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Multi-scale assessment of wetland hydrological function at a wet grassland in southeast England

David Jonathan Mould

UCL

Thesis submitted for the degree of Doctor of Philosophy
I, David Jonathan Mould, confirm that the work presented in this thesis is my own. Where information has been derived from other sources, I confirm that this has been indicated in the thesis.
Abstract

An examination of the hydrological functioning of a wet grassland in southeast England is undertaken within the context of scale issues. The importance of wetland hydrological functioning is demonstrated, alongside highlighting a gap in the literature of how scale issues affect contemporary wetland hydrological science. The subsequent assessment of hydrological functioning at a field site at Otmoor, Oxfordshire, is thus undertaken at both the field and catchment scales. At the field scale, an intensive field campaign over 18 months establishes the dominant hydrological processes as being precipitation and evaporation, the latter with losses of up to 5 mm day\(^{-1}\) driving an unexpected diurnal fluctuation in soil water levels. The dominant hydrological function was surface water storage, showing potential conflict with current land management practice of raising water levels for wetland restoration purposes. The impact of scale issues on numerical models is assessed through utilisation of a multi-scale model. At the catchment scale, the wetland's impact was assessed through increasingly complex numerical models, ultimately an event-based hydrodynamic model, and is shown to be significant to flood management downstream at Oxford. Decreasing peak flood flows through flood storage was the dominant function, as dictated by surface topography, whereas other online floodplain areas within the catchment increase time to flood peaks by attenuating flow through surface roughness, confirming the importance of wetlands to river flow. Surface roughness was therefore shown to be critical for wetland behaviour at different scales for different wetland types, indicating the importance of scale to wetland hydrological processes. The significance of initial model conceptualisation was demonstrated, and several recommendations were made for modelling procedures in order that scale issues be incorporated and prevented from causing complications in future modelling work. These include taking an iterative approach to increasing understanding through modelling, and linking models of different scales.
Acknowledgements

"Science is a group effort. Any one person only makes a small contribution."

Professor Stephen Hawking, BBC Today programme, 30/11/2006

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Simon, now I'm done, shall we get out on the bikes?
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Chapter 1

Scale Issues and Wetland Hydrology
1.1 – *Introduction*

The aim of this work is to investigate the role wetlands play in determining the hydrological behaviour of the catchment in which they are located. Due to the broad range of wetland impacts from local to basin-wide, this may be investigated at several different scales. It is proposed that there are implications for the outcome of research according to the scale at which the investigation is carried out.

This work is principally one of hydrology. It focuses on one wetland site, Otmoor located 5 km northeast of Oxford within the catchment of the River Cherwell. Chapter 1 initially introduces wetlands, the science of wetland hydrology and the methods of assessing the impact of a wetland on its catchment, namely wetland hydrological functioning. This approach identifies and quantifies dominant net hydrological transfers and their temporal variations, culminating in behaviour which characterises the wetland in the landscape according to its impacts.

As the hydrological assessment of Otmoor is undertaken within the context of scale issues, the second half of Chapter 1 discusses the relevant literature and methodologies which enable the appropriate scale framework to be developed for the subsequent investigations. The outcome of this review is the adoption of two independent scales for the hydrological assessment: the field scale and the catchment scale.
Chapter 2 describes a comprehensive investigation of the hydrological functioning of the Otmoor. The site is one of the UK’s most extensive semi-natural wetlands. Large areas of the wetland have in recent years been reclaimed from agricultural use, restored to wet grassland, reed beds and riparian areas through the use of water level management, the installation of surface foot drains, and extensive grazing. Much of this work has been done under the auspices of the Royal Society of the Protection of Birds (RSPB), whose primary objective has been the creation of a wide range of habitats for different bird species. An intensive field hydrological monitoring programme within the site was undertaken through two growing seasons and one intervening winter (May 2005 to August 2006). Data collection included surface water levels across the site, water table levels, soil moisture, meteorological conditions and direct evaporation measurements. Chapter 2 describes the use of these data to evaluate the major hydrological processes operating at the site and so the likely impact of the wetland on the wider catchment.

Since these hydrological investigations are set within the context of scale issues, there is a requirement to understand the fundamental principles of scale and scaling theories. One way in which scale issues are investigated is the assessment of the utility of high temporal resolution monitoring at the field scale, and so the benefits provided by more frequent visits and the use of monitoring technology. Furthermore, data collected during the field scale hydrological monitoring campaign are used to drive an energy- and water budget model at
different spatial scales. This part of the study is used as a tool to investigate scale effects in hydrological modelling.

Chapter 3 investigates the hydrology of the Otmoor site through the development of increasingly complex numerical models. Firstly a simple spreadsheet-based model is used to quantity the volumes involved in each major wetland process at a relatively crude level. The simplicity of this model enables a contrasting wetland site, Tadham Moor (Somerset), to be directly compared to Otmoor in terms of hydrological components, and so dominant hydrological processes.

The development of a complex hydrodynamic model of the River Cherwell catchment, including Otmoor, provides the culmination of Chapter 3. The model is used to investigate the behaviour of the Otmoor through the simulation of several scenarios in which the configuration of the model elements used to represent the wetland are modified. As the model coverage includes the floodplains adjacent to the River Cherwell, the behaviour of these wetlands is also investigated and contrasted to Otmoor.

Chapter 4 draws conclusions from the outcome of the work undertaken at different scales, showing the benefits of each and differences between the two. Recommendations are provided for future assessment of wetland hydrological functioning and the utilisation of numerical models with scale issues in mind. It is concluded that the knowledge gained through the assessment of wetland hydrological functioning at different scales at Otmoor is transferable and beneficial to research at other wetland sites.
1.2 – Wetland Hydrology

1.2.1 – Defining Wetlands

There are numerous and varied definitions of wetland environments, and much debate in the scientific community as to exactly what properties characterise a wetland. The definition adopted by the Ramsar Convention (a global treaty to provide a framework for wetland conservation) is as follows (Ramsar, 2004 p6):

"Areas of marsh, fen, peatland or water; whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish or salt, including areas of marine waters, the depth of which at low tide does not exceed six metres [and may include] riparian and coastal zones adjacent to the wetlands and islands or bodies of marine water deeper than six metres at low tide lying within the wetlands."

Mitsch and Gosselink (2000) have determined three primary components that are included in many wetland definitions:

1. Wetlands are distinguished by the presence of water, either at the surface or within the root zone;
2. Wetlands often have unique soil conditions that differ from adjacent uplands;

3. Wetlands support vegetation adapted to the wet conditions (hydrophytes) and are characterised by an absence of flooding-intolerant vegetation.

Hydrology is generally accepted to be the dominant factor in determining wetland distinctiveness; Gilman (1994, p5) agreed in his wetland definition which reflects his work on establishing the links between hydrology and vegetation:

"Wetlands are distinguished from other terrestrial habitats by having a significant excess of water for a large proportion of the time. This excess water imposes an important control on the natural vegetation."

Importantly, the ecotone nature of the hydrology required to produce wetland environments is often found at the boundary of rivers, lakes and other water bodies across the landscape. Hollis and Acreman (1994, p351) recognised this and placed it centre stage in their definition of wetlands:

"Wetlands are transitional environments between terrestrial and fully aquatic ecosystems which normally include floodplains, fens, bogs, shallow lakes and salt marshes."

This transitional nature not only makes wetlands particularly difficult to define, but also to delimit, and so causing the wide range of definitions and imprecision of some examples such as that of the Ramsar Convention. The sum of the above definitions and conditions of wetland environments may be summarised...
as being seasonally, periodically or permanently inundated zones forming the transition between terrestrial and aquatic environments.

1.2.2 – Wetland Degradation and Recovery

It is only in recent decades that wetlands have been accepted as a valuable component of the landscape, primarily due to the diverse habitat niches they provide. Wetlands were believed to have little or no value until this change in attitudes around the 1970s (Hollis and Thompson, 1998). The previous view of wetlands as wastelands (Acreman, 2000) stems from their lack of population, low perceived productivity and inaccessibility (Haslam, 2003). Reports of disease and strange lights (probably from natural methane production) prompted these vast empty tracts of the British landscape to develop myth and prevent settlement. Their place in history was established as a place to avoid, or in which to be banished. Indeed the potential for drainage or conversion is clearer to the untrained eye than the vast benefits wetlands can offer. Humans do not naturally adopt the precautionary principle and consequently wetlands have suffered greatly and continue to do so through modern times (Acreman, 2000).

This unfortunate image of wetlands through history led to the drainage of wetlands being seen for centuries as a progressive act, improving the land (Baldock, 1984). The need to feed increasing populations through the 20th
Century led to growing pressures on all habitats whilst mechanisation increased the efficiency of making such changes (Maltby, 1986). The sensitivity of wetlands to changes in hydrological regime left them susceptible to relatively small levels of river basin management (Gilman, 1994; Gavin, 2003a).

The benefits that wetlands provide for society have, until recently, been ignored (Maltby, 1986). In recent years the benefits wetlands bring have been highlighted, and many of these have fiscal value, such as tourism for bird watching and products such as reeds for thatching. Appreciation for these benefits has motivated the conservation movement to act for wetland protection and sustainable use (Thompson and Finlayson, 2001), and this movement became increasingly powerful from the 1970s onwards, resulting in a raft of legislation to protect wetlands, occurring at all levels of governance from local to global (Everard, 1997). Wetland conservation is embodied globally by the Ramsar Convention (Ramsar, 1971; Davis, 1994). This has given wetlands the status of being the only ecosystem with a dedicated global protection body (Maltby, 1986) and collaboration with United Nations Educational, Scientific and Cultural Organisation (UNESCO) and national governments gives credibility to protected wetland sites.

This and other global protection measures are summarised in Table 1.1. European legislation has also afforded protection to wetland areas in recent years, most notably with the introduction of the Water Framework Directive
which included innovative measures to bring holistic management of European water resources. This and other European legislation and measures which protect wetlands are listed in Table 1.2. At a national level in the UK, much legislative protection of wetlands is driven primarily by overarching European law, but each member state has unique interpretations of these. The UK Biological Action Plan (UKBAP), the UK's response to the Rio Earth Summit in 1992, also drives UK land management policy. The wetland management measures in place in England and Wales, such as Water Level Management Plans, are summarised in Table 1.3. Also included here are some more local initiatives for sympathetic land management at the local or regional level, driven by national policy.
<table>
<thead>
<tr>
<th>Measure</th>
<th>Participants</th>
<th>Protection afforded and effectiveness</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Convention on Wetlands of International Importance especially as Waterfowl Habitats (The Ramsar Convention) 1971</td>
<td>138 member states</td>
<td>Protection from development and outside influences limited for 1,675 wetland sites constituting 1.51M km² globally. Aim is conservation and wise use of wetlands to aid sustainable development. Gives credible and recognisable conservation status to a wetland site.</td>
<td>Maltby, 1986; Ramsar, 2004; Ramsar 2007</td>
</tr>
<tr>
<td>The Convention on Conservation of Migratory Species of Wild Animals (The Bonn Convention/CMS) 1979</td>
<td>101 member states</td>
<td>Protects a wetland if it supports an endangered migratory species. Action only when a species is endangered enough to be placed on Appendix 1 (threatened with extinction).</td>
<td>Everard, 1997; DEFRA, 1999; CMS, 2007</td>
</tr>
<tr>
<td>UN Convention on Biological Diversity (Agenda 21) 1992</td>
<td>190 parties</td>
<td>Goal of nations to achieve sustainable development whilst maintaining economic growth. In UK this manifested as UKBAP and in turn ESA and CS schemes in England and Wales (now Environmental Stewardship). Improved and sympathetic management of land, wetland restoration and protection.</td>
<td>Glowka, 1994; CBD, 2007</td>
</tr>
</tbody>
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Table 1.1 – Global wetland conservation measures
<table>
<thead>
<tr>
<th>Measure</th>
<th>Protection afforded to wetlands and effectiveness</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>EC Directive on the Conservation of Wild Birds (The Birds Directive) 1979</td>
<td>Protects endangered birds and their habitat including wetland areas. Protected areas classified as Special Protection Areas (SPAs) and contribute to the Natura 2000 network. Legally binding throughout the EU.</td>
<td>Acreman and Miller, 2003; JNCC, 2007</td>
</tr>
<tr>
<td>EC Directive on the Conservation of Natural Habitats and of Wild Fauna and Flora (The Habitats Directive) 1992</td>
<td>Protects endangered habitat types, including wetland areas. Protected areas classified as Special Areas of Conservation (SACs) and contribute to the Natura 2000 network. Legally binding throughout the EU.</td>
<td>Acreman and Miller, 2003; JNCC, 2007</td>
</tr>
</tbody>
</table>

Table 1.2 – European wetland protection measures
Abstraction licences response to 1963 Water Resources Act, CAMS now manages these through a consultation process involving all stakeholders. No licence issued if danger of “river flows, groundwater levels or water levels in wetlands” falling below the “minimum level required for the conservation of the aquatic environment”. Limitation is knowledge of what minimum levels are for wetland sites, and so effectiveness will vary. The principle aims of the process are 1) To make water resources and licensing information publicly available; 2) To provide a consistent approach to water resources management whilst balancing abstractors’ and the environment’s needs; 3) To encourage public involvement; 4) To manage time-limited licences; 5) To facilitate licence trading. CAMS must meet requirements of more recent EU legislation (see Table 1.3), e.g. providing catchment assessment needed for the development of River Basin Management Plans under the WFD.

An Environment Agency (EA) programme aiming to manage flood risk from a top-down approach, so from the river catchment perspective. CFMPs are based on detailed knowledge of the flood generation and dissipation mechanisms. Wetlands and natural floodplain function restoration are tools used to reduce flood risk and economic incentives are available to change poor land management practice where flood risk may be increased.

Programme of protection for listed habitats and species. At least 11 priority habitats are wetland environments, and many species utilise wetlands areas, most notably the Bittern (Botaurus stellaris) which resides in reedbeds.

Table 1.3 – Wetland protection measures in England and Wales
Countryside Stewardship scheme (CS)*

Introduced by MAFF in 1987 alongside ESA in response to the 1986 Countryside Act; similar to ESA but across the wider countryside. For benefit of more general nature conservation as opposed to specific target species. As with ESAs, CS is important in the implementation of UKBAP, yet was voluntary to the land owner or manager. 263,000 ha assigned (2.8% of agricultural and open grazed land in England). Deemed success, especially in lowland areas where response time is shorter.

Environmentally Sensitive Areas scheme (ESA)*

Focussing on 22 areas of less common habitat threatened by agriculture in order to support UKBAP species. Management prescriptions (decided upon by stakeholders) set in tiers of increasing protection and enhancement. Increasing tier levels have more extensive grazing, lower fertiliser application levels and higher minimum water levels (and so likely to represent wetland coverage). Payments made to farmers for this sympathetic management in return for revenue losses. Varied success, but 1M ha under agreement consuming £43M in 2002/3. General decrease in breeding wader numbers in ESA areas, with low uptake of higher tiers possible reason: where tier 1 only prevents further decline higher tiers would enhance habitat. Habitat restoration only effective for ~3 years due to lack of sustainable invertebrate supply. Increased ditch water levels less effective due to low hydraulic conductivities.

*CS and ESA schemes have now merged to form the Environmental Stewardship programme.

Water Level Management Plans (WLMP)

Managed by DEFRA, Water Level Management Plans (WLMP) balance the water level requirements in wetlands for multiple stakeholders. These include EA, English Nature, and local Internal Drainage Boards among others, and so account is taken of flood risk management, conservation and agriculture. The plans depend primarily on management by incentive schemes such as ESA and CS for implementation.

Local legislation

E.g. The Norfolk and Suffolk Broads Act 1988.

Table 1.3 – Wetland protection measures in England and Wales (continued)
1.2.3 -- Components of the Wetland System

There are several physical components of wetland systems that combine to create and maintain the wetland environment and give it ecosystem structure (Simpson, 2002). Each wetland is unique and its appearance and processes vary according to the local environment and pressures placed on it.

1.2.3.1 -- Hydrology

It is widely acknowledged that hydrology determines wetlands as distinct environments. For example, Hollis and Thompson (1998, p9) summarise:

"Hydrology modifies and determines the nature of wetland substrate and, together, these jointly allow specific ecosystem responses."

As discussed above, wetlands typically form on the fringes of more distinct water bodies such as rivers, lakes and oceans. These features may exchange water, sediment, nutrients and biological matter with wetlands, and their water regime may drive that of the wetland. This has an effect which is summarised in the commonly used model of wetland hydrologically-driven fluxes from Wicker et al. (1982). Figure 1.1 shows how the hydrology of a wetland modifies and determines the chemical and physical properties of the substrate and in turn determines its specific ecosystem response which may be in the form of sediment dynamics and nutrient fluxes, which sequentially directly impact upon flora and fauna composition. As wetlands represent the aquatic edge of terrestrial systems and the terrestrial edge of aquatic systems, small changes in
hydrology can result in significant biotic changes, in all other components of the wetland system including soils, vegetation and fauna (Mitsch and Gosselink, 2000).

Figure 1.1 – Impact of wetland hydrology on physical and chemical components (from Wicker et al., 1982)
A key feature of wetlands, the presence of water in the soil column, has several important impacts on processes that operate in wetland systems. Saturation from either water retention or high rates of water supply precludes aeration and so slows the process of decomposition, important for both nutrient cycling and chemical status. It also has implications for hydrological feedbacks, as a saturated soil column cannot accept more water and so for example cannot further buffer flood waters from an adjacent river.

Other components of the water cycle are intricately connected to wetlands, as hydrology is acutely connected to climate: rainfall and evaporation are often the primary source and loss respectively of water for a wetland system. Hydrology determines soil profile development and habitat provision for flora and fauna. Water and its movement dictate wetlands as landscape features and habitats, linking other wetland components.

1.2.3.2 – Soils
A key component of any wetland system is its soil structure, and this is often determined by the hydrology of the site. Wetlands can indeed be formed by the poor drainage of a soil allowing a build up of water, and at the very least the soil hydraulic properties are of great importance to wetland maintenance (Mitsch and Gosselink, 2000). The drainage or otherwise will influence the hydrological regime of the site and so impact upon other wetland components.

The presence of waterlogging in soils, a key feature of wetlands, has important repercussions for flora and fauna living in them. Primarily, oxygen diffusion rates in water are 10,000 times slower than rates in the atmosphere, and so
waterlogging restricts the availability of oxygen to the underground organs of plants (Wheeler, 1999). This in turn attracts plants which can tolerate anoxic conditions, and wetland areas therefore have a continuum of characteristic vegetation, from aerobic plants in dry areas and at the edges to anaerobic plants in poorer draining areas (Armstrong et al., 1995).

The high water content of wetland soils and subsequent low oxygen levels has an effect on the decomposition of plant material. Anaerobic conditions retard the breakdown of organic matter, and so wetland soils often develop very slowly (Haslam, 2003). An obvious example is the accumulation of peat through decomposition of vegetation, a process which facilitates the sequestration of carbon.

Saturated soils also develop distinct chemical signatures. Wetlands fed solely by precipitation have low nutrient inflows and are known as ombrotrophic, whereas a spring-fed wetland may be particularly nutrient rich, due to the water containing nutrients leached from the catchment (Haslam, 2003). Low redox potentials associated with low oxygen levels described above are associated with a reduced chemical status of some ions. In a mineral soil the chemical status may be seen clearly through the colour of the soil profile. A characteristic feature of saturated mineral soils is the development of a blue-gray colour as due to gleying, which is the result of chemical reduction of iron driven by the low redox potentials (Mitsch and Gosselink, 2000). Under aerated conditions, iron remains oxidised and red or brown in colour, allowing analysis of historic hydrological conditions from inspection of soil profile colour.
Wetland soils vary markedly across the continuum of wetland types. An organic wetland soil has a low bulk density (dry weight per unit volume) and so more pore space to hold water than a mineral soil (Bromley et al., 2004). The soil type impacts upon nutrient status, with the cation exchange capacity (sum of exchangeable cations: generally analogous to nutrient content) increasing with organic content. Mineral soils are not cation poor, but have more nutrients that may be unexchangeable, or chemically bound to the soil matrix. The reduced state of many nutrients under saturated conditions can increase their availability to plants, but so much so that toxicity may occur (Wheeler, 1999). It is clear that the development and dynamics of wetland soils and the interaction with hydrology and vegetation is a complex area of science.

1.2.3.3 – Climate
Climate is fundamental to wetland formation and status. As discussed above, wetlands often form in areas of excess water, and the prevalent weather conditions play a major role in water availability and so the water budget of a wetland. Precipitation is the key input for many key wetland types, and for some it is the sole input, such as in the ombrotrophic blanket bogs of upland Britain (Holden et al., 2004). Precipitation also drives river flow and its variability, and so indirectly supplies floodplain wetlands; similarly so for groundwater-sourced wetlands with a longer lag time (Acreman, 2004).

Evaporation is driven by a combination of climatic variables (Herbst and Kappen, 1999) including solar radiation, surface temperature, wind speed and direction and relative humidity (Shuttleworth, 1993; Roberts, 1999), and is a major conduit
for wetland water loss (Gavin, 2001). As wetlands exist in and contribute to particularly heterogeneous landscapes, meteorological conditions can change over short distances (van der Tol et al., 2003). As such, the monitoring of meteorological conditions is essential to successful understanding of the hydrological system of any wetland. As wetlands commonly have either water at the land surface, or vegetation linking the atmosphere to the water just beneath the surface, evapotranspiration has been shown in some circumstances to be equal to or greater than that of open water (Gavin and Agnew, 2000; Acreman et al., 2003a) and so may be inferred to have the potential to increase water loss from the terrestrial system, although this is not always the case.

1.2.3.4 – Vegetation

The unique hydrology of wetland sites attracts water-tolerant hydrophytic species such as many sedges (†Typha sp.), rushes (†Juncus sp.) and reeds (†Phragmites sp.) (Haslam, 2003). These form in communities according to the conditions created by a combination of hydrology, soil properties and micrometeorology, and so form a subtle continuum across a wetland site, a process producing spatial heterogeneity known as zonation (Kotowski et al., 2001). In time this manifests as succession (Kotowski, 2006), whereby the community composition reacts to changing conditions in any of the other wetland physical components. These changes also represent other ecological pressures such as competition, herbivory, disturbance, site management and water quality, but hydrology and the other wetland physical components described here are dominating factors (Kotowski, 2006). The niche conditions found at wetland sites encourage rare species and communities, often attracting conservation status for wetland areas.
Subsequently, much ecological research is undertaken in wetlands and this has often lifted their profile, such as at Redgrave and Lopham Fen, Suffolk, UK (Smith, 2006).

Transpiration from vascular plants contributes to the loss of water from wetland systems, and it is particularly difficult to separate this distinct process from evaporation (Callaghan, 2007). The two are commonly amalgamated and the term evapotranspiration used to describe the combined water loss. The interaction of plants and their physiology is a complicating and important factor in the estimation and calculation of evapotranspiration (Roberts, 2007), and vegetation composition thus has an important feedback to the hydrology. As plants form a direct connection between the soil water and the atmosphere when active (Tyree, 1999), van den Honert used the analogy of electrical conductivity as far back as 1948. Inversely to the direct effect of plants on climate, climate has a large influence on vegetative composition through temperature and sunlight gradients, wind stress, frost and through manipulating levels of available water (Wheeler, 1999). Wetland vegetation also has a long history of providing resources for Man; reeds have long been used for thatching and home building and many plant species are useful to humans (Haslam, 2003).

Vegetation has an important impact on flow caused by its higher surface area and so larger resistance to flow than a conventional river channel, retarding the flow of water across a wetland site relative to a channel. This effect can be described as higher surface roughness, and when combined with a shallow topographic gradient often associated with wetlands, results in low flow velocities.
(Baker et al., in press) and forms an important interaction between hydrology, vegetation and soils. A lower velocity activates deposition of sediment (Knighton, 1998) and so can initiate an accretion of sediment layers from flood events in wetland areas. As well as slowing the flow of water, vegetation can alter water quality by absorbing nutrients and pollutants dissolved and suspended in the flow (Wilson, 2007). Indeed, sediments deposited on floodplain meadows acted as a traditional fertiliser before river flows were so regulated and regular flooding was more common (Gowing, 2006).

1.2.3.5 – Fauna
A wetland soil’s anaerobic nature also attracts particular soil fauna, and may be rich in invertebrates (Ausden et al., 2001), which may in turn be prey for larger mammals and bird species. As such, wetlands are widely renowned for large and varied bird populations, and many wetlands are protected, restored or created for this purpose. In Europe, wading birds, such as Lapwing (Vanellus vanellus), Snipe (Gallinago gallinago) and Redshank (Tringa totanus) probe the soil with long beaks for prey, and so a damp, soft wetland soil surface provides feeding areas (Milsom et al., 2002).

Many such waders have declined in abundance with the loss of wetland habitat in Britain over recent decades (Ausden et al., 2001), and now much wetland restoration is funded through schemes aimed at providing habitat for wading birds. In the UK, groups such as the Royal Society for the Protection of Birds (RSPB) and the Wildfowl and Wetlands Trust (WWT) specialise in purchasing land and restoring or creating wetland habitat for specific bird species. Government policy is aimed at increasing biodiversity in general, such as the
ESA scheme (recently incorporated into the Environmental Stewardship scheme and managed by Natural England and sister agencies). Often driven by European policy for habitat and species conservation targets, such schemes are often undertaken in wetland areas using management tools such as raising water levels, which creates appropriate habitat and subsequently attracts target species (DEFRA, 2004).

1.2.3.6 - Summary
Several components combine to produce a wetland system: hydrology, soils, climate, vegetation and fauna. The hydrology is commonly regarded as central to determining wetlands as distinct from other landscape units, and directly influences other components significantly to also create unique features in each. All components interact and must be considered if a holistic assessment of wetland management is to be undertaken.

1.2.4 – Wetland Water Transfer Mechanisms

Above it has been established that hydrology is essential to wetland systems. The way in which water moves into, out of and through wetland systems is also fundamental to the processes occurring in wetlands. A description follows of the principal methods of water movement in wetland systems, referred to as water transfer mechanisms after Acreman (2004). Some mechanisms of water transfer combine with stores, making the distinction between stores and fluxes vague; an example of this would be lateral transfer in the water Table. This section will introduce each water transfer mechanism.
1.2.4.1 – Precipitation

Wetlands primarily occur in areas of water surplus (Acreman, 2004), the source of which can often be high rainfall (Baker et al., in press). Rainfall is therefore fundamental to any attempt to characterise the hydrology of a wetland site, especially through quantification of a water balance or turnover rate (Hollis and Thompson, 1998). Rainfall is the primary water source for many wetland types, but may be the only hydrological input to many ombrotrophic upland bogs.

Although inter-annual variation may be great, rainfall generally has well-defined seasonal patterns, especially in the UK where winters generally have higher rainfall totals (frontal in origin) and summers have less frequent but intense (convective) rainfall. The fate of rainfall is determined largely by vegetation structure within wetland systems, and varies in ratio between interception from vegetation (and evaporation thereof), stemflow (literally running down the stems of plants to the ground surface) and throughfall directly to the surface (Mitsch and Gosselink, 2000; Callaghan, 2007). After collection on the surface, destinations include direct evaporation, infiltration into the soil profile (for lateral or vertical transfer or loss through transpiration from plants), or overland flow.

1.2.4.2 – Evaporation

Due to the very nature of wetland sites facilitating a close interaction between surface water and the atmosphere, evaporation can be a significant conduit for water loss from wetland environments; water is often at or near the surface, access to water is not a limitation as it is in fully terrestrial environments. Transpiration, the loss of water from plants to the atmosphere, is very difficult to distinguish from evaporation, and so the two are commonly combined and
termed evapotranspiration. Indeed, water loss through evapotranspiration can in some cases be higher than that of open water, as transpiration from plants exacerbates total water loss (Gavin and Agnew, 2000; Acreman et al., 2003a).

Many factors affect the rate of evapotranspiration from wetland sites, but in general hot, dry and windy conditions encourage evaporation (Callaghan, 2007). It is therefore primarily meteorological conditions that dictate evapotranspiration levels to a great extent. The vegetation cover and type is also important, as different vegetation types transpire at different rates, and deeper rooted plants such as *Phragmites* sp. can access lower water tables (Roberts, 1999).

1.2.4.3 – Runoff

Runoff from any upland catchment areas can bring water to or take water from a wetland site. Runoff can be induced by two processes: 1) infiltration excess, also known as Hortonian runoff, whereby rainfall rate exceeds the infiltration capacity of the soil, and 2) saturation excess, also known as Dunne runoff, whereby the soil becomes saturated and cannot absorb any more water (Kubota and Sivapalan, 1995).

Water may contain nutrients collected as it passes over the surface or through shallow substrates, and so lead to the development of minerotrophic (nutrient rich) wetlands such as fens (Kotowski et al., 2001). Runoff quantity and quality can be difficult to measure directly due to its diffuse nature and complex sources.
1.2.4.4 – Groundwater Interaction

Wetlands may occur at, or facilitate the development of, sites where surface water interacts with groundwater (Acreman, 2004). Groundwater recharge occurs where surface water passes into a regional aquifer; groundwater discharge vice versa. As the process happens beneath the surface and over a wide and indistinct spatial coverage, groundwater interaction is difficult to measure, complicated by the extremely long residence times for some regional aquifers. As with runoff, water from a groundwater source can have a variable nutrient status.

Some wetlands are dominated by groundwater interaction (Gilvear et al., 1997), but more frequently it may be a smaller component of the water balance (Acreman, 2004). The properties of the soil profile are important in determining the level of groundwater interaction. Wetlands can be isolated from the regional groundwater system by a substrate with a very low hydraulic conductivity, but spring flow may bypass this impermeability. Another important factor is the potentiometric head of water in the wetland relative to that of the groundwater table (Mitsch and Gosselink, 2000), as this will affect the rate of water transfer.

1.2.4.5 – Flood Event Inundation

Floodplain wetlands exchange water, sediment and nutrients with the river system during a flood event, with the amount depending on the topographic characteristics and position of the wetland, as well as the flood magnitude (Hughes and Rood, 2001). Sediment deposited will have been entrained by the high energy of the river during high flows, and may be deposited on the
floodplain as the water slows on the wider floodplain (Knight and Shiono, 1996; Knighton, 1998), which may bring significant nutrient enrichment to the wetland system (Gowing, 2006).

1.2.4.6 – Subsurface Storage

Wetland subsurface water storage is typically very high, as the water table is frequently at or near the surface, and often increased by surface storage ditches that are created to hold water for summer use or to drain excess water (Gilman, 1994). Water can be transferred either laterally or vertically whilst being stored in the soil profile, depending on any water level gradients that may be present within a wetland area, or from surrounding terrestrial or aquatic areas. The inherent heterogeneity of soil properties (Wood, 1995) leads to stark differences in levels of both storage and transfer rates in the soil medium across even short distances (Ward and Robinson, 2000).

1.2.4.7 – Other Water Transfer Mechanisms

Coastal wetlands may have tidal as well as riverine influences, and this may have a significant impact on the water balance over a short period of time. An input of brackish or saline water will also have a large impact on water chemistry and so floral and faunal composition, with halophytic species typifying coastal wetland environments (Mauchamp et al., 2002). The hydrological regime will likely be dominated by the tidal ebb and flow, and this will affect other components of the regime such as groundwater discharge, which may only
happen at low tides depending on water table elevation (Mitsch and Gosselink, 2000).

As contemporary wetlands are highly managed environments, there is frequently a water management programme (Mould, 2007), where water can be efficiently pumped on or off site to different stores within the site. Weirs and gates can also manage water levels using gravity, often feeding a dense network of surface drains and ditches which may be also act as wet fences to control stock. The maximum distance in-field that the water levels in the ditches have an effect is highly dependent on the soil properties. Peat soils have a generally high porosity and subsequently higher hydraulic conductivity, and the ditch water level may be reflected in the field up to 20 m away (Gilman, 1994). A soil with a higher clay content and so lower hydraulic conductivity will transfer water laterally less readily, with the consequence that the ditches have much less impact on in-field water levels (e.g. Thompson et al., 2004). As such, the water levels in the field centre are left to be dictated by rainfall and evapotranspiration (Acreman et al., 2007).

Figure 1.2 summarises the water transfer mechanisms that may influence the hydrological balance of an inland freshwater wetland system. Every wetland is unique in terms of location on the landscape and relative to other surface water bodies, and so may utilise any combination of these transfers and stores. It is therefore important when beginning to assess the hydrology of a site where water transfer mechanisms are likely to be present, to determine which can be excluded (for example due to soil type preventing groundwater interaction), and
which have an unknown status. This may provide guidance for further investigations; it would be naive to use such a broad classification to determine specific hydrological functions present. This may follow the risk-based approach advocated by Acreman (2004), whereby a conceptual model of site hydrology is started using maps and other remote data in the form of a desk study. This is developed with a site visit, advice from experts (local farmers, site managers, hydrogeologists etc), hydrological monitoring and if data are sufficient, hydrological modelling. With each step, the conceptual model of the wetland hydrology is developed and improved until the required level of understanding is reached for the application, dependent upon financial and other resources available.

![Figure 1.2 — Possible water transfer mechanisms in a freshwater wetland](image)

### 1.2.5 — Wetland Hydrological Regimes

Each water transfer mechanism detailed above has its own temporal regime. For example, rainfall may be seasonally dominant (e.g. North Kent Marshes;
Hollis and Thompson, 1998), tidal inflow will be highly regular (e.g. Severn Estuary tidal wetlands; EN and CCW, 2005) and groundwater inflow is likely to be more stable (e.g. Redgrave and Lopham Fen, Suffolk; Smith, 2006). As such, the combination of water transfer mechanisms which constitute a specific wetland system will produce a distinct hydrological regime. This will be true for every wetland site, and the complexity will be determined by the number of water transfer components composing the wetland's hydrology. Each hydrological component will have one of the following temporal characteristics, making wetlands characteristically dynamic landscapes.

The cycle of the seasons through the year is caused by the Earth's axis being tilted relative to its orbital plane, exposing alternate parts of the planet more directly to the Sun's rays over the year. Precipitation reflects climatic conditions, and in Britain this is exhibited by higher precipitation totals in the winter months from frontal weather systems. However, convectional summer storms can produce intense rainfall events and so high daily peaks (Kay et al., 2006a). Incoming radiation and so temperatures also vary on a seasonal cycle and peak during the summer months: this cycle of available energy is reflected in the regime of several water transfer mechanisms. Evaporation is an important example, as it is a primary loss of water in wetland systems (Gavin, 2001), and peaks accordingly. In many wetlands evaporation can be equal to or greater than precipitation levels in summer months (Hamilton et al., 1997; Gavin and Agnew, 2004); this imbalance in energy through the year, coupled with a disparity in rainfall, results in a significant increase in available water in winter months. Accordingly, many wetlands display higher water levels in winter,
exemplified by Wicken Fen in Cambridgeshire, UK, which exhibits water levels 80 cm higher than in summer months (Gilman, 1994).

A second strong cycle in available energy is driven by the rotation of the Earth, manifesting itself as day and night. Incoming radiation peaks at midday, as the sun reaches its zenith, and this has a strong effect on the water regime, again directly affecting evaporation and transpiration losses, but on a daily basis. Amongst others, Wetzel (1999) has reported a wetland water table exhibiting a distinct diurnal periodicity in levels due to the water demands from evapotranspiration during daylight hours, with head recovery during darkness as demand is removed. Hays (2003) and Frahm (2007) have been able to model evaporation levels from interrogation of similar diurnal traces.

Longer term trends, including groundwater level drift (monthly to decadal) and climate change (over centuries) can affect wetland hydrology (Kotowski et al., 2001). Rainfall changes influence groundwater levels, but with lag times of months and years, and groundwater levels affect interaction with surface water, and so groundwater recharge and discharge (Krause and Bronstert, 2004). Many wetlands have been degraded or lost due to abstraction for irrigation and public supply lowering groundwater levels and preventing hydrological transfer to wetland sites (e.g. Las Tablas de Damiel, Spain: Llamas, 1989; Redgrave and Lopham Fen: Harding, 1993).

The rhythm of ebb and flow of the tides directly affects wetlands lying at or near the coast. Higher relative sea levels prevent outflow and vice versa. Dependent
on the wetland topography, there may also be a twice-daily inflow of saltwater, impacting upon both the hydrology and chemical balance of the system. Such a strong chemical influence will have a direct effect upon vegetation able to colonise such an area, and an example of a tidal-influenced wetland is Ramsar-designated Alberfera on the Balearic island of Mallorca (Howe, 1989).

Many events occur randomly, with no particular periodicity. Rainfall, although dominant in one season, occurs in unpredictable events, and as such flooding and associated inundation from overbank events to wetland systems may occur at any time. A similar situation occurs with drought events.

The different impacts of water level between these temporal regimes are shown in Figure 1.3, with the y axis of temporal regime length being logarithmic. Also shown is an example of a combination of temporal regimes, where each water transfer mechanism has an impact upon the unique hydrological regime of the (theoretical) wetland in question.

Wetlands are relatively dynamic ecosystems when compared to other landscape forms (Maltby, 1986), as they are transitional landscapes, both temporally and spatially. Temporally, wetlands often form the intermediate landscape after disturbance and before climax vegetation has been achieved; spatially, wetlands form the transition between terrestrial and aquatic ecosystems, and so form riparian habitat (Haslam, 2003). Being transitional, wetlands often experience sedimentation or erosion, and due to this, a site may change dramatically during the course of a single storm, which can transport vast amounts of sediment.
Over longer time scales vegetation and sediment build-up creates uplift and so can create bogs and floodplain wetlands respectively, and at least dramatically changing the hydrological dynamics of the wetland. Wetlands rarely represent vegetative climax in a landscape. As spatially transitional landscapes, wetlands often buffer terrestrial ecosystems from disturbance by water, though this very disturbance regime prevents climax ecosystems to develop (Haslam, 2003).

Figure 1.3 – The effect of various temporal hydrological regimes in wetlands on water level
1.2.6 – Wetland Hydrological Functioning

Physical, chemical and biological processes occur widely in wetlands as in other environments, and are defined by Simpson (2002) as changes that occur naturally. It has been demonstrated that wetlands perform a range of functions which result from the interaction between processes operating and ecosystem structure, which is the combination of wetland components described in Section 1.2.3 (Maltby et al., 2005; Simpson, 2002). Wetland functions may result in several groups of benefits, namely goods (e.g. reeds for thatching, fish), services (e.g. improving water quality) and attributes (e.g. birds to watch), collectively referred to as values, which society may utilise for its own means. The overarching group of wetland functions may be split into several subgroups including physical hydrological (hereafter ‘hydrological’) and water quality functions. Not all functions are positive, and many hydrological functions indeed have little effect, but the functions wetlands perform have in many cases determined their use, especially as flood defence and water quality buffers. For example the function of reducing nutrient loading has led to wetlands being coined ‘the kidneys of the landscape’ (Mitchell, 1994), and as such wetlands have frequently been used as buffers for the treatment of waste water (Fisher and Acreman, 2004).

Simpson (2002) highlight that there is some confusion in the literature as to the definition of ‘function’ and ‘value’, but the definitions described above shall be adopted here. There are also varied groupings of functions used by different authors, including combinations of hydrological, biological, chemical, ecological and societal. Importantly to wetland hydrological science, the combination of
different wetland water transfer mechanisms, working under various wetland hydrological regimes, provides each wetland with signature hydrological functioning. A wetland function is therefore the result of interactions between wetland components (Thompson and Finlayson, 2001), and a wetland performs a hydrological function by changing the characteristics of the flow from the upstream catchment to the downstream outflow (Mould and Acreman, in preparation), which may be beneficial or detrimental to human concerns. Wetland functions, as opposed to the values of a wetland, are those attributes or uses that humans may have no opinion of; those which are unambiguous (Lewis, 2001). Wetland benefits are summarised in Table 1.4.

Hydrological functions that a wetland may carry out can be utilised by Man for the management of a river basin, as well as being a natural and integral part of the hydrological cycle. Functions may be visible from many water transfer mechanisms and there is a distinction, and indeed some ambiguity, as to whether a function occurs because the wetland is present, or the wetland exists because the function occurs (Bullock and Acreman, 2003). One example of this juxtaposition is the typical groundwater discharge site, where it is unclear whether the wetland has formed around an existing discharge site, or if discharge has been facilitated by the wetland's existence. There is often a lack of differentiation in the literature on this issue (Bullock and Acreman, 2003), as studies of groundwater interaction are in reality very difficult to quantify, due to the diffuse and hidden nature of the phenomenon (Hayashi and Rosenberry, 2002).
### Table 1.4 - Benefits of wetlands

Wetlands have traditionally been thought of as acting as a ‘sponge’ and Lewis (2001) epitomises this assumed behaviour with his summary of three key hydrological functions of wetlands as 1) temporal spreading of flow, which leads to moderation of discharge volume and velocity; 2) spatial spreading of flow, which leads to moderation of velocity and 3) maintenance of contact between groundwater and surface water, which leads to exchange. The broad array of literature which takes measurements of hydrological functioning has shown these assumptions to be incorrect (Bullock and Acreman, 2003; Mould and Acreman, in preparation), and for wetlands to have a varied impact on flow.
The role of groundwater is very dependent on local conditions, both at the wetland site and below ground in any connected regional aquifer. For example, Choi and Harvey (2000) found almost exclusive groundwater recharge at a site in Florida, USA, but with short yet distinct periods of groundwater discharge. However, other studies showed clear trends in groundwater flows, such as Wolski's (2002) work in the Okavango delta showing consistent groundwater recharge. Similar findings were demonstrated by Hayashi et al. (1998a; 1998b) in Saskatchewan, Canada and Logan and Rudolph (1997) in Argentina. In Europe, Wassen et al. (1990) found groundwater discharge to predominate in a semi-natural mire in Biebrza National Park, northeast Poland and Gilvear et al. (1997) found a similar situation in a 'hydrologically complex' wetland in Norfolk, UK. Weng et al. (2003a) and Brunet et al. (2003) both found evidence of groundwater discharge at French wetland sites. The delicate dependence of some wetland sites on groundwater levels is demonstrated by the example of Redgrave and Lopham Fen on the Norfolk-Suffolk border in the UK, as described by Harding (1993) and Smith (2006). A local groundwater abstraction was decreasing local aquifer water levels and endangering the habitat and nationally endangered fen raft spider (*Dolomedes plantarius*); removal of the abstraction in 1999 led to a rapid recovery of both water levels and the spider population.

Wetlands are traditionally regarded as beneficial to managing flood events, acting to reduce peak flows and time to peaks by absorbing water during times of excess and releasing it slowly like a 'sponge'. Some numerical hydrological studies do agree with this view, such as Hillman's (1998) study of a floodplain
wetland in Alberta, Canada demonstrating attenuation in both volume and timing. Hardy et al. (2000) concurred with analysis from a floodplain wetland in Devon, UK, which increased the time to peak flow. However, some wetlands may display hydrological functioning characteristics that may exacerbate flooding. One notable example is that of McCartney's 2000 work which concluded that the Zimbabwean wetland site principally generated storm runoff. Another illustration of this is Waddington et al.'s (1993) work which concluded that the groundwater-linked site in Ontario, Canada generated significant storm runoff.

The 'traditional' view of wetlands absorbing high flows has the consequence of releasing the water collected over a longer period of time and thus maintaining the base flow, or low flow of a river. Examples of this can be found in the scientific literature, such as Raisin et al.'s 1999 work in Victoria, Australia, where a baseline discharge was found to persist. However, there is evidence to the contrary, and examples include Spieksma's (1999) work in Germany where a raised bog did not sustain low flows, and Burt's 1995 work in a Lancashire (UK) peatland which produced minimal summer baseflow. Gilvear et al. (1993) found that the high levels of summer evapotranspiration prevented significant water volumes from draining to the river network, and so interrupting any base flow maintenance. Gibson et al.'s work, also in 1993, drew similar conclusions but for different reasons: the wetland site in the Northwest Territories, Canada, frequently froze and so prevented outflow.

A 'typical' wetland would theoretically reduce flow variability through the above functions (Acharya, 2000). Gilvear et al. (1993) showed that wetlands could increase flow variability through decreasing in baseflow generation. Bucher et al.
(1993) concurred, with modifications to the floodplain wetland preventing it from acting to decrease flow variability. Some evidence does show wetlands to decrease overall flow variability of river flows, often through floodplain wetlands reducing flood event volume (Leblois and Sauquet, 2000; Brunet et al., 2003, both France).

Bullock and Acreman, in their 2003 review of the global scientific literature on wetland hydrology, concluded that published hydrological literature contrasted the traditional view of wetlands acting as a sponge and always to reduce floods, promote groundwater recharge and regulate river flows. They concluded that this may be true for some wetland types such as floodplains, but not for others, such as headwater wetlands, which have a more varied effect, including many which increased flood peaks. In addition, the published literature suggested that many wetlands increase annual evaporation rates and rather than regulate river flow, may indeed decrease low flows and so increase flow variability. Mould and Acreman (in preparation) primarily concurred with Bullock and Acreman (2003), in their European-focused literature review, concluding that wetlands predominantly reduce flows, often through exacerbating evaporation levels and so facilitating the loss of water from the river basin system. Furthermore, European wetlands in the literature studied consistently diminished low flows and floodplain wetlands were seen to contribute water, primarily through facilitation of groundwater discharge (Mould and Acreman, in preparation). Another important finding was that although floodplain wetlands may generally decrease the downstream impact of floods; headwater wetlands increase both the magnitude and shorten the time to peak of the flood events studied.
With such a back catalogue of contrasting wetland hydrological functioning assessments, it can appear at first to be a confusing picture emerging from the relatively new field of wetland hydrological sciences. However, it is proposed herein that the opposite is in fact true, and this confusion is indicative of a very important feature of wetland hydrology and the functioning it provides to wider catchment river flows. It is the very nature of wetlands described in the preceding sections which creates this variability: the site specific nature of wetland hydrology is fundamental to this issue. There is therefore no use in applying a broad theory, such as that of the ‘sponge’, or indeed arguing the opposite is true, on the basis of one or a few studies. It seems essential that any attempt to describe the hydrological functioning of a wetland site is useless without direct measurement and quantification of water transfer mechanisms, and establishment of any hydrological regime that may be operating. The science of wetland hydrological functioning has continued to establish itself in recent years as integral to the understanding of wetland systems as a whole (Mould, 2007), and in particular research on flood storage and the flood alleviation potential of wetlands concentrates on hydrological functioning. This must continue and be extended for wetlands to be adequately managed in a catchment context and for them to contribute as tools to any such catchment management. Their potential through hydrological and other functioning is great in this area, but only with appropriate understanding of their hydrological systems and responses. Continued protection of wetlands as landscape units, of their functions and values, and their conservation for future generations necessitates proper understanding.
1.3 – Scale Issues and Hydrological Science

1.3.1 - The Reason for Scales Issues

Although the concept of the hydrological cycle taught at a fundamental level is a straightforward one in essence, it is an extreme simplification. It represents an unknown number of physical processes and interactions, fluxes, stores and feedbacks, incorporating all fragments of the biosphere from the high atmosphere to the deep lithosphere. Tchiguirinskaia et al. (2004) highlighted the obvious: that a common feature of these processes and phenomena is their huge variability over a wide range of space and time scales, and it is widely accepted that hydrological processes operate at an extensive range of scales (Blöschl and Sivapalan, 1995). Klemes (1983) proposed that this included eight orders of magnitude in both space and time, from molecular interaction to the well known global hydrological cycle, first proposed by Edmund Halley in 1750. Doodge (1988) has suggested this range might be as high as 15 orders of magnitude.

Interactions force the nonlinearity of these simultaneous processes to propagate through the physical world we see and study with alarming speed (Tchiguirinskaia et al., 2004). Six key issues have been highlighted by Schulze
(2000), based on work by Harvey (1997) and Bugmann (1997) whereby differences in scale may be introduced to the hydrological system:

1. Spatial heterogeneity in surface processes;
2. Non-linearity of responses in the natural world. These can be divided between episodical (e.g. rainfall), cyclical (e.g. evaporation) and ephemeral (e.g. flood events);
3. Processes require threshold scales to occur;
4. Dominant processes change with changing scale;
5. Development of emerging properties with scale, e.g. edge effects change with scale;
6. Disturbance regimes.

The term 'scale' refers to a rough indication of the order of magnitude of the work in question (Blöschl and Sivapalan, 1995), rather than a specific value (Schulze, 2000). Wood (1995, p89) summarises the reason for the introduction of scale issues to hydrological sciences succinctly, and indicates that item 1 above is imperative:

"The complex heterogeneity of the land surface through soils, vegetation and topography, all of which have different length scales, and their interaction with meteorological inputs that vary with space and time, results in fluxes of energy and water whose scaling properties are unknown."

Scale issues are not new to science, and have been incorporated in many other fields (Blöschl and Sivapalan, 1995) such as meteorology (Avissar, 1995),
geomorphology (de Boer, 1992) and soil science (Hillel and Elrick, 1990). Hydrology appears to be somewhat more complicated due to the comprehensive coverage across all scales that the field encompasses, and the legacy of hydrologists focussing on the catchment scale processes (Wilby and Schimel, 1999).

1.3.2 - Problems and Complications of Scale

1.3.2.1 - Introduction
Cammeraat (2002, p1201) suggested that scale issues pose "one of the major challenges in the fields of physical geography, hydrology and ecology". Recent appreciation of scale issues would suggest this to be developing, including the publication of special issues of several major journal titles (e.g. Journal of Hydrology, volume 217, issue 3-4, 1999; Hydrological Processes, volume 18, issue 8, 2004) with scale themes.

Bergkamp (1998) identifies a primary problem which arises when dealing with scale. Processes are commonly studied intensively and the interaction of processes are understood as a result of this only at distinct scales. However, there is often no synthesis of observations taken at different scales.

Systems in the natural environment are organised hierarchically (Schulze, 2000). For example in a river basin, many upslope areas flow into slightly fewer, but larger hillslopes; these in turn flow into fewer, but larger sub-catchments, and so
forth through catchments to basins. We must remember that these different scales are not discreet and unconnected from each other, but that there are transfers and feedbacks which occur within each scale and also across overlapping scales (Harvey, 1997). For example, lateral groundwater flow is a local process affecting the location of certain wetlands (e.g. Lopham and Redgrave Fen, Norfolk), but it is also a regional process transferring water between basins and supplying drinking water, such as the Thames basin aquifers.

Scientists, including wetland hydrologists, often work across spatial and temporal scales without regard to the problem of scaling, nor appreciating the implications of it on their work (Schulze, 2000). The problem of scaling is therefore tremendously important to scientists on the whole, and as water accesses and affects every facet of the physical environment from the soil pore to the open ocean, these issues are especially pertinent to hydrologists. Several issues transcend hydrological work and will be discussed further below. These include scaling (upscaling and downscaling), and the relationship between spatial and temporal scales.

1.3.2.2 – Scaling

Scaling is the transfer of information across scales (Blöschl and Sivapalan, 1995) and is required most commonly due to discrepancies in the scales at which data are collected and at which predictions are needed (Western and Blöschl, 1999). It is clear that properties important to hydrological sciences change readily across many spatial scales, as exemplified by the notoriously high spatial
heterogeneity of hydraulic conductivity in wetland environments. Such spatial heterogeneity causes the fundamental differences in hydrological response between hillslopes, sub-catchments and basins, and there is often a need in hydrology to change from one scale to another, either transferring results of investigations or using observations as an input to a model. Refsgaard (2007a) summarises that scaling is basically a question of how to handle heterogeneity of physical properties at different spatial and temporal scales. Information is lost when scaling occurs (Blöschl and Sivapalan, 1995): aggregation of data is essential with downscaling and some form of extrapolation or infilling is required for upscaling. This is the fundamental reason why the correct scale must be chosen from the outset when embarking upon data collection, so as to minimise scaling requirements.

Bierkens et al. (2000) reviewed the methodology of scaling, and some standard terminology was established which was rational and appropriate for the study of scale. As the current work is predominantly hydrological research, the more common terms will be maintained as associated with hydrological modelling. This terminology is summarised in Table 1.5 and accounts for the domains of both space and time.
A key point is that each of these parameters can be scaled, and it is important to realise which you are dealing with. Figure 1.4, below, shows the processes involved with each, and the terminology used.

It becomes clear from Figure 1.4 that upscaling and downscaling are very distinct processes which are only relevant to changes in grid size. However, in hydrology these terms are used more generally across the literature as indicating a scaling of any of the parameters from Table 1.5, consequently the current work will use the terms more generally to remain consistent with hydrological work.

<table>
<thead>
<tr>
<th>Scale Term</th>
<th>Definition</th>
<th>Hydrological Modelling Term</th>
</tr>
</thead>
<tbody>
<tr>
<td>Extent</td>
<td>Area or period over which observations are made.</td>
<td>Study area</td>
</tr>
<tr>
<td>Coverage</td>
<td>Ratio of sum of areas for support unit to the extent.</td>
<td>Resolution</td>
</tr>
<tr>
<td>Support</td>
<td>Largest area or interval for which the property of interest is considered homogenous.</td>
<td>Grid size</td>
</tr>
<tr>
<td>Support unit</td>
<td>Units of observation.</td>
<td>Point</td>
</tr>
</tbody>
</table>

Table 1.5 – Parameters which may be scaled, adapted from Bierkens et al. (2000).
Figure 1.4 – Scaling different components of data (from Bierkens et al., 2000).

1.3.2.3 – Upscaling

Upscaling is the process of transferring information from a given scale to a larger scale (Blöschl and Sivapalan, 1995), and in effect it is the process of extrapolation to a coarser scale (Schulze, 2000). Bugmann (1997) highlighted that upscaling itself can be grouped according to methodology, distinguishing between implicit upscaling which accounts for scale-dependent features and so requires methods to be more exact, and explicit upscaling which assumes a certain level of representation and may rely on numerical simulations to scale up the response to change.
Schulze (2000) and Harvey (1997) summarise the various approaches available to upscaling, shown in Table 1.6. They are listed in order of increasing complexity and are accompanied by key problems and uncertainties. Only the second, third and fourth options involve any form of tangible upscaling, as the other options facilitate working around the problem.

<table>
<thead>
<tr>
<th>Method of Dealing with Upscaling Need</th>
<th>Associated Problems</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ignore problem</td>
<td>No solution; error of commission (Haufler et al., 1997).</td>
</tr>
<tr>
<td>Lumped model with effective parameters</td>
<td>Parameters may change with conditions, calibration invalid for periods other than calibration data coverage.</td>
</tr>
<tr>
<td>Distributed model with unit representation</td>
<td>Interactions between units/pixels unknown.</td>
</tr>
<tr>
<td>Parameterise unit interactions</td>
<td>Edge effects between units.</td>
</tr>
<tr>
<td>Create new model</td>
<td>Workload increased; avoiding problem.</td>
</tr>
<tr>
<td>Model at fine enough resolution</td>
<td>Not a solution: different processes will dominate at different scales and so pseudo-process representation prevails.</td>
</tr>
</tbody>
</table>

Table 1.6 – Upscaling methods, adapted from Harvey (1997) and Schulze (2000)

Bierkens et al. (2000) approach the problem of upscaling through a systematic procedure using questions about the problem, as shown in Figure 1.5. Some form of averaging is typically required, either of input or output variables, and where this is the case, details are given on the methodologies available (shown in blue). This leads the user to the most appropriate method of upscaling.
Some analogies can be seen between Figure 1.5 and Table 1.6, yet the two are distinguished by Bierkens et al.'s (2000) work being focussed on upscaling during modelling, and so having a more applied utility. It is also more complicated, involving several phases and so several opportunities for upscaling which may not be mutually exclusive during a given model construction.

Issues such as these and the general complexity and uncertainty of upscaling led Schulze (2000, p196) to conclude that "The problem of upscaling thus remains a largely unsolved one." However, appropriate upscaling techniques may be applied to different hydrological modelling situations, and the mere fact that a modeller is aware of upscaling issues will lead to any results from model use being taken within the appropriate context and with a realistic appreciation for model uncertainty.

1.3.2.4 – Downscaling
A disaggregation of information from coarse to a finer scale is known as downscaling (Schulze, 2000). The process attempts to reconstruct the variation of a property (Bierkens et al., 2000) within known point values. Hostetler (1999) summarised downscaling methods in hierarchical order according to increasing complexity and cost. This work dealt with climate data, but scaling issues are generally transferable across disciplines; Table 1.7 shows this work and some of the problems involved.
<table>
<thead>
<tr>
<th>Method of Downscaling</th>
<th>Associated Problems</th>
</tr>
</thead>
<tbody>
<tr>
<td>Use nearest point values</td>
<td>No effective downscaling. Error to assume large scale parameters adequate for interpretation of impacts at finer scales.</td>
</tr>
<tr>
<td>Append nearest point value to present day values (contemporary auxiliary)</td>
<td>Intercorrelation between datasets involved.</td>
</tr>
<tr>
<td>Interpolation from point values</td>
<td>Complex for complicated systems; unwanted smoothing of local variation.</td>
</tr>
<tr>
<td>Statistical methods</td>
<td>Depends on statistics chosen.</td>
</tr>
<tr>
<td>Model specific downscaling techniques</td>
<td>Depends on chosen techniques.</td>
</tr>
</tbody>
</table>

Table 1.7 – Downscaling methodology, adapted from Hostetler (1999)

Bierkens *et al.* (2000) conceptualised the task of downscaling within the modelling environment as a decision making process. As with upscaling, the methods proposed are organised according to the properties of the data and the modelling methodology: Figure 1.6 shows this procedure for downscaling from scale two (S2) to scale one (S1).
Here a mechanistic model refers to one which describes the variation of the parameter, and this of course requires further information. Empirical functions would require statistical analysis and fine scale auxiliary information, necessitating data collection or retrieval from a database (Bierkens et al., 2000).

It is clear that some of the methods utilised under Hostetler’s (1999) framework fall within the conceptual process of Bierkens et al. (2000). However, as the latter deals solely within the environs of modelling, the two are not open to synthesis, or much of the former will be lost, as some of these techniques will be inappropriate for modelling applications.
1.3.2.5 – Linkages Between Temporal and Spatial Scale

Blöschl and Sivapalan (1995) describe the concept of characteristic velocity, introduced by Haltiner and Williams (1980). This is a pattern shown in almost any physical process, and proposes that there is a generally constant ratio between characteristic length and time scales as these scales change. As length scales increase, time scales increase correspondingly (Figure 1.7). A process occurring on a small spatial scale will operate on a shorter time scale. This has a consequential effect on the variability, and this leads to an important link between variability and spatial scale: generally the smaller a process is spatially, the more variable it will be in time and the more spatially

![Figure 1.7 - Commensurate scales (adapted from Wilby and Schimel, 1999)](image-url)
heterogeneous it will be. An example of this is a convectional storm, as discussed in Schulze (2000), whereby within such a local phenomenon, rapid changes in rainfall intensity occur. In contrast to this would be larger scale frontal rainfall, with more homogenous rainfall intensity.

On reflection this seems instinctive: it takes a shorter time for a smaller change to take place. Blöschl and Sivapalan (1995) summarise that the effect of this is that the overall term ‘scale’ can represent both temporal and spatial scales. This relationship leads to a higher predictive power when spatial and temporal scales are analogous. Wiens (1989) relayed the relationship between scales and predictability through Figure 1.8.

![Figure 1.8 - Predictive power from different spatial and temporal scales (from Wiens, 1989)](image)

Increased temporal resolution without increased spatial resolution results in merely an apparent increase in predictive power. Of course more information is
better for a model, but any increase in predictive power will be limited, beyond which it will be merely superficial. The correct information (and scale) is of a higher priority, and hence it is important that as scaling is carried out either spatially or temporally, consideration must be given to the other, or the effects of the deficiency assessed if the other is not scaled.

It is clear from Figure 1.8 that the most utility for prediction comes when spatial and temporal scales are similar. This complements and vindicates the concept of characteristic velocity.

1.3.2.6 – Summary

Ultimately it is more appropriate to have the scales at which you are investigating and those at which your findings are to be applied to be commensurate, and to measure model inputs at this scale also. Investigating at too coarse a scale misses vital patterns and processes and so explanations for the observations. Investigations at too fine a scale are interrupted by patterns emerging from larger scale processes and perturbations at finer scales (Cammeraat, 2002).

In addition, it has been highlighted that as models incorporate data from a range of sources, an assessment should be undertaken as to the relationship between the datasets and any effect of contradicting scales appreciated. Any mismatch between scales of inputs may manifest to produce increased uncertainty in results, which may be acceptable for certain applications and unacceptable for others. The practicalities of model construction and development are often dictated by data availability and quality. The model developer is required to be
aware of such issues and include this in model application and assessment of uncertainties.

1.3.3 – Methods of Conceptualising Scale and Scaling

To assist with the comprehension of scale issues, several concepts have developed within the research community. Fundamentally, Bloschl and Sivapalan (1995) highlighted the emergence with empiricism of the difference between the process scale and the observation scale. From this, the development of distinct process, observation and modelling scales have become integral to conceptualising scale.

1.3.3.1 – Process Scale

All natural phenomena exhibit a dominant scale of operation, which is known as the process scale (Bloschl and Sivapalan, 1995). The term natural scales refers to length or time scales that are more likely to occur than others; the remainder, where processes are less likely to occur is termed the spectral gap (Bloschl and Sivapalan, 1995). Many hydrological processes have a natural scale due to inherent scales in the drivers of the processes, and so the process and natural scales are inherently linked. An example of this in the physical world is annual and diurnal cycles of the solar regime dominating the temporal dimension of hydrology, as shown commonly through the establishment of diurnal cycles and longer term seasonal trends in localised water table, as discussed in Section 1.2.5. This leads to dominant natural process scales, in this case the seasonal,
larger scale (~2 m) trends in water level and daily, smaller scale fluctuations (~0.1 m). The concept of characteristic velocity is thus also incorporated, with larger spatial variations associated with the longer periodicity and vice versa.

1.3.3.2 – Observation Scale

In contrast to these natural regimes is the scale at which the same processes are observed (Blöschl and Sivapalan, 1995; Schulze, 2000). Several fundamental limitations to observation including cost, data collection logistics, the technology of measurement and associated accuracies, or mere ignorance and/or assumption-making may cause the observation scale to be different from the process scale. The number of samples able to be taken is finite in any research project however well it may be funded (Blöschl and Sivapalan, 1995). All data collection inherently involves sampling and filtering (Cushman, 1984), whereby selected units represent surrounding areas, and so information is missed. Ideally, science should observe processes at the scale at which they work (Blöschl and Sivapalan, 1995), but this is seldom possible as the process and observation and scales therefore rarely coincide. It is difficult to measure the bias introduced by the scale of observation quantitatively, although Western and Blöschl (1999) attempted this with some success with soil moisture data through the use of geostatistical frameworks.

Advances in recent decades of remote sensing technologies have added new depth to the observation scale, forcing scale issues to light. Data are now available across the globe at an increasing resolution and accuracy for both direct hydrological measurements such as soil moisture, but also analogous
datasets such as topography, from which secondary data such as slope and wetness can be derived. This higher quality data only increases the appreciation of heterogeneity amongst variables directly related to the hydrological sciences (Wood, 1995).

1.3.3.3 - Interaction Between Process and Observation Scales

Cushman (1984) discussed the interface between these two key scales. Their relationship is important when conceptualising the impact of data collection on the outcome of research, as Figure 1.9 summarises. It can be seen that processes larger than the observation scale appear as trends in the data; processes on a scale smaller than the observation resolution appear as noise. Only when process and observation scales are commensurate does the observation adequately reflect the processes occurring (Cushman, 1984).

![Figure 1.9 – Comparing observation and process scales](from Blöschl and Sivapalan, 1995)
We have seen above that threshold values may indeed be a cause of scale issues. Wood (1995, p103) suggests that thresholds may also be of use in solving scale issues and in particular the interaction between the process and observation scales:

"An important research objective in scaling hydrological variables is the determination of the threshold scale where statistical representations of smaller (sub-grid) areas can replace actual patterns of variability."

Wood (1995) goes on to suggest that the "threshold scale" is likely to be in the region of 5-10 km² for many hydrological parameters.

1.3.3.4 – Modelling Scale

A model is a representation of reality (Kirkby et al., 1993), and predicts the behaviour of a physical system based on laws integrated into the model code, and based on known physical laws: another scale important to the current work is the modelling scale. Models are commonly utilized and well established in the hydrological sciences to improve understanding and extrapolate responses outside of observed conditions, and the modelling scale is the scale at which the process or system is represented (Blöschl and Sivapalan, 1995). The generally accepted modelling scales are shown below, in Tables 1.8 (space) and 1.9 (time), adapted from Blöschl and Sivapalan (1995) and Beven (1989).
It should be noted that the term 'model' can be used to describe a dataset which describes a parameter across a region, such as a digital elevation model (DEM). More commonly the term is used to describe a complete descriptive tool, with inputs, calculations and outputs, used to predict how a system will respond to changing conditions.

<table>
<thead>
<tr>
<th>Scale</th>
<th>Approximate Spatial Extents</th>
</tr>
</thead>
<tbody>
<tr>
<td>Point / micro</td>
<td>1-10m</td>
</tr>
<tr>
<td>Plot / field</td>
<td>10-100m</td>
</tr>
<tr>
<td>Reach / hillslope / meso</td>
<td>100m – 1 km</td>
</tr>
<tr>
<td>Catchment</td>
<td>1-100km</td>
</tr>
<tr>
<td>Basin</td>
<td>100-1000km</td>
</tr>
<tr>
<td>Continental or global / mega</td>
<td>&gt;1000km</td>
</tr>
</tbody>
</table>

Table 1.8 – Spatial modelling scales

<table>
<thead>
<tr>
<th>Scale</th>
<th>Approximate Temporal Extents</th>
</tr>
</thead>
<tbody>
<tr>
<td>Event/short term</td>
<td>&lt; 1 second - 1 week.</td>
</tr>
<tr>
<td>Monthly</td>
<td>&gt; 1 week – several months</td>
</tr>
<tr>
<td>Seasonal</td>
<td>&gt; 1 month - 1 year</td>
</tr>
<tr>
<td>Medium term</td>
<td>1 - 10 years</td>
</tr>
<tr>
<td>Long term</td>
<td>10 - 100+ years</td>
</tr>
</tbody>
</table>

Table 1.9 – Temporal modelling scales
Blöschl and Sivapalan (1995) summarised scaling within a modelling perspective simplistically. With modelling including variables, parameters and inputs in its conceptualisation of reality, complications due to scale are inherent. It became common in the early years of modelling in hydrology to scale only one of the above components and consequently assume that model outputs will be strong across scales.

Within a modelling context, Band and Moore (1995) defined scaling as the extension of small scale process models, which may be directly parameterized and validated, to larger spatial extents. Indeed, many scaling issues arise from a need to develop the understanding from ‘point’ scales, using existing experiments and models. Complexity arises as this scaling is undertaken and estimations of the model parameters and process computations are expanded over the heterogeneous land surface (Band and Moore, 1995), which is often known less well than the study area.

Modelling complexity has been appreciated recently by the distinguished modeller Jens Christian Refsgaard (2007a). In a frank admission of overconfidence, Refsgaard (2007a) assessed the reality of a comment in previous work, whereby Refsgaard and Storm (1995, p810) had proclaimed that modelling would improve considerably in coming years:

“MIKE-SHE is applicable on spatial scales ranging from a single soil profile to larger regions”.

60
Although appreciating at the time that a number of fundamental scale problems would need to be “carefully considered” to achieve such a holistic modelling environment, Refsgaard (2007a, p61) later admitted:

“I do not believe any longer that a universally applicable code and modelling methodology is theoretically realistic, and certainly not feasible in practice. The main reason for this is the scaling problems. Because scaling is interlinked with modelling concepts, I therefore do not believe it will ever be feasible to derive a universal scaling theory of practical applicability.”

Such a sea change in attitude reveals the growth in appreciation of scale issues over the past decade. As such, the incorporation of scale and its effects on modelling and applications of modelling are becoming established in contemporary modelling practice (Quinn et al., 2004).

Important to many modelling-scale issues is the different groups of models available. A lumped model takes no account for the spatial variability of physical properties and processes, and effectively this is an averaging of parameters over a certain area. Conversely, a distributed model has parameter values which change across the modelled domain and so have a varied contribution to model output. A physically based model has a grounding in well-founded laws and relates inputs and outputs directly to these laws, whereas a conceptual model may just be based on empirical relationships between inputs and outputs.

Band and Moore (1995) described the development of hydrological models since the 1970s as initially taking two distinct trajectories. Firstly, the optimism of
rapidly increasing computing power enthused many hydrologists to increase the physical complexity of models and incorporate realism with the determinism of the time. Simultaneously, models were developed which spatially distributed the representation of key hydrological processes. From the 1990s the development of spatial datasets of high enough resolution that could describe the heterogeneity of the grid surface developed which, together with the advent of remote sensing techniques, resolved this disparity somewhat. The latter group advocated the use of 'effective' parameters, which were unique to the calibrated conditions and not transferable neither spatially nor temporally. One such modeller was Beven (1989), who argued that the limitations in both understanding of physical processes and representation of parameters from poor spatial data coverages determined that the lumped methodology was the best available. Today this is still the case, and there has been a more recent re-acceptance of 'lumped' models, with partially 'effective' parameters, such as the use of the Probability Distributed Model (PDM) for rainfall-runoff modelling (Moore, 2007).

After model construction, data from monitoring can be used to calibrate a model, which involves changing model parameters in order to make the model predict accurately for the specific site or region. Validation uses independent data to verify that this process has been successful, before the model can be used for simulation which is in effect extrapolation to use at times with no observed output with which to compare model outputs. This logical framework for model development (Figure 1.10) is outlined by Refsgaard (2007a), but in general terms is widely accepted through the scientific community. Although this methodology
conceptual model are arbitrary and open to user interpretation. Field data collected is open to issues discussed Section 1.3.3.2. Model development again relies on field data and so is again vulnerable to scale issues from observations.

Model conceptualisation is fundamental to subsequent model construction (Acreman 2004), and therefore it must be inferred also to model results. Thus it is essential to research the system to be modelled as thoroughly as possible and plan how the system is to be modelled logically. As Acreman (2004) concludes, information on the behaviour of the system (provided by a model) would enable improved understanding and so conceptualisation; therefore an iterative process would be ideal.

Limitations include uncertainty of extrapolation outside the calibration periods, as conditions here may be very different and so results may in reality be different from those simulated. Beven (1989) critiqued the over-reliance on modelling, something which has since become more commonplace. He highlighted issues such as the calibration and validation elements of the procedure are dependent on observed results, and so subject to the same error as those data. Also, there is a risk of over-parameterization, leading to a loss of meaning of the calibration parameters used as inputs to the model; this is more so with physically based models which regard parameters as true values. Related to this is interdependence between calibrated parameters, even when there are few parameters being calibrated.
The work of Bromley et al. (2004) is important to wetland hydrology because it was the first applied investigation which appreciated scaling issues and incorporated them into wetland hydrology. Unfortunately, no work since has developed the lead and undertaken similar investigations into either other parameters, or $K$ at other sites.

Much general hydrological work has been undertaken to assess the impact of scale issues on hydrology (as shown in Section 1.3.3.4), but little other analysis of wetlands has been carried out with regard to scale issues. An example of the impact of a lack of appreciation of scale issues is Maltby et al.’s (2005, p149) conclusion that:

"Functional analysis of wetlands has tended to focus on the impacts of individual large wetlands. The cumulative impact of many small wetlands within a river basin requires further research."

This, it is proposed, is a basic weakness in contemporary wetland hydrological science, given the fundamental importance of scale effects and their ubiquitous nature, and the increasing consideration for scale shown in the wider hydrological sciences, including modelling. Wetlands have been consistently shown to affect the hydrograph of rivers, sometimes dramatically, and in a variety of fashions (e.g. Bullock and Acreman, 2003). If this is the case with the current coverage of remnant wetlands, the composite effect of smaller yet numerous wetlands across the landscape must plausibly be as, if not more important. Although wetland science is currently regarded as a multidisciplinary
work, one more discipline must be mastered before it can be regarded as comprehensive: that of scale.

A common theme running through scale issues is the linkage to heterogeneity. Many parameters important to hydrology exhibit heterogeneity even at very small scales, and so the scale of observation appears to be highly dependent upon the application. The debate on spatial heterogeneity and its representation at various scales for the understanding of hydrologic processes has been on-going for many years now and will no doubt continue to do so. Becker et al. (1999) summarise the two key methodological questions of the field to be (a) optimum representation and (b) the scaling of point measurements to catchment models. I would argue from the literature above that the issue of conceptualisation comes first and foremost, requiring increased priority amongst everyday hydrological modelling, as a focus on scale issues even during conceptualisation may prevent many problems of scale being incorporated into research projects and hydrological modelling.

In wetland hydrology this is formidable task; Bergström and Graham (1998, p255) concluded that "one can easily get paralysed by the magnitude of the scale problem." Wetlands vary tremendously across the temporal and spatial scales, possibly more so than any other landscape due to their transitional nature. Augment this with the fact that we discern at the observation scale to explore the process scale, and manage at the policy scale at best. These processes are investigated at the modelling scale whilst we stretch the limits of the data collected through extrapolation (amongst other methods). Standard
procedure dictates that these outputs are subject to coupling with other data which have undergone similar yet perhaps more extreme extrapolation, such as climate change scenarios.

In conclusion, wetland hydrology as a field of expertise must accept scale issues as an integral factor in their work, just as climate change has become accepted in recent years as central to forward thinking wetland science.
1.5 – Conclusions

A review of knowledge in the fields of both wetland hydrology and scale issues has been enlightening: there are large gaps in the knowledge of both areas, and a void of knowledge associating the two. Wetland hydrological functioning is an emerging method of wetland hydrological assessment which characterises hydrological impact of a site through the analysis of significant water transfer mechanisms. Wetland hydrological functioning is site-specific, and so knowledge is not transferable from one site to another. This is a scale issue itself, as there is heterogeneity across a region, crucially dependent on the scale of observation. Such an analysis may be undertaken at any wetland site for determination of the likely impact of the wetland on local and more regional water resources, and is being increasingly used for such purposes in the UK and further afield.

When models are used to assess wetland hydrological functioning, scale becomes more entrenched as an issue, as datasets and model codes can have inherent working scales themselves. This is typified by the differences between lumped and distributed models, and the continuum of complexity between the two.
The understanding of the impact of scale issues on the hydrological sciences in general has been improving, and there have been numerous and wide ranging assessments of the impact of different scales of data on modelling, in particular for rainfall-runoff modelling. There has been a general agreement across the field that there is indeed an impact, and that it is important that scale issues are examined from the outset of a project in order that data collected are of a relevant scale for any model that may be constructed.

However, the integration of scale issues in wetland hydrological science has been very poor, bar very few individual examples including papers from Bromley et al. (2004) and Liu and Cameron (2001). As such, it is important that scale issues are advanced further up the agenda across the wetland science board.
1.6 – Hypotheses and Research Structure

The review of the literature has determined issues outlined in the Conclusions as particularly important to contemporary wetland hydrological science. As such, the following hypotheses are proposed for further examination:

1. Scale issues will be evident at each scale of observation of wetland hydrology;
2. The field scale site-specific nature of a wetland's hydrology is crucial to any hydrological functioning at the catchment scale;
3. Scale issues will be important for and so impact upon the assessment, through monitoring and modelling, of wetland hydrological functioning; the successful comprehensive assessment must therefore be scale dependent.

In order to examine these hypotheses, it is the plan of this research to assess the hydrological function of a wetland site, and in light of the apparent gaps in the literature, it is clear that the methodology used will need to incorporate scale issues in this assessment. As information is lost as scaling takes place, and the process of scaling seems to be incredibly reliant on an existing expert knowledge at both scales, a solution seems to be to assess the hydrology of a wetland site at two scales, thus removing any requirement for scaling of results. The plot and
catchment scales lend themselves to concurrent analysis as they allow independence in observation from each other, whilst not being at either ends of the scale spectrum and so close enough to assess the impacts on one another.

Firstly, a plot scale (10-100 m) analysis will determine the hydrological functioning of a wetland site at Otmoor, Oxfordshire, UK. Monitoring will take place at a high temporal resolution in order to match the small spatial scale as dictated by the likely characteristic velocity. This plot scale analysis will aim to determine the processes dominating the hydrology, and so which are likely to be important at larger scales, and will form Chapter 2 of the current work, as shown in Figure 1.15.

Information from the plot scale work will be used as input to an analysis of the wetland's behaviour in a catchment context (that of the River Cherwell), and the results will allow a comparison of the important processes at each scale. To place in context the hydrological functioning of the wetland at the catchment scale, a simple water storage model and a complex hydraulic model of the catchment will be constructed, the development and use of which will be described in Chapter 3. For the simple storage volume model, two Otmoor will be directly compared to a very different peat wetland, which is foreseen to have very different storage components and so likely to have a different impact on the catchment. Unnecessary data collection and analysis will be avoided and focus will be given to dominant water transfer mechanisms and stores, and so this analysis will complement the first assessment at the plot scale. The two scales
of wetland hydrological functioning assessment are shown below (Figure 1.15) in an adaption of the earlier Figure 1.7 describing characteristic velocity.

Having gained an insight into the hydrological functioning of the wetland at both the plot and catchment scales, conclusions can be drawn as to any relationships or interactions between the two. Furthermore, an assessment of the work with regard to scale issues in light of the newly highlighted deficiencies of the majority of contemporary wetland hydrological work; conclusions will form Chapter 4.

It is anticipated that roughness will be an important concept across the spatial scales for wetland hydrological functioning, especially so as its definition can be
somewhat open to interpretation. Roughness is primarily used to describe the resistance to flow in a river or across a floodplain, and in hydraulic models of river systems it is estimated numerically and commonly forms a calibration parameter. If the bed of a river has a high roughness, water flowing over it will be attenuated more and the system will take more time to remove water from the catchment. Roughness can be estimated at the point scale, with a detailed assessment of a sample of sediment and vegetation (e.g. Ackers, 1991; Sear et al., 2002), but is often applied at the reach or catchment scale in modelling environments with an estimated initial value and used as a calibration parameter (Pappenberger et al., 2005). On a slightly larger scale, roughness may be considered as the spatial change in bed and floodplain topography, and so interpreted as the continuum of pools and riffles which combine to incorporate varying bed conditions. This analysis would fall between the plot and reach scales. Cowan (1956) incorporated several scales of elements into his assessments of river channel roughness. These elements ranged from incorporating a factor for the size of the channel cross section, one for obstructions, one for vegetation and other flow condition regulators and finally for the meandering of the river channel. As such the roughness accounted for by Cowan (1956) covered the very small point scale factors up to the reach scale of meanders.

Extending this process, roughness could conceptually be extended to incorporate the long section of a river from source to mouth, which is directly related to catchment scale topography. By loosening the accepted definition of roughness, other concepts may be included under its umbrella. For example,
the level of surface water storage in a wetland may be determined by roughness which may incorporate the depth and frequency of surface water bodies. This concept will be investigated in the course of wetland hydrological assessments at both the plot and catchment scales.

The wide ranging representation of roughness over many scales is represented in Figure 1.16, which gives spatial and temporal scales for the development and impact of each. Bed roughness (1) affects river flow from the point scale up to a few metres spatially, and temporally from seconds up to possibly attenuation over an event hydrograph. Surface water storage at a wetland site (2) may typically be from 10 to 100 m in size, and store water from a rainfall event (over an hour) for up to several months. Pool-riffle sequences on the bed of a river can be as low as about 10 m apart and occur in a stretch of river up to the reach scale; they can affect the flow over several hours and take months to develop. Meanders (4) are known to affect larger stretches of river from the reach to most of a catchment, and may affect the flow hydrograph at event scale but may take decades to develop. The long section of a river (5) affects the entire catchment and river basin, and takes many centuries to develop but may affect the flow of the river almost down to the event scale.
Figure 1.16 – Representation of roughness at different scales (adapted from Wilby and Schimel, 1999). 1: river channel bed roughness; 2: surface water storage; 3: river pool-riffle sequences; 4: meander structure of river; 5: catchment scale river long section.

The current study will determine which of items 1-5 from Figure 1.16 are most important to determining wetland hydrological function at a wetland site.
Chapter 2

Plot Scale Assessment of

Wetland Hydrology
Chapter 1 identified a gap in the knowledge of combining increases in understanding of wetland hydrology at different scales. This chapter will describe the development of plot scale wetland hydrological understanding at a wetland field site, Otmoor. Chapter 3 will assess the hydrology of the site at the catchment scale, using information obtained in the current chapter.

Section 2.2 will introduce the field site and describe the methodology of the hydrological monitoring and analysis undertaken. Results will be provided in Section 2.3, and conclusions on the hydrological functioning of the wetland site drawn in Section 2.4, along with implications for the subsequent catchment scale work.

The hydrological research was undertaken at a site established for a project funded by the UK Department for Food, Environment and Rural Affairs (DEFRA), and undertaken primarily by the Centre for Ecology and Hydrology (CEH). That research aimed to investigate the effect of shallow surface foot drains, known as grips, on the hydrology, its effect on invertebrate distribution (as prey for birds) and subsequent bird utilisation of foot drains for feeding and nesting. Vegetation was monitored, and soil analysis was undertaken for development of hydrological understanding. The ‘Grips’ project has three sites in a transect
across southern England, in Norfolk, Oxfordshire and Somerset, on different soil types.

Hydrological monitoring of the sites entailed installation of a dense network of hydrological monitoring stations, sampled monthly and the siting of an automatic weather station (AWS) and stageboards to assess surface water levels. The current work has developed this research at one of the three sites, Otmoor, Oxfordshire, through increasing the number of hydrological parameters measured and the temporal resolution of sampling, enabling the determination of hydrological functioning at the plot scale, and so assessment of research hypotheses.

The Grips project and the current work thus complement each other. The Grips project has provided broad background understanding with several years of hydrological, botanical, and invertebrate sampling, along with laboratory analysis of soil properties (Acreman et al., 2008). The current work has furthered the hydrological investigations at Otmoor significantly, providing a weekly sample run for a limited number of hydrological variables and adding to the on-site instrumentation. Together the financial and field sampling resources have enabled Otmoor to be one of the most hydrologically instrumented and sampled wetland research sites in the UK.
2.2 – Methodology

2.2.1 – The Study Site

2.2.1.1 - Location

Otmoor is a wetland site located 5 km northeast of Oxford, and approximately 50 km northwest of London, UK (Figure 2.1). Otmoor lies in a basin (4 km across), with higher ground surrounding a 1,600 ha expanse of semi-natural land in a clay basin surrounded by the "seven towns of Otmoor" (Rackham, 2000; Bloxham, 2005): Beckley, Noke, Oddington, Charlton-on-Otmoor, Fencott, Murcott, and Horton-cum-Studley, and with Islip to the west (Figure 2.2). Figure 2.3 shows the topographic setting of Otmoor.

![Figure 2.1 - Location of Otmoor within England](image)
Figure 2.2 – The Otmoor area; © Crown copyright
Ordnance Survey 2007

Figure 2.3 – DEM of the Otmoor area; NEXTMap Britain™ elevation data from Intermap Technologies
2.2.1.2 – History

The Otmoor area has a varied and interesting history, first recorded in Roman times (AD 43-400) due to the (limited) drainage and construction of a Roman Road running north-south across the site, thought to be part of a wider transport link between Dorchester Abbey, 30 km to the south, and Alcester, 80 km to the north (Ponting, 1991; Rackham, 2000; Otmoor, 2006). Between 1006 and 1011 Otmoor appeared in a charter allocating land by King Æthelred II. King Edward the Confessor was born in nearby Islip in 1005 (Otmoor, 2006).

There is little record of Otmoor from the middle ages to the 19th Century, but folklore and other evidence tells an interesting story of the importance of Otmoor, including the 1955 weather vane placed on Charlton church depicting locals with webbed feet to help them through the marshes, as myth suggested. Otmoor was known to be predominantly common land managed as water meadows to support the rearing of geese by locals. The locals claimed that 'Our Lady of Otmoor' had ridden a circuit round the moor while an oatsheaf was burning, and given the area inside it to the people of Otmoor in perpetuity (Bloxham, 2005).

The Otmoor riots were sparked when local landowner Sir Alexander Croke tried to evict the people and fenced off the land in 1829, resulting in public uprising, as described by Bloxham (2005). Early attempts to drain the land failed, and actually caused widespread flooding of Otmoor. Subsequent work was continually disrupted by local men during night time raids, destroying hedges, fences and embankments on the River Ray. Their anguish inspired the following rhyme (Oxford Times, 2006):
The fault is great in Man or Woman,
Who steals the Goose from off a Common;
But who can plead that man's excuse,
Who steals the Common from the Goose.

The goose mentioned refers to the Aylesbury ducks and geese bred on the common land before the exclosures (Oxford Times, 2006). The disturbances peaked on September 6th, 1830, when about 1,000 people walked the seven mile circumference of Otmoor destroying fences, only to be read the Riot Act. After refusing to disperse, 66 men were arrested and put onto wagons destined for Oxford gaol. A large mob attacked the escort on the outskirts of the City of Oxford, allowing the prisoners to escape. The situation slowly calmed and people were allowed to bring their animals back to the meadows. Some work was undertaken to drain the land for agriculture, along with the canalisation and redirection of the River Ray through Otmoor (Armstrong et al., 2000), creating the New Ray (Figure 2.4).

By the 1970s intensive agriculture had drained 25% of Otmoor (Otmoor, 2006; RSPB, 2006) through the installation of electrical pumping stations and dredging of drainage channels through the site. A more recent threat emerged in the early 1980s with the development of the M40 motorway linking the cities of London and Birmingham, and only a public enquiry and an active campaign from local residents (see Alice's Meadow, 2006) saved Otmoor's bisection.
There is a Ministry of Defence (MOD) rifle range on the site, and the Royal Air Force used the area as a practice bombing range from 1932 through World War II (Otmoor, 2006). Much of the MOD land has been left largely undisturbed and has consequently become a haven for floral species and pond life. This area has subsequently acquired Site of Special Scientific Interest (SSSI) designation for its rich biodiversity of grassland wildflowers, and is managed as a hay meadow with seasonal grass cutting.
At present the RSPB owns 267 ha of Otmoor (Figure 2.5), and after acquiring the land in 1997, management has existed within the Upper Thames Tributaries ESA (DEFRA, 2006). This has included wetland restoration, including 180 ha of wet grassland (Figure 2.6) on which the research plot lies, achieved through increasing water levels, the creation of surface foot drains and the introduction of extensive grazing. There has been success in the management objectives of attracting waterfowl (e.g. 3,600 teals, Anas crecca, and 3,900 wigeons, Anas penelope in 2002) and breeding waders (e.g. 60 pairs of lapwing, Vanellus vanellus, and 20 pairs of redshank, Tringa totanus in 2002; RSPB, 2006). A 22 ha reedbed has been engineered including 5 km of new ditches, and 150,000 young reeds were planted in a new reedbed location in 2006, together successfully attracting 20,000 starlings (Sturnus vulgaris) to roost in the same year (RSPB, 2006). Work has also begun on the newly acquired land to restore even more extensive areas of wet grassland on the southern edge of the site: see Figure 2.6.

Figure 2.5 – Otmoor from Noke Hill
Otmoor is one of the most extensive restoration projects in the UK, and managed proactively as prime habitat for endangered bird species, but in association with the Environment Agency to enable flood storage during very wet periods. The Otmoor basin is known to flood dramatically periodically, as shown starkly in Figure 2.7, an aerial photograph taken during the 2000 floods. Lying on a major
tributary to the River Cherwell, the Otmoor area has been cited as having the potential to protect urban areas in Oxford from river surges (Pearce, 2007).

Figure 2.7 – Otmoor flooded during winter 2000 from northwest; thumbnail shows paths of Old and New Rays as shown in Figure 2.4
Haslam (2003) defines a wet grassland simply as a grassland with soil flooded or at least waterlogged for enough of the year to influence species composition. Wet grasslands are a specific category of wetland environment as defined by Gavin (2001, p35):

"Wet grasslands encompass semi-natural floodplain grassland, washland, water meadows, lakeside wet grasslands, and wet grasslands with intensive water level management on drained soils."

Incorporating the traits of wider wetlands as defined in Section 1.1.1 such as the water table being at or near the surface, unique soil conditions and hydrophytic vegetation, wet grasslands have the added characterisation of intense management. Historically this would have been through grazing by geese, sheep or cattle, or through the manual cutting of grasses for bedding or storing as animal feed (Gowing, 2006). Importantly, a lack of management would enable vegetational succession to occur, allowing woody plants and shrubs, followed eventually by trees and so the loss of the environment. This dynamic nature of wet grasslands typifies the transitional nature of wetlands in general, as described in Section 1.1.5.

2.2.1.3 - Hydrological Setting

Otmoor lies on the floodplain of the River Ray (Figure 2.8), which rises in Grendon Underwood, Buckinghamshire (Armstrong et al., 2000), and joins the River Cherwell at Islip, Oxfordshire. The Cherwell confluences with the River Thames in turn, 10 km upstream from Oxford; key descriptors of the three rivers are shown in Table 2.1 (Marsh and Lees, 2003).
Figure 2.8 – Hydrological drainage network from Otmoor

<table>
<thead>
<tr>
<th>River</th>
<th>Gauging station</th>
<th>Catchment area</th>
<th>Mean flow</th>
<th>Mean annual rainfall</th>
<th>Mean annual runoff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ray</td>
<td>Islip</td>
<td>290</td>
<td>1.83</td>
<td>674</td>
<td>198</td>
</tr>
<tr>
<td>Cherwell</td>
<td>Oxford</td>
<td>907</td>
<td>5.18</td>
<td>666</td>
<td>180</td>
</tr>
<tr>
<td>Thames</td>
<td>Farmoor</td>
<td>1,607</td>
<td>15.63</td>
<td>804</td>
<td>306</td>
</tr>
<tr>
<td>Thames</td>
<td>Kingston upon</td>
<td>9,948</td>
<td>77.87</td>
<td>719</td>
<td>247</td>
</tr>
</tbody>
</table>

*Kingston upon Thames not shown on Map: 157 km downstream of the Cherwell-Thames confluence

Table 2.1 - Key parameters of the Rivers Ray, Cherwell and Thames (from Marsh and Lees, 2003)
Mean annual runoff is the notional depth of water in millimetres over the catchment equivalent to the mean annual flow as measured at the gauging station (Marsh and Lees, 2003), and is the consequence of many factors including abstraction for public or industrial supply, groundwater interaction, reservoir storage, or other regulation. The comparison of rainfall to runoff therefore indicates any management, regulation and water use in the upstream catchment.

2.2.1.4 – Geological Context

Below Otmoor Kellaway limestone underlies a deposit of mid-Jurassic mudstone (Oxford Clay), the boundary of which lies at between 37 and 50 m below the surface of Otmoor. This impermeable clay supports a thin (2 to 3 m) layer of alluvially reworked clay at the surface (Mould, 2007; Hughes, pers. comm.), as shown in the profile in Figure 2.9.

![Geology of the Otmoor area](image)

**Figure 2.9 – Geology of the Otmoor area**
2.2.2 – Hydrological Monitoring

2.2.2.1 - Introduction

In order to comprehensively investigate the hydrology of the research plot at Otmoor, a hydrological monitoring programme was established to measure all relevant water transfer mechanisms. The objective was to elucidate the hydrological functioning of the instrumented plot, the concept of which was described in Section 1.1.6.

Figure 2.10 shows the location, Figure 2.11 the field site area and Figure 2.12 the research plot layout, with grips installed in 2004. A wider network of ditches across the site, allowing intensive management of water levels, drain through a pipe to the ring ditch shown to the south of the site in Figure 2.11. Water levels in this ditch are managed by the RSPB, who may pump water onto or off of the site, and so dictate levels in the on-site ditch network.
Figure 2.10 – Otmoor research field site location;
© Crown copyright Ordnance Survey 2007

Figure 2.11 – Otmoor field site
The Grips research designed the layout of the grip network to have varying spaces between the grips, allowing the effect of different spacings on hydrology to be determined. The distances can be seen in Figure 2.12 to be 5, 10 and 20 m, and the instrumentation is centred on this research objective, placed in transects across the different spacings. The grips were dug by rotary machine, (Figure 2.13) and were ~2 m wide and ~40 cm deep. For scientific robustness the Grips project required a control site (without grips), shown to the east of the treatment area in Figure 2.11. The 5 m grip spacing is shown in Figure 2.14.
Figure 2.13 – Grips being dug at Otmoor

Figure 2.14 – Grips: 5 m spacing
2.2.2.2 – Automatic Weather Station

As described in Section 1.1.4, precipitation and evapotranspiration can be the major hydrological input and output respectively to a wetland site. Evapotranspiration may be estimated from meteorological measurements, and so an automatic weather station (AWS) is central to any wetland hydrological monitoring project, also measuring precipitation.

Standard meteorological parameters were monitored by an AWS (Didcot Instruments, Abingdon, UK; Figure 2.15) located centrally on the research plot, as shown in Figure 2.16. These parameters are listed in Table 2.2. The instruments were positioned at 1.5 m above the ground in contrast to the more common 2 m, as low as possible so as not to act as a perch for crows, which predate the nests of RSPB target bird species. This was a condition of sighting the instrument within the RSPB reserve, but did not affect the operation of instruments. All instruments of the AWS are powered by a solar panel and battery, and output is logged in a storage module to be downloaded at intervals of up to three months. Data are averaged for both hourly and daily outputs.
Figure 2.15 – Automatic weather station on site at Otmoor

Figure 2.16 – Location of automatic weather station on site at Otmoor
Penman (1948) devised the estimation of evaporation from open water surfaces using standard meteorological variables. An important assumption of this approach is the lack of water limitation, explaining the common term potential evaporation, $E_p$. This estimate of potential evaporation from a reference (or standard) crop can be modified, and used to estimate actual evaporation at a site, $E_a$, using a crop factor. A crop factor is the ratio of $E_a/E_p$ for a reference crop grown under similar conditions (Cain, 1998; Burgin 2006).
Penman's (1948) equation for potential evaporation is as follows:

\[ E_p = \frac{\Delta(R_n - G) + \gamma 6.43 f(u)(e_a - e_d)}{\gamma + \Delta} \]

Equation 2.1

Where

- \( E_p \) = potential evaporation \( \text{kJ m}^{-2} \text{s}^{-1} \)
- \( \Delta \) = gradient of saturation vapour pressure with temperature \( \text{kPa} \text{ °C}^{-1} \)
- \( R_n \) = net radiation \( \text{Wm}^{-2} \)
- \( G \) = ground heat flux \( \text{Wm}^{-2} \)
- \( \gamma \) = psychometric constant \( \text{kPa} \text{ °C}^{-1} \)
- \( f(u) \) = Empirical wind speed function \( \text{N/A} \)
- \( e_a \) = saturation vapour pressure \( \text{kPa} \)
- \( e_d \) = actual vapour pressure \( \text{kPa} \)

The equation combines available energy and aerodynamic properties, and reflects the dryness of the air through vapour pressure deficit \((e_a - e_d)\). The saturation vapour pressure \((e_a)\) and actual vapour pressure \((e_d)\) are commonly calculated as follows:
\[ e_a = 6.108 \exp \left( \frac{17.27T_a}{T_a + 237.3} \right) \]

Equation 2.2

\[ e_d(T_d) = 6.108 \exp \left( \frac{17.27T_d}{T_d + 237.3} \right) \]

Equation 2.3

Where

\[ e_d(T_d) = \text{actual vapour pressure at the dewpoint temperature} \] kPa

\[ T_a = \text{dry bulb temperature} \] °C

\[ T_d = \text{dew point temperature} \] °C

Where measurements of dew point temperature are not made, the wet bulb depression is used to calculate \( e_d \): 

\[ e_d = e_a(T_w) - \gamma(T_a - T_w) \]

Equation 2.4

Where

\[ e_d = \text{actual vapour pressure} \] kPa

\[ e_a(T_w) = \text{saturation vapour pressure} \] kPa

at wet bulb temperature

\[ T_a - T_w = \text{wet bulb depression} \] °C
In cases where wet bulb temperature is not available, it is possible to calculate \( e_d \) using daily minimum temperature (\( T_{\text{min}} \)) values instead of dewpoint temperature (\( T_d \)). This is possible as daily minimum temperature is approximately equal to dew point temperature (Burgin, 2006; Harding, pers. comm.). In this instance, \( T_d \) is substituted with \( T_{\text{min}} \) in Equation 2.3:

\[
e_d(T_{\text{min}}) = 6.108 \exp \left( \frac{17.27 T_d}{T_{\text{min}} + 237.3} \right)
\]

Equation 2.5

Penman’s (1948) work was developed by Monteith (1965) to take account of the properties of vegetation, and so transpiration in addition to evaporation (Gavin, 2001). The result is the widely used Penman-Monteith equation:

\[
E_p = \frac{\Delta (R_n - G) + \rho_a C_p \left( e_a - e_d \right)}{\frac{r_s}{r_a} \left( \Delta + \gamma \frac{r_s}{r_a} \right)}
\]

Equation 2.6

Where

- \( \rho_a \) = mean air density \( \text{kgm}^{-3} \)
- \( C_p \) = specific heat capacity of air \( \text{MJkg}^{-1}\text{C}^{-1} \)
- \( r_a \) = aerodynamic resistance \( \text{sm}^{-1} \)
- \( r_s \) = bulk surface resistance \( \text{sm}^{-1} \)
The aerodynamic resistance, $r_a$ (Equation 2.7), and the bulk surface resistance, $r_s$ (Equation 2.8), are important components of calculating the transfer of heat and water from the crop surface to the atmosphere. $r_a$ describes the resistance to air flow over the surface, and $r_s$ describes the resistance of water flow through vegetation to the atmosphere.

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

**Equation 2.7**

Where

- $z_m = \text{height of wind speed measurement}$, m
- $d = \text{zero plane displacement of wind profile}$, m
- $h = \text{crop height}$, m
- $z_{om} = \text{roughness parameter for momentum}$, m
- $z_{om} = 0.123 \, h$
- $z_h = \text{height of air temperature and humidity measurements}$, m
- $z_{oh} = \text{roughness parameter for heat and water vapour}$, m
- $z_{oh} = 0.1 \, z_{om}$
- $k = \text{von Karman constant}$, 0.41
- $u_z = \text{mean wind speed at height } z_m$, ms$^{-1}$
Where $r_s = \frac{r_i}{LAI_{active}}$

Equation 2.8

The leaf area index (LAI) represents the leaf area (upper side only; $m^2$) per unit area of soil ($m^2$), and so is effectively dimensionless. The bulk stomatal resistance is the average stomatal resistance of an individual leaf (Allen et al., 1998).

Equation 2.7 assumes neutral stability conditions, with temperature, atmospheric pressure and wind velocity distributions following near-adiabatic conditions with no heat exchange (Allen et al., 1998). As well watered surfaces (such as wetland environments) have small levels of sensible heat exchanged, and primarily latent energy exchange (evaporation), the equation is typically appropriate in such environments.
2.2.2.3 – Solent

Although commonly used, the above equations for estimating evapotranspiration are limited by the fact that they only estimate potential evapotranspiration, $E_p$. One method which can be employed to measure directly the evapotranspiration at a site is the eddy correlation-energy balance (ECEB) technique. This employs expensive instrumentation (~£20,000), but is regarded as perhaps the most reliable of direct measurement techniques, as it relies on a minimum of theoretical assumptions and has a high accuracy with errors in the order of 5-10% (Shuttleworth, 1993).

Reynolds averaging is employed, which assumes that atmospheric entities have mean ($\bar{x}$) and fluctuating ($x'$) components, such that:

$$x = \bar{x} + x'$$

Equation 2.9

The vertical and horizontal components of the instantaneous wind are:

$$uw = (u' + \bar{u})(w' + \bar{w})$$

Equation 2.10

$$uw = u'w' + u\bar{w}' + \bar{u}w' + \bar{u}\bar{w}$$

Equation 2.11

$$uw = \bar{u}\bar{w} + u'w'$$

Equation 2.12
The eddy correlation technique measures the surface heat flux, using rapid response measurements of vertical wind fluctuations \( w' \) and potential temperature fluctuations \( \theta' \). Instantaneous potential temperature \( \theta \) is conserved if the negligible effects of radiation and molecular energy are neglected:

\[
\frac{D\theta}{Dt} = 0
\]

**Equation 2.13**

Then Reynolds averaging discloses the rate of change of horizontal mean potential temperature:

\[
\frac{D\bar{\theta}}{Dt} = \frac{\partial w' \theta'}{\partial z} = \frac{1}{\rho C_p} \frac{\partial H}{\partial z}
\]

**Equation 2.14**

Then,

\[
H = \rho C_p \overline{w' \theta'}
\]

**Equation 2.15**

Where \( H = \text{sensible heat flux} \quad \text{W m}^{-2} \)

A sonic anemometer can be used to measure \( w' \), and a platinum resistance thermometer to measure \( T' \), which is almost identical to \( \theta' \).
The sensible heat flux, $H$, is then inserted into Equation 2.16, a simple energy balance, to find the residual latent heat flux. A net radiometer measures $R_n$ and a ground heat flux plate $G$.

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

Equation 2.16

In order to measure evaporation on site accurately, the eddy correlation-energy balance technique was employed. A sonic anemometer (Solent R3, Gill Instruments, Lymington, UK; Figure 2.17) was used to measure $\nu'$, and a platinum resistance thermometer to measure $T'$. The sensible heat flux, $H$, calculated as described above, is then inserted into Equation 2.16, an energy balance, to find the residual latent heat flux. A net radiometer and a ground heat flux plate complete the parameters required to close Equation 2.16. The instrument uses significantly more power than the AWS, reflected in the size of the solar panels supplying the electricity (4 X 50 W; storage in 4 X 18 v lead acid batteries). Although comprising many components, the system has adopted the name ‘Solent’ from its most distinguishing feature, and its siting at Otmoor is shown in Figure 2.18.
Figure 2.17 – ‘Solent’, incorporating sonic anemometer (top), net radiometer (on horizontal arm), and to the right control box and solar panels

Figure 2.18 – Solent location on Otmoor site
2.2.2.4 - Surface Water Storage: Stageboards

Surface water levels are monitored by basic metric stageboards. The water levels are simply read against the boards, which were levelled to Ordnance Datum using existing Datum benchmarks on site and a surveyor's level. Two stageboards (with labels SB1 and SB2) were installed on the study site (one of which, SB1, is shown in Figure 2.19), one in each half of the grip network to identify any transfers across the research plot. Two others placed nearby measured water levels in the ditches surrounding the plot (SB3) and the main storage channels on site (ring ditch; SB4), with the locations shown in Figure 2.20.

Differences between water levels at stageboard locations can reveal important information about the runoff water transfer mechanism, as water levels may be used to determine any direction of flow both across the research plot and between the research plot and the wider ditch network on site, and the stageboard locations were chosen for this reason. It was also important to compare water levels in the grips against in-field water table levels, as this would indicate any lateral flow between the water stored in the grip network and the adjacent soil profile. Stageboards were also used to calibrate water level recorders (detailed below).
Figure 2.19 – Stageboard one (SB1)

Figure 2.20 – Stageboard locations

- Otmoor ditch network
- Research plot grips
- Access bridleways
- Stageboard locations
2.2.2.5 – Subsurface Storage: Dipwells

Depth to the shallow water table was monitored by a series of dipwells. These are plastic tubes (1.5 m long, 35 mm diameter) inserted into a hand-augured hole 1.5 m deep and 4 cm in diameter. Regular 1 mm screening (slots) in the tube allowed the free flow of water in order that the level of water in the tube represented the water table, as shown in Figure 2.21. Sand was placed around the dipwell to secure it and prevent clay blocking the sides; a clay seal was inserted around the top of the dipwell to prevent surface water entering the augured hole, and a cap placed on the top to prevent debris falling into the well. The water level in the dipwell was measured using a diptone, an electrical water sensor attached to a tape measure.

Figure 2.21 – Dipwell installation
One of the primary research foci of the Grips research projects was to determine the maximum distance away from a grip where the hydrology is influenced by the grip and its water level. For this reason a systematic network of dipwells was installed across the site (Figure 2.22). Three different distances between grips were tested (5, 10 and 20 m), and dipwells were placed at increasing distances from the grip centres (1.1, 1.5, 2.5, 5 and 10 m, maximum grip spacing permitting). Three repetitions of a transect across the three grips spacings were installed across the site. Due to smaller grip spacing preventing higher distances from grip, repetitions were installed in higher grip spacings to maintain statistical robustness of the 5 and 10 m dipwells. A control was installed consisting of a further thirteen dipwells in an adjacent, non-gripped area. Each of the three transects were split and labelled according to distance, making nine minor transects between grips, with three of each across 20, 10 and 5 m. The transect labels and locations are shown in Figure 2.22, and the numbers of each distance from grip of monitoring stations shown in Table 2.3. A control transect consisted of 13 dipwells equally spaced 10 m apart across the control area, making a total of 97 monitoring stations.
Figure 2.22 – Dipwell transect locations

<table>
<thead>
<tr>
<th>Transect ID</th>
<th>1.1 m</th>
<th>1.5 m</th>
<th>2.5 m</th>
<th>5.0 m</th>
<th>10.0 m</th>
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<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
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</tr>
<tr>
<td>C</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
<td>N/A</td>
</tr>
<tr>
<td>D</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
<td>N/A</td>
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<tr>
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<td>2</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
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<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
<td>N/A</td>
</tr>
<tr>
<td>G</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>H</td>
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<td>2</td>
<td>2</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>I</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
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<td><strong>18</strong></td>
<td><strong>18</strong></td>
<td><strong>18</strong></td>
<td><strong>12</strong></td>
</tr>
</tbody>
</table>

Table 2.3 – Number of monitoring stations at each distance from grip for each transect
2.2.2.6 – Subsurface Storage: Pressure Transducers

Pressure transducers (Diver DI240, Van Essen Instruments, Delft, The Netherlands; Figure 2.23), were used to record water levels. They were placed at 1.1, 5 and 10 m from the grip within a 20 m grip spacing (see location map Figure 2.24). A further pressure transducer was placed in the foot drain adjacent to stageboard one to monitor surface water levels.

![Water level pressure transducer](image)

The pressure transducers were placed in piezometers. These were plastic tubes (2 m deep, 35 mm diameter) screened for the lower metre and unscreened for the top metre. This, and a sealing of clay surrounding the unscreened section, allows the effect of surface water to be excluded and only the water table level to determine the water level within the well (Figure 2.25a). The pressure
transducers log data internally, and were calibrated during monthly site visits by checking depth to water manually.

Figure 2.24 – Pressure transducer locations

The objective of the hydrological monitoring during this plot scale research was to determine the hydrological functioning of the plot at the smallest possible spatial and temporal scales, and this was taken into account when programming the pressure transducers. No assumptions were made as to the temporal resolution that was required, and so they were programmed to record water levels every ten minutes, a frequency far greater than conventional wetland sampling, usually every two weeks at most.
A pressure transducer recording barometric pressure placed in a dry well was used to remove the effect of atmospheric pressure from those recording water levels. The dry well was otherwise identical to the water level pressure transducer installations as described above, but not screened and sealed at both top and bottom, with small air holes to allow changing air pressure to be reflected inside the tube (Figure 2.25b).

Figure 2.25 – Pressure transducer installations for a) water levels in piezometer and b) barometric pressure transducer in dry well
2.2.2.7 - Surface Capacitance Insertion Probe

The Soil Capacitance Insertion Probe (SCIP) was developed by the Institute of Hydrology (IH, now integrated into CEH) for the measurement of soil moisture content in the unsaturated zone. This is done indirectly by measuring the soil dielectric constant (Evans, 1994), which is closely dependant on soil moisture content and, to a lesser extent, soil type. Each SCIP unit is calibrated in the laboratory using liquids (of known dielectric constant) and air, and unit-specific coefficients for conversion of output to volumetric soil moisture determined. The SCIP will measure total water content in the soil, including any water adsorbed (chemically bound) to the soil structure and so unavailable to plants. The SCIP assesses the dielectric constant of soil using two rods placed into the soil profile, and the rods are only able to measure at either 5 or 10 cm below the surface. The unit is shown in Figure 2.26.

The SCIP was used at every monitoring station adjacent to the dipwell during sample runs. The variable depth of measurement was utilised at different types of visit, with both 5 and 10 cm depths used during full ‘Grips’ visits, but just ten cm readings taken during the more frequent intermediate visits, to increase the speed of the intermediate sampling run. More details of the differences between the two types of site visit are detailed below in Section 2.2.2.10.
2.2.2.8 – Soil Profile Analysis

Soils profiles can reveal information concerning the hydrological history of the soil, since the saturation status determines the availability of oxygen, and so dictating the reduction or oxidation status and subsequently colour of soil minerals (Mitsch and Gosselink, 2000). As clay soils have a very high mineral content, notably iron, they are particularly open to this analysis; iron has strong colour changes with oxidation status, being brown to red in the presence of ferric
oxide (oxidised) which is reduced to an obviously blue ferrous oxide through lack of oxygen. By drilling through the soil profile and comparing the colours present to standard colour charts (Munsell Color, 1990), much information about previous water table levels can be retrieved. Furthermore, the presence or otherwise of carbonate nodules in the soil reveals if the sediment has been deposited by alluvial floodwater, which has a high carbonate content which precipitates as carbonate nodules over time.

For this study, an electric percussion drill (04.19.SD, Eijkelkamp, Giesbeek, The Netherlands) was used to drill to a depth of 4.1 m (below ground surface, 58.0 maOD), with a sample chamber 4 cm wide allowing inspection of each sample for colour and carbonate presence on the surface after extraction. The depth of each changing colour and so oxidation status is recorded along with photographing each layer and transition. The location of the drilling site is shown in Figure 2.27. Only one sample was taken due to equipment availability; this was thought sufficient as the upper 1.5 m was known to be consistent across the site from repeated dipwell drilling, and lower than 1.5 m was unlikely to vary much across the research plot due to decreased surface influence. Also, data were available for comparison from a drilled core taken for the RSPB in 1997.
2.2.2.9 - Guelph Permeameter

Hydraulic conductivity, $K$, is a key variable for understanding soil water movement, and field saturated $K$ in the unsaturated zone was measured using a Guelph permeameter (Figure 2.28). The method utilises the Mariotte siphon principle to maintain a constant head of water in an augered hole (Soilmoisture, 1987). By measuring the amount of water required from a graduated vessel to maintain the head at a steady state rate of flow, calculation of $K$ is possible. Gavin (2001) utilised the Guelph permeameter extensively and detailed the many comparisons made with the more traditional auger hole method, and slug tests. Bromley et al. (2004) detail issues related to scale concerning the measurement of $K$, including the conclusion that point measurements such as
the Guelph permeameter may underestimate plot scale $K$ by up to two orders of magnitude through missing larger scale preferential flow pathways, as described in Section 1.4. Three repetitions were taken on a single day in different locations across the plot, selected at random but away from grips and spoil from grip digging that would interfere with measurements; see Figure 2.29. With such a non-conductive soil, the time taken to reach equilibrium was lengthy (~1.5 hours) and three repetitions was considered representative.

Figure 2.28 – Guelph permeameter
2.2.2.10 - Sampling Frequency

In order to add value to the comprehensive, yet modest temporal resolution of sampling provided by the Grips research project, a high frequency sampling regime was planned. Weekly sample runs covering fewer parameters were added to the monthly Grips sample runs with all parameters. Table 2.4 summarises the parameters covered, and each parameter was recorded at all 91 hydrological monitoring stations.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Instrument</th>
<th>Logging permanently</th>
<th>Full Grips sample run</th>
<th>Intermediate sample run</th>
<th>Measured once</th>
</tr>
</thead>
<tbody>
<tr>
<td>Meteorological parameters</td>
<td>AWS</td>
<td>✓</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Evaporation</td>
<td>Solent</td>
<td>✓</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Surface water</td>
<td>Stageboards</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td></td>
</tr>
<tr>
<td>Water table height</td>
<td>Dipwells</td>
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<td>✓</td>
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<td></td>
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<td>Water table height</td>
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<td>Soil moisture at 5 cm</td>
<td>SCIP 5 cm</td>
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<td>Guelph permeameter</td>
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</tbody>
</table>

Table 2.4 – Data collected on different sample runs

Grips sample runs began in Spring 2004, and finished in Spring 2007. The intermediate weekly sample runs stretched from June 2005 until July 2006, covering two growing seasons. Figure 2.30 shows the dates of samples for both types of sample runs.
Certain instruments provided the ability to log results, which was particularly important as it added the utility of being able to demonstrate the response of parameters to rainfall. It means the parameters involved are not restricted to weekly or even monthly readings. This was the case with the AWS, Solent and pressure transducers. This sampling regime ensures that as many temporal scales of sampling were covered as possible. This ranged from pressure transducers recording at a sub-hourly frequency, to the sampling programme being spread over three years.

2.2.3 - Multi-Scale Model

In order to enable further investigation of scaling effects with direct consideration of the Otmoor study site, a multi-scale model was utilised. The work was a sensitivity analysis exercise, looking at the effect of different scales of input data to an existing model estimating evapotranspiration. Attention was drawn to the model code during a seminar at CEH Wallingford (Ellis, pers. comm.) reviewing...
the utility of an energy and water balance model used at the global scale. The links between this global scale model of evapotranspiration and the work undertaken at Otmoor on evapotranspiration at the field scale were obvious within the context of scale issues. Richard Ellis stated that he was looking for field data to validate the global scale model, and hence collaboration in this area provided important opportunities for the present study, whereby in return for validation data, model runs could be undertaken with different driving data, in order to investigate the effects of running the model at different scales and directly addressing many of the scale issues highlighted earlier. Therefore a mutually beneficial relationship was established, whereby Richard Ellis' knowledge of the model code and expertise in large scale water fluxes were complemented by the current work's comprehensive hydrological dataset for the Otmoor site, together with developing field scale evaporation understanding and time to process results.

The model used was that developed by the Joint Centre for Hydro-Meteorological Research (JCHMR, based at CEH Wallingford), called the Joint UK Land Environment Simulator (JULES), as described by Essery et al. (2001) and JCHMR (2007). This model can be used to calculate the surface energy balance, and so vertical fluxes of water, at sub-diurnal time steps, and is driven by meteorological data. The code is based on that of MOSES (Met Office Surface Exchange System; Cox et al., 1999), originally designed to represent the land surface in meteorological and climate models; applications now include wider water resource issues (Choudhury et al., 1998) and global scale wetland identification (JCHMR, 2007). The performance of JULES (previously known as
MOSES 2.2) has been assessed in climate simulations by Essery et al. (2003), where it compared well with observed climatological data.

JULES is a model designed to calculate water, energy and CO$_2$ fluxes at a global scale by representing the atmosphere, soil and vegetation systems, as described by Cox et al. (1999) and shown graphically in Figure 2.31. Importantly for the current application, the total moisture flux from the land surface to the atmosphere comprises the components of evaporation from water intercepted by the canopy (labelled 1), transpiration from vegetation (2) and evaporation from the soil (3). The snow component was seen as negligible due to latitude, and the subsurface runoff considered to be less important due to low hydraulic transmissivities.

JULES allows distribution of variables within grid cells, enabling the representation of sub-grid heterogeneity of meteorological parameters (such as temperature, short- and longwave radiative fluxes, sensible and latent heat fluxes, ground heat fluxes and canopy moisture), and local physical parameters such as vegetation cover and type. This 'mosaic' structure improves estimation of model outputs within the existing working resolution of the model. Parameters for nine surface types are held within the model, and a combination of these can form the grid square coverage.
Water and carbon budgets are intricately linked, and JULES has been designed to model both. Fundamental to the flux of both between the Earth's surface and atmosphere is vegetation and its activity, primarily through stomatal conductance (Essery et al., 2001). Photosynthetic activity enables the stomata to open and water and carbon dioxide to be exchanged between plants and the atmosphere, providing the pathway for transfer; the amount of flux is also dependent on the gradient of each between the different stores. For temperate (C3) grass water-atmosphere interaction, JULES utilises work from Jacobs (1994) and Cox et al.

The surface energy balance for each tile includes fluxes of sensible and latent heats, and moisture. Soil surface properties also impact upon the land-atmosphere moisture flux, as the heat flux into the ground (combining radiative fluxes below vegetation canopies and conductive fluxes for any unvegetated areas) is parameterised as a function of surface layer thickness and temperature. Aerodynamic resistance for sensible and latent heat fluxes is calculated as a function of temperature, humidity and windspeed, and for transpiration, surface resistance is the reciprocal of canopy conductance (Essery et al., 2001).

Hydrology is modelled using the partitioning of precipitation into interception, throughfall, runoff and infiltration, as described by Gregory and Smith (1990). Soil moisture is modelled in JULES using a finite difference approach to the Richards (1931) equation, with vertical discretization assigned by the user. Saturated zone water fluxes are modelled using the Darcy equation (Fetter, 1994), and so are directly dependent upon soil hydraulic conductivity (\(K\)) and soil water suction (or depth to saturated zone). McGuire et al.’s (1992) model is used in the unsaturated zone, whereby respiration increases with soil moisture content until an optimum threshold of moisture is reached, when respiration begins to decrease (JCHMR, 2007).
Three model runs were set up, each with a different spatial scale, in order to investigate the model's response to changes in scale of input data. Firstly, data were extracted for the grid square in which Otmoor lies from a simple global scale run (scale 1). This uses meteorological inputs from the Global Soil Wetness Project 2 (GSWP2, 2007), an ongoing research activity contributing to the Global Energy and Water Cycle Experiment (GEWEX). The data have a one degree spatial resolution and three hour temporal resolution, and are designed to be used with global scale models. LAI and soil properties (hydraulic conductivity, soil profile data) are estimated from a satellite-derived global dataset within JULES.

A second run (scale 2) used input data from the automatic weather station (AWS) at Otmoor, to investigate changes in output associated with finer spatial scale resolution meteorological inputs, but the other data input to the model remained as scale 1. As the Otmoor AWS did not carry a fully comprehensive list of parameters (this was not the anticipated application), limited data were taken from the AWS at CEH Wallingford, 24 km away. As these parameters (humidity and atmospheric pressure) are those which vary only slightly across such distances, it was thought that this method of increasing data coverage was appropriate. LAI data collected at Otmoor under the Grips project was interpolated to provide an estimate of change through the season, but soil properties were kept as for scale 1.

Lastly, local surface parameters from investigations in Chapter 2 were used to calibrate the model for a point scale spatial resolution, again using the Otmoor
AWS for meteorological data. This model run (scale 3) required data on initial soil moisture conditions, vegetation height, land surface cover, LAI, and soil properties. In effect, the scale 3 model represented a model of a flux tower, modelling water fluxes for only one point spatially. The scale 3 model was split into two sections, with one representing Otmoor with its actual properties of heavy clay soil (3a), and another with a more hydraulically transmissive, silty soil (3b), in order to assess the sensitivity to soil parameters. Actual evapotranspiration measurements from the Solent formed scale 4, completing the continuum of increasingly fine resolution, and so increasingly small scale models, as summarised in Table 2.5.

<table>
<thead>
<tr>
<th>Model Run</th>
<th>Meteorological Input</th>
<th>Output</th>
<th>Detail</th>
</tr>
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<tbody>
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<td>GSWP2</td>
<td>1 degree grid cell</td>
<td>Global scale model run</td>
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<tr>
<td>2</td>
<td>Otmoor AWS</td>
<td>1 degree grid cell</td>
<td>Global scale but with local meteorological input</td>
</tr>
<tr>
<td>3a</td>
<td>Otmoor AWS</td>
<td>Point</td>
<td>Local scale with measured soil parameters</td>
</tr>
<tr>
<td>3b</td>
<td>Otmoor AWS</td>
<td>Point</td>
<td>Local scale with varied soil parameters</td>
</tr>
<tr>
<td>4</td>
<td>Solent</td>
<td>Point</td>
<td>Taken as accurate measurement of evapotranspiration</td>
</tr>
</tbody>
</table>

Table 2.5 – Details of scale levels in the multi-scale model of evaporation
For Scale 1, GSWP2 meteorological data were not available for 2006, and so the model was run for the ten years 1986 to 1995 and results averaged for each calendar day from 100 to 150 over the ten years. This should reveal seasonal and diurnal cycles in the energy budget for the wetland from the global scale model. The scale 2 and 3 models were run from 10th April 2006 for 150 days, in order to cover the growing season, which has the highest evapotranspiration levels.
2.3 – Results

2.3.1 – Hydrological Monitoring

2.3.1.1 - AWS

Data from the AWS will be used for two ends. Firstly, many other hydrological parameters measured on site will need to be placed in the context of prevailing conditions, as they respond to available energy and water which is dictated by meteorological conditions, and for this end a summary of meteorological data follows. Secondly, data collected by the AWS will be used to estimate evaporation, as outlined in Section 2.2.2.2.

The daily data are presented (Figure 2.32) from installation of the unit at Otmoor on 31/07/2004. Gaps in the data appear from 18/08/2004 to 09/10/2004 and 17/03/2005 to 12/05/2005 due to malfunction and return for the addition of soil temperature probe respectively. The trace of daily-averaged temperature data (Figure 2.32a) shows a clear seasonal trend as expected. Long term (1971-2000) average mean, minimum and maximum data from the Oxford meteorological station (6 km away) are shown, and the data largely fall within these bounds. There are regular extrusions beyond the average minimum and maximum values, but this might be expected due to the more exposed nature of
the Otmoor site when compared to more urban setting for the Oxford meteorological station, and the averaged nature of the Oxford data.

**Figure 2.32** – Otmoor meteorological data from August 2004 to December 2006; a: temperature (average data from Oxford also shown), b: net radiation, c: rainfall
Net radiation (Figure 2.32b) data also show a clear seasonal trend, with a much greater variation in summer as expected due to the large dependence on (highly variable) cloud cover. Net radiation is a key parameter, as it indicates the amount of energy available in the locality, which is used for evapotranspiration. Daily totals of rainfall are shown (Figure 2.32c), and the general trend is for the highest individual totals to occur during the summer months, caused by intense convectional events, with the maximum being 61 mm on August 10\textsuperscript{th} 2004. Frontal rainfall leads to more consistent rainfall totals during winter months.

Evaporation may be estimated from meteorological data using the methodologies outlined in Section 2.2.2, namely Penman (1948) and Penman-Monteith (Monteith, 1965) methods. After inspection of the data, it was found that the wet bulb thermocouple psychrometer was faulty from installation to replacement on 13/07/2006. The erroneous data are shown in Figure 2.33. Firstly it may be seen that wet bulb temperature values are consistently only $-0.5 ^\circ C$ below the dry bulb values; there is usually a far greater difference, normally several degrees during daylight hours. Secondly, at times the wet bulb temperature actually rises above that of the dry bulb temperature (difference on graph is negative), which is physically impossible.
This problem necessitated the utilisation of Equation 2.5 to calculate actual vapour pressure, $e_d$, rather than the more conventional Equation 2.3. This method was verified by Burgin (2006), using data from another of the Grips field sites, Pawlett Hams, Somerset, shown in Figure 2.34, a scatter plot of the two methods for the dates 10/05/2005 to 31/10/2005. An $r^2$ value of 0.94 is obtained confirming that the $T_{min}$ method is appropriate for use, although it was not possible to determine which method was most accurate, as no record of actual evaporation is taken at the Pawlett Hams site.
A replacement wet bulb thermocouple psychrometer was installed at Otmoor on 12/07/2006. Figure 2.35 shows evaporation calculations plotted against each other with the distinction of before and after this date. It can be seen that previous to the newly installed wet bulb thermocouple psychrometer, evaporation estimates using $T_d$ were underestimated significantly. After installation of the new instrument, the two methods match up very well, confirming use of the $T_{min}$ method as justified for the whole period of Otmoor data collection.

Figure 2.34 – Comparing $T_d$ and $T_{min}$ methods using Pawlett data
Having verified that the method of calculating $E_p$ using $T_{\text{min}}$ is accurate and likely to provide a good estimation of evaporative loss at the Otmoor plot, Figure 2.36 shows $E_p$ over the period of 2005-2006. There is a clear seasonal trend and in the summer months a great inter-daily variation of evapotranspiration level, caused by variation in meteorological variables such as solar radiation and wind speed. The maximum daily evaporation was estimated to be 5.9 mmday$^{-1}$, which is relatively high when compared to previous estimates of potential evaporation.
from wetland sites. Gasca-Tucker et al. (2007) found a maximum actual evaporation rate ($E_a$) of 4.6 mm day$^{-1}$ from a wet grassland site at the Pevensey Levels in Kent, UK, and Gavin (2001) found a maximum potential evaporation ($E_p$) of 4.1 mm day$^{-1}$ at another wet grassland site at Elmley Marshes, again in Kent. A rate of 6.9 mm day$^{-1}$ has been observed by Smid (1975) from a 'vigourously growing dense reed stand', but this is unusually high. Acreman et al. (2002b) also found a high $E_a$ rate from a wet grassland in southwest England of 5.8 mm day$^{-1}$. Standard error for Penman Monteith measurements has been estimated as 0.15 mm day$^{-1}$ (Sumner and Jacobs, 2005).

Figure 2.36 – $E_p$ estimated using the $T_{min}$ method at Otmoor
2.3.1.2 – Solent

A summary of evapotranspiration as measured by the Solent is shown in Figure 2.37, giving daily averages from the date of installation in May 2005 to December 2007. The expected annual pattern of high evapotranspiration in summer months is shown, with a maximum $E_a$ of 5.1 mm day$^{-1}$ which is slightly lower than the $E_p$ estimated from AWS data, but still comparable with previous studies on wet grasslands. Also of note is the highly variable nature of evapotranspiration within short timeframes, which is expected as evaporation is affected greatly by highly variable meteorological parameters.

Hourly Solent data are plotted in Figure 2.38, for a period of May 30th 2006 to 19th July 2006. This is a summer period with high evapotranspiration levels and a strong diurnal pattern, and so reveals much useful information on the daily hydrological functioning of the wetland evaporation system, such as condensation (evaporation levels < 0) occurring during the hours of darkness. It can be seen that when daytime maxima are lower, night time minima are also less extreme, leading to a smaller daily range. It is thought this is due to increased cloud cover reducing both energy available during the day and insulating during the early hours to limit condensation. The large effect this has can be seen on both the daily totals and hourly detail as shown in Figures 2.37 and 2.38 respectively.
Figure 2.37 – Daily Solent data 2005-2006; red lines show range of Figure 2.38

Figure 2.38 – Hourly Solent data summer 2006
When daily $E_p$ (estimated from AWS data using $T_{\text{min}}$) and $E_a$ (Solent) are compared directly (Figure 2.39), there is a generally strong relationship between them. Although difficult to distinguish between the daily point measurements, the general trends of the two datasets are very similar, together with the ranges and level of inter-daily variation. It is known (Shuttleworth, 1993) that the Solent data are of a higher accuracy (5-10%), as reflected by it being regarded as actual evaporation, $E_a$.

Figure 2.39 – Daily $E_p$ (AWS) and $E_a$ (Solent) presented as time series
A scatter plot of the same data (Figure 2.40) reveals more information about the relationship between the two methods of evaluating evaporation on site at Otmoor. There is a good fit between the two datasets ($r^2$ of 0.71), but there is a spread of points especially towards the middle range (~3 mm day$^{-1}$). Closer to the higher end of the range, $E_p$ is higher than $E_a$, suggesting some limitation due to available water at the site. This might occur if the water table drops far enough below the surface to be disconnected from the atmosphere and thus preventing its loss, or similarly deep enough to be beyond the reach of plant roots for inclusion in transpiration component of water loss. The apparent low number of points in Figure 2.40 is due to both methods recording to one decimal place, and so data stacking up on one another.

![Figure 2.40 – Daily $E_p$ (from AWS; using $T_{min}$) and $E_a$ (Solent); solid line shows 1:1 fit](image-url)
2.3.1.3 – Surface Water Storage: Stageboards

The stageboard data, shown in Figure 2.41, confirms that the grip system is not as deep as the wider ditch network: the grips (SB1 and SB2) quickly dry out in the summer months. The broken blue horizontal line in Figure 2.41 represents the level at which this occurs, 58.320 m aOD. Water levels in SB1 often fall more quickly than those at SB2. Also, during high water levels, SB1 is consistently higher than SB2, and together, these trends indicate a flow of water southwards, from SB1 to SB2, and towards the ring ditch and so ultimately off of the site.

Water levels at SB3 (ditch network) are generally higher than at SB4 (ring ditch). There is a turnpipe (shown in Figure 2.42) which allows water to flow off of the site from the wider ditch network to the ring ditch (from SB3 to SB4) if higher than 58.72 m aOD. Otherwise, water remains in the ditch system to feed the network of smaller ditches, including the grips on the research plot. As Figure 2.41 shows, the level of the turnpipe is rarely reached, and only in early 2007 did this occur when it appears that the flow was reversed and the ring ditch was supplying water to the site through the turnpipe. Therefore the slower route through the ditch and grip network seems more likely to be used, and runoff is not such a large proportion of water loss from the wetland system. This is a facet of the level at which the turnpipe is set, and as wetland restoration is underway, the RSBP want to keep water on site for the most part, and hence its only occasional utilisation.
Figure 2.41 – Stageboard data; black lines show years, red lines show range of Figure 2.43

Figure 2.42 – Turnpipe at SB3 (flows to SB4)
In order to demonstrate the differences in water levels and the processes they indicate, Figure 2.43 shows a shorter period of water level data from November 2005 through to March 2006. Although there are times when the water in the wider ditch network (SB3) fall below that of SB1 and/or SB2, this appears to be after periods of rainfall; the smaller volume of the grip network would cause water levels to rise faster. During the winter months, the ditch network (SB3) has a higher water level than the grip system. Water therefore enters the research plot, flows through the site and is lost to the ring ditch (SB4) through the connecting pipe. The marginally lower values for SB4 when compared to SB1 and SB2 (usually 1-2 cm) confirm this to be the case, but the small differences suggest the loss of water to be negligible.

![Figure 2.43 - Stageboard data: November 2005 to March 2006](image_url)
Data from SB1 are compared to data from a single dipwell (A1, close to a grip) in Figure 2.44. This demonstrates clearly that the grips dry out at a level which is comparatively shallow, whereas the range available to soil water levels is greater. This has implications for comparative storage in the two systems, although of course a rise in level in the soil column only includes soil pore spaces. When water levels are above the level of the grip bed, there is more variation in the soil water levels than in the grips system. This is likely to be the result of the pore space issue, whereby any change in volume of water in the soil has a greater change in height due to the relatively small pore space in which water can sit, and a low hydraulic conductivity restricting the speed of response toward equilibrium. A particularly important soil property here is the specific yield, defined by Fetter (1994) as the ratio of the volume of water that drains from a saturated soil owing to the attraction of gravity to the total volume of soil. If a soil has a low specific yield, there will be a larger change in water level for the same volume of water loss. Therefore, these data indicate a low specific yield in the soil, which is characteristic of clay soils such as that at the Otmoor plot.

The grips were dry for most of the summer of 2005 and indeed through Autumn, from early July until early November when the winter rainfall increased water levels across the site. The dipwell data show this seasonal trend more effectively, including the lower extent of the water levels, which were similar in both 2005 and 2006. These data must be placed in the wider context of the regional drought of 2005 and 2006, which consisted of deficiencies in both summer and winter rain, and which lasted until the winter of 2006-2007, to the end of the data presented.
2.3.1.4 – Subsurface Storage: Dipwells

Using dipwell data and looking firstly at the transects across the 20 m spacings between grips, some clear trends emerge. Figure 2.45 shows the dipwell data from transect A, in two halves to clarify the large amount of data, with the western half of the transect (labelled A1) in 2.45a and the eastern half (A2) in 2.45b. Sampling rates increase to weekly between May 2005 and May 2006, shown by vertical purple lines. Broken horizontal lines indicate the ground surface at each dipwell location, with colours corresponding to the dipwell data traces.
Figure 2.45 – Dipwell data from transects A1 (a) and A2 (b); vertical red lines indicate range of high frequency sampling.
It is clear from Figure 2.45 that the water levels respond seasonally, in a similar manner to the grip and ditch water levels. Winter levels are typically 1.3 m (maximum 1.5 m) above those in summer, and are at or near the surface. This has implications for the hydrological functioning of the site, as water levels are high during winter months when most rainfall occurs, there is little subsurface flood storage available. However, during summer months, when water levels fall significantly below the surface, large amounts of storage are available. Assuming a specific yield of 3% (Fetter, 1994), a water table 1.5 m below the surface provides potential storage for 45 mm of rainfall in the soil profile; the grip and wider ditch network across the research plot and the site respectively, allow further storage. Data collected reveal that rainfall events of 45 to 65 mm are common with strong convectional conditions at this site, and occur frequently over timescales of several days in winter months through frontal rainfall. This wetland system would be able to store such amounts of water easily without conveying it quickly to the wider catchment river network, if the storage was available which is only the case in the summer months. This storage capacity is somewhat dependent upon adequate infiltration capacity, a process likely to inhibited by the clay structure (although macropore development may complicate this issue).

Significantly, Figure 2.45a shows a clear pattern during summer months of dipwells nearest the centre of the field (10 m from the grip) having water levels typically 25 cm lower than those near to the grip (1.1 m). Also of note is the consistent decrease in water levels as distance from the grip increases, suggesting a significant lateral transfer of water through the soil.
During any rainfall events, water will collect in the grips (the lowest areas topographically), and infiltrate into the soil profile beneath the grips. The higher water levels in the areas close to the grips during wet periods indicate a transfer of this water away from the grip to the centre of the field. This situation indicates that the grips are acting to irrigate the field when water is available. Water levels in the grips are maintained by the wider ditch network, as established in the previous section detailing stageboard data. This pattern is less obvious in the transect A2 (Figure 2.45b). Indeed, water levels between dipwells never vary by more than 5 cm, with no pattern of one area being consistently higher or lower, even during the water stressed summer months. Such differences between two half-transects only metres apart suggests a large influence of soil heterogeneity. The grips either side of the two halves of transect A are of similar depths and had similar water levels throughout the study period, so would not have led to this discrepancy with a homogeneous soil.

During the winter months water levels across the transect are more consistent. In transect A2, however, there is a clear trend of water levels being higher in the centre of the field than nearer the grips. This indicates a situation whereby the grips will be draining the field. This is shown clearly in Figure 2.46, which shows detail from mid-December 2005 to early February 2006. For transect A2, especially when water levels are at or just a few centimetres below the ground surface, a strong process of draining from the field is evident. This is slightly less clear in transect A1, with data from the dipwell 2.5 m from the grip not following this trend.
Figure 2.46 – Dipwell data from transects A1 (a) and A2 (b) from mid-December 2005 to early February 2006.

Similar patterns can be seen on the remaining 20 m spacing transects, B (Figure 2.47) and I (Figure 2.48). However, the summer pattern is less clear on these transects, due partly to many dipwells becoming blocked with fine silt. Data either side of these gaps can give an indication of possible levels during these periods, such as across transect I2 during the summer of 2006 (Figure 2.48b). What remains very clear across the 20 m transects A, B and I is the consistently higher water levels near to the grips during the winter months, especially during very wet periods, indicating irrigation.
Figure 2.47 – Dipwell data from transects B1 (a) and B2 (b)
Figure 2.48 – Dipwell data from transects 11 (a) and 12 (b)
One factor that may affect interpretation of data from dipwells is the difference in topographic level at the dipwell location. For this reason ground heights at each dipwell have been shown on the figures. It may be seen that ground levels do not vary greatly, and so this effect should be minimal, and importantly less than the differences described above.

The 10 m spacing transects have no dipwell 10 m from the grip, as the maximum distance from grip is 5 m. Figures 2.49 to 2.51 show the data collected from dipwells in the 10 m spacings, with fewer clear patterns emerging than at the 20 m spacings. This in itself is revealing, as the threshold for a clear difference between grip and field centre may be above 10 m.

During summer periods there is little consistency between transects. For example, transect C shows a clear pattern of lower water levels towards the centre of the field in the second half of the transect (C2, Figure 2.49b), yet no such pattern (and indeed the reverse in the summer of 2006) across transect C1 (Figure 2.49a).

Data from the summer periods in transects D (Figure 2.50) and H (Figure 2.51) are limited due to the silting up of the dipwells preventing readings. Winter water levels reveal a similarly confused situation, in contrast to the trends shown at the 20 m spacings. However, it is true that immediately after a rise in water levels the centre of the fields shows a higher water level indicating the drainage of water to the grip system and away to the wider drainage network.
Figure 2.49 - Dipwell data from transects C1 (a) and C2 (b)
Figure 2.50 – Dipwell data from transects D1 (a) and D2 (b); red lines indicate range of Figure 2.52.
Figure 2.51 - Dipwell data from transects H1 (a) and H2 (b)
Figure 2.52 shows the data from late January to April 2005 for Transect D. As the water levels rise to above the ground surface, there is no obvious trend or major difference across either of the two half transects. In section D2, dipwells close to the grips start this period with generally higher water levels, and as the water level across the site rises, the water level in the centre of the field becomes higher than that near to the grip, but then this situation is reversed once more. Similar inconsistencies are apparent across transects C (Figure 2.49) and H (Figure 2.51), but, the absolute differences between dipwells are much lower. This is intuitive where there is a shorter distance between the grips and centre of the field.

![Graphs of dipwell data for Transects D1 and D2](image)

Figure 2.52 – Dipwell data from transects D1 (a) and D2 (b) from late January to April 2005
The 5 m spacing transects have dipwells only at 1.1 m, 1.5 m, and 2.5 m from the grip. Dipwell data from transects E (Figure 2.53), F (Figure 2.54) and G (Figure 2.55) are shown, but unfortunately transect G suffers significantly from a lack of data due to dipwells silting up. Water levels are seen to be very close across the 5 m transects for much of the year. Only during the summer months, especially during 2006, is there a significant difference between water levels. In these instances, it is the 2.5 m dipwells at the centre of the field which are generally higher, suggesting a draining of water to towards the grip. However, as water levels are lower than even the bed of the grip at this time, this is obviously not to be lost via runoff. It is possible that increased evaporation rates, induced by bare soil of the grip bed which cracks to a depth of several cm during prolonged dry periods, may be acting as a conduit for water loss. This effect can be seen clearly across transect E1 and E2 (Figure 2.53) together with G1 (Figure 2.55). In contrast the remaining transects (F) and half-transect (G2) show little difference across their extents during the summer months.
Figure 2.53 - Dipwell data from transects E1 (a) and E2 (b); red lines indicate range of Figure 2.56
Figure 2.54 - Dipwell data from transects F1 (a) and F2 (b)
Figure 2.55 – Dipwell data from transects G1 (a) and G2 (b)
During winter months, a confused picture again emerges, as exemplified across transect E, with detail from winter (late-January to April 2005; Figure 2.56) across transect E1 showing levels close to the grip to be consistently lower, whereas across E2 they are higher. Importantly, the differences do not fall into the bounds of error measurement, as dipwell measurements have been tested in an experimental laboratory set-up and shown to be accurate to the nearest cm; the differences here are still greater than 3 cm. However, in the case of transect F2 (Figure 2.54b), any difference may be influenced by a difference in topography across the (half-) transect, which are of the same magnitude as differences in water level. As water level represents a state of equilibrium between downward gravitational pull and upward draws on water (including capillary action and osmotic pressure), a lower topography would dictate lower water levels (Ward and Robinson, 2000), and for this reason the points cannot be compared directly.

![Diagram of transect E1 and E2](image)

Figure 2.56 – Dipwell data from transects E1 (a) and E2 (b) in 2005.
To investigate any transfer of water across the whole site (and not just across transects), dipwells from the extents of the research plot were compared. Figure 2.57 shows water level data from dipwells A2 (northwest of site) and 111 (southeast), both 2.5 m from their respective grips. Water levels in dipwell 111 are consistently higher than those in A2 suggesting there is a transfer of water from southeast to northwest across the research plot, at least in the sub-surface medium. This contrasts with evidence from the stageboards, which indicated a gradient from north to south.

![Figure 2.57 - Dipwell data across the site from northwest to southeast](image)

Figure 2.57 – Dipwell data across the site from northwest to southeast

horizontal lines show ground surface level

Figure 2.58a shows water levels from dipwells A1, B1 and G1, moving from north to south and remaining close to the grips at the 1.1 m dipwell. No obvious patterns emerge, suggesting little transfer in either direction. A similar situation
is depicted in Figure 2.58b, showing dipwells in a north-south plane but in the
centre of the fields (dipwells A6, B6 and G3).

Figure 2.58 - Dipwell data for stations across the site
Results are more elucidating when looking east-west across the site. Figure 2.59 shows evidence from dipwells 1.1 m from their respective grips in transects to the north of the site (stations A1, C1 and E1). During dry periods there is a clear and obvious trend of water levels to the east (E1) being significantly higher than the others. Strangely the centre field (transect C) shows lowest water levels by up to 20 cm. During these times water levels are some way beneath the grip beds, and so the grips are unlikely to affect water flow.

![Graph showing water level data](image)

**Figure 2.59 – Dipwell data for stations West to East across the site near to grip**

This summer trend is repeated at stations 2.5 m from the grips (A3, C3 and E3), as shown in Figure 2.60. During the winter at these stations there is a more discernable trend with stations in the west having higher water levels than those in the east, suggest an easterly transfer.
The general trend when looking at evidence for a water flux across the research plot is for transfer from north and west to south and east, so from the wider ditch network to the ring ditch and away from the site. During high water levels in winter there are often no obvious trends as to the efficacy of the grips, as water levels are uniform and often above the ground surface. During summer there is evidence of a loss of water from the plot through the grip system, likely to be facilitated by the lower ground surface allowing a more active link between atmospheric draw and deeper water resources. Significantly this would mean that the grips were acting to initiate water loss, when in fact their primary objective is to re-wet the area through increased water management. However, evidence for this is mixed which suggests high heterogeneity in soil properties allowing variable transfer of water. A key conclusion from the analysis of dipwell data is that significant differences can only be seen in summer, suggesting that
during high water levels in winter months, there is little runoff from the site, again probably due to management. In summer, the grips become dry and runoff approaches zero; it may be inferred that the dominant water transfer mechanisms at this time are precipitation and evaporation.

2.3.1.5 – Subsurface Storage: Pressure Transducers

Figure 2.61 shows the data from the pressure transducer placed at 10 m from the grip, and the manually recorded dipwell data from monitoring station A6, also 10 m from the grip within the same 20 m spacing; the transect of pressure transducers was close to transect A, as shown above on the Otmoor plan (Figure 2.24). Figure 2.61 shows only data from the recording closest to midday, and so is in effect daily data. The water levels from the two methods agree closely, with the same trends shown (with a sharp decrease in levels after mid May), yet obviously the higher temporal resolution of data from the pressure transducer reveals more detail about the water levels in this area, such as the rise in late May.
The surface water level in the grip and water table 10 m from the grip during the period from 30th May to 5th July 2006 are shown in Figure 2.62a, and associated rainfall is shown in Figure 2.62b. It is clear that the water table falls rapidly through the early summer after heavy rain in late May. The rapid rise in surface water level on 15th June and shortly afterwards in the field centre was due to managed pumping of water onto the site. Such an event is similar in nature to an extreme rainfall event and has identical consequences for water levels.
As summer develops, the grip generally has a lower water level than at 10 m from the foot drain, and so acts to drain the area. The reverse is the case in the few days after a rainfall or pumping event where surface water from the foot drain is irrigating the centre of the field, and the effects are visible even at 10 m from the foot drain, as shown after 1st and 15th June 2006.
Most striking from the results shown is the identification of a strong diurnal fluctuation of water table levels, of up to 10 cm. As the water level drops through the summer months, the diurnal fluctuation in water levels begins to weaken, particularly so when water levels drop to approximately 0.70 to 0.75 m below the ground surface. Although common in many more permeable soils, and especially wetland soils where evapotranspiration is not water limited (Gilman, 1994; Wetzel, 1999), this fluctuation was unexpected at Otmoor. It was assumed that the heavy clay structure of the soil profile would have a low hydraulic conductivity and impede the vertical water fluxes required for this process.

Water loss via evapotranspiration is well known to follow a similar diurnal pattern, as it is driven by diurnal cycles in temperature and light levels (Snyder and Boyd, 1987; Price, 1994). As evapotranspiration is a major conduit for water loss in wetland environments (Gavin, 2001), it can be inferred that this process drives the diurnal cycle in water levels on site.

This regime is only observed when the water level is below the ground surface; above this the process occurring is open water evaporation, and any effect will be dampened by the extensive area under water. The regime also develops in the field drain water level, once this level has dropped below that of the drain bed for the same reasons. Also of note is the cessation of this regime once the water
level approaches the limit at which phreatophytic plants can access the water, as determined by their maximum root depth and known as the extinction depth. This level, as indicated by the 10 m water level in Figure 2.62a, is about 0.70 to 0.75 m (58.00 to 57.95 maOD) below the ground surface, where the regime begins to weaken in strength.

The diurnal fluctuation was present in all soil water levels recorded, and the surface water level. Together with the linkage between the foot drain and the field centre, this implies that both vertical and lateral hydraulic conductivities are greater than the low values expected from the soil profile, advocated by its high clay content. The fluctuation suggests an active vertical water flux in the soil profile, and subsequently that evaporative loss may not be limited by the soil hydraulic properties. It is suggested that the low specific yield may exacerbate this effect by leading to a larger vertical change in water levels for a given volume change.

As a transect perpendicular to the grip was used, it is possible to examine the effect of increasing distance from the grip on water levels. Figure 2.63 shows the water level data from the grip, along with 1.1 m and 10 m from the ditch. As the water levels are dropping from inundation levels, water levels are significantly higher in the centre of the field (10 m) than in the grip, suggesting that the grip is facilitating the drainage of the site as discussed above. The water levels at 5 m from the ditch also reflect this situation, sitting between those at 10 m and in the ditch, confirming this to be the case. Evidence to further suggest that the lateral
transfer of water may be observed is the timing of reactions to the pumping event on the 15\textsuperscript{th} June, where the pressure transducer 1.1 m from the grip responds 4 hours after the rise in grip water level, whereas the time lag is 14 hours 10 m from the grip. This continuum of increasing response times as distance from the water source at the grips increases, implies that water is indeed transferring through the soil, despite the assumed low hydraulic conductivities.

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{water_level_data}
\caption{Water level data from pressure transducers from 30\textsuperscript{th} May to 5\textsuperscript{th} July 2006; horizontal lines represent ground surface levels.}
\end{figure}

By 8\textsuperscript{th} June, ditch water levels fall below the bed of the grip (water levels are measured below bed level although labelled 'grip water level'), and so irrigation stops and drainage begins as indicated by relative water levels between grip and
field. Grip water levels then quickly drop to be level with those in the field centre and continue to fall. A diurnal regime is established here, yet begins to weaken beneath the field itself, as water levels have reached the effective extinction depth of 0.75 m, as discussed above. After 26th June there is minimal rainfall and water levels continue to drop. The ditch water level and that 1.1 m from the ditch centre follow each other closely with the 10 m level a few cm higher, possibly due in part to the difference in ground surface levels. This suggests that even at this depth in a heavy clay soil, water may still be transferring towards the ditch and lost through the shorter path to the atmosphere created by the lower grip bed level.

The utility of logging instruments such as the pressure transducers described above is shown forthrightly Figures 2.62 and 2.63 above, but when put into the context of conventional wetland hydrological sample rates, the differences are clearly evident. Figure 2.64 shows the increase in information gleaned from an increasing sample rate for wetland hydrological monitoring. If only visited monthly (Figure 2.64a) with traditional dipwell technology, a common practice across the UK at important wetland sites, one may only notice that the water levels have dropped from late May to early July 2006. Few wetland monitoring programmes have the resources to sample dipwells on a weekly basis, and Figure 2.64b reveals that perhaps there would be little benefit in doing so, as again, only broad trends in water levels are shown. A daily sample rate (Figure 2.64c) begins to reveal the subtle changes on site, but only with an hourly sampling frequency (Figure 2.64d) can the true intricacies of the hydrological
response to rainfall (and pumping) events be seen. The data are also much more useful for further examination of hydrological functioning. Frequent sampling is becoming more accessible through the use of telemetry, although this requires initial expensive capital investment.

Figure 2.64 - Comparing wetland hydrological sample rates
2.3.1.6 - Using Pressure Transducer Data to Estimate Evaporation

It is possible to evaluate evaporative loss from a wetland system through interrogation of the diurnal water regime curve. Gilman (1994) developed Equation 2.17 from White (1932), for use in peat soils in southwest England. Hays (2003) estimated water loss from Saltcedar trees (*Tamarix* sp.) which had established a similar regime in the USA, using Equation 2.18, a method which has been tested extensively by Frahm (2007) in a wetland environment (groundwater-fed fen) in northern Germany. This method differs from that of Gilman (1994) by having a variable time component which is adjusted to the maximum and minimum values, and so the period of recharge is not fixed at 4 hours.

\[
E_s = \frac{S}{100} (24r + s)
\]

Equation 2.17

\[
E_f = \frac{S}{100} \left( (H_1 - L) + \frac{(H_2 - L)T_2}{T_1} \right)
\]

Equation 2.18
Where:

\( E_g \) is the transpiration loss from the groundwater body over the 24 hour period from midnight to midnight (as developed by Gilman (1994));

\( E_f \) is the transpiration loss from the groundwater body over the 24 hour period from midnight to midnight (as used by Frahm (2007));

\( S \) is the specific yield of the soil, expressed as percentage; %

\( r \) is the hourly rate of recharge between the hours of midnight and 04:00; mmh\(^{-1}\)

\( s \) is the net fall in water table over the 24 hour period; mm

\( H_1 \) is the maximum water level during the target day; mm

\( H_2 \) is the maximum water level during the subsequent day;

\( L \) is the minimum water level during the target day; mm

\( T_2 \) is the time between \( H_2 \) and \( L \); h

\( T_1 \) is the time between \( H_1 \) and \( L \); h
Specific yield is the ratio of the volume of water that drains from a saturated soil owing to the attraction of gravity to the total volume of the soil (Fetter, 1994), and is often expressed as a percentage. A value of 3% was used for specific yield, corresponding to typical values for clay soils stated in the literature (e.g. Fetter, 1994). Both methods assume that the recovery experienced during the hours of darkness (when plants are inactive and transpiration may be assumed to be zero) is constant throughout the day, and therefore the loss during active hours is in addition to this.

The evaporation calculated using this approach for the period 31st May to 19th June 2006 are plotted in Figure 2.65 and compared with the Solent technique. It can be seen that evaporation estimates derived from the $E_s$ (Equation 2.17) method follow broadly those obtained through the Solent. It was found that this method was only successful on days of zero precipitation. Errors were large on days when rain was recorded (e.g. 13th June), and when the diurnal water level fluctuation was smaller, such as between days 10th and 12th June.

The $E_f$ (Equation 2.18) method follows observed evaporation more closely during the period analysed, and responds more appropriately during times of rainfall. One notable exception is the final day of the series showing an unexpected increase, due to a poorly defined fluctuation.
A scatter plot of the same data (Figure 2.66) reveals that the models generally underestimate evaporative water loss from the wetland system, with an $r^2$ fit of only 0.42 and 0.11 for $E_g$ and $E_f$ respectively, showing a poor statistical correlation between modelled and observed.

The values for evaporation calculated through the interrogation of the diurnal water table fluctuation rely heavily on the value used for the specific yield, $S$. Above a typical value of 3% has been taken from the literature, but a sensitivity
analysis on this principal parameter reveals that even a small change in $S$ will

Figure 2.66 – Scatter plot of evaporation data from modelling $(E_g$ and $E_f)$ and actual (Solent) methods

have a marked effect upon modelled water loss. Figure 2.67 shows the results of model runs for the $E_g$ method with $S$ values of 2, 4 and 8%, which show consistent changes of -33%, +33% and +167% respectively, reflecting directly the percentage changes in $S$ during sensitivity analysis runs. The relatively good fit of the model, albeit visually rather than statistically, clearly relies on accurate estimation of $S$, and the value chosen from the literature, 3%, was most appropriate.
The methods demonstrated here of modelling evaporation through the interrogation of a diurnal water table fluctuation have shown promise of utility. When compared to actual evaporation as measure with the Solent technique, both $E_g$ and $E_f$ methods give a good broad agreement with observed, if not reflecting the sensitive nuances of the surface micrometeorological conditions. Although the $E_f$ has a poorer statistical fit than $E_g$, it has a more consistent response to the prevailing conditions, and so draws more confidence from the user. Both methods rely heavily on the value chosen for specific yield, which is very low in a clay soil such as that seen at Otmoor. In a more porous soil profile,
the sensitivity may be lost somewhat, leading to an increase in both the accuracy and usefulness of the technique, as demonstrated by Frahm (2007).

2.3.1.7 – Soil Moisture – SCIP

For dates when both 5 cm and 10 cm data were collected, a scatter plot (Figure 2.68) reveals that soil moisture is higher at the greater depth of measurement: the vast majority of points fall above a 1:1 fit line. This is likely to be due to water loss at the surface from evaporation, leading to a drying out of the surface layers.

![Scatter plot of 5 and 10 cm SCIP data](image)

Figure 2.68 – Scatter plot of 5 and 10 cm SCIP data
Figure 2.69 shows SCIP and dipwell data from one monitoring station, A1. This again confirms that soil moisture nearer the surface (5 cm) is consistently lower than slightly deeper into the soil profile (10 cm). Also important from Figure 2.69 is the good agreement of SCIP readings with dipwell readings through the seasonal change: during the wetter winter months, soil moisture levels are generally higher.

Figure 2.69 - SCIP (a) and dipwell (b) data for monitoring station A1
The relationship between dipwell and SCIP data is explored further in Figure 2.70, where the two datasets are plotted against each other with data from monitoring station A1 highlighted (red). A clear relationship is evident, whereby soil moisture decreases with decreasing soil water levels as expected. A quadratic regression line (incorporating all data: \( y = 30.42x - 0.26x^2 - 889.95 \)) demonstrates clearly how the soil moisture levels off with high water levels (despite only having a poor statistical fit). There is a large variation in soil moisture as the SCIP instrument detects water adsorbed to the clay particles and so is available neither to plants nor for drainage.

Figure 2.70 – Water level (dipwell) plotted against soil moisture (10 cm) for all monitoring stations with station A1 highlighted; quadratic regression for all data shown
Despite this good relationship, during the summer months soil moisture levels do not drop as dramatically as water levels. The heavy clay nature of the soil at Otmoor includes water that is adsorbed to cations within the soil matrix (as described in Section 2.2.2.7); this water is unavailable and not lost even when the clay is cracked and apparently completely dry. The SCIP technique will continue to register this moisture, and so soil moisture levels recorded at Otmoor are never likely to drop below about 0.2 m$^3$m$^{-3}$. Within this observational range of soil moisture (0.2 to 0.9 m$^3$m$^{-3}$), there is a high variability of soil moisture levels between observations, especially during summer months. This suggests that soil moisture responds quickly to prevailing meteorological conditions, such that even a small rainfall event will lead to significant increases in soil moisture levels. Similarly a short dry spell will lead to much lower soil moisture levels, especially in the upper parts of the soil profile. Figure 2.69 suggests that the localised water table will respond with a longer timeframe, reflecting more the seasonal differences between rainfall and precipitation.

Soil moisture data, although revealing, are perhaps less important than otherwise might be because of the low specific yield of the soil. As so little of the soil is taken up by mobile water, any changes in soil moisture has a small volumetric effect, and as such is not important in terms of wetland hydrological functioning.
2.3.1.8. – Guelph Permeameter

Results from three Guelph permeameter experiments were averaged to give a hydraulic conductivity ($K$) of $1.28 \times 10^{-4} \text{ cms}^{-1}$, with a standard deviation of $6.7 \times 10^{-5} \text{ cms}^{-1}$. This value is high relative to values given in the literature (such as Fetter, 1994) for clay soils, in the range of $10^{-9}$ to $10^{-6} \text{ cms}^{-1}$. However, it is low relative to other soil types which have a lower clay content and are more free-draining.

The relatively low hydraulic conductivity of the soil type at Otmoor is what prevents the free-draining of the soil, as shown in the stageboard, dipwell and pressure transducer data discussed in previous sections of this chapter. However, vertical and lateral fluxes were far from zero, when the $K$ values from the literature would suggest fluxes approaching zero. It must be noted that point measurements of $K$ must be taken within the context of associated scale issues. As previously noted, Bromley et al. (2004) concluded that point measurements (such as those taken by the Guelph permeameter) are often required to be increased by up to two orders of magnitude in order to be representative of the wider site, as preferential flow pathways (such as soil cracking in clay soils) are likely to be unrepresented: no evidence of such features was apparent on the sample date. With this in mind, it is possible that $K$ could be as high as $0.01 \text{ cms}^{-1}$. This would explain the evidence of lateral and vertical water fluxes demonstrated in this chapter, especially using pressure transducers.
2.3.1.9 – Soil Profile Analysis

An electric percussion drill was used to drill to a depth of 4 m in order that information on the previous aeration status might be revealed, as described in Section 2.2.2.8. The findings are summarised in Table 2.6, and combined with evidence from other sources. This includes data from boreholes held in the British Geological Survey (BGS) archives which have provided supplementary information on geology of the local area, revealing that this region has Kellaway limestone underlying approximately 37-50 m of mid Jurassic mudstones, namely Oxford Clay. The Otmoor area has alluvially reworked clay at the surface, of varying depths but typically 2-3 m (Hughes, pers. comm.). Furthermore, a drilling test was undertaken on behalf of the RSPB in 1997, to depths similar to this drill (~4 m). This test found almost identical results to the drilling undertaken here (Lambert, pers. comm.), with variations between layer depths of up to 15 cm. As only one core was drilled for the current study, this provided some repetition, and confirmed that this drilling location (shown in Section 2.2.2.8) was indeed representative of the Otmoor site in general.

The upper 0.66 m (Layer A) is based on the alluvial clay, but contains organics mixed through plant activity and agricultural practice in recent years. Dense root networks can be seen. Layer B (0.66 to 1.50 m) contains mottled areas of both grey-blue and brown. As this layer is saturated for much of the year, oxidation is limited in most areas. However, plant roots and cracking during the summer
<table>
<thead>
<tr>
<th>Horizon</th>
<th>Max. depth</th>
<th>Colour</th>
<th>Carbonate presence?</th>
<th>Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.66</td>
<td>Brown, mixed</td>
<td>No</td>
<td>Upper layer with dense root networks; much organic material</td>
</tr>
<tr>
<td>B</td>
<td>1.50</td>
<td>Brown and blue</td>
<td>Yes</td>
<td>Mixture of grey-blue and brown clays. Browns indicate aeration from macropores and plant root channels at times of low water levels</td>
</tr>
<tr>
<td>C</td>
<td>2.03</td>
<td>Blue</td>
<td>Yes</td>
<td>Heavy clay with little signs of aeration. Bottom of this layer is maximum extent of localised (or perched) water table</td>
</tr>
<tr>
<td>D</td>
<td>2.78</td>
<td>Dark brown</td>
<td>Decreasing</td>
<td>Oxidation dominated. Historically dry layer, below localised water table</td>
</tr>
<tr>
<td>E</td>
<td>37-50</td>
<td>Blue</td>
<td>No</td>
<td>Start of <em>bona fide</em> marine clay. Not saturation, but lack of air due to depth prevents oxidation and so blue colour persists</td>
</tr>
<tr>
<td>F</td>
<td>?</td>
<td>N/A</td>
<td>N/A</td>
<td>Kellaway limestone</td>
</tr>
</tbody>
</table>

Table 2.6 – The soil profile at Otmoor

months allow oxygen-rich air to reach these layers, causing the mottled brown effect. Carbonate nodules can be seen to have precipitated at this level; it is
thought that during the formation of the alluvial deposits, floodwater would have deposited carbonates, which precipitates as nodules over time. The abrupt transition between Layers A and B is highlighted in Figure 2.71 with the red arrow, with A to the left and B to the right. Evidence of the carbonate precipitates can be seen to the right of the sample on the bottom edge of the sample chamber opening (indicated by blue marker). Fine roots were found in the sample to a maximum depth of 0.85 m (into layer B), but not reproduced well photographically.

Layer C (1.50 to 2.03 m) is blue, indicating a reduced state and lack of air. This is a generally saturated layer, too deep for plant roots and macropores to aerate. The bottom of this layer indicates the lower extent of the localised or perched
water table. Again, this layer shows carbonate nodules indicating an alluvial reworking of the clay. The transition between layers B and C is more gradual than that between A and B, as shown in Figure 2.72, which also demonstrates the palpable differences between the brown colours associated with oxidation of the clays (layer B) to the left, and the strong blue shading of the reduced clays (C) to the right.

Layer D (2.03 to 2.78 m) is dark brown, indicating a historically dominant role of air, and a lack of water. This has a decreasing carbonate nodule density, suggesting the start of transfer from alluvially deposited and reworked clays to the deeper marine clay. The upper extent of layer D is believed to be the maximum depth of the localised water table, as there is evidence of persistent
aeration here. The water table is known to be local due to the clay nature of the Jurassic mudstone substrate preventing access to any deeper aquifer. Layer E (2.78 to 37+ m) is a layer of increasingly blue clays, without carbonate precipitates. This blue colour originates not from water excluding the access of air, but of the sheer depth and distance from the atmosphere precluding oxygen from being present, and this layer is mid-Jurassic mudstone in origin, referred to commonly as Oxford Clay. The transition between layers D and E is shown in Figure 2.73, with the obvious brown colouration of the aeration-dominated layer D to the left, and the start of the limit of aeration manifesting as the increasingly blue colour to the right. This marine clay layer (E) is believed to be up to 50 m deep, and is underlain by Kellaway limestone to unknown depths (Hughes, pers. comm.).
2.3.1.10 – Vegetation Composition

An analysis of the vegetation structure has been undertaken as part of the Grips research project, involving determination of species present and those which were dominant (Acreman et al., 2008). The grassland community was dominated by the grass *Agrostis stolonifera*, with up to 1/3 coverage and the rush *Juncus articulates*, again with approximately 1/3 cover. Minor species included other rushes (*J. conglomerates* and *J. effuses*), buttercup (*Ranunculus repens*) and the Meadow Fescue (*Festuca pratensis*). This combination of species was assessed as closest to the OV28a community (*Agrostis stolonifera-Ranunculus repens*, open inundation pasture) under the National Vegetation Classification system (Rodwell, 1992) in 2004 at the start of the Grips research project (Dueñas, pers. comm.). Although a branch of the open habitat category, this community is similar in sward structure to communities more commonly associated with wetland environments, those of the mesotrophic grasslands MG10 (*Holcus lanatus-Juncus effuses*, ill-drained rush pasture) and MG13 (*Agrostis stolonifera-Alopecurus geniculatus*, inundation grassland). Both of these communities are associated with periodic waterlogging and surface inundation, and so composed primarily of hydrophilic plants. A more recent survey in summer 2006 revealed that the change in management over recent years at Otmoor has had an impact on community structure, undergoing succession towards more typical wet grassland vegetation of the MG10 community (Dueñas, pers. comm.), most likely due to the consistently high water levels creating unsuitable conditions for OV28a species and allowing succession of MG10.
2.3.2 – Multi-Scale Model

Table 2.5 is repeated in order that the differences between the various model scales are clear.

<table>
<thead>
<tr>
<th>Model</th>
<th>Meteorological Output</th>
<th>Meteorological Input</th>
<th>Detail</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>GSWP2</td>
<td>1 degree grid cell</td>
<td>Global scale model run</td>
</tr>
<tr>
<td>2</td>
<td>Otmoor AWS</td>
<td>1 degree grid cell</td>
<td>Global scale but with local meteorological input</td>
</tr>
<tr>
<td>3a</td>
<td>Otmoor AWS</td>
<td>Point</td>
<td>Local scale with measured soil parameters</td>
</tr>
<tr>
<td>3b</td>
<td>Otmoor AWS</td>
<td>Point</td>
<td>Local scale with varied soil parameters</td>
</tr>
<tr>
<td>4</td>
<td>Solent</td>
<td>Point</td>
<td>Taken as accurate measurement of evapotranspiration</td>
</tr>
</tbody>
</table>

Table 2.5 – Details of scale levels in the multi-scale model of evaporation

Daily averages over ten years for scale 1 model results are shown in Figure 2.74, and show a clear seasonal trend through the year peaking at 3.55 mm day\(^{-1}\) on 13/06/2006, a time when evapotranspiration would be expected to peak. The majority of the total evapotranspiration is composed of transpiration from vegetation in the summer months, when vegetative activity is highest (fraction shown on Figure 2.74 as the broken line; second y axis), comprising up to 75%
of total evapotranspiration. Outside of the growing season transpiration drops considerably, forcing total evapotranspiration rates to fall. Evaporation from soil increases through the summer months, but not to the extent of transpiration. Evaporation from interception also increases as vegetative cover increases and interception of rainfall increases accordingly.

When compared to measured total evapotranspiration from the Solent on site at Otmoor (Figure 2.75), modelled total evapotranspiration from the scale 1 model fitted the data well despite the temporally different meteorological data used to run the model. The maximum daily value of 3.55 mmday$^{-1}$ was broadly comparable to 4.80 mmday$^{-1}$ as measured by the Solent. The 10 year-averaged global scale model shows less variance and more stability, reflecting the averaged nature of the data. An example year, 1995 (light green, Figure 2.75) has been added, and shows that the model does represent high daily variability. This corresponds to variability shown by the Solent method, where small scale changes in local conditions such as windspeed and temperature which will force total evapotranspiration to vary significantly between days. The relative robustness is demonstrated by the close annual means of 1.56 and 1.81 mmday$^{-1}$ for modelled and actual evapotranspiration respectively. The lower mean of 1.47 mmday$^{-1}$ for the 1995 data is likely a facet of the year's conditions. Overall, model performance was of a very high quality when considering the difference in scale of the model code development and foreseen application, as well as meteorological data drivers.
Figure 2.74- Components of the 10 year daily averages of the scale 1 model of Otmoor evapotranspiration (ET)
Figure 2.75 – Comparing Scale 1 modelled evapotranspiration daily averages with observed (Solent)
Hourly output from the scale 1 model run shows a similar high quality, even when compared to 2006 actual measurements of evapotranspiration, and again, both the 10 year average (dark green) and 1995 (light green) data are shown (Figure 2.76). Notably, modelled total evapotranspiration very rarely falls below zero, whereas the data from the Solent show this to be common at Otmoor in the summer as the area is very exposed and susceptible to sharp temperature drops and so condensation at night. JULES is not capable of modelling dew formation, hence its absence from modelled data (Ellis, pers. comm.). In essence this is a facet of the global scale nature of the model, which is not designed to take account of such subtle processes at the field scale. The maximum values for the data points (shown in broken horizontal lines of respective colours) reveal important features of the data. The maximum observed data is high (0.597 mmhour\(^{-1}\)) reflecting high local variability. The 1995 maximum is lower (0.47 mmhour\(^{-1}\)), showing that although a different year was being modelled and so is not directly comparable, the global scale model might not be able to reflect these local small scale variations. The 10 year average has a much lower maximum (0.36 mmhour\(^{-1}\)), as would be expected.
The results from the scale 2 model have the advantage over the scale 1 model of being modelled from 2006 data from the AWS sited at Otmoor, and component...
results from the scale 2 model are shown in Figure 2.77. Total evapotranspiration is estimated to be very high, up to 13.56 mm day\(^{-1}\). The majority of this is from the transpiration component, with the interception and soil components having a relatively minor role. This suggests that the Otmoor AWS output is not representing the atmospheric system well for the 1 degree grid square for which the model is being run. Loss from the interception and soil components have maxima of 3.84 and 2.85 mm day\(^{-1}\) respectively, which are close to total actual daily maxima whilst forming a much smaller fraction of total evapotranspiration. As there are no observed measurements of these components at Otmoor, and as it seems intuitive that transpiration forms the bulk of water loss, it must be assumed here that the component ratios are close to accurate, and that water loss from each component in the scale 2 model seems to be overestimated by a similar magnitude.

Figure 2.78 shows the daily output from the scale 2 model with observed evapotranspiration data (from the Solent; Figure 2.78a), together with estimated soil moisture and observed precipitation (Figure 2.78b). This reveals that the soil moisture component of model output is very responsive to rainfall, both wetting quickly after a rainfall event and drying quickly during dry periods. This suggests that the soil parameters within the model are overly transmissive, allowing a conductive connection between the soil water and the atmosphere, which would explain the higher modelled evaporation rates, which can be seen clearly against observed evapotranspiration in 2.78a.
Figure 2.77 – Daily totals of scale 2 model components

When looking at an hourly resolution (Figure 2.79), the scale 2 model shows consistent overestimation of evapotranspiration throughout the day. As with the scale 1 model, condensation was not modelled, as there is no dew function within the model (Ellis, pers. comm.). It is also clear that the lack of dew is not the cause of the overestimation of daily values: daytime hourly peaks remain consistently higher than observed. It is clear that the combination of Otmoor AWS data within a global scale model has not been successful.
a - Evapotranspiration

Figure 2.78 - Scale 2 model daily output compared to observed (Solent; a) with modelled soil moisture and observed precipitation (both b)
Figure 2.79 - Scale 2 model hourly evapotranspiration output compared with observed (Solent)
The scale 3 models (a and b) have soil parameters more closely related to the observed situation at Otmoor, and so is more directly comparable with actual measurements on site. The daily totals (Figure 2.80) again show that transpiration forms the bulk of total evapotranspiration loss. Results from this scale of model run suggests more appropriate water loss from the system than the scale 2 model, with up to 12 mmday$^{-1}$ of total evapotranspiration loss. Although more than double the observed maximum, the evapotranspiration levels are consistently lower than those modelled in scale 2.

Figure 2.80 – Daily totals of scale 3a model components

Figure 2.81 demonstrates that the scale 3a model has a consistently lower total evaporation output when compared to scale 2, suggesting that the former is closer to actual levels. When compared to actual evapotranspiration
measurements Solent method (Figure 2.82), this overestimation of total evapotranspiration by the scale 3a model can be clearly seen. Water loss is overestimated significantly during wet periods, suggesting the error is an underestimation of rainwater routing through infiltration, possibly due to the very low hydraulic conductivity values assumed from the clay soil selected in the model. Soil moisture during these periods can be seen to be close to saturation, and so water will be freely available at the surface for interaction with the atmosphere. The scale 3a model could be said to be a poor representation of local evapotranspiration at Otmoor, with neither patterns of high and low evapotranspiration reflecting meteorological conditions, nor the magnitude of results representing actual measurements at a small scale when rainfall occurs.

Figure 2.81 – Scale 2 and scale 3a total evapotranspiration model outputs
Figure 2.82 – Scale 3a model daily output compared to observed (Solent; a) with modelled soil moisture and observed precipitation (both b)
For model 3b, the parameters of the model are changed so that a more silty soil is represented, in contrast to the very heavy clay structure present at Otmoor and the scale 3a model. The output (Figure 2.83) changes somewhat, and evaporative loss is increased due to the higher conductivity of the soil and so more free connection between the water table and the atmosphere. It might be expected that a silty soil would be much more amenable to free drainage and so dry out further; the former appears more important than the latter. It is during wetter periods that the two models 3a and 3b are more congruent (although not closer to observed values), suggesting that the differences are caused by access to water when it is deeper in the soil profile during drier spells. This is logical, as with a more transmissive soil (3b), water will be more readily available and so evapotranspiration levels will be higher. This is confirmed by soil moisture output from models 3a and 3b, with levels appearing similar during wetter periods when modelled evapotranspiration is also more agreeable between the two models. However, soil moisture for model 3b never fell below 0.5 m$^3$m$^{-3}$, with only a small difference from the 3a model with clay soils. Overall though, it may therefore be concluded that the model is sensitive to soil parameters, and so calibration to the site would be possible, although this is not the objective of the current work.
Figure 2.83 – Comparing scale 3a and 3b model daily outputs with evapotranspiration (a), soil moisture and precipitation (both b)
In conclusion, the scale 1 model shows remarkable correlation with observed values at Otmoor, especially considering model runs were not undertaken for 2006 due to restrictions of observed data availability. This suggests that even a model designed and applied at a global scale can produce results directly comparable with observed values taken at the field scale, when all model parameters are of a similar scale. With levels of accuracy demonstrated here, such a model could be realistically used to predict approximate values for evaporation at wetlands sites, for example in the development of conceptual water balances in the early stages of study at a field site.

Using observed meteorological data (scale 2 model) changed the output of the model, with a dramatic decrease in accuracy compared to observed values. It is thought that the observed meteorological parameters do not work well with other parameters in the model, including LAI and soil properties, which were taken from global scale datasets. Significantly, the predictive power of the model increases when these parameters were changed to be more congruent with the field scale observed AWS data. Although these showed a poorer fit than the scale 1 model, the scale 3 model showed greater utility that that of scale 2. It is inferred that this was because more parameters within the model were of a similar scale.

It may therefore be concluded that results deteriorate not only as the model is forced to predict at a scale for which it was not designed, but as parameters within the model demonstrate increasing disparity of scale. Using mixed scales of parameters yielded the poorest results, even when compared to
comprehensively global scale (scale 1) and predominantly field scale parameters (scale 3). This is a surprising result and one that both demonstrates the utility of the global scale model and the precautions and caveats needed when applying a model at unfamiliar scales.

It has also become clear that several interrelated factors determine the amount of evapotranspiration modelled, including soil properties and subsequent soil moisture levels. Therefore changes in model structure, such as the implementation of a more silty soil in model scale 3b, may have a large impact on model results.

The application of the multi-scale model of evapotranspiration has revealed many important points about hydrological modelling and scale issues: it is important to use a model at the scale for which the code was designed and with appropriate driving data. Model code should be chosen carefully and with the application very much in mind, requiring careful planning of the project and all modelling outputs that may be needed in later phases.
An intensive monitoring regime exploring the hydrology of Otmoor has been undertaken. Significant progress towards increasing the understanding of the dynamic hydrological regime and subsequent hydrological functioning of the site has been demonstrated. There was no evidence of significant runoff from the site during times outside very high water levels. This, coupled with a lack of groundwater interaction, indicates a hydrological system dominated by precipitation and evapotranspiration as the primary source and outlet for water respectively.

Recent micro-management has included the installation of shallow surface ditches, or grips, across the immediate research plot, and these have been shown to irrigate during and after rainfall (or pumping) events when they contain sufficient water. In contrast, when the grips have dried out in the summer, they act as a conduit for further water loss. This may be due to the ditches decreasing the distance from the surface to the water table, allowing easier access to the atmosphere for soil water. Significant lateral flow has been observed in certain dipwell cross sections, but not consistently, indicating heterogeneity of soil conditions across the site.
Evaporation has been estimated from meteorological data, measured directly via the eddy correlation-energy budget method, and modelled from daily water loss measurements. All methods compare well and show high evaporation levels, occasionally over 5 mm\textpermm, resulting in a significant loss of water from the wetland system in summer months, and subsequently to a sharp decline in water levels across the research plot to as deep as 1.5 m below ground.

There was a large range of water levels in the years of observation, with summer levels generally 1.3-1.5 m below winter levels, which were generally at or above the surface. This has important implications for flood storage capacities, and therefore issues of site management, as the flood storage potential of the wetland is only available in summer months, the time of year it is least likely to be required. During winter this storage is not available, and so water will be transferred via runoff directly to the river network. The recent intense water level management, directed at attracting specific species of rare breeding waders, may be keeping water levels higher for longer over winter and spring, but cannot prevent the sharp declines in water level over the summer months. The change in water level regime has been reflected in the vegetation community, which has shown evidence of succession towards a typical MG10 wet grassland over recent years (Dueñas, pers. comm.).

Soil moisture broadly follows water levels, but its measurement through dielectric means is limited by the high adsorbed water content present in the clay soils. However, soil moisture is not a major store at Otmoor, as the specific yield is very low. This also has implications for flood storage at Otmoor, as water levels
can be kept near surface level without a significant loss of storage, and so giving Otmoor the potential to be managed with both conservation and flood storage in mind.

Evidence from drilling has shown the observed maximum depth of the localised water table to be little more than 2 m. Furthermore, signs of intermittent aeration suggest that the observed contemporary regime of a seasonal water table depth fluctuation (the level of the top of this localised water table) between near the surface and 1.5 m below is representative of the historic situation.

The high frequency sample rate of logging instruments has increased the understanding of site hydrology significantly from that available via conventional manual site visits alone. The pressure transducers placed on site have revealed the presence of a diurnal water table fluctuation during the summer, driven by high water demand from evapotranspiration. This phenomenon was unexpected due to the high clay content of the soil profile at the site and inferred low hydraulic conductivity, which would prohibit such high vertical water fluxes; it is suggested that macropore flow facilitated by cracks in the soil (to a depth of 30 cm or more) allows connection with the atmosphere to a higher depth. It is suggested that the low specific yield in the soil could exacerbate this effect, as a small change in water volume brings about a larger change in water table elevation. Even so, water transfers appear greater than those suggested by the point measurement of hydraulic conductivity using a Guelph permeameter of $1.28 \times 10^{-4} \text{ cms}^{-1}$. Interrogation of the diurnal fluctuation has allowed prediction of evaporative loss with good results on dry days with a clear fluctuation, and
may be used as a substitute where no observed evaporative loss is available, but is highly sensitive to the value used for specific yield.

Although dependent upon heterogenic soil conditions, the pressure transducers have also shown a significant lateral flow of water between the surface drain and field centre. When water levels were high, the grips acted to irrigate the fields, but drained water from the fields when water levels dropped below those in the field centres. The grip beds appeared to act as a conduit for water loss from the soil system during times of low water levels in summer months, and it is thought that the shorter distance between water table and atmosphere facilitated a less resistive path, drawing water from the field centre.

The elucidation of the diurnal water table fluctuation has implications for scale issues within the context of the research hypotheses. The pressure transducers were only set to record at such high temporal resolution to complete the characteristic velocity associated with the small spatial scale of the plot scale monitoring programme. Despite this, they have proven that the observation scale can be brought closer to the process scale with the removal of assumptions and the application of appropriate technology.

The water transfer mechanisms referred to in Section 1.1.4 have been identified at Otmoor as being dominated by precipitation and evaporation. Floodwater is
known to inundate Otmoor on occasion, with the amount dependent upon the size of the flood event. Management of the site by the RSPB involves pumping water onto the site when water levels are below target levels and water resources are available. Shallow groundwater (localised water table, up to 2 m below the surface) storage has been shown to be important, limited by the low specific yield, and is only available in summer when water levels are low. The associated subsurface storage and therefore transfer is minimal due to soil properties. Figure 2.84 shows the water transfer mechanisms relevant to the Otmoor wetland system; deeper, regional groundwater interaction is absent. Runoff has also been demonstrated to be minimal, although floods of a long return period have not been observed during the monitoring period.

![Figure 2.84 - Water transfer mechanisms present at Otmoor](image)

The scale of hydrological regime is varied, ranging from a water table fluctuation of up to 10 cm at the daily scale to a seasonal variation of 1.5 m. The combination of transfer mechanisms and regime dictate that the hydrological functions present at Otmoor are surface flood storage and the exacerbation of
evaporative water loss. The effect of each function has been assessed at only the plot scale, and is as yet unknown at the catchment scale.

Scale issues have been further investigated through a multi-scale model of evapotranspiration at Otmoor. It has been demonstrated that the model in question (JULES) provided most accurate output when a) it was used at the scale for which it was designed and b) the scales of input data were commensurate. This has important implications for modelling, especially in wetland environments.
Chapter 3

Catchment Scale Assessment of Wetland Hydrology
3.1 – Introduction

Chapter 2 described the development of understanding of a wetland site at the field scale. Aspects of the site’s hydrology were revealed including the above ground and subsurface storage, evaporation, and fine scale hydrological dynamics. This work made no inferences about what effect these factors might have upon the wider context of the wetland within the catchment. This chapter will assess the impact of the same wetland site, Otmoor, on river flows within the catchment in which Otmoor lies, that of the River Cherwell. This will be done with the use of numerical models of increasing complexity. Acreman (2004) advocates a gradual development of conceptual understanding as information on the wetland site is elucidated, and this approach will be adopted with the modelling procedure.

A primary conclusion from Chapter 2 was the importance of the Otmoor wetland as a potential store of flood water. In light of this, a simple conceptual model will calculate the flood storage capacity, and relate this to a standard measure of flood magnitude seen in the catchment. This model will be referred to as ‘level one model’, and will enable an assessment of the relative importance of the wetland to wider catchment hydrology in simple terms, this work is described in Section 3.3.
Secondly, a more complex model will be developed. This will take the form of a hydraulic model incorporating a rainfall-runoff component, allowing time-varying analysis of the hydrological effect of the wetland during flood events; this development will be described in Section 3.4. Information obtained from the level one model will be used to inform the priorities for development of this ‘level two model’, and so it is conceptualised as an evolution of the more simple, volume-based level one model. The level two model, once developed, will be queried to assist the understanding of the Otmoor wetland system within the context of catchment scale hydrology.

In order to place the two models in the context of general hydrological modelling, a review of different types of models and model codes will be undertaken in Section 3.2. This exercise will also allow an informed choice of model code for the rainfall-runoff and hydraulic models for the level two model.
3.2 – Review of Model Types and Codes

3.2.1 – Introduction to Modelling

A numerical model is a representation of reality (Kirkby et al., 1993), and is based on mathematical equations to provide an interface enabling data input, equation solution and output (Booker, 2004). Models are routinely utilised across the environmental sciences to increase the utility of contemporary understanding, as the behaviour of the modelled system may be investigated through extrapolation beyond the boundaries of observed behaviour.

Models of wetland systems are particularly useful as wetlands are sensitive to natural and man-induced hydrological change (Gilvear et al., 1992). The delicate interdependence of the components of wetland systems (as described in Section 1.1.3), particularly the dependence of vegetation and soils on hydrological conditions (Hollis and Thompson, 1998, Wassen et al., 1996), is key to the distinct nature of wetlands, and many models attempt to predict the changing water level in a wetland caused by changes in surrounding areas (Wassen et al., 1990). Examples include changes in groundwater levels (e.g. Gilvear et al., 1993), river flow (e.g. Mauchamp et al., 2002), land management (e.g. Acreman et al., 2007) and climate change (e.g. Schulze, 2000).
It should be noted that there is no proven composite model for predicting the transfers through a wetland system and into the river channel. Bates *et al.* (1996, p244) sum this up well, stating what exactly is required to resolve the greatest constraint on combined modelling:

"Currently the ability to fully model coupled hydraulic and hydrological processes during flood events does not exist, yet this is an essential prerequisite to a more complete understanding of many catchment processes."

Wetland modelling frequently focuses on flood events, as they are important in the management of wetland systems, as wetlands can provide significant flood storage potential (Acreman *et al.*, 2007). However, the management of wetlands and their bearing on the river catchment system is equally important during times of low flow, as it is during more 'normal' flow. Knight and Shiono (1996) conclude that the river channel and floodplain (including wetlands) elements should be treated as one system, although with significant changes in parameter values at bank full discharge. The basis for this inference is that rivers do not simply 'burst their banks', but the floodplain is utilised periodically as the discharge becomes too high for the main channel to contain, and as such channel geometry complexity increases significantly (Knight and Shiono, 1996). This natural flow characteristic should not be separated just because it occurs infrequently, as flooding is an integral element of the action of the river system and its associated drive towards dynamic equilibrium (Knighton, 1998).

Contemporary hydrological science is often unable to effectively model wetland systems, partly due to the late start of wetland hydrological science relative to other fields, and partly due to the inherent complexity of wetland systems, which
as Chapter 2 has highlighted, may involve more facets of the hydrological cycle than any other. The former may be due to an historical lack of appreciation for wetlands and their benefits and values, but today's climate of change and associated increased flood risk (summer 2007, for example) might put wetlands to the forefront of land use philosophy and drive a focussed need for development of comprehensive wetland models.

More recent developments in the coupling of hydrological and hydraulic modelling components has shown utility in enabling a more comprehensive and so more powerful wetland modelling environment. An example of this work is that of Thompson (2004) and Thompson et al. (2004), whereby the MIKE-SHE (distributed hydrological: Graham and Butts, 2005) and MIKE-11 (in-channel hydraulic: Havnø et al., 1995) models were successfully applied to wet grasslands in southern England.

The current work cannot utilise a combined model, as this application requires catchment perspective which cannot be effectively provided by any contemporary model code that also allows field scale analysis. Although in theory a distributed field scale hydrological model could be extrapolated, the problems associated with scaling highlighted in Chapter 1 are to be avoided where possible. A hydraulic model is required, coupled with an effective rainfall-runoff model in order to provide robust flow boundary conditions. As such, rainfall-runoff models will be summarised, followed by hydraulic models. In-field hydrological models (including sub-surface dynamics) are often utilised for wetland hydrological
modelling. Although not used in the current work, such models will be detailed as many seminal wetland studies have used in-field hydrological models.

3.2.2 – Rainfall-Runoff Models

A model which predicts river flow from a measure of rainfall is termed a rainfall-runoff model, and is essentially a mathematical representation of the link between the meteorology and hydrology of the study area (Knight, 2006). The development of rainfall-runoff models has largely been driven by the requirement for streamflow data for the ungauged catchment (Young et al., 2006b). Several groups of rainfall-runoff models have evolved, with the primary difference being the complexity with which differences in physical characteristics across the catchment are portrayed. A lumped model takes just one value for each parameter for the entire model extent, whereas a fully distributed model gives spatially-varying account of each parameter at a spatial resolution determined by the model code. Lying between these lumped and fully distributed models sit semi-distributed models, which may have a limited degree of spatial variability for one or more parameters.

Young et al. (2006a) discuss the merits of the two extremes, with lumped models being preferred due to the smaller number of parameters needing to be identified, but with a consequence of higher perceived simulation error. Lumped models have been successfully utilised extensively for various applications in the UK for many years (NERC, 1975; Houghton-Carr, 1999; Kjeldsen, 2007; Moore, 2007), but key limitations of catchment-averaged, or lumped, rainfall-runoff
models are summarised by Young et al. (2006a). Calibrated parameters are often a function of forcing data and the calibration scheme (including measures of fit), and model parameters may therefore have little physical relevance.

Physically-based models attempt to accurately represent the physical processes occurring across the landscape. In the case of a rainfall-runoff model this would entail a model taking into account the intricate processes of rainfall interception, infiltration, through-flow and conversion of all components to river flow. In addition, data on these processes and physical descriptors would need to be spatially distributed across the catchment in question. Kokkonen (2003) highlights that even physically-based rainfall-runoff models are unsatisfactory, as measured parameters are nearly always calibrated in order to fit observed flows and so are changed from measured values. Kokkonen (2003) continues, adding that often the problem is exacerbated by the difference in scale between field measurement and model, leading to inappropriate model inputs.

Moore (1985; 1999) suggested that semi-distributed models be utilized in order to take account of some of the spatial variations of hydrological characteristics across the catchment. Moore (1999, p149), an advocate of the intermediate approach, summarises:

"...the complex hydrological response of river basins is best represented by models which represent the components of runoff production and translation in a conceptual manner. This is not to decry the utility of simple black-box models for certain applications but derives from experience gained on a number of river basins that conceptual models, when properly constructed, are able to better reproduce the nonlinear behaviour of the rainfall-runoff process."

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Alongside the distribution of physical characteristics is the complexity of the model, with basic models having few calibrated parameters, and complex models having many parameters available. Although in theory a more complex model should be a more realistic representation of reality, in practice parameter-sparse models limit interdependence between parameters which again, may lead to them having little physical relevance (Kay et al., 2006b) and limiting any value in extrapolation to other events (Bell, pers. comm.).

A wide variety of model codes are available and a selection of several rainfall-runoff modelling codes follows. Although not exhaustive, it summarises some of the established codes regularly utilised and established in contemporary hydrology. The selection of models summarised below (Table 3.1) demonstrate the flexibility of modern rainfall-runoff modelling, with a range of spatial representation techniques and applications. The variety of codes vary from the simple yet dexterous PDM (in which the number of parameters calibrated is variable) to the fully distributed and complex MIKE-SHE model code, which integrates rainfall-runoff modelling with distributed in-field hydrological modelling (see Section 3.2.4).

Many rainfall-runoff models have standalone utility, rather than existing merely to produce boundary conditions for hydraulic models (Crooks and Naden, 2007) and predict flow in the ungauged basin (Young et al., 2006b). This is exemplified by the climate change work undertaken by Kay et al. (2006a).
<table>
<thead>
<tr>
<th>Model Code</th>
<th>Details</th>
<th>Spatial Representation</th>
<th>Reference</th>
<th>Application Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grid-to-Grid (G2G)</td>
<td>Grid-based approach is an extension of PDM (see below) whereby rainfall and routing of water are estimated at all grid squares across the domain.</td>
<td>Fully distributed</td>
<td>Bell and Moore, Moore, 1998a</td>
<td>1998b; Bell et al., 2007</td>
</tr>
<tr>
<td>MIKE-SHE</td>
<td>Integrated rainfall-runoff and hydraulic routing model.</td>
<td>Fully distributed</td>
<td>Henriksen et al., 2003; Graham and Butts, 2005</td>
<td>2004</td>
</tr>
<tr>
<td>Probability Distributed Model (PDM)</td>
<td>Utilises conceptual water stores, represents non-linearity of soil moisture storage through a probability distribution, to allow for time-varying contribution to runoff.</td>
<td>Lumped</td>
<td>Moore, 1985; 1999; 2007; Kay et al., 2006b; Young et al., 2006a; 2006b</td>
<td></td>
</tr>
<tr>
<td>Time-Area Topographic Extension (TATE)</td>
<td>Based on response functions approach, with a time-area convolution calculating distance from river channel network, and topographic catchment configuration.</td>
<td>Lumped</td>
<td>Calver, 1996</td>
<td>Kay et al., 2006</td>
</tr>
</tbody>
</table>

Table 3.1 – Rainfall-runoff model codes
3.2.3 – Hydraulic Models

McCartney (2001, p4) defines a river flow model as providing the following to a resource manager:

"A quantitative expression of the time-variant interaction of various hydrological and hydraulic processes that occur within river catchments. Such models are collections of physical laws and empirical observations, written in mathematical terms and combined in such a way, as to produce a set of results based on a set of assumed inputs."

Channel flow routing models commonly derive from the St.-Venant equations for open channel flow and their simplifications, and can be discretised into hydraulic models based on flow hydrodynamics and the simpler formulations referred to as hydrological routing methods. The latter are based on simple mass balance storage, and usually link channel storage or water level to flow discharge (Moore et al., 2006b). The field of study focussing on the in-channel processes in rivers is therefore covered under the umbrella of the science of hydraulics; hereafter a hydraulic model refers to a representation of reality for the physical processes occurring in the channel of a river, stream, or man-made channel. Also included in the scope of hydraulic processes is that of floodplain surface flow during flood events, as the floodplain is essentially an extension of the channel when flooded, as discussed above.

Within a hydraulic model, the mathematics used can be complex. The velocities, levels and volumes of flow are calculated in the river, across the floodplain and
at the interaction between the two. These processes are byzantine, with many factors needing to be accounted for or averaged, including turbulent eddies, a river's response to fine scale changes in bed conditions (e.g. Wilcock and McArdell, 1993), and representation transcending multiple scales: using observations at the cross section scale to provide information to models which are used to give answers to questions at a reach or basin scale (Kirkby, 1996).

McCartney (2001) explains that the Navier-Stokes equations, concerning the conservation of mass and momentum (Knight and Shiono, 1996), balance the rate of water level rise or fall with changes in storage (Gomes Lopes et al., 2004), allowing hydraulic models to calculate the propagation of a hydrograph downstream. The continuity element of the equation expresses the conservation of mass, whereby the inputs to the system at the upstream boundary condition are equal to the outputs at the downstream boundary condition. The momentum equations represent fundamental laws of dynamics, in a form for fluids (Hervouet and van Haren, 1996), balancing the forces of inertia, diffusion, gravity and friction (Gomes Lopes et al., 2004).

The basic equations of hydrodynamics are well known and widely used in hydraulic modelling. Hervouet and van Haren (1996) summarise succinctly, stating that the only problem remaining is how to solve them! These equations are 1D non-linear hyperbolic partial differential equations (Gomes Lopes et al., 2004), and subsequently the solution requires the existence of unique and stable conditions.
The main difficulty in solving the Navier-Stokes equations comes from the fact that turbulence is a parameter in the equation, and that there are several non-linear terms present that require very high computational power. Hence the complexity of calculation increases dramatically with both increasing number of dimensions and study area (Hervouet and van Haren, 1996).

The St.-Venant (1871) approximations of the Navier-Stokes equations are widely used, as these average the 3D Navier-Stokes equations over depth. In effect this decreases the number of dimensions from three to two (e.g. Hervouet and van Haren, 1996; Knight and Shiono, 1996), and models that use this method are known as fully dynamic. There is obviously a loss of information and so predictive power when using this step, but models may be more readily applied and more practical to use. Several assumptions are made when using the St.-Venant equations (Beven and Wood, 1993):

- The water is incompressible and of constant density and viscosity;
- the streamlines are straight and in the downstream direction;
- the flow is gradually varied in both space and time, so that the amplitude of surface waves is very much smaller than the wavelength;
- the bed slope is small and the bed fixed;
- the addition of momentum associated with lateral inflows is small relative to that of the main channel.

As stated, steep slopes are not valid for expression under the St.-Venant solutions. This derives from the simplification process from the Navier-Stokes
equations, which assumes the vertical velocity of water is negligible, and this would not be the case for slopes greater than 1:10 (Hervouet and van Haren, 1996). A further limitation of the St.-Venant equations is the very fact that they are depth averaged. Thus they are not of use to the analysis of, for example, solute transfer if the solute is transported only in either the upper or lower regions of the channel (Hervouet and van Haren, 1996). Another approximation of the Navier-Stokes partial differential equations is the kinematic wave method. This is more basic and assumes that gravitational acceleration and friction are in balance, so has no change on the wave form as it moves downstream (Hervouet and van Haren, 1996).

The friction of the bed and banks of the river system, and that of the floodplain, is a key parameter in the modelling of hydraulic flow. Resistance is directly proportional to friction, and so resistance laws are generally adopted for open channel flow (Bates et al., 1996; Hervouet and van Haren, 1996). There are several numerical approximations to roughness, and often the model input is interchangeable between the measurements by Chezy (1769; from Heschel, 1897), Manning (1891) and Darcy-Weisbach (Darcy, 1857). For a summary of the differences between roughness estimation methods, refer to Chow (1959) and Knighton (1998).

There are a large number of model codes available to use for hydraulic modelling, and all are based on the above theories, differing in user interface and data input and output. A summary of some more common codes is provided in Table 3.2, with examples of application.
<table>
<thead>
<tr>
<th>Model Code</th>
<th>Details</th>
<th>Model Type</th>
<th>Reference</th>
<th>Application Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>ISIS (now IWRS)</td>
<td>Industry standard model widely used for river flow modelling. Versatility shown by range of applications including river flow (Acreman et al., 2003b), river channel evaporation (McKenzie and Craig, 2001) and floodplain modelling (Acreman et al., 2002a).</td>
<td>1D hydraulic model</td>
<td>Gomes Lopes et al., 2004</td>
<td>Acreman et al., 2002a; McKenzie and Craig, 2001</td>
</tr>
<tr>
<td>HEC-RAS</td>
<td>Similar to IWRS, also using Saint Venant 1D hydraulic flow approximations, widely utilised for general catchment scale flow modelling.</td>
<td>1D hydraulic model</td>
<td>Pappenberger et al., 2005</td>
<td>Pappenberger et al., 2005</td>
</tr>
<tr>
<td>MIKE-11</td>
<td>Hydraulic component of the MIKE suit of models. Often utilised for wetland applications due to ability to represent weirs and penning boards, and stable with small channels.</td>
<td>1D hydraulic model</td>
<td>Havnæ et al., 1995</td>
<td>Thompson, 2004; Thompson et al., 2004</td>
</tr>
</tbody>
</table>

Table 3.2 - Hydraulic model codes
Model Details

**Code**

<table>
<thead>
<tr>
<th>Model</th>
<th>Details</th>
<th>Model</th>
<th>Reference</th>
<th>Application</th>
</tr>
</thead>
<tbody>
<tr>
<td>Landscape</td>
<td>Developed from 1970s to model landscape evolution over geological timescales.</td>
<td>NA</td>
<td>Coulthard, Coulthard et al., 2001</td>
<td>Coulthard et al., 2000; Coulthard et al., 2005</td>
</tr>
<tr>
<td>Working over the basin scale spatially and at least tens of thousands of years temporally, models such as CEASAR have utility in understanding different processes from conventional hydraulic models (above). For example, soil creep would be negligible at the 'hydraulic scale', yet fundamental at the 'geomorphic scale' (Coulthard, 2001), in direct contrast to bed friction.</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Table 3.2 – Hydraulic model codes (continued)

### 3.2.4 – In-Field Hydrological Models

In-field hydrological models, as opposed to hydraulic, cover aspects of water outside of the river channel. This includes atmospheric exchanges, soil water storage and transfers, and deeper aquifer processes. The floodplain is included in the hydrological sphere when not inundated, as water moves laterally and
vertically beneath the surface, and is not subject to hydraulic free-flowing conditions. An in-field hydrological model is used to predict lateral and vertical fluxes of water and subsequent changes in water levels within an area, elements which are important in wetland environments.

The scale of this application is usually of the order of the reach or slope (McCartney 2001), but up to the catchment scale, and usually works on a water budget basis, calculating in- and outflows to and from the system in question. A rainfall-runoff model, by contrast, works at a larger scale of sub-catchment or basin, traditionally with the primary aim of creating boundary conditions for hydraulic models, although as discussed above, have now developed to be useful models in their own right. In-field hydrological models are widely used in wetland hydrological science for predicting water level change, often in order to quantify the impacts of change induced by climate change or changes in water management. Table 3.3 summarises popular model codes and provides examples of the use of in-field hydrological models in wetland environments.
<table>
<thead>
<tr>
<th>Model Code</th>
<th>Details</th>
<th>Reference</th>
<th>Application</th>
</tr>
</thead>
<tbody>
<tr>
<td>DITCH</td>
<td>Using Darcy's law for saturated flow, DITCH models just one water level for the field, negating the unsaturated zone. Does not require parameters for hydraulic conductivity and specific yield, as the software is designed to be simple to run.</td>
<td>Armstrong, 2000 (based on Youngs et al., 1989)</td>
<td>Armstrong and Rose, 1999; Gavin, 2003a</td>
</tr>
<tr>
<td>MODFLOW</td>
<td>Widely regarded as the industry standard and used extensively globally. Allows 3D modelling of the soil profile using a finite element discretisation, and is highly dependent on hydraulic conductivity values.</td>
<td>McDonald and Harborough, 1988</td>
<td>Bradley 1996; 2002; Restrepo et al., 1998; Bradford and Acreman, 2003</td>
</tr>
<tr>
<td>MIKE-SHE</td>
<td>Comprehensive and extensively used modelling package, covering subsurface, overland and in-channel flow, Thompson et al. (2004) found drawbacks with floodplain flow. MIKE-SHE is versatile and couples well not only with sister programs (see MIKE-11, above), but with other software such as Daisy (Boegh et al., 2004)</td>
<td>Refsgaard and Storm, 1995</td>
<td>Havne et al., 1995; Al-Khudhairy and Thompson, 1997; Boegh et al., 2004; Christiansen et al., 2004; Thompson, 2004; Thompson et al., 2004</td>
</tr>
<tr>
<td>ZOOM</td>
<td>UK-developed groundwater-focused model with local grid refinement algorithm, enabling the user to incorporate a higher spatial resolution in an area of interest. Similar in nature to MODFLOW, ZOOM is yet to be tested in wetland environments, although becoming more widely used with time.</td>
<td>Bear, 1979; University of Birmingham, 2001</td>
<td>Jackson, 2001</td>
</tr>
</tbody>
</table>

Table 3.3 - In-field hydrological model codes
3.2.5 – Conceptual Understanding of the Modelling Process

Modelling procedure has established as a standardised process as models have become more widely used in environmental sciences over recent decades. As discussed in Chapter 1, Refsgaard (2007a) has documented a standardised protocol for model development (Figure 3.1) incorporating widely accepted
phases such as calibration and validation, but also formalising more imprecise steps such as code selection. Not all of Refsgaard’s (2000) steps are elaborated upon below, as some such as purpose definition require little or no clarification.

3.2.5.1 – Model Construction

This step is the design and development of the model, such that it works efficiently and represents the real world conceptually yet accurately. This is perhaps the most arbitrary phase of model development, and relies on judgements to be made by the developer in order for the model to run efficiently. However, despite the apparently unempirical nature of this step, Bartlett (2007) explains that as long as judgements are documented and justified methodically, and uncertainties are highlighted, model construction should be repeatable and scientific in nature.

3.2.5.2 – Calibration

Calibration is undertaken for a period of time during which the outcome of the modelled variable is known, such as river flow at a gauged location, and the difference between the observed and modelled values can be seen. During calibration, parameters within the model are adjusted so as to find those values which give the best fit between observed data and modelled output. Parameters are changed and the model re-run, and the process repeated until the model represents reality to within specified tolerances.
The method of finding values for such parameters may be imperfect. For example, $K$ values (see Chapter 2) for soils are often measured using a soil core in a laboratory. This gives only a point scale result and may miss macropore flow, subsequently underestimating $K$; errors of two orders of magnitude are common for this particular parameter (Bromley et al., 2004).

In order for the model to be as robust as possible, calibration data must cover as wide a range of conditions as the model will be utilised for as a tool post-development; if this is done, a user will have confidence that the model will be effective in similar scenarios. For example, a model predicting wetland water table fluctuations must be calibrated for both summer and winter conditions as precipitation and evaporation are very different across the hydrological year.

Refsgaard and Storm (1995) observe that the number of parameters changed during calibration should be kept to a minimum. Therefore a modeller should choose those parameters which have less certainty associated with them or which have known to be erratic in previous models. As discussed above, $K$ values have high associated uncertainty and are regularly varied (often upwards) during the calibration procedure by up to two orders of magnitude (Bragg, 1991; Bromley et al., 2004; Thompson et al., 2004) and therefore will usually be calibrated. For hydraulic models, a common calibration parameter is roughness, as again, uncertainties are often high (e.g. Chormański et al., 2004).
3.2.5.3 – Validation

Validation documents that the site-specific model gives sufficiently accurate predictions (Refsgaard, 2000). This involves using the calibrated model (with calibrated parameters) to simulate conditions for a period of time both for which further observed data are available and which is independent from calibration data. As the calibration and validation steps are data intensive, during studies where data is sparse there can be a tendency to forego the validation process in order to maximise the calibration period. This must be avoided where possible although logistics and fiscal restraints do not always allow for intense data collection. A split sample technique can be adopted in place of formal validation where data are very limited, as described by Klmeš (1986).

3.2.5.4 – Sensitivity Analysis

An important step, and one that is often sacrificed due to time and fiscal restraints, and omitted by Refsgaard (2007a), is sensitivity analysis of input parameters. By changing certain parameters of the model by known amounts and analysing the change in output, a modeller can examine the sensitivity of a model to parameters. Sensitivity analysis is important as the confidence in parameter values may not be high, especially as their values can be changed by the calibration process from those determined in the field. Therefore, sensitivity analysis can direct the user to parameters that would benefit from higher confidence and so can drive data collection.
3.2.5.5 – Simulation

Using the calibrated and validated model for prediction is often the explicit aim of the model application (Refsgaard, 2000). The simulation of changes from the calibrated set up can be tested, to represent changes that could be made in the real world (e.g. climate change), and the effect of such changes seen in model output. Confidence in the model must be gained through transparent model construction, calibration and validation.

3.2.5.6 – Linking Models

A model representing a system may compose several models representing various subsystems. This can be the case with modelling water movement in wetlands, where no single model which can sufficiently represent both in-channel and in-field soil processes, and these may be intricately linked in many wetlands.

Many of the difficulties in linking models in this way are summarised by Thompson et al. (2004). An in-field hydrological model will have limitations in representing flow through channels and ditches commonly found in wetland environments, and similarly, a hydraulic model will lack the capability to model the flow through soil systems. The need to link such models discloses difficulties in enabling different models to communicate, and indeed how this is to be affected. Furthermore calibration becomes increasingly complicated as parameters are required to be changed in all separate models, and issues of when information is exchanged emerge. Despite these issues, some success has been observed in recent years, notably with the MIKE-SHE/MIKE-11 models.
being dynamically coupled for use in a wet grassland system in southern England (Thompson, 2004; Thompson et al., 2004). This system represented flow in both drainage ditches across the site using the hydraulic MIKE-11 model, as well as in-field hydrological processes using the MIKE-SHE hydrological code, with a dynamic link between the two exchanging information after each time step. Output from each standalone model assisted calibration and validation.

A new system, OpenMI (Tindall, 2005), acts as a linking mechanism for data exchange between models. Within OpenMI, issues such as spatial representation, time step, measurement units and initialisation conflicts are resolved by controlling each model code with standardised procedures and centralised management of model runs. By allowing models to integrate and represent increasingly complicated hydrological systems, this approach should enable greater modelling power in future years.
Chapter 2 increased the field scale understanding of the hydrology of the wetland site at Otmoor. An intense sampling and data collection regime was able to describe many of the intricate details of the hydrology at the field scale. Local meteorology was monitored, evaporation measured directly at a very fine scale, water levels across the site were measured manually on a weekly basis and monitored automatically every 10 minutes, and soil moisture was also observed. Many facets of the local hydrological regime were elucidated, and a model was developed to estimate evaporation from a strong diurnal water regime discovered at the site. These advances in the hydrological understanding were placed within the context of the locality, such as the specific heavy clay soil.

To assess the impact of the wetland’s hydrological functioning on the wider hydrology of the River Cherwell catchment, a simple conceptual model (termed level one) is developed. As Chapter 2 concluded that primary hydrological function of the wetland site within the catchment is to act as a flood storage area, the focus of the model will be on flood storage volumes. A standard measure of flood volume likely to inundate the catchment on a regular basis will be calculated using standard peaks over threshold (POT) techniques (Section
3.3.2.1). This will be directly compared to the capacity of storage in the wetland (Sections 3.3.2.2 and 3.3.2.3), which will be evaluated from data held in digital elevation models (DEMs) and from knowledge obtained in Chapter 2. A similar procedure will be undertaken at another wetland site in the UK, Tealham and Tadham Moors (Section 3.3.3), which is a contrasting wetland dominated by peat soils, and likely to have different flood storage properties. As Tadham has not thus far been studied in the current work, a brief introduction will be given in Section 3.3.3.1. The analysis and direct comparison of such contrasting wetland sites should reveal the importance of physical differences between sites, and the implications of this on any wetland hydrological functions that may affect catchment scale hydrology.

3.3.2 – Otmoor Level One Model

3.3.2.1 – Calculating Typical Flood Event Flow and Volume at Otmoor

Standard POT analysis (Robson and Reed, 1999) was undertaken on daily flows from the River Ray at Islip (west of Otmoor, location shown in Figure 3.2) gauging station. $Q_{med}$ (the peak flow of the median annual maximum flood) will be calculated to enable the height of flood for this standard event to be determined. However, as flood storage is the primary component of the analysis of the level one model, this method has been adapted to calculate a value of $V_{med}$ (the volume of the median annual maximum flood), as utilised by Acreman et al. (2007).
Figure 3.2 – Islip gauging station location map (indicated by yellow dot);
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Daily flow data for the Ray at Islip were obtained for the hydrological years 1996 to 2004 inclusive, with no data available for the calendar year 2004. The periods of October to December 1996 and January to September 2005 were taken as one complete hydrological year, as accepted in standard flow analysis practice in the UK (Robson and Reed, 1999). This left eight complete hydrological years of flow data available for analysis, comparable with the Tadham level one model (Section 3.3.3). Table 3.4 details peak flows by rank for the data period.
Table 3.4 – Peak flows by rank for the River Ray
at Islip for the hydrological years 1996-2004

$Q_{med}$ is given by:

$$Q_{med} = wQ_i + (1-w)Q_{i+1}$$

Equation 3.1

Where:

- $Q_{med}$ = peak flow of the median annual flood $\text{m}^3\text{s}^{-1}$
- $w$ = weight (as given by Robson and Reed, 1999) 0.147
- $Q_i$ = peak flow of the $i$ ranked event $\text{m}^3\text{s}^{-1}$
- $Q_{i+1}$ = peak flow of the $i +1$ ranked event $\text{m}^3\text{s}^{-1}$
Standard POT methodology requires $w$ and $i$ to be taken from reference tables to ensure appropriate statistical weighting (Robson and Reed, 1999), and are given as 0.147 and 6 respectively. $Q_{\text{med}}$ is therefore calculated as 11.84 m$^3$ s$^{-1}$.

$V_{\text{med}}$ is required to compare storage volumes at Otmoor with a standard measure of flood event volume. As $Q_{\text{med}}$ calculations manipulate single values of daily flows, it is straightforward to extract peak flow data. However, with $V_{\text{med}}$, the volume of the entire storm event must be considered (Acreman et al., 2007), not merely the 'peak volume' (a direct conversion from peak flow). This creates added complication as the start- and endpoints of the event must be methodically evaluated. This was done in the current study by using the 10 percentile value (the flow which is maintained in the river for 10% of the time; see Marsh and Lees, 2003). For the River Ray at Islip, this value is 3.1 m$^3$s$^{-1}$ (Marsh and Lees, 2003), and distinguished the beginning and end of each event (non-inclusive). The same data were used as for calculating $Q_{\text{med}}$, above, with eight years of data. Table 3.5 shows the event volume data by rank.
<table>
<thead>
<tr>
<th>Rank</th>
<th>Volume</th>
<th>Event Start</th>
<th>Event End</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>43,475,616</td>
<td>28/10/2000</td>
<td>28/12/2000</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>36,408,096</td>
<td>18/12/1998</td>
<td>02/02/1999</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>23,936,256</td>
<td>21/12/2002</td>
<td>16/01/2003</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>22,793,184</td>
<td>22/01/2001</td>
<td>21/02/2001</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>22,551,264</td>
<td>03/11/2002</td>
<td>07/12/2002</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>17,808,768</td>
<td>09/04/1998</td>
<td>30/04/1998</td>
<td>i</td>
</tr>
<tr>
<td>7</td>
<td>16,464,384</td>
<td>24/12/1997</td>
<td>22/01/1998</td>
<td>i+1</td>
</tr>
<tr>
<td>8</td>
<td>15,174,432</td>
<td>25/01/2002</td>
<td>16/02/2002</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.5 – Peak volumes by rank for the River Ray at Islip for the hydrological years 1996-2004

\[ V_{med} = wV_i + (1-w)V_{i+1} \]

Equation 3.2

Where:

- \( V_{med} \) = volume of the median annual flood \( m^3 \)
- \( w \) = weight (as given by Robson and Reed, 1999) \( 0.147 \)
- \( V_i \) = volume of the \( i \) ranked event \( m^3 \)
- \( V_{i+1} \) = volume of the \( i+1 \) ranked event \( m^3 \)
Again, $w$ and $i$ were taken from reference tables (Robson and Reed, 1999), and are given as 0.147 and 6 respectively, and $V_{med}$ is subsequently calculated as 16,662,008 m$^3$.

3.3.2.2 – Calculating Subsurface Storage Capacity at Otmoor

Subsurface storage has been shown in Chapter 2 to be important for hydrological functioning, dependent upon the time of year. In order to calculate the subsurface storage for the Otmoor area, a geographic information system (GIS) was constructed containing Ordnance Survey (OS) background maps, a digital elevation model (DEM; sourced from NextMAP and LiDAR, see Section 3.3.2.3 for further details) and manually digitised features such as the Otmoor boundary and river networks, sourced from OS mapping. The GIS gave the Otmoor area as 18.906 km$^2$.

Considering that most storm events of significant magnitude in volume terms are frontal rather than convectional, and so occur in winter months, soil water levels across Otmoor are likely to be high, limiting the subsurface storage potential severely. For the purposes of this study an intermediate water level will be assumed in order to calculate the potential of subsurface storage in the context of $V_{med}$. Although the water table is often at or above the surface during winter months when the majority of flood events occur, a value of 0.4 m below ground would reflect a typical autumn water level, as observed over the autumn sampling of 2004 to 2006.
Chapter 2 also explicated information on the specific yield (the volume of water storage per unit volume of soil) of the soil at Otmoor. Although this is applicable only for the research plot, it is assumed here that this value of 3% is representative of the Otmoor area. The same extrapolation warning is true for the autumn water table depths above. Combining these factors leads to a calculation of subsurface storage, as in Equation 3.3:

\[ S_s = A \Delta h S_y \]

Where:

- \( S_s \) = subsurface storage = \( m^3 \)
- \( A \) = area = \( m^2 \)
- \( \Delta h \) = change in height of storage = \( m \)
- \( S_y \) = specific yield of the soil = \( \% \)

The equation, with area, height to water table and specific yield as discussed above gives a subsurface storage value of 226,872 m³.

3.3.2.3 - Calculating Above Ground Flood Storage Capacity at Otmoor

The flood storage of the Otmoor area is a key element of its hydrological functioning. The above ground component of this will be calculated from DEM data. The GIS is used to calculate the volume beneath a theoretical horizontal plane (at 0.1 m vertical intervals, from 57.8 to 59.0 maOD) within the Otmoor area using ArcMap GIS software’s 3D Analyst (version 9.1, ESRI Inc., Redlands,
USA) for two different DEMs. The first DEM was NEXTMap Britain™ (Intermap Technologies; henceforth referred to as NEXTMap), which has a spatial resolution of 5 m, a vertical accuracy of 0.5 m and is produced using interferometric synthetic aperture radar. Complete coverage of Great Britain is anticipated, and licensed to CEH under a NERC agreement allowing access to the entire Cherwell catchment. Whilst comparing remotely sensed imagery for the identification of glacial landforms, Smith et al. (2006) found NEXTMap performed well, largely due to its fine spatial resolution, only failing on small scale drumlin features. The same authors conclude that the importance of scale-appropriate datasets with regard to the application is great. One major limitation of NEXTMap is the lack of penetration through forest canopies (Dowman et al., 2003), which should not be a problem at Otmoor, although hedges are detected. Figure 3.3 shows the NextMAP data for the Otmoor area.
The second DEM is LiDAR (light-induced direction and ranging) data, obtained for the Otmoor area from the Environment Agency (EA) through an existing agreement for ongoing research on the Cherwell catchment. LiDAR data (Otmoor area shown in Figure 3.4) has a finer spatial resolution (2 m grid cell) and accuracy (published specification ±0.10 to ±0.15 m), and is regarded within the field as the contemporary market leader in digital topographic mapping datasets, forming ideal reference data (Smith et al., 2006). French (2003) details that LiDAR data are invaluable in revealing subtle variations in topography typical of estuary and floodplain areas, and that the accuracy and resolution are close to present limits of numerical model representation. It is expected that due to the finer resolution and higher accuracy, the LiDAR coverage should be of increased utility than the NEXTMap data and offer enhanced results for calculating wetland water storage. Post-processing creates a digital terrain model (DTM) version by passing the raw digital surface model (DSM) through a filtering routine to strip out vegetation and buildings (Dowman et al., 2003). This is an automatic procedure searching for unnatural slope gradients, and interpolating resulting gaps from surrounding data.

Figure 3.5 shows the Otmoor water storage volumes calculated from the DEMs at different water levels. It can clearly be seen that the storage potential appears to be greater when the NEXTMap dataset is used, and that the difference begins as low as 57.8 maOD. At key water levels this is an important difference; at 58.2 maOD there is a factor of 6 between storage calculated from LiDAR and NEXTMap data. 58.2 maOD is a level at which the River Ray, adjacent to Otmoor, will reach on a regular basis.
Figure 3.4 - LiDAR data for the Otmoor area;

Figure 3.5 - Comparing storage using different DEMs
In order to investigate the reasons for this discrepancy further, maps were produced of water storage at Otmoor, again at flood level intervals of 0.1 m, shown in Figure 3.6. It is evident from the LiDAR images that there are significant areas of 'no data', where there are apparent gaps in the data coverage. It has been calculated that these represent 0.14 km$^2$, or 0.74% of the Otmoor coverage. On further inspection, these seem to correspond to a similar coverage of the NEXTMap data at approximately 58.0 m aOD. One known limitation of the LiDAR technique is its lack of recognition of water surfaces (French, 2003); it is therefore likely that these areas of 'no data' are in fact inundated areas, and so represent water levels at the time of flying. Therefore, although constituting only a relatively small area, these areas could be particularly important to water storage as they represent the lowest areas and so where water will congregate. On this basis, the LiDAR data were post-processed by replacing 'no data' areas with a level of 57.9 m aOD, assuming that 'no data' areas are all surface inundation and that the surface topography at these points is known to be a maximum of 57.9 m aOD. The maximum accuracy of LiDAR is known to be +/-0.1 m, and so current water level minus 0.1 m was seen as the most appropriate substitute level. Although somewhat undermining LiDAR's high accuracy, it was felt justified in order to utilise the finer resolution of LiDAR data. The results (Figure 3.7) show that some of the discrepancy between the DEMs has been resolved, but parity is not approached. This is likely to be due to some areas of inundation being lower than 57.9 m aOD, leading to storage volume continuing to be underestimated. On this basis it is decided that NEXTMap data be used for above-ground storage, as it is felt that
Figure 3.6 - Comparing LiDAR and NEXTMap at different elevation thresholds; blue shows area below threshold, purple above and white areas of no data.
The conventional $Q_{med}$ (Robson and Reed, 1999) calculated above for the River Ray at Islip is 11.84 m$^3$s$^{-1}$. Using a flow-stage relationship, it is possible to estimate the stage of the river at Islip for a $Q_{med}$ event, and by applying the difference in elevation between Islip and Otmoor, the likely stage upstream at Otmoor. A flow-stage relationship is determined from flow data supplied with concurrent stage data.
For flows up to 15 m$^3$s$^{-1}$, the fit is a third-order polynomial:

$$h = -0.005439Q + 0.02160Q^2 - 0.0008419Q^3 + 56.76$$

Equation 3.4

For flows above 15 m$^3$s$^{-1}$, a linear fit is applied, calculated from the largest flow on record (20.41 m$^3$s$^{-1}$, measuring a stage of 58.858 maOD):

$$h = 0.02976Q + 58.2506$$

Equation 3.5

This methodology captures well the essence of the flow-stage data, and provides a robust model to estimate stage at Islip from flow data, with an $r^2$ of 0.86. The data of flow and stage are daily averages from 1997 to 2003 inclusive, and are shown in Figure 3.8. Also shown is the above flow-stage relationship. A flow of 11.84 m$^3$s$^{-1}$ ($Q_{med}$) translates to a stage at Islip of 58.33 maOD.
To translate this upstream to Otmoor, a point on the River Ray half way through its journey across Otmoor is chosen, and the difference in elevation between here and Islip is applied to the stage of the River at Islip. This method is not without limitation, as shown in Figure 3.8 there is a significant phenomenon of the River Ray 'backing up' during high flow on the River Cherwell. Therefore stages at Islip are likely to be somewhat unusually higher for a given flow relative to the stage at Otmoor when this occurs (> ~25 m$^3$s$^{-1}$ on the Cherwell at Islip). A $Q_{med}$ flow should not be high enough to coincide with this phenomenon, although this cannot be discounted altogether. No robust method can be found to avoid this, and considering the raison d'être of the level one model, this method is assumed satisfactory: a quick assessment is the main objective of the model.
The difference in topography between the gauging station at Islip (57.92 maOD) and Otmoor (58.10 maOD) is 0.18 m. Adding this difference to the stage at Islip of $Q_{med}$ (58.33 maOD), gives 58.51 maOD. Extracting the volume of storage from the NEXTMap stage-volume curve gives 720,000 m$^3$.

### 3.3.2.4 – Calculation of Evaporation Loss at Otmoor

In order to estimate the effect of water loss by evaporation on the flood storage, a calculation was made of loss during an average length flood event, using Equation 3.6:

$$E_{V_e} = (E_{V_d}dA)1,000$$

**Equation 3.6**

Where:

- $E_{V_e}$ = evaporation over the wetland during the flood event m$^3$
- $E_{V_d}$ = daily evaporation mm day$^{-1}$
- $d$ = length of event days
- $A$ = area km$^2$

A value for wetland evaporation was taken from the analysis undertaken in Chapter 2. As almost all flood events in Table 3.5 (analysis of $V_{med}$ for the Ray at Islip) occurred in winter months, a range of 0.5-1.0 mm day$^{-1}$ was seen to be representative of winter evaporation; the lower value was taken as evaporation will be lower during rainfall events. The GIS gave the Otmoor area as 18.906
km$^2$, and the length of the event was taken as the average of those in Table 3.5: 33 days. The volume of evaporation is calculated as 311,949 m$^3$.

3.3.2.5 – Calculation of Ditch Storage at Otmoor
Surface ditches are used at Otmoor to increase surface water and so marginal habitat for breeding waders, and to manage surface water movement across the site. The GIS has calculated the length of ditches to be 15 km and a survey of ditches at 10 points both close to the research plot and near access tracks across the site gave an average width of 2 m and depth of 0.3 m. This gives a total storage volume of 9,000 m$^3$, assuming a rectangular cross section (so being an overestimate).

3.3.2.6 – Otmoor Model 1 Results
It was thought relevant to compare the results of Otmoor flood storage calculations with the $V_{med}$ (Table 3.7) of the River Cherwell downstream at Oxford; $Q_{med}$ (Table 3.6) was also calculated in order to put later results into context. Using the same methodology as for the River Ray at Islip, $V_{med}$ was found to be 64,453,430 m$^3$, 386% that of the Ray at Islip (which obviously forms part of the $V_{med}$ at Oxford). The largest events are shown in Table 3.7. The 10th percentile value for the Cherwell at Oxford is given as 7.2 m$^3$s$^{-1}$ by Marsh and Lees (2003), and this value delineates the start- and endpoints of flood events. The criteria for ensuring flood peak independence were required for this analysis (as some ambiguity was present for these events), and are as follows (Robson and Reed, 1999):
• Separation of flood peak by at least three times the average time to peak;
• The minimum discharge in the trough between peaks must be less than two thirds the discharge of the first peak.

If peaks are thus classified as independent but flow does not drop to less than the 10th percentile between events, the minimum flow between events is taken as the delimiting point.

<table>
<thead>
<tr>
<th>Rank</th>
<th>Flow</th>
<th>Date</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>m³ s⁻¹</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>90.87</td>
<td>11/04/1998</td>
<td></td>
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<td>2</td>
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<td>10/12/2000</td>
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<tr>
<td>3</td>
<td>44.04</td>
<td>27/12/1999</td>
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<tr>
<td>4</td>
<td>42.69</td>
<td>15/02/2001</td>
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<tr>
<td>5</td>
<td>38.04</td>
<td>05/01/2001</td>
<td></td>
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<td>6</td>
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<td>06/01/1998</td>
<td>i</td>
</tr>
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<td>7</td>
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</tr>
<tr>
<td>8</td>
<td>32.08</td>
<td>22/01/1999</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.6 – Peak flows by rank for the River Cherwell at Oxford for the hydrological years 1996-2004
Table 3.7 – Peak volumes by rank for the River Cherwell at Oxford for the hydrological years 1996-2004

\[ Q_{med} \] is therefore calculated as 33.74 m³s⁻¹, and \[ V_{med} \] as 64,453,430 m³ using Tables 3.6 and 3.7 respectively. Table 3.8 outlines the potential water storage at Otmoor to the \[ V_{med} \] at both Islip and Oxford. The total flood storage calculated in Table 3.8 may be expressed per unit area, which equates to 0.067 m³m⁻².
### Table 3.8 – Results for Otmoor level 1 model

<table>
<thead>
<tr>
<th>Measure</th>
<th>Flood Event</th>
<th>Subsurface Above</th>
<th>Ditch Network</th>
<th>Evaporation</th>
<th>Total</th>
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</thead>
<tbody>
<tr>
<td>Ray at Islip:</td>
<td>m³</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>16,662,008</td>
<td>226,872</td>
<td>720,000</td>
<td>9,000</td>
<td>311,949</td>
</tr>
<tr>
<td></td>
<td>%</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N/A</td>
<td>1.36</td>
<td>4.32</td>
<td>0.05</td>
<td>1.87</td>
</tr>
<tr>
<td>Cherwell at Oxford:</td>
<td>%</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N/A</td>
<td>0.44</td>
<td>1.12</td>
<td>0.01</td>
<td>0.48</td>
</tr>
</tbody>
</table>

3.3.2.7 – Otmoor Model 1 Conclusions

It must be noted that for larger floods than $V_{med}$, the above ground storage at Otmoor will obviously increase as the stage of the flood at Otmoor increases; this is not the case with the subsurface storage component. It can be seen from the results of the level one model that for a $V_{med}$ event, Otmoor has significantly more storage capacity above ground than below. The heavy clay nature of the soil at Otmoor dictates that the specific yield is particularly low, estimated to be 3%. Therefore even an increase in the depth of the subsurface store (such as during a drier time of year, when the localised water table is significantly lower) will make an impact on the total store. To give some indication of the impact of this, if the water table were at its maximum recorded depth of 1.3 m below the
surface, the subsurface storage component would increase to 737,334 m³ (4.43% of $V_{med}$ at Islip), and so be comparable to the above ground storage during a $V_{med}$ event. However, as the soil is heavy clay, access to lower layers may not be available in the timescale presented by summer convectional events, which are likely to be prevalent during times of such water table elevations. On this basis it is felt that the subsurface storage component is not fundamental to the hydrological functioning of the Otmoor wet grassland wetland system in a catchment context, and need not necessarily be included in a full hydraulic model of the river network. Similarly, evaporation is not a major storage component for a winter flood event (1.87% of $V_{med}$ at Islip), although would increase greatly for a summer event such as that seen in July 2007. Despite the extensive ditch network at Otmoor, again this is small compared to the above-ground storage (0.05% and 4.32% of $V_{med}$ at Islip respectively), showing their utility for management (through movement) of surface water and creation of habitat, and not storage.

Total wetland storage during a typical flood event has been calculated to be 7.61% of the volume of that event at Islip. This is a significant volume and could possibly be the difference between the river bursting its banks and remaining in-channel. Obviously, once the storage has been used, it will take time to dissipate from the wetland system. Therefore the storage would only be available for a primary rain event and not for subsequent events occurring soon afterwards. The implications for this phenomenon will be explored using the hydraulic model in Section 3.4. The storage has been shown to be 1.97% of a typical flood event in Oxford, a city prone to flooding.
This level of flood storage, when considered as originating from a single wetland site, is important for flood defence managers. Importantly, the amount of storage has been shown to be significantly affected by the water level on the wetland site previous to the storm event. This information has implications for the management of the wetland, in today's climate of wetland restoration and widespread raising of water levels to achieve such aims (RSPB, 2006). It was shown in Chapter 2 that the management of Otmoor has a large impact on the hydrological behaviour of the site at a small scale, and this work has shown that the water level management will also greatly affect the behaviour of the wetland at the catchment scale, and may be used as an effective tool by flood defence managers. The impact of 18.9 km² of floodplain in the Thames basin has been demonstrated as having an effect downstream, and so the vast cumulative coverage of wetlands across the Upper Thames basin has huge potential for more effective management. This issue will be explored in Section 3.4 with the use of a hydrodynamic model of the Cherwell catchment.

3.3.3 - Tadham Level One Model

3.3.3.1 - Introduction to Tealham and Tadham Moors

In order to compare the contributing components to hydrological functioning of different wetland systems, the level 1 model methodology will be applied to a second wetland site. The North Drain is a 30 km² sub-catchment of the River Brue in Somerset, southwest England (see location map, Figure 3.9). The
catchment is dominated (88%) by Tealham and Tadham Moors (referred to hereafter as Tadham, see detail map, Figure 3.10), a peat wetland area overlying marine clay, and part of the wider Somerset Levels and Moors wetland system. The area is particularly flat, with only a small contributing upland to the north (Figure 3.11).

Figure 3.9 – Map showing location of Tadham within England
Figure 3.10 – The Tadham area;
© Crown copyright Ordnance Survey 2007

Figure 3.11 – DEM of the Tadham area; NEXTMap Britain™ elevation data from Intermap Technologies
The site (Figure 3.12) has a long history of management, with peat extraction beginning during Roman times (DEFRA, 2006). Today the area has Site of Special Scientific Interest (SSSI) and Special Protection Area (SPA) status, and 290 km$^2$ of the Brue catchment are managed under what was previously the Environmentally Sensitive Areas (ESA) scheme, now under the umbrella of Natural England’s Environmental Stewardship programme (Natural England, 2007). The scheme offers payment to farmers to raise winter water levels in order to create wetland habitat and attract breeding waders such as Snipe (\textit{Gallinago gallinago}) (Tucker and Petterson, 2003). The system uses graduated tiers, with increasing severity of water level prescriptions and their duration; Tier 3 prescriptions, the most stringent, dictate that winter water levels should not be less than mean field level. The effect of raising water levels was thus the focus of the calculations.

With such a high water storage component in a single catchment, it is felt that if the impacts of such storage could not be demonstrated at Tadham, it would be somewhat undermined as a factor in catchments with less wetland coverage. This analysis of flood storage at Tadham contributed to the work by Acreman et al. (2007).

The clay-based Otmoor wet grassland and peat-dominated Tadham areas have very different water storage components. Otmoor, as shown above, has subsurface and above-ground storage components. In addition, Tadham has a dense network of ditches (significantly more substantial than those at Otmoor), installed both to manage water levels and act as wet fences, which provide
significant water storage potential. Water levels at Tadham are managed by a series of gates and pumps, the latter primarily at the pumping station at the downstream end of the North Drain catchment. Water levels are restricted from rising above mean field level, and consequently, above-ground storage is not of particular importance.

![The Tadham wetland site](image)

Figure 3.12 – The Tadham wetland site

A GIS was created for the North Drain catchment, including the catchment boundary (determined from flow directions derived from a DEM), ditch network data (digitised from 1:25,000 OS maps) and areas of land recently under Tier 3 conditions (from local authority payment records). Two scenarios were developed for the flood storage calculations. Firstly it was deemed important to calculate storage with contemporary Tier 3 coverage, and secondly with complete Tier 3 coverage across the peat covered component of the North Drain catchment.
### 3.3.3.2 - Calculating Typical Flood Event Volumes at Tadham

A similar methodology to that used in the Otmoor level one model will now be applied to the Tadham wetland area. Flow data were not available at the North Drain outlet, as there is no conventional gauging station at the site. However, records were available for a pumping station at the catchment outlet; the sole connection for the North Drain catchment to the River Brue. Records were kept of natural drainage and pumped water, the total of which formed the basis for the present work. Data were only available in daily volumes, necessitating manipulating the standard $Q_{med}$ calculation to its $V_{med}$ form (Acreman et al., 2007).

The threshold at which to delimit the start- and endpoints of the flood event was somewhat arbitrary, set at 160,000 m³day⁻¹ (1.85 m³s⁻¹), yet reflected the level at which pumping was required and so water surplus was being removed by pump operators. Table 3.9 shows the resulting event volume data by rank.

<table>
<thead>
<tr>
<th>Rank</th>
<th>Volume</th>
<th>Event Start</th>
<th>Event End</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>12,112,356</td>
<td>18/12/1999</td>
<td>13/01/2000</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>5,267,993</td>
<td>02/01/1998</td>
<td>15/01/1998</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>4,780,755</td>
<td>06/12/2000</td>
<td>18/12/2000</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>4,476,182</td>
<td>24/09/1999</td>
<td>06/10/1999</td>
<td>$i$</td>
</tr>
<tr>
<td>7</td>
<td>4,240,915</td>
<td>14/01/1999</td>
<td>31/01/1999</td>
<td>$i+1$</td>
</tr>
<tr>
<td>8</td>
<td>3,713,878</td>
<td>18/04/2000</td>
<td>29/04/2000</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.9 – Peak volumes by rank for the North Drain at Tadham
As with the Otmoor calculations, Equation 3.2 was used, and the coefficients $w$ (0.147) and $i$ (rank 6) were taken from the Flood Estimation Handbook (Robson and Reed, 1999); $V_{med}$ was subsequently given as 4,275,499 m$^3$.

3.3.3.3 – Calculating Subsurface Storage Capacity at Tadham

The volume of water able to be stored in the soil column can be large, especially for peat soils such as those at Tadham, as peat has a generally high specific yield of up to 20% (Armstrong, 1993; Gilman, 1994; Price et al., 2003). The method for calculating this storage used here distinguishes between areas where water levels are likely to be higher (nearer the ditches under the Tier 3 raised water levels), and those under normal conditions (in the centre of fields and all areas not under Tier 3 conditions). It was determined from several years' monitoring at CEH's wetland research facility located within the Tadham area, that winter water levels outside the raised water level scheme areas were on average 0.57 m below the soil surface, leading to Tier 3 (winter water levels at mean field level) removing that amount of potential flood storage ($z_d$; Figure 3.13, a cross section through a typical ditch and adjacent field). The same dataset indicated that water levels in the centre of the fields with perimeter ditches under Tier 3 prescriptions was 0.2 m lower than that of the ditches ($z_f$; Figure 3.13) and that the lateral extent of the ditches' effect ($E_d$; Figure 3.13) was approximately 10 m. A specific yield ($S_y$) value of 20% was taken from the literature cited above.
The GIS gave the total length of ditches under Tier 3 prescription \((L_3)\) as 10,978 m, and field area with the same conditions \((A_3)\) as 681,753 m\(^2\). The soil water storage for current T3 is calculated using Equation 3.7.

\[
S_3 = L_3 2E_d z_d S_y + (A_3 - L_3 E_d S_y) z_f
\]

Equation 3.7

Where:

- \(S_3\) = soil water storage under current T3 \(m^3\)
- \(L_3\) = length of ditches under current T3 \(m\)
- \(E_d\) = distance of lateral percolation into field \(m\)
- \(z_d\) = change in ditch water level storage of T3 \(m\)
- \(S_y\) = specific yield \(\%\)
- \(A_3\) = area of fields under current T3 \(m^2\)
- \(z_f\) = change in field water level storage of T3 \(m\)
Equation 3.7 gives $S_3$ as 125,149 m$^3$. The soil water storage for complete T3 coverage is given by Equation 3.8, which gives $S_c$ as 3,015,619 m$^3$.

$$S_c = A_c z_d S_y$$

**Equation 3.8**

Where:

- $S_c = \text{soil water storage under complete T3}$ (m$^3$)
- $A_c = \text{area of fields under complete T3}$ (m$^2$)
- $z_d = \text{change in ditch water level storage of T3}$ (m)
- $S_y = \text{specific yield}$ (%)

### 3.3.3.4 – Calculating Ditch Storage Capacity at Tadham

Ditches are important at Tadham for managing water levels and acting as wet fences to control stock. Their potential for storing flood water is also important, and contributes to the wetland's overall flood storage capacity. Information on the dimensions of ditches has been collected as part of CEH's wetland research at Tadham, and this will be used in conjunction with data from the GIS to estimate the role of the ditches in flood storage. The Tadham GIS reveals that across the 29.98 km$^2$ of catchment (26.45 km$^2$ of which is peat covered), there are 308.6 km of ditches, 10.98 km of which are under Tier 3 prescriptions. Ditches are almost uniformly 3 m wide; this consistency indicates a good degree of maintenance. As with soil water storage, the amount of vertical storage lost is 0.57 m compared to land outside the improvement scheme.
Under recent Tier 3 coverage, this would remove 18,772 m$^3$ of flood storage, and if Tier 3 prescriptions were extended across the peat component of the North Drain catchment, this would increase to 527,743 m$^3$ (utilising all of the 308,622 km of ditches), as shown in Table 3.10.

If the scenario of complete Tier 3 coverage were to become reality through policy change, it could be expected that the stage of North Drain would also be raised in line with that of the ditches and in-field water tables within the wider catchment, and on this basis the North Drain component was calculated. Only partial Tier 3 coverage would not initiate raising the level of the North Drain. The GIS shows that the length of the North Drain is 10.26 km long, and has an average width of 6.87 m. Assuming a change in stage concurrent with that of the ditches, 0.57 m, a storage capacity of 40,169 m$^3$ would be lost to such a scenario.

<table>
<thead>
<tr>
<th>Length</th>
<th>Width</th>
<th>Vertical Change</th>
<th>Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>m</td>
</tr>
<tr>
<td>m</td>
<td>m</td>
<td>m</td>
<td>m$^3$</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Ditches</th>
<th>10,978</th>
<th>3</th>
<th>0.57</th>
<th>18,722</th>
</tr>
</thead>
<tbody>
<tr>
<td>Complete T3</td>
<td>308,622</td>
<td>3</td>
<td>0.57</td>
<td>527,743</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>North Drain</th>
<th>10,258</th>
<th>6.87</th>
<th>0.57</th>
<th>40,169</th>
</tr>
</thead>
<tbody>
<tr>
<td>Complete T3</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3.10 – Tadham level 1 model storage ditch storage results
3.3.3.5 – Calculation of Evaporation Loss at Tadham

In order to estimate the effect of water loss by evaporation on the flood storage, a calculation was made of loss during an average length flood event, using Equation 3.6, as with the Otmoor calculations. Calculation of evaporation loss was not included in the work published by Acreman et al. (2007).

The average evaporation was taken as that for Otmoor, 0.5 mm day$^{-1}$, as it was felt that this would be representative of winter wetland evaporation, which is much less variable than that in summer. The length of event was taken as 17 days (the average of those in Table 3.9 to calculate $V_{med}$ in the North Drain), and the area of wetland at Tadham was calculated as 26.45 km$^2$ as in Section 3.3.3.3. The evaporation loss over an average event was calculated as 224,825 m$^3$, and would be the same for both Tier 3 coverage scenarios. This loss is effectively extra storage, as it is water lost from the system so allowing more in. The value calculated for Tadham is less than that for Otmoor, due to the significantly shorter length of average event.

3.3.3.6 – Tadham Model One Results and Conclusions

A summary of the calculations of storage lost across the Tadham peat wetlands to two raised water level scenarios is shown in Table 3.11. The first scenario is the recent coverage of Tier 3 uptake, and the second complete coverage across the peat component of the catchment.
### Table 3.11 – Tadham level 1 model total storage results

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Soil storage</th>
<th>Ditch storage</th>
<th>North Drain</th>
<th>Evaporation</th>
<th>Total</th>
<th>% $V_{med}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>m$^3$</td>
<td>m$^3$</td>
<td>m$^3$</td>
<td>m$^3$</td>
<td>m$^3$</td>
<td></td>
</tr>
<tr>
<td>Current T3</td>
<td>59,232</td>
<td>18,722</td>
<td>N/A</td>
<td>224,825</td>
<td>302,779</td>
<td>7</td>
</tr>
<tr>
<td>Complete T3</td>
<td>3,015,619</td>
<td>527,743</td>
<td>40,169</td>
<td>224,825</td>
<td>3,808,356</td>
<td>89</td>
</tr>
</tbody>
</table>

When expressed as storage per unit area, current T3 is 0.011 m$^3$m$^{-2}$, and complete T3 coverage as 0.144 m$^3$m$^{-2}$. This compares with the Otmoor value of 0.066 m$^3$m$^{-2}$.

Even with Tier 3 coverage at its present 0.68 km$^2$, the amount of potential flood storage taken up by the raised water levels is 7% of $V_{med}$, an increase that would have a notable impact on downstream flow during storm events. This value is dominated by evaporation loss, importantly for the water resource management at Tadham.

The hydrological impacts of wetland restoration are demonstrated starkly by the introduction of a scenario whereby Tier 3 coverage is increased to cover the entire peat-covered component of the North Drain catchment. The loss of flood storage is estimated to be 89% of $V_{med}$, with soil storage dominating storage, although the North Drain catchment does represent an unusual situation of having 88% of the catchment constituting peat wetland. Even so, it has been
proven that raising water levels to achieve environmental objectives may have serious implications for integrated catchment management, such as flood storage, in this case on the River Brue, downstream.

### 3.3.4 – Model 1 Conclusions

The analysis of storage capacities of two wetland systems have been compared to a standard measure of flood event. At Otmoor, a site studied in detail in Chapter 2, flood storage capacity during a typical flood event has been shown to be 7.61% of an annual flood at nearby Islip on the River Ray, and 1.97% of the flow further downstream at the flood-prone city of Oxford. At Tadham, the impact of raising water levels under agri-environment management practices has been shown to considerably reduce the flood storage potential of the wetland: by 7% of $V_{med}$ at the North Drain outlet under current management, with the potential for 89% of $V_{med}$ should the scheme be extended to complete coverage. This particularly large (potential) value reflects the extensive wetland coverage within the Tadham catchment (88%); Otmoor constitutes a much smaller component of the Ray catchment (1%).

The wetland storage values expressed per unit area were calculated as 0.066 m$^3$m$^{-2}$ for Otmoor, 0.011 m$^3$m$^{-2}$ for current T3 at Tadham and 0.144 m$^3$m$^{-2}$ for potential T3 at Tadham. Otmoor has a greater storage than Tadham in its current state due to the dominance of the above surface storage, a feature which is absent from Tadham. The storage at Tadham rises above that of Otmoor.
when T3 conditions are applied across the site, due largely to the high specific yield.

Although relatively elementary calculations, the results are stark: these wetlands are large complete wetland units by British standards, but represent only small areas of larger river basins. The behaviour and flood storage potential of these wetlands of different type demonstrate the importance of wetlands within the concept of integrated river basin management. Also demonstrated is the impact of management at the sites – increasing water levels does significantly decrease flood storage potential.

The work has highlighted the importance of the assessing the hydrological functioning of wetland systems across the UK. Although flood storage is only one particular function, it has been shown to be key at the catchment scale. Another possible side effect of increasing water levels under wetland restoration schemes is the subsequent increase in evaporation likely to occur as water becomes more directly linked with the atmosphere by sitting closer to wetland surfaces (Gasca-Tucker et al., 2007), adding further to water resource demands across the catchment.
3.4 – Level Two Model

3.4.1 – Introduction

The range of model types available to estimate the flow of water through a river system has been demonstrated in Section 3.2, and a simple model utilised in Section 3.3 to assess storage volumes at two wetland sites. In this section a hydraulic model will be developed to predict river flows within the Cherwell catchment, which can then be queried to provide information about the effect the wetland system at Otmoor has on catchment flows.

The model codes for the rainfall-runoff and hydraulic models used for the current work will be selected in Section 3.4.2. A major component of the model is a rainfall-runoff model, which translates rainfall across the catchment to flow within the river network. It is then the hydraulic model per se which routes the water within the river network to calculate water flow and levels throughout the river and floodplains. Section 3.4.3 describes the rainfall-runoff model development, including calibration and independent validation as described by Refsgaard (2000).

Several steps are required to construct a comprehensive hydraulic model, including importing river cross section data, which are essential to accurately
compute river flow along a significant stretch of river. Section 3.4.4 describes this methodology, including converting and importing data from surveyed cross sections of the river (Section 3.4.4.1). The model is to be primarily utilised for the assessment at high flows, during which water levels in the cross sections rises and so forcing water to spread out of the river channel and across the floodplain. In order to be able to estimate flow during these periods, river cross sections are required to be extended out of the river channel and on to the wider floodplain. The methodology for this task is described in Section 3.4.4.2. Further development of model configuration and calibration of the model is described in Sections 3.4.4.3-9, including the addition of weirs and spill units.

The primary calibration parameter used is roughness of the river channel and floodplain surfaces, and due to the dependence on this one parameter, a sensitivity analysis of roughness is undertaken (Section 3.4.5). The final calibration of river channel and floodplain roughness is shown, together with model performance for calibration events (Section 3.4.6). Section 3.4.7 shows the validation of the model using events independent from the calibration procedure.

Once the model has been successfully calibrated and validated, Section 3.4.8 investigates the likely effect of different modelling strategies, and Section 3.4.9 will utilise the model to assess the impact of making changes to the river channel and floodplain areas. Section 3.4.10 will show an assessment of the data used, helping to place the modelling in the context of scale issues. Conclusions will be drawn in Section 3.4.11 on the utility and effectiveness of the level two model.
3.4.2 – Choice of Model Code

After a full review of model types and codes (Section 3.2), the selection of rainfall-runoff and hydraulic model codes used in the current work will be outlined and justified below.

3.4.2.1 – Rainfall-Runoff Model

The software chosen to model the rainfall to streamflow conversion was the Probability Distributed Model (PDM), the theoretical background to which is described by Moore (1985; 1999; 2007). PDM forecasts flow from rainfall at the basin scale (Moore, 1999), by estimating the distribution of water stores in the soil profiles across the catchment using a probability distribution (Moore, 1985). This distribution is key to calibration, as a distribution incorporating deeper stores will enable more water to be stored in the conceptual soil column and so delay its conversion to flow. The water is then transmitted through different pathways (fast surface runoff, slower through flow and high lag time baseflow), the composition of which is determined by calibrated parameters (Moore, 1999).

The model is designed for operation at a short time interval (sub hourly), and observed flow measurements can be incorporated so as to allow calibration. An auto-calibration function is available which reduces the objective function, a quantitative measure of difference between observed and modelled flows (CEH, 2005b). This function is used during the final phase of calibration after manual calibration has taken place, and ‘fine tunes’ parameters within a small range of the manually calibrated values, allowing the minimisation of the objective
function and maximisation of fit. The objective function chosen was the root mean square error (rmse), which gives reasonable weight to errors at high flows (CEH, 2005b), which supports the primary application of assessing the impact during flood events.

In practice, an empirical rainfall-runoff model attempts to match the rainfall profile to the modelled flow via mathematical relationships. These relationships will be based upon the physical processes occurring in the catchment, and parameters calibrated represent physical characteristics of the catchment. Examples of these are the maximum and minimum depth of soil stores, and magnitude of groundwater recharge. It is therefore important for successful model extrapolation that these parameters are as close as possible to those characteristics they represent. However, as no data are being used in order to make these estimates, it is important that sensible estimations are made based on user-experience.

PDM is widely used among the contemporary hydrological sciences community. Kay et al. (2006a; 2006b) used a simplified version of PDM to assess climate change impacts on UK flood frequencies; Young et al. (2006b) utilised PDM to provide a tool for the prediction of ungauged catchments in the UK. The version of PDM used in the current work will be v1.50, incorporated into the InfoWorks RS v7.51 model shell. Later versions are available separately, yet in a format less compatible with InfoWorks RS.
3.4.2.2 – Hydraulic Model

As the exclusive application is as a catchment scale model, the 1D approach is entirely appropriate. More detail would require spatially distributed inputs and subsequently an overwhelming data collection programme. Data for a 1D model could be sourced from existing surveys and previous research projects on the same area. A 1D hydraulic model will not model subsurface storage volumes, but these have been assessed in the level 1 model as less important at Otmoor, particularly in winter months when water levels are high. Furthermore, the level 1 model has concluded that evaporation is not a major factor in determining wetland hydrological functioning at timescales as small as flood events, as it accounts for only a relatively small amount of the volume of an event. The catchment scale nature will also average out any effects evaporation might have. As such, the level 2 model is not required to incorporate evaporation at a significant level of detail. It is therefore included in the rainfall-runoff model as a standard yearly sine curve with an annual average of 2 mm/day\(^1\), a value used industry-wide (Bell, pers. comm.).

The software chosen was InfoWorks RS v7.51, (hereafter IWRS; Wallingford Software Ltd., Wallingford, UK). Based on the ‘ISIS’ model engine, the code is widely used in contemporary river hydraulic studies (e.g. Acreman et al., 2002a; McKenzie and Craig, 2001), and utilises the St.-Venant approximations of flow (see Section 3.2.3). These are adequate at this level of detail, and when used with a high quality DEM, provide the potential for highly accurate flood mapping (Lamb, 2007).
IWRS is regarded as an industry standard for generic river hydraulic modelling, and its applications are widespread. IWRS is also well established within CEH as a modelling environment, and known to be reliable and flexible: the word of colleagues is reassuring. IWRS is also a standard and commonly used tool at the UK Environment Agency; see Section 3.2.2 for examples of IWRS application in research.

PDM is included within the IWRS software package, allowing expenditure on model codes to be minimised. Despite this, development of the two models was undertaken separately, with the final PDM rainfall-runoff component providing boundary conditions for the hydraulic model component, as shown in Figure 3.14. This systematic approach ensures independent calibration of the rainfall-runoff and hydraulic model components.

![Conceptualisation of the model construction](image)

*Figure 3.14 – Conceptualisation of the model construction*
3.4.3 – Rainfall-Runoff Modelling

3.4.3.1 - Introduction

It is essential that any hydraulic model has a rainfall-runoff component that converts rainfall across the target catchment to river flow in the channel. This is done for all sub-catchments in the target river basin which contribute flow to the main channel. It is then the hydraulic model which calculates the flow velocities and levels within the channel to accurately predict river flow at any point in the river.

The objective of the rainfall-runoff model is therefore to predict flows for the gauged and ungauged sub-catchments across the Cherwell catchment. The model will be calibrated for each gauged sub-catchment. These calibrations will be used to produce boundary conditions for all gauged and ungauged sub-catchments, as described later in the section.

Calibration was to be done not only to maximise fit between observed and modelled flows, but so as to embed confidence with the user for extrapolated events. Therefore, parameter values that are as realistic as possible will be used.

3.4.3.2 – Model Construction

Data were provided by the EA as part of a memorandum of understanding with CEH. 15 minute flow and level data were available for six gauging stations across the River Cherwell catchment, as shown in Figure 3.15. These were the
River Cherwell at Banbury, Enslow and Oxford; the River Ray at Grendon Underwood and Islip; and the River Sor at Bodicote. Data were provided from 1997 to 2003 inclusive, and continuity was good with only limited periods of poor coverage. 15 minute rainfall data were available from two recording rain gauges, again shown on Figure 3.15. These were at Grimsbury water treatment works to the southeast of Banbury, and Bicester. As with the river flow data, continuity was generally good within the 1997-2003 coverage provided, as shown in Table 3.12. A further rain gauge is available at Osney, Oxford, but this was seen as too far from the river flow gauges to be utilised in the rainfall-runoff modelling, as well as lying outside the Cherwell catchment.

![Figure 3.15 - Location of flow and rainfall gauges; see Table 3.12 for key](image-url)
<table>
<thead>
<tr>
<th>Gauge Type</th>
<th>Key</th>
<th>Gauge Location</th>
<th>Data Coverage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flow</td>
<td>1</td>
<td>Cherwell at Banbury</td>
<td>99.8</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>Sor at Bodicote</td>
<td>99.9</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>Ray at Grendon Underwood</td>
<td>87.1</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>Cherwell at Enslow</td>
<td>99.9</td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>Ray at Islip</td>
<td>93.5</td>
</tr>
<tr>
<td></td>
<td>6</td>
<td>Cherwell at Oxford</td>
<td>75.8</td>
</tr>
<tr>
<td>Rainfall</td>
<td>1</td>
<td>Banbury</td>
<td>98.2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>Bicester</td>
<td>98.9</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>Oxford</td>
<td>98.8</td>
</tr>
</tbody>
</table>

Table 3.12 - Key to Figure 3.15; coloured labels refer to gauge labels

Examination of the data from the River Ray at Islip revealed an unexpectedly flashy response at the gauging station. Several reasons might cause an increase in the height of flow ‘spikes’ in data (Marsh, NRFA, pers. comm.), including sediment blocking ultrasonic transducers, floodplain flow or pulsation from a nearby weir (there is one located just upstream of the gauging station).

One method of verifying the data is to compare runoff with a catchment of similar geography (primarily geology, land use and size). Runoff is the notional depth of water in millimetres over the catchment equivalent to the mean flow as measured at the gauging station (Marsh and Lees, 2003). The River Thame at Wheatley

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provides a suitable comparison, and was known by the National River Flow Archive (NRFA, based at CEH Wallingford) to be a reliable gauging station. Runoff was calculated from 15 minute data using Equation 3.9 (Figure 3.16).

\[
r = \frac{0.9Q}{A}
\]

Equation 3.9 (from Marsh and Lees, 2003)

Where:

- \( r \) = runoff, mm
- \( Q \) = discharge, \( m^3s^{-1} \)
- \( A \) = catchment area, \( km^2 \)

The results of the comparison of runoff for the Ray at Islip with the Thame at Wheatley show a broadly analogous response between the two catchments. The River Thame has one large peak towards the end of the year not seen in the Ray, but some differences are to be expected as rainfall will vary to some extent between the catchments. Generally, the runoff is similar in both magnitude and response time between the catchments. This verifies, at a general level, the Islip data as suitable for use as observed data for calibration of the model 2 output.

However, some doubts remain for the Islip data at very high flows, due to a known phenomenon at the site of severe backing up from the confluence with the River Cherwell. As the stage of Cherwell rises at Islip, the Ray fails to discharge its water into the larger channel, and indeed under extreme
circumstances there may be a negative discharge in the Ray at Islip (water flowing from the River Cherwell upstream into the River Ray).

Table 3.13 shows the events for which the rainfall-runoff model was calibrated. The dates were chosen to represent the largest flood events during the period of data availability. The hydraulic model was primarily to be used to model the effect of flood events across the Cherwell catchment, and so the rainfall-runoff model must also be calibrated primarily for flood events.
Low flow prediction is important, however, and so the events also contain low flow periods. Furthermore, PDM is more stable during the calibration process when the events begin with low flow (for the calculation of the baseflow component of river flow).

Optimally, all sub-catchments would have been calibrated with the same events. However, for the Easter 1998 floods no data were available for the River Ray at Grendon Underwood. Also, event 3 was not a high magnitude event at this station. Therefore events four and five were chosen to supplement event one here.

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Start Date</th>
<th>End Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>06/12/1997</td>
<td>15/02/1998</td>
</tr>
<tr>
<td>2</td>
<td>01/04/1998</td>
<td>15/05/1998</td>
</tr>
<tr>
<td>3</td>
<td>01/10/2000</td>
<td>01/02/2001</td>
</tr>
<tr>
<td>4</td>
<td>23/11/1999</td>
<td>10/01/2000</td>
</tr>
<tr>
<td>5</td>
<td>28/03/2000</td>
<td>03/05/2000</td>
</tr>
</tbody>
</table>

Table 3.13 – Rainfall-runoff model calibration events

As previously noted, Refsgaard and Storm (1995) observed that the number of parameters changed during calibration should be kept to a minimum. Therefore a modeller should choose those parameters which have less certainty
associated with them and which are known to have been sensitive in previous models (Bell, pers. comm.). PDM has the flexibility to allow certain parameters to be left out of the calibration procedure at any stage. Parameters which are unknown can be changed to provide the best fit, and these values are only changed within a feasible parameter space. These boundaries were chosen from those suggested in the software literature (CEH, 2005a), and in collaboration with software developers (Bell, pers. comm.). Examples of these parameters are the maximum soil storage capacity depth and the baseflow time constant; Table 3.14 describes each variable (CEH, 2005c).

Parameters which are thought to be representative of catchment characteristics, or which provide final stage adjustment factors, are initially excluded from modification during calibration and are changed after those described above. Examples of these parameters are the rainfall factor and a time delay; both of which adjust for unrepresentative rain gauge locations. The Ray at Islip required minimal changing of these parameters, the Ray at Grendon Underwood slightly more so due to the increasing distance from the Bicester rain gauge. For the remaining stations only negligible changes were necessary.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Unknown/Final stage</th>
<th>Details</th>
<th>Starting value</th>
<th>Min</th>
<th>Max</th>
</tr>
</thead>
<tbody>
<tr>
<td>rainfac</td>
<td>Rainfall factor</td>
<td>Final stage</td>
<td>Adjusts rainfall volume</td>
<td>1</td>
<td>0.45</td>
<td>1</td>
</tr>
<tr>
<td>cmin</td>
<td>Minimum store capacity (mm)</td>
<td>Final stage</td>
<td>Affect time</td>
<td>0</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>cmax</td>
<td>Maximum store capacity (mm)</td>
<td>Unknown</td>
<td>and onset of runoff and rate of wetting</td>
<td>75</td>
<td>5</td>
<td>250</td>
</tr>
<tr>
<td>b</td>
<td>Spatial variability of store</td>
<td>Final stage</td>
<td>wetting</td>
<td>0.5</td>
<td>0.1</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>capacity function</td>
<td></td>
<td>up</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>be</td>
<td>Actual evaporation function</td>
<td>Unknown</td>
<td>Changes level of disparity between seasons</td>
<td>2.5</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>k1</td>
<td>Reservoir time constant</td>
<td>Unknown</td>
<td>Controls hydrograph peakiness</td>
<td>10</td>
<td>1</td>
<td>100</td>
</tr>
<tr>
<td>kb</td>
<td>Baseflow time constant</td>
<td>Unknown</td>
<td>Controls length of recession</td>
<td>50</td>
<td>0</td>
<td>500</td>
</tr>
<tr>
<td>kg</td>
<td>Groundwater recharge time</td>
<td>Unknown</td>
<td>Controls aquifer recharge rate</td>
<td>(10^5)</td>
<td>0</td>
<td>(10^5)</td>
</tr>
<tr>
<td>bg</td>
<td>Recharge function</td>
<td>Final stage</td>
<td>Controls sensitivity of recharge rate to soil dryness</td>
<td>1.5</td>
<td>0</td>
<td>15</td>
</tr>
<tr>
<td>tdlv</td>
<td>Time delay</td>
<td>Final stage</td>
<td>Shifts hydrograph along time axis</td>
<td>0</td>
<td>0</td>
<td>20</td>
</tr>
</tbody>
</table>

Table 3.14 – Rainfall-runoff model parameters within PDM
3.4.3.3 – Calibration

The calibration procedure consisted of a manual adjustment of parameters, followed by utilisation of PDM's automatic calibration function, as used by Young et al. (2006). The first stage of this process reduces visible error between observed and modelled flow, trading off different aspects of the model fit (CEH, 2005a). Subsequently an automatic algorithm is employed to search feasible parameter space in order to minimise an objective function (CEH, 2005a; Young et al., 2006) and maximise model fit, with final values shown in Table 3.15. Table 3.16 summarises the statistical fit for each station and for each calibration event. Figures 3.12 to 3.15 show the observed and modelled flow for each of the stations.

Two measures of fit will be used to assess model performance. The coefficient of determination \( r^2 \) is a statistical measure of how well the modelled values for flow and stage approximate the observed data points, and so is a simple dimensionless measure of correlation. It is a measure of the proportion variability of observed output that is accounted for by modelled output, and so provides a useful indication of how closely the two datasets relate to each other.

The root mean square error (RMSE) is a measure of the differences between values predicted by the model and the observed values. The method aggregates residuals into a single measure of predictive power, and uses the same units as the data (so m³s⁻¹ for example). The RMSE will be lower for a better fit in contrast to a higher value for \( r^2 \). RMSE is more sensitive than some
other measures to large residual points: the squaring process gives disproportionate weight to very large errors. For the current application, this is appropriate as a good fit is required in all sections of the hydrograph but especially during flood peaks, when model error and so residuals are likely to be higher.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Cherwell at Banbury</th>
<th>Ray at Grendon Underwood</th>
<th>Ray at Islip</th>
<th>Sor at Bodicote</th>
</tr>
</thead>
<tbody>
<tr>
<td>rainfac</td>
<td>0.98</td>
<td>0.85</td>
<td>0.91</td>
<td>1.00</td>
</tr>
<tr>
<td>cmin</td>
<td>6.37</td>
<td>3.09</td>
<td>0.05</td>
<td>0.00</td>
</tr>
<tr>
<td>cmax</td>
<td>66.32</td>
<td>131.49</td>
<td>97.57</td>
<td>247.83</td>
</tr>
<tr>
<td>b</td>
<td>0.46</td>
<td>0.71</td>
<td>0.99</td>
<td>0.50</td>
</tr>
<tr>
<td>be</td>
<td>1.55</td>
<td>3.66</td>
<td>3.27</td>
<td>1.01</td>
</tr>
<tr>
<td>k1</td>
<td>14.35</td>
<td>9.22</td>
<td>75.45</td>
<td>18.84</td>
</tr>
<tr>
<td>kb</td>
<td>43.07</td>
<td>256.10</td>
<td>72.20</td>
<td>266.96</td>
</tr>
<tr>
<td>kg</td>
<td>49557</td>
<td>98187</td>
<td>49864</td>
<td>31630</td>
</tr>
<tr>
<td>bg</td>
<td>2.17</td>
<td>1.15</td>
<td>1.54</td>
<td>1.50</td>
</tr>
<tr>
<td>tdly</td>
<td>0.01</td>
<td>0.01</td>
<td>0.00</td>
<td>0.00</td>
</tr>
</tbody>
</table>

Table 3.15 – Final calibration parameters for rainfall-runoff model
Figure 3.17 – Cherwell at Banbury calibration events
Figure 3.18 - Ray at Grendon Underwood calibration events
Figure 3.19 - Ray at Islip calibration events
Figure 3.20 - Sor at Bodicote calibration events
During calibration, several issues were highlighted that require further analysis. These are described in the following subsections.

### 3.4.3.4 - Effect of Rainfall Station Distribution

Figure 3.10 shows that the two rain gauges are distributed well across the Cherwell catchment. The Banbury station is close to the river flow stations at both Banbury and Bodicote; the Bicester station is close to the two river flow stations on the River Ray.

To objectively investigate the representation of the two rain gauges, a double mass curve (Wilson, 1974) was plotted (Figure 3.21), covering data from the years 1997 to 2003. This method plots the cumulative rainfall at one station against that of another, and any previously unnoticed effects of rain gauge readings (poor location, unit error etc) would be highlighted. It is clear that the
cumulative rainfall totals are very close; $r^2$ values between the two are >0.999, and it may be concluded that both rain gauges are reliable.

Moore et al. (2006a) evaluated the method of coupling point rain gauge data with more spatially distributed radar data for use with rainfall-runoff models, concluding that standalone rain gauge data perform well with the PDM software. With a fully distributed model, such as G2G (see Section 3.2.2), benefits can be seen from having a greater spatial distribution of rainfall, but for applications such as the current work, rain gauge data alone provides high quality boundary conditions for a model (Moore et al., 2006a).

Figure 3.21 – Double mass curve for Bicester and Banbury rain gauges, daily data 1997-2003
As the rainfall data represent the spatial coverage of river flow data well and are reliable, there is no need to construct a catchment average hyetograph. This would combine recording rain gauge data with daily (total rainfall) data closer to the location required (Jones, 1983). Such a procedure would be required if the rain gauge locations were some distance from the river flow gauges or coverage was poor.

The two recording rain gauges used in this study will obviously have different profiles for each event. To investigate the sensitivity of the calibrated model to differences in rainfall, a simple test was run. Using the calibrated model for the Ray at Islip, the input data was changed from the station used to calibrate the model (and for which data will be used to run the model to produce boundary conditions for the hydraulic model), Bicester, to the alternative gauge at Banbury. The effect of a different rainfall profile is shown visibly in Figure 3.22. The alternative rain gauge (Banbury, Figure 3.22c) had a very similar profile to the default rain gauge (Bicester, Figure 3.22b) up to approximately 1,700 hours during the event, when rainfall intensity was considerably higher at Banbury. This has an obvious effect on the flow at the Islip gauging station in the hours following this rain. This finding justifies the use of two rain gauges in the Cherwell catchment. Were the rainfall-runoff model to be calibrated on only one rain gauge, river flow gauges at the extremities of the catchment, such as Banbury in the north, would have a less reliable output than with a partially distributed rainfall input.
3.4.3.5 – Effect of Initial Conditions

The importance of initial conditions to the success of model outputs is required to be investigated, as initial conditions for ungauged catchments and theoretical events will be unknown. The PDM software takes initial conditions during a calibration session as a measure of baseflow, and therefore, it is important to begin calibration events with low flow conditions.

During calibrated runs, such as those used to produce boundary conditions for the hydraulic model, the initial condition is important. If not correct, the model
will take a long time, possibly six months to a year or more to recover sufficiently to represent realistic flow in the river system based on experience. A sensitivity exercise was undertaken to help identify the importance of initial conditions. The Sor at Bodicote was run using an extended event 4 (15/10/1999 to 31/01/2000), so as to create varied conditions. The output of the calibrated model was compared with initial conditions taken from the following:

- observed initial conditions,
- 25% higher than observed,
- 50% higher than observed and
- 50% lower than observed.

The model outputs are shown in Figure 3.23. It is clear that the initial conditions have an important impact upon model output accuracy, as shown by $r^2$ values: fit decreases as initial conditions vary from observed. The effects of initial condition choice are dampened as the event progresses. This effect can be explained by the model converging better after a period of 'warming up'. It was thought necessary to investigate the effects of a warm-up period to confirm this.

The absolute error of each scenario from the observed is shown in Figure 3.24, confirming convergence after model warm-up. It also shows that errors are greatest during periods of high flow. These findings are important as the rainfall-runoff model will be used where the initial conditions are not known, and furthermore the model will also be predicting flow for ungauged sub-catchments and for theoretical events.
Figure 3.23 - Testing initial conditions (IC) of rainfall-runoff model

Figure 3.24 - Absolute errors of rainfall-runoff model with different initial conditions
3.4.3.6 – Effect of Warm-Up Periods

Theoretically a warm-up period allows the model to reach equilibrium, and thus provide more realistic conditions, before the required simulation period begins. It was deemed necessary to investigate the effect of a warm-up period on the model output. Warm up periods allow initial conditions to vary from observed before the target period begins, and initial conditions are crucial to the model run (Bell, pers. comm.).

A test event was initialised, using an extended event 4 (as above), in order to experiment with different warm-up periods. The gauging station at the Sor at Bodicote was chosen at random to investigate this, and the calibrated model was run with different warm-up periods ranging from zero to 180 days. Figure 3.25 shows the modelled output with warm up periods of zero, one day, 30 days and 180 days (observed initial condition used for all).

![Figure 3.25 - Testing warm-up period for rainfall-runoff model](image.png)
Figure 3.26 shows graphically the absolute error (3.26a) of each of these warm-up scenarios from the observed flow (3.26b). It is clear that a short warm-up period has little or no effect on modelled output. The 1 day warm-up trace follows almost exactly the zero-warm-up trace in Figure 3.25. A longer warm up period increases the difference from zero-warm-up. This may be explained by the initial conditions becoming increasingly diverged from the observed initial condition. It must be assumed that this observed initial condition would give maximum accuracy.

---

**a - Absolute Errors From Observed**

![Graph showing absolute errors from different warm-up periods](image)

**b - Sor at Bodicote Observed**

![Graph showing Sor at Bodicote Observed flow](image)

Figure 3.26 – Absolute errors from different warm-up periods (a), with Sor at Bodicote Observed flow (b)
This finding emphasises the importance of careful estimation of initial conditions. Also of note is the convergence of different warm-up periods as the event continues, a feature clearly shown in Figure 3.25, where the traces converge in the latter stages of the event. The absolute errors from observed (Figure 3.26b) also converge.

However, there is little absolute difference in error between a short and particularly long warm-up period. Shorter warm-up periods have a lower absolute error from observed flow, as the initial conditions are set by the observed flow and not allowed to diverge when the event begins. Without this difference in initial conditions, there would not be any difference between runs using different warm-up periods. It is therefore deemed unnecessary to have a warm-up period for gauged catchments. It can be seen that accurate estimation of initial conditions is more important than allowing the model to warm up.

### 3.4.3.7 - Ungauged Catchments

Not all sub-catchments in the Cherwell catchment being modelled are gauged. The rainfall-runoff model is required to predict flows for these, and there are several methods that can be used for transfer of a calibrated model to uncalibrated sites, as summarised by Moore et al. (2006b). The most straightforward is the simple scaling method, whereby the nearby gauged (and so calibrated) forecast is subject to adjustment, usually by catchment area, for the target location. Secondly, a model transfer would involve confidence that the catchments are so similar (including area, terrain, soil and geology) that scaling...
for catchment area difference is unnecessary. Next, it is also possible to relate model parameters to catchment properties, and establish regression relationships between the two in order to estimate possible parameter modification for the target catchment. More complex again, site similarity approaches use a measure of site similarity which is compared to a pooling group of gauged sites, and parameters modified with a distance-weighting function (Moore et al., 2006b).

The method adopted in the current work will use a simple scaling methodology, factoring by area, but with choice of surrogate catchment determined by analysis of soil type and geology. An assumption is made that the flow characteristics will be similar, and the primary disparity will be the catchment size difference, affecting the amount of runoff produced. Lamb (2007) concludes that only where real-time forecasting (short term prediction) is required from a model is there a tangible benefit in fully distributed rainfall-runoff modelling.

The only option available to validate the method of rainfall-runoff modelling for ungauged catchments is to use the methodology to predict for a sub-catchment for which data are available. This may be done for a catchment within the gauged model, only as long as the modelled hydrograph remains independent from the observed data. For this procedure the Sor at Bodicote and Ray at Islip gauging stations were utilised: firstly the Sor was treated as an ungauged sub-catchment and the Ray as a surrogate, and subsequently vice versa. This method simply multiplies the model’s flow output by the appropriate area factor.
As shown earlier, initial conditions are crucial to model success. In order to methodically calculate initial conditions for ungauged catchments it has been assumed that a factoring by area of the initial condition from the surrogate sub-catchment will be appropriate. For example, if using the Ray at Islip (290.10 km\(^2\)) as a surrogate for Bayswater Brook (18.58 km\(^2\)), the observed flow at the start of the modelled event will be factored by the proportionate area (0.064) to calculate an initial condition for Bayswater Brook.

Figure 3.27 shows the validation procedure for transforming the Sor at Bodicote (87.7 km\(^2\)) modelled flow to the Ray at Islip (209.1 km\(^2\)). Figure 3.28 shows the validation procedure for transforming the Ray at Islip modelled flow to the Sor at Bodicote. Fit is not as high as during calibration; this is to be expected as the model parameters have not been calibrated for these catchments. Flow is largely predicted within the magnitude expected, although flood event peaks have a higher error, as shown in Figure 3.27 and 3.28.

As these catchments both have observed flow, it was possible to have both a factored (by area) initial condition and an observed initial condition. It is clear that the factored method performs better, yet only marginally for the Ray to Sor conversion. This implies that the procedure for predicting flow for ungauged sub-catchments is robust, even when using factored initial conditions.
Figure 3.27 – Rainfall-runoff model validation: Sor to Ray

Figure 3.28 – Rainfall-runoff model validation: Ray to Sor
This methodology depends highly upon the choice of surrogate catchment for the ungauged site. Several measures of similarity could be used, including similarity of catchment area, land use and distance between catchments. Most important, however are the geological characteristics, which reflect most strongly in the characteristic hydrograph for the catchment by changing flow routing pathways (Bell, pers. comm.; Lamb, 2007).

Data for deep and superficial geology were obtained from the British Geological Survey. These were incorporated into a Geographic Information System (GIS) for the Cherwell catchment in order that surrogate sub-catchments be assigned from visual inspection of similarity.

The catchment of the Cherwell at Oxford and that of its sub-catchments are shown in Figure 3.29. Table 3.17 references the numbers in Figure 3.29 to the sub-catchments. Note that the Ray at Grendon Underwood is only used as a surrogate, and not needed in the model as it is incorporated into the Ray at Fencott Bridge.
Figure 3.29 - Subcatchments in the Cherwell; see Table 3.17 for key to area numbers

Figure 3.30 shows the superficial geology with sub-catchments overlaid. It is clear that little further than the extensive alluvial deposits across the River Ray floodplain to the south of the catchment are shown, as there are few significant superficial deposits in the catchment. This confirms that the River Ray gauging station at Grendon Underwood has a similar superficial geology to that at Fencott Bridge downstream. Figure 3.31 shows the deep geology; again, the Ray at Grendon Underwood has a similar coverage for the remainder of the Ray, and so using this as a surrogate for the Ray at Fencott Bridge seems appropriate.
<table>
<thead>
<tr>
<th>Number</th>
<th>Sub-catchment</th>
<th>Modelled</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Cherwell at Banbury</td>
<td>Yes</td>
</tr>
<tr>
<td>2</td>
<td>Farthinghoe Stream</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Sor</td>
<td>Yes</td>
</tr>
<tr>
<td>4</td>
<td>Unspecified 1</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Charlton and Kings Brooks</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Ockley Brook</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Swere</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Deddington Brook</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>Unspecified 2</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>Ray at Grendon Underwood</td>
<td>Yes</td>
</tr>
<tr>
<td>11</td>
<td>Ray at Fencott Bridge</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Ray at Islip</td>
<td>Yes</td>
</tr>
<tr>
<td>13</td>
<td>Unspecified 3</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>Unspecified 4</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Bayswater Brook</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.17 – Key to Figure 3.29

Bayswater Brook has a large coverage of Oxford clay (green), extending from the Ray catchment. Other coverage in the Bayswater Brook catchment is limestones, which are also represented in the Ray, although to a lesser extent. The catchment bears little similarity to others in the northern reaches of the Cherwell, and so it seems most appropriate to use the River Ray at Islip as a surrogate for Bayswater Brook.
Bedrock geology in the Deddington Brook catchment is dominated by Charmouth mudstone (purple). It is clear that the same is true for the Cherwell at Banbury. Secondary coverage in both includes Whitby mudstone (brown), and therefore the Cherwell at Banbury is to be used as a surrogate for Deddington Brook. Farthinghoe Stream also shares much of the geological properties with the
Cherwell at Banbury. Its geology is again dominated by Charmouth mudstone (purple) and Whitby mudstone (brown), with some Marlstone limestone (red).

The ungauged River Swere, adjacent to the River Sor, shares many bedrock characteristics with its neighbour. Both contain extensive coverages of Marlstone limestone (red) and Whitby mudstone (brown). Although the upper reaches of the Swere contain some other variations of limestone (light green and
yellow), the Sor should be an excellent proxy for the Swere. Charlton Brook and Kings Brook were combined, as they confluence with the Cherwell less than 200 m from one another, and were again similar in geological structure to the Sor.

For Ockley Brook the story is different. The majority of the catchment is underlain by White limestone, which is sparse across the remainder of the Cherwell catchment. Although some White limestone lies within the Ray at Islip watershed, this catchment is overwhelmingly dominated by mudstone, and so is inappropriate for representing Ockley Brook. Most appropriate seems to be the Sor at Bodicote, owing to the Charmouth mudstone (purple) and Whitby mudstone (brown) in the lower reaches of Ockley Brook. As the Ray upstream of Fencott Bridge is included with the model for that area, the remaining Ray catchment, with a contributing area between Fencott Bridge and Islip, is added downstream, just upstream of Islip. The Ray at Islip is an obvious surrogate for this input, which is labelled as Lower Ray.

The remainder of the catchment lies outside established subcatchments, but does contribute to the river flow. This area, predominantly adjacent to the river channel, has been discretized into areas labelled Unspecified 1-4, and are added at their downstream extents, with the most appropriate rainfall-runoff model used as a proxy. Table 3.18 shows the final surrogate sub-catchments for the ungauged sites. The two lower areas were judged to be most similar to the Ray at Islip, the most northerly to the Cherwell at Banbury, and the remaining area in the central region was modelled using the Sor at Bodicote.
<table>
<thead>
<tr>
<th>Ungauged Sub-catchment</th>
<th>Surrogate catchment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ray at Fencott Bridge</td>
<td>Ray at Grendon Underwood</td>
</tr>
<tr>
<td>Bayswater Brook</td>
<td>Ray at Islip</td>
</tr>
<tr>
<td>Deddington Brook</td>
<td>Cherwell at Banbury</td>
</tr>
<tr>
<td>Farthinghoe Stream</td>
<td>Cherwell at Banbury</td>
</tr>
<tr>
<td>Swere</td>
<td>Sor at Bodicote</td>
</tr>
<tr>
<td>Ockley Brook</td>
<td>Sor at Bodicote</td>
</tr>
<tr>
<td>Charlton and Kings Brooks</td>
<td>Sor at Bodicote</td>
</tr>
<tr>
<td>Unspecified 1</td>
<td>Cherwell at Banbury</td>
</tr>
<tr>
<td>Unspecified 2</td>
<td>Sor at Bodicote</td>
</tr>
<tr>
<td>Unspecified 3</td>
<td>Ray at Islip</td>
</tr>
<tr>
<td>Unspecified 4</td>
<td>Ray at Islip</td>
</tr>
<tr>
<td>Lower Ray</td>
<td>Ray at Islip</td>
</tr>
</tbody>
</table>

Table 3.18 – Surrogate catchment assignments

3.4.3.8 – Validation

A formal validation procedure was undertaken to assess the soundness of the calibration procedure and its results. In order to be independent from the calibration procedure, the validation used event data that were not used during calibration. Although the events of highest magnitude were used for calibration, large events remained. These are shown in Table 3.19, and are labelled as events six and seven.

The results for the validation procedure for all calibrated stations are shown in Figures 3.32 to 3.35. A summary of the statistical fit for all events (calibration
and validation) at all stations is shown in Table 3.20. The fit compares favourably with that obtained by a similar study using the PDM model code on the River Cherwell (Moore et al., 2006a) which averaged an $r^2$ of 0.74.

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Start Date</th>
<th>End Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>01/01/2002</td>
<td>05/04/2002</td>
</tr>
<tr>
<td>7</td>
<td>05/10/2002</td>
<td>15/01/2003</td>
</tr>
</tbody>
</table>

Table 3.19 – Rainfall-runoff model validation events

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Cherwell at Banbury</th>
<th>Ray at Grendon Underwood</th>
<th>Ray at Islip</th>
<th>Sor at Bodicote</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$r^2$</td>
<td>RMSE</td>
<td>$r^2$</td>
<td>RMSE</td>
</tr>
<tr>
<td>1</td>
<td>0.93</td>
<td>2.25</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2</td>
<td>0.88</td>
<td>3.94</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>3</td>
<td>0.78</td>
<td>1.99</td>
<td>0.65</td>
<td>0.35</td>
</tr>
<tr>
<td>4</td>
<td>N/A</td>
<td>N/A</td>
<td>0.53</td>
<td>0.40</td>
</tr>
<tr>
<td>5</td>
<td>N/A</td>
<td>N/A</td>
<td>0.76</td>
<td>0.41</td>
</tr>
<tr>
<td>Average</td>
<td>0.862</td>
<td>2.73</td>
<td>0.65</td>
<td>0.39</td>
</tr>
<tr>
<td>6</td>
<td>0.734</td>
<td>1.01</td>
<td>0.65</td>
<td>0.21</td>
</tr>
<tr>
<td>7</td>
<td>0.64</td>
<td>3.76</td>
<td>0.65</td>
<td>0.44</td>
</tr>
<tr>
<td>Average</td>
<td>0.68</td>
<td>2.39</td>
<td>0.65</td>
<td>0.325</td>
</tr>
</tbody>
</table>

Table 3.20 – Rainfall-runoff model fit for calibration and validation events
Figure 3.32 – Cherwell at Banbury validation events

Figure 3.33 – Ray at Grendon Underwood validation events
Figure 3.34 - Ray at Islip validation events

Figure 3.35 - Sor at Bodicote validation events
3.4.3.9 – Rainfall-Runoff Model Conclusions

PDM, a lumped model code, was chosen to model rainfall-runoff for the target catchment. This was done with high temporal resolution data supplied by the EA. In order to have confidence with the prediction of theoretical events and those predicted by climate change scenarios, it was essential that the calibration procedure was undertaken not merely to gain greatest fit between observed and modelled flows, but to provide realistic parameters reflecting catchment hydrological properties. High $r^2$ values were obtained during calibration and these were validated successfully, again with high $r^2$ values which compared well with those obtained from similar studies (e.g. Moore et al., 2006a).

Several issues were highlighted during calibration that required further investigation. Subsequently, the user may have a high degree in confidence in using this calibrated model in synchronisation with a hydraulic model of the target catchment.
3.4.4 – Hydraulic Model Construction

Having validated the rainfall-runoff component, the next phase of composite model development is the construction of the hydraulic model component. To obtain calibration, several stages of development were required to be undertaken in order to stabilise the model and achieve fit with observed data. These include adding notional weirs to the Rivers Cherwell (standard height, determined by adjacent cross section elevation), and the Ray (calibrated height), adding cross sections to the Cherwell, adding a spill unit (calibrated height) to connect the Otmoor storage unit, and adding a Priessmann slot to the river channel. Each stage will be explained below.

Some changes that needed to be made to the model structure in order to achieve stability and reasonable run times require personal judgements to be made by the model developer. Bartlett (2007) discusses these judgements, commenting that they need to be explained and justified in order that the method is as repeatable as possible. He concludes that models are not designed to replace experience and judgement of a user, and that a hydrologist’s experience is better than any model 90% of the time.

3.4.4.1 – Importing Cross Section Data

Cross sections were obtained from the EA, based on surveys for the River Cherwell carried out primarily in 1993, and supplemented by later surveys in the north of the catchment (near Banbury) in 2002. Coverage of cross sections limited the spatial extent of the model. A coverage south of Banbury (78% of the total river length) was possible, with only the upper reaches not included in the
full hydraulic model. However, these upper reaches are included in the Cherwell at Banbury component of the rainfall-runoff model.

Cross section data from the EA were converted from an ‘eeby’ format to the *.dat format required for use in IWRS. This conversion was carried out by a program written specifically in Visual Basic language. The output provided a simple text file containing cross chainage and corresponding bed level at surveyed points across the river, markers for delineating channel banks and bed, and initial roughness values. There were 650 cross sections for the 60 km of river from Banbury to Oxford, an average of 92 m between cross sections. It is essential in a hydraulic model to provide enough density of cross sections downstream for stability of the model; otherwise there may be a large vertical drop between cross sections which may instigate model instability.

Data for cross sections on the River Ray are older, having been surveyed in the late 1970s. They were used in a previous modelling study of Otmoor (Acreman et al., 2002a), and were supplemented by contemporary surveys in 2001 by CEH staff. As they were used in a previous study using similar software, the format of the River Ray cross sections was already the desired *.dat format.

Each cross section typically covered between 30 and 40 m width in total, increasing as the river becomes wider downstream. Included were the river channel, banks, and the immediate adjacent land. As water levels easily rise above this level at high flows, these cross sections needed to be extended across the wider floodplain, and this process is outlined in Section 3.4.4.2.
A systematic processing of cross sections was required in order to make sure each was broadly representative of its corresponding reality. Often errors were incorporated from survey error, data input error, data conversion or data import. Where this error could be fixed manually, either by removing cross chainage points or interpolating between points, this was undertaken. Where cross section data was obviously erroneous and unable to be interpreted, the cross section was omitted from the model.

Structures such as bridges and weirs were too complex to be converted from the 'eeby' format into usable data, and such cross sections were also omitted from the model. Often in these areas, particularly where weirs were present on the river, an inspection of the long section (downstream cross section of the river) revealed many areas of sharp changes in slope. Smoothing of the long section is essential to gaining model stability, although obviously not where river slope is steep in reality. This is undertaken by inserting a notional weir (usually appropriate where weirs are missing from the model) or interpolating between existing cross sections, for which an automated algorithm exists within IWRS. Such methodology is standard practice among the IWRS user community, and is indicative of the need for a certain level of human judgement in the initialisation of hydraulic models (Bartlett, 2007). As such, all judgments made will be highlighted and explained in the text of this methodological description: it is anticipated that the method used will be fully reproducible as per contemporary scientific theory.
Some typical cross sections from the upper, middle and lower reaches are shown in Figures 3.36 a, b and c respectively. In the upper channel, near Banbury (locations shown in Figure 3.37), the channel is approximately 15 m wide with a well defined floodplain, likely Quaternary river terrace in nature, as shown from cross section number 1.102. Further downstream, cross section C1.039 is typical of the river geometry with a 15 m wide channel and wider floodplain component. In the lower reaches of the Cherwell towards Oxford, cross section E1.019 shows the significant size of the channel, now nearly 35 m across. The locations of the left and right banks and bed are also shown in Figure 3.31; these were specified in the data supplied by the EA. All three example cross sections described here will be used as examples again in Section 3.4.4.2 for extending model coverage across the floodplain.

Figure 3.36 – Example cross sections
3.4.4.2 – Extending Model Coverage

The extent of cross section data laterally across the floodplain was not sufficient to accommodate any flow significantly above the river’s banks. As this study is specifically investigating the effect of wetland systems on flood events, it was necessary to extend the model across the floodplain component of the river...
system. This was done using two very distinct methods for different areas of the catchment. For the majority of the channel, cross sections were simply extended across the floodplain using DEM data, as IWRS has functionality to automatically extend cross sections across a given elevation model. The DEM used was NEXTMap (see Section 3.3.2.3 for details), as coverage of the entire River Cherwell catchment was available and accuracy and resolution appropriate for such an application. The width of extension across the floodplain was determined by floodplain width, as measured from DEM coverage in the GIS, and applied in reaches of similar floodplain extent (Figure 3.38). The example

Figure 3.38 – Map of cross section (XS) extensions
cross sections shown in Section 3.4.4 (Figure 3.36 a-c) are shown in their extended form (Figure 3.39 a-c), and the effect of this method of cross section extension is immediately evident. The ability to model floodplain flow is evident from the extensions of the cross sections, and the limit of the floodplains are evident, showing the restricted nature of the pre-extended (surveyed) cross sections.

![Figure 3.39 - Example cross section extensions; Figure 3.37 above shows locations](image)

At Otmoor, the hydrology is complex, involving circular flow and depends intricately on very small scale subtleties across the floodplain's topography, leading to it not being appropriate for 1D model representation. The system was
modelled using the multi-channel 1D approach by Acreman et al. (2002a), and was very complex (many weirs and sluices), unrepeatable and yielded large uncertainties in output.

A solution to this problem was found by representing Otmoor as a single storage unit, similar to a reservoir, connected directly to the river. IWRS allows an area to be defined and storage volume for a given stage to be extracted from a DEM, a procedure which is fairly simple and so relevant to the catchment scale philosophy of the model. The Otmoor storage unit was linked to the River Ray with a spill unit, the height of which formed part of the calibration procedure (see Section 3.4.4.6). Having Otmoor represented as a storage area gave the resulting hydrograph of the River Ray a shape that was very close to that of the observed data, giving confidence that this was the correct route to take for Otmoor.

This methodology was therefore taken forward to be included in the calibrated and validated model, but a ‘cross section extension’ method (as used across the remainder of the Rivers Cherwell and Ray floodplains) of representing Otmoor was investigated later (Section 3.4.8.2) for completeness. Another investigation examined the effect of using the higher resolution and higher accuracy LiDAR DEM (see Section 3.3.2.3 for details) to calculate the storage volume in comparison to NEXTMap data (Section 3.4.8.1).
3.4.4.3 – Preissmann Slot

Model stability is often poor at low flows, when insufficient water is available to fill the model’s river channel network and some areas may become dry. This is common at the start of a model run, especially when initial conditions are low. A Preissmann slot, a small theoretical channel cut into the bed of the river (Figure 3.40), was added to the entire river network at an early stage of model development to enable flow of water at very low flows and so provide computational stability. The dimensions of the Preissmann slot are 0.1 m wide and 1 m deep, and it is not used during stages higher than a few centimetres above the bed (Wallingford Software, 2007). The addition of a Preissmann slot is standard procedure in the development of hydraulic models to achieve stability (Bartlett, 2007).

![Figure 3.40 – Example of a Priessmann slot applied to a model channel cross section](image-url)
3.4.4.4 – Adding a Weir and Spill Units to the River Cherwell

Notional weirs are added to the river where there is a steep section of thalweg, often caused by the absence of a conventional weir (see Section 3.4.4.1 for reasons). A weir, usually 0.1 m above the upstream cross section’s bed and a similar width to the cross section, holds up the flow and prevents water levels falling below the bed level, leading to instabilities. Importantly, at higher flows the notional weir has no effect on flow.

By adding a spill unit, a model user is able to bypass an unstable section of river as flow is effectively conveyed immediately from upstream of the difficult area to downstream of it. To compensate for the loss of length of river, a similar length is added to nearby river reaches. Figure 3.41 shows the locations of the spills (total of six) and a single notional weir on the river network.
3.4.4.5 – Adding Weirs to the River Ray

Weirs exist on the River Ray to manage water levels in the river and so hold back flow. They form an integral part of the water management of the Otmoor area, and so are essential to a hydraulic model of the area. Such weirs will have impacts on the flow at all discharges, and so a notional weir was inappropriate. A broad-crested weir was used (Bos et al., 1989), as this was the most simple form of non-notional weir available.

Experimenting revealed that the weirs added to the Ray had little bearing on the shape of the hydrograph, but consistently raised water levels upstream of the structure. It was therefore a matter of calibrating the height of the weirs to the baseflow level, as at low flows it is possible to see the minimum stage of the river, and this was possible on the weir located approximately 100 m downstream of the Islip gauging station. Another weir was located upstream, just below the Otmoor wetland area, raising water levels in the river through Otmoor. Here the weir height adjustments comprised a combination of information on river channel geometry, known water levels at Otmoor and examining the effect on flows at the gauging station site compared to observed values. The final calibrated heights for the upstream (close to Otmoor) and downstream (close to Cherwell confluence; see map Figure 3.42) weirs were 57.90 and 56.65 maOD respectively. A long section of the River Ray area of the model (Figure 3.43) reveals the effect on flow of the weirs during low flows in calibration event E.
Figure 3.42 – Map of Ray with weirs

Figure 3.43 – Long section of Ray with weirs; water level at low flow shown
3.4.4.6 – Adding Spill Unit to Otmoor Storage Area

Otmoor is represented in the model as a storage area, the volume of which is calculated from a DEM. The volume of the storage area itself is not calibrated, but the level of storage capacity utilised at any time is highly dependent on the connection between the storage area and the river network. The connection is a spill unit, effectively a weir-type object with editable cross section to allow changeable flow over the spill. Flow is two-way, so that water flows from the river to the reservoir when the river stage is higher than the reservoir stage and the spill height is reached; flow is reversed when the reservoir stage is higher than the river stage, draining the storage unit. The shape of the spill unit is important, as it determines how much water can be transferred at various stages, and is usually a wide ‘U’ shape, allowing increasing flow of water with increasing stage. This situation is consistent with the process being modelled, whereby the river will transfer more water to the floodplain (Otmoor) with a higher stage above the river banks.

A range of minimum spill heights were investigated using 0.1 m increments and utilizing information from the DEM about bank height as well as local knowledge gained during the field component of this study. Fit was assessed at the Islip gauging station for flow and stage; the best fit was given by a spill height of 58.0 maOD. This level appeared logical, as this corresponds to the lowest lying areas of Otmoor.
3.4.4.7 - Adding Cross Sections to the River Cherwell

Cross sections need to be added where there is a steep slope on the long section, in order to maintain a certain frequency of cross sections for each unit of vertical drop in river bed level downstream. Simple rules of thumb were adopted from Samuels (1990) including cross sections not being more than twenty times the channel width apart. Where necessary, IWRS can create further cross sections to account for insufficient numbers of surveyed cross sections by interpolating between the geometry of the cross sections up- and downstream of the problem area. This is an automated process, although user input is required on the number of cross sections required, and this is chosen on the basis of model output error reports from previous model runs. A total of 35 cross sections were required in 11 locations (Figure 3.44).

Figure 3.44 – Map of interpolated cross sections added to Cherwell
3.4.4.8 – Development of Oxford Flow-Stage Relationship

During the calibration process, a problem evolved whereby fit was difficult to obtain at Oxford for both flow and stage, the primary reason being a lack of full cross section data for the gauging station site. It was likely that flow was being predicted well, but the geometry of the cross section used for data output was thought to be affecting the fine detail of the river stage here. Therefore, it was decided to apply a flow-stage relationship to achieve optimum model output of both flow (direct model output) and stage (indirect calculation).

As flood events were the primary focus of the model, daily averages of flow were thought to miss the very high flow periods during large magnitude events, and on this basis, 15 minute data were used to derive the flow-stage relationship. Due to the large number of data points (35,040 per annum), only two years' data were used, but to provide a broad range of data, a relatively dry year (1999) and a relatively wet year (2000) were used. To supplement this at very high flows, data from the Easter 1998 floods were added to the dataset to provide a full range of flow and stage data.

A plot of the data (Figure 3.45) shows a particularly high range of stages for low flow. This is due to intense management of gates and weirs on the River Thames to manage flows, resulting in a wide range of backing-up conditions to the River Cherwell (Marsh, NRFA, pers. comm.). Although the Oxford gauging station is 0.93 km upstream of the Thames confluence, the flat topography of this floodplain area and subsequent low bed slope exacerbates the effects of backing-up.
There is also more variation than expected at high flows, in the form of several 'tails', seemingly from separate events. This is partly explained through hysteresis, where antecedent conditions may dictate stage at a given flow. Furthermore, inaccuracies in the measurement may cause inconsistencies for two reasons (Marsh, NRFA, pers. comm.). Firstly, this site utilises ultrasonic sensors within the channel, usually on the banks of the channel, which do not directly measure out of bank, or floodplain flow. Secondly, high sediment content in the river water, often associated with high flows, may disrupt ultrasonic communication and provide different output for a given flow.

![Figure 3.45 - Cherwell at Oxford flow-stage relationship; vertical dashed line shows separation between Equation 3.10 and 3.11 for fitted model](image)

A fit was applied to the data (also shown in Figure 3.45) comprising two components: a 3rd order polynomial for flows up to 47 m$^3$s$^{-1}$ (Equation 3.10) and
a linear fit for flows greater than 47 m$^3$s$^{-1}$ (Equation 3.11). This flow-stage relationship describes the data well, and delivers the broad themes of the low and high flows, whilst taking into account the known data from very high flows.

$$h = -0.009071Q + 0.001728Q^2 - 0.00002221Q^3 + 54.84$$

Equation 3.10

$$h = 0.005258Q + 55.678$$

Equation 3.11

Where:

\[Q\] = flow \hspace{1cm} \text{m}^3\text{s}^{-1}

\[h\] = stage \hspace{1cm} \text{maOD}

3.4.4.9. – Initial Conditions

Initial conditions are important for the stability of the model in the early stages of a model run (Bartlett, 2007). Of course it is impossible for a modeller to know the stage and flow in every cross section at the start of each event. For this reason initial conditions were created from the results of a steady state model run, whereby the model was run with the same start times in an iterative mode at the same time step until convergence of the code is achieved and the model is stable. The results were then used as initial conditions for the unsteady model run. Although this is not part of the calibration procedure per se, this is an important step in moving towards creating a stable ‘unsteady mode’ model which can be used as a tool. The use of steady state results is common practice in hydraulic modelling for creating initial conditions (Wallingford Software, 2007).
3.4.5 – Sensitivity Analysis

Calibration was primarily undertaken by changing the parameter of bed roughness. As Pappenberger et al. (2005) discuss, all hydraulic modelling packages focus on the calibration of roughness and channel geometry, which together are considered to have the most important impacts on modelled hydraulic flow. In the current study, channel geometry is considered as an observed parameter, and roughness as an unknown which requires calibration. Although the aim of parameter calibration is to obtain a reasonable value for each parameter, the value chosen is likely to incorporate uncertainty and error from other, non-calibrated parameters, together with that from boundary conditions and incorporated from assumptions made in order to make the modelling process feasible. Therefore a reasonable parameter space is often specified before calibration begins, in order that the calibration process does not change parameter values inappropriately.

Several different measures of resistance to flow are available, ranging from the attenuation parameter proposed by the 1975 NERC Flood Studies Report to the Darcy-Weisbach (Darcy, 1857) and Manning’s estimations of uniform flow. Manning’s $n$ was taken as the measure of roughness (Manning, 1891) due both to its simplicity ($n$ is dimensionless) and its ease of use within the software, IWRS. $n$ is derived from rearranging the following Equation which is used to estimate depth-averaged velocity in natural channels (Knighton, 1998):
\[
\bar{u} = \frac{R^{\frac{1}{3}} s^2}{n}
\]

Equation 3.12

Where:

\[
\begin{align*}
\bar{u} & = \text{depth-averaged velocity} = \text{ms}^{-1} \\
R & = \text{hydraulic radius} = \text{m} \\
s & = \text{channel slope} = \text{mm}^{-1} \\
n & = \text{roughness coefficient} = \text{dimensionless}
\end{align*}
\]

\( n \) is an estimate of resistance to flow (Chow, 1959). The IWRS code allows the specification of different \( n \) values for each cross section, but as there was no data available for this and the model was catchment-wide in nature, an estimate of roughness for the catchment as a whole was thought sufficient and so the hydraulic model was not set up to be fully distributed for roughness. The roughness is disaggregated into channel and floodplain components, as floodplain roughness is generally accepted as being higher than the river channel due to taller vegetation being present.

A range of sensible values was taken from the literature. Chow (1959) gives values of Manning's \( n \) for many types of surface, summarised in Table 3.21.
<table>
<thead>
<tr>
<th>Type of Surface</th>
<th>Typical $n$ Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glass</td>
<td>0.010</td>
</tr>
<tr>
<td>Finished concrete channel</td>
<td>0.012</td>
</tr>
<tr>
<td>Clean earth</td>
<td>0.018</td>
</tr>
<tr>
<td>Short grass</td>
<td>0.030</td>
</tr>
<tr>
<td>Natural channel (clean, no riffles/pools)</td>
<td>0.030</td>
</tr>
<tr>
<td>Natural channel (winding, some weeds and stones)</td>
<td>0.045</td>
</tr>
<tr>
<td>Dense weeds, high as flow depth</td>
<td>0.080</td>
</tr>
<tr>
<td>Natural channel (very weedy, deep pools, stands of timber)</td>
<td>0.100</td>
</tr>
<tr>
<td>Dense woodland</td>
<td>0.100</td>
</tr>
</tbody>
</table>

Table 3.21 Typical values for Manning’s $n$ for different surfaces (from Chow, 1959)

As Manning’s $n$ was the primary method of calibration for the hydraulic component of the model, sensitivity analysis was undertaken on this parameter. From the information in Table 3.21, a parameter space for $n$ was derived, and a grid of possible combinations for channel and floodplain $n$ was assembled. Each combination in the table was tested (with the values applied to all cross sections) on a pre-calibrated version of the model run for three calibration events (Table 3.22), with the results shown in Table 3.23. Results are compared to observed flow at Oxford, measured using $r^2$, and averaged over three events. The calibration events were chosen on the basis of data availability (observed flow and stage at Islip and Oxford; rainfall input at Bicester and Banbury), and providing high magnitude events whilst leaving the largest for validation (Figure 3.46).
<table>
<thead>
<tr>
<th>Event</th>
<th>Start</th>
<th>End</th>
<th>Duration</th>
<th>Maximum Flow at Oxford</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>hours</td>
<td>m$^3$s$^{-1}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>10/12/1999</td>
<td>18/12/1999</td>
<td>192</td>
<td>21.54</td>
</tr>
<tr>
<td>D</td>
<td>28/10/2000</td>
<td>05/11/2000</td>
<td>192</td>
<td>42.53</td>
</tr>
<tr>
<td>E</td>
<td>04/12/2000</td>
<td>24/12/2000</td>
<td>480</td>
<td>45.61</td>
</tr>
</tbody>
</table>

Table 3.22 – Calibration events

Floodplain $n$

<table>
<thead>
<tr>
<th>$r^2$</th>
<th>0.02</th>
<th>0.03</th>
<th>0.04</th>
<th>0.05</th>
<th>0.06</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.02</td>
<td>0.11</td>
<td>0.10</td>
<td>0.11</td>
<td>0.19</td>
<td>0.09</td>
</tr>
<tr>
<td>0.03</td>
<td>0.20</td>
<td>0.19</td>
<td>0.27</td>
<td>0.29</td>
<td>0.33</td>
</tr>
<tr>
<td>0.04</td>
<td>0.27</td>
<td>0.32</td>
<td>0.38</td>
<td>0.42</td>
<td>0.47</td>
</tr>
<tr>
<td>0.05</td>
<td>0.32</td>
<td>0.40</td>
<td>0.45</td>
<td>0.52</td>
<td>0.56</td>
</tr>
<tr>
<td>0.06</td>
<td>NA</td>
<td>0.53</td>
<td>0.54</td>
<td>0.61</td>
<td>0.67</td>
</tr>
</tbody>
</table>

Table 3.23 – Grid of sensitivity analysis $r^2$ results
Figure 3.46 - Flow at Oxford: calibration events
The results show that the effect of Manning's $n$ on the flow in the modelled channel is marked, although the difference of a change of just 0.01 on either the catchment floodplain or channel roughness would not dramatically change the results. The results of the lowest (0.02 channel and 0.02 floodplain) and highest (0.06 channel and 0.06 floodplain) total roughness values show clear differences in fit (Table 3.23). However, differences between close roughness values are small, with a difference in $r^2$ of only 0.02 with a floodplain roughness increase from 0.04 to 0.05 (channel roughness 0.03; Table 3.23).

Although in-channel conveyance will increase at higher flows with increased velocity, changing channel or floodplain roughness affects different components of the hydrograph; channel roughness primarily affects low flows and floodplain roughness only affects higher flows. Using a floodplain roughness of 0.04, the effect of changing channel roughness from 0.02 to 0.06 is clearly visible (Figure 3.47a), with a sharp difference in peak flow and decreased time to peak. The sharp oscillations in the run with low channel roughness are a reflection of instabilities in the model (flow is being routed down the river channel far too quickly), but the broad implications of the change are clear. The effect of changing floodplain roughness in a similar fashion is not quite as great (Figure 3.47b), as the retardation of flow caused by higher roughness only has an effect at higher flows, when floodplain inundation has occurred and constitutes a large enough proportion of the flow to impact the hydrograph. Whereas differences in channel roughness of 0.02 and 0.06 can be seen at flows as low as 3 m$^3$s$^{-1}$, a flow of 10 m$^3$s$^{-1}$ is required for similar floodplain roughness differences to be noticeable.
3.4.6 – Model Calibration

Now that some awareness has been gained of the model's response to its primary calibration parameter, calibration can be undertaken using roughness as measured by Manning's $n$. Events used for calibration are outlined in Section 3.4.5 (Table 3.22 and Figure 46). Five events were initially chosen with expectation that some would be unsuitable for lack of calibration data, and this was the case with events A and B (not shown).

Sear et al. (2002) undertook development of a model of the River Cherwell, investigating the impact of the floodplain along several discreet reaches of the river. The modelling took the form of three separate models of short (~5 km)
river reaches, and incorporated field reconnaissance of bed material size to
determine roughness values for input into the model. Although roughness was a
calibration parameter, field surveys of sediment yielded less uncertainty with the
parameter than for the current study, and so the tolerances for changing the
roughness for calibration purposes were smaller. The current work is modelling
the Cherwell catchment as a whole, and so identification of roughness at such a
small scale is both unrealistic practically and unnecessary, as variation and so
uncertainty across the catchment will be higher than variation within small
reaches. Roughness, as the primary calibration parameter, will incorporate the
uncertainties in other areas of the modelling, such as assumptions incorporated
into the modelling.

The final values for floodplain roughness were 0.05 for the channel and
floodplain components of the system, applied to all cross sections. This
representation of the river system is a simplification, and has no spatial
distribution downstream of changes in roughness, for example as land use
changes. For this reason the channel and floodplain components are broad
spatial averages, and serve to calibrate the model for the current application.
Although not usual to have channel and floodplain $n$ values as the same, it is not
unrealistic due to this broad catchment scale representation. Another
application, where model output would be required at many points along the river
network for instance, might require spatially distributed values for roughness.
For the current application, the lumped methodology is entirely appropriate and
calibrated model output fits observed data well (Figures 3.48 to 3.51), and
different values for channel and floodplain $n$ provided an inferior fit.
The final calibrated model was deemed to be a good representation of flows on the Rivers Cherwell and Ray, and the fit with observed data for the calibration events was generally very good (Figures 3.48 to 3.50 by event, Figure 3.51 composite; Table 3.24). Fit was measured using the standard $r^2$ and root mean square error (RMSE) statistics. It must be considered that modelled flow is highly dependent upon rainfall input to the model, and this is done through two rain gauges across the catchment, which may not be representative of catchment-wide rainfall, and so this may be a source of error for modelled flow for any given event.

Flow at Oxford is well modelled, although with a slight overestimation of peak flow. However, this was deemed satisfactory, as peak flow is known to be underestimated by the gauging station (Marsh, pers. comm.). The shape of the modelled hydrograph is also analogous with observed data; the timing of peak flow corresponding well together with the gradient of rising and falling limbs. The combined $r^2$ for the fit is 0.87, although the RMSE was not so impressive with 4.88 (largely due to the error in event D, with an RMSE of 8.85).
Figure 3.48 – Calibration: Event C
Figure 3.49 – Calibration: Event D
**Figure 3.50 - Calibration: Event E**

- **a - Flow at Oxford**
  
  \[ r^2 = 0.95; \text{RMSE} = 3.17 \]

- **b - Stage at Oxford**
  
  \[ r^2 = 0.93; \text{RMSE} = 0.18 \]

- **c - Flow at Islip**
  
  \[ r^2 = 0.72; \text{RMSE} = 1.88 \]

- **d - Stage at Islip**
  
  \[ r^2 = 0.94; \text{RMSE} = 0.27 \]
Stage at Oxford was calculated using the flow-stage relationship described in Section 3.4.4.8, using the modelled flow as the input. Again, the modelled and observed data fit well, with an overall $r^2$ fit of 0.83 and RMSE of 0.17.

Modelled flow at Islip (on the Ray) fitted well with observed data, with an $r^2$ of 0.67 and RMSE of 2.01. The observed flow data from the River Ray at Islip, although validated in general (Section 3.4.3.2), reflect the backing up effect from the River Cherwell at high flow, manifesting in sharp oscillations in flow which are indicative of not entirely trustworthy data. The modelled flow represents the broad response of the river at this site, and is likely to be reliable, as the magnitude of flow is correct and rising and falling limbs of the hydrograph either side of the oscillations match well with observed data. Although the statistical analysis of the modelled fit is unlikely to be high due to the oscillations in flow of the observed data, the representation of the flow is acceptable.
The fit of stage at Islip ($r^2$ of 0.93; RMSE 0.24) was particularly good, with both absolute error low and the model reproducing stage well at all phases of the flood events. Although the model did underestimate stage at peak flow (by approximately 30 cm during Event E), any change in calibration of roughness would make the error in Event C (an overestimation) larger, and greatly affect fit at Oxford.

<table>
<thead>
<tr>
<th>Event Measure</th>
<th>Oxford Flow</th>
<th>Oxford Stage</th>
<th>Islip Flow</th>
<th>Islip Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Event C $r^2$</td>
<td>0.89</td>
<td>0.64</td>
<td>0.90</td>
<td>0.86</td>
</tr>
<tr>
<td>Event D RMSE</td>
<td>2.16</td>
<td>0.10</td>
<td>2.38</td>
<td>0.15</td>
</tr>
<tr>
<td>Event E $r^2$</td>
<td>0.87</td>
<td>0.87</td>
<td>0.62</td>
<td>0.93</td>
</tr>
<tr>
<td>Event E RMSE</td>
<td>8.85</td>
<td>0.19</td>
<td>1.92</td>
<td>0.24</td>
</tr>
<tr>
<td>Average $r^2$</td>
<td>0.95</td>
<td>0.93</td>
<td>0.72</td>
<td>0.94</td>
</tr>
<tr>
<td>Average RMSE</td>
<td>3.17</td>
<td>0.18</td>
<td>1.88</td>
<td>0.27</td>
</tr>
</tbody>
</table>

Table 3.24 – Calibration results
3.4.7 – Validation

3.4.7.1 – Verification of Model Performance

Having been calibrated, the model was validated using data independent of the calibration events. The events used are shown in Table 3.25 and Figure 3.51, and the model results shown in Figures 3.53 to 3.55 by event and 3.56 combined; Table 3.26 then summarises the statistical measures of fit. Event G was a notable event on the River Cherwell catchment, caused by a stationary weather front resulting in prolonged localised rainfall and significant flooding over the Easter period of 1998 (Knight, 2006). Widespread disruption was caused by the flooding, which had a return period of greater than 100 years (Acreman et al., 2002a), with 1,500 people evacuated from their homes, up to five fatalities, and £500-700 million of damage (Knight, 2006). This event should therefore provide solid validation of the model, as the magnitude is significantly outside that of the calibration events.

<table>
<thead>
<tr>
<th>Event</th>
<th>Start</th>
<th>End</th>
<th>Duration (hours)</th>
<th>Maximum Flow at Oxford (m³s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>F</td>
<td>11/02/1997</td>
<td>03/03/1997</td>
<td>480</td>
<td>21.52</td>
</tr>
<tr>
<td>G</td>
<td>07/04/1998</td>
<td>21/04/1998</td>
<td>336</td>
<td>97.30*</td>
</tr>
<tr>
<td>H</td>
<td>20/01/2002</td>
<td>10/02/2002</td>
<td>501</td>
<td>30.74</td>
</tr>
</tbody>
</table>

Table 3.25 – Validation events

*Estimated at >100 m³s⁻¹ by a spot check (Marsh, pers. comm.)
It should be noted that the Banbury raingauge was disabled by the intense rainfall during the Easter 1998 floods. Although the data have been verified and are believed to be representative of the event, no data are available for after the event for several weeks. It was decided that it was justified using event G as a validation event due its high magnitude, despite the subsequent uncertainty after the peak rainfall, so as to provide robust validation for the model. Also, the event can be used in subsequent investigations, and being such a large event would provide information on how the wetlands and wider catchment behave during very high flows.

The validation procedure has verified that the model is able to represent effectively flow in the Cherwell catchment. Flow at Oxford shows a very good fit, with a combined $r^2$ of 0.89, and good visual fit. The peak flows are timed well, although slightly retarded as peak flow in event F and the dominant peak in event G are somewhat later than the observed data. The peak modelled flow of 80 $\text{m}^3\text{s}^{-1}$ is a reasonable relative error from the 100 $\text{m}^3\text{s}^{-1}$ observed, and performance is remarkably good considering the magnitude of this 1 in 100+ year event. Underestimation of flow for this event is predictable, considering the localised rainfall which produced flood event. The falling limb of the hydrograph is somewhat steeper than anticipated in event G, but late event performance is very good in events F and H. Stage at Oxford has a good fit, although variation is small, even for the data-sparse event G (it was felt that event G was essential for validation purposes despite this lack of peak stage data, as discussed above). Performance during the extensive event H was particularly impressive,
and it is unlikely that any model, however well calibrated, would have picked up the small scale variations in stage observed during event F.

The results for modelled flow at Islip verified that the model was very proficient at reproducing flows on the River Ray, and indeed capable of reproducing with some competency the 'backing-up' effect experienced at the Cherwell confluence. Event F shows a good reproduction of flows, although peak flow is underestimated by some 25%. Particularly pleasing is the response of Islip flows during event G, where after an initial build up of flow after approximately 50 hours, the backing up effect is shown in the modelled response with both the timing and magnitude fitting observed data well. Fit during later phases of the event are not quite as good, although quality of observed data under these particularly high flows is known to be variable, but an $r^2$ fit of 0.88 is excellent during such a large event. Again, the total volume of the event is underestimated, probably due to the limited spatial representation across the catchment of the rain gauges due to localised rainfall. Fit during event H is very good, with an $r^2$ of 0.93 and a low RMSE, as flow is reproduced well with timings of large increases in flow very close to observed data.

Fit of stage at Islip is less impressive, although the focus of calibration was on flow, with the peak in event F being late by approximately ten hours. The shape of response to event G is a good trace of observed data, although stage is significantly underestimated for a considerable time, especially on the falling limb of the flood event. Event H shows a generally good fit, with timings of peak stage close to observed.
For further validation, modelled flows on the River Cherwell were compared to observed flows at Enslow (Figure 3.57), 22.8 km upstream of Oxford and 11.6 km upstream of the confluence with the River Ray. This is not possible for stage, as no details of the cross section are available. Flow is obviously of a lower magnitude than downstream at Oxford, but the shape of the hydrograph is very similar. Model performance at Enslow is again good, although not as close to

<table>
<thead>
<tr>
<th>Event Measure of Fit</th>
<th>Oxford Flow</th>
<th>Oxford Stage</th>
<th>Islip Flow</th>
<th>Islip Stage</th>
<th>Enslow Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>F</td>
<td>r²</td>
<td>0.86</td>
<td>0.61</td>
<td>0.85</td>
<td>0.76</td>
</tr>
<tr>
<td></td>
<td>RMSE</td>
<td>1.81</td>
<td>0.09</td>
<td>1.00</td>
<td>0.13</td>
</tr>
<tr>
<td>G</td>
<td>r²</td>
<td>0.89</td>
<td>0.70</td>
<td>0.88</td>
<td>0.79</td>
</tr>
<tr>
<td></td>
<td>RMSE</td>
<td>13.16</td>
<td>0.21</td>
<td>3.64</td>
<td>0.67</td>
</tr>
<tr>
<td>H</td>
<td>r²</td>
<td>0.94</td>
<td>0.92</td>
<td>0.93</td>
<td>0.91</td>
</tr>
<tr>
<td></td>
<td>RMSE</td>
<td>3.17</td>
<td>0.08</td>
<td>1.41</td>
<td>0.17</td>
</tr>
<tr>
<td>Average</td>
<td>r²</td>
<td>0.89</td>
<td>0.81</td>
<td>0.80</td>
<td>0.84</td>
</tr>
<tr>
<td></td>
<td>RMSE</td>
<td>6.86</td>
<td>0.12</td>
<td>2.13</td>
<td>0.36</td>
</tr>
</tbody>
</table>

Table 3.26 - Validation results
observed as at Oxford. Flow is underestimated for the first two peaks during event F, yet overestimated in the later and larger peak between 300 and 400 hours. The shape of the modelled hydrograph for event G is not as close to the observed data as at Oxford; the more rounded, less flashy model response reflects calibration at Oxford and so wider catchment response. There is a general underestimation of flow during event H, contrasting with the Oxford and Islip gauging stations, although the shape of the modelled hydrograph is generally commensurate with observed flow.

Fundamentally, flow is represented well at the Enslow gauging station despite it being completely independent of the calibration procedure. Flood event magnitudes are similar to those of observed data, and hydrograph shape broadly reflects the upper catchment response to rainfall.

3.4.7.2 – Model Uncertainty
In order for the uncertainty associated with the modelled output to be appreciated, one must determine the uncertainty associated with each of the model's component inputs. Firstly, the rainfall data used to drive the rainfall-runoff model and hydraulic model will have error associated with it. Although all tipping bucket rain gauges are calibrated, this will be to a certain level of accuracy which depends upon installation (level ground) and maintenance (blocked funnels are, for example, a common problem). Uncertainty from this source is unknown, but likely to be minimal, of the order of 4-5%.
Errors in river flow estimation are often not accounted for, and can be appreciable (Marsh, NRFA, pers. comm.). Gauged flows in the UK, although accurate compared to many other countries as a result of relatively well-resourced hydrometric authorities, well maintained equipment, highly trained staff and relatively small discharges and velocities, do have a significant level of uncertainty associated with them. These inaccuracies are increased at both low flows (due to sensitivity of level measurements) and high flows (due to, for example, flow above the gauging structure or increased sediment disrupting ultrasonic signals). Recent advances in spot measurements using mobile acoustic Doppler current profilers has greatly increased the measurement of high flows, and one such measurement was included in the Cherwell at Oxford dataset and had a significant impact on the extension of the stage-discharge relationship developed. However, the conventional time series used to calibrate the rainfall-runoff and hydraulic models are somewhat limited in high flow accuracy, yet are likely to give a broadly indicative quantification of flows as a worst case scenario. No study has quantified absolute error at a general level, but it is estimated that up to 10% error would be representative.

Cross section data were regarded as observed data in the development of the hydraulic model. However, there will be measurement accuracy in the surveying process (~5 mm), and the cross section is likely to have undergone some geomorphological development from flow since surveying in the late 1970s (River Ray) or late 1990s-2000s (River Cherwell). Although these changes cannot feasibly be assessed quantitatively in the scope of this project, some error will have been introduced from this source.
The extension of cross sections from topographic datasets has been demonstrated to be effective. The error in the DEM datasets (0.5 m vertical accuracy for NEXTMap, 0.15 m for LiDAR) is very low compared with previous generations of comparable data. Even so, it has been demonstrated that small differences level lead to large differences in volumes of water stored on the wetland site (see Section 3.3.2.3).

The mathematics of the 1D hydraulic model code includes several important assumptions, outlined in Section 3.2.3, which make the necessary calculations feasible. The model is a representation and simplification of reality and although optimised to produce best results, cannot reproduce every nuance of the actual hydrodynamics of the river system's response to a rainfall event.

The primary calibrated parameter in the chosen configuration is channel roughness. This was not a measured parameter, but a parameter space was selected from values provided in the literature and a sensitivity analysis was undertaken on the effect of changes in roughness (Section 3.4.5). There is subsequently confidence that appropriate values have been assigned, but an issue exists whereby the calibration parameter can reflect uncertainty in other areas of the model. For example, errors in cross sectional geometry might need to be compensated through roughness, but the user would be unaware of this unless the calibrated parameter fell outside of the designated tolerances.
Despite these issues and uncertainties, the model has been assessed numerically for fit with observed data for each calibration and validation event (six in total). The model fit was good statistically and visually, with average statistical fits for flow (over the calibration and validation events) of $r^2$ of 0.81 and RMSE of 3.9. Importantly, as the results (Sections 3.4.6, calibration and 3.4.7, validation) demonstrate, there is no obvious difference in fit between the events that would suggest a bias towards a certain flow condition or event type. Furthermore, two measures of statistical fit have been used: $r^2$ and RMSE (as discussed in Section 3.4.3.3), allowing comprehensive assessment of model fit.

The combined model has been calibrated using the roughness parameter to provide a tool which converts rainfall to river flow and storage, and is likely to be accurate to within 10% of the equivalent gauged flow at any time.
3.4.8 – Hydraulic Model Construction – Investigations

During model development, some issues arose that required decision making by the model developer. Although the choices made were justified and believed to be correct, they could have implications for model output, and so have introduced a degree of uncertainty to the results. For completeness, alternative options from those chosen for the final construction and calibration will be investigated in this section.

3.4.8.1 – Investigation 1: DEM Changes

After the issues arising from DEM uncertainty during the development of the level one model at Otmoor (Section 3.3.2.3), it was thought necessary to evaluate the effect of using a different DEM in the same vein of the level one model. Again, LiDAR data was used as a substitute for NEXTMap for calculating the storage capacity of the Otmoor reservoir unit in the level two model, the methodology for which is described in Section 3.4.4.2.

As discussed under the level one model (Section 3.3.2.3), the LiDAR DEM was processed to remove a significant area of ‘no data’ coverage and replace it with a level of 57.9 maOD. This increased the volume of storage, but not to be comparable to that of the NEXTMap DEM. The flood storage was recalculated for the replacement DEM and the model run with no other changes so that the direct impact of DEM replacement may be quantified; Figure 3.58 shows modelled flows and stages, allowing the impact of the change to be investigated.
It can be seen that there is a noteworthy impact on flow from the change in storage indicated by Figure 3.58. As expected, the greatest impact can be seen on the River Ray (Figure 3.58 c&d), with any changes diluted on the Cherwell, although still visible at Oxford (3.58 a&b). The decrease in storage produces a flashier response on the Ray, with flood peaks arriving sooner and being of higher magnitude. Flow peaks are up to 20% greater during event F, and the difference is so distinct that during event G the flow on the River Ray is much stronger and holds its own at the confluence with the Cherwell, leading to significantly less backing up on the Ray at Islip (minimum flow of 2 m³s⁻¹ compared to -0.5 m³s⁻¹). The consequence downstream at Oxford is visible yet not extreme, with an increase in peak flow of about 1 m³s⁻¹ for each event and slight decrease in time to peak.

The changes implemented highlight the need for accurate representation of the floodplain whilst utilising the DEM-volume method of calculating floodplain storage. The model was calibrated with the NEXTMap DEM, and could theoretically be recalibrated for the LiDAR DEM to again replicate observed flow effectively. However, the running of this scenario has shown the importance of differences between DEMs: it is essential that appropriate data are used for the model being calibrated. Although it is possible that the LiDAR data could be the more accurate representation of topography at Otmoor, the initial ‘no data’ areas precipitate enough doubt to suggest the NEXTMap data to be more reliable. A key issue here is confidence of the model developer and user in the model and its representation of reality.
3.4.8.2 – Investigation 2: Contrasting Floodplain Representation

During model development, it was decided to represent the Otmoor floodplain system as a storage area (much like a reservoir) with a single connection to the river via a spill unit. This was justified by local knowledge of the system and previous attempts to model the complicated flow through a variety of small channels across the floodplain (Acreman et al., 2002a). It was thought that the reservoir system represented Otmoor in a simple yet largely realistic fashion, as the stage of floodwater at Otmoor is known to follow that of the river through a well established connection between river channel and floodplain.

In order to test this decision, Investigation 2 attempted to represent Otmoor with a contrasting methodology, using conventional cross sections extended with the technique utilised for the other floodplain areas in the catchment. The storage area was removed and cross sections extended across Otmoor, with floodplain roughness $n$ set initially to 0.05, as in the rest of the calibrated model. Validation events (F, G and H) were re-run. The results (Figure 3.59) show that the River Ray's response becomes faster, with higher peaks and less time to peak, with some effect downstream at Oxford. The flow is not decreased nearly as much as the storage area method, as water is routed much faster to the River Cherwell and catchment outlet. On this basis it was thought that the roughness on the floodplain component of the cross sections should be increased. This was done, firstly to 0.07 and subsequently to 0.10, but the changes made very little difference to the flow, with no significant retardation. A roughness of greater than 0.10 is thought unrealistic for a conventional floodplain roughness, although
the complicated flow pathways of Otmoor might act to slow the floodplain conveyance further than any point measurement of \( n \). On this basis a value of 0.20 for \( n \) was used (also shown on Figure 3.59), and although a difference was seen the roughness was still not high enough for flow to be commensurate with the calibrated model. This infers that the Otmoor floodplain acts as a bulk storage area, and not to attenuate flow through increased roughness. This vindicates the decision during model construction to represent the Otmoor floodplain system with a reservoir storage system, as the extended cross section method would have required floodplain roughness values outside the range of normal floodplain roughness as discussed above.

### 3.4.9 – Hydraulic Model Scenarios

Now calibrated and validated, the hydraulic model of the Cherwell catchment can be interrogated to assess the impact of land use and catchment management changes on catchment hydrology. The scenarios have been chosen to reflect historic and potential future changes across the catchment, and based on other work undertaken in the past.

#### 3.4.9.1 – Scenario 1: Embankment on River Cherwell

In a study investigating the effects of channel geometry changes on flow, Acreman et al. (2003b) modelled an embankment along a 5 km reach of the Cherwekk from Somerton Bridge to Upper Heyford (Figure 3.60), where the floodplain is typically 2 km wide. The work concluded that the theoretical
Scenario 1 modelled an embankment on the River Cherwell in a similar fashion to Acreman et al. (2003b), but for an increasingly long stretch of river, from the top of the model at Banbury to Oxford. The method adopted in the current work was to clip the floodplain sections of the river cross sections at the left- and right bank markers (even with an embankment between the channel and floodplain, the 1D code would still use the floodplain for routing water), and raise the level of these markers to 115 maOD, a level far above any feasible river stage. Although implausible today, significant river canalisation has taken place historically (see Section 3.4.9.2 for River Ray embankment), and was thought of as improvement in terms of water resource management and flood defence (Haslam, 2003). Contemporary management advocates a reversal of these techniques and a return to natural flood defences of floodplain storage and flow retardation, but many rivers have been heavily managed on very large scales (e.g. the Mississippi River; Remo and Pinter, 2007), and it is not out of the realms of possibility that similar management could have occurred on the Cherwell.

The embankment was established in cumulative phases downstream, as shown in Figure 3.61. Table 3.27 summarises the phases of embankment, with distance embanked and number of model nodes (cross section (XS) locations) in the embanked section. The model was run again using the validation events F, G and H.
The results of Scenarios 1a-f (Figure 3.62), show that embankment of the River Cherwell would have a notable effect on the river flow and stage. There is a continuum of increasing difference with increased embankment, as one would expect, and the general trend is for a decreased time to peak flow, but with little change in peak flow itself. The largest modelled event, G, shows the biggest impact, and the later results shown focus on this event: flow and stage data at Oxford and Islip are detailed in Figure 3.63; detail for peak flow and time to peak flow for event G are given in Table 3.28; and these data are displayed in Figure 3.64.

Only when Scenario 3f is reached, and the embankment covers the entire modelled component of the River Cherwell, do peak flows begin to be influenced by the embankment. Significantly, it is only in this scenario where the embankment reaches the confluence of the River Ray, and during event G (detail shown in Figure 3.63) this fact ensures that the backing up effect at Islip is exacerbated appreciably by causing an increase in stage of the Cherwell and so preventing the Ray from emptying. Indeed, after steady increases in the backing up effect from the calibrated model through to Scenario 1e, 1f then produces a jump in negative flow on the Ray from -4.80 m³s⁻¹ to -45.19 m³s⁻¹. This suggests that after this comprehensive embankment, the River Ray is acting as a large scale outlet for the river system during high flows, rather than a source of water. The effect of this is seen downstream at Oxford, where peak flows remain largely unchanged, until Scenario 1f is reached and the flow peak flow drops by 32% from 79.67 m³s⁻¹ to 54.01 m³s⁻¹.
Stage at Oxford is very stable following the establishment of embankments, showing only the decrease in time to peak, with change in peak levels being only negligibly affected. Stage at Islip closely reflects the flow conditions, and the continuum of decreased time to peak with increasing embankment. Only the final Scenario, 1f, shows significant variation from this, reflecting the very significant backing up effect described above, and stage under these conditions reaches 60 m aOD, 3.2 m above base flow conditions.

Table 3.28 shows the impact on both parameters of the embankments, and these are shown graphically in Figure 3.64, both for the largest event, G. It can clearly be seen that up to Scenario 1e the peak flows remain largely unchanged, but the time to peak flow is considerably reduced at Oxford with increasing embankment.

With the final, comprehensive, embankment the very large decrease in peak flow is offset by the extremely large minimum flow at Islip, where negative (reverse) flows of -45.19 m$^3$s$^{-1}$ are registered. Again, this reiterates the fact that during very high flows, the River Ray has become a conduit for water loss from the Cherwell system rather than an inflow, caused by the dramatic increase in stage on the River Cherwell at the confluence with the Ray. There is little change in time to minimum flow at Islip, although it regularly occurs well before the maximum flow at Islip, and of course the former is a main driving force for the development of the latter, as a blockage in the outlet of the Ray causes a build up of water in the Ray system which flows out later when the stage in the
Cherwell has dropped sufficiently. Importantly, the time of the minimum flow at Islip broadly corresponds to the time of maximum flow on the Cherwell at Oxford, which seems logical as it is primarily the maximum flow on the Cherwell (at Islip) which in turn causes the backing up effect and so the minimum flows on the Ray at Islip.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Peak Flow m$^3$s$^{-1}$</th>
<th>Time to Peak hours during event</th>
<th>Peak Flow m$^3$s$^{-1}$</th>
<th>Time to Peak hours during event</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calibrated</td>
<td>11.61</td>
<td>151.50</td>
<td>80.45</td>
<td>118.25</td>
</tr>
<tr>
<td>1a</td>
<td>11.74</td>
<td>149.75</td>
<td>80.40</td>
<td>116.50</td>
</tr>
<tr>
<td>1b</td>
<td>11.88</td>
<td>146.75</td>
<td>79.83</td>
<td>115.00</td>
</tr>
<tr>
<td>1c</td>
<td>12.16</td>
<td>143.25</td>
<td>79.82</td>
<td>112.25</td>
</tr>
<tr>
<td>1d</td>
<td>12.62</td>
<td>139.50</td>
<td>81.60</td>
<td>108.75</td>
</tr>
<tr>
<td>1e</td>
<td>13.26</td>
<td>131.50</td>
<td>79.67</td>
<td>99.25</td>
</tr>
<tr>
<td>1f</td>
<td>20.82</td>
<td>182.50</td>
<td>54.01</td>
<td>78.75</td>
</tr>
</tbody>
</table>

Table 3.28 – Key data from event G Scenario 1 model runs
roughness may not be realistic for the wide floodplain areas taken offline by this scenario.

These findings contrast with the work of Acreman et al. (2003b), where both peak flow and time to peak flow were reduced by a 5 km stretch of embankment in the middle reach of the river. This model was of a more simple construction, of a shorter reach and as such with fewer lateral tributary inflows, although these were provided by gauged data rather than a rainfall-runoff model. The model in the current work uses modelled inflows, which are known to be underestimated in the north of the catchment (where the storm event was primarily situated) due to limitations of the rainfall-runoff modelling process (see Section 3.4.3). Due to these differences, the model developed by Acreman et al. (2003b) has a larger inflow from the north of the catchment and less further downstream, and any floodplain areas would be available to attenuate flow for the primary flood wave propagating downstream. The model developed by the current work has more distributed tributary inflows and so floodplains across the catchment will be utilised early by multiple, simultaneous inflows, with little remaining floodplain storage for the following primary flood wave propagating from the north of the catchment.

3.4.9.2 – Scenario 2: Embankment on River Ray

A similar scenario to the embankment on the Cherwell was run, but this time with the River Ray the target. The embankment of the Ray has been undertaken in part over history; through Otmoor the current river resides in an anthropogenic channel, the New Ray, after works during the 1830s. The Old Ray still exists,
but the vast majority of the water is routed through the hydraulically efficient New Ray. Scenario 2 is initiated here to predict the effects of complete canalisation of the New Ray through the Otmoor area, and so no floodplain existing here. The contemporary effect of the flood storage component of Otmoor should be evident, as it is utilised regularly to store excess water from the river channel.

As with Scenario 1, the floodplain storage component was removed (in this case the Otmoor storage area), effectively embanking the river, and the validation events (F, G and H) were re-run (Figure 3.65). The effect of the loss of floodplain storage is immediately visible, and much greater than the changes in Scenario 1. During event F, a much quicker response can be seen at Islip, and flood peaks 65% greater than the calibrated model. At Oxford the change is noticeable, with the flood peaks arriving eight hours earlier although not different in magnitude. Flood stages are not dramatically different at either Islip or Oxford. The event with the highest peak flow, G, shows great differences from the calibrated model. At Islip peak flows are nearly 200% greater and arrive earlier, and the typical flow regime of increased yet stable flow disappears, to be replaced with a responsive and fluctuating hydrograph. The backing up effect is greatly reduced, and the minimum flow increases from 0.5 m³s⁻¹ to 8 m³s⁻¹ at the time of this effect being at its strongest in the calibrated model. This is reflected downstream at Oxford, where peak flow is increased by 16% from 80 m³s⁻¹ to 93 m³s⁻¹, and stage increases by 5 cm. Event H shows perhaps the most dramatic change at Islip, where the River Ray responds with great speed, and the peak flow increases from 14 to 24 m³s⁻¹, and there are now three peaks over 15 m³s⁻¹. Peak stage increases by 20 cm to 58.25 maOD at Islip. At Oxford the hydrograph is
transformed by the behaviour of the River Ray, reflecting the increased response of the catchment to rainfall with a decrease in the time to peak of 35 hours, increase in peak flow of 15% and increase in maximum stage of 10 cm.

The impact of losing the flood storage capacity of Otmoor has been illustrated, demonstrating that the 1,600 ha of wet grassland and grazing marshes provide an invaluable function by protecting downstream areas including the city of Oxford from flooding. This highlights the danger of previous threats to Otmoor, such as the building of the M40 motorway during the 1990s. Also, the utility of wetland areas for flood storage in general has been demonstrated and the importance of assessing the hydrological function of wetland sites can only be stressed from the results shown.

3.4.9.3 – Scenario 3: Simultaneous Embankments

The previous two scenarios have detailed the embankment of the River Cherwell and the major floodplain component of the River Ray sub-catchment. It is clear that both embankment schemes would have a notable effect on flow across the catchment, but if the temporal change in hydrographs seen for each event correspond, the effect may exacerbated.

The same changes were made to the model construction as detailed in sections 3.4.9.1 and 3.4.9.2, embanking both the Ray at Otmoor and the Cherwell. For the Cherwell embankment, Scenario 1d (from Banbury to Somerton Bridge) was chosen as this was a intermediate level embankment and did not interfere structurally with the Cherwell-Ray confluence. Events F, G and H were then run,
with results shown in Figure 3.59, together with results from single embankments in Scenarios 1 and 2.

It is immediately apparent that there are two groups of traces shown in the results (Figure 3.66), as the Scenario 1d (Cherwell embankment only) results generally follow that of the calibrated model but with a decreased time to peak; Scenario 3 results are generally more commensurate with those of Scenario 2 (Ray embankment only). This suggests that the Ray embankment has significantly more impact on river flows than that of the Cherwell embankment, and even with both embankments in place, the Ray continues to dominate differences from calibrated flow.

At Oxford during event G, Scenario 3 peak flow (93.33 m$^3$s$^{-1}$) is very similar to that of Scenario 2 (91.21 m$^3$s$^{-1}$), but the time to peak is decreased by the embankment on the River Cherwell by 7.5 hours, which would make a significant difference to flood defence managers by allowing less warning time. The same effects on the hydrograph are occurring in events F and H, but the magnitude of the impact is smaller due to the events being smaller. Stage at Oxford closely follows flow conditions, with the corresponding increase in stage for Scenario 3 being only 1.1 cm when compared to Scenario 2. One interesting change in the shape of the hydrograph can be seen on the rising limb of the event G main peak, where Scenario 2 (only Ray embanked) displays a slowing of the rate of increase in flow, which is not seen during Scenario 3. This is caused in Scenario 2 by the sheer speed in responsiveness to rainfall, whereby a relatively short break in rain is reflected in the flow at Islip. This phenomenon is displayed
during Scenario 3, but the higher discharge on the Cherwell at this time, caused by the faster response of the Cherwell because of the embankment, means that the effect is hidden. This is a change in the hydrograph initiated by the changing times to peak flow and subsequent synchronisation of flood peaks, exacerbating the filling of the embanked channel at an already high flow time, and so also increasing flooding extent downstream of the embankments, in this case at Oxford.

At Islip, flow is highly affected by the double embankment during the relative trough between peaks, as described above. As discussed during Scenario 2 (Section 3.4.9.2), the magnitude of the flow on the Ray, as dictated by any embankment or other management, directly influences the backing up, as the phenomenon is a function of the relative strengths in flow on the two rivers. If the Ray has a low flow compared to the Cherwell, backing up will be large. However, the Ray embankment here produces high flows on the Ray preventing backing up initially, but conveying all water from the Ray system quickly and so causing backing up later in the hydrograph, although the trough in flow does not manifest in negative flows in this instance. It appears that this situation would only benefit areas downstream of the confluence of the Ray and Cherwell at Islip, 11.2 km north of Oxford; upstream areas on the River Ray would be disadvantaged through substantial backing up.

There is no doubt that both embankments impact upon the catchment’s response to heavy rainfall, and that in certain circumstances the combination of multiple embankments (or, by inference, any other river or land use changes)
may increase this impact above the combined impact of the changes. The synchronisation of the flood waves from sub-catchments is known to increase flooding potential in general (Holden, 2005), and changes such as these across the River Cherwell catchment have been shown to generally increase peak flows and decrease times to peak flow, culminating in an increased flood risk potential downstream at Oxford.

3.4.9.4 – Scenario 4: Restoration of River Ray

In contrast to Scenario 2, the restoration of the River Ray should improve the connection between the river and the floodplain area of Otmoor. The aim is to predict the effect of the floodplain on flow before the canalisation and diversion of water through the New Ray in the 1830s. Restoration would remove all embankments and man-made channels, and leave only the Old Ray as a conduit for flow through Otmoor, the channel of which is much shallower and narrower than that of the New Ray. The flow path of the Old Ray is also 325 m longer than that of the canalised New Ray (Figure 3.67).

Cross sections for the Old Ray were digitised from surveys undertaken in the 1970s and supplemented by GPS surveys carried out by CEH staff for the work detailed by Acreman et al. (2002a). These were imported into IWRS and connected to the channel of the Ray above and below the divergence and re-convergence of the New and Old Rays respectively. The weirs installed as part
The results from the model (Figure 3.68), again compared to the calibrated model, indicate a typical increase in floodplain storage. The hydrograph at Islip is flattened, becoming much less responsive to rainfall and almost flat and completely unresponsive to event F. The impact on flow downstream at Oxford is a lower peak flow by 10% and slight delay in time to peak. The larger event G showed a similar effect, with a much lower flow for the entire duration of the event at Islip. The lower flow on the Ray allows the River Cherwell at their confluence to increase the magnitude of the backing up effect on the River Ray, decreasing the minimum flow from -0.5 to nearly -5 m³s⁻¹. At Oxford this change is not quite so noticeable, possibly because the greatest intensity of rainfall fell in the northern parts of the catchment and not in the River Ray sub-catchment. The peak flow is largely unaffected and only small changes in rising and falling limbs of the hydrograph are perceptible. The response to Event H shows a large difference, as flow at Islip is retarded considerably, being significantly lower for the first 400 hours of this, the longest event. As the storage capacity of Otmoor nears its limit past 400 hours, the response begins to trace that of the calibrated model representing current conditions. This is shown in the stage of water on Otmoor during the model (Figure 3.69). Although flow at Oxford mirrors this effect and is lower until about 400 hours, peak stage is mostly unaffected due to the recovery to calibrated conditions once the Otmoor storage reaches capacity and can no longer influence flow.

This scenario, reflecting the likely behaviour of the Otmoor floodplain area before intense management and canalisation began in the 19th Century, contrasts markedly with the previous scenario of increased management and improved
which the flood storage potential of Otmoor could be removed is immediately after a large rainfall event, when all storage is being utilised and so unavailable to further rainwater which may fall.

Therefore, Scenario 5 will investigate the impact of having one very large rainfall event following another. As no event with these characteristics exists in the data held, an event was instead created. This theoretical event was based on the largest event in recent times in the Cherwell catchment, the validation event of Easter 1998, during which the vast majority of the rainfall fell on the 9th April. As shown in Figure 3.69, the stage of the Otmoor reservoir after Event G was high, falling from a peak of nearly 59.0 m\(\text{aOD}\) (a water height of nearly 1 m across many areas of Otmoor). If significant rainfall fell at this point, as Otmoor is starting to drain, flood storage would not be available and flood peaks are unlikely to be retarded as before. Otmoor is not the sole water storage component across the Cherwell basin; all ground will be saturated and unable to store any more water.

Scenario 5 will copy the rainfall from the 9th April, and feed it into the system again during the 12th April, as shown in Figure 3.70. As noted previously, the Banbury raingauge was disabled by the rain towards the end of the Easter 1998 floods, but the data are thought important enough to include in this scenario.
The results from Scenario 5a (Figure 3.74) show significant differences from Scenario 5. The increased delay in time to the repeated rainfall event is reflected in the flow on the Rivers Ray and Cherwell. The peak flow at Islip on the Ray is decreased from 20.27 m$^3$s$^{-1}$ (Scenario 5) to 15.60 m$^3$s$^{-1}$ (Scenario 5a) through a further 48 hours delay in rainfall repetition. At Oxford, the peak flow of the River Cherwell drops from 111.15 m$^3$s$^{-1}$ (Scenario 5) to 85.08 m$^3$s$^{-1}$ (Scenario 5a), an 85% return to the modelled peak flow for the event of 80.45 m$^3$s$^{-1}$.

The data of stage at Otmoor during Scenario 5a (Figure 3.75) show that the wetland responds readily to rainfall. With 96 hours between rainfall peaks (Scenario 5a), stage at Otmoor drains from 58.95 (peak of modelled event) maOD to 58.77 maOD before it begins to rise again. This is in contrast to Scenario 5, where only 48 hours rest between rainfall peaks means that Otmoor has only just reached its peak stage when the (theoretical) rain falls again and levels begin to rise further. This important difference between the two Scenarios means that the fall in peak stage is 0.19 m, from 59.23 maOD in Scenario 5 to 59.04 maOD in Scenario 5a. This difference of 0.19 m in stage is estimated to be equivalent to 3,787,000 m$^3$ of water storage, obviously an important amount and one which will have a noticeable effect downstream at Oxford as shown above.
rainfall event has almost as much effect on flow as full embankment and subsequent complete loss of the wetland. This is a noteworthy result and demonstrates the importance of the Otmoor wetland and its hydrological functioning on the river flow in both the River Ray sub-catchment and wider River Cherwell catchment. This point is reiterated when flow at Oxford is considered, as flows produced from repeated rainfall (Scenario 5) have a higher maximum flow (111.15 m$^3$s$^{-1}$) than the flows from the embankment scenario at Otmoor (Scenario 2; 91.21 m$^3$s$^{-1}$), likely to be due to an increased soil moisture level across the catchment and so increased rate of runoff. This highlights the importance of the wetland areas across the wider catchment. A combination of the two scenarios has only a marginal further increase of flow at Oxford (to 114.01 m$^3$s$^{-1}$).

As expected, the combined scenarios’ (embankment and nested event, purple line) trace follows the embanked results (Scenario 2) until the effect of change in rainfall profile begins to manifest at about 120 hours on the River Ray and 125 hours on the River Cherwell. The flow on the Ray then increases at a rate similar to the initial event, but to a significantly higher flow of 40.59 m$^3$s$^{-1}$ compared to Scenario 2’s (albeit earlier) peak of 26.80 m$^3$s$^{-1}$ with an embankment only composition. This demonstrates that although storage loss during a previous event does have a large impact on flow as shown above, when the river is embanked and storage unavailable, there is still an impact of having subsequent events, suggesting that the antecedent flow conditions in the river and other soil stores across the catchment remain important to determining the river’s response during an event.
3.4.10 – Assessment of Data Used

Table 3.29 summarises the data used in hydraulic (level two) model development, and gives the scale of the extent and resolution of each variable. The labels for groups of scales are given according to those used in Tables 1.8 (spatial scales) and 1.9 (temporal scales). Rainfall data are taken at two point locations to represent the catchment, and although they have a poor spatial resolution, the temporal resolution of 15 minutes and accuracy of 0.01 mm is very high. This is appropriate for event-based hydraulic modelling. The flow data again have poor spatial resolution, but the locations of gauges are very directed, being at systematic positions such as towards the end of sub catchments. The temporal resolution is fine scale, and the data cover the extents of the flood events modelled, so again are appropriate.

<table>
<thead>
<tr>
<th>Dataset</th>
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<td>Flow</td>
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<td>Model output</td>
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<td>Cross section</td>
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<td>DEM (NEXTMap)</td>
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<td>DEM (LiDAR)</td>
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Table 3.29 – Hydraulic Model Data Assessment
Model output is flow and stage at key locations across the catchment, given at a temporal resolution of 15 minutes. This was chosen with specific regard to the temporal resolution of the rainfall inputs, as well as to be directly commensurate with observed flow data for assessment of output. Cross section data have a high enough resolution to be used for modelling, and are likely to have been undertaken with this application in mind. Their extent, although not covering the entire catchment, has a high enough coverage for good quality modelling to be undertaken. Cross sections are unlikely to change significantly over medium scale time periods, and will therefore need updating only on a decadal-scale (medium-term) basis, as cross sectional geometry may be changed by large scale events that may entrain and subsequently deposit large amounts of sediment, but which occur only infrequently as the characteristic velocity would suggest. This suggests that the River Ray cross sections, surveyed in the 1970s, may be somewhat out of date and confidence in these might be beginning to fade, although they were verified by CEH in the early 2000s as part of work by Acreman et al. (2002a). River Cherwell cross sections, surveyed in the 1990s and early 2000s are much more recent and completely applicable to hydraulic modelling. Survey methods will have improved significantly with technology in recent decades, such as with the advent of GPS. This may not necessarily increase accuracy of data, but will increase speed of work and so perhaps resolution of cross sections.

DEM data have improved considerably over the last ten years: the modelling methodologies adopted in the current work would not have been possible even
five years earlier. NEXTMap data have complete catchment coverage and high spatial resolution of 5 m. LiDAR data have a higher spatial resolution (2 m), balanced by a slightly smaller extent as only the Otmoor area within the Cherwell catchment were available. Temporally, these datasets have only been acquired once thus far due to financial and time restrictions. As with cross sections, this is appropriate, as the landscape is unlikely to change much over decadal time periods.

In summary, no particular problems were uncovered by assessing the spatial and temporal scales of the hydraulic model input data, and scales were appropriate for the scale of model constructed. An update of River Ray cross sections is now due, and should be undertaken before further models of the river are constructed.

It would seem appropriate for this analysis to be done as part of good practice in modelling, and incorporated into contemporary modelling protocols such as those described by Refsgaard (2007a) and Harmoniqua (Old et al., 2005).
3.4.11 – Level Two Model Conclusions

A full hydraulic model of the River Cherwell catchment including the River Ray sub-catchment (and its associated Otmoor wetland system) has been developed, incorporating the rainfall-runoff component which produces flow boundary conditions. The model has been successfully constructed, calibrated and validated using industry standard techniques. During calibration and validation, fit with observed data has been good, although representation of the extremely large Easter 1998 floods (event G) was underestimated, most likely due to unrepresentative rainfall data.

The system has been utilised for the analysis of various investigations and scenarios to assess both the sensitivity to different modelling techniques, and the impact of changes to the wetland and wider catchment. For the former, it has been shown that the model is sensitive to methodology applied during model construction, including the DEM used to calculate floodplain storage potential and alternative representation of the floodplain system. The calibrated model uses a storage reservoir system to represent Otmoor, as it was thought this most closely resembles the actual processes, particularly when viewed from a catchment scale perspective. It was shown that in using conventional cross sections, extended by up to 4 km across the floodplain wetland, roughness values required to calibrate the model would have been far outside the tolerances of normal observed roughness values. This proves that the initial conceptualisation of the river network is important, although in this case it appears to have been correct.
The wetland components on the Rivers Cherwell and Ray have behaved very differently in the modelled situations whilst investigating using embankments and restoration. The floodplain on the River Cherwell has acted as a typical online wetland, increasing the time to peak flow during the flood event through increased friction as the flow has expanded onto the floodplain, but having little or no impact on flood event peak flows or volumes. In contrast, Otmoor has functioned purely as a storage component, decreasing the flood event’s peak flow and the timing on event H significantly.

Although an important finding, caution is required as the modelling approach used for each was different, although justified and vindicated above. It is possible that the modelling methodologies chosen for each could have contributed somewhat to this outcome. For the floodplain component of the River Cherwell, 1D hydraulic model cross sections were extended across the floodplain. For Otmoor, a reservoir storage unit was used, and the cross section extension method was proved to be unsuitable as it required floodplain roughness parameters far outside normal and reasonable parameter space. This verifies the initial model conceptualisation and subsequent methodologies for modelling different wetland components within the catchment, and reinforces the need for suitable and thoughtful conceptualisation, and its importance within the structure of wetland hydrological functioning assessment, as advocated by Acreman (2004).

This is also an important distinction between wetland types, and one that Bullock and Acreman touched upon in their 2003 paper reviewing the literature on wetland hydrological functioning. For floodplain wetlands, they concluded that wetlands
differ in their impact on catchment flow, decreasing flood peaks or flood event volumes, or increasing the time to peaks. Key in this literature review, and an updated version (Mould and Acreman, in preparation), was that there was no evidence of wetlands lying within the same catchment behaving with such differences as those displayed in the current study.

It has been demonstrated that both the embankment and restoration of floodplain components have a large impact upon flows both near to the site and downstream at Oxford. As limited canalisation of the River Ray through the Otmoor wetland area has been undertaken in the past, modelling of both restoration and full embankment has been possible. Furthermore, testing has been carried out to investigate the effect of simultaneous embankment on both the Rivers Ray and Cherwell, concluding that only if flood pulses from each were dictated by specific rainfall events, would problems be exacerbated by synchronisation of hydrographs.

In Scenario 4, storage at Otmoor was more readily utilised by the River Ray with a longer flow path and more natural channel geometry. This had a large effect on flow, resulting in more of the storage at Otmoor being used during a flood event. However, as Scenario 5 has demonstrated, once the wetland system contains water, the storage for flood water is removed; the wetland being full has a similar effect to hydraulically disconnecting the wetland using an embankment. Scenario 5a has established that the storage levels recover with a rest between rainfall events as the wetland drains.
The model remains as a tool which can be used in future for further work which might be required for the understanding of the Rivers Cherwell and Ray catchments. As the RSPB continue to develop and manage the Otmoor wetland for their particular objectives, which focus on creating bird habitat (RSPB, 2006), this may be feasible over the coming years. Another application might be to investigate climate change impacts across the Cherwell and Ray catchments, although it would need adapting to be continuous simulation rather than event-based to gain maximum use (Kay et al., 2006c).

The level two model could be further complemented by coupling with an in-field hydrological model (using a code such as ZOOM or MODFLOW, for example). This would develop the model of Otmoor from being a simple reservoir coupled to the River Ray via a spill unit of calibrated height to being a full three dimensional multi-component hydrological model representing the evaporation directly, as well as vertical and lateral water transfers. Such a model could be coupled with the catchment-wide hydraulic model using an interface such as OpenMI (Tindall, 2005).
Chapter 2 described the development of understanding of the hydrology of Otmoor through an intense data collection campaign and field scale analysis of data. This understanding has been supplemented in the current chapter by developing an understanding of the hydrology of Otmoor within the context of the wider catchment. A particular focus on flood storage has been developed, as this was highlighted in the field scale work as being important to the hydrological functioning of the site.

The simple level one model has demonstrated efficacy at providing an insight into the flood storage potential of Otmoor, and this has been directly compared to a contrasting wetland site, Tadham in Somerset, southwest England. Evident from the model output was the obvious differences between the wetland sites, particularly in the dominant storage components; Otmoor has little subsurface storage and almost all is above ground on the conventional floodplain system, whereas Tadham has a much larger subsurface store for rainfall and runoff and is managed to prevent significant above ground water storage. Interestingly, also evident are the similarities between the sites, as both provide significant flood storage and subsequent protection for downstream communities from flooding. From personal experience both sites provide excellent complementary habitats, targeted towards similar breeding wader populations and wider conservation targets.
The level two hydraulic model has been shown to be a significant development of level one, not only as it added a temporal dimension to the hydrological analysis of Otmoor, but also it was a much more robust style of model. Although requiring significantly more investment in development, the benefits have been obvious, with important insights into how the Otmoor wetland system behaves within a catchment-wide setting. It has become evident that the wetland is important not only for its habitat and conservation properties, but the flood storage potential at the site is very large by UK standards, and if managed effectively could help protect downstream areas such as the City of Oxford from flooding. However, flood storage potential of the site is decreased, as it is on the Tadham peatlands, by raising water levels, a practice which has been undertaken in recent years by the RSPB and others to improve habitat for breeding waders. The conservation and flood defence policies of different stakeholders across the Cherwell catchment appear at odds with one another, but a balance struck between the two (whereby water levels are raised in areas or to mean field height allowing above ground storage to be available) should provide parity.
Chapter 4

Conclusions
4.1 – Introduction

A review of the literature in Chapter 1 concluded that wetland hydrological functioning is fundamental to a wetland’s behaviour both locally and in the wider catchment. It also highlighted a gap in the field of wetland hydrology relating to an appreciation of scale issues and their impact on the outcome of hydrological science in wetland environments. The following hypotheses were proposed:

1. Scale issues will be evident both at each scale of observation of wetland hydrology;
2. The field scale site-specific nature of a wetland’s hydrology is crucial to any hydrological functioning at the catchment scale;
3. Scale issues will be important for and so impact upon the assessment, through monitoring and modelling, of wetland hydrological functioning; the successful comprehensive assessment must therefore be scale dependent.

This study has thus worked at two very distinct spatial scales and a wide range of temporal scales, and assessed the links between them in order to investigate scale effects. The first was the field scale, described in Chapter 2, which used high temporal resolution monitoring to collect data on the hydrological response to rainfall and pumping events. A multi-scale model normally used to model
Chapter 2 assessed the hydrological functioning of the wet grassland site at Otmoor through an intensive hydrological monitoring programme over 18 months. It was determined that the hydrology on site is dominated by precipitation and evaporation, with pumping also being important when instigated by the land managers. Runoff was observed only during times of very high water levels, and regional groundwater interaction was precluded by an impermeable Oxford Clay substrate. Evaporation was estimated and measured using several techniques, and water loss through this conduit shown to be high, up to 5 mm day$^{-1}$ in summer months.

Surface drains, or grips (Figure 4.2), have been installed to increase management of surface water levels and to provide marginal habitat for breeding waders. These features have been shown to irrigate the field centres when water is available to fill them, but to facilitate water loss through evaporation during summer months by allowing a shorter connection to the atmosphere for soil water. The high evaporation and consequently high seasonal range of water levels across the site, generally 1.5 m during each year, means that flood storage is a key potential hydrological function of the wetland. The sub-surface storage is only available during summer months, as during winter, water levels
4.3 – Catchment Scale Conclusions

Chapter 3 has used increasingly complex numerical models to assess wetland hydrological functioning, and at a simple level compare different wetland sites. Although both wetland areas have shown large flood storage potential, it has been demonstrated that wetlands with different structures and management can have strongly contrasting storage components, with Tadham (Somerset) dominated by subsurface and ditch storage and Otmoor (Oxfordshire) dominated by above-ground storage.

A full hydraulic model of the River Cherwell has allowed further investigation of the hydrological functioning of Otmoor, with the disclosure of the reservoir-nature of the flood storage there. The wetland has acts as a reservoir, reducing the volume of water in the river during peak flow, but having little or no impact on time to peak flow, indicating that the primary impact is through bulk storage and not attenuation through increased roughness, although some important effect on the time to peak flow was seen during event H. In contrast to Otmoor, the online floodplain components of the River Cherwell valley have displayed flood wave attenuation, increasing the time to peak flow and not peak flows in the river channels.
Although the differing results from the two wetland types are important, some reservation should be applied as their initial conceptualisation and methodology within the model structure was different, although the methods used were justified and believed to be correct. The online storage component was modelled using extended cross sections and the Otmoor component was modelled using a reservoir system. This confirms the importance of initial conceptualisation, as advocated by Acreman’s (2004) risk-based approach involving iterative development of model complexity.
4.4 – Linking Scales

Several instances of scale issues emerged during the plot scale analysis, including a range of spatial scales of water level change from 10 cm at a daily frequency to 1.5 m between seasons. Indeed, were it not for a targeted approach to scale issues, the important diurnal fluctuation in water tables would not have been observed.

A multi-scale model was utilised to investigate scale issues directly. Using the JULES code (Essery et al., 2001; JCHMR; 2007) designed to model water and energy balances at the global scale, its traditional application was adapted and the model was driven with different scales of data. The model worked effectively at the coarse scale for which the code was intended, using global scale data, predicting evaporation levels at Otmoor with high accuracy. Also, when the model used predominantly field scale data (meteorological, vegetation and soil properties), model performance was adequate, although not as good as with global scale data. However, when the scales of driving data were mixed, with field scale meteorology and global scale soil parameters, model performance dropped substantially.
During catchment scale investigations, it has been shown that attenuation through roughness is less important at Otmoor for its impact on downstream river flow, but that bulk storage is dominant. However, as the opposite has been shown to be true for the online floodplains of the River Cherwell valley, roughness seems important at the catchment scale as a concept. As both surface roughness and bulk storage depend on variations in surface topography, albeit at different scales, it may be concluded that roughness in the general sense is implicitly relevant to wetland hydrological functioning across the spectrum of scales, from micro to plot to catchment and beyond. It was anticipated after the literature review in Chapter 1 that roughness would be important to wetland hydrology across scales, as it can influence many facets of how wetlands interact with the wider hydrological system, from surface storage volumes to different river channel roughness values causing different floodplain inundation extents. Figure 4.3 shows a key figure from Chapter 1, with different scales of roughness and their likely impact on catchment hydrology. The most important roughness component for the Otmoor wetland system has been shown to be surface water storage sitting in surface undulations (2), as this dominates the wetland hydrological functioning at the plot scale (shown in Chapter 2) and impacts very strongly on the downstream flow at the catchment scale (Chapter 3), and so being fundamental to the wetland’s behaviour across the spectrum of scales. Also important was the river channel roughness (1), and this was used as a calibration parameter for the hydraulic model. The channel roughness will affect the channel conveyance and so level of floodplain (including Otmoor) inundation. However, changes in channel roughness are unlikely to vary dramatically over any but the shortest distances, and changes will be averaged
The multi-scale model used in Chapter 2 to investigate the scale effects of evaporation modelling provided some important results for scale issues. The primary conclusion was that the datasets used to drive a model need to be of commensurate scale, or model performance can drop considerably. The model worked very well at the very coarse scale for which the code was designed, and also well when used in conjunction with predominantly plot scale data. However, when scales of driving data were mixed, model performance fell sharply. This is logical when compared to other scaling theories introduced in Chapter 1, notably that of characteristic velocity, whereby spatial and temporal scales are required to be commensurate (Haltiner and Williams 1980; Blöschl and Sivapalan, 1995). Furthermore, Wiens (1989) stated that the predictive power of a model is improved when the spatial and temporal scales of data used are analogous, which again seems intuitive. The finding from the multi-scale model stated above reinforces these logical steps that can be taken to avoid problems associate with scale issues.

Scaling has been avoided as far as possible in the current study so as to investigate effects of scale independent of scaling. Some scaling was unavoidable for the level 2 model, in order to be comprehensive with the coverage of the rainfall-runoff model; the gauged (and so calibrated) catchments were scaled by area to estimate the ungauged sub-catchments. However, this method was essential to enable the construction of a comprehensive hydraulic model for the Cherwell catchment and the methodology was justified.
In assessing the models developed in Chapter 3 of the current work, it has been found that generally they are of appropriate scale in terms of model code choice, input data and application. One exception to this is the discovery that the River Ray cross sections should be updated if further hydraulic modelling is to take place. That is not to say that existing modelling (including in the current work) has been undermined, as any inaccuracies in river channel representation are likely to be relative to other inaccuracies and uncertainties in the models, both rainfall-runoff and hydraulic. The exercise was beneficial for uncovering limitations of data and model structure, and such procedures should be incorporated into existing modelling protocols such as those proposed by Refsgaard (2007a) and Old et al., (2005). This is likely as scale issues become more widely recognised across general wetland hydrology scientific practice, and not just in the domain of journal special issues.
4.5 - Recommendations

When modelling wetland hydrology, it has been shown that scale issues are incredibly important, especially when choosing model code and creating or searching for driving data. Data must be appropriate for both the model code and the application. In terminology highlighted in Chapter 1, the modelling scale must be as close as possible to the process scale. Contemporary models are becoming increasingly data intensive, and observations undertaken in order to drive models must at a scale which reflects the processes occurring. Hence for successful, reliable and therefore transferable modelling to be undertaken, the process, observation and modelling scales must be similar in nature. Hulme (2007) recognised the need for regular interaction between field monitoring and modelling, whereby a cycle is developed of focussed monitoring to improve modelling, which in turn can focus monitoring needs.

This work has assessed the hydrology of a wetland at two distinct scales and assessed the links between those scales. At each scale, the work has been approached in a manner typical of recognised contemporary wetland hydrological science, at the field and catchment scales respectively. However, working at both scales has enabled the elucidation of different information about the wetland. Scale issues were appreciated from the outset and it was recognised that the understanding from each would be enhanced by results from
the other. Few wetland studies have appreciated this fact, and almost none undertaken work at multiple scales. This project has systematically worked from a small (field) scale to the large (catchment) scale, assessing which processes are important at the wetland in question and which processes may be disregarded at the larger scale study. This has facilitated improvement through having the ability to learn from previous models. As the field scale work was undertaken first, followed by the catchment scale model development, information has enabled better catchment scale model development. For example, the knowledge that alluvial flood storage dominated the wetland hydrological functioning allowed the catchment scale models to focus on this and not groundwater interaction. Had the reverse been the case, the models developed in Chapter 3 would have been very different, focusing on finding areas of groundwater-surface water interaction, possibly by utilising water chemistry techniques (e.g. Musgrave, 2006), inferring more regional aquifer interaction. Also, the model code used would have been very different from a 1D hydraulic model, such as a regional-scale in-field hydrological model (see Section 3.2.4).

It would therefore seem sensible that if resources allow, an iterative process would occur whereby catchment scale understanding would also feed into further developments of the field scale work, before the cycle begins again to improve catchment scale understanding. One foreseen application is information on thresholds for stages of floodplain inundation from the River Ray on to the Otmoor floodplain being useful for the development of a distributed hydrological model of the Otmoor wetland, a task which would enhance the current work.
whether through upscaling or downscaling. Model results are not always suitable to be interpreted for use at other scales nor scaled for this purpose. As NRC (1991, p143) summarised:

"Linking and integrating hydrological laws at different scales is not yet fully addressed".

Furthermore, whilst discussing modelling, Beven (1995, p263) concluded that it is:

"...unlikely that any general scaling theory can be developed due to the dependence of hydrological systems on historical and geological perturbations."

Both of these statements are as true today as they were thirteen years ago. For these reasons it is concluded that the future of successful wetland hydrological modelling is likely to lie with the continued improvement of separate models for different scales. This must be embedded in good practice alongside increased appreciation of scale issues and assessment of modelling procedures with such issues in mind. It is the linking of these congruent models which will provide the utility of trans-scale modelling, and so investment must be focussed on improvement of universal model linking approaches, such as the OpenMI scheme (Tindall, 2005), which aims to link different model types of disparate origin. The linking of distinct groups of models, such as the MIKE suite of codes (e.g. Thompson, 2004; Thompson et al., 2004) is also a direction to follow, whereby a comprehensive suite of models covering all aspects of wetland
hydrology are linked, and utility improved. This methodology is sympathetic with the conclusions drawn by Becker and Braun (1999, p251).

"There are land surface units differing in their hydrological behaviour and the controlling land surface characteristics so significantly from the surrounding environment as, for instance, surface water bodies, wetlands, irrigated and shallow groundwater areas ... that they need to be modelled separately... The models or their parameters should not be extrapolated/regionalised across the aforementioned land surface discontinuities."

This suggests that a balance is required to be struck, whereby an appropriate technique is adopted. If the landscape is homogeneous at the target scale, then a lumped modelling approach will be sufficient, and Becker and Braun (1999) agree that this is appropriate sometimes. However, where large differences are likely to be noticed or have an impact at the modelling scale, or indeed the policy scale at which results are to be utilised, then a distributed model is required. Where a distributed model is inappropriate due to the limitations of scale which would be required, then it is proposed that a linked modelling approach is the only feasible option at present.

Otherwise there is a need to develop models which are more flexible and which can deal with different scales of operation simultaneously, rather than require scaling of datasets, parameters or outputs to take place. This can be done in two ways, the first being high resolution grids which are aggregated in regions of low target resolution, yet retain high resolution in areas which may require it; such an approach is adopted by Ewen et al. (1999). The second method is to
employ a so called 'object oriented' approach through local grid refinement at areas of high heterogeneity or anisotropy. This has been utilised by Jackson (2001) in the development of the ZOOM model code, and shows great potential in application to wetland modelling.

The following recommendations are made for future work not only assessing the hydrological functioning of wetland environments, but other hydrological assessment.

1. Scale issues should be incorporated into standard wetland hydrological assessment (at all scales);
2. Modelling should be undertaken using appropriate scales, with effective linking between models where possible;
3. An iterative process of improving conceptual understanding between scales will benefit model development;
4. Multi-scale modelling is not necessarily required, but sensible and logical steps should be taken to appreciate scale issues:
   a. Keep spatial and temporal scales similar (characteristic velocity);
   b. Drive models with data of similar scales, in order to maximise their predictive power.
Chapter 5

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