Post Drought Soil Water Recharge

A Study of the Processes of Recharge

Institute of Hydrology

National Groundwater & Contaminated
Land Centre Project NC/06/12

Post Drought Soil Water Recharge

A Study of the Processes of Recharge

J Finch

Research Contractor:
Institute of Hydrology

Environment Agency
Rivers House
Waterside Drive
Aztec West
Bristol
BS12 4UD

National Groundwater & Contaminated Land Centre Project
NC/06/12

Publishing Organisation:
Environment Agency
Rio House
Waterside Drive
Aztec West
Almondsbury
Bristol BS12 4UD
Tel: 01454 624400 Fax: 01454 624409

Dissemination status
Internal: Released to Regions
External: Released to Public Domain

Statement of use
This report summarises the findings of research carried out in the Pang and Tern catchment into the changes in soil water at the end of a dry summer and the implications for estimating groundwater recharge using simple water balance models. The information within this document is for use by EA staff and others involved in managing water resources.

Research contractor
This document was produced under a National Groundwater & Contaminated Land Centre Project by:

Institute of Hydrology
Maclean Building
Crowmarsh Gifford
Wallingford
Oxon
OX10 8BB

Tel: 01491 838800 Fax: 01491 692424

Environment Agency's Project Manager
The Environment Agency's Project Manager for this project was:
Ms Sarah Evers - Environment Agency, National Groundwater & Contaminated Land Centre

National Groundwater & Contaminated Land Centre Project NC/06/12

EXECUTIVE SUMMARY

Recent studies have shown that when the models currently used to estimate potential direct groundwater recharge correctly estimate the magnitude of the soil moisture deficits, they consistently show the onset of drainage downward from the soil profile earlier than is observed, resulting in an overestimation of recharge. Three key questions need to be investigated in order to understand the causes of this discrepancy and correct it:

1. What is the depth of soil from which evaporation takes place and what is the soil moisture deficit above which actual evaporation falls below potential?
2. How quickly does evaporation return to potential following rainfall after a prolonged period of stress?
3. Is preferential flow through macro-pores or fissures bypassing the soil moisture store?

In order to provide answers to these questions, this study was set up with the following objectives:

- To collect a detailed data set of changes in soil moisture that occur at the end of a prolonged dry summer as the soils wet up in the autumn;
- To analyse these data to achieve a conceptual model of the processes involved in the wetting up;
- To review how well the changes in soil moisture can be duplicated by simple soil models and to suggest how these could be improved;
- To test the impact of spatial heterogeneity on spatially distributed estimates of recharge.

Soil moisture has been the key element of this study because drainage, and thus recharge, cannot generally take place whilst a soil moisture deficit exists. Hence it is important that the water balance models, used to estimate direct groundwater recharge, are able to predict the duration (and magnitude) of deficits accurately. Thus, this project has made measurements of soil moisture and tested models against these data.

The first objective was achieved by a programme of field measurements carried out in the Pang and Tern catchments during the autumn and winter of 1997. These catchments were selected because they were sited on the two most important UK aquifers, the Chalk and Permo-Triassic sandstones respectively. In addition some suitable measurements (soil water content, soil water potentials, groundwater levels, actual evaporation) had already been made in these catchments. Measurements of soil moisture were made, using neutron probes, in 10 access tubes in the Pang catchment and at 9 sites in the Tern. Measurements of evaporation were made at a site in each catchment.

The second objective was achieved when the data were analyzed to develop conceptual models of the processes involved in the water balance at the sites. These showed that soil moisture deficits developed to depths significantly greater than the anticipated maximum rooting depth. The effect was especially marked on the Chalk soils where the deficits extended to depths in excess of 3m.

Evaporation from grass was limited by soil moisture availability once the deficits exceeded 70 to 100 mm. The measurements of evaporation showed that there were no indications of a lag between significant rainfall and an increase in evaporation.

No evidence could be found for bypass flow occurring. Runoff and interflow were significant at the sites on low permeability substrates and can be generated when there is a soil moisture deficit present when particularly heavy rainfall occurs, possibly over several days.

The study has confirmed the importance of taking into account interception losses for woodland by demonstrating that the soil moisture deficits developed under woodland can not be explained by evaporation through the leaves alone.

The third objective was achieved by making a comparison between the ability of the FRL and MORECS soil water models to simulate the observed soil moisture deficits and the implications for estimating groundwater recharge were estimated. Important differences have been highlighted. Generally, the multi-layered formulation of the FRL model gave it an advantage over the single layer MORECS soil water model. This was particularly relevant in relation to simulating growing crops and the transition from a vegetated surface to bare soil at harvest. This is likely to be important as a significant part of the aquifer outcrops is cultivated with annual crops. The FRL model also has an advantage in simulating the development of deficits beneath the rooting zone. The difference between estimates of mean annual groundwater recharge by FRL and MORECS soil water models varied between 5 and 10% according to the combination of soil class and land cover type. Changing the depth of the rooting zone and the fractional available water content in both models resulted in differences in the estimated mean annual groundwater recharge of up to 15%.

Both models share the limitations of not considering runoff or interflow with the result that, when these processes are important, they will seriously over estimate recharge if used without further development. They also both assume that water content greater than field capacity drains out within a day, which is only true of freely draining soils. There is uncertainty in how to represent evaporation from bare soil and what values should be used as little research has been carried out on this subject.

It is noted that Chalk soils are exceptional because rainfall can percolate through the soil profile in the presence of a soil moisture deficit, with the result that drainage, and hence recharge, can occur potentially in the summer and autumn.

Heterogeneity within a unit mapped as a single soil series in these trials resulted in variations in mean annual groundwater recharge of at least 5%. When considering the spatial representation of different soil types, the choice of aggregation scheme is crucial because an inappropriate choice can result in differences in annual groundwater recharge of at least 5%.

The following recommendations are made:

- A groundwater recharge model should include a multi-layered soil water model.

- The process of water being drawn up from beneath the rooting depth should be explicitly represented in the soil water model.
- Release of water to drainage over periods greater than one day should be incorporated in the soil water model.
- A runoff model should be incorporated in a groundwater recharge model.
- Partition of soil drainage between recharge and interflow should be incorporated in a groundwater recharge model (this will involve consideration of the soil and substrate hydraulic properties and the slope of the land surface).
- A canopy interception model should be included for deciduous and coniferous woodland.
- Research should be carried out into an appropriate model formulation to represent the process of rainfall percolating through the Chalk soil profile in the presence of soil moisture deficits.
- Research should be carried out to determine an appropriate model formulation and the values of parameters to represent evaporation from bare soil.
- Research should be carried out into the impact of deciduous woodland on recharge for a variety of soil types. This should include the potential changes in soil hydraulic properties, the rooting depth and an appropriate model representation.

CONTENTS

LIST OF FIGURES

LIST OF PLATES

LIST OF TABLES

1. INTRODUCTION

Recent studies, (Finch et al. 1995), have shown that when the models currently used to estimate potential direct groundwater recharge use the correct soil hydraulic properties, and thus correctly estimate the magnitude of the soil moisture deficits, they consistently show the onset of drainage downward from the soil profile earlier than is observed, resulting in an overestimation of recharge. This effect is especially marked on Chalk soils. The models assume that recharge occurs as drainage downwards out of the soil profile once the water content of the soil exceeds a particular value (field capacity), i.e. when a soil moisture deficit does not exist. Thus it is crucial that these models predict the soil water content, particularly the periods when drainage downward does occur, with sufficient accuracy for reliable estimates of groundwater recharge.

The implication of the models predicting the early end of soil moisture deficits is that the models are 'losing' water, for which there are two possible explanations:

1. the actual evaporation is being underestimated by the models and hence they underestimate the soil moisture deficits
2. water is migrating downwards from the soil layer, as bypass flow or gravity drainage, and is not recognised by the model.

Three key questions need to be answered:

1. What is the depth of soil from which evaporation takes place and what is the soil moisture content above which actual evaporation falls below potential? These values play a significant part in determining the maximum soil moisture deficit that can develop in any given combination of land cover and soil type prior to wetting up.
2. How quickly does evaporation return to potential following rainfall after a prolonged period of stress? Grass responds to severe water stress by the leaves dying and so it probably takes some time for new leaves to grow before evaporation can return to the 'normal' where as needliferous trees are capable of responding more rapidly.
3. Is preferential flow through macro-pores or fissures bypassing the soil moisture store? There is a growing volume of literature which shows that water can drain through the soil layer whilst there is a soil moisture deficit. However, it is not clear under what circumstances this actually occurs.

In order to provide answers to these questions, this study was set up with the following objectives:

- To collect a detailed data set of changes in soil moisture that occur at the end of a prolonged dry summer as the soils wet up in the autumn.
- To analyse these data to achieve a conceptual model of the processes involved in the wetting up.
- To review how well the changes in soil moisture can be duplicated by simple soil models and to suggest how these could be improved.

Chapter 2 gives a brief introduction to the water balance and defines the terms used. The role of evaporation in the energy balance is described in chapter 3, as is the effect of soil moisture on limiting evaporation. The methods used to measure evaporation and soil moisture are outlined in chapter 4 whilst the field measurements programme is described in chapter 5. The measurements are described and interpreted in chapter 6 and a conceptual model of the processes involved in the water balance is developed in chapter 7. This chapter includes a comparison between the conceptual model developed and those of the MORECS and FRL soil water models. In chapter 8 the results of tests of the MORECS and FRL soil water models against the data are described and these are discussed in the light of the conceptual model. The issue of spatial heterogeneity is discussed in chapter 9 and conclusions and recommendations are made in chapter 10.

2. RECHARGE AND THE WATER BALANCE

Simple water balance models, such as those described by Grindley (1969) and Hough and Jones (1998) have been widely used for estimating groundwater recharge in the UK. These models combine a model of evaporation with a model of soil moisture. This approach has the advantage that it can be used to provide time series estimates of direct groundwater recharge from readily available meteorological data and that areal estimates of recharge can be produced from these estimates by applying simple procedures for spatial aggregation, either directly (Rushton and Ward, 1979) or as part of a full catchment water balance model (Wilby et al. 1994). An additional advantage is that, as the models are causal, they can be used to predict groundwater recharge resulting from different scenarios, e.g. impact of land use change.

Simple water balance models calculate direct recharge using a water balance equation which is of the form:

$$P = E + I + R_o + R_i + D + \Delta S$$

where:

P	rainfall (mm)
E	actual evaporation (mm)
I	interception loss (mm)
R_o	surface runoff (mm)
R_i	interflow (mm)
D	bypass flow (mm)
ΔS	change in soil water content (mm)

Recharge is the sum of the bypass flow and positive values of the change in soil water content, once the soil water content exceeds the water content of the soil at which drainage can occur. Negative values of the change in soil water content represent an increase in soil moisture deficit. It is implicit in this statement that there is no upward movement of water from groundwater to the soil store, i.e. the water table is at depth.

It is important that these models should represent the processes involved to an acceptable level of accuracy because they estimate direct groundwater recharge by subtracting 'losses', such as evaporation, from rainfall. In addition, the water content of the soil is derived by keeping a running 'balance' with the result that the models are only initialised in the winter when the soil water content returns to field capacity.

The models generally use a daily time step because meteorological data is readily available as daily values. In addition Howard and Lloyd (1979) demonstrated that large errors can occur unless the accounting period for the soil water balance is less than 10 days because short periods of infiltration are often masked by the averaging effect of input data with longer increments.

The components of the water balance are shown schematically in Figure 2.1. Actual evaporation, E, is water transferred from the soil to the atmosphere by direct evaporation from the soil pore space and soil water drawn into plants via the roots and transpired through the leaves. Evaporation from the soil pore space is not a significant factor if the soil is totally covered by a vegetation canopy, e.g. permanent pasture. This is because the downwelling radiation (solar and thermal radiation striking the land surface), which is the primary source of energy for evaporation, does not impinge on the soil.

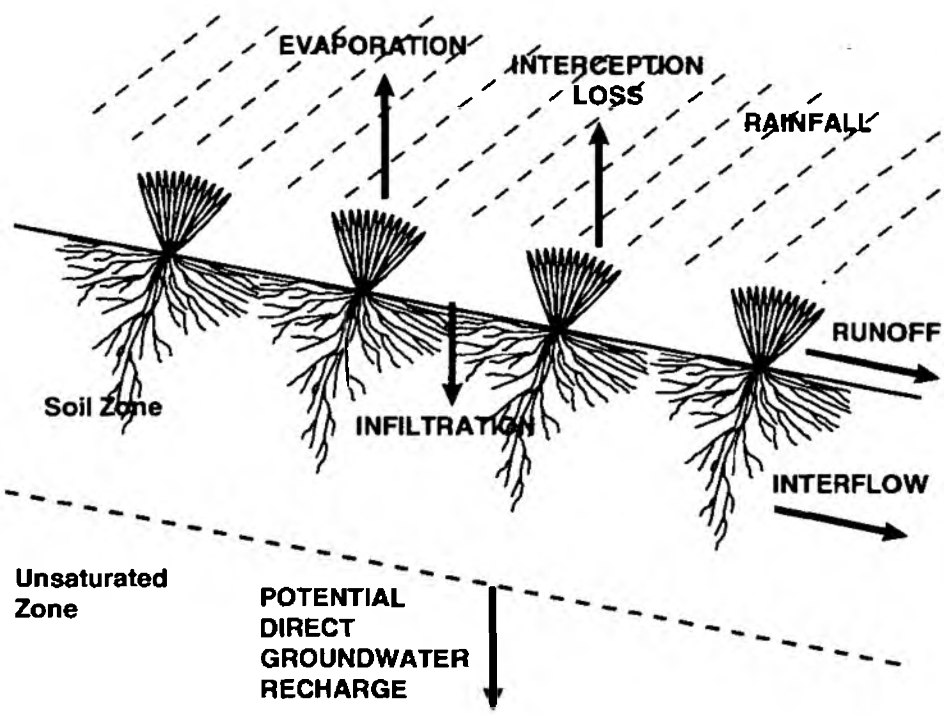


Figure 2.1 Schematic representation of the water balance

Interception loss is water evaporated directly from the surface of vegetation during and directly after rainfall.

Runoff is used here to mean the movement of water on the land surface, i.e. overland flow. Models of direct groundwater recharge generally assume that this water is not available for recharge, i.e. it enters a surface water course. Interflow is water that moves laterally through the soil profile and eventually enters a surface water course. In some models these are combined into a single term.

Bypass flow is rainfall that reaches the soil surface and is then conveyed into the unsaturated zone without entering the soil store and thus is not available to plants. It should be noted that the definition of the soil used by a hydrologist is different to that of a soil scientist. A hydrologist defines the soil as the zone, extending down from the surface, within which water can be lost to evaporation, i.e. the zone in which soil water deficits can develop. A soil scientist, on the other hand, defines a soil in terms of properties such as mineralogical composition etc. and thus the soil depth is the depth to unweathered bedrock. In the case of the soils developed directly in the Chalk, these differences in definition result in a soil scientist defining a shallow soil whilst a hydrologist will define it as a deep soil.

The soil water store is a very important component in the water balance because it's properties determine the partition of the precipitation that has infiltrated between evaporation, runoff, interflow and recharge. It is for this reason that correctly understanding the processes involved and being able to simulate the changes in soil water content are considered essential in order to produce reliable estimates of groundwater recharge.

The water balance is concerned with the conservation of mass. However, one of the terms, evaporation, can also be considered within the context of conservation of energy.

3. EVAPORATION AND THE ENERGY BALANCE

3.1 The energy balance

Much of the following is a summary of the description given by Shuttleworth (1993). The energy balance at the land surface takes the form of:

$$R_n = \lambda E + H - G$$

To be rigorous, there are other terms that define the energy temporarily 'stored', energy absorbed by biochemical processes and loss of energy associated with horizontal air movements but these are sufficiently small to be ignored for most practical purposes.

The net radiation, R_n, is the net input radiation at the surface, i.e. the difference between the incoming and reflected solar radiation, plus the difference between the incoming long-wave (thermal) and outgoing long-wave radiation.

The latent heat flux, λE, is the energy used to convert water from the liquid phase, in the soil and vegetation, into the gaseous phase. It is the product of the mass of water evaporated, E, with the latent heat of vaporisation of water, λ.

A portion of the radiant energy input to the earth's surface is not used for evaporation but warms the atmosphere in contact with the ground and then moves upward. It is referred to as 'sensible' heat flux, H, because it changes air temperature, a property of the air that can be easily measured.

The soil heat flux, G, is the heat conduction into the soil. It varies on both a diurnal and annual cycle. With dense vegetation, little radiation reaches the ground and heat storage can often be neglected.

3.2 The Penman-Monteith model of evaporation

A description of many of the evaporation models used in hydrology can be found in Shuttleworth (1988). The discussion here refers to the Penman-Monteith model as it forms the basis of the MORECS system and has been used with the FRL soil model, (Finch et al. 1995).

The model of Penman (1948) is the original model of evaporation, based on the energy balance. However, it incorporates some empirical factors that are based on calibration against data from grass lysimeters in the south of England and so, strictly speaking, it is only valid for these conditions. It can only be used for estimating evaporation from other land cover types by using empirical 'crop factors'. Grindley (1967) coupled a very simple soil water model with the Penman model of evaporation in order to estimate soil moisture deficits.

The Penman-Monteith evaporation model (Monteith 1965) is a development of the Penman model and is gaining wide acceptance for estimating evaporation for operational hydrology because it combines a physically based approach with a pragmatic requirement for data. It is sometimes referred to as a 'big-leaf' model as it assumes that the overall effect of the whole canopy on the energy fluxes above the canopy can be approximated by assuming that all the elements that make up the vegetation are exposed to the same microclimate. It assumes that all the energy available for evaporation is accessible by the plant canopy, and water vapour diffuses first out of the leaves against the surface resistance, r_s, and then out into the atmosphere above against the aerodynamic resistance, r_a. Meanwhile, the sensible heat, which arises outside rather than inside the leaves, only has to diffuse upward against the aerodynamic resistance. These resistances are parameters that must be specified for each vegetation type but are independent of the soil type and climate. Thus the Penman-Monteith model is capable of estimating evaporation from different land cover types.

The FAO now recommend evaporation be calculated in the form given by Allen et al (1994) the model is:

$$\lambda E = \frac{\Delta(R_n - G) + 86.4\rho c_p \dfrac{(e_a - e_d)}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_s}{r_a}\right)}$$

where:

E	evaporation (kg m^{-2} d^{-1})
G	soil heat flux (MJ m^{-2} d^{-1})
R_n	net radiation flux density at the surface (MJ m^{-2} d^{-1})
c_p	specific heat of moist air (kJ kg^{-1} °C^{-1})
e_a	saturation vapour pressure (kPa)
e_d	saturation vapour pressure computed at dew point (kPa)
r_a	aerodynamic resistance (s m^{-1})
r_s	bulk surface resistance of the vegetation canopy (s m^{-1})
Δ	slope of the saturation vapour pressure curve (kPa °C^{-1})
λ	latent heat of vaporisation (MJ kg^{-1})
γ	psychrometric constant (kPa °C^{-1})
ρ	atmospheric density (kg m^{-3})

the aerodynamic resistance is calculated as:

$$r_a = \frac{\ln\left(\dfrac{z_h - d}{z_{om}}\right)\ln\left(\dfrac{z_h - d}{z_{oh}}\right)}{k^2 U_z}$$

where:

U_z	mean wind speed at height z (m s^{-1})
d	zero plane displacement of wind profile (m)
k	von Karman constant
z_h	height of air temperature and humidity measurements (m)
z_m	height of wind speed measurement (m)

z_{oh} roughness parameter for heat and water vapour (m)
z_{om} roughness parameter for momentum (m)

d is taken to be $2/3h_c$, where h_c is the mean vegetation height. The roughness lengths are as given by Monteith and Unsworth (1990):

$$z_{om} = 0.123h_c$$

$$z_{oh} = 0.1z_{om}$$

The Penman-Monteith model was originally conceived as requiring meteorological data at short time intervals, typically hourly, in order to reflect the diurnal cycle. However, it is now apparent, for example Finch and Harding (1998) and Stockle and Jara (1998), that reliable estimates of evaporation can be obtained by using the daily total (24 hour) net radiation and daily averages of the air temperature, wind speed and vapour pressure deficit.

Several terms are used in describing different evaporation rates. One of these involves the concept of a reference crop. It is defined as an idealised grass crop with a fixed crop height of 0.12 m and a surface resistance of 70 s m^{-1}. It is implicit in this definition that the evaporation is not limited by the availability of soil water. The evaporation from the reference crop is termed the reference transpiration (Cain et al. 1998). A term often found in the literature is potential evaporation. Unfortunately, this term has been applied to a variety of conditions with the result that there is no clear definition. In general, it can be taken that the term potential evaporation refers to evaporation rates for a specific land surface (e.g. open water, coniferous woodland etc.), a given evaporation model and that evaporation is not limited by the availability of soil water. Thus, it is important to understand the context in which this term is used. Finally, the terms evaporation or actual evaporation are taken as the 'true' evaporation from a given land surface and thus may be limited by soil water availability.

3.3 Soil water

Infiltration is the process of water entry into a soil from rainfall. Soil water movement is the process of water flow from one point to another within the soil. The two processes cannot be separated as the rate of infiltration is controlled by the rate of soil water movement below the surface and the soil water movement continues after an infiltration event, as the infiltrated water is redistributed. The soil water movement also controls the supply of water for plant uptake and for evaporation at the soil surface. The soil properties affecting soil water movement are hydraulic conductivity and water-retention characteristics. These soil hydraulic properties are closely related to the soils physical properties such as particle-size and morphology.

The water-retention characteristic of the soil describe the soil's ability to store and release water and is defined as the relationship between the soil water content and the matric potential (soil suction). Matric potential is a measure of the energy status of water in the soil and is a component of total soil water potential that includes gravitational and osmotic potentials.

The hydraulic conductivity is a measure of the ability to transmit water and depends upon both the properties of the soil and the water content. The hydraulic conductivity is a non-linear function of volumetric soil water content and varies with the soil texture. Runoff is generated when the rainfall rate exceeds the hydraulic conductivity of the soil at the surface. Soil water drainage can be considered as being split into vertical and horizontal components, recharge and interflow. The relative proportions are a function of both the slope of the land surface and the difference in hydraulic conductivity between the soil and the underlying substrate. In the case of direct recharge of aquifers there is usually little difference in the hydraulic conductivities and so interflow is the minor component.

A useful concept is given by Wellings and Bell (1982) which is of the 'zero flux plane' (ZFP). In the summer, upward movement of water in the soil occurs in response to evaporative demands. This creates an upward soil water potential gradient near the surface, while a downward potential gradient persists lower in the profile. Between these two zones is a plane where the gradient of total soil water potential is zero - the 'zero flux plane'. The rates of change of soil water above and below the ZFP represent evaporation and drainage respectively. In the autumn, as the evaporative demand decreases and becomes less than the input from rainfall, the ZFP is eliminated and drainage occurs throughout the profile. It should be noted that this concept implies that drainage from the soil profile occurs even when a soil moisture deficit exists.

3.4 The effects of soil water on evaporation

Soil saturated by rain or irrigation first drains, i.e. potential recharge occurs, until the remaining water held by surface tension on the soil particles is in equilibrium with gravitational forces causing drainage. It is then defined as being at field capacity with a fractional water content of θ_f. The drying proceeds with little soil water restriction until the soil moisture falls to θ_d when the evaporation rate begins to fall. It continues to fall until a wilting point is reached, when the soil water content is θ_w, at which point the evaporation rate effectively ceases. These two soil water contents represent progression from the evaporation controlled by meteorological conditions to it being controlled by the soil. However the actual nature of the progression and the processes involved in evaporation at the wilting point are poorly understood. Soil moisture deficits are defined as relative to the soil water content at field capacity. However, it should be remembered that the field capacity is actually a point on the curve describing the water retention characteristics of a soil rather than a clear break when drainage ceases.

The amount of accessible soil water available to plants depends on their rooting depth, which can of course change as the vegetation grows, a point of particular significance for annual crops. This cannot be simulated by a soil water model consisting of a single layer, such as used in MORECS, but requires a multi-layer model.

Soil water models used for modelling the water balance, such as FRL and the soil model in MORECS (Finch et al. 1995), often work in terms of soil water content. However, plants are more sensitive to soil water potential and some of the differences reported in the literature in the functions linking the soil water content to reducing the evaporation may be caused by the variability in the water retention characteristic of different soils. In the Penman-Monteith model, to be strictly correct, the effect of soil water limiting evaporation should be applied to the surface resistance, r_s, as is done in MORECS. However, in a multi-layer soil model it is not clear how the soil water content of the various layers should be used to reduce evaporation and so a pragmatic solution is adopted in the FRL model which is that the potential evaporation, calculated using the Penman-Monteith model, is reduced to actual evaporation using a soil water stress term for each layer, determined by the water content of that layer.

3.5 Interception loss

The amount of rainfall actually reaching the ground surface, and thus infiltrating into the soil, is largely dependant upon the nature and the density of the vegetation cover. This cover intercepts part of the falling rain and temporarily stores it on its surface, from where the water is either evaporated back into the atmosphere, interception loss, or falls to the ground. The remainder of the falling rain misses the vegetation canopy and falls directly on the soil (free throughfall).

Interception by the canopy is normally greatest at the beginning of a storm because the dry surfaces of the vegetation prevent a large proportion of the rainfall from reaching the ground. As the leaves become wetter, the weight of water eventually overcomes the surface tension by which it is held and further additions from rainfall are almost entirely offset by water droplets falling from the edges of leaves. Even during rainfall, a significant amount of water may be lost by evaporation from the leaf surfaces with the result that there is some further retention of water to make good this evaporation loss. During prolonged rainfall, the interception loss may be closely related to the rate of evaporation so that meteorological factors affecting the latter become significant. Whilst rain is falling, the net radiation tends to be low due to the presence of clouds and hence the wind speed becomes the significant factor. Thus interception losses are higher from trees than short, uniform vegetation, such as grass or crops, because the trees have a greater surface roughness.

An example of a canopy rainfall interception model, comparable to the daily form of the Penman-Monteith evaporation model in its degree of rigour, is described by Gash et al. (1995) It is an analytical interception model which considers that for any storm there are a series of distinct phases, beginning with a period when the rainfall is less than a threshold value necessary to saturate the canopy. This is followed by a period of saturation and then a period of drying out after rainfall ceases. For a small storm, insufficient to saturate the canopy, the interception, I, is calculated as:

$$I = cP_G$$

where:

c canopy cover

P_G storm rainfall (mm)

For a storm which saturates the canopy the interception is:

$$I = cP'_G - cC_c + (c\overline{E}_c / \overline{P})(P_G - P'_G) + T$$

where:

C_c canopy capacity per unit area of cover (mm)

$\overline{E}_c$ mean evaporation rate during rainfall (mm d^{-1})

P'_G threshold rainfall necessary to saturate the canopy (mm)

$\overline{P}$ mean rainfall rate (mm d^{-1})

T evaporation from trunks (mm)

The evaporation from trunks for storms which saturate the trunks is:

$T = S_t + p_t P_G$

where:
S_t trunk storage capacity (mm)
p_t proportion of rainfall diverted to stemflow
and for storms which do not:

$T = C_c + p_t P_G$

It is assumed that only one storm occurs per day and so P_G is equal to the daily rainfall. The mean evaporation rate during rainfall can be calculated from the Penman-Monteith model, for days when significant rain fell, by setting the surface resistance, r_s, to zero. It should be noted that whilst interception losses occur, 'normal' evaporation from the leaves ceases as all the available energy is used to evaporate the water on the surface of the vegetation.

In MORECS a much simpler scheme is used whereby a proportion, f, of the rainfall is intercepted by the vegetation canopy:

$I = Pf$

the proportion is a function of the leaf area index (LAI), L, of the canopy:

$$f = 1 - (0.5)^L$$

The interception is constrained to not exceed the capacity of the canopy which is assumed to be 0.2 mm per unit LAI. This term ignores the evaporation of intercepted water whilst rain is falling and so it is multiplied by an additional factor, which varies between 1.0 in winter and 2.0 in summer, in an attempt to compensate for the greater evaporation rates of summer.

4. MEASUREMENT METHODS

4.1 Measuring evaporation

Two instruments, relying on different principles, have been used to measure evaporation during this project.

4.1.1 Eddy correlation

The IH Hydra, deployed in the Tern catchment (Plate 1), uses the eddy correlation technique (Shuttleworth et al. 1988).

Plate 1 The IH Hydra sited adjacent to the Bacon Hall soil moisture monitoring site in the Tern catchment

Near the surface, the mean wind motion is parallel to the ground, so that the mean vertical wind is zero. However, turbulent eddies within the body of the moving air cause fluctuating movements, both towards and away from the land surface. On average these fluctuations produce no net vertical movement of the air. There are similar turbulent fluctuations in the specific humidity of the air. If upward fluctuations are simultaneously moister then moister than air will be carried away from the ground and when they are both simultaneously downwards then drier than air will move toward the ground. These conditions occur during evaporation. When the vertical direction of wind and vapour pressure deficit movement are opposite then condensation occurs. Thus the evaporation rate can be determined by measurements of these vertical movements.

Many practical problems are involved in using the eddy correlation technique. The fluctuations in the wind speed, humidity and temperature can occur over a broad range of time scales and so the sensors used must have a rapid response. The sensors must be co-located so that they measure the same moving air, yet they should ideally have limited size and be carefully positioned so as not to interfere with the air movements they measure. Despite these difficulties, the eddy correlation technique is the preferred micro-meteorological technique because it is a direct measurement with minimum theoretical assumptions, in particular, a simple quality check can be made using the energy balance because the fluxes are measured independently.

The Hydra can provide routine evaporation measurements with an accuracy of between 5 and 10 percent. Care needs to be taken in selecting the site for the instrument to ensure that the wind is parallel to the ground and that the measurements are of the land surface of interest. This is achieved by ensuring that the instrument is sited on a relatively level site and that there are no objects significantly different in height, e.g. buildings, trees etc., nearer than about 300 m. Measurements can not be made when the instrument is wet, i.e. during rainfall or when dew is present, as this interferes with the vertical wind speed sensors on the instrument. This is not a major problem as the measurements made are the average fluxes at hourly intervals. This project is concerned with daily values and so it is not difficult to either delete or interpolate values when significant loss of data occurs due to this problem.

For this project, the hourly values recorded by the Hydra were aggregated to daily values using the method described by Finch and Harding (1998):

- all hourly values not coded as error free were rejected;
- when the latent heat flux measurement was missing it was set to zero when the net radiation was less than 10 W m^{-2} ;
- any days where all the flux values for more than half the daylight hours were missing were rejected;
- for the remaining days, where an hourly value of the latent heat flux was missing it was estimated from the hourly net radiation assuming the same Bowen ratio ($H/\lambda E$) as that of the next hour with measured fluxes;

Where a daily value of the latent heat flux was missing it was estimated from the daily net radiation by assuming the same Bowen ratio as the next day with a measured value so long as they were no more than five days apart.

4.1.2 Bowen ratio method

A Campbell Scientific Bowen Ratio Machine was deployed in the Pang catchment (Plate 2).

Plate 2 The Bowen ratio machine at the soil moisture monitoring site 10 in the Pang catchment

This method is also known as the energy balance method. The turbulent diffusion processes responsible for the transport of water vapour and sensible heat through the atmosphere are very similar, but they differ from those responsible for the transport of momentum. It is therefore plausible to assume that the aerodynamic resistance which restricts the flow of water vapour and relates that flow rate to the difference in vapour pressure, Δe, between two particular heights, is numerically equal to the resistance which relates the flow of sensible heat to the temperature difference, Δt, between the same two heights. Therefore, the Bowen ratio, β, which is the ratio of the sensible heat to the latent heat, is directly related to the ratio of the differences in temperature and humidity measured between any two heights. Thus, by measuring at the same time the net radiation and the soil heat flux, the evaporation rate can be calculated from:

$$E = \frac{R_n - G}{1 + \gamma(\Delta t / \Delta e)}$$

Measurements made are the time average values, typically over intervals of 20 minutes.

The method has the advantage that, by using the same instrument to measure humidity at both measurement levels, systematic offsets in calibration can be eliminated. A standard Bowen Ratio system, marketed by Campbell Scientific, was used in this study.

Similar care, to that for eddy correlation instruments, must be taken in choosing the site for Bowen ratio measurements, i.e. avoiding proximity to tall objects. The main problem with the method is that the sign of the sensible heat flux often changes in the evening and morning, resulting in the temperature difference being zero and thus measurements are inaccurate at this time. In addition, the combination of low values of the difference between net radiation and soil heat flux at night and the fact that the sensible and latent heat fluxes are commonly in opposite directions suggests that night time values are unreliable. However, this is not a serious problem for this project as it is concerned with daily values and the evaporation at night is small enough to be neglected. The Bowen ratio method can be used to determine evaporation to an accuracy of between 5 and 10 %.

Essentially the same method as used for Hydra data was employed to aggregate the 20 minute data to daily values

4.2 Measuring soil water

During this project, measurements of the soil water content have been made using a neutron probe, Plate 3. The following is taken from the detailed description given by Bell (1987).

The instrument consists of a probe containing essentially a radioactive source of fast neutrons and a slow neutron detector. This is connected via a cable to a pulse counter ('ratescaler'). The whole is contained within a transport shield. In use, the transport shield is fitted over the protruding upper end of an aluminium access tube which has been positioned vertically in the soil. The probe is lowered to successive measurement depths by means of the cable; a depth indicator and clamp mechanism operating on the cable are mounted in the transport shield. The fast neutrons emitted by the source collide with nuclei of atoms in the soil. It is collisions with the nuclei of hydrogen that are predominantly responsible for scattering and slowing these neutrons to produce a cloud of slow (thermal) neutrons that is sampled by the slow neutron detector. The electrical pulses from the detector as a result pass up the cable to the counter which displays the mean count rate which can be translated into soil water content using a calibration curve appropriate for the soil.

The method is subject to a variety of sources of error. One of these is the statistical counting error that arises from the randomness of radioactive decay in the source which generates the neutrons. This error is reduced to an acceptable level by counting for a sufficiently long time to achieve the minimum required precision. Another potential source of error is in the accuracy of the calibration curve used for the soil. This tends only to be a problem if absolute values of soil moisture are required. The slopes of the calibration curves are very similar and so changes in soil water with time, as was required in this project, can be determined very accurately. Slow ageing of the components of the probe are accounted for by regularly taking measurements in a laboratory standard, generally a drum full of water.

The volume of the soil to which the measurement applies varies according to the soil water content and ranges from about 0.15 m in wet soils to 0.3 m in dry soils. Thus readings are generally taken at intervals of either 0.1 or 0.2 m in depth. A further consequence of this is that it is difficult to obtain reliable readings in the top 0.2 m of the soil as there is a loss of both fast and slow neutrons into the air. Another potential source of error are discrepancies in the actual depth of the probe from the depth that it is assumed that the measurement is being made at. These are only significant in the presence of steep gradients in the soil water content.

With careful field practices it is possible to obtain measurements of soil water with an accuracy of 1% using a neutron probe.

Plate 3 Measuring soil water content in the Pang catchment using a neutron probe

5. FIELD MEASUREMENTS

The study has made use of data sets for two areas where measurements of soil moisture were already being made. These are the Pang catchment, representative of the Chalk aquifer, and the Tern catchment, representative of sandstone aquifers, Figure 5.1. Additional measurements were made in the catchments to supplement the existing data sets.

5.1 The Pang Catchment

The catchment of the River Pang is located on the Berkshire Downs and has an area of 175 km^2, an average annual rainfall of about 700 mm and annual runoff of 177 mm. The hydrogeology is principally unconfined Chalk with approximately 15% covered by impermeable deposits of clay and alluvium. The river has a baseflow index of 0.87 indicating that 87% of the runoff is derived from groundwater and that the river is in close hydraulic contact with the aquifer. The catchment is predominantly rural with agriculture being restricted to cereal crops and grassland. There are numerous abstraction boreholes, for municipal supply and irrigation, throughout the catchment.

IH installed neutron probe access tubes in this catchment during February 1997 for a project commissioned by the European Space Agency which finished in September 1997 . There are three sites, each with three access tubes and include land covers of grass, cereals and woodland, Table 5.1. The locations of the sites are shown in Figure 5.2. Weekly monitoring began at the beginning of February. An IH soil moisture station had also been installed at one of the grass sites, recording hourly values of rainfall, soil temperature, soil water content and potential. A Campbell Scientific Bowen Ratio System and a neutron access tube were installed at the beginning of July in a field of winter wheat in the north of the catchment, site 40.

For this project, tri-weekly measurements were made at the 10 neutron probe access tubes for three months, beginning in mid-September to give the increased temporal resolution of changes in soil moisture required to meet the objectives of the project. Soil moisture measurements have continued during 1998 at weekly intervals.

Table 5.1 Soil moisture measurement sites in the Pang catchment

Name	Easting	Northing	Soil Series	Land Cover
Site 10 tube 1	451030	181730	Andover	Winter barley
Site 10 tube 2	451320	181870	Andover	Winter barley
Site 10 tube 3	451750	182070	Andover	Winter barley
Site 20 tube 4	451610	179450	Hornbeam	Grass ley
Site 20 tube 5	451740	179380	Andover	Grass ley
Site 20 tube 6	451930	179220	Andover	Permanent grass
Site 30 tube 7	452140	172430	Wickham	Deciduous woodland
Site 30 tube 8	452190	172480	Wickham	Permanent grass
Site 30 tube 9	452180	172610	Wickham	Permanent grass
Site 40 tube 10	451680	184670	Andover	Winter wheat, barley

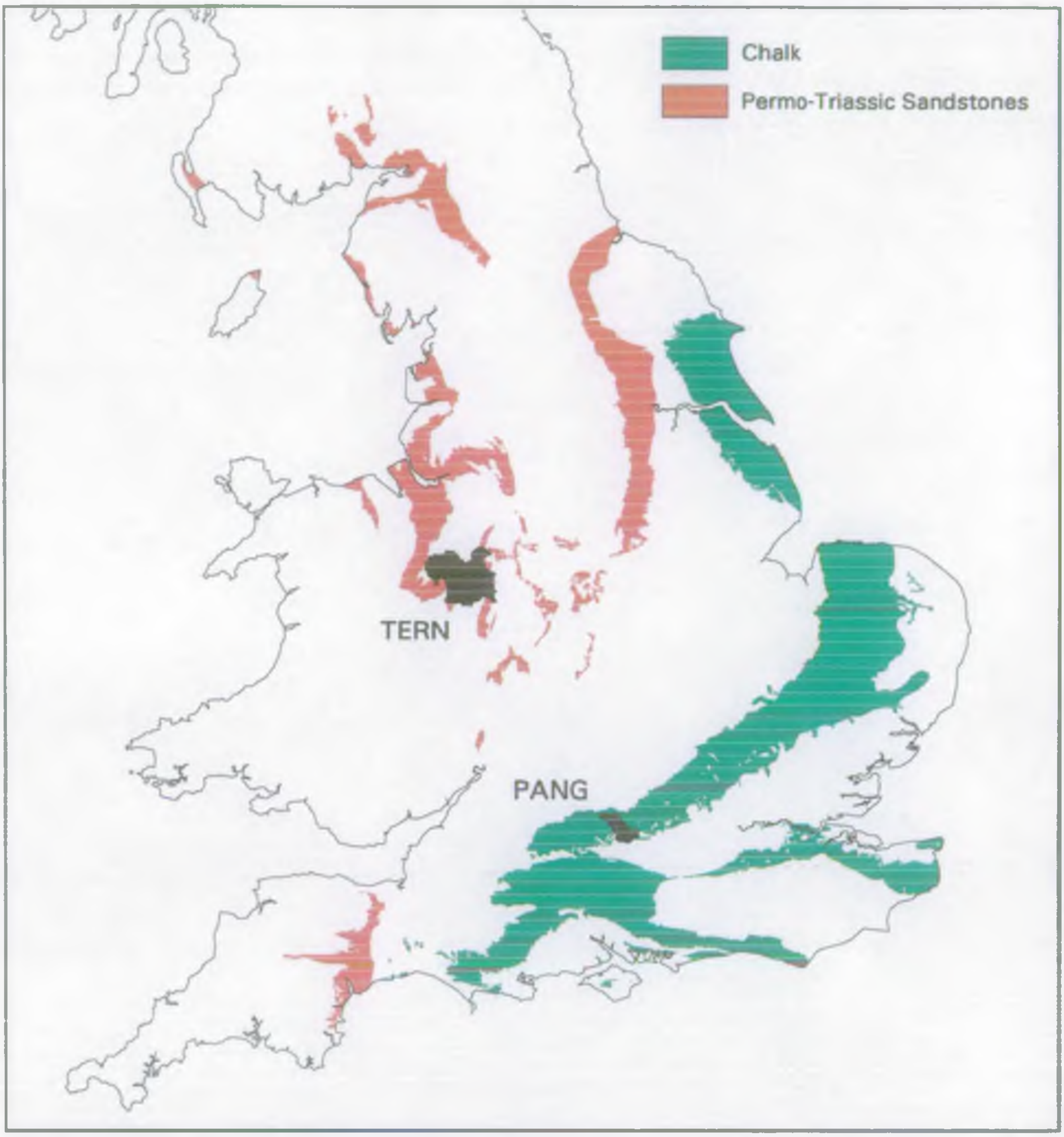


Figure 5.1 Locations of the Pang and Tern catchments

Rainfall measurements were available from a tipping bucket raingauge at a site near Compton (Grid reference 451440 179684). This record began in February 1998 and was merged with data from manual raingauges located at Compton to generate the time series required for modelling.

Additional data is available from the site of the Institute of Hydrology, a few miles north-east of the catchment. The measurements made there include rainfall, meteorological variables, soil moisture and actual evaporation at a site in a field of permanent grass.

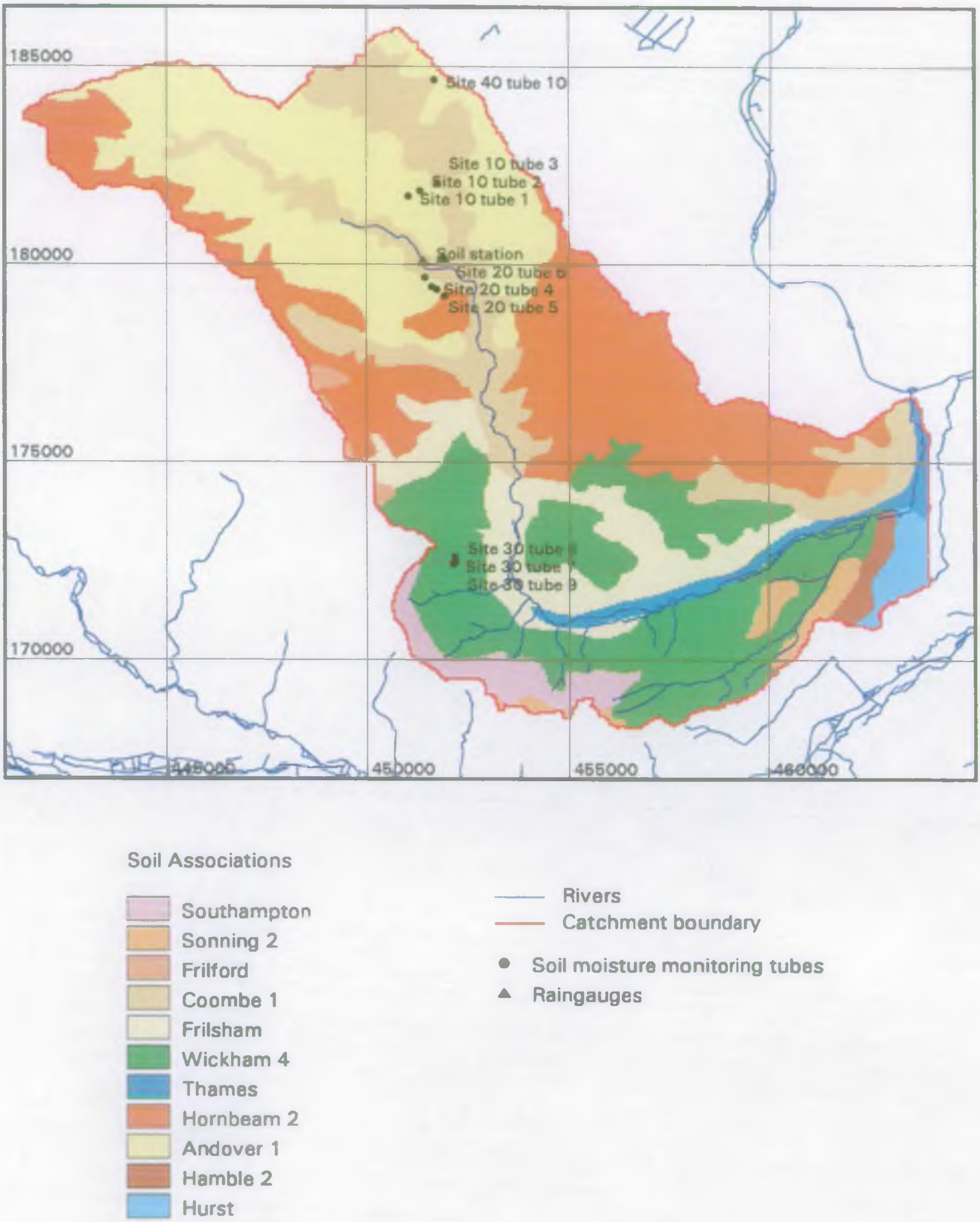


Figure 5.2 Location of soil moisture measurement sites in the Pang catchment

5.2 The Tern Catchment

The catchment of the river Tern is in Shropshire and has an area of 876 km^2. The average annual rainfall is about 700 mm and the runoff 256 mm. Regional groundwater abstraction occurs for public water supply and regulation of the river Severn. The catchment generally has a low relief and is predominantly agricultural. The geology is of mixed glacial deposits overlying Triassic series. The baseflow index of the river is 0.69 indicating that a significant fraction of the flow is derived from groundwater.

Soil moisture is monitored at thirteen sites in the Perry and Tern catchments by the Environment Agency at approximately fortnightly intervals. Most of these sites have two neutron probe access holes, a raingauge and a tubewell. In the past, measurements have also been made of soil water potential but these have been discontinued. Seven of the sites have a land cover of grass whilst three are heathland and three arable. The fortnightly interval between measurements is insufficient to give the detail required of the soil wetting during the autumn and so the frequency was increased to tri-weekly during the period from mid-September to mid-December 1997 at nine sites with permanent land covers in the Tern catchment, Table 5.2. The locations of the sites are shown in Figure 5.3.

Table 5.2 Soil moisture measurement sites in the Tern catchment

Name	Easting	Northing	Soil Series	Land Cover
Hodnet Heath Control	361890	326560	Crannymoor	heathland
Hodnet Heath Wet	361920	326240	Crannymoor	heathland
Hodnet Heath Deep Wet	361940	326240	Crannymoor	heathland
Bacon Hall	365280	323710	Newport	permanent grass
Windy Oak	360740	323040	Isleham	permanent grass
Bowling Green	362500	325600	Redlodge	permanent grass
Stoke Bridge	363380	327840	Hall	permanent grass
Heathbrook	363030	328830	Altcar	permanent grass
Shawbury Heath	353750	320180	Wick	pine plantation

There are no measurements of fluxes routinely made in the catchment and so an IH Hydra was deployed for a period of 6 months, starting in September 1997, to acquire measurements of actual evaporation at a grass field adjacent to the soil moisture measurement site at Bacon Hall. An Automatic Weather Station (AWS) was also deployed to give additional detail on the wind direction and independent measurements of the net radiation, rainfall and soil heat flux

Daily values of rainfall and meteorological variables were obtained from the Met. Office for the station at Shawbury.

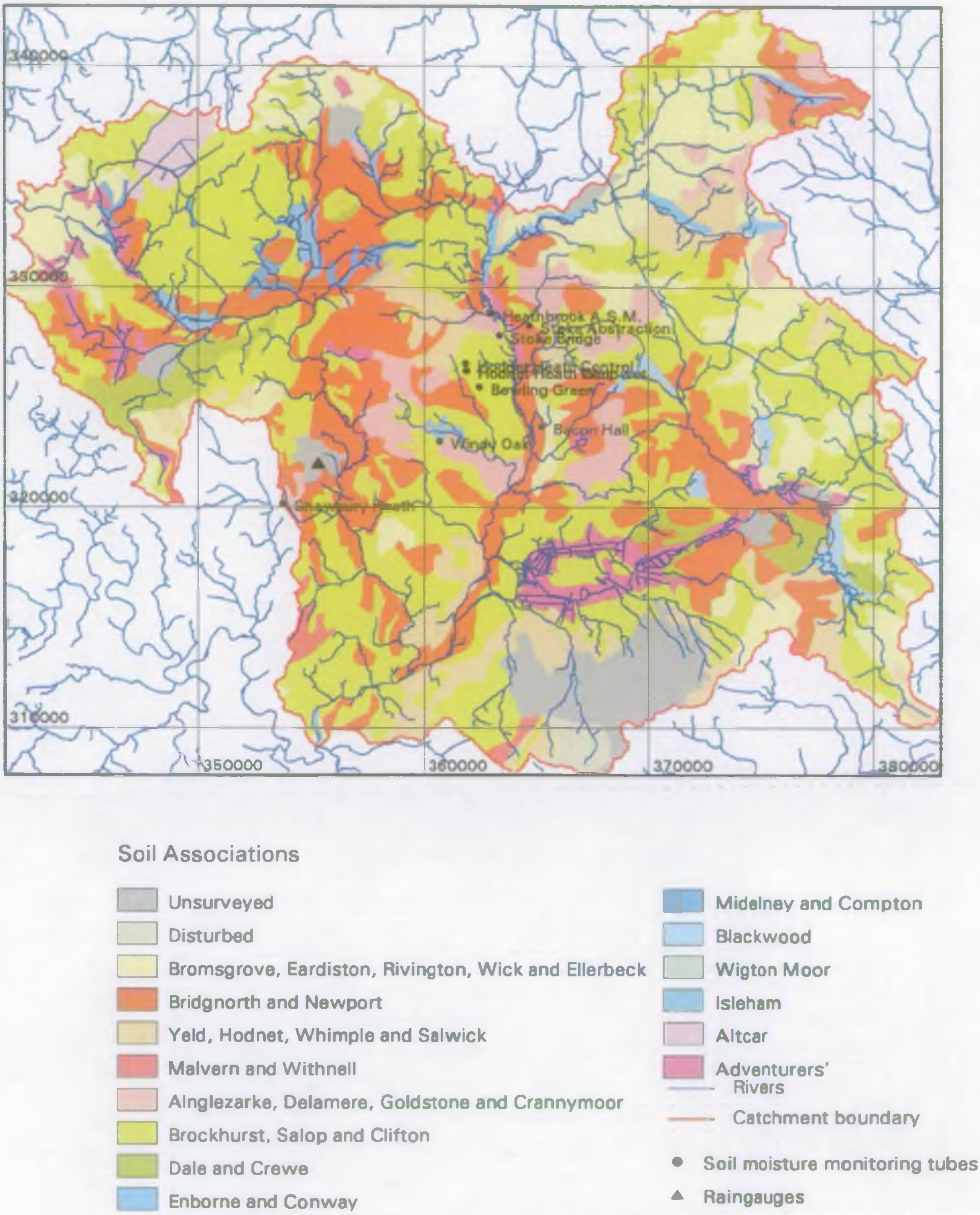


Figure 5.3 Location of soil moisture measurement sites in the Tern catchment

6. FIELD DATA ANALYSIS

This section describes the analysis and interpretation of the field measurement. Each site is analysed and the processes present at the site are highlighted. This is achieved by a schematic for each site which shows the processes present. Figure 6.1 shows the processes potentially present.

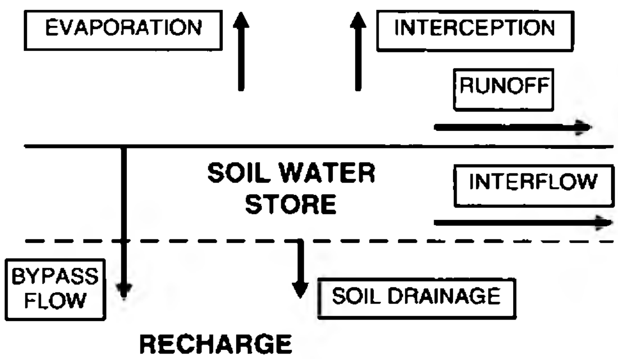


Figure 6.1 Schematic of water balance processes

6.1 Pang Catchment

6.1.1 Summary of results

- During 1997, soil moisture deficits developed to depths significantly greater than the anticipated maximum rooting depth in the Andover, Wickham and Hornbeam soils for a land cover of grass. The greatest depth was in the Andover series where the depth exceeded 3 m.
- Evaporation from grass was limited by lack of soil moisture once soil moisture deficits exceeded between 70 and 100 mm, depending on the soil type.
- There is no evidence for any lag in evaporation increasing following the wetting up of the upper part of the soil profile.
- There is no evidence for bypass flow occurring
- Runoff or interflow occur on the Wickham soil series but not at the sites on other soil series.
- Evaporation from winter wheat decreases significantly for a period prior to harvest.
- Deciduous trees can change the soil hydraulic properties of the Wickham soil series
- Evaporation from deciduous woodland during the summer is significantly higher than reference transpiration.

6.1.2 Meteorological conditions

The rainfall during 1997 was highly variable through the year. A dry January was followed by a very wet February with rain falling on almost every day. Very little rain fell during March and the first half of April but was followed by heavy rainfall through late April and the first half of May. Through the following summer there was a tendency for wet periods of 7 to 10 days to be followed by relatively dry periods of 10 to 20 days. Heavy rainfall characterised the last two months of the year and the first half of January 1998. The remainder of January and the rest of February were characterised by little rainfall.

6.1.3 Grass

Andover Series Soils

Two of the soil moisture monitoring access tubes, numbers five and six, are located in the Andover soil series, which is a thin soil directly overlying the Chalk. Tube six can be analysed with the most confidence as it is in a field of permanent grass which is grazed and thus represents the simplest situation of a permanent cover of short vegetation which approximates to the concept of a 'reference crop'.

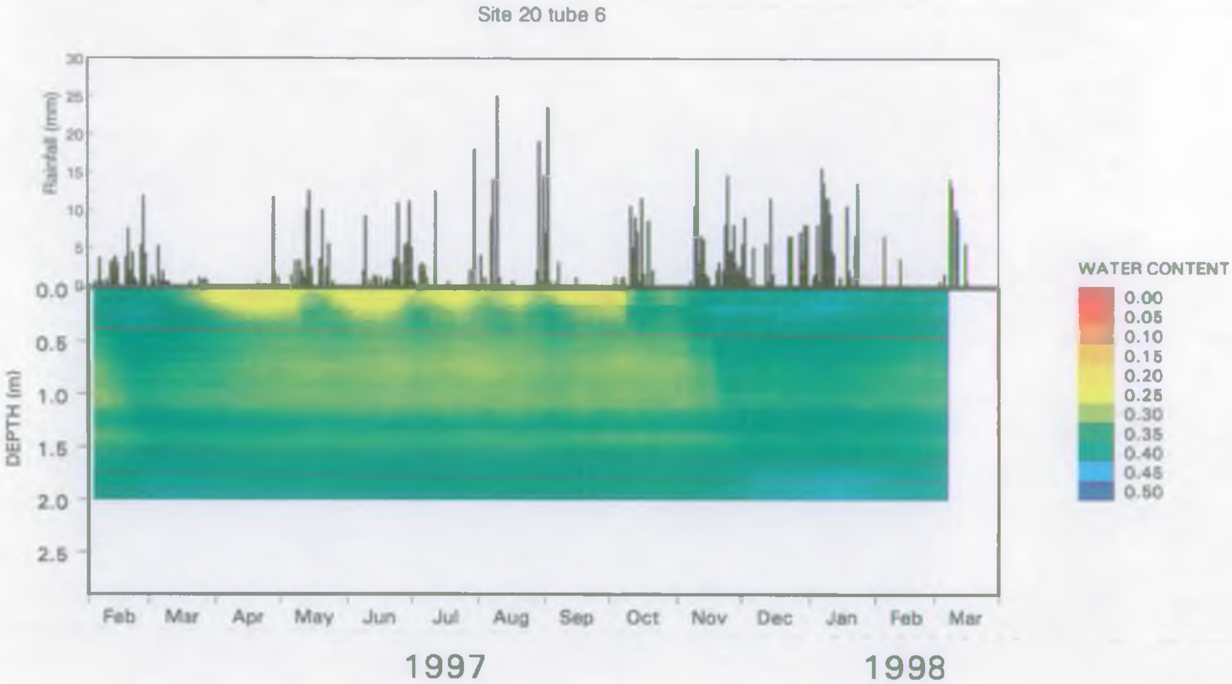


Figure 6.2 Changes in soil water content at tube 6

The variation in soil water content at tube 6 is shown in Figure 6.2. The soil profile probably did not return to field capacity during the winter of 1996/97. However the heavy rainfall during February 1997 ensured that it was only in the deeper part of the profile, probably below one metre, that a deficit continued. However, increases in water content are observed through the whole of the measured soil profile. During March and April, evaporation resulted in the progressive drying of the profile with decreases in water content occurring at all depths.

The heavy rainfall in May resulted in the top of the soil profile wetting up but there was no effect on the lower part of the profile. The changes in the soil water content can be attributed to evaporation and rainfall alone until the end of May. This is shown by the cumulative water balance, Figure 6.3. The cumulative change in soil water content can be assumed to be identical to the cumulative actual evaporation, provided soil water drainage does not occur. A sequence of the daily change in the soil water content was generated by linearly interpolating the water content, measured using the neutron probe, between dates when readings were taken and then subtracting the rainfall. The result shows excellent agreement with the reference transpiration.

There is a consistent difference between the two data sets in late-May to early-June. Measurements at Wallingford showed an identical effect which was attributed to the evaporation dropping below potential in response to the developing soil moisture deficit, (Finch and Harding, 1998).

Following heavy rainfall in mid-June, evaporation increased but a second period of evaporation below the reference value began in the first week of July and was sustained until the beginning of October. Heavy rainfall, together with decreasing evaporative demand as a result of decreasing net radiation, then restored the soil water content to a level that could support the full capacity for transpiration by the vegetation. An examination of the soil moisture data showed that the soil water contents when the actual evaporation began to fall below the reference transpiration were between 670 to 680 mm.

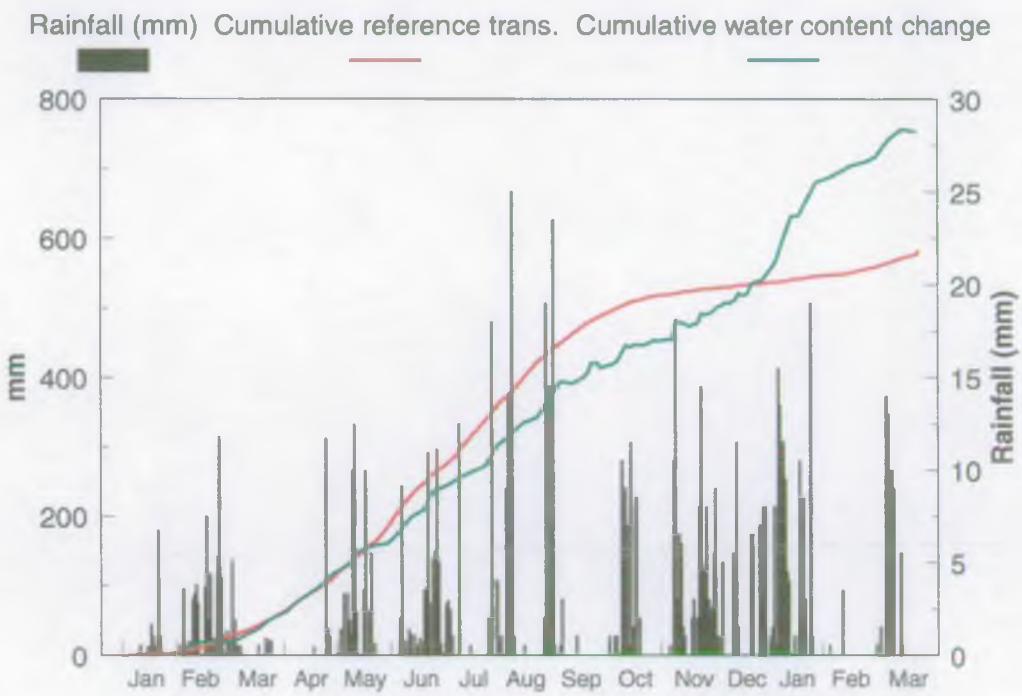


Figure 6.3 Reference transpiration and changes in soil water content at tube 6

The rapid increase in the evaporation rates in response to rainfall in June and again in October suggest that the vegetation is able to react rapidly to the wetting up of the soil profile. In particular, only the upper portion of the soil profile, where the majority of roots occur (about 0.2 m), needs to show a significant increase in soil water content. A note of caution should be introduce here. During the summer of 1997, the vegetation showed no sign of senescence, probably as a result of periodic rainfall. This might not be the case after a summer with a more prolonged dry period.

Decreases in the soil water content during the summer were observed in the deepest measurements taken, at 2 m, and significantly deeper than roots would be expected to penetrate, about 0.8 m. These could be attributed to either downward drainage or upward movement of water induced by the decrease in water content in the top part of the profile. The latter explanation is preferred as the results early in the year show that the changes could be fully explained by evaporation. It is also in accord with results made by other researchers, (Wellings and Bell, 1980) and (Gregory, 1989), who have shown that upward movement of water in the Chalk, in response to the potentials developed in the root zone, can occur at depths of 3 m or more. This suggests that the measurements were not taken deep enough to measure the full extent of the soil moisture deficits that developed during 1997.

Prolonged rainfall early in October restored the evaporation rate to that of reference transpiration. However the observed changes in soil water content from November onward imply an increase in evaporation to well above that of reference transpiration if this were the sole source of water loss from the observed soil profile. As this was not observed at any other time of the year, nor in the measurements made at Wallingford, it is unlikely that this is the cause of the loss. The explanation must be sought in the form of one or more other processes. The most likely processes involved are:

- runoff or interflow
- bypass flow
- drainage downwards out of the soil profile
- wetting up of the soil profile deeper than measured

The first process is considered unlikely as runoff or interflow are rarely recorded in catchments where Chalk soils dominate. The standard percentage runoff, SPR, for such catchments is typically 2 % (Boorman et al. 1995). Bypass flow is also considered to be unlikely as there is no evidence of its occurrence earlier in the data set and the loss of water appears to be a continuous process rather than episodic which would be expected from bypass flow. It is not possible from the data alone to decide between drainage downward and wetting up of the soil profile deeper than measurements were made. On balance, the last explanation is preferred, i.e. that the water goes to replenish deficits in the deeper soil profile. There is evidence for this in the soil moisture data which shows that the water content of the deepest measurements increase from mid-November onwards implying that water does migrate downward through the profile before field capacity is attained. A summary of the processes is shown in Figure 6.4.

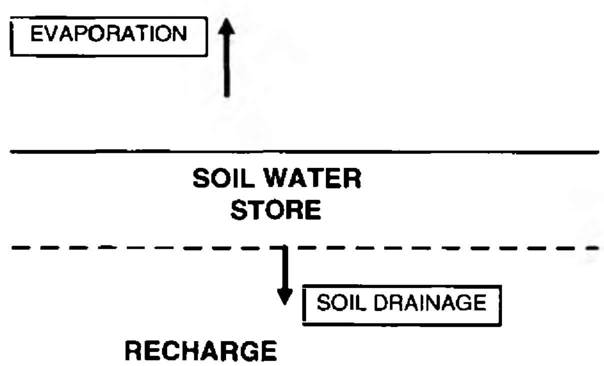


Figure 6.4 Schematic diagram of the processes for grass on Andover Series Soils

The mass balance shows that major losses of water are observed during the period from the last week of December through to the end of January. The most likely explanation for these losses seems to be that this marks a period of drainage from the bottom of the soil profile resulting in recharge. Approximately 160 mm of water are lost during this short period.

The analysis given above also serves in all but minor details for the measurements at tube 5. However, this tube is sited in a field of grass ley. The conclusion drawn is that grass ley can be considered as reacting identically to permanent grass for all practical purposes. This is surprising as the management of the ley involved taking a silage crop on 7 July after which the ley was grazed by sheep. It may be due to the reduction in the vegetation canopy occurring during a period when the soil moisture deficits were already limiting evaporation so the loss of canopy did not have as much impact as might be expected.

Wickham Series Soils

Two neutron access tubes are sited in the Wickham Series soils which are sandy loams on the Reading Formation. The Reading Formation is Lower Eocene in age and consist of clay, sand and loam with local seams and lenses of flint pebbles overlain by highly irregular alternations of stiff, more or less plastic, structureless clays with sand and loam. They are fluviatile in origin. The tubes are located in a field of permanent grass which was periodically grazed by cattle during 1997. For all practical purposes, there are no differences between the data sets collected from the two tubes.

No significant changes in soil moisture are observed at depths greater than 1.2 m which is interpreted as the maximum depth from which water was lost to evaporation. The pattern of changes in soil moisture during 1997, Figure 6.5, is very similar to that observed at the sites on the Andover series soils. It is likely that the soils returned to field capacity for a short period at the end of February. For the first part of the year, until the end of May, the changes in soil moisture can be attributed to evaporation and rainfall alone. In early June the actual evaporation fell below reference transpiration as a result of the developing soil moisture deficit. Following heavy rainfall in mid-June, evaporation increased but a second period of evaporation below the reference value began in the first week of July and was sustained until the beginning of October. Heavy rainfall, together with decreasing evaporative demand as a result of decreasing net radiation, then restored the soil water content to a level that could support the full capacity for transpiration by the vegetation. The soil moisture deficits which occurred when the actual evaporation began to fall below the reference transpiration were about 90 to 100 mm.

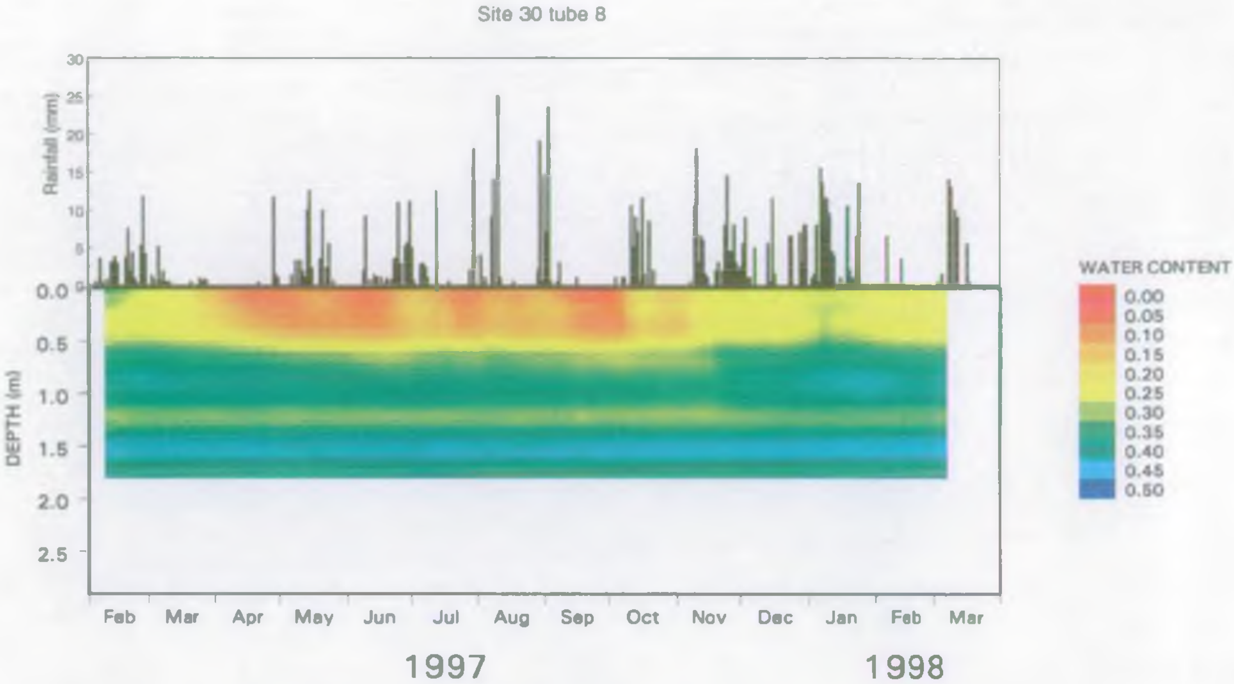


Figure 6.5 Changes in soil water content at tube 8

As was the case with the Andover Series soils, prolonged rainfall early in October restored the evaporation rate to that of reference transpiration. The sequence of wetting up of the soil profile seems to be different in that wetting up proceeds very distinctly from the top downwards, although it is not necessary for a layer to fully return to field capacity before wetting up of the layer below begins. Thus the rainfall at the beginning of October resulted in significant increases in water content down to a depth of 0.6 m but with the top 0.3 m returning to around field capacity. The rainfall early in November resulted in the main changes in soil water content occurring between 0.3 and 0.6 m. The soil profile seems to have returned to field capacity by the beginning of December.

The observed changes in soil water content from November onward imply an increase in evaporation to well above that of reference transpiration if this were the sole source of water loss from the observed soil profile. It also occurs in February, unlike the measurements on the Andover Soil series. The most likely processes to explain this loss are:

runoff or interflow
bypass flow
soil drainage downwards

Given the nature of the soils, it is very unlikely that bypass flow would be the cause of the loss as no cracks in the soils had been observed during the summer and there is no suggestion in the literature that bypass flow occurs in soils such as these. Given that the loss of water occurs very early in the wetting up period, it seems unlikely that water is draining down out of the profile. Thus runoff or interflow are considered the most likely explanation for the loss of water. If it is assumed that evaporation takes place at the rate of reference transpiration and no soil drainage takes place, then the amount of water lost to runoff or interflow in the period 5 November to 1 December is 23 mm for tube 8 and 40 mm for tube 9, from a rainfall total of 94 mm.

The heavy rainfall in January resulted in the soil water content exceeding field capacity for some time in the zone between 0.3 and 1.2 m. The water temporarily stored in this zone amounted to about 25 mm and the soils returned to field capacity relatively quickly suggesting that this water drained downwards out of the profile. A summary of the processes occurring at these sites is shown in Figure 6.6.

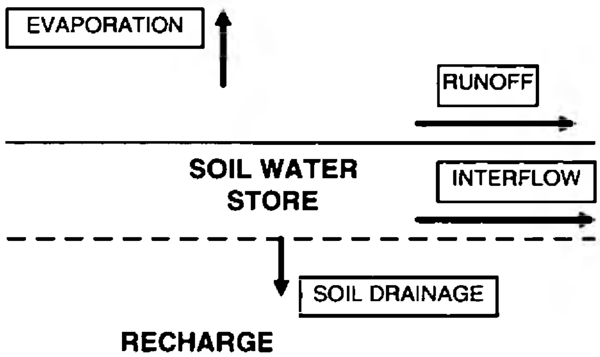


Figure 6.6 Schematic diagram of processes for grass on Wickham Soil Series

Hornbeam Series Soils

One neutron access tube, site 20 tube 4, is located in the Hornbeam Series soils which are a stagnogleyic paleo-argillic brown earth developed on the Clays-with-flints. The Clay-with-flints are rather stiff, brown or red-brown, slightly sandy clays containing angular flints. Their probable origin is a mixture of the insoluble residue of Chalk-with-flints and Lower Eocene deposits that have been re-arranged as the underlying Chalk was dissolved. It is located in the field adjacent to one of the tubes in the Andover series and the land cover is grass ley.

During 1997, the same trends are observed in the data as at the other sites with the soil moisture limiting evaporation during the summer. Evaporation became limited by soil moisture deficits once these exceeded 60 mm. Seasonal changes in soil moisture are limited to the top 1.7 m of the soil profile although this is deeper than would be expected that the roots would have penetrated, Figure 6.7. However, there is a difference in the autumn as the changes in soil moisture can be explained by evaporation and rainfall alone until the last part of November, significantly later than the tubes sited in other soils.

The rainfall in the first half of October returned the top of the profile to field capacity and penetrated down to a depth of about 0.5 m. The heavy rainfall in November resulted in rapid increases in water content down to 1.5 m so that the entire profile had returned to field capacity early in November. Thus water entering the soil profile is only lost once the entire profile has returned to field capacity. This suggests that runoff or interflow are not an important process at this tube but that the excess water drains downwards as recharge to the aquifer. A summary of the processes is shown in Figure 6.8.

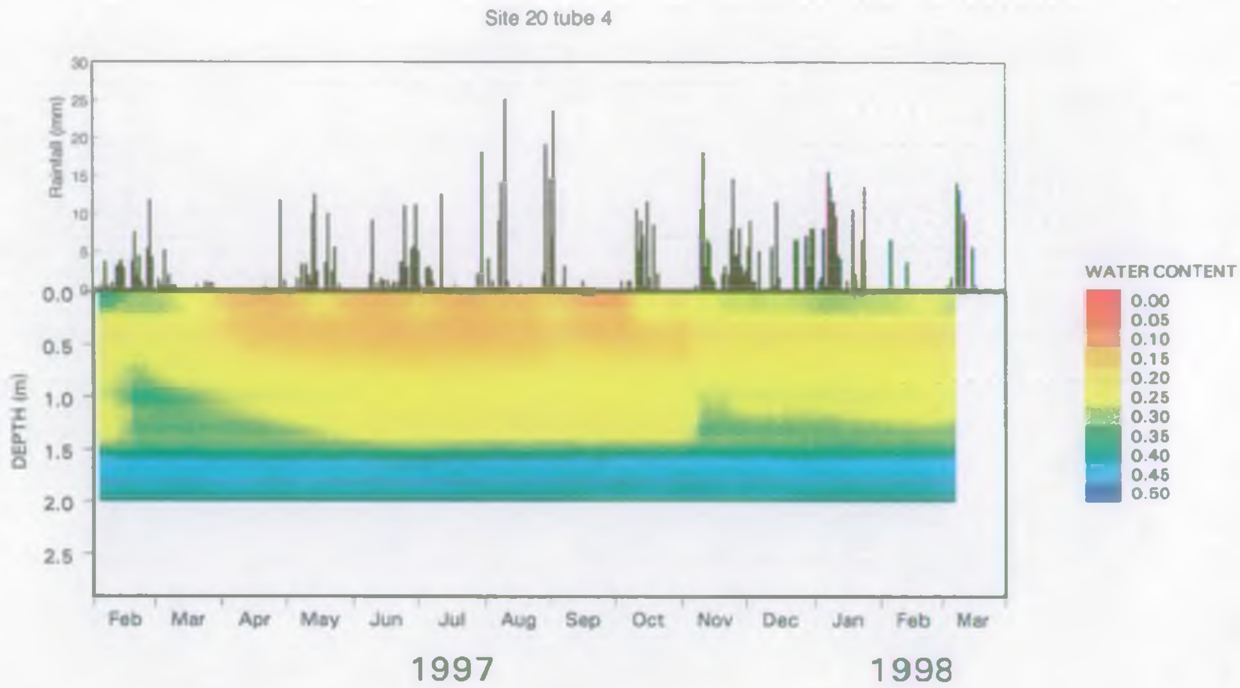


Figure 6.7 Changes in soil water content at tube 4

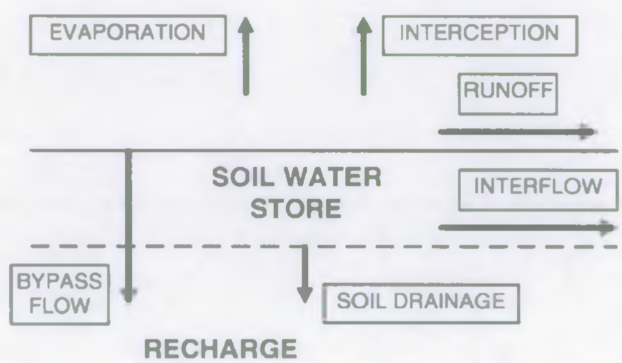


Figure 6.8 Schematic diagram of the processes for grass on Hornbeam Series Soils

6.1.4 Cereals

Four neutron access tubes are located in fields with a land cover of cereals. The field where the three tubes of site 20 are located was under barley in 1997 and 1998. Tube 10 is located in a field which was under winter wheat in 1997 and was replaced by winter barley. All the neutron access tubes are in the Andover series soils. The analysis has been confined to site 10 as the Bowen Ratio System was located here and data from this gives a guide to the evaporation losses.

The data acquired by the Bowen Ratio System is not continuous. A lengthy gap occurs when the system was removed during harvest and seeding to allow unrestricted access by farm machinery. Other gaps occur due to a variety of problems such as wires being damaged by rodents and intermittent problems with the data logger. However, the record does allow the changes in evaporation to be observed. In July 1997, Figure 6.9, the majority of the net radiation was driving evaporation but, in the period immediately proceeding harvest, the situation had reversed and comparatively little energy was being converted to evaporation. In October, following seeding but before leaves had appeared at the surface, the day time sensible heat flux was low or negative and evaporation, from the soil, and the soil heat flux were dominant. Despite the emergence of leaves, evaporation remained low through the rest of the autumn and through the winter virtually no evaporation was measured. In March 1998 the net radiation was increasing steadily and evaporation was again dominant as the vegetation canopy continued to develop.

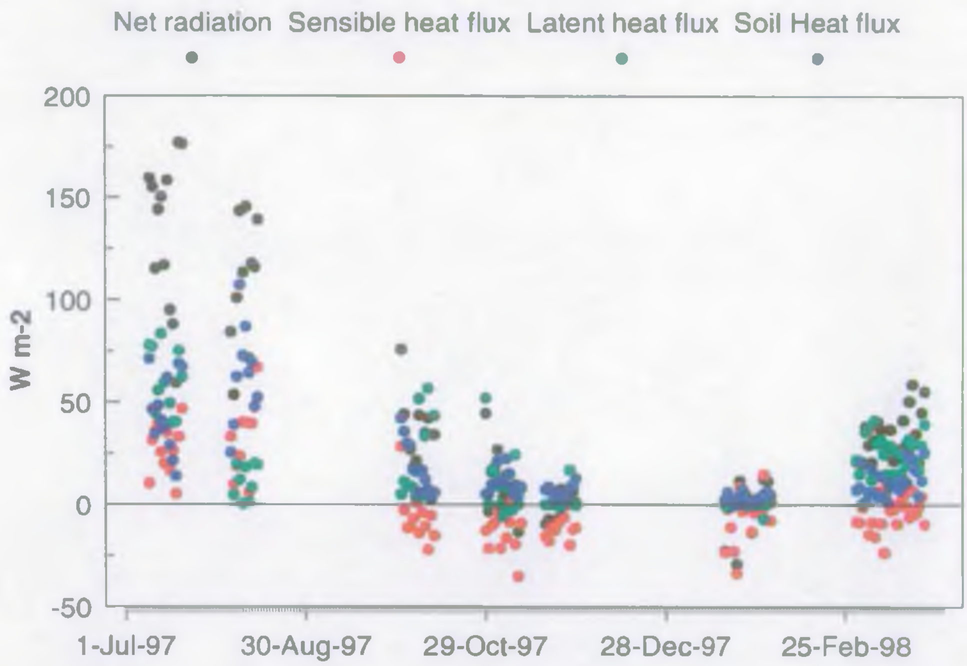


Figure 6.9 Daily daylight hours fluxes measured at tube 10

The soil moisture data, Figure 6.10, clearly shows changes in water content through the profile measured, 2.8 m, which suggests that soil moisture deficits developed to a depth greater than 3 m during 1997. The data also shows that there is no difference in the processes occurring to those observed with a vegetation cover of grass. Figure 6.11 shows a schematic diagram of the processes occurring at this site.

6.1.5 Deciduous Woodland

A single neutron probe access tube is located in a small patch of deciduous woodland on the Wickham Series soils. The data for this tube, Figure 6.12, is very different to that from the two tubes in the adjacent field with a cover of permanent grass. The water contents are markedly lower throughout the whole of the profile. There are two distinct horizons of higher water content that can be recognised in the data from this tube and that from both the tubes in the grass field which suggests that the differences are not due to a change in the soil type. The implication is that the trees have had an effect on the soil hydraulic properties.

The second major difference is that major soil moisture deficits are developed throughout the measured profile of 3 m. An increase in the soil water content at the bottom of the profile was only observed in the second part of April 1998.

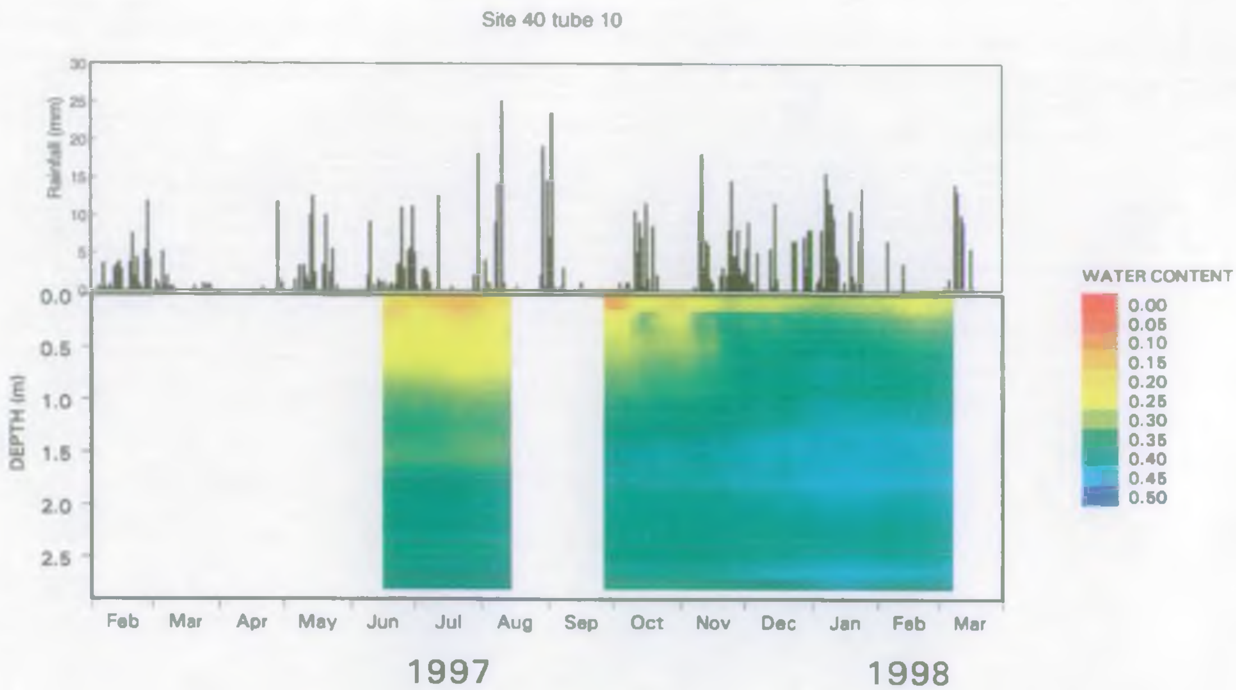


Figure 6.10 Changes in soil water content at tube 10

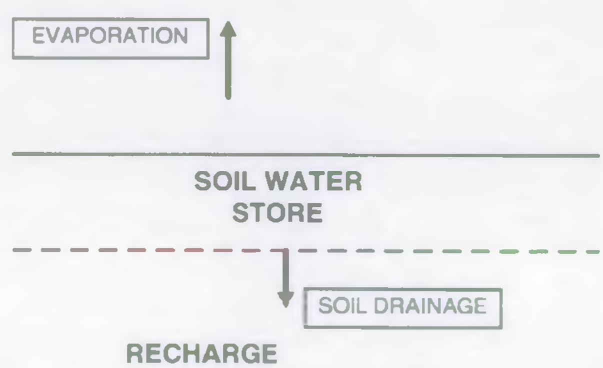


Figure 6.11 Schematic diagram of the processes for a site of cereals on Andover Series Soils

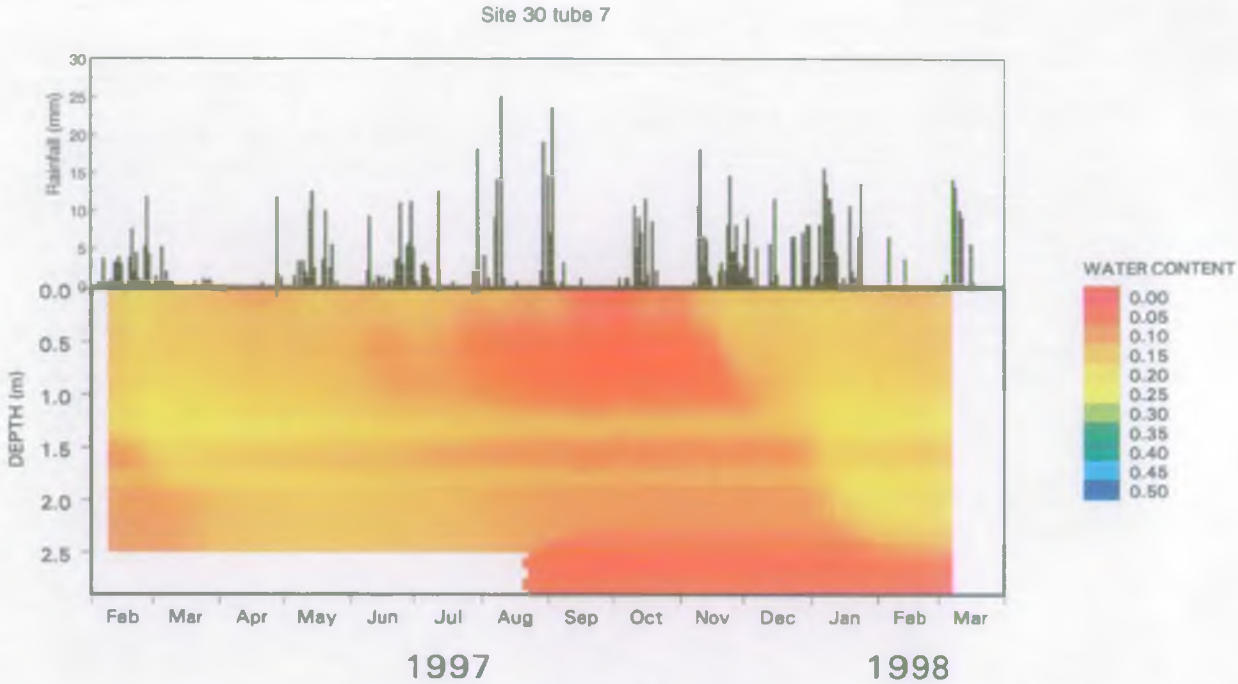


Figure 6.12 Changes in soil water content at tube 7

The evaporation and interception losses during the summer of 1997, shown by the changes in soil water content, were significantly higher than reference transpiration, Figure 6.13, (i.e. significantly higher than those for the adjacent grass) in part contributing to the size of the soil moisture deficits. (The cumulative change in soil water content can be assumed to be identical to the cumulative actual evaporation, provided soil water drainage does not occur.) The trees have no leaves during winter and there is very little understory with the result that evaporation during winter is solely from the soil. Thus, evaporation rates were very low from November 1997 to March 1998, i.e. less than reference transpiration. The changes in soil water content during the winter are unlikely to be due evaporation and the persistence of the deficits at depth suggests that it is also unlikely that drainage downward from the soil profile took place. Therefore it seems likely that runoff or interflow were occurring from the autumn onwards, possibly from when the near surface soil water content returned to field capacity at the beginning of November. The combined losses due to evaporation and interception are significantly greater than those for the adjacent grass sites (Tubes 8 and 9).

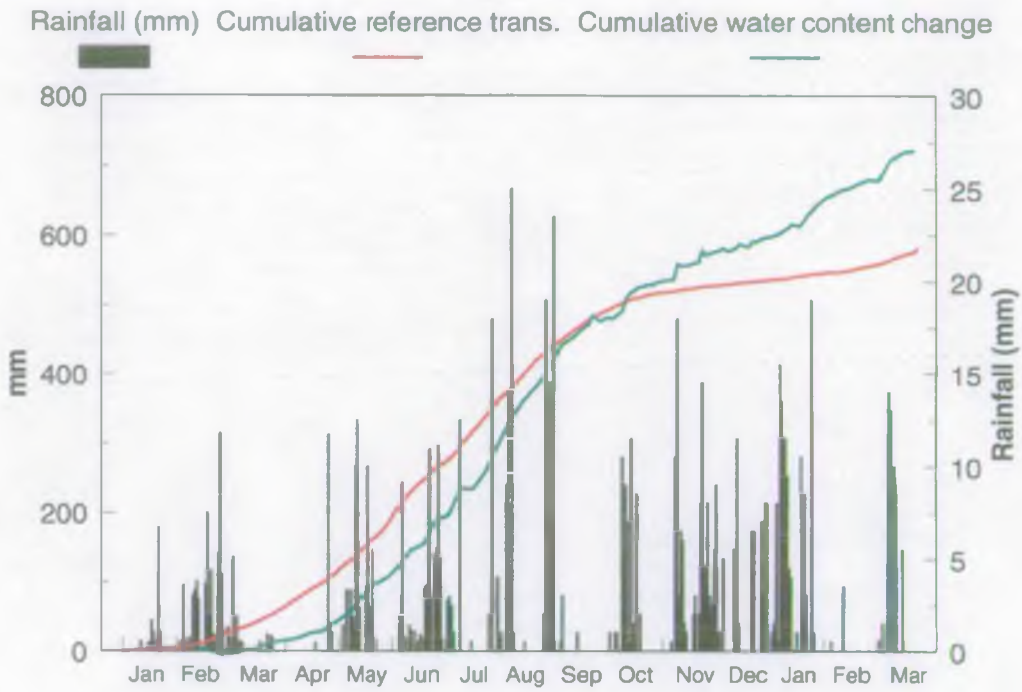


Figure 6.13 Reference transpiration and changes in soil water content at tube 7

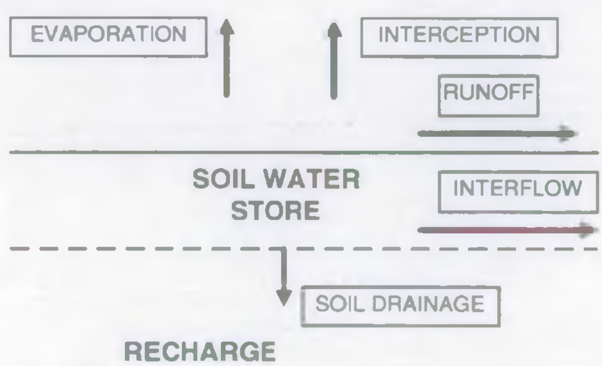


Figure 6.14 Schematic diagram of processes for a site of deciduous woodland on Wickham Series Soils

6.2 Tern Catchment

6.2.1 Summary of results

- The Windy Oak and Heathbrook sites are riparian so that recharge occurs rapidly to the water table in response to rainfall and it is unlikely that evaporation is limited by soil moisture at these sites.
- Most of the soils exhibit water draining downwards through the soil profile after rainfall.
- Soil moisture deficits developed to depths greater than rooting depth at the Bacon Hall and Stoke Bridge sites.
- Net radiation estimated from sunshine hours was significantly above the measured values

- Runoff or interflow probably occurred at the sites on the Redlodge and Hall soils in response to particularly heavy rainfall.

6.2.2 Grass

Newport Series Soils

These soils, brown slightly stony loamy sand or sand developed in Boulder Clay overlying the Lower Mottled Sandstone, occur at the Bacon Hall site which is also where the Hydra and AWS were deployed.

The Hydra and AWS proved very reliable. No data were lost due to instrument malfunction although some data were lost from the Hydra in late December 1997 and mid-January 1998 due to prolonged periods of fog and overcast skies resulting in insufficient power from the solar panels being stored in the battery.

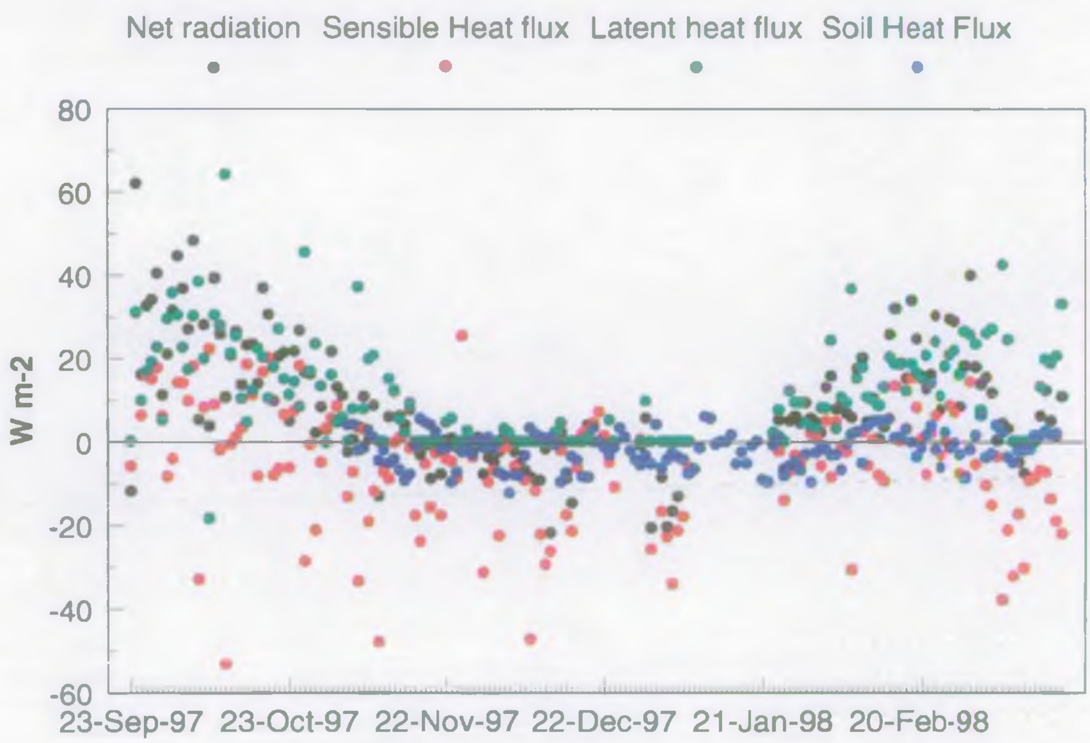


Figure 6.15 Daily daylight hour fluxes measured over grass at Bacon Hall

The resulting daily fluxes are shown in Figure 6.15. The decreasing net radiation through the autumn and subsequent increase is clearly shown, as is the extended period of very low net radiation that began on 15 November 1997 and continued until 20 January 1998. Generally, the latent heat flux is the dominant flux showing that evaporation was active throughout the period. There are some days when the latent heat flux exceeds the net radiation as a result of increased evaporation due to high wind speed or vapour pressure deficit. It is reflected in the daily evaporation values, Figure 6.16, which also illustrates the extended period of low evaporation.

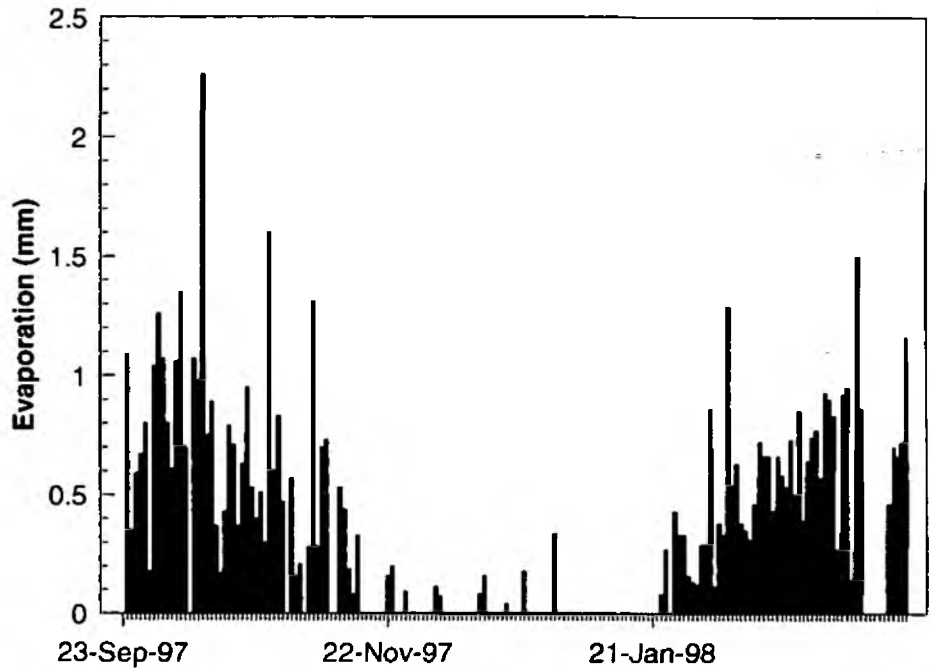


Figure 6.16 Daily evaporation values measured at Bacon Hall

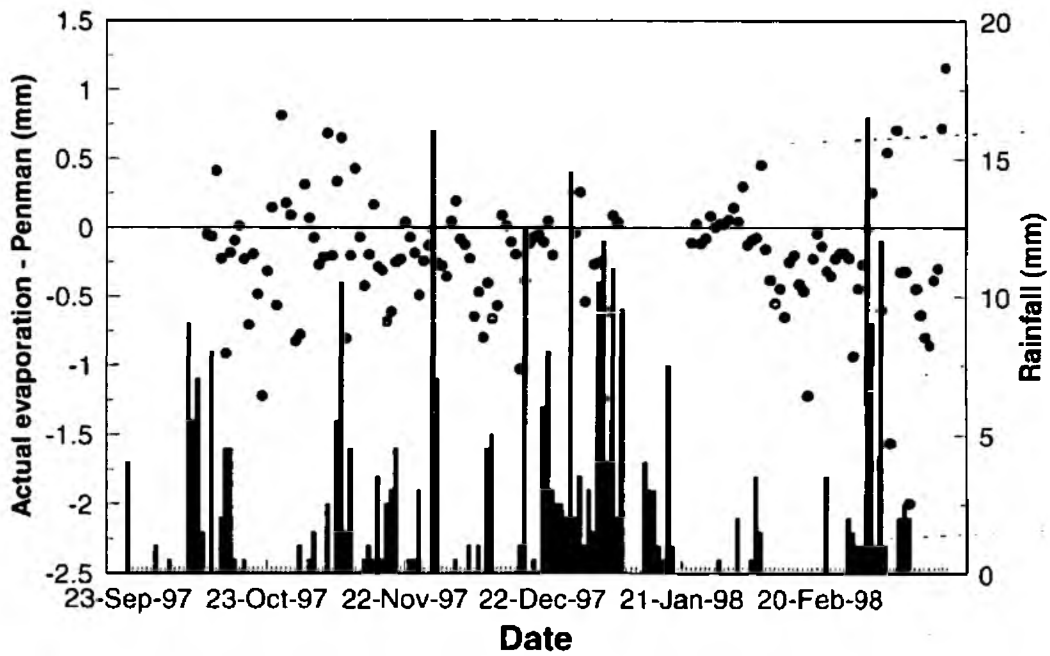


Figure 6.17 Difference between measured and Penman evaporation at Bacon Hall

There is no consistent indication of any change in evaporation following rainfall, Figure 6.17. However this graph does show that estimates of evaporation derived using the Penman evaporation model consistently exceed the measured values.

The Penman evaporation was calculated using the hourly values from the AWS data. Reference transpiration was calculated using the daily meteorological values obtained from the Met. Office for the observation site at Shawbury. This was found to be consistently much higher than the actual evaporation whereas it had previously been found that the values agreed very well for Wallingford. The reason for this seems to be that the net radiation calculated from sunshine hours was consistently higher than that measured. For the period 24 September to 31 December 1997, the cumulative net radiation measured by the automatic weather station was 1002 Wm^{-2} whilst that calculated from sunshine hours, using the measured albedo, was 1173 Wm^{-2}, a difference of 17%. As the difference between the measured and calculated solar radiation, is 6.4 % part of the problem is the estimation of the net long wave radiation. The most likely reason being that the coefficients used were inappropriate to the conditions in the Tern catchment. This suggests that considerable care needs to be taken in using values of net radiation derived from measurements of sunshine hours.

The soil moisture data shows changes in soil moisture down to and including 1.6 m which again suggests that water can be lost to evaporation from depths significantly in excess of the expected maximum rooting depth. The water table is at a depth of 2 m and so there is a possibility that water could be drawn up from the saturated zone but this is considered unlikely as the measurements at a depth of 1.8 m show no significant variations.

Rainfall in the first half of October resulted in the top of the profile returning to field capacity and increases in water content are observed down to a depth of 1 m. There was some drying out of the upper layers before heavy rainfall in November resulted in significant increases in water content at depths between 1.2 and 1.6 m so that the entire profile had returned to field capacity by the end of November.

The changes in soil moisture, in the soil profile measured, observed during the first part of October can be accounted for by evaporation and rainfall alone. This is shown in Figure 6.18 which compares the observed soil water contents with those predicted by a simple water balance model using evaporation and rainfall as inputs:

$$\theta_i = \theta_{i-1} + P_i - E_i$$

where

E_i actual evaporation on day i (mm)
P_i rainfall on day i (mm)
θ_i soil water content on day i (mm)
θ_{i-1} soil water content on day i-1 (mm)

The close agreement between the modelled and observed values confirms that the measured evaporation alone can explain the 'loss' of water from the soil profile. However, in the second half of October there is loss of mass in the days following the last rainfall event. It is unlikely to be due to runoff or interflow as it follows the rainfall and is interpreted as downward drainage out of the soil profile. A detailed examination of the soil profiles from the 15 October to the 31 October 1997 shows a decrease in water content from the surface down to a depth of 1.1 m but a small increase at subsequent depths. This suggests that, although loss of water from the soil profile to evaporation is taking place in the upper part of the profile, lower down, water is draining downwards despite the layers not being a field capacity. Changes in the soil water through November and up to the middle of December can be fully accounted for by evaporation and rainfall and so it is difficult to understand why the rainfall in October should result in drainage other than because of the particularly intense event on 8 October. The soil profile returned to field capacity early in December and so drainage out of the soil, and hence recharge, occurred from early December 1997 onwards. The variations in soil water content can be explained without taking into account runoff or interflow.

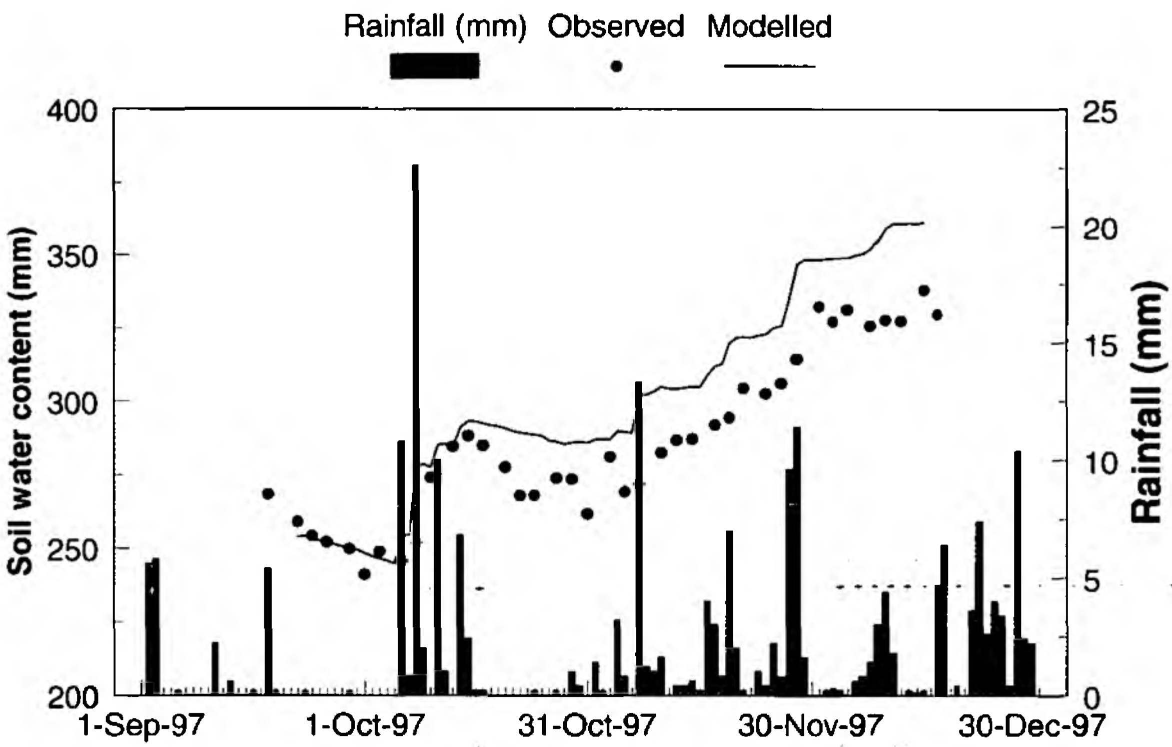


Figure 6.18 Comparison of modelled and observed soil water contents at Bacon Hall

The interpretations above are confirmed by the depths to the water table measured in the tubewells at the site. Both the shallow and deep tubewells show essentially constant levels until the beginning of December when there is a rise in the water levels. A summary of the processes at this site are shown in Figure 6.19.

Isleham Series Soils
These soils, permeable sandy and peaty soils developed in Alluvium, occur at Windy Oak. This site should probably be considered riparian as the water table lies at a depth of about 1 m below ground level and so, given the evidence from the other sites that water can be drawn up from depths greater than the expected maximum rooting depth, it is probably shallow enough that the vegetation can extract water from the saturated zone for transpiration.

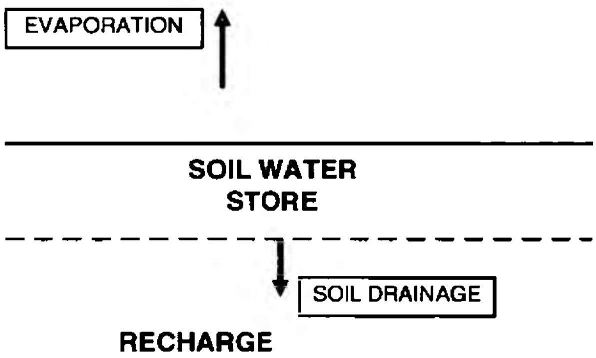


Figure 6.19 Schematic diagram of the processes for a site of grass on Newport Series Soils

Changes in soil moisture are observed down to a depth of 0.8 m and a strong soil moisture deficit had developed over the summer in the top 0.4 m. However, the rainfall early in October 1997 was sufficient to return the entire soil profile to field capacity and the tubewell shows that the water table reacted to this rainfall. The lack of rainfall during the second part of October resulted in some drying out of the upper part of the soil profile which matches the evaporation at this time. The rainfall at the beginning of November 1997 returned the soil to field capacity and subsequent rainfall resulted in recharge shown by the rise in the water levels. There are no indications of runoff or interflow taking place although it is impossible to be certain that this did not occur. A summary of the processes at this site are shown in Figure 6.20.

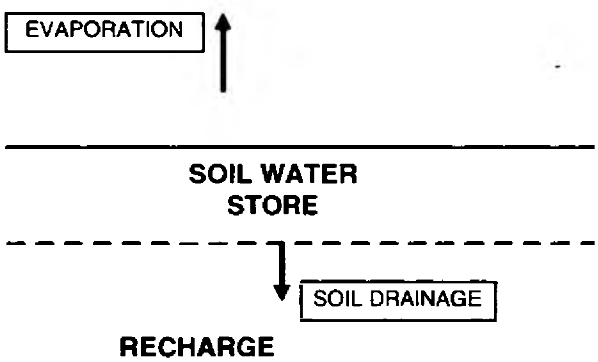


Figure 6.20 Schematic diagram of processes for a site of grass on Isleham Series Soils

Redlodge Series Soils

These soils, which tend to be reddish brown stony sands developed in Boulder Clay, occur at the Bowling Green Site. The enclosure with the neutron probe access tubes is under grass but the surrounding field had been planted with barley during 1997 so the results from this site need to be treated with some caution as, at the time of the measurements, the field would have been essentially bare soil. However, the water table is at a depth of around 3.8 m so there is no link between the groundwater and the zone affected by evaporation.

The observed change in soil moisture as result of the rainfall in the first half of October 1997 is too small to be explained by rainfall and evaporation alone, Figure 6.21. There is no change in the water table at this time so the most probable explanation is that runoff or interflow have occurred. This may be as a result of the particularly heavy rainfall, 22.6 mm, on 8 October. The changes in soil moisture observed at this site from the mid-October until the middle of November 1997 can be accounted for by rainfall and evaporation, assuming the same evaporation measured at Bacon Hall.

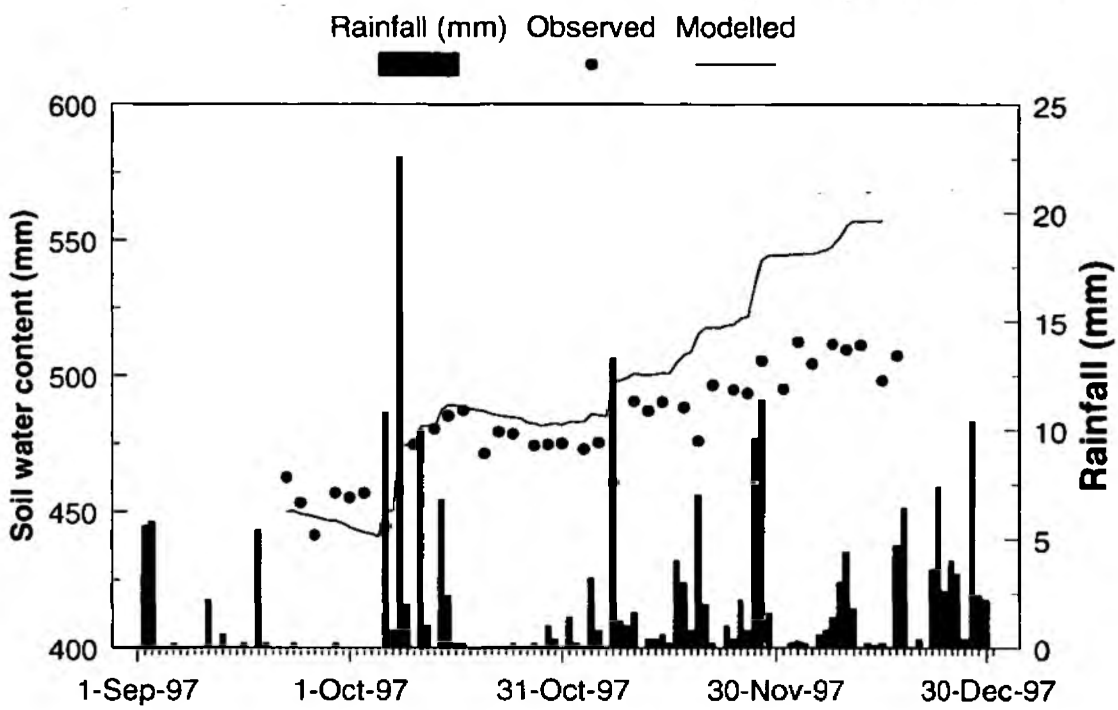


Figure 6.21 Comparison of observed and modelled soil water contents at Bowling Green

The rainfall in early October resulted in increases in water content down to a depth of 1.6 m with the main changes occurring near the surface and between 0.6 and 1.2 m. The rainfall in November had a similar effect but in December there were distinct increases in soil water content in the whole of the lower part of the profile. This appears to be a progressive migration of water from the upper part of the profile into the lower part as successively deeper layers show increases in soil water content at later dates. The water table shows a decrease in depth from 10 November 1997 onwards, which coincides with the period when the simple soil water model no longer describes the changes in soil water well, and so this date can be taken as the onset in recharge. Figure 6.22 shows a schematic diagram of the processes occurring at this site.

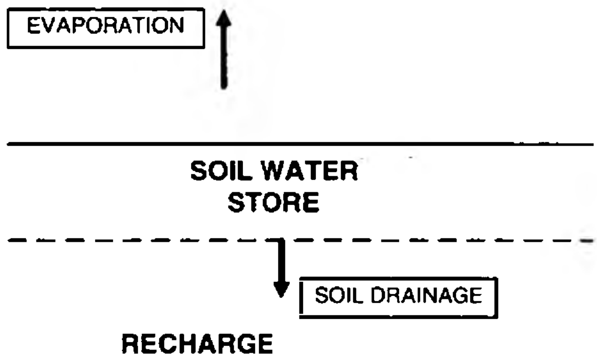
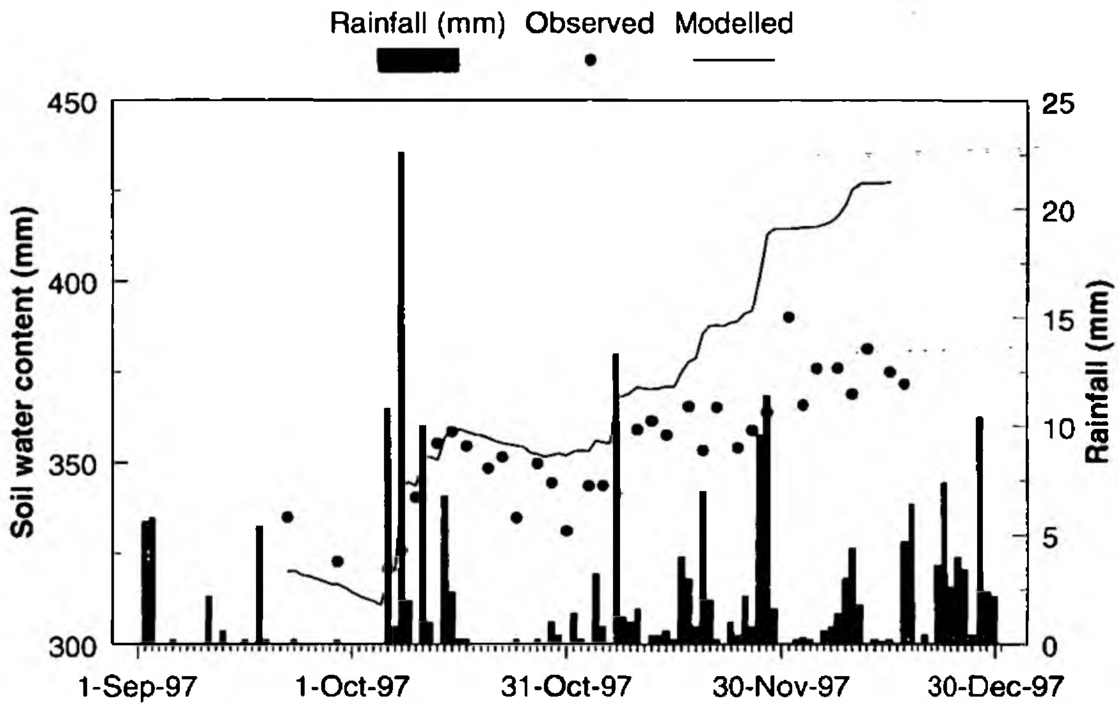


Figure 6.22 Schematic diagram of processes occurring for a site of grass on Redlodge Series Soils

Hall Series Soils
These soils, brown coarse loamy over non-calcareous gravely soils formed on the First Terrace, occur at Stoke Bridge. This site is located in an area of grass adjacent to a field of sugar beet. There is a possibility that the site may be located on rubble from the demolition of a nearby building.

Again, the observed change in soil moisture as result of the rainfall in the first half of October 1997 is too small to be explained by rainfall and evaporation alone and so the 'loss' is attributed to runoff or interflow, Figure 6.23. As at Bacon Hall, the second half of October is characterised by a reduction in the soil water content greater than can be accounted for by evaporation, It seems possible that there has been some drainage out of the soil profile. Changes in soil moisture are observed down to a depth of 1.2 m but there are no changes deeper. However, there is a rise in the water table in the tubewell of about 0.05 m at this time so drainage downward is taking place.

Figure 6.23 Comparison of observed and modelled soil water contents at Stoke Bridge



The rainfall in October increased substantially the water content from the surface down to a depth of 1.1 m with some increases at depths down to 1.6 m. This was followed by a reduction in the water content down to 1.1 m before the rainfall in November resulted in a substantial rise in the water content and drainage out of the soil profile began, shown by an increase in the water levels in tubewell from 6 November onwards. A more marked increase in water levels in the tubewell began in early December in response to the continuing rainfall. The onset of recharge is illustrated by the soil water contents predicted by the simple model diverging from the observed values from the middle of November onwards. A summary of the processes at this site are shown in Figure 6.24

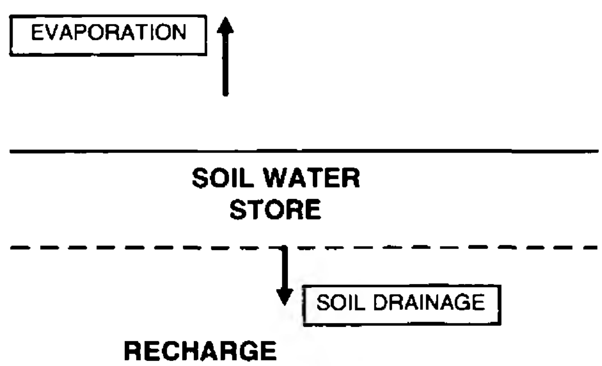


Figure 6.24 Schematic diagram of processes for a site of grass on Hall Series Soils

Altar Series Soils

These soils, earth eu-fibrous peat soils formed in Alluvium, occur at the Heathbrook site. This is very close to a water course and the water level in the tubewell is generally at a depth of less than 1 m so it is likely that conditions here are riparian. This is confirmed by the water levels that respond immediately to the rainfall at the beginning of October 1997. With the exception of the top 0.1 m, the water contents of the soil are high and soil moisture deficits are only developed in the top 0.3 m. The processes are summarised in Figure 6.25

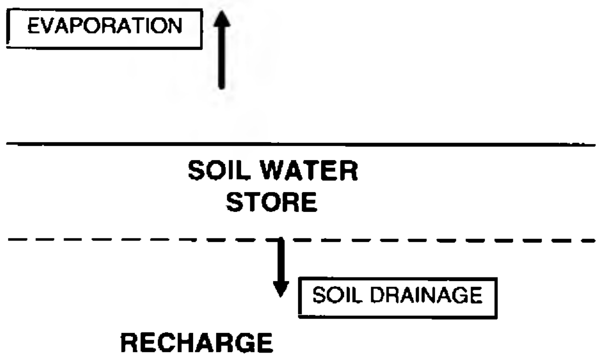


Figure 6.25 Schematic diagram of the processes at the site of grass on Altar Series Soils

6.2.3 Heathland

There are three sites on heathland, all on the Crannymoor Soil series which is a brown stoneless, loamy sand developed in sands and gravels overlying the Lower Mottled Sandstone. Hodnet Heath Control is in an area of sparse trees whilst Hodnet Heath Wet and Deep Wet are located in a treeless area.

At Hodnet Heath Control, changes in soil water content are observed down to a depth of 0.8 m. Below this there are no changes and the water level in the tubewell remained constant, at around 1.94 m, throughout the period of observation. However, the main changes in soil water content occur between the surface and a depth of 0.5 m. There are reductions in the water content in the second half of October but these are too small to be fully accounted for by evaporation. There are small increases in the water contents below 0.5 m in the second half of October, when there was no rain, which suggest that water is migrating downwards, although possibly in quantities too small to be detected in the water levels in the tubewell. The heavy rainfall in November increases the water content of the observed soil profile to a maximum by mid-November after which there are few indications of increases. This suggests that recharge should be taking place but there is no response in the water levels.

At Hodnet Heath Wet the water levels in the tubewells respond rapidly to all the rainfall events during the period of observations. The change in soil water content during the first half of October is fully accounted for by evaporation and rainfall but there is a reduction in soil water content in the second half of October well in excess of evaporation which suggests drainage of the soil profile downwards. This is confirmed by the soil moisture data, albeit close to the measurement uncertainty. A similar pattern is observed at the nearby Hodnet Heath Deep Wet site but the water levels in the tubewell show no response to rainfall at this time. The changes in soil water content at this site occur between the surface and a depth of 1.2 m. Although there are changes at greater depths, these do not seem to show any particular pattern suggesting that the zone between 1.2 m and the water table acts solely as a transfer with little effect on the actual water contents. The water levels in the tubewell show that recharge began in response to the rainfall on 6 November 1997.

6.2.4 Coniferous Woodland

The sole site on this land cover type is at Shawbury Heath where the soils are the Wick series, unmottled brown coarse loamy soils formed in sands and gravels overlying the Lower Mottled Sandstone.

Variations in soil water content that can be attributed to rainfall can be recognised down to a depth of 1.8 m where there is a layer of particularly low water content. Between 2.0 and 2.5 m the soil water contents show a trend of increasing water content with time. No trend can be recognised in the changes in water content deeper than 2.5 m. Between the surface and 1.8 m depth, the soil responds rapidly to the rainfall events in October suggesting that rainfall percolates rapidly through the soil profile. Between 0.5 and 1.8 m the rainfall in October results in a rapid rise in water content after which the water contents return to around the values observed before. A response to later rainfall events cannot be recognised. Between the surface and a depth of 0.5 m a response to rainfall events is clearly recognisable. This is interpreted as the October rainfall raising the water content of the soil profile above field capacity, after which there was drainage downwards and most of the soil profile returned to field capacity.

There are long term trends in the soil water data from this site which make it difficult to interpret the data. These are illustrated by the deepest measurements made, 3 to 3.6 m, shown in Figure 6.26. There is a clear trend of decreasing water contents after 1994. This is surprising as the values and lack of seasonal changes suggest that these values are below the water table. Thus the data from this site needs to be treated with caution.

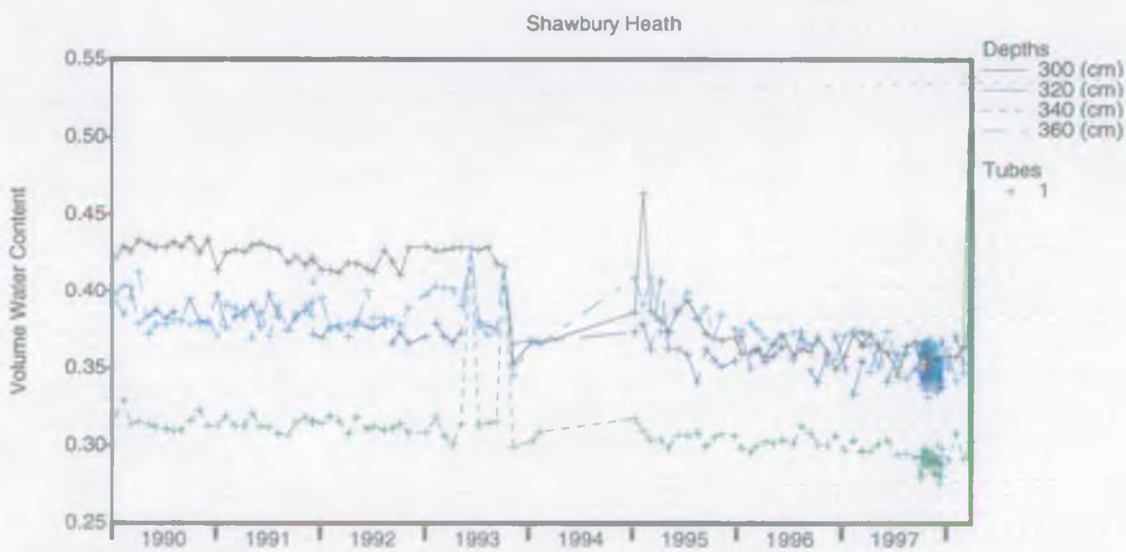


Figure 6.26 Changes in soil water content at depth for Shawbury Heath

7. CONCEPTUAL MODELS

In this chapter, conceptual models of the processes involved in the water balance at the sites. are developed. How these models are represented in the MORECS and FRL models is then discussed.

7.1 The Pang catchment

The measurements made confirm the conclusion of Finch and Harding (1998) that, in the presence of a significant soil moisture deficit limiting evaporation, grass is able to respond rapidly to rainfall infiltrating into the near surface soil and increase the evaporation rate. The increased evaporation measured cannot be attributed to interception as it persists for several days. This occurs on widely differing types of soil and so can be accepted as a general process.

7.1.1 Soil water storage and drainage

On the Chalk soils, the Andover and Coombe series, the observations confirm previous studies that soil moisture deficits are developed to depths significantly deeper, in excess of 3m, than roots have been recorded to reach. Gregory (1989), working in a Chalk soil at a site quite close to the Pang catchment, recorded that roots from cereal crops are restricted to the top 0.8 m, although occasionally roots were found down to 1.0 m in cracks. The phenomena has been observed in other soils, but the effect is particularly marked in Chalk soils and is a function of the exceptional nature of the Chalk. Wellings (1984), in a comprehensive study at a Chalk site in Hampshire, concluded that a field capacity water content could not be defined because rainfall infiltrating into the soil drained through the soil profile, and hence recharge, took place even when a soil moisture deficit was present in the summer and autumn. This occurs because the hydraulic conductivity of the matrix changes little with falling soil-water potential with the result that water continues to flow easily through the soil as the water content decreases. This also implies that soil water can be drawn up from depths significantly greater than the rooting depth in response to the potentials developed near the surface as water is extracted from the soil by roots.

7.1.2 Bypass flow

There is no evidence from this study that bypass flow occurs at any of the sites on the Chalk soils as it has been shown that the water balance can be explained without recourse to this process. This agrees with the conclusions of other studies using soil physics measurements, e.g. Wellings and Cooper (1983) and Wellings (1984), which demonstrated that despite the presence of small fissures in the Chalk, flow was through the matrix. Bypass flow is only likely to be a factor when these fissures are widened, by acidic drainage from soils such as the Wickham Series (Reading Formation), resulting in karstic features being developed in the Chalk. Features such as these have been recognised in the southern part of the Pang catchment, (Banks et al. 1995).

7.1.3 Runoff and interflow

Similarly, there is no evidence of runoff or interflow occurring on the Chalk soils because the mass balances add up without requiring contributions from these processes. Increasing slope has little effect on the proportion of rainfall going to interflow because the hydraulic conductivities of the soil and the rock are similar. The implication for recharge is that recharge rates are high because all the effective precipitation is available for recharge. Therefore, the conceptual model for the Chalk soil is that rainfall on these soils goes to either drainage, and thence recharge, or evaporation. Nevertheless, in response to the high evaporative demands during summer, soil water is drawn up from significant depths, potentially in excess of 3 m, into the root zone from which it is lost to evaporation. However, rainfall infiltrating into the soil travels through the soil profile in the presence of a soil moisture deficit with the result that drainage (recharge) can take place in the summer and autumn.

There is no evidence of runoff or interflow occurring on the Hornbeam soils and so the processes are similar to those on the Chalk soils. Again, soil water is drawn up from depths greater than roots are expected to reach, 1.6 m, but this is not as great as on the Chalk soils. Without measurements of the soil water potentials, it is not possible to give a definitive statement about whether drainage occurs whilst a soil moisture deficit exists, but it is considered unlikely from the soil water content data. It implies that the soil wets up from the top downwards and that recharge will only occur when the entire profile has returned to field capacity. Thus the processes involved in the water balance of this soil are comparatively straight forward.

The Wickham soils exhibit evidence of significant runoff or interflow. This appears to occur when prolonged rainfall results in the upper part of the soil profile having a very high water content implying that the storage capacity of these soils is sufficiently low that it is periodically exceeded by the rainfall. This is probably exacerbated as water is temporarily stored in the soil profile The result is that rates of recharge and interflow are comparatively low. Water is lost to evaporation from depths up to 1.2 m under grass, which is only slightly greater than the anticipated rooting depth. Thus these soils represent a more complex situation with drainage in the soil occurring much more slowly than either of the previous soil series with the result that the processes of runoff and interflow can occur.

7.1.4 Interception

The discussions above apply to the processes in the soil and essentially assume that the land cover is permanent grass. There is no need to modify the conceptual models for other land cover types except that the rooting depth and thus the maximum depth soil water is capable of being drawn up from will differ. However, it also shows that, for grass, the water balance can be understood with losses to the atmosphere confined to evaporation alone, i.e. that interception losses are not significant. This is confirmed more rigorously by the work of Finch and Harding (1998) at a site nearby, albeit on a different soil. The measurements of evaporation over the cereal crop suggest that interception losses can also be neglected over this land cover type. However, interception losses can not be ignored where the land cover is woodland as is shown by the magnitude of the soil moisture deficits developed under deciduous trees.

7.2 The Tern catchment

The conceptual model of the Hall, Newport, Redlodge and Isleham series soils is similar to that of the Hornbeam soils in the Pang catchment. There is no evidence of interflow or runoff occurring so the situation is comparatively simple as changes in the water balance are confined to gains from rainfall and losses to recharge or evaporation. Soil water is drawn up from depths of up to 1.6 m which is greater than roots of grass can be expected to penetrate, around 0.8 m. There is some evidence of decreases in the soil water content in the days following rainfall during the winter, which suggests that drainage in the soil profile can take a few days. However, the evidence is not compelling and so it can generally be assumed that drainage though the soil profile is comparatively rapid. Particularly heavy rainfall events, exceeding 20 mm, may result in drainage, either to interflow or to deeper parts of the soil profile.

The Crannymoor series soils present an anomaly as the water levels in the tube wells do not show recharge occurring when the soil water measurements suggest that it should. At this stage the explanation seems to be that these soils show considerable heterogeneity with very localised perched water tables and thus, at the very localised spatial scale of the measurements, no clear picture emerges. It is possible that models operating at the larger spatial scale with soil hydraulic properties representative of this scale may be able to reliably estimate recharge but further work would be required to demonstrate this.

7.3 A comparison between the MORECS and FRL soil water models and the conceptual models

7.3.1 Summary

Both the MORECS soil water model and the FRL model have the advantage of simplicity with minimal requirements for data. They can be expected to provide reliable estimates for a variety of land covers on free draining soils. However, because runoff and interflow are not represented in either model, the results from the models will be less reliable when these processes cannot be neglected, because rainfall that goes to runoff or interflow is not available for direct groundwater recharge. This could have a significant impact on estimates of direct groundwater recharge.

Both models are formulated on the assumption that soil moisture deficits only develop in the root zone. However there is clear evidence that soil moisture deficits can develop at depths much greater than the root zone, especially in the Chalk soils. The consequence of neglecting this process is to underestimate the magnitude of the soil moisture deficits that develop. This will result in over-estimating direct groundwater recharge because the models will require less rainfall to replenish these deficits than happens in reality. However, it is possible that the effect of this omission can be minimised by increasing the rooting depth (and thus the available water content). The FRL model is likely to be more successful in this because it is possible to reduce the proportion of roots in the bottom layer(s). Potentially, this factor could have a major impact on estimates of direct groundwater recharge

The multi-layered nature of the FRL model has some advantages over the single layer used in the MORECS soil water model. It allows the model to rapidly increase evaporation in response to rainfall despite the presence of soil moisture deficits at depth. The growth of crops is more easily simulated and the migration of water downwards through the soil profile is explicitly included in the model formulations.

7.3.2 Types of soil water model

Soil water models can be divided into two classes; those that use conservation of mass (soil water content) and those that use conservation of energy (matric potentials and hydraulic conductivity). The FRL model and the MORECS soil model, (Finch et al. 1995), simulate the soil water content. This is a pragmatic solution determined by the availability of data. Models based upon energy need to work with discrete time steps less than a day, typically an hour, and so require meteorological data of this time interval. They also require information on the hydraulic conductivity of the soil and how this changes as a function of soil water content. Measurements of soil hydraulic properties are rarely available and so they are usually estimated from measures of texture using regression models. These models are termed 'pedo-transfer functions'. As hydraulic conductivities vary over several orders of magnitude whilst critical water contents (saturation, field capacity and wilting point) vary by factors of about 2, the latter are less sensitive to errors in the pedo-transfer function. Finally, models of the energy are based on the Richards' equation, (Darcy's law combined with the law of conservation of mass) which is a partial differential equation and thus not trivial to solve. A recent attempt to provide a robust solution to the equations is given by Miller et al. (1998). However, the price of the simplicity of the FRL and MORECS soil models is that they do not attempt to fully describe the physical processes involved.

7.3.3 The MORECS soil water model

There are some differences between the soil water model used in MORECS 1.0 and MORECS 2.0 and so the following is a description of the model used in the latter and is taken from Hough and Jones (1998)

The model takes the soil to be a single reservoir which has only one parameter the 'available water content' which is a measure of the water available for evaporation by vegetation (and bare soil). Water is extracted from this reservoir for evaporation at the rate of potential evaporation until the soil moisture deficit exceeds the proportion, g, of the available water content that defines the 'easily available water'. Once the soil moisture deficit exceeds this value then the bulk surface resistance (of the Penman-Monteith evaporation model, see section 3.2) is increased from the value for the land cover when water is freely available, by multiplying it by a factor, a, calculated as:

$$a = \frac{2.5}{1 - \frac{SMD - gAWC}{(1-g)AWC} - 1.5}$$

where

SMD the soil moisture deficit

AWC the available water content

The form of this function is illustrated in Figure 7.1, for a value of g of 0.4. It can be seen that it only has an effect on the bulk surface resistance, and therefore evaporation, at large soil moisture deficits. The model will overestimate soil moisture deficits and thus underestimate recharge if the deficits used by the model are larger than those that actually result in a reduction in evaporation below the potential rate. This is because the model will overestimate the evaporation rate. MORECS uses a value of 0.4 for *g* for all vegetation types. There have been a number of experiments designed to measure this parameter reported in the literature, e.g. Jamieson et al. (1995). These studies have tended to make simultaneous measurements of soil water content and evaporation and fitted a simple model to the data.

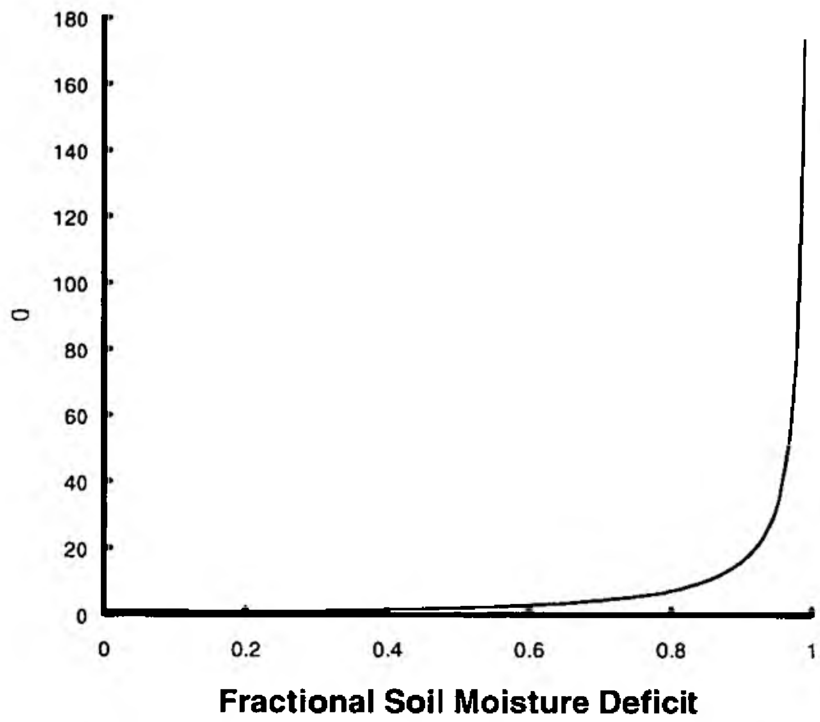


Figure 7.1 The MORECS model of surface resistance as a function of soil moisture deficit

The crucial parameter in the MORECS soil water model is the available water content, *AWC*. The concepts involved have not changed between versions 1.0 and 2.0, rather the terms used have been related to matric potential characteristics of soil in version 2.0. The available water content is defined by conceiving of the water content between the soil water contents at field capacity and wilting point being divided into two types on the basis of the matric potentials. The first type is held at low matric potentials and thus is readily extracted by vegetation, the easily available water. The second type is held at much higher potentials, the restricted available water, and thus is extracted with difficulty. The concept is that roots going down to a depth, which varies according to the vegetation type, have access to both types of water whilst those going deeper only have access to the easily available water. This definition seems to be, in part, an attempt to reflect the decreasing density of roots with depth which cannot be simulated easily with what is effectively a single layer model. In order to simulate growing crops, the available water content is assumed to increase from the bare soil value at emergence to the maximum for that crop on that soil at the date of maximum leaf cover.

There is no provision in MORECS for surface runoff as it assumes that all rainfall, other than interception losses, infiltrates into the ground. This is potentially a significant problem in estimating direct recharge on soils for which runoff is an important process because runoff is rainfall not available for recharge.

Infiltration goes to reduce the soil moisture deficits and, once the soil water content has returned to field capacity, is then taken as the 'hydrologically effective precipitation'. No differentiation is made between recharge and interflow and thus it can only be used, without further modelling, to estimate recharge for soils where interflow is not a significant process.

At the end of each time step, i.e. each day, the model sets the water content to field capacity if the water content exceeded this value. The excess is allocated to 'hydrologically effective precipitation' and thus drainage takes place within a day, which implies that the soils are free draining. Although this is true of some of the soils in the two catchments, it would not apply to either Chalk soils or clayey soils.

The use of what is effectively a single layer model results in some limitations. The first is that it will not be able to simulate the rapid increase in evaporation following rainfall in the presence of a soil moisture deficit developed throughout the root zone. This may result in the soil moisture deficit being underestimated and hence recharge overestimated. The second limitation is that the method used by the MORECS model to simulate growing crops is crude and it is not clear what the implications are for estimates of recharge from this except that there must be some concern for their accuracy in this situation.

The definition of the available water content is made on the basis of the rooting depth of vegetation and no account is explicitly taken of water being drawn up from below the maximum rooting depth, a process that the conceptual model suggests is very important in Chalk soils and is significant in some other soils. This is likely to have a significant effect on estimates of groundwater recharge as much larger soil moisture deficits will be developed in these soils than are predicted by the model. The result will be that the model will significantly over-estimate the groundwater recharge as it will require less of the rainfall to replenish the soil store.

Bypass flow is not represented in any way in the MORECS model. This would only be a limitation in areas of the Chalk soils where bypass flow occurs, i.e. where karstic features are present. Finch (1998) has suggested that estimates of mean annual groundwater recharge are not particularly sensitive to this parameter and so its absence may not be crucial in terms of the quantity of recharge. However, it will have a significant effect on the timing of recharge events.

There is no provision in the MORECS model for the slow drainage of water, as occurs in the Chalk soils, in the presence of a soil moisture deficit. This is because the model is based around the concept of field capacity which does not really work in the Chalk soils. However, this may not be a crucial failing if the objective is to estimate the annual recharge to the Chalk aquifer. It may be possible to define an effective field capacity which represents the soil water content below which drainage is negligible. Although the model will not predict the timing of the recharge correctly, it may be reasonably accurate in terms of the quantity as the amount of recharge will still be determined dominantly by the amount of rainfall lost to evaporation.

In conclusion, the MORECS soil water model will perform best on free draining soils, in which runoff and interflow are not important factors, and when soil water can not be drawn up from depths significantly greater than the rooting depth. In fact, these conditions can be taken as typical of many soils that overly aquifers which are near the surface. This may explain why estimates of direct groundwater recharge from MORECS have been found to be useful in these situations.

7.3.4 The FRL soil water model

The concept in the FRL soil model is that the root zone is divided into four layers; water movement between layers is based on a capacity approach. If the inflow to the first layer exceeds its field capacity then the excess water drains down to the second layer and so on for each of the layers. Water in excess of the field capacity of the bottom layer is considered available for groundwater recharge. The effective rainfall (infiltration) is taken as the inflow to the top layer.

$$P_e = P_f(1-b) \qquad P_f < P_c$$

where:

P_f daily rainfall after interception loss has been subtracted (mm)
P_c minimum rainfall required before bypass flow occur (mm)
P_e effective rainfall, i.e. infiltration (mm)
b fractional proportion of rainfall assigned to bypass flow.

Plant roots take up water at the rate of potential evaporation as long as there is no water stress, i.e. the water content is equal to the field capacity. The form of the stress function has been changed from that given by (Finch et al. 1995). The ratio between the current water content and the difference between the water contents at which evaporation falls below potential and wilting point is considered as a stress factor, S, and used to reduce the potential water uptake rate to actual uptake rate for each layer:

$$S_j = \frac{(\theta_j - \theta_{w,j})Z_j}{\theta_{d,j} - \theta_{w,j}} \qquad \theta_{w,j} < \theta_j < \theta_{d,j}$$

$$S_j = 0 \qquad \theta_j \le \theta_{w,j}$$
$$S_j = 1 \qquad \theta_j \ge \theta_{d,j}$$

$$E_{a,j} = E_p S_j R_j$$

where:

E_{aj} actual root water uptake for the jth layer (mm)
E_p potential total root water uptake (mm)
Z_j layer thickness for the j^{th} layer (mm)
R_j fractional proportion of roots in the j^{th} layer
θ_j current fractional soil water content for the j^{th} layer
$\theta_{d,j}$ fractional soil water content at which evaporation falls below potential for the j^{th} layer
$\theta_{f,j}$ fractional soil water content at field capacity for the j^{th} layer
$\theta_{w,j}$ fractional soil water content at wilting point for the j^{th} layer

The contribution of each soil layer to the total root water uptake, and hence the actual evaporation, depends on its root density. The distribution of active roots in a normal soil is approximately triangular in shape, the greater concentration being near the surface. The model allows the root development of a growing crop to be simulated by changing the contribution made from each layer.

The flow of water, F_j, downwards into each layer can then be calculated as

$$F_j = P_e \qquad j = 1$$

$$F_j = F_{j-1} - (\theta_{fc,j-1} - \theta_{j-1})Z_{j-1} \qquad j > 1$$

$$F_j = 0 \qquad F_{j-1} < (\theta_{fc,j-1} - \theta_{j-1})Z_{j-1}$$

Thus the potential direct groundwater recharge is the influx to the fifth layer.

FRL shares several of the limitations of the MORECS soil water model because both use the concept of soil water capacities. There is no provision made in FRL for surface runoff (although it is possible to add a sub-model to do this, see Finch, 1998). Therefore there is the potential for significantly overestimating groundwater recharge, for soils on which runoff is a significant component, because runoff represents rainfall not available for recharge.

Similarly, no provision is made for separating drainage into interflow and recharge because it is assumed that all movements of water are vertical. Again, this is likely to result in overestimates of recharge.

The FRL model, parallels the MORECS soil water model by setting the water content to field capacity if the water content exceeds this value in a day. This will result in similar problems with soils that are not free draining. The consequence of this is linked to the lack of a runoff component to the model as it is likely to be important if the water content of soil stays significantly above field capacity after one rainfall event so that, should a second heavy event follow closely, there will be insufficient storage within the soil for the rainfall and runoff will result.

The FRL model is specifically designed to deal with changes in soil water in the root zone. Therefore it does not recognise the process of water being drawn up from the soil beneath the root zone. The result will be to underestimate the size of the soil moisture deficits and hence overestimate the recharge. However, because the model is multi-layered, it may be more successful than the MORECS soil water model in being 'adjusted' to simulate these conditions. This could be achieved by defining the rooting depth to be the maximum depth that soil moisture deficits within in 'average' years. The proportion of roots in the deepest one (or two) layer(s) could be set to a small number, 1-2 percent, so that evaporation losses from these layers will be small. The result will be large deficits in the upper portion of the soil profile and small deficits in the lower part.

The multi-layered nature of the FRL model allows it to simulate the rapid increase in evaporation following rainfall, despite the presence of a soil moisture deficit at depth. This is because the evaporation from the top layer will increase as a result of the decrease of the soil moisture deficit in this layer. Similarly, the model will allow the progression of a 'wetting front' downwards through the layers. However, there is no provision for drainage of water in the presence of soil moisture deficits but, as with the MORECS model, this may not be a major problem if the magnitudes of the soil moisture deficits is correctly simulated as it will have more impact on the timing of recharge than on the quantity.

A further advantage of using a multi-layered approach is that more confidence can be placed in the simulation of growing crops. This is because the increase in the rooting depth is a parameter, rather than being implied, as in the MORECS soil water model.

The FRL model is likely to perform best in the same situations as the MORECS soil water model. These will be free draining soils in which runoff and interflow are negligible and where soil moisture deficits develop in the rooting zone.

8. MODELLING

8.1 Summary of results

- The Chalk soils clearly show that water can be drawn up from several metres below the root zone and this process can be simulated, to some degree, using the FRL model with the low values for the root proportions in the bottom two layers.
- Water is drawn up from beneath the rooting zone in soils developed on the Clay-with-Flints in the Pang catchment and the Boulder Clay in the Tern. However, the effect is not as marked as in Chalk soils, but it must be simulated by the soil water model in order to obtain reliable estimates of recharge.
- A bulk canopy resistance of 100 s m^{-1}, as used by MORECS, is appropriate for bare soil.
- The FRL model is better suited than the MORECS model to modelling the change from evaporation from the vegetation canopy to that from bare soil at harvest because it is multi-layered.
- The strategy for reducing evaporation from potential to actual used by the FRL model is more successful than that used by the MORECS model, mainly because of the multi-layered nature of the former.
- The FRL soil water model is more successful than the MORECS model at simulating water drawn up from beneath the root zone, due to the facility for the rate water is extracted for evaporation being altered in different layers by changing the proportion of roots in the layers.
- Water is not drawn up from depth in the Wickham soil series but runoff and/or interflow definitely occur. Because neither of the soil water models incorporate these processes they are not appropriate for estimating recharge in this situation without an additional module to partition drainage between interflow and recharge.
- Differences in estimates of the mean annual recharge of up to 15% can arise from the different rooting depth, in the case of the FRL model, or available water content, in the case of the MORECS model, used.

8.2 Modelling strategy and parameters used

The purpose of the modelling has been to test the FRL and MORECS soil water models, both the formulation of the models and the parameters used, against the observed data to determine their effectiveness for estimating groundwater recharge. The models have been run to simulate the data from three of the sites in the Pang catchment and two in the Tern.

For each site, the FRL model has been run with four different rooting depths:

1. The rooting depth was set to the 'conventional' value, i.e. the rooting depth anticipated for the vegetation type, 0.8 m for grass and 1.0 m for winter wheat. This is effectively the value that has been used in the past, e.g. Finch et al. (1995) and is implicit in MORECS.
2. The model rooting depth has been set to an arbitrary depth, greater than the actual rooting depth, such that it was anticipated that the predicted soil moisture deficits would correspond to the observed values for the entire soil profile.
3. The rooting depth was set to the maximum depth that soil moisture deficits were observed to occur.
4. The rooting depth was set to the maximum depth that soil moisture deficits were observed to occur. However, the fractional proportion of roots in the bottom two layers were set to

very low values, between 0.01 and 0.05, and the proportions in the upper two layers increased to compensate. This strategy was designed to test whether the effects of upward movement of water from below the rooting zone could be simulated by this crude adjustment.

The proportion of roots in successive layers, from the surface downwards, has been taken as: 0.6, 0.25, 0.1, 0.05, as in Finch et al. 1995, for 1, 2 and 3

A similar strategy has been followed for the MORECS soil water model to allow the results from this soil water model to be directly compared with those from the FRL model. However the MORECS soil water model consists of a single layer and so the equivalent available water content to that used for the FRL model was calculated. This was achieved by summing the fractional water content (water content at wilting point subtracted from that at field capacity) multiplied by the layer thickness for the four layers of the FRL model:

1. Corresponding to FRL 1
2. Corresponding to FRL 2
3. Corresponding to FRL 3

No attempt has been made to optimise any of the values for the model parameters. This is because this project is seen as potentially contributing to the Environment Agency's requirement for a national method for estimating groundwater recharge. Therefore, the values, such as fractional available water content, are those that an operational system might be anticipated to use. The values for field capacity have been obtained for the appropriate soil series from the Soil Survey's SEISMIC database (SSLRC, 1995). This database lists, for every soil series, the average value of various hydraulic properties and thus is a useful source of the parameters required by the models. It should be noted that the values given for the soil hydraulic properties are not derived directly from measurements but have been estimated, by the Soil Survey using a pedo-transfer function (a regression model), from measurements of soil texture, bulk density and organic content.

Both soil water models require an input of a daily time series of potential evaporation. For MORECS, this is calculated using a form of the Penman-Monteith evaporation model. The vegetation canopy parameters are set in the model and are given by Finch et al. (1995). The potential evaporation input to the FRL soil model for a land cover of grass was also calculated using the form of the Penman-Monteith evaporation model described in Section 3.2. The vegetation canopy parameters (height and bulk canopy resistance) used were those given for calculating the potential evaporation of a reference crop (which can be considered as grass) adopted by the FAO (Allen et al. 1994). The height of the crop is given as 0.12 m and the bulk canopy resistance as 70 s m^{-1}. In the MORECS evaporation model the height is given as 0.15 m and the bulk canopy resistance varies through the year between 40 and 80 s m^{-1}.

In order to estimate the evaporation from cereals, MORECS uses the Penman-Monteith model but changes the bulk canopy resistance and height with time in order to change the contribution of evaporation from the bare soil and the crop. This is necessary because the Penman-Monteith model assumes a closed vegetation canopy. A more realistic description of the land surface is provided by the Shuttleworth-Wallace model (Shuttleworth and Wallace, 1985), which is a derivative of the Penman-Monteith model. The Shuttleworth-Wallace model describes the energy partition of a sparse canopy by conceiving of the land surface as being a mixture of bare soil and vegetation. It requires essentially the same parameters as the Penman-Monteith model and so has been used to provide the input evaporation time series to the FRL model for a land cover of winter wheat with the same vegetation canopy parameter values as used by MORECS. Thus the maximum height of the crop is 0.8 m and the bulk canopy resistance for the full canopy (LAI of 5.0) is 40 s m^{-1}. The time sequence for the changes in the crop are that the crop emerges on 1 October and continues to grow to a height of 0.8 m and an LAI of 1 until 1 December. These remain constant until 17 March of the following year when growth restarts. Harvest occurs on 20 August.

8.3 Pang Catchment

8.3.1 Winter wheat on Andover soil series

This combination occurs at site 10 and represents a cereal crop growing on a soil developed on the Chalk. In addition to the soil moisture measurements, data from the Bowen Ratio Machine has provided values of daily evaporation from this site. The fractional available water contents used for each layer in the FRL model, starting from the surface, are: 0.240, 0.190, 0.190, 0.190. The rooting depths for the model runs of FRL and the corresponding available water contents for MORECS are given in Table 8.1.

Table 8.1 Soil water model parameters for site 10

	Rooting depth (m)		Available water content (mm)
FRL 1	1.0	MORECS 1	203
FRL 2	2.5	MORECS 2	506
FRL 3	2.8	MORECS 3	567
FRL 4	2.8		

The soil moisture deficits predicted by the FRL model generally correspond well with the observed values, Figure 8.1, with the exception of that with a rooting depth of 1.0 m. Although the observed soil moisture deficits for 1.0 m are duplicated fairly well in the middle of the year by the FRL model with a rooting depth of 1.0 m, case 1, there is poor correspondence in the autumn. In particular, the soil is predicted to return to field capacity much earlier than is observed. This confirms that it is important to consider the full depth that soil moisture deficits are developed in. The soil moisture deficits of the full soil profile are predicted well by the three other rooting depths, such that there is little to chose between them as they all predict the returning of the soils to field capacity at about the same time as the observed data. The model with the greatest rooting depth and the 'standard' proportion of roots in each layer predicts the greatest soil moisture deficits and thus tends to correspond best with the observed values. The evaporation model is underestimating the evaporation rates in the autumn because, during dry periods in September and October, the observed soil moisture deficits increase at a faster rate than those predicted. This is confirmed when the observed and predicted values of evaporation are examined, Figure 8.2. An improvement could be achieved by reducing the bulk resistance of the soil from the value of 500 s m^{-1} used.

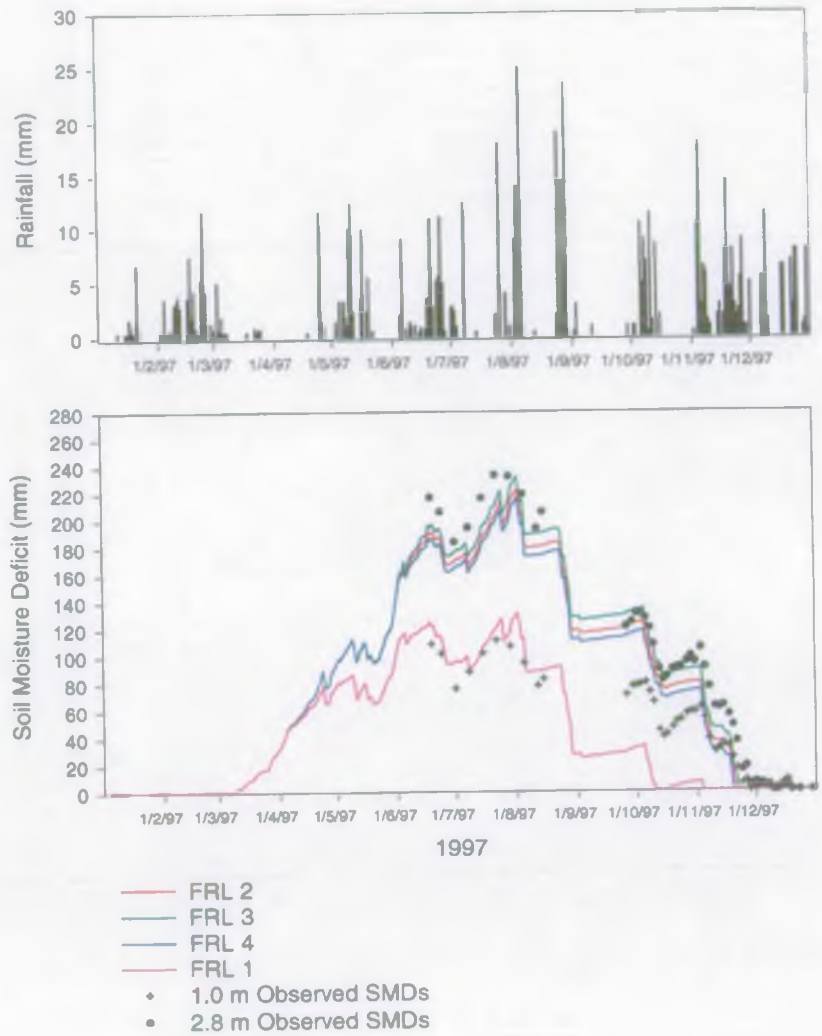


Figure 8.1 Observed soil moisture deficits and those predicted by FRL

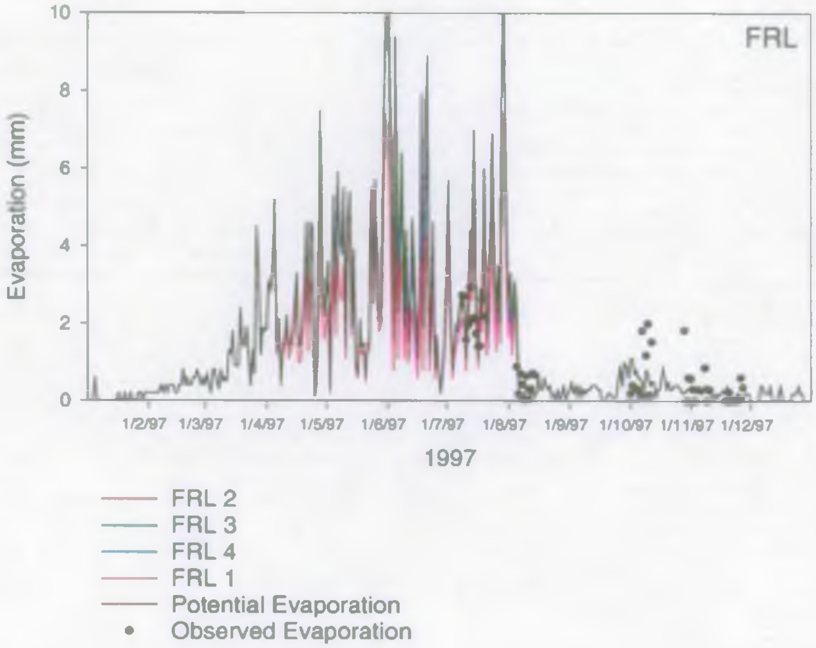


Figure 8.2 Observed evaporation and that predicted by FRL for site 10

The rapid drop in evaporation rates at the time of harvest is duplicated well by the model. However, the daily values show more variability than is observed and the values predicted are sometimes considerably higher than those measured. This appears to be due to the formulation of the aerodynamic resistance in the evaporation model and needs further investigation. Figure 8.2 shows that a significant reduction in evaporation due to lack of soil water is only predicted for the a rooting depth of 1.0 m. Thus the crop was well supplied with water for evaporation.

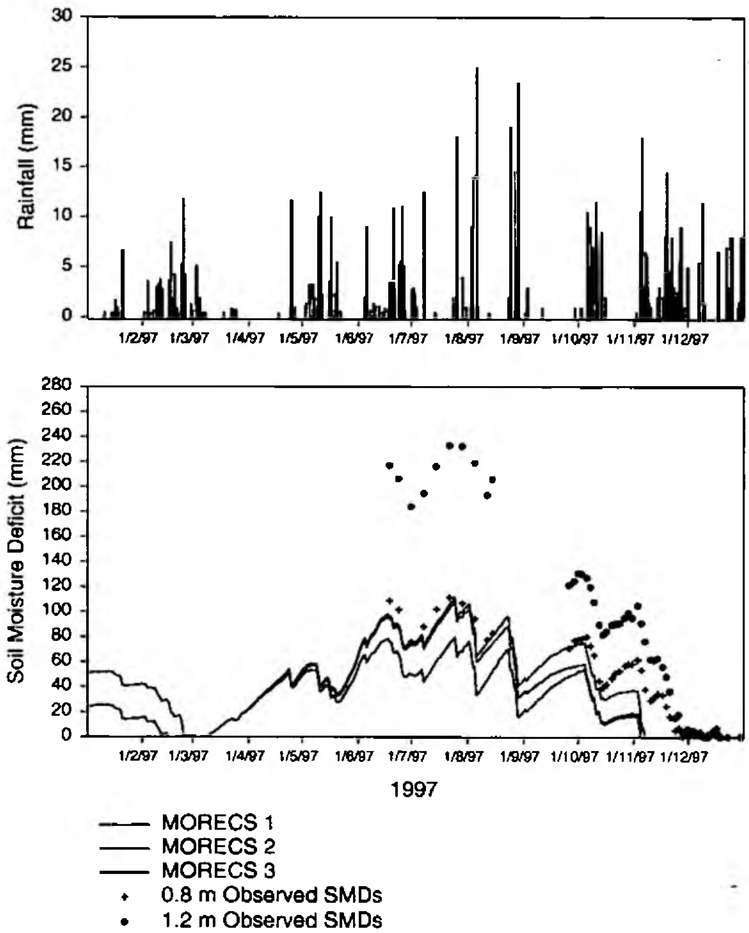


Figure 8.3 Observed soil moisture deficits and those predicted by MORECS for site 10

The soil moisture deficits predicted by MORECS are significantly lower than those predicted by the FRL model, Figure 8.3. This appears to be as a result of the lower potential evaporation rates predicted by the MORECS evaporation model for the fully developed wheat canopy. The result is that the observed soil moisture deficits for the top metre are duplicated well until the beginning of October but the return to field capacity of the soils is predicted to occur about three weeks too soon. The magnitude of the soil moisture deficits developed over the full soil profile are not duplicated well. However, the correspondence of the observed evaporation rates in July and those predicted by MORECS is extremely good, as is that in the autumn and early winter.

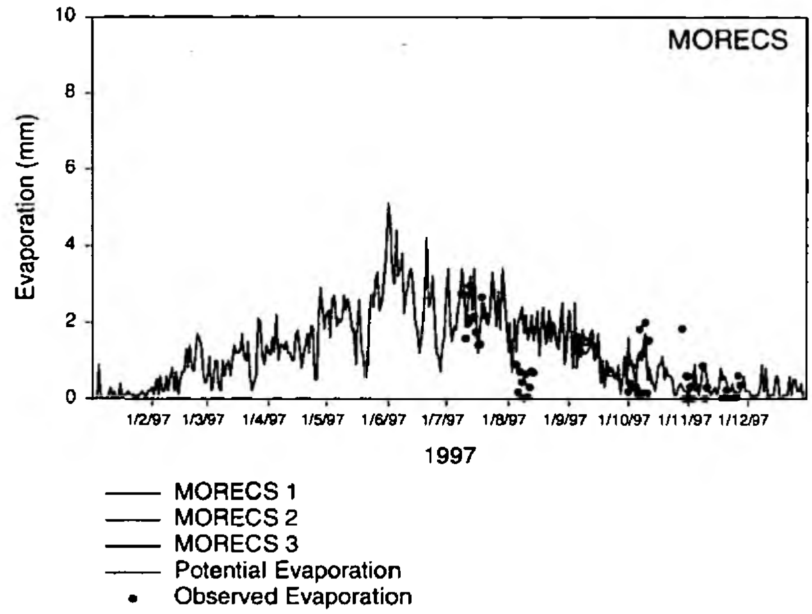


Figure 8.4 Observed evaporation and that predicted by MORECS for site 10

The evaporation rates predicted by the MORECS model do not change very much either prior to harvest or immediately after, whilst the observed values show a decline to very low values in August. This is could arise from two factors in the MORECS model. The first is that the value, used in MORECS, for the bulk canopy resistance of bare soil is 100 s m^{-1}, which is comparatively low, resulting in high evaporation rates. However, given that the values of evaporation predicted for this condition agree well with the measured values, it is probably representative of this site. The second factor is that the reduction from potential to actual evaporation in response to soil moisture deficits is necessarily computed from the whole of the soil profile, as the soil water model is effectively a one layer model. In practice, it is generally accepted that evaporation from bare soil occurs down to a depth of 0.1 to 0.2 m., e.g. Yamanaka et al (1998). Evaporation will be severely restricted at the time of harvest due to the removal of the vegetation canopy and because the zone from which bare soil evaporation can occur will have developed a large soil moisture deficit. Thus, a multi-layered soil water model is required to simulate the evaporation from annual crops.

The values of the mean annual recharge for the period 1988 to 1997 predicted by the FRL model are all higher than might be expected for the catchment, Table 8.2. This is due to the evaporation model under-estimating the evaporation from bare soil, which is particularly crucial in the winter months. Reducing the bulk canopy resistance for bare soil to the value used by MORECS reduces the mean annual recharge by about 140 mm, with the result that the values are less than those predicted by MORECS and more in line with what might be anticipated.

For this site, the different sets of parameters, for the two models, result in estimates of mean annual recharge varying by about 30 mm, i.e. around 10 % for FRL and 15% for MORECS,

Table 8.2. Thus it is important to represent the soil water accurately. This is likely to be even more important in drought years as the error in determining the date when the soils return to field capacity will be a larger fraction of the duration, in the winter, when recharge occurs. For the FRL model, the model runs with the largest rooting depth value, FRL 3, give the best agreement with the observed soil moisture deficit data and so these are preferred for Chalk soils. This results in the lowest estimate of groundwater recharge,

Table 8.2 Predicted 10 year mean annual recharge (mm)

	Pang Site 4	Pang Site 9	Pang Site 10	Tern Bacon Hall	Tern Stoke Bridge
FRL 1	251	249	364	195	214
FRL 2	236	209	331	200	177
FRL 3	230	219	328	166	187
FRL 4	183	240	337	190	175
MORECS 1	243	240	232	136	135
MORECS 2	230	227	205	116	105
MORECS 3	208	216	199	74	90

8.3.2 Grass on Hornbean soil series

This combination is found at Site 4 and represent the situation of a soil developed on the Clay-with-Flints. The fractional available water contents used for each layer in the FRL model, starting from the surface, are: 0.205, 0.154, 0.136, 0.136. The rooting depths for the model runs of FRL and the corresponding available water contents for MORECS are given in Table 8.3

Table 8.3 Soil water model parameters for Site 4

	Rooting depth (m)		Available water content (mm)
FRL 1	0.8	MORECS 1	126
FRL 2	1.0	MORECS 2	158
FRL 3	1.7	MORECS 3	268
FRL 4	1.7		

The soil moisture deficits predicted by the FRL model for cases 2 and 4 agree well with the observed values, Figure 8.5. All four runs show soil moisture deficits developing earlier than is observed. This may be as a result of the grass ley having been sown the previous autumn so of the canopy had not developed fully by the spring of 1997. The values predicted by case 1, with the value for the rooting depth set to the anticipated rooting depth of grass, are greater than are observed, again suggesting that the process of water being drawn up from depth is important at this site. However, the predicted values do not correspond well with the deficits observed over the whole soil profile, resulting in the model predicting a return to field capacity approximately 10 days earlier than occurred. The changes in soil water content predicted during dry periods in September and October correspond well to the observed values, demonstrating that the evaporation predicted by the model is accurate. The case of simply using a value for the rooting depth corresponding to the maximum depth soil moisture deficits are observed to develop, case 3, results in an overestimation of the soil moisture deficits. This suggests that the process of water being drawn up from below the rooting depth is not as important in these soils as in the soils developed on the Chalk Good agreement with the observed soil moisture deficits is achieved by either using a rooting depth 20% greater than 'conventionally' accepted, or reducing the proportions of the roots in the last two layers to 0.01 each. Either of these strategies would give acceptable results.

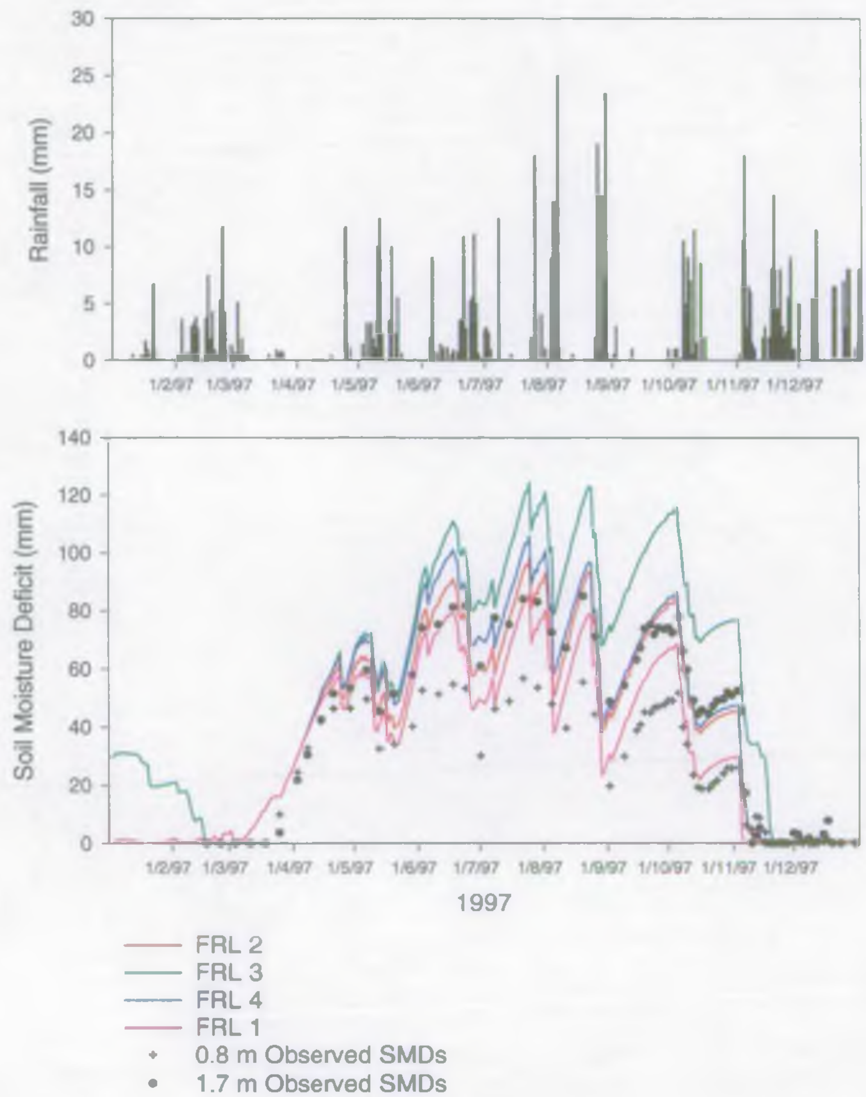


Figure 8.5 Observed soil moisture deficits and those predicted by FRL for Site 4

There is no significant difference in the soil moisture deficits predicted by the three different available water contents used with the MORECS soil water model, Figure 8.6. This is a consequence of the values of potential evaporation predicted by the evaporation model being lower than that used with the FRL model. Thus evaporation is at the potential rate throughout the year because the soil moisture deficits are insufficient to cause any significant reduction. This is unlikely to be correct because, for the same period and at a site approximately 10 km.

away, Finch and Harding (1998) observed that evaporation from a grass pasture was reduced in response to soil moisture deficits, albeit on a different soil series.

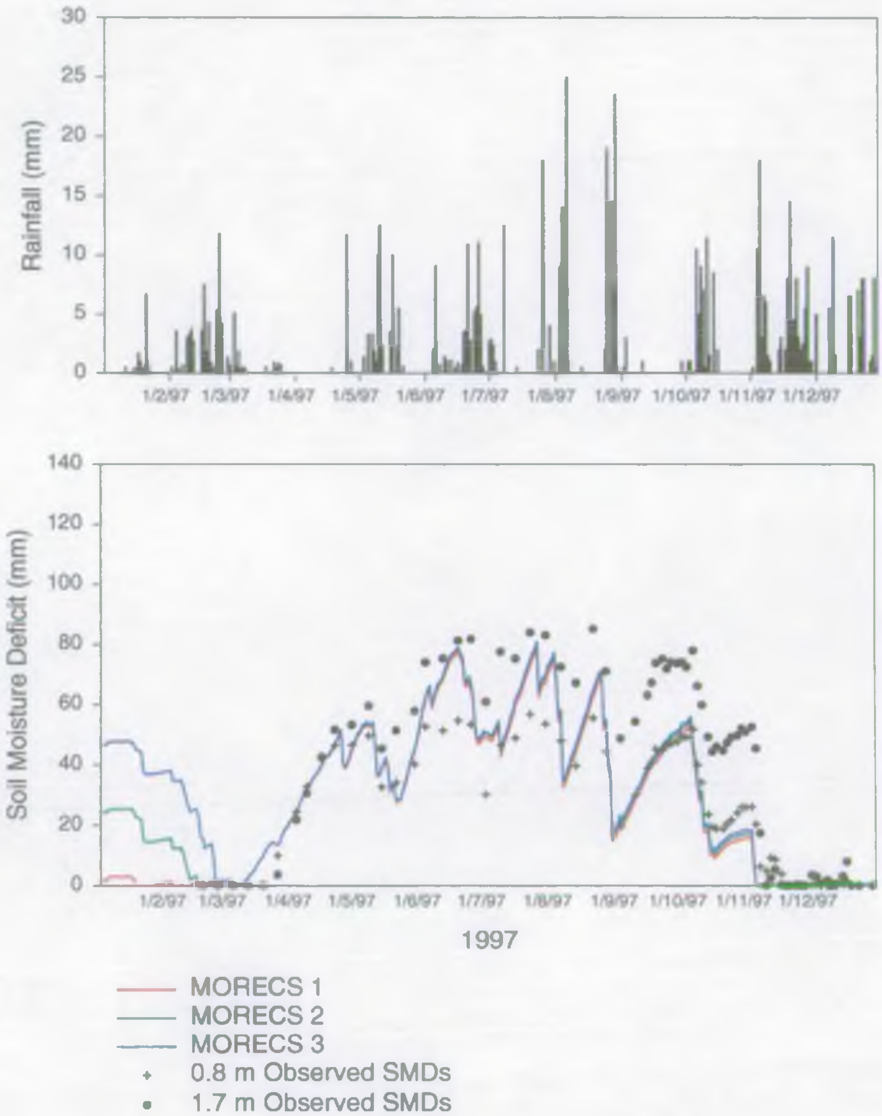


Figure 8.6 Observed soil moisture deficits and those predicted by MORECS for Site 4

The soil moisture deficits predicted by the MORECS model agree well with the observed values for the rooting zone, 0.8 m, until June. The values predicted by the model then exceed those observed until the end of August. It seems likely that this is the period when actual evaporation fell below potential. It illustrates one of the differences between the FRL and MORECS soil models. The FRL model reduces evaporation below potential at a lower total profile soil moisture deficit than the MORECS model because the effect of soil moisture deficits is applied to each layer separately. This results from the proportion of roots in each layer being used to determine the proportion of the total potential evaporation that is contributed from that layer. Thus, soil moisture deficits develop rapidly in the top most layer resulting in a reduction in the actual evaporation. For the data at this site the FRL model reduces evaporation below potential when the total profile soil moisture deficit exceeds between 50 and 55 mm. This is slightly less than the value of 60 mm suggested from the analysis of the observations (see section 6.1.3). In comparison, the MORECS model has not reduced evaporation at soil moisture deficits that exceed 80 mm.

The soil moisture deficits predicted by the MORECS model return to field capacity about two weeks earlier than the measurements of soil water content show. This results from the model underestimating the magnitude of the soil moisture deficits developed in the total profile. The impact of this is seen in the estimates of the 10 year mean annual recharge, Table 8.2. The values predicted by MORECS are consistently larger than those by the FRL model, by between 5 and 10%. There is a range of 10% in the estimates of recharge from the different MORECS runs and 27% for the FRL model runs. The model runs that best duplicated the observed soil moisture deficit values were FRL 2 and 4. These estimate mean annual recharge values at 236 and 183 mm respectively. The latter case is the preferred choice as it is closer to the conceptual model of this site.

8.3.3 Grass on Wickham soil series

This combination is found at Site 9 and represent the situation of a soil developed on the Reading Formation. The fractional available water contents used for each layer in the FRL model, starting from the surface, are: 0.207, 0.152, 0.156, 0.156. The rooting depths for the model runs of FRL and the corresponding available water contents for MORECS are given in Table 8.4.

Table 8.4 Soil water model parameters for Site 9

	Rooting depth (m)		Available water content (mm)
FRL 1	0.8	MORECS 1	134
FRL 2	1.0	MORECS 2	168
FRL 3	1.2	MORECS 3	201
FRL 4	1.2		

The observed soil moisture deficits are poorly predicted by all the FRL model runs, Figure 8.7. There is good agreement between the modelled and observed values for the first half of the year. Subsequently, cases 2 and 3, representing the deepest rooting depth, significantly overestimate the values compared to the observed data but do get the date of returning to field capacity correct, i.e. they get the right answer but for the wrong reasons. Cases 1 and 4, representing a shallow rooting depth and few roots in the lower half of the profile respectively, simulate the observed data well with the exception of the two weeks prior to the soils returning to field capacity. The most likely explanation for this poor agreement is that the assumptions implicit in the model have been broken, probably by runoff or interflow occurring when the soil moisture deficit in the total profile is small, less than 30 mm. This may be an exceptional case as it occurs during a period when rainfall occurs on every day for a week, with two large rainfall events early in the period. However, it does show that runoff and/or interflow are processes that can occur at this site. This is not surprising given the nature of the Reading Formation which has layers of clay.

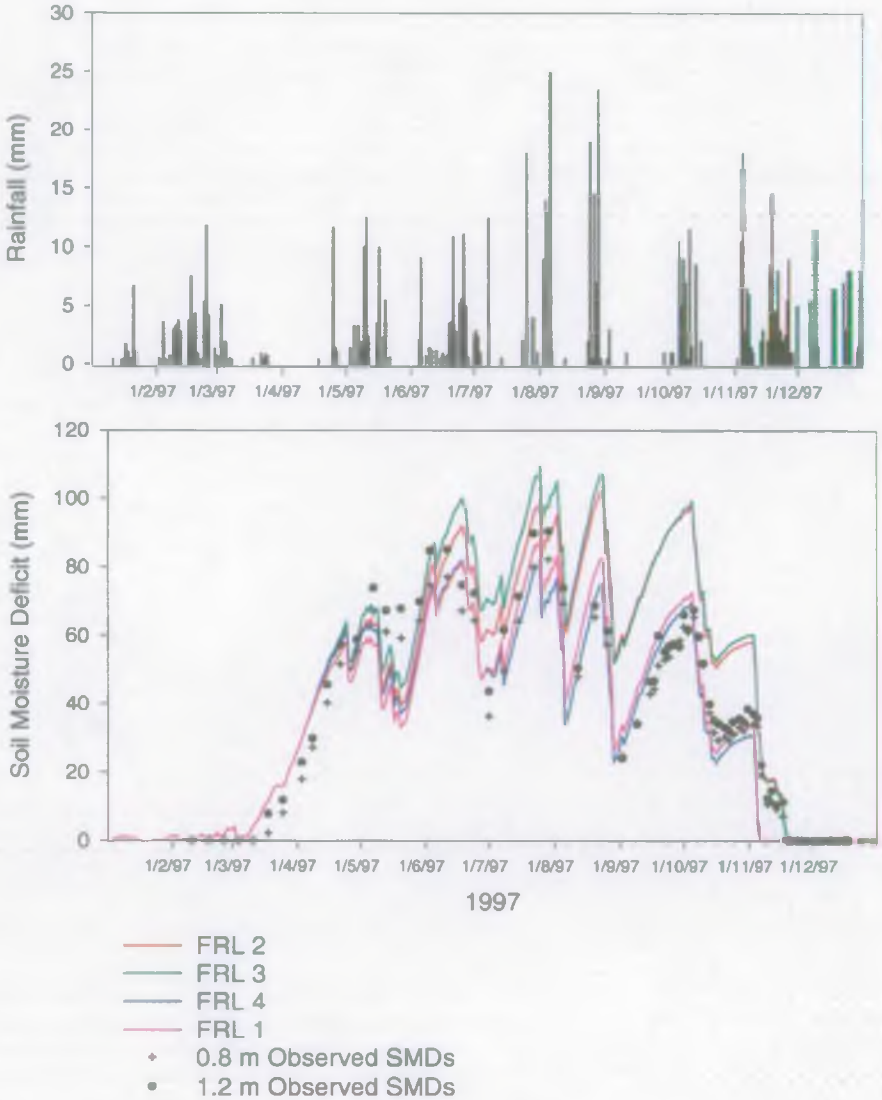


Figure 8.7 Observed soil moisture deficits and those predicted by FRL for Site 9

A feature of the observed soil moisture data is that the differences between the values for the profile down to 0.8 m. and those down to 1.2 m., differ in the main part of the summer and only by 5 to 7 mm. This suggests that the process of water being drawn up from beneath the rooting zone is not very significant at this site.

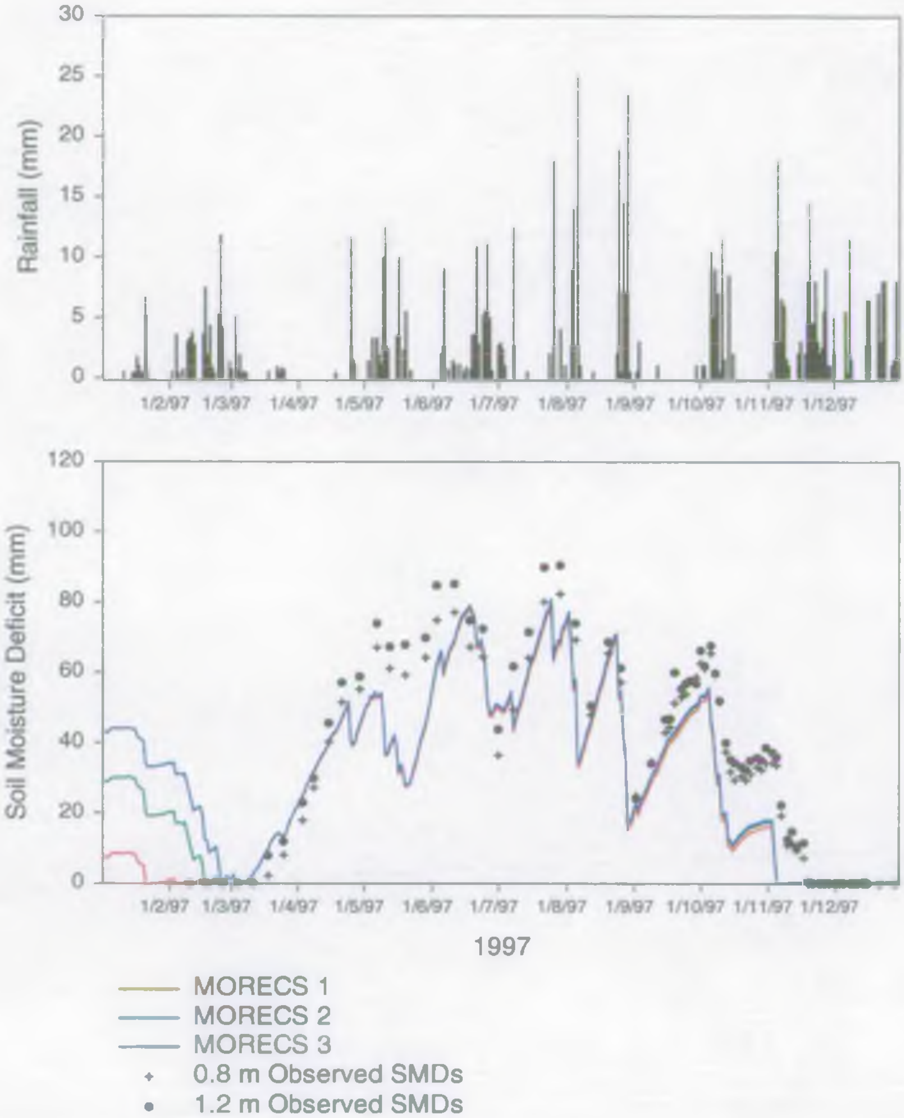


Figure 8.8 Observed soil moisture deficits and those predicted by MORECS for Site 9

The soil moisture deficits predicted by the MORECS model show poor agreement with the observed values, Figure 8.8. There are no differences between the soil moisture deficits predicted by the different available water contents used with the MORECS model. As at Site 4, this is because the model predicts that the evaporation throughout the year continues at the potential rate and is unaffected by the soil moisture deficit developed, as discussed in section 8.1.2. The result is that the model predicts that the soil water content return to field capacity earlier than is observed by about two weeks.

The values of 10 year mean annual recharge predicted by the FRL and MORECS models, at this site, are very comparable, Table 8.2. The differences in the parameters result in estimates that vary by about 19%. However, they must be considered as significantly over-estimating the groundwater recharge that occurs at this site since interflow and/or runoff are likely to be important processes. This illustrates the importance of taking into account these processes in any operational system.

8.4 Tern catchment

8.4.1 Grass on Newport soil series

This combination is found at the Bacon Hall site and represent the situation of a soil developed on the Boulder Clay. In addition to the soil moisture measurements, values of water levels from two piezometers are available, and of daily evaporation from the Hydra. The fractional available water contents used for each layer in the FRL model, starting from the surface, are: 0.170, 0.157, 0.092, 0.092. The rooting depths for the model runs of FRL and the corresponding available water contents for MORECS are given in Table 8.5.

Table 8.5 Soil water model parameters for Bacon Hall

	Rooting depth (m)		Available water content (mm)
FRL 1	0.8	MORECS 1	102
FRL 2	1.0	MORECS 2	128
FRL 3	1.2	MORECS 3	205
FRL 4	1.2		

The observed soil moisture deficits are poorly predicted by all the FRL model runs, Figure 8.9. It occurs throughout the period when soil moisture deficits are developed so it is unlikely that this is due to interflow and/or runoff occurring. The most likely options are therefore that either evaporation is being over-estimated or that the value used for the fractional available water content of the soil is too high. The predicted values of evaporation agree well with the values measured by the Hydra until mid-November, Figure 8.10, which suggests that, through the summer months, the values predicted by the model are likely to be appropriate. This is also confirmed by the calculations of the simple water balance for the site, see section 6.2.2, and by examining the changes in soil water content during periods of negligible rainfall, e.g. September. Thus, by a process of elimination, the most likely explanation is that the fractional available water content, obtained from the SEISMIC database, is too high for this site.

The general trends of the soil moisture deficits predicted by the FRL model agree quite well with the observed data. The discrepancies become significant in July and August as a result of an extended period of rainfall which results in the soils returning to field capacity. This is not duplicated by the models and hence the soil moisture deficits predicted by the model are larger than observed from this period on. This results in the models predicting a return to field capacity later than is observed, with a consequent delay in the onset of recharge. A short period of recharge in mid-October is not duplicated by the models. This is shown when the changes in water level at the deeper piezometer at this site are examined, Figure 8.11. Although there is scatter in the points, there are suggestions of a rise in the water table in the middle of October, but the onset of prolonged recharge in the middle of November is clearly marked and agrees with the observed soil water contents returning to field capacity at this time. This also suggests that runoff and/or interflow were not occurring during the period that measurements were being made.

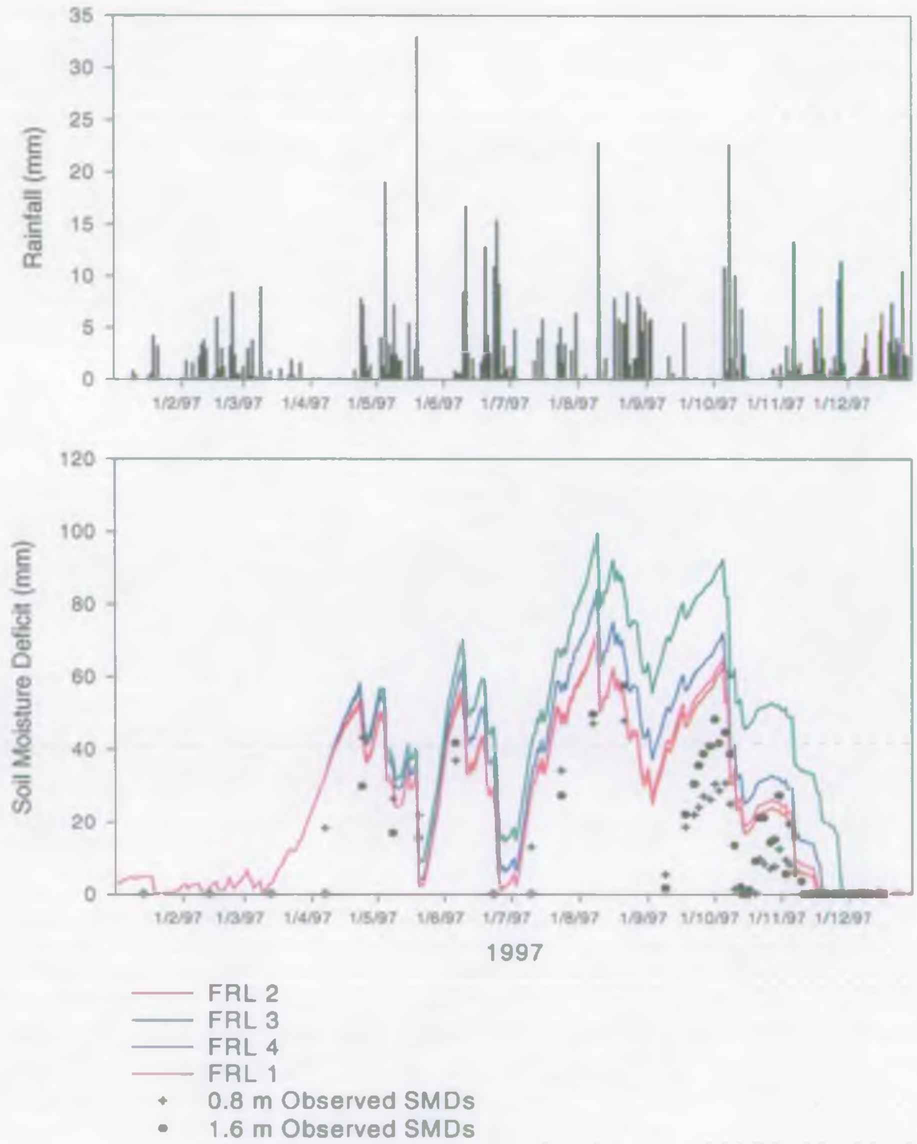


Figure 8.9 Observed soil moisture deficits and those predicted by FRL for Bacon Hall

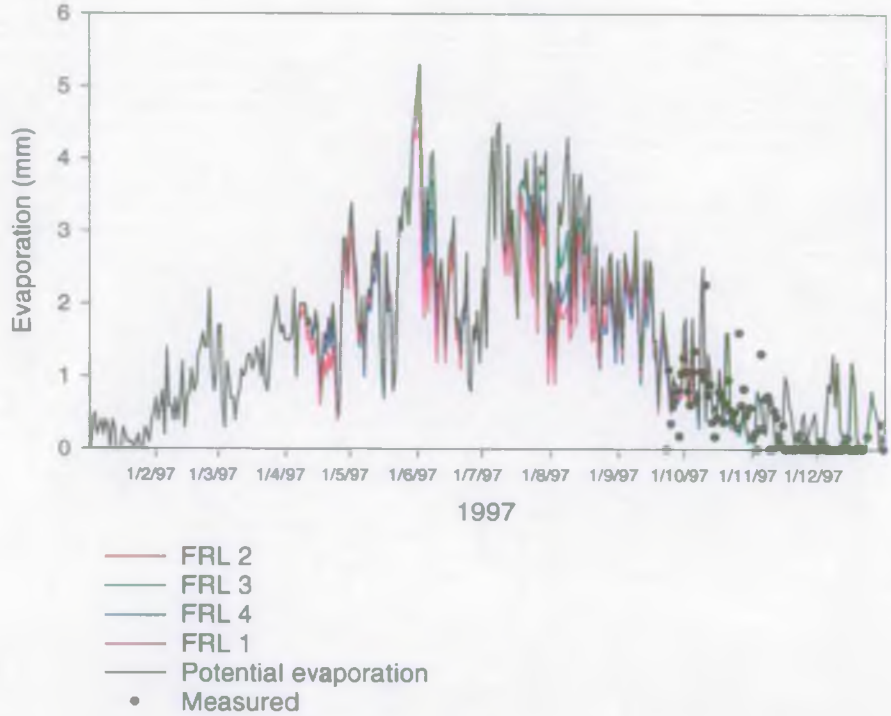


Figure 8.10 Observed evaporation and that predicted by FRL at Bacon Hall

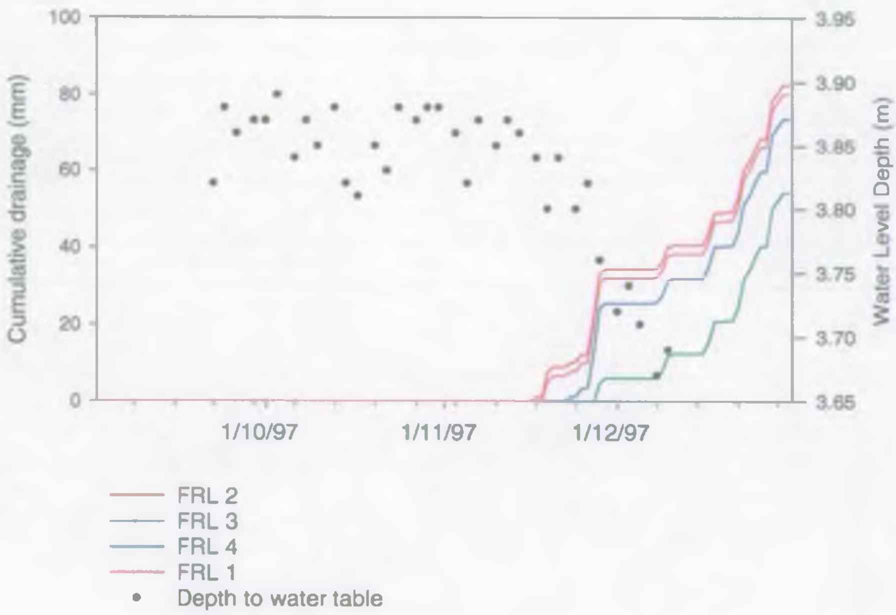


Figure 8.11 Soil drainage predicted by FRL and observed water levels at Bacon Hall

The MORECS soil water model poorly reproduces the observed soil moisture deficits, Figure 8.12. This is in part due to the values used for the available water contents being too high, as with the FRL model. However, the MORECS evaporation model predicts evaporation rates higher than those predicted by the evaporation model used with the FRL model and tending to be higher than those measured by the Hydra, Figure 8.13. The combination of these two factors results in large soil moisture deficits being predicted by the MORECS model. This is particularly a problem for case 3, with the highest available water content, as this predicts that the soils did not return to field capacity over the winter of 1996/97 resulting in very large deficits throughout 1997 and no recharge.

The impact of these factors on estimates of mean annual groundwater recharge are illustrated in Table 8.2. The estimates range from 74 to 200 mm, a considerable level of uncertainty. In practice, it is likely that the higher values are nearer to the actual situation because the model runs have consistently over-estimated the duration of soil moisture deficits.

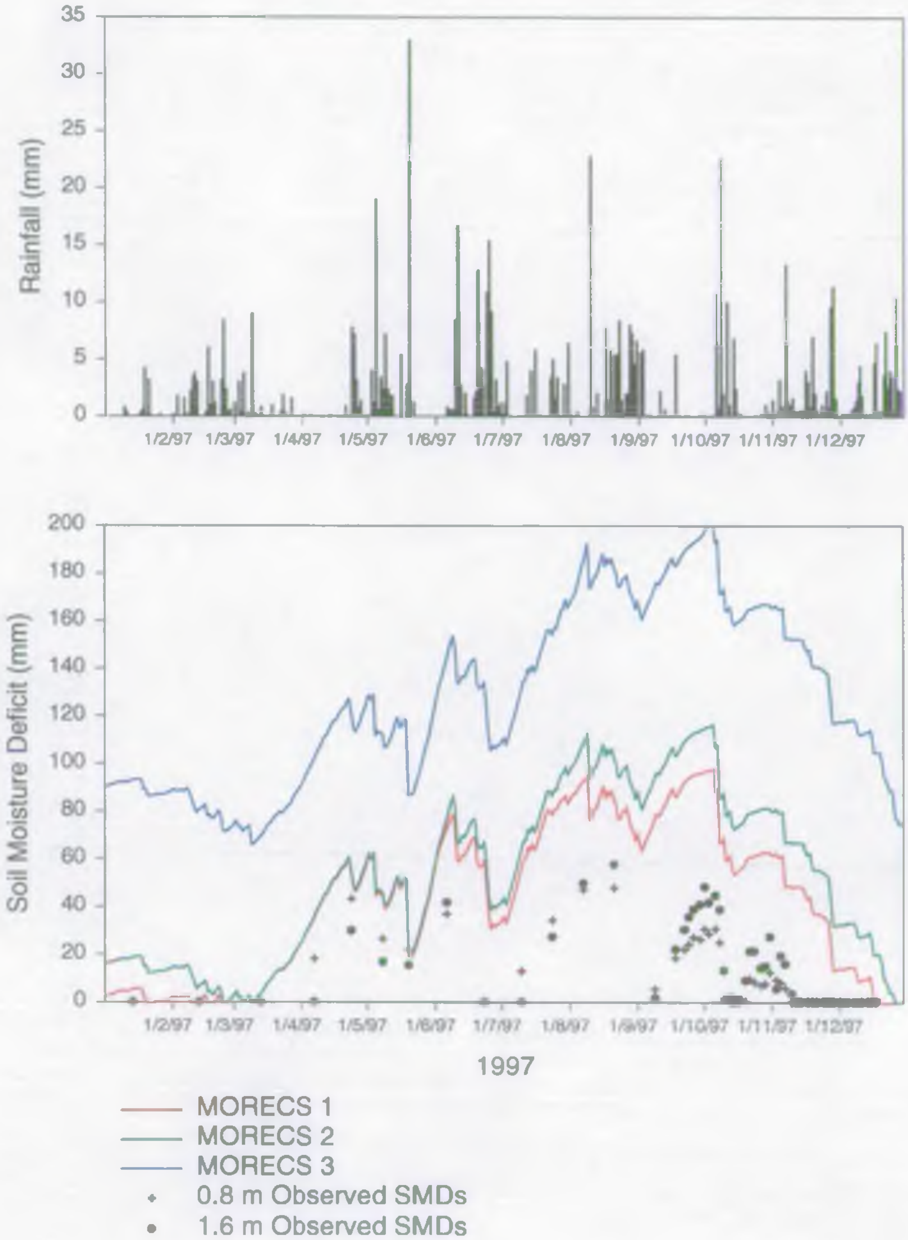


Figure 8.12 Observed soil moisture deficits and those predicted by MORECS at Bacon Hall

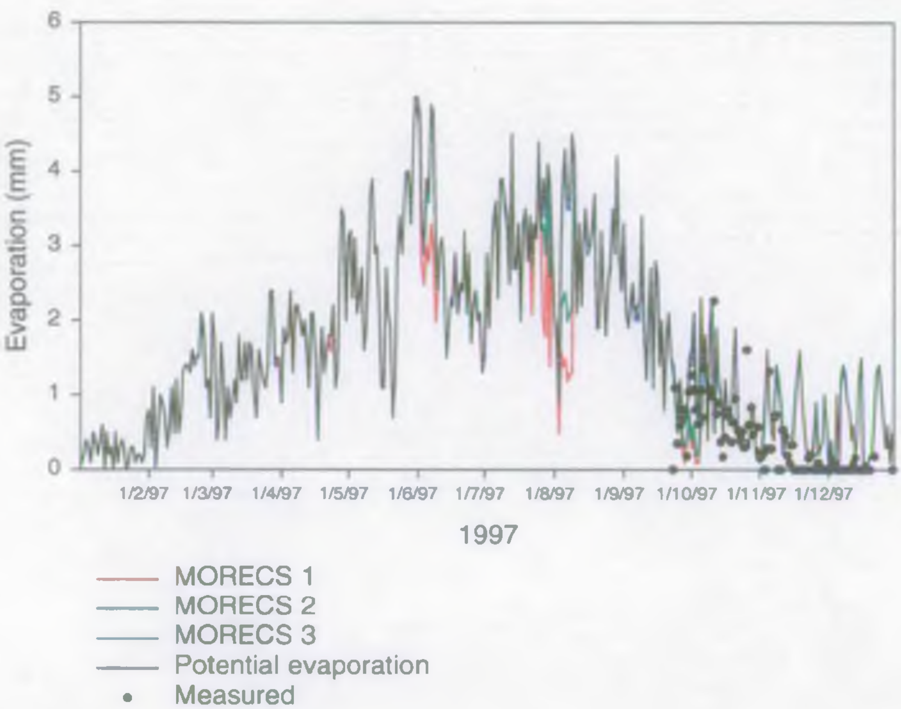


Figure 8.13 Observed evaporation and that predicted by MORECS at Bacon Hall

8.4.2 Grass on Hall soil series

This combination is found at the Stoke Bridge site and represent the situation of a soil developed on the First Terrace deposits. The fractional available water contents used for each layer in the FRL model, starting from the surface, are: 0.193, 0.151, 0.114, 0.114. The rooting depths for the model runs of FRL and the corresponding available water contents for MORECS are given in Table 8.6

Table 8.6 Soil water model parameters for Stoke Bridge

	Rooting depth (m)		Available water content (mm)
FRL 1	0.8	MORECS 1	104
FRL 2	1.0	MORECS 2	143
FRL 3	1.2	MORECS 3	172
FRL 4	1.2		

The FRL model reproduces the observed soil moisture deficits well during the first half of 1997, Figure 8.14. However, from July onwards, all the model runs overestimate the soil moisture deficits. This cannot be explained by inappropriate values for the available water content as this would result in discrepancies throughout the year. For the same reason, it cannot be explained by the evaporation being consistently overestimated. The most likely cause would seem to be a reduction in the vegetation canopy in the middle of the summer. The site is a small plot of grass in an enclosure between the road and a field, in which sugar beet was being grown in 1997. If the grass had been allowed to grow long in the first part of the year and was subsequently mown then the evaporation would be reduced after mowing. The observed increase in soil moisture deficits, during a period in September when there is relatively little rainfall, is generally the same as predicted by the model. This suggests that, at this time, the evaporation rates predicted by the model are in agreement with the observations of soil moisture.

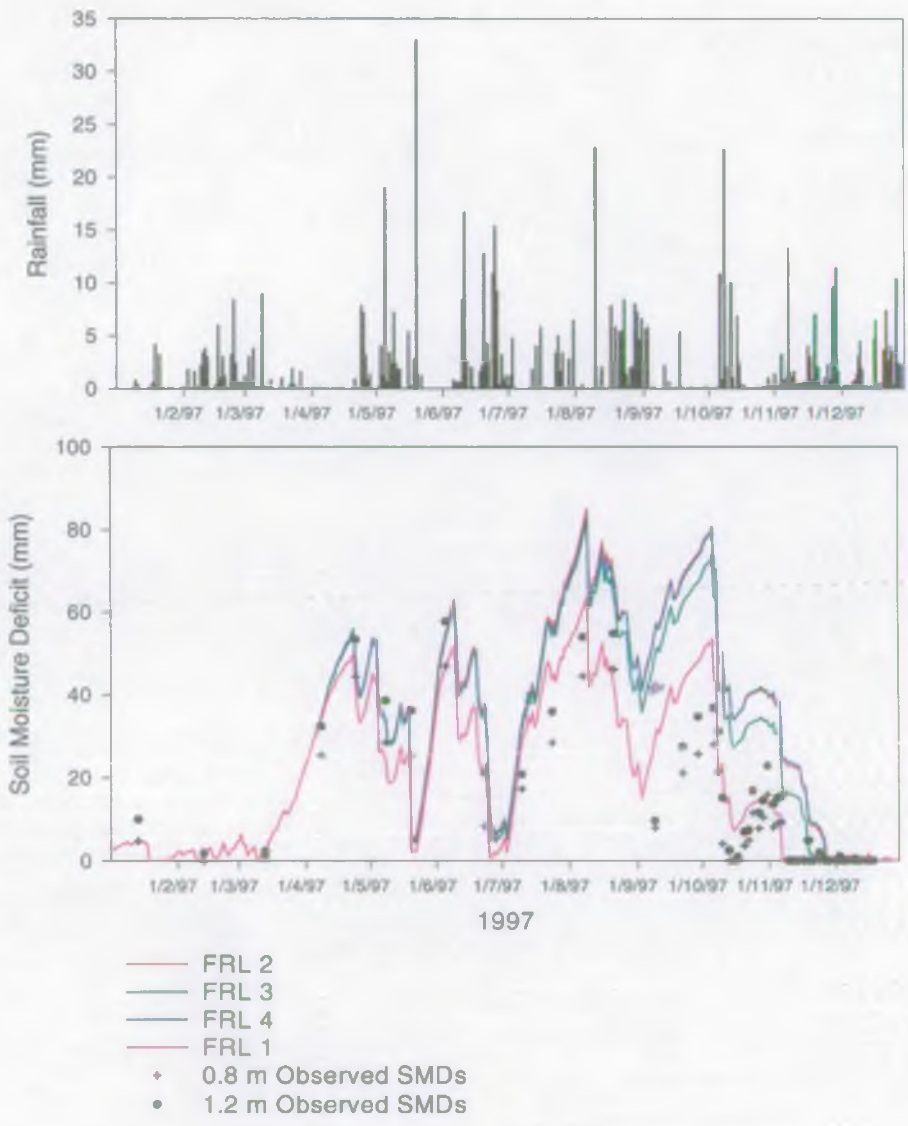


Figure 8.14 Observed soil moisture deficits and those predicted at Stoke Bridge by FRL

The date of the soils returning to field capacity is predicted best by case 1, i.e. with the rooting depth set to that of grass. The model predicts the start of drainage at about the same time as a major increase in water levels are observed in the piezometer, although it does not reproduce the small event in October, Figure 8.15. However, although the model simulates the return to field capacity correctly, it is as a result of the evaporation being incorrectly estimated during the summer and so this is not necessarily the correct model to be used for this site. In fact, given that this case poorly simulates the soil moisture deficits developed over the whole soil profile for the first part of the year, it is likely to be the least model accurate, in terms of groundwater recharge, if evaporation is correctly modelled.

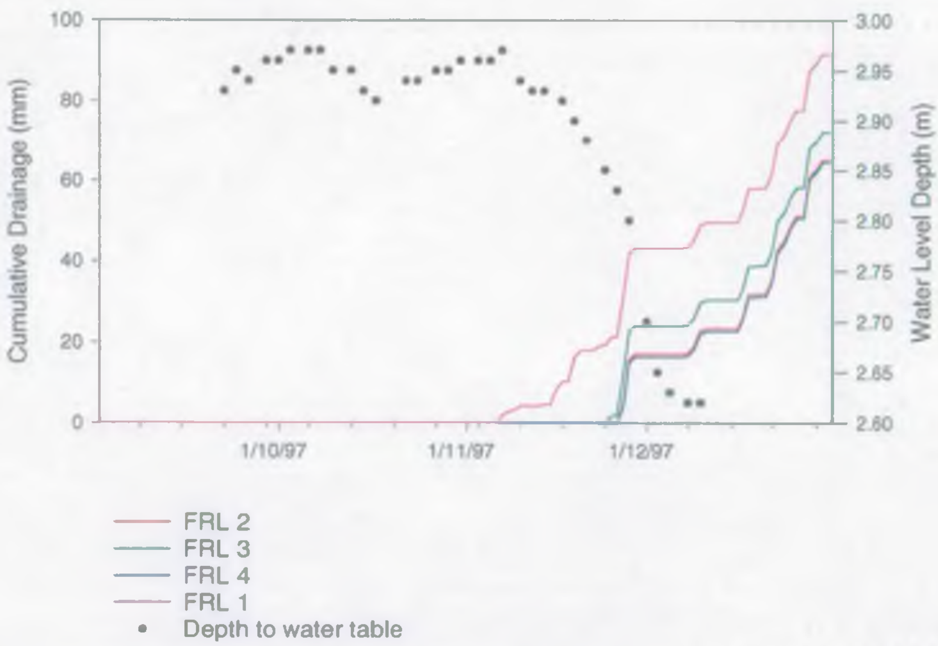


Figure 8.15 Soil drainage predicted by FRL and observed water levels at Stoke Bridge

The agreement between the soil moisture deficits predicted by the MORECS soil water model is poor, Figure 8.16. This is partly because, for two of the cases, the model predicts that the soils do not return to field capacity over the winter of 1996/97. However, for the run where the soils do return to field capacity, i.e. with the lowest available water content, the predicted soil moisture deficits are significantly over-estimated. This results from the high evaporation rates throughout the year predicted by the evaporation model. Consequently the date of the return to field capacity is seriously over-estimated.

The effect of the discrepancies in the models on the predicted 10 year mean annual recharge, Table 8.2, is a large degree of variation with estimates ranging from 90 to 214 mm. The lowest values are almost certainly wrong as they are as a result of the MORECS model predicting that the soils do not always return to field capacity over winter. The model most likely to give the correct results is FRL case 4 as this corresponds best to the conceptual model for the site. This gives a value of 182 mm and is the lowest predicted by the different rooting depths used with the FRL model.

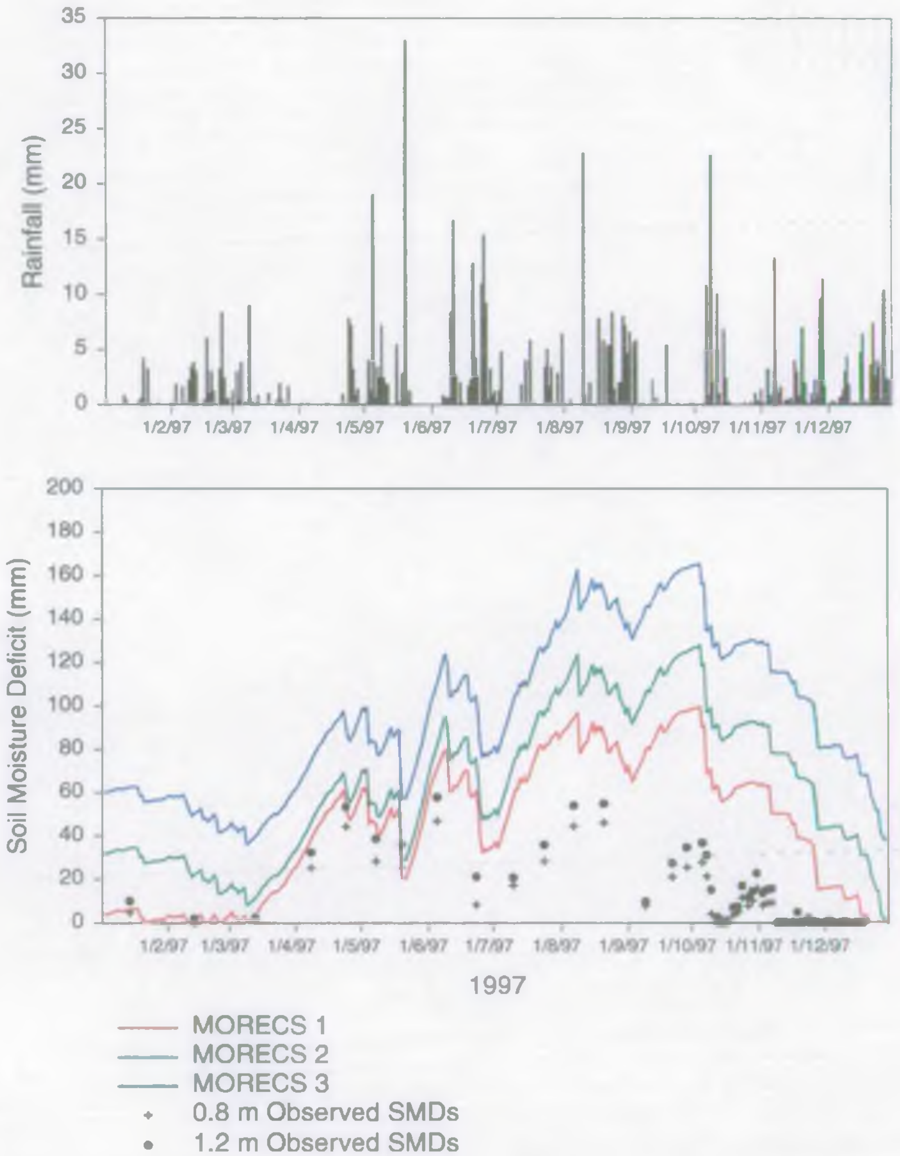


Figure 8.16 Observed soil moisture deficits and those predicted by MORECS for Stoke Bridge

9. SPATIAL HETEROGENEITY

9.1 Summary of results

- Variations in the texture of a mapped soil series units are likely to result in variations in mean annual groundwater recharge of at least ±5%.
- Aggregating soil series of the Pang catchment using the area weighted mean available water content produces values of the mean annual groundwater recharge that do not varying significantly from those achieved by mosaicing.
- Other aggregation methods overestimated the mean annual groundwater recharge by around 5%.
- These results are only applicable to freely draining soils directly overlying an aquifer as runoff and interflow were not considered.

9.2 Background

The preceding analysis and modelling have considered a one-dimensional situation, i.e. that the vegetation, soils and the atmosphere are homogeneous laterally for an infinite distance. This is obviously not a realistic representation, particularly in terms of the land surface, because the landscape is composed of a wide variety of vegetation types growing on disparate soil types, differing at a variety of spatial scales. These differences must be considered in formulating a spatially distributed water balance model for estimating groundwater recharge as some form of strategy is necessary in order to represent this spatial heterogeneity without excessive data and computational time.

All spatially distributed models assume that there is a representative value (often the mean or median value) that, at a particular spatial scale, can be used for a parameter. An example is in modelling the evaporation from an area of coniferous woodland. Single values for the height, stomatal conductance and leaf area index for the whole woodland are generally assumed. It is clear that there will be variations between individual trees and, in the case of the last parameter, within an individual tree. Thus, at these scales, there will be variations in the evaporation but the presumption is that the evaporation of the whole woodland will approximate to the sum of the individual components. This is often referred to as aggregation. It has the advantage of minimising the computation required but risks not giving representative results if the 'real' parameters show a high degree of variability in a systematic manner.

The alternative to aggregation is mosaicing. In this strategy, it is assumed that the spatial variety can be represented by a series of units (tiles), within each of which conditions are homogeneous. An example is if the evaporation from a grid cell (e.g. the 40 x 40 km squares of MORECS) is being modelled. The land cover within the grid cell could be assumed to consist of a series of discrete, homogeneous units (fields, patches of woodland etc.). The evaporation from each of these units can be modelled separately, assuming representative values for the necessary parameters. The evaporation of the grid cell is then assumed to be approximated by the area weighted sum of the evaporation from each tile. This method requires more computation than aggregation but has a better prospect of representing the effects of strong spatial heterogeneity, provided each tile is approximately homogeneous.

In practice many models combine these two methods. For example, the MORECS 40 km grid model uses aggregation for the soil available water content and mosaicing, over 14 different land cover types, for the vegetation canopy parameters. Thus, the median available water content is calculated for each model grid square from all measurements of the available water content within that square. Within each square, the water balance of the different land cover types is calculated separately, using the median available water content, and then summed to give the evaporation, and hydrologically effective precipitation. There are two major assumptions implicit in this strategy. The first is that the distribution of the vegetation types is randomly distributed over the range of soil available water contents in the grid square. The second assumption is that the hydrological processes operating in each soil type are the same. These assumptions are highly questionable from a hydrological perspective but stem from the original concept of MORECS, whose prime purpose was the calculation of evaporation and soil moisture. Thus MORECS does not simulate runoff and does not apportion the hydrologically effective precipitation (soil drainage) between recharge and interflow. In practice, land cover is not random across soil types at this scale because the use man makes of the soil is determined in part by its hydrological characteristics. An example of this is seen in the Pang catchment where the woodland tends to occur on the agriculturally unproductive soils developed on the Reading Formation. The hydrological processes are not equally proportioned. Again, the Pang catchment provides an example as runoff and interflow are negligible on the Chalk soils whereas theses processes are dominant on the soils developed on the clays of the Reading Formation.

9.3 Sensitivity analysis

The sensitivity of a simple water balance model, incorporating the FRL soil water model, to land surface parameters has recently been reported by Finch (1998). He concluded that the model was not particularly sensitive to the vegetation canopy parameters other than to consider them under major land cover types, e.g. annual short vegetation, woodland etc. The model was most sensitive to the soil parameters. Therefore, this study has concentrated on the implications of using an aggregation strategy for the soil available water content, which is the prime parameter required by the FRL model (see section 7.3.4) and the MORECS soil water model. Aggregation at two scales has been examined: within a soil series and within a catchment.

The analysis has been carried out using the FRL soil model with potential evaporation calculated for a land cover of permanent grass, see section 8.2.

In order to ensure consistence of results, the available water content (θ_f-θ_w) has been calculated using the simple pedo-transfer function of Clapp and Harnberger (1978) using the equations of Saxton et al (1986):

$$A = \exp(-4.396 - 0.0715 P_{clay} - 4.880 \times 10^{-4} P_{sand}{}^{2} - 4.285 \times 10^{-5} P_{sand}{}^{2} P_{clay})100$$

$$B = -3.140 - 0.00222 P_{clay}{}^{2} - 3.484 \times 10^{-5} P_{sand}{}^{2} P_{clay}$$

$$\theta_f = \exp((2.302 - \log(A)) / B)$$

$$\theta_w = (1500 / A)^{1/B}$$

where:

P_{sand} Percentage sand grain size
P_{clay} Percentage clay grain size

The texture measures for the various soil series were obtained from the Soil Survey's SEISMIC database (SSLRC, 1995).

9.3.1 Within soil series variations

The analysis has been carried out for the Hornbeam Series soils. These are developed on the Clay-with-Flints in the Pang catchment and are taken to represent a 'typical' soil of the lowlands, i.e. recharge is the dominant process. The SEISMIC database contains both the mean and the standard deviations for the texture measures. These were used to calculate the available water content for each of the four layers of the FRL model for the mean texture measures and also for one standard deviation either side of the mean. The 25 year mean annual recharge was then calculated using the meteorological data from Wallingford and the rainfall data from Compton, Table 9.1

Table 9.1 Available water contents and estimated recharge for within soil series heterogeneity

FRL layer	Mean	P_{sand} + SD P_{clay} + SD	P_{sand} + SD P_{clay} - SD	P_{sand} - SD P_{clay} + SD	P_{sand} - SD P_{clay} + SD
1 AWC	0.232	0.205	0.267	0.266	0.201
2 AWC	0.214	0.194	0.235	0.239	0.187
3 AWC	0.196	0.188	0.215	0.207	0.179
4 AWC	0.227	0.188	0.179	0.133	0.246
Recharge (mm)	244.2	253.1	233.2	232.8	255.4

It can be seen that changes in the texture measures by one standard deviation result in changes in the mean annual groundwater recharge of around 10 mm (5%). Thus this demonstrates that, within an area mapped as a single soil series, variations in the soil texture are likely to result in changes in the modelled mean annual groundwater recharge of around ±5% in the majority of cases (68%) but there will be sites where the variation is even larger. In practice, there is no scope for representing the variation within a soil series in any other way than the aggregated value. This is because the smallest unit generally mapped by the Soil Survey is the soil association. It is in effect the occurrence of a soil series but with small inclusions of other soil series which do not warrant mapping separately at that scale. Thus the actual variability in groundwater recharge within a unit mapped as a soil association will be greater than given here due to these inclusions. Thus, it is possible to state that the variation in mean annual groundwater recharge estimated within a mapped soil unit is likely to be a minimum of ±5%.

9.3.2 Within catchment variations

The Pang catchment has been used to test the results from a variety of aggregation strategies. There are 11 soil associations mapped in the Pang catchment, Table 9.2, representing a range of conditions from the highly permeable soils on the Chalk to the soils developed on the low permeability Reading Formation and thus represent a range of soil types.

Table 9.2 Pang catchment soil associations and available water contents

Soil Association	Geological Unit	% area	FRL layer available water content 1	2	3	4
Andover	Chalk	21.9	0.257	0.257	0.257	0.257
Coombe	Valley bottom Chalk	14.9	0.260	0.249	0.254	0.254
Frilford	Reading Formation	0.5	0.124	0.123	0.122	0.121
Frilsham	Chalk	9.1	0.218	0.191	0.191	0.191
Hamble	London Clay Formation	0.9	0.281	0.274	0.282	0.282
Hornbeam	Clay-with-Flints	23.6	0.232	0.214	0.196	0.227
Hurst	River and Valley Gravel	1.3	0.158	0.153	0.120	0.109
Sonning	Bagshot Formation	2.1	0.169	0.139	0.131	0.115
Southampton	Plateau Gravel	2.8	0.151	0.128	0.079	0.107
Thames	Alluvium	2.3	0.145	0.165	0.214	0.170
Wickham	Reading Formation	20.6	0.227	0.220	0.226	0.249

The available water content was calculated from the mean texture measures for each soil from the SEISMIC database. The 25 year mean annual recharge was then calculated using the meteorological data from Wallingford and the rainfall data from Compton. A variety of procedures were used to estimate the average direct groundwater recharge for the catchment:

1. Area weighted mean recharge - the recharge estimated for the individual soils multiplied by the fraction of the catchment they occupied and then summed, i.e. the mosaicing method.
2. Mean available water content - the mean available water content of all the soils was used to parmeterise the FRL model.
3. Median available water content - the median available water content of all the soils was used to parmeterise the FRL model.
4. Area weghted mean available water content - the mean of the available water content of the soils multiplied by the fraction of the catchment they cover was used to parmeterise the FRL model.
5. Mean texture - the mean of the texture measures was computed and then the available water content calculated for input to the FRL model.
6. Area weighted mean texture - the mean of the texture measures of the soils multiplied by the fraction of the catchment they cover was computed and then the available water content calculated for input to the FRL model.

The average recharge for the catchment calculated using each method is given in Table 9.3. The area weighted mean recharge is assumed to be the most reliable and as such is the standard against which the other methods are judged. There is no difference in the results between the methods using the texture measures and those using the available water content. This confirms that the pedotransfer function is linear and thus there is no preference to calculating the average texture measure over the mean available water content derived from these texture measures. Not weighting the calculation of the mean by the fraction of the catchment occupied by the soils results in overestimates for the mean annual recharge of about 12 mm (5%).

Table 9.3 Mean annual recharge for the Pang catchment estimated using different aggregation schemes

Aggregation Procedure	Mean Annual Recharge (mm)
Area weighted mean recharge	243
Mean AWC	253
Median AWC	251
Area weighted mean AWC	243
Mean texture	255
Area weighted mean texture	242

The results show that there is no significant between using the mosaicing method and an aggregation method based on the area weighted mean available water content. Thus, a significant decrease in the computational time could be achieved by the latter with no loss of accuracy. However, this conclusion needs to be applied with caution as this has not been an exhaustive trial. The first limitation is that only a land cover of grass has been considered and so the conclusions do not necessarily apply to other land cover types. The second, and more serious limitation, is that the model does not consider either runoff or interflow. The implications of this are demonstrated by the results for the Wickham soils, which are developed on the Reading Formation and thus recharge can be expected to be low, as the model predicts a mean annual recharge of 244 mm. Both the FRL and MORECS soil model can only be used to estimate direct groundwater recharge on freely draining soils overlying an aquifer without the inclusion of additional models for runoff and interflow. Clearly, these conditions do not hold for some of the soils in the Pang and Tern catchment. The inclusion of additional models for runoff and to partition the soil drainage between recharge and interflow are necessary before more definitive tests of aggregation procedures can be made. Nevertheless, this study has shown that the selection of an inappropriate aggregation procedure can result in errors in estimating groundwater recharge of around 5% and that this is probably a minimum value. The inclusion of runoff and interflow models is likely to increase this value.

10. CONCLUSIONS AND RECOMMENDATIONS

This study was set up with the following objectives:

- To collect a detailed data set of changes in soil moisture that occur at the end of a prolonged dry summer as the soils wet up in the autumn.
- To analyse these data to achieve a conceptual model of the processes involved in the wetting up.
- To review how well the changes in soil moisture can be duplicated by simple soil models and to suggest how these could be improved.
- To test the impact of spatial heterogeneity on spatially distributed estimates of recharge.

Soil moisture has been the key element of this study because drainage, and thus recharge, cannot generally take place whilst a soil moisture deficit exists. Hence it is important that the water balance models, used to estimate direct groundwater recharge, are able to predict the duration (and magnitude) of deficits accurately. Thus, this project has made measurements of soil moisture and tested models against these data.

The objectives of this study have been achieved by carrying out a programme of field measurements in the Pang and Tern catchments. The conclusions, in the form of the conceptual model developed and the results of comparing the soil moisture deficits predicted by the MORECS and FRL soil models are given below.

10.1 Conclusions

This study has again shown the exceptional nature of Chalk soils, such as the Andover and Coombe series. Soil moisture deficits can develop to much greater depths, in excess of 3m under cereals, in these soils compared to other soils. The unsaturated hydraulic properties of the Chalk are such that the hydraulic conductivity of the matrix changes little with changes in soil water potentials. The result is that water continues to flow easily through the soil as the water content decreases. Water can be drawn up (capillary rise) from below the rooting depth (0.8 to 1 m for grass and cereal crops) in response to increasing potentials in the root zone as a result of water being extracted by plants for evaporation, Figure 10.1. In addition rainfall is able to percolate through the soil profile in the presence of a soil moisture deficit making the concept of field capacity not truly valid.

The implications for estimating direct groundwater recharge from simple water balance models are profound. The lack of agreement between the observed soil moisture deficits and those predicted by the models in terms of the soils wetting up and returning to field capacity can be explained by the depths to which the deficits are developed. In the past, the models have been used with the assumption that deficits only occur in the root zone (0.8 to 1 m for grass and cereal crops). Hence, they have underestimated the total soil moisture deficit, and so predict that less rainfall is required to make good these deficits, with the result that they over estimate recharge. The error in estimating mean annual groundwater recharge is likely to be in the region of 10-15%. The effect is sufficiently large that it can not be convincingly duplicated by a single layer soil model such as used by MORECS. A multi-layered model such as FRL can be made to approximate the observed soil moisture deficits by reducing the proportions of roots in the bottom three layers to low values (0.01 to 0.05) but this could reduce the advantage of the model in terms of evaporation from a bare soil phase of the land cover, i.e. cereal crops which are often grown on Chalk soils. Ideally, the process needs to be explicitly incorporated in a model although the representation should not be overly complex.

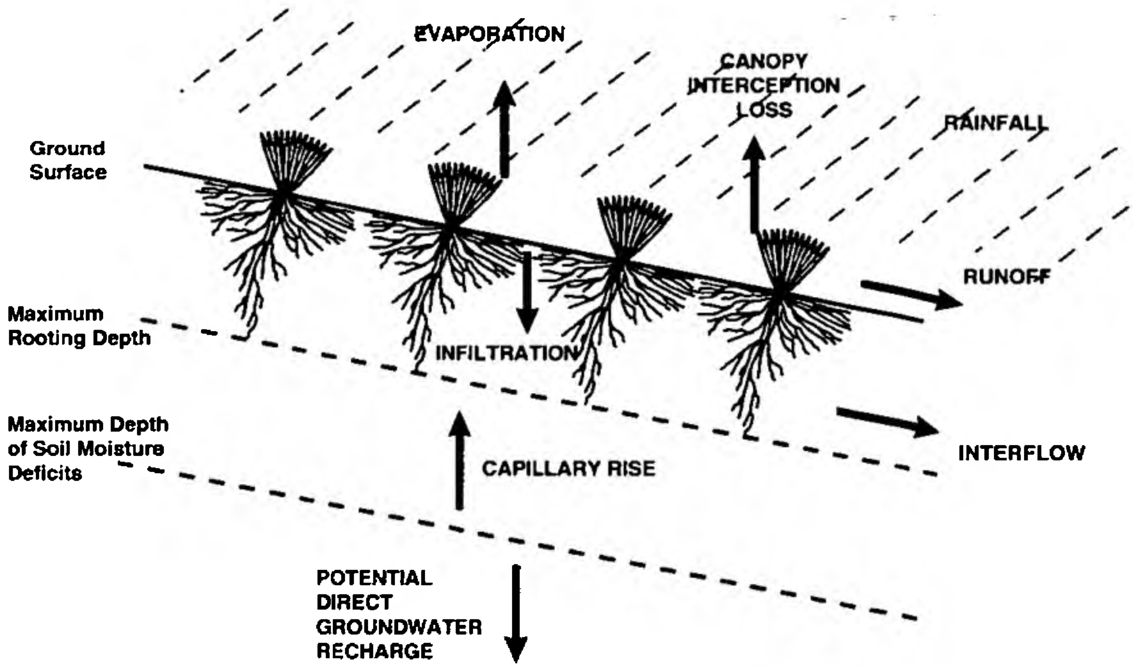


Figure 10.1 Schematic diagram of the conceptual model of direct groundwater recharge

Other soils, such as the Hornbeam series in the Pang catchment and the Newport and Hall series in the Tern, exhibit a similar feature of soil moisture deficits being developed below the rooting zone. However, the effect is less marked with soil moisture deficits not extending below a depth of 1.6 m, i.e. between 0.5 and 0.8 m below the rooting depth. The result will still be that models will overestimate the mean annual groundwater recharge but by less, probably around 5%.

This study has shown that there is no need to invoke the concept of bypass flow to explain the water balance of Chalk soils at the sites in this study. Hence, it has been suggested that this process may only operate where the soil water has been made acid by overlying soils and widened fissures in the Chalk by dissolution and so is a localised phenomena.

It is noted that Chalk soils are exceptional because rainfall can percolate through the soil profile in the presence of a soil moisture deficit, with the result that drainage, and hence recharge, can occur potentially in the summer and autumn.

Care is needed to ensure an accurate value is used for the available water content of the soil and the rooting depth. Incorrect values can result in the mean annual groundwater recharge being wrong by as much as 15%.

A multi-layered soil water model has some significant advantages over a single layer model in the particular cases of land covers of annual crops and bare soil. The first advantage is in simulating the development downwards of the roots. This can be readily achieved by changing the proportions of roots in the layers whilst in a single layer model it has to be done by changing the available water content. However, this is unlikely to have a major impact on estimates of groundwater recharge because, in the UK, soil moisture deficits will rarely develop, during the growing season, to an extent that they will limit growth or evaporation. This applies to the climatic conditions of the last 25 years and climate change may make exceptional years, such as 1976 and 1996, more frequent.

A more important feature of the multi-layered model is in terms of simulating the impact of harvest when there is an abrupt transition to evaporation from the soil when the vegetation canopy is removed. The multi-layered model allows the soil moisture deficit throughout the profile to be retained whilst limiting evaporation to only occur from the surface layer. Thus the rapid increase in evaporation following heavy rainfall from the near surface soil can be simulated, whilst retaining the soil moisture deficit at depth. This cannot be achieved convincingly with a single layer model. This is likely to be important given that large areas of the outcrops of the major aquifers are used to grow annual crops.

A second issue relating to bare soil that needs to be addressed is what value(s) should be used for the bulk canopy resistance (in the Penman-Monteith evaporation model and other similar ones, see section 3.2) for bare soil. The measurements of evaporation made in this study have shown that the value of 100 ms^{-1} used in MORECS is more appropriate than the value of 500 ms^{-1} used with the FRL model. The difference in the values resulted in a difference in the estimated mean annual groundwater recharge of around 60%. However, little research has been carried out on whether the concept of a bulk canopy resistance is valid for bare soil and, if it is, what values it takes and how it can be incorporated within a soil water model.

The study has confirmed that the evaporation from grass can increase rapidly after rainfall, despite the presence of a significant soil moisture deficit. A multi-layered soil water model.is able to simulate this more easily than a single layer model. Evaporation from grass was limited by soil moisture availability once the deficits exceeded 70 to 100 mm.

Both the FRL and MORECS soil water models do not take any account of runoff or interflow. This is acceptable for the high permeability soils developed on the Chalk and does not seem to be a serious limitation for the others such as the Hornbeam, Hall and Newport series. However, it is a serious limitation if these models are applied to other, less permeable soils. Runoff is generated if the storage capacity of the soil at the surface is exceeded by the rainfall. This can happen if the soils are nearing saturation following heavy rainfall, possibly over a few days. This can be included more easily in a multi-layered model than a single layer model. Incorporating a module that partitions the soil drainage between recharge and interflow will be more of a challenge as it will require information on the slope of the land and the hydraulic properties of the substrate below the hydrological soil layer.

Another feature that is likely to be important in these soils is the rate of drainage through the soil. Both MORECS and FRL assume that water content in excess of field capacity drains away within a day. The data from some of the sites involved in this study have shown that this is not necessarily true. This feature is important as the slower drainage of the soil could result in more runoff. Because runoff and interflow represent water unavailable for direct recharge it is important to be able to accurately quantify them.

The data from the site in the Pang catchment consisting of both grass and deciduous woodland has emphasised the importance of taking into account interception losses when the land cover is woodland. Interception losses may result in the mean annual groundwater recharge under deciduous woodland being half those under grass. At this site, it appears that the presence of woodland has resulted in the field capacity of the soils being significantly reduced. However, there is only a small change in the available water content and so the impact of the change in the soil hydraulic properties on estimates of groundwater recharge is small. It should be pointed out that these conclusions are based on data from a single soil moisture tube and more work is needed to understand the impact of lowland forestry on groundwater recharge.

The spatial variability within a soil series is likely to result in a variability of mean annual groundwater recharge by a minimum of ±5%. Similarly, the selection of an inappropriate aggregation scheme can lead to errors of at least 5%. For example, for the soils present in the Pang catchment, use of a mean available water content resulted in an overestimate of around 5 %, whilst the error was insignificant for an aggregation scheme based on the area weighted mean available water content.

The study has raised some concerns about the estimation of net radiation from records of sunshine hours which may need some further study. For the period 24 September to 31 December 1997, there was a difference of 17% between the cumulative net radiation measured by the automatic weather station and that calculated from sunshine hours.

10.2 Recommendations

It is estimated that the predictions of groundwater recharge could increase in accuracy by approximately 10 to 20% if the following recommendations are acted upon.

- A groundwater recharge model should include a multi-layered soil water model.
- The process of water being drawn up from beneath the rooting depth should be explicitly represented in the soil water model.
- Release of water to drainage over periods greater than one day should be incorporated in the soil water model.
- A runoff model should be incorporated in a groundwater recharge model.
- Partition of soil drainage between recharge and interflow should be incorporated in a groundwater recharge model (this will involve consideration of the soil and substrate hydraulic properties and the slope of the land surface).
- A canopy interception model should be included for deciduous and coniferous woodland.
- Research should be carried out into an appropriate model formulation to represent the process of rainfall percolating through the Chalk soil profile in the presence of soil moisture deficits.
- Research should be carried out to determine an appropriate model formulation and the values of parameters to represent evaporation from bare soil.
- Research should be carried out into the impact of deciduous woodland on recharge for a variety of soil types. This should include the potential changes in soil hydraulic properties, the rooting depth and an appropriate model representation.

11. REFERENCES

Allen, R.G., Smith, M., Pereira, L.S., Perrier, A. 1994. An update for the calculation of reference evaporation, ICID Bulletin, 43, 35-92.

Banks, D., Davies, C., Davies, W. 1995. The Chalk as a karstic aquifer: evidence from a tracer test at Stanford Dingley, Berkshire. UK, Q.Jl.Eng.Geol., 28, S31-S38

Bell, J. P., 1987. Neutron probe practice, Report 16, NERC, Institute of Hydrology, Wallingford, 51 pp

Boorman, D. B., Hollis, J.M., and Lilly, A., 1995. Hydrology of soil types: a hydrologically based classification of the soils of the United Kingdom, 126, Institute of Hydrology

Cain, J.D., Batchelor, C.H., Gash, J.H.C., Harding, R.J. 1998. Comment on the paper 'Towards a rational definition of potential evaporation' by J.P. Lhomme, Hydrology and Earth System Science, 2, 137-138.

Clapp, R.B., Hamberger, G.M. 1978. Empirical equations for some soil hydraulic properties, Water Resour.Res., 14, 601-604.

Finch, J.W. 1998. Estimating direct groundwater recharge using a simple water balance model - sensitivity to land surface parameters, J.Hydrol., 211, 112-125.

Finch, J. W., Calver, A.F., Harding, R.J., and Ragab, R., 1995. National system for groundwater recharge assessment, R&D Project Record 499/1/A, Environment Agency, UK, 110 pp

Finch, J.W., Harding, R.J. 1998. A comparison between reference transpiration and measurements of evaporation for a riparian grassland site, Hydrology and Earth System Science, 2, 101-108.

Gash, J.H.C., Lloyd, C.R., Lachaud, G. 1995. Estimating sparse forest rainfall interception with an analytical model, J.Hydrol., 170, 79-86.

Gregory, P.J. 1989. Depletion and movement of water beneath cereal crops grown on a shallow soil overlying chalk, J.Soil Sci., 40, 513-523.

Grindley, J. 1967. The estimation of soil moisture deficits, Meteorol.Mag., 96, 97-108.

Grindley, J. 1969. The calculation of evaporation and soil moisture deficit over specified catchment area, Hydrological Memorandum 28, Meteorological Office, Bracknell, UK, 10 pp

Hough, M.N., Jones, R.J.A. 1998. The United Kingdom Meteorological Office rainfall and evaporation calculation system: MORECS version 2.0 - an overview, Hydrology and Earth System Science, 1, 227-239.

Howard, K.W.F., Lloyd, J.W. 1979. The sensitivity of parameters in the Penman evaporation equations and direct recharge balance, J.Hydrol., 41, 329-344.

Jamieson, P.D., Francis, G.S., Wilson, D.R., Martin, R.J. 1995. Effects of water deficits on evapotranspiration from barley, Agricultural and Forest Meteorology, 76, 41-58.

Miller, C.T., Williams, G.A., Kelley, C.T., Tocci, M.D. 1998. Robust solution of Richards' equation for nonuniform porous media, Water Resour.Res., 34, 2599-2610.

Monteith, J.L. 1965. Evaporation and environment. In: 19th Symposium of the Society of Experimental Biology, Cambridge University Press, Cambridge, UK, pp. 205-234.

Monteith, J.L., Unsworth, M. 1990. Principles of environmental physics, Edward Arnold, London, 291 pp

Penman, H.L. 1948. Natural evaporation from open water, bare soil and grass, Proc.Roy.Soc., A193, 120-145.

Rushton, K.R., Ward, C. 1979. The estimation of groundwater recharge, J.Hydrol., 41, 345-361.

Saxton, K.E., Rawls, W.J., Romberger, J.S., Papendick, R.I. 1986. Estimating generalized soil-water characteristics from texture, Soil Sci.Soc.Am.J., 50, 1031-1036.

Shuttleworth, W.J. 1988. Evaporation models in hydrology. In: Land surface evaporation: measurement and parameterization, Schmugge, T.J. and Andre, J. (Eds.), Springer-Verlag, New York, pp. 93-120.

Shuttleworth, W.J. 1993. Evaporation. In: Handbook of Hydrology, Maidment, D.R. (Eds.), McGraw-Hill, New York

Shuttleworth, W.J., Gash, J.H.C., Lloyd, C.R., McNeil, D.D., Moore, C.J., Wallace, J.S. 1988. An integrated micrometeorological system for evaporation measurements, Agric.For.Meteorol., 43, 295-317.

Shuttleworth, W.J., Wallace, J.S. 1985. Evaporation from sparse crops - an energy combination theory, Quart.J.Roy.Met.Soc., 111, 839-855.

SSLRC, 1995. SEISMIC user manual, Soil Survey and Land Research Centre, Silsoe, Beds., 105 pp

Stockle, C.O., Jara, J. 1998. Modeling transpiration and soil water content from a corn (*Zea Maize* L.) field: 20 min vs. daytime integration step, Agric.For.Meteorol., 92, 119-130.

Wellings, S.R. 1984. Recharge of the upper Chalk aquifer at a site in Hampshire, England 1. Water balance and unsaturated flow, J.Hydrol., 69, 259-273.

Wellings, S.R., Bell, J.P. 1980. Water and nitrate fluxes in unsaturated upper chalk at Bridgets experimental husbandry farm, Winchester, UK. In: The influence of man on the hydrological regime with special reference to representative and experimental basins - Proceedings of the Helsinki symposium, IAHS-AISH, pp. 309-314.

Wellings, S.R., Bell, J.P. 1982. Physical controls of water movement in the unsaturated zone, Q.Jl.Eng.Geol., 15, 235-241.

Wellings, S.R., Cooper, J.D. 1983. The variability of recharge of the English Chalk aquifer, Agric.Wat.Man., 6, 243-253.

Wilby, R., Greenfield, B.J., Glenny, C. 1994. A coupled synoptic-hydrological model for climate change impact assessment, J.Hydrol., 153, 265-290.

Yamanaka, T., Takeda, A., Shimada, J. 1998. Evaporation beneath the soil surface: some observational evidence and numerical experiments, Hydrol.Processes, 12, 2193-2203.

12. ANNEXE 1 - List of symbols

A	pedotransfer function parameter
AWC	available water content (mm)
B	pedotransfer function parameter
C_c	canopy capacity per unit area of cover (mm)
D	bypass flow (mm)
E	evaporation (kg m^{-2} d^{-1})
$\overline{E}_c$	mean evaporation rate during rainfall (mm d^{-1})
E_{aj}	actual root water uptake for the j^{th} layer (mm)
E_p	potential total root water uptake (mm)
F_j	flow of water downwards from the j^{th} layer of the FRL model
G	soil heat flux (MJ m^{-2} d^{-1})
I	interception loss (mm)
L	leaf area index
P	rainfall (mm)
$\overline{P}$	mean rainfall rate (mm d^{-1})
P_{sand}	percentage sand grain size
P_{clay}	percentage clay grain size
P_f	daily rainfall after interception loss has been subtracted (mm)
P_c	minimum rainfall required before bypass flow occur (mm)
P_e	effective rainfall, i.e. infiltration (mm)
P_G	storm rainfall (mm)
P'_G	threshold rainfall necessary to saturate the canopy (mm)
R_j	fractional proportion of roots in the j^{th} layer
R_n	net radiation flux density at the surface (MJ m^{-2} d^{-1})
R_o	surface runoff (mm)
R_i	interflow (mm)
S_j	soil water stress term for the j^{th} layer of the FRL model
S_t	trunk storage capacity (mm)
SMD	soil moisture deficit (mm)
T	evaporation from trunks (mm)
U_z	mean wind speed at height z (m s^{-1})
Z_j	layer thickness for the j^{th} layer (mm)
a	factor used by MORECS to scale surface resistance
b	proportion of effective rainfall going to bypass flow
c	canopy cover
c_p	specific heat of moist air (kJ kg^{-1} $^{o}C^{-1}$)
d	zero plane displacement of wind profile (m)
e_a	saturation vapour pressure (kPa)
e_d	saturation vapour pressure computed at dew point (kPa)
f	proportion of rainfall intercepted by the vegetation canopy
g	proportion of available water content consisting of easily available water
h_c	height of vegetation canopy (m)
k	von Karman constant
p_t	proportion of rainfall diverted to stemflow
r_a	aerodynamic resistance (s m^{-1})
r_s	bulk surface resistance of the vegetation canopy (s m^{-1})

z_h	height of air temperature and humidity measurements (m)
z_m	height of wind speed measurement (m)
z_{oh}	roughness parameter for heat and water vapour (m)
z_{om}	roughness parameter for momentum (m)
Δ	slope of the saturation vapour pressure curve (kPa $^{\circ}C^{-1}$)
ΔS	change in soil water content (mm)
Δe	vapour pressure gradient (kPa m^{-1})
Δt	temperature gradient ($^{\circ}C$ m^{-1})
λ	latent heat of vaporisation (MJ kg^{-1})
γ	psychrometric constant (kPa $^{\circ}C^{-1}$)
ρ	atmospheric density (kg m^{-3})
θ_d	soil water content at which evaporation begins to be limited
θ_f	soil water content at field capacity
θ_w	soil water content at wilting point
θ_j	current fractional soil water content for the j^{th} layer
$\theta_{d,j}$	fractional soil water content at which evaporation falls below potential for the j^{th} layer
$\theta_{f,j}$	fractional soil water content at field capacity for the j^{th} layer
$\theta_{w,j}$	fractional soil water content at wilting point for the j^{th} layer

13. ANNEXE 2 - The field sites

13.1 Pang Catchment

Neutron probe access tubes were installed at three sites in the Pang catchment early in 1997 as part of a study to determine the feasibility for monitoring the spatial distribution of soil moisture from data acquired by spaceborne synthetic aperture radars. Subsequently a fourth site was established in the north of the catchment for the purpose of determining the water balance of cereal crops on Chalk soils. A Campbell Scientific Bowen Ratio machine was also located at this site as part of that study. The OS grid references of the sites are given in Table 14.1.

Table 14.1 Location of soil moisture measurement sites in the Pang catchment

Name	Easting	Northing
Site 10 tube 1	451030	181730
Site 10 tube 2	451320	181870
Site 10 tube 3	451750	182070
Site 20 tube 4	451610	179450
Site 20 tube 5	451740	179380
Site 20 tube 6	451930	179220
Site 30 tube 7	452140	172430
Site 30 tube 8	452190	172480
Site 30 tube 9	452180	172610
Site 40 tube 10	451680	184670

13.1.1 Site 10

Plate 4 The location of Site 10

Three neutron probe access tubes were installed at this site in a field which is located in the north of the catchment The site is on Andover Series soils, a thin soil developed on the Chalk, Plate 4. Winter barley had been sown in 1996 and was harvested in August 1997. All three tubes were removed immediately prior to the harvest but were re-installed at the same positions after another crop of winter barley had been planted in August of the same year. Measurements ceased in July 1998, immediately prior to harvest, when the tubes were removed. The site is on a north-east facing slope with tube 1 near the crest of slope, tube 2 in the middle (towards the left side of Plate 4) and tube 3 near the bottom. All three tubes were installed to a depth of 2 m.

13.1.2 Site 20

There are three neutron probe access tubes at this site. The first (4) is located on the crest of a small hill in a small, unmapped patch of Hornbeam Series soils developed in Clay-with-Flints. The field was sown with a two year ley of grass towards the end of 1997. A cut of silage was taken in August 1997 after which sheep grazed the field until January 1998 when the field was ploughed in order to grow a crop of field maize. The second tube (5) is located in the same field but mid way down the hill slope to the east of the first tube, Plate 5. It is located in an Andover Series soil. The third tube (6) is the most easterly and is sited on rising ground in a field of permanent pasture, grazed by cattle in the summer, on an Andover Series soil, Plate 6. All three tubes were initially installed to a depth of 2 m but tube 6 was deepened to a depth of 3.8 m on 22 March 1998. The tubes were installed early in 1997 and measurements began on 3 February 1997. Measurements at tubes 4 and 5 ceased on 8 December 1998.

Plate 5 Location of tube 6 at Site 20

Plate 6 Location of tube 6 at Site 20

13.1.3 Site 30

The soils at the site are sandy loams of the Wickham Series which are developed in the Reading Formation. The Reading Formation consists of clay, sand and loam with local seams and lenses of flint pebbles overlain by highly irregular alternations of stiff, more or less plastic, structureless clays with sand and loam. Two access tubes (8 and 9) were located in a grass field and one (7) in an adjacent area of woodland Plate7. The tubes were spaced approximately 100 m apart. The three tubes were originally installed to a depth of 2 m , but the tube in the woodlands was deepened to 3 m on 9 September 1997 .

Plate 7 Location of Site 30

The field was grazed by cattle during the summer months. It had been seeded with a four year ley of mixed rye grass and clover in September 1994. The woodland consists of mature oak (*Quercus robur*), larch (*Larix decidua*) and sweet chestnut (*Castanea sativa*) and the neutron probe access tube was sited between sweet chestnuts and larches. The land surface slopes gently down from the woodland, through the field, towards the river Pang which lies about one kilometre away.

Measurements began on 10 February 1997 and ceased on 11 August 1998 when the access tubes were removed.

13.1.4 Site 40

This site is located in a large field in the very north of the catchment on an Andover Series soil developed in the Middle Chalk, Plate 8. There is a single access tube, with a depth of 2.8 m, which was installed in June 1997. Measurements began on 19 June 1997. In order to allow access for farm machinery to harvest the crop of winter wheat, the tube was removed in late August and re-installed as close to the original location as possible in late September, after the field had been sown with a crop of winter barley. The winter barley was harvested in August 1998 and the field has since been used for pigs.

Plate 8 Looking south from the location of Site 40

13.2 Tern Catchment

Soil water monitoring sites were established in the Perry and Tern catchments in the early 1970's as part of the pilot hydrogeological investigations for the Shropshire Groundwater Scheme. In 1981, following a Public Inquiry, the then Severn Trent Water Authority was granted a licence to promote the scheme. In response to concerns from the farming community and environmental pressure groups, the inspector's decision included the need to continue with both groundwater level and soil moisture monitoring. As a result, additional soil monitoring sites were established in the 1980's. The sites tend to be on flat, low lying ground with a shallow water table. Generally, these sites were instrumented with two neutron probe access tubes, a piezometer, a set of tensiometers and a raingauge located in a fenced area. Measurements are no longer taken from the tensiometers although they have not been removed from the sites. Readings are generally taken at intervals of two to three weeks. The OS grid references of the sites used in this study are given in Table 14.2

Table 14.2 Location of soil measurement sites in the Tern catchment

Name	Easting	Northing
Hodnet Heath Control	361890	326560
Hodnet Heath Wet	361920	326240
Hodnet Heath Deep Wet	361940	326240
Bacon Hall	365280	323710
Windy Oak	360740	323040
Bowling Green	362500	325600
Stoke Bridge	363380	327840
Heathbrook	363030	328830
Shawbury Heath	353750	320180

13.2.1 Hodnet Heath Control

This site is located in an area of sparse mixed woodland in an area of rough grass, Plate 9. There are two neutron access tubes and a tubewell on a Crannymoor Series soil. The soil is a brown stoneless, loamy sand developed in sands and gravels overlying the Lower Mottled Sandstone. The adjacent land is drained by the farmer which has lowered the water table. Measurements are taken down to a depth of 2.2 m and began in January 1972.

Plate 9 The Hodnet Heath Control soil water monitoring site

13.2.2 Hodnet Heath Wet and Deep Wet

These two sites are located in an area of heathland, consisting of rough grassland and heathers, on a Crannymoor Series soil, Plate 10. The soil is a brown stoneless, loamy sand developed in sands and gravels overlying the Lower Mottled Sandstone. The wet site has two neutron probe access tubes and a tubewell whilst the Deep Wet has a single neutron probe access tube and a tubewell. Neutron probe readings are taken down to a depth of 1.8 m at the Wet site and 4 m at the Deep Wet site. Measurements began in January 1972.

Plate 10 Location of the Hodnet Wet and Hodnet Deep Wet sites

13.2.3 Bacon Hall

The Bacon Hall site is located in a grass field, near the top of a small rise in the topography, in a corner formed by a road and the access road for the farm. It is on a Newport Series soil; brown slightly stony loamy sand or sand developed in Boulder Clay overlying the Lower Mottled Sandstone. There are two neutron probe access tubes, two tubewells, a tensiometer set and a raingauge. The tubewells are at depths of 2 m and 4 m and show different water levels. Measurements began in July 1983. The field is permanent grass which is grazed and cut for silage, Plate 11. Measurements are taken down to a depth of 2.8 m and began in July 1983

Plate 11 The Bacon Hall soil water monitoring site

13.2.4 Windy Oak

The site is located in a grass field, Plate 12, which is grazed, in an Isleham Series soil. There are two neutron probe access tubes, a tubewell, a tensiometer set and a raingauge. Measurements are taken down to a depth of 2.0 m and began in September 1983. The water level is generally close to the surface.

Plate 12 The soil moisture measurement site at Windy Oak

13.2.5 Bowling Green

This site is near the middle of a field and is on a Redlodge Series soil, reddish brown stony sands developed in Boulder Clay. There are two neutron probe access tubes, a tubewell, a tensiometer set and a raingauge, Plate 13. A further tubewell is at the roadside style. The field was originally grass but currently alternates between cereals and grass on an annual basis. It was cultivated for barley in 1997. Measurements are taken down to a depth of 2.4 m and began in October 1983. The ground level in the instrument enclosure is above that of the surrounding field.

Plate 13 Location of the soil moisture measurement site at Bowling Green

13.2.6 Stoke Bridge

This site is located in a small area of grass between the road and a cultivated field, Plate 14. The measurement compound is against the boundary with the field. The soil is Hall Series, brown coarse loamy over non-calcareous gravely soils formed on the First Terrace. Sugar beet was grown in the field in 1997. At some time, an old pumping station, adjacent to the site was demolished and there is a possibility that the site is located where some of the rubble from this was buried. There are two neutron probe access tubes, a tubewell, a tensiometer set and a raingauge. A continuously monitored borehole is adjacent to the site. Measurements are taken down to a depth of 2.0 m and began in January 1984.

Plate 14 Location of the Stoke Bridge soil moisture measurement site

13.2.7 Heathbrook

This site is in a low lying grass field adjacent to a stream, Plate 15. It is on the Altcar Series soil, earthy eu-fibrous peat soils formed in Alluvium. There are two neutron probe access tubes, a tubewell and a raingauge in the enclosure. In addition, there are two shallow tubewells, one in the enclosure and the other 20 m away on the edge of the a ditch. Measurements are taken down to a depth of 2.4 m and began in January 1984.

Plate 15 Location of the Heathbrook soil moisture measurement site

13.2.8 Shawbury Heath

This site is located in a pine plantation, near the edge, Plate 16. It is on a Crannymoor Series soil, a brown stoneless, loamy sand developed in sands and gravels overlying the Lower Mottled Sandstone. There are two neutron probe access tubes, a tubewell and a raingauge at this site. A shallow tubewell is situated in the headland of the corner of a field about 50 m. away. Measurement are taken down to a depth of 3.6 m and began in November 1970.

Plate 16 Location of the Shawbury Heath soil moisture measurement site

14. ANNEXE 3 - Data file formats

14.1 Soil moisture

The soil moisture data is supplied as a standard SWIPS data set

14.2 Evaporation

The data from the Hydra is supplied in a standard Excel for Windows 95 (version 7.0a) spreadsheet file. The first sheet gives the column headings. The data is organised as hourly values in order of increasing date and time with a sheet for each month. The columns are:

Column A
Hydra number - day and month

Column B
Year

Column C
Day of the month

Column D
Hour measurements start

Column E
Average hourly temperature (°C)

Column F
Variance of temperature during the hour(°C)

Column G
Average hourly humidity (gm/m^3)

Column H
Variance of humidity during the hour (gm/m^3)

Column I
Vertical wind speed (m/sec)

Column J
Variance of vertical wind speed during the hour (m/sec)

Column K
Horizontal wind speed (m/sec)

Column L
Variance of horizontal wind speed during the hour (m/sec)

Column M
Friction velocity, u* (m/sec)

Column N
Sensible heat flux, H (W/m^2)

Column O
Latent heat flux, λE (W/m^2)

Column P
Net radiation, R_n (W/m^2)

Column Q
Energy closure $(H+\lambda E)/R_n$ It is an attempt at obtaining the hourly recovery ratio but without using any soil heat flux information. The night time values are difficult to interpret although the components will contribute to the daily recovery ratio. The day time values tend to increase during the day. At the start of a typical sunny day much of the energy will go into heating the cold soil, leaving this ratio too low. Also towards the

end of daylight the warm soil will be contributing energy to enhance evaporation and so the ratio will become too large.

Column R

z/L (measurement height, z, Obukhov length, L) is an indication of atmospheric stability. Neutral conditions are when turbulence is caused predominately by wind shear and very little by thermal buoyancy. Stable conditions are when the wind drops and the temperature increases with height, usually because the ground is radiating heat into space and cooling the adjacent air. Unstable conditions are predominately caused by thermal buoyancy. Values between -0.05 and +0.05 describe neutral conditions with other negative values describing stable conditions and other positive values describing unstable conditions. Stable and unstable conditions have little to do with the structural uniformity of the site. Wind shear on the other hand is affected by the site.

Column S

When neutral conditions are present and the wind speed is greater than 1 m/sec, the ratio of vertical wind speed variance to friction velocity, sW/u*, can be an indication of the structural uniformity of the site (or is the upwind fetch sufficient?). It is generally accepted that a value of 1.3 in these circumstances means that there is little upwind disturbance of the streamlines of wind. However, because the Hydra Mk2 uses a mechanical anemometer, the measured variances of horizontal windspeed are attenuated relative to the variances of vertical windspeed. This reduces u* and increases sW/u*. The Mk2 is more likely to give an average value in these conditions of 1.5.

Column T

Channel 3 volts

Column U

Aerodynamic resistance (s m^{-1})

Column V

Hydra status code, this is decoded as follows:

0 no error

9999 value missing

The error code is in the form of four integers ABCD. The integers are in the range of 0 to 4 Values higher than this imply that the code has been subtracted from 9000 and indicate that the data, although less than perfect, can be used. The error code is decoded by:

A This is a software generated code which indicates that the FQ function (calculated from the I.R. hygrometer output) has been unacceptably large

A=0 no error

A=1 error < 1% of the hour

A=2 error < 5% of the hour

A=3 error > 5% of the hour

B This is a combination of hardware and software generated error for the hygrometer cooler/heater and the time constant in the filter of the running mean. When the cooler/heater of the hygrometer is not operational, an error flag is set on the hygrometer board. Note: disconnecting the cooler on the top of the board or at the plug does not cause the flag to be set. If the all OK voltage in Position 6 (status channel) is > 0.18 volts then this may be erroneously seen as a cooler fault, being included in the 0.45

volt bin. The running mean filter time constant is reduced to give rapid adjustment after start up or after a status 3 error on the hygrometer or sonic anemometer. This sets the fast time constant flag.

B=0 cooler/heater on and normal time constant
B=1 cooler/heater on but fast time constant
B=2 cooler/heater off but normal time constant
B=3 cooler/heater off and fast time constant

C This is a hardware generated error indicating that the hygrometer is in error when the Position 3 voltage (hygrometer channel) is > ±5 volts. This may be a temporary error due to water on the lens, or it may be a permanent fault.

C=0 no error
C=1 hygrometer in error < 1% of the hour
C=2 hygrometer in error < 5% of the hour
C=3 hygrometer in error > 5% of the hour

D This is the same as C but indicates an error in the sonic anemometer which is normally caused by excessive water on the transducers but is much less sensitive to water than the hygrometer.

D=0 no error
D=1 sonic anemometer in error < 1% of the hour
D=2 sonic anemometer in error < 5% of the hour
D=3 sonic anemometer in error > 5% of the hour

15. ANNEXE 4 - SHROPSHIRE SOIL WATER DATA

by J D Cooper

15.1 Background

The Shropshire Groundwater Scheme was conceived in the early 1970s to augment river flows in the River Severn. The scheme was opposed by a number of groups, particularly local farmers, who were concerned that the lowering of the water table, as a result of pumping, would affect crops adversely. Consequently, a soil water monitoring scheme was set up for the then Severn Trent River Authority. This comprised a large number of neutron probe access tubes in fields over the catchment areas of the Tern and the Roden. Several of these access tubes had nests of mercury manometer tensiometers associated with them. Installation and early monitoring of the network was carried out by Dr Graham Fry of Laurence Gould Consultants Ltd.

Initially, the monitoring was in areas unaffected by pumping, as there was a long delay before a licence to operate the scheme was granted in 1981. A condition of the licence was that monitoring should continue in order to detect any changes in the soil water conditions as a result of the pumping. To this end, the network includes sites which are not within the pumped area and so could act as a control.

Over time, there have been a number of different operators of the measuring equipment, different data processors, different neutron probes, different data processing systems and different bodies responsible for collecting, processing and evaluating the data. Records are incomplete, but the chief events have been:

- Monitoring of a number of sites in the proposed Groundwater Scheme area in the mid to late 1970s by a number of people under the auspices of Severn Trent Water Authority, the National Farmers Union, Aston University and others. These are not of interest in the present investigation, but may be for a more wide-ranging study. Some records are known to be kept at the University of Aston, Birmingham, Department of Civil Engineering (Dr P Hedges is the contact).
- A Soil Survey, carried out by the then Soil Survey of England and Wales in 1982. Based upon this survey, Laurence Gould Consultants Ltd proposed the establishment of 18 monitoring sites. Unfortunately, it is not clear how these relate to the sites actually established.
- 1982/83. Installation of neutron probe access tubes (and tensiometers at some sites), followed by the commencement of readings by Severn Trent staff (often students), under the supervision of Dr Fry. Two neutron probes (designated 1 and 2) were used for this. These are believed to have been Pitman Model 225 (Wallingford) probes, obtained from elsewhere. They must have been bought originally before 1977, since D A Pitman ceased production of neutron probes then. At some point a third probe designated 3) was borrowed. Early data was, apparently, unreliable and measurements survive only from late 1983, by which time the teething problems appear to have been solved. The height of each tube was measured and the depth counter reset at each so that it showed the actual measurement depth. The setting of the depth counter at each individual tube was recorded on the field sheets. Processing of the data was carried out initially on a Commodore PET microcomputer. The initial depth counter setting was transferred to the computerised record at this stage.

- At an unknown date, the processing system was changed to one based on dBase on an IBM PC or compatible. The PET records were transferred to this system, but the depth counter information appears not to have been transferred along with them. The original paper records were disposed of at some point and, more recently, have been kept for only one year after data collection.
- 1988/1989. The original neutron probes were disposed of in mid-1989 and replaced by two new ones (designated 4 & 5). In the intervening period, another probe (designated no 3) was used from 27 July 1989 to 14 November 1989. This may have belonged to the Soil Survey and Land Research Centre, who carried out the readings at this time.

15.2 Problems with Use of Data

Problems with the use of the data were identified during a study into groundwater recharge by IH in 1994, which used the data collected from the area. A discontinuity was found in the records in 1989, which appeared to coincide with the purchase of two new neutron probes and also with changes in management arrangements and personnel working on the monitoring. This part of the current project was prompted by a desire to track down the cause of the discontinuity and, if possible, to provide a means of correcting the data, so that the data series would be homogeneous.

The opportunity has been taken to filter out many one-off inconsistencies in the data. Recommendations are also made for operation of the data collection programme in the future to improve the quality of data collection.

15.3 Approaches to Investigation of Causes of the Problem

The major change in the data occurred during 1989. Figure 15.1 illustrates the discontinuity at Bacon Hall, whilst Figure 15.2 shows data for the same period at Bowling Green. The discontinuity appears more subtle at some sites than at others, whilst the presence of significant inter-annual variations in weather and soil water patterns makes it difficult to be certain that any differences are real.

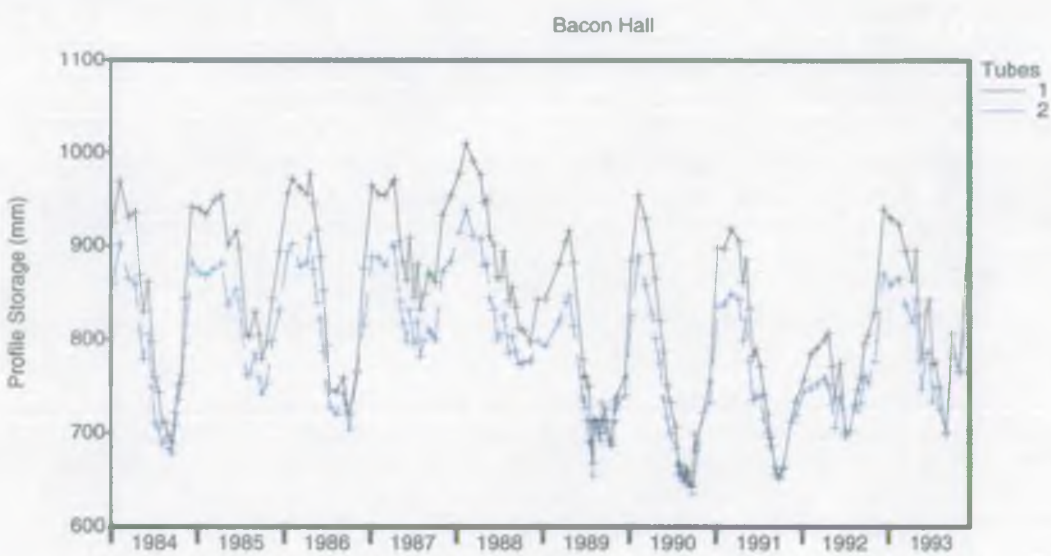


Figure 15.1 Changes in soil water content at Bacon Hall

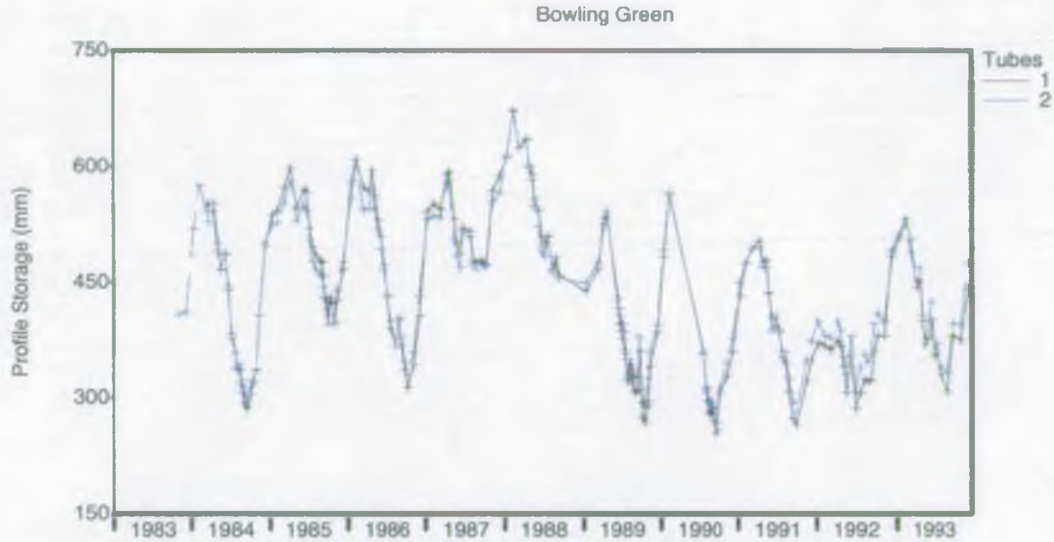


Figure 15.2 Changes in soil water content at Bowling Green

Investigation of the problem involved:

- detailed examination of soil water content records at a number of sites. Whole profile water content, that at individual depths, over a range of depths and successive water content plots against depth were all examined. The most revealing plots were found to be at those depths where there was a strong gradient of water content, but where the water content was expected to be stable with time because it was below the water table;
- the record of water counts was examined to see if this gave any clues to the cause of a discrepancy. Use of an inappropriate water count value would lead to incorrect calculations of water content, more or less proportional to the water count error;
- key personnel who had been associated with the monitoring in the past were contacted and asked to recall their experiences and procedures;
- attempts were made to track down the original field records from the readings.

15.4 Likely Causes

The possible causes of a discrepancy considered were:

- the differences observed are a true reflection of the changes in soil water at the various sites and any apparent discrepancy is, in reality, caused by drier weather conditions obtaining since 1989.
- an incorrect allowance was made for the different depth placement necessary for the new probes used since 1989.
- the water count procedure used results in a different value to the appropriate one. This might apply to the pre- or the post-1989 data.
- some of the tubes may have been repositioned, leading to a discontinuity in the data, although the trends will be expected to be comparable.

15.4.1 Changes in weather conditions

There is some evidence that climate is getting drier in this area of the country. This could well lead to lower average soil water content. However, it would be expected that, in the driest summers, there would be a limiting water content that most soils would dry to by evaporation from vegetation. Similarly, in the wettest winters, the soil would not get very much wetter than a "field capacity" value, other than for very short periods. These limits might be expected to be reached more frequently in the case of dry summers and less frequently for wet winters, but, nevertheless, they should not be very different when they are reached. This is not supported by the evidence. At both Bacon Hall and Bowling Green, similarly dry conditions were experienced in the summers of 1984 and 1986. These may well be close to the dry limit. However, even drier limiting values were recorded in 1990 and 1991, by some 50 mm over the depth of the profile in each case. In the case of winter conditions, excluding 1988 (which appears to have been unusually wet), the "field capacity" values seem to be some 50 mm lower post -1989 than before. Whilst this can hardly be described as conclusive evidence that the effect is not climatic in origin, it makes such an explanation look unlikely.

15.4.2 Depth Placement Problems

The newer "IH II" probes, manufactured by the Didcot Instrument Company Ltd., purchased in 1989 and used from 12 December that year, differ a little from the D. A. Pitman Ltd. Model 225 "Wallingford" probes used up to 27 July 1989. The centre of sensitivity of the Pitman probe (i.e. the point at which measurements are assumed to be made) is 120 mm above the bottom of the probe casing. A socket in the bottom of the transport shield fits over the top of the access tube, so that the upper 40 mm of the access tube are inside the transport shield. The centre of sensitivity of the probe is thus 80 mm above the top of the access tube when the probe is on it and it is locked in its shield. The probe, therefore, needs to be lowered 80 mm before the measuring point is level with the top of the tube and then a further distance equal to the height of the access tube above ground plus the desired measurement depth. Different users have different ways to locate a probe consistently at the same depth. The procedure adopted in the Shropshire Groundwater Scheme was to measure the distance from the ground surface to the bottom of the transport shield (which should be level with the bottom of the probe casing when it is locked in its shield) and then to adjust the depth counter on the probe such that it indicated the actual depth of measurement. Different tubes vary somewhat in their height above ground, so that a slightly different number would be set for each tube. As an example, if the measured distance from the ground surface to the bottom of the transport shield were 150 mm, the top of the tube would be 150 + 40 mm = 190 mm above ground. Lowering the probe by 80 + 190 mm = 270 mm would bring the measuring point level with the ground surface. The depth counter actually indicates in cm and finer adjustment is not practicable. To indicate the actual depth of measurement, the depth counter would need to set initially at - 270 mm (-27 cm). The depth counters used roll over from 9999 to 0000 (cm) as they pass through zero and so the depth counter would be set at 9973. Each tube had its own depth meter setting to ensure consistent placement of the measuring point and it is understood that preprinted field data collection forms had this information on them and, indeed, that the initial depth counter readings were entered into the computer processing system, at least in the early years. Unfortunately, these records do not appear to have survived, so that checks are not possible retrospectively.

This procedure is perfectly reliable and certainly no less prone to error than other methods. In common with other procedures, the likely problems come with incorrect setting of the depth counter, mixing up different tubes, poor training of operators and not ensuring that the probe is locked in its shield correctly at the start of readings. Other problems which might occur include the height of the tube changing over time, as a result of its being forced further into the ground by repeated setting of the probe on it; damage to the tube or external interference; backlash in the depth placement mechanism, and differences between probes. The Pitman probe is more prone to this last problem, as the position of its depth placement mechanism with respect to the transport shield is adjustable. All probes will, however, have manufacturing tolerances and some differences between the characteristics of different examples of nominally identical depth counters.

When the new probes came into use, the operators were faced with the problem of how to make allowances for the different depth placement requirements. This coincided with a change in the way that the operation was managed and, crucially, in the personnel carrying out the work. They opted to remeasure the ground surface to bottom of the transport case distance and then to recalculate the depth counter settings using the same procedure as before, but using the newer dimensions. The Didcot IH II probes have the same depth socket as the older Pitman probes, but the centre of sensitivity is 20 mm lower, hence the depth counter is set initially at a slightly larger number than previously. In the example quoted above, the initial depth counter reading would be 9975. Since ground is rarely completely flat, it is quite plausible that a second measurement after some years would result in a difference of a cm or two in the measurement, because of natural bumpiness, faunal activity, erosion, hummocks forming, etc. It is also possible that at some sites, the tube would have moved down or (less likely) up in the soil.

If a record of the actual depths at which the probe had been set in the period up to 1989 were available, then this could be compared with the new depths and any discrepancies identified. Unfortunately, although the depths were entered onto the then computer system, neither copies of these files nor the original field data sheets have been retained. Nor do there seem to be any copies of reports detailing this information still extant. This comparison does not, unless one of these sources comes to light, therefore, seem possible.

To test the hypothesis that a consistent difference in depth placement is responsible for the reduced water contents, it was reasoned that if depths in the soil could be found at which the water content would be expected to remain constant (or would return to a consistent level seasonally) and if the gradient of water content with depth were steep, a marked difference between the measurements pre- and post- 1989 would be expected. Equally, if depths could be found at which water content is stable and the gradient with depth small, then no difference would be expected and some other possible explanations (such as consistent problems with water counts) could be excluded.

The best stable water content measurements are expected to be below the water table, particularly at sufficient depth that water table fluctuations do not make any of the measurement volume around the depth of interest unsaturated. The area of the measurements is one where shallow groundwater is prevalent and so it seemed entirely possible that tubes penetrating below the water table would exist. Several sites have dipwells, so these records were examined initially in conjunction with the depth of access tubes to identify likely candidate sites. Bacon Hall proved to have the most promising combination. Two dipwells on the site show markedly different depths to groundwater (approximately 3.8 m and 1.45 m). Tensiometer records support the shallower depth (Macdonald, 1996). There is some seasonal fluctuation, of around 0.3 m. The lowest reading depth of the tubes at Bacon Hall is 2.8 m, so it was expected that there would be an adequate amount of data to test the hypothesis.

Figure 15.3 shows sequential depth profiles of water content, illustrating that there is, indeed, little variation observed at depths expected to be below the water table, but also that there is a substantial gradient of water content at depths between 1.8 m and 2.2 m, as well as a fairly constant (with both time and depth) water content between 2.4 m and 2.8 m.

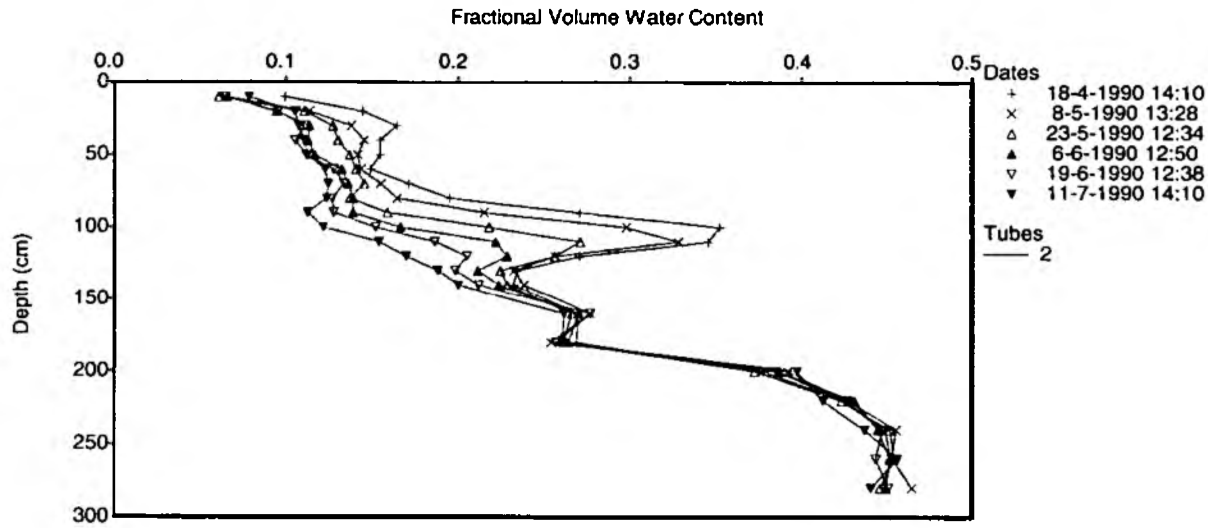


Figure 15.3 Soil moisture content profiles at Bacon Hall

Figure 15.4 shows the water content v time record for depths of 1.6 m, 1.8 m, 2.0 m and 2.2 m over the period 1984 to 1993. There is little change in water content at either 1.8 m or 2.2 m (which are both close to the edge of the steep section), but at 2.0 m, there is a clear change from a volumetric water content of close to 0.425 to one of about 0.385. The gradient of water content is increasing strongly downwards at this depth, so that a reduction of water content implies that the probe is set higher in the profile. The behaviour at 1.6 m is instructive. This is only a small distance below the zone of fluctuation of the water table. In winter, spring and most of the summer, water content up to 1989 was fairly constant at about 0.437, but in early autumn in most years, it dipped for a short time, apparently to a limiting value of about 0.263 After 1989, the upper (well below water table) level dropped to 0.380 and the limiting (presumably unsaturated) level was also lower at about 0.224 Perhaps more tellingly, the length of the period each year when the water content was below its highest level is significantly longer (almost 50% of the time, rather than occasional dips) after 1989, consistent, again, with the measurement point in the profile being higher than previously. Lastly, data for Bacon Hall show a reduced water content since 1989. The overall trend of water content with depth is for it to increase with depth. If the reading depths are shallower than previously, then it would be expected that lower water content would be estimated. This is indeed the observation and all three means of identifying the direction of the discrepancy seem to agree.

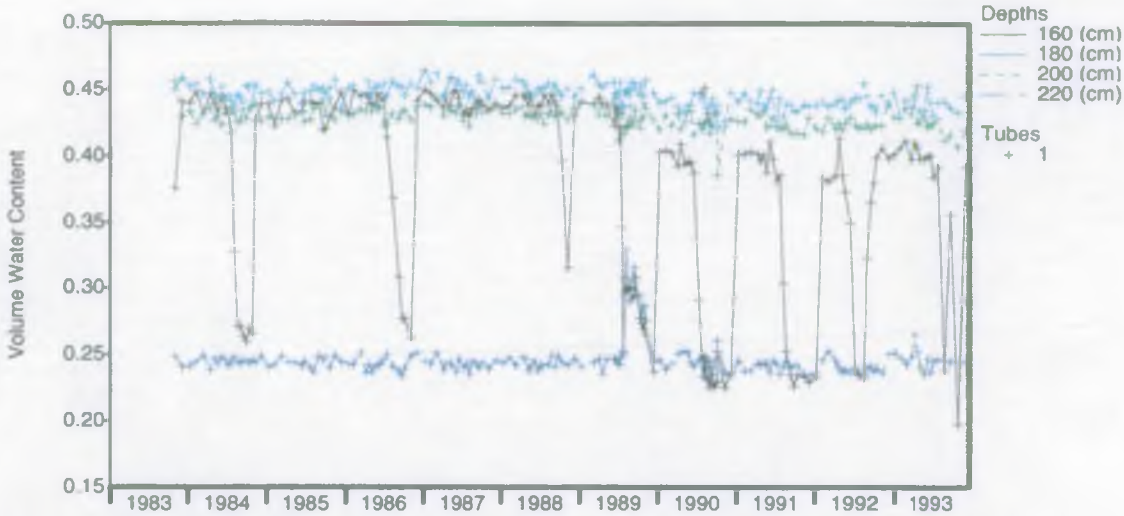


Figure 15.4 Changes in soil water content at selected depths, Bacon Hall

It needs to be appreciated that the neutron probe integrates over a substantial volume (typically a sphere of radius 150 mm). This has the effect of enabling it to "see" changes in water content some distance above or below the nominal measuring depth and also to smear out gradients somewhat. The actual gradient in water content represented by the data of Figure 15.3 is, therefore, much sharper than shown (quite likely very nearly a step change at a lithological interface). It also means that any attempt to quantify the depth difference between the pre- and post-1989 data is very hazardous.

15.4.3 Water Count Differences

As stated above, the stability of water content with time at some depths where this would be expected, rules out inconsistencies in the water count procedure as a possible cause of the problems. Nevertheless, the water count record has been examined to attempt to pinpoint any areas where there may be problems with the data and to aid production of a more homogeneous data set. Figures 15.5 - 9 give the details of water counts. The procedure appears to have been to perform 10 water counts for 64 seconds on each occasion that the probe has been used and to use the mean of this value as the relevant water count for the readings on that day.

The theory of Poisson statistics states that the expected standard deviation of a set of N events is $\sqrt{N}$. This means that the standard error for a set of n readings with a mean count rate R, taken over a period t is $\sqrt{(R/nt)}$ Bell (1987). With R typically 900 counts per second for a Pitman/DICO neutron probe, n = 10 and t = 64 s, the standard error of the mean is about 1.2 cps. Changes from one occasion to another of this sort of magnitude are of no significance and the adoption of a long-term mean would be a better choice. Alternatively, where there is a long-term trend in water count, a line fitted through the data is often a more appropriate choice.

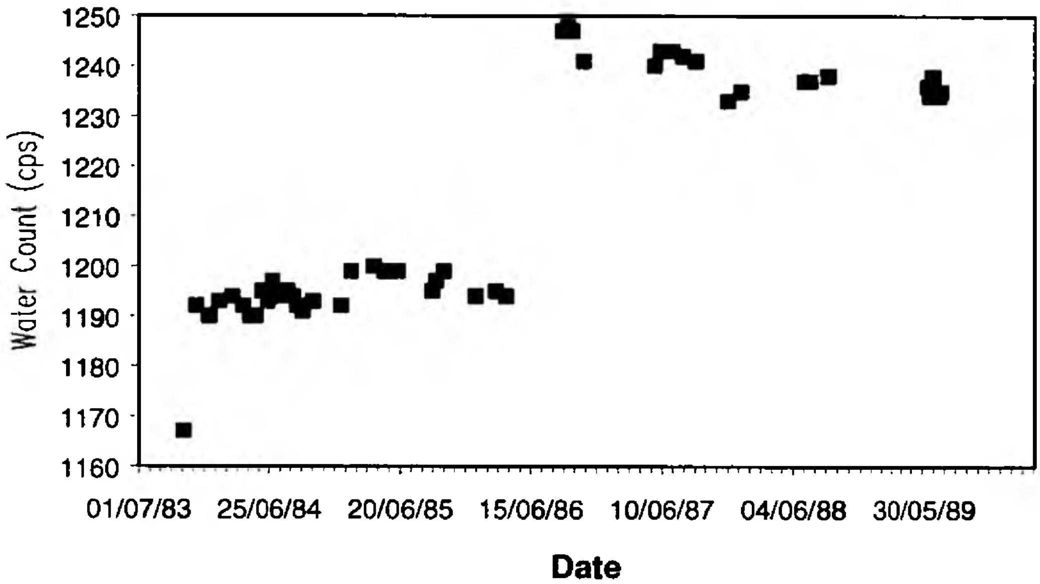


Figure 15.5 Water counts for probe 1

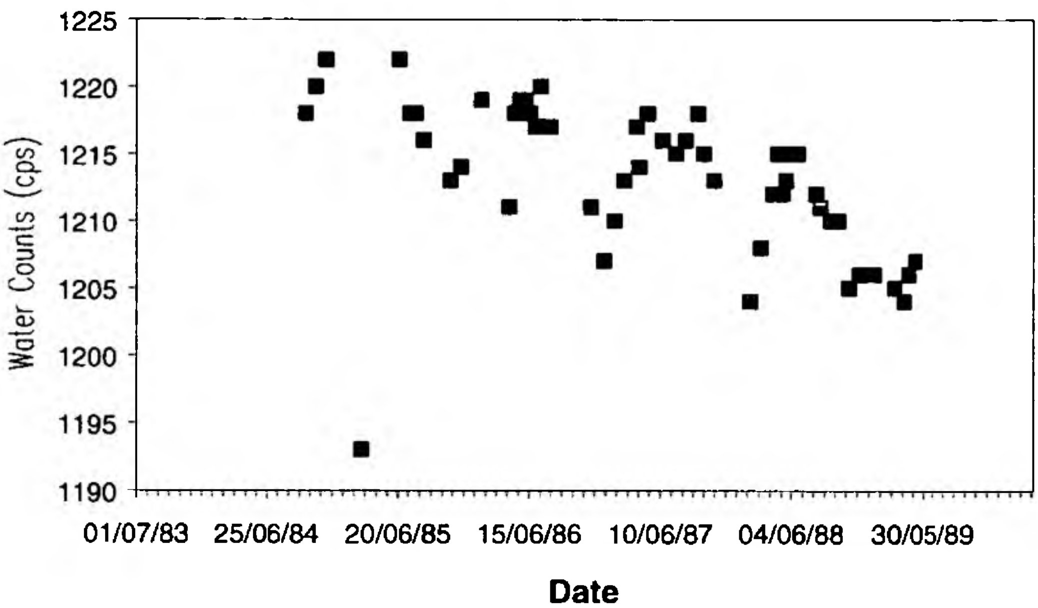


Figure 15.6 Water counts for probe 2

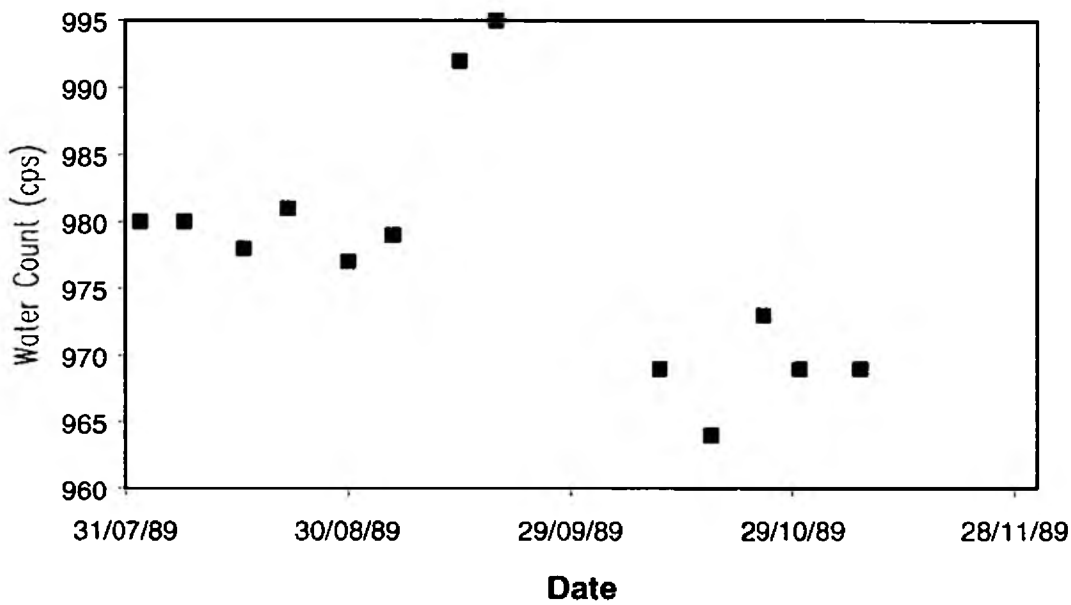


Figure 15.7 Water counts for probe 3

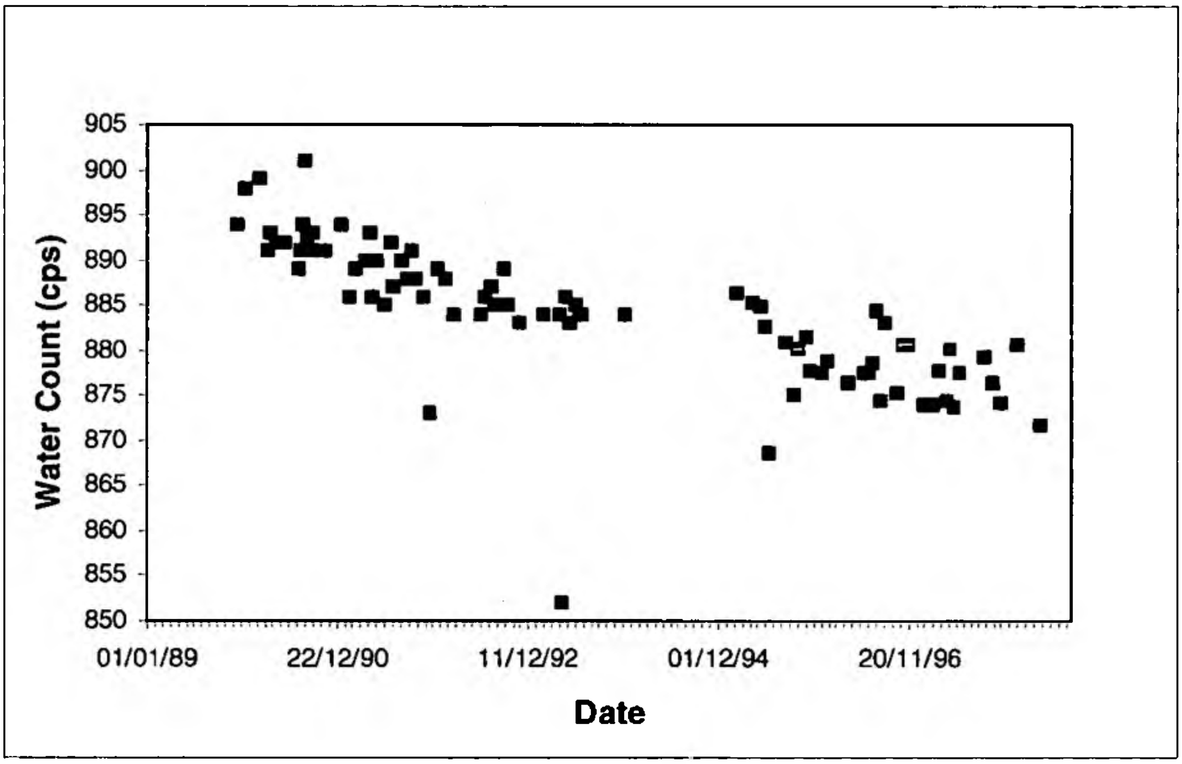


Figure 15.8 Water counts for probe 4

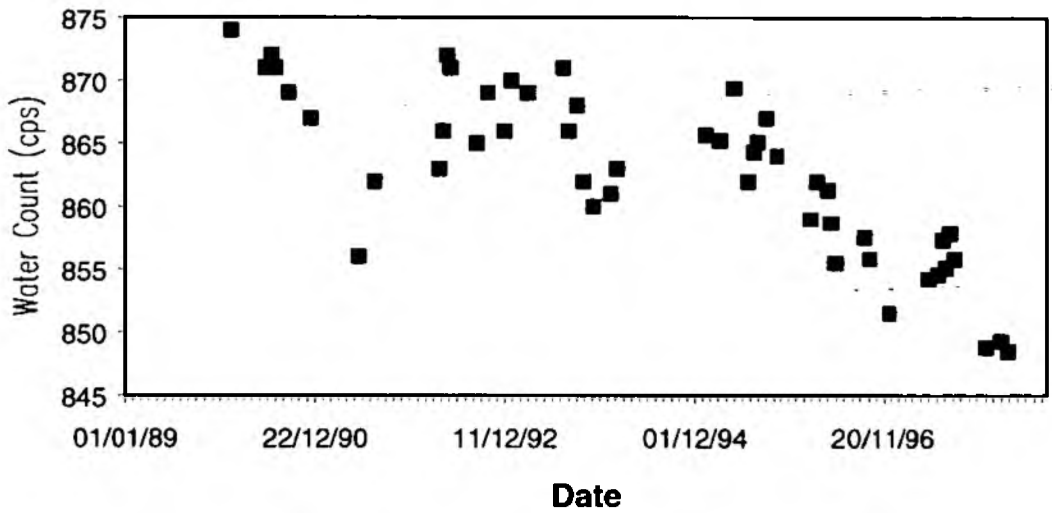


Figure 15.9 Water counts for probe 5

Apart from the smoothing of the water counts, suggested above, there are a number of inconsistencies in the data. Both probes 1 and 2 show some erratic behaviour early in the data record. This may be inexperience by operators or teething problems with new probes (although it seems unlikely that they were, in fact, new). Apart from this, Probe 1 starts with a stable water count, which then starts to rise slowly. There is then a discontinuity in the data, before it resumes a stable reading, but much higher than previously. This may have been a consequence of a breakdown and repair. There is a four-month gap in the water count record for this probe in 1986, which lends support to this suggestion. Two readings for Probe 2 at the beginning and 3 readings on 12/13/14 March 1985 may have been wrongly attributed to Probe 1. This may not lead to a problem with the data record for those days, though, if the same error had been made in recording the probe used for the readings (i.e. the water count may be right even though the probe number had been recorded wrongly). Some seasonal fluctuations appear in the water count record for Probe 2. This is probably a temperature effect. The middle part of the record for Probe 3 seems odd. The probe may have been failing (there is a 2-week gap in the record which would be consistent with a repair). The odd reading of 1023 on 20 September may have been a different probe, perhaps because the original was being repaired. After 6 October 1989, the readings are stable, but at a lower level than previously, which suggests that it was a different probe or had been repaired. Probes 4 and 5 have consistent water counts, albeit with more scatter than simple statistics would suggest and, particularly in the case of Probe 4, a number of outliers, which might be transcription errors.

15.4.4 Repositioning of Tubes

None of the people interviewed has any recollection of any tubes having to be replaced or moved. The suggestion that the cause of the pre- and post- 1989 discrepancy is due to an inconsistency in depth placement between these periods seems to explain the differences and there seems no reason to pursue this suggestion further.

15.5 Recommendations for Future Users of the Data

Future use of the data relies on taking a view of the reliability of the data. A data set of this length (now 14 years) is unique in the UK. The quality of most of the data is excellent. With the amount of data collected, it would be unreasonable to expect no problems or inconsistencies. Some clearly exist, the most glaring one being that there is a discontinuity between data collected before 1989 and that collected after. Some of the data collected in the period that the readings were subcontracted to the Soil Survey and Land Research Centre are of lower quality, but, in the main, still remarkably good. There are also some occasions when there are large changes in the data, apparently over a few hours. It is not clear whether these are genuine changes, perhaps as a result of heavy rainstorms, or are erroneous.

An aim of the investigation had been to produce a homogeneous record. This was on the assumption that an unambiguous cause for the discrepancy could be found and that the "correct" data could be recalculated. Although the cause of the problem has been established with high probability, quantification of it is very uncertain. The difference in depth location between data in the early and later part of the record probably varies from one place to another, as the height of the ground surface is a matter of subjective judgement. This is a field where consistency is probably more important than absolute accuracy and there is no reason to prefer one set of interpretations over the other. It is suggested that in using the data, the two periods be treated separately. For instance, if running a soil water balance calculation, the calculation should not run through the time that the two sets of data change over, but it be calculated up to 14 November 1989 and then separately from 12 December 1989. This assumes that the period from 27 July to 14 November 1989, when SSLRC were doing the readings "belongs" to the earlier data period. There is evidence that this is so, but (probably because of unfamiliarity with the area and possibly inexperience of some operators) there are more outlying points in the soil water record in this period than most others and, for critical work, it should be treated with some caution.

15.6 Recommendations for Future Data Collection and Processing

A considerable amount of effort was expended in this investigation tracing people who had worked on the project in the past to try to reconstruct what had happened. This would have been eased considerably if original data records had not been destroyed. The main recommendation is, therefore, that, whatever else may be disposed of, original data sheets should be kept. It may be acceptable, on the grounds of convenience, to microfilm or keep scanned copies of these on CD-ROM, but it needs to be established that they are perfectly clear before the originals are destroyed. This recommendation applies equally to water count records.

The IH SWIPS software offers a comprehensive system to store, process, quality control, edit and present soil water data and this should be kept up. Tensiometer data should be transferred to it as well.

Data collection procedures at present seem to be more than adequate. Consideration needs to be given to keeping the same level of quality and consistency of data collection in the future during inevitable personnel and management changes.

15.7 Acknowledgements

The assistance of many people who have worked on the Shropshire Groundwater Monitoring Network or with the data is gratefully acknowledged. Without their frank and free sharing of information and recollections, no progress with this part of the project would have been possible. The principal people whose help has been instrumental are: Ken Parker and Kevin Voyce of the Environment Agency, Shrewsbury; Rick Ireland of Severn Trent Water; Simon Taylor of the Environment Agency, Southern Area; Graham Fry, formerly of Laurence Gould Consultants Ltd.,; Andree Carter of the Soil Survey and Land Research Centre; Malcolm Reeve and Arthur Thomasson, both formerly of the Soil Survey and Land Research Centre; Peter Hedges of Aston University; David Macdonald of BGS and Jon Finch of IH.

15.8 References

Bell, J.P. 1987. Neutron probe practice. *Inst. Hydrol. Rep. 19.*, Wallingford, UK. 51 pp.

Macdonald, D.M.J. 1996. Estimating groundwater recharge through glacial till at Bacon Hall, Shropshire. *British Geological Survey*

16. ANNEXE 5 - DATA SOURCES

16.1 FRL input parameters

The parameters required by a multi-layered soil water model, such as FRL are:

1. The maximum rooting depth for the vegetation cover type.
2. The proportion of roots in each layer.
3. The fractional available water content (the difference between the soil water contents at wilting point and field capacity) for each layer.
4. The proportion of the available water content that is freely available for uptake by plant roots.

16.2 Data sources

Realistically, the only sources of rainfall and meteorological data for a national system for estimating groundwater recharge are the Met. Office and the Environment Agency itself. This is especially true if a historic time series of 30 years is required. Although, in terms of raingauges, the historic data set is good enough to represent the spatial variability of rainfall, this may not be true if it is planned to run a model in near real time, e.g. updating on a weekly basis. In this case the same problem that MORECS has will be encountered. The number of raingauges whose data is available in this time scale is of the order of 200 and so the spatial variability is not well represented. Data from the Met. Office's network of weather radar may be a suitable replacement but this would have cost implications.

There are similar concerns over the network of meteorological stations. The number of meteorological stations has been steadily declining. Although temperature and vapour pressure can be expected to show low spatial variance, the same may not be true of wind speed or downwelling solar radiation. MORECS currently has access to data from about a hundred sunshine recorders in the UK. The distribution of these is biased towards coastal towns and airfield (which tend to be sited on flat, low lying ground). The Met. Office has deployed about 70 solarimeters but the data from these is not available to MORECS and their distribution is again biased. Earth Observation data is capable of supplying the required information but it will take time and resources to develop.

The situation with regards to soil hydraulic properties and their spatial distribution is better The only source is the Soil Survey and Land Research Centre (SSLRC). This organisation has several data sets which are potentially of use to the Environment Agency. There is a database, called LandIS, which includes a National Catalogue of Soils that contains a wide variety of properties, including water retention properties. These are derived from replicated, undisturbed, volumetric, horizon samples of at least 200 cm^3. There is also the SEISMIC database that provides estimates of the hydraulic properties for all soil series, on a horizon basis, derived from texture measures using a pedotransfer function. Finally, there are the 1 : 100 000 soil maps of the UK which are available in digital form. The precise combination of these data sets that will achieve the objectives of the Environment Agency cannot be defined until the detailed specification for a national system for estimating groundwater recharge has been defined. However, it is likely that some form of aggregating hydrologically similar soil types together within model cells would be advantageous. This may take the form of considering the soil and the substrate and would reflect the relative importance of the processes of recharge, interflow and runoff. It is likely that, due to their exceptional properties, the Chalk soils should not be aggregated with any other type of soil.

There is a limitation to the SSLRC's data, in terms of the depths at which measurements were made. The data was collected on the basis of the soil horizons. The result is that the Chalk soils are often defined as thin soils, typically 0.25m, whereas hydrologically they are deep soils. The measurements were only made on the shallow horizons and so it is very unlikely that these can be taken as representative of the full range of depths of interest to hydrologists. Pedotransfer functions cannot be used to define the hydraulic properties of these deeper soils as they are not composed of grains derived from weathering. It is also not certain that the sample sizes used by the SSLRC are large enough to represent heterogeneity resulting from microfissures. Thus it will be necessary for the Environment Agency to commission a programme to make the necessary measurements. It will also need to investigate whether this is also true of soils developed on limestones.

Information on the land cover and vegetation canopy parameters is also readily available. The Institute of Terrestrial Ecology's Land Use GB data set was derived from Earth Observation data in the early 1990's and provides information at 25 m resolution. Unfortunately, it defines all annual crops as tilled ground, which limits its use. However, a new data set will be produced for the year 2000 and this will distinguish various annual crops. Thus this is likely to be the best data set for the Environment Agency to use. The parameters required by the model can only be defined once the decision is made as to which model will be used However, they are likely to be available in the literature. It has been pointed out above that the exception is likely to be the bulk canopy resistance to be used for bare soil. Consideration should also be given to how the phenological development of annual crops etc. is handled. In MORECS it is achieved by assuming fixed dates for emergence, canopy closure, harvest etc. Although this is simple to implement it makes no allowance for annual differences due to weather. Therefore it may be preferable to implement a simple phenological model driven by meteorological data. These are currently been developed for agricultural yield monitoring (the models can be constrained by products derived from Earth Observation data to take account of local variations in seeding date, fertiliser application etc.)

17. ANNEXE 6 - COMPARISONS OF OBSERVED AND PREDICTED EVAPORATION

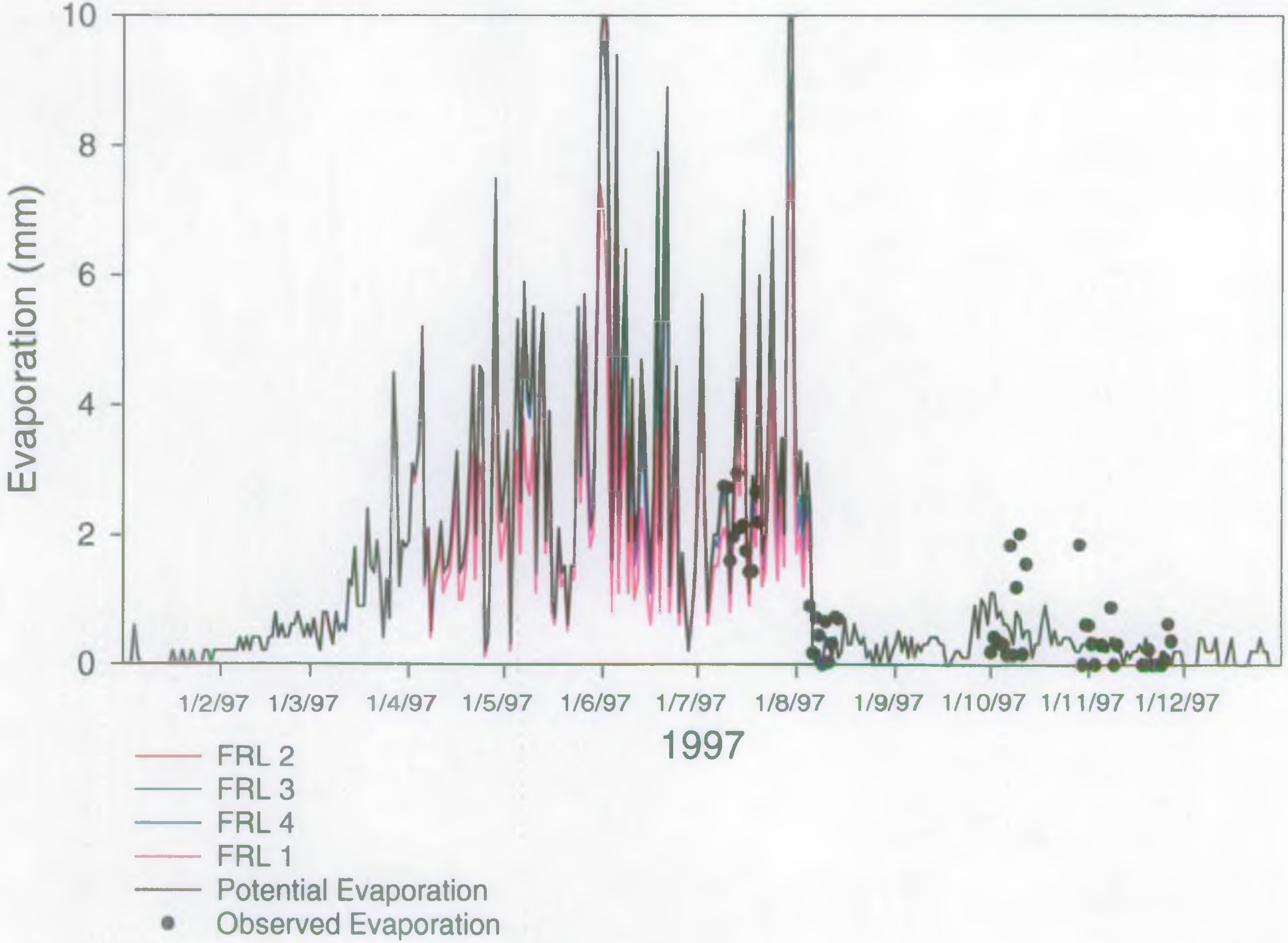


Figure 17.1 Observed evaporation and that predicted by FRL at Site 10

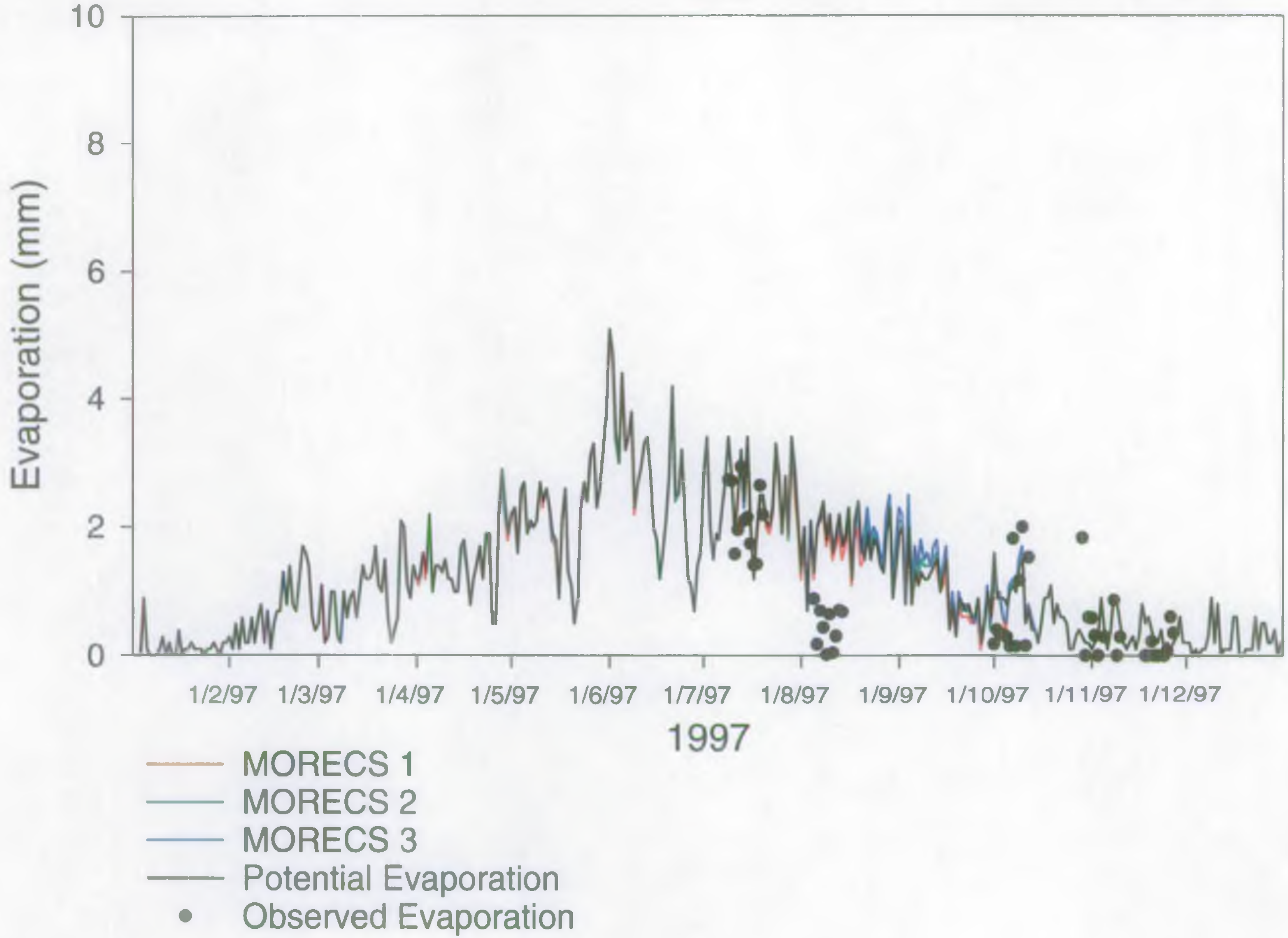


Figure 17.2 Observed evaporation and that predicted by MORECS at Site 10

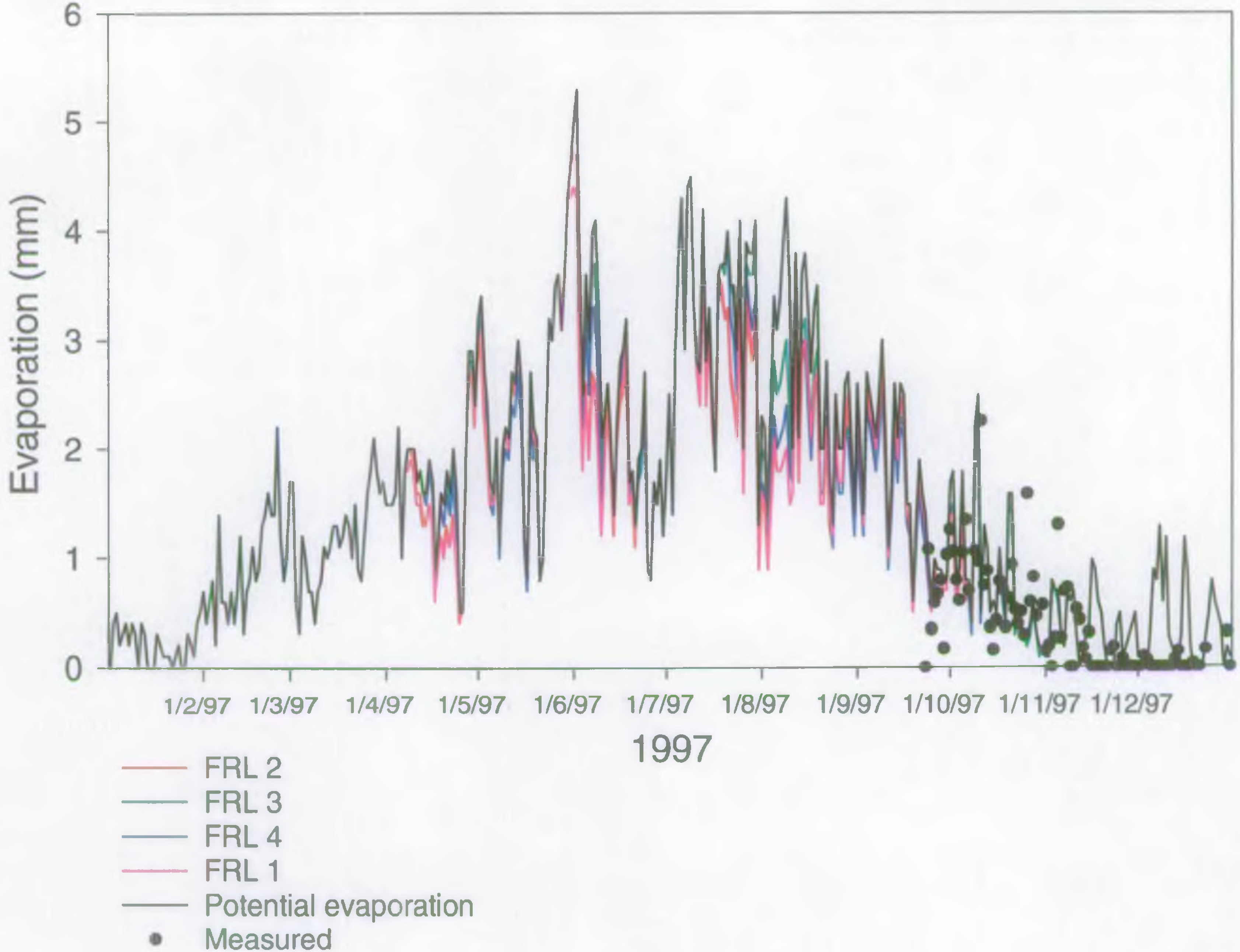


Figure 17.3 **Observed evaporation and that predicted by FRL at Bacon Hall**

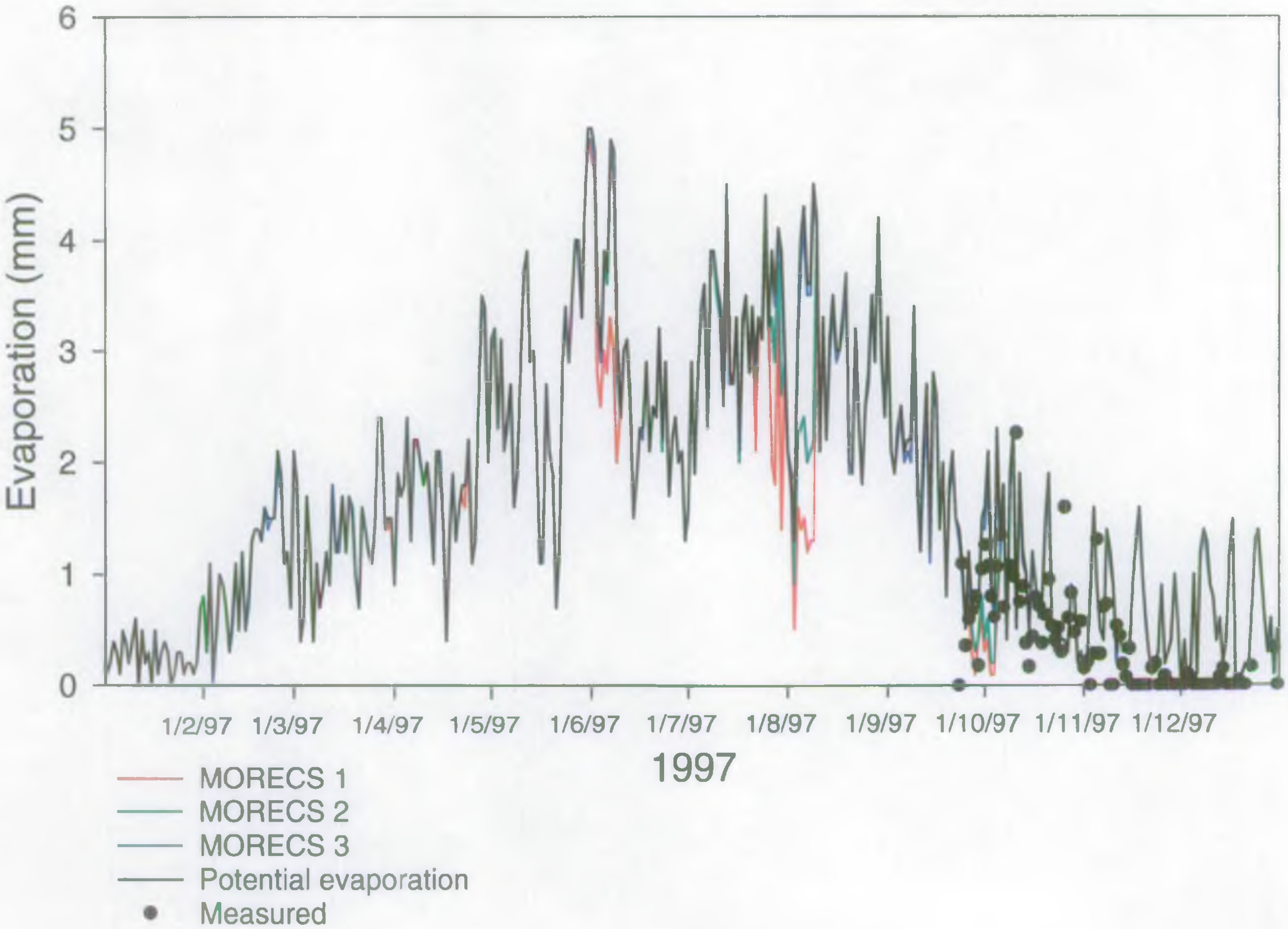


Figure 17.4 Observed evaporation and that predicted by MORECS at Bacon Hall