THE HYDROLOGY OF A FLOODPLAIN WETLAND, NARBOROUGH BOG, LEICESTERSHIRE

Thesis submitted for the degree of
Doctor of Philosophy
at the University of Leicester

by

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March 1994

Abstract

A combination of fieldwork and numerical modelling is used to examine the hydrology of a floodplain wetland, Narborough Bog in Leicestershire. The hydrogeological conditions which maintain floodplain wetlands are considered by describing floodplain hydrostratigraphies and deriving a simplified model of wetland hydrology.

The hydrological processes which provide water inflow and outflow to a wetland system are reviewed. The mathematics of subsurface water flow are described to provide the background for application of a full groundwater model to the site. The processes are considered by reviewing studies on wetland hydrology.

Regular monitoring of field water tables was undertaken, from November 1990 to June 1993; and the spatial and temporal relationship of these records to rainfall, evapotranspiration and river stage are described. Regression models and a response function are used to quantify the relationship of water tables to meteorological parameters, and also to examine the extent of temporal variations in model explanation.

Experiments investigating water flow through in-situ peat deposits and alluvial sediments are described. These included an artificial flood experiment and the study of infiltration through an isolated peat column. The results enable approximate values for hydraulic parameters to be estimated for organic and alluvial deposits.

The groundwater model MODFLOW was used to develop a calibrated transient model, the ability of which to replicate water table responses to isolated recharge and evapotranspiration events was examined. The results enable an assessment of the significance of influent and effluent water flows, and the contribution of overbank water flow to Narborough Bog. Suggestions for further refinement to the model are advanced. The model is used to derive approximate water budgets for 1991 and 1992 to demonstrate the sensitivity of Narborough Bog to periods of drought, and examine the current significance of the river to site hydrology.

Acknowledgements

This thesis was funded by a NERC studentship, with additional financial help from an English Nature equipment grant. I am grateful for the permission of the Leicestershire and Rutland Trust for Nature Conservation to work at Narborough Bog, and for the interest of Pete Asher, and members of the local management committee.

I would also like to thank Dr Tony Brown for his supervision and guidance during this study. During the course of this study I have received advice from a number of people. The experiment on the peat column was undertaken in cooperation with Dr. J.A. van den Berg, from the Dept of Physical Geography, Utrecht University. A short visit to Utrecht, provided an opportunity to discuss groundwater modelling packages, and was financed by NERC. I would also like to thank Dr. Dave Furbish, for his advice on time-series methods, which provided the background for the response function models discussed in chapter 5.

I am very grateful for the encouragement of Dr. Jeff Warburton throughout this study, and for his comments on the thesis. Bill Hickin and Dr Mitch Langford, helped with some of the computing. Figures 2.2, 4.1 and 4.7 were drawn by Kate Moore, and the aerial photograph of Narborough Bog given on page 84 was taken by Ian Paterson. I would also like to thank Mr and Mrs Cox for providing a home for the automatic weather station in their garden for the duration of the study.

Ruth Tarrant assisted with the field survey of Narborough Bog, and I would like to thank Dr. Brown, and my parents for continuing the measurements at Narborough, while I was temporarily absent from Leicester. Dr Brown also helped with the stratigraphic survey at Narborough.

The thesis was completed with the aid of the facilities of the Geography Department at Leicester, and I am grateful for the help of Dave Orme. Lastly, I would like to thank Irene, for her encouragement over the last year.

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Notation

Chapter 2		Units
ET	Evapotranspiration.	mm
P	Precipitation.	mm
q_{ov}	Overbank discharge.	m^3/s
q_t	Tributary stream discharge.	m^3/s
q_e	Effluent river seepage	m^3/s
q_i	Influent river seepage	m^3/s
q_r	Return flow	m^3/s
f	Infiltration	m^3
I	Infiltration rate	m^3/s
ΔS	Change in water storage	m^3
Chapter 3		
ϕ	Potential	The units of potential
ϕ_g	Gravitation Potential	can be a volume, mass,
ϕ_p	Pressure Potential	or weight:
ϕ_o	Osmotic Potential	(N/m^2); (J/kg) or (m)
ϕ_v	Velocity Potential	
h	Hydraulic head	m
Ψ	Pressure head	m
z	Elevation head	m
p	Fluid pressure	N/m^2
ρ	Mass density	kg/m^3
γ	Weight density	-
q	Specific discharge	m/s
k	Hydraulic conductivity	m/s
g	Acceleration due to gravity	m/s^2
∂	Partial derivative	-
θ	Volumetric water content	-
ε	Porosity	-
β	Fluid compressibility	m^2/N
α	Aquifer compressibility	m^2/N
S_s	Specific storage coefficient (= $\rho g(\alpha+\beta)$)	1/m
∇	Grad (Laplacian operator)	-
θ'	Degree of saturation (= θ/ε)	-
i	Hydraulic gradient	-

The Hydrology of a Floodplain Wetland, Narborough Bog, Leicestershire.

Chapter 1

Introduction

1.1. OBJECTIVES OF STUDY.

This thesis describes an investigation into the hydrology of a floodplain wetland, Narborough Bog, in central England. Narborough Bog is a designated Site of Special Scientific Interest, managed by the Leicestershire and Rutland Trust for Nature Conservation. The site was formerly one of the best local examples of alder/willow vegetation, with a reed-bed, and adjacent wet meadows. However, changes in ecology have been observed at Narborough Bog recently, particularly in the area of reed-bed, where there has been an increase in the abundance of species preferring a drier habitat.

This study examines the hydrology of the site at Narborough Bog, in order to investigate the extent of recent hydrological changes, and variations in the balance between water inflows and outflows. Although the study examines one floodplain site, the results have wider applicability as the balance of between water flows from different sources determine the characteristics of individual floodplain wetlands. Quantification of the importance of these water fluxes and development of a representative hydrological model can thus help wetland conservation work to be directed more efficiently. The information is also required to assess the impact of any future hydrological changes.

The significance of the thesis relative to other work on wetland hydrology lies in the nature of the approach adopted to the study at Narborough Bog. While there have been many studies of water-table fluctuations at various time-scales, most notably by Godwin (1931), the potential resulting from applying deterministic hydrological models to wetland sites has yet to be investigated as part of an integrated fieldwork and modelling study.

The approach has the advantage that monitoring of contemporary processes provides a record of fluctuations in water flux, whilst the modelling techniques enable assumptions concerning the flow of water through a wetland system to be investigated.

The aim of the study is to understand the hydrology of Narborough Bog, through monitoring contemporary processes, and also by developing a hydrological model for the site. Associated with this the work entailed the production of a data set comprising water-table and meteorological measurements within the area of Narborough Bog. The thesis seeks to provide a complete picture of the hydrology of the wetland site, and to this end specific objectives were identified as follows:

- **To derive a conceptual model describing the interaction of hydrological processes for a floodplain wetland.**

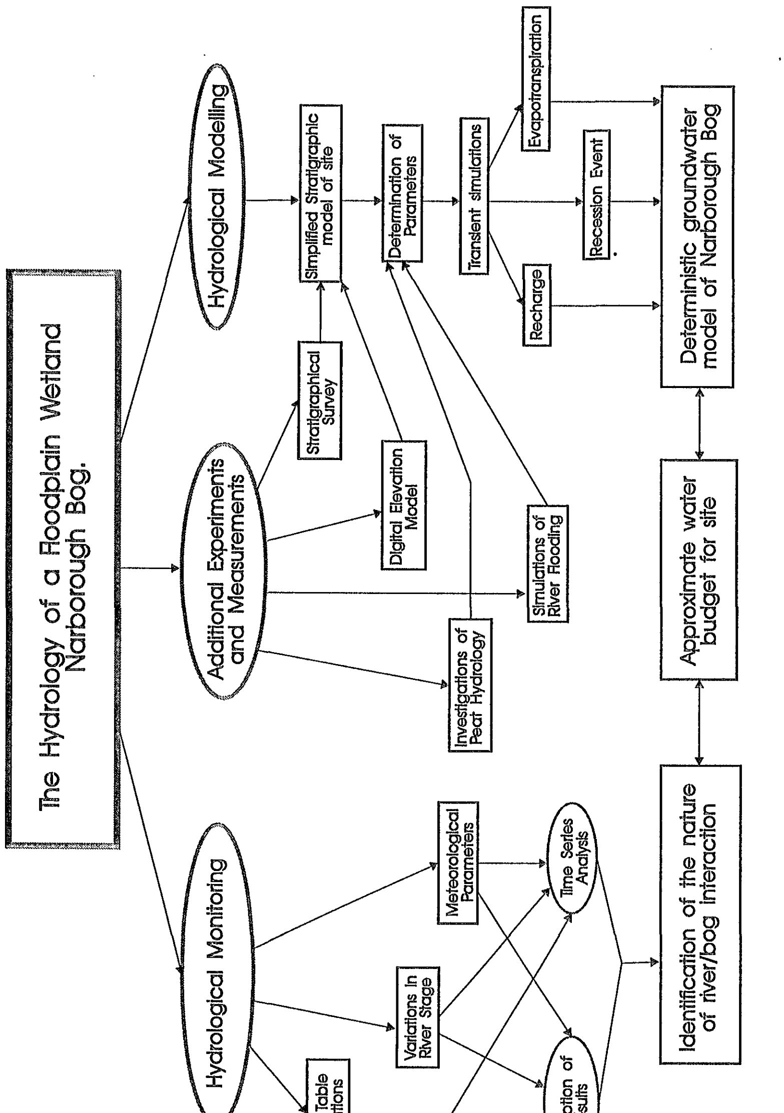
- **To examine the relationship of fluvial processes to the wetland at Narborough Bog; and to establish the balance of influent and effluent flows and overbank flooding.**

- **To produce a detailed stratigraphic diagram and consider possible simplification of the stratigraphy to represent permeability variations most effectively.**

- **To develop a deterministic computer model to represent the hydrology of Narborough Bog.**

The way in which these objectives are integrated within the context of the thesis is illustrated in Figure 1.1. The structure of the research project is centred upon the monitoring of contemporary hydrological processes at Narborough Bog, including measurements of water-table variations, meteorological variables, and river stage. These data are required to assess

Figure 1.1. Flow diagram illustrating the structure of the research project, and the interaction of different components.



whether there has been a significant lowering of the water-table at Narborough Bog. The results provide both a local base point to assess future change, and also help to describe the annual and seasonal cycle of changing water-tables. Here, time series techniques are used, to produce a stochastic model describing water-table fluctuations at Narborough. This method enables a preliminary examination of hydrological processes within the wetland, and particularly the response time which is required for floodplain water-tables to react to changing precipitation and evapotranspiration levels.

The hydrological concepts which underlie exchanges of water between riparian wetlands and a river system are hard to assess quantitatively. The relationship between these wetlands and rivers is not constant, and river water can form an inflow or outflow to the wetland system. It is thus important to study variations in fluvial processes, principally overbank flooding, but also influent and effluent flows and the significance of these processes to the hydrology of Narborough Bog. The importance of individual hydrological processes to the water budget of the study area, may then be considered.

Floodplain deposits are extremely heterogeneous, and differ markedly in their ability to conduct water. A complete stratigraphic diagram of the field site was required. The stratigraphy of the alluvial deposits provides information on the historical development of the wetland site, and would also be necessary before proceeding to study the variability of the hydraulic conductivity of the sediments at Narborough Bog. The implications of any heterogeneities in the alluvial deposits for the magnitude and direction of subsurface water flows are investigated. An emphasis is placed upon the processes of water flow through peat, as it has been suggested that Darcy's Law should not be applied in an unmodified form to peat deposits. Hence, hydraulic conductivity should not be used uncritically to describe water flow through peat deposits. Experimental investigation of detailed water flow through peat helps to justify the approach taken in the development of a numerical hydrological model for the site and also the choice of hydrological parameters.

The development of a deterministic computer model capable of explaining the observed patterns in hydrological behaviour represented an essential application of the water-table records which would permit calibration of a groundwater model for a specific area of the field site. The groundwater flow model MODFLOW, which has been developed by workers in the United States Geological Survey (McDonald and Harbaugh, 1988), was chosen for this study. The model enables the operation of specific hydrological processes to be isolated, for example areal recharge and river seepage. This enables the theoretical understanding of site hydrology to be tested.

Results from the modelling simulations, and also from the qualitative analysis of water-table records are then compared, both to assess the fit of the model, and also as a framework for describing the hydrology of Narborough Bog. In this section the importance of variations in water flow through the different wetland deposits are considered and a tentative water budget for the site is derived.

1.2. BACKGROUND.

Wetlands formerly covered extensive areas of both lowland and upland Britain, but have been lost at an increasing rate through developments in land drainage. An increasing amount of evidence suggests that individual wetlands may have an important hydrological role in regulating surface and groundwater resources (Carter, 1986), and controlling water quality (Hemond, 1980). Furthermore, wetlands serve as refugia for rare fauna and flora (Everett, 1989; Wheeler, 1984; 1988). Consequently the benefits derived from wetland conservation are becoming more widely appreciated. However wetland characteristics, their initial development, and hence the success of any conservation measures reflect the interaction between their geomorphology, hydrology and ecology. Wetland conservation therefore requires a sound understanding of processes, and in particular how wetland characteristics may be related to the balance between inputs and outputs of water to a wetland system.

This thesis considers the specific example of one alluvial wetland, namely Narborough Bog, which lies on the floodplain of the river Soar in Leicestershire. Although floodplain wetlands, such as Narborough Bog, represent only one type of floodplain landform, they illustrate the interdependence of geomorphology, hydrology and ecology. Initial wetland development occurs through a suitable combination of geomorphology and hydrology. Floodplains by definition are liable to occasional overbank flooding, and in certain areas the deposition of fine sediments with low permeability may impede subsurface drainage with the potential to produce locally perched aquifers. Small wetland areas may develop in areas of impeded drainage, for example where depressions intersect the floodplain groundwater-table, such as meander swales or river cutoffs. Larger wetlands may form in backswamps where flood waters are held in the area between levees and river terraces or the floodplain edge.

The study site at Narborough Bog is not a unique wetland and consequently the results from the investigation discussed here have wider applicability. Large areas of floodplain in lowland Britain were formerly covered by alluvial wetlands (Wilcox, 1933), of which only a small proportion remain intact. Substantial areas have been drained for agriculture, particularly in East Anglia; however several important wetland sites remain including the Amberley Wildbrooks on the floodplain of the river Arun in West Sussex, the site of Otmoor beside the river Ray in Oxfordshire, Wilden Marsh in Worcestershire, the Halvergate Marshes in Norfolk, and the Derwent Ings in South Yorkshire (Purseglove, 1988). Other significant wetland sites include the Washlands of the Fens and parts of the Somerset Levels, for example along the floodplain of the river Parrett (Williams, 1990a); Moseley Bog in Birmingham, the Wiltshire valleys, and also along the river Perry in Shropshire. However, few wetlands have survived unchanged, for example unsuccessful attempts were made to drain Otmoor in the 19th Century, and the remaining wetlands are vulnerable to hydrological change. Land drainage has been important in this respect, but river regulation also has significant implications for floodplain wetland systems which depend upon the water inflows provided by occasional overbank river floods.

The persistence of bog and fen lands relies upon the maintenance of a hydrological balance between water inflows and outflows to a wetland area. Consequently, the receipt of sufficient water inputs, whether from periodic high river flows, precipitation or groundwater flow, to supply evapotranspirative demand and any water outflows towards the river or to the groundwater body, is crucial. A suitable combination of meteorological and hydrostratigraphical conditions is therefore required as wetland occurrence is determined partly by topographic effects and partly by the relative geometry of floodplain sediments which have widely varying hydraulic conductivities.

It is generally accepted that there has been a global net loss of wetlands, which reflects consistent pressure to develop and drain land for agriculture. Precise figures for wetland loss, and particularly the loss of floodplain wetlands, are hard to obtain, however Maltby (1989) considers that 84% of the original total of lowland raised bog in Britain was lost over the period 1850-1978. The rate of loss of other lowland wetland types is indicated by the total land area which has been drained for agriculture in England and Wales since the 1800s, which has been estimated at 4.7 million hectares (Maltby, 1989).

Recognition of the conservation value of wetlands has prompted the introduction of statutory controls to restrict wetland drainage in several countries. The most comprehensive legislation exists in the United States, where the US Army Corps of Engineers has been given regulatory powers on rivers and wetlands under the Clean Water Act of 1977 (Williams, 1990b). In Britain wetland conservation is hindered by the absence of a structured conservation policy, and conflict between the National Rivers Authority which oversees water management and bodies responsible for land drainage which include regional land drainage committees and Independent Drainage Boards (Purseglove, 1988; Williams, 1990b).

In general, this loss of wetland habitat is occurring at the same time as increasing awareness of the importance of these sites in a wide range of environmental processes, including regional hydrology and hydrochemistry, which is in addition to their potential as sources of palaeoenvironmental information. The hydrological role of wetlands is incompletely understood at present; although evidently they may account for significant water storage, and under certain conditions groundwater recharge and stabilisation of river flows. This needs to be qualified by considering the effect of variations in wetland location within the hydrogeological framework of the floodplain (Carter and Novitzki, 1988).

The role of contemporary riverine wetlands as hydrochemical buffers appears to be more important than their hydrological role in sustaining the base flow of rivers at low flow conditions. Hence, the potential use of wetlands for water purification has been investigated and it seems that riparian wetlands may have an important function in taking up nitrates and phosphates (Burt and Haycock, 1992). In this way, conservation of alluvial wetlands may be justified as a result of improvements in river water quality.

Floodplain wetlands have additional value as they may provide palaeoenvironmental information which presents a record of vegetation change in lowland areas throughout the Holocene. This may represent the only surviving record of historical change in lowlands which lack lake and raised bog systems (Brown, 1988).

Floodplain wetlands have a rich ecology, and are highly sensitive to external hydrological changes that may affect the quantity and quality of water inflows. Consequently, individual wetland systems should not be viewed in isolation from hydrological processes occurring within the wider drainage basin, and particularly those processes affecting the water level of the regional and floodplain aquifer. Conservation of sites will therefore only be successful if both the local hydrology and also the interaction of wetlands with external hydrological processes are understood (Newson, 1992). Unfortunately quantifying the ecological response of a wetland system to a range of hydrological changes is extremely difficult. The problem is further complicated by the need to identify those elements of the ecological response that are associated with continued natural ecological change. This limits the ability to isolate the individual hydrological processes that may contribute to changes of wetland ecology and demonstrates the importance of maintaining hydrological studies over a long time period to obtain sufficient data to draw meaningful conclusions.

1.3. STRUCTURE OF THESIS.

Following this introductory chapter, chapter 2 considers how the development of floodplain wetlands may be viewed within the context of the geomorphology and sedimentology of floodplains. Processes responsible for floodplain construction are described and their hydrogeological implications discussed. A conceptual hydrological model for a floodplain wetland is then proposed, and the range of processes providing water inflows and outflows are discussed in detail. Chapter 2 concludes by outlining the principles of a water budget approach to a wetland system, which enables the relationship between the range of hydrological processes to be quantified.

The thesis emphasises the underlying hydrological principles both in defining the factors governing saturated and unsaturated water flow, and also in determining quantities of water loss through application of an evapotranspiration model. Chapter 3 provides much of the theoretical background to the study, by giving the derivation of the hydrological flow equations used later in the study. The chapter is focused towards processes relevant to wetland hydrology by considering the specific case of water flow through peat deposits. It has been suggested that the usual equations describing water flow have to be modified for the case of flow through peat deposits, and hence the limitations of this approach are discussed in detail. Modelling of evapotranspiration from wetland systems is also considered here with greater discussion of the variety of techniques used to determine evapotranspiration estimates than the summary given in chapter 2. This provides a background to understanding the development of the evapotranspiration model used in this study.

In the fourth chapter a full description is given of the study area, Narborough Bog, including its geomorphology, geology and stratigraphy. This is followed by a discussion of the fieldwork programme which begins by summarising the aims of the monitoring programme. The measurement programme, which included readings of water-table variations, and unsaturated water content, is then outlined, and the instrument design is discussed.

The fifth chapter forms the first discussion chapter of results and describes the records of variations in floodplain water-tables. The chapter begins by summarising some results from a previous period of data collection at the site, before presenting results from the three year research programme. The results are considered at different time-scales; the results for different years are compared; the annual cycle of water-table variations is examined; and specific events are studied in detail. The importance of spatial differences in water-tables is also considered. The results from the chapter are then drawn together using different time-series and regression-based models to help identify the hydrological processes responsible for the observed variations, and the possibility of temporal variations in individual processes is considered.

The sixth chapter presents the results from two specific experiments which were conducted to investigate particular elements of the hydrology of the site. The chapter begins by analysing records of overbank flooding, to assess the current contribution of flood waters to the hydrology of the site. The results of a controlled flood experiment are then described. A small area of Narborough Bog, near the river Soar, was flooded by pumping water from the river. The results provide an indication of variation in infiltration rates, and help to assess the hydrological implications arising from deposition of heterogeneous sediments within the wetland environment. In addition, a detailed experiment was undertaken to study the processes of water flow through peat. This was designed to determine water flow through an in-situ column of peat, to clarify processes of infiltration which would help identify a suitable approach for subsequent hydrological modelling simulations. Finally, further experiments are described in which hydraulic conductivity was determined using conventional laboratory and piezometer tests.

The seventh chapter presents the results of the application of the groundwater model MODFLOW. An introduction to the modelling technique is given first, followed by a short summary of the MODFLOW package. The application of the model to simulate specific events, consisting of steady decline in the water-table, isolated recharge events and periods of varying evapotranspiration, is discussed. The potential of the model to represent time-dependent problems is also considered.

In chapter 8 the results from chapters 5, 6, and 7 are drawn together and summarised. The picture of the wetland system obtained from the work is discussed and the hydrological implications raised by specific elements of the study are described. The data are interpreted within the framework provided by the water budget approach, introduced in chapter 2. Comparative water budgets are produced for the years 1991 and 1992, and records of precipitation for recent years are considered to examine the range in hydrological conditions experienced by the wetland. The chapter then proceeds to assess the contribution of the thesis relative to other work published on the subject. In addition the chapter considers the implications for management of Narborough Bog which arise directly from the results of the thesis.

Finally, chapter 9 concludes by summarising the results of the thesis.

Chapter 2

Literature Review

Scope of Chapter

The previous chapter outlined the objectives of this study which seeks to model the hydrological behaviour of a small floodplain wetland. In this chapter the roles of both historical and contemporary hydrological processes in determining the character of floodplain wetlands are discussed. Floodplain hydrogeology reflects the combination of a variety of historical processes responsible for floodplain construction, in addition to the contemporary river regime. Hence, after first giving a definition of a floodplain, this chapter describes the general sedimentology and geomorphology of floodplains before placing the location of floodplain wetland landforms within the context of a generalised floodplain hydrostratigraphy. A conceptual model for the hydrology of a floodplain wetland system is then described, and the operation of the range of processes responsible for initial wetland formation and maintenance is outlined. A framework for studying the wetland is provided by describing the application of a water budget approach to a floodplain wetland system. This requires an examination of the variety of water inputs, outputs, and storages to the wetland. Finally the relationship between wetland vegetation and hydrology is discussed, using the ecology of the reed, *Phragmites australis*, as an example.

2.1. INTRODUCTION.

Floodplains may be described using either geomorphological or hydrological criteria. Geomorphologically, the floodplain is an alluvial landform created by sequences of fluvial erosion and sedimentation. These processes may be associated with the contemporary river regime, or inherited from previous hydrological and sedimentological controls (Nanson and Croke, 1992). The hydrological definition considers the floodplain to be land adjacent to a river which is flooded within a certain return period (Leopold *et al.*, 1964). This may include land not considered to be floodplain under the geomorphological definition . The frequency with which different floodplains are flooded varies widely, enabling the identification of floodplains as either hydrologically active or inactive. Inactive floodplains may retain the characteristic floodplain geomorphology, although only experiencing flooding during extreme runoff conditions, while active floodplains are flooded at regular intervals. Inevitably, environmental processes will determine the frequency with which floodplains are inundated, while agricultural and urban development on floodplains also affect spatial patterns of river flooding.

2.2. HYDROGEOLOGY OF FLUVIAL DEPOSITS.

An investigation of the hydrology of a floodplain wetland system requires that the significance of a combination of surface and sub-surface water flows is recognised. Surface water flux is determined by environmental factors such as river regime and the frequency and magnitude of flooding in addition to topography; however, the magnitude of subsurface water flow reflects the ease with which water flows through deposits, and the direction of the hydraulic gradient. In this section, the processes of floodplain construction are described, concentrating upon the hydrological characteristics of the sedimentary assemblage produced. The following section will then discuss examples of floodplain environments at a larger scale.

Floodplains may develop through a combination of different processes, whether lateral point-bar accretion and vertical accretion during overbank floods in meandering river systems, or deposition within a braided river environment (Nanson and Croke, 1992). Deposition of fine and coarse sand and gravel within a point bar environment and erosion of outer banks enables a river to continue meandering. Overbank sedimentation during floods produces the deposition of a variety of sediment sizes. Coarse sediments, such as cross-bedded sands, are deposited first to produce levees as water velocity decreases with the initiation of overbank flooding (Price, 1973; Bates *et al.* 1992). Adjacent to the river the deposits are typically silty while at greater distances from the river the flow velocity decreases and fine silts and clays are deposited in a backswamp environment. The braided river depositional environment reflects periodic shifting of river channels, and localised sequences of aggradation and incision.

The significance of the combination of the floodplain depositional processes, described briefly above, is that the sedimentary deposits have widely varying permeabilities. This is illustrated in Table 2.1, which summarises the hydraulic conductivities of alluvial deposits. Internal heterogeneities of a deposit severely restrict the extent to which characteristic summaries of conductivity can be given; however, the extreme variation between gravel with a hydraulic conductivity of between 3×10^{-4} and 3×10^{-2} m/s and clay with conductivity ranging from 1×10^{-11} to 4.7×10^{-9} m/s is clear. These differences, which extend over several orders of magnitude, are close to

Table 2.1. Hydraulic conductivities of selected alluvial deposits (from Domenico and Schwartz, 1990).

Material	Hydraulic Conductivity (m/s)
Gravel	$3 \times 10^{-4} \rightarrow 3 \times 10^{-2}$
Coarse Sand	$9 \times 10^{-7} \rightarrow 6 \times 10^{-3}$
Medium Sand	$9 \times 10^{-7} \rightarrow 5 \times 10^{-4}$
Fine Sand	$2 \times 10^{-7} \rightarrow 2 \times 10^{-4}$
Silt, loess	$1 \times 10^{-9} \rightarrow 2 \times 10^{-5}$
Till	$1 \times 10^{-12} \rightarrow 2 \times 10^{-6}$
Clay	$1 \times 10^{-11} \rightarrow 4.7 \times 10^{-9}$
Unweathered marine clay	$8 \times 10^{-13} \rightarrow 2 \times 10^{-9}$

the maximum range for geological materials, and help explain the variability of rates of water transmission through alluvial deposits. They also demonstrate the significance of understanding the mechanisms under which localised clay sediments of low permeability are deposited, and which alter the subsurface flow net considerably.

2.3. FLOODPLAIN HYDROGEOLOGY.

The previous section indicated the amount of variation which has been observed in the permeability of alluvial deposits of gravel, sand, silt and clay. These deposits may have a very complicated distribution within floodplains; however, their location may be interpreted within the framework provided by the depositional models for floodplain construction. Thus the coarsest deposits, with highest permeability, are deposited as bed load. Point bar accretion produces layers of sand with intervening silt and clay, deposited on meander swales. Overbank deposition produces a sequence of deposits varying from sand in levees to silt and clay within backswamps.

The effect of these permeability variations on subsurface water flows is related to the combined sequences of floodplain construction and erosion which have occurred at any location. Individual floodplains acquire distinctive characteristics arising from variations in the combination of processes of vertical erosion and aggradation (scour and fill) and lateral erosion and deposition. Classifications of floodplains have therefore tended to adopt either a morphological or genetic approach. Morphological classifications consist mainly of describing floodplain landforms (Lewin, 1978); however, a genetic approach provides a more useful framework as it will include the importance of interaction between floodplains and river systems. Nanson and Croke (1992) produced a floodplain classification derived from considerations of available energy. Further factors they identified as important included: valley confinement, channel cutting and filling, accretion within a braided system, point-bar accretion, and overbank accretion.

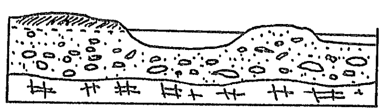
In Figure 2.1. a selection of generalised floodplain cross-sections are shown, illustrating the variety of stratigraphies associated with different floodplain morphologies. The important point to consider is the relationship of either an alluvial aquifer, composed of sand and gravel deposits, or a regional aquifer, with the river system. The floodplain environments are considered on the basis of the expected magnitude of the subsurface water exchange with the river system.

Firstly, the cross-section in A illustrates the stratigraphy associated with a braided river floodplain. These consist of wide alluvial deposits composed predominantly of silt and gravel, although there is frequently a fining-up sequence arising from limited overbank sedimentation. The hydraulic conductivity of the sediments is high, and hence the magnitude of subsurface water flows may be large. Although true braided rivers are rare in Britain, Ferguson and Werritty (1983) have examined channel changes in the 'wandering gravel-bed' River Feshie in the Cairngorms which is confined by valley walls in three reaches and is actively reworking glacial gravel deposits. Variation in the river's sedimentary load will determine the sedimentary composition of the floodplain with consequences both for bank stability and for the transmission of water through the floodplain sediments. The stratigraphy which arises from deposition in a braided environment is considered here because many lowland British floodplains overlie gravel deposits, derived from fluvial deposition during sequences of glaciation earlier in the Pleistocene. These deposits potentially represent a significant regional aquifer, with large transmission rates of subsurface water.

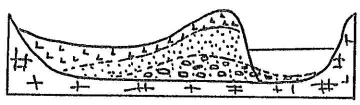
Secondly, in cross-section B, the profile for a confined sandy floodplain is given. Here, the floodplain accumulates through overbank deposition of sand and silt with large well-defined levees adjacent to the channel. In this situation it is possible for the river to maintain hydrological contact with the regional aquifer. Subsurface water flows are concentrated within the basal layer of coarse sand and gravel in the floodplain. Where suspended sediment transport

Figure 2.1. Generalised floodplain stratigraphies arising from variations in the sequence of erosion and deposition (from Nanson and Croke, 1992).

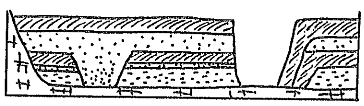
A Braided River Floodplain



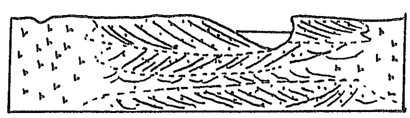
B Confined Sandy Floodplain



C Cut and Fill Floodplain



D Lateral Migration, Backswamp Floodplain



is significant, stable, cohesive banks may be produced, which are resistant to erosion. Consequently, deposits are dominated by the products of overbank sedimentation, supplemented in certain areas by levee sands, while periodic levee failure may produce crevasse splay deposits and perhaps cause a readjustment of channel direction.

In the third cross-section, C, the stratigraphy for a cut and fill floodplain is illustrated. This floodplain reflects the balance between periods in which sediment is deposited through accretion in abandoned channels and overbank sedimentation, with alternating phases of channel avulsion. Channel mobility produces a complicated distribution of coarse and fine sediments which affects the subsurface hydrology. Layers of fine sediment of low permeability within the floodplain restrict soil-water drainage and produce high water-tables. There is consequently only limited exchange with the river system.

Finally, the fourth profile, D, shows a lateral migration floodplain, with backswamp deposits distant from the river channel. This is produced by accumulation of the point bar as discussed above. Sand and gravel are deposited at the base, with a fining upwards sequence of sand and silt which may be reworked by secondary currents. This process only occurs in areas adjacent to the channel in situations where the river has limited power for lateral migration. At greater distances from the channel, floodplain construction takes place through sedimentation of fines in a backswamp environment. In this floodplain type, lateral subsurface water flow through the point-bar deposits will be limited, and exchanges with the river will depend upon the depth of fine deposits underlying the river channel. Variations of this model are characteristic of floodplains in lowland Britain where lateral channel migration is limited. For example, little channel movement is seen in the chalk areas of southern England, nor in the Broads of eastern England, enabling the accumulation of extensive peat deposits (Lambert *et al.*, 1960; Rose *et al.*, 1980). This lack of river movement may reflect low stream power due to the small differences in relief in large drainage basins (Brown, 1987), or the strength of river banks from local vegetation such as alder or willow (Ferguson, 1986).

The discussion above demonstrates how the subsurface hydrology of floodplains varies widely and also illustrates how classical models of floodplain evolution have developed. These typically considered lateral accretion to be the dominant depositional process with vertical accretion only contributing a fine mantle from overbank flood deposits at the surface (Wolman and Leopold, 1957; Stene, 1980). Floodplain characteristics are thus influenced by changes in valley morphology arising from erosion, and hence the resistance of the valley walls and floor to erosion has an important influence upon the character of floodplains. River base level and regime will also be important and similarly the sedimentary load. Variations in slope and sedimentary input along the long profile of a river enable a river to be braided in upper reaches while meandering downstream.

The importance of different factors to floodplain planform is illustrated by the work of Burrin and Scaife (1984) on the river Ouse in Sussex. In lowland Britain interpretation of floodplain development is complicated by the need to identify separate aggradational processes; for example, Holocene valley aggradation which followed climatic change and rising sea level, and also floodplain construction arising through anthropogenic activity, particularly land clearance, during the Neolithic. Burrin and Scaife (1984) described the interaction of different processes responsible for floodplain construction by identifying processes which they considered to be either exogenic or endogenic in nature. Exogenic factors include environmental processes, sea and base level changes, tectonic activity, anthropogenic activity and climatic activity, all of which are external and independent of the fluvial system, while endogenic factors are the internal responses of the floodplain system itself, which would include the effects of sediment deposition and consequent feedback and complex response effects.

In summary, differences in the rates of water movement over small distances in floodplain deposits can often be explained using the conceptual models of the sedimentology of meandering and braided floodplain types described above which provide a framework explaining the deposition of particular sediment types. Braided rivers typically occur in areas where there are extensive deposits of coarse-grained sand or gravel with high hydraulic conductivity, while finer deposits accumulate in dead water in abandoned channels. Meandering rivers also have quantities of these sediments but the proportion of silt and clay, which infill depressions and abandoned channels in the floodplain, and which have a lower hydraulic conductivity, is greater. Of particular importance for the magnitude of water flow are the characteristics of sediment lining the channel perimeter. Here, drapes of fine sediment, of low hydraulic conductivity, may significantly reduce the connectivity between river and aquifer.

Internal variation in hydraulic conductivity within a deposit is frequently significant in both a vertical direction, due to varying grain size associated with the typical sequences of upward fining, and horizontally, due to lateral changes in the process of floodplain construction. Frequently therefore, the value obtained for the hydraulic conductivity of a floodplain deposit in the field varies depending on the direction of measurement, indicating that the deposit is anisotropic. This reflects deposition and settling of particles which do not have a uniform, spherical shape, and also arises through layering which occurs with superimposition of different sedimentary materials having varying hydraulic properties.

Although the models describing floodplain construction provide some measure of explanation of variation in floodplain deposits, considerable interpolation is required to reconstruct floodplain stratigraphy from isolated bore-hole data. River channels frequently change location within the floodplain so that periodic episodes of floodplain deposition followed by erosion may produce markedly heterogeneous deposits, with consequent differences in rates of water transmission. As a result of substantial differences in the hydraulic conductivity of the individual deposits, the accuracy of interpolated stratigraphies from isolated data points has considerable implications for the success of groundwater modelling within floodplain deposits. Specific statistical techniques, such as kriging, have been developed to produce interpolated surfaces between isolated points where stratigraphy has been recorded, while in some cases stochastic models have been applied to provide a consistent means of applying the results of sparse data coverage (Anderson and Woesner, 1992; Peck *et al.*, 1988, chapter 5).

2.4. FLOODPLAIN WETLANDS.

A range of processes may contribute to the development of wetland areas within floodplains especially in situations where the water-table remains close to the surface, and significant quantities of water are added through overbank flooding. The hydrological properties of alluvial deposits are important in determining the direction of subsurface drainage, but also significant may be the flow of tributary streams which may locally raise the water-table enabling wetland formation. However, wetland extent may be limited by the presence of river terraces, and modern farming practices which frequently require drainage of wetland systems. In this section a general introduction to wetland systems is given before proceeding to look in detail at the sedimentology and hydrology of floodplain wetland landforms.

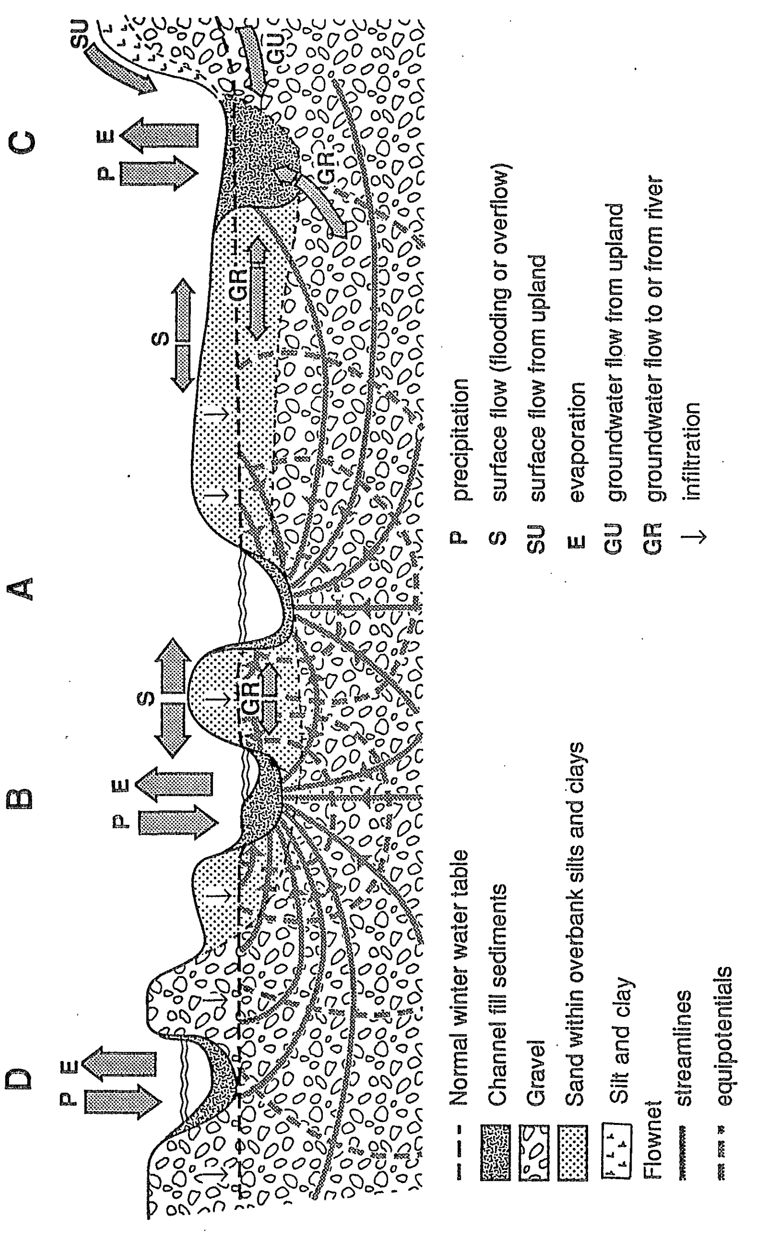
2.4.1. Classification of Wetland Systems.

The words mire, marsh, fen, bog, and swamp have often been used interchangeably in the wetland literature with varying definitions adopted in different countries. The problem was addressed by Tansley (1939) and more recently by Gore (1983) whose definitions will be summarised here. Marshes are generally considered to be non-peat forming systems which are waterlogged on a regular basis, although Moore (1987) considers that the term has been widely used to the extent that a strict definition can no longer be applied. Mire is a broad descriptive term for land having a high water content and includes all habitats in which peat accumulation is made possible by slow organic decomposition (Moore, 1984), while swamps are systems which are permanently under water. The definition of fen and bog and the difference between the two systems has been the subject of much debate. The distinguishing criteria normally used include wetland shape, chemistry and ecology. Typically bogs are mineral-poor and acid while fens vary from alkaline to neutral pH and mineral-rich. Moore and Bellamy (1974 p. 80) have simply defined the difference as being that fens are mires influenced by water from outside, while bogs receive only water from precipitation. These terms are not exclusive however, and have often been applied differently around the world. Thus applying these definitions strictly it is debatable whether Narborough Bog should be considered as a floodplain mire or fen; disregarding the question as to whether peat is currently accumulating at the site the definition would rely upon the significance of external water flows in determining the site hydrology.

2.4.2. Floodplain Sedimentology and Wetland Location.

A schematic section across a floodplain is given in Figure 2.2. showing the distribution of different alluvial deposits, and also possible locations for different wetland types. Specific wetland types may occur within a floodplain and can be differentiated initially on the basis of the level of their interaction with a river system, which may either be through surface flows such as overbank flooding, or subsurface flows, for example seepage to or from a river. In the figure three sediment classes are recognised: a gravel base, channel fill sediments and sand with overbank silts and clays. The wetland types which may develop include: the river itself (A in Figure 2.2.), a river cutoff (B), a backswamp (C) and a perched aquifer (D). For the wetland to be maintained, a balance between water inflow and outflow is required. All wetlands receive inputs of water through precipitation with evapotranspiration providing the principal water output. The magnitude of the latter may be close to the maximum potential evapotranspiration as water will probably not be a limiting factor because of the shallow water-table. In these conditions, evapotranspiration will follow a marked seasonal variation, due to prevailing

Figure 2.2. Possible wetland locations within floodplains, and their differing relationship to the river (A) including river cutoff (B): backswamp (C) and perched aquifer (D).



meteorological conditions and the annual vegetation cycle. Absolute quantities of precipitation will vary depending on a range of factors including aspect and altitude. The operation of these processes will be examined in further detail in the following sections.

The variety of possible wetland sites is illustrated by McKay *et al*'s (1979) study of the hydrogeology of wetlands within the floodplain of the Mississippi River in southern Illinois. In this area the valley fill sediments, mainly sands and gravels, are 35 m thick along the line of the valley. Surface geomorphology includes ridges and swales, natural levees, and old channels abandoned by chute and neck cutoff. The hydrogeology of the region is such that several wetland systems have been able to develop; primarily due to water inflows from the adjacent river. Deeper groundwater flow does not appear to be a significant water source. For example, McKay *et al.* (1979) discusses the hydrology of the La Rue Swamp which lies in an abandoned river cutoff. Two sources of groundwater inflow were identified: discharge from the alluvial aquifer and discharge from bedrock outcropping locally. Preliminary measurements suggested that the discharge of the latter, although continuous, was too small to be significant.

The widespread occurrence of peat deposits in lowland Britain demonstrates that historically wetlands covered a significantly larger area than at present. East Anglia has large areas of fenland that have been drained and are now used intensively for arable cultivation, while certain river valleys in southern England have an oligotrophic fen community, termed a valley bog by Tansley (1939), the ecology of which has been discussed by Rose (1953) and Wheeler (1984). Ombrogenous bogs are known to have evolved in Britain, however, generally these are no longer active and are not discussed here.

In floodplain areas, wetlands are likely to form where the land is periodically flooded and fine sediments, of low hydraulic conductivity, are deposited. Where extensive sand and gravel deposits occur influent river seepage may maintain the groundwater-table, with additional water inflow derived from overbank flows which are prevented from draining back to the river by levees. Such backswamps are thus limnogenous, as the river represents the principal water source. The peat produced is classified as telmatic, being formed at the water-table due to plants growing under conditions of periodic flooding. The landform assemblages associated with river floodplains provide appropriate opportunities for peat accumulation under conditions of high water content. Surface depressions on floodplains produced by former flood channels, sloughs, cut-offs, chutes, swales, flood basins and backswamps all may provide conditions required for the development of small wetlands. The development of peat is helped by deposition of fine silt and sand from overbank flows which have a lower permeability and produce anisotropic conditions. The variety of environments in which peat accumulates is seen in the floodplain of the river Severn where Brown (1983) describes peat deposits across the entire floodplain on the Perry and Ripple in a terrace palaeochannel at Hartlebury Common, in an overdeepened terrace channel at Ripple Brook, and a cut-off meander bend at Callow End. Rose *et al.* (1980) described typical peat backswamp deposits from the floodplain of the river Gipping which must have accumulated in an environment characterised by limited discharge variation and high rates of infiltration. The preservation of these deposits also indicates a long period of channel stability.

Geology is also important in two respects in determining the characteristics of peat deposition. Firstly, there is a hydrological effect as the permeability of the underlying sediments affects the direction and magnitude of water fluxes through the organic deposit. Secondly, there is the chemical effect upon nutrient supply which influences the resulting vegetation cover.

2.4.3. Hydrology of Floodplain Wetlands.

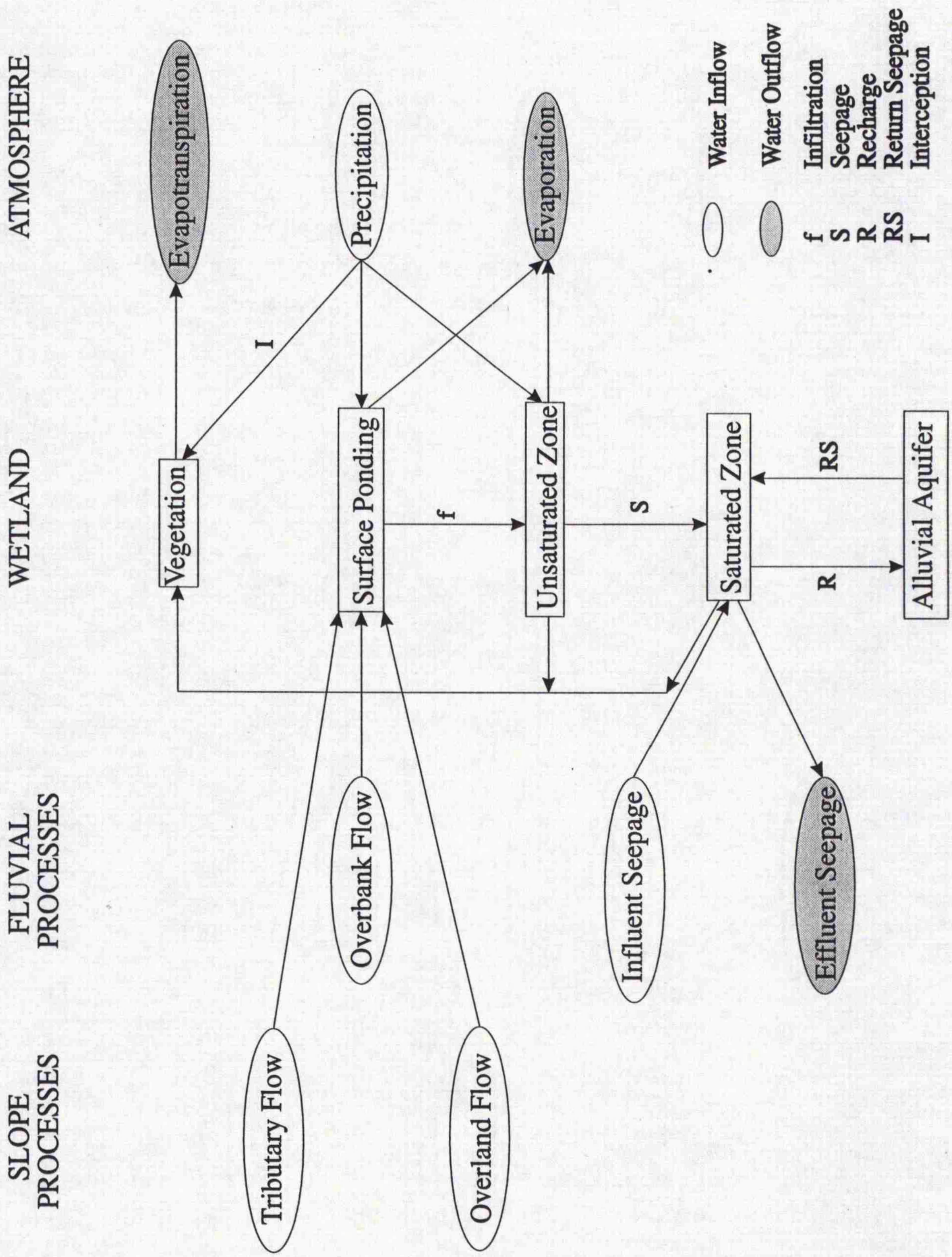
Wetland characteristics reflect both floodplain hydrogeology and also the operation of contemporary hydrological processes. The manner in which variations in fluvial processes contribute to the natural variability of floodplain sedimentology has already been discussed, in addition to the variability of the magnitude of subsurface water flows through floodplain deposits. In this section, the operation of specific hydrological processes which influence the quantities of water present within floodplain wetland systems are described. These processes are introduced with the aid of a simple conceptual diagram of a wetland system, before introducing a water budget model for a typical alluvial wetland.

The wetland environment can be considered in a similar form to the drainage basin in the hydrological cycle (eg. More, 1969), whereby distinct inputs and outputs of water to the wetland system are identified and within the wetland itself there are internal processes responsible for water movement between different areas of water storage. The internal form and combination of water inputs and outputs vary widely between wetland types and can be used for purposes of wetland classification. The principal processes contributing to water inflows and outflows for a floodplain wetland are shown in Figure 2.3. These processes can be classified as either hydrological or atmospheric in nature. Atmospheric inputs comprise precipitation; atmospheric outputs are evaporation and evapotranspiration. Hydrological processes account for the interaction of the wetland with an external hydrological system, principally surface water bodies and groundwater flow. These inputs of water include surface water flows arising from overbank river flow, and subsurface seepage either from a river (influent seepage) or from the wetland (effluent seepage), the direction of movement depending upon near-stream hydraulic gradients. Similarly, the relationship of the wetland to an underlying alluvial aquifer can be one of either local recharge or discharge, depending upon the direction of groundwater flow. In the case of riparian wetlands, limited flows of water are possible through an alluvial aquifer, the direction of movement of water being determined by the hydraulic gradient between the wetland water-table and river stage. The operation of hydrological processes within the wetland system and the balance between the different processes varies widely between wetland types and on a seasonal or annual time-scale as a result of the evapotranspiration cycle, the distribution of precipitation through the year, or specific water inflows arising from, for example, snow-melt (Carter and Novitzki, 1988).

Hydrological modelling of the interaction between wetland water inflows and outflows also requires study of the internal structure of a wetland and, in particular, identification of possible areas of water storage. A proportion of precipitation will be intercepted by vegetation where it may either evaporate or pass to the ground. Water at the surface may either infiltrate into the unsaturated zone, or directly to the saturated zone, from where it may contribute to groundwater recharge. Vegetation can extract water from both the saturated and unsaturated zones, from which water is lost directly through evapotranspiration. Saturated water storage is most important in terms of volume of water stored; however at the surface, storage of water may also be significant, either at times of high water-table, when any further water inputs are stored at the surface, or during overbank flooding. This water may remain at the surface or be lost as a surface outflow as flood waters recede.

The distinction between the unsaturated zone, the saturated zone and groundwater is not clear-cut. The unsaturated zone and saturated zone are separated by the intermediate capillary fringe, or tension saturated zone. Within the saturated zone, water is stored as shallow groundwater, the quantity of water storage being reflected by the position of the water-table. The groundwater term is used to identify possible interaction with groundwater flow at different scales. The influence of groundwater flow and recharge and discharge areas at a regional scale is probably not significant for this study directed at a floodplain wetland system; however the character of particular wetlands can be affected by localised patterns of groundwater flow. For example, Schot *et al.* (1988), Schot (1990), and Wassen *et al.* (1990a) discuss the impact of groundwater recharge from the sandy ridge the Het Gooi on the Naardermeer wetland on the floodplain of the river Vecht, in the Netherlands. Historically, this recharge contributed significant quantities of water which had

Figure 2.3. Flow diagram illustrating the nature of hydrological relationships for a floodplain wetland.



a low dissolved load but groundwater extraction has increased the flow of eutrophic water affecting the vegetation composition.

In the following sections the operation of the hydrological processes introduced above will be discussed in more detail. These include precipitation, evapotranspiration, subsurface water exchanges between the river and wetland, and overbank river flooding.

i. Precipitation

The contribution of precipitation in the hydrology of floodplain wetlands has generally been disregarded, with the exception of lowland raised mires; however over the annual time-scale precipitation may represent the largest water input to a wetland water budget. The relative importance of precipitation will vary spatially since in river marginal areas the relative contribution from river flooding may be of greater significance.

Most work on water-table response to precipitation has been in conjunction with attempts to produce time series models of the relationship between precipitation and water-table response and has not been concerned specifically with wetlands; however some of the conclusions discussed are also applicable to wetland systems (eg. Armstrong, 1988; Rennolls *et al.*, 1980; Viswanathan, 1983, 1984). Evidently, the importance of the precipitation contribution to recharge depends upon the characteristics of individual rain events in addition to preceding moisture conditions. Sophocleous and Perry (1985) studied natural recharge from precipitation over 19 months in Central Kansas. They concluded that intense rainstorms of short duration contributed less to groundwater recharge than longer duration rain events. This was a result of seasonal effects and particularly the magnitude of the pre-event moisture deficit within the catchment; precipitation of short duration was only sufficient to replenish the near surface water deficit with the result that recharge only occurred in late winter or spring. It was also found that the majority of the spring recharge was subsequently lost through evapotranspiration.

Freeze and Banner (1970) drew attention to how the quantity of recharge depends upon the combination of unsaturated and saturated conditions. They studied recharge processes within the Good Spirit Lake drainage basin in Saskatchewan and were able to identify only a limited number of important groundwater recharge events. Three separate hydrological events were discussed: a summer rainstorm, a summer recession period, and the effects of the spring snow-melt period. The rainstorm produced a reversal in the near-surface hydraulic gradient towards the river, with a rapid water-table rise, which was attributed to the infiltration of water as fracture flow. During the recession event a small amount of rainfall was sufficient to provide some groundwater recharge because of the shallow water-table and high moisture content. In the snow-melt event, a localised increase in the water-table occurred through recharge in depressions which produced a hydraulic gradient through the gravel deposits of the site.

In some situations it is hard to identify the quantity of water-table change which arises specifically from precipitation infiltration. For example, Rasmussen and Andreasen (1959) used a network of 25 observation wells to study water-table recharge arising from local precipitation and infiltration through an unconfined aquifer of homogeneous sand deposits overlying Miocene blue clay in the Beaverdam Creek basin, Maryland. Differences were observed in the magnitude of the response recorded in the dipwells to different rain events: the greatest variations in the water-table position being observed at measuring sites the furthest distance from the river. The relationship of baseflow to the position of the water-table was investigated and a rating curve produced. Rasmussen and Andreasen (1959) also studied the relationship of rainfall to water-table change and noted temporal patterns in divergence from the expected best-fit line. They considered that the varying position of the capillary fringe might be a possible mechanism for some of these observations, and also significant might be the effects of varying exchanges through the underlying aquifer. These mechanisms are described in more detail in chapter 3 (section 3.1).

ii. Evapotranspiration.

Evapotranspiration includes water loss through vegetation transpiration and evaporation from soil and vegetation surfaces. The quantity of water loss accounted for by evapotranspiration is significant and it is probably the most important source of water loss in wetland environments, principally because the water-table lies close to the surface, the capillary fringe extends to the surface, and wetland plants are not generally subject to levels of water stress (Duever, 1988). Phreatophytic vegetation, which extracts water directly from the saturated zone below the water-table, can withdraw water directly from groundwater storage during the day, producing a diurnal fluctuation in shallow water-tables (Godwin, 1931; Meyboom, 1967). In groundwater discharge areas phreatophytes are able to survive without an extensive root system; however they require the overnight recovery of the water-table. At this time plant stomata are closed, loss through evapotranspiration is minimised, and groundwater inflow may compensate for the daytime depletion to some extent. Influent river flows are thought responsible for nocturnal recharge in the Assiniboine River Drainage Basin of Saskatchewan (Meyboom, 1967). In some situations the daily water loss exceeds water inflow so that daily fluctuations are superimposed upon a falling base level. The precise form of the fluctuations will reflect the degree of hydrological isolation or connectivity of the system. Any short-term water-table fluctuations due to evapotranspiration are virtually absent over the winter and gradually develop through the spring at a parallel rate to plant growth.

Early studies of daily water-table variations (notably White, 1932 and Troxell, 1936) derived equations to describe the quantity of water lost, based on the rate of change of the water-table over time. Gatewood *et al.* (1950) undertook comprehensive investigations of water loss by phreatophytes in the Lower Safford Valley of Arizona which included use of lysimeters where plants were isolated in drums enabling isolation of other influences on the water-table.

Many studies have recorded the nature of such variation over varying time scales. Godwin (1931) studied the hydrology of Wicken Fen using adapted stage recorders to record water level change continuously. He noted a characteristic diurnal oscillation which was attributed to the time-dependent loss of water through transpiration. Similar results were recorded by White (1932) in Escalante Valley, Utah, and Gatewood *et al.* (1950) in the Lower Safford Valley, Arizona. Troxell (1936) discusses how the variation of the water-table through time is an accumulative record of the rates of groundwater inflow and transpiration use. Differences in water-table variation under different crops should therefore be attributable to different transpiration losses (given comparable hydrogeologic conditions), while Heikurainen (1963) noted two periods of water-table decline each day which corresponded to the timing of stomatal opening. Calculations of the quantity of water lost through evapotranspiration from phreatophytes in a controlled area prompted further study of the water balance, to determine if water availability could be increased by replacing deep-rooted species which were extracting water from depths as great as 30m.

White (1932) produced graphs showing the typical form of the diurnal water-table fluctuations under different types of plant cover; however only tentative conclusions can be drawn concerning their respective evaporative characteristics as the quantity of recharge is related to local hydrogeological conditions, and frequently uniform cover of a particular species is not found. Some differences in evapotranspiration rates between plants are to be expected for different species as a result of the physiological characteristics of particular wetland species. *Phragmites communis* provides a very good example. Smid (1975) examined stands of *Phragmites* in a fish pond in Czechoslovakia: his results indicated very low levels of moisture stress, no stomata closure, and hence no plant control over evaporation loss. The mechanisms of evaporation from *sphagnum* have also been studied in detail but will not be considered here as *sphagnum* is either absent or a minor component of most alluvial wetlands. Bryophytes and liverworts have no physical barrier to water loss through evapotranspiration; however vascular plants have cuticles with stomata which act as an effective barrier. Stomata are thought to have the greatest effect on evapotranspiration in plants with small leaves.

In general, wetland systems should not be subject to conditions of water stress for prolonged periods of time and consequently studies have frequently assumed that rates of wetland evapotranspiration should be close to the maximum potential evapotranspiration. However, this should not be assumed as evapotranspiration loss also depends upon the presence of a low humidity layer of air above an evaporating surface and a combination of processes may limit actual rates of evapotranspiration. For example, Linacre *et al.* (1970) compared evapotranspiration from a 3000 hectare wetland in Australia with evaporation from a nearby lake and found that wetland evapotranspiration was lower than that over the lake surface, and hence below the maximum potential rate. They attributed this to differences in albedo between the two surfaces and also to modification of the local wind patterns by stands of *Phragmites*.

Ingram (1983) describes how most work on wetland evapotranspiration has used indirect methods to determine water loss, for example treating evapotranspiration as the residual in the water balance equation. Energy balance approaches have been used but in this case studies are restricted by the limited extent to which wetland areas are sufficiently large to have their own micro-climate.

Although vegetation type and rooting depth influence the depth from which water may be extracted for evapotranspiration, the position of the water-table at any time is also important. For example, Romanov (1968) examined the water relations of dwarf shrubs on raised bogs in the Soviet Union. He found that the root system penetrated between 10 and 20 cm, and that evapotranspiration decreased sharply when the water-table fell below 40cm.

iii. Subsurface Exchanges between River and Wetland Systems.

Riparian wetlands have been classified as limnogeneous recognising that they receive substantial quantities of water inflow from rivers and lakes. Outflows of water also occur which may be important in sustaining the discharge of rivers at times of low flow. Several studies have documented the effect of river flows upon adjacent wetlands and the hydrological mechanisms involved are influent and effluent flows, which are described first. These subsurface flows are best considered separately to water inflows arising from overbank river flow which are discussed in section iv below.

Exchanges of water between groundwater and surface water bodies occur continuously under natural conditions. The magnitude of any flows depends on the hydraulic characteristics of floodplain sediments, in particular their ability to transmit flows quickly, and also on the size of the hydraulic gradient between the river and local water-table. The direction of water flow may change over short time-scales and during the course of a rain event, the controlling factors being the difference in hydraulic head between the river surface and the water-table. Influent flows refer to water flow from the river to the groundwater body, while effluent seepage refers to the reverse seepage of groundwater to the river. Different models can describe the relationship between river level and the water-table in temperate environments as illustrated in Figure 2.4. (Ward and Robinson, 1990). While effluent conditions may dominate in many situations as groundwater drains into a river (situation A), it is possible that after rainfall, or during snow-melt, river levels may rise more quickly than the water-table so that influent seepage replaces the previous process of effluent flow (situation B).

In most situations the majority of groundwater flow between the river and floodplain groundwater occurs through the river bed and not the river banks, as the river bed represents a greater area of the wetted perimeter, and the bed sediments have a greater hydraulic conductivity (Sharp, 1977). The river bank and bed tend to have differing sedimentological and hence hydrological characteristics, with smaller sediment sizes and higher clay contents in the bank material (Schumm, 1960; Sharp, 1977). Singh (1968)

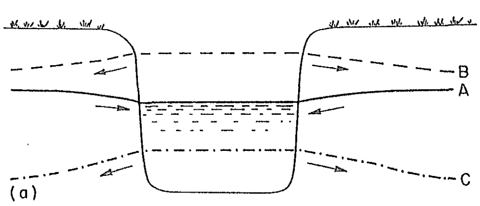


Figure 2.4. The varying relationship between river and groundwater: effluent flow (A), influent flow (B and C). (Ward and Robinson, 1990).

discusses how the relationship of water-tables and river levels also depends upon the degree to which the river is entrenched within the local aquifer. The implications of the sedimentology of river margins are that infiltration of river water into the underlying aquifer is limited if fine sediments are deposited on the river bed. River flow will limit deposition and high flows scour the river bed; however where flows are controlled, for example above dams, deposition will occur and infiltration will be limited (Kazmann, 1947). Such conceptual models of groundwater-river response are supported by field observations. For example, drawdown forces associated with seepage of water from a river into an underlying aquifer can inhibit the initiation of sediment movement; however the effects are likely to be small as deposition of a mud seal will restrict infiltration and mantle the river-bed sediments (Harrison and Clayton, 1970).

At peak river flow water may be temporarily stored along the channel margins as 'bank groundwater', whereby a large rise in river stage induces influent flow for the duration of the flood peak, the direction of flows being reversed as the flood subsides. This mechanism has been used to explain downstream moderation of a flood wave (e.g. Todd, 1955). It is possible that the occurrence of bank groundwater, may present a problem where rivers have high pollutant concentration which could contaminate groundwater supplies (Heij, 1989). The actual variation of the water-table in space and time will depend upon antecedent moisture conditions, particularly the position of the water-table and capillary fringe, river levels, the hydraulic properties of local sediments, and the intensity and duration of rainfall.

Granneman and Sharp (1979) found that the response of the floodplain water-table to a given river stage of the lower Missouri River varied between rapidly fluctuating to a slow response depending upon the proximity of the river and also the presence of clay plugs which restrict subsurface water flow. In some areas, down-valley flow was observed while elsewhere areas of consistently high water levels were found. The alluvial deposits were typically 30m thick, comprising c.5m of surface clay and silt, overlying c.20m sand on a gravel base. The deposits therefore had increasing hydraulic conductivity with depth. Significantly, of the four possible contributory sources to local groundwater - direct precipitation, overbank flooding, flow from bedrock aquifers, and infiltration from tributary streams - Granneman and Sharp (1979) considered percolation from sustained high river stage the most important. In certain situations it has proved possible to identify a simple linear relationship between the height of the water-table and river stage (Bell and Johnson, 1974), although there remains the problem of isolating water-table changes which arise from changes in river stage and those due to other processes, for example evapotranspiration.

Sophocleous *et al.* (1988) studied the hydraulic characteristics of alluvial deposits of the Arkansas River near Great Bend in Kansas where the near-surface deposits were coarse sands and gravels with isolated lenses of thin clay. Monitoring of the response of floodplain water-tables to drawdown induced by pumping was undertaken and it was found that the alluvial aquifer could support very high rates of water transmission (aquifer transmissivity c. 1800

m^2d^{-1}). River levels decreased during the time of pumping. Sophocleous (1991a) discusses further studies on the same aquifer. Areas of rapid water-table response to river levels were identified which he considers to arise through pressure pulses induced in palaeochannel deposits of coarse sand and gravel which have a high diffusivity. These pressure waves could theoretically be transmitted a distance of several tens of kilometres depending upon aquifer characteristics.

The mathematics of pressure induced changes in floodplain water-tables, analogous to the tidal effects discussed by Freeze and Cherry (1979), were described by Werner and Noran (1951). Evapotranspiration is considered to induce a short-term harmonic wave, while precipitation is approximated by an annual harmonic form. They consider the results of Troxell (1936) who recorded sinusoidal fluctuations of the Santa Ana River, and Prinz (1923) who looked at the floodplain of the River Elbe where variations in river levels induced related fluctuations in the water-table at distances varying between 0.3km and 4.6km.

Meyboom (1967) noted that a rapid rise of the water-table alongside the Qu'Appelle River in Saskatchewan tended to produce an immediate increase in streamflow which he attributed to pressure changes acting on soil water with the advance of the wetting front. The type of near-river vegetation is important in determining the nature of the relationship between river stage and the floodplain water-table. Phreatophytes extract their water directly from the water-table and hence rely on the overnight recovery of the water-table to survive. The consequent depression of the water-table due to transpiration losses from phreatophytes may cause streamflow depletion through direct interception of groundwater which would otherwise augment river discharge via effluent flow, and also by inducing influent river flow due to depression of the near-river water-table.

Specifically within a wetland area, Hurr (1983) studied the degree of connection between wetland areas on the floodplain of the Platte River, Nebraska, to river flow. Around the Mormon Island area much of the lowland is seasonal wetland. Water-tables were monitored over a 7 month period to examine the aquifer-river relationship and investigate the possibility of local water management schemes altering site hydrology. Local alluvium consisted of interfingering beds of clay, silt, sand and gravel. The local water-table was normally below river level as a result of high evapotranspiration losses and the consequent hydraulic gradient indicated that conditions of influent river seepage normally prevailed. Water-table position varied from the surface to 2.5 metres below the surface during the study. However, water-tables responded rapidly to changes in river flow. Evapotranspiration produced a diurnal variation in the water-table which induced smaller diurnal fluctuations in the flow of one of the river channels.

iv. Overbank Flooding.

Overbank flooding is the most immediately visible of the hydrological processes providing water flows onto a wetland site. Studies of overbank flooding have included qualitative description of processes operating (Lewin *et al.*, 1979; Hughes, 1980), mapping of floodplain areas inundated (Popov and Gavrin, 1970; Brown *et al.*, 1987), and modelling of flood water flows (Gee *et al.*, 1990; Bates *et al.*, 1992).

Inundation of the floodplain surface commences, especially along the line of old river channels and ditches, during rises in river stage as river banks are breached and overspilling of banks occurs. Waters from breach flow are ponded within topographic hollows such as old river cuttoffs, sloughs and backswamps. Flow from general overbank flows may be more important if river stage continues to rise but depends upon the degree to which bank vegetation or other structures restrict flow. For example, overbank flows may be confined by terraces, levees or man-made structures such as road and rail embankments (Lewin *et al.*, 1979). These limit drainage of water back to the river during recession flow and may also influence the direction of surface water drainage. Deposition of fine sediment in backswamp areas will restrict the infiltration rate, which decreases rapidly away from the channel thereby leaving standing water at the floodplain surface.

Hughes (1980) describes how the characteristics of an overbank flood vary depending on the combination of flood hydrology (eg. shape of hydrograph; magnitude of peak discharge; and antecedent moisture conditions), and floodplain type (relief and roughness). Under certain conditions there is a significant hysteresis effect whereby flood waters are retained within depressions so that the return flow of water back to the river is delayed. In many cases flood events coincide with conditions of high local water-table arising from precipitation or snow-melt (Popov and Gavrin, 1970) so that the infiltration rate is low and waters remain at the floodplain surface. Here the degree to which surface features modify the direction of surface flow may be important in determining the actual characteristics of a flood event.

The importance of a variety of internal wetland properties in determining surface water flows has been discussed by Hammer and Kadlec (1986). Surface water fluxes depend both on surface topography and surface permeability. Vegetation will also have a significant influence on water flows which can be modelled in a similar way to the Manning bed roughness, and a model incorporating vegetation roughness has been developed for the case of a subtropical marsh (Shih and Rahi, 1982).

The relative contribution of water inputs from overbank flows will depend largely upon the infiltration rate. The magnitude and duration of flood water storage at the wetland surface will depend upon minor variations in surface topography, the rate of infiltration and the flooding regime of the river. Given the great variability in the hydraulic conductivity of alluvial deposits, infiltration rates are likely to be highly variable. Lateral movement of water may therefore be rapid as soon as the river flood peak passes. Deposition of fine sediments during flood events will also reduce infiltration rates, thus limiting future infiltration during overbank events.

River flooding is also an important factor determining the ecology of floodplain areas. Mitsch and Rust (1984) examined variations in overbank flood events for the Momence Wetlands, in Illinois, consisting of forested bottomlands beside the Kankakee River. The flood series was reconstructed for the period 1917 to 1978 and it was found that the number of days of flooding per year varied from none in 13 years to a maximum of over 100 days in one year. There was only a poor correlation between the growth of water-tolerant species and flooding frequency due to the complexity of positive and negative relationships in riparian ecosystems, particularly in relation to competition from other species. Mitsch (1978) outlines how the nutrient supply derived from river flooding has contributed to the development of a highly productive, 30 hectare cypress swamp at Heron Pond adjacent to the Cache River, Illinois.

Gosselink and Turner (1978) describe how wetland types can be considered to occupy distinctive niches which are adapted to a particular combination of climate, geology, or hydrology. The specific hydrological processes which are significant include: water source, velocity, and the frequency and regularity of water inundation.

v. Fluctuations in Floodplain Water-tables.

In this section variations in saturated water content, as indicated by water-table fluctuations within floodplain sediments, are summarised. The importance of unsaturated variations are considered in the following chapter. An examination of the relationship between water-tables within floodplain sediments and river stage and surface water bodies suggests that the important factors will include (McKay *et al.*, 1979):

1. *Hydraulic Properties of the alluvial aquifer.*
2. *River Stage.* If there is sufficient water flow through the alluvial aquifer, (and hence also depending upon the hydraulic conductivity of the river bed and aquifer), the height of river flow provides both a local base level for

the groundwater-table, and a water source during short-term flood peaks.

3. *Distance from River.* The potentiometric surface of the water-table, graded from the river, determines the magnitude of fluctuating water-tables, which are typically greatest near the river.

4. *Geometry of the valley.* The geology of the valley enclosing the floodplain determines the direction of regional groundwater flow, eg. Gatewood *et al.* (1950).

5. *Small tributary streams.* Infiltration from streams may increase the local height of the water-table.

The floodplain water-table represents the depth at which aquifer fluid pressure is equal to atmospheric pressure. The position of the water-table reflects the interaction and balance of a combination of hydrological processes, a summary of which is given in Table 2.2. The factors of most significance for floodplain wetlands have been discussed above.

In the floodplain environment the natural processes which determine water-table position, namely precipitation, evapotranspiration, and water exchanges between a floodplain aquifer and the river, were discussed in the previous section; however other processes, for example agricultural irrigation and drainage, groundwater well extraction and drainage for engineering works, may be of local significance.

Changes in atmospheric pressure can produce large fluctuations in wells which penetrate confined aquifers, with an increase in atmospheric pressure producing a decrease in water-table height. In unconfined aquifers only small fluctuations have been observed in water levels in response to increases in pressure increases. Peck (1960) attributes this to the effects of pressure changes on air bubbles trapped in the unsaturated soil-moisture zone. As atmospheric pressure increases the pressure exerted on the air bubbles also increases and the bubbles deform to occupy less space. Water is consequently drawn up from the saturated zone into the unsaturated zone producing a fall in the observed

Table 2.2. Mechanisms producing Fluctuations in the Water-table (from Freeze and Cherry, 1979; p.230).

	Unconfined	Confined	Natural	Man-induced	Short-lived	Diurnal	Seasonal	Long-term	Climatic influence
Groundwater recharge	✓		✓				✓		✓
Air entrapment	✓		✓		✓				✓
Evapotranspiration	✓		✓			✓			✓
Bank storage	✓		✓				✓		✓
Tidal effects	✓	✓	✓			✓			
Atmospheric Pressure	✓	✓	✓			✓			✓
External loading		✓		✓	✓				
Earthquakes	✓		✓		✓				
Pumpage	✓	✓		✓				✓	
Well injection		✓		✓				✓	
Artificial recharge	✓			✓				✓	
Irrigation and Drainage	✓			✓				✓	✓
Geotechnical Drainage	✓			✓				✓	

position of the water-table. External loading can also create water level changes in confined aquifers, analogous to the effects of barometric pressure. For example, Jacob (1939) has described fluctuations in groundwater level induced by passing trains on water level in a well in Long Island, New York.

As a final comment to this section it is useful to add that human activity can alter floodplain water-tables. Gray and Foster (1972) described the processes affecting groundwater levels below the City of London, as determined from the drilling of boreholes. Although the principal sources of water and groundwater sinks can be identified, construction works along the margins of the river Thames have complicated quantification of water exchanges along the river bank. Furthermore the water-table may be depressed by pumping during large construction projects and along Underground routes. Significant infiltration is restricted to the vicinity of the main parks.

2.4.4. Floodplain Wetland Water Budgets.

Previous sections have described the principal inputs and outputs of water which may be available to a floodplain wetland. In this section the interaction of these water sources and sinks will be considered as an introduction to a water budget approach taking the field site at Narborough Bog as an example. The adoption of a water budget requires firstly the identification and recognition of the main water sources, storages and losses, secondly the recognition of linkages and transfers, and thirdly the quantification of these processes. Understanding the balance of water input and output flows, and quantifying their interrelationships is a necessary pre-requisite for hydrological modelling approaches which seek to investigate the effects of variations in hydrological flows.

For the purposes of a water budget inputs and outputs of water are brought together in a mass-balance type equation. This involves assumptions of continuity and conservation of mass and therefore requires that the flow chart in Figure 2.3. has identified all the important processes. At its simplest the water balance equation can relate water inputs to water outputs and any change in the quantity of water stored within the wetland (either within the saturated zone producing an increase in water-table height, or the unsaturated zone):

$$\textit{wateroutputs} = \textit{waterinputs} \pm \textit{storedwater} \tag{1}$$

A representative area of Narborough Bog is given in Figure 2.5. with the location and direction of the principal hydrological processes shown. As discussed in the previous section the main water inputs are precipitation (P), influent river seepage (q_i), and contributions from overbank flooding (q_{ov}), while tributary stream flow (q_t) and subsurface flows (q_{sub}) may be locally important. Outflows are evapotranspiration (ET) and effluent river seepage (q_e). Drawing these processes together gives the mass balance equation:

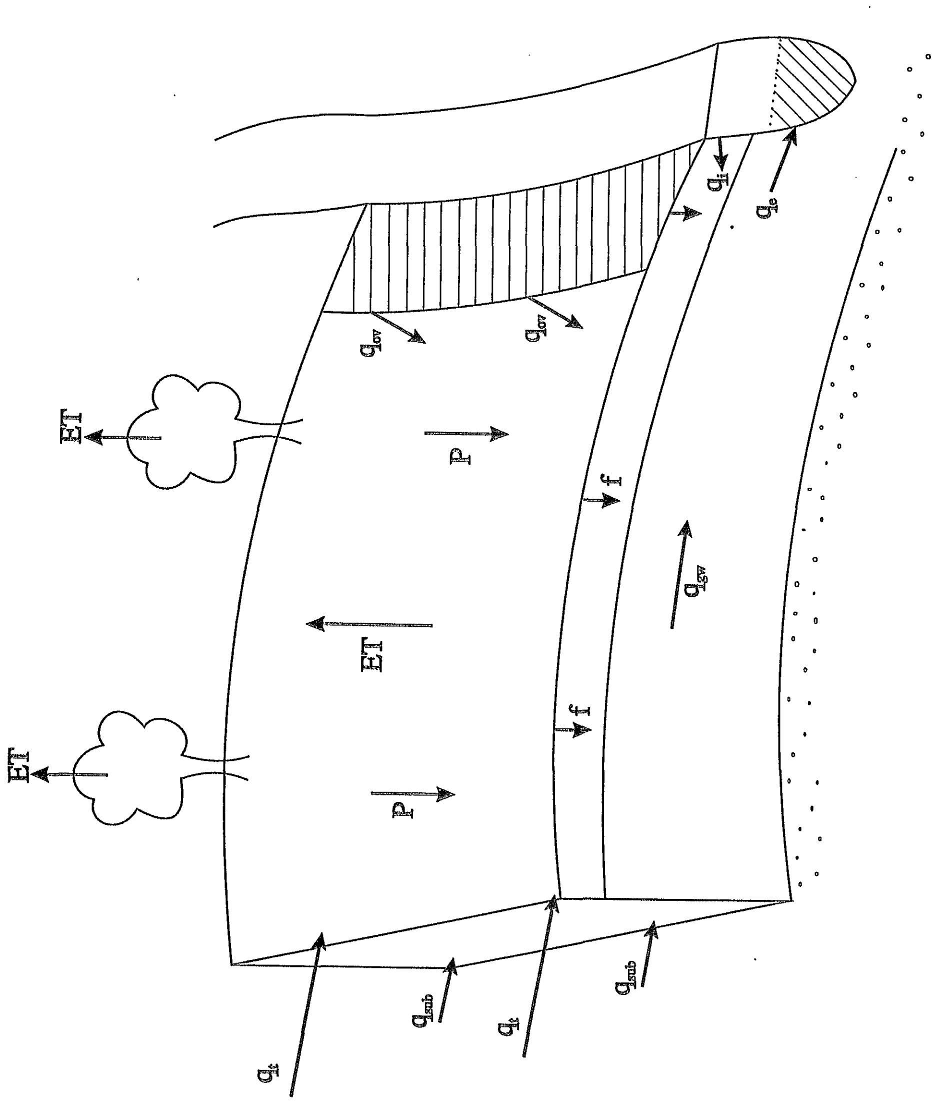
$$P + q_i + q_{ov} + q_t + q_{sub} - q_e - ET = \Delta S \tag{2}$$

Continuity of water flows is assured in the equation by maintaining a term ΔS representing changes in water storage within the area. Stream or river flow onto a wetland would clearly represent open channel flow; however in certain cases the actual contribution to local water storage will depend upon infiltration rates over the floodplain surface. The relationship of a wetland to an adjacent river may be dominated by influent or effluent seepage for much of the time, with diffuse surface flow occurring as river stage increases and some overbank flooding takes place.

Further expansion of equation 2 is possible; for example, some surface return flow to the river is likely following overbank flow, so that the actual contribution from overbank river flows can be described using the following equation:

$$\begin{aligned} q_{ov} &= f - q_r \\ f &= I \cdot H \end{aligned} \tag{3}$$

Figure 2.5. Schematic diagram of hydrological processes within a floodplain wetland. See text for explanation of symbols.



This indicates that the volume of water provided by overbank river flow (q_{ov}) either infiltrates to the water-table (f) or is lost through return flow to the river (q_r). The volume of seepage is determined by the Infiltration rate (I) and the hydraulic head (H) of the standing water. The actual contribution of overbank floods to the water budget is thus dependent upon the characteristics of the flood hydrograph and the residence time of each flood peak as discussed in section 2.4.3.iv.

As Figure 2.5. indicates, the importance of the various hydrological processes will vary spatially, the area flooded during overbank flood events forms only a small proportion of the overall study area, while the effects of seepage tributary streams may similarly be of local importance. However, using a water budget enables some consideration to be given to the relative importance of individual processes.

Water budgets are useful in calculating nutrient fluxes and in predictive models assessing the impact of external changes, for example changes in river regime or precipitation distribution. However, the accuracy of measurements must be considered, and the ease of determining different components varies widely so that, for example, the importance of interception has frequently been assessed by examining the residual of the water balance equation. This term will also include the total error which can be enormous as error budgets rarely accompany water or nutrient budgets (Carter, 1986).

Wetland characteristics are largely related to the balance between the terms in the water budget equation, principally because the corresponding nutrient fluxes influence the floristic composition of the system. Thus ombrotrophic bogs, which are sustained by precipitation, are nutrient poor, while wetlands through which significant regional groundwater flows are nutrient rich with a more diverse flora.

There have been few published wetland water budgets due to the difficulty in measuring all the hydrological processes. Koerselman (1989) produced a water budget for a small groundwater-fed fen in the Netherlands. Here, measurements were aided by the small area of the study site (0.3 hectares), but calculations of water flows were complicated by the variability of hydraulic conductivity. Boeye and Verheyen (1992) also looked at water flows within a small catchment; however they considered that regression analysis based upon water level measurements could provide sufficient indication of some internal water fluxes. For example, they determined changing water storage by the rate of change of water levels between measurement days, while drainage from the wetland was considered to be proportional to the difference between water levels within the wetland compared with an adjacent ditch network.

The problems generally associated with determining wetland water budgets have been summarised by LaBaugh (1986), in which quantification of groundwater and evapotranspiration loss frequently raises difficulties. In addition, a considerable range of techniques have been used to measure wetland water fluxes, and there has been little assessment of measurement error.

2.5. LOWLAND FEN VEGETATION.

The relationship between the composition of lowland fen vegetation, hydrology, and general environmental factors, is difficult to quantify. Wheeler (1984) has described how the significant characteristics may include variation in base saturation and pH; nutrient enrichment; water depth; water movement; and the effects of recent management whether for reed production, livestock grazing, or peat digging. In this section, the manner in which environmental characteristics affect the common reed, *Phragmites australis*, is used as an example. This is because the original motivation for the study lay in the observation of changing species composition of the reed-bed at Narborough Bog. The observations are therefore particularly directed towards the nutrient-rich floodplain fens.

Haslam (1972) describes how reed-beds may cover large areas of floodplain wetlands in East Anglia along the Waveney, Bure and Yare Rivers. *Phragmites australis* forms one stage in the process of hydroseral succession covering the transition from aquatic plants, to reedswamp, to increasing bush cover as frequency of flooding decreases, finally producing a fen carr. In dry areas meadowsweet (*Filipendula ulmaria*) or willowherb (*Epilobium hirsutum*) may become established. Typically *Phragmites* is the dominant species when the land is flooded to a depth exceeding 1m, although *Phragmites* is also physiologically capable of extracting water when the water-table lies over 1m below the surface. However, under drier conditions, invasion by other species such as *Filipendula* or *Epilobium* can be significant (Haslam, 1970), and these may out-compete the original wetland vegetation. This is therefore an example of a natural process of vegetation change, which is frequently accelerated when falling water-tables, which produce the drier conditions enabling other species to become established, also cause decomposition of organic deposits at the surface. This contributes to a nutrient-rich environment, suitable for a wider variety of plants. While a dense cover of *Phragmites* is able to smother most competitors, once established the invasive species can only be controlled by either summer flooding or changes in nutrient availability. *Phragmites* normally requires the maintenance of stable conditions, and although a period of sustained summer flooding may reduce the cover of herbaceous plants, one dry summer may lower the density of reed cover (Haslam, 1972).

It is frequently very difficult to establish a direct correlation between changes in hydrological conditions and vegetation cover. The problem is compounded in floodplain areas where firstly, the hydrological characteristics of surface deposits vary widely, and secondly, the frequency of inundation by flood waters also differs. In some of the large bottom-land forests of the United States theoretical relationships have been identified between vegetation cover, and height enabling vegetation cover to be used as a measure of river inundation (eg. Mitsch and Rust, 1984; Gosselink *et al.* 1981). However classifications of this nature are significantly harder in small floodplain wetland systems, where the differences in relief are less, and also where there has possibly been some ecological control applied.

An alternative approach to considering the linkage between vegetation and hydrology is to relate the nutrient content of wetlands, which arises from groundwater flow, to species content. Wetland hydrology studies have typically concentrated upon monitoring easily visible surface water fluxes; however, there is increasing recognition of the contribution of groundwater chemistry. Thus Wassen and Barendregt (1992) related the occurrence of productive fens of the *Phragmites* and *Filipendula* association to the inflow of polluted surface water, while low productive fens receive most water input from nutrient poor groundwater. Similarly in Poland, Wassen *et al.* (1990b) identified vegetation gradients in a valley mire system which were related to the combination of groundwater flow and flooding.

2.6. SYNTHESIS.

This chapter has introduced a variety of material not necessarily specifically concerned with wetlands. Published papers which discuss studies on precipitation, evapotranspiration and surface and subsurface exchanges of water between rivers and floodplains were also considered. The following chapter describes some of the hydrological processes in greater detail. However, in this section some of the studies of wetland hydrology are summarised to demonstrate their variations in approach.

In Table 2.3. further details on a selection of seven papers on wetlands, which have been introduced in this chapter, are tabulated. The papers all consist of field based studies of wetlands in four countries: the UK, Belgium, the Netherlands, and the United States. There has been little work published on alluvial wetlands in other countries although the valley mire at Biebrza,

Table 2.3.Summary of selected papers on wetland hydrology.

Paper	Wetland type and size	Monitoring Period	Comments
Boeye and Verheyen (1992)	Groundwater fen in Belgium; 25 ha.	Two periods of 10 - 11 months with weekly water-table measurements.	Applies techniques of linear regression to model the water balance as a function of rainfall, evapotranspiration and groundwater discharge.
Godwin (1931)	Wicken Fen, Cambridgeshire, c. 130 ha.	Two and a half years using a modified thermohydrograph to monitor water levels, supplemented by hourly measurements on two days in August.	A significant paper which considered seasonal water-table fluctuations in addition to response to rainfall and evapotranspiration. The relationship of water-tables to flow in drains was also considered.
Hammer and Kadlec (1986)	Porter Ranch peatland, Michigan, 700 ha.	Water depths monitored over two years - some data recorded continuously.	Good illustration of a wetland model incorporating surface water flux based upon a DEM of the site.
Heliotis and DeWitt (1987)	Northern Michigan peatland; 182 ha.	Two periods of 8 months data collection with continuous water-table and rain gauge records.	This illustrates the value of continuous records; and considers the peculiarities of peat hydrology. No consideration of the regional hydrology of the wetland system.
Hurr (1983)	Floodplain wetland adjacent to Platte River, Nebraska, 2590 ha.	Weekly measurements of 54 dipwells; 6 monitored continuously over an 8 month period. Two river stage recorders used.	A stratigraphic diagram is not given but the area appears to be dominated by recent alluvial deposits, with little organic material. The paper thus presents an extreme example of a wetland in good contact with a river.
Koerselman (1989)	Small fen, of 0.3 ha, in the Vechtplassen area of the Netherlands.	Weekly measurements of dipwells over two and a half year period.	Application of water budget approach. Includes estimation of hydraulic conductivity. Simplistic relationship to water positions in drains.
McKay (1979)	Three wetlands on the Mississippi floodplain in southern Illinois.	12 piezometers were installed; data are only given for two days.	The paper describes the development of small wetlands in a meandering river system where deposits vary from fine silts and clays to coarse sands; no organics were recorded. The processes of subsurface flow are reviewed.

Poland, described by Wassen *et al.* (1990b), represents an interesting field area.

The summary of papers places the early work of Godwin (1931; 1932) in context. This work provided a clear indication of spatial and temporal variation in water-tables at Wicken Fen, the value of which is enhanced by the lack of hydrological studies on British wetlands in the following years. The table demonstrates the extreme range in the size of wetlands studied. The wetlands range from 2590 hectares of floodplain wetland in Nebraska (Hurr, 1983), to the fen of 0.3 hectares in the Netherlands studied by Koerselman (1989). The papers by Hammer and Kadlec (1986) and Heliotis and DeWitt (1987) indicate the direction that wetland studies have taken in the United States. Here, field data have been collected for input into deterministic models of wetland hydrology. This represents a progression from the hitherto largely descriptive studies of wetland water-tables and is aided by the increase in use of data loggers and also, in North America, by the preservation of large wetland areas. However, the increase in modelling requires that the intricacies of water flow through wetland deposits is understood in detail and hence Heliotis and DeWitt (1987) investigated water flow through peat.

In Europe, the scale of studies is generally smaller and has been directed by the need to conserve those wetlands that survive; thus Boeye and Verheyen (1992) and Koerselman (1989) attempted to produce water budgets of individual wetlands. In both cases only weekly measurements were completed and the hydraulic conductivity of deposits was not measured directly.

In summary, there has been considerable difference in the scope of wetland studies and, as Table 2.3. indicates, there has been a similar range in the resolution of data obtained, both in the timing between readings and the dipwell spacing. None of the papers include a full stratigraphic diagram, detailed water-table data and direct measurement of hydraulic conductivity. It was to address these shortcomings identified above that the present study was designed.

Chapter 3

Wetland Hydrology

Scope of Chapter

Chapter 2 provided a broad overview on floodplain hydrogeology and introduced the subject of floodplain wetland hydrology. The summary of the water budget approach given in Section 2.4.4. provides the theoretical background to the hydrological study of Narborough Bog. In this chapter the quantitative techniques, introduced by the water budget section, are developed further. The chapter begins by reviewing the conditions of water storage within a wetland soil profile on a seasonal basis and then investigates the physics of subsurface water flow. These introductory hydrological concepts are then placed within the context of the study at Narborough Bog by considering the example of water flow within peat deposits. The chapter ends by examining the modelling of evapotranspiration from wetland systems. This section is included here because a physical understanding of evapotranspiration modelling is important before considering the Narborough Bog study further.

3.1. SOIL WATER HYDROLOGY.

The classifications of floodplain hydrostratigraphy, discussed in section 2.2., illustrated how the susceptibility of floodplains to flooding varies in relation to topography and river regime, while the sedimentology of floodplain sediments will determine water infiltration rates and the magnitude of subsurface water flows. Significant quantities of surface and sub-surface water may be stored within the floodplain environment. Surface water storage may occur at a variety of scales either through overbank flooding or heavy precipitation, in association with impeded drainage.

Subsurface waters can be distinguished as either water held in the saturated layer below the water-table which forms shallow, phreatic, groundwater, or water within the overlying unsaturated, vadose, layer. Different forces are responsible for water distribution within a vertical profile, including a combination of gravity, capillary forces, adsorption and, in conditions of uneven solute distribution, osmosis forces apply. Figure 3.1. gives a schematic water profile showing the different locations of water storage. The two zones are separated by the water-table, which is the surface at which the internal fluid pore pressure equals atmospheric pressure. The vadose zone is further subdivided into the soil water zone, an intermediate vadose zone, and capillary water. The capillary zone or fringe is an area within which all soil pores are water-filled and water is held through forces of surface tension on the water films surrounding soil particles. It extends from the water-table to the intermediate zone by a distance dependent upon the pore-size distribution of the deposit. The intermediate zone is an area of predominantly downward movement of water to the capillary fringe and the water-table. However, in summer, upward water movement may occur through surface evapotranspiration, which produces a reversal in the vertical moisture potential gradient. The soil zone is subject to processes of soil formation and may experience rapid change in water content through rainfall or evapotranspiration.

The relationship between hydrology and ecology in wetlands, and hence the role of hydrology in determining soil structure, has encouraged the use of a distinct terminology to describe a wetland profile. Ingram (1978) describes how Ivanov (1953) suggested the division of organic wetland soils into an upper layer of variable water-table and high hydraulic conductivity, and a lower layer in which the water content does not vary through time, and of low hydraulic conductivity. Ingram introduced the terms acrotelm and catotelm to describe the upper and lower layers respectively which has been widely adopted.

The vertical moisture distribution varies temporally in proportion to the balance between the various sources of soil moisture as illustrated theoretically in Figure 3.2. for a northern, temperate, environment. For the first three months of the year, water infiltrates directly to the water-table, producing an increase

Figure 3.1. Theoretical water profile, illustrating possible locations of water storage (from Domenico and Schwartz, 1990).

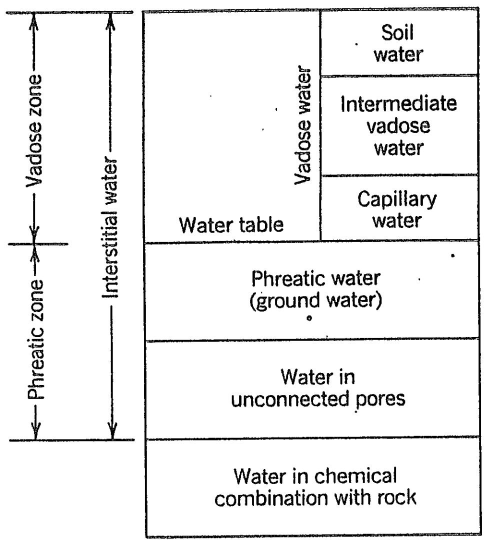


Figure 3.2. Annual cycle in vertical moisture distribution (from Wellings and Bell, 1982).

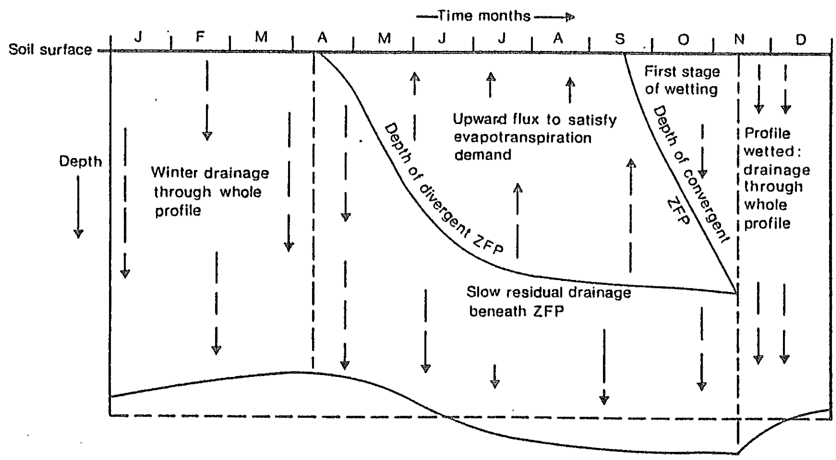
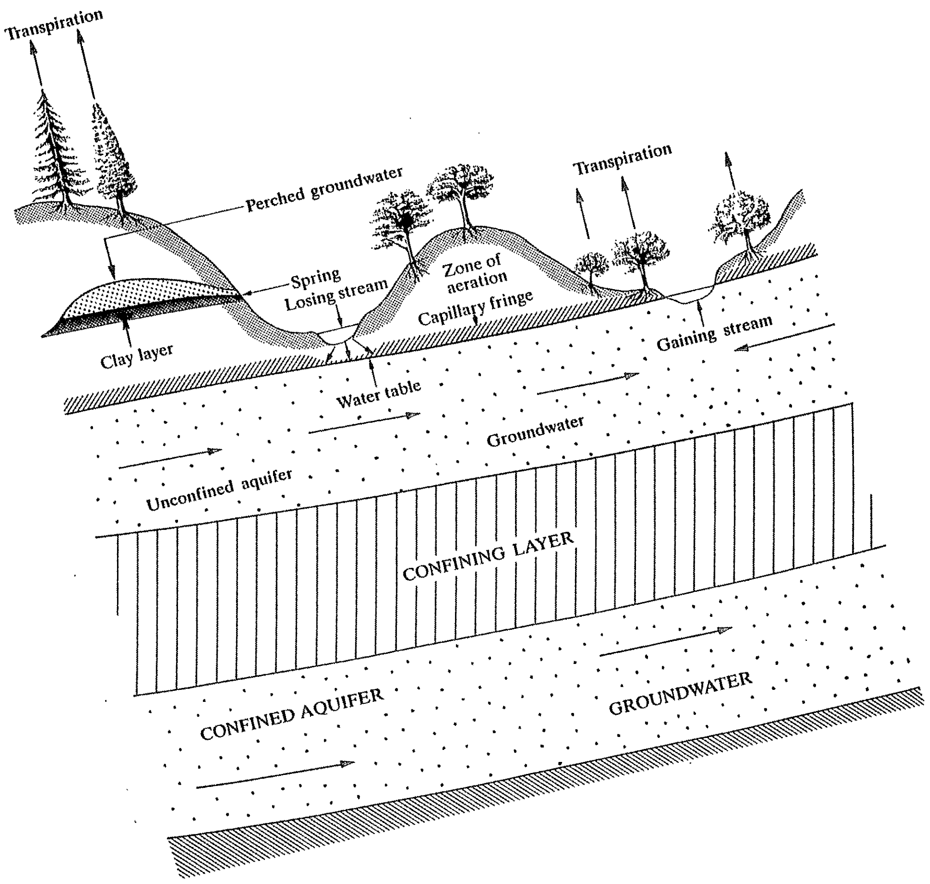


Figure 3.3. Schematic diagram illustrating different groundwater environments from Bras (1990).



in water-table height. From April, evapotranspiration begins and causes an upward movement of soil water, the zone of upward movement increases in the early summer as the evapotranspiration flux increases. A theoretical line, the zero flux plane, can be identified which separates areas of upward water flow from residual drainage to the water-table. In September, water accumulates at the surface and slowly infiltrates to the water-table, initially leaving an intermediate unsaturated area. The area of downward movement increases steadily until by mid-November infiltration of water is able to proceed directly to the water-table.

The position of the water-table within wetland systems changes through time in response to quantities of precipitation and evapotranspiration loss. It is therefore possible for temporary wetlands to be formed in a variety of locations during rain events as a result of intense precipitation and impeded drainage at depth. Essentially the wetland forms through a process analogous to Betson's partial area streamflow generation model (Betson, 1964), in which storm flow is understood to be generated regularly by the surface saturation of a small proportion, representing perhaps 10%, of an upstream drainage basin. Dunne and Black (1970) described how subsurface water flows could control the size of these contributing areas which were generally topographically low wetlands adjacent to the stream channel. Field measurements recording changing hydraulic heads in a small wetland on the Canadian Shield are described by Buttle and Sami (1992). Buttle and Sami used a combination of an isotopic and hydrometric approach, and examined whether the groundwater ridging hypothesis could be used to explain variation in hydraulic heads; however they found that some of the hydrograph peak consisted of surface water displaced within the wetland itself.

The physical processes governing the occurrence of subsurface water, therefore varies significantly, reflecting the interaction of seasonal variation in meteorological processes, with the hydrostratigraphy. The latter may vary over short distances as illustrated in the schematic cross-section in Figure 3.3. which shows a selection of aquifer types. Groundwater consists of water held in the saturated portion of soils and rocks. A material that contains water in appreciable quantities is known as an aquifer, while formations that transmit water at lower rates are aquitards, and aquicludes act as a barrier to water flows. Aquifers may be classed as either confined or unconfined. Unconfined aquifers have the water-table as the upper groundwater boundary where porewater pressure is equal to atmospheric pressure. In confined aquifers the upper boundary is provided by a less permeable bed. Groundwater is therefore subject to a pressure greater than atmospheric, so that if a well is sunk into the aquifer groundwater will rise to a height termed the potentiometric surface. Perched aquifers are a particular type of unconfined aquifer where the aquifer lies at some height above the main groundwater body, as a result of the local outcrop of an impermeable bed of sediment.

Considerable research has been undertaken on the interaction between water profiles, surface water, and precipitation. Much of the work has been within small catchments, and has examined the contribution of groundwater to runoff generation. The work is discussed in this section because of its applicability to the processes of water infiltration in an environment where the water-table lies close to the surface, and because the rapid increase in water-tables adjacent to streams and river requires a hydrological explanation. Separation of groundwater and surface water, with particular chemical and isotopic signatures, has enabled water derived from different sources, and event and pre-event water to be determined (Pinder and Jones, 1969; Sklash *et al.* 1986). Monitoring variations of the environmental isotopes oxygen-18, deuterium and tritium, in two small catchments below 5km^2 in area on the Canadian Shield, Sklash and Farvolden (1979) found that in most rain events the hydrograph was dominated by groundwater discharge. The only exception was following very high precipitation when the relative contribution of groundwater declined. To explain this observation, they advocated a mechanism of groundwater ridging entailing a rapid rise in hydraulic head along the perimeter of water discharge areas. Seemingly small quantities of infiltrating water, can produce a rapid water-table response. Such water exerts extra pressure on air above the capillary fringe; to compensate for this and obtain an equilibrium state the water-table rises an amount proportional to the depth of rain penetration and to the distance from the surface to the top of the capillary fringe. This has been termed the Lisse effect (Hooghoudt, 1947). The Lisse effect on the capillary fringe has been used to explain anomalously high fluctuations in the water-table. In some floodplains, the water-table lies close to the surface and any infiltrating water may convert the near-surface tension-saturated capillary fringe into a pressure-saturated zone, which has been termed a groundwater ridge (Ragan, 1968). This increases the hydraulic gradient both towards and away from the stream, providing a mechanism for the early displacement of groundwater into the river and increasing the groundwater discharge area. The groundwater may either enter the river or stream directly or flow over the surface first along a seepage face.

The capillary fringe therefore provides a mechanism whereby a small increase in water input may produce a rapid increase in the water-table. The response is most rapid in situations where the capillary fringe extends to the surface. The depth of the capillary fringe will vary along a slope profile, and especially near a stream where the water-table response may be most immediate (Abdul and Gillham, 1984; 1989). Abdul and Gillham (1984) produced a laboratory model consisting of a narrow glass box containing six tensiometers to demonstrate the effect of the capillary fringe depth on the rate of tensiometer response to rain. At their field site near Camp Borden in Ontario, on a sandy aquifer, Abdul and Gillham (1989) observed the production of a groundwater mound from small amounts of precipitation which enabled rapid runoff of event water via overland flow.

Within wetlands identification of the importance of a capillary fringe is complicated by the variety of processes which determine the response of wetland water-tables. For example, Heliotis and DeWitt (1987) studied the water-table response to precipitation in a wetland in Michigan, and identified three distinct types of response which they termed the Wieringermeer type, the Lisse type, and the storage response type following Meyboom's terminology (Meyboom, 1967). The mechanism of the Lisse response has been described above; the Wieringermeer effect occurs when the capillary fringe reaches the ground surface and the addition of a small amount of rainfall produces a rapid increase in water-table height followed by an equally rapid decline. The storage response type occurred at moderate rain intensities and produced a change in water-table position which was predictable given the specific yield of the peat deposit.

There is some difference of opinion over the significance of the capillary fringe to wetland hydrology. In a review of the hydrology of freshwater wetlands, Duever (1988) considered that the pore size structure of wetland deposits produced a pronounced capillary fringe, which he adjudged to be the major factor explaining the development of some wetlands. Boelter (1966) recorded a capillary fringe which was between 20 and 40 cm thick. However, Heliotis and DeWitt (1987) compared variations in water response from three wetland sites and concluded that a critical water depth existed below which a rapid rise in the water-table did not occur. They found this to be between 9 and 14cm. It is most likely that the capillary fringe will be of variable importance, as its size varies in a complex pattern in relation to the position of the water-table (Ingram, 1992). The size of the tension saturated zone increases as the water-table falls, and decreases as the water-table rises close to the surface.

The significance of the capillary fringe mechanism for providing rapid water-table rises within a wetland environment is likely to be limited by the variable pore sizes of wetland deposits. Gillham and Abdul (1986) consider that a capillary fringe effect, whereby infiltration of precipitation converts the capillary zone to saturated water, may produce a water-table mound in areas immediately adjacent to first order streams. However, the magnitude of the effect is heavily dependent upon the pore size characteristics of a deposit, and the water-table depth will determine whether capillary rise influences runoff generation. For example, Zaltsberg (1988) describes how the capillary fringe may range from 6.50-12.00 m in a sandy clay, but only 2-4cm in a coarse sand.

3.1.1. The Mechanisms of Water Flow.

Water movement between points occurs as a consequence of spatial differences in the potential energy of water, whether due to differences in pressure or temperature. Water flow acts to equalise any difference in potential. The concept of hydrological potential was first introduced by Buckingham (1907); however modern work on water potential largely follows Hubbert (1940) who defined potential as a physical quantity capable of measurement at every point in a flow system. The properties of potential are such that flow always occurs from regions of high to low potential, regardless of the direction in space. Fluid potential (Φ) is the mechanical energy per unit mass of fluid and consists of a gravitational potential (Φ_g) due to elevation, and a pressure potential (Φ_p). Additional elements sometimes included are a velocity potential (Φ_v), which may be important for marginal floodplain soils at breaks of slope and an osmotic potential (Φ_o), possibly significant for fine clay soils:

$$\phi = \phi_g + \phi_p + \phi_o + \phi_v \tag{1}$$

Generally the velocity and osmotic potential are considered negligible as velocities of water movement are small and Φ_o is relevant only over very short distances, thereby presenting an important scale distinction. Hydraulic potential therefore comprises the sum of the gravitational and pressure potentials. The gravitational potential (w_1) is equal to the work required to lift a unit of mass (m) through height z (i.e. w_1 = mgz), and the pressure potential (w_2) accounts for the work done to increase fluid pressure from p_o to p for volume V:

$$w_2 = m\int_{p}^{p_o} \frac{V}{m}\delta p = m\int_{p}^{p_o} \frac{\delta p}{\rho} \tag{2}$$

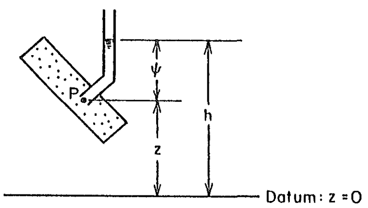


Figure 3.4. Illustration of hydraulic head, pressure head, and elevation head (from Freeze and Cherry, 1979).

Pressure potential varies with respect to the water-table and in particular between the saturated and unsaturated zones. At the water-table pressure is atmospheric (p_o), and increases with depth below the water-table as indicated by readings from piezometers. Above the water-table and where water is held within soil pores under forces of surface tension the pressure potential decreases. To simplify the mathematical treatment pressures are considered with respect to atmospheric pressure so that above the water-table soil water is held at negative pressures.

Fluid potential (Φ) is equal to the sum of w_1 and w_2:

$$\phi = gz + \int_{p}^{p_o} \frac{\delta p}{\rho} \qquad (3)$$

For incompressible fluids the density (ρ) is not a function of pressure (p) so that:

$$\phi = gz + \frac{p - p_o}{\rho} \qquad (4)$$

This may be simplified further by considering the manometer illustrated in Figure 3.4. (from Freeze and Cherry, 1979). At point P the pressure p is equal to:

$$p = \rho g \Psi + p_o \qquad (5)$$

where Ψ is equal to the pressure head, z is the elevation head, and h the hydraulic head.

Given that $\Psi = h - z$ and substituting equation (4) into equation (3) gives:

$$\phi = gh \qquad (6)$$

This last equation has the fundamental result that fluid potential at any point is equal to the hydraulic head multiplied by the acceleration due to gravity. However, the potential may be represented in terms of three different system characteristics: namely as a per unit mass, per unit volume, or per unit weight:

a) Energy per unit mass. $\phi = p/\rho = gh$

b) Energy per unit volume, or pressure. $p = \rho\phi = \rho gh$

c) Energy per unit weight, or hydraulic head. $h = p/\rho = \phi/g$

3.1.2. Hydraulic Conductivity and Darcy's Law

Most work on fluid flow relies upon the work of Darcy who examined the flow of water through a column of sand (Darcy, 1856). He observed that the rate of water flow between two points was proportional to the difference in hydraulic potential and inversely proportional to the distance, indicating a direct linear relationship to the hydraulic gradient (i). The hydraulic gradient is equal to the difference in hydraulic head (h_1 - h_2) divided by the horizontal distance (x_{12}). Darcy's Law can be summarised by the equation:

$$q = \frac{Q}{A} = -K(\theta)\frac{h_1 - h_2}{x_{12}} \qquad (7)$$

where $K(\theta)$ is a constant termed the hydraulic conductivity which varies with the water content and has dimensions of velocity, LT^{-1}. The negative sign in the equation indicates that water flow is in the direction of decreasing potential. Hydraulic conductivity depends upon viscosity (η) and hence is also a fluid characteristic. In some cases the intrinsic permeability (k) is specified instead where $k = K\eta/\rho g$ (Marshall and Holmes, 1988, p. 209). This allows the density and viscosity of the fluid to be included in the calculations.

Darcy's Law can be expressed in other terms also. For example, other indicators of potential can be substituted for hydraulic head, while the scale of investigation can be expanded to three dimensions through inclusion of unit vectors **i**, **j**, **k** in the directions x, y, and z:

$$q = -\left(K_x\frac{\partial\phi}{\partial x} + K_y\frac{\partial\phi}{\partial y} + K_z\frac{\partial\phi}{\partial z}\right) \qquad \textbf{(8)}$$

Darcy's Law implicitly contains certain assumptions including homogeneity, implying uniform hydraulic conductivity throughout the deposit; hydraulic continuity; and free drainage. Darcy's Law has certain limitations, although it is conceptually simple and valid for water flow under a variety of specific conditions. Hydraulic conductivity is considered to be a mean value for a given deposit, indicating that questions of scale may be important. As Freeze and Cherry (1979, p. 69) discuss, application of Darcy's Law requires that a deposit is represented as a continuum. The law may not be applicable for microscopic studies, as at this scale differences between the average linear water flow path through a medium and the actual flow path through pore spaces become significant, and inter-particle forces of attraction are important. This illustrates why the question of water flow through peat deposits is a problem. Darcy's statement of fluid flow is true for water flow through a granular based material, and where water flow is laminar. Where water flows through pipes, the Reynolds number provides an indication of when flow becomes turbulent, at which time water flow paths become irregular. Darcy's Law may also be limited where water flow velocity is small as a threshold potential gradient may be required for water flow to begin (Marshall and Holmes, 1988, p.85).

There are therefore several considerations when assessing the applicability of Darcy's Law. The actual magnitude of groundwater flow is highly variable, and the Darcian equation for water flow represents macroscopic water velocity through a cross-section consisting of a solid matrix and interstices; however the interstices will vary in shape, width and direction. Furthermore, the rate of water flow is greatly influenced by unsaturated flow as the replenishment of groundwater depends on downward movement of soil water through infiltration. For the hydrologist to appreciate the actual variation in groundwater flux, application of the simple Darcy equation is inadequate and the importance of the mechanisms of water infiltration, the velocity of head transmission, and other characteristics of the aquifer such as the coefficient of storage have to be considered. A discussion of the flow equations derived from Darcy's Law is required for assessing the limitations of a modelling approach as Darcy's Law provides the conceptual basis for groundwater modelling.

3.1.3. Flow through Saturated and Unsaturated Soils

Although Darcy's Law was developed using experimental results under saturated conditions, the equation can be applied in the case of unsaturated flow. In saturated soils the hydraulic conductivity at a point remains constant through time as all pore spaces are water-filled. In unsaturated soils hydraulic conductivity changes as a function of matric potential. Where water content changes through time, the continuity equation has to be applied in addition to Darcy's Law. This equation ensures that the rate of change of water content is equal to the difference between the flux of water that enters a volume and which exits, plus any sources of water located in the volume and minus any sinks. A storage term has to be used, which can be expressed mathematically as:

$$\frac{\partial\theta}{\partial t} = -\nabla \cdot q + sources - sinks \tag{9}$$

where $\partial\theta/\partial t = -\nabla \cdot q$ is the continuity equation and $\nabla \cdot q$ is known as the divergence of q:

$$\nabla \cdot q = \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} \tag{10}$$

Ignoring the action of any sources or sinks of water, Darcy's Law can be applied to equation 10:

$$\begin{aligned} \frac{\partial \theta}{\partial t} &= \nabla \cdot K \nabla \phi \\ &= \frac{\partial}{\partial x}(K_x \frac{\partial \phi}{\partial x}) + \frac{\partial}{\partial y}(K_y \frac{\partial \phi}{\partial y}) + \frac{\partial}{\partial z}(K_z \frac{\partial \phi}{\partial z}) \end{aligned} \tag{11}$$

For unsaturated flow, the direction of flow can be determined from Darcy's Law; however, complications are introduced by variations in matric and piezometric potential. The matric potential, or soil water tension is negative above the water-table, while the piezometric pressure is positive above the water-table. The physics is simplified by using the water-table as a reference elevation, where both the gravitational and pressure potential are zero. At increasing heights above the water-table the gravitational potential increases, but matric potential becomes increasingly negative, through upward directed moisture fluxes under the action of surface evapotranspiration, although at times immediately following rain events water may infiltrate vertically through the profile.

Unsaturated flow can be more readily understood by resolving flow into horizontal and vertical components (Ward and Robinson, 1990):

a) For simple horizontal flow only the matric potential drives the moisture flux:

$$q = -K(\theta) \frac{\partial \phi}{\partial x} \tag{12}$$

where K(θ) indicates that hydraulic conductivity is dependent on the water content, and is negative to show that water flux is directed to areas of decreased potential.

b) For vertical flow the gravity component is important giving:

$$q = -K(\theta)\frac{\partial(\psi - z)}{\partial z} \\ = -K(\theta)\left(\frac{\partial\psi}{\partial z} - 1\right) \qquad (13)$$

3.1.4. Infiltration.

The vertical passage of water into a soil profile is a complex process, which varies spatially and temporally. Infiltration varies in relation to several factors: the hydraulic conductivity and potential gradient at the surface, and also the rate of increase of total water storage in the soil profile. Under natural conditions the infiltration rate is frequently determined either by the rate of rainfall or snow-melt at the surface, however, infiltration below a flooded surface will attain a steady value termed the infiltration capacity. Early work on infiltration by Horton (1933) considered the example of semi-arid areas, where soils were noted to have a distinct infiltration capacity. The rate of infiltration was determined by the relationship of rainfall intensity and the infiltration capacity, and overland flow occurred where rainfall intensity exceeded the infiltration capacity. Translocation of soil particles during a precipitation event may contribute to a reduction in infiltration rates through the course of a storm, although where an impeding crust occurs the infiltration rate may only increase after a threshold value for the surface hydraulic head has been exceeded. Following the Hortonian model, infiltration rates (f_o) are initially high but fall to a limiting value (f_c) when the soil is saturated, after a certain time (t_o) when surface water stores become full, so that infiltration capacity (f_p) is given by:

$$f_p = f_c + (f_o - f_c)\exp^{[-k(t-t_o)]} \qquad (14)$$

The constant, k, should vary depending upon the infiltration characteristics of a soil type, however, it apparently has no physical meaning (Marshall and Holmes, 1988, p. 119).

In practice, overland flow rarely occurs by this mechanism apart from the specific case of semi-arid areas, since cultivation practices or desiccation cracks may provide a suitable route for percolating water, while vegetation enables larger quantities of water to be stored at the surface. The complexity of infiltration processes is illustrated by the work of Bodmin and Colman (1944) who identified a number of soil zones through which water passed. The immediate surface area is saturated during precipitation; below this lies a transition zone where water content decreases considerably from top to bottom. Below the transition zone is an area of transmission where water moves with little change in water content; water content then decreases sharply in the wetting zone, which is formed through the downward passage of a wetting front. The mechanics of water movement are complicated by changes in pressures during the course of infiltration. Advance of a wetting front over a wide area towards the water-table can trap significant quantities of soil air; the resulting increase in pressure will provide some resistance to water movement.

Drainage of water to the water-table may be treated approximately by the Green-Ampt method. Childs (1969) describes how Green and Ampt (1911) envisaged the advancing wetting front to comprise a well-defined surface at which the soil pressure is H_f, while the pressure head at the surface is H_o. If l is the depth to which the water front has infiltrated in time t, using Darcy's Law the rate of change of the water content with time is:

$$\frac{\delta Q}{\delta t} = K_{sat} \frac{(H_o + l - H_f)}{l} \quad (15)$$

This equation presents the advantage that infiltration rates can be determined easily from experimental data.

The infiltration process has been modelled mathematically by Philip (1957) who derived analytical functions, based on the Richards' soil moisture equation, to describe the variation of the infiltration rate with time following the start of infiltration. He recognised that infiltration followed a polynomial function, so that given a curve of known form, the infiltration rate (f_t) at time t could be determined from:

$$f_t = \frac{1}{2} S t^{-\frac{1}{2}} + B \qquad \textbf{(16)}$$

Here, S and **B** are coefficients, which are functions of soil water diffusivity and initial and surface moisture contents. **S** represents sorptivity, indicating the ability of the soil to absorb water, while **B** is equal to the hydraulic conductivity over long time intervals (Marshall and Holmes, 1988, p.139).

3.1.5. Macro-pores.

The process of infiltration and water flow through soils can be greatly affected by the presence of continuous openings in field soils, termed macro-pores by Beven and Germann (1982). These structures in the soil profile may be produced by biological processes, for example by the action of burrowing animals, earthworms and plant roots, but macro-pores are also produced by desiccation cracking, and from erosion by subsurface water flows.

The possibility of flow through macro-pores provides a mechanism for rapid infiltration of water through the soil, thereby contributing to a fast increase in the water-table and in river hydrographs following large precipitation events. Table 3.1. presents some figures for published observations of water flow rates through macro-pores given by Beven and Germann (1982), with comparative data for matrix flow obtained from Table 2.1. Flow rates through macro-pores are comparable to rates of flow through a gravel aquifer, and are several orders of magnitude greater than water velocities through clays. Macro-pore flow represents a distinct exception to the Darcian approach to subsurface water flow, and limits the application of the infiltration models discussed above. It is therefore important to identify the conditions under which macro-pore flow is likely to be significant. It seems to be most important in clay-type soils, which are susceptible to desiccation cracking.

Table 3.1. Recorded Rates of Soil Water Movement.

Average maximum recorded velocity through macro-pores from Beven and Germann (1982; Table 3)	0.0082 m/s
Gravel (mean value from Table 2.1)	0.003 m/s
Medium Sand (Table 2.1)	0.00025 m/s
Clay (Table 2.1)	2.4×10^{-9} m/s

Rainfall intensity will also be important as macro-pores will only conduct water when the soil potential exceeds the attractive forces which are responsible for water retention in the soil profile. The possibility of macro-pore flow thus needs to be considered within the floodplain environment; where although the hydraulic conductivity of the deposits varies significantly, water flow through macro-pores may increase the effective hydraulic conductivity of some alluvial deposits, particularly clays.

3.2. PEAT HYDROLOGY.

An understanding of the various processes of water flow through peat is important for a number of reasons. The magnitude of water flows, and variations in chemical water quality partly determine the ecology and geomorphology of wetland types, whether an upland raised bog, or floodplain mire. Early studies investigated the hydrology of surface peat deposits and considered possible techniques of land drainage for agriculture, however, the example of water flow through peat also exemplifies the limitations of Darcy's Law to describe subsurface flow at the micro-scale. These studies will be considered here since developing a hydrological model describing water flow through floodplain peat deposits requires an appreciation of the hydrological relations of peat soil and an ability to determine the quantity of mobile water (George, 1975).

Summaries of peat hydrology are complicated by the general use of the term peat to describe partially decomposed organic matter, with a high water capacity. Internal variability of a peat deposit can be considerable, while differences between individual peat deposits reflect the potential range in botanical composition of the peat. There are corresponding differences in morphology, complexity and texture of peat. General peat classifications have been based upon the position of the deposit relative to the water surface, the manner of deposition, nutrient supply and topography as described briefly in Section 2.4.2. (West, 1977). A peat classification system based upon differences in peat structure has been described by MacFarlane and Radforth (1968), and this is summarised below as it provides an introduction to the physical process of water flow through peat.

Peat structure varies depending on the morphology and arrangement of individual organic elements at different scales. Structural variability determines the water relations of the peat, and the magnitude of water retention or expulsion. Typically an inorganic soil is composed of soil horizons, with varying aggregation and pore space, in which any water present can be identified as either static water, which is absorbed onto mineral grains, or mobile water, held within interconnected pores. In contrast, a well humified peat is composed of densely packed hollow organic particles, of which 10% by weight at saturation may consist of organic matter the remainder being water and solutes. The majority of the water is not free to move, but is closely bound to old plant cells, or to hygroscopic colloids, and only the water held within large pores is mobile and will flow according to Darcy's Law.

The primary elements of peat are the remains of its original botanical components, while secondary elements are formed as a consequence of subsequent physical and chemical decomposition. Different structural models should be applied for peats with a high colloidal fraction compared with fibrous peats. Examining the internal structure of a non-woody fibrous peat under a microscope, MacFarlane and Radforth (1968) identified a hollow central axis and a spiral of leafy appendages. The outside diameter of the tubes varied from 0.24mm when wet to 0.216mm when dry, representing a reduction in volume of 19% between the wet and dry states. This suggests a possibility of varying water flow with water content, and also variable water flow through peat deposits, either through the peat matrix or structural voids (i.e. macro-pores).

Internal heterogeneity of peat has obvious hydrological implications as undecomposed moss peats can contain more water in easily drained pores and release 50-80% of their water content to drainage. Decomposed and herbaceous peats retain more water in smaller pores. Bryophytes, such as mosses, are permeable at the molecular level, while vascular plants, for example *Phragmites australis*, are protected by a waxy external structure; they do not decompose readily and are not permeable. Peat deposits also exhibit anisotropy, as their permeability varies with the direction of measurement due to the preferential orientation of incompletely decomposed vegetation. Mathematical modelling of flow through peat is therefore complicated as it is necessary to account for the possibility of differing flow through the peat matrix and macro-pores. This is in addition to any problems in measuring hydrological parameters such as the hydraulic conductivity, which describe the permeability of peat deposits.

Other complicating factors include the wide variation in the characteristics of peat itself and variation due to the different techniques used to determine hydraulic conductivity. Ingram (1983 p.134) outlines how the hydraulic conductivity of peat varies with the botanical composition of the peat, the degree of humification, bulk density, fibre content, porosity, water yield content, and surface loading. The relationship of these variables to hydraulic conductivity is summarised in Table 3.2. Rycroft *et al.* (1975) summarised the results of several studies of hydraulic conductivity, and some of the data is given in Table 3.3. Peat is similar to other soils as problems of scale have been observed when comparing laboratory and field measurement, possibly arising from local heterogeneites in the deposit but also the considerable problems in

Table 3.2. Relationship of material properties of peat to Hydraulic Conductivity.

Property.	Effect on Hydraulic Conductivity (K).
1. Botanical composition	Sphagnum peat is the least permeable; *Carex* and *Phragmites* peat is the most permeable.
2. Humification.	K is inversely proportional to the degree of Humification.
3. Bulk Density.	K is inversely proportional to Bulk Density.
4. Fibre content.	Fibre content is inversely proportional to Humification and hence is directly proportional to K.
5. Porosity	Porosity is proportional to humification, and inversely proportional to K.
6. Water yield	Water yield is inversely proportional to K.
7. Surface loading	Loading is inversely proportional to K and acts to increase Bulk Density.

Table 3.3. Variation in published hydraulic conductivities of peat deposits.

Reference	Peat Type	Hydraulic Conductivity (m/s)
Boelter (1965)	Moss peat 15-25 cm deep	3.8×10^{-4}
	Moss peat 45-55 cm deep	1.04×10^{-5}
	Wood peat 35-45 cm deep	4.96×10^{-5}
	Herb. peat 70-80 cm deep	7.5×10^{-8}
Chason and Siegel (1986)	moss peat 3 m deep	5.0×10^{-5}
Sturges (1968)	Decom. peat 46 cm deep	2.8×10^{-9}
	Decom. peat 91 cm deep	1.8×10^{-9}
Ingram (1967)	Fibrous mire 50 cm deep	9×10^{-10}
	Fibrous mire 100 cm deep	6×10^{-10}
Dai and Sparling (1973)	Sphagnum peat 50 cm deep	6×10^{-5}

obtaining a representative sample of peat (Chason and Siegel, 1986).

Where macro-pore flow is important, application of in-situ tests presents certain advantages in enabling account to be taken of the structure and distribution of macro-pores through a sample. Macro-pore flow may be important for much water flow through a peat body under natural conditions of infiltration. However, where there is a dense vegetation cover, interception and the presence of rootlets may reduce the hydraulic head at the surface so that water flow may occur entirely through the peat matrix.

Saturated hydraulic conductivities in peat soils have normally been determined in the field using modifications of the seepage tube method described by Luthin and Kirkham (1949) for horizontal hydraulic conductivity, and the unlined auger hole method for vertical conductivity. Dai and Sparling (1973) used a piezometer method to determine hydraulic conductivity on a spaghnum peat and sedge peat. The sedge peat showed greater stratification, and had hydraulic conductivities significantly greater in the horizontal, as opposed to the vertical direction. This arises through the preferential orientation of the constituent materials during peat accumulation (Vos, 1982).

The degree of peat decomposition helps determine the hydrological characteristics. Undecomposed peats have larger pores which enable faster rates of water movement, and various correlations can be established between water retention and bulk density and fibre content (Boelter, 1969; 1975). Boelter (1965) compared laboratory and field methods to determine hydraulic conductivity on various peat deposits: a sphagnum moss peat, decomposed moss peat with wood, decomposed peat, and herbaceous peat. The hydraulic conductivities obtained from laboratory measurements were found to be an order of magnitude greater than field measurements on the same peat type, which Boelter attributed to water flow at the core perimeter.

One problem in the measurement and comparison of peat conductivities is that the characteristics of peat vary considerably with depth, and conductivity is itself inversely proportional to bulk density. The results of Chason and Siegel (1986), on peat from a raised bog in Minnesota, apparently contradict this, as they found no correlation of hydraulic conductivity with depth. However, in this case the peat profile had a consistent structure, and they confined themselves to sampling deep deposits. Korpijaakko and Radforth (1972) observed that humification and dry density had the closest correlation to hydraulic conductivity. Working on a sphagnum peat from a raised bog in New Brunswick, they attributed a decreasing effect of depth on hydraulic conductivity to an increasing humic acid content.

Several studies have considered that water flow through peat does not follow Darcy's Law. Ingram *et al.* (1974) found that the hydraulic conductivities of humified peat, obtained from seepage tube tests, were time dependent and increased with hydraulic head, suggesting a departure from Darcy's Law. However, Hemond and Goldman (1985) concluded that Darcy's Law remained valid as any non-Darcian behaviour could be explained due to the elastic properties of peat. Possible departures from Darcy's Law may arise where peat has a constant structure, but water discharge varies with hydraulic gradient. Alternatively there may be some variation in peat structure with hydraulic head due to changes in pore geometry. The latter processes may arise as fluid flow produces a frictional drag on particles with consequent re-orientation and blocking of pore spaces. The implication is that steady state formulae describing water flow are inappropriate due to the elastic storativity of the peat, whereby water is taken up or lost through expansion or contraction of the peat matrix. Brown and Ingram (1988) undertook a series of seepage tube tests on well-humified peat and obtained repeatable results which appeared to indicate reversible pressure-dependent storage effect within the peat.

In conclusion, there are limitations to the use of Darcy's Law for peat deposits, however, Rycroft *et al.* (1975) and Hemond and Goldman (1985) agree that flow through the acrotelm, or the upper layer of unhumified peat, follows

Darcian principles. Hemond and Goldman argue further that the Law may still be applied in humified peats, under conditions of a small hydraulic gradient and constant effective stress.

3.3. EVAPOTRANSPIRATION.

The characteristics of evapotranspiration from wetland systems have already been considered in Section 2.4.3.ii. In this section, possible methods to estimate evapotranspiration are summarised and examples given of wetland studies which adopted different approaches to determine quantities of evapotranspiration. The Penman equation, which was chosen for use in this study, is also discussed.

The quantity of water lost through evapotranspiration from a floodplain wetland is hard to determine accurately. Direct measurements of evapotranspiration loss are possible but are time consuming and expensive requiring either a detailed micro-meteorological approach, or the use of vegetation chambers (eg. Farrington *et al.* 1990). Consequently, where only limited data are available an approximate indication of evaporation has frequently been obtained through theoretical calculations, using evaporation pans, or as the residual term in the water balance equation.

Evaporation occurs when the air vapour pressure lies below the saturated vapour pressure. The latter is temperature dependent but evaporation also requires an energy input, provided by solar radiation, to supply the latent heat of vapourisation, equal to 2.47×10^6 J/kg at 10°C. The simplest case of evaporation is from an open water body (E_o), whereas under a vegetation cover the quantity of water lost through transpiration (E_t) also becomes important. Transpiration is the water loss through stomata in the leaves or lenticels in bark, and occurs when the vapour pressure in the leaf is greater than atmospheric vapour pressure. The amount of transpiration will vary with vegetation type and soil water conditions, and may also include some evaporation from intercepted precipitation. Evaporation is largely determined by climatic conditions and in particular the moisture characteristics of a prevailing air mass. This has facilitated the application of large scale empirical evaporation models based on commonly monitored meteorological parameters, as exemplified by the Thornthwaite (1948) model which estimates evapotranspiration using monthly mean temperatures.

Evaporation is directly proportional to net radiation, the balance of incident and reflected shortwave and longwave radiation. However, also important is the moisture carrying capacity of the lower layers of the atmosphere. The saturation deficit of an air body is the difference between actual vapour pressure and the saturation vapour pressure, and determines the quantity of water vapour that can be taken up. The capability of the air to absorb water is therefore temperature dependent and increases with temperature. Where the lower atmosphere is at saturated vapour pressure, evaporation will cease unless processes of turbulent mixing of air brings drier air into contact with the evaporation surface. Evaporation is thus also proportional to wind speed which encourages more complete mixing of air and is inversely proportional to surface roughness. Where surface friction is substantial a boundary layer of air may be formed that limits the dispersal of humid air pockets.

These are the main meteorological factors governing evaporation. The precise amount of evapotranspiration loss is limited by water availability and in this respect the capability of plants to extract water is important. A physiological limit is placed on evaporation by the individual plant stomata, through which water is lost through transpiration, and which close under conditions of water stress thus reducing evaporation.

Potential evapotranspiration is the combined water loss from evaporation and transpiration under conditions of maximum water availability. Boundary effects due to surface roughness are minimised by restrictive study of evaporation from a short green crop, free of moisture stress. More recent modifications to the term include the wet-surface evaporation which is likely to be the maximum possible evaporation as the vegetation cover is wet and water deficiency is likely to be insignificant. It thus represents the maximum potential loss of water under given meteorological conditions. In agricultural areas it is unlikely that the soil surface will be entirely covered by vegetation and the significance of evaporation loss from bare soil needs to be considered. Milly (1984) has related evaporation from the soil to temperature distribution through the soil profile, which produces a corresponding variation in the coefficients of water transport. Actual evaporation will depend on the upward moisture flux from the water-table; this may be considerable where the capillary fringe intersects the ground surface. Keen (1927) noted that the capillary fringe was only important when the water-table was within 35 cm of the surface in coarse sand, 70 cm in fine sand, and 85 cm in heavy loam soil. Thus in floodplain wetland areas actual evaporation levels may approach the maximum theoretically possible due to the high level of the water-table.

Evaporation may be measured by a variety of methods. For instrumented drainage basins the water loss through evaporation can be determined using a water budget approach. Here, evapotranspiration represents the difference between net inflow (I) and outflow (O) of water, with changes in internal water storage (ΔS) also accounted for. The open water evaporation is given by:

$$E = I - O - \Delta S \tag{17}$$

Evaporation can also be measured crudely by noting the change in water level of pans of water, various standards of which have been developed and which are described in meteorological manuals.

Evapotranspiration may also be measured on test plots using a lysimeter which involves the isolation of a parcel of land and determination of changes in weight of a soil body; however the possibility of short term process changes makes this approach only suitable for longer term studies. Other approaches include lysimeters which consist of hydrologically isolated tanks of soil with vegetation covers. By monitoring precipitation and changes in drum weight an estimate can be made of transpiration loss. Dooge (1975) discusses studies which considered the relationship of evapotranspiration figures as measured by a lysimeter to calculations based on meteorological data, and reports a good correlation for wetland areas (eg. Bay; 1966).

An alternative method was adopted by Farrington *et al.* (1990) who used ventilated chambers to monitor evapotranspiration on a groundwater mound in western Australia. Estimates of evapotranspiration obtained from the ventilated chamber method showed a close correspondence with the amount suggested by monitoring of diurnal water-table fluctuations, and amounted to over 800 mm in the year 1987-1988.

3.3.1. Hydrometeorology.

Two particular methods have been used for evapotranspiration calculations using detailed meteorological data. The mass transfer method and the vapour flux approach both examine the upward flux of air from an evaporating surface, while the energy balance method considers the radiation balance and the energy available for evaporation processes.

i.a. Mass transfer method.

The calculation of evapotranspiration by mass transfer methods uses a bulk aerodynamic equation. This theoretically comprises the factors responsible for removing water vapour from the evaporating surface, and has been termed the Dalton evapotranspiration formula:

$$E = f(u)\ (e_s - e_d) \tag{18}$$

where f(u) is a function of wind speed, e_s is the saturation vapour pressure of water at mean surface temperature, e_d is the actual vapour pressure in the air so that (e_s - e_d) is the saturation deficit. The component f(u) is composed of empirical coefficients developed for particular meteorological conditions.

Eisenlohr (1966) used a formula of this type to determine evaporation over an annual cycle from a water body containing hydrophytes. It was found that the function f(u) fluctuated during the growing season, in response to vegetation growth.

i.b. Vapour flux method.

Evapotranspiration calculations using the vapour flux method assume, as with the mass transfer method, that quantities of evapotranspiration are proportional to removal of moisture laden air (Shaw, 1988). However, here measurements are taken at two points above the evaporating surface to determine the flux of moisture laden air using an equation of the form:

$$E_o = \frac{0.623 \mathrm{K}^2 \rho (u_2 - u_1)(e_1 - e_2)}{p(\ln z_2 / z_1)^2} \tag{19}$$

where e_1 and e_2 are vapour pressures at heights z_1 and z_2, u_1 and u_2 are the corresponding wind speeds at the two heights, p is atmospheric pressure, ρ is air density and K is von Karman's constant (equal to 0.41).

ii. Energy Budget method.

The energy available for evaporation (Q_{Eo}) is given by:

$$Q_{Eo} = Q_s - Q_{rs} - Q_l - Q_c \pm Q_g \pm Q_v \tag{20}$$

where Q_s is short-wave solar radiation, Q_{rs} is reflected short-wave radiation, Q_l is long-wave radiation from the water body, Q_c is the sensible heat transfer to the air, Q_g is the change in stored energy and Q_v is the energy transfer between water and bed.

For terrestrial studies the energy budget is rewritten as follows:

$$R_n = H + L \cdot E_a + G \qquad (21)$$

where R_n is net incoming radiation, L is latent heat of vaporisation, H is convection of sensible heat and E_a is evaporation of water. G, the conduction of heat into the soil, is often disregarded as it is small and much is lost by re-radiation at night, however, van den Berg (1989) demonstrated that measurements should be included in certain conditions, particularly where the moisture content of the soil varies significantly.

Actual evapotranspiration (E_t) is then calculated using the Bowen ratio method:

$$E_a = \frac{(R_n - G)}{L(\beta + 1)} \; (mm\,day^{-1}) \qquad (22)$$

Here, β is the Bowen ratio which is the ratio of sensible and latent heat. The ratio is determined by measuring the differences in air temperature and water vapour pressure over a given vertical distance:

$$\beta = \gamma \cdot \frac{k_h}{k_m} \cdot \frac{\Delta T}{\Delta e} \qquad (23)$$

where γ is the psychrometer constant, k_h and k_m are the eddy diffusivities of sensible heat and water vapour, and ΔT and Δe are respectively the vertical differences in temperature and vapour pressure.

Linacre (1976) considers that the Bowen ratio method provides a more accurate estimate of evapotranspiration, as measurements are made over natural vegetation at two heights. The method was successfully used by Priban and Ondok (1986) to estimate evapotranspiration from a willow carr wetland. Smid (1975) also used the Bowen ratio method to determine evapotranspiration from a reed swamp. In this case net radiation was not measured directly but was estimated from the albedo, short-wave radiation, and long-wave radiation.

iii. Combination of the energy budget and aerodynamic approach.

The widely used Penman (1948) method for calculating evapotranspiration uses four meteorological parameters: duration of sunshine; mean air temperature; mean air humidity and mean wind speed. Several variants of the original method have been developed to accommodate different combinations of input data, however, the technique recognises that evapotranspiration is driven by radiation. The limit placed upon evapotranspiration loss is provided by the rate at which moisture-laden air is removed from the evaporating surface.

In its earliest form, the Penman method required three particular inputs: firstly a measure of solar radiation (combining sunshine duration and mean air temperature); secondly the saturation deficit of the air at screen level, air temperature and vapour pressure (mean air humidity); and thirdly the run of wind at 2m above the ground (mean wind speed).

Evapotranspiration is then calculated as a combined function of radiation, relative humidity and wind speed:

$$Evp = f(radiation) + f(relative\ humidity) + f(wind\ speed) \tag{24}$$

The full equation to determine evapotranspiration is of the form:

$$f(radiation) = \frac{\Delta}{\Delta - \gamma}\left[R_a(1-r)\ (a+b)\frac{n}{N}\right] \tag{25}$$

$$f(humidity) = -\frac{\Delta}{\Delta + \gamma}\left[\sigma\, T_a^4 (0.56 - 0.092\sqrt{e_d})(0.10 + 0.90\frac{n}{N})\right] \tag{26}$$

$$f(wind\ speed) = \frac{\gamma}{\Delta + \gamma}\left[0.35(1 + \frac{u}{100}(e_a - e_d)\right] \tag{27}$$

where:

a and b are regression terms for the number of hours of sunshine for the study region.

Δ = slope of the saturation vapour pressure curve at mean air temperature.
γ = constant of the wet and dry bulb psychrometer equation.
R_a = theoretical incoming short wave radiation.
r = reflection coefficient (albedo).
n = hours of sunshine.
N = theoretical duration of sunshine.
σT^4_a = black body radiation at mean air temperature.
e_d = mean vapour pressure.
e_a = saturation vapour pressure at mean air temperature.
u = run of wind at 2m.

The Penman equation has been widely used in the United Kingdom since 1948. The equation has been used as the basis for calculations of groundwater recharge, although the time period adopted has variously been ten day, or monthly means (Howard and Lloyd, 1979). The method assumes efficient transfer of heat and water vapour, transfer of sensible heat into the ground is neglected, and the equation is unsuitable for estimates of evapotranspiration over short time periods.

Various modifications have been made to the Penman equation to improve its capabilities to estimate evapotranspiration under different conditions. Among these the application of the Priestley and Taylor's (1972) simplification of the Penman formula has been examined within a wetland context. Priestley and Taylor (1972) eliminated wind speed and the humidity deficit from the equation, and considered that the average evaporation rate from a vegetated surface was a function of Δ, the slope of the saturation vapour pressure curve, and γ, the constant of the wet and dry bulb psychrometer equation:

$$Evp \propto 1.26 \frac{\Delta}{\Delta + \gamma} \qquad \textbf{(28)}$$

Stagnitti *et al.* (1989) assessed the accuracy of a further modification of the Penman equation, the Penman-Monteith model, which incorporates the turbulent transfer component and available energy. They investigated a small forested wetland of 2.3 hectares in the eastern United States, and compared evapotranspiration estimates obtained from the Penman-Monteith formula with estimates from a water budget study. Although the results were considered satisfactory, it was concluded that the formula requires calibration for different application which represents a major limitation.

3.4. CONCLUSION.

This chapter has included detailed descriptions of the hydrological processes which have been observed within wetland systems. A physical understanding of the processes is required before proceeding to describe the detailed hydrology of Narborough Bog. Principal areas considered in the chapter include an introduction to the mathematics used to describe the processes of subsurface water flow between points; an introduction to the specific problems raised when looking at water flow through peat deposits; and an examination of the variety of techniques that have been used to determine evapotranspiration loss from wetland areas. The chapter indicates the variety of methods that have been applied on studies of wetland hydrology, which need to be understood before undertaking a comparative analysis of results from different studies of wetland hydrology later in the thesis. In particular, the methods introduced in this chapter will help address the major question of how typical the field-site at Narborough Bog is of other floodplain wetlands. Consequently, before considering hydrological processes in further detail, the field-site are described in the next chapter.

Chapter 4

Field Area and Measurement Programme

Scope of Chapter

The last two chapters have provided the theoretical background to this study of the hydrology of the floodplain wetland at Narborough Bog. This chapter describes the geomorphology of the field-site and surrounding floodplain and then introduces the measurement programme. The geomorphology of the area is described, and the formation and preservation of the wetland at Narborough is placed within the context of the sedimentology and geomorphology of the surrounding Soar floodplain. The geology of the surrounding Soar floodplain is then described prior to summarising the ecology of the site, and discussing the work undertaken at Narborough Bog by the Nature Trust, who own the site. Finally in this section, the rationale governing the selection of the field-site at Narborough is considered.

The latter part of the chapter describes the field measurement programme. The objectives of the field-work are outlined and the installation of apparatus to monitor variations in saturated and unsaturated water contents are described.

4.1. INTRODUCTION.

Narborough Bog is 9.5 hectares of mixed woodland, reed-bed and meadow which lies on the floodplain of the river Soar 7 km to the south west of Leicester (map reference SP 549 979) at altitude 63 m a.s.l. The site is a nature reserve which has been owned by the Leicestershire and Rutland Trust for Nature Conservation since 1975 and has been designated a Site of Special Scientific Interest since 1956. The official justification for the original citation included the size of the "natural" reed-bed at Narborough, which is of local significance and forms one of the largest reed-beds in the county, and the surrounding areas of wet woodland and meadow with their characteristic flora and fauna.

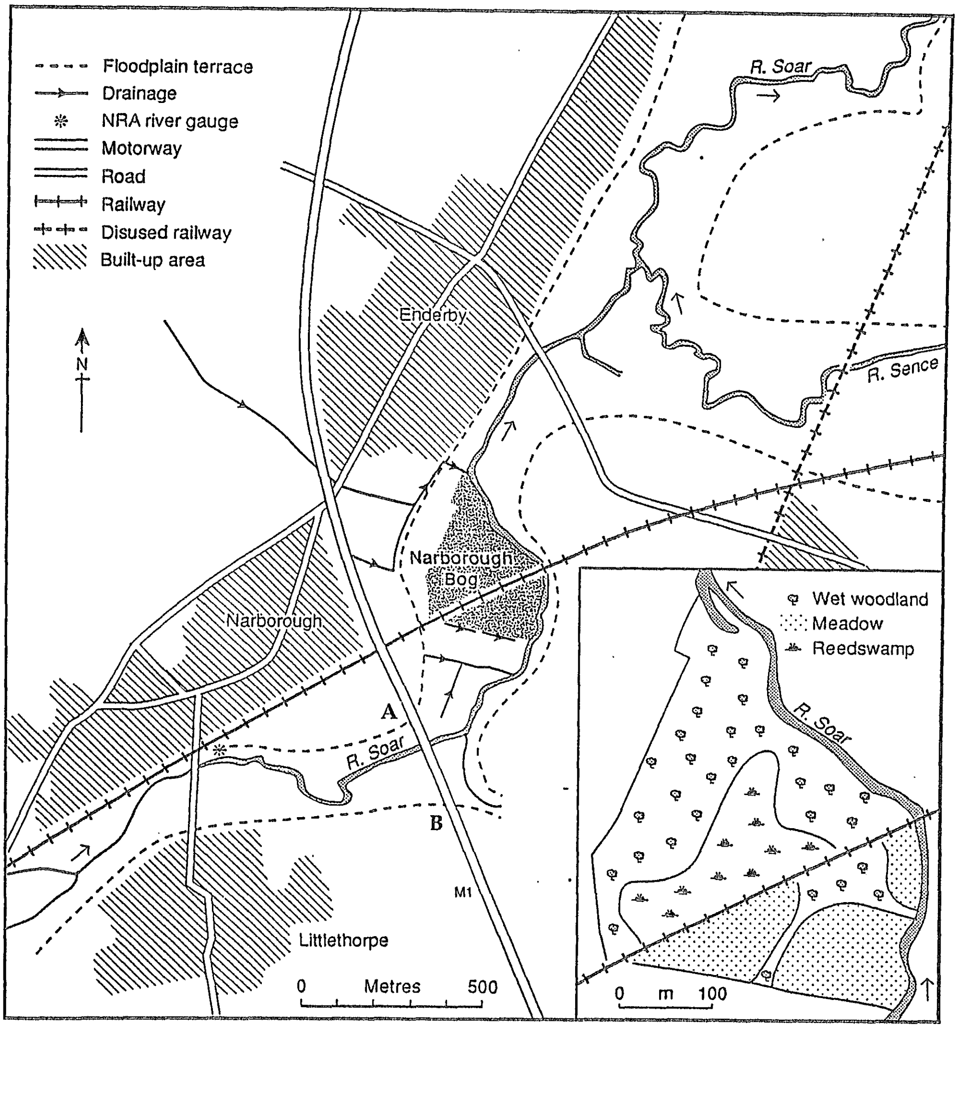
An aerial view of Narborough Bog, looking East (taken in May 1992) is shown in Plate 4.1. and the location of the wetland at Narborough Bog in relation to the Soar floodplain and adjacent villages is shown in Figure 4.1. In the picture Narborough Bog lies to the west of the river Soar, which meanders across the photograph from right to left. The proximity of a variety of land-uses to the reserve is evident; these include allotments, a sports field and agricultural land, in addition to several villages, which include Whetstone and Narborough at the top and bottom of the photograph respectively. A railway line and embankment, which were constructed in 1864, cross the site. To the West of Narborough Bog, the M1 motorway comes to within 100 m of the Reserve boundary.

Within the small area covered by the Narborough Bog reserve, several distinctive ecological zones can be identified. These are indicated in the inset figure in Figure 4.1. and can also be seen in Plate 4.1. The majority of the reserve consists of an alder-willow woodland, which extends along the margins of the river Soar and the boundary with the sports field, to the west. In the centre of the reserve is an area of former reedswamp which lies beside the railway embankment. The southern area of the reserve, below the railway line, is composed of three small meadows which are separated by small areas of woodland. These meadows are all that remain of a considerable number of small fields which were formerly characteristic of this area of the Soar floodplain.

The degree of variety in the wetland flora of Narborough within these different habitats provides the principal argument for the continued conservation of Narborough Bog. Several studies have documented the flora of the site; Wade (1919) described the floras of the Narborough and Aylestone Bogs whilst the flora of the wet woodland at Narborough was discussed within

Plate 4.1. Aerial photograph of Narborough Bog, viewed from the West, on 17th May 1992.

Figure 4.1. Location map showing the position of Narborough Bog with respect to the Soar Floodplain. The letters **A** and **B** refer to the cross-section in Fig. 4.2.



books on the flora of Leicestershire, by Horwood and Noel (1933) and Primavesi and Evans (1988). A substantial range has also been observed in the fauna of Narborough, and particularly significant is the local population of *Lepidoptera*.

4.1.1. Sedimentology and Geomorphology of the Soar floodplain.

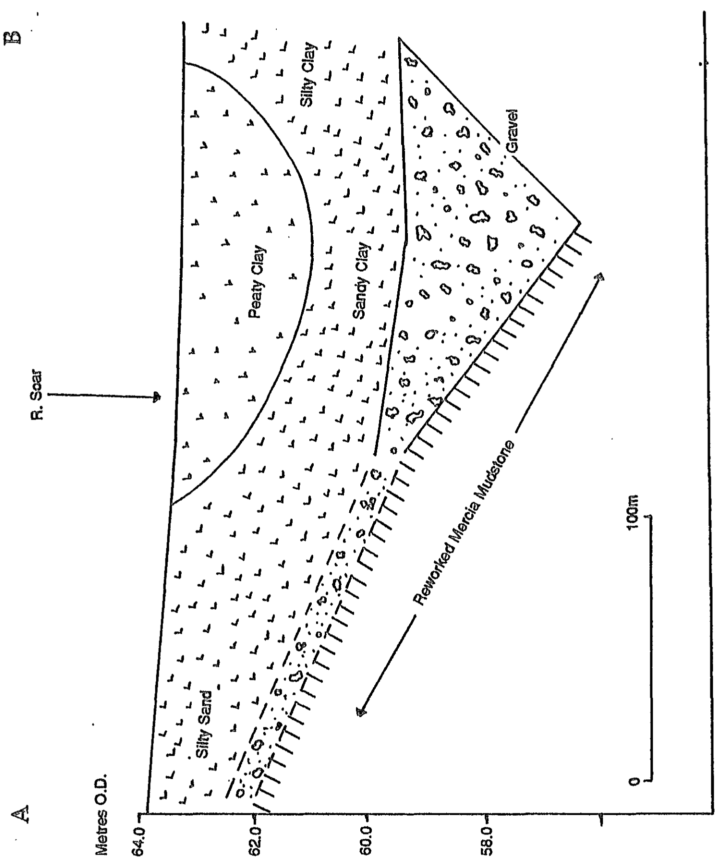
The evolution of the Soar floodplain near Narborough Bog, and the combined sequences of local fluvial aggradation and degradation, have had an important effect upon providing the hydrogeological environment required for the development of the wetland site. The location of different alluvial deposits, and in particular any gravel deposits, are important when discussing the significance of subsurface water fluxes which are considered later in the thesis.

The depositional history of the Soar floodplain near Narborough reflects a complex history of development through the Pleistocene, which has been described by Rice (1981). The interpretation given here is open to re-examination as the status of the Wolstonian glaciation has been questioned (Rose, 1987). The modern drainage net of the river Soar is closely related to the flow of the Proto-Soar, which developed in the period following withdrawal of Wolstonian ice, c. 150,000 years BP. At this time the river began to incise into glacial deposits from the Wolstonian glaciation and through to the underlying local bedrock, the Mercia Mudstone. This process of downcutting by the Soar produced river terraces which contain gravels of the Oadby and Thrussington units, which are interbedded in places with sand and gravel that accumulated during the Wolstonian ice advance. Among the latter deposits are the Bagington-Lillington sands and gravels which were deposited along the line of the Proto-Soar. Following the end of the interglacial conditions, colder conditions in the Devensian c. 100,000 years BP brought a period of alternating incision and degradation, and at this time coarse gravels were deposited under a braided flow regime. During the Devensian, the elevation of the river channel was below its present elevation, indicating that there have been more recent phases of fluvial aggradation. This is discussed later when considering the near-surface stratigraphy of Narborough Bog.

The stratigraphy of the Soar floodplain near Narborough Bog has been further complicated by the identification of a glacial furrow in the bedrock directly underlying the site. Rice (1981) identified three possible glacial furrows in this area from bore-hole records. The furrows consist of linear erosional features in the Mercia Mudstone which have been infilled by a mixture of glacial and water-born deposits. Rice calls the main feature the Narborough furrow, which is approximately 200 m wide and 25 m deep and closely follows the current line of the Soar floodplain from the south west to the north east of Figure 4.1. The possible significance of the Narborough furrow for the hydrogeology of the site is discussed in more detail later, however, it would seem that from a hydrological viewpoint, the deposits which infill the furrow have similar characteristics to the local bedrock, as they are typically of low permeability.

The depth of the Devensian gravel deposits underlying Narborough Bog has particular significance for the groundwater hydrology of the site as they represent deposits which potentially have a significantly higher permeability than the overlying organic deposits. Although their depth could not be determined directly using a hand auger, an indication of the depth and extent of the deposits is provided from boreholes drilled in association with the motorway construction in 1965. These data have been used to plot a cross-sectional stratigraphy of the Soar floodplain which is given in Figure 4.2. The line crosses the river Soar approximately 400 m upstream of Narborough Bog, close to the point where the Whetstone Brook flows into the river Soar. Only one borehole was sufficiently deep to reach the Mercia Mudstone bedrock, which was recorded at a depth of 24 m, however, the shallow records used here provide an indication of the relative deposition of alluvial deposits in this area. Deposits of clayey peat overlie variable depths of gravel, on a base of reworked Mercia Mudstone, which is apparently associated with the glacial furrow mentioned above. At this point the river Soar is flowing along the western

Figure 4.2. Stratigraphy across the Soar floodplain along a transect **A B** where the motorway crosses the river as shown in Figure 4.1. The variable thickness of gravel deposits across the floodplain is apparent.



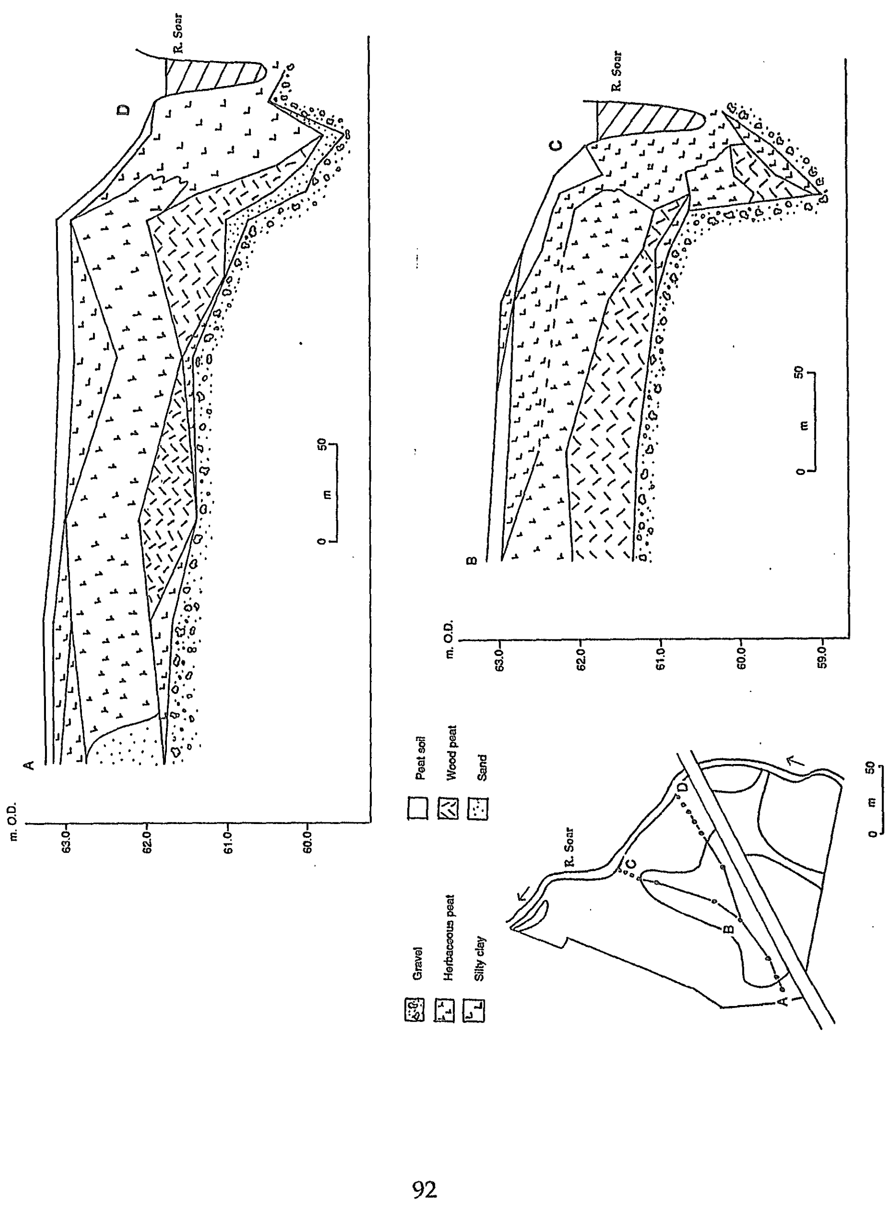
margin of the floodplain, and the gravel deposits are relatively thin, at c. 0.8m; however the gravel deposits thicken towards the centre of the floodplain, reaching a maximum depth of 3m. Additional indications of the nature of the gravel deposits are provided by a survey of sand and gravel resources in the Narborough area undertaken by Engineering Geology (1983). The report discusses results obtained by supplementing existing borehole data with resistivity surveys and additional boreholes in areas of poor coverage. It was concluded that the maximum thickness of the Devensian gravels was between 2 and 3 metres, decreasing slowly towards the edge of the floodplain.

4.1.2. Geomorphology of Narborough Bog.

The development of the wetland at Narborough Bog is discussed here using records of stratigraphy obtained by collecting cores using a Russian or Dutch hand auger along two transects identified in Figure 4.3. These transects extend for 340 m from the river Soar, through approximately 70 m of alder woodland and for the whole length of the reed-bed to the edge of the Nature Reserve. A representation of the stratigraphy along the two transects is given in Figure 4.3. Over the area sampled, all deposits were found to overlie gravels of Devensian age, with the exception of the core at the western end where an impenetrable thick red clay was encountered. The depth from the surface to the gravel was typically 1.8 m, as the gravel surface closely follows surface topography. The only exception was within a distance of 30-50 m, from the current position of the river Soar where the depth to the gravels increased by 0.9m, before rising under the current river channel. This most probably indicates the location of either a palaeochannel, or buried valley. Under the current channel, gravel deposits lie within 40-50 cm of the river-bed, and are separated by a silt-clay layer.

Above the gravel base there is a wide band of dark brown wood peat, with in some cases, near the river, an intervening layer of grey silt clay. Frequent *Alnus* twigs were found within the wood peat, the thickness of which increases with distance from the river to a maximum of 0.8m. The wood peat

Figure 4.3. Stratigraphy of near surface deposits at Narborough Bog along three transects.



is almost certainly a product of deposition within a wetland environment in the early to mid Holocene. Pollen analysis of the basal wood peat by Dr A.G. Brown (pers. com) has also revealed early Boreal spectra dominated by *Pinus* and *Betula*. In the upper area of the reed-bed an herbaceous phragmites peat, c. 1m thick, overlies the brown wood peat, which is approximately 0.7m thick at this point. The wood peat is absent from one point in the stratigraphy which corresponds to dipwell **5n**, however, here it is possible that the stratigraphy is complicated by the flow of a small stream at some point. Although speculative, this would explain the 0.4m thick grey brown clay near the top of this particular core, which could have been deposited during local flooding around the stream margins.

The deposits which lie above the wood peat consist of a substantial depth of herbaceous peat within the reed-bed which reaches a maximum depth of 0.8m. Within 110-120m of the river the silt/clay concentration of the herbaceous peat increases and in places is overlain by a grey brown silty clay. This represents sedimentation during various episodes of overbank flooding, and which remains significant during current flood events. The grey silty clay deposits adjacent to the river are more extensive in the northern transect. This is probably a result of the lower surface elevation in this area in comparison with the southern transect, with consequently greater depths of overbank flooding permitting greater sedimentation of fine deposits.

Over much of the area covered by the two transects, a brown peat soil 20-30 cm in depth was found, with a depth increasing locally to 45cm under the woodland. This is an important indicator of current hydrological conditions at the site, suggesting oxidation of the surface peat deposits under conditions of lower moisture content.

The surface topography at Narborough Bog maintains a gentle gradient towards the river Soar. This is illustrated in Figure 4.4. which shows a three-dimensional plot of a rectangular area extending into the reed-bed from the

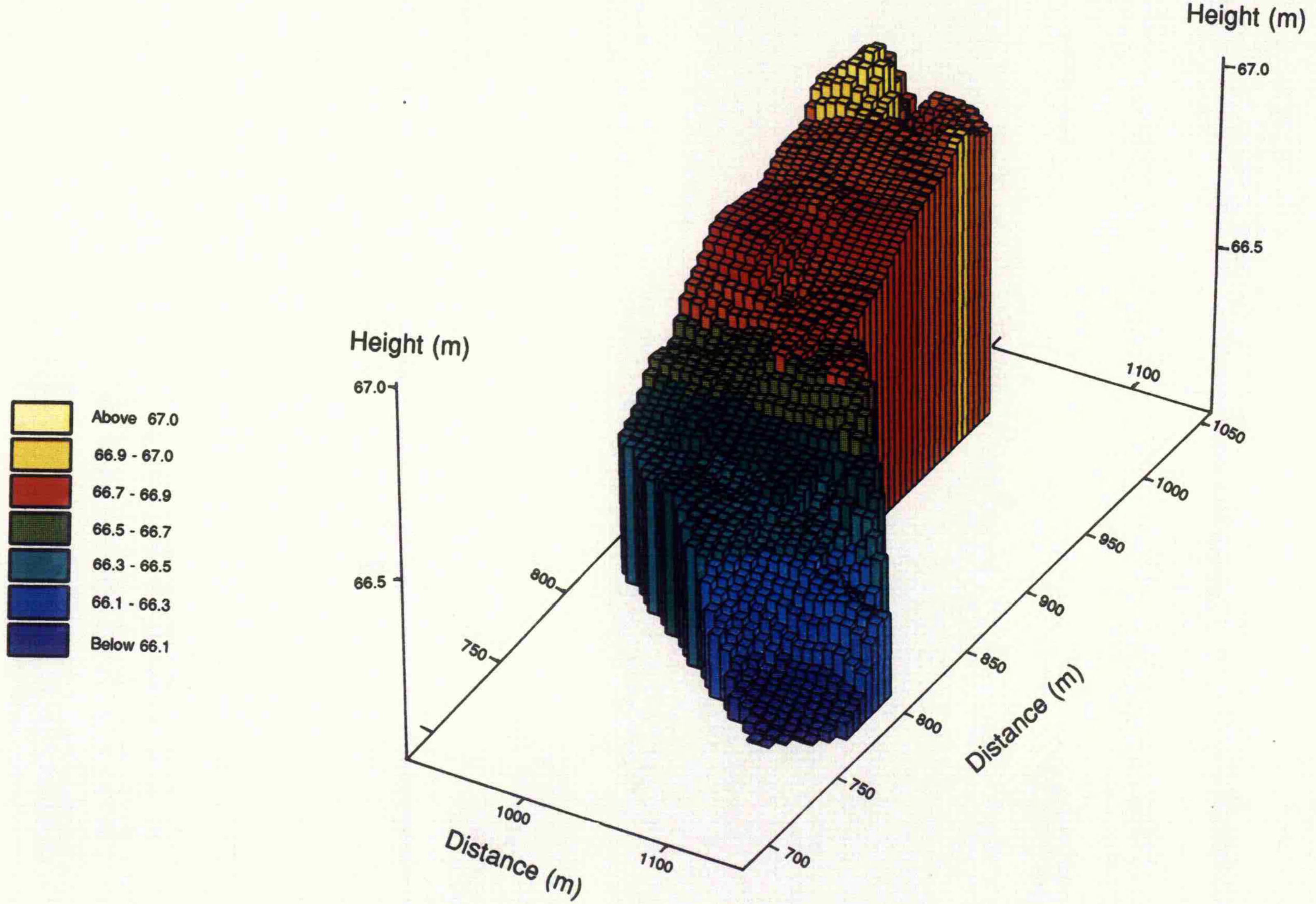
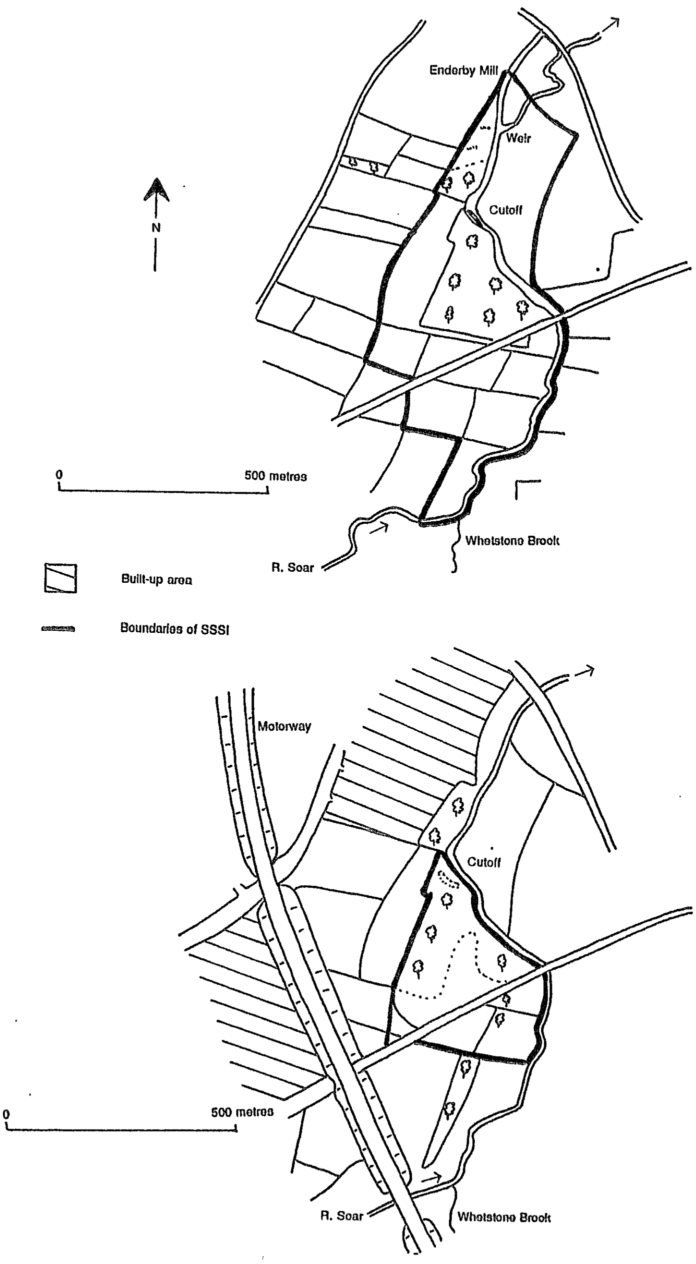


Figure 4.4. Topographic plot of the reed-bed at Narborough, illustrating the decrease in surface elevation near the river (in foreground).

Figure 4.5. The Soar floodplain near Narborough redrawn from Ordnance Survey maps surveyed in 1885 (top) and 1966.



river. The diagram was obtained by a process of bilinear interpolation from transects of 482 individual survey points using the UNIMAP software package. The area to the southwest of the reed-bed rises sharply as a result of the dumping of a variety of waste material, principally crockery, during construction of a nearby hospital in 1904, to provide a causeway over the railway embankment. Within the reed-bed, the surface slopes at an angle of between 0.05° and 0.1°, but increases to c. 2° along the river margins and within the area of a fringing woodland. A network of 'rigg and furrow' extends through the lower area of the reed-bed and woodland which possibly once formed part of a water management scheme to control the frequency of river water inundation on the reed-bed, probably as an osier bed as described below.

Another feature of interest is a small river cut-off which can be identified in the northern section of the reserve. This is currently isolated from the river, and is only identifiable by a slight surface depression, although standing water is frequently held within the depression in periods during or following high river stage. The development of the cutoff is revealed by examination of the series of Ordnance Survey maps covering the area. Two illustrations, which are redrawn from maps surveyed in 1884-1885 and 1966, are given in Figure 4.5. The information shown here is discussed in detail later, however, in the early maps the main channel of the river Soar divides in the northern corner of Narborough Bog and flows around an island in the channel. In a map showing revisions from 1928, the feature is shown as a river cutoff, while in the most recent map, surveyed in 1966, the cutoff is outlined and is isolated from the channel.

The preservation of the herbaceous and wood peat deposits at Narborough shown in Figure 4.3. indicates a lack of meander migration in this area of the Soar floodplain, with a prolonged period of bank stability, with the exception of the river cutoff discussed above. This contrasts with the behaviour of the river Soar downstream of Narborough Bog, where several meander cutoffs can be identified in the field. However, local stability of the river is also suggested by the mature riparian vegetation at Narborough Bog.

Generally the deposits at Narborough Bog reveal a sequence of recent deposition from overbank floods of fine silts and clay in the area adjacent to the river Soar, with a simpler accumulation of fen and wood peat within the reed-bed. Peat accumulation requires high water-tables, which may have been maintained at Narborough through a combination of river flooding, and also aided by the presence of a small stream flowing into the river Soar at this point. The silt and clay deposits adjacent to the river provide a hydraulically resistant barrier behind which stable hydrological conditions are maintained and seasonal water-table variations controlled. The deposits also confine the river, which in consequence would have insufficient power for lateral erosion, thereby preserving the peat deposits.

The surface slope at Narborough indicates gradual sedimentary deposition through settling out of fine sediments from overbank flooding. In addition, the limited extent of overbank flooding is demonstrated by the width of the clay deposits near the river. The initial development of the wetland at Narborough Bog almost certainly occurred within a backswamp environment. High water-tables were maintained at the site, as the return flow of flood waters was prevented by levees along the river margin. The flow of a small stream through the wetland would have also provided significant further quantities of water. There is evidence of a palaeochannel 50-70 m west of the current channel, infilled with wood peat at the base, which most probably dates from the early Boreal.

The recent history of Narborough Bog includes some local management of water levels which is indicated by the rigg and furrow topography, but observations of changing wetland ecology apparently suggest that the relationship of the wetland with the river has changed through continued peat accumulation so that the site is now more dependent upon precipitation. The extent to which this has occurred is considered later in the thesis when the fieldwork results are examined.

4.1.3. Hydrology of the River Soar and Narborough Bog.

The characteristics of the wetland at Narborough Bog need to be placed within the context of the hydrology of the river Soar, its annual regime, and the local climate. Consequently in this section, the hydrology of the river Soar is discussed and features affecting the surface hydrology of the area surrounding Narborough Bog are described. In addition some meteorological records collected locally are summarised.

The river Soar has a total length of 64 km, and drains an area of 1165 km^2, of which approximately 190 km^2 lies upstream of Narborough Bog. The Soar rises near the village of Sharnford on the Leicestershire - Warwickshire border, and flows through Leicester and Loughborough before joining the river Trent at Sawley. Significant tributary rivers include the river Sence which flows into the Soar near Blaby, downstream of Narborough Bog, and the river Wreake which joins the river Soar at Cossington. In Figure 4.6.A discharge records from a gauge maintained on the river Soar by the National Rivers Authority at Littlethorpe are summarised. The data cover the period from 1983 to 1990, and consist of mean monthly discharges, with bars indicating the extreme maximum and minimum daily discharges over the period. Mean monthly discharges are greatest in January and December at 2.7 and 2.2 cumecs respectively, and fall to a minimum in July and August of 0.5 cumecs. There is less variability in the minimum monthly discharge; 0.6 cumecs was recorded in January and February, and 0.2 cumecs in July and August. The maximum discharges demonstrate clearly the annual variation in flows. During the period 1983-1990 the greatest discharge was recorded in January (17.7 cumecs), monthly maximum discharges then fell within the range 10.9-12.1 cumecs until June, after which they fell to under 5 cumecs. Maximum discharges then increased from October to December, from 8.8 cumecs to 14.5 cumecs.

Measurements of precipitation and temperature were collected at a local weather station at Braunstone, 3 km North of Narborough, over the period 1959-1978. Two graphs showing mean monthly precipitation and temperature are given in Figure 4.6.B and 4.6.C. Precipitation is evenly distributed throughout the year. Mean annual precipitation over the period was 643mm, with a minimum monthly mean precipitation of 40mm in October, and a maximum mean precipitation of 64mm in August. In the same period, the mean annual temperature at Braunstone was 9.4°C, with a maximum mean temperature of 16.0°C in July and August, and a minimum mean temperature of 3.3°C in January.

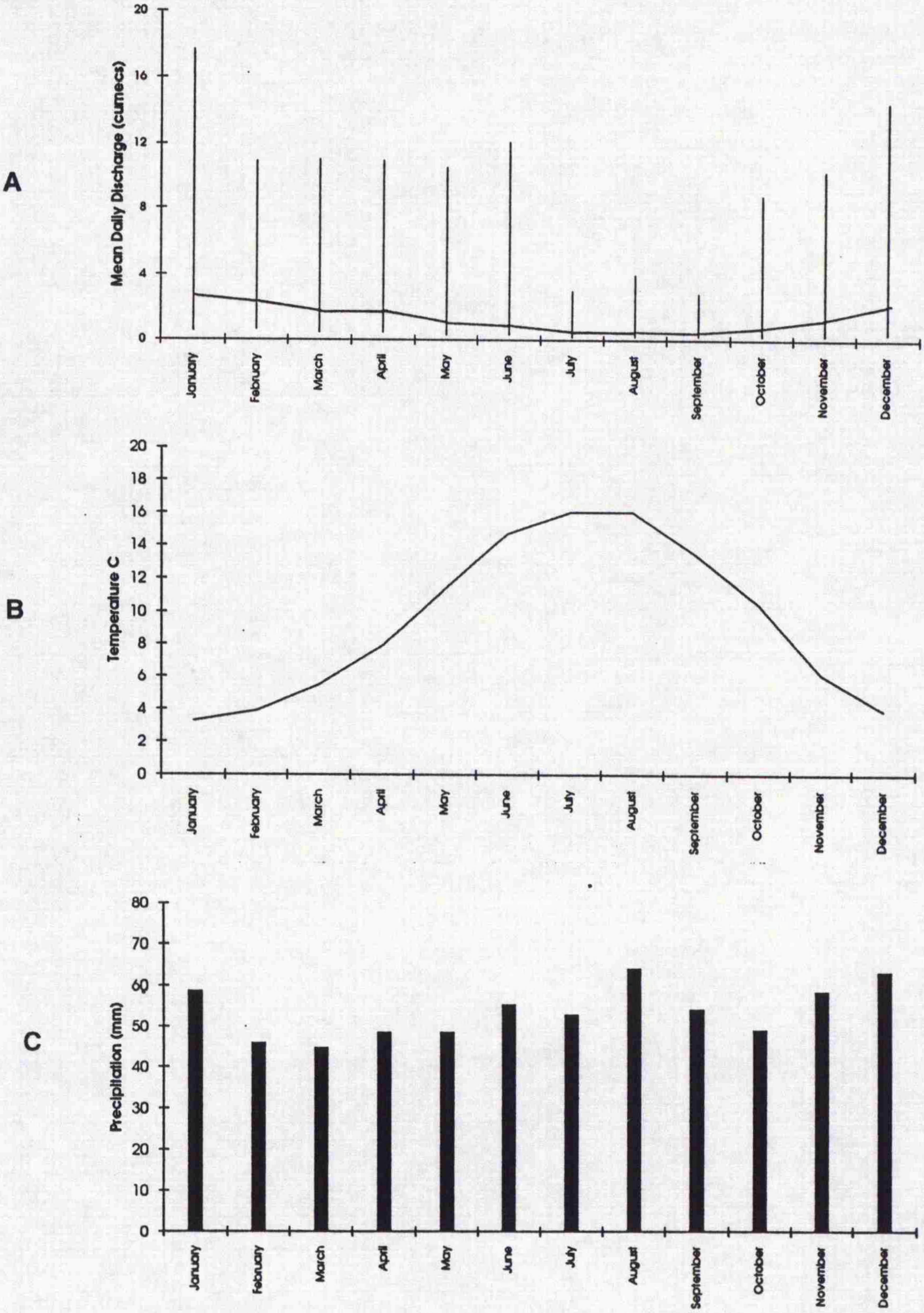
Some of the data described above is considered further in the following chapters. In chapter 6 (section 6.1) the frequency of overbank discharges is examined, while in Appendix IV the recent precipitation record is used to identify the timing and frequency of periods of drought. There have been several documented examples of work within the catchment which are likely to have affected the flow of the Soar locally, influencing both river levels and also the magnitude, duration, and pattern of overbank flooding. In the mid-1960s a weir across the Soar, downstream of Narborough Bog at the site of Enderby Mill, was removed. The weir was originally used to maintain the head of water across the Mill, and directed water into a mill race. Removal of the weir produced an approximate decrease in the stage of the river at low flow conditions of 1.5 metres (Baker, unpubl.). One implication is that the situation of a weir at this point on the Soar is likely to have been responsible historically for the maintenance of a local increase in river stage near Narborough Bog, with a consequent increase in local groundwater levels. For example, Jarrett (1987) mentions that a mill was recorded at this point on the Soar in the middle of the 13th Century. The relationship of the water-table within Narborough Bog to river levels are discussed later in the thesis, however, the increase in low river stage would most probably have enabled the maintenance of suitable conditions for peat accumulation.

At a similar time to the removal of the weir, a flood prevention scheme was undertaken downstream on the river Soar when a series of meander bends were removed and the course of the river straightened. More recently the floodplain of the Soar has again been modified to control flooding between Leicester and Loughborough. These works are likely to have affected the regime of the Soar upstream at Narborough because the aim of this work was to increase the speed of passage of flood peaks down the river. The objective is to ensure that flood waters from the Soar catchment pass into the river Trent before the flood peak from waters draining the upper area of the Trent drainage basin, and the river Derwent arrive. These changes are also likely to produce a sharper falling limb on flood hydrographs within the Soar drainage basin.

Several roads cross the Soar floodplain near Narborough and Enderby as shown in Figure 4.1. and which can also be identified in Plate 4.1. In addition to the motorway and railway which have been referred to above, two roads cross the river Soar between the villages of Narborough and Littlethorpe, and between Enderby and Whetstone (Whetstone lies at the top of the inset in Figure 4.1.). These roads are possibly significant in modifying the pattern of overbank flooding in the local floodplain. The construction of the railway is likely to have been particularly significant in modifying the area of Narborough Bog flooded during high river flows.

An indication of the direction of lines of surface drainage around Narborough Bog is shown on Figure 4.1. A small stream flows from higher ground to the north west, and has been culverted around Narborough Bog and now discharges directly into the Soar. The stream appears to be groundwater-fed, and some discharge is maintained throughout the year. It is most likely that the stream formerly flowed onto the site, and contributed to the conditions for peat accumulation at Narborough, however, the only indication of this is the stratigraphy collected in the transects across the reed-bed, discussed in section 4.3. The stream was probably first controlled when the surrounding fields were enclosed in 1752 (Jarrett, 1987), as there is no record on local maps of it occupying a different course. A second stream only flows intermittently,

Figure 4.6. Hydrological data from the Soar basin: **A.** Mean daily discharge of the Soar 1983-1990 plotted as a line graph with bars showing extreme maxima and minima daily discharge; **B.** mean monthly precipitation and **C.** mean daily temperature at Braunstone, 1959-1978.



carrying storm runoff from the motorway to the river. Improvements to the culvert and the field ditch in which both streams flow into the river were undertaken at the time of the motorway construction in 1963.

The fields surrounding the reserve formed an extensive area of wet meadows at the time of Wade's (1919) survey; however they have all been drained recently. A field to the west of Narborough Bog, now used as a football pitch, was drained in 1968, and 12 inch drainage pipes were used to drain the meadows to the South in the mid 1970s.

4.1.4. Ecology of Narborough Bog.

A summary of the flora observed at Narborough Bog in recent years is given in Appendix 1. The principal elements of the ecology can be identified by defining three distinct communities within the reserve boundaries, the approximate extent of which are outlined in Figure 4.1. and in more detail in Figure 4.7. These consist of woodland, reed-bed and wet meadow. At present the woodland area has arguably the most interesting flora of Narborough Bog. The crack willow *Salix fragilis* is found throughout this area; however there is also a considerable range of other trees including hawthorn *Crataegus monogyna,* elder *Sambucus nigra,* field maple *Acer campestre,* ash *Fraxinus excelsior,* alder *Alnus glutinosa,* sallow *Salix cinerea* and oak *Quercus robur.* There is also a varied herbage layer which includes herb paris *Paris quadrifolia,* guelder rose *Viburnum opulus* and cleavers *Galium aparine.*

The reed-bed was formerly dominated by the common reed *Phragmites australis* but in this area the invasion of other species is most marked; these include meadowsweet *Filipendula ulmaria,* great willow herb *Epilobium hirsutum,* and in the upper area of the reed-bed, grasses *Poa travilis* and the field thistle *Cirsium arvenze.* The reduced abundance of *Phragmites australis* appears to arise through increased competition from these species as there is a vigorous stand of *Phragmites* on the railway embankment adjacent to the reed-bed.

The conservation value of Narborough Bog lies in the fact that it is the only remaining natural wetland within the Soar floodplain. The occurrence of *Paris quadrifolia* within the wet woodland suggests a possibly ancient history for the site which has been allowed to evolve naturally until very recently. In the wet meadow area to the South, a rich meadow flora has been recorded in recent years; this includes the southern marsh orchid *Dactylorhiza praetermissa*, burnet saxifrage *Pimpinella saxifraga* and meadow saxifrage *Saxifraga granulata.*

In addition to the varied flora of the reserve, some records of fauna have been collected which provide a further indication of the value of the site and a summary of the main peculiarities is given briefly below.

A survey of *Arachnida* in July 1968 revealed a varied number of species at Narborough Bog, including two which were particularly rare. These were *Salorca dicevos*, an extremely rare monitor species indicative of damp woodland with stable vegetation cover; and *Baryphyma pratensis*, a rare species characteristic of damp riverside meadows.

In the period November 1991 to March 1992, a survey of the small mammal population was undertaken by Mr. H. Ball, who found an extremely high population of the long tailed field mouse, *Apodemus sylaticus*, in traps throughout the reserve. Also observed in the reed-bed were the common shrew, *Sarex saraneus*, and the vole, *Micratus arvalis*.

A survey of *Lepidoptera* in July 1991 found several rare wetland species, Round winged muslin, *Thumatha senex*, not known elsewhere in Leicestershire, and the Flame wainscot moth, *Senta flammea*. Ornithological interest lies in the populations of reed bunting *Emberiza schaeniculus* and sedge warbler, *Acrocepholus schoenobaenus*, although these two birds do not appear to have nested at Narborough in recent years.

Narborough Bog still has a varied wetland flora, and some rare wetland plants remain in small areas of wet ground, however, significant changes in the ecology have occurred since Horwood and Noel's (1933) survey when Narborough Bog was considered the main site indicative of the natural wetland flora found along the Soar floodplain. Among individual species lost since this time are marsh pennywort *Hydrocotyle vulgaris*, common water dropwort *Oenanthe fistosa*, early purple orchid *Orchis mascula*, the southern marsh orchid *Dactylorhiza incarnata* and twayblade *Listera ovata*. Furthermore there have been changes in the abundance of those species remaining; for example, there were 31 spikes of the southern marsh orchid in May 1993, while over 100 were observed in 1973. The extent of general changes within the reed-bed at Narborough is illustrated in Plates 4.2. and 4.3. which show the upper area of the reed-bed in February 1984 and February 1992 from the same location. The pictures indicate some encroachment of the sallow bushes on the reed-bed over the eight year period, but also evident are changes in the vitality of the *Phragmites*. Invasion of the reed-bed by grasses is noticeable in the foreground while the *Phragmites* stems are less abundant, and have fewer seed-heads (inflorescence) in 1992.

4.1.5. Management and Development of the Nature Reserve.

As discussed above, the preservation of the site at Narborough Bog is unusual in that it represents an isolated example of the natural vegetation of the Soar floodplain. It is also fortunate that some well-documented ecological surveys of the reserve are available. However, the ecology reflects both the hydrological conditions of the site and also the management practices. In this section the recent management work undertaken by the Nature Trust is described, and some information on the development of the site which has been obtained from historical records is outlined.

The earliest indication of the use of the current area of Narborough Bog is provided by examining the extent of the field network, which pre-dates the construction of the railway in 1863, as field boundaries and ditches can be traced on both sides of the railway embankment. The network of 'rigg and

Plate 4.2. The reed-bed at Narborough Bog, looking northeast, February 1984.

Plate 4.3. The reed-bed viewed from the same point in February 1992. Note the reduced inflorescence.

furrow' in parts of the reed-bed and woodland may also represent elements of a localised water management system. The most probable cause for this is the use of the land as either an osier bed, or floated meadow, although Baker (unpubl.) considers that there were also attempts at the surface drainage of the northern area of the reserve. Fitzrandolph and Hay (1977) describe how the floodplains of the river Trent and its tributaries, including the Soar, were formerly of national importance for osier growing. There was also a large basket making firm in Leicester which supplied baskets for the local hosiery industry.

Baker (unpubl.) mentions that the northern woodland was planted with the crack willow *Salix fragilis* at around 1900, possibly for use as a game covert, which would explain the occurrence of *Salix fragilis* together with *Salix viminalis*, as only the latter species is normally associated with osier beds. Quantities of reed were cut from the reed-bed for thatching until the early 1900s, while horses and cattle have been grazed within the southern meadow area until comparatively recently.

Changes in local land-use identifiable on Ordnance Survey maps of the area, from when the land was first surveyed in 1884-85 until the last survey in 1966, are summarised in Figure 4.5. The maps indicate the extent of changes in the urban area, and illustrate the impact of the construction of the motorway. The changes in built-up area include the spread of housing from the villages of Enderby to the North, and Narborough to the south west. The maps provide a record of changes near the Enderby Mill which were briefly discussed above. In Figure 4.5. the position of the mill, the weir across the Soar, and the mill race are shown, in the map of 1884-85.

The changing characteristics of the land surrounding the current nature reserve are also indicated by a revision of the area designated a Site of Special Scientific Interest. The area covered in 1956 is outlined in the top diagram in Figure 4.5. This drawing is based upon the Ordnance Survey map of 1885, but there was little change in the period to 1956. The original SSSI extended directly upstream of the Enderby Mill and included a large field on the east of the river Soar which formed part of Wade's (1919) study. A substantially larger area was designated a SSSI in comparison with the area currently managed as a Nature Reserve. The construction of the motorway in 1963 removed the south west corner of the area, and the recreation ground to the west was first ploughed in 1965, and then drained early in 1968. Consequently, when the designation of Narborough Bog was reviewed in 1981, the area notified was decreased to the outline shown in the bottom drawing in Figure 4.5. This change in area of the SSSI consists of a decrease from just over 26 hectares in 1956 to 9.5 hectares in 1981, so that the current SSSI directly corresponds with the area of land owned by the local Nature Trust.

The whole area currently designated a SSSI was purchased by the Leicestershire and Rutland Trust for Nature Conservation in 1975, who had leased 19 hectares annually since 1967. Acquisition of the land was justified by a desire to protect the reedswamp and alder/willow woodland. Management of the reserve by the Nature Trust has several specific objectives: to maintain the alder/willow association and restrict the growth of non-native species such as Sycamore; to prevent further deterioration of the *Phragmites* reed-bed; to maintain the diversity of the marsh area; and to encourage research on the existing fauna.

There has been no recent maintenance work at Narborough Bog to control the hydrology, although at one time the removal of surface deposits within the reed-bed was suggested as a means of providing an improved wetland habitat for the flora and fauna. Two small ponds were excavated in the 1970's and 1980's to investigate whether this was practicable. Water levels remained stable within the ponds; however excavation would require the removal of a large volume of sediment and would destroy the sedimentary record, and the suggestion seems to have been abandoned.

4.1.6. Rationale for the selection of the field-site at Narborough Bog.

The wetland nature reserve at Narborough was selected for this study because of the opportunity it presented to undertake a local investigation of a wetland system. Section 1.2. described how wetlands formerly covered extensive areas of floodplains in lowland Britain (Wilcox, 1933) of which the site at Narborough Bog would seem to comprise an isolated remnant. There are few local examples of floodplain wetlands, as opposed to flood meadows which are more numerous in the Soar floodplain. The nearest example is Moseley Bog in Birmingham; other British examples were given in section 1.2.

As will become clear from the following section, equipment constraints required regular visits to the site for monitoring water-table changes during periods of precipitation and evapotranspiration and a site close to Leicester was therefore desirable. It would not have been possible to maintain frequent visits over an extended period if a field site had been chosen further from Leicester.

The small size of Narborough Bog does not represent a problem as the reserve lies in an intermediate position in relation to the other wetlands summarised in Table 2.3. Indeed the small size enables a dense distribution of points to monitor water-tables thereby providing information on water flux at a micro-scale. The problem with the study of this size of wetland is that hydrological processes outside the boundaries of the site are likely to exert a significant influence on the direction and magnitude of water fluxes; here river flooding is likely to be of greatest interest. In this respect, the effects of local construction work in the floodplain should be considered. The railway line which crosses the reserve at right angles to the river Soar appears to lie directly upon the peat deposits of the wetland. These deposits will have been compressed by the hardcore embankment, thereby decreasing their permeability and reducing lateral seepage. More significant is likely to be the effect of the embankment on the pattern of overbank flooding as floodwaters were directed towards the river following construction of the railway. Consequently there would have been a reduction in the flood frequency of the upper reed-bed.

4.2. FIELD MEASUREMENT PROGRAMME.

4.2.1. Introduction.

Measurements of hydrological and meteorological variables were made at Narborough Bog throughout the three year period of this study from 1991-1993. The intention was to obtain field data which recorded the response of the wetland system described above to a full range of hydrological and meteorological conditions. These would encompass two complete years, 1991 and 1992, and would therefore include a collection of precipitation events at different times of the year, summer evapotranspiration, and also occasional overbank river floods. Although one of the original justifications for this study was the concern about falling water-tables at Narborough Bog, it was accepted that the monitoring programme would not be of sufficient length to resolve these questions completely, and consequently an emphasis was placed upon understanding the contemporary hydrology of the site. In this way it was hoped that a conceptual understanding of the site hydrology would enable an assessment of the potential stress which the wetland would experience under a range of natural hydrological conditions.

The rationale for wetland hydrological studies has been briefly summarised by Mitsch and Gosselink (1986 p. 86). They describe how useful data can be obtained very simply by studying wetland systems, if measurements of water level change are collected and precipitation and evapotranspiration estimates obtained. The instrumentation summary given by Freeze and Banner (1970) also provides a good background to the hydrological requirements of the field programme. Freeze and Banner (1970) described a combined laboratory and field approach to monitor groundwater recharge and discharge and outlined the benefits of integrated hydrological measurement about the position of the water-table.

The use to which the field data would be put determined the initial form of the measurements. There were essentially two main purposes:

1. To obtain a record of saturated and unsaturated water contents

throughout the reed-bed area at Narborough Bog, which would provide a qualitative indication of site hydrology.

2. To provide data which would be suitable firstly to help the formulation of a deterministic hydrological model of Narborough Bog, and secondly to assess the accuracy of the model.

4.2.2. <u>Dipwell Installation and Monitoring.</u>

The concept of hydraulic head as a measure of water potential was introduced in chapter 3. Measurement of spatial variation in hydraulic head is an important means of determining the directions of water flow, as discussed by Reeve (1986). The height of the water-table was measured at several locations at Narborough Bog using a network of dipwells. The network was designed to produce a sufficient spatial distribution that would enable the plotting of groundwater level contours, and spacing was therefore partly determined by the expected hydraulic gradient, which was suggested by the results of some previous water-table monitoring at the site. However, the location of dipwells was also constrained by access problems and particularly the need to avoid excessive walking over certain areas of the nature reserve.

The location of all the dipwells used in the study is shown in Figure 4.7., and their characteristics are summarised in Table 4.1. Here the coordinates of each dipwell, the length and diameter of the well, and comments on their location are given. Dipwells were installed on several occasions both before and during this study. An isolated dipwell, numbered 3o, was installed by Dr. A.G. Brown in 1984 in the centre of the reed-bed area. In May 1985 six dipwells were installed by Mr. W. Moffat, numbered 0, 2, 3o, 4o, 5, and 6. This comprised the dipwell network which existed prior to this study, namely a transect of dipwells: 0, 2, 3o, 4o and 6 with two additional dipwells on either side of the central line. All the dipwells had a base at c. 1.25m below the ground and had holes drilled in the bottom of the pipes.

Two groups of dipwells were installed in November 1990 to provide greater spatial resolution. Firstly, additional dipwells were located along the original transect, concentrating particularly on the area near the river which was not covered by the existing dipwells. These dipwells, numbers 1, 4n, 5n, 7, 8, 9, 10 and 11, were of outside diameter 5cm. They vary in total length, and rest either on wood peat or gravel as summarised in Table 4.1. A second transect of dipwells, numbers M1, M2, M3 and M4, was located in the wet meadow south of the railway line to examine whether there was any difference in hydraulic gradient over the small distance between the two transects. All dipwells were installed using a Dutch auger to produce a hole of slightly greater diameter than the plastic tube. The hole was then packed with fine shingle of diameter c. 5mm, and the dipwell inserted with more shingle packed around the sides. Dipwells were secured by consolidating the dipwell sides with sediment produced from augering the initial hole. Heights of the dipwells were surveyed to an accuracy of ±1mm, and later related to Ordnance Datum.

The dipwell network in November 1990 thus consisted of 15 dipwells centred on a transect through the reed-bed, and a second transect of 4 dipwells in the southern meadow. At this time it was anticipated that a two-dimensional profile model would be used to examine the relationship of water-tables within Narborough Bog to distance from the river, for which this coverage of dipwells would be sufficient. Four further dipwells, numbers 12, 20, 21 and 30 were installed at the margins of the central transect to provide additional information on the representativeness of the main dipwells. However, it became clear that a three-dimensional model would permit greater analysis of water-table variation. Consequently in June 1991 additional dipwells were installed to provide a second transect of dipwells at right angles to the river. These dipwells, numbers 21, 23, 24 and 25 were sufficient length to lie on the underlying gravel aquifer. Three further dipwells, numbers 26, 27 and 28) were installed to a depth of 50cm below the ground adjacent to existing dipwells (nos. 3o, 5o and 23). This was with the intention of investigating the extent of depth dependent changes in hydraulic head to help validate the application of a full three-dimensional model during the modelling phase of the study. The

Figure 4.7. Site diagram of Narborough Bog, indicating location of numbered dipwells, tensiometers, and the stage recorder.

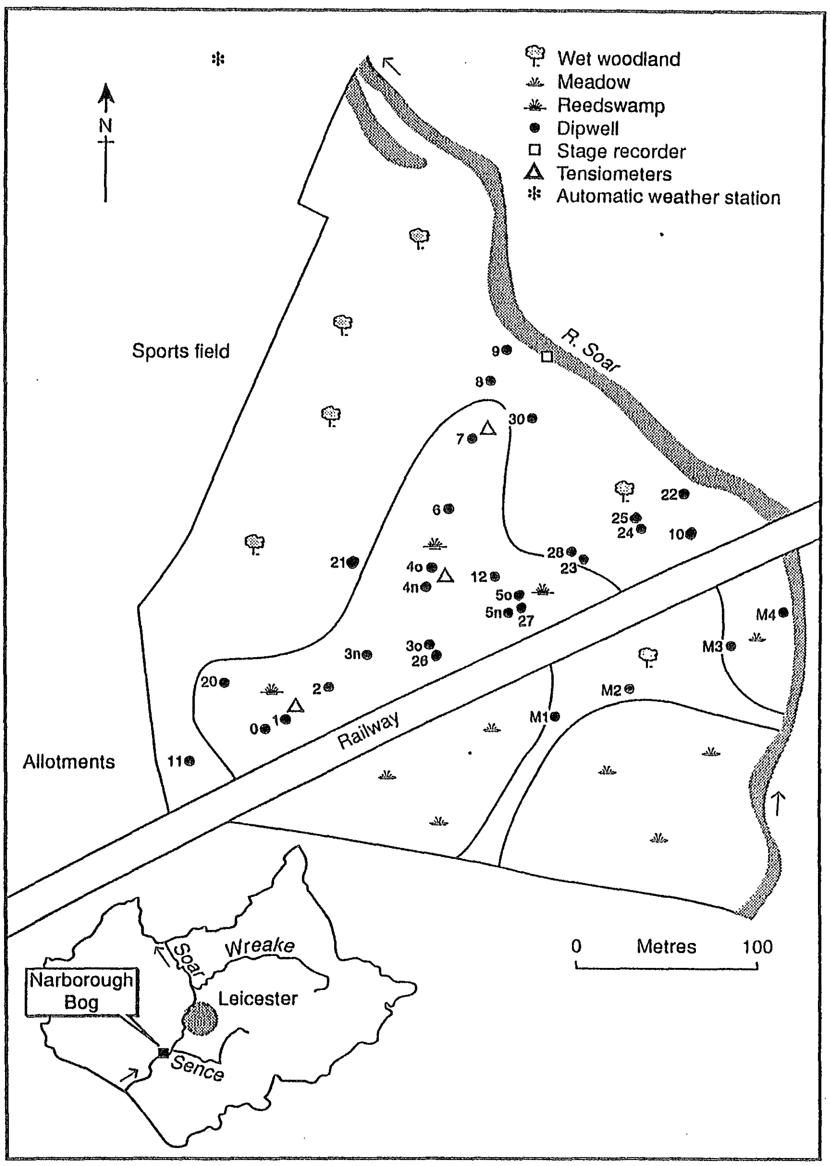


Table 4.1. Characteristics of individual dipwells.

no.	Date installed.	Coordinates.	Comments.
0	May 1985 (M)	1011.164 984.636 63.540	Installed by Mr. W.S. Moffat in a transect at right angles to the river Soar. The wells consist of grey plastic tubing, 5cm O.D., installed to a depth of 1.25m below the ground surface. Holes were drilled in the base of the tube and gravel was used to infill the base.
2	May 1985 (M)	1013.471 940.841 63.690	
3o	May 1985 (M)	1014.253 896.554 63.544	
4o	May 1985 (M)	1042.978 864.814 63.366	
5o	May 1985 (M)	1019.151 833.583 63.250	
6	May 1985 (M)	1070.014 835.153 63.390	
3n	1984 (B)	1015.155 921.988 63.377	4cm OD. white pipe; c. 1.7 m long.
1	Nov. 1990 (CB)	1010.808 967.743 63.403	Installed early in this project to supplement the existing transect. They comprise 4cm OD. white pipes with holes drilled in the bottom 25 cm, and shingle packed at the sides. Dipwells 1, 4n and 5 were of 1.25m length, and rest on wood peat. Dipwells 7, 8 and 9 were 1.25m long, but have greater quantities of clay, and hence were less responsive to water infiltration. Dipwells 10 and 11 were 1.5m long, and were installed at the margins to increase coverage. No. 11 lies on a thick red clay layer and falls outside the area of peat coverage.
4n	Nov. 1990 (CB)	1039.661 870.120 63.220	
5n	Nov. 1990 (CB)	1018.420 841.174 63.088	
7	Nov. 1990 (CB)	1092.455 816.335 63.210	
8	Nov. 1990 (CB)	1127.873 784.981 63.006	
9	Nov. 1990 (CB)	1128.804 755.254 62.133	
10	Nov. 1990 (CB)	1010.682 725.145 62.954	
11	Nov. 1990 (CB)	1020.640 1039.548 63.729	
M1	Nov. 1990 (CB)	929.999 877.658 63.525	Transect installed to study the meadow area. The tube type, diameter and installation was as above. M1 was 2.25m long and lies on gravel near the river. M2, M3 and M4 were 1.8m long, over gravel.
M2	Nov. 1990 (CB)	932.413 824.083 63.288	
M3	Nov. 1990 (CB)	941.012 746.572 62.833	
M4	Nov. 1990 (CB)	945.407 703.710 62.891	
12	April 1991 (CB)	1037.644 827.663 63.459	Installed to increase the dipwell coverage at the margins of the existing transect. They were all 5cm OD. 20 was 1.25m long and lies on wood peat; 21 was 1.60m long, resting on gravel; 12 and 30 were both 1.80 m long, resting on gravel.
20	Nov. 1990 (CB)	1054.234 954.510 63.436	
21	June 1991 (CB)	1044.868 899.256 63.660	
30	July 1991 (CB)	1080.708 778.036 63.160	
22	June 1991 (CB)	1033.782 711.854 62.231	Introduced to provide a further transect for 3-D coverage of water-tables. All were 5cm OD. No. 22 rests on sand and gravel 1.3m deep near the river; Dipwell nos. 23 and 24 are in a line away from the river; no. 23 is 2.27m deep, and no. 24 2.58m deep. Both lie over gravel. At the same time some shallow dipwells, nos. 25, 26, 27 and 28, were installed to a depth of 0.5m and lie at the bottom of the herbaceous peat, positioned next to existing dipwells.
23	June 1991 (CB)	1035.162 772.627 63.066	
24	June 1991 (CB)	1036.014 747.776 62.654	
25	June 1991 (CB)	1035.872 747.656 62.641	
26	July 1991 (CB)	1013.906 896.516 63.401	
27	July 1991 (CB)	1018.833 833.384 63.316	
28	July 1991 (CB)	1034.888 772.412 62.947	

depth of the water-table below the top of individual dipwells was measured using an electrical dipstick, with an approximate accuracy of ±2 mm. The dipwells were evacuated periodically to ensure that they maintained a reasonable hydraulic contact with the surrounding deposits. The dipwells were monitored as an entire network, so that water-table depths were measured at a consistent frequency between dipwells. Detailed monitoring commenced in November 1990 and continued until June 1993. The interval between readings varied depending upon hydrological conditions, but dipwells were read at least twice a week, and before and immediately after rain events. Thus in 1990 the network was read a total of 25 times during November and December. In 1991 readings were taken on 188 occasions, and in 1992 on 173 days. Over the two calender years, 1991 and 1992, water-tables were thus read at a mean interval of two days which reflects the greater frequency of readings during periods of rainfall and overbank flooding.

4.2.3. Measurement of River Stage.

Variations in stage of the river Soar at Narborough Bog were measured during the study to examine the relationship between water-tables and river levels, and to provide a record of flood levels. A stilling well was installed in the river Soar at the site shown in Figure 4.7. in January 1991. The equipment consisted of a 20 cm diameter float with counterweight which was attached to an Ott stage recorder mounted in a box above. The recorder consisted of an analogue drum chart which advanced 4.8 cm every 24 hours. The instrument was checked at monthly intervals, at which time the clock was rewound and the chart removed and scanned manually to produce data in a digital format.

A stage board was also mounted in the river; this was read as part of the routine measurement programme, and provided a method for checking the accuracy of the stage recorder. The heights of both the stage board and stage recorder were surveyed to enable direct correlation with the dipwell measurements.

4.2.4. Tensiometers.

Soil moisture suction, or tension, was measured at several sites adjacent to the dipwell points. Tensiometers consist of a porous cup, filled with dearated water, which is placed in close contact with the soil matrix. Cassell and Klute (1986) consider the use of dearated water to be essential. It was obtained by boiling water for over five minutes and then minimizing the water surface area during cooling. Soil water is able to move in both directions through the porous cup to equalise any differences in potential. One end of the tensiometer is composed of a vacuum seal, and changes in pressure which follow water movement are observed using a pressure gauge. Two types of tensiometer were used during the study. Six Soil Moisture Equipment Corporation model 2900F soilmoisture probes were installed at Narborough. Two different lengths were used, 45cm and 60cm, and were positioned in three locations as indicated in Figure 4.7. Some additional tensiometers were constructed from porous cups which were read manually using a hand-held pressure meter.

4.2.5. Meteorological Measurements.

Meteorological data were collected both to provide an accurate indication of precipitation totals for Narborough Bog, and also for input into an evapotranspiration model, to assess evapotranspiration loss during the study.

The choice of a location for an automatic weather station presented a problem as occasional trouble with vandalism had been reported at Narborough. An ideal position would have been in the middle of the reed-bed, however, this would have been too conspicuous, and also evapotranspiration models typically require temperature and relative humidity to be measured at a fixed height above the ground or vegetation surface. This would have been difficult within the reed-bed where *Phragmites* and *Epilobium* grow to heights approaching 2m. Consequently, the automatic weather station was located in the garden of a house close to Narborough Bog. The garden was adjacent to an area of alder-willow woodland beside the river Soar which formed part of the original area designated a Site of Special Scientific Interest, so that measurements should be reasonably representative of the study area.

The weather station was composed of sensors for measuring temperature, relative humidity, wind speed, wind direction, incoming radiation and precipitation. Measurements were recorded by data logger which was downloaded at weekly intervals. The logger was programmed to measure temperature and relative humidity every ten minutes, and all other parameters every twenty seconds. This was necessary to include possible short term fluctuations. Data were summarised every fifteen minutes and summary hourly and daily characteristics which included maximum and minimum temperature and totalised rainfall were also obtained.

Temperature and relative humidity were measured by a Rotronic MP100 probe at a quoted accuracy of $<\pm 1\%$ RH at 25°C, precision $<0.5\%$ RH, while temperatures were measured to an accuracy of above ±0.2°C. The sensors were mounted within a radiation shield, which was mounted to the steel tripod of the weather station.

Wind speed was measured using a Vector A100R anemometer, with an accuracy of 1 per cent ±1 ms^{-1}. The anemometer was installed on the top arm of the weather station tripod. Wind direction was monitored using a Vector W200P windvane which comprised a 358° potentiometer. Wind directions can be recorded at wind speeds upto 75 ms^{-1} with an accuracy of ±1°. The instrument was mounted on the top arm of the tripod, and was orientated to the North using a prismatic compass. Solar radiation was measured using a Li-Cor LI200SZ pyranometer, which produced a voltage output on a scale ranging from 0 to 10 mV. The readings were multiplied by a constant to give flux density (kJm^{-2}). Precipitation was measured using a tipping bucket rain gauge, with a mechanism calibrated to tip for every 0.204mm total precipitation.

Meteorological data collected by the automatic weather station, were used to estimate potential evapotranspiration from the Penman equation. A full description of the procedure is given in Appendix III, where the selection of the Penman method is justified. The derivation of the computer program, which was written specifically for this purpose, is also discussed in detail. Evapotranspiration figures quoted in the following chapters were obtained using this method, and hence it is important to consider the accuracy of estimates. While evapotranspiration is ideally measured directly in the field, rates of water loss are proportional to radiation levels, and so the results should indicate the magnitude of daily and seasonal variations in water loss.

4.2.6. Conclusion.

The previous sections have described the nature of the routine measurements which were undertaken at regular intervals during this study. Their relationship to the overall structure of the project can be seen from Figure 1.1. in which the importance of additional experiments and hydrological modelling simulations is clear. Measurements of water-table, soil water tension, river stage, and meteorological variables were used both to determine the current hydrological processes at Narborough Bog, and also for the calibration of a groundwater model of the site. In the following chapter the results of the monitoring programme are described to facilitate a discussion of the site hydrology.

Chapter 5

Results of Water-table Monitoring

Scope of Chapter

This chapter introduces results from the water-table monitoring programme, the aims of which were discussed in the previous chapter. The chapter begins by outlining hydrological records from Narborough Bog which were collected in 1985-86, before this study commenced. The data obtained during this three year study are then presented and analysed using summary graphs for meteorological data and dipwell records. Various formats are used including time series plots, daily transects, and 3-dimensional plots of the water-table surface. Sections also consider the relationship of water-tables with the reed-bed (section 5.3.1.), variations in vertical water fluxes, examined by studying observations for hydraulic heads from dipwells at varying depths (section 5.3.2.), and the records of unsaturated water content obtained from tensiometer measurements (section 5.3.3.). The results, are considered at annual, seasonal and event scales, and illustrate how a variety of processes are together responsible for the pattern of water-table fluctuations at Narborough Bog, through space and time.

The qualitative description of results is followed by time series analysis of the records of selected dipwells. The analysis aims to quantify the relationships between water-table response and meteorological variations. The characteristics of individual events are also considered in order to examine whether the importance of particular hydrological processes, for example infiltration, varies significantly through time.

5.1. PREVIOUS HYDROLOGICAL RECORDS FROM NARBOROUGH BOG.

Chapter 4 highlighted how observations of a changing ecological balance within the reed-bed area at Narborough Bog have raised concern about the hydrology of the site. Early studies at Narborough Bog were principally concerned with describing the ecology of the site, however, the possibility of ecological changes in the reserve as a result of modifications to hydrological processes was raised, for example, by Walpole (1972). Concern initially arose in the period following the removal of a weir across the Soar downstream of Narborough Bog in the 1960s, which produced a fall in low flow river stage. Unfortunately it is very hard to quantify the consequences of this work as local records of river stage only extend back to 1971.

In the 1980s, ecological changes within the reed-bed became more evident and as a result some local monitoring of water-table levels was undertaken. Six dipwells were installed in the reed-bed at Narborough Bog in May 1985 as described in section 4.2.2. These consisted of 5cm diameter plastic tubes with holes drilled near the base, which were installed to a depth of c. 1.25m below the surface. Their locations are indicated in Figure 4.7. (numbers 0, 2, 3o, 4o, 5o, and 6). The depth of the water-table below the dipwell top was measured at intervals varying from one to four weeks from early June 1985 to the end of May 1986, and the full data-set for three dipwells, numbered 0, 3o, and 6 is given in Figure 5.1., with precipitation plotted for the same time period. Although the temporal resolution is insufficient for the data to indicate the detailed water-table response to precipitation events over the time period, the annual cycle in water levels can be clearly identified. In particular, the extent to which summer water-tables are lowered by evapotranspiration is evident. This accounts for the significant difference in water-tables over the period which covered a range of 45cm for dipwell 0 and 55cm for dipwell 6. During the monitoring period minimum water-tables were observed on 23rd September 1985, and maximum water-tables on 3rd February 1986.

The data also indicate some spatial trends in the water-table. The dipwells are located in a transect towards the river; the water-tables recorded from dipwells 0 and 3o were not significantly different, while nearer the river the water-table at dipwell 6 was on average 0.2m lower. This implies that the

Figure 5.1. Water-table observations recorded in 1985-86, with daily precipitation plotted below at the same scale.

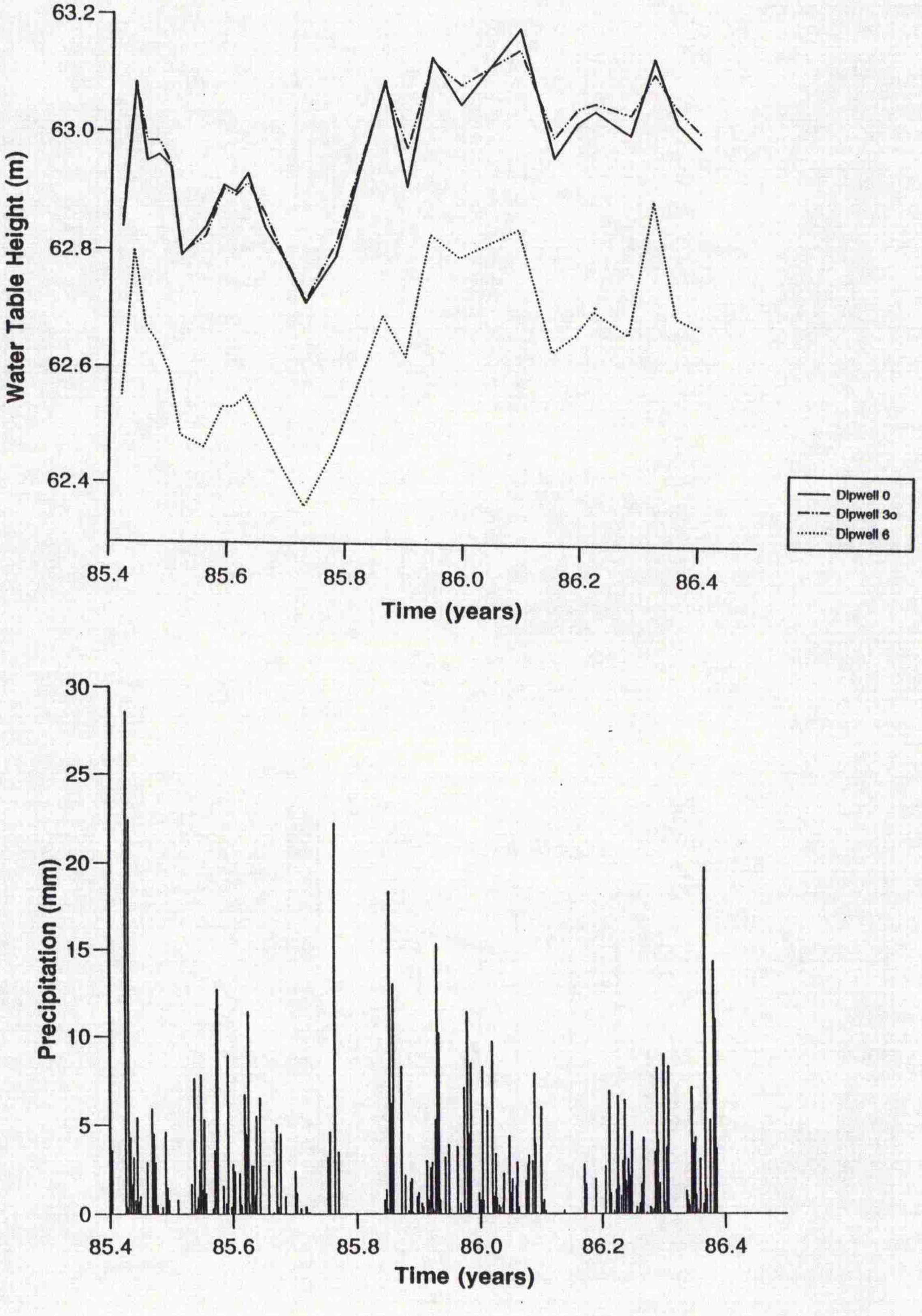


Table 5.1. Comparison of hydrological data for the years 1985-86, 1991-92 and 1992-93.

	1985 - 1986	1991 - 1992	1992 - 1993
Dipwell 0	62.953	62.865	62.935
Dipwell 2	63.032	62.749	62.924
Dipwell 3o	62.964	62.710	62.820
Dipwell 4o	62.801	62.609	62.722
Dipwell 5o	62.683	62.484	62.601
Dipwell 6	62.623	62.440	62.556
Mean	62.843	62.643	62.759
St. Dev.	0.152	0.149	0.147
Precipitation	666.4 mm	497.5 mm	732.37 mm

direction of water-table drainage was towards the river (i.e. from dipwells 0 and 3o towards dipwell 6) during the times when measurements were taken.

As part of this thesis, records of water-table depth were taken from the same dipwells used in the period 1985-86. A summary for the three periods 26 May 1985 - 25 May 1986; 26 May 1991 - 25 May 1992; 26 May 1992 - 25 May 1993 is given in Table 5.1. The tabulated data consist of mean water-table levels at the six dipwells, and cumulative precipitation over the period. Although few conclusions can be drawn from the data due to the infrequent measurement interval in 1985-86, generally the results indicate a consistent relationship in the relative water-table positions between individual dipwells. The relationship of water-table height to precipitation in the three years is harder to identify; higher water-tables were observed in 1985-86 compared with 1992-93 although precipitation was lower. This most probably reflects differences in the timing of readings and a direct comparison of the water-table data would only be possible if the results from 1991-1993 were filtered to produce a similar measurement interval to 1985-86, however, significant periods of high or low water conditions might be lost if this were undertaken.

Differences between the three periods were investigated further using a two-tailed difference of means test. Comparing the data from 1985/86 with 1991/92 produced a t-statistic of 2.304. This was significant at the 5% level but was insufficient to reject the null hypothesis of no difference between the means at the p=0.01 level for 10 degrees of freedom. Similarly, comparing 1985/86 with 1992/93 gave a t-statistic of 0.971; and for 1991/92 and 1992/93 the t-statistic was 1.362. This suggests that while the tabulated data indicate differences in water-table heights for the three periods, the difference is not significant and probably indicates spatial variation in the pattern of water-table response.

While of some interest in indicating the broad annual cycle of water-tables at Narborough Bog, the records should ideally be viewed together with precipitation and evaporation data to determine the response of the local water-table to individual hydrological events. Certain events may not be apparent from the data record due to the occasionally long measurement interval which ranged from one to four weeks. The lack of detail in the data demonstrate the importance of maintaining regular water-table measurements, within a flexible framework in which readings are taken at varying intervals, before; and after individual rain and flood events.

5.2. DIPWELL SELECTION.

In chapter 4 the characteristics of the entire dipwell network were summarised. The monitoring network consisted of dipwells installed on several occasions, with the initial intention of only instrumenting one transect at right angles to the river Soar. Additional dipwells were then introduced to enable a three-dimensional groundwater model to be applied. In this discussion chapter of results from the water-table monitoring it is necessary to be selective

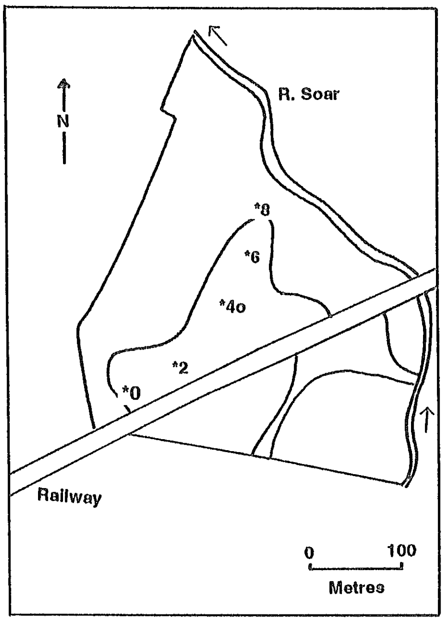


Figure 5.2. Location of dipwells considered in detail in this chapter.

in choosing dipwells to examine in greater detail, because of the amount of data available. Thus in this chapter the records from dipwells 0, 2, 4, and 6 are used to characterise the water-table fluctuation with distance from the river. Their location is given in Figure 5.2. Some data are also used from other dipwells, for example when looking at the variation in hydraulic head with depth; where necessary their locations are shown in Figure 4.7.

Those dipwells selected for further consideration were installed in May 1985, and are unlikely to suffer from any disturbance to the ground which is inevitable after dipwell installation. The dipwells have an identical diameter and depth, and all rest upon the underlying gravel aquifer.

5.3. ANNUAL WATER-TABLE RECORDS 1991 AND 1992.

The seasonal pattern of water-table variations has been termed the hydroperiod by Mitsch and Gosselink (1986, p. 58), who present examples of hydrographs for several different wetland types. Distinct spatial and temporal variations in water levels should reflect the interaction of atmospheric and hydrological processes and the physical characteristics of the study area, which together determine water-table height (Section 2.3). A representative view of site hydrology therefore requires analysis of both the spatial and temporal variations in water-tables. Figures 5.3. and 5.4. show three-dimensional plots of water surface in 1991 and 1992. The graphs were obtained using interpolated data from six dipwells (nos. 8, 7, 6, 4, 3, and 2). Time is plotted along the x axis in Julian Day format whereby January 1st is equal to JD 1, January 31st equals JD 31 and December 31st JD 365. Distance from the river, in metres, is plotted along the y axes and water-table height on the z-axis. The dipwells are situated progressively further from the river and, although there is some distortion and individual detail is lost, the data illustrate both the relationship of water-tables to distance from the river, and the nature of the annual hydroperiod.

A comparison of Figures 5.3. and 5.4. reveals a significant difference in form which reflects differences in climatic conditions between the two years, 1991 and 1992. The predominant direction of water-table drainage remains towards the river throughout the period, and water-tables vary with a greater annual range in the upper area of the reed-bed. The most notable difference between the data for the two years is the summer depletion of the water-table in 1991 which produced a pronounced trough in the water surface centred on JD 250 (7 September) in Figure 5.3. This summer lowering of the water-table was completely absent in 1992 due to a more even precipitation distribution throughout the year. Over the two years, major precipitation events can be identified as individual ridges in the water surface, consisting of rapid increases followed by decreases in the water-table. The only exception to this was a period of snow-melt in 1991 when slow infiltration of snow-melt water produced a gentle mound in the water-table centred on JD 100 (10 April) in Fig.

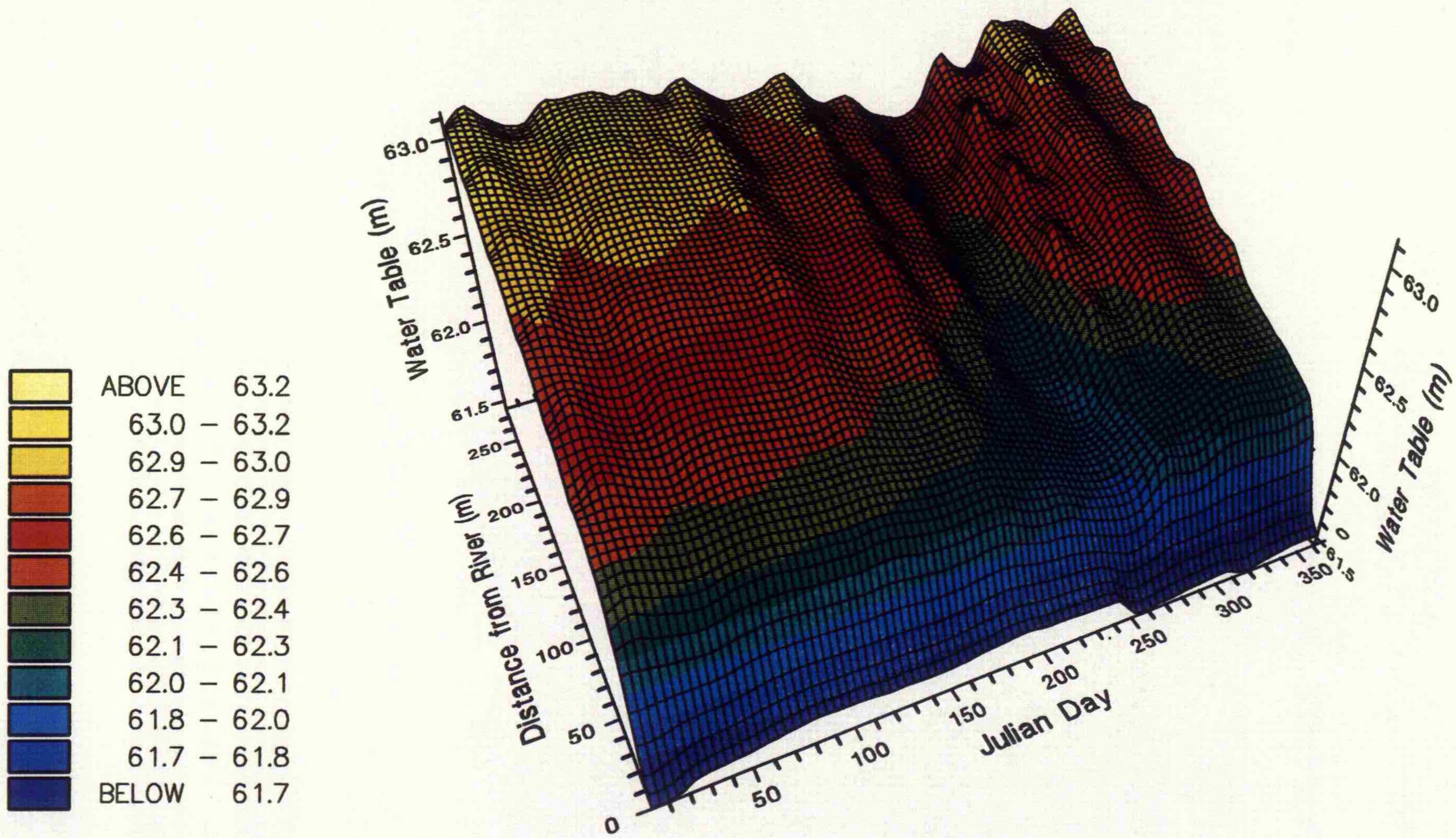


Figure 5.3. Three-dimensional graph showing interpolated water surface for 1991. The summer depression of the water table appears as a trough centred on Julian Day 250.

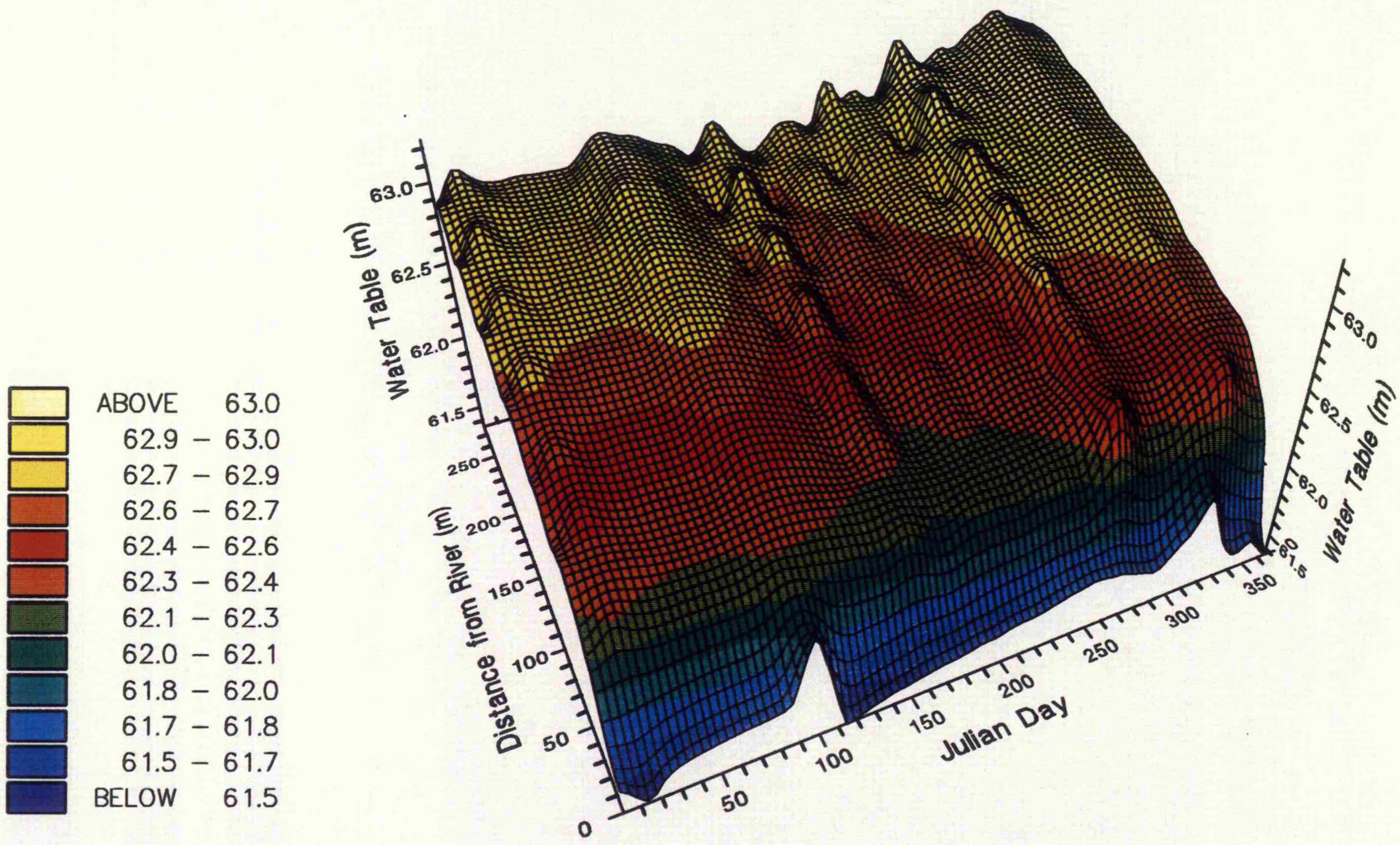


Figure 5.4. Three-dimensional plot of water surface for 1992.

5.2. In December 1992, water-tables remained high in the upper area of the reed-bed due to steady rainfall on a saturated catchment. In general, the magnitude of water surface fluctuations increases with distance from the river; this may partly reflect the dampening effect of the river, but also corresponds to the differences in stratigraphy which were discussed in Section 4.1.3. Typically water-tables are more responsive to precipitation in the area covered by peat deposits as opposed to the silt-clay area adjacent to the river.

The three-dimensional plots show that the dominant direction of drainage is towards the river. However, there are isolated occasions when a reversal of the hydraulic gradient occurs. Several significant reversals in gradient occurred in water-tables adjacent to the river in 1992, on JD 15 (15 January), JD 280 (7 October) and JD 350 (16 December). These were all days of overbank flooding and reflect infiltration of water lying over a small area adjacent to the river as a result of overbank flooding. The localised changes in hydraulic gradient are discussed in more detail later in the chapter when the changes in gradient are also plotted over time, enabling clearer identification of individual events.

The relationship of the water-table to variations in meteorological parameters can be seen in Figures 5.5. and 5.6. for 1991 and 1992. Here data from three dipwells are used (nos 2, 4o, and 6), which are located in a transect across the reed-bed at distances of 238m, 151m, and 111m respectively from the river Soar. Precipitation and evapotranspiration are also plotted at the same time-scale, using data from the automatic weather station at Narborough. Daily precipitation totals are shown, with smoothed daily evapotranspiration obtained using a modified Penman equation (chapter 4 and appendix III). The consistent difference in the water-table height for the three dipwells indicates a general decline in the water surface towards the river. The relationship between the hydrological record of the three dipwells also remains consistent over the time period, indicating a similar response to different meteorological events.

Figure 5.5. Water-tables recorded for dipwells 2, 4o and 6 in 1991 with precipitation and evapotranspiration.

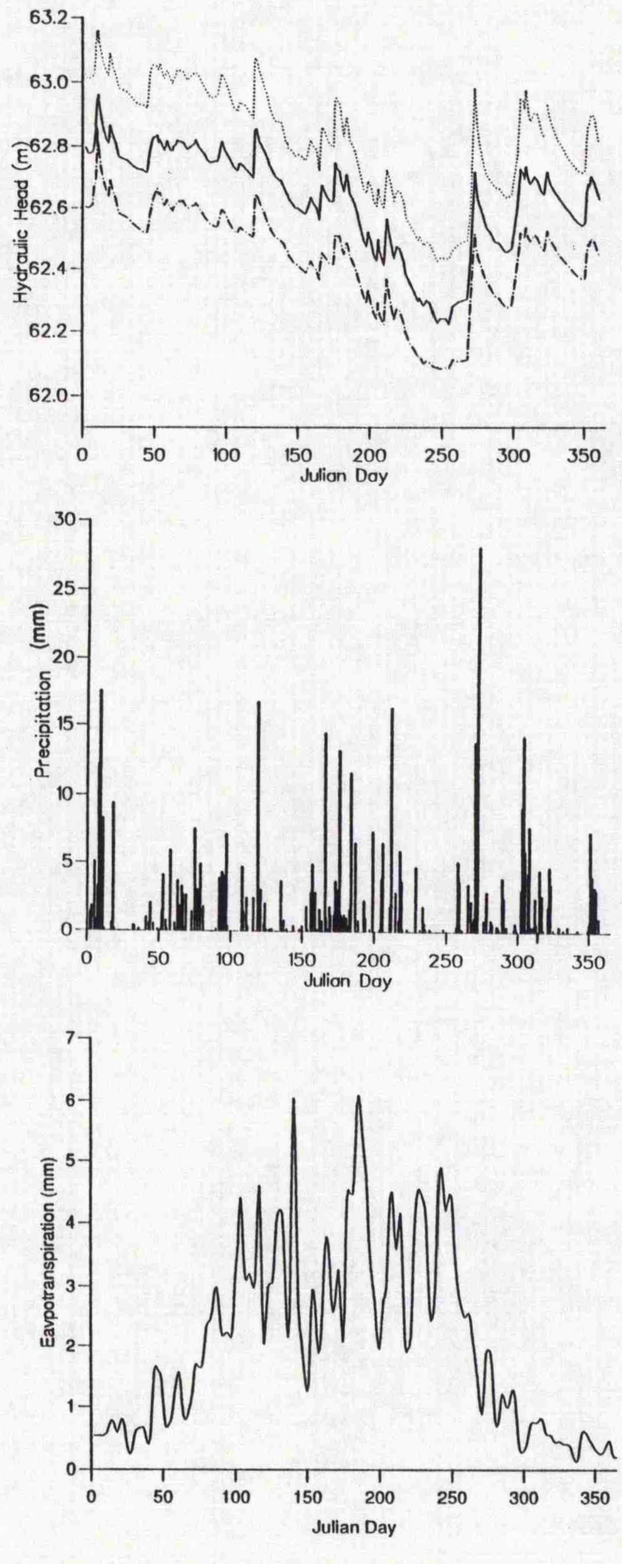


Figure 5.6. Water-tables observed for dipwells 2, 4o and 6 in 1992 with meteorological data for comparison.

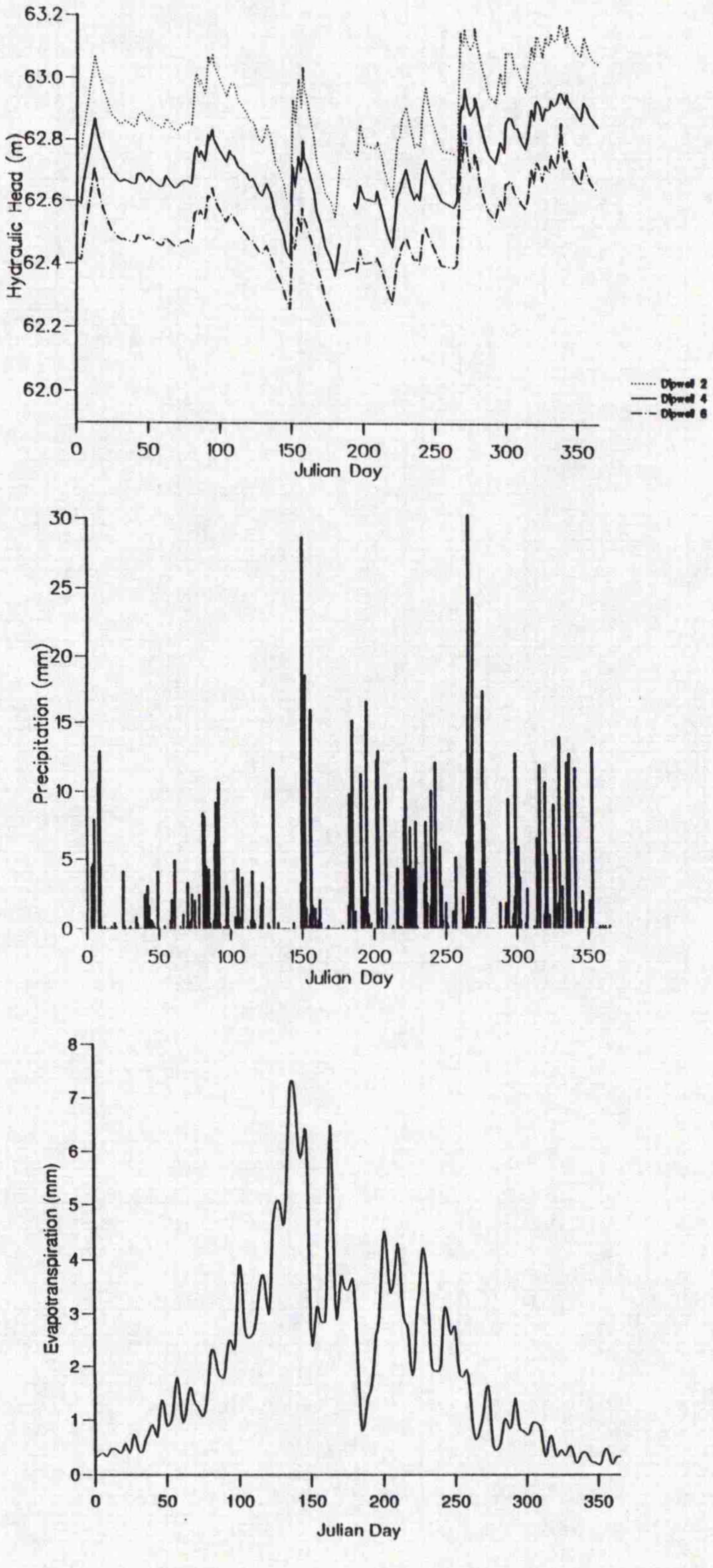


Table 5.2. Summary statistics for 1991 and 1992.

	1991	1992
Dipwell 2 (m) Mean	62.85	62.94
Maximum	63.16	63.18
Minimum	62.43	62.55
Standard Deviation	0.17	0.15
Dipwell 4o (m) Mean	62.66	62.74
Maximum	62.96	62.97
Minimum	62.22	62.36
Standard Deviation	0.16	0.14
Dipwell 6 (m) Mean	62.47	62.55
Maximum	62.86	62.85
Minimum	62.09	62.20
Standard Deviation	0.16	0.14
Precipitation (mm) Mean	1.37	2.08
Maximum (per day)	28.15	30.19
Total	497.7	759.7
Evapotranspiration* (mm) Mean	2.07	2.15

* missing data in 1992 from JD 143 to 161.

The water-table at Narborough in 1991, shown in Figure 5.3. and 5.5., was initially high in early January as a result of several days of heavy rainfall, a total of 39mm falling between January 8 to 11. There was a further rainfall induced peak in the water-table on January 18 when 9mm of rain fell, after which the water-table fell gradually until infiltration of water from snow-melt produced an increase around February 15. Two months of well-distributed moderate rainfall followed which produced fluctuating high water-tables with a sharp peak on April 29 following 16mm rainfall. After this point evapotranspiration increased significantly and water-tables fell, reaching a minimum level on September 7 (JD 250), although the falling trend was partially interrupted by several days of moderate rainfall between mid-June and early July (JD 160-185).

The period of low summer water-tables ceased in early September when a total of 49mm rainfall fell over the three days September 27 - 29 (JD 270-272); however the water-table then fell steadily due to a combination of continuing evapotranspiration loss and an absence of any large precipitation events. A further concentration of rainfall in late October to early November (JD 304-308) brought another increase in water-tables although they fell again due to the lack of rainfall from mid November to mid December (JD 323-349). Significantly, during this period of relative drought, substantial water-table decline occurred despite low levels of evapotranspiration. The implications of this for the hydrology at Narborough Bog are discussed in detail later, however, it implies continued water loss through seepage.

The annual hydrographs for the same three dipwells for 1992, shown in Figure 5.4. and 5.6., are very different from the preceding year. Again the water-table was high in early January due to high rainfall of 23mm on January 8-9, which followed 16mm rainfall over the period 3-5 January. The water-table then fell rapidly and there was no sustained rainfall until mid March when there were several days of moderate rainfall of around 10mm in magnitude. In early April (from JD 90) several days with rainfall of c. 5mm occurred at regular intervals, however, water-tables fell as water loss through evapotranspiration increased. Three days of high rainfall at the end of May and beginning of June (28mm on JD 150; 18mm on JD 152; and 16mm on JD 156) produced sharp peaks in the water-table which then fell quickly due to a relative drought in mid to late June. Throughout the remainder of the year, the water-table fluctuated about a high level due to the succession of moderate precipitation events at regular intervals. The concentration of heavy rainfall around 23 September (totalling 30mm) corresponds with the time when evapotranspiration diminished, thereby limiting the amount of water-table decline. The month of December is characterised by a sequence of days of moderate rainfall in the range of 10 - 15mm, producing a high monthly rainfall total of 66mm. Under these conditions, the water-table remained high with only small fluctuations between rain events.

The two years, 1991 and 1992, provided markedly different weather conditions which can be seen by comparison of the data summarised in Table 5.2. The most significant difference between the two years was in precipitation, which totalled 497mm in 1991 and 759mm in 1992, as measured by the automatic weather station at Narborough. Examining the longer term precipitation series over the last 20 years, which is summarised in Appendix IV, reveals that 1991 was one of the driest years, while 1992 was one of the wettest years over this medium time-scale. This variation in the annual precipitation totals in 1991 and 1992 provides a good basis to examine the annual pattern of water-table fluctuation at Narborough under extreme weather conditions. There are also differences in the timing of periods of high evapotranspiration loss. Evapotranspiration was generally low in early summer 1991 (JD 140 - 160) but remained high through to JD 250, while in 1992 peak evapotranspiration occurred early in the summer, c. JD 150, and fell more quickly than in the previous year. For all three dipwells, mean water levels were higher in 1992 by between 0.08m and 0.09m and the standard deviation lower. Maximum water levels were similar in the two years for all three dipwells, however, in 1991 the minimum water-table level was lower by 0.08m (dp 2) 0.14m (dp4) and 0.11m (dp6). These differences in minimum water levels between dipwells indicate an increase in draw-down with distance from the river. The total variation in the water-table, as given by the difference between maximum and minimum water-tables, over the period was between 0.73m and 0.77m.

The impact of the varying meteorological conditions is illustrated in

Figure 5.7. where the annual hydroperiod for dipwell no. 0, at the head of the reed-bed, is plotted for 1991 and 1992, on the same graph. A number of differences can be identified between the two years. In the first three months, water levels are lower in 1992, due to a general shortage of precipitation after the rain event on 9th January. In the period from JD 90 to JD 220, conditions are similar as water levels fluctuated about a falling base level in response to precipitation. From JD 220 (7 August) the graphs diverge due to several days of high rainfall in 1992, which were sufficient to maintain water levels at a consistently higher level in 1992 throughout the remainder of the year.

Several general conclusions can be drawn from the these graphs before considering the results in more detail. Water levels are clearly dependent upon total precipitation and rainfall distribution throughout the two year period. This is further illustrated in Figure 5.8., where the difference between water-table height on consecutive days is plotted with precipitation, for the whole of 1991. The daily data were produced by compiling a daily mean position of the water-table from the available record and then subtracting the height for day **t** by the height for day **t-1**, thereby de-trending the variation in water levels, and producing a negative value for days of falling water-tables. In both years, changes in water-tables exhibit an obvious dependence upon the occurrence of precipitation, demonstrating how the distribution of precipitation is important in determining the general form of water-table variation. The only exception was the increase in the water-table which occurred from February 15 1991 (JD 46); this followed a period of 20 days of low and declining water-tables, and was due to water infiltration during the snow-melt event mentioned above. In addition to precipitation, the magnitude of water loss through evapotranspiration is important and contributes a strong seasonal component to the water-table time-series. Differences in evapotranspiration are responsible for the greater draw-down of the water-table over the summer which is especially apparent in the annual record of 1991.

While Figure 5.8. indicates the sensitivity of water levels to precipitation,

Figure 5.7. Comparative water-table heights for dipwell 0 in 1991 and 1992.

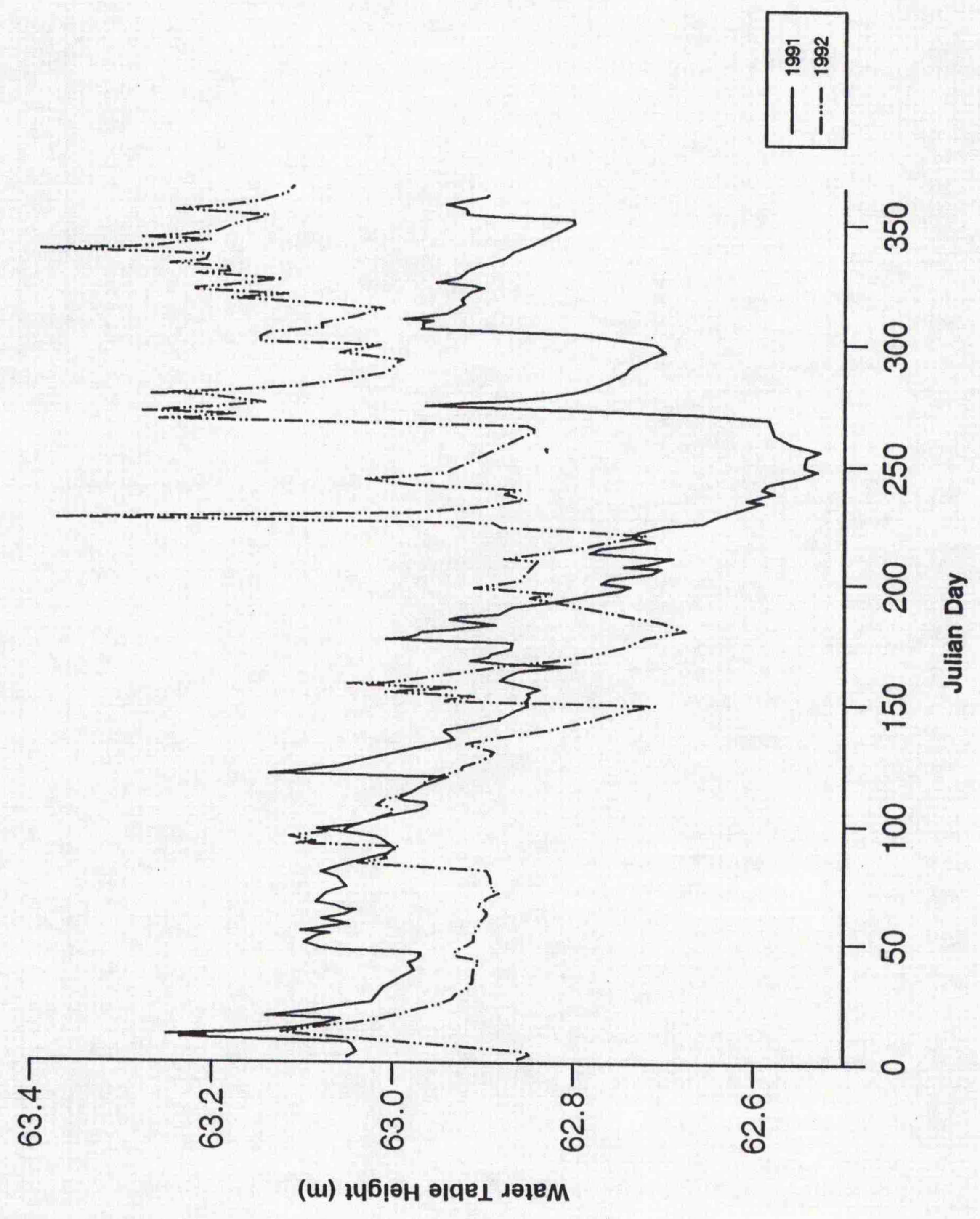


Figure 5.8. The relationship between precipitation and daily water-table change for dipwell 0 in 1991 (top) and 1992.

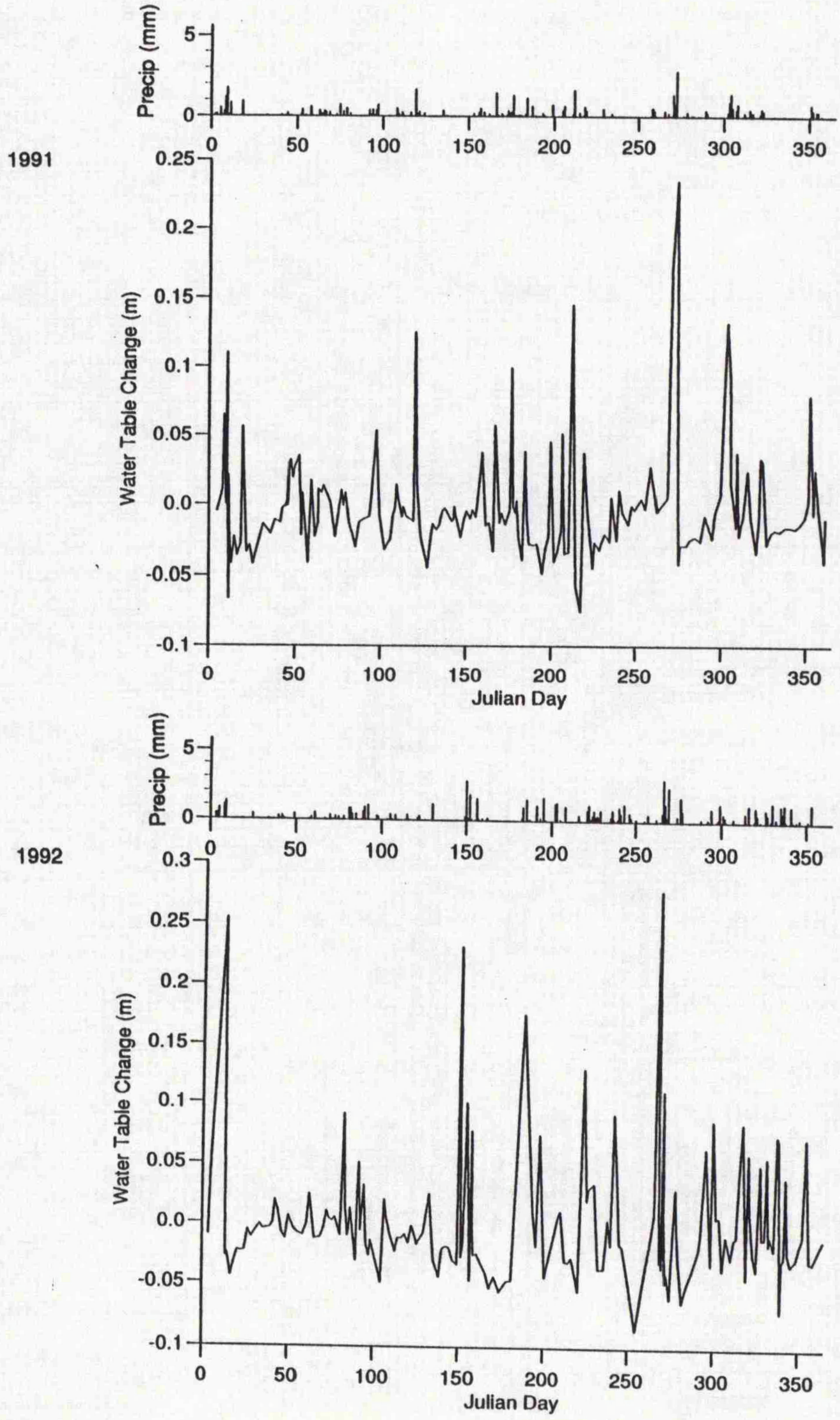
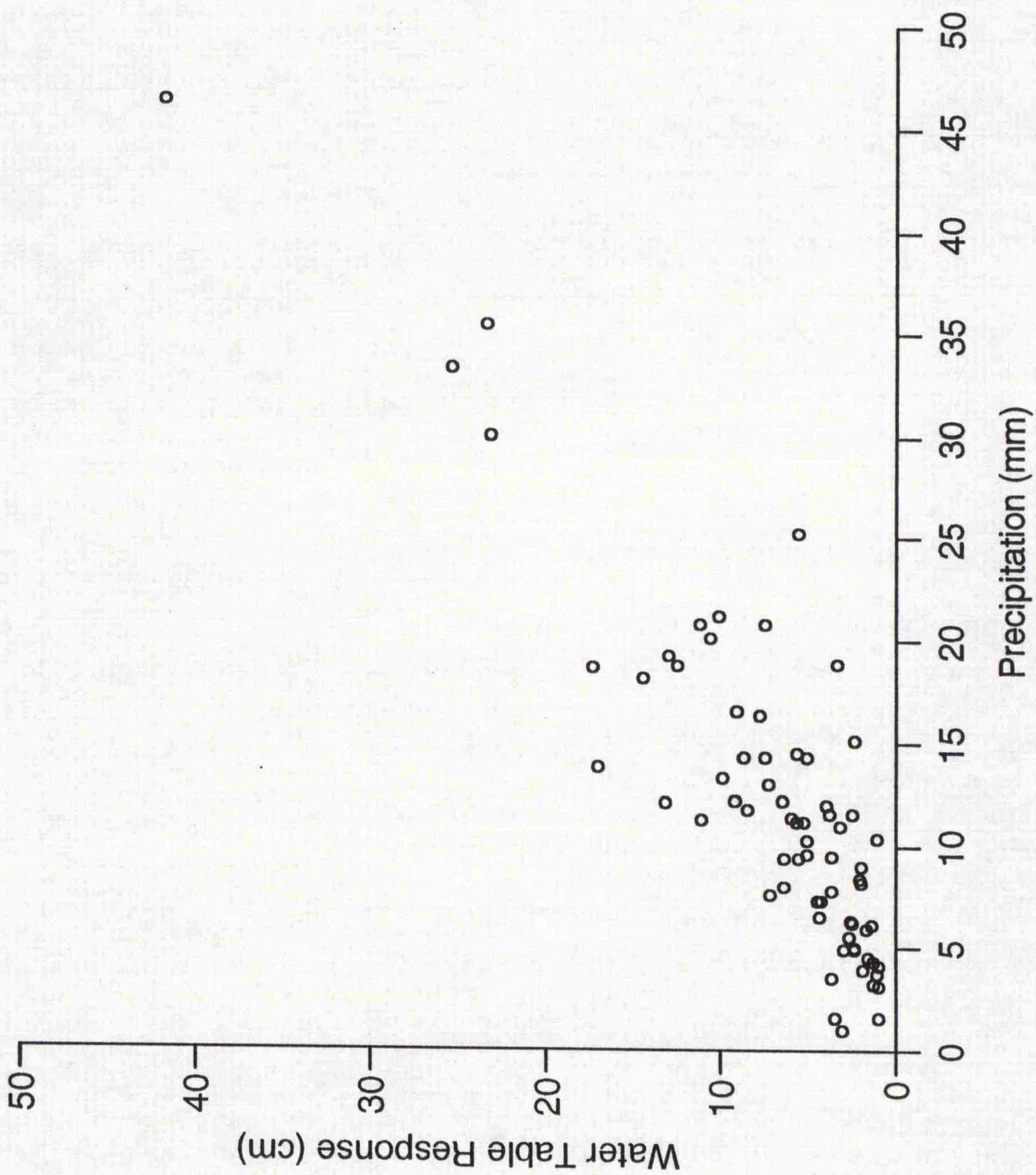


Figure 5.9.	Scatter plot of water-table response and precipitation in 1991 and 1992 for dipwell 0.



it is difficult to identify a definitive relationship between precipitation and water-table response. This is demonstrated further in Figure 5.9. where data of the same format used in Figure 5.8. are shown in a scatter plot of change in water-table against rainfall totals, for all rain events above 5mm in 1991 and 1992. From the graph it is possible to recognise a maximum water level rise for a particular rainfall, however, there is considerable scatter below the line. This reflects differences in preceding moisture conditions, and varying proportions of rainfall interception in response to the duration and intensity of rain events.

Further details of transverse variation in water-tables are given in Figure 5.10.A and 5.10.B where the gradient of the water-table towards the river is shown in profile for days of maximum and minimum levels in 1991 and 1992. The maximum water-table in 1991 occurred on January 10 (JD 10), and in 1992 on December 3 (JD 337), while minimum water levels were September 4 (JD 247) and June 29 (JD 180). At mean water level, the water slope was 0.2m per 100m between dipwells 2 and 4, and 0.5m per 100m between dipwells 4 and 6. Water slope bears a close relationship to surface topography, although the water-table gradient increases in the river marginal area which corresponds with the coverage of sediments having a high silt-clay content. Water-table gradients in the upper area of the reed-bed are similar under both high and low water-table conditions. In both years, the days of maximum water-table occurred during overbank river flooding which explains the observation of water-tables lying above the ground surface beside the river. Heavy rainfall also produced a local rise in the water-table so that standing water occurred in some topographic hollows near the top of the reed-bed, at a distance of 280m from the river.

5.3.1. Relationship of Wetland Water-tables to River Stage.

The observations of water-table variation, which have been discussed above, have concentrated upon an examination of how movements of the water-table are caused by environmental influences, principally precipitation, but also evapotranspiration. The significance of the contribution of river water has not

Figure 5.10. Two-dimensional water profile for days of annual maximum and minimum water levels in 1991 (A) and 1992 (B), with river stage indicated (*).

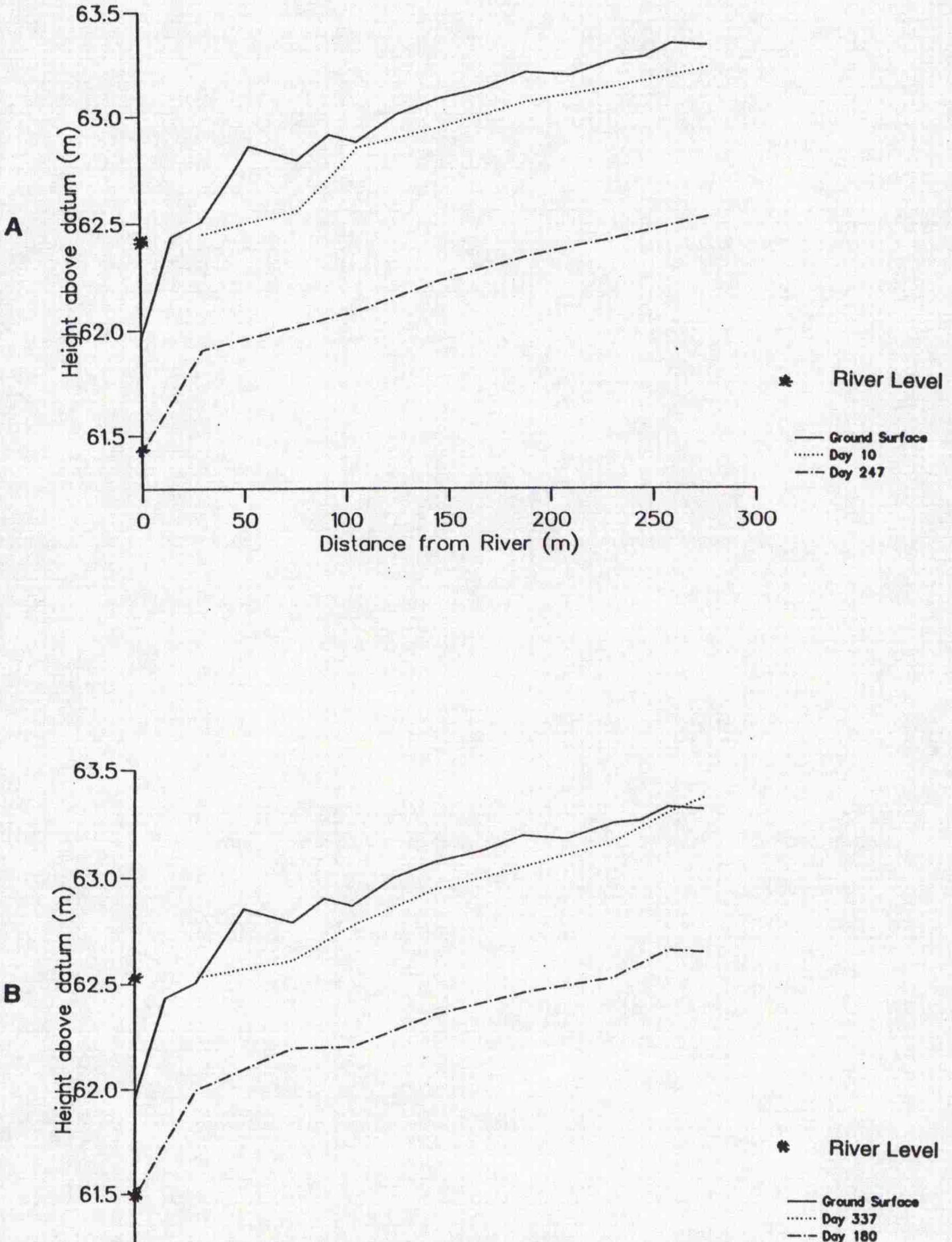
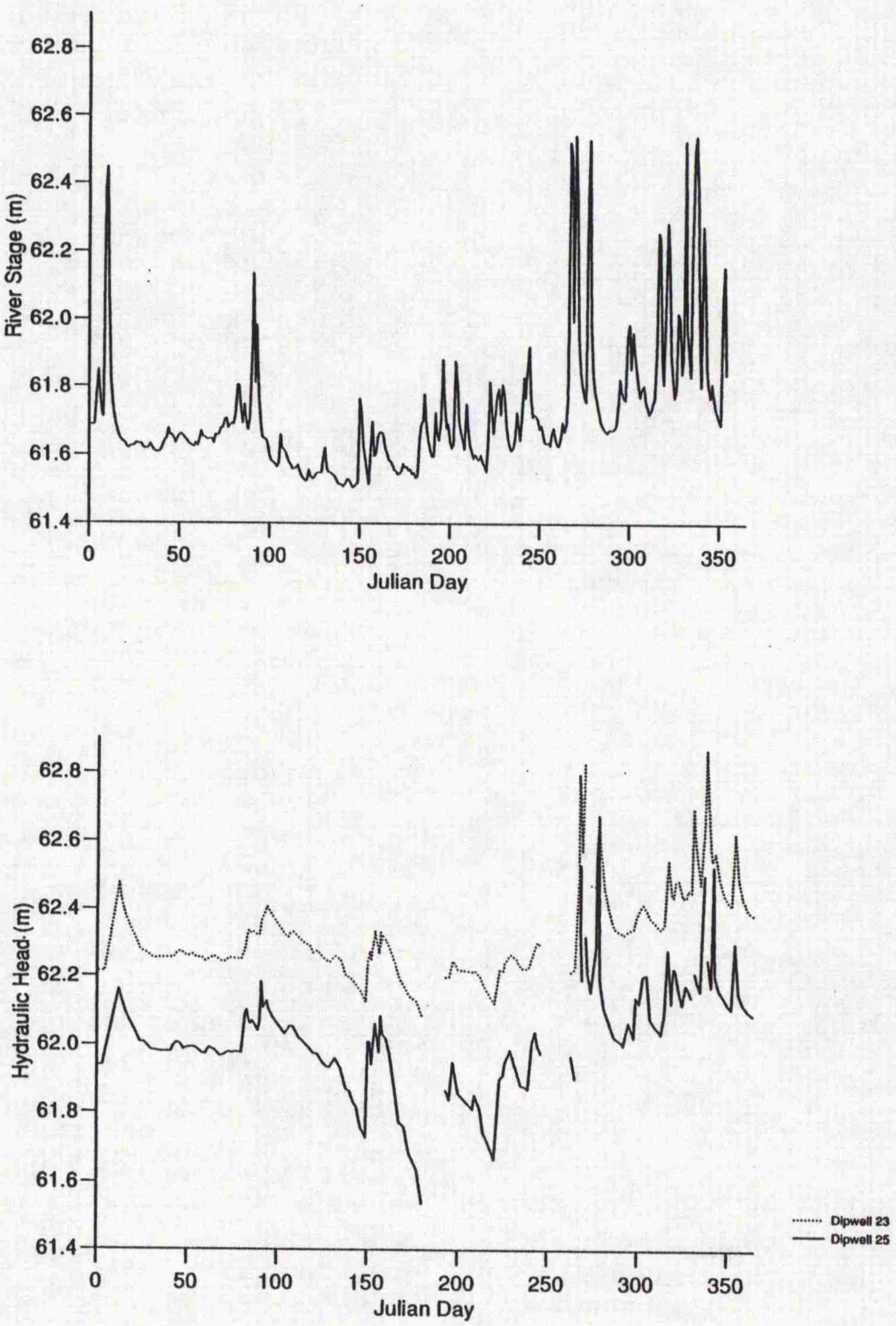


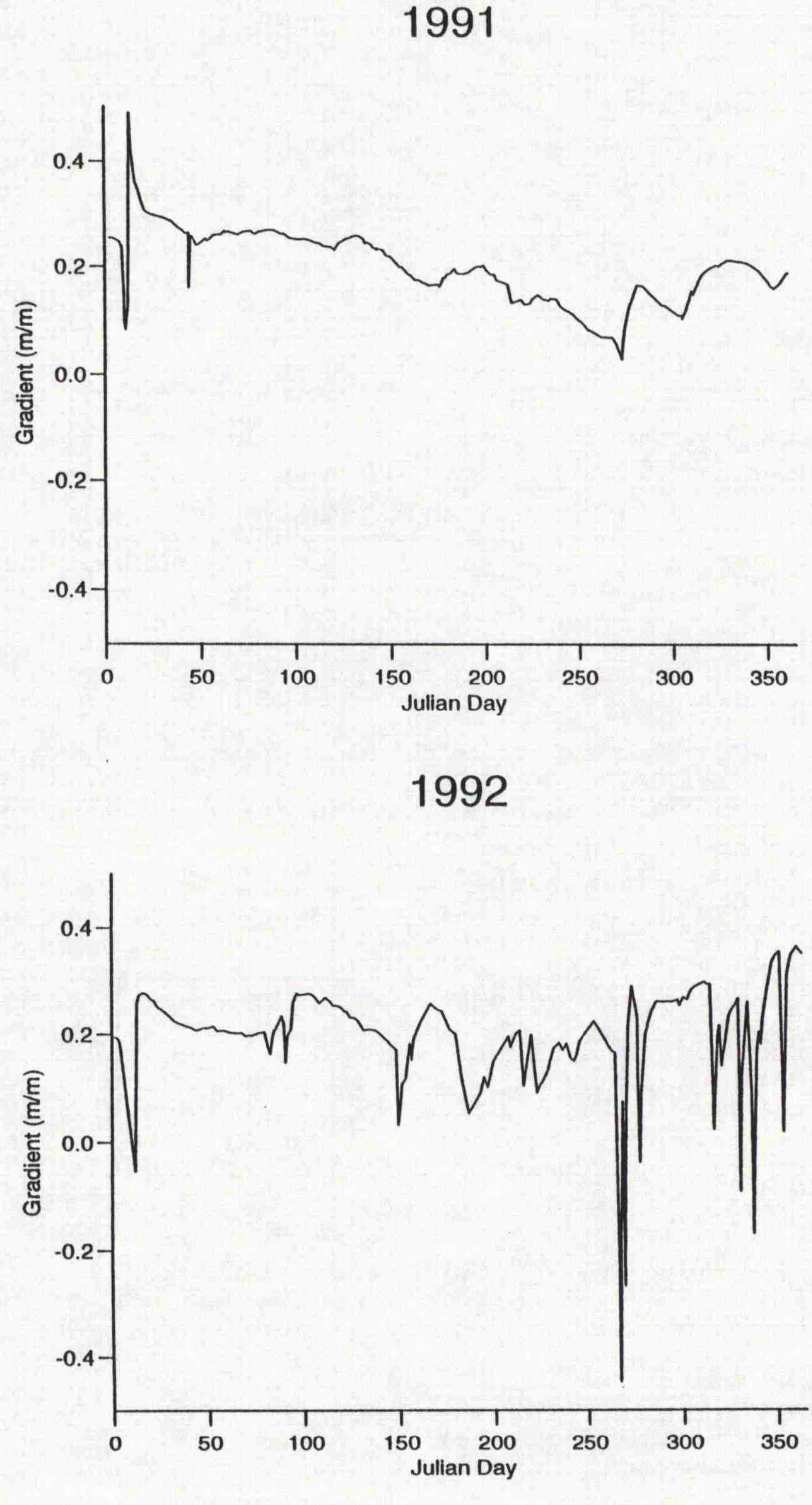
Figure 5.11. Time series plots of river stage, and hydraulic head of dipwells 23 and 25 in 1991. The data are plotted at the same scale and indicate differences in the response to particular rain events.



been considered, although fluvial processes were certainly an important factor influencing the development of Narborough Bog as a floodplain wetland, as was suggested by the stratigraphical evidence discussed in chapter 4. The theoretical contribution of river water to the water budget was discussed in chapter 2, where the importance of river flows in maintaining subsurface influent and effluent flows was described, in addition to overbank flooding.

The relationship between river stage and the water-table recorded in two dipwells at different distances from the river is illustrated in Figure 5.11. Here the upper graph shows river stage in 1992 plotted as height above Ordnance Datum, at the same scale used for the two hydrographs for dipwells 25 and 23. These dipwells are situated progressively further from the river at distances of 35.9m and 60.9m respectively. Identification of a causal relation between the wetland water-table and river stage is difficult as both dipwells and river stage respond to the same precipitation event. Only in large catchments, where river stage increases through precipitation in a separate sub-catchment, would differences in behaviour be expected. Although stage varies substantially, the maximum response of the hydrograph is determined by the floodplain cross-section, presenting an effective upper limit upon response. Comparison of the two graphs reveals how water-tables remain higher than river stage for the majority of the time, however, examining the main periods of high stage reveals some differences in the direction of water drainage. Unfortunately field readings were not collected during the first flood event of the year on 9-10 January; however for the second event, which consisted of two local hydrographs peaks, on 1 and 3 April (JD 91 and JD 93), river stage reached a level intermediate between the hydraulic head of dipwells 23 and 25. The level at dipwell 25 remained below river stage, but recorded a sharp increase, presumably reflecting a combined response to the infiltration of river water and precipitation. In contrast in some of the flood events later in the year, notably 5 October (JD 278), the levels of the two dipwells were identical to river stage, indicating that an equilibrium position of the water-table had been reached.

Figure 5.12. The hydraulic gradient between dipwells 7 and 8 in the two years 1991 (top) and 1992. A reversal of drainage direction occurs during overbank flooding. This is clearly evident in the late 1992.



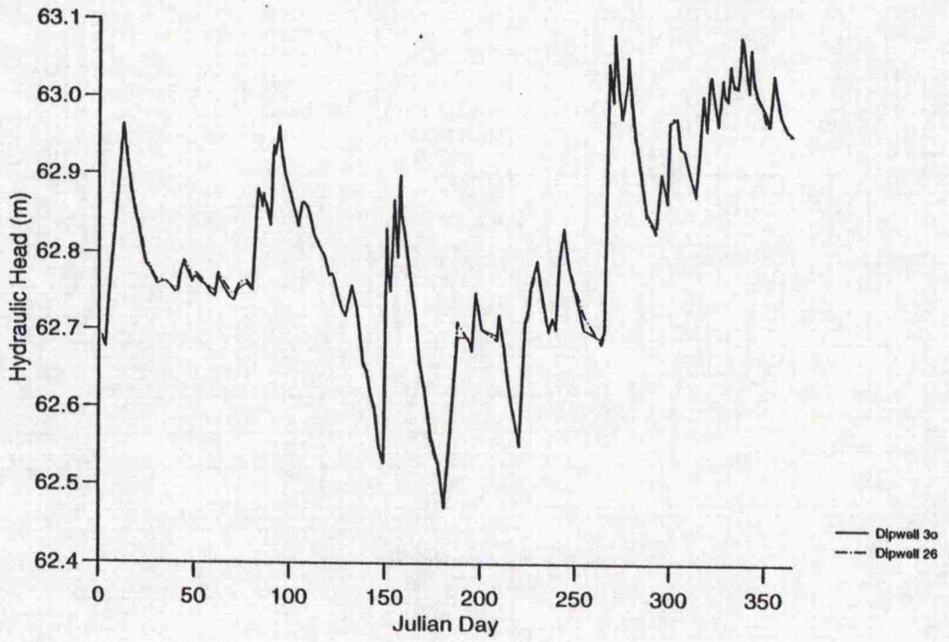
The short periods of time during which the direction of subsurface water flow was reversed are examined in more detail in Figure 5.12., in which the water-table gradient between dipwells 7 and 8 are plotted for both 1991 (top) and 1992 (bottom). The graphs provide further evidence of the extent of the hydrological differences between the two years. While the maximum gradient occurring is similar in both years, at between 0.3 and 0.4 m/m, in 1992 several events produce a reversal in the direction of water drainage. This indicates water flow from the area near the river onto the reed-bed. In 1991 the two major rain events at JD 10 and JD 270 produced large reductions in gradient, but it remained positive, implying water flow from dipwell 7 to 8, throughout the year. In 1992 several events occurred in which the lower area of Narborough Bog was flooded and which produced a reversal in the water-table gradient. These occurred on 12 January (JD 12), 24 and 27 September (JD 267 and 270), 5 October (JD 278), 27 November and 4 December (JD 331 and 338).

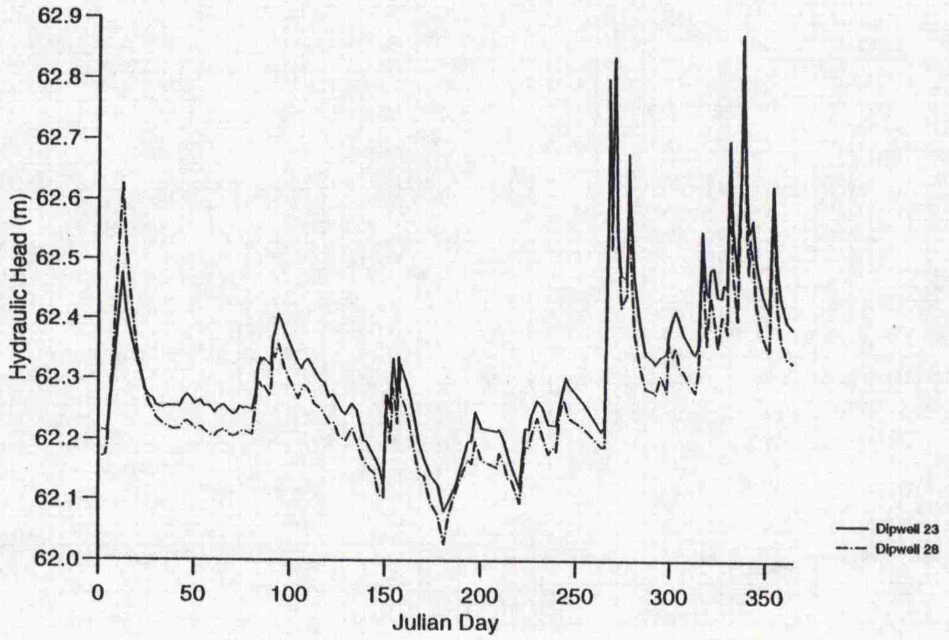
5.3.2. Spatial variation in vertical water flux.

In chapter 4 it was mentioned that at selected sites dipwells were installed at different depths, so that readings of water-table depth would provide an indication of the variation of hydraulic head with depth. Several shallow dipwells were installed at sites adjacent to selected dipwells, which extended to a depth of 50 cm below the ground surface, with holes drilled over the bottom 15cm. They therefore provide an indication of hydraulic head integrated over a depth of 35-50 cm, in contrast with the majority of dipwells which were approximately 1.80m long, and recorded hydraulic head immediately above the gravel layer. Some spatial differences were observed in the response of the dipwell nests as indicated in Figure 5.13., where hydrographs are plotted for dipwells 3o and 26 (A) and dipwells 23 and 28 (B) for 1992.

Figure 5.13.A indicates that there is little difference between the head observed at dipwells 3o and 26, suggesting no substantial vertical water flux, and hence the maintenance of equilibrium conditions through the profile. In

Figure 5.13. Variations in vertical water flux, at dipwells 3o and 26 (top), 23 and 28 (bottom) in 1992. The difference in head of dipwells 23 and 28 indicates a predominantly upward water flux.





contrast, Figure 5.13.B, showing the hydrographs for dipwells 23 and 28 records consistent differences in hydraulic head between the two soil depths. At this site the hydraulic head of the shallow dipwell no. 28 is generally 40 cm below the level of dipwell 23. This indicates the maintenance of an upward water flux for much of the time, due to a significant hydraulic gradient between the two depths. The direction of subsurface drainage was reversed in the first major flood event on 9-10 January, indicating the dominant effect of downward drainage after a large precipitation event of 23mm rainfall over two days. However, in the rain events towards the end of the year the hydraulic head of the two dipwells reached an equilibrium position.

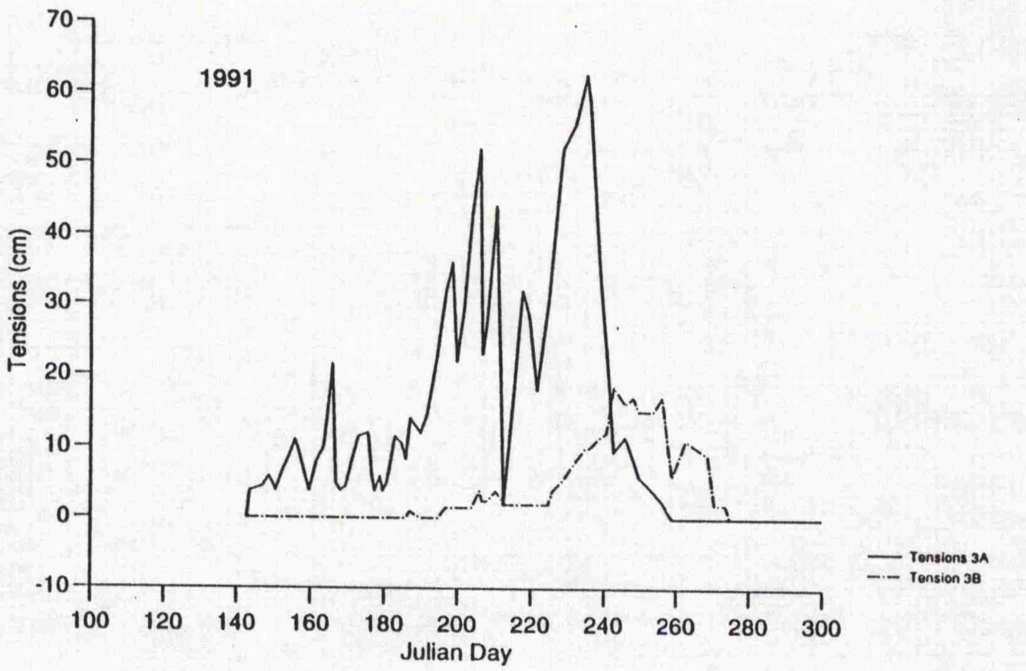
These spatial differences in hydraulic head almost certainly arise as a consequence of variations in local stratigraphy, and particularly the deposition of a clay layer near the river, which extends to the site of dipwells 23 and 28. The clay deposit acts as a semi-confining unit, which is sufficient to produce a hydraulic head difference of 40 cm in response to the direction of local subsurface water drainage. Within the reed-bed, variations of vertical hydraulic conductivity within the peat deposits are too small to produce a measurable difference in hydraulic head, although some minor differences in the response of dipwells 3o and 26 were observed.

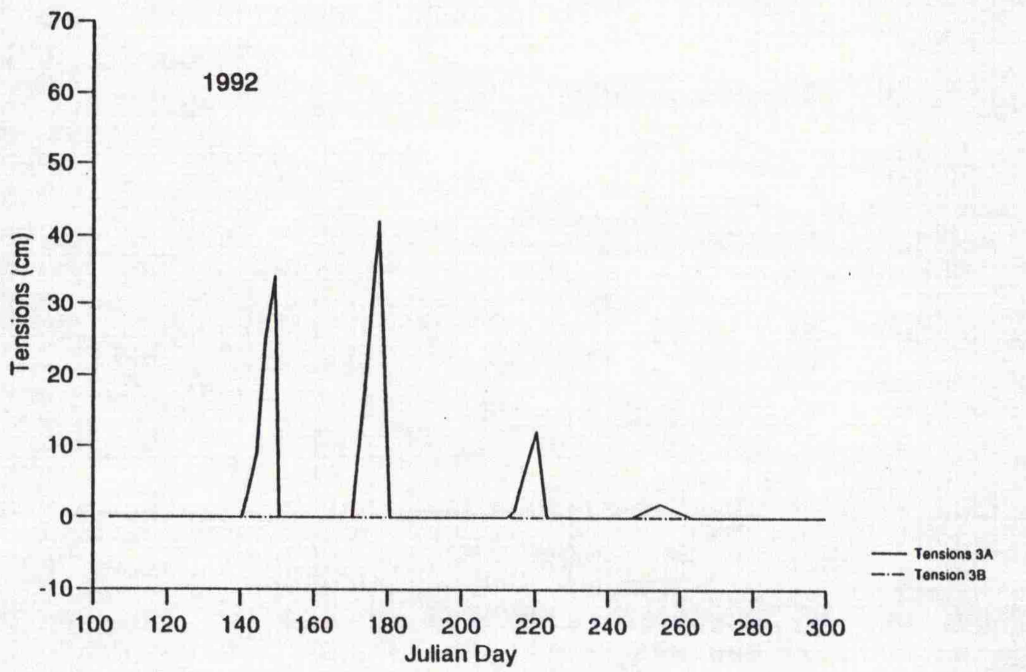
5.3.3. Variation in Unsaturated Water Content.

All the results introduced in the sections above have considered temporal variations in the water-table observed in the dipwell network. Consequently, water fluxes through the unsaturated zone have been neglected; this partly reflects the difficulty of obtaining measurements in the unsaturated zone in a peat soil. However, some results were obtained from the tensiometer nests installed at three locations in the reed-bed (shown on Figure 4.7.).

The results for one tensiometer nest are given in Figure 5.14. for 1991 and 1992, indicating differences between the two years which parallel the variation in the water-table results for the two years discussed above. The graphs record

Figure 5.14. Soil water tensions recorded at points 3A and 3B, 1991 and 1992.





tensions at two depths, 30cm and 60cm below the surface (tensiometers 3A and 3B respectively). In 1991 tensions near the surface varied considerably over a range of 62cm to 0cm soil suction, as a result of low summer precipitation and a consequent depression of the water-table. The fall of the water-table made soil tensions extremely sensitive to atmospheric changes, as tensions decreased following precipitation and increased with evapotranspiration. At a depth of 60cm, soil tension only increased slightly in the early summer, for example up to 2cm on JD 190, however, tensions increased considerably in late summer, as the water-table reached a minimum position. Tensions increased to a level between 10 and 20 cm soil suction, although decreasing near the surface. This behaviour implies the isolation of an unsaturated area between the water-table and percolation of rainwater through the peat matrix. Elevation of the water-table occurs therefore through preferential routing of water, while localised low permeability clay deposits produce some hydrologically isolated areas.

In 1992 different meteorological conditions produced a significantly different temporal pattern of soil tensions. At a depth of 60cm no change in tensions was observed, and a reading of 0cm was recorded throughout the summer. At 30cm depth, tensions increased sharply for short periods of time of high evapotranspiration, reaching peaks of 34cm, 42cm, and 12cm; however the duration of these events was too short for the maintenance of high soil water tensions in the intermediate period.

5.4. REGRESSION MODELLING OF Water-table FLUCTUATIONS.

The discussion above demonstrates the general dependence of water-table levels upon the annual precipitation distribution. Additional processes contributing to fluctuations in the water-table were considered in chapter 2; these may include groundwater recharge, near stream bank storage, tidal effects, atmospheric pressure, external loading, pumpage, agricultural irrigation and drainage. It is very difficult to isolate the effects of individual processes, and therefore, regression modelling will be used to consider the extent to which water-table position may be predicted on the basis of the dominant variables, precipitation and evapotranspiration alone. This should clarify the importance of these factors and indicate whether the operation of other processes has to be invoked to provide further explanation. The data set used comprises average daily values for water-table position for two dipwells in 1991 and 1992. Dipwell no. 6 is located towards the centre of the reed-bed, and no. 8 close to the river in an area partially covered by woodland. The data were obtained by taking a running average from the regular measurements of water-table position which were generally taken at an interval of three to four days during stable conditions, and every one to two days at times of rain events or high evapotranspiration. The data were therefore produced by a simple process of linear interpolation.

Time series methods have previously been applied in the analysis of near-surface groundwater levels, notably by Rennolls *et al.* (1980) and Viswanathan (1984). Box-Jenkins models, predicting linear response systems, have also been described by Armstrong (1988), while Youngs *et al.* (1989) discuss a more applied approach to water-table modelling by looking at the development of water-table profiles between a ditch network in response to precipitation and evapotranspiration. One advantage presented by time-series methods is that detailed characterisation of aquifer properties is not required initially, and the results should indicate which hydraulic properties have greatest significance. This requires detailed examination of the modelling results to determine the reasons for any deviation in the fit of particular models.

A simple regression model of water-table change and precipitation gives a poor explanation of 7% for the Narborough data set, as this takes no account of the importance of soil moisture conditions and depth to the water-table (cf. Figure 5.9.). These factors will determine the proportion of precipitation that is taken up within the unsaturated zone and which therefore does not contribute to an increase in water-table height. Furthermore, precipitation alone is a poor indicator of the rate of fall of the water-table which depends upon evapotranspiration, which is itself significantly higher when the water-table lies close to the surface. Interception storage may also be significant at certain times of the year, and is likely to vary spatially.

Rennolls *et al.* (1980) outlined a simple autoregressive model relating water-table position to water level on the preceding day and precipitation:

$$h_i = a_1 h_{i-1} + a_2 R_i + e \qquad (1)$$

where h_i is the water level on day i, R_i is rainfall on day i, and e is random error. Rennolls *et al.* (1980) give some explanation of what they consider the two constants in the regression equation to represent. The constant a_1 they term the proportional decrease in the water-table per day, and a_2 a measure of the drainable pore space of the soil (i.e. the pore space in the unsaturated zone which is filled with infiltration of precipitation producing an increase in the water-table). Taking this model as a starting point, regressing water-table height against daily rainfall and preceding water-table position produced the range in model equations given in Table 5.3. for dipwells 6 and 8 in 1991 and 1992.

The considerable difference in the fit of the model over the two years raises several questions which need to be considered. In 1991 the approach produces a good degree of fit for dipwell 6, indicating that water-table levels are highly autocorrelated. The degree of explanation is less for dipwell 8, where the rate of recession of the water-table, indicated by coefficient a_1, is less rapid. This is most probably due to different stratigraphy, and a higher silt-clay concentration around dipwell 8. In 1992 the different equations indicate a non-stationary response in the water-table, especially for dipwell 6. Examining the days on which the residuals from the model are unusually high (i.e. above a z score of 4.0) indicates a rapid increase in water-tables at this point through overbank flood events: the greatest outlier of z=7.20 occurred on 27 September (JD 270) due to a large flood event. The model is similarly unable to account for the rapid rate of water-table recession following these events. The different

Table 5.3. Results for simple autoregressive model applied in 1991 and 1992.

Dipwell, Year	Equation	R^2
Dipwell 6, 1991	$h_i = 0.995\ h_{i-1} + 0.0011\ R_i$	97.4%
Dipwell 8, 1991	$h_i = 0.998\ h_{i-1} + 0.0015\ R_i$	93.7%
Dipwell 6, 1992	$h_i = 0.995\ h_{i-1} + 0.0029\ R_i$	94.7%
Dipwell 8, 1992	$h_i = 0.998\ h_{i-1} + 0.0020\ R_i$	92.5%

fit of the model can therefore be explained partially by the differing significance of water inflow from overbank flooding, and also time of year.

An improved fit to the model is possible if precipitation (R_i) is substituted by a parameter representing lagged effective precipitation (essentially precipitation minus evapotranspiration) (W_i). However, the increase in explained variation is relatively minor for the whole data-set, increasing by 0.5% to a total of 97.8%.

The simple model described above has been criticised for failing to include any measure of soil moisture variation through time, apart from the autoregressive water-table component, as only current daily rainfall is used. A possible solution was outlined by Viswanathan (1983) who introduced an auto-regressive rainfall model which covers rainfall on the previous seven days:

$$h_i = a_i h_{i-1} + a_2 R_i + a_3 R_{i-1} \cdots + a_9 R_{i-7} + a_{10} + e_i \qquad (2)$$

where $\mathbf{a_x}$ are best-fit coefficients, $\mathbf{h_i}$ is water-table position at day **i**, and $\mathbf{R_i}$ is precipitation at day **i**.

The model thus implies that the water-table on any day is determined by rainfall on the seven preceding days, in addition to water level on the previous day. The parameters $\mathbf{a_1}$ to $\mathbf{a_8}$ were also assumed to be time dependent

Table 5.4. Results of the Autoregressive precipitation models

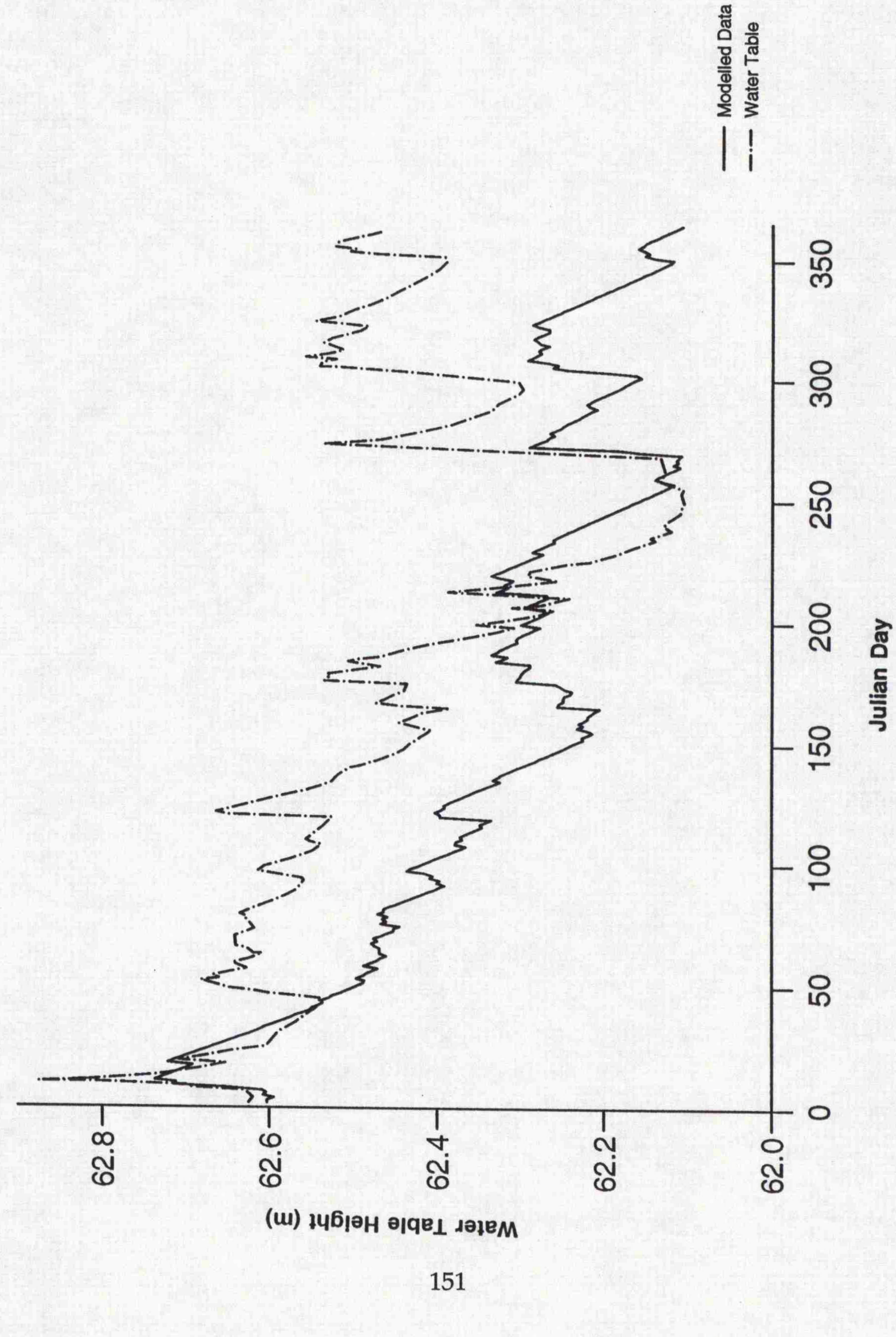
Dipwell, Year	Model Equation	R^2
Dp 6, 1991	$h_i = 0.92 + 0.99\ h_{i-1} + 0.959\ R_i + 3.93\ R_{i-1} + 0.531\ R_{i-2} - 0.141\ R_{i-3} - 0.565\ R_{i-4} - 1.13\ R_{i-5} + 0.621\ R_{i-6} - 1.17\ R_{i-7}$	98%
Dp 8, 1991	$h_i = 2.02 + 0.97\ h_{i-1} - 0.367\ R_i + 1.97\ R_{i-1} + 1.23\ R_{i-2} + 0.708\ R_{i-3} + 0.852\ R_{i-4} - 1.86\ R_{i-5} - 0.286\ R_{i-6} + 0.225\ R_{i-7}$	94.6%
Dp 6, 1992	$h_i = 1.98 + 0.97\ h_{i-1} + 1.64\ R_i + 2.68\ R_{i-1} + 1.66\ R_{i-2} - 1.13\ R_{i-3} - 0.25\ R_{i-4} + 0.85\ R_{i-5} - 0.304\ R_{i-6} - 0.752\ R_{i-7}$	96.1%
Dp 8, 1992	$h_i = 9.02 + 0.86\ h_{i-1} + 2.27\ R_i + 3.52\ R_{i-1} + 3.84\ R_{i-2} - 2.39\ R_{i-3} - 0.394\ R_{i-4} + 1.99\ R_{i-5} - 1.06\ R_{i-6} - 0.889\ R_{i-7}$	81.3%

by exponentially weighting past observations which enables the identification of any seasonal differences in the importance of individual parameters. For example, it is possible that seasonal interception storage by vegetation may be responsible for a variation in the time of water-table response following precipitation, and hence temporal changes in the parameters $\mathbf{a_1}$ to $\mathbf{a_8}$. For the Narborough data-set the method was first considered by taking the parameters to be time-invariant and applying a simple multiple regression. The results are summarised in Table 5.4. (with precipitation and water-table height both in metres).

Examining the coefficients of the four equations given in the table it is clear that the dominant statistical term remains the water-table position on the previous day with precipitation on the preceding day and on the current day, the only significant additional influences on water-table location. It seems that most recharge to the water-table occurred in the preceding 48 hours, and also that after this period the water-table fell. Using the file incorporating evapotranspiration loss in addition to rainfall produced no increase in explanation, R^2 remaining 98.2%.

All the above equations incorporate an autoregressive term, whereby the water-table position on any day is related to the level on the preceding day; this

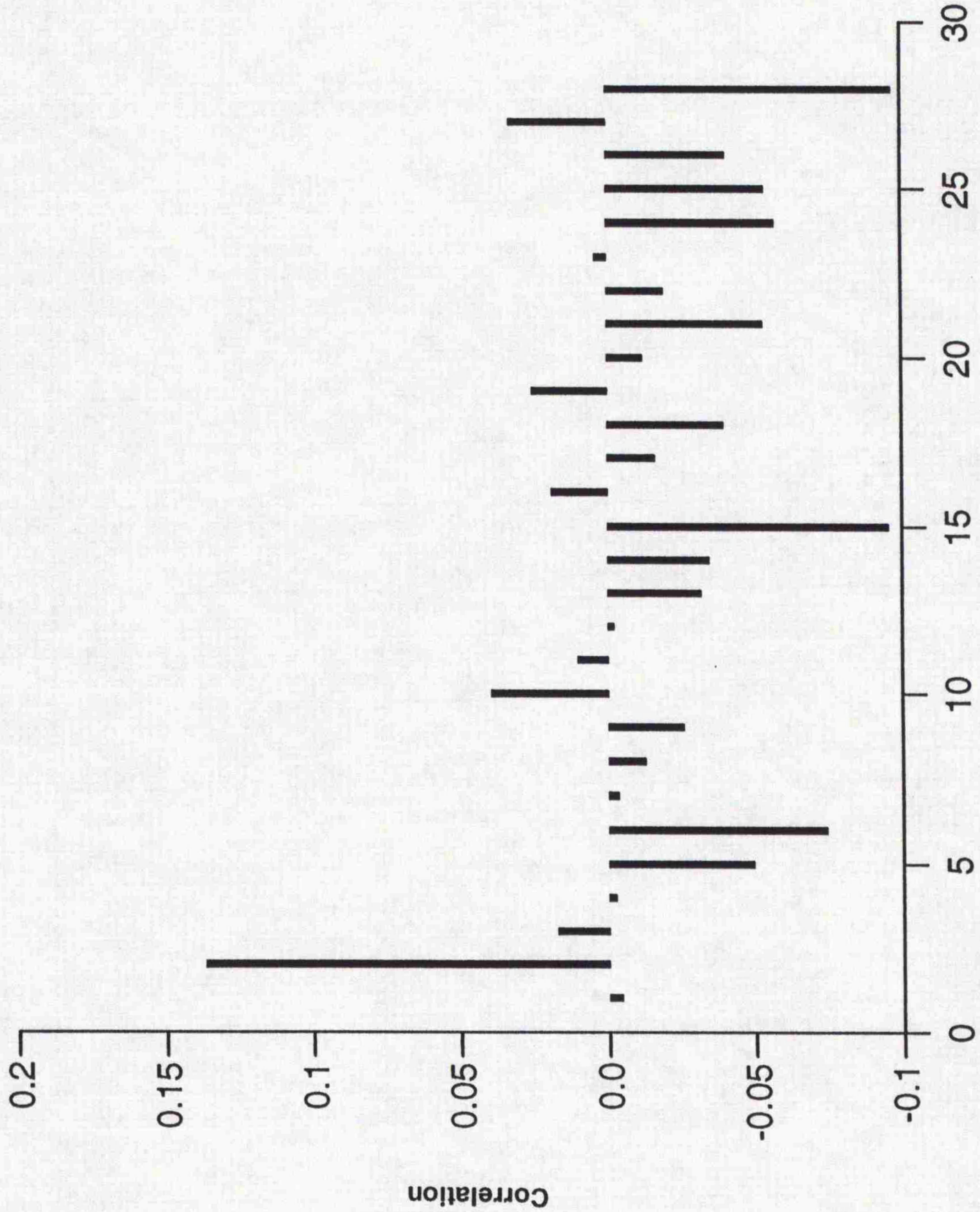
Figure 5.15. Comparison of observed water-table variation with results from the autoregressive model, ($h_i = a\, h_{i-1} + b\, W_{i-1}$) for dipwell no. 6 in 1991.



limits the usefulness of the model in forecasting water-table levels for a long time period where only the initial position of the water-table is known. This is illustrated in Figure 5.15. for the example of dipwell no. 6 in 1991 where a modification of equation 1 ($h_i = a\ h_{i-1} + b\ W_{i-1}$) is used. Here W_{i-1} represents the effective precipitation on the previous day (precipitation minus evapotranspiration). The constants a and b were determined by simple regression; the model was applied using predicted values of h_i to input into the equation to give the next value of h. While the response of the two data series to rain events is similar, there is a rapid divergence centred on JD 40-50 which corresponds to the delayed snow-melt induced increase in water-table height. The modelled results were unable to identify this, possibly due to the difference in timing between the melting of snow in the rain gauge, and the time of infiltrating water reaching the water-table. As soon as water-table heights begin to diverge the effects are accentuated at later time intervals, although the two data-sets converge in late July (after JD 200) as evapotranspiration produces a water-table fall. The limitations of the approach are that the coefficients describing water-table drawdown and precipitation response are annual means and thus cannot represent seasonal variation. Improvement of the model to include consideration of any snow-melt component represents a considerable problem which has been widely noted, for example by Lammala and Tattari (1988).

The above equations were developed in a deterministic manner based upon a certain understanding of water-table - rainfall relationships. Box and Jenkins (1970) describe procedures for fitting time series models to continuous data series, which are appropriate where prior knowledge of these relationships cannot be assumed. The autocorrelation function enables identification of the most appropriate stochastic model. The two main types of model are the autoregressive and moving average models. In autoregressive models the value of the time series is dependent on the prior observed value plus a random shock. Moving average series are dependent only on present and past shocks to the system.

Figure 5.16. The autocorrelation function for dipwell 6.



The autocorrelation function of the water-table time series has a smooth exponential form which decreases slowly towards zero, indicating some autocorrelation. However, the autocorrelation of the first-differenced series, shown in Figure 5.16., has only one significant correlation at lag 2 (0.14). Taking the significant value for cutoff as $\pm 2/\sqrt{n}$ for n=360 any value over 0.106 is significant at the 95% level. The value is thus only just significant and there is very little correlation.

The application of a simple autoregressive model to the water-table variation gives:

$$h_i = 1.01 + 0.986\,h_{i-1} \qquad (3)$$

where the degree of explanation is 97.5% (sum of squares of the residuals: 358.86).

Fitting a second order autoregressive model gives no increased explanation to the data set (R^2 = 97.5%, SSQ: 361.41):

$$h_i = 1.06 + 0.993h_{i-1} - 0.0073h_{i-2} \qquad (4)$$

The possibility of applying an autoregressive moving average (ARIMA) model to the water-table time series was investigated. The data have a strong autoregressive component, and it was thought that a seasonal component could be introduced to simulate the evapotranspiration cycle. However, although the application of several ARIMA models was attempted, they will not be discussed here because the method is not suitable for forecasting. If prediction of the data series is required it would be necessary to fit an ARIMA type model to an independent data set, representing water deficit (i.e. Precipitation minus Evapotranspiration), and then to examine the cross correlation between the residuals of this model and the same ARIMA model applied to the dependent data set, namely the water-table position. A satisfactory ARIMA model could not be found for the data set and so a forecast of water-table position was made upon the value of the water deficit and water-table position on the previous day.

5.5. TRANSFER FUNCTION SIMULATIONS OF RAIN EVENTS.

While the autoregressive models described above have a good statistical fit with the data-set, this reflects the dependence of water-table position upon the position of the water level on the preceding day of measurement. Autoregression models indicate the generalised pattern of water-table response over an annual cycle, and the autoregressive component simply provides a measure of the mean proportional decrease in water-table which arises through seepage to the river and evapotranspiration. Also important is the temporal change in individual processes which could not be considered satisfactorily using time-dependent coefficients, as discussed above. Consequently the possibility of deriving an impulse response function to describe aquifer response to a unit precipitation event was investigated. A linear system model can be used to examine the relationship between individual input and output data series and hence infer properties of the system itself. The use of this technique has been advocated by Bruen and Dooge (1984) to derive unit hydrographs, and variations in water-tables have also been analysed by Armstrong (1988) using the method. Furbish (1991) isolated the component of water-table variation arising through changes in atmospheric pressure using a response function approach. The impulse response function describes the response of a system to a perturbation of unit size, and as such may be taken as an indication of the efficiency with which the wetland system at Narborough removes water. This would be expected to vary within an annual cycle, depending upon the operation of certain seasonal hydrological processes. For example, one limitation of the approach is that the possibility of water loss through evapotranspiration is not explicitly modelled. The quantity of evapotranspiration should be apparent from seasonal differences in the degree of fit of the model, particularly at times of high evaporation. A measure of evapotranspiration will therefore be provided from the residuals of the model. This method of determining evapotranspiration is open to the criticisms discussed in chapter 3, however, the data are useful in enabling comparison with the estimates of potential evapotranspiration derived from the Penman equation.

The technique involves an analysis of a simple, isolated, rain event for which an impulse response function is derived. Initially the water-table is in an equilibrium state at the beginning of the event, and remains stationary after rainfall commences and infiltration occurs. The water-table then rises and subsequently falls in response to the rain event. The relationship between rainfall and the water-table is described by an equation of the form:

$$H(T) = \int_{0}^{t_m} R(\tau) P(T-\tau) \, d\tau \qquad (5)$$

where H(T) is the water-table height, P(T) is precipitation, at time T, and R(τ) is the response function. These measurements of water-table and precipitation form discrete time series in that they represent daily summaries. The relationship forms a stochastic process as it will change through time in a probabilistic manner as a result of varying conditions of soil moisture, or quantities of water intercepted by vegetation. The data usually consist of a sequence of values so that the discrete equivalent of this equation is used:

$$h(t) = \sum_{k=1}^{m} r(k) p(t-k+1) \qquad (6)$$

Programming sub-routines to determine the form of the response function (R(τ)) for particular rain events have been published by Bruen and Dooge (1984) based on Farden (1976). The program considers the transformation of effective rainfall to water-table change, and adopts a smoothing process which is dependent upon a parameter RI, termed the regularisation parameter. The response function is estimated by a least-squares process from an input and output data series. Taking an individual rain event the response function is convoluted with the rainfall data to assess the degree of fit of the model and to optimise the parameter RI.

The impulse response function obtained is then used to simulate the annual cycle of water-table variations. Application of an impulse response function in this way depends upon an assumption that the relationship between precipitation and water-table response is time invariant. Thus the results indicate the fluctuations in the water-table which would be expected if the hydrological conditions during the one event were applicable for the entire period. Here, the assumption will not be valid, as the lag time before the water-table responds to rainfall is partly determined by interception storage and evapotranspiration loss will vary. The success of the model therefore indicates the significance of these two processes, which both vary seasonally and are difficult to separate, as interception is related to vegetation development as well as the characteristics of individual rain events.

The procedure was investigated for an isolated rain event on 29 April 1991 when 16.7 mm of rain fell, with 2.0 mm falling the following day. In the preceding week, there was one day of rainfall, 2.4 mm on 26 April. The characteristics of the event can be seen in Figure 5.17.; precipitation is shown in 5.17.A, the water-table response to the event is given in 5.17.B for dipwell 2, and the impulse response function obtained is in 5.17.C. Dipwell 2 was selected in this case because it exhibited the greatest response to rain events of dipwells 2, 4 and 6 (cf. Figure 5.6.). The modelled response, obtained by convoluting the impulse response function with rainfall, is also given in Figure 5.17.B. The modelled response has a very good fit with the actual data; which is to be expected as the event is used for calibration purposes, of more interest are the graphs which give the modelled response for the annual cycles in 1991 and 1992, shown in Figure 5.18.A and 5.18.B. As explained above, these were obtained by multiplying the precipitation series for both years by the coefficients from the response function.

Figure 5.17. Linear Response Function model for rain event on 26 April, 1991. A, B, C at dipwell 2.

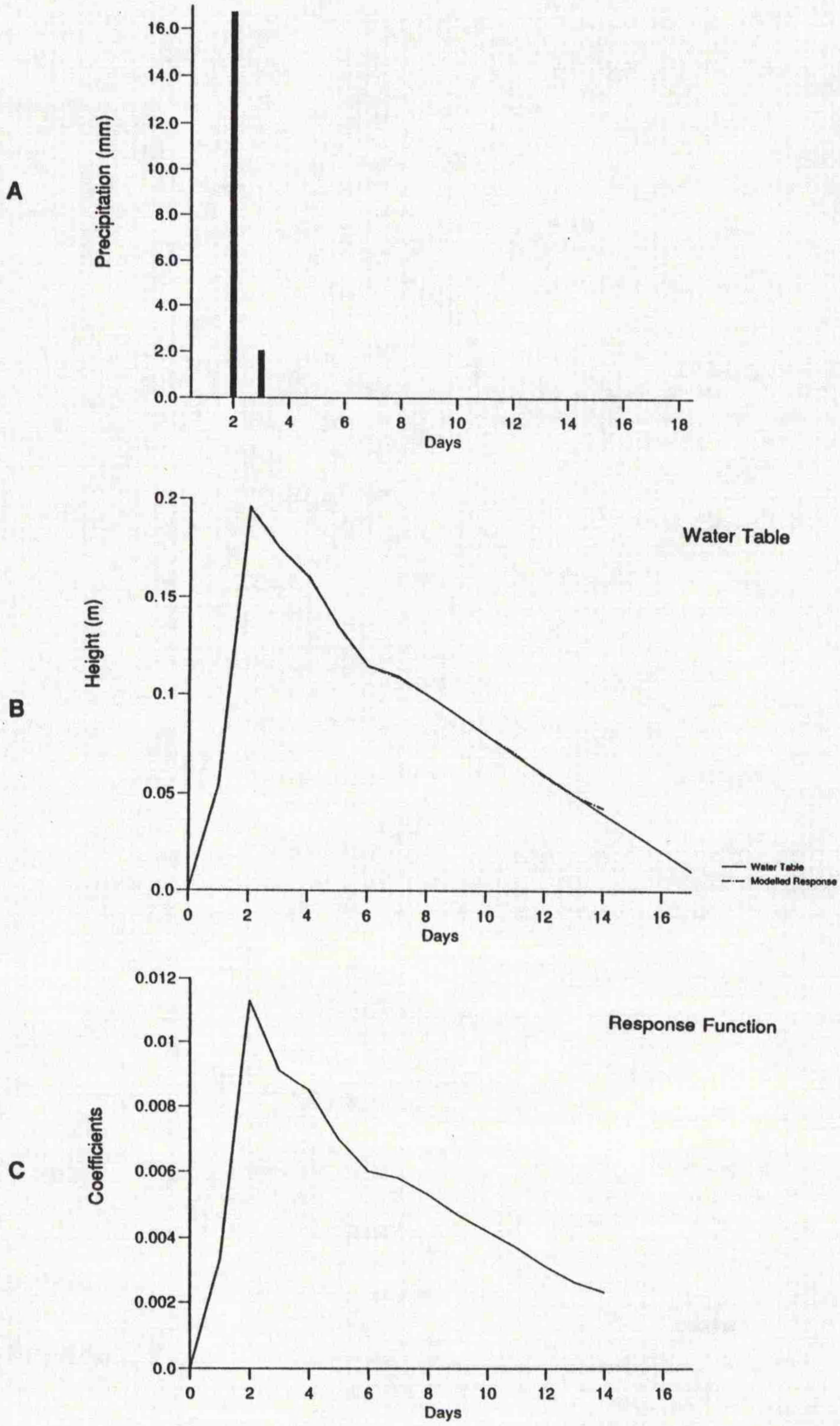
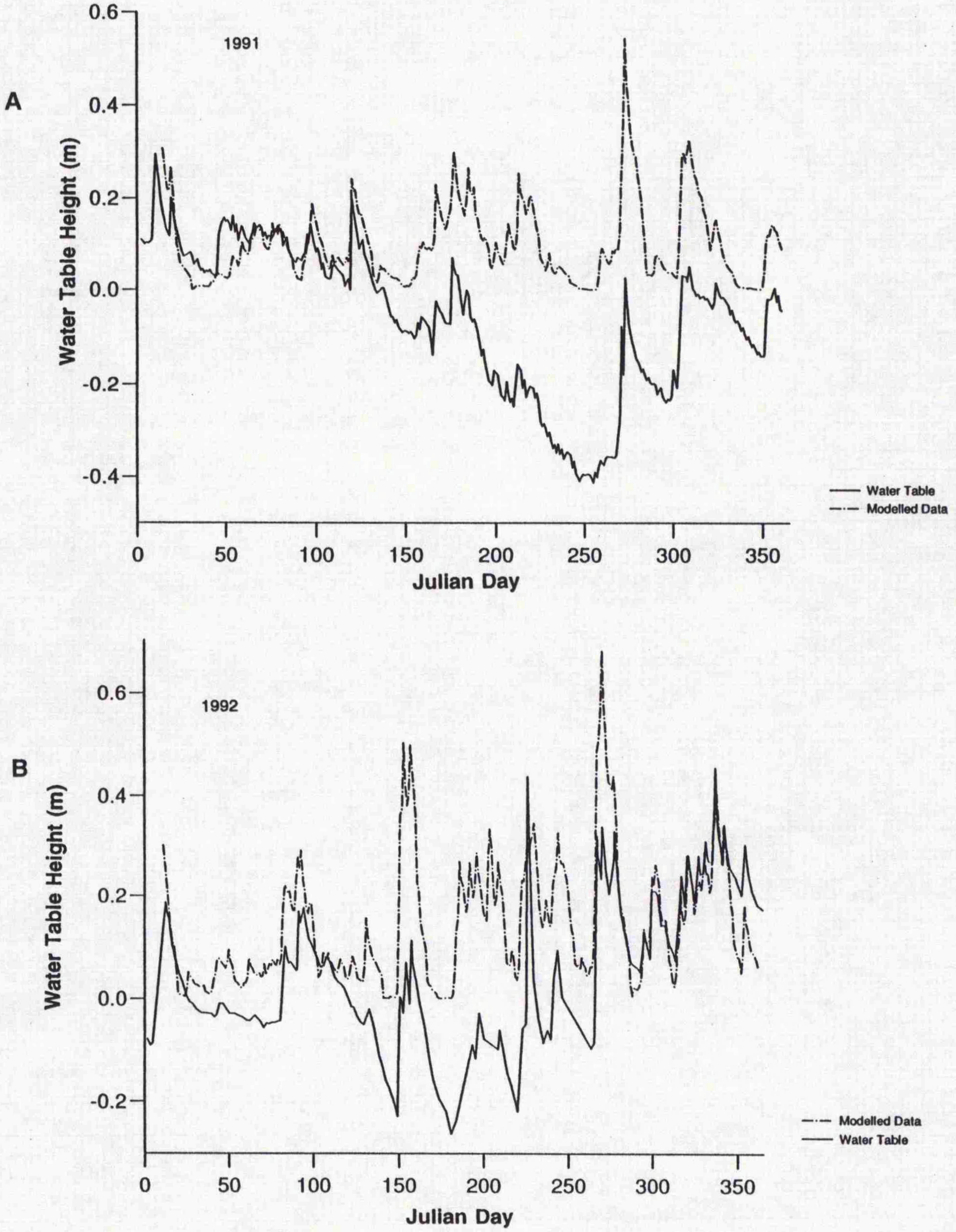


Figure 5.18. Application of the impulse response function for 26 April 1991 for the entire data-set in 1991 (A) and 1992 (B).

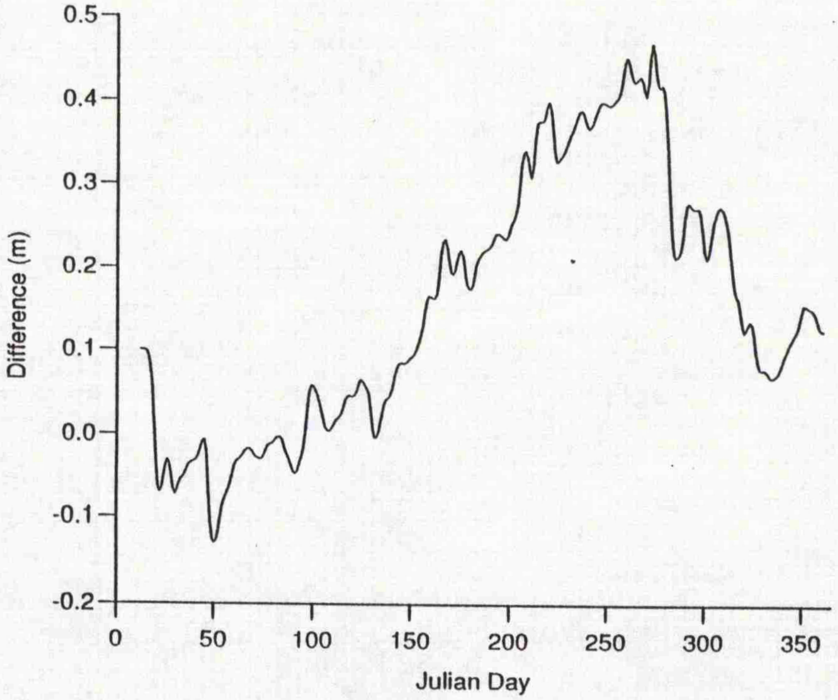


In 1991 there is initially a good correspondence between the water-table and the modelled data, however, the model is unable to represent the snow-melt event. The water-table response to the rain event of 29 April (JD 119) is slightly overestimated probably due to the small amount of preceding rainfall which was not used in the calibration. Over the summer, from 30 May onwards (JD 150), water-table responses to rainfall are over-predicted, particularly the event on 31 July (JD 212) when 16 mm of rain fell. This is partly due to interception, however, the model cannot represent the draw-down arising from evapotranspiration. Consequently the modelled data remain high and do not fall below base level, which was the minimum level recorded during the calibration event. The large rain event on 29 September (JD 272) is well represented, however, in the modelled data the falling limb for this event and those following to the end of the year, is larger than in the observed data. This is probably due to the seasonal reduction in evapotranspiration, which reduces the daily draw-down of the water-table.

The 1992 results, shown in Figure 5.18.B, initially have a reasonable fit, although there the modelled data remains slightly higher due to low rainfall in the period from 10 January to 22 March (JD 10 to 81). Again there is a difference over the summer through evapotranspiration, and summer rain events are over-predicted as in 1991. The most noticeable difference between the data series occurs on 12 and 13 August (JD 225 and 226), when the water-table increases dramatically from -0.05m to 0.436m, but is not apparent in the model results. This is probably due to a temporary problem with the recording rain gauge which led to under-recording of rainfall during this event. During the latter part of 1992, from 27 September onwards (JD 270) the model is able to predict changes in water level accurately.

The residuals of the model, produced by differencing the model results and the observed fluctuations of the water-table, are given in Figure 5.19.A and 5.20.A for 1991 and 1992. Smoothed evapotranspiration for the two years is reproduced in Figure 5.19.B and 5.20.B to consider the dependence of the

Figure 5.19. The residuals from application of the response function in 1991, and evapotranspiration (bottom).



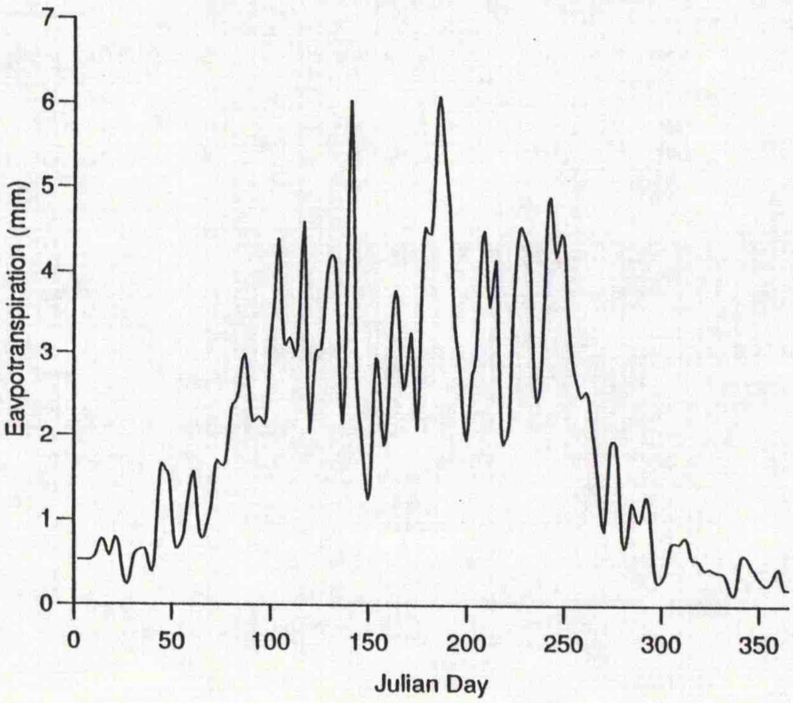
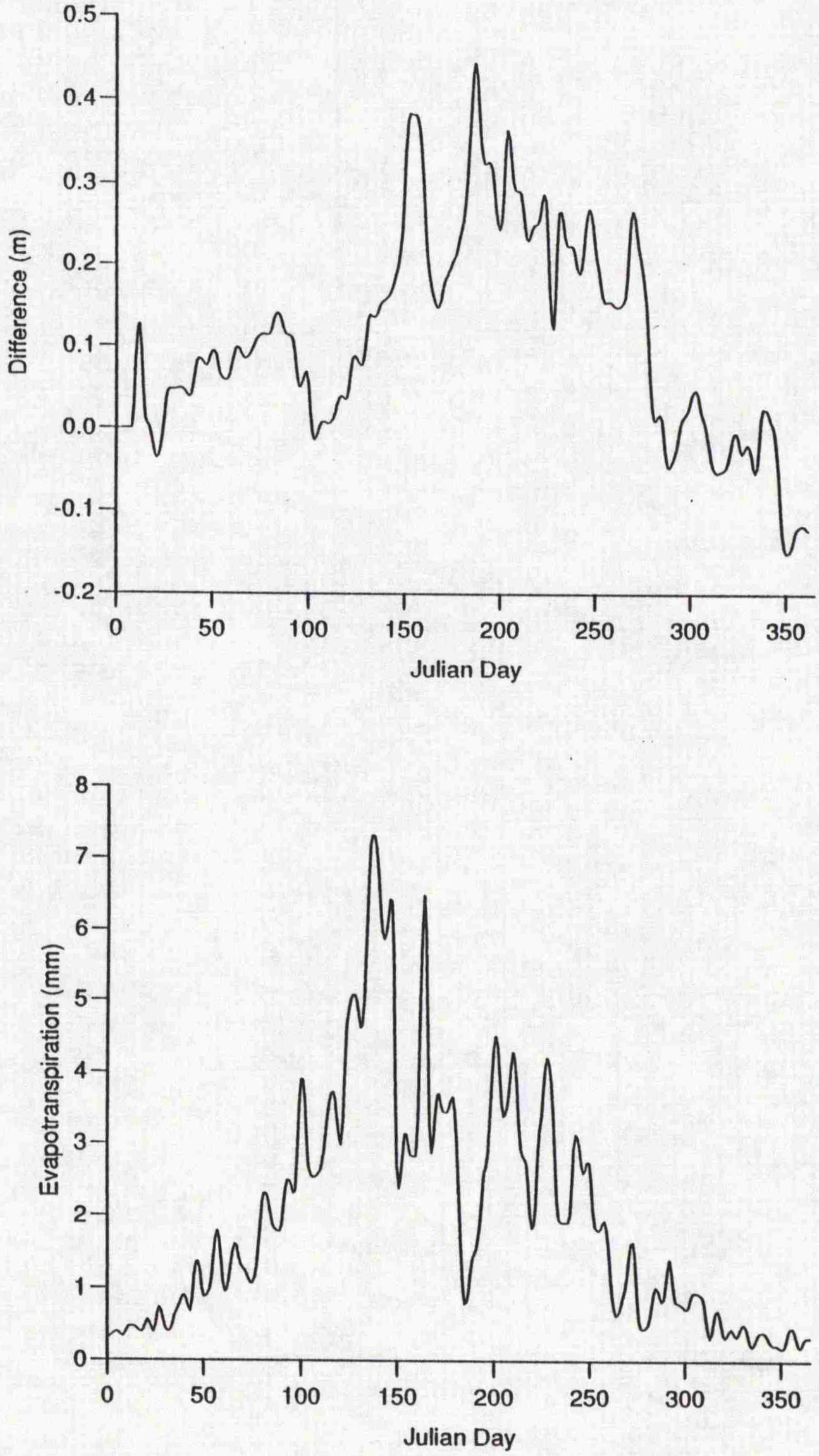


Figure 5.20. Residuals for the response function in 1992 and evapotranspiration.



degree of fit upon evapotranspiration totals. In both years the residuals are composed of fluctuations due to rainfall, superimposed upon a general curve which reflects the quantity of evapotranspiration. Consequently, differences in the appearance of the two curves is a result of variation in the timing of evapotranspiration. Thus in 1991 the asymmetry in the plot of residuals shown in Fig. 5.16.A reflects the continuation of high levels of evapotranspiration until September (JD 250), while in 1992 evapotranspiration levels are significantly lower. In 1992 also, the sensitivity of the modelled data to large summer rain events is responsible for the pronounced peaks in the residuals, for example the events on 29 May (JD 150) and 3 June (JD 185).

5.5.1. Conclusions from Quantitative Analysis.

The results discussed above, provide confirmation of the general dependence of water levels on precipitation and evapotranspiration. These processes interact in a varying manner through time to determine the fluctuations of water-tables at Narborough Bog. Water-tables vary on different time-scales, showing evidence of a distinct annual cycle, but also responding distinctly to individual precipitation events. A simple examination of the graphs of water-table fluctuations with precipitation and evapotranspiration confirms the degree to which water levels are dependent upon precipitation, and this is confirmed by the mathematical analysis. The good explanations provided by auto-regression models arise due to high levels of autocorrelation. Thus these results provide little information apart from indicating the magnitude of the daily fall in the water-table. However, applying an impulse response function to the annual data-set and considering the residuals from the fit of the model enables comparison of the nature of the water-table response to precipitation throughout the year. In this way the temporal effect of different processes, including interception and evapotranspiration, can be appreciated. Obviously the techniques present certain problems, and at the level given here cannot be applied in an unmodified form for forecasting purposes. However, the indications of wetland hydrology they provide can be applied further in the development stages for a deterministic model, which is discussed in the following chapter.

5.6. CONCLUSIONS

This chapter has introduced some of the results of the monitoring programme to provide a background against which the hydrological experiments discussed in the following chapter, can be placed. Although the historical evidence is only of limited extent, the twenty year time series of precipitation and river stage give a record of temporal variations in the hydrological processes which the wetland has experienced. The water level data, which has been collected during the annual cycles in 1991 and 1992, demonstrate how water-tables respond to the hydrological processes in different ways throughout the year. Essentially water-tables reflect the balance of precipitation and evapotranspiration, however, spatial differences in the water surface indicate the direction of water drainage. The predominant relationship between wetland water-tables and the river Soar is one of effluent seepage for much of the year, but with short time periods in which the direction of drainage is reversed. Application of a uniform impulse response function also reveals temporal changes in the hydrological processes, and it is the method used to resolve these questions which is the main challenge of the hydrological modelling techniques discussed in the following chapter.

Chapter 6

Hydrological Experiments

Scope of Chapter

The previous chapter considered the results obtained from the programme of water-table monitoring, however, the data were not used to assess the importance of river overbank flooding towards the current balance of water fluxes at Narborough. This chapter introduces and discusses the results of two detailed experiments which were designed to investigate particular hydrological processes at Narborough Bog. The chapter begins by considering the input of water from overbank flooding, as indicated by stage data collected from gauges on the river Soar, adjacent to Narborough Bog. In the following section, the results of an experiment, which artificially flooded a small area adjacent to the river, are considered.

In the remainder of the chapter, the results of an investigation into water flow through an isolated peat column are presented. This demonstrates the significance of scale considerations when modelling water flow through peat, and illustrates how the magnitude and direction of water flows depend upon the hydraulic conditions imposed by the site stratigraphy described in chapter 4.

6.1. THE ROLE OF OVERBANK FLOODING.

Overbank river flooding was almost certainly of great significance in the historical development and maintenance of the wetland at Narborough Bog as discussed in chapter 4. However, precise figures describing the contribution of water inflows from overbank floods are difficult to obtain and hence to compare with the total amount of water received from precipitation. The inflow of water onto the wetland from overbank sources will depend firstly upon the frequency and duration of overbank floods, and secondly upon the infiltration characteristics of the flooded ground. Hammer and Kadlec (1986) discussed theoretical surface movement of water within a wetland environment. Water depths during flood events are likely to be highly variable due to varying surface topography, with corresponding differences in the hydraulic head. Surface characteristics are critical in determining the direction of overbank floods, and it is difficult to obtain accurate surface elevations in areas of dense vegetation cover. At Narborough Bog, the occurrence of a 'rigg and furrow' topography was described in chapter 4. In certain areas this modifies the direction of surface water flow during flooding, and also provides a focus for infiltration.

It should be possible to determine the importance of overbank flooding by studying stage records of the river Soar, and examining the relative timing of individual flood events. A representative study of the distribution of overbank floods ideally requires a data-set extending over 30 years, which is not available for this section of the river Soar, however, a data-set from 1971 to the present was used to provide an indication of any changing frequencies of flooding. The National Rivers Authority has maintained gauging stations on the river Soar at Narborough, from 1971 to 1984, and at Littlethorpe, from 1981 to the present. A correlation was established between the discharge records for the two locations, and a data file of discharge over the full period 1971 to the present compiled. The number of overbank flood events over this time was identified by the exceedance of a discharge of 6.0 cumecs. This figure was chosen by comparison of overlapping data from the Littlethorpe gauge and records collected at the stage recorder installed at Narborough Bog. The discharge data are plotted in Figure 6.1. as the cumulative departure from the long term mean daily discharge, and individual overbank events are plotted, as flood stage against time, in Figure 6.2. The data indicate the extent of the seasonal variability of discharge which is typically highest in spring, and also demonstrates the effects of low flows in the period 1976 to 1980. In particular, the annual discharge peak was absent in 1976. The consequences of this temporal pattern are evident in the timing of overbank flood events in the years 1971-1993 shown in Figure 6.2. The figure illustrates the degree to which flood events were clustered over the period, most recently in late 1992, but also in late 1987 and in early 1986.

Clustering of overbank flood events through time can be illustrated in a more quantitative approach by statistical analysis of the river stage records. This demonstrates how the likelihood of a flood event varies depending on the length of time since a previous event. McGilchrist *et al.* (1968) applied a Markov model to daily stage data, and examined temporal variations in the time interval between flood events of a particular magnitude. In Figure 6.3. the frequency distribution of the number of days between flood events is plotted for successive overbank events in the period from June 1971 to May 1993. Over this period there were a total of 85 overbank flood events, of which 35 events occurred within 20 days of a previous flood. As the number of days separating floods increases the likelihood of a flood event decreases, 15 events occurring within 21-40 days of a flood, and 8 within 41 and 80 days. However, a significant number of floods (17), took place at an interval greater than 151 days. The greatest interval between floods was 616 days, between 9 March 1975 and 19 December 1976, while large intervals also occurred from 25 March 1988 to 25 February 1989 (336 days), and 18 February 1977 to 22 January 1978 (325 days).

The frequency plot of timing between overbank events shown in Figure 6.3. is approximately a log-normal distribution. Figure 6.4. gives a cumulative probability plot for overbank floods separated into 32 class intervals as described above, with the theoretical log-normal distribution superimposed. The latter was obtained using the formula given by Bras (1990). The data are first transformed by taking natural logarithms of the frequency class Y to produce a logarithmic class X of mean m_X and variance σ_X, with the probability density function given below:

$$f_Y(y) = \frac{1}{y\sigma_X\sqrt{2\pi}} \exp\left[-\frac{1}{2}\left(\frac{\ln y - m_X}{\sigma_X}\right)^2\right] \qquad (1)$$

Figure 6.1. Departure of river flows from long term mean, 1971-1993.

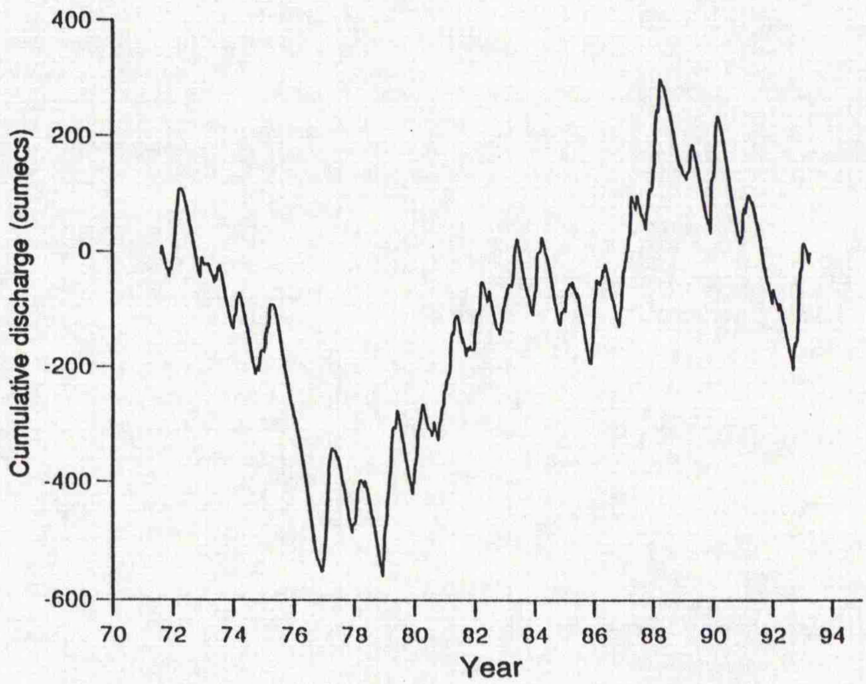
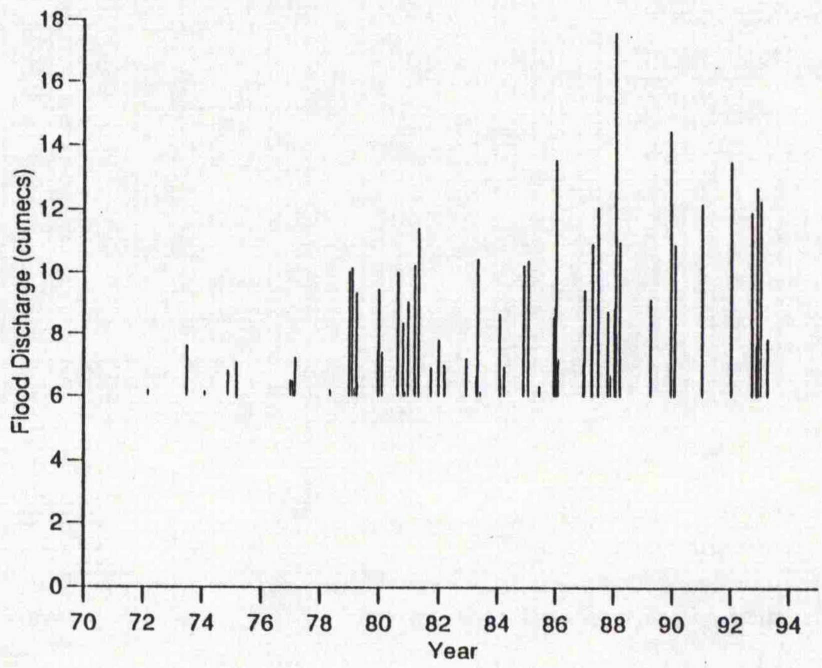


Figure 6.2. Timing of overbank flood events 1971-1993.



The differences between the two cumulative distributions shown in Figure 6.4., are not significant using the Kolmogorov - Smirnov test and do not exceed 0.17, however, they are clustered in the early class intervals of 1 to 5, corresponding to a flood interval of upto 100 days. This indicates that flood events are more likely to occur within 100 days of a previous flood, than would be predicated using a log-normal distribution. Conversely within the class interval of 10 to 13 (100 days to 160 days) flood events are less likely than expected from the theoretical distribution, most probably as a result of greater water storage within the catchment.

These results reveal a considerable range in the timing between overbank floods; which suggests that at present river flooding can only be a minor contributor to the water budget of the site. Clustering of flood events may produce certain periods when river flooding was more significant, for example in December 1992, however, the area flooded directly remains a small percentage of the total wetland area. In the Soar catchment, floods also tend to occur after heavy precipitation, when the water-table is also high thus limiting the hydraulic gradient between river stage and the local floodplain water-table.

6.1.1. <u>Summary.</u>

The implications of the above results are that, although the wetland at Narborough Bog almost certainly developed initially through water inflows from a variety of processes, the current flooding regime of the river Soar indicates that currently water from overbank flooding does not represent a consistent water source. Absolute quantities of water input are also determined by the total area flooded, the residence time of the water, and the infiltration capacities of the ground beside the river which is periodically flooded. The relative importance of river flooding is placed within the context of the processes contributing water later in the thesis. However, it is clear that while in certain years flooding is important in providing water inflows, for much of the time the reserve area remains more vulnerable to the distribution of

Figure 6.3. Frequency distribution of overbank flood occurrence.

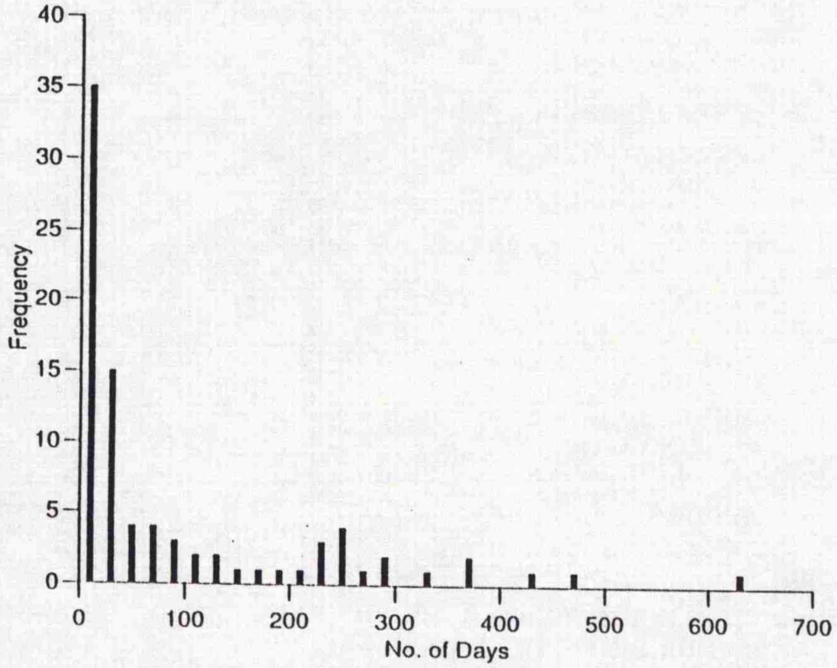
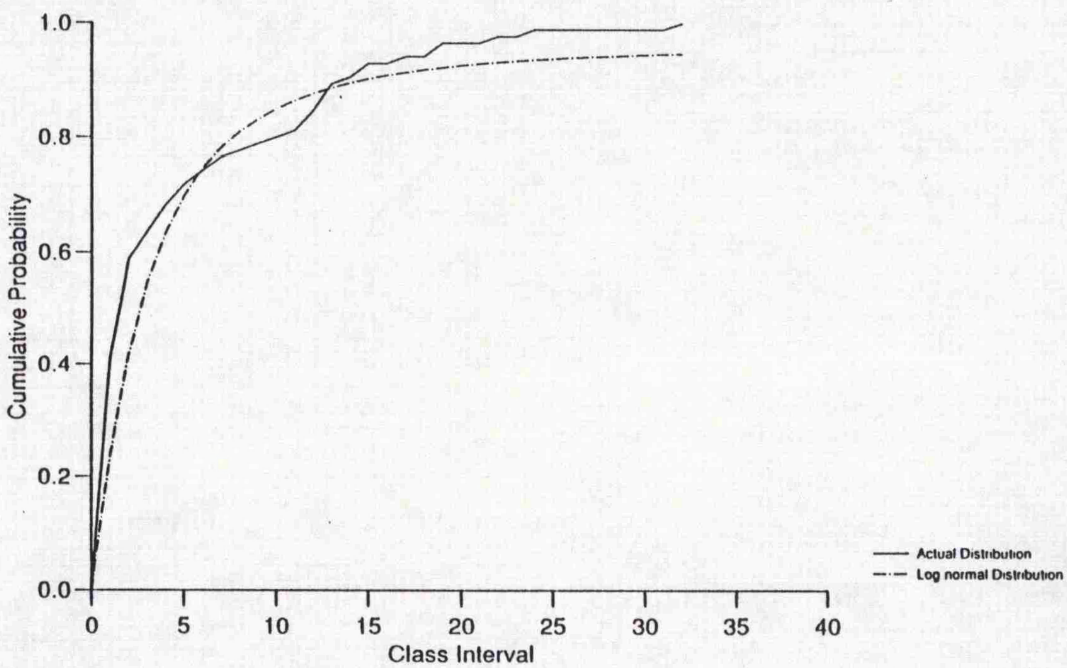


Figure 6.4. Comparison of flood frequency with log-normal distribution.



medium precipitation events.

In the following section, the implications of variations in water infiltration rates are considered first by simulating a small flood event and monitoring the resulting changes in hydraulic head. Several overbank flood events were studied during the routine monitoring programme and the observations of reversed hydraulic gradient in a localised area near the river were briefly mentioned in chapter 5. The study of dipwell records provides an incomplete picture of actual hydrological processes, as the period of flood water inundation was frequently only 2-3 days, and there is a lack of detailed hydrological data for the initial period when river stage rises and when water levels recorded at dipwells may change to attain a new equilibrium position with respect to river stage. Manual reading of water-tables contributed to a lack of data of sufficient detail to interpret changing hydrological processes at a short time scale, and consequently a controlled experiment simulating overbank flooding over a small area was conducted. This experiment is described in the next section.

6.2. ARTIFICIAL FLOOD EXPERIMENT.

The experiment consisted of pumping water for short periods of time, from the river Soar into a topographic hollow with a surface area of $8.4m^2$. Extra dipwells were installed within and around the area to be flooded, with the objective of determining the effective infiltration characteristics of floodplain sediments with high silt/clay contents which lie in the vicinity of the river. The results of the experiment would also demonstrate whether controlled flooding of areas of Narborough Bog by pumping river water would be a practical method of providing additional water inflows to the nature reserve at times of water shortage.

The existence of a network of 'rigg and furrows' within the reed-bed and extending into the surrounding woodland at Narborough Bog was mentioned in chapter 4. These extend close to the river and the experiment was designed to make use of this topography by partially flooding one furrow. Although, for practical reasons, the experiment is only testing infiltration characteristics under the furrow, the results should indicate infiltration rates over a wider area, due to lateral subsurface water flow. A 5.5m length of the furrow was isolated by sealing the ends of the hollow with timber and clay, enabling a constant head of water to be maintained within the area by pumping water from the river. The furrow lay within 5m of dipwell no. 8 (referred to as dipwell 8* in this chapter), and was within 34 metres of the river. A contour plot of the area is given in Figure 6.5. in which the area flooded is shown, and the location of dipwell measuring points are superimposed. A transverse section across the ditch is given in Figure 6.6. The section illustrates how the depression was approximately 20cm deep, with a spacing of 3m between adjacent ridges.

Two flood experiments were conducted; the first was used to test the experimental design, which was changed slightly for the second experiment. Dipwells were installed in lines of three perpendicular to the direction of the furrow, as indicated in Figure 6.5. In each transect the middle dipwell was situated in the centre of the flooded hollow, while the surrounding dipwells were in the centre of the adjacent ridges. In the first experiment six PVC dipwells, 1.3m long, diameter 5cm, and numbered 1 to 6 were installed to a depth of 1m, with holes drilled in the bottom 25cm of the dipwell. Dipwells 2 and 5 were in the centre of the hollow, while dipwells 1,3,4 and 6 were in the middle of the adjacent ridges. For the second experiment two additional transects, each of three dipwells, were installed, numbered 7 to 12. These were shallow dipwells, 75cm, installed to a depth of 50cm. Here, dipwells 8 and 13 were in the middle of the hollow, dipwells 7 and 9 were on the ridge. The second transect, dipwells 10, 11 and 12, were outside the flooded area, and were installed to study the lateral spread of infiltration.

Figure 6.6. shows the location of two sets of dipwells of different lengths, and their relationship to the existing dipwell, no. 8*. A simplified stratigraphy

Figure 6.5. Topographic surface for flood experiment and location of dipwells.

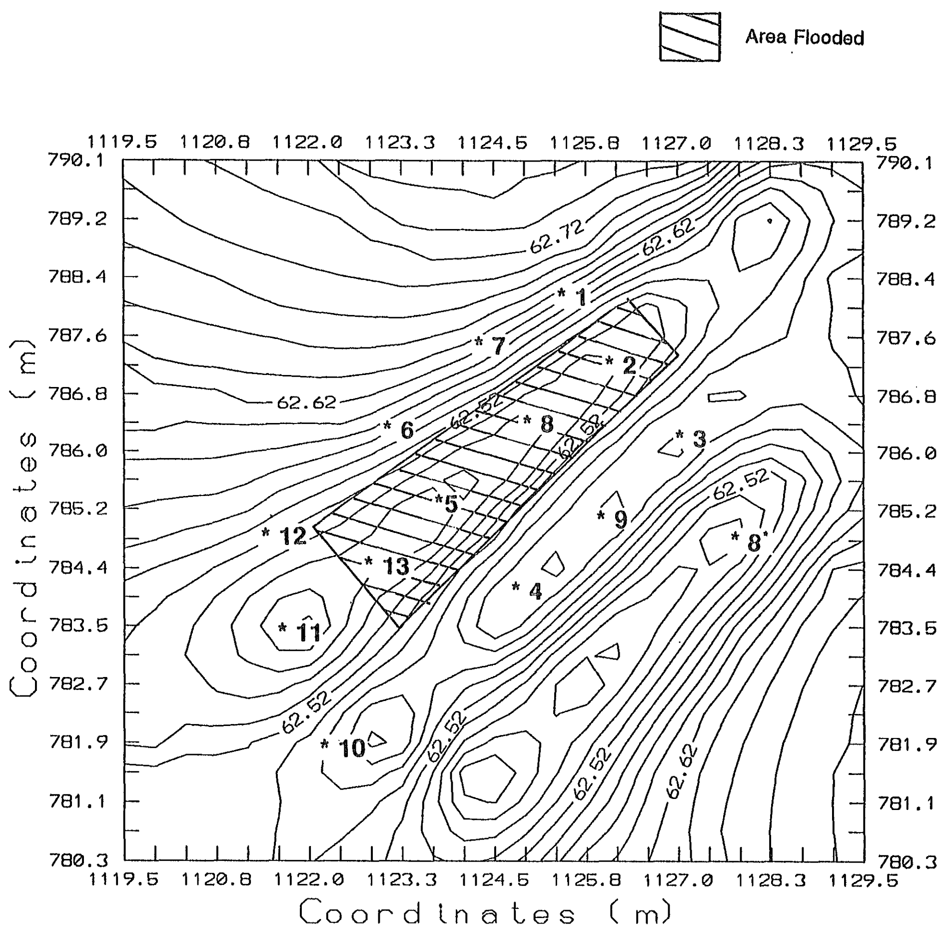


Figure 6.6. Section across flooded area, indicating stratigraphy and dipwell location.

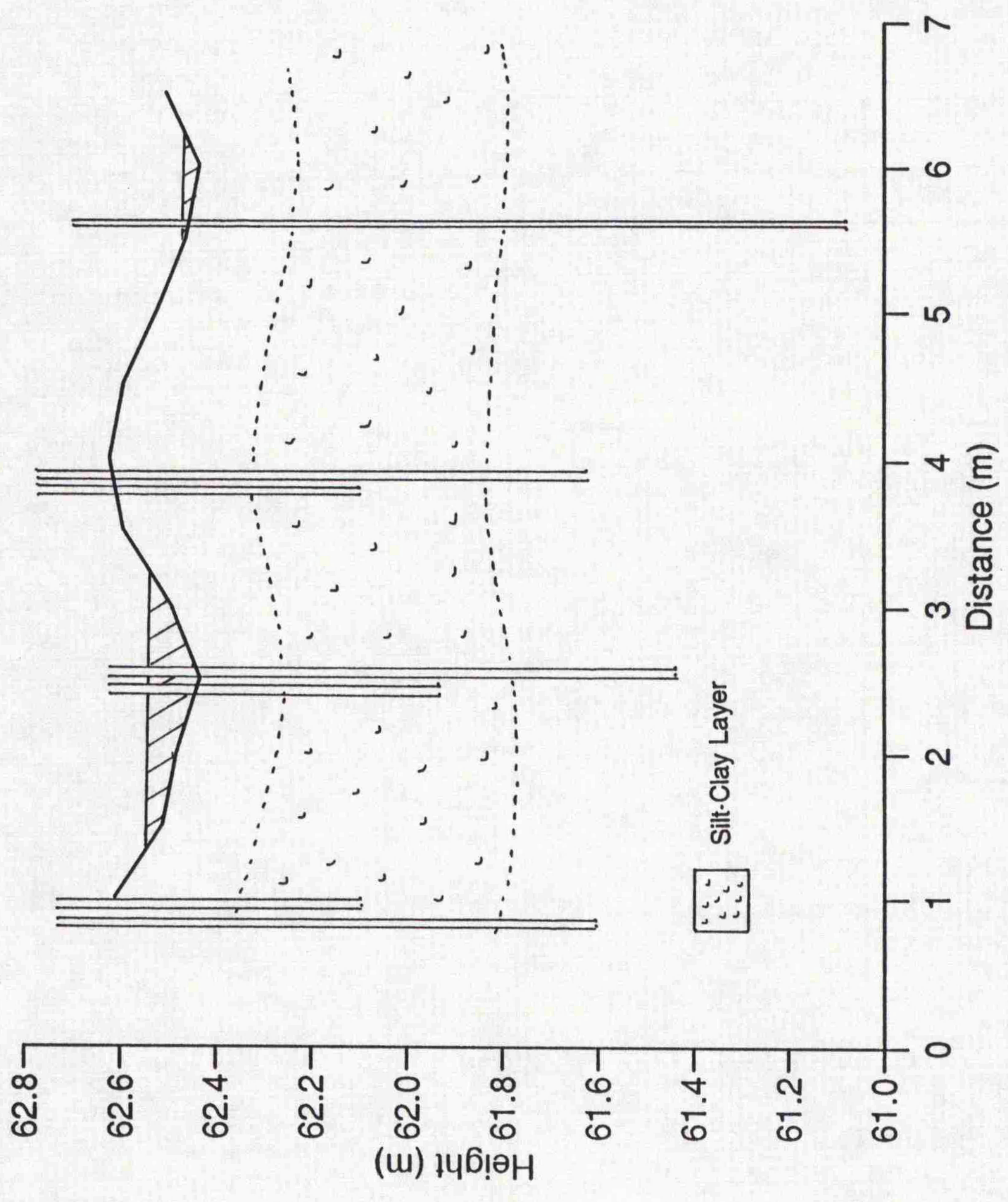


Plate 6.1. A view of the area flooded in experiment 2, looking towards the central axis of dipwells. The yellow delivery hose is visible across the picture.

is illustrated, demonstrating how the profile consists of a brown organic soil, overlying a band of thick grey/yellow clay, with an herbaceous brown grey organic silt below. The depth of flooding in the experiment is also illustrated. In the second experiment, standing water appeared in the adjacent hollow which is shown in Figure 6.6.

Plate 6.1. shows the flooded area during experiment two, which was taken from a position in the top right section of Figure 6.5., looking towards the impounded area. The line of four dipwells along the centre of the hollow can be identified, with the yellow hose which delivered the pumped water from the river (to the left of the picture).

i. Experiment 1.

The first experiment was completed on May 6^{th} 1993. A two-inch diaphragm pump was used to pump river water onto the impounded area at a constant rate of 30 litres per minute for 65 minutes. Flooding of the ditch commenced at 13:25, and continued until 14:30. The dipwells (numbers 1 to 6; 8*) were monitored at regular intervals before, during and after the flood experiment, to see whether the local water surface reached an equilibrium state in response to water infiltration.

A time-series plot showing the response of dipwells 1, 2, 3 and 8* is given in Figure 6.7., with the period of water pumping indicated by the shaded box. The graph indicates how the dipwell response to infiltration of ponded river water was limited. Dipwells 1 and 8* were most responsive, recording a parallel increase in hydraulic head. In three of the dipwells, numbers 1, 3 and 8*, a slight decrease in the head was observed in the initial period of flooding when infiltration of surface water commenced. This appears to be a genuine feature, not due to any inaccuracy in determining water levels, and may arise as infiltration of water over a wide front increases the pressure of soil air in the underlying unsaturated zone, which produces a corresponding decrease in the water level. The hydrographs of dipwells 1 and 8* indicate a water-table

Figure 6.7. Water-table response to flooding, experiment 1.

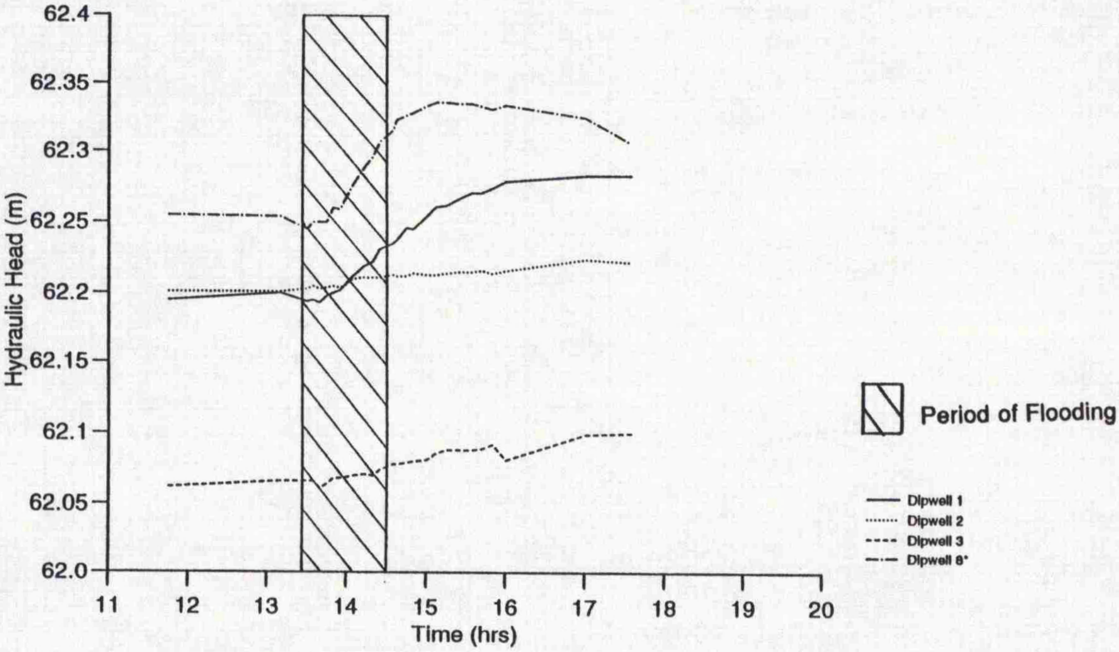
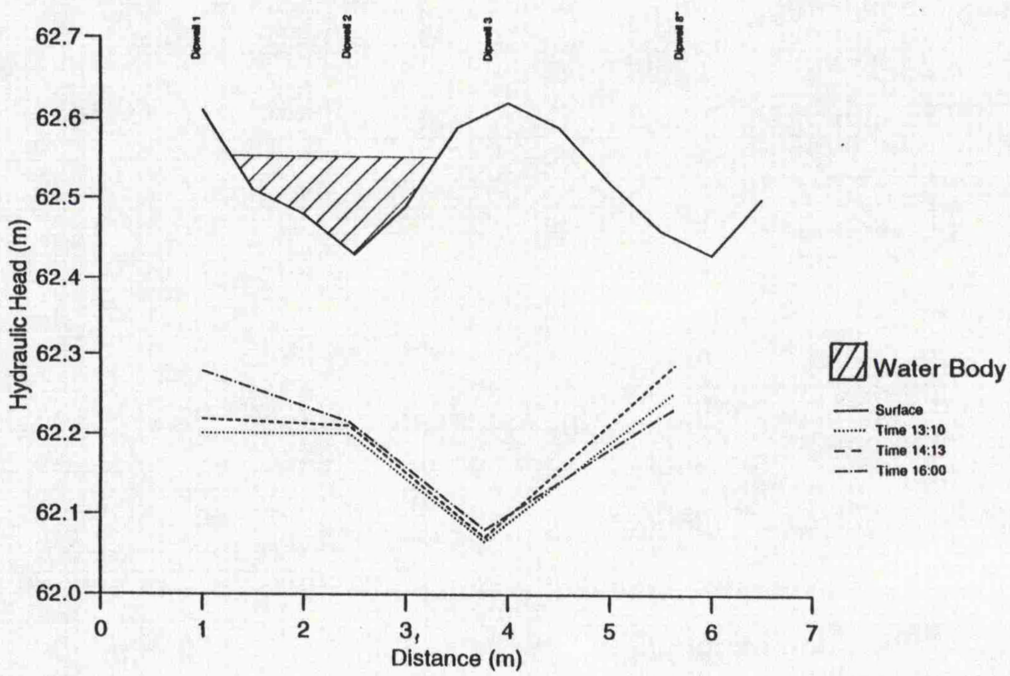


Figure 6.8. Water profile in response to flooding, experiment 1.



depression before pumping began, however this mainly reflects the timing of measurements. The subsequent increase in hydraulic head of dipwells 1 and 8* was similar, although comparatively small. The maximum observed rate of change of these dipwells was 0.07m per 30 minutes which was only maintained for the period while water was pumped. After pumping of water ceased, the head of water began to fall in dipwell 8*, however, dipwells 1,2 and 3 continued to show a slight increase in hydraulic head. This indicates much slower infiltration rates as flooding occurred through the development of a small perched water-table, with low water flux through an underlying clay layer.

The results are presented in a different form in Figure 6.8. where water profiles for one transect of dipwells 1, 2, 3 and 8*, are plotted for three time intervals: at 13:10 before the experiment began, at 14:13 during the experiment, and lastly at 16:00. The overlying surface is also plotted, and the maximum depth of the flooded area indicated. The graphs demonstrate that the initial water surface at 13:10 was not horizontal, the water surface was a constant 62.2m between dipwells 1 and 2, then fell to 62.07m at dipwell 3, and rose to 62.25m at dipwell 8*. In the second profile, at 14:13 after 13 minutes of ponded infiltration, dipwells 1 and 2 record a similar increase in the water surface of 0.07m, little change was observed at dipwell 3 (0.01m), while hydraulic head at dipwell 8* increased by 0.03m, thereby maintaining the previous hydraulic head configuration. By the time of the third profile at 16:00, ponding of water had ceased 3 hours before and there was little change in the readings at dipwells 2 and 3; however, the water surface at dipwell 1 had increased by 0.06m and at dipwell 8* by 0.05m.

The spatial variability of the response of dipwells to infiltration during the experiment is indicated further in Figure 6.9. In this figure a three-dimensional bar graph is used to show change in hydraulic head which was recorded for the seven dipwells, numbers 1,2,3,4,5,6 and 8*, at 16:00. The results clearly indicate differences in the sensitivity of response of the dipwells,

Figure 6.9. Bar chart indicating the change in hydraulic head for experiment 1.

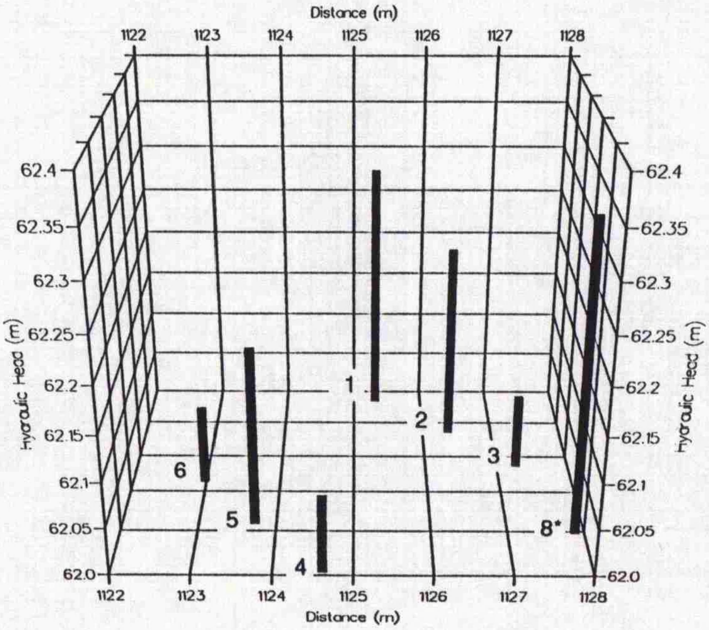
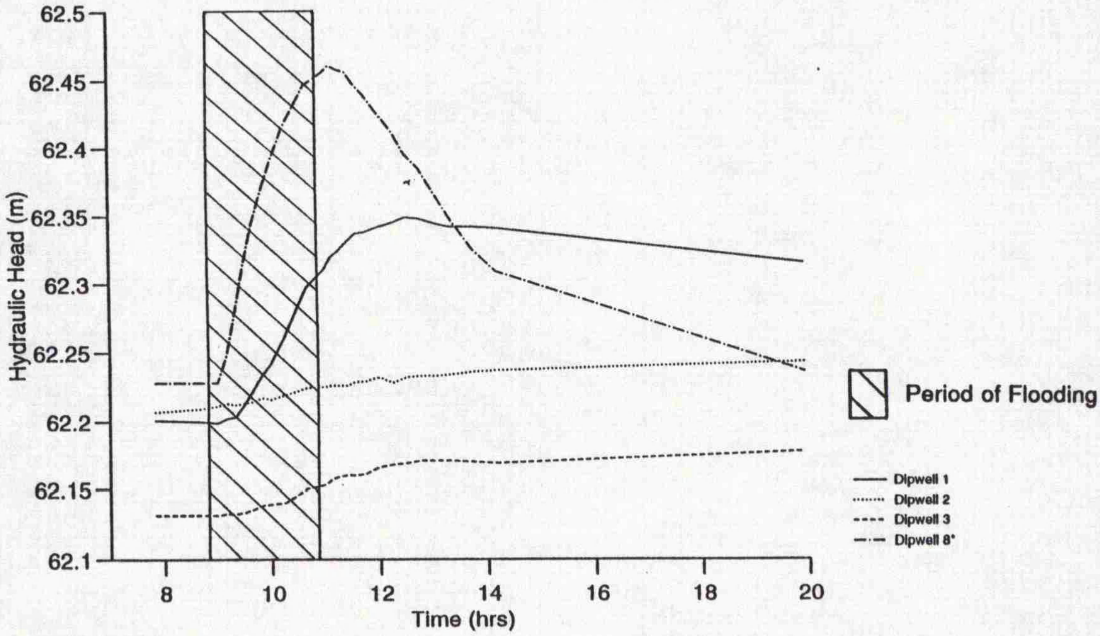


Figure 6.10. Time series plot, dipwells 1,2,3,8*, in experiment 2.



although all were installed by a consistent technique. Three dipwells, nos. 3, 4 and 6, showed only a very small change in head, while the greatest increase occurred at dipwell 8*. During the experiment no dipwell reached a hydraulic head close to the level of ponded water, despite the substantial volume of water pumped onto the area. The surface water level did not vary, indicating a substantial lateral flow rate, slow infiltration rates through deposits of low permeability, and also variations in water flux. This indicates that the rate of water pumping exceeded the rate of water infiltration through the underlying deposits. Differences between the response of dipwells thus most probably reflect variations in stratigraphy at the base of individual dipwells, and the relative location of clay lenses with respect to the dipwell base. Horizontal clay layers will act both as a barrier to vertical infiltration while water is ponded at the surface, and will also direct water laterally with the possibility of a greater increase in hydraulic head at distant dipwells.

ii. Experiment 2.

The tentative conclusions outlined above were used to improve the instrument design for the second experiment. The possibility of extended lateral water flow was investigated by supplementing the dipwell network, and installing two further dipwell transects. These consisted of shallow dipwells, 0.75m long, which would also indicate differences in the hydraulic gradient over small vertical distances. The experiment was conducted on 10th May 1993, when water was pumped at the same rate onto the area used in experiment 1, for 115 minutes from 08:55 until 10:50. Hydraulic head measurements were taken at regular intervals as in the previous experiment.

Time series plots recording the response in hydraulic head are given in Figure 6.10. and 6.11. for two transects of dipwells. Figure 6.10. shows the records for the four dipwells illustrated in Figure 6.8., and demonstrates how the form of the two graphs differs significantly. The longer monitoring period in the second experiment revealed a sinusoidal response of dipwell 8* to the growth and decay of ponded infiltration, as the observed hydraulic head first

Figure 6.11. Time series plot, dipwells 7,8,9, experiment 2.

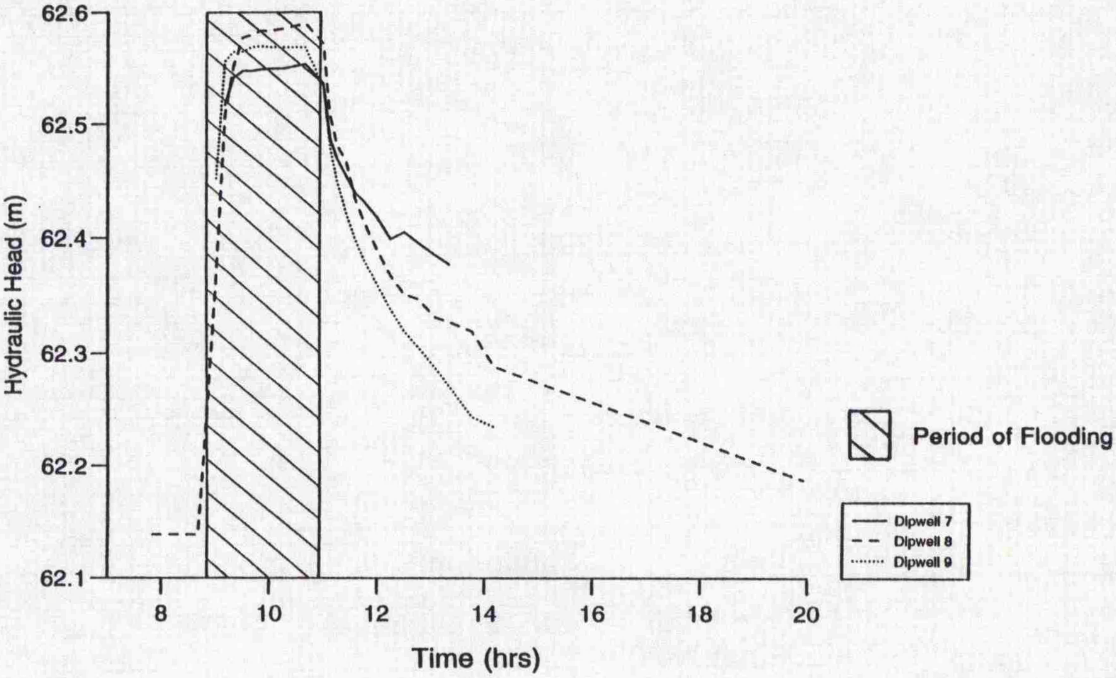
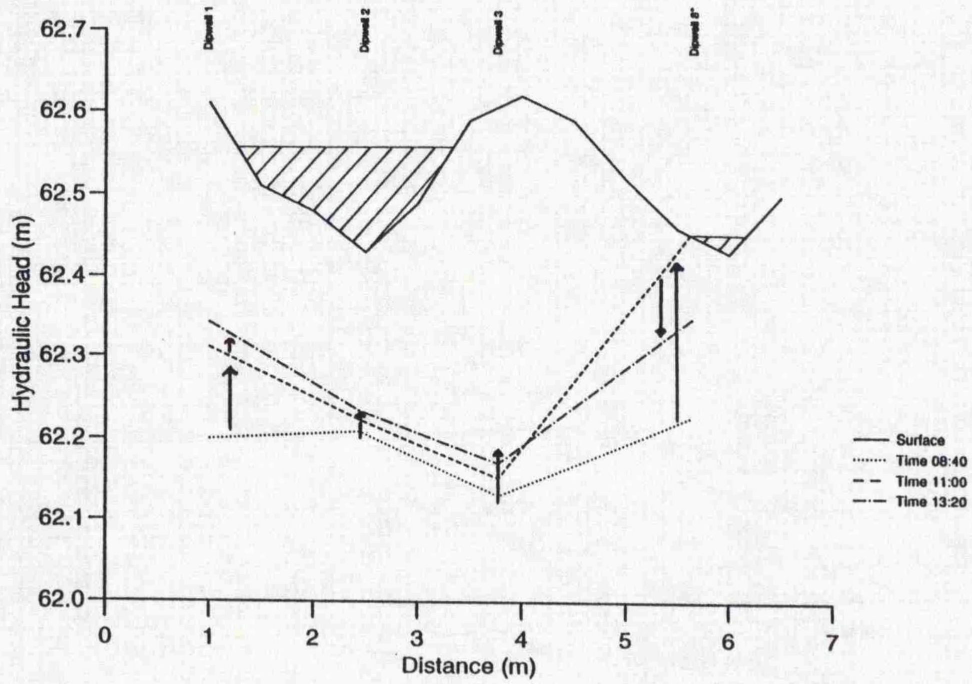


Figure 6.12. Water profile, dipwells 1,2,3,8*, experiment 2.



increased then decreased symmetrically. Dipwell 8* responded more quickly to the onset of flooding, with the water head increasing rapidly at a maximum rate of 0.08m per 30 minutes, peaking at 11:00 after pumping had ceased and then decreasing rapidly. The head at dipwell 3 also increased significantly, but at a lower rate of 0.035m per 30 minutes. The maximum hydraulic head was reached at this dipwell at 12:30, and then decreased at a lower rate than dipwell 8*. Rates of increase of head for dipwells 2 and 3 were again lower, at 0.005m per 30 minutes, however, the readings of hydraulic head continued to increase at a similar rate until measurements ceased at 20:00.

Changes in hydraulic head are shown in Figure 6.11. for the transect of shallow dipwells, numbered 7, 8 and 9. An initial rapid increase in hydraulic head was observed, followed by a decrease in head, which represents a different response to the dipwells shown in Figure 6.10. The three dipwells in this transect also rapidly reached a state at which a stationary water-table was observed during the experiment, which decreased quickly following the end of water pumping. This indicates close hydraulic contact with the ponded surface water and the water surface reached in the dipwells is approximately the same as the surface water level. However, there is a maximum difference of 5mm between the surface at dipwell 8 and that at dipwell 7, indicating differences in the degree of contact of the dipwells. The highest water surface was observed at dipwell 8, in the centre of the ditch, indicating a sloping water-table centred on the ponded area.

Further detail of the variations in hydraulic head are shown in Figures 6.12. and 6.13. which give water profiles along the two transects discussed above, for three times: at 08:40 before pumping began, at 11:00 as water-tables began to fall after pumping ended, and finally at 14:10. The first profile in Figure 6.12. shows a water surface which is approximately in equilibrium at the beginning of the experiment, although as in experiment 1, the water-table at dipwell 3 was again lower. As time proceeded the greatest increase in the water surface was observed at dipwells 1 and 8* at the margins of the flooded

Figure 6.13. Water profile dipwells 7,8 and 9 in experiment 2.

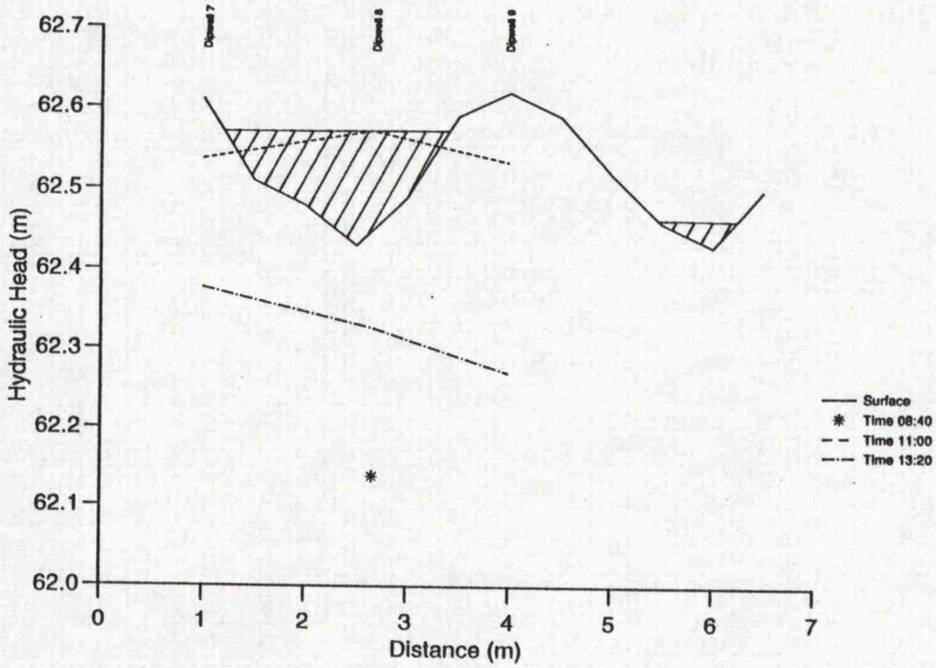
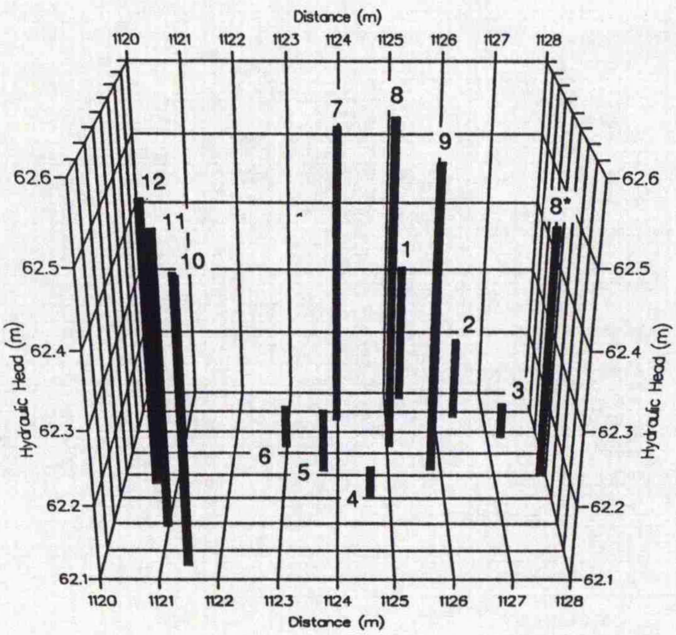


Figure 6.14. Bar graph of dipwell response, experiment 2.



area. At 11:00, the hydraulic head had risen 0.11m at dipwell 1, 0.02m at dipwells 2 and 3, and 0.23m at dipwell 8*. Seepage flows from the main furrow flowed into the hollow beside dipwell 8*, producing an increase in head to the local water level. By 14:00 dipwells 1, 2 and 3 had risen by similar amounts of c.0.03m thereby maintaining a constant gradient, while at dipwell 8* the water-table decreased by 0.15m. The arrows in Figure 6.12. indicate differences in the timing of dipwell response to infiltration. The measurements at dipwells 1, 2 and 3, showed a steady increase in hydraulic head through time. At dipwell 8* there was an initial large increase in head, which was followed by a fall, as pumping of water ceased and surface water was no longer ponded at the surface. This indicated that the wetting front had passed through the soil profile.

In the second profile for the shallow dipwells, 7, 8 and 9 shown in Figure 6.13., the water surface is significantly different. Initially hydraulic head was only obtained for dipwell 8, as the water-table lay below the base of the other dipwells. However, after pumping began, there was a rapid increase in hydraulic head as the water in the three dipwells rose rapidly, and formed a water-table mound centred on the ditch by 13:20. At this time the head in dipwell 8 was equal to the depth of impounded water, while the head in dipwells 7 and 9 was 0.3m lower.

The variability of the results for the second experiment are illustrated in Figure 6.14. where the hydraulic heads for all dipwells at 11:00 are plotted using a bar graph. Generally the greatest increase in water levels was observed in the two transects of shallow dipwells, although the deeper dipwell at 8* also responded quickly to infiltration. The considerable differences between the dipwell transects demonstrate the possible importance of local variations in permeability.

6.2.1. Discussion.

Results from the two experiments are summarised in Table 6.1. which gives comparative figures for the hydraulic gradients for the time periods illustrated in the profile plots above. These again demonstrate the scale of the difference in gradient between dipwells in the same transect, as was evident from the graphs above. However, the table also clarifies the relationship between the dipwells of different length used in the second experiment. For example, the gradient between dipwells 2 and 8, which both lie within the centre of the flooded area, indicates the size of the hydraulic gradient between water flooded at the surface as an upper perched water-table, and the underlying water level. Thus at time 11:00 in experiment 2, a gradient of 0.1792 m/m was observed which provides the mechanism for the water flux through the clay layer. At 14:10 the gradient between the same dipwells had fallen to 0.0263, showing how lateral water flow limited the time over which flood conditions were maintained. The lateral extent of the water flows can also be seen from examining the last collection of gradients between dipwells in the transect 4,5 and 6 and the transect 10, 11 and 12. As Figure 6.5. indicates, the latter transect is about 1.5m outside the flooded area, however, here also a large water-table gradient developed during flooding, for example 0.1157m/m between dipwells 6 and 12 at 11:00, which decreased to 0.0549m/m at 14:10. Unfortunately here the data are limited as the initial water-table position lay below the bottom of some of the shallow dipwells.

The two experiments reveal considerable differences in dipwell response during the period of localised surface flooding. These differences were observed over very small horizontal and vertical distances and demonstrate the importance of scale when determining variations in hydraulic head. In a normal overbank flood there will be a much larger area flooded, and the hydrological response may be different, however, the experiment emphasises the small-scale differences in permeability of alluvial sediments. In chapter 2 the magnitude of the range in hydraulic conductivity of alluvial deposits was described; the results demonstrate how this extreme range may determine the characteristics of water infiltration from overbank floods, and also influence the direction of lateral subsurface water fluxes. Rates of water movement are very

Table 6.1. Hydraulic Gradients in Flood Experiment.

Experiment 1		Gradient	
Time	Dipwell 1:2	Dipwell 2:3	Dipwell 3:8*
13:10	0.0003	0.0504	0.0499
14:13	0.0028	0.0522	0.0576
16:00	0.0220	0.0504	0.0671
Experiment 2			
08:40	0.0031	0.0291	0.0257
11:00	0.0307	0.0261	0.0804
14:10	0.0362	0.0254	0.0371
Experiment 2	Dipwell 1:7	Dipwell 2:8	Dipwell 3:9
08:40	*	0.0366	*
11:00	0.1289	0.1792	0.1815
14:10	*	0.0263	0.0310
Experiment 2	Dipwell 4:10	Dipwell 5:11	Dipwell 6:12
08:40	*	*	*
11:00	0.0826	0.0862	0.1157
14:10	0.0259	0.0178	0.0549

* - water-table below the level of one dipwell.

low through clay deposits; however small scale heterogeneities in a deposit account for significantly greater water fluxes than would otherwise be possible. Structural discontinuities in a clay deposit provide one mechanism for faster routing of water; however also important will be the localised occurrence of more permeable deposits, for example herbaceous peat, and also root systems which can provide areas in which faster routing of water is possible. These mechanisms need to be invoked to explain the experimental observations, in which a constant rate of water application maintained by pumping river water into the ditch produced no variation in the head of surface water. Hydraulic head measurements will also vary, reflecting the head differences across sedimentary lenses of differing hydraulic characteristics.

There have been few studies of areal infiltration, however, general results support the high variability in infiltration rates, which the results described above indicate. Tricker (1981) found hourly infiltration rates in a 3.6km^2 catchment to vary from zero to 256 cm/h. Wilcock and Essery (1984) studied infiltration within a 15.7km^2 lowland catchment in Northern Ireland, and recorded a dependence of infiltration on organic matter, with some seasonal variability.

6.2.2. Determination of hydraulic parameters.

Although the study of ponded infiltration revealed very low rates of water movement to the deep piezometers, which were located below a low permeability clay layer, substantial infiltration occurred during the experiment as the surface head of water does not change with time despite continued application of surface water. Accurate determination of the contribution of water provided by overbank events will therefore be complicated by the development of a perched water-table at the surface, which will be followed by a slow trend towards equilibrium conditions, as the hydraulic head in dipwells within clay deposits adjusts to the flooded water at the surface. Whether this is achieved will depend upon the residence time of flood water, and the nature of surface water flow through the wetland. However, the time-series of dipwell response may be used to determine the hydraulic conductivities of intervening deposits using standard formulae for piezometer tests. Measurement of hydraulic conductivity, by pump evacuation of monitoring tubes, uses the rate of change of water level in relation to an applied hydraulic gradient. In the flood experiment the water ponded at the surface represents a constant head of water (for the duration of pumping), and the dipwell response is proportional to the permeability of the intervening deposits.

Kirkham (1946) used the following equation to obtain hydraulic conductivity:

Table 6.2. Calculated hydraulic conductivities from the flood experiment.

Dipwell	Initial water table (m)	Depth to base of dipwell (m)	Tube Radius (m)	Head diff at time 1. (m)	Head diff at time 2. (m)	Time interval (s)	Shape factor	Hydraulic Conductivity (m/s)	Hydraulic Conductivity (m/day)
Experiment One									
8*	62.259	1.800	0.02	0.324	0.268	2380	30.0	3.3397×10^{-9}	2.885×10^{-4}
1	62.202	1.000	0.02	0.381	0.348	2380	26.9	1.7783×10^{-9}	1.537×10^{-4}
3	62.068	1.000	0.02	0.515	0.505	2380	26.9	3.8488×10^{-10}	3.325×10^{-5}
Experiment Two									
8*	62.228	1.800	0.02	0.355	0.136	5980	30.0	6.7207×10^{-9}	5.807×10^{-4}
1	62.199	1.000	0.02	0.384	0.286	5980	26.9	2.3018×10^{-9}	1.989×10^{-4}
3	62.131	1.000	0.02	0.452	0.433	5980	26.9	3.3547×10^{-10}	2.899×10^{-5}
8	62.138	0.500	0.02	0.445	0.076	900	26.9	9.1735×10^{-8}	7.925×10^{-3}

$$K = \frac{\pi r^2}{A_{(r,d,s)}} \frac{\ln h_i/h_j}{t_j - t_i} \quad (2)$$

where r is the radius of the dipwell; h_i and h_j are the differences between the initial water-table depth and the depth of water in the pipe, at times t_i and t_j; and A is a shape factor determined by the depth of the dipwell base below the water-table (d), proximity to either an impermeable, or infinitely permeable boundary (s) and tube radius (r). The data used to calculate hydraulic conductivity are given in Table 6.2. Conductivity was determined for dipwells 8*, 1 and 3, in experiment 1, and again in experiment 2, when the shallow dipwell number 8 was also used. The shape factor was obtained from a table given by Amoozegar and Warrick (1986); it was assumed that the dipwells were situated a given height above an infinitely permeable deposit, the underlying gravels. The time interval is equal to the period of pumping, except in the case of dipwell 8, which responded rapidly to water infiltration, where it equals the time to equilibrium.

The results indicate the extent of variability in hydraulic conductivity over the small area flooded. The greatest conductivity of 7.925×10^{-3} m/day was recorded at dipwell 8, while the least responsive dipwell, number 3, had a conductivity of c. 3×10^{-5} m/day. Comparison of results for experiments 1 and 2 reveal the amount of error in the calculations; this is substantial for dipwell 8*, where the conductivity obtained in experiment 2 was twice the value for the first experiment, however, the error in the other measurements is less (25% at dipwell 1, and 13% at dipwell 3).

6.3. EXPERIMENTS ON PEAT HYDROLOGY.

6.3.1. Introduction.

Studies investigating the nature of peat hydrology were reviewed in section 3.2., in which the limitations of applying Darcy's Law at different scales were indicated. Peat deposits are highly heterogeneous, with considerable horizontal and vertical differences in permeability within a particular deposit, which is additional to substantial variations between peat of different types. Much of the stratigraphy of Narborough Bog, with the distinct exception of the area adjacent to the river, consists of a combination of herbaceous peat overlying dark wood peat (chapter 4). The hydrological characteristics of these sediments are therefore an important element in determining the dynamics of water flow through the wetland system.

Peat hydrology has frequently been studied by either taking cores for laboratory analysis, or using limited in-situ tests such as the auger-hole or seepage tube test. In this section, experimental tests are described which were undertaken on an isolated column of peat, which enabled the hydrological effects arising from deposition of different peat types to be assessed. Variations in water flux through a peat column were investigated by controlling the flux of water using a surface crust of hydraulically resistant composition. The use of a crust of this type to determine hydraulic conductivity has been described by Hillel and Gardner (1969; 1970a; 1970b). The method involves isolating a column of soil which is instrumented with tensiometers at different depths. An artificial crust consisting of varying proportions of gypsum and fine sand (60-250 μm) is applied at the surface of the column, over which a constant head of water is maintained. Tensiometers are used to monitor the process of water flow through the column as indicated by the changing soil water tension.

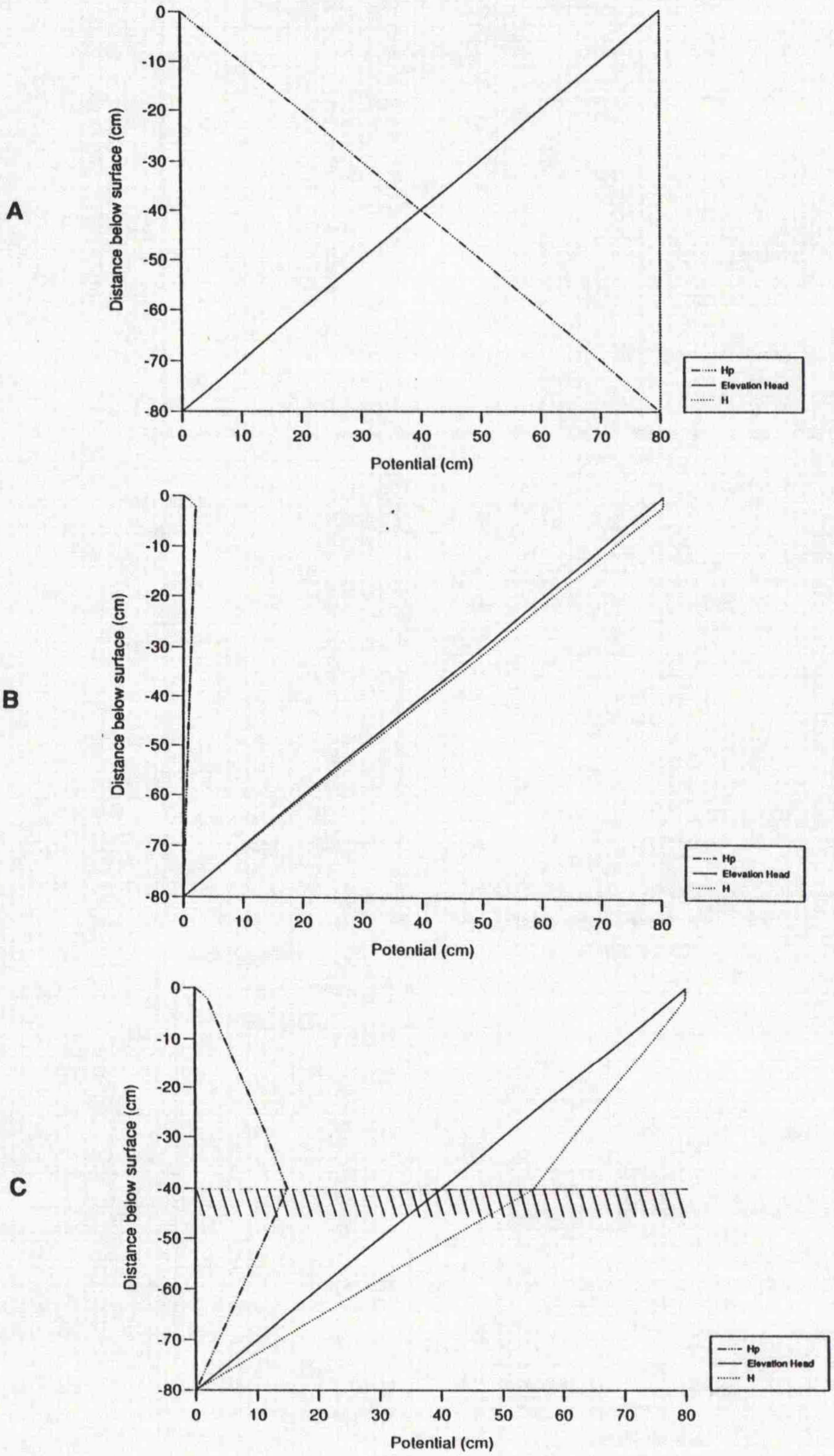
6.3.2. Theoretical Infiltration.

The hydrological theory to the process of water infiltration was discussed in chapter 3, section 3.1.4. The mechanics of infiltration are reviewed here, with respect to changes in hydraulic head, pressure head, and elevation head, to provide a framework for the analysis of the experimental results. Infiltration of water occurs in proportion to the gradient in hydraulic head between two points in accordance with Darcy's Law. Water flow will therefore vary depending upon changes in hydraulic head (h), which is the sum of elevation head (z) and pressure head (h_p), as discussed in chapter 3. Under conditions of no water flow, hydraulic head will be unchanged with depth through the soil profile, thus pressure head will increase with depth as the elevation head decreases. This situation is shown schematically in Figure 6.15.A for the example of a homogeneous soil column. Here, hydraulic head is unchanging with depth remaining a constant 80cm, while elevation head and pressure head are directly inversely proportional to each other.

If infiltration of water begins through the ponding of a 2cm depth of water at the surface the head distribution changes considerably, as illustrated in Figure 6.15.B. Elevation head is unchanged; however water flow occurs through the profile due to a uniform gradient in hydraulic head, which varies from 80cm at the surface to 0cm at a depth of 80cm. The pressure head initially increases with depth as a result of surface ponding of water, but then decreases constantly through the profile as determined by the difference between the hydraulic and elevation head. In this example, flow would occur due to a hydraulic gradient of 80cm/80cm (ie. gradient of 1).

The head distribution during infiltration into a soil profile consisting of two layers, the upper layer having a greater permeability, is more complex. A theoretical graph of variation of elevation head, pressure head, and hydraulic head with depth for this example is shown in Figure 6.15.C. Elevation head decreases constantly with depth, as in the previous example, but the gradient of hydraulic head, and hence pressure head varies in proportion to the change in permeability. Water flux is maintained by the gradient of hydraulic head across the column; a constant flux of water through the profile is assumed. Consequently a greater gradient is required in the bottom layer, of lower permeability, to maintain the same water flux, and consequently two lines of different gradient are required to represent hydraulic head variation through the profile. As a result field observations of pressure head will increase with depth in the upper layer, before decreasing in the layer below.

Figure 6.15. Theoretical hydraulic head for conditions of A. no flow; B. flow; and C flow through a two-layered profile.



6.3.3. Experimental procedure.

The area chosen for the experiment was in the centre of the reed-bed at Narborough Bog, where the stratigraphy consisted of 20cm peat soil, above 1m herbaceous peat, overlying 0.6m wood peat. The choice of location enabled the study of water infiltration through peat of varying stratigraphy, which was representative of the stratigraphy of a wider area of Narborough Bog, with the exception of the river marginal area. Preparations were undertaken in mid-September 1991 when the water-table was c. 106cm below the surface. Vegetation was removed over a $9m^2$ area. A metal infiltration ring (height 25cm; diameter 30cm) was driven into the peat surface to a depth of 13cm. A column of peat was excavated around the ring to a depth of 104cm below the ring top. The column was slightly conical in shape to increase stability, the circumference increasing from 107.5cm at 29.5cm depth, to 131.0cm at 74.0cm, which is equivalent to a side slope angle of 85.2°. Tensiometers were installed at three positions. T_{upper} (T_u) at 29.5cm below the surface, T_{middle} (T_m) at 53.0cm, and T_{lower} (T_l) at 74.0cm. The porous cups of the tensiometer were packed into holes drilled in the peat using a sand/loess mixture to maximise contact between the peat matrix and porous cup, and were sealed using pellets of bentonite clay. Visible macro-pores ending at the margin of the column were also sealed with bentonite. A small constant head of water (c. 2cm) was maintained at the surface of the column using a 1 litre Mariotte bottle, which was secured to the infiltration ring. A schematic illustration of the column is given in Figure 6.16., which shows the location of the tensiometers and likely water flow paths through the column which would develop during water infiltration. The peat column is illustrated in Plate 6.2., during the first experiment.

The variable stratigraphy of the peat column is summarised in Table 6.3. A monolith of peat was collected and bulk density and organic mater content determined. Bulk density was calculated at 4.5cm intervals to a depth of 45cm. Organic matter content was determined by the modified Wakley-Black method of wet oxidation.

Figure 6.16. Schematic diagram of peat column.

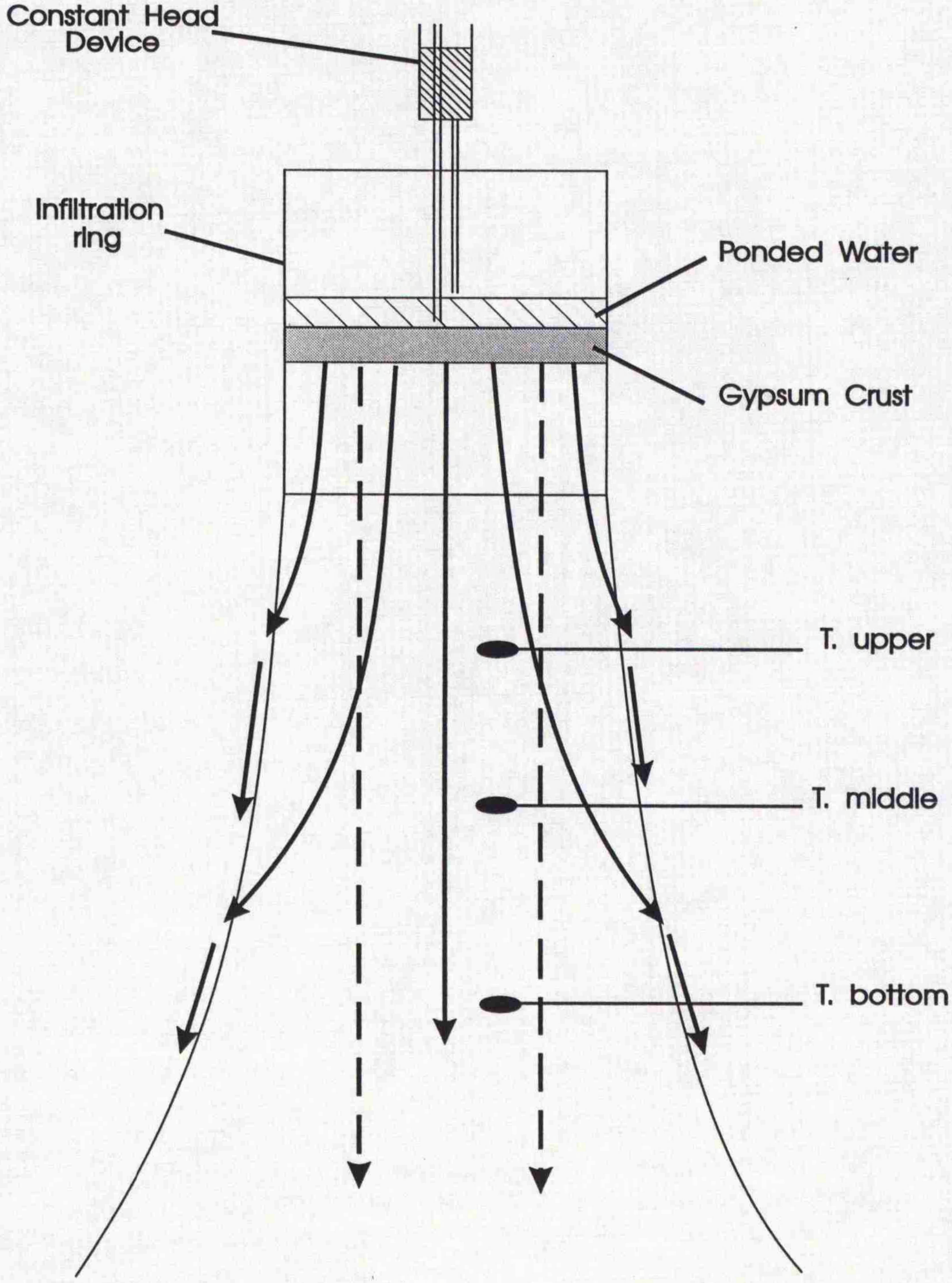


Plate 6.2. The peat column, after experiment 2. The Mariotte bottle is at the top of the picture mounted on the infiltration ring. Bubbles from expelled soil air are visible around the column margins.

Table 6.3. Characteristics of Peat Column.

Depth	Characteristics	Bulk Density (mg/m^3)	Organic Content
0 - 4 cm	Friable peat soil		
4 - 14 cm	Peaty silty clay	0.325	26-29%
14 - 26 cm	Herbaceous Phragmites Peat	0.160	47-49 %
26 → cm	Dark Wood Peat		51-55%

A total of four experiments were completed on the column as described below. These used crusts of differing hydraulic resistance to control the water flux through the column, while the tensiometers were used to monitor differences in soil tension. The first experiment investigated the rate of water flow through the column, with no surface crust. In the following three experiments crusts of varying hydraulic resistance were produced by modifying the proportions of sand and gypsum in the crust. In the second experiment a crust, consisting of 10% gypsum and 90% sand by weight was used, while in the third and fourth experiments a greater proportion of gypsum was used. Here, the quantities are specified by volume, which was intended to compensate for the gypsum occupying pore spaces within the sand.

i. Infiltration without a crust.

An initial infiltration experiment was undertaken to examine the interaction of water flows through the different peat layers under conditions of unimpeded flow. A total of 53 litres of water was poured onto the peat surface within the infiltration ring over a period of 2.5 minutes, representing an equivalent water flux of 0.3 litres/second. The rate of water infiltration was so rapid that the surface application of water could not be maintained and the experiment was terminated.

Rapid transport of water occurred through the column through macropores which appeared to transmit much of the flow as water was discharged at the column margins. This is illustrated in Plate 6.2. which shows the peat column at the end of the experiment. Air bubbles can be identified at the sides of the column, which indicate the expulsion of soil air during downward passage of the wetting front.

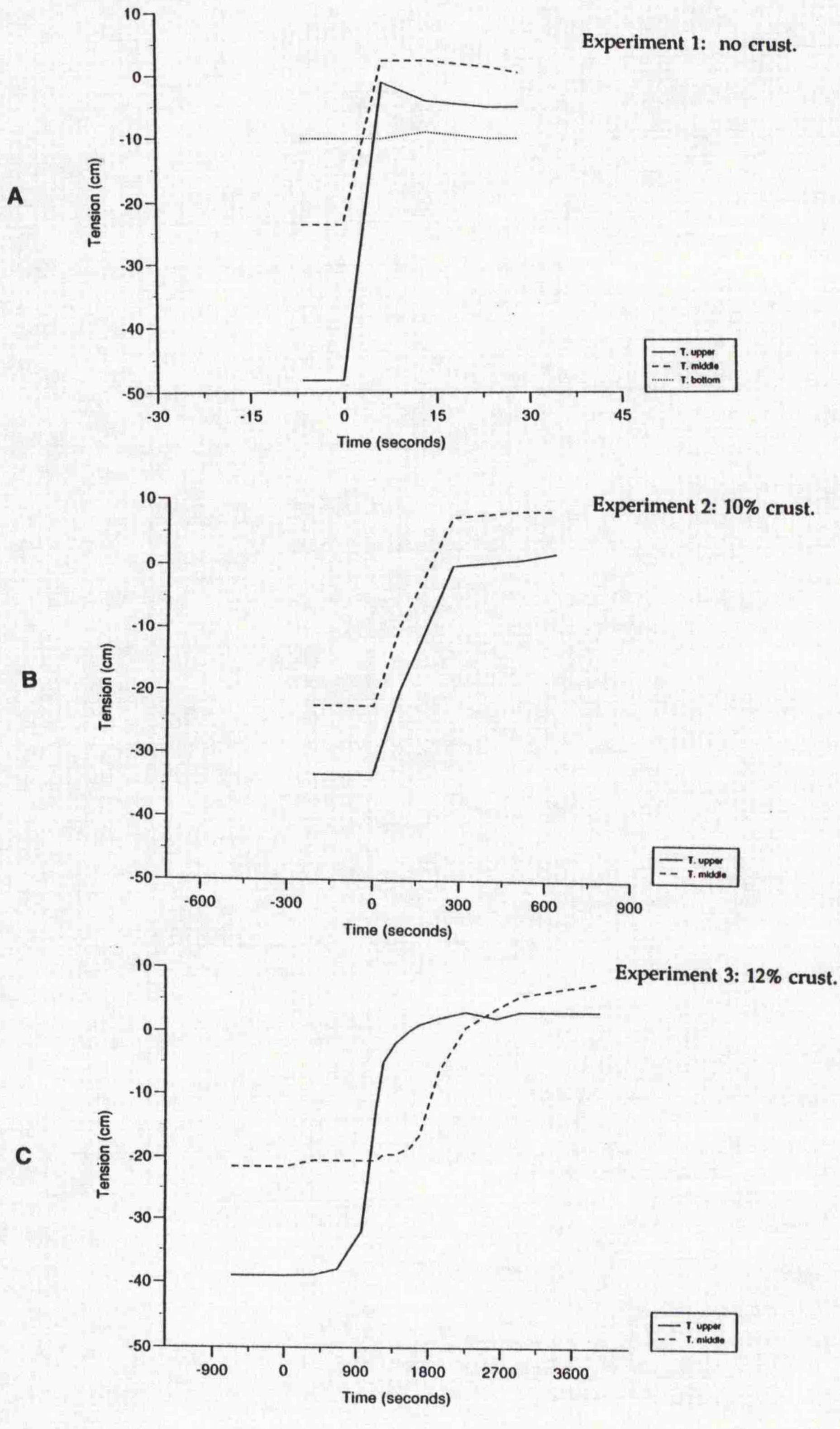
The tensiometers were monitored before and after the experiment and the readings, corrected for elevation head and drift, are shown in Figure 6.17.A. The data are also summarised in Table 6.4. The tensiometers recorded a slight increase in suction caused by drying of the column, before the experiment began at time t=0, T_u recorded a tension of -48cm, at T_m -23.5cm, and T_l -10cm. Following application of water at the column surface the top two tensiometers recorded a decrease in tension within 10 seconds, T_u fell from -48cm to -1.0cm, and T_m from -23.5cm to +2.5cm, while T_l initially remained at -10cm but decreased slowly with time. This indicates significant lateral water flow from the column so that positive water pressures were only recorded towards the top of the column. After the application of water had ceased the tensions at T_u and T_m decreased at a constant rate, T_u to -6.0cm, T_m to -0.5cm, while T_l decreased at time t=0.5 to -4cm.

ii. Gypsum Crust (10% gypsum and fine sand).

Following the ponded infiltration experiment, the surface of the column was levelled firstly by adding gravel coarser than 4mm in diameter, and intermediate hollows were infilled with gravel of diameter 2-4mm. A gypsum crust was then applied to the surface consisting of 90% sand (0.5-2mm in diameter) and 10% gypsum. The crust covered the peat surface completely and the mixture was well compacted to prevent water flows bypassing the crust. The final crust thickness was c. 1cm over the column surface.

The experiment began 22 hours after the first experiment ceased allowing time for drainage of water. Soil water tensions are shown in Figure 6.17.B, and

Figure 6.17. Tensiometer time series, for experiment 1, no crust (A); experiment 2, 10% crust (B); and experiment 3, 12% crust (C).



the results are summarised in Table 6.4. Before water was added to the column tensions of T_u and T_m were stable at -33.5cm and -22.5cm respectively. Tensions decreased sharply as the infiltration of water began, T_u recording a potential of 0cm and T_m, +7.5cm at time 300 seconds after the experiment began. Subsequently, both tensiometers varied only slightly at between 0cm and ±2cm for T_u and +7.5cm to +10.5cm for T_m until ponding of water ceased at time 1440 seconds. At this point tensions became more constant and decreased slightly until measurements ceased at time 48 minutes when T_m was stable at +2.5cm and T_u at -2.5cm. During the experiment the flux of water was maintained at an equivalent rate of 2077 cm/day. Readings at tensiometer T_l were continued, but did not vary significantly and are not shown on the graph. This is probably a consequence of a local increase in the water-table following the previous experiment and reflects the much lower hydraulic conductivity in the wood peat at the base of the column.

iii. Gypsum Crust (12% gypsum).

The experiment was repeated 24 hours later with a crust consisting of 12% gypsum by volume. Tensions recorded during the experiment are shown in Figure 6.17.b. Readings were initially stable, T_u at -39.0cm, and T_m at -21.5cm at time 0 seconds. A head of water was introduced over the soil column at time t = 0 and was maintained until t = 5040 seconds, when ponding of water ceased. As the Figure shows, T_u first responded to water infiltration as tensions decreased slowly to -38.0cm at 660 seconds and then more rapidly to -32.0cm at 900 seconds and -5.0cm at 1200 seconds. The wetting front passed the tensiometer at T_m, 600 seconds later than tensiometer T_u, indicating the time taken for water to travel the 23.5cm separating the two tensiometers. Thus between 1700 and 1800 seconds, soil water tension at T_m decreased from -18.5cm to -6.5cm. Tensions at T_m then decreased further until positive water pressures were recorded, +6.5cm at 2700 seconds. At this time the flux of water was 205 cm/day.

Table 6.4. Summary statistics for the column experiment.

experiment	q_z cm/day	$T_{wf,u}$ sec	$T_{wf,m}$ sec	$h_{p,u,i}$ cm	$h_{p,u,f}$ cm	$h_{p,m,i}$ cm	$h_{p,m,f}$ cm
no crust	- 4320	24	24	- 48.0	- 1.0	- 23.5	+ 2.5
10 % gypsum	- 2080	300	300	- 33.5	- 1.5	- 22.5	+ 9.0
12 % gypsum	- 205	1320	2460	- 39.0	+ 3.0	- 1.5	+ 7.0
20 % gypsum	- 100	660	?	- 2.0	+ 1.0	- 1.0	+ 2.5

where q_z is flux of water; $T_{wf,u}$ and $T_{wf,m}$ are the times of passing of wetting front for the upper and middle tensiometers; $h_{p,u,i}$ and $h_{p,u,f}$ are the initial and final pressure heads for the upper tensiometer; and $h_{p,m,i}$ and $h_{p,m,f}$ are the corresponding pressure heads for the middle tensiometer.

iv. Gypsum Crust (20 %).

The crust used in experiment III was removed and a crust composed of 990g sand (2-4mm diameter) and 161g of gypsum was added to the gravel base at the top of the peat column. Precipitation in the preceding week totalled 30mm and produced a rise in the water-table of 45cm. Experimental conditions were therefore significantly different to those in the other experiments but it was decided to continue to determine whether unsaturated conditions would persist given a very small water flux applied at the column surface.

Limitations with equipment only permitted the monitoring of one tensiometer at any time. Consequently, before water was applied to the surface, tensions at T_u and T_m were recorded and the pressure transducer was left attached to the top tensiometer until the reading had stabilised following the passage of the wetting front through the peat column.

Tensiometer readings were initially stable, T_u at -2cm and T_m at -1cm. The top tensiometer, T_u decreased to 0 at time 900 seconds after a head of water was first applied to the top of the column, and stabilised at +1cm 600 seconds later. The pressure transducer was connected to T_m 3660 seconds after the experiment began, at which time the reading was +2.5cm. The flux of water through the column remained constant at 100.8 cm/day.

6.3.4. Discussion.

The results from the four experiments are summarised in Table 6.4. The table gives values of hydraulic conductivity obtained by applying Darcy's Law and using the uncorrected values of water flux applied at the column top; also given are the times for the wetting front to pass the upper and middle tensiometers, and the initial and final pressure heads for the upper and middle tensiometers. Recordings of controlled infiltration through crusts of differing hydraulic resistances have been used to calculate hydraulic conductivities. In the experiments, and especially for those with a high water flux, a significant quantity of water bypassed the peat matrix and passed quickly through the column within structural voids, or macro-pores. Without accurate determination of the quantity of water lost, calculations of hydraulic conductivities using these data must be open to question. Hydraulic conductivities were computed for the four experiments and values of 4320, 2080, 205, and 100 cm/day were recorded, on the same column of peat. The values were thus directly proportional to the water flux which was allowed to flow through the column.

One problem is with the possibility of translocation of sedimentary particles through the column, which might block macro-pores during the course of the experiment. Peat is also a compressible medium; permeabilities and porosity will therefore vary depending upon the history of compression. In this experiment any translocation and blocking of pores are most likely to have occurred during the first experiment, when a very high water flux was applied to the surface. Significant differences were observed in the results of the following experiments, indicating that the operation of other mechanisms, in addition to translocation of soil particles, has to be invoked to explain variations in the potential gradient.

A further question is whether a different model for subsurface water flow should be applied to the upper layer of peat due to a higher clay content, with attendant possibility of volumetric changes. Hemond and Goldman (1985) considered that peat structure varied with hydraulic head through changes in pore geometry, however, they concluded that Darcy's Law remained valid. The important point was that observations which deviated from Darcy's Law occurred where unrealistic conditions were imposed, for instance a very large hydraulic gradient. Under natural, steady state conditions with low hydraulic gradient Darcy's Law may well still apply.

An alternative framework for interpreting the results, separate from the concept of hydraulic conductivity, is provided by analysing the temporal variations in potential in comparison with the theoretical hydraulic gradients for a two layered soil, discussed in section 6.2.2. and shown in Figure 6.15. This information may be used to speculate whether different mechanisms of water flow through the peat matrix may explain the variable results. Figures 6.18. and 6.19. show the variation in elevation head, pressure head, and hydraulic head at selected time intervals for experiments 2 and 3, with gypsum crusts of 10% and 12%, and are given in the same format as the graphs in Figure 6.15. In both graphs the gradient of hydraulic head indicates that the initial conditions consist of downward drainage in the top peat layer, although in the bottom layer the hydraulic head between depths of 52cm and 72cm is approximately constant, suggesting equilibrium conditions of no water flow and saturation. Within a short time the water flux is sufficient to produce downward water flow throughout the column. In Figure 6.18. this change occurs within 180 seconds, and by 300 seconds the variation in hydraulic head has assumed a concave profile, similar to the curve shown in Figure 6.15. In Figure 6.19., under a smaller water flux, this transition takes longer; from t1 to t4 (0 seconds to 1680 seconds) the upper gradient in hydraulic head becomes increasingly steeper to a maximum gradient of 55cm potential per 20cm depth at time t4. Tensions in the middle tensiometer then decrease sharply, recording a small positive pressure at 3920 seconds, producing a more uniform potential gradient through the column, with a slightly concave profile.

In both Figure 6.18. and 6.19. the final variation in pressure head

Figure 6.18. Variation of pressure head and hydraulic head with time in experiment 2.

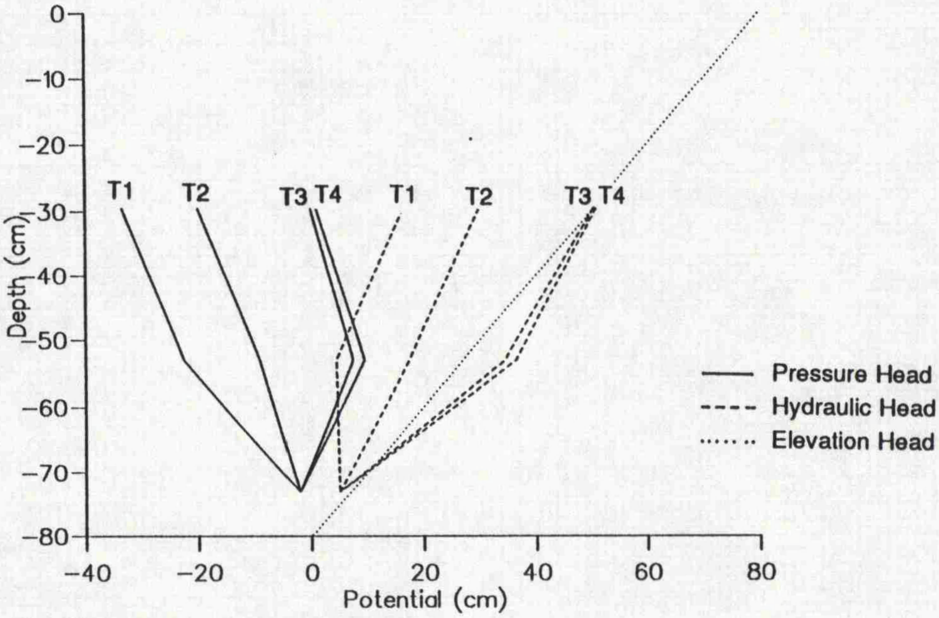
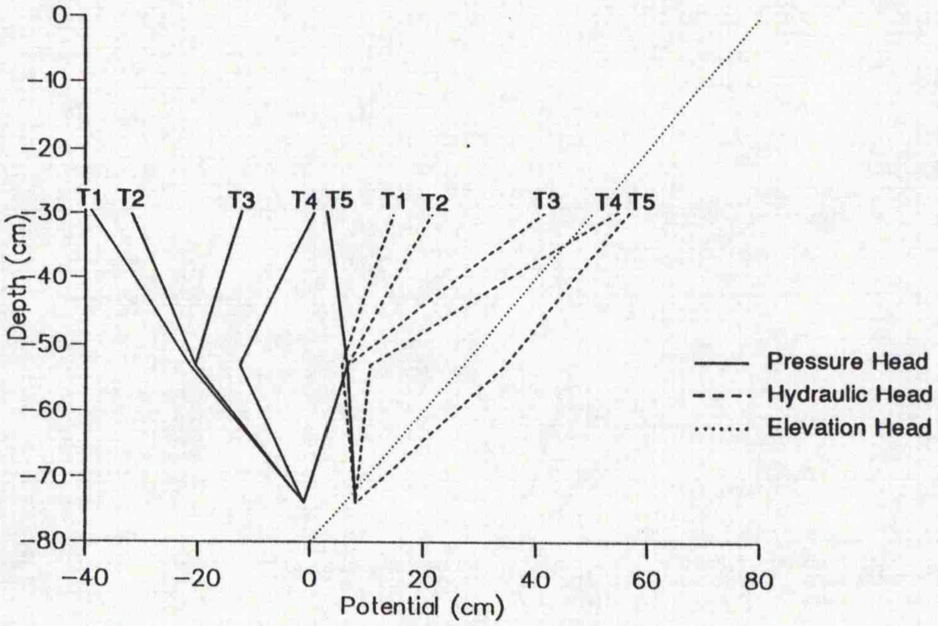


Figure 6.19 Temporal variation of pressure head and hydraulic head in experiment 3.



indicates an initial increase in pressure head with depth in the upper layer followed by a decrease with depth in the lower area; this explains the higher pressures recorded in the middle tensiometer, compared with the top tensiometer, at the conclusion of all four experiments (Table 6.3.).

The observations of changing hydraulic heads can therefore be explained by the varying differences in permeability through the peat profile. The implication is that studies on a small scale, such as described here, provide an inadequate representation of larger scale hydrological processes which a wetland system will experience. The imposition of large water fluxes is unrealistic in this case, as the natural equivalent, overbank flooding, has greater areal extent, thus reducing the significance of lateral water flow through macro-pores as observed at the column sides in the experiment.

The results also indicate the limitations of published methods to determine the hydraulic conductivity of peat deposits. Vertical and horizontal differences in the composition of peat make it very difficult to identify a suitable average permeability, and raise questions concerning the scale of the investigation. Hydrological modelling at a large scale requires information on broad patterns of water-table response, in which case Darcy's Law may remain valid. However, at the micro-scale the Darcian assumptions may not be valid. This can be illustrated with reference to Figure 6.20. (from Domenico and Schwartz, 1990), which shows how observations of hydraulic conductivity may vary with the scale of study and the size of the representative volume. At the micro-scale conductivities are highly variable, although they may become more stable as volume increases. Beyond a certain point there is a discontinuity as hydraulic conductivity changes, in this case possibly through differing proportions of water flow through macro-pore structures in the peat. In the experiment the extreme variation in calculated hydraulic conductivities was found to be proportional to the applied water flux; this determined the scale of water flow through the peat matrix and is thus analogous to the considerations of volume size illustrated in Figure 6.20.

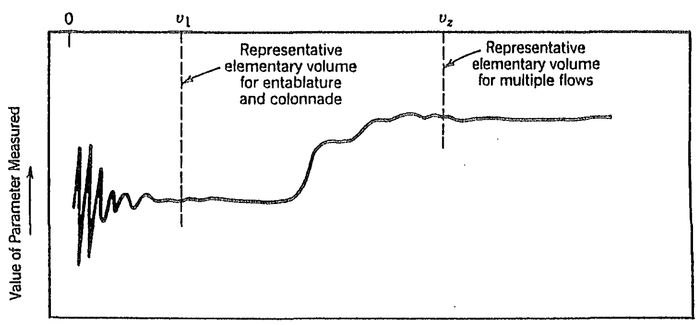


Figure 6.20 Possible change in hydraulic conductivity with scale of measurement (from Domenico and Schwartz, 1990).

6.4. LABORATORY MEASUREMENTS OF HYDRAULIC CONDUCTIVITY.

Although greatest emphasis was placed upon measurements of peat hydrology in the field as described in the experiment above, hydraulic conductivity was also determined by laboratory studies of water flux through samples of peat. Cores of wood peat of length 10 cm were collected from a depth of 70cm in a soil pit at Narborough near dipwell 12.

The cores were maintained in a saturated condition, and a constant head device was used to apply a small head of water at the top of the core, having secured the base with cheesecloth. The flux of water through the core was measured, and hydraulic conductivity could then be calculated using Darcy's Law:

$$q = K\frac{\delta H}{\delta x} \qquad (3)$$

where q is the water flux, K is hydraulic conductivity, and $\delta H/\delta x$ is the gradient of hydraulic head.

The water used for the experiment was a dearated 0.005 M $CaSO_4$ solution, as recommended by Klute and Dirksen (1986). This limited the possibility for dissolved gases to come out of solution and block soil pores, which would produce a decrease in hydraulic conductivity with time.

The results are given in Table 6.5. for the three samples of wood peat on which experiments were undertaken. There was some variability in measured hydraulic conductivity both within and between the sample cores, and it appeared that a certain time was required before a stable water flux could be maintained through each core. Calculated values for hydraulic conductivity varied by a factor of 6 from 0.034 to 0.197 m/day, indicating significant variation although the cores were collected from the same deposit. At the scale

Table 6.5. Measurements of K_{sat} on samples of wood peat.

Time. (hrs)	Infiltration Rate (ml/s)	Saturated K (m/day)
Sample 1.		
6.00	0.0029	0.078
21.00	0.0050	0.135
mean	0.0040	0.107
st. dev.	0.0009	0.023
Sample 2.		
6.00	0.0057	0.155
21.00	0.0070	0.192
26.00	0.0070	0.197
45.00	0.0053	0.154
53.00	0.0054	0.145
mean	0.0061	0.167
st. dev.	0.0007	0.022
Sample 3.		
8.00	0.0022	0.061
24.00	0.0012	0.034
32.00	0.0026	0.073
48.00	0.0025	0.070
mean	0.0021	0.060
st. dev.	0.0006	0.015

of the individual sample, there were differing proportions of vegetation roots which would account for some of the variability, modifying the bulk density and influencing the water flow through the core.

6.5. PIEZOMETER TESTS TO DETERMINE HYDRAULIC CONDUCTIVITY.

The results of the column experiment illustrate the wide variability of rates of water flow, especially within a flow system in which water may be routed quickly through a macro-pore network, or more slowly through the peat matrix. Thus, before considering how parameters may be derived for hydrological modelling, in this section the application of an additional method of determining hydraulic conductivity is described. The study of water flow through wetland deposits has traditionally been undertaken using variations of the seepage tube, or auger hole method. Both methods consist firstly of depressing the water-table, and then monitoring the rate with which the water-table returns to its initial position. Hydraulic conductivity is determined by considering the flow dynamics which govern the rate of recovery. In the auger hole method the hole remains unlined, thus producing a measure of vertical hydraulic conductivity, while the seepage tube approach uses a lined hole, giving horizontal hydraulic conductivity at a depth equal to the base of the tube.

The equation to calculate hydraulic conductivity was introduced in section 6.2.2. (equation 2). Hydraulic conductivity was determined at two dipwells, numbers 6 and 2, using the seepage tube method. These dipwells had the same diameter and were installed at the same time, enabling an investigation of spatial variation in the rate of water-table recovery. Water was evacuated to a depth of 1.5 m using a hand-held bilge pump, and the recovery of the water-table monitored. The water-table rise was extremely rapid, requiring initial measurements at a 30 second interval. A time-series plot showing the change in water-table depth for the two dipwells is shown in Figure 6.21. The results are summarised in Table 6.6; the shape factor was

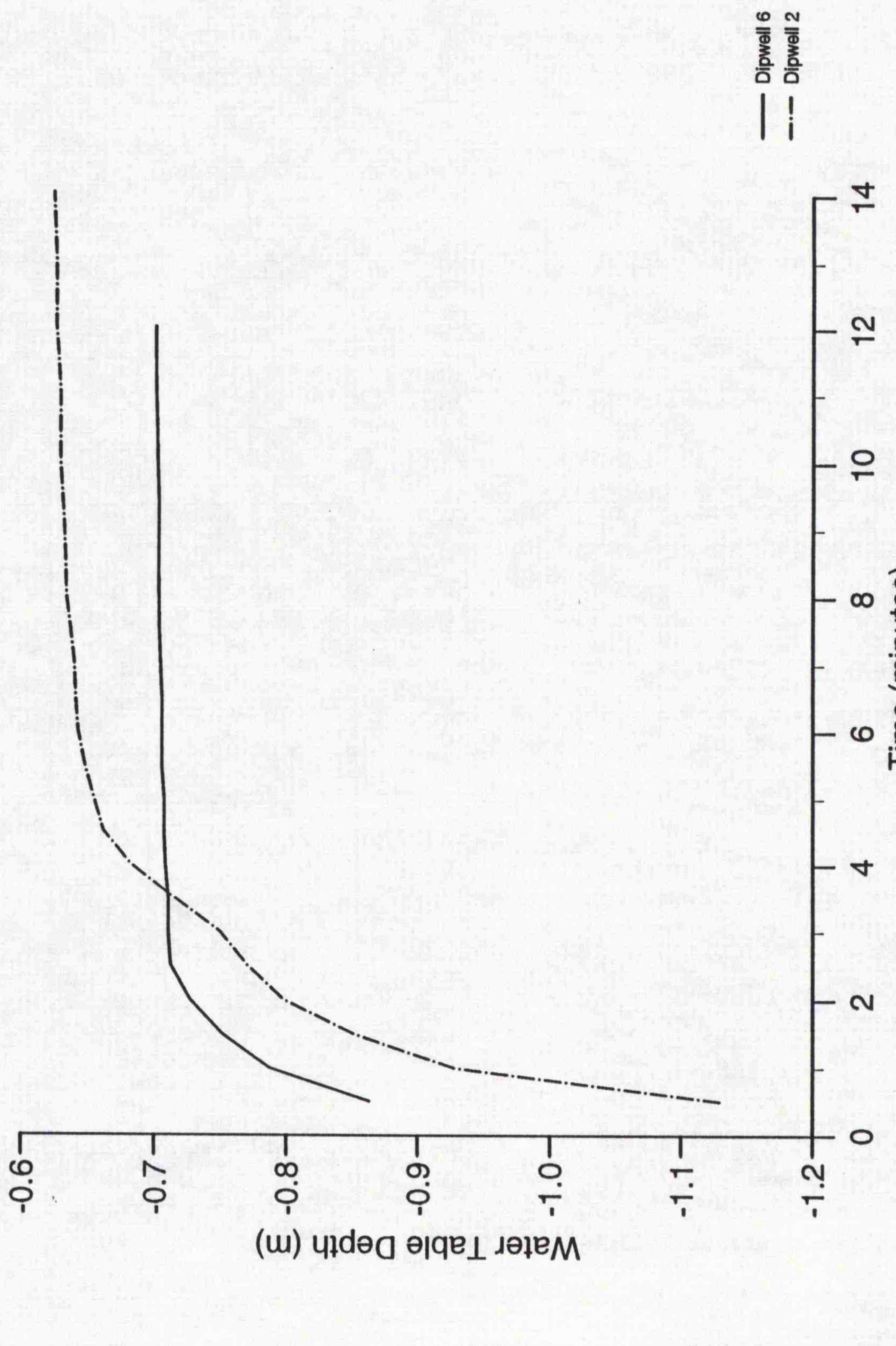
Table 6.6. Summary Data for Seepage tube experiment.

	Dipwell 2	Dipwell 6
Initial water-table depth (m).	0.610	0.699
Depth to base of dipwell (m).	1.630	1.303
Radius of tube (m).	0.024	0.024
head difference at time 1 (m).	0.519	0.164
head difference at time 2 (m).	0.051	0.013
time interval (s)	270	150
Shape factor	30.8	30.8
Hydraulic Conductivity (m/s)	5.049×10^{-7}	9.929×10^{-7}
Hydraulic Conductivity (m/day)	0.0436	0.0858

determined using a table given by Amoozegar and Warrick (1986; Table 29-2). It was assumed that the dipwells overlaid a permeable boundary as they were installed on top of the alluvial gravel deposits. The difference in head between water in the two tubes and the water-table was determined from the initial measurement, to the point at which the curve of water-table change the rate of change stabilised. This occurred at 4.5 minutes at dipwell 2, and 1.5 minutes at dipwell 6.

The results from the test indicate hydraulic conductivities at the two dipwells sites of between 0.086 and 0.044 m/day. These values lie within the range of hydraulic conductivities measured in the laboratory which were summarised in Table 6.5., although they are lower than the mean values recorded. There is potentially a larger error in determining hydraulic conductivity by this method. The initial rise in water-tables was very rapid, and the water level within the dipwells was lowered to differing amounts in the two tests, thus affecting the difference in heads (h_i) at the first time interval.

Figure 6.21. Water-table recovery curve for dipwells 2 and 6 in the seepage tube test.



6.6. CONCLUSION.

This chapter has discussed several experiments of water flow through sedimentary assemblages at Narborough. These consisted of two studies which examined spatial variations in water flux, and also a summary of the results of applying two recognised techniques to determine hydraulic conductivity. The chapter began by looking at the stage records of the river Soar, and considered the changing frequencies of overbank flooding over the last twenty years. The first experiment then examined the processes likely to operate during water infiltration in an overbank flood event. This is followed by a study of the detailed pattern of water flow through a peat column consisting of two identifiable layers of different hydraulic conductivity.

The results from the study of controlled flooding indicated the possible significance of small variations in stratigraphy, in which clay layers alternated with an herbaceous peat soil and peat lenses at different depths. The hydraulic conductivity of these sediments may differ by several orders of magnitude, and the implications this presents for the direction of water flow need to be considered. Dipwell response to flooding was variable, illustrating the difficulty in obtaining a representative measure of water-table in this area. The data revealed that areas of saturation could be locally concentrated, as low permeability clay deposits increase the time required for water-tables to reach an equilibrium state.

The study of water infiltration through the peat column demonstrated the scale problems in measuring hydraulic conductivity, however, the hydraulic head results were shown to follow the expected relationship for theoretical infiltration into a layered soil profile. A useful measure of hydraulic conductivity is therefore hard to identify, which reflects both the nature of Darcy's Law describing an empirical relationship, and also differences in the characteristic processes of water flow through peat deposits.

The last two experiments provide a clearer indication of what values of hydraulic conductivity may be used to describe water flow through the wood peat layer. The laboratory results indicate large variability in the conductivity of samples which were collected within a $1m^2$ area, however, there is less variation in the values obtained from the seepage tube test. It is significant that while the seepage tube test is measuring horizontal hydraulic conductivities, the laboratory cores were collected vertically and so represent vertical hydraulic conductivities. Anisotropy is normally assumed in the horizontal and vertical directions in peat deposits, however, the wood peat was compressed, with little internal structure preserved, and it is possible that the peat samples were disturbed when sampling.

The aim of the specific studies, discussed in this chapter, has been to examine the variability of water flows through different deposits at Narborough Bog, before proceeding to apply a groundwater model to the site in chapter 7. The hydraulic conductivities of deposits have varied from a maximum rate of $2 \rightarrow 4$ m/day (Table 6.4.) for peat in the column experiment, with no crust and 10% gypsum crust. The flood experiment revealed much smaller conductivities in silt/clay deposits adjacent to the river, which were within the range 7.9×10^{-3} to 2.9×10^{-5} m/day (Table 6.2.). The piezometer and laboratory tests were undertaken on wood peat deposits. In this case, hydraulic conductivities of between 0.0439 m/day and 0.169 m/day were obtained. Estimates of conductivity based upon these values are used in the development of the groundwater model in the following chapter.

Chapter 7

Groundwater Modelling

Scope of Chapter

This chapter discusses the hydrological modelling of water-table response at Narborough Bog. The chapter begins by introducing groundwater modelling techniques, and then reviews selected modelling studies in floodplain and wetland environments. The groundwater modelling package, MODFLOW, is described, and the program structure and input data requirements are detailed. The chapter then proceeds to develop a groundwater model of Narborough Bog. The production of a generalised hydrogeology of the site is described first, and then attempts to calibrate the model to steady state and transient conditions are described. Finally, the model results are compared with the field data of water-table variations, to assess the accuracy of the hydrological simulations.

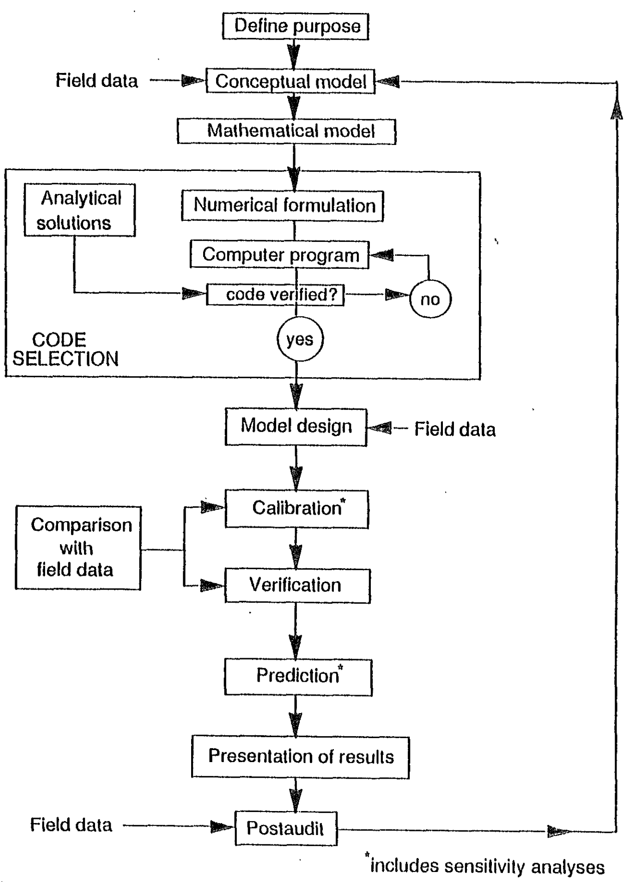
7.1. INTRODUCTION

In chapter 5, the response of water-tables at Narborough Bog to periods of precipitation and evapotranspiration was examined. The quantitative analysis demonstrated that water-tables respond in a predictable manner to different events, however, the techniques used in chapter 5 provide little information on the processes of water flow within the heterogeneous deposits at Narborough. The experiments described in chapter 6 illustrate the variability of water flux through different deposits at Narborough Bog, which complicate the interpretation of how water flows through the wetland system. In this chapter a three-dimensional groundwater model, MODFLOW, is applied to the field-site at Narborough Bog to investigate variations in water flow through different deposits. The ability of the model to reproduce water-table fluctuations in response to precipitation and evapotranspiration are considered, and the results are used to assess the flow of water within the wetland. In addition, a comparison of model predictions with field data may indicate whether all components of the water budget have been identified, and also if representative values of hydraulic conductivity are being used. A fully calibrated groundwater model should enable prediction of the hydrological effects of any future change in individual components of the water budget, within a given error band, thereby permitting better management of groundwater resources.

Successful hydrological modelling is dependent upon the input of satisfactory field data, as illustrated by the flow chart in Figure 7.1. (from Anderson and Woessner, 1992). The Figure illustrates how field data should be used to derive a generalised 'conceptual' model of the area. The box covering code selection includes elements such as model verification which have normally already been undertaken in the most widely used models. Model output should then be compared with field results, when the accuracy of the conceptual model of the field site and choice of flow parameters can be assessed. A process of optimisation then follows in which a validated, calibrated model is considered to be the final objective.

Two types of mathematical approach have been applied to groundwater analysis, namely the analytical and the numerical model. Analytical models attempt to solve the equations governing water flow, by simplifying the representation of hydraulic properties and the initial and boundary conditions. The results illustrate the dependence of water flow upon generalised water flow and soil characteristics which have been specified for the field site. In contrast, numerical models use mathematical flow equations to describe the physical forces which determine subsurface water flow. Numerical models are less constrained by the choice of the initial and boundary conditions. In a comparison of the two different modelling techniques, Peters (1987) considered that analytical models are most useful for simple lithology and boundary conditions, while numerical models are more appropriate in areas of complex lithology or where the effects of interactive processes need to be considered.

Figure 7.1. Flow chart illustrating the steps to be followed in a groundwater modelling study (from Anderson and Woessner, 1992).



In numerical modelling, a discrete representation of the flow field is produced. The region is divided into a number of cells, each having given hydraulic properties, and a node within each cell where hydraulic head is defined. Water flow is determined by equations derived from a combination of Darcy's Law and the continuity law, which specifies that the rate of flow into an individual cell equals the rate of change of water storage within the cell. Two types of numerical method have been widely used, finite difference and finite element analysis. These differ in the type of grid which is superimposed on the study area. In the finite difference approach, the two dimensional study area is described using a rectangular x,y grid, with constant cell dimensions. In the finite element method, a flexible grid replaces the rigid x,y grid, enabling a more complicated representation of the field area. In this method the nodes form the corner points of either an irregular triangular, or quadrilateral network. This enables the possibility of a smaller network with a corresponding reduction in computing time, and may better represent the data available.

Groundwater models have been applied both within a steady-state, and a transient time-frame. Transient simulations are used to solve time-dependent problems. This requires introducing a factor accounting for internal changes in water storage, described by storativity, either as specific storage (S_s), storage coefficient (S), or specific yield. Specific storage is most commonly used; it is defined as the volume of water released from storage by a unit volume of aquifer under a unit decline in hydraulic head (Freeze and Cherry, 1979, p. 58). Lowering of water levels within an unconfined aquifer involves the release of water from storage.

7.2. EXAMPLES OF GROUNDWATER MODELLING STUDIES.

Within floodplains, modelling the interaction between groundwater flow and rivers presents special problems, as their relationship may vary through time. The river may represent either a source of groundwater recharge, or groundwater discharge, depending upon the near-stream hydraulic gradient. A brief description of influent and effluent flow processes in relation to wetlands was given in section 2.4.3.iii. The implications of differences in hydraulic conductivity are illustrated by Freeze and Witherspoon (1966, 1967), who considered the combined effects of localised deposits of sand and gravel with high hydraulic conductivity and anisotropy. Although the papers are mainly theoretical, they demonstrate the practicability of applying numerical techniques within a floodplain environment, and illustrate how groundwater flow varies depending upon water-table configuration and stratigraphy.

Field data indicating alterations in these conditions are sparse, however, several theoretical solutions have been published. For example, Todd (1955) used a Hele-Shaw flow model to look at seepage flows at the Sacramento River, while Cooper and Rorabaugh (1963) described the mathematics underlying temporary bank storage of groundwater, and variation in groundwater flow with distance from the river. An early groundwater model, which specifically considered wetland hydrology, was described by Sander (1976), who constructed a physical analogue of a small peat bog using electrical components. However, models of this type have been superseded by more sophisticated deterministic hydrological models, and a selection of papers using these approaches are summarised in Table 7.1. The papers include studies using three different groundwater models. Gilvear *et al.* (1993), Hensel and Miller (1991) and McNamara *et al.* (1992) all used MODFLOW for investigations of wetland hydrology. Siegel (1981; 1983), and Sophocleous (1991a) used an earlier version of MODFLOW (Trescott *et al.*, 1976) which had a two-dimensional structure. Schot and Molenaar (1992) and Wassen *et al.* (1990a) used a two-dimensional profile model FLOWNET.

Several points are raised by this selection of published studies. Firstly, modelling techniques have been applied at a considerable range of scales, from a 200km transect described by Siegel (1983), to a 12 hectare wetland (McNamara *et al.* 1992). The resolution of individual cells varies similarly, from a spacing

Table 7.1. Summary of Papers including a Groundwater Modelling element.

Paper	PACKAGE		SCALE	COMMENTS
Gilvear *et al.* (1993)	MODFLOW	Steady State	i. 17.0 x 8.0 km. (15 x 17 nodes) ii. 3.5 x 3.0 km. (1050 ha) (250m and 125m)	Simulation of Groundwater contribution to a small wetland. Includes sensitivity analysis of effects of varying parameters.
Hensel and Miller (1991)	MODFLOW	Steady State	40 x 60 nodes, spacing 28m	Study of wetlands on the floodplain of the Des Plaines River, Illinois, to examine effect of wetland creation on effluent flows.
McNamara *et al.* (1992)	MODFLOW	Steady State	20 x 20 grid (12 hectares)	Modelling of water flow through a wetland in a small kettle hole, New York, USA.
Yager (1986)	MODFLOW	Steady State and Transient	42 columns x 50 rows (176 ha)	Examination of variable relationship between groundwater flow and infiltration from Susquehanna River, New York.
Schot and Molenaar (1992)	FLOWNET Profile model	Steady State	270 columns x 16 rows	Study of changing hydrology of wetlands on the Vecht floodplain from 14th Century to present.
Siegel (1981)	Trescott	Steady State		Simulation of two-dimensional water flow in the Glacial Lake Agassiz peatlands.
Siegel (1983)	Trescott	Steady State	200 km transect	Studied the relationship of water-table mounds under raised bogs to adjacent fen systems.
Sophocleous (1991)	Trescott	Steady State	4.3 x 6.7 km	Two-dimensional groundwater model used to study water flux between surface water and groundwater in the Great Bend Prairie, Kansas.
Wassen *et al.* (1990a)	FLOWNET	Steady State	Cells: 250m (Horizontal); 5m (vertical)	Simulation of effects of changing water managements on wetlands on the Vecht floodplain.

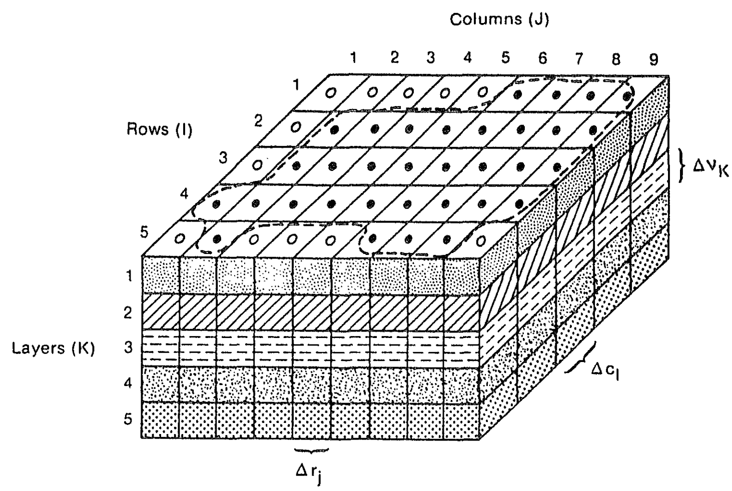
of 28m to dimensions of 1km^2. Secondly, all the studies sought to obtain steady state solutions to groundwater flow problems, and the possibility of transient effects is not considered, except by Yager (1986). Thirdly, only Yager (1986) assessed model error by calculating the root mean square (RMS) error between field hydraulic head and model simulations, although packages such as MODFLOW include a calculation of the water balance error within the model output. Fourthly, only Gilvear *et al.* (1993) and Yager (1986) describe the results of a sensitivity analysis whereby the effects of variations in parameter values were investigated. Lastly, although not indicated in Table 7.1., several of the papers did not determine hydraulic conductivity in the field, but used values published in the literature. It is debatable whether this is sufficient given the variable permeability of wetland and floodplain deposits. Indeed, the validity of categorising heterogeneous wetland deposits by constant flow parameters required for hydrological modelling has still to be investigated satisfactorily.

In summary, the studies described above mainly described theoretical simulations on the basis of monthly water-table observations. In this chapter the detailed measurements of water levels, which were described in chapter 5, are used to assess the development of a groundwater model for the field site at Narborough Bog.

7.3. DESCRIPTION OF MODFLOW.

MODFLOW is a quasi-three dimensional groundwater model developed by McDonald and Harbaugh (1988) and used extensively by the United States Geological Survey. The program uses a finite difference method whereby a rectangular, two-dimensional grid, of **x** rows and **y** columns, is superimposed over the study area as illustrated in Figure 7.2. A vertical dimension is provided using layers to represent aquifers, which may be confined, unconfined, or a combination of both. MODFLOW represents a quasi-three dimensional model because, while the **x** and **y** dimensions of the grid are specified, the dimensions of the vertical grid are derived indirectly from flow

Figure 7.2. Configuration of sample flow block, illustrating notation, and the approach to describing water flux (from McDonald and Harbaugh, 1988).



Explanation

---- Aquifer Boundary

● Active Cell

o Inactive Cell

Δr_J Dimension of Cell Along the Row Direction. Subscript (J) Indicates the Number of the Column

Δc_I Dimension of Cell Along the Column Direction. Subscript (I) Indicates the Number of the Row

Δv_K Dimension of the Cell Along the Vertical Direction. Subscript (K) Indicates the Number of the Layer parameters which are proportional to layer thickness. Thus layer transmissivity, which equals hydraulic conductivity multiplied by layer thickness, is given. Consequently, the model is most appropriate for simple examples where distinct aquifers can be identified, particularly hydrostratigraphic layers in which flow properties remain constant (Maxey, 1964). The principal limitation of the finite difference method is the requirement that significant variations in hydraulic conductivity only occur in directions parallel to the coordinate axes. This implies in the direction of the x and y axes.

MODFLOW has a modular structure, comprising ten optional packages which consider either specific hydrological processes or details of water flow. Input for each module is specified in individual files, which are distinguished by the file extension. The consequent structure is illustrated by the flow diagram in Figure 7.3. in which sample file names are given. Further details on the individual packages are given in Table 7.2., where functions of the modules are outlined. The Basic Package (narl.bas) forms the core of the model; here the dimensions of the grid are defined and the choice of individual packages specified. The Block-centred Flow Package (narl.bcg), and the Solver Package (narl.sor) are also central, as they determine water flow between nodes and solve the iterative equation for hydraulic head. Inputs of water are defined by the Recharge Package (narl.rch) and, under certain conditions, the River Module (narl.riv). Water outputs consist of evapotranspiration (narl.evp) and the river (narl.riv). The format of results is defined by an Output Control Module (narl.oc), while a listing of results is sent to a specific file (narl.lst), and contour plots of either head or drawdown within individual layer are stored in additional files (narl.hd and narl.dd).

Cells within each layer are classified as either variable head, constant head or inactive cells. Only the hydraulic head for variable head cells is calculated in the model. Inactive cells are used to define the study area by identifying areas of the rectangular grid which lie outside the dimensions of the study area, thereby enabling irregular areas to be modelled using a rectangular

Figure 7.3. Illustration of model structure and file notation.

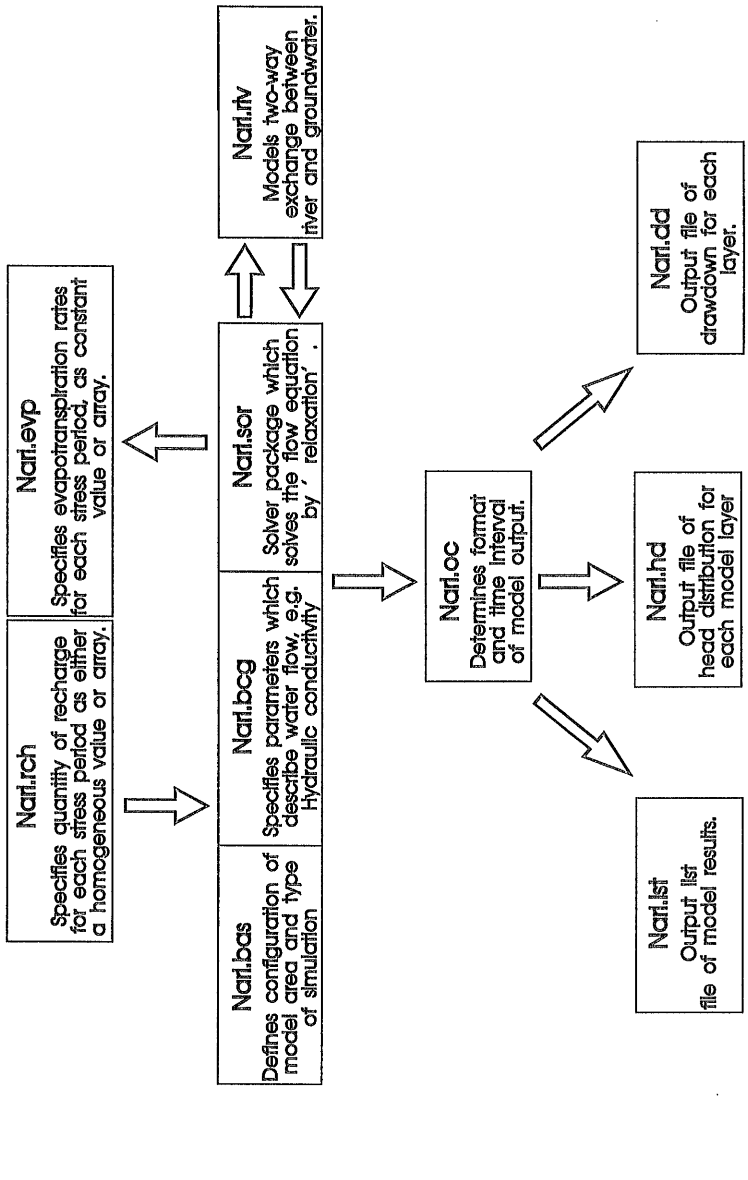


Table 7.2. Characteristics of MODFLOW modules.

PACKAGE	COMMENTS
Basic Package	This is the core of MODFLOW; the model size, simulation type, number of stress periods, and hydrological options are specified. The boundary conditions and initial hydraulic head values are input.
Block-centred Flow Package	In this section water flux between individual cells are calculated. In transient simulations, the flow of water to and from storage is calculated. Vertical flow is determined from a V_{cont} array, which is obtained from the vertical hydraulic conductivities of adjacent layers.
River Package	River leakage is calculated from a head dependant relationship between river stage, and the adjacent water-table. Water flow is proportional to the hydraulic gradient and conductance of river bed sediments. The conductance term, C_{RIV}, is calculated as the product of the hydraulic conductivity of the river bed sediments and the surface area of the cell, divided by the thickness of the sediments.
Recharge Package	Here, the water-table rise through recharge is calculated. Recharge may either be applied to the unconfined layer at the surface of the model, or a specified underlying layer.
Evapotranspiration Package	Evapotranspiration is calculated by assuming linear rates of water loss between two water-table elevations. Evapotranspiration is assumed to equal the specified rate, at a maximum evaporation surface, and ceases when the water-table falls to a cutoff depth.
Solver Package: Strongly Implicit Process.	The simultaneous equation governing the hydraulic head distribution is solved by iteration. The process consists of applying repeated sweeps across the model grid, and applying the equation at every node where the head is unknown. This continues until the residual head is below a specified error.

grid. Constant head cells are cells for which the head is specified initially and whose head does not change during the simulation; these are used for certain types of boundary and for surface water features such as lakes where a constant head value may be assumed.

Vertical and horizontal water flows within the model are calculated from Darcy's Law and the continuity equation, whereby the change in water storage is given by the sum of subsurface flows and net flow to, or from, the surface. The net water inflow represents the difference between water loss, for example through evapotranspiration, and replenishment flow (eg. from precipitation, stream seepage). Hydraulic head at different time intervals is generated at nodal points lying in the centre of grid cells. Water flow is calculated using the standard differential equation, governing groundwater flow in three-dimensions, derived from Darcy's Law (chapter 3, equation 11). Flow parameters, such as hydraulic conductivity, are specified for individual nodes. Hydraulic head is calculated at individual nodes, and thus a series of simultaneous equations, which equal the number of nodes, have to be solved, In this study, the equations are solved by a process called relaxation in the Strongly Implicit Procedure Package. The equation is applied in a consistent process across the model grid, varying the hydraulic head at unknown nodes until the difference in head between iterations is below a specified error.

The model may be used at time-scales of seconds, minutes, hours, days or years, as specified in the Basic Package. The modelling period is divided into stress periods, during which external stresses are constant, and time steps, which are intervals within stress periods. Thus, variations in recharge and evapotranspiration may be varied at fixed intervals. Units of length are not specified, and hence hydraulic conductivity and cell dimensions must be defined using consistent units.

The choice of MODFLOW reflected the versatility of the package, and its ability to consider most of the components of the water budget identified at Narborough Bog, including precipitation and evapotranspiration. Also, the river module enabled both influent and effluent flow to be modelled, while the model could be used in both steady-state and transient conditions. The limitations of MODFLOW are that it is unable to incorporate interception, and that it only considers processes within the saturated zone.

7.4. GROUNDWATER MODELLING AT NARBOROUGH BOG.

7.4.1. Applying the model.

The aims of this section are:

1. To simplify the hydrostratigraphy of Narborough Bog for modelling purposes.

2. To obtain steady-state solutions for a selection of events, the results from which can then be compared with field data.

3. To obtain time-dependent solutions, describing the response of water-tables to a combination of water inflow and outflow.

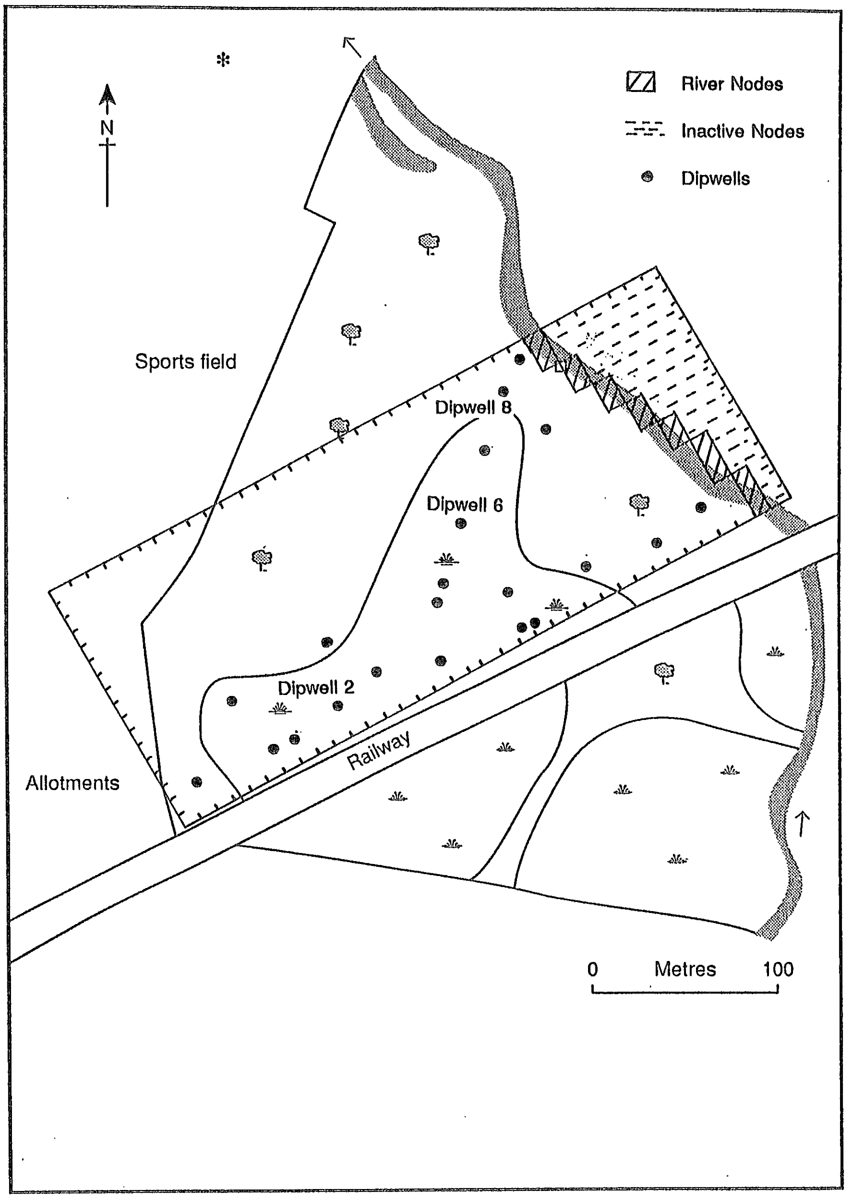
7.4.2. Conceptual Model of Narborough Bog.

Modelling simulations were undertaken using a rectangular grid, centred upon two transects of dipwells which extend from the river Soar through the reed-bed at Narborough Bog. The model grid is illustrated in Figure 7.4., where the relationship of the model grid to existing dipwells and the river Soar can be seen. The grid had dimensions of 360m x 150m, thus representing an area of 5.4 hectares, which comprises 57% of the total area of Narborough Bog. It was decided to use individual model cells of constant size 10m x 10m, hence the grid was composed of a total of 36 rows and 15 columns. This choice reflected the distance between dipwells, which were situated to represent variations in water-table gradient.

The location of individual dipwells within the model area is significant, as the dipwells are used both to define the initial water surface and also provide the control points at which model results are compared with field data. It is clear from Figure 7.4. that dipwell location was not determined with a fixed finite difference grid in mind and reflected the combination of access and an original intention to apply a two-dimensional model along an axis perpendicular to the river. However, by considering the results mainly at dipwell points, greater confidence may be placed in the results.

The stratigraphy of the area covered by the model grid is complex, as discussed in section 4.1.1. and shown on Figure 4.3. Results from the simulated flood experiment discussed in section 6.2.1., indicated that differences in the stratigraphy were responsible for a variable dipwell response to flooding. This illustrates the importance of correct hydrostratigraphic representation. For modelling purposes, the variability in stratigraphy was simplified by

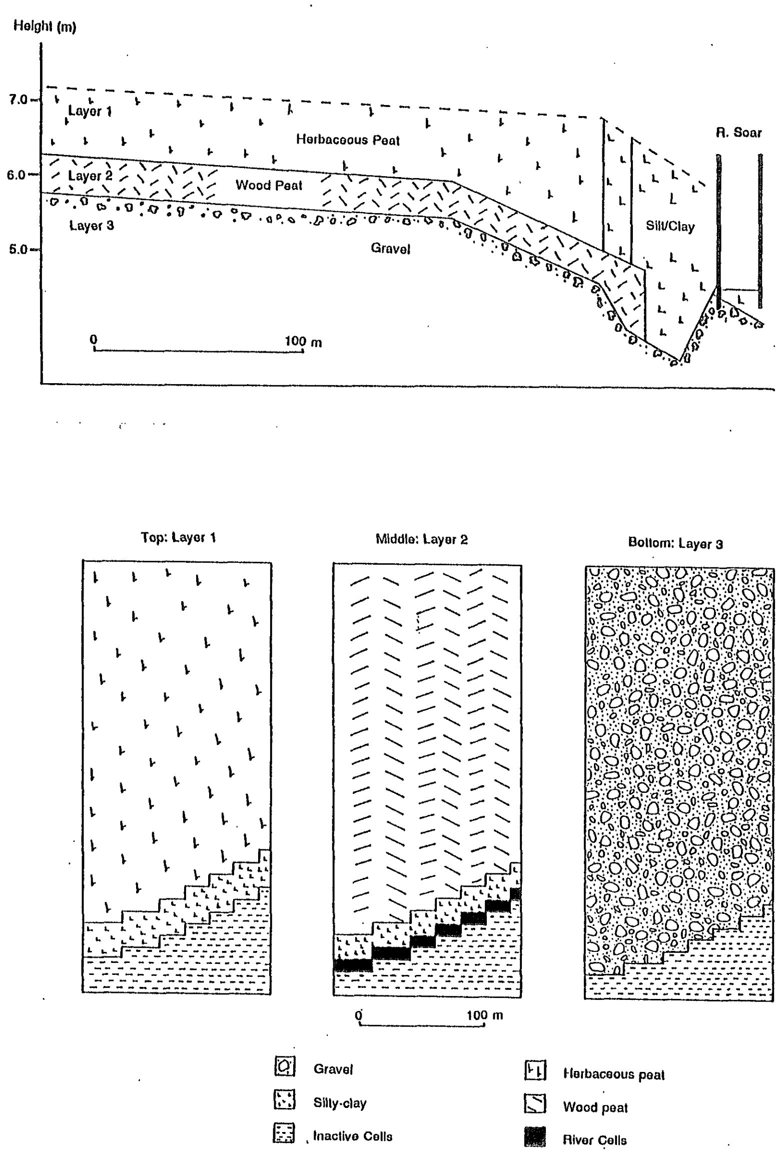
Figure 7.4. Model grid superimposed upon site diagram.



aggregating deposits on the basis of their presumed hydrogeological characteristics. This was straightforward within the reed-bed area, which consisted of a basal gravel aquifer overlain by layers of wood peat and herbaceous peat. In the area adjacent to the river, the alluvial deposits were aggregated, although the hydrological characteristics of the sediments differ. This was necessary as parameters were required to describe mean water flux through the sediments, at the scale of a 10m x 10m cell. It was beyond the scope of the model application to consider variations at a smaller scale. Taking account of these factors, the model representation is shown in Figure 7.5., where the horizontal and vertical variation is shown by a cross-section and plan views of individual layers.

The cross-section in Figure 7.5. indicates how a gravel aquifer was envisaged to underlie the whole of the model area, this comprised Layer 3. Layer 2 consisted of a wood peat layer, with a 20m band of aggregated silt-clay deposits alongside the river. The main component of Layer 1 was herbaceous phragmites peat, with a more extensive band of 40m of silt-clay layer. The river was envisaged as flowing above the gravel layer, separated by an intervening silt-clay layer. Any water fluxes, whether comprising influent or effluent flows, only passed through the river-bed, and not adjacent model cells in layers 2 and 3. This is indicated by the thick line which defines the river bank in Figure 7.5. As discussed above, the flux of water through the river-bed is a function of the hydraulic gradient between river stage and the hydraulic head of water in the basal gravel layer, and the hydrological characteristics of the river-bed sediments. The plan views of the three layers indicate how inactive cells were used to provide a more realistic outline for the field area. A line of cells in layer 2 were classed as river cells within MODFLOW. The gravel aquifer, in layer 3, was classified as a confined layer, requiring transmissivity as an input parameter to characterise water flow. The wood peat and silt-clay bands in layer 2 were classed as either confined or unconfined depending upon hydrological conditions, and the water level in layer 1. The top layer, layer 1, was considered to form an unconfined aquifer. Figure 7.5. also illustrates how

Figure 7.5. Hydraulic representation of Narborough Bog, with cross-section (top), and views of layers 1, 2 and 3.



the river was represented by a line of river cells which crossed the model grid at a slight angle, and which is reflected in the location of adjacent silt/clay deposits.

Boundaries were specified for the top, bottom and sides of the modelled area. The upper boundary was represented by the river and water-table, which would receive recharge through precipitation and lose water through evapotranspiration. The bottom and sides of the model were considered to form a no-flow boundary. Surface elevation is not required by MODFLOW, and therefore there is no theoretical upper limit to which water-tables might rise. This is indicated by the dotted top line on Figure 7.5. Consequently, there are possible limitations in the ability of the model to represent periods of high water-tables when field observations recorded surface ponding of water.

Representation of heights, both of the hydraulic head of each layer and also the top and bottom elevations of each layer, presented an initial problem. The model requires input of heights above an initial datum, and to avoid confusion with specifying heights above Ordnance Datum, all heights were given with respect to a local base level below the lowest point of layer 3. Elevations of the top and bottom of individual layers were specified by an array which enabled details of the surface topography and stratigraphic data to be used when developing the simplified three-dimensional hydrogeology of the field-site. The bottom elevation of layer 1 was specified, the top and bottom of layer 2, and the top elevation of layer 3. Here, the bottom elevation was determined within the model calculations by default, through the specification of transmissivity. The thickness of the gravels was not determined in the field, but previous stratigraphic surveys of the Soar floodplain around Narborough Bog, discussed in chapter 4, indicated an approximate thickness of a gravel layer of between 3 and 4 metres in this area.

7.4.3. Summary of the Calibration Process.

Simulations were undertaken at a daily time interval, which appeared most appropriate given the frequency with which readings of field water-table height were undertaken. The unit of length selected was 1m and consequently all data values given below are described in these units (m/day). The results therefore only consider water-table changes which occur over this time-scale; fluctuations over a seasonal or hourly scale are not modelled here.

Initial hydraulic conductivities for the simple hydrogeologic layers of the model were obtained using the experimental results which were summarised in section 6.6., and are reproduced in Table 7.3. The standard measurements of hydraulic conductivity produce an estimate of vertical permeability; horizontal conductivity was estimated by assuming an anisotropy factor of 10. This helps to represent the greater magnitude of water flux in the horizontal direction.

Certain additional data were required. Firstly, it was necessary to specify the conductance of the river-bed sediments (C_{RIV}). As summarised in Table 7.2., C_{RIV} equals the hydraulic conductivity of the river-bed sediments multiplied by the area of the river cell, and divided by the thickness of the sediments. Augering the centre of the river channel revealed a 40 cm thick silt/clay layer, which directly overlay gravels as described in chapter 4. The dimensions of individual cells were 100 m^2 (LW in eq.3), the hydraulic conductivity (K) of the sediments was taken as 5.4 x 10^{-5} m/day (from Table 7.3) and the thickness of

Table 7.3. Original Hydraulic Conductivities from section 6.3.

Deposit	Hydraulic Conductivity (m/day)
Herbaceous Peat	2 $\rightarrow$ 4 m/day
Wood Peat	0.0439 $\rightarrow$ 0.169 m/day
Silt-clay deposits	2.9 $\rightarrow$ 7.9 x 10^{-5} m/day

the layer (M) was 0.4m. This produced an initial estimate for C_{RIV} of 0.0135 m^2/day. Secondly, it was not possible to measure the transmissivity of the gravels underlying Narborough Bog, and an estimate was required. Using published hydraulic conductivities as a guide (Domenico and Schwartz, 1990) an initial, uniform transmissivity of the gravel of 300 m^2/day was chosen.

It was initially intended to develop a steady state solution to simulate field water-table variations during periods of recharge and evapotranspiration. The normal groundwater modelling procedure is that a calibrated steady-state model is then used to define the starting conditions for a transient simulation, considering time-dependent problems. Although a steady state model was calibrated successfully for a period of falling water-tables, when recharge and evapotranspiration rates were small, a moderate recharge event could not be simulated satisfactorily. Steady state simulations produce an equilibrium water-table profile in response to different stresses, and do not consider the changes in internal water storage, which are associated with recharge. At Narborough Bog the quantity of water storage will vary as water-table fluctuations occur constantly. Water is released from storage when the water-table falls, and is taken-up from storage during recharge following precipitation. Generally, the application of a steady state model is justified when water-table fluctuations are small in comparison with the total vertical thickness of an aquifer, and the water-table configuration remains constant (Freeze and Cherry, 1979, p.194). This assumption does not appear valid at Narborough, and consequently it was decided to develop a transient groundwater model for the site.

For a transient model an additional parameter describing water storage within individual hydrogeological layers is required. In confined deposits this parameter equals the specific storage of the deposit, multiplied by layer thickness; in unconfined deposits it is equivalent to specific yield. In unconfined aquifers, a unit fall in the water-table produces a release of water equal to S_y per unit area, where S_y is the specific yield. Estimates of yield were obtained by analysis of water-table records from individual dipwells, and are given in Appendix V for 1991 and 1992 (Table A.1). The data illustrate how the yield at most dipwells was between 0.3 and 0.4, but decreased at dipwells near the river to a minimum of 0.067 at dipwell 7.

During calibration, the ability of the groundwater model to replicate water-table fluctuations observed during well-defined hydrological events was considered. Three periods were selected:

1. A stable winter period, 3-12 February 1991, in which no precipitation occurred, evapotranspiration was negligible and temperatures were below zero. During this time water-table changes will represent the adjustments required for wetland water levels to achieve an equilibrium state, with respect to river stage and any additional groundwater flow.
2. A winter period, 21-30 March 1992, dominated by an isolated rain event and with negligible evapotranspiration. In this example, water-tables should increase sharply, as precipitation infiltrates, before falling subsequently. These observations would be superimposed upon any lower amplitude background changes.
3. A summer period, 15-24 May 1992, consisting of constant evapotranspiration with little precipitation.

The parameters identified in Table 7.3. were used to define the initial model, and were altered by 'trial and error' to produce model results which corresponded more closely with field observations. Table 7.4. summarises the significance of individual parameters and outlines the stage at which it became necessary to modify the value. In Table 7.5. the final values for parameters are given after calibration.

Field water-tables from the first day of each simulation were used to define the initial starting heads. An interpolated water surface, based upon a 10m x 10m grid, was obtained by kriging using the SURFER package. The file was modified for input to MODFLOW. Initial heads were required for the

Table 7.4. Summary description on the significance of individual model parameters.

Layer 1	Notes
Hydraulic Conductivity	no effect during calibration; the original input values were maintained.
Vertical Conductance	only became important during the final simulation; when recharge was applied to layer 2 the water-table failed to rise sufficiently into layer 1. Vcont was reduced so that the effect of a water-table rise within layer 1 could be included during initial calibration.
Yield	determined the water-table drawdown from evapotranspiration. The high yield reflected the open pore structure of the herbaceous peat within layer 1. Yield was varied spatially to allow for spatial differences in water-table rise.
Layer 2	
Hydraulic Conductivity	limited effect during calibration.
Vertical Conductance	this determines the interaction of layers 1 and 2 with the gravels of layer 3, and the size of daily water-table fall. The value was determined during early simulations.
Yield	determined the water-table rise due to recharge. A value of 0.2 was initially chosen, for the wood peat of layer 2. However, as this layer was assumed to be confined the value was reduced by a factor of ten, and calibrated to reproduce the observed water-table rise from precipitation.
Layer 3	
Transmissivity	values of transmissivity were varied spatially to reproduce the water-table configuration. This was considered to represent a large, lens-shaped, gravel aquifer underlying the model area. Use of a constant value in simulations produced unrealistic clustering of increases and decreases in the water-table.
Yield	this was found to be unimportant. A value of 0.3 was initially chosen, but was readjusted to 0.003 as layer 3 is confined.
River-bed conductance	this was initially determined approximately using measurements of the hydraulic conductivity of silt/clay deposits. The value was adjusted to reproduce observed water-table drawdown adjacent to the river, and reduced further following error analysis.
Recharge	considered equal to precipitation measured at site. Recharge was applied uniformly to layer 2.
Evapotranspiration	obtained from the Penman formula using data from the automatic weather station. Water loss was considered uniform and only applied to layer 1.

Table 7.5. Summary of input parameters for MODFLOW.

Layer 1	Unconfined	
Bottom Elevation	6.7 → 3.75 m	
Thickness	1.2 m	
Yield	0.22→0.18	
Hydraulic Conductivity Phragmites Peat Silty-Clay	Horizontal 2.5 m/day 0.1→0.01m/day	Vertical 0.2 m/day 0.09→0.01m/day
Layer 2	Unconfined/Confined	
Bottom Elevation	6.2 → 2.952 m	
Thickness	0.5 m	
Yield	0.27→0.12→0.015	
Hydraulic Conductivity Wood Peat Silty-Clay	Horizontal 0.6→0.4 m/day 0.3→0.01 m/day	Vertical 0.01 m/day 0.05 m/day
Layer 3	Confined	
Thickness	Specified indirectly through transmissivity.	
Yield	0.003	
Transmissivity Gravels	700→10 m^2/day	

three hydrogeologic layers of the model, and were assumed to be identical in the three individual layers at the beginning of each period. While this assumption may be justified in the upper reed-bed, near the river piezometers recorded an upward water flux (section 5.2.2.). The degree to which the model simulations were able to replicate this water flux would therefore provide an indication of the validity of the model representation.

Generally, horizontal hydraulic conductivities did not have a significant influence on the results and thus in layer 1 the horizontal hydraulic conductivity of the herbaceous phragmites peat was specified as 2.5m/day, and the silt-clay deposits between 0.1 and 0.01m/day. In layer 2, the hydraulic conductivity of the wood peat was between 0.6 and 0.4m/day, and the silt-clay deposits between 0.3 and 0.1m/day. The greater range in the hydraulic conductivity of silt/clay in layer 2 enabled a gradation in the horizontal variation of conductivity.

The vertical conductivity of layer 2 was adjusted during the first calibration period, as it determines the vertical water flux to the gravels of layer 3. The same simulation also revealed distinct spatial clustering of residuals from the model which were reduced through using an array to represent variations in the transmissivity of the gravel aquifer. This enabled the dimensions of a gravel lens to be represented, as illustrated in the cross-section across the Soar floodplain given in Figure 4.2.

The remaining parameters, the yield of layers 1 and 2, and the vertical conductivity of layer 1, were adjusted to reproduce water-table fluctuations from recharge and evapotranspiration. Estimates of recharge and evapotranspiration were obtained using data from the automatic weather station at Narborough. Recharge was considered to be equivalent to precipitation, although differing amounts of rainfall will constitute recharge. This reflects variation in interception, and differences in unsaturated water contents. For evapotranspiration, the estimates obtained from the Penman equation were used. Evapotranspiration was considered to be greatest when the water-table was at the ground surface, and decreased linearly to a cutoff depth corresponding to the bottom of layer 2.

Considerable problems were experienced in obtaining accurate responses to these two types of event. The amount of water-table change was determined by the yield in the layer to which recharge or evapotranspiration was applied. However, there were significant differences in the size of the water-table change for equivalent amounts of recharge and evapotranspiration. Initially recharge and evapotranspiration were both assumed to act on layer 1, and the yield of layer 1 was calibrated to reproduce the response for one rain event. Thus 8.4mm of rainfall in March 1992 produced a predicted water-table rise of +0.07m at dipwell 2, which was within 0.001m of the observed value.

However, the model predicted a water table fall of -0.035m after 5mm of evapotranspiration in mid-summer 1991, while the field readings indicated a fall of only 0.012m.

These results suggest the model was predicting comparable responses to precipitation and evapotranspiration, in proportion to the yield of layer 1, whereas the field data indicate a variation in the magnitude of water table change through evapotranspiration. This may be explained by vertical differences in the yield, and particularly between the wood peat in layer 2 and herbaceous peat in layer 1. These are associated with the distinction between the acrotelm and catotelm identified in mire hydrology by Ingram (1978). Wood peat in the catotelm consists of a compressed peat, with small pore structure and low yield, whilst herbaceous peat in acrotelm consists of large pores with a high yield.

The considerable differences in hydraulic properties through the vertical profile suggest that the model may be calibrated for both recharge and evapotranspiration by considering that they apply to different layers, with the yield of the two layers calibrated to either recharge or evapotranspiration. Thus, if **recharge was applied to layer 2**, rather than layer 1, the smaller yield of the wood peat would enable a greater increase in the water table height for a given quantity of rainfall. Conversely, the water lost through **evapotranspiration could be extracted solely from layer 1**, which has a higher yield. These assumptions may be justified hydrologically as rainwater infiltration produces an increase in the water table, which is determined by soil moisture conditions immediately above the water table. The magnitude of the water table change will thus depend on the hydraulic characteristics of sediments in the zone within which water tables vary, and may also be influenced by the capillary fringe. Water loss through evapotranspiration occurs by a different process, depending partly on the physiological ability of plants to extract water. However, although plants extract water from shallow depths to supply transpiration, upward water fluxes are induced by surface evaporation so that hydraulic characteristics near the surface may be more significant in determining water table change, in comparison with recharge.

In this way, the model was initially calibrated for evapotranspiration by adjusting the yield of layer 1 to replicate the rate of water table fall under varying evapotranspiration rates. The model was then calibrated for the recharge event by adjusting the yield of layer 2. At this stage, the vertical conductivity of layer 1 was modified, as this controlled the rate at which the water table rose into the top layer after recharge.

In the following section the response of the model to the three calibration periods is discussed in detail. The ability of the model to predict water table fluctuations in additional recharge and evapotranspiration events is also considered, to investigate the validity of the assumption that recharge and evapotranspiration may be applied separately to different layers.

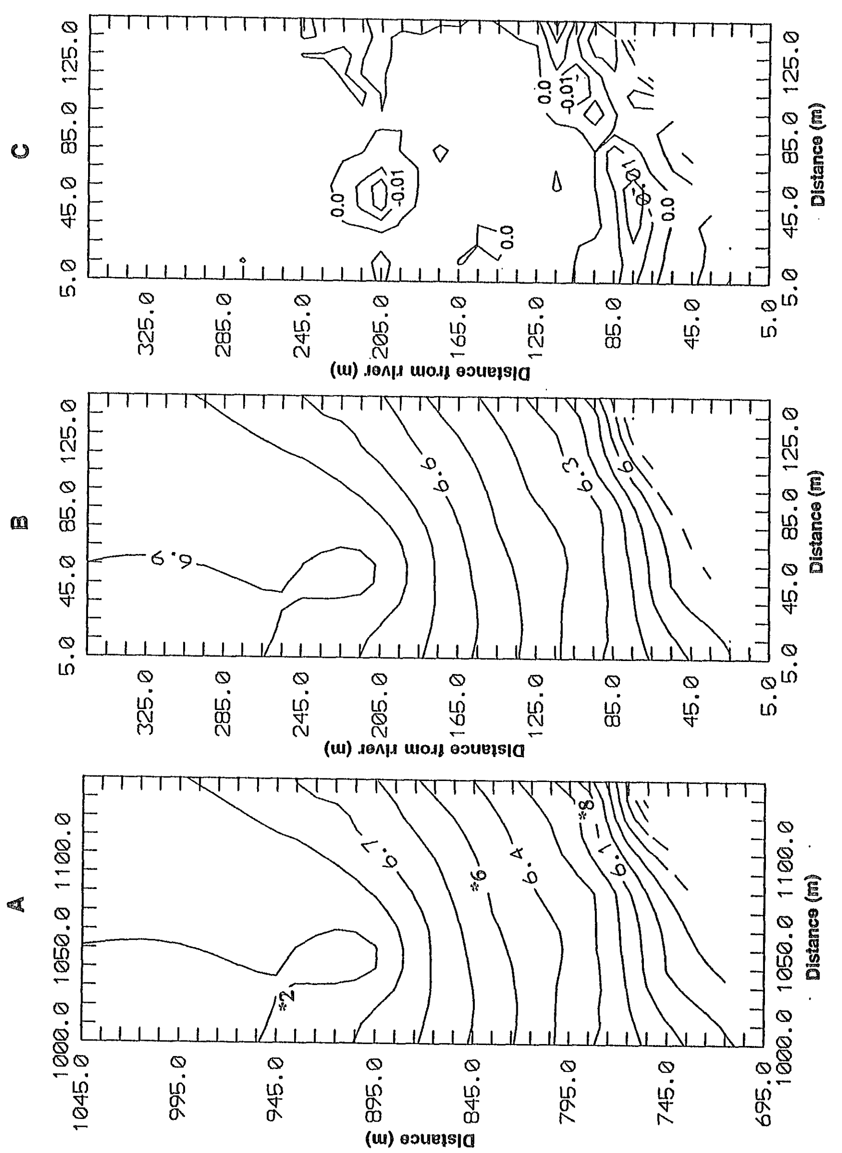
7.4.4. Results of Transient Simulations.

i. Stable conditions: no precipitation or evapotranspiration.

3-12 February 1992.

The model results for the first period of no precipitation and evapotranspiration from 3-12 February 1992 are discussed first. During this period, air temperatures remained below zero which induced freezing of the peat surface and limited vertical water fluxes through the profile. The initial hydraulic head of layers 1, 2 and 3 were obtained from field data for February 3rd. The starting heads are shown in a contour plot in Figure 7.6.A, while the model predictions for layer 2, for the following day are given in Figure 7.6.B. These plots indicate the ability of the model to replicate field data over a daily time interval. In Figure 7.6.C the difference between field heads, and the model results are given. This data was obtained directly from the MODFLOW output; negative numbers indicate a model water table lower than the observed water table. Different coordinate axes are used for the field and model data, however the results are directly comparable. Two principal areas of difference may be

Figure 7.6. Contour plots of results for first calibration period, 3-12 February 1991. **A**: Starting heads from field data on 3 February; **B**: Model results for 4 February; **C**: Residuals between model results and field data.



identified: firstly, a 40m wide band alongside the river where slightly larger rates of drawdown occurred than suggested by the field data; and secondly, at coordinates (205.00,50.00) where the plot of residuals indicates a local increase in the water table. The latter feature corresponds to the location of one dipwell which was recording a water table level lower than surrounding dipwells. This was responsible for the convoluted form of the 6.9m isoline in the water table contour plots of Figures 7.6.A and B and indicates a local measurement problem with the data.

The model predicted the hydraulic head at each node within the grid; these data were used to determine the daily water table fall which were then compared with records from three dipwells, numbers 2, 6 and 8. These dipwells are located at different positions within the reed-bed area at Narborough, as indicated on Figures 7.4. and 7.6. In Table 7.6. the daily water table change for the three dipwells recorded in the field is given, together with the model predictions at grid locations corresponding with the dipwell locations. Water tables were only measured in the field on days indicated by an asterisk in column 1 of Table 7.6., and intervening values were determined by interpolation. Also shown in Table 7.6. are two indications of model error,

Table 7.6. Comparison of observed daily water table drawdown with simulated results for first calibration, 3-12 February 1991.

JD	Error %	Σ error %	Stage (m)		Dip. 2 (m)	Model Dp. 2	Dip. 6 (m)	Model Dp. 6	Dip. 8 (m)	Model Dp. 8
34	2.07	2.07	5.613		-0.005	-0.0002	-0.005	+0.0005	-0.006	-0.0018
35 *	3.47	2.62	5.609		-0.006	-0.0004	-0.005	+0.0005	-0.005	-0.0015
36	4.83	3.14	5.612		-0.005	-0.0004	-0.004	+0.0005	-0.001	-0.0001
37	5.73	3.59	5.586		-0.005	-0.0005	-0.003	+0.0007	0.0	+0.0014
38 *	6.84	4.03	5.670		-0.005	-0.0005	-0.004	+0.0007	-0.001	+0.0027
39	7.41	4.41	5.699		-0.001	-0.0005	-0.004	+0.0008	-0.003	+0.0039
40	7.83	4.74	5.631		-0.001	-0.0006	-0.005	+0.0009	-0.002	+0.0048
41 *	8.99	5.10	5.610		-0.001	-0.0006	-0.004	+0.0009	-0.003	+0.0053
42 *	9.64	5.43	5.598		-0.001	-0.0007	-0.003	+0.0010	-0.001	+0.0057
43	9.67	5.71	5.600		-0.004	-0.0007	0.0	+0.0009	-0.001	+0.0058

equivalent to the error in the water balance calculations output by MODFLOW. Here release of water from storage is considered as inflow; the error is obtained as the difference between total inflow and outflow divided by inflow (McDonald and Harbaugh, 1988). The tabulated error indicates differences in the water budget, firstly for individual days, and secondly the cumulative error in the water balance for the entire model simulation. This error is distinct from differences in the response of the model in comparison with field data. The latter is considered here qualitatively by comparing tabulated data.

The field results for the three dipwells indicate some spatial variation in the rate of water table decline during this period of stable conditions, especially for dipwell 8 near the river. Several mechanisms may explain this, including local upward groundwater flow induced by topography, and also differences in permeability. Initially field observations indicated a constant water table fall of approximately -0.005m, which decreased to c. -0.001m at dipwells 2 and 8, while at dipwell 6 the water table decline remained within the range 0.005-0.003m. In comparison, the model results predicted a steady water table decline of -0.0002 $\rightarrow$ -0.0007m at dipwell 2, +0.0005 $\rightarrow$ +0.001m at dipwell 6, and -0.0001 $\rightarrow$ +0.0058m at dipwell 8. The latter feature apparently reflects the occurrence of increased deposits of lower permeability in the area near the river which produces a very small upward water flux at this point.

The results indicate some problems with the ability of the model to predict conditions of falling water tables. This is evident from the increase in error as the simulation progresses. The initial criterion for acceptance of the model was an error in the water balance of below 5%, although Anderson and Woessner (1992) consider that only an error of below 1% is satisfactory. The results on subsequent days suggest that the large error reflects the sensitivity of the water budget to calculations in the flow of water, to and from storage, and hence reflects the accuracy with which storage values are specified for different layers.

However, the cumulative error is below 5% for the initial 7 days of the simulation, and the field data should also be interpreted within the error margins for the initial measurements of water table position. Water tables were read to within an accuracy of ±2mm in the field, however the surveying of dipwell heights includes some additional error of ±2mm. It is possible that some of the assumptions which justified the initial description of conditions for this period are not satisfied. The approach assumed no evapotranspiration or recharge, and this appeared valid as temperatures were below zero for much of the period. As the surface froze any water flux through the unsaturated zone above the water table should be limited. However, there is also the question of whether any pressure changes arising from surface freezing might produce dipwell observations which are unrepresentative of water table conditions for a 10m x 10m grid cell.

These results provide no indication of the ability of the model to simulate large changes in the water table which arise through recharge and evapotranspiration. However, their significance is that changes in the model are small over a daily time-scale, and hence the results demonstrate the stability of the model constants.

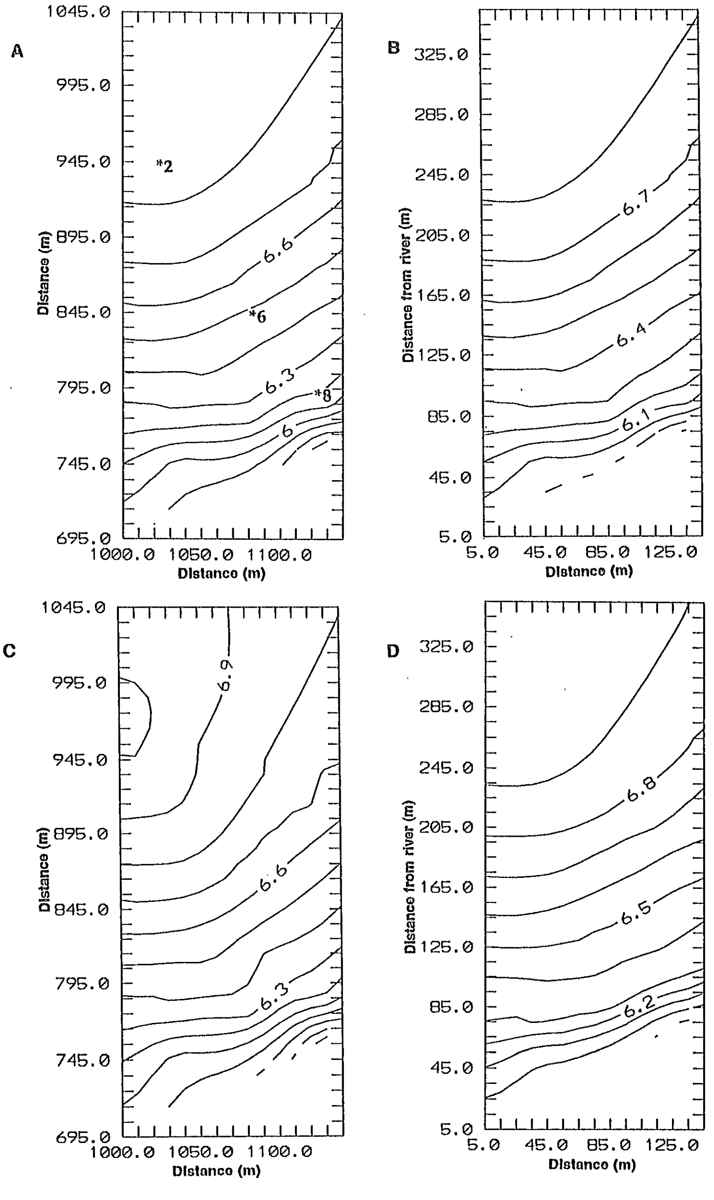
ii. Isolated Recharge through Precipitation.

a 20-29 March 1992.

The model response to recharge was calibrated for an isolated precipitation event where the initial water table was in an equilibrium position, indicating no delayed water table response to a previous event. The period selected was 20-29 March 1992, (JD 80 to JD 89), when a total of 32.5mm of precipitation fell. Starting heads for the model run were input for 21st March, and the responses of the model were compared with field data over the following ten days.

As discussed in the previous section, recharge was applied directly to layer 2. The storage values for this layer were adjusted during calibration and

Figure 7.7. Contour plot of model results for recharge event 20 March 1992 (JD 80). **A**: initial water table starting heads (JD 80); **B**: model results JD 81, **C**: field water tables on JD 85; and **D**: model results for JD 85.

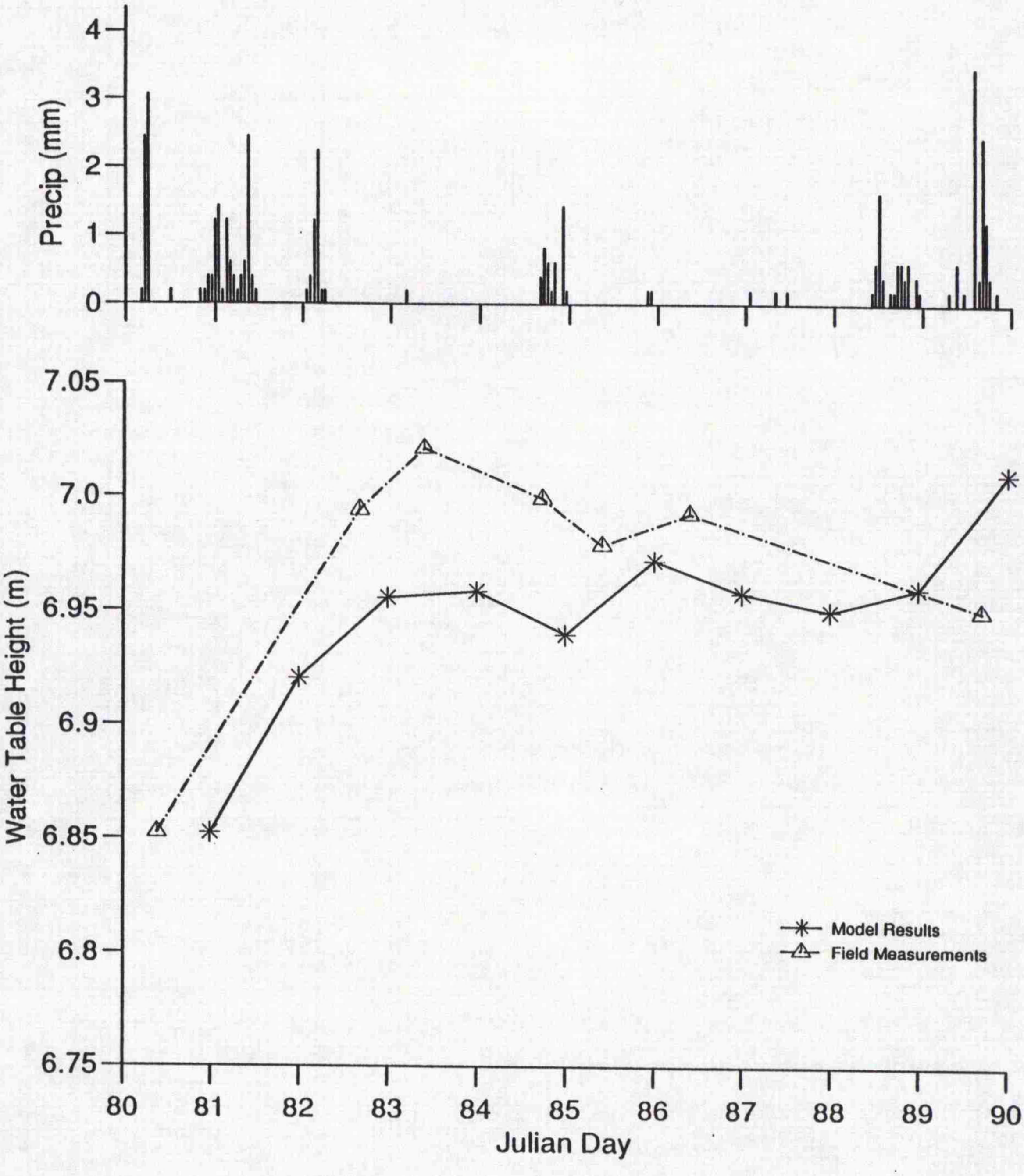


compared with field observations from one period of recharge. River stage varied by 0.275m over the ten day period, and consequently the stage values specified in the river module were adjusted daily to equal the measured readings. This enables a more accurate representation of the variation in hydraulic gradients near the river. In Figure 7.7. a selection of contour plots for the recharge event are given. These consist of the initial heads in plot A; the model predictions for day 1 in plot B; the field heads recorded on day 5 of the model simulation (corresponding to JD 84) in plot C; and the model predictions for day 5 in plot D.

The contour plots indicate how the geometry of the flow system is preserved during the model simulations. Plots A and B in Figure 7.7. reveal a good correspondence after one day, however, there are some differences between the two contour plots by day 5, after a total of 21mm precipitation. These plots illustrate how errors in the model are cumulative, and several areas of divergence can be identified. The field data in plot C record a larger water table gradient near the river compared with plot D, for example near (1120.0, 750.0). Also there are some boundary effects along the left and top side of plot D in which the angles of the contour lines are less than in plot C. This most probably reflects data uncertainty, as there are few dipwells in this area and interpolation is consequently a problem.

In Figure 7.8. the time series of field and model water table variation for dipwell 2 is given, together with a bar graph of hourly precipitation. The relationship of field readings to precipitation events is very significant. The field readings comprise measurements at a given time, while the model predictions represent mean daily forecasts. The two readings may therefore differ in the relationship to precipitation, which explains the difference between model and field results on 25 March (JD 85). On this day 4.3mm precipitation occurred, however, this followed the field water table reading, while the model predicts a water table response which includes the effect of this rainfall. Thus there are limitations in using daily precipitation totals, especially where a rain

Figure 7.8. Time-series plot of water table simulation and field results for dipwell 2, with precipitation, for the period from 20-29 March 1992 (JD 80 - 89).



event occurs overnight. It is significant for the success of the calibration that the field measurements on 22 and 23 March (JD 82 and 83) occurred after precipitation. The frequency of field measurements is also insufficient to determine the increase in water table height following the rain events on JD 89, mainly because the preceding period of water table decline cannot be determined.

A greater indication of the accuracy with which the model simulates water table changes in areas of data coverage is available from the data summary in Table 7.7. Here, the water tables predicted by the model are given from 20-29 March 1992, with observed water table data for comparison. Also shown is precipitation, measured at the automatic weather station, and river stage. The field observations provide a clear record of an increase in water tables arising from the first precipitation event; this varied from +0.071m at dipwell 2, to +0.046m at dipwell 6, to +0.021m at dipwell 8. The response of the model for the first two days was used for calibration of the yields used for layer 2, explaining the good correspondence between the model predictions and field results. The model then predicts a low increase in the water table on 22 March (JD 82) at dipwells 2 and 6, although the daily precipitation was similar. Similarly the water table rise which was predicted on 23 March (JD 83) was too small, although the decrease in water tables on JD 84 was predicted

Table 7.7. Daily Water Table Change for first recharge event, 20-29 March 1992.

JD	Error %	Σ error %	Prp (mm)	Stage (m)	.	Dip. 2 (m)	Model Dp. 2	Dip. 6 (m)	Model Dp. 6	Dip. 8 (m)	Model Dp. 8
80*	1.48	1.48	0	5.663		-0.004	-0.0007	-0.003	+0.0002	0.0	-0.0002
81	0.72	0.94	8.4	5.719		0.071	+0.0683	0.046	+0.0376	0.021	+0.0266
82*	0.91	0.93	8.2	5.938		0.071	+0.0350	0.045	+0.0277	0.021	+0.0233
83*	1.68	1.07	4.5	5.880		0.027	+0.0029	0.016	+0.0088	0.017	+0.0105
84*	4.93	1.32	0.2	5.787		-0.022	-0.0188	-0.010	-0.0075	-0.012	-0.0022
85*	2.03	1.42	4.3	5.793		-0.014	+0.0325	-0.005	+0.0161	-0.009	+0.0149
86*	6.09	1.64	0.4	5.814		0.014	-0.0144	0.013	-0.0042	0.003	-0.0014
87	8.21	1.87	0	5.760		-0.014	-0.0039	-0.010	-0.0027	-0.004	+0.0010
88	7.70	2.10	0.6	5.772		-0.015	+0.0036	-0.011	+0.0014	-0.003	+0.0040
89*	1.76	2.05	6.1	5.866		-0.014	+0.0470	-0.010	+0.0256	-0.004	+0.0217

satisfactorily. The model also recorded slightly larger increases in the water table near dipwell 8, which reflects a predicted upward water flux in this area near the river, induced by topography where layer 2 and 3 are confined by low permeability silt/clay deposits at the surface. This observation is supported by field readings near the river which indicated an upward water flux (section 5.2.2). The differences between model and field results from 25 March (JD 85) reflect variations in the timing of field readings, as identified in Figure 7.8. This emphasises the importance of accuracy of the field data used for calibration.

The error in the water balance is given daily and as a cumulative amount in Table 7.7. The magnitude of error is generally less than in the last data set, with individual daily errors below 1% on JD 81 and 82, when precipitation was greatest. Error is greatest on days in which no precipitation occurred, with a maximum error of 8.21% on JD 87, however the cumulative error over the period is only 2.05%. The error reflects the release of water from storage as the water table falls, and is thus related to the high yield in layer 1. In theory, the storage parameters for layer 2, determined by calibration during the recharge event described above, will include the effects of storage within the unsaturated zone. The magnitude of the soil water deficit will determine the proportion of precipitation which can be stored within the unsaturated zone, and which is therefore not available for recharge.

b. 15-23 December 1991.

The recharge event described in the previous section was used to calibrate the model; the accuracy of the calibration was tested by considering a second period of rainfall from 15 to 23 December 1991 (JD 349 to 357). This will clarify whether the volumetric water content of peat varies significantly. During this period evapotranspiration should be insignificant, and there were no large rain events in the preceding week which would introduce a delayed water table response. As in the previous simulations, initial starting heads were input from field measured heads observed on 15 December (JD 349), and recharge was considered equivalent to precipitation. The period consisted of

Figure 7.9. Time series of model and field results for recharge event no. 2 at dipwell 2, 12-23 December 1991 (JD 349 - 357), with hourly precipitation.

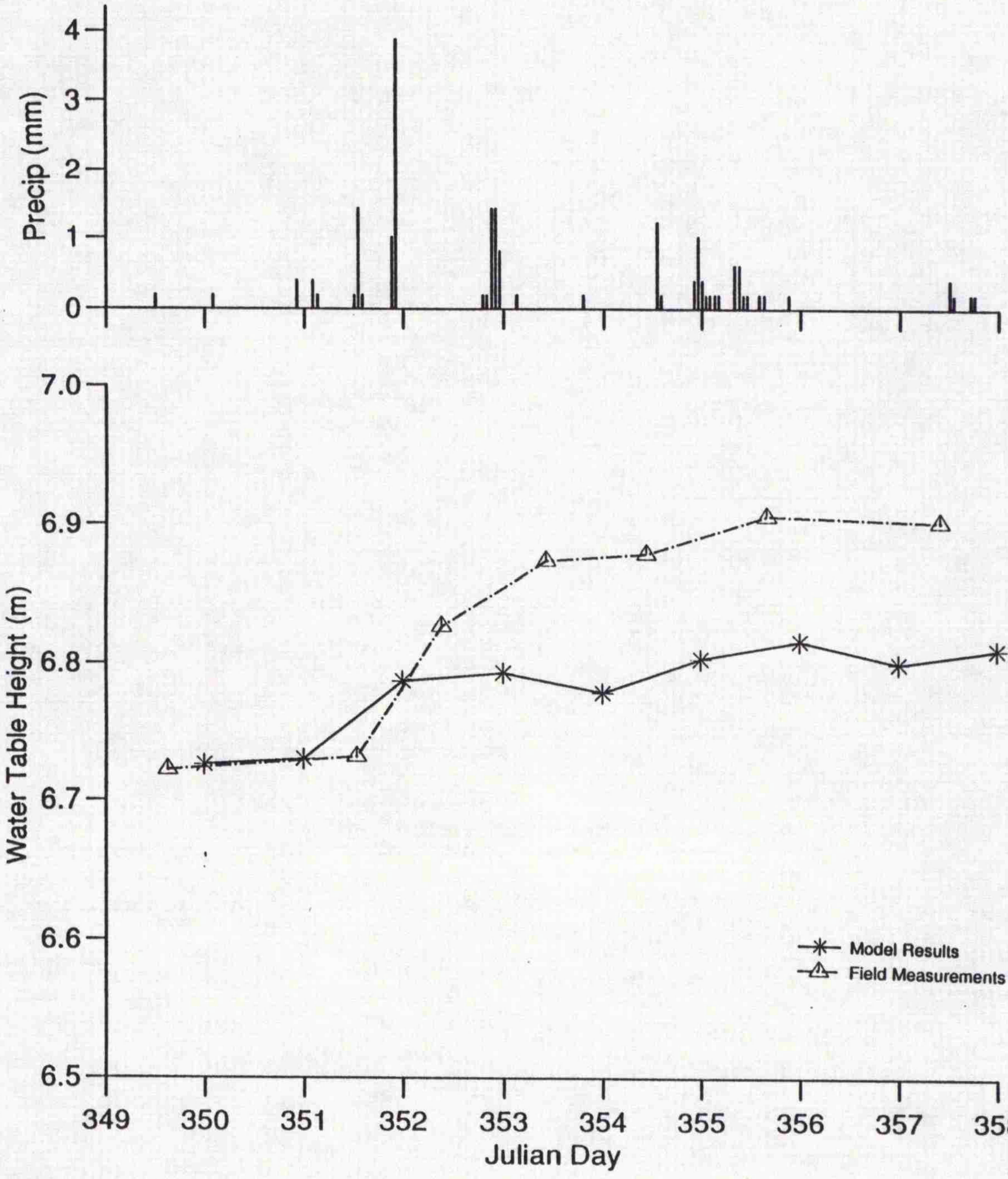


Table 7.8. Daily water table change for recharge event, 12-23 December 1991.

JD	Error %	Σ error %	Prp mm	Stge (m)		Dip. 2 (m)	Model Dp. 2	Dip. 6 (m)	Model Dp. 6	Dip. 8 (m)	Model Dp. 8
349*	0.02	0.02	0.2	5.568		-0.002	+0.0022	-0.003	+0.0025	0.0	+0.0003
350	0.00	0.01	0.6	5.575		0.005	+0.0044	0.001	+0.0040	0.002	+0.0020
351*	1.17	0.65	7.3	5.632		0.005	+0.0582	0.001	+0.0333	0.002	+0.0236
352*	1.95	0.99	4.1	5.667		0.096	+0.0059	0.079	+0.0118	0.005	+0.0117
353*	5.19	1.37	0.4	5.677		0.047	-0.0143	0.034	-0.0037	0.006	+0.0002
354*	2.69	1.59	3.3	5.651		0.005	+0.0245	-0.007	+0.0134	0.009	+0.011
355*	2.90	1.76	3.1	5.674		0.027	+0.0126	0.025	+0.0104	0.009	+0.0103
356	8.34	2.05	0	5.669		-0.002	-0.0131	-0.002	-0.0039	0.005	+0.0007
357*	6.20	2.29	1.02	5.649		-0.002	+0.0072	-0.002	+0.0038	0.004	+0.0052

two days of low rainfall of 0.2mm and 0.6mm followed by two days of moderate rainfall of 7.3mm and 4.1mm. In Figure 7.9. a time series plot, comparing field and model results with hourly precipitation, is given for dipwell 2. The results are summarised in the same format as previously in Table 7.8. for the nine day period.

The field data in Figure 7.9. record a varying relationship to daily precipitation totals, as the water table increased slightly in response to 0.6mm rainfall on 16 December (JD 350), but showed no change after 7.3mm rainfall on the following day. This is because the rainfall occurred after the field measurements, and demonstrates the significance of timing, as discussed in the previous event. The model results are initially similar to the field data; for example the model predicts an increase in the water table of +0.0060m at dipwell 2 which corresponds to an observed rise of +0.0050m. However, the results then diverge as the model results indicate an immediate increase in water tables of +0.0643m at dipwell 2 on JD 351, with a smaller increase of 0.0367m the following day. The error arising from the event on 16 December is therefore cumulative, and accounts for the subsequent difference in the level of the field and model data, although the rates of change of the two data-sets are comparable.

The ability to simulate differences between the response of the three dipwells for this event appears to be satisfactory when comparing dipwell 2 and 8, as the water table rise at dipwell 8 is less than dipwell 2. However, comparing the results of dipwells 2 and 6 indicates that the model predicts a smaller increase in the water table at dipwell 6 than observed in the field. The error in the water balance, determined by MODFLOW, is again low. The cumulative error over the period was 2.29%, however as before the largest errors were associated with days of no rainfall.

iii. Isolated Evapotranspiration with no Precipitation.

a. 15-24 May 1992.

The model was calibrated for water table fluctuations from evapotranspiration during the period 15-24 May 1992 (JD 135-144). Potential evapotranspiration on 15 May (JD 135) totalled 6.4mm and was followed by 4 days of evapotranspiration of c.7.3mm. The rate of water loss decreased slightly towards the end of the period with a minimum of 4.5mm on 22 May (JD 142). Over the period from JD 135 to 144, water tables fell by varying amounts, averaging 0.027m at dipwell 2, 0.014m at dipwell 6, and 0.001m at dipwell 8 (Table 7.9). Water tables on 15 May (JD 135) were used to define the initial hydraulic head.

In Figure 7.10. a selection of four contour plots comparing the field and model results for the period are given. Plot A shows the initial field water table; plot B gives the model results for day 1 of the simulation (JD 135); plot C shows the field readings for day 6 of the simulation (JD 140); and in plot D the corresponding model predictions are given. Plots A and B reveal a similarity of the model surface on day 1 and water tables obtained from field measurements. Some difference between model and field heads were observed by 20 May (JD 140), in plot D the water table gradient near the river is too large, and three of the contour lines (6.55m, 6.50m and 6.30) are significantly distorted. The latter suggests that some modification to the spatial distribution of yield in layer 1 is required, however there is insufficient field data to justify

Figure 7.10. Contour plot of results for first evapotranspiration event, 15-24 May 1992. **A**: Field water table, 15 May; **B**: Model results, 15 May; **C**: field water table 20 May (JD 140); **D**: model results for 20 May.

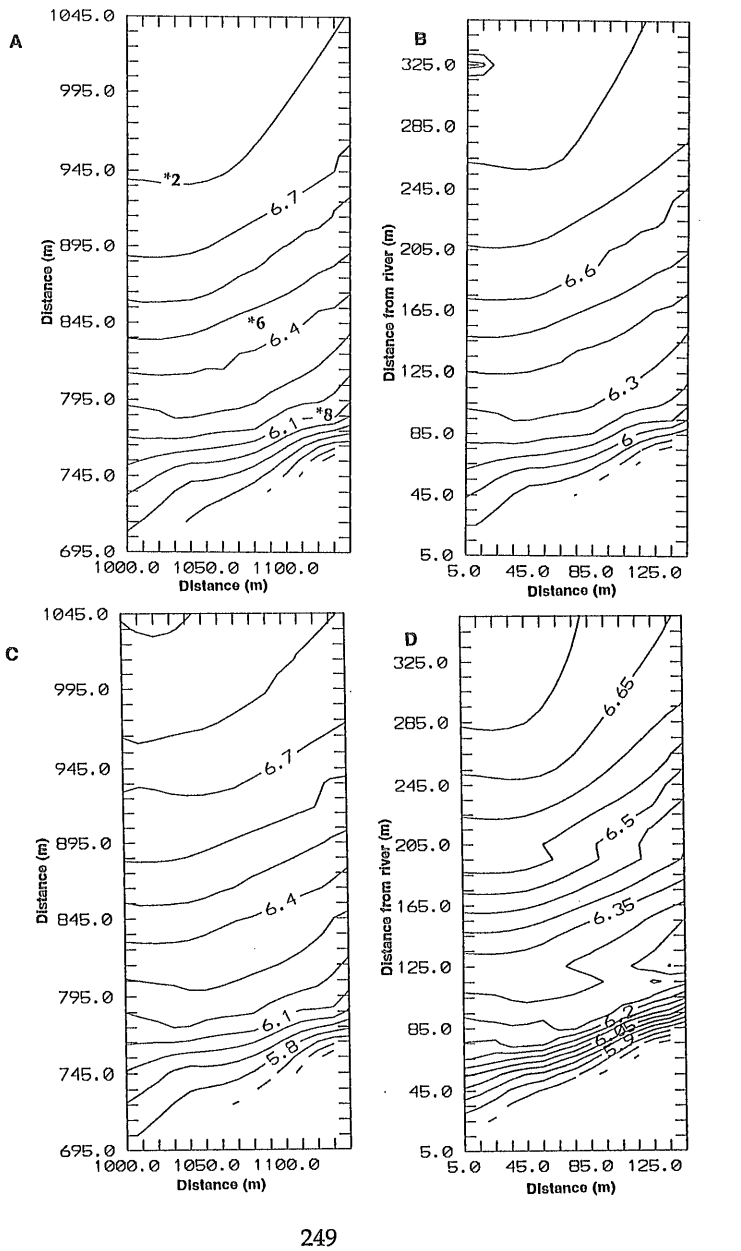


Figure 7.11. Time series plot of model and field results for evapotranspiration 15-24 May 1992 at dipwell 2.

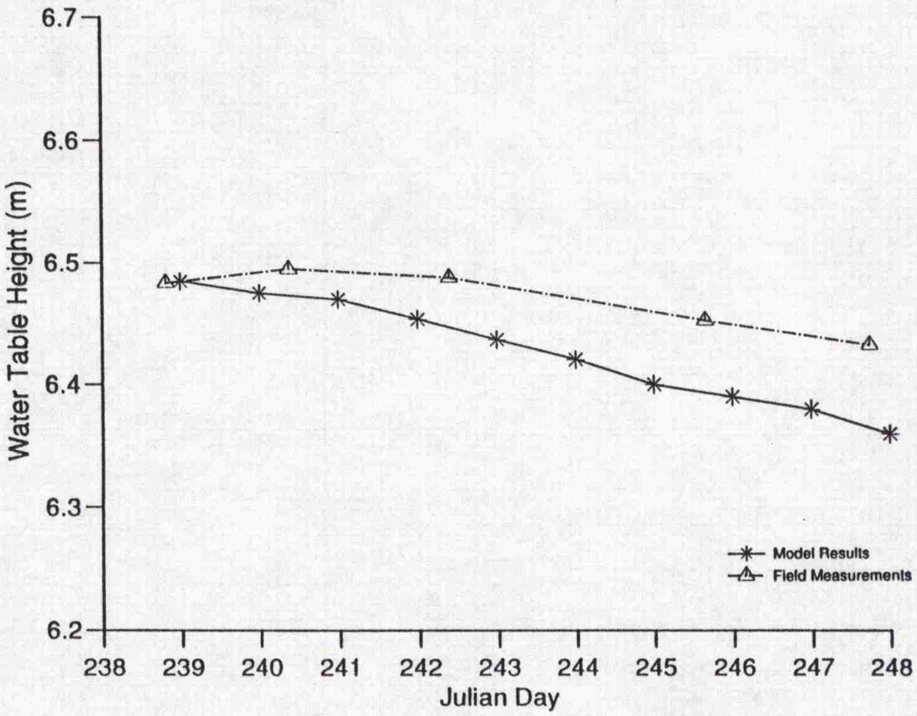
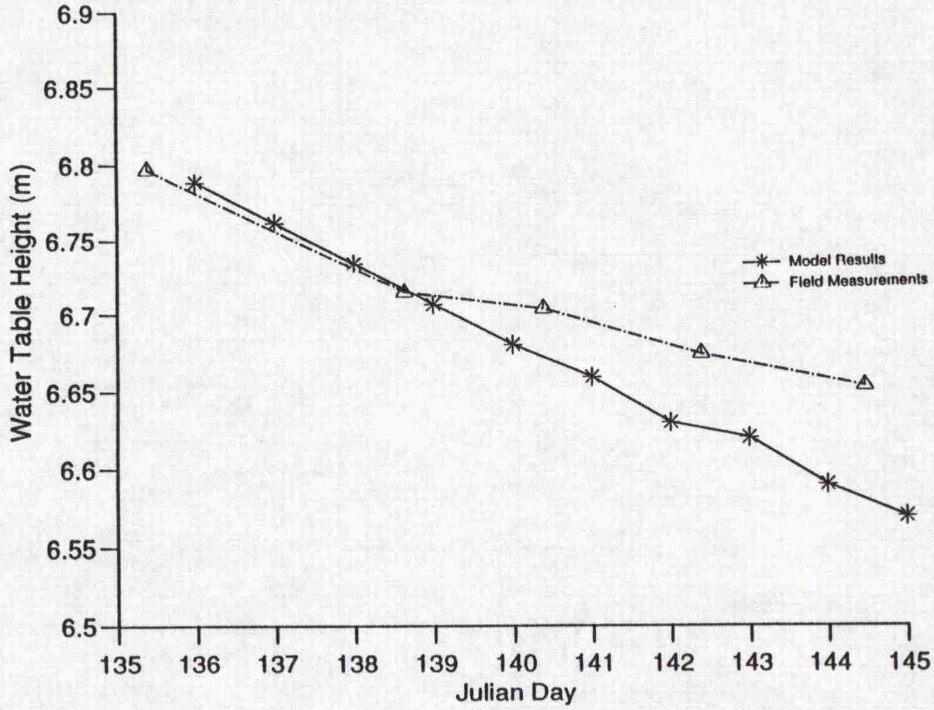


Figure 7.12. Time series plot comparing model and field results for 26 August to 4 September 1991 at dipwell 2.



any modification.

In Figure 7.11. field and model results are plotted for dipwell 2 for the period. The diagram illustrates how the model and field results are similar until 18 May (JD 138), but then decline at different rates, illustrating how errors are cumulative. The five field readings of water table were taken at different times, and thus differ in their timing with respect to the daily evapotranspiration peak. Readings 1, 3, 4 and 5 were taken in the morning, before the radiation peak, while reading number 2, on 18 May, was taken in the afternoon, after the radiation peak. It is possible that the pronounced diurnal radiation cycle will affect the timing of water table drawdown, thus placing certain constraints upon the field data.

In Table 7.9. further detail on the differences in the model and field results are given. The model provides an accurate indication of observed water table change for the three days JD 136, 137 and 138. During this period, estimated potential evapotranspiration was approximately 7.3mm, and the water table fell by -0.027m at dipwell 2, -0.015m at dipwell 6, and -0.007m at dipwell 8. However, the model overpredicted the water table fall for the

Table 7.9. Daily Water Table fall for first period of Evapotranspiration, 15-24 May 1992.

JD	Error %	Σ error %	P.Evp (mm)	Stge (m)		Dip. 2 (m)	Model Dp. 2	Dip. 6 (m)	Model Dp. 6	Dip. 8 (m)	Model Dp. 8
135*	0.65	0.65	6.4	5.730		-0.014	-0.0224	-0.010	-0.0118	-0.003	-0.0011
136	0.16	0.42	7.3	5.716		-0.027	-0.0270	-0.015	-0.0167	-0.007	-0.0010
137	1.29	0.68	7.3	5.722		-0.028	-0.0270	-0.014	-0.0176	-0.008	-0.004
138*	1.06	0.77	7.4	5.711		-0.027	-0.0272	-0.015	-0.0176	-0.007	+0.0006
139	0.55	0.73	7.3	5.713		-0.005	-0.0267	-0.010	-0.0169	-0.006	+0.0013
140*	1.14	0.79	6.8	5.711		-0.006	-0.0248	-0.011	-0.0153	-0.005	-0.0007
141	2.44	0.97	6.2	5.718		-0.015	-0.0225	-0.010	-0.0133	-0.005	-0.0023
142*	3.24	1.14	4.5	5.704		-0.014	-0.0164	-0.009	-0.0092	-0.005	-0.0028
143	2.37	1.26	6.7	5.715		-0.011	-0.0234	-0.010	-0.0123	-0.007	-0.0028
144*	3.03	1.39	5.6	5.712		-0.010	-0.0199	-0.010	-0.0107	-0.006	-0.0028

following three days, although potential evapotranspiration totals were similar. The decrease in model and field water tables are more similar on JD 142 when the potential evapotranspiration total fell to 4.5mm.

b. 26 August to 4 September 1991.

A second period of reduced evapotranspiration was examined to see whether changes in field water tables could be reproduced under different conditions. A ten day period from 26 August to 4 September 1991 (JD 238 to 247) was selected. Calculated potential evapotranspiration was 2.9mm for the first two days, then fell to 1.5mm, before rising to approximately 5mm for the remainder of the period. The field heads recorded on 26 August were used to define the starting conditions. In Table 7.10. the results of the simulation are given in the same form as previously, with a time series plot, comparing the field and model results for dipwell 2 in Figure 7.12. The Table reveals some differences between the field and model records over the period. The initial water table fall of 0.01m at dipwell 2, 0.007m at dipwell 6, and 0.006m at dipwell 8 on 26 August is predicted accurately, however field readings suggested that the water table then rose for the following two days, although potential evapotranspiration of 1.5-2.9mm occurred. From 30 August (JD 243)

Table 7.10. Daily Water Table fall for second period of Evapotranspiration, 26 August to 4 September 1991.

JD	Error %	Σ error %	P.Evp (mm)	Stge (m)		Dip. 2 (m)	Model Dp. 2	Dip. 6 (m)	Model Dp. 6	Dip. 8 (m)	Model Dp. 8
238*	1.45	1.45	2.9	5.636		-0.010	-0.0086	-0.007	-0.0027	-0.006	-0.0005
239	2.31	1.82	2.9	5.614		0.006	-0.0098	0.005	-0.0043	0.002	-0.0004
240*	3.75	2.25	1.5	5.615		0.006	-0.0052	0.005	-0.0018	0.001	+0.0001
241	2.22	2.24	5.0	5.612		-0.004	-0.0164	-0.003	-0.0079	-0.003	+0.0004
242*	2.53	2.30	4.9	5.571		-0.003	-0.0167	-0.003	-0.0090	-0.002	+0.0005
243	2.75	2.38	5.0	5.570		-0.012	-0.017	-0.005	-0.0092	-0.005	+0.0006
244	2.95	2.46	4.7	5.565		-0.012	-0.0161	-0.004	-0.0084	-0.006	+0.0007
245*	3.84	2.60	3.7	5.563		-0.012	-0.0126	-0.005	-0.0062	-0.005	+0.0008
246	3.49	2.69	4.3	5.563		-0.010	-0.0144	-0.001	-0.0066	-0.008	+0.0008
247*	4.23	2.81	3.7	5.561		-0.010	-0.0125	0.0	-0.0054	-0.009	+0.0008

model and field results are similar, although the model predicted an increase in water table near the river.

Figure 7.12. provides greater detail on the differences between field and model results. With the exception of the initial period of a slight increase in water tables, the model and field water tables apparently behave in a similar way, which illustrates that the calibration value of yield in layer 1 for the previous evapotranspiration period appears valid. There remains the question of explaining the observations of an increase in field water tables upto 28 August (JD 240). There was a small quantity of rainfall on 28 September (JD 240), but only 0.2mm. Adjusting the evapotranspiration total by this amount would reduce some of the difference. However, it is clear that the water table fall is over-predicted. There are two clear explanations for these observations; firstly, the results are dependent upon the accuracy of evapotranspiration estimates, which were not tested during this project. It is possible that the Penman estimates are inadequate under certain conditions; the model is driven by radiation and may not give sufficient account of the factors which limit water loss. Secondly, the evapotranspiration totals calculated by MODFLOW may be modified by varying the elevations of the surfaces of maximum water loss and cutoff depth. These surfaces were both specified within the evapotranspiration module, and affect how the actual evapotranspiration totals are calculated, in relation to the water table position.

7.4.5. Sensitivity Analysis.

The purpose of a sensitivity analysis is to quantify the error in a calibrated model which arises from uncertainty in the estimates of aquifer parameters, stresses and boundary conditions. There is additional uncertainty in the geometry of the field system and in the way the hydrogeology is represented, which provides further complexity. For the analysis discussed in this section, individual flow parameters were changed in a systematic way to examine the resulting effect. For this purpose, the model was subjected to a hypothetical five day event consisting of no recharge or evapotranspiration on day 1; 2mm recharge on day 2; 4mm recharge on day 3; 2mm evapotranspiration on day 4; and 4mm evapotranspiration on day 5. The effects of varying the following parameters by +25% and -25% of calibrated values were considered for:

Layer 1: Hydraulic conductivity; Vertical Conductance; Yield.

Layer 2: Hydraulic conductivity; Vertical Conductance; Storage Coefficient.

Layer 3: Transmissivity.

As for the transient simulations described above, the results were interpreted with reference to the three dipwell locations; dipwell 2, 6 and 8. The analysis consisted of undertaking a total of 14 simulation runs to determine the sensitivity of parameters to individual events. The results are summarised in Table 7.11., and indicate how results from the groundwater model are maintained when the input parameters are varied. Hydraulic conductivity had little effect upon the final results, and the most important variables appeared to be transmissivity, vertical conductance, and yield. Transmissivity determines the rate at which water moves through layer 3, and produces an upward movement of heads near the river. The vertical conductance of layer 1 was important in determining how a water table rise from recharge, to layer 2, moved upwards into the top layer. In layer 2, increasing the vertical conductance reduced the increase following recharge, as it increased the flow of water into layer 3. Adjustments of storage coefficients are harder to assess, as the response to recharge and evapotranspiration reflects the combination of storage in layers 1 and 2. Thus increasing yield in layer 1 reduces the water table rise through recharge, and reduces the fall due to evapotranspiration. However, decreasing the yield in layer 1 produces a large water table increase through recharge, the effects of which remain during subsequent evapotranspiration.

This analysis indicates some stability in the model response, as differences in the model output are mainly below 0.001m as input parameters

Table 7.11.Calibration Summary of model parameters.

Layer 1	Notes
Hydraulic Conductivity	This variable had little influence on the final head distribution. When increased by 25% the head at dipwell 6 was 0.0007m lower than the calibrated value, and when decreased by 25% it was 0.0006m higher. There was no change at the other dipwells.
Vertical Conductance	When increased dipwell 2 rose 0.0015m less than the calibrated value after 2mm recharge, and when decreased by 25% rose 0.002m higher. The effects were similar, although not as large at the other dipwells.
Yield	An increase of 25%, produced a smaller rise from recharge - 0.002m at dipwell 2. During evapotranspiration of 4mm dipwell 2 fell 0.003m more than the calibrated amount. A decrease of 25%, produced a 0.002m large rise from recharge at dipwell 2, and a fall of 0.0047m higher for evapotranspiration.[1]
Layer 2	
Hydraulic Conductivity	There was no effect of either increasing or decreasing this variable by 25%.
Vertical Conductance	Increasing Vcont by 25% reduced the water table increase during recharge. Near the river, the increase in water tables due to upward groundwater flux was reduced by 0.0007m on day 2. Reducing Vcont by 25% produced a reversed effect.
Storage Coefficient	An increase of 25% in yield, reduced the water table rise through 2mm recharge by 0.0005m at dipwell 2, there was no effect on evapotranspiration.
Layer 3	
Transmissivity	An increase of 25% in transmissivity, had no effect at dipwell 2, a total increase in head of 0.0007m at dipwell 6, and 0.005m at dipwell 8. The reduction of 25% produced a reversed effect of a decrease in heads at dipwell 8.

[1] This is because the larger increase in water tables during recharge, produces correspondingly high evapotranspiration rates.

are varied by 25%. The parameters which produced the greatest change were the yield and vertical conductance of layer 1. However, in none of the examples did the difference from the calibrated value exceed 0.003m.

7.5. DISCUSSION.

The results of the simulations discussed above reveal differences in the accuracy with which water table fluctuations could be modelled during recharge and evapotranspiration events. The model is able to produce realistic simulations in response to specified input conditions of recharge and evapotranspiration, however, the accuracy and timing of the field data must be considered. Some of the difference between field and model data is an artifact of the time when field readings were taken, with respect to either a precipitation event or daily evapotranspiration. There is also the question of spatial coverage of dipwells. In Figure 7.5. the distribution of dipwells within the model grid was illustrated. There are areas of sparse dipwell cover, with a consequent problem of data interpolation, however, the model results are mainly examined using tabulated comparisons of field and model data at known points, and time-series plots, so that undue emphasis is not placed upon results from marginal areas of the grid.

The accuracy of the results inevitably depend upon the assumptions about the hydrogeologic system, in which water flow within the modelled area is described using a series of differential equations, and parameters are used to represent water flow through the subsurface deposits. It is assumed that parameters such as hydraulic conductivity, vertical conductance and yield remain constant within individual layers, which is questionable for herbaceous and wood peat deposits. Increasing compression of the peat with depth will change the hydraulic properties of the sediment, which explains the differences in hydraulic conductivity of herbaceous and wood peat. It is argued here that much of the problem with producing a precise water table response to a given event arises through incorrect identification of the boundary between layers 1 and 2. The two layers were separated using sedimentological criteria, based on the stratigraphical diagram in Figure 4.3., whereas ideally hydrological criteria should form the basis for subdivision.

The main problem during the modelling simulations was how to reconcile large differences in water table response to precipitation and evapotranspiration events of the same size. Obviously there are certain problems which cannot be resolved by MODFLOW, for example, variations in interception, and differences between potential and actual evapotranspiration. The solution applied here consisted of directing recharge and evapotranspiration to different layers. Recharge was applied to layer 2, which has a low storage coefficient, and consequently produces a larger change in the water table than evapotranspiration, which is extracted from layer 1. A justification for this was given in section 7.4.3. The magnitude of the water table rise through recharge is determined by the hydraulic characteristics of sediments in the vicinity of the water table, whilst evapotranspiration is driven by water loss at the surface. However, there are problems, principally in the ability to model certain recharge events, as on the second day of recharge a larger proportion of the water table change lies within the boundaries of layer 1. Here, the higher yield in layer 1 produces a smaller change in the water table.

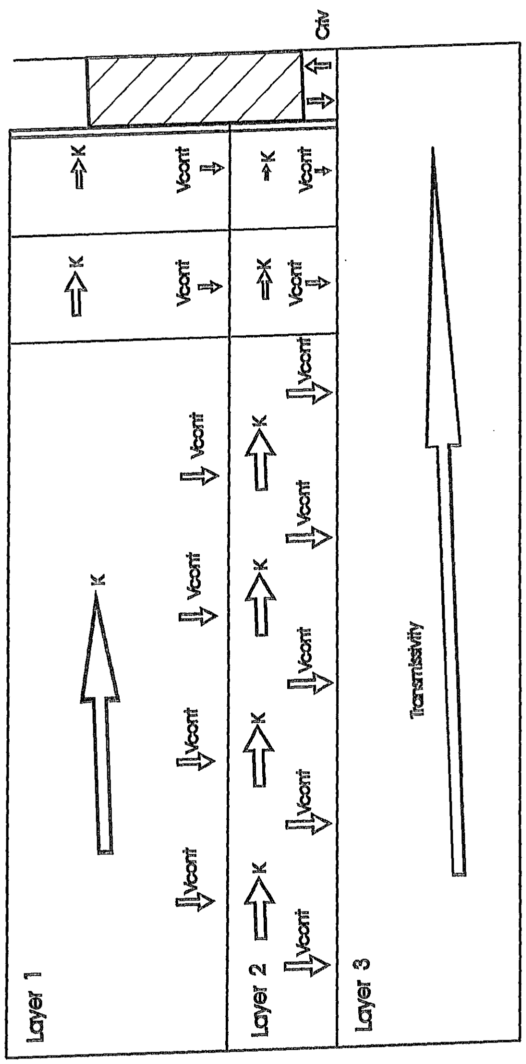
At Narborough Bog spatial variations in the water surface are also of interest, and the varying stratigraphy of the site suggests that interactions of water flow between the peat deposit and the underlying gravel aquifer, and also with the adjacent silt-clay deposits, may be significant. Siegel (1988) considers that questions of water flow through peat cannot be divorced from an examination of water flux through wider sedimentary deposits. This is supported by the results introduced above, and in particular the dependence of the water table distribution upon the transmissivity of the gravel deposits. This is illustrated schematically in Figure 7.13. which illustrates how MODFLOW envisages water flow to occur, using the specified values of hydraulic conductivity, transmissivity and vertical conductance as a guide. In the diagram the length of the individual arrows is drawn approximately proportional to the magnitude of each parameter. While the visual effect of high rates of water transmission through the gravels in layer 3 is evident, the spatial pattern of water tables reflects the balance between the input and output of water and the redistribution of water through the area as a whole. Thus it is significant that layer 3 is confined by low permeability sediments in layer 2, so that the water table in layer 2 is determined firstly by the river level, and the conductance of the river bed sediments, and secondly by steady water flux through layer 2 following recharge.

It may be possible to improve the model, to produce greater correspondence between field and model results. Firstly, it is assumed that yield remains constant throughout the year. The calibrated value for yield is proportional to water storage in the unsaturated zone; this will vary in relation to the position of the water table. It is beyond the scope of MODFLOW to incorporate variations of this type, and consequently more accurate simulations might be obtained if the model is calibrated separately for conditions of high and low water table. Secondly, the significance of the correct identification of the boundary between layers 1 and 2 has already been identified, and a re-evaluation of the dimensions of these two layers may produce more accurate predictions. In addition, it might be necessary to introduce a further layer at the surface to enable differences within the herbaceous peat layer to be represented better.

7.6. CONCLUSION.

This chapter has concentrated specifically on the application of a groundwater model to the wetland system at Narborough Bog. Although there are differences between the observed and predicted water tables, the results demonstrate the processes of water flow through the wetland, and illustrate the importance of varying hydraulic properties in the three hydrogeologic layers

Figure 7.13. Cross-section indicating different input parameters. The arrows indicate the approximate magnitude of each parameter.



used in the model. A perfect correspondence between field and model water tables cannot be expected when applying a groundwater model such as MODFLOW, as this assumes a simple hydrogeologic framework.

One major conclusion which follows from the model simulations is identification of the importance of sufficient detail in the field data which are used for calibration. The measurements of field water table are accurate, but water tables vary over short time periods, and increased temporal resolution is required. Given more detailed data it should be possible to recalibrate the model at an hourly interval, thereby enabling shorter-term fluctuations to be considered. Nevertheless, the results indicate the potential of the modelling approach; they confirm the balance of water flows within individual model layers, and hence may provide a useful initial point for further hydrological modelling of the hydrological response of the wetland at Narborough Bog to future hydrological changes.

Chapter 8

Discussion

Scope of Chapter

This chapter brings together the results from the previous chapters; the records from the field monitoring programme, the results of specific experiments, and hydrological modelling simulations. The chapter begins by applying the groundwater model derived in chapter 7, to produce a comparative water budget for 1991 and 1992. In the following section, the results are described in more detail. The chapter concludes with a short section considering some of the implications of the results for the hydrological management of Narborough Bog and similar wetland sites.

8.1. WATER BUDGET OF NARBOROUGH BOG, 1991 AND 1992.

Chapter 2 focused upon the water budget approach to wetland hydrology and the range of processes which contribute to floodplain/bog water inflows and outflows. In this section, water budgets for Narborough for 1991 and 1992 are estimated. As described in chapter 5, there were substantial differences in the total precipitation of the two years, and thus the results should indicate the potential range in stress which the wetland may experience. This is used to introduce the main section of the chapter which summarises the hydrological processes at Narborough Bog identified during this thesis.

A possible water budget equation for a wetland site was derived in Section 2.4.4. and is given again below:

$$P + q_i + q_{ov} + q_t - q_e - ET = \Delta S \qquad (1)$$

where precipitation (P), influent river seepage (q_i), and infiltration from overbank river flows (q_{ov}) represent the principal water inflows; and evapotranspiration (ET) and effluent seepage (q_e) the main water outflows.

Tributary river flow (q_t) is here considered to be negligible, although it may have formerly been significant.

Of the variables described briefly above, only precipitation and potential evapotranspiration were measured directly during the field monitoring programme, however, the water budget calculated by MODFLOW, may be used to indicate differences in the flux of recharge, evapotranspiration, and river seepage for the two extreme hydrological years, 1991 and 1992. Ideally, a complete hydrological budget would be calculated by specifying daily values of recharge, evapotranspiration, and river stage, as for the short simulation periods discussed in chapter 7. However, as the predicted water-tables are subject to cumulative error which will affect the simulated response to individual events, an alternative approach using annual daily mean data for the two years, is described here. Thus recharge was calculated from the annual mean daily precipitation for both 1991 and 1992, while evapotranspiration, and river stage were obtained from annual mean daily values of potential evapotranspiration, and river stage. Initial water-tables were obtained by producing an interpolated water surface from annual mean values of daily water-tables obtained from field readings of the dipwell network over the year. In Table 8.1. the variables used as input to the model in 1991 and 1992, are given, as obtained by the procedure above.

Two simulation runs were completed, for the years 1991 and 1992, each of two days length. For the first day in each simulation, no precipitation or evapotranspiration was applied, which allowed the model to achieve an approximate equilibrium state before applying the annual mean daily precipitation and evapotranspiration totals in day two. The contrasting water budgets obtained in 1991 and 1992 from the model simulations, are shown in Table 8.2. The components of the water budget consist of volumetric terms for the whole model area, and thus comprise areally integrated precipitation, evapotranspiration, and river seepage, in units of m^3. In both years the main component of the water budget consisted of water flows to and from internal

Table 8.1. Input values for water budget study.

Variable	1991	1992
Mean daily Precipitation (mm)	1.37	2.08
Mean daily Evapotranspiration (mm)	2.07	2.15
Mean river stage (m)	5.513	5.613
Mean height, dipwell 2.	6.821	6.958
Mean height, dipwell 6.	6.466	6.550
Mean height, dipwell 8.	5.881	6.101

Table 8.2. Output water budget for 1991 and 1992.

Year	Water Inflow. m^3		Water Outflow m^3		Error %
1991	Storage Recharge	174.20 64.80	Storage Evapotranspiration River Leakage	148.62 85.58 8.52	-1.54%
	Total:	239.00	Total:	242.72	
1992	Storage Recharge	158.43 98.38	Storage Evapotranspiration River Leakage	146.42 90.91 22.81	-1.28%
	Total:	256.82	Total:	230.15	

storage. A rise in water-table occurs through addition of water to storage, while a falling water-table, includes the release of water from storage within the soil profile. These values depend upon the accuracy with which layers of high and low yield are identified in the wetland soil. In both simulations the error in the water budget is low. The error values were obtained directly from the MODFLOW output and represent the calculated difference between water inflow and outflow to the model area. The balanced budget occurs, despite the lower precipitation total in 1991, and reflects larger water inflow from storage in this year. In the budget, water flow into storage is considered to remove water from the flow system, thus representing an outflow (McDonald and Harbaugh, 1988). The larger water inflow from storage in 1991 therefore represents the release of water from storage to supply evapotranspiration requirements, during a period of insufficient precipitation.

Although some problems with the groundwater model remain, as discussed in chapter 7, the results provide a further indication of the hydrological balance within the wetland system. Release of water from storage, enables water inflow to equal water loss, while the model is also able to differentiate between the total evapotranspiration losses in 1991 and 1992, on the basis of differing water-table depths. The total evapotranspiration loss is also lower in 1991, partly because of the lower water-table, which averaged 0.147m lower at the three dipwells given in Table 8.1. as evapotranspiration is assumed to be directly proportional to water-table height. In addition, the significance of the interrelationship between water-tables and river stage is evident from the different totals of river leakage in 1991 and 1992. The size of influent and effluent river flows is determined by the hydraulic gradient between river stage, and the adjacent water-table. In both years, effluent seepage remains dominant, but is significantly larger in 1992 compared with 1991 (22.81m^3 as opposed to 8.52m^3). This can be explained by the higher hydraulic gradient between the wetland and river in 1992, and reflects a larger response of wetland water-tables to precipitation, compared with river stage. Thus the mean water-table was 0.147m higher in 1992, while river stage was only 0.100m higher, producing an increase in the hydraulic gradient, with a corresponding increase in river leakage.

The main conclusion from the water budget analysis is that, in volumetric terms, the water flux derived from precipitation and evapotranspiration is more significant than river seepage, confirming how the wetland is sensitive to variations in meteorological conditions. Clearly there is a fine balance between inputs and outputs of water, which illustrates how the wetland is vulnerable to sustained periods of low precipitation.

Clearly, the possibility of error in the original data should be considered when analysing the results from quantitative studies of this nature. Gardner *et al.* (1980) discussed the specific example of a wetland system, and used a Monte Carlo simulation to demonstrate the effects of variability in individual values. In some circumstances a stochastic approach is necessary to incorporate errors which result from error in parameter determination and also natural variability. Calculations of actual evapotranspiration present the greatest source of error of the data collected at Narborough Bog. It is generally recognised that potential evapotranspiration calculated using the Penman equation produces an overestimate of actual evapotranspiration loss, and most studies of wetland hydrology have been unable to resolve the problems satisfactorily. Koerselman and Beltman (1988) describe how lysimeters filled with peat and vegetation were used to establish a regression equation between calculated potential evapotranspiration and actual evapotranspiration for a small fen in the Netherlands. While Newson (1994) outlines how evapotranspiration loss in small wetlands may be influenced by edge effects. The common assumption that wetland evapotranspiration is not limited by water supply, and consequently that water loss equals calculated potential evapotranspiration, may therefore not be justified.

8.2. SYNTHESIS.

The results of the water budget analysis, described above, confirm that there is a fine balance between the total water inflow and outflow to Narborough Bog. This was clearly demonstrated in chapter 5, where the difference in the annual hydroperiod between 1991 and 1992 reflected the combined effects of high summer evapotranspiration and precipitation shortage in lowering the water-table significantly. In the same chapter regression equations were applied to describe the relationship of precipitation and evapotranspiration quantities to water-table fluctuations. The results illustrate the limitations of attempting to produce generalised equations at an annual time-scale, but they also help to identify circumstances when other factors should be incorporated into a full deterministic model.

In this respect, the benefits derived from developing a full groundwater model of the site are clear. A hydrological model of Narborough Bog enables the qualitative relationships identified in chapter 5, between water-table fluctuations and precipitation and evapotranspiration, to be tested. The model also highlights circumstances in which additional processes contribute to water flux. However, this approach requires more detailed understanding of water flow through the deposits at Narborough Bog, and hence the experiments described in chapter 6 were necessary to determine the modelling strategies which were most likely to be successful. The experiments revealed large differences in the rates of water flow, both within the area adjacent to the river and also within an isolated column of peat. Values which were chosen for hydraulic conductivity can thus only be estimates, especially as it appears that the applicability of Darcy's Law depends upon the scale of measurement. Hydraulic conductivities of wood peat determined by piezometer test and by the constant head method in the laboratory show some correspondence, indicating that the model may be most successful by representing the extent of differences in the permeability of wetland deposits at Narborough Bog.

In many respects the results of water-table fluctuations are similar to the annual cycle described by Godwin (1931) at Wicken Fen. He sub-divided the annual hydrological cycle into four time periods: these comprise a period of fluctuating high water-tables dominated by variations in precipitation from January to early June; a period of falling water-tables through high rates of evapotranspiration from early June to mid-September; an increase in water-tables as transpiration ceases from mid-September to late-October; and a return to fluctuating high water-tables in November and December.

At Narborough Bog the water-table record is slightly different, and where it is possible to generalise it seems there is an initial period of high fluctuating water-tables related to rainfall from January to March, followed by a period from April to mid-September when the effects of evapotranspiration are important but where water-tables are determined by the precipitation distribution. From mid-September to December evapotranspiration is low, and precipitation is the dominant process. The results differ from Godwin's (1931) description, as data from 1991 and 1992 indicate that significant water-table depression also occurs when evapotranspiration is low and there is a shortage of precipitation. This indicates continued seepage of water through rapid water flow through the gravel aquifer, and controlled by the stage of the river Soar. The MODFLOW simulations were able to replicate the spatial variation in water-table fall, and demonstrate the significance of the large permeability differences between the herbaceous and wood peat, the silt and clay, and the underlying gravel deposits.

The water budget results given in section 8.1. indicated that water outflow due to effluent seepage from Narborough Bog is considerably smaller than evapotranspiration losses. In chapter 5, the analysis of water-table fluctuations compared with river levels was generally inconclusive, due to difficulties in isolating the water-table change in response due to river level, from precipitation infiltration. Currently, the river level appears to have a minor influence upon the hydrology of the wetland, with the exception of periods of overbank flooding. However, river stage may be more significant in the summer, when the water-table is low. A low water-table will reduce the drawdown through evapotranspiration and increase the relative importance of river level, as the hydraulic gradient between river stage and the adjacent water-table determine water loss through effluent seepage.

Some studies of floodplain wetland ecosystems have observed a clear relationship of water-tables to variations in river flow. For example, Hurr's (1983) study of the Mormon Island nature area in Nebraska discussed in chapter 2, where fluctuations in the water-table parallel changes in the stage of the Platte River. However, the environment is significantly different from the situation at Narborough Bog. The Platte River has a pronounced snow-melt regime, while the underlying aquifer consists of lenses of sand and gravel between 36.5 and 47m thick, with correspondingly high transmission rates. Other studies investigating the hydrology of floodplain wetlands have characteristics which distinguish them from Narborough Bog. The nearest examples of studies which have examined the effects of flood frequency are those which consider bottomland hardwood forests in the United States (eg. Mitsch, 1978; Bedinger, 1981). More comprehensive hydrological studies have been undertaken on larger wetland systems such as the Okefenokee Swamp (Hyatt and Brook, 1984), and the distinctive Pocosin wetland type composed of evergreen bogs found on poorly drained terraces along the Atlantic coast (Daniel, 1981).

There are few studies which consider the specific example of the hydrology of floodplain mires, in which exchange of water between a wetland and river system are likely to be important. Grieve *et al.* (In Press) studied variations in hydraulic head along a dipwell transect in the Insh Marshes, Inverness, and considered the relationship to the stage of the river Spey and groundwater flow. Here, the water-table gradient remained towards the river, but groundwater flow derived from the valley side slopes complicated the pattern. Theoretically, floodplain ecosystems should be vulnerable to changes in river regime, and the implications arising from changes in river management practices have been discussed by Klimas (1988). Several studies of floodplain forests in the southern United States have revealed how distinct forest communities have developed which are adapted to variations in the timing and duration of river flooding. The processes which have been observed at Narborough Bog are analogous to this, albeit at a considerably smaller scale, and the example of alluvial wetlands such as Narborough have not been examined to great depth in the literature. The ecological consequences of changes in flooding timing and frequency are complex but reflect the position of individual species along a flooding gradient.

One major point which arises from examining studies of wetland hydrology is the considerable variation in the frequency with which measurements are made. For example, Grieve *et al.* (In Press) measured water-tables every two weeks, while McNamara *et al.* (1992) only took readings every month. This is despite the observations of Godwin (1931) who observed water-table variations over a daily time-scale, based upon hourly measurements. During this study measurements of dipwell levels were undertaken at varying intervals, determined principally by the timing of precipitation. It was necessary to use this data carefully when attempting to calibrate the groundwater model at daily intervals, however, maintaining the data collection period for two years ensured that results from sufficient hydrological events were available for analysis.

Although Godwin's (1931) paper discusses the results of work undertaken over sixty years ago, it demonstrates the lack of floodplain wetland hydrology studies in Britain in comparison with other countries, and particularly the United States. Several recent publications have discussed developments in wetlands, for example Mitsch and Gosselink (1986) and Nachtnebel and Kovar (1991). However, wetland systems vary widely, and floodplain wetlands represent a distinct example where the characteristics of the wetland reflect the balance between fluvial processes and changing meteorological variables. It is believed that the observations of changing ecology observed at Narborough Bog, which included an increase in coverage of dryland species, reflect changes in the importance of river flooding. Quantitative evidence supporting this statement is not available, but the stratigraphic data discussed in chapter 4 suggest that the fen and wood peat at Narborough Bog, accumulated in a backswamp environment. The analysis of stage records of the river Soar since 1971, in chapter 6, illustrated how the frequency of river flooding varies. Overbank floods typically correspond with periods of high and sustained precipitation, when water-tables are high. Infiltration of floodwaters is also limited by low permeability silt/clay deposits adjacent to the river, and the increase in surface slope away from the river limits the area of Narborough Bog which is flooded regularly.

There have been relatively few studies on British floodplain wetlands since Godwin's (1931) study, and although there has been wider appreciation of the importance of wetland preservation this has yet to be reflected in a noticeable increase in investigations of wetland hydrology. The paper by Gilvear *et al.* (1993) indicates the direction in which wetland studies are developing. The approach consists of numerical hydrological simulations which seek to clarify the significance of individual hydrological processes for wetland initiation and maintenance. Similar studies have investigated the interaction of wetland systems with groundwater flow at a variety of scales, for example the papers by Hensel and Miller (1991) and McNamara *et al.* (1992) which both applied a groundwater model to investigate specific aspects of wetland hydrology. However, these studies differ from the Narborough study, in that they lack a comprehensive field database, to describe hydraulic parameters and to calibrate the groundwater model. The results discussed here demonstrate the advantages of a combined field monitoring and modelling approach, especially in a situation where water-tables respond rapidly to changing environmental conditions.

The project has thus concentrated upon a combination of fieldwork monitoring to determine which hydrological processes are dominant, and in order to develop the hydrological model. Thus the representation of the site at Narborough Bog is based initially upon an understanding of how water was flowing through the peat deposits. The experiments revealed large differences in water flow through the sedimentary units, and it is consequently difficult to identify appropriate parameters to describe water flow for hydrological modelling. Klemes (1986) has drawn attention to the growth of computerised hydrological models which are assessed on their ability to fit data, and considers the retrogressive effects of models that work well. The important point is that modelling progresses through stages which can be clearly justified conceptually, and similarly there should be a physical explanation for differences between model and field results.

In chapter 2 the use of different terms to describe wetland types was outlined. Having reviewed the hydrology of Narborough Bog, the correct term for the site can now be considered. In many respects it would seem that Narborough can be considered to be a bog, as precipitation appears to be the principal influence upon water-table levels. However, river flooding was almost certainly important in determining the development of the wetland site, and thus initially Narborough Bog probably represented a floodplain fen. Currently, the variable frequency of river flooding, described in section 6.1., indicates that river flooding no longer represents a major water source to the site. It is also debatable whether Narborough Bog currently represents a true wetland site. Oxidation of the surface peat deposits indicates that there are extended periods of time when the water-table is below the surface, Narborough Bog is thus only flooded after concentrated precipitation, and consequently the reserve is vulnerable to extended periods of precipitation shortage, especially where this coincides with a period of high evapotranspiration.

Reviews of wetland hydrology indicate the difficulties in generalising the relationship between wetland systems and regional groundwater flow, and river levels. Carter (1986) described how the hydrological characteristics of individual wetlands are derived from a combination of the water budget, water chemistry, the water regime and boundary conditions. One of the frequently stated benefits of wetlands is their role in providing flood storage, reducing the peak of the flood hydrograph and also sustaining base flow. The results, suggest that Narborough Bog does not store significant quantities of flood water, although steady water loss through effluent flow indicates the potential of wetlands to maintain river discharge during low flow conditions.

Theoretical studies of the hydrology of upland raised mires have applied groundwater models to describe the shape of raised mires (Ingram, 1992). In these wetland types the formation of the mire is determined by the water balance, and a groundwater mound is formed in the catotelm. The topography of the mire reflects the characteristics of the groundwater mound, which can be calculated by applying Darcy's Law to analyse subsurface water flow. Particularly important is the rate of water seepage through the wetland deposits. The significance of the groundwater mound hypothesis is that it relates the shape of the mire to hydraulic processes, which can be applied to the floodplain wetland at Narborough Bog. The difference at Narborough is that precipitation is insufficient for a raised bog to develop, while the river also provides a local base level to which the wetland water-table adjusts.

The processes of water flow through the peat deposits at Narborough are similar to those described by Ingram (1992) for *Sphagnum* bogs. The wetland is increasingly permeable towards the surface which prevents surface water flow, while in dry weather the low water-table lies within low permeability deposits thus reducing the magnitude of water flows. A substantial volume of water is stored within the deposits at Narborough Bog, however this water is only released at slow rates, due to the hydraulic properties of peat, in particular its high water retention properties, and low hydraulic conductivity of wood peat. Consequently, the hydrology of the wetland is dominated by variations in rainfall and evapotranspiration.

8.3. MANAGEMENT IMPLICATIONS

Although this study is not directly concerned with assessing possibilities of hydrological management at Narborough Bog, certain general conclusions follow from the work which are discussed here. In addition, several questions arise from the results of water-table measurements, and the modelling simulations. Regular visitors to Narborough Bog have described how there have been recent changes in the ecology of the reserve, and particularly within the reed-bed area. The significance of ecological change in wetlands, to hydrology and the possibilities for conservation, have been described by Newson (1994). There is considerable difficulty in quantifying ecological damage to stressed environments, which is unfortunate as changes in ecology frequently provide the earliest indication of the hydrological stress experienced by a wetland. This change may be irreversible, and reflect a delayed ecological response to conditions of water shortage. However, wetlands also represent transient features in the process of hydroseral change, so that ecological changes arise from natural causes. In this respect, conservation may require the control of successional change, as plant associations provide the main conservation interest. The threats to these sites may be direct, and readily identifiable, or alternatively peripheral, producing progressive damage to the wetland (Newson, 1994).

Identification of the precise cause for the ecological changes at Narborough Bog is therefore difficult because of the complicated interaction between hydrology and ecology at the site. It is most probable that the current ecological changes are a result of a significant change in the level of flow of the river Soar and the pattern of overbank flooding which occurred in the 1960s, following the removal of a weir on the Soar at Enderby Mill immediately downstream of Narborough Bog. In chapter 4 it was suggested that the presence of the weir may have been historically significant in maintaining river stage at Narborough, thereby helping the preservation of high water-table conditions at the Bog. The effect of river levels on water-tables within Narborough was investigated using the groundwater model for the period of high evapotranspiration in May 1992, and increasing the river stage by 0.5m to represent a higher stage above the weir. There was a negligible effect upon water-tables within the wetland except close to the river, where the stage was sufficiently high to induce influent flow. The results indicate that the river has mainly a localised effect on determining water levels, while the upper-reed-bed is effectively isolated hydrologically.

The diversion of a small tributary stream which formerly flowed directly onto the reserve would have reduced the amount of water inflow, thus contributing to the water deficit over a long period of time. Although the date on which this occurred cannot be determined, the loss of water from this source exacerbates the effect of water shortages during periods of low precipitation. It is also probable that construction of the railway line across the floodplain had a significant effect in modifying the local pattern of flooding, and directing the surface flow of floodwaters. Overbank floods would have been diverted away from the upper reed-bed at Narborough, as flow became concentrated at the point where the railway crosses the river Soar.

At the scale of Narborough Bog the problem of increasing water flows has similarities with attempts that have been made to control the water-table within archaeological sites. Areas such as the Somerset Levels and the Wash Fenlands have a large number of archaeological sites, and Corfield (1993) has described how archaeologists have begun to monitor hydrological processes. For example, a 500 m length of the Sweet Track in the Somerset Levels was hydrologically isolated by constructing a clay bund and the water level maintained while the surrounding land was drained as a result of peat cutting (Coles, 1990). At a larger scale the Royal Society for the Protection of Birds has had some success in hydrological management at some of its reserves (Burgess and Hirons, 1990).

There have also been several detailed models of soil physics developed which describe procedures to control water-tables precisely, including Youngs *et al.* (1989; 1991). However, these studies have essentially approached the question from the side of land drainage theory. This consists of the control of water level within a ditch network and the use of equations describing seepage flow to determine the corresponding spatial variation of the water-table. At Narborough Bog there are practical limits on the source of water, such that it is the problem of water source, in addition to management, which presents the greatest problem.

The current hydrological monitoring programme has indicated a clear dependence of the local water-table upon precipitation, and the limited impact of overbank river flows. The result is that the reserve is susceptible to any prolonged period of precipitation shortage, especially during the summer when considerable water loss through evapotranspiration occurs.

The ecological consequences of variations in the hydrology are harder to identify. For example, *Phragmites australis* is able to extract water from depths as great as 1.5m below the surface (Haslam, 1970). It is therefore most likely that ecological changes have arisen due to the increasing mineralisation of peat deposits within the reed-bed which has provided a relatively dry, nutrient rich environment in which the number of competitors to the *Phragmites* has increased.

The nature of the ecological changes that have taken place recently seems to indicate a continued transition towards a fen carr vegetation type, however active ecological and hydrological management of the site should enable some maintenance of the former character of the reed-bed and adjacent wet woodland area of Narborough Bog. Ideally the management would probably include a combination of the steps outlined below:

i. Ecological Management

Where reed-beds are harvested, the crop is cut in winter every one or two years. It is possible that mowing the reed-bed at a particular time of the year will limit the herbaceous cover. The *Filipendula* plant association, described by Wheeler (1984), closely resembles the current species composition of the reed-bed, and is restricted by mowing or grazing. If this strategy were adopted care would have to be taken to restrict the number of sites at which the cuttings were burnt, as increased nutrient levels near fire sites would be beneficial for the herbaceous, non-wetland vegetation.

ii. Increase of Tributary flow onto reed-bed.

The diversion of a tributary stream which formerly flowed onto the reserve was mentioned in chapter 4. This is a groundwater-fed spring which flows throughout the year, flow levels increasing considerably during large rain events. It is possible that a small culvert might be constructed to re-direct the stream onto Narborough Bog, thereby providing a year-round supply of water. This would have a beneficial impact on the water-table both towards and away from the river.

iii. Pumping of River Water

The value of periods of river flooding for preserving wetland ecology and maintaining reed-bed areas has been described by Duffey (1971) and Haslam (1972). At Narborough Bog pumping of water from the river Soar could provide additional water inflows at times of greatest water shortage. This would enable increased control over water-table levels which appears to be the only solution to water shortages during periods of low precipitation. However, the possibility of changing water quality should be considered, as high nutrient levels tend to be associated with low river flow conditions when pumping would be required. This will not affect *Phragmites*, but might increase the cover of non-wetland vegetation, if the water-table is not maintained at a sufficient level. The water budget calculations in section 8.1. indicate the approximate balance between water inflow and outflow. Under the extreme conditions in 1991, the difference between mean daily totals of evapotranspiration and recharge was approximately $20m^3$, demonstrating that the water deficit could be easily reversed by pumping small amounts of water onto the wetland at critical periods. Conversely, Table 8.1. indicates that in 1992 mean daily water inflow through recharge exceeded that lost from evapotranspiration, which suggests that pumping of water from the river would not necessarily be required every year.

8.4. Conclusion.

This chapter has described a sample water budget for 1991 and 1992, indicating the differences in importance of individual water inflows and outflows under varying meteorological conditions. The following section then discusses some of the results of the thesis within the context of selected published work on wetland hydrology. There have been few studies which have combined a detailed field monitoring programme with analysis using numerical modelling techniques, and it is hoped that the results demonstrate the advantages in adopting this combined approach. In section 8.3. the possibilities for hydrological management are briefly considered. In chapter 9, the principal conclusions of the thesis are summarised.

Chapter 9

Conclusions

In this chapter the main conclusions of the hydrological study at Narborough Bog are summarised and considered against the initial objectives of the thesis which were introduced in chapter 1. Possible directions for further work which follow from the work discussed here are also identified.

The development of **a conceptual model describing the interaction of hydrological processes for a floodplain wetland** formed the basis of chapter 2. Floodplain wetlands are subject to a range of water inflows and outflows and before proceeding to the modelling simulations it was necessary to consider the circumstances under which particular water fluxes might be important. The schematic diagram in Figure 2.2. was used to illustrate how the relationship of wetlands with rivers may vary. Two diagrams, Figure 2.3. and 2.5., show the direction of possible water flows, indicating how wetlands may receive water inflow from a variety of sources. These include flow from adjacent slopes in addition to precipitation, evaporation and evapotranspiration. Finally in chapter 2 the discussion of wetland water budgets indicates how the conceptual model can be applied in field investigations.

The relationship of fluvial processes to the wetland at Narborough Bog, and the balance of influent and effluent flows and overbank flooding was considered at several points in the thesis. Firstly, the water-table data were used to aid a qualitative examination of the relationship between river levels and the wetland water-table. Differences were observed in the relationship depending upon the moisture characteristics of the catchment; generally water-tables exhibited a greater response to precipitation although near the river overbank flooding produced a sharp local increase in the water-table height. Due to difficulties in measuring influent and effluent flows the water budget calculations produced by the groundwater model were used to produce approximate values for seepage under extreme conditions for the years 1991 and 1992. Results from the simulations demonstrate the dominance of effluent seepage, with the contribution from influent flows only affecting an area immediately marginal to the river. The water budget calculations undertaken using the calibrated model illustrate the significance of the balance between inputs of precipitation and water outflows of evapotranspiration and effluent seepage. Currently the river Soar has a minor effect on the hydrology of Narborough Bog: water-tables are relatively isolated by the effects of river stage variations, and more vulnerable to fluctuations in rainfall and evapotranspiration. This is believed to be the case with many alluvial wetlands in the United Kingdom, as a result of land drainage. The contribution of the river through overbank flooding is also limited, as revealed by analysis of the frequency of overbank flows at Narborough. Water flow from this source is of varying importance and mainly serves to exaggerate the existing contribution of precipitation.

Description of the stratigraphy of Narborough Bog, to clarify the history of the wetland and to consider possible simplification of the stratigraphy to represent permeability variations most effectively was considered in chapters 4 and 7. The stratigraphic survey of Narborough Bog provides important information on the initial development of the wetland, which would seem to be as a backswamp of the river Soar. Accumulation of peat was facilitated by the local stability of the river channel and high water-tables. The latter were probably maintained by a combination of river flooding and discharge of a small stream, while a barrier of silt/clay deposits adjacent to the river and stream provided the stable conditions required for continued accumulation of organic deposits and a rising water-table.

The stratigraphic data provided central information in developing the simplified hydrostratigraphic diagram required before applying the groundwater model. Experiments examining water flow through silt-clay deposits near the river and through an isolated peat column revealed large variations in water flux suggesting different modes of water flow, determined by the hydraulic head applied at the surface. The results indicated the importance of large differences in the permeability of alluvial deposits, but the modelling simulations indicate that the simplifications of hydrostratigraphy produced an acceptably low level of error.

The development of a deterministic computer model to represent the hydrology of Narborough Bog was one of the major results of the thesis and is described in chapter 7. Although field observations indicated large differences in the rate of water flow over short distances, the calibrated model of the site was able to simulate water-table responses to recharge and evapotranspiration events. Large variations in the fluctuation of water-tables, in response to rainfall and evapotranspiration, would seem to reflect vertical differences in water storage, which support the acrotelm and catotelm models of the wetland profile. Significantly, the model was calibrated for transient state simulations. It is therefore capable of simulating water-table fluctuations within a wetland environment and the associated movements of water to and from storage.

Several general points also follow from the work described in this thesis. Oxidation of the surface deposits indicates that Narborough Bog is no longer an actively forming wetland. The results indicate the extent of seasonal variation in water levels and demonstrate the vulnerability of the site to sustained precipitation shortage, especially during periods of high evapotranspiration. The study was fortunate to include the two years of 1991 and 1992 with a large difference in total precipitation, thus illustrating the extremes of moisture available to wetland vegetation. Application of simple quantitative regression models demonstrate the limitations of applying time-invariant solutions when examining annual and seasonal variations in water flux.

The study at Narborough provides several indications of the areas in which further study is required. The work has demonstrated the value of adopting a combination of process and modelling studies, however, it is unrealistic to expect this approach to be widely adopted in studies of wetland hydrology. It would therefore be useful if there were further investigation of wetland hydrogeology, possibly entailing the production of generalised hydrostratigraphies. These might then be tested using recognised modelling procedures, as described here, so that the circumstances in which individual water fluxes vary might be better understood.

The work has also revealed limitations in the measurement of certain features of the water budget. Determination of actual evapotranspiration, as opposed to potential evapotranspiration, remains difficult. It would be useful if the evapotranspiration from small alluvial wetlands was measured directly, rather than indirectly using formulae such as the Penman equation. This becomes more significant when potential evapotranspiration data are used for calibration purposes, as discussed here in chapter 7.

The study demonstrated how the direction and magnitude of subsurface seepage might be modelled using packages such as MODFLOW. There remains however, a lack of field studies considering the determination of the permeability of river-bed sediments. This mainly reflects difficulties both in measuring permeabilities in-situ, and also in obtaining samples of river-bed deposits. However, such data would provide valuable confirmation of the parameters obtained by calibration in groundwater models.

The groundwater model developed in chapter 7 introduced several new techniques to the study of wetland hydrology which require further refinement. In particular, there are outstanding difficulties in representing the manner in which the hydraulic conductivity of peat varies as a function of depth. This represents a significant problem if simple groundwater models, developed using the principle of layered aquifers, are to be applied to wetlands. The models are also limited in their restriction to the saturated zone, whilst moisture storage within the unsaturated zone may be an important component in determining the response of water-tables. Here it would seem that greater instrumentation of the zone within which water-tables fluctuate would help improve the validity of numerical modelling simulations. There also remains the possibility of applying the modelling approach at different scales in an attempt to forecast the seasonal variation in water-tables and also to examine the timing of precipitation infiltration at an hourly scale. For the latter purpose it would be necessary to increase the temporal resolution of measurements by using data logging techniques.

Finally the study has emphasised the comparative lack of comprehensive studies of wetland hydrology. Application of general monitoring procedures in combination with detailed stratigraphic surveys should lead to a fruitful reconsideration of the principle features of wetland hydrology, and how the magnitude of individual water fluxes varies in different environments.

APPENDIX I

Species List, Narborough Bog

The ecology of Narborough Bog outlined below is a summary of information obtained from Primavesi and Evans (1988) and the site description produced by the Nature Conservancy Council in association with the designation of Narborough Bog as a Site of Special Scientific Interest.

The northern woodland is dominated by the crack willow *Salix fragilis*, with the hawthorn *Crataegus monogyna* and elder *Sambucus nigra* locally abundant. Also present are alder buckthorn *Frangula alnus*, rowan *Sorbus aucuparia*, field maple *Acer campestre*, ash *Fraxinus excelsior*, alder *Alnus glutinosa*, sallow *Salix cinerea*, and oak *Quercus robur*. The herb layer includes the guelder rose *Viburnum opulus*, cleavers *Galium aparine*, rough meadow grass *Poa trivialis*, stinging nettle *Urtica dioica*, cuckoo-pint *Arum maculatum*, enchanters nightshade *Circaea lutetiana*, herb-Paris *Paris quadrifolia* and foxglove *Digitalis purpurea*.

The reed-bed is composed of the common reed *Phragmites australis*, great burnet *Sanguisorba officinalis*, and tufted vetch *Vicia cracca*. Although once dominant (eg. Wade, 1919), the reeds are now facing increasing competition from other plants including meadowsweet *Filipendula ulmaria*, great willowherb *Epilobium hirsutum*, stinging nettle *Urtica dioica*, cleavers *Galium aparine* and grasses including *Poa trivalis*.

Beside the river, the vegetation includes hemlock *Conium maculatum*, field thistle *Cirsium arvenze*, Himalayan balsam *Impatiens glandulifera*, goldilocks buttercup *Ranunculus auricamus*, greater celandine *Chelidanium majus*, and brooklime *Veronica beccabunga*.

The southern meadow area has a distinct flora including the southern marsh orchid *Dactylorhiza praetermissa*, fen bedstraw *Galium uliginosum*, wild angelica *Angelica sylvestris*, ragged robin *Lychnis flos-cuculi*, devil's bit scabius *Succisa pratensis*, burnet-saxifrage *Pimpinella saxifraga*, and marsh thistle *Cirsium palustra*. Yellow rattle *Rhinanthus minor*, marsh valerian *Valeriana dioica*, and cowslip *Primula veris* have been recorded but not observed recently.

Meaningful comparison of changes in the flora composition since Wade's 1914 study is made difficult by the preservation in certain areas of the Reserve of permanently water-logged conditions. In two areas ponds have been excavated, one to the southwest of the reed-bed beyond the spoil heap, and the other half-way down the reed-bed. In the first pond, in particular, water-cress *Nasturtium officinale*, yellow iris *Iris pseudacorus*, water figwort *Scrophularia auriculata*, marsh marigold *Caltha palustris*, and water parsnip *Berula erecta* are found. Furthermore, the study only gives limited evidence as to the precise location of the observation and does not indicate species abundance.

English Nature are currently funding regular ecological surveys of the reserve to construct a data-base from which it should be possible to identify changes in species abundance.

APPENDIX II

Auger Stratigraphies

In this appendix the stratigraphies obtained from hand augering along transects at Narborough Bog are summarised. The locations of the auger sites are indicated on Figure 4.3. while the numbered dipwell locations are shown on Figure 4.7.

Dipwell 0.

0-30 cm.	Peat soil, some sand.
30 cm →	Increasing clay content, iron rich, some cinders, occasional stones.
40-50 cm	Clay, some peat at base. Remains of Phragmites.
50-80 cm	Increasing peat content.
80-100 cm	Herbaceous peat.
c. 140 cm	Gradual boundary from herbaceous peat to silt-clay.

Dipwell 3n

0-28 cm	Brown, partially oxidised peaty soil.
28-108 cm	Fibrous *Phragmites* peat.
108-116 cm	Dark black/brown wood peat.
116-180 cm	Dark brown wood peat.

Dipwell 4.

0-30 cm	Friable peat soil.
30 cm →	Peaty silty clay.
c. 50cm	Some Phragmites remains.
60-80 cm	Darker peat; still herbaceous; some roots.
80-100 cm	At 100 cm some gritty clay, dark peat.
100-120 cm	Dark wood peat

Dipwell 7.

0-10 cm	Silty clay soil
10-40 cm	Silty clay, some stones, and *Phragmites* stems.
50-135 cm	Silty herbaceous peat.
135-180 cm	Gradual boundary to darker wood peat.
180-200 cm	Coarse sand lying on gravel.

Additional core midway between dipwells 7 and 8.

0-25 cm	Peaty crumbly black brown soil
25-52 cm	Stiff banded grey brown silty clay, fibrous organic remains.
52-145 cm	Silty herbaceous peat.
145-162 cm	Organic silt with wood, banded medium coarse sand, occasional pebbles.
162-180 cm	Banded silty grey clay, large wood fragments, fine sand, on gravels.

Dipwell 8.

0-20 cm	Herbaceous peat soil.
20-40 cm	As above but increasing *Phragmites* remains.
40-100 cm	Silty clay, herbaceous.
100-120 cm	Brown silty peat, fibrous.
120-160 cm	Silty peat.
160-170 cm	Sandy silt overlying gravels.

Additional core 7m from dipwell 8, near River.

0-14 cm	Grey brown, slightly organic silty soil.
14-20 cm	stiff, grey yellow, brown mottled silt clay.
(20-29 cm)	reddish-brown micro-aggregate organic rich soil (ant chamber).
29-58 cm	thick silty clay, yellow becoming increasingly grey, wood/root fragments.
58-96 cm	Boundary to light brown, grey, yellow herbaceous peaty silt, becoming grey/green after 70 cm, herbaceous until 96 cm.
96-100 cm	Abrupt boundary to grey/black silty clay with herbaceous phragmites roots and stems.
100-177 cm	Thick grey blue clay with occasional charcoal, abundant phragmites roots and stems.
177-200 cm	Abrupt boundary, brown grey organic silt with abundant root and stem fragments, small fragments of wood, mostly herbaceous, increasingly brown and organic with depth.
200-260 cm	Brown very peaty silt, small undecomposed wood fragments. At 237-244 cm silty grey band mixed with wood debris.
260-308 cm	Gradual boundary to compact brown wood peat.
308-330 cm	Organic rich sandy silt, small stones lying over impenetrable gravel.

Additional core 14m from dipwell 8, near River.

0-55 cm	Black peaty soil.
55-110 cm	Brown grey silt, some fine sand.
110-136 cm	Soft grey silty clay.
136-165 cm	Grey brown organic rich silt, phragmites remains.
165-200 cm	Gradual boundary to brown fibrous peaty silt, banded with occasional thin sand layers, varied humification.
200-280 cm	Dark brown wood peat.
280-300 cm	Mixed organics, silt clay wood/plant fragments and hazelnuts.
310 cm	Gravel base.

Dipwell 9.

0-20 cm	Stiff grey brown silty clay, occasional sand layers.
20-40 cm	As above, plastic, grey brown silt, fine sand.
40-60 cm	Grey sandy silt with organics including wood, sand horizon at 40-55 cm.
60-80 cm	Grey brown silty clay with reddish fine sand.
80-125 cm	Black organic mud.
125-231 cm	Black silt mud grading to thick grey silty clay with shells.
at 241 cm	Gravel.

Dipwell 5n

0-20 cm	Brown peat soil.
20-40 cm	Silty clay, some small stones.
40-137 cm	Gradual boundary to herbaceous peat.
137-150 cm	Boundary to organic silt and clay lying on gravel.

Dipwell 22.

0-14 cm	Friable, crumb structured clay/silt; some fine sand and rootlets.
14-41 cm	Plastic clay/silt; fine sand; light brown; no plant remains; Wetness 3/4; Very stiff.
→ 67 cm	As above.
67 cm →	Black/brown orange mottled clay with root fragments.
108 cm →	Increasingly sandy with gravel.
132 cm →	Impenetrable sand and gravel.

Dipwell 23.

0-18 cm	Humified silty peat; Crumbly structure; contemporary root structure; Dry (wetness 3); No stratification.
18-50 cm	Brown/black peaty silt; Wetness 2; Plastic; Humification 3; Fragments of twigs and roots; Stratification 2/3.
50-60 cm	As above.
60-90 cm	Brown/dark brown silty peat; traces of sand; Humification 2; Wetness 2; Plastic; Stratification 2.
90-174 cm	As above; increasing wood content to a wood peat at 105 cm.
→ 186 cm	Increasing sand content.
197 cm	Top of gravel
226.5 cm	Maximum depth; angular sand and gravel; some silt and clay.

Additional core lying between dipwells 23 and 24.

0-13 cm	Black brown fibrous peaty soil.
14-25 cm	Brown peaty soil with root fragments.
25-50 cm	Silty clay / brown grey becoming increasingly grey, including brown organic soil and red plant fragments.
50-52 cm	Thick grey silty clay.
52-125 cm	Light brown, occasionally black compact silty herbaceous peat, abundant root fragments.
125-138 cm	Gradual boundary to grey silty peat with charcoal, root fragments, increased organic content.
138-265 cm	Abrupt boundary to compact wood peat, large wood fragments upto 5cm in diameter.
265-280 cm	Gradual boundary to green/grey silty peat with wood, extending to gravel.

Dipwell 24.

0-31 cm	Brown peaty soil; root fragments; crumbly structure; increasingly organic with depth; non-stratified.
31-50 cm	Orange-brown silty-clay; Stratified 1; no plant remains; mottled, stiff, plastic; Wetness 3/4.
→ 76 cm	As above.
76 cm →	Gleyed with root fragments; including grit and clay.
100-119 cm	Very wet.
119-200 cm	Stiff, grey silty clay; some plant fragments.
200-241 cm	Wood peat; Brown, unhumified; Wetness 2/3; Plastic, elastic 1/2; Wood and roots.
241 cm	Sand, silt and clay, with pebbles (angular sub-rounded).
257.5 cm	Gravel.

APPENDIX III

Evapotranspiration Model

The daily record of meteorological data recorded by data logger and measured using a Campbell Scientific CM10 weather station at Narborough Bog, was used to calculate daily evapotranspiration loss based upon the Penman equation. A vax VMS Basic program was written, and is given below:

```
5  OPTION ANGLE = DEGREES
   OPTION SIZE = REAL DOUBLE
10 DECLARE STRING CONSTANT FORMAT_STRING = &
   "### ##.###"
20 INPUT 'input filename';F$
25 INPUT 'output filename';OF$
30 READ LATUDE,SC
35 DATA 56.633, 1.94
36 OPEN F$ FOR INPUT AS FILE #1
37 OPEN OF$ FOR OUTPUT AS FILE #2
65 INPUT #1,JD,RH,T,IRRD,WS
66 IF (EOF) THEN 999
68 IRRAD=(IRRD*60*24)/59
   S1=(((T*.6136820929*(10^-12))+(.2034080948*(10^-09)))&
   *T+(.3031240396*(10^-07)))*T+(.2650648471*(10^-05))
85 S2=((S1*T+(.1428945805*(10^-03)))*T+(.4436518521*(10^-02)))&
   *T+(6.107799961)
90 SVP=S2*0.750061505
     D=23.45*COS((JD-172)*(180/184))
     H1=-(TAN(LATUDE)*TAN(D))
100 OPEN "TEMPH1.DAT;1" FOR OUTPUT AS FILE #3,  &
      ALLOW MODIFY,   &
      ACCESS MODIFY
110 PRINT #3, H1
115 CLOSE #3
120 EXTERNAL SINGLE FUNCTION MTH$ACOS
   DECLARE SINGLE H2,HS
INPUTROUTINE:
   OPEN "TEMPH1.DAT" FOR INPUT AS FILE #4
   INPUT #4, H2
   CLOSE #4
   IF (H2 < -1) OR (H2 > 1)
   THEN
       PRINT "Invalid cosine value"
      END IF
130 HS=MTH$ACOS(H2)
132 HR=(HS/PI)*180
     R1=(1440*SC)/(59*PI)
```

```
150 RA=R1*((HS*SIN(LATUDE)*SIN(D))+(SIN(HR)*COS(LATUDE)*COS(D)))
160 N=(((IRRAD/RA)/0.75)-0.18)/0.55
    ED=(RH/100)*SVP
180 RR=(((T*T*T*T)*0.000000122535)/59)*&
 (0.56-(0.09*SQR(ED)))*(0.1+(0.9*N))
190 H=IRRAD-RR
    EA=0.35*(SVP-ED)*((WS/100)+1)
    DELTA=(SVP/(273.16+T))*(6790.498/(273.16+T)-5.02808)
220 EVP=((DELTA/0.00066)*H+EA)/((DELTA/0.00066)+1)
    PRINT #2 USING FORMAT_STRING,JD,EVP
    GOTO 65
999 END
```

The program was derived with the aid of several published listings of evapotranspiration programs principally Young (1963); and Chidley and Pike (1970). Saturation vapour pressure was determined using the algorithm outlined by Lowe (1971).

The choice of the method to calculate potential evapotranspiration was limited as only incoming radiation was measured continuously throughout the period of study.

The Penman equation normally quoted is:

$$E = \frac{(\frac{\Delta}{\gamma} H + E_a)}{(\frac{\Delta}{\gamma} + 1)} \; mm/day \qquad (1)$$

where:

γ = constant of the wet dry bulb hygrometer.

$E_a = 0.35\ (e_a - e_d)\ (1 + u_2/\ 100)$ mm/day

e_a = saturation vapour pressure of water at T_a.

e_d = saturation vapour pressure of water at dew point.

u_2 = mean daily wind speed (miles/day).

Δ = slope of the saturation vapour pressure curve at mean air temperature T_a.

This is determined by the equation (from Young, 1963):

$$\Delta = \frac{e_a}{(273 + T_a)} \left[\frac{6463}{273 + T_A} - 3.927 \right] \qquad (2)$$

H = net radiation = A - B.

A = incoming short wave radiation.

$= R_a\ (1\text{-}r)\ (\ 0.18 + 0.55\ n/N)$

B = long wave out radiation.

$= \sigma T_a^4\ (0.56 - 0.092\ \sqrt{e_d})\ (0.10 + 0.90\ n/N)$

R_a = theoretically calculable amount of radiation received at earth's surface as given by:

$$R_a = 1440R(h\sin L\sin D + \sin h\cos h\cos D)\ mm/day \qquad (3)$$

L = Latitude.
R = Solar Constant.
D = sun's mean daily declination.
$h = \cos^{-1}(-\tan L \tan D)$
r = reflection coefficient.
n/N = ratio of actual to possible hours of sunshine.
$N = 24\,h / \pi$
σT^4_a = theoretical black body radiation.

The choice of the Penman equation as the method to estimate evapotranspiration loss was determined principally by the range of meteorological data which were available. The small field site is unlikely to possess a significant local climate by itself, and for security reasons the weather station was located in the garden of a house adjacent to Narborough Bog, as indicated on Figure 4.7. The Bowen ratio method requires measurements of temperature and humidity at two heights above the ground surface which was not possible. In addition the dependence of the Penman equation largely on a measure of incident radiation was an advantage, especially as it is unlikely that evapotranspiration rates would be moisture limited at Narborough where the water table lies close to the surface.

APPENDIX IV

Precipitation Records

In chapter 6 the discharge records of the river Soar were discussed from two gauges, which had been maintained by the National Rivers Authority at Narborough and Littlethorpe. The data were used to consider the frequency of flooding at Narborough Bog, and also provide an indication of climatic variation since 1971. Parallel records of precipitation have also been collected at the Narborough Sewage Works at Grid Reference SP 549966, and altitude 74.0m, over the period from 1972 to the present. These records were not considered in the main body of the thesis, because to a certain extent they provide similar information to the discharge data. However, the data are given here as they indicate the variability in total precipitation which has important implications for water levels at Narborough Bog.

The precipitation data are summarised in two graphs. In Figure A.1., annual precipitation totals are given for the period of data coverage, while in Figure A.2. the frequency of droughts within this period is identified, using the definition of drought as 15 consecutive days, none of which has rainfall exceeding 0.3mm (Shaw; 1988).

Figure A.1. demonstrates the magnitude of the difference in rainfall totals for 1991 and 1992, which correspond with the main water table monitoring period for the thesis. 1991 had one of the smallest rainfall totals in the period 1972 to 1992, while 1992 was the wettest. The contrasting graphs illustrating water table positions in the two years discussed in chapter 5 (eg. Figures 5.2. and 5.3.) thus provide a good indication of the difference in observed water levels under extremes of rainfall. Perhaps of greater significance than total precipitation is the timing of rainfall and its relationship to seasonal water loss through evapotranspiration. Figure A.2. indicates the temporal clustering of drought periods over the period. The drought of 1976 can be identified, as can a large period of water shortage in the previous year. In both years droughts occurred in mid-summer. Smaller droughts were also experienced in the summers of 1977, 1979, 1982, and 1984. There also appears to be a greater incidence of drought in the period from 1989 to the end of the record. This is likely to be important for ecological response, as the effects of drought will be cumulative and vegetation may take several years to recover from water stress. Equally one year of low precipitation, and low water tables, may enable species perferring a dry habitat to become established.

Figure A.1. Annual precipitation totals from Narborough, 1972 to 1992.

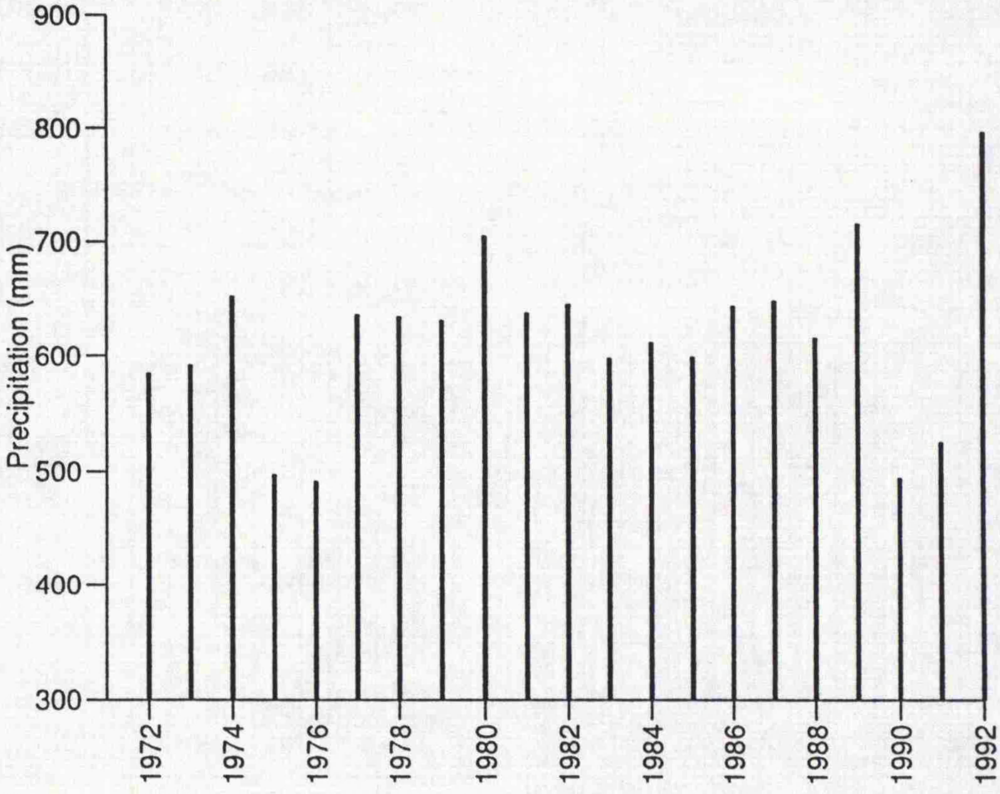
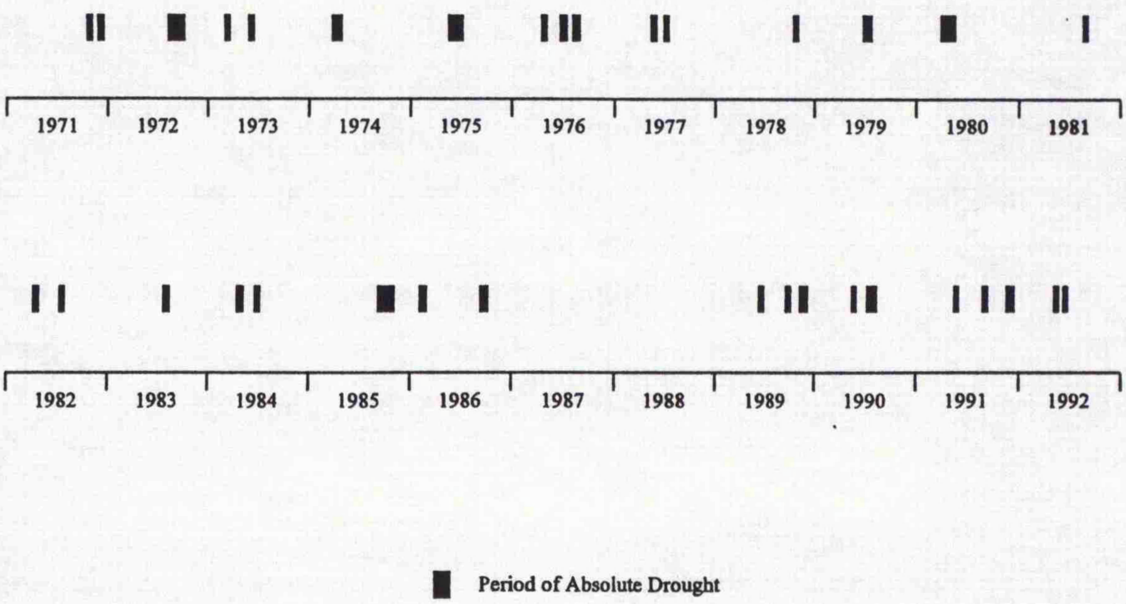


Figure A.2. Timing of periods of absolute drought from 1971 to 1992.



APPENDIX V

Estimates of Yield

Estimates of water storage were required to obtain a transient solution to the groundwater model. As discussed in chapter 7, storativity is the volume of water released from storage by a unit volume of aquifer under a unit decline of hydraulic head (Freeze and Cherry, 1979). For an unconfined aquifer, this equals specific yield. Yield is normally obtained from published values for modelling studies, and does not vary significantly. However, Boelter (1970) describes how the water yield of peat varies from 0.08 to 0.57, and therefore it was decided to undertake some simple tests to determine approximate initial values for yield. The water table records from a selection of dipwells were used to estimate the spatial variability of yield.

In chapter 5, multiple regression models were used to describe the time-series of water table variation. Rennolls *et al.* (1980) related water levels on individual days to preceding water table position and rainfall (equation 1, chapter 5):

$$h_i = a_i h_{i-1} + a_2 R_i + e$$

where a_1 represents the proportional decrease in the water table per day, and a_2 is a measure of the drainable pore space of the soil. This equation was used to obtain an indication of yield for dipwells 0, 1, 2, 3n, 3o, 4o, 5o, 6, 7, 8, 12, 23 and 25 in both 1991 and 1992. The results are given in Table A.5.

The tabulated data indicate only small variations in estimated yield. Over much of the area of Narborough Bog, yield is within the range 0.30 to 0.40; however the effects of localised clay deposits are apparent in the lower yield for dipwells 7 and 8 near the river.

Table A.1. Estimates of yield from Regression Model

Dipwell	1991	1992
1	0.310	0.319
3n	0.403	0.380
3o	0.369	0.337
4o	0.376	0.321
5o	0.367	0.327
6	0.323	0.280
7	0.097	0.067
8	0.131	0.194
30	0.325	0.209
12	0.441	0.305
25	0.378	0.287
23	0.241	0.398

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