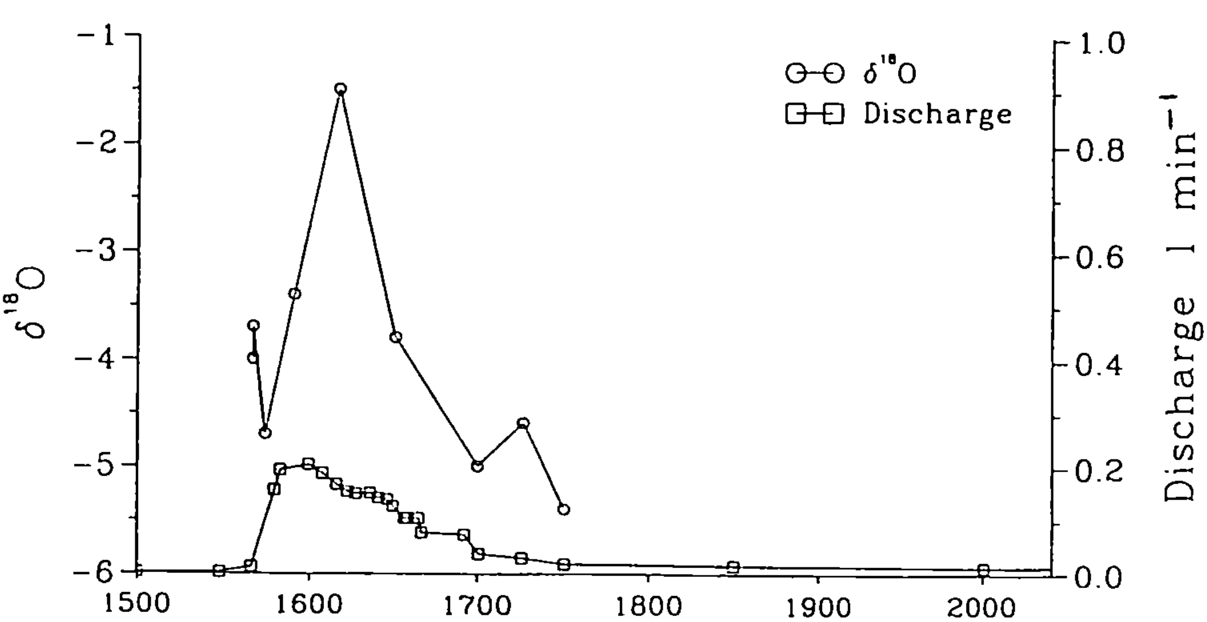

Figure 7.8 Tracer concentrations in old mole drain discharge (11.6.91)

Examination of runoff shows that the first flush of water from each drain had similar stable oxygen concentrations, although the old mole response was lagged by 23 minutes. Two explanations are offered. First, initial water moved rapidly and turbulently down the mole slot and cracks. This bypass flow washed old water from ped faces, but quickly exhausted this supply. As macropore walls became wetted, conductivity increased and opportunities for mixing of old matric water and new macropore water arose. Second, wetter segments of the matrix had a higher hydraulic conductivity and acted like a stack of tipping buckets (Figure 7.9). As applied water fulfilled the storage capacity in these zones, soil water spilled into lower soil volumes. This would explain the initial decline in $\delta^{18}O$, as the large volume of stored old water diluted incoming tracer at first. Maximum $\delta^{18}O$ values were attained towards the end of peakflow as tracer water became a more important component of the total water content, but even the maximum value attained of $\delta^{18}O$ -1.1 was considerably lower than that of the tracer being applied. The second explanation would also account for the sometimes rapid matric potential and soil moisture response reported in Sections 7.22 and 7.2.4.

The new mole hydrograph recession was more marked than that of the old mole drain because it obtained its water more directly from the surface and hence the residence and lag times were lower. The final equilibrium value was lower in old mole runoff because pathways from the wetted soil volumes were better established, so there was greater opportunity for old water to contribute to it. The old mole profile had experienced structural amelioration: it held more water prior to tracer application and had the highest hydraulic conductivity (Sections 7.2.2 and 7.2.4).

Soil water samples were extracted from the block throughout the experiment (Section 3.4.2). Before application of the tracer, soil water was $\delta^{18}O$ -5.8. A vacuum was applied to cups immediately before tracer application and subsequent irrigation events. Samples were then obtained daily, with a vacuum applied for a maximum of 18 hours. Soil water samples showed little evidence of uniform infiltration and percolation of labelled water in the upper 0.1 m. However, samples from other depths clearly contained tracers, e.g. profile 1 - 0.2, 0.4 and 0.6 m; profiles 5 and 6 - 0.4 m (Table 7.5). Explanation is attempted in Section 7.2.6, where other data are also considered.

Figure 7.9 a Schematic presentation of hydraulic zones on drained soil at Rowden

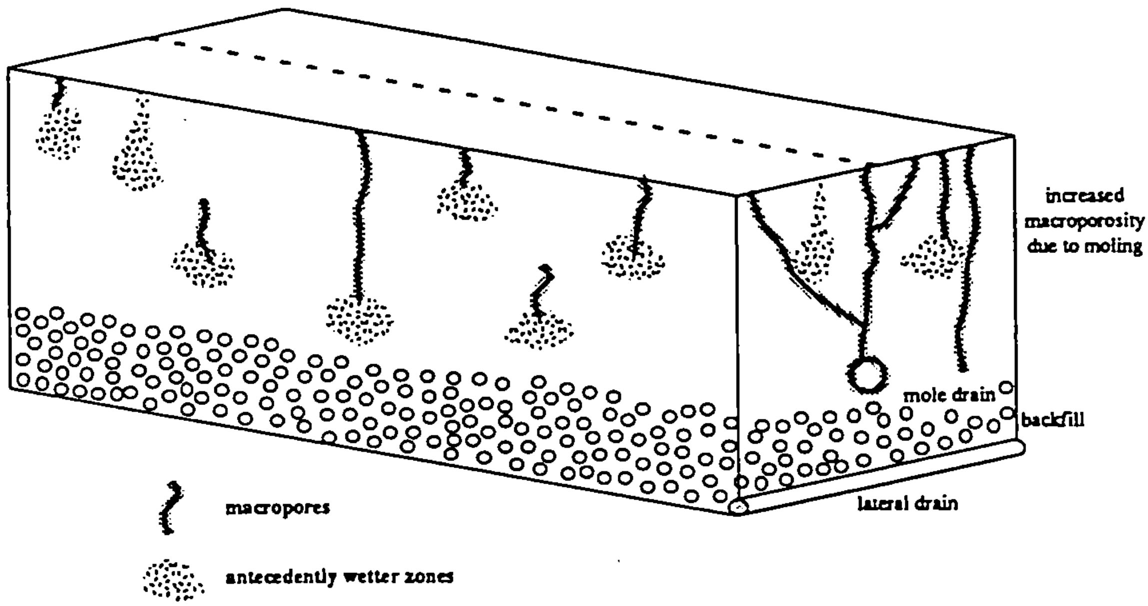

Figure 7.9 b Schematic presentation of flowpaths on drained soil at Rowden.

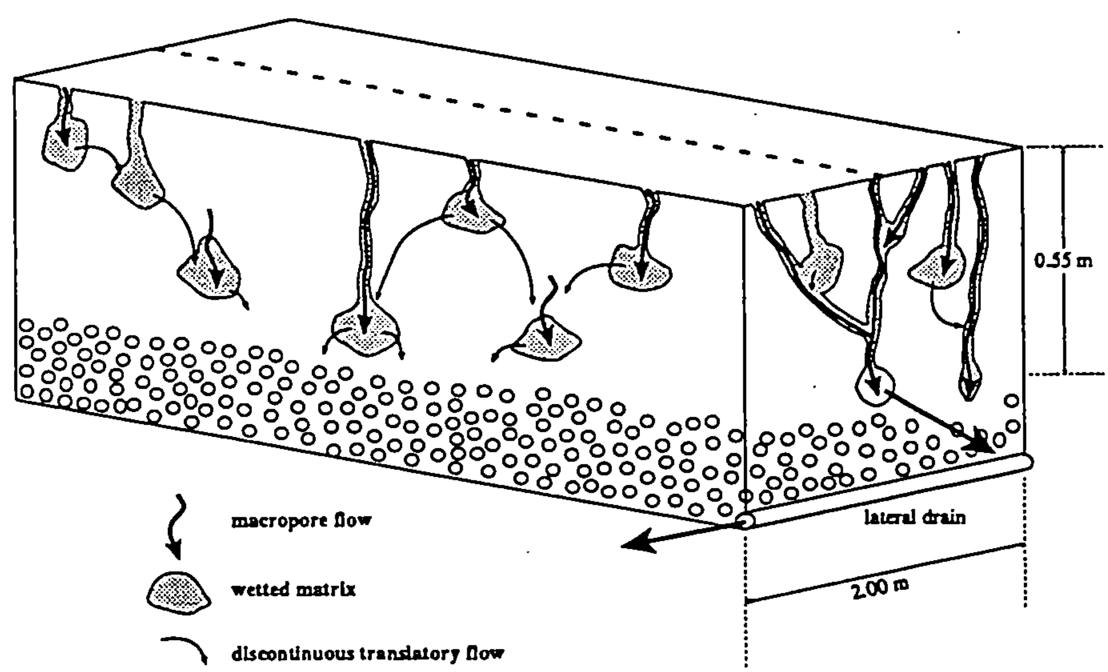

Table 7.5 Soil water composition $\delta^{18}O$.

	Distance across transect					
	0.3 m	0.8 m	1.0 m	1.3 m	1.8 m	2.3 m
Date &	Profile					
depth	1	2	3	4	5	6
10.6.91	-5.8					
11.6.91						
0.1 m	-5.7	-5.8	-5.2	-5.7	-3.2	-4.7
0.2 m	-3.2	-5.6	-2.8	-5.8	-2.5	-5.5
0.4 m	-3.7		-5.7	-5.8	-3.1	-2.6
0.6 m	-3.6	-5.8				

7.2.6 Summary of hydrological and tracer results.

Consideration of hydrometric and tensiometric data at the most detailed level of this nested study enabled a number of conclusions regarding the hydrological character of the block to be drawn:

1. discharge was concentrated during periods of irrigation, when application rates exceeded the infiltration capacity of the matrix and macropore flow was generated. Not all macropore flow was conveyed to the mole drains;
2. the soil matrix was unsaturated. The wettest profile was above the old mole, and the wettest segment was 0 to 0.1 m;
3. results for matric potential and soil moisture contents were similar, indicating wet and dry zones even though they look at different volumes of soil;
4. soil moisture content at 0.4 to 0.6 m changed considerably more than that in the horizon above.

The short lag time before generation of drainflow, the sharp recession curve (Section 7.2.1), the spatially confined nature of soil matric potential response (Section 7.2.2) and the distribution of soil moisture (Section 7.2.4) all provide evidence for effective bypass flow. However, evidence of matric response from tensiometers, T.D.R. and soil water suction samplers leads to the following proposition:

5. Wetted zones became hydrologically connected, so that new and old water mixed in the matrix. These soil volumes provided discharge during simulations and low runoff was sustained for around 8 hours.

The distribution of different soil hydraulic zones and related water pathways is schematically presented in Figure 7.9.

7.3 Analysis of simulation data (13.6.91).

7.3.1 Introduction.

The block continued to be irrigated daily (Table 7.1) and samples of matric and drain water were collected to determine the nature and extent of tracer water movement. This section examines data for 13.6.91, to illustrate the fate of tracer water applied several days earlier.

7.3.2 Hydrograph description.

Two storms of 115.5 litres and 75.5 litres (11.55 and 7.55 mm) were simulated. The first application began at 1355 h. Once more, response was rapid with new and old mole discharge commencing at 1416 h and 1435 h respectively (Figure 7.10). New mole runoff peaked at 1445 h (0.9 l min^{-1}). Application ceased at 1435 h (application rate of 2.9 l min^{-1} or rain of 17.3 mm hr^{-1}). By 1630 h, discharge had declined to 0.1 l min^{-1}. Commencing at 1805 h, the second irrigation event ended at 1855 h. The new mole responded at 1815 h and in spite of a lower intensity of application (1.9 l min^{-1} or 11.3 mm hr^{-1}) response was faster due to higher antecedent soil moisture. The rising limb was less steep, leading to a peak discharge of 0.4 l min^{-1}. As on 11.6.91, runoff declined quickly when irrigation ceased on 13.6.91.

Cumulative discharge from the new mole was 20 litres following the first application and this rose to just under 40 litres by 2000 h. However, discharge from the old mole was only 10 litres. Only 26.2 % of the applied water was retrieved in drain discharge 13.11.91.

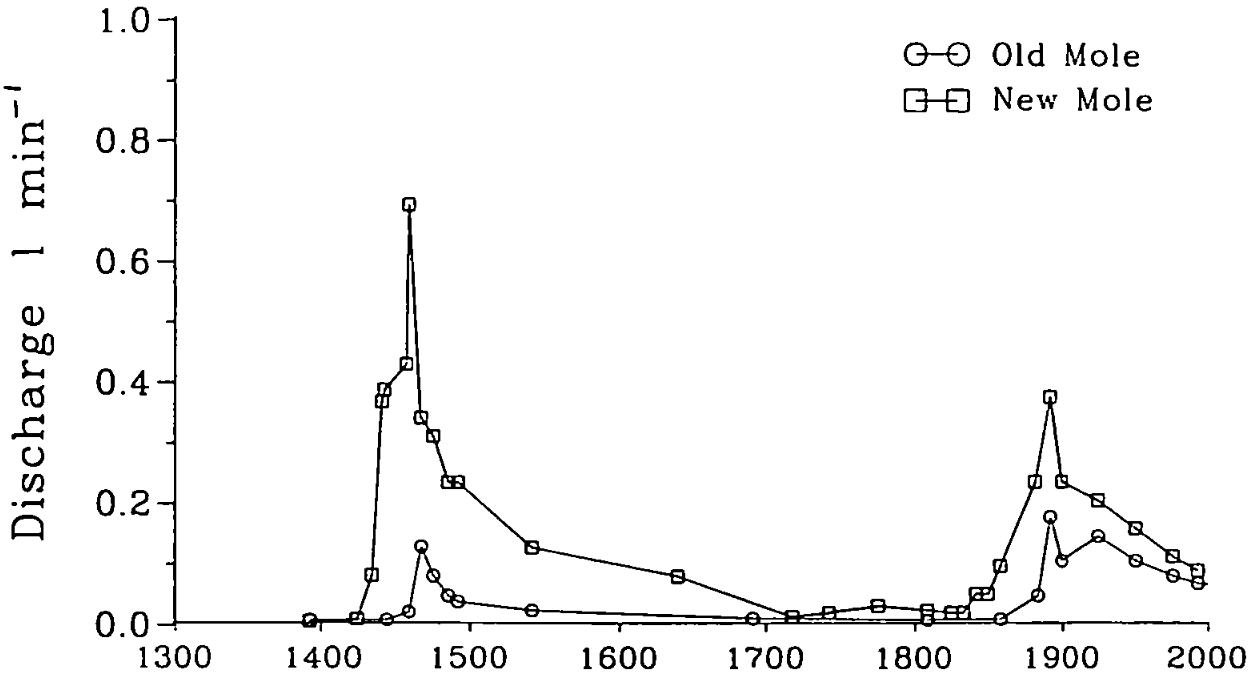

Figure 7.10 New and old mole drain hydrographs (13.6.91)

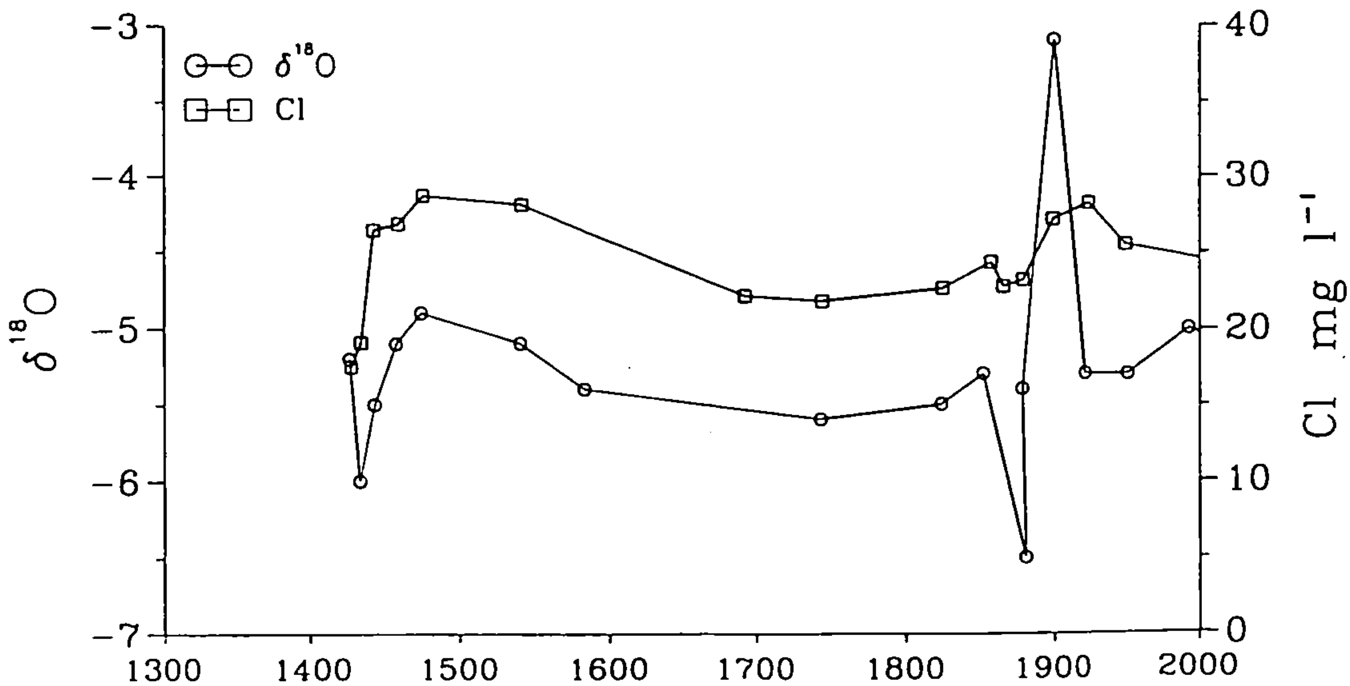

Figure 7.11 Tracer concentrations in new mole drain discharge (13.6.91)

7.3.3 Matric potential: spatial and temporal variability.

On 11.6.91, matric potential observations suggested that the response of the unsaturated soil to irrigation was highly variable, often being related to antecedent moisture conditions. Figure 7.12 shows that matric potentials continued to be variable, ranging from 58 cm H_2O in profile 1, to near saturation at several points above the moles (4 cm H_2O). Tensiometers identified some persistent wet sites (e.g. 0.2 and 0.4 m in profile 3). Matric potential continued to reduce slowly at other sites (0.6 m in profile 3 and 4) demonstrating that water was being absorbed by the matrix. Surface tensiometers above the mole still responded before those at some distance from it. Other parts of the matrix remained unresponsive (0.1 m and 0.4 m in profile 5).

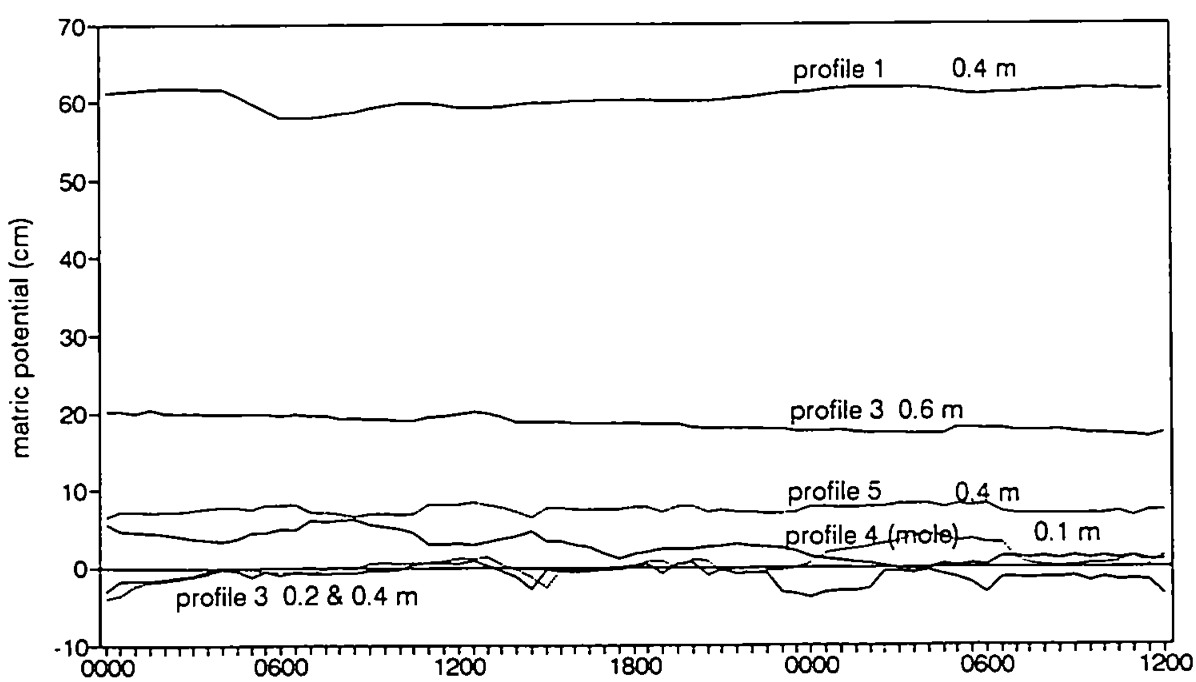

Figure 7.12 Matric potential 13.6.91

Therefore, observed matric potential and runoff generation appeared to continue to be unrelated as preferential flow continued to operate through the matrix and macropores. Thus, the five concluding points of Section 7.2.2 also provide a valid summary for the lysimeter on 13.6.91.

7.3.4 Soil moisture content: spatial and temporal variability.

Soil moisture contents continued to be variable on 13 and 14.6.91. Highest values occurred on different days for different profiles (e.g. profile 1 1600 h, 11.6.91; profile 2, 14.6.91). Moisture content generally decreased with depth, but as wetting proceeded 0 to 0.2 and 0 to 0.4 m segments in profiles 2, 4 and 7 became markedly wetter. In contrast, profiles 6 and 8 contiued to show little evidence of wetting.

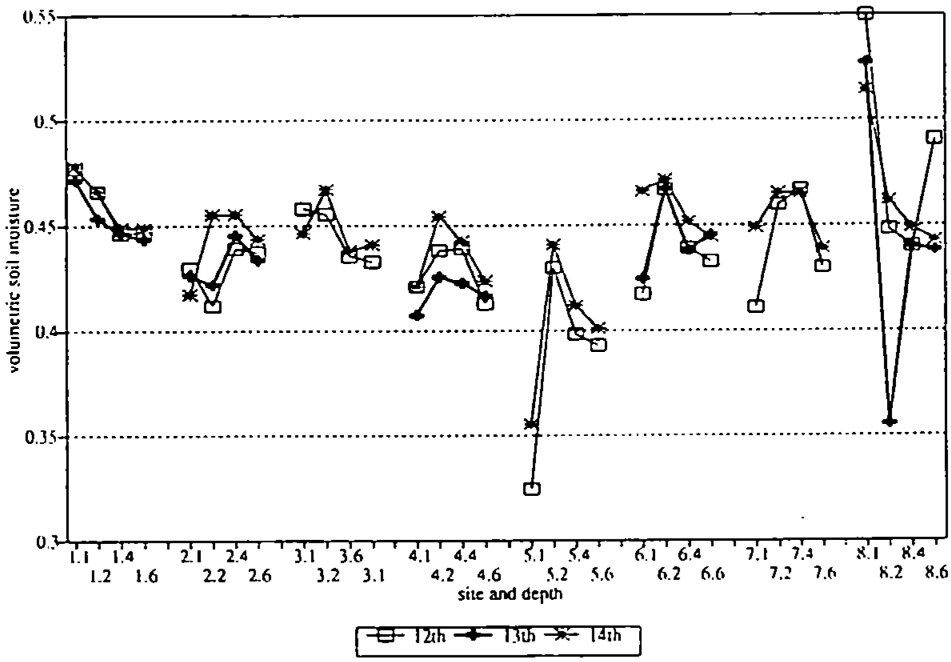

Figure 7.13 Moisture content for soil profiles along drain-centred transect (13.6.91). (e.g. profile 3, 0.4 m = 3.4)

7.3.5 Analysis of water composition.

In contrast to new mole discharge composition for 11.6.91 (peak $\delta^{18}O$ -1.1 and minimum $\delta^{18}O$ -4.9 in baseflow; Section 7.2.1), Figure 7.11 shows that values on 13.6.91 ranged from $\delta^{18}O$ -4.9 (peakflow) to $\delta^{18}O$ -5.6 (baseflow). Stable oxygen concentrations on the rising limb of the first simulation were largely between $\delta^{18}O$ -5.5 and -4.9. After peak discharge at 1445 h, values fell back slowly to $\delta^{18}O$ -5.5, but still largely followed the hydrograph shape.

In the second simulation (starting 1805 h), new mole drainflow $\delta^{18}O$ rose immediately to $\delta^{18}O$-5.5 (1815 h) and -5.0 (1835 h). While irrigation continued ($\delta^{18}O$ -5.8), a drop to $\delta^{18}O$ -6.5 (1840 h) was observed. As this was followed by a value of $\delta^{18}O$ -5.4 (1848 h), the accuracy of this result may be questioned. Shortly after peak drain discharge, the isotopic signature peaked at $\delta^{18}O$ -3.1, before falling to $\delta^{18}O$ -5.3, in line with the recession curve.

Soil water composition results are presented in Table 7.6. They illustrate the highly variable distribution of tracer within the block. There was little evidence of uniform infiltration and percolation of labelled water in the upper 0.1 m, whilst samples from other depths clearly contained tracers, e.g. profile 1 at 0.2, 0.4 and 0.6 m; profiles 5 and 6 at 0.4 m. Temporally, several cups sampled high concentrations of irrigation water on 11.6.91 but showed that much more background water was at these locations later in the experiment, (e.g. profiles 1 and 3 at 0.2 m). Water at other sites changed more gradually and some sites showed little variation of soil water composition throughout the monitoring period (e.g. profile 1 at 0.1 m).

It should be remembered that imposed suctions were greater than the matric potential, so sampled water was a mix of 'stored' and 'moving' water. However, the two water bodies were inextricably linked. The former continued to contribute labelled water from the 11.6.91 irrigation to percolating water on 13.6.91. Maximum tracer concentrations at peakflow demonstrated that water pathways were persistent and that rapid mixing continued. These observations have ramifications for solute movement and pollutant transport.

Table 7.6 The influence of tracer application on soil water composition δ¹⁸O.

Date & depth (m)	0.3 m	0.8 m	1.0 m	1.3 m	1.8 m	2.3 m
10.6.91	-5.8					
11.6.91						
0.1	-5.7	-5.8	-5.2	-5.7	-3.2	-4.7
0.2	-3.2	-5.6	-2.8	-5.8	-2.5	-5.5
0.4	-3.7		-5.7	-5.8	-3.1	-2.6
0.6	-3.6	-5.8				
12.6.91						
0.1	-5.6			-5.8		-3.3
0.2	-4.0		-3.5			-5.8
0.4			-5.8	-5.5		
0.6				-5.1		-5.6
13.6.91						
0.1	-5.8	-3.4	-5.5			-3.1
0.2	-4.3	-4.7	-3.7			-4.7
0.4						
0.6		-5.9	-5.8			-5.7
14.6.91						
0.1	-5.6				-3.9	
0.2	-4.3	-3.5		-3.6	-4.1	-5.2
0.4		-4.3			-3.7	
0.6						-5.8
15.6.91						
0.2	-4.7	-3.5	-4.2	-3.9		-4.5
0.4		-5.1	-4.5			-5.6
17.6.91						
0.2		-4.9	-4.1	-4.5		
0.4		-5.5	-4.9			
19.6.91						
0.2	-5.7	-5.2		-4.6	-4.3	-4.9
0.4		-5.5				-5.8

7.4 Chloride tracer results.

The variation (timing and amplitude) of chloride content in drain water mirrored that of δ ¹⁸O (Figures 7.11 and 7.14). The strength of the chloride tracer signal was similar to that of the isotopic tracer. The composition of soil water was determined much more intensively in terms of chloride content than stable oxygen content. Chloride concentration results were similar to those for stable oxygen: some sites were dominated by old water (normally sites with low antecedent matric potential). Surface concentrations were relatively unaffected by incoming water. Tracer was observed at other sites after a matter of days, whilst some areas showed no indication of the tracer throughout the monitoring period, indicating that not all of the soil block was hydrologically active. However, the data substantiated the conclusions drawn in earlier sections and suggest that:

1. chloride was behaving as a conservative tracer, acting in a similar manner to stable oxygen.
2. chloride movement was highly variable in magnitude and timing;
3. distribution of tracers was more widespread than matric potential measurements would have suggested.

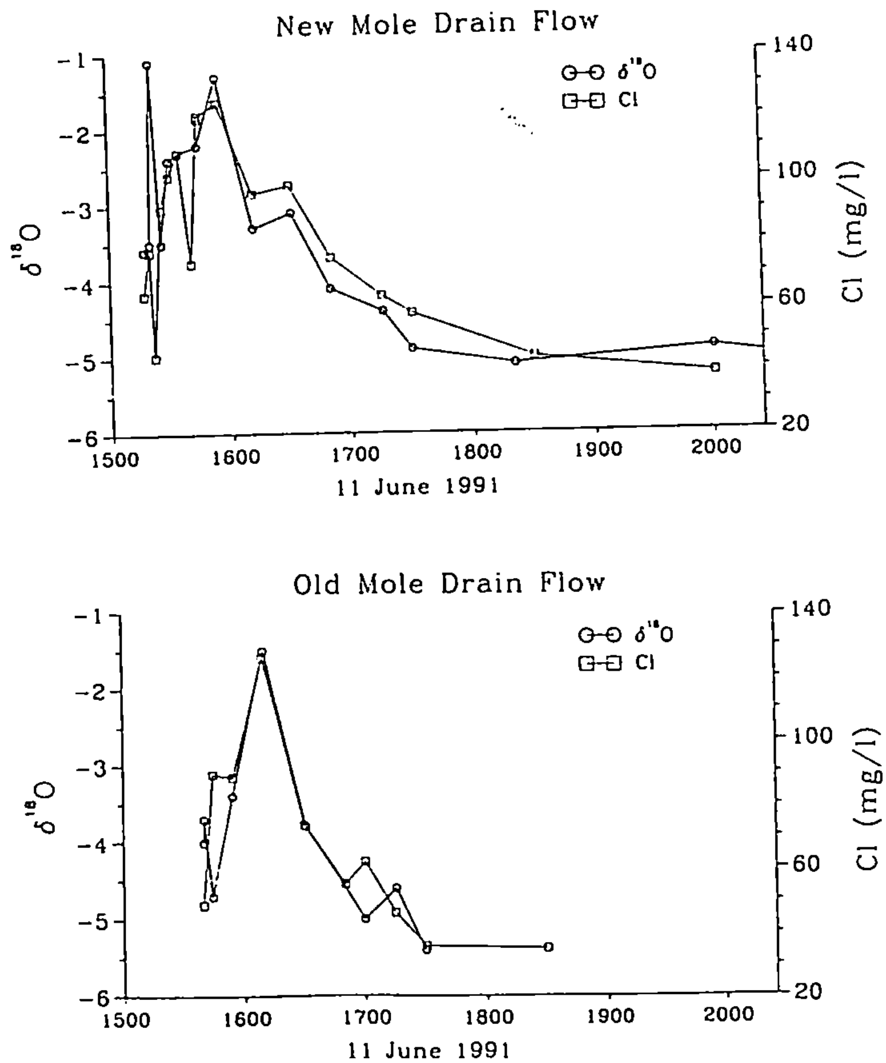

Figure 7.14 Chloride tracer concentrations in soil block drainflow (11.6.91)

Table 7.7 The influence of tracer application on soil water composition (Cl mg l^{-1}).

Date & depth (m)	0.3 m	0.8 m	1.0 m	1.3 m	1.8 m	2.3 m
10.6.91	23					
11.6.91						
0.1	40	26	43	31	98	58
0.2	96	31	100	20	112	40
0.4	75		25		97	111
0.6	88	24		24		
12.6.91						
0.1	24			22		96
0.2	73		90			22
0.4			22	32		
0.6				49		134
13.6.91						
0.1	22	93	28			102
0.2	71	62	80			60
0.4						
0.6		22	22			33
14.6.91						
0.1	30				75	
0.2	69	87		85	71	52
0.4		70			85	23
0.6						
15.6.91						
0.2	61	78	72	62		59
0.4		51	63			28
17.6.91						
0.2		48	73	64		
0.4		40	52			
19.6.91						
0.2	32	28		58	50	41
0.4		32				25

7.5 Synthesis of physical, isotopic and chemical results.

The tracer experiment was conducted in an attempt to characterise water movement in a small soil block. In particular, the relative importance of matrix and macropore water for drainflow generation was to be ascertained. Analysis of samples allowed the characterisation of isotopic signatures of drainflow and soil water. The experimental results will also be used to validate the assumption that new and old water isotopic signatures are distinct and remain so.

7.6 Conclusion.

The tracer experiment was conducted in an attempt to characterise water movement in a poorly structured soil with enhanced macroporosity, to expand upon the discussions of Chapters 5 and 6. From observations of soil hydrological response, drain hydrographs and soil and drainflow composition, it can be concluded that:

1. soil moisture content and matric potential were variable before tracer application;
2 water did not infiltrate uniformly, but largely exploited areas of high soil moisture content and low matric potential;
3. no water table was observed, therefore drain discharge was generated by either macroporeflow or preferential matric flow. Evidence of wetter zones and response in the matrix point to preferential flow. The spatially sporadic occurrence of wetted pockets of soil suggests that field capacity could be exceeded in these areas, enabling them to generate throughflow (Figure 7.9). Where wetted soil pockets became connected, they would be able to generate rapid drain discharge: a form of discontinuous translatory flow;
4. with regard to the new water-old water controversy, the rapid generation of drainflow and absence of a water table suggest that macropore flow is in operation. However, analysis of tracer concentrations showed that runoff which reached the drain early in the irrigation events and at peak dicharge always included both old and new water. Turbulent water movement through macropores was able to pick up water from ped faces and micro- and meso-pores which lined fissures and the mole channel. The matrix and macropores are inextricably linked similar conclusion to that drawn by Dowd *et al* (1991). During subsequent irrigation

events when macropore flow and translatory flow were at a maximum, it might have been expected that labelled water would be diluted. The opposite is true, at peak discharge stored labelled water was released from the matrix which surrounded the macropores and which constituted the wetter zones indicating that flowpaths were persistent.

Results indicate that considering macropores and the matrix as separate entities is an oversimplification. Runoff generation is more complicated than simple bimodal models would suggest. Further, classic views such as those advocated by Sklash *et al.* (1976, Section 1.6) assume that water conveyed by macropore and translatory flow retains a discrete isotopic signature. Evidence from this experiment refutes this idea. Attempts by investigators such as McDonnell (1990) to further demarcate hydrological pore functions and water pathways add little to the work of authors such as Sklash. This is because they are based on the compartmentalisation soil water pathways and soil pores rather than treating the soil and soil water movement as a continuum.

CHAPTER 8: SYNTHESIS AND CONCLUSIONS.

8.1 Introduction.

Previous work on Rowden Moor highlighted the operation of different pathways on undrained and drained soil (Hallard, 1988). It was suggested that overland flow operated on undrained sites, while lateral subsurface flow dominated on drained fields even though it was not always directed to the mole drains. The mechanisms of subsurface runoff generation were not understood, but were of interest as the study site is the location of nitrate modelling and management experiments.

As a continuation of work at the Rowden site, runoff generation mechanisms were identified. Their spatial and temporal variability was investigated using a combined hydrometric, tensiometric, isotopic and chemical tracing approach. Water movement was studied at three scales: the field (1 ha), the plot (10 m^2 within the field) and the lysimeter (10 m^2 isolated soil block). Work at these scales represents a nested approach to the study of soil water from the landscape to the soil profile. In the one hectare field, soil pathways were unknown and the soil was essentially treated as a black box, information regarding the distribution of outputs formed the basis for inference about soil water pathways. Plots were located within fields and enabled point observation of soil hydrological response. However, the plots were not hydrologically isolated and field runoff generation could not be directly related to observations of soil response. The lysimeter allowed characterisation of the parameters monitored at the other two scales, and supplemented the flow data with tracer information. The major findings of this study are reviewed below in terms of soil conditions and hydrology (Section 8.2), solute movement (Section 8.3) and suggestions for further work (Section 8.4).

8.2 Undrained and drained hydrology.

Over the drainage season, the volumes of runoff from undrained and drained fields were similar, yet the routing of water was very different (Chapter 5). On the undrained field all runoff was conveyed by overland flow and shallow subsurface flow even though the soil was only saturated from January to March. By contrast, on the drained field 60 % of precipitation left the site via the weirs and 97 % of this was drainflow. Therefore, mole drainage was very effective in diverting soil water to subsurface pathways. To describe and account for this, a more detailed consideration of the soil and its hydrology was undertaken.

The information in Chapter 4 describes how soil properties varied within profiles due to the distinct A/Bg boundary. The upper 0.2 m of the soil had a higher porosity and saturated hydraulic conductivity, this being more marked on the drained site. Matric potential and volumetric soil moisture contents varied anisotropically on the undrained plot. However, important differences in pedological and hydrological characteristics (Chapters 4 and 6) were observed under different land management.

On the drained plots mole installation resulted in greater connectivity of the soil surface and subsurface and had four implications for drained plots. First, some of the subsurface soil and the drains responded to rainfall relatively quickly. Second, the 0.6 m soil profile was generally drier throughout the drainage season as indicated by higher matric potentials and lower moisture contents. However, moisture contents were higher in the drained topsoil (encouraged by improved structure, better rooting, more faunal activity). Third, soil moisture profiles varied isotropically . Fourth, the magnitude of soil hydrological response was greater on the drained plot and this was reflected in peakier field hydrographs.

The isotropic variation of moisture content on the drained plot and the lysimeter was confirmed by measurements of moisture content but not matric potential results (Chapter 6). It was not possible to identify a spatial effect due to mole drain location across the drained plot or lysimeter transects. At the micro scale, soil heterogeneity complicated matric potential patterns. However, in larger volumes of soil (as monitored by the T.D.R.) a significant influence was detected. Therefore, observations of soil hydrological behaviour could lead to very different conclusions depending on the parameter which is characterised, the technique and instrumentation employed and the scale of study. Further work might focus on the definition of an appropriate scale of investigation for a heterogeneous soil such as that at Rowden.

The drained soil was observed to be more conductive (Sections 4.2.2 and 4.2.5) representing the cumulative impact of drainage on soil hydraulic properties and soil water movement. Greater hydraulic conductivity can be attributed to increased porosity through a range of pore sizes (Section 4.2.2). However, structural amelioration did not overcome the low permeability of the soil so that large proportions of water appeared to bypass the upper soil and were transmitted to Bg and C horizons. Similar observations were made at Rowden by Hallard (1988) who discussed drained field hydrology and Deeks (1995) who reported on a 1 m^3 mole drained lysimeter experiment.

The variable hydraulic character of soil reported for all sites meant that water did not move uniformly through either undrained or drained soil. On fields and the lysimeter, the bulk of the soil was unsaturated when discharge was first generated. Soil was not even at field capacity when field discharge was initiated in October, thus the threshold for runoff generation was the filling of interped spaces rather than satisfaction of total soil moisture deficit. On cessation of drainflow soil moisture contents were higher in March than at the start of flow in October demonstrating hysteretic soil water behaviour.

Throughout the duration of the drainage season and the lysimeter experiment, discrete wet and dry zones were simultaneously identified. When considered with the hysteretic behaviour reported above, this observation led to the suggestion of two runoff generation mechanisms on plots and the lysimeter (e.g. Sections 6.2.4 and 7.2.6):

- macropore flow conveyed the bulk of water to depth and allowed local wetting;
- preferential flow in finger-like zones occurred (Tamai *et al.* 1987; Steenhuis and Parlange, 1988; Roth *et al.* 1991).

Runoff generation mechanisms were elucidated through a tracer experiment (Chapter 7) in which isotopic and chemical determinations of runoff and matric water were coupled with hydrometric and tensiometric observations. The tracer experiment substantiated the theory that macropore flow was responsible for substantial runoff generation. Macropore flow almost definitely operated on the lysimeter as the mole slot and fissures were still open.

However, three observations during the lysimeter experiment showed that 'classic' bypass macropore flow could not be accepted as the only significant runoff generation mechanism:

1. isotopic analysis indicated that event water travelling rapidly through fissures and mixed with old water; labelled water concentrations continued to be highest at peak drain discharge during subsequent irrigation;
2. widespread occurrence of tracers in the matrix was observed;
3. no wetting front or water table was identified, but discrete, wet pockets of soil were monitored and responded relatively quickly to irrigation.

These results did not contradict any field or plot observations.

If 'classic' bypass flow was the dominant runoff generation mechanism, irrigation would not result in substantial alteration of matric potential; water would reach the drain with negligible isotopic or chemical alteration and matric water composition would be little changed (Section 1.7). Thus, movement of water along macropores was observed to occur at Rowden but the process was not as staightforward as many workers would suggest (McDonnell, 1990; Beven 1991. Fissure walls were a source of old water for the first macropore flow. As flow became established, macropore walls wetted and the matrix and macropores became connected, allowing rapid mixing of old and new water (Youngs and Leeds-Harrison, 1990; Johnson *et al.*, 1993). This was a source of labelled water during subsequent irrigations.

A second significant runoff mechanism was a form of discontinuous translatory flow through wetted zones. The heterogeneous physical and hydrological nature of the soil was consistently noted. Hydraulic variability within profiles was reinforced as wetter soil has a higher conductivity and provides a preferential flowpath. A positive feedback mechanism is established. Discrete wetted zones received preferential flow until their storage capacities were overcome and they in turn generated discharge. These areas then became sources of matric flow for other wetted zones (Figure 7.9). Therefore, rapid hydrological response to irrigation in a few apparently discrete volumes of soil resulted in mixing of new and old water and runoff generation. These wetted areas were also capable of generating the observed delayed flow, as water continued to drain from these pockets (for up to 12 hours). Labelled water was concentrated in the matrix at sites where it would be available for transport during subsequent drainflow generation events. Thus, it is proposed that a continuum of mechanisms operate through a wide range of pore sizes and that pathways are far from discrete.

Monitoring of matric potential, soil moisture content and water composition failed to identify continuity of results (i.e. gradients) between sampling points within horizons and profiles. The degree of variability in this structured soil means that it may be more useful to characterise the variability and attempt to identify trends or thresholds than to compartmentalise the profile and active water pathways. Interpolation of nets or flowpaths between sampling points which are 0.5 m apart in horizons and 0.1 to 0.2 m apart vertically may be inappropriate.

An understanding of the nature of field hydrology is a prerequisite for the explanation of catchment hydrology and streamwater quality. The implications of hydrological observations for solute transport are discussed below.

8.3 Implications for solute transport.

This study has shown that macropore flow cannot be regarded as a 'bypass' mechanism, in which the matrix and water contained therein play a relatively passive role. Water which moves through macropores is intimately associated with that in the surrounding matrix. Further, even in mole drained, heavy soils such as the Hallsworth series, the matrix may also provide the 'framework' or 'habitat' for rapid runoff generation, if antecedent soil moisture and hydraulic conductivity are sufficiently high. While mole drainage at Rowden has greatly increased artificial macroporosity this has also led to structural amelioration of the matrix (Section 4.2.5). So water is diverted from overland

and near surface routes to subsurface pathways and matric wetting is encouraged.

There are strong relationships between land management practices, the abundance of macropores and the infiltration rates of soils (Edwards *et al.*, 1984) and many workers have found that macropores represent a major pathway for solute movement (Shufort *et al.*, 1977; Tyler and Thomas, 1977; Schaffer *et al.*, 1979), including rapid transport of surface applied nitrate to drainage systems (Smettem *et al.*, 1983). Results from the isotopic tracer experiment indicate that previous assumptions with regard to nitrate location (Hallard, 1988) were simplistic. Hallard argued that the role of macropores in the leaching process largely depended on the source of nitrate. If nitrate was easily contacted by rainfall and infiltrating water before macropore flow became operational then macropores would transmit both water and solute more rapidly to depth than uniform displacement approaches would predict (Smettem *et al.*, 1983). Barraclough *et al.* (1983) and Hallard (1988) believed that if nitrate was relocated within soil peds then macropore flow would produce inefficient leaching. Both contact with the nitrate source and residence time would be insufficient for runoff to equilibrate with nitrate in the soil profile. However, it is the drained fields, where rapid macropore routing of water to the drains is so crucial to the successful operation of the drainage scheme, that experience greatest nitrate leaching losses. The lysimeter experiment conclusively demonstrated that a high degree of matrix-macropore interaction occurred and that the peds and macropores should not be viewed as two separate domains. On the basis of the tracer results (Chapter 7), maximum nitrate concentrations which occur at hydrograph peakflow at Rowden (Scholefield *et al.* 1995) do not represent unbuffered contributions from the soil surface but maximum mixing with the matrix at a time when a larger proportion of the soil contributes runoff.

This study emphasises the need for more detailed and controlled comprehensive experiments in order to improve our understanding of field hydrology and nitrogen flows on grassland. At the catchment scale, the significance of structure for nitrate leaching was demonstrated by Scholefield and Rodda (1994). They found that prediction of nitrate leaching at this scale was most successful when soil structure was taken into account.

However, the extent of interaction between matrix and macropores is controlled by soil hydrological conditions and in turn determines the water quality and must be observed and explained at the micro scale.

The foregoing discussion has wider implications. Kirkby (1978) noted that soil and hillslope hydrological processes were frequently inferred from observation of hydrograph characteristics and streamwater quality. Deduction of hillslope runoff generation mechanisms using streamwater quality data assumes that water moves through discrete pathways and retains a distinct chemical and isotopic character. This assumption remains central to more recent work, for example, the widely employed technique of hydrograph separation (Beven, 1991; McDonnell, 1991).

The tracer experiment on the drained lysimeter (Chapter 7) was designed to examine whether water from different events remained discrete. The lysimeter experiment conclusively demonstrated that the peds and macropores should not be viewed as two separate domains. It may now be appropriate to move away from the compartmentalist perspective and recognise the continuum of flowpaths within soil profiles and on hillslopes. Hooper *et al.* (1990) advocate an approach whereby representative mixing zones are identified. Recognition of interaction is fundamental. In each zone hydrological and chemical mechanisms are defined as 'critical' in terms of catchment runoff. The identification of an appropriate scale of investigation will be an important element of such work if it is to be successful.

8.4 Future research requirements.

This study involving the comparison of hydrological response characteristics of undrained and drained fields has identified several aspects of mole drainage scheme hydrological performance which require further attention. These are all concerned with the routes and rates of water movement to the mole and lateral drain systems and are likely to have important implications for the overall impact of the drainage system on both downstream hydrology and nitrate leaching.

Macropores have been identified by previous studies as an important pathway of water movement (Beven and Germann, 1982). A similar mechanism to that proposed by McDonnell (1988), combining macropore flow with old water displacement (see Chapter 7), is propounded. However, McDonnell's supposition was that macropores provided routes for new water transport and topographically wetter hillslope zones were sites of old water displacement. This study showed that there were inextricable links between the matrix and macropores. Thus water moving through the matrix or macropores cannot easily be classified as new or old water and the classic two-domain approach to water flow in macroporous soils must be modified for Rowden.

Water within the matrix and that conveyed by macropores are not separate entities. There is physical, chemical and isotopic mixing/diffusion of 'different bodies' of water, even when most water moves rapidly through cracks and fissures (i.e. classic macropore flow). 'Bypass flow' is, therefore, a misleading term. Water does not 'bypass' the soil, but interacts with soil and soil water, even when transmitted rapidly, and over short distances. This is an obvious area for future work. More detailed investigations of the role of different pore sizes in the transmission of water under different hydraulic conditions could further elucidate the mechanisms of runoff generation, and the significance for solute movement.

At Rowden only a fraction of the subsurface discharge from drained fields actually moves through the soil via mole drain system (Section 6.2.1). Further research is required to clarify these findings and to determine more precisely how the remainder of the water moves through the soil to the lateral drain network. A characterisation of the zones of preferential matric flow may be useful. Physical soil characteristics, such as texture and total porosity and connectivity, determine moisture content and matric potential. Soil structure, which integrates a number of other soil properties, determines the direction and rate of water movement. Identification of the nature and thresholds of controlling soil properties may be an appropriate focus for future work.

Finally, the possibility of identifying one component of soil hydrologic response as an indicator of field runoff generation could be investigated. This would be especially useful for distributed models such as S.H.E. (Bathurst, 1986; Ragab and Cooper, 1992; Reeves and Beven, 1992) if one value could describe a complex heterogeneous catchment area.

APPENDIX 1 Conversion of A20 values to Andreasan pipette equivalent.
Dept. Geographical Sciences, University of Plymouth.

It can be argued that no one technique gives the absolute correct answer when it comes to particle size analysis. This is particularly true when assessing the amount of clay in a sample. However, the Andreasan pipette technique (British Standard 1377, 1992) is still widely accepted as the yardstick, especially when comparing results from other workers.

The Fritsch Analysette 20 always under estimates the amount of clay in a sample. This it does consistently and a relationship (figure 1) has been developed between the two techniques to enable calculation of the pipette equivalent.

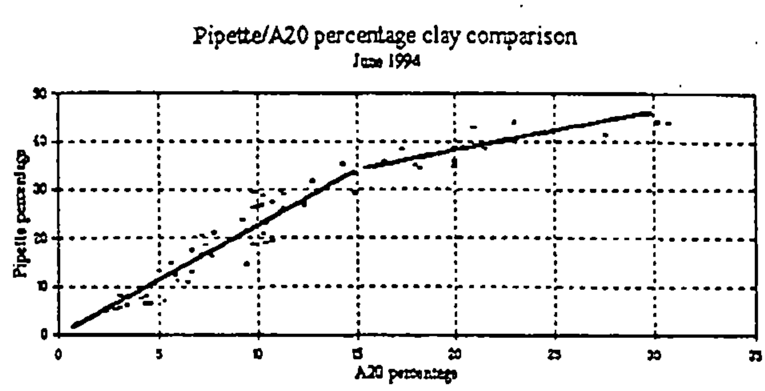

Figure 1

From the graph it will be apparent that two relationships exist; one for A20 values *less* than 15 percent, and another relationship for values greater than 15 percent. The degree of correlation for values less than 15 percent is 0.95, and for greater than 15 percent it is 0.723. These two relationships can be expressed as regression equations.

<u>For A20 values *less* than 15 percent:</u>

pipette equivalent = -1.3825 + 2.4515 x A20 clay percentage

<u>For A20 values *greater* than 15 percent:</u>

pipette equivalent = 23.744 + 0.7224 x A20 clay percentage

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