HYDROECOLOGICAL MONITORING AND MODELLING OF RIVER-FLOODPLAIN RESTORATION IN A UK LOWLAND RIVER MEADOW

By

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A thesis submitted for the degree of Doctor of Philosophy

SIGNED DECLARATION

I, Hannah Marie Clilverd 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

Channelization and embankment of rivers has led to major ecological degradation of aquatic habitats worldwide. River restoration can be used to restore favourable hydrological conditions for target processes or species. This study is based on rarely available, detailed pre- and post-restoration hydrological data collected from 2007–2010 from a wet grassland meadow in Norfolk, UK. Based on these data, coupled hydrological/hydraulic models were developed of pre-embankment and post-embankment conditions using the MIKE-SHE/MIKE-11 system. Fine-scale plant and chemical sampling was conducted on the floodplain meadow to assess the spatial pattern of plant communities in relation to soil physicochemical conditions. Simulated groundwater levels for a 10-year period were then used to predict changes in plant community composition following embankment-removal. Hydrology was identified as the primary driver of plant community composition, while soil fertility was also important. Embankment removal resulted in widespread floodplain inundation at high river flows and frequent localised flooding at the river edge at lower flows. Subsequently, groundwater levels were higher and subsurface storage was greater. The restoration had a moderate effect on flood-peak attenuation and improved free drainage to the river. Reinstatement of overbank flows did not substantially affect the degree of aeration stress on the meadow, except along the river embankments where sum exceedance values for aeration stress increased from 0 m weeks (dry-grassland) to 7 m weeks (fen). The restored groundwater regime may be suitable for more diverse plant assemblages. However the benefits of flooding (e.g. propagule dispersal, reduced competition) may be over-ridden without management to reduce waterlogging during the growing season, or balance additional nutrient supply from river water. The results from this study suggest that removal of river embankments can increase river-floodplain hydrological connectivity to form a more natural flood-pulsed wetland ecotone, which favours conditions for enhanced flood storage, plant species composition and nutrient retention.

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Chapter 1: River-floodplain habitats and functions

1.1 Introduction

Natural riparian river-floodplain ecosystems are strongly influenced by disturbances due to regular flooding events (Poff *et al.* 1997; Naiman and Décamps 1997; Stanford 2002). They form highly dynamic ecotones (i.e. transitional zones) between terrestrial and aquatic environments that are characterized by high habitat heterogeneity, primary productivity and biodiversity (Grevilliot *et al*. 1998; Ward 1998; Gowing *et al*. 2002a, Woodcock *et al*. 2005). These conditions are driven by the strong hydrological connections between rivers and their floodplains. These in turn facilitate the exchange of water, sediments, organic matter and nutrients that are fundamental in shaping floodplain structure (e.g. plant community assemblages) and function (e.g. riparian production and nutrient retention) (Triska *et al*. 1989; Ward and Stanford 1995; Poff *et al*. 1997; Grevilliot *et al*. 1998; Pringle 2003). In floodplain habitats, fluctuations in the soil water regime, associated with strong exchanges of water with the adjacent river, are important for the creation of a dynamic and varying physical environment (Poff *et al*. 1997; Robertson *et al*. 2001). This variety exerts a strong influence upon species composition, and the creation and maintenance of high biodiversity in floodplain habitats (Ward 1998; Freeman *et al*. 2007).

Lowland wet grassland, the habitat type that characterises the site investigated in this thesis, is defined as grassland growing at sites below 200 m that is subject to periodic freshwater flooding or waterlogging (Jefferson and Grice 1998). A wet grassland’s hydrological regime is one of the most important factors determining the plant communities that are present (Gowing *et al.* 1998; Silvertown *et al.* 1999; Castelli *et al.* 2000; Kennedy *et al.* 2003; Dwire *et al.* 2006; Araya *et al.* 2011). Most commonly, the vegetation structure of floodplain grasslands is influenced by variations in water table depth and by the magnitude-frequency characteristics of flood events (Poff *et al.* 1997). These in turn control the oxygen status in the root zone (Wheeler *et al.* 2004; Barber *et al.* 2004). Soil nutrient availability, local and regional plant species pools and the resultant seed availability also have important effects on wet grassland plant community composition and are indirectly linked to river-floodplain hydrology (Bedford *et al.* 1999; Kalusová *et al.* 2009).

Rivers and their connected riparian zones are widely recognised for the ecosystem services they provide, which are of ecological, commercial and societal value. They include the provision of habitat, flood water storage, nutrient attenuation, the creation of aesthetically pleasing open spaces, and the maintenance of biodiversity (Hill 1996a; Forshay and Stanley 2005; Ward *et al.* 2002; Naiman *et al*. 2010). These services are, however, dependent on strong hydrological links via overbank and subsurface flow that have, in many cases, been disrupted by anthropogenic modifications to rivers and floodplains over the past few centuries (Ward *et al*. 1999; Zedler and Kercher 2005; Kondolf *et al*. 2006).

An estimated 50 **–** 60% of wetlands have been lost worldwide (Davidson 2014). This is largely attributed to the drainage of floodplains and riparian areas for agricultural and urban development, to water abstraction, and to pollution (Russi *et al*. 2013). In England and Wales, over 40% of the total river length is classified as severely modified (Environment Agency 2010), where, due to alteration of the natural flow regime, the overbank flow that historically was a regular occurrence is now regularly prevented, therefore severely limiting the hydrological connectivity between rivers and their floodplains. As a consequence, the transfer of water, sediment, and nutrients to floodplains has been strongly impeded (Tockner *et al*. 1999; Wyżga 2001; Antheunisse *et al*. 2006). This has led to major ecological degradation of numerous aquatic ecosystems (Erskine 1992; Petts and Calow 1996; Nilsson and Svedmark 2002; Pedroli *et al*. 2002).

River embankments are engineered to limit overbank flows onto the floodplain in order to protect adjacent land from flooding. However, river embankment can severely impact flood defence downstream. Embankments lead to increased channel volume and flow depth and reduced resistance to flow, which in turn results in higher flow velocities, decreased contact time of water with sediments that is important for the nutrient filtering capacity of aquatic environments, and increased downstream transport of water (Darby and Simon 1999; Gilvear 1999). The importance of providing ‘room for rivers’ has become apparent given the recent extreme weather patterns and severe flooding in the UK and elsewhere (Hooijer *et al.* 2004; DEFRA 2004; Wilby *et al*. 2008; Met Office 2015a; Met Office 2015b). The projected higher magnitude and increased frequency of extreme hydrological events due to climate change (Wilby *et al*. 2008; Thompson 2012; IPCC 2014) contributes to mounting concerns over the future management of the nation’s rivers and floodplains (Wade *et al*. 2013; Royan *et al*. 2015; NRFA 2016).

River restoration involving the removal of river embankments is an increasingly popular management technique being used to re-establish river-floodplain connections and restore a more natural, dynamic, flood-pulsed hydrological regime (Acreman *et al*. 2003; Blackwell and Maltby 2006; Pescott and Wentworth 2011). The aims of these restoration works are often multifaceted and include enhanced floodplain biodiversity, improved nutrient-attenuation capacity, and the provision of temporary storage of flood water (Muhar *et al*. 1995; Bernhardt *et al*. 2005). Hydrology, in terms of water quantity (duration, depth/extent and frequency of floods) and quality (supply of nutrients and dissolved oxygen), is an important driver of floodplain biodiversity and nutrient-attenuation capacity (Silvertown *et al*., 1999; Baker and Vervier 2004; Forshay and Stanley 2005; Dwire *et al*. 2006). Hence, river restoration that aims to create favourable hydrological conditions for floodplain biota and the biogeochemical cycling of nutrients is also central to the legislative plans of governing bodies which aim to achieve good ecological and chemical status of European waters (Water Framework Directive 2000/60/EC).

1.2 Research rationale, aims and objectives

The study was carried out at Hunworth Meadow on the River Glaven, UK. It focusses on the removal of river embankments along a 400 m reach of the River Glaven in the framework of a restoration scheme. A thorough understanding of hydrological processes and their consequences (e.g. frequency, extent, and duration of waterlogging and overbank flows) is essential for predicting changes in wetland function and subsequent response patterns of floodplain biota, and a variety of ecosystem services, following restoration activities. There is a need, therefore, for integrated, process-based wetland restoration research, in order to inform and improve the success of future restoration efforts. The effects of river restoration on ecohydrological processes are complex, and are often difficult to determine if there is insufficient monitoring conducted before and after the restoration works (Kondolf 1995; Darby and Sear 2008). Consequently, an important objective of this thesis was to establish a rigorous hydrological monitoring programme before and after the restoration that is commonly lacking in river restoration projects, in order to document important baseline pre-restoration conditions against which the major effects of the restoration works could be determined. These data were used in conjunction with hydrological modelling to better understand the long-term effects of river restoration activities under a variety of hydrological conditions.

The principal aim of this thesis is to advance our general understanding of river-floodplain hydrological processes and the impact of river restoration on floodplain soil water regimes, soil chemistry, the floodplain plant community composition, and flood-peak attenuation by conducting a detailed hydroecological analysis of a floodplain restoration (river embankment removal) scheme.

The significance of enhancing river-floodplain interactions, i.e. hydrological connectivity (via embankment removal), on floodplain functioning was addressed with data from an extensive field sampling campaign. This included two years of pre-restoration hydrological and chemical data, and 1.5 years of post-restoration hydrological data. These data are used to address the following research questions.

(i) What is the hydrological and biogeochemical regime of an embanked-river floodplain?

(ii) What is the measured hydrological response to embankment removal?

To better understand the long-term impacts of restoration projects on river processes and associated floodplain ecosystem services (e.g. flood water storage, biodiversity, and water quality), hydrological/hydraulic modelling is undertaken using the MIKE SHE/MIKE 11 system (Thompson *et al.* 2004; DHI 2007a) to simulate the effects of river restoration activities under a variety of hydrological conditions. Coupled surface-groundwater MIKE SHE/MIKE 11 models of pre-embankment and post-embankment conditions at Hunworth Meadow are constructed to simulate the hydrological impacts of embankment removal. Over three years of river discharge and meteorological data, and observed groundwater elevations, are used to, respectively, parameterise and calibrate/validate the models.

Following model calibration, pre- and post-restoration hydrological conditions are simulated for the same period to enable the effects of embankment removal alone to be assessed. The simulation period for this assessment is the decade

2001 – 2010. The MIKE SHE/MIKE 11 simulations are used to address the following research questions:

(iii) What are the effects of embankment removal on key components of river-floodplain hydrology (water table elevation, frequency and extent of floodplain inundation, flood-peak attenuation)?

(iv) How will embankment removal impact river-floodplain hydrology under a range of expected river flow conditions?

Plant species have individual tolerance ranges to aeration stress in the root zone that results in niche-segregation along fine-scale hydrological gradients (Silvertown *et al.* 1999, Araya *et al.* 2011). Fine scale botanical, chemical, and topography data were used to assess the relationships between spatial plant distributions, soil fertility, and soil moisture and oxygen status of the root environment in response to river flow alterations. Using a novel oxygen optode technique, direct measurements of oxygen status in response to changing hydrological conditions are conducted to better understand the use of water-table position as a proxy for aeration stress in plants. A cumulative stress index described by Gowing *et al.* (1998), based on the position of the water table, is employed to predict the aeration stress in the rooting zone of plants and account for spatial patterns in wet meadow plant community composition. Furthermore, niche and habitat-suitability models of plant sensitivity to soil moisture regime and simulations of water table elevation from the coupled MIKE SHE-MIKE 11 hydrological-hydraulic models are used to predict aeration stresses in the rooting zone of plants and the effects of river restoration on plant community composition. The work is undertaken to address the following research questions:

(v) What are the importance of soil moisture and nutrient status in predicting the composition of plant communities on a disconnected floodplain meadow?

(vi) What is the relationship between water table depth and oxygen content in the root zone?

(vii) What are the likely long-term impacts of floodplain restoration on the vegetation?

In summary, this thesis seeks to investigate whether the removal of physical barriers along rivers (i.e. embankments) can re-establish hydrological linkages between the river channel and floodplain that promote a more dynamic, flood-pulsed hydrological regime, a major aim of river restoration schemes globally. With the questions stated above, this thesis addresses the implications of river embankment removal on river processes and associated ecosystem services (e.g. flood water storage, biodiversity, water quality), information that can be used to direct and inform future planning and management of river restoration schemes in the UK and further afield.

1.3. Thesis structure

This thesis is structured into nine further chapters. Chapter 2 provides a multidisciplinary review of concepts and research in floodplain hydrology and biogeochemistry, and details the importance and recognised qualities of floodplains in terms of ecosystem services, the principles and application of river restoration, and the modelling tools that can be used to quantify the hydrological impacts of river restoration. Chapter 3 provides a site description of Hunworth Meadow and the River Glaven catchment, and sets out the aims of the river restoration and the techniques used to implement them. Following this, there are three main sections that present the original research undertaken, each addressing the different topics and questions introduced above. The first section comprises Chapters 4 – 5, and focuses on field survey and monitoring and the use of the resulting data to investigate the hydroecological characteristics of

Hunworth Meadow. Chapter 4 details the physical (topographical and hydrological) and chemical monitoring conducted at the site. Chapter 5 describes the hydrological and biogeochemical regimes of the original embanked river floodplain and the initial responses to embankment removal.

The second main section focuses on modelling. The MIKE SHE-MIKE 11 hydrological-hydraulic model setups are detailed in Chapter 6. This includes specifics of the model parameterisation, development, and calibration (manual and automatic procedures) of the MIKE SHE groundwater and the MIKE 11 surface water models. A sensitivity analysis is also included as an initial step in the calibration process to select the most sensitive model parameters for inclusion in the model calibration. Furthermore, details are provided of an impact assessment method used to simulate pre- and post-restoration conditions for the same extended period and directly assess the impact of the restoration. The modelling results for the pre- and post-restoration models are presented in Chapter 7, which includes analysis of the performance of the models, and simulations of hydrological consequences of the embankment removal for a variety of river-flow conditions (i.e. high and low flows).

The final main section focuses on floodplain vegetation. The vegetation survey methods are outlined in Chapter 8. Fine scale chemical sampling and comprehensive laboratory analyses for determination of plant-available nutrients are described, which includes information of a new method of oxygen analysis in soil air using oxygen optodes (based on fluorescence quenching) (Bittig and Körtzinger 2015). In addition, data inputs used to calculate an aeration stress index (presented by Gowing *et al.* 1998) for predicting plant sensitivity to waterlogging is described. Chapter 9 presents the results from spatial analyses of plant communities in relation to soil physicochemical conditions. This chapter couples the water table simulation results from Chapter 7 with the soil aeration index to predict plant community change associated with the floodplain restoration.

The final chapter summarises the key hydroecological conclusions of the study, and the implications for river restoration practices. It also proposes areas of further research.

The work detailed in this thesis has already been published in / prepared for submission to peer reviewed journals. An account of the field hydrological monitoring and the hydrological responses to the restoration scheme that is derived from the work presented in Chapters 4 and 5 has been published in *Hydrological Sciences Journal*:

Clilverd, H.M., J.R. Thompson, C.M. Heppell, C.D. Sayer, and J.C. Axmacher, 2013. River-floodplain hydrology of an embanked lowland Chalk river and initial response to embankment removal. Hydrological Sciences Journal 58(3): 1-24.

A paper detailing the development and application of the MIKE SHE / MIKE 11 models and their use to assess the impacts of the restoration scheme (Chapters 6 – 7) has been published in *River Restoration and Applications*: http://onlinelibrary.wiley.com/doi/10.1002/rra.3036/abstract.

Clilverd, H.M., J.R. Thompson, C.M. Heppell, C.D. Sayer, and J.C. Axmacher, in press. Coupled hydrological/hydraulic modelling of river restoration and floodplain hydrodynamics. River Restoration and Applications.

A third paper which focuses on the vegetation and soil oxygen status research which is presented in chapters 8-9 has been prepared for submission to *Journal of Vegetation Science*.

Chapter 2: Hydrological, chemical, and ecological characteristics of floodplain environments

2.1 Introduction

This chapter details the hydrological, biogeochemical, and biological characteristics of floodplains and the ecosystem services that they can provide. It considers the history of river channel modification and the associated impacts on floodplain structure and function. The restoration methods employed to enhance and rehabilitate degraded riverine habitats are described, as well as the monitoring techniques and modelling tools used to assess the success of restoration works.

2.2 Floodplain and riparian zone hydrology

2.2.1 Conceptual models of floodplain and riparian zone hydrology

A floodplain can be defined as an area of land composed of alluvium that is periodically inundated by stream or river water (Bren 1993). A riparian zone is the area of land adjacent to streams and rivers, and can range in extent from a narrow band of land between a headwater stream and hillslope to an expansive floodplain that borders a large river (Naiman and Décamps 1997). Traditionally the riparian zone only included the vegetation immediately next to the river channel, but more recently the definition has widened to include a larger area of land alongside the river channel, which often includes the floodplain (Burt *et al*. 2010). In this thesis, the terms floodplain and riparian zone are used interchangeably to describe the area of land adjacent to rivers and stream that periodically floods.

Riparian zones are often described as being at the interface between terrestrial and aquatic environments (Gregory *et al*. 1991; Triska *et al*. 1993a; Vázquez *et al*. 2007; Mayer *et al*. 2010). Many floodplains and riparian zones can be classified as wetlands where the land surface is saturated with water long enough during the year to have a dominant influence on soil biogeochemistry and vegetation (Hill 2000). High water table levels can result from: (1) an excess of water in response to precipitation, which can reach floodplains via surficial and deep groundwater pathways (Figure 2.1); (2) catchment controls on infiltration and runoff, such as topography, geology, soil permeability, and land cover; and (3) possible inputs from overbank inundation (Brinson 1993; Haycock *et al*. 1997; Hill 2000; Jencso *et al*. 2010).

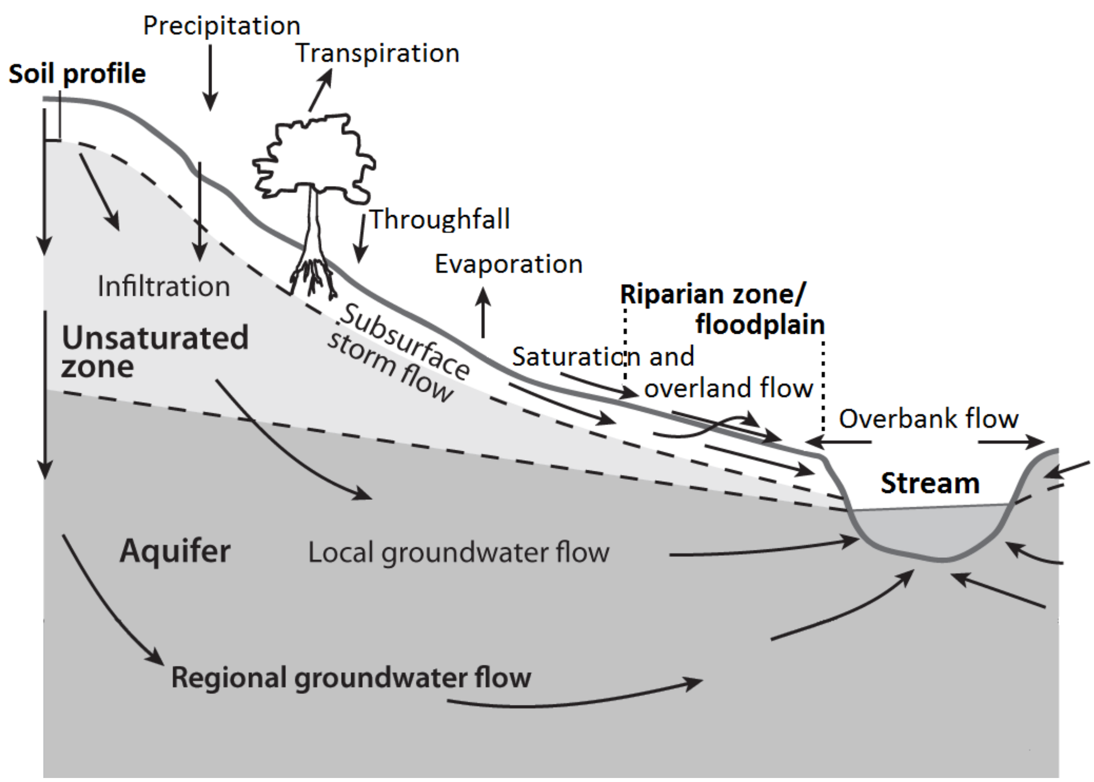


Figure 2.1: Conceptual model of the basin hydrological cycle amended from Lohse *et al*. (2009) indicating the storage and movement of water from upslope to a stream channel.

A water balance for a riparian zone (assuming it is a distinct storage unit) can be defined as follows:

$$Over_{Hill} + Sub_{Hill} + Precip + GW_{Inflow} + Stream_{Seep} + Over_{Stream} - Over_{RZ} - GW_{Discharge} - ET - Perc \pm \Delta Storage = 0 \quad (2.1)$$

These components are expressed as:

INPUTS

A) Overland flow from hillslope to the riparian zone ($Over_{Hill}$)

B) Subsurface flow from hillslope to the riparian zone (Sub_{Hill})

C) Precipitation (*Precip*)

D) Groundwater inflow (GW_{Inflow})

E) Seepage from the stream channel through the bank ($Stream_{Seep}$)

F) Overbank flow from the stream to the floodplain surface ($Over_{Stream}$)

OUTPUTS

A) Overland flow from the riparian zone to the stream ($Over_{RZ}$)

B) Subsurface discharge from the riparian zone to the stream ($GW_{Discharge}$)

C) Evapotranspiration from the Riparian Zone (*ET*)

D) Percolation from riparian zone into aquifers below (*Perc*)

Temporal and spatial changes in the importance of these processes influence the inputs, outputs and storage in the riparian zone. High water table levels are likely for much of year in riparian zones due to their topography (low flat gradients) and location in the landscape (between hillslopes and streams), which results in inputs from the adjacent slopes and stream channel (Hill 2000; Mitsch and Gosselink 2007) (Figure 2.1). Fine-grained alluvial sediments and accumulated organic matter on the floodplain help to sustain waterlogged conditions (Richardson *et al*. 2001). Even above the water table, the soil is likely to remain close to saturation due to capillary action (Richardson *et al*. 2001).

Given their prominent position in the landscape between hillslopes and streams and rivers, riparian zones can moderate and buffer the delivery of water from the surrounding land to river channels. Consequently, riparian zones sustain stream baseflows in interstorm periods, and attenuate downstream flood peak discharges during storm events (Gregory *et al*. 1991; DeLaney 1995; Hill 2000).

Riparian zones differ in the capacity to buffer stream flows based on a number of physical and biological characteristics, such as landscape position, soil porosity, saturation, and organic matter content, and density and type of vegetation (Gregory *et al*. 1991; Tabacchi *et al*. 2000; Jencso *et al*. 2010). Fluctuation of the water table (hydroperiod) above the soil surface is unique to each wetland type. The frequency (recurrence interval), and intensity (duration and area) of flooding can be used to classify wetland type (Figure 2.2). Wet woodlands, wet meadows, marshes, and fens are a sequence of vegetation types that are influenced by an increasing duration of flooding. Riparian wet meadow grasslands, the focus of this thesis, are defined by episodic flooding that can vary in area and depth of inundation (Keddy 2010). These grasslands occupy a relatively narrow space between swamp (lower boundary) and marsh (upper boundary) wetland types along the water level continuum (see Figure 2.2).

Riparian zones have also been classified using hydrogeological models, which can be used to explain the mechanisms that control spatial and temporal variation in surface soil saturation and biogeochemistry (Gilvear 1989; Devito and Hill 1997; Hill 2000). Some upland areas and riparian zones are underlain with impermeable superficial geology, such as clay and dense till, which restricts the downward flow of water and results in shallow aquifers (Figure 2.3b-d) (e.g. Allen *et al*. 2010; MacDonald *et al*. 2014). In this hydrogeologic setting, the water table is likely to fluctuate seasonally and interact considerably with surface soils, which can provide suitable soil water conditions for wet meadow grasslands, and can promote favourable redox conditions for rapid removal of nitrate as groundwater flows through it (Figure 2.3b-d) (Hill 2000; Wheeler *et al*. 2004).

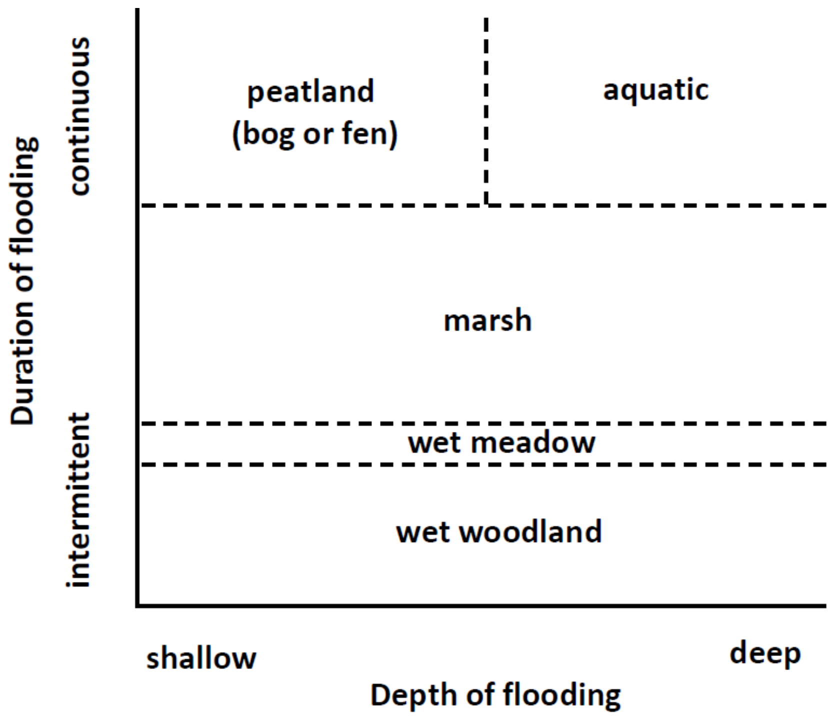


Figure 2.2: The principal kinds of wetlands related to duration and depth of flooding (after Brinson 1993, cited in Keddy 2010).

At one end of the hydrogeologic gradient riparian zones are located in landscapes where groundwater fluctuation is very limited, because of shallow soils overlying impermeable materials (Figure 2.3a). The model suggests these wetlands only discharge water during large floods, and thus have limited effect on stream baseflow chemistry, but can produce large flushes of elements during storm flows. In landscapes where groundwater flows through more extensive, but shallow flowpaths, groundwater fluctuations are more varied (Figure 2.3b-d). At the other end of the hydrogeologic gradient, riparian zones with deep permeable sediments connected to thick aquifers have much more stable water tables and redox patterns. Groundwater can bypass surface soils and vegetation at depth to the channel, and thus have a limited effect on stream chemistry (Figure 2.3e).

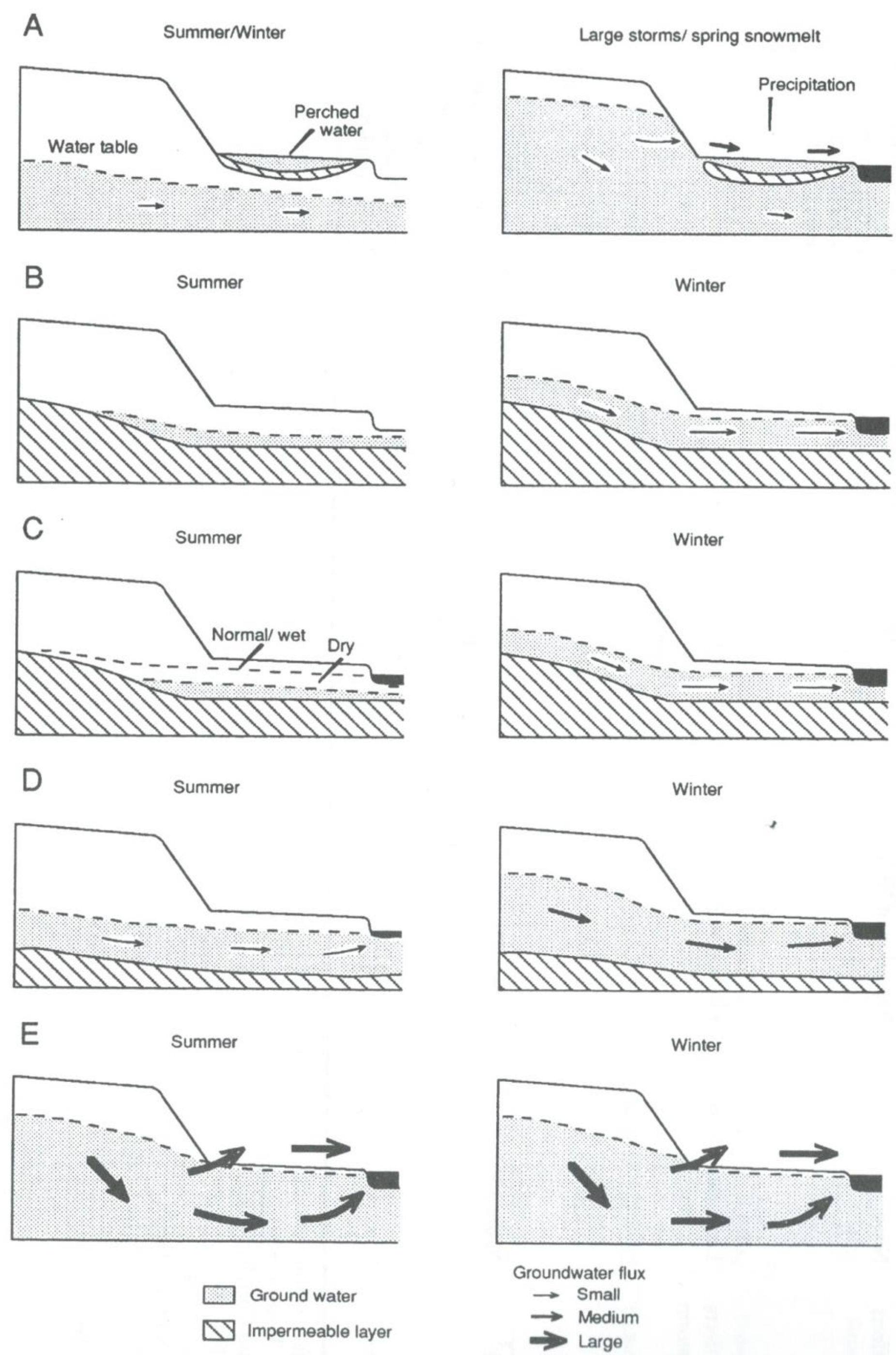


Figure 2.3: Conceptual hydrogeologic model illustrating varying groundwater flow systems in riparian zones. (A) Perched aquifer riparian zone. (B) Thin aquifer riparian zone. (C) Thin aquifer-rain dependent riparian zone. (D) Intermediate aquifer riparian zone. (E) Thick aquifer riparian zone (from Hill 2000).

2.2.2 The hyporheic zone

The hyporheic zone is the region beneath and adjacent to rivers and streams, which contains both groundwater and surface water (Triska *et al*. 1993a; White 1993; Boulton *et al*. 1998). While river banks separate rivers from their floodplains by limiting surface interactions, at depth the mixing of surface water and groundwater in the hyporheic zone can connect biological and chemical processes that are occurring in the river with the surrounding sediments, and vice versa (Jones and Holmes 1996; Crenshaw *et al*. 2010; Williams *et al.* 2010). These interactions result in a dynamic near-river environment that is characterised by enhanced productivity and biogeochemical activity (Findlay 1995; Hedin *et al*. 1998; Morrice *et al*. 2000), and is described as an ecotone between the aquatic and terrestrial environments (Valett *et al*. 1997; Boulton *et al*. 2010; Williams *et al*. 2010). It is important, therefore, to consider the hyporheic zone when studying the hydrological and chemical regimes of near-river environments.

The degree of mixing between surface and subsurface water, and the residence time of surface water in the hyporheic zone has been investigated using conservative tracers. Using such techniques, Triska *et al.* (1989) define the hyporheic zone as the saturated sediment containing 10 – 98 % advected surface water (Figure 2.4). This study was conducted on gravel bars of a pristine third-order stream in California; they found that in porous soils the hyporheic zone extended more than 10 m from the channel, with stream water comprising 44% of flow at their sample wells. In contrast, Stanford and Ward (1993) delineate the hyporheic zone in a biological context, as a saturated zone hydrologically connected with the channel, and accessed by lotic-dwelling macro-invertebrates. This definition can extend the hyporheic zone hundreds of meters from the channel (Stanford and Gaufin 1974). In less porous soils (e.g. organic, sandy loams), however, the hyporheic zone is more likely to extend in the order of tens of centimetres, rather than tens of metres from the river (see Hedin *et al*. 1998). The spatial and temporal variability of the hyporheic zone means that definitions are often based on specific research questions. In this study, therefore, the hyporheic zone is defined as the saturated sediments hydrologically connected to the river channel, characterized by chemical gradients (e.g. in dissolved oxygen, ammonium, nitrate, and dissolved organic carbon) (see Figure 2.4).

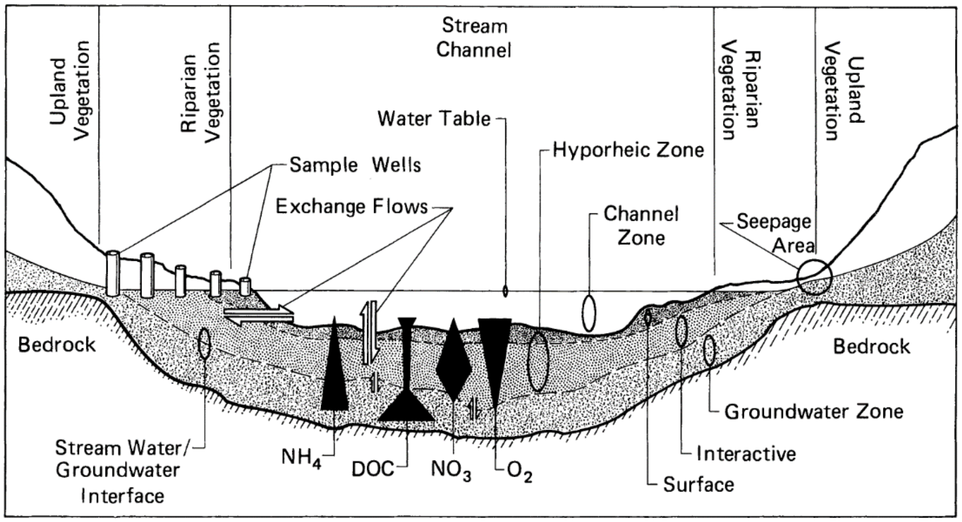


Figure 2.4: Conceptual model of the groundwater-surface water interface (from Triska *et al*. 1989). Three zones are delineated: a channel zone containing surface water, a hyporheic zone, and a groundwater zone. The hyporheic zone is characterized by chemical gradients in NH_4^+, DOC, NO_3^-, and O_2.

2.2.3 Hydrological connectivity and its importance

Riparian zones lie at the terrestrial-aquatic interface, and as such are highly connected to rivers and streams at a range of spatial and temporal scales through the exchange of water and matter (Tockner and Stanford 2002). In a natural state riparian ecosystems form highly dynamic regions that support a range of diverse microhabitats and species, which are maintained by an active balance due to regular floods that continuously reshape the river channels and their banks, and deliver water, sediment and nutrients onto the floodplain (Junk *et al*. 1989). Hydrological connectivity refers to three different vectors of transport: (1) longitudinal, upstream-downstream connectivity, which links reaches along the river; (2) vertical, hyporheic-surface water exchanges between sediments beneath the river and the overlying water column; and (3) lateral overbank connections between the river and its floodplain (Figure 2.5) (Stanford and Ward 1993; Ward and Stanford 1995; Fisher *et al*. 1998; Stanford 2002).

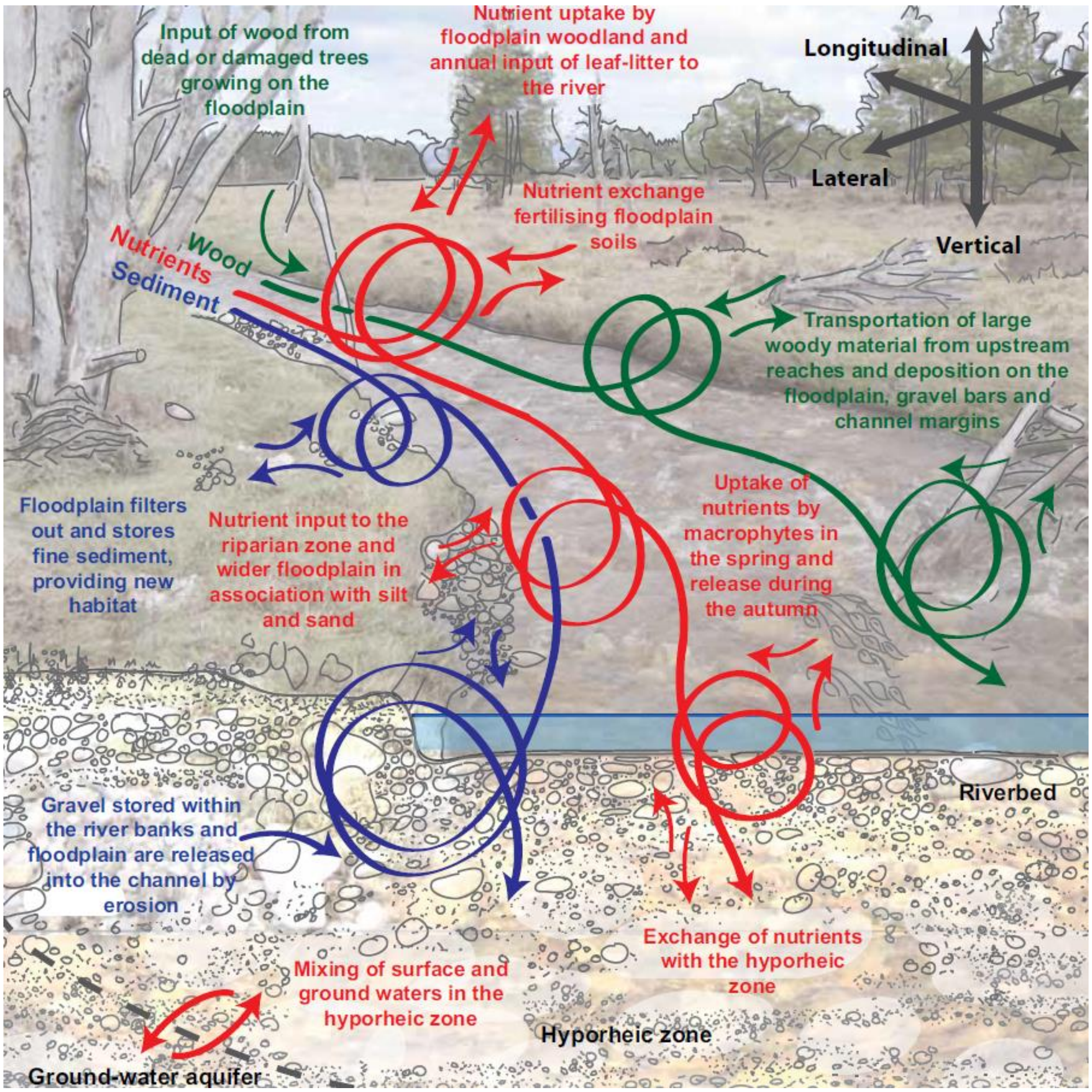


Figure 2.5: Examples of longitudinal, vertical, and horizontal linkages important for sustaining healthy river ecosystems. Red arrows denote nutrients, green arrows signify woody material, and blue arrows denote sediment fluxes. From Perfect *et al*. (2013).

Studies into the importance of longitudinal linkages in rivers and the observable changes in physical conditions from headwaters to mouth resulted in the development of the River Continuum Concept (Vannote *et al*. 1980). This model describes rivers as longitudinally integrated systems and demonstrates how factors in the upper catchment (temperature, organic matter inputs, transport, utilisation, and storage) should vary in a predictable way, and are closely connected to biological processes (respiration, nutrient cycling, biotic assemblages) in the lower catchment. For instance, leaves that enter wooded upland streams provide organic matter (energy) for microbes and macroinvertebrates further downstream. As organic matter is broken down and transported downstream, there is a shift in structural and functional components of streams, i.e. from macroinvertebrates that live on coarse organic particulate matter upstream to plankton that filter fine organic particulate matter downstream. Thus, to understand what is happening at a point along a river, an appreciation of upstream processes and linkages is important (Vannote *et al*. 1980; Minshall *et al*. 1985; Ward 1989; Stanford 1998).

Streams and rivers are also connected by vertical linkages, which result in the exchange of sediment, organic matter, nutrients and oxygen between the river bed and water column. Another organising concept of flowing water, directly linked to vertical connectivity, is the Nutrient Spiralling Concept, which describes the average uptake length and spiralling (release) length travelled by a plant nutrient (usually either N and P) as water travels downstream (Newbold *et al.*1983). This measure is used to determine nutrient retention within a stream, i.e. the shorter the spiral, the higher the efficiency of nutrient retention of a stream. Spiral length is strongly correlated with stream discharge, with short uptake lengths in small streams due to shallow depths and high sediment-surface to water-volume ratios (Peterson *et al*. 2001). Short spiral lengths also indicate nutrient limitation and high biochemical demand of the benthos (Ensign and Doyle 2006). This is typical of healthy streams, which are generally associated with high channel complexity, available organic matter, low nutrient concentrations, and unaltered hydrology (Grimm *et al*. 2005; Meyer *et al*. 2005).

Lateral, subsurface exchanges of water between a river and floodplain sediments is a relatively slow, but constant water-transfer mechanism (generally in the order of cm day^{-1}). In permeable floodplain sediments, subsurface flow of water from the river towards the floodplain is an important mechanism that can facilitate removal of river nutrients via plant assimilation and denitrification, typically 1−2 m from the river in the hyporheic zone where a strong redox gradient exists and nitrate-rich river water or groundwater intersects with alluvium that is rich in organic matter (Triska *et al*. 1989; Jones and Holmes 1996; Hedin *et al*. 1998; Burt *et al*. 1999). Likewise, floodplain interception of shallow subsurface flow from hillslopes can be important for the removal of agricultural fertilisers in groundwater and protection of water quality in water courses (Vidon and Hill 2004a; Billy *et al*. 2010). The degree of hydrological exchange between rivers and their floodplains varies widely and is a function of river-floodplain geomorphology, the magnitude-frequency characteristics of river discharge, and sediment porosity (Triska *et al*. 1993a; Boulton *et al*. 1998; Dahm *et al*. 1998).

Precipitation and rising groundwater and river levels influence the exchange between groundwater and surface water on floodplains, either through up-welling of groundwater or down-welling of the surface water into the aquifer, which serves to extend the hyporheic zone vertically and laterally. Burt *et al*. (2002) present the basic pattern of cross-valley flow direction in floodplain sediments (Figure 2.6), which indicate that during baseflow conditions, water levels are maintained by groundwater discharges from floodplain sediments into the river. During within-bank flood pulses, water levels in the channel are above those in the floodplain, which acts to reverse the hydraulic gradient and direct subsurface flow from the channel into the floodplain. Following flood peaks, as river levels decline, interaction between the surface-water and groundwater interface is maintained in both directions until a hydraulic gradient towards the stream is reinstated (Figure 2.6). Although these controls are fairly well understood, subsurface exchange is thought to be more complex in some settings due to morphology of the channel, valley floor, and hillslopes, as well as during periods of hydrological change (Larkin and Sharp 1992; Woessner 2000; Sophocleous 2002). Indeed, Burt *et al*. (2002) acknowledge that antecedent conditions, local rainfall and runoff, and flood stage collectively complicate the basic pattern presented in Figure 2.6.

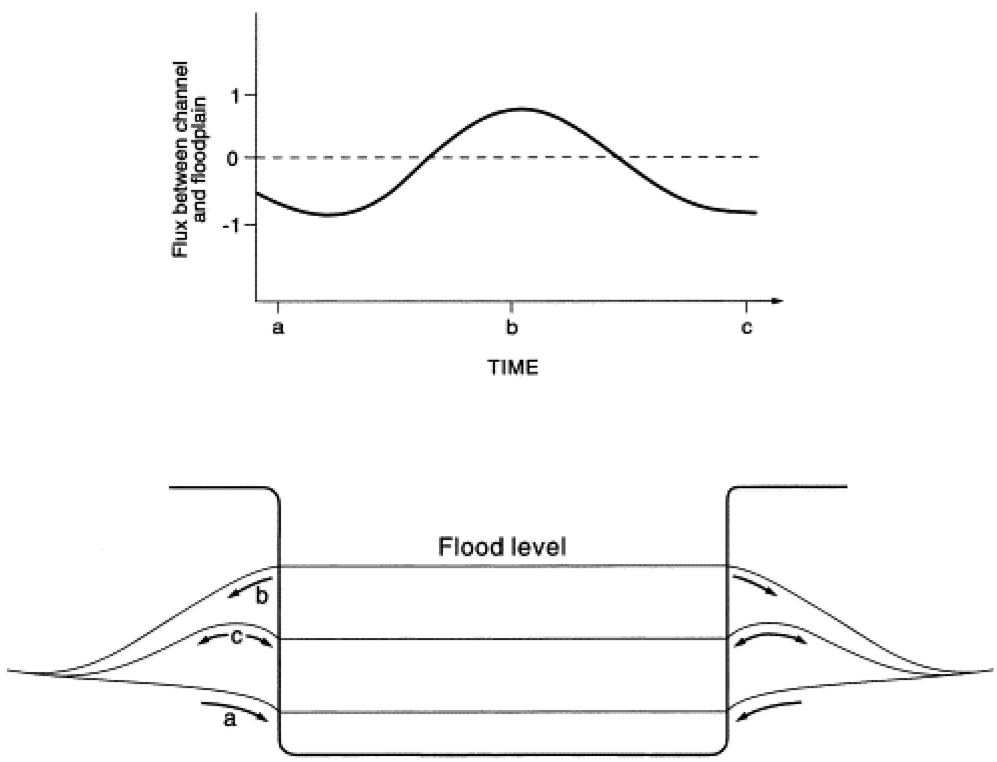


Figure 2.6: Variation in bank storage. At time (a) flow is directed from the floodplain to the river (baseflow), at time (b) a flood peak is passing and flow is directed into the banks; at time (c) the peak has passed and the bank-storage ridge is draining. Cited in Burt *et al*. (2002) based on Dingman (1994).

Voltz *et al*. (2013) present results from salt-tracer injections in streams, which indicate that in steep valley floors surface-subsurface exchange is extensive.

In their study, intrusion of salt-labelled stream water into the riparian aquifer occurred throughout the baseflow recession period (Figure 2.7), and very little change in the water table gradients during high-magnitude stream discharge was observed. This was attributed to a dominant down-valley gradient, and river stage in the channel acting as a boundary condition to the riparian water table. In addition slight fluctuations in hydraulic gradients were observed due to diurnal cycles of evapotranspiration. Patterns of riparian hydraulic gradients and surface-groundwater exchange drive important hydrological and biogeochemical processes in riparian sediments, such as bank storage (Burt *et al*. 2002) and nutrient removal (Dahm *et al*. 1998; Hedin *et al*. 1998) that can result in flood-peak attenuation and improved water quality downstream (Burt and Pinay 2005; Chen and Chen (2003); Harvey and Gooseff 2015). For example, Hedin *et al.* (1998) and McClain *et al*. (2003) demonstrate that near-river environments are hotspots of biogeochemical activity and nutrient transformation, e.g. nutrient removal via denitrification, where hydrological flowpaths converge with substrates or missing reactants (discussed further in Section 2.3.1).

Overbank flow forms a second, more episodic mechanism which can often inundate large parts of a floodplain (Malard *et al*. 2006). Bankfull discharges, which typically occur every 1-2 years in natural systems, are often assumed to control the form of the channel (Darby and Simon 1999), whereas more regular discharges maintain the channel form and smaller scale features such as gravel bars and bedforms (Gordon *et al*. 2004). Overbank flow substantially enhances the intrusion of river water and accompanying particles into floodplain sediments and the underlying groundwater, such that overbank flow may represent a major source of nutrients to floodplain plants and microbes (Triska *et al*. 1989; Jones *et al*. 1995, Pinay *et al*. 1995; Schade *et al*. 2002; Baker and Vervier 2004). In addition overbank flow and the storage of water and sediments on the floodplain can reduce flooding pressures downstream (e.g. Acreman *et al*. 2007). For instance, in a modelling study of a small lowland stream in California conducted by Hammersmark *et al*. (2008), overbank flows in a connected river-floodplain reduced river discharge downstream by up to 25%, whereas no differences in inflow and outflow occurred in the incised river channel scenario (Figure 2.8).

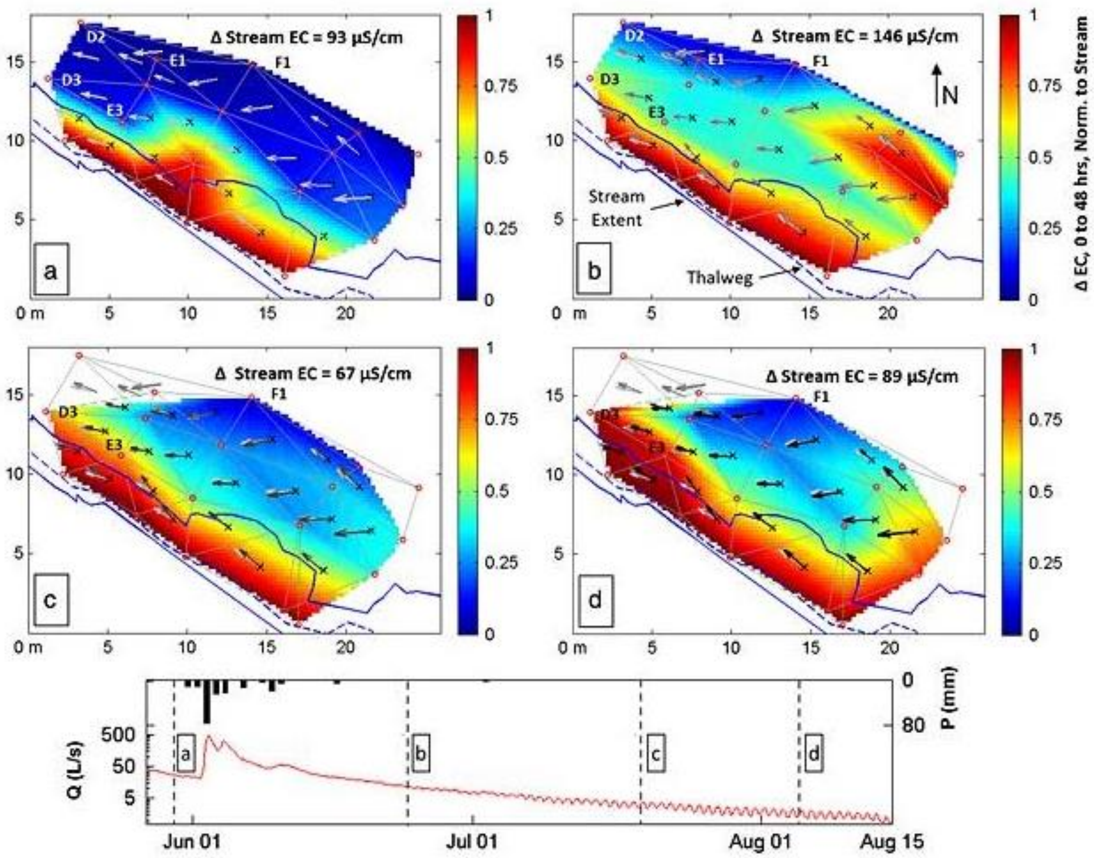


Figure 2.7: Riparian hydraulic gradient and stream-groundwater exchange dynamics in a steep headwater valley. Change in fluid electrical conductivity (EC) of four 48 h constant-rate salt tracer injections is shown. Gradients at each time step are displayed as arrows The salt tracer was injected about 40 m above the upstream-most well transect. Flow is from bottom right to top left. Hydrograph (Q) in red and hyetograph (P) in black are presented in the bottom panel. From Voltz *et al*. (2013).

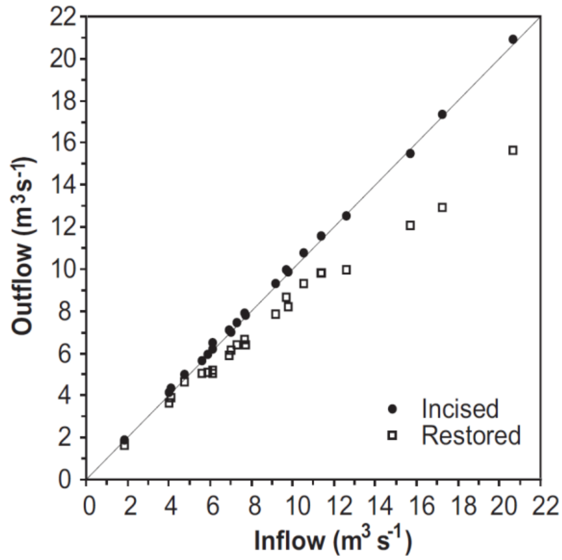


Figure 2.8: Comparison of flood peak inflow/outflow values for incised and restored conditions (from Hammersmark *et al*. 2008).

Rivers and their connected riparian zones are increasingly recognised for the flood defence services they can provide. Recent severe flooding in England, (December 2015 was the wettest and mildest December and 2013/2014 was the wettest winter on record in the UK) and continental Europe (2013 was one of the wettest summers on record in central Europe) (Met Office 2015ab), are forcing priorities in floodplains to change from a focus on development and agricultural production towards the allocation of space along rivers for natural flood water storage and management of flood risk (Hooijer *et al.* 2004; DEFRA 2004; Wilby *et al*. 2008). This is also, in part, being driven because of the likelihood of more frequent and intense rainfall events associated with a warmer climate (Jenkins *et al*. 2010; Murphy *et al*. 2010).

Floodplain storage of water can significantly attenuate many flood peaks (e.g. Ludden *et al*. 1983; Beechie *et al*. 2013; Acreman and Holden 2013). Where river-floodplain connectivity is intact, overbank water is slowed by vegetation on the floodplain (e.g. Dixon *et al*. 2016), which leads to storage of water in ditches and ponds, groundwater recharge, and evaporation, and consequently a reduction in flood flows in the main channel (Anderson *et al*. 2006). Wide floodplains and meandering river channels, in particular, increase flow resistance and have a strong effect on flood water retention and propogation of the flood-wave (Zevenbergen *et al*. 2010).

Woltemade and Potter (1994) examined flood-peak attenuation in major streams of the Grant River watershed, Wisconsin, using the MIKE 11 hydraulic modelling system (see Section 2.6). Overbank flows stored water on the floodplain and attenuated downstream flood peaks (Figure 2.9a-d). Flood events with high peak discharge were most attenuated by overbank storage, for example the events shown in Figure 2.9b and c, were attenuated by 34.1% and 17.4%, respectively. Flood peaks were also reduced during small overbanks events, which resulted in 12.4% flood-peak attenuation (Figure 2.9a). However, a larger, prolonged event was only attenuated 2.9% (Figure 2.9d), which was attributed to limited floodplain capacity for large volumes of water in long-duration events.

Similar reductions in flood peaks were reported in the River Cherwell, Oxford (Acreman *et al*. 2003), and in Bear Creek, Northern California (Hammersmark *et al*. 2008) following river-floodplain reconnections (i.e. embankment removal).Thus natural management strategies that can maintain the self-regulating properties of floodplains such as flood water and sediment storage, may help to protect downstream areas from flooding. Sustainable flood risk management and measures that work with nature are being encouraged through legislative requirements of the Water Framework (Directive 2000/60/EC) and Floods Directives (Directive 2007/60/EC). These policies entail sustainable flood risk measures that work with nature, and allow floodplains to provide flood water storage as well as deliver multiple ecosystem services such as preserving biodiversity, improving water quality, and providing recreational areas.

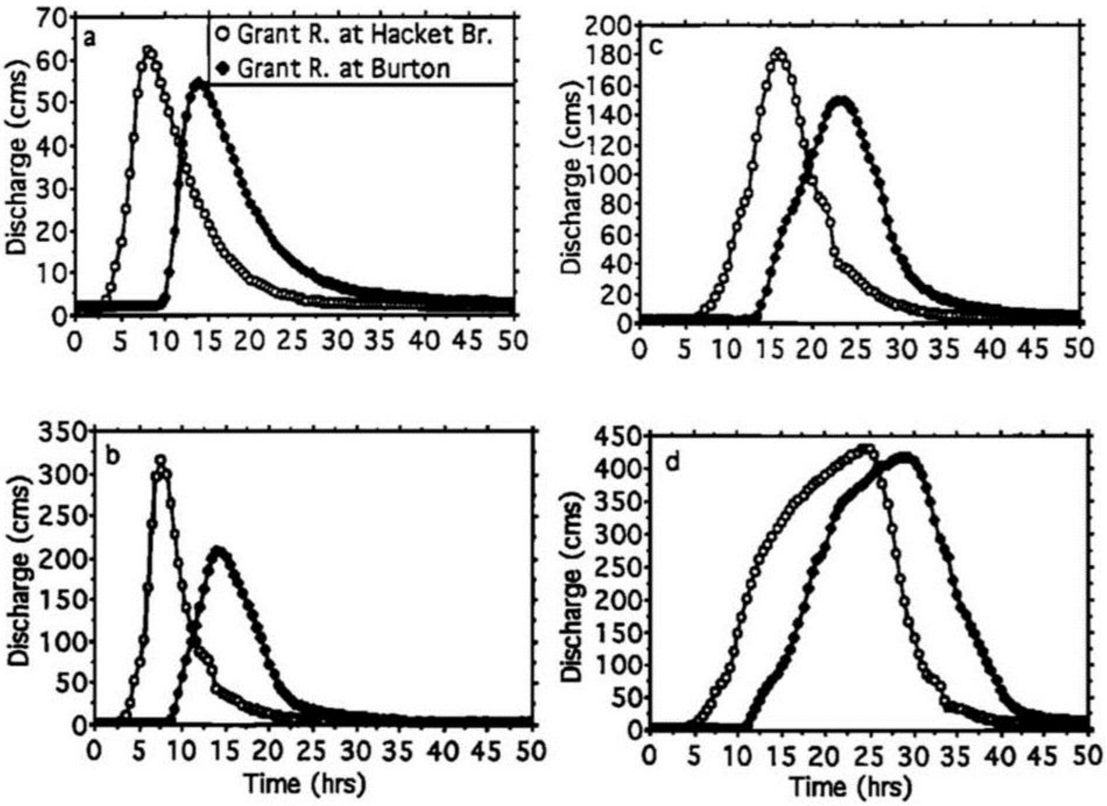


Figure 2.9: Upstream and downstream hydrographs for different precipitation events (amended from Woltemade and Potter 1994).

2.3 Riparian zone biogeochemistry

2.3.1 Biogeochemical transformations in riparian sediments

Surface-subsurface hydrological exchange in riparian zones is important for the supply of dissolved oxygen, nutrients and dissolved organic carbon to hyporheic flowpaths, and microbially active sediments (Jones *et al*. 1995; Pinay *et al*. 1995). Microbial metabolism in hyporheic and groundwater sediments is dependent on the supply of electron acceptors (O_2, NO_3^-, Mn^{4+}, Fe^{3+}, SO_4^{2-}, CO_2) and donors (DOC, CH_4, HS^-, Fe^{2+}, Mn^{2+}, NH_4^+, H_2), which are used according to the yield of free energy (Champ *et al*. 1979; Rysgaard *et al*. 1994; Findlay 1995; Hedin *et al*. 1998; Morrice *et al*. 2000). In oxic environments (e.g. at the river water-sediment interface), oxygen is the dominant electron acceptor used in respiration (Table 2.1). As the residence time of water increases along subsurface flowpaths and demand for oxygen exceeds its supply, anaerobic bacteria use alternative electron acceptors in a predictable sequence (Table 2.1) (Champ *et al*. 1979; Duff and Triska 1990; Holmes *et al*. 1996; Morrice *et al*. 2000).

Table 2.1: Sequence of microbial redox reactions (from Hedin *et al*. 1998), arranged according to decreasing yield of free energy for conditions of decreasing versus increasing electron activity (pE). (a) Energies are calculated per mole available organic matter for reduction reactions and (b) per mole O_2 for oxidation reactions.

Process	Reaction	Free Energy (kJ)
a) Decreasing pE		
1. Aerobic respiration	$CH_2O + O_2 \rightarrow CO_2 + H_2O$	-501
2. Denitrification	$CH_2O + (^4/_5)\ NO_3^- + (^4/_5)\ H^+ \rightarrow CO_2 + (^2/_5)N_2 + (^7/_5)H_2O$	-476
3. Sulphate reduction	$CH_2O + (½)SO_4^{2-} + (½)H^+ \rightarrow (½)HS^- + H_2O + CO_2$	-102
4. Methanogenesis	a) $CH_2O \rightarrow (½)CH_4 + (½)CO_2$	-93
	b) $(½)CO_2 + 2H_2 \rightarrow (½)CH_4 + H_2O$	-66
b) Increasing pE		
5. Methane oxidation	$O_2 + (½)CH_4 \rightarrow (½)CO_2 + H_2O$	-408
6. Sulphide oxidation	$O_2 + (½)HS^- \rightarrow (½)SO_4^{2-} + (½)H^+$	-399
7. Nitrification	$O_2 + (½)NH_4^+ \rightarrow (½)NO_3^- + H^+ + (½)H_2O$	-181

Aerobic respiration and denitrification produce similar energy yields per mole of organic matter oxidized (-501 and -476 kJ free energy, respectively; Table 2.1), hence when anoxia develops denitrification is generally the first anaerobic respiration to occur. Under permanently anoxic conditions, obligate anaerobic bacteria are involved in sulphate reduction and methanogenesis. These processes produce approximately five times less free energy (-102 and -66 kJ, respectively; Table 2.1) than denitrification and therefore are restricted to anoxic sediments where higher energy yielding electron acceptors have been depleted (Champ *et al*. 1979).

2.3.2 Nitrogen and phosphorus biogeochemistry

In near-stream and river environments where sediments are well oxygenated, nitrification, the oxidation of ammonium to nitrite and nitrate by nitrifying bacteria, can result in low ammonium concentration relative to nitrate (Figure 2.10) (Jones and Holmes 1996; Hedin *et al*. 1998; Morrice *et al*. 2000). As hydrological exchange decreases and oxygen is consumed along hyporheic flowpaths, ammonium concentration increases in the absence of nitrifying bacteria (Figure 2.10) (Triska *et al*. 1993b). Denitrification can then cause nitrate concentration to decrease along subsurface flowpaths as it is reduced to nitrous oxide and dinitrogen gases by facultative anaerobes, which can switch between aerobic and anaerobic respiration as environmental conditions change (Figure 2.10) (Hedin *et al*. 1998; Hill 2000). This can result in high rates of nitrate removal in anoxic regions at the river-sediment interface, where supply of nitrate and dissolved organic carbon (DOC) is great (Triska *et al*. 1989; Dahm *et al*. 1998; Hedin *et al*. 1998; Pinay *et al*. 1998; Krause *et al*. 2008).

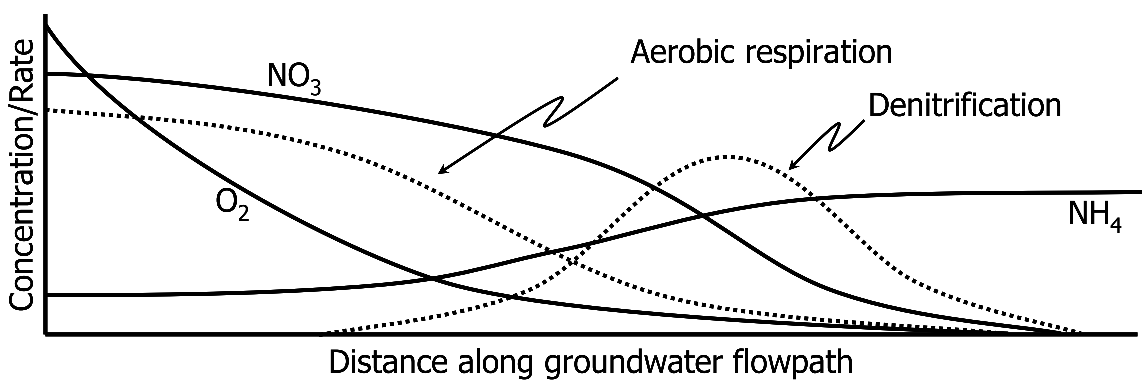


Figure 2.10: Conceptual model of metabolic processes along subsurface flowpaths (from Jones 1994).

Plant uptake, microbial immobilisation, and denitrification are the primary mechanisms accounting for the removal of nitrate from groundwater. Assimilation of nitrogen by plants and microbes stores nitrogen temporarily in biomass. These stores of nitrogen are returned to the available soil nitrogen pool when plants and/or bacteria die and are mineralised. Dissimilatory pathways include nitrification (the conversion of ammonium to nitrate), dissimilatory nitrate reduction to ammonium, and denitrification. Denitrification is of particular interest in eutrophic environments as it results in the permanent loss of nitrogen from ecosystems to the atmosphere in the form of nitrous oxide and dinitrogen gases (Duff and Triska 2000).

The efficiency at which nitrate is removed from riparian zones is known to differ temporally as well as spatially, in relation to variations in soil temperature (Haycock and Pinay 1993) and hydrological pathways (residence time and contact between soil with groundwater) (Hill 1996a; Ocampo *et al.* 2006). During periods of high river flow, denitrification and plant uptake of nitrogen by vegetation can be a substantial sink for nitrogen (Lowrance *et al*. 1997; Schade *et al.* 2002; Banach *et al.* 2009). However, the relative importance of these processes is likely to shift with periods of flooding that can increase denitrification and limit plant activity (Baker and Vervier 2004; Shabala 2011), or during the summer when plant growth and uptake of nutrients is high (Hefting *et al.* 2005), and denitrification may be less important due to lower groundwater levels and increased soil aeration.

Landscape features such as topography, geology, soil and vegetation types in addition to human inputs influence the amounts of phosphorus reaching riparian zones and streams (Gburek and Sharpley 1998; Mengistu *et al*. 2014). Phosphorus is often the limiting nutrient for plant growth (Schulze *et al*. 2005). There are four main reasons for this: (1) the major source of phosphorus is in rocks, hence the release of phosphorus into ecosystems is controlled by the slow process of weathering; (2) unlike nitrogen, there is no atmospheric form of phosphorus and inputs via precipitation are negligible; (3) terrestrial vegetation intercepts most phosphorus; and (4) adsorption to sediments make phosphorus unavailable to plants (Hendricks and White 2000; Kalff 2001).

Dissolved inorganic P is considered bioavailable, whereas organic P must undergo transformation (Figure 2.11). Unlike nitrogen, there is no ‘permanent’ gaseous route for P to be removed from floodplain soils, hence P retention refers to the storage of P in forms that will not easily be released under normal environmental conditions i.e. either in organic forms in plant or microbial biomass, or in inorganic forms occluded in minerals (Dunne and Reddy 2005). The literature largely focuses on the retention of inorganic P, which is controlled by abiotic adsorption onto soil surfaces, and precipitation reactions between P ions and cations such as aluminium, iron, calcium, or magnesium, and biotic microbial immobilisation, and plant uptake (Figure 2.11) (Vought *et al*. 1994; Dunne *et al*. 2005; Reddy and DeLaune 2008). However wetland, including floodplain, soils is often associated with low mineral and high organic matter soils. Thus in many wetlands, where anaerobic conditions persist and decomposition is slow, organic soils provide a long-term storage for P (Roberts *et al*. 2012).

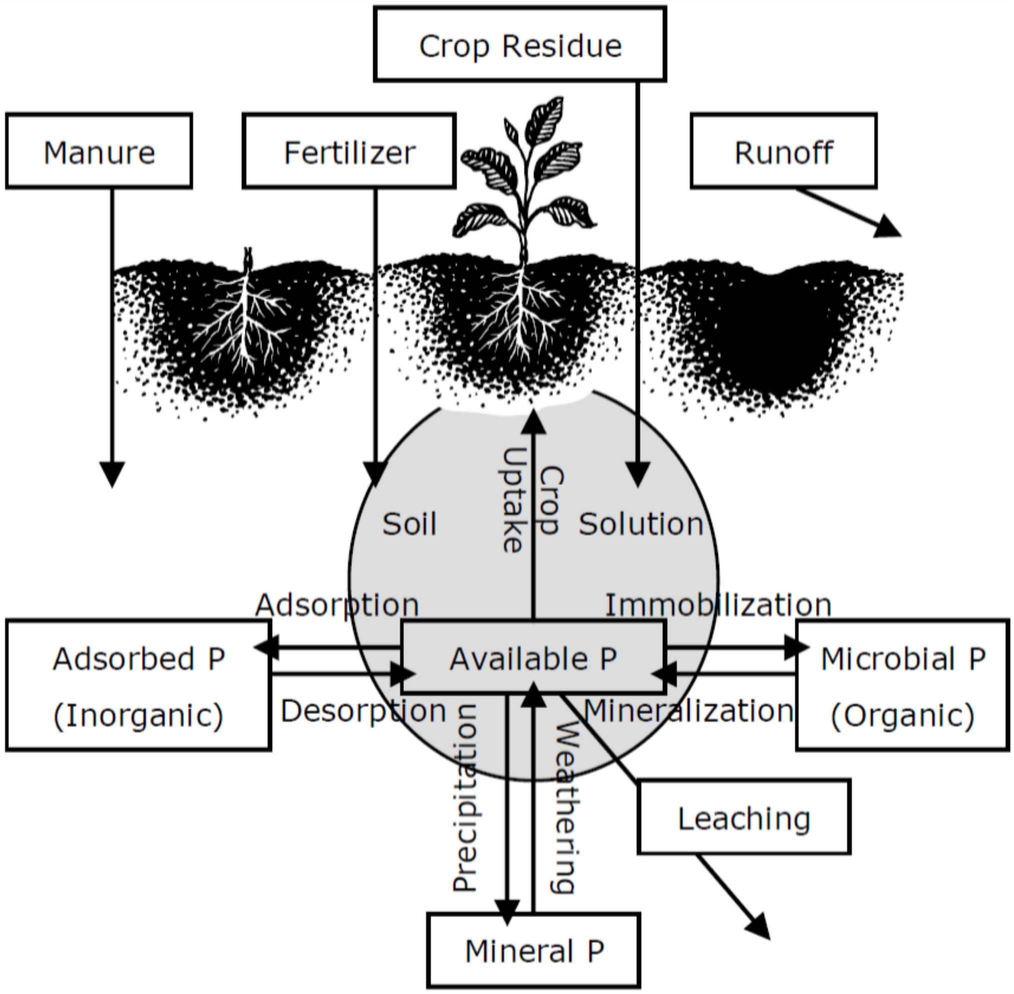


Figure 2.11: Forms and interactions of phosphorus (from Hyland *et al*. 2005).

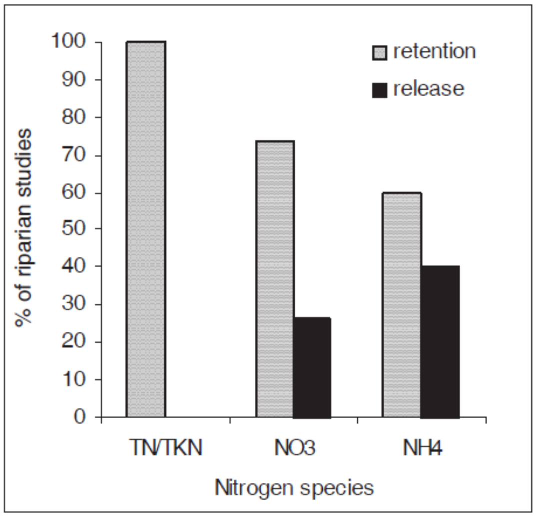
2.3.3 Water quality functions and management of floodplains

Many studies of riparian zones from a number of locations around the world (e.g. USA, Canada, England, France, Denmark, and New Zealand) have reported that riparian areas and floodplains are natural effective nutrient sinks that can substantially reduce nutrient concentrations of water that flows through them. In this way that can act to mitigate the effects of artificial fertilizer runoff on surface water bodies (Lowrance *et al*. 1984; Dosskey 2000; Brusch and Nilsson 1993; Vidon and Hill 2004b; Schade *et al*. 2002; Sabater *et al*. 2003; Billy *et al*. 2010). As such, floodplains are often described as having kidney-like functions within the catchment (Fisher and Acreman 2004; Mitsch and Gosselink 2007; McJannet *et al*. 2010). The functional role of the riparian zone as a nutrient filter is due to several important features. For instance, and as described above, despite occupying a small area of the landscape, riparian zones are prominently located to intercept water and nutrients as they move between terrestrial and aquatic environments (Gregory *et al*. 1991). Positioned at the land-water interface, riparian zone soils are often saturated, which leads to anoxic conditions. As described above, these conditions are conducive to denitrification, the microbially mediated reduction of nitrate to nitrous oxide and dinitrogen gas. Furthermore, riparian soils are usually rich in organic carbon, which is the dominant electron donor in many environments for heterotrophic processes such as denitrification, and can be more important than oxygen status in determining the rate of denitrification (Vervier *et al*. 1993; Holmes *et al*. 1996; Hedin *et al*. 1998; Hill *et al*. 2000).

A study in southern Ontario, Canada, of stream riparian sites on glacial till, found that >90% of nitrate was removed from groundwater flowing from agricultural land into the riparian zone. This was most rapid (i.e. within the first 15 m of riparian zone) at sites with sandy loam soils overlying a shallow 1-2 m confining layer. A greater width of riparian buffer zone for nitrate removal was required in sites with sand and cobble sediments (Vidon and Hill 2004a). In contrast, Burt *et al*. (1999) found that despite significant potential for denitrification (e.g. anoxia and DOC supply) at a floodplain site along the River Thames, UK, the hydrology of the site was inappropriate as water bypassed the riparian zone through gravel lenses beneath the floodplain. Thus, hot spots of denitrification within riparian zones are attributed to key landscape variables such as slope, sediment texture and depth of confining layers on hydrological pathways that link supplies of nitrate and dissolved organic carbon (Vidon and Hill 2004b).

Changes in land use management that optimise the nutrient filtering capacity of riparian zones to reduce diffuse pollution are of great ecological and commercial interest. Using the USGS modular 3-dimensional finite difference groundwater flow model MODFLOW (e.g. Chiang and Kinzelbach 1993), coupled with the MT3D nitrate transport model (Zheng and Wang 1999), Krause *et al*. (2008) found that substantial improvements to the ecological status and water quality of a lowland stream could be achieved through changes in land management. The simulations involved the optimisation of natural buffer zones as well as changes to crop types, such as the extensive use of grasslands on hydromorphic soils, increased set-aside, and an increase in the proportion of deciduous to coniferous forest plantations; and changes in farming methods, such as crop rotation, and a reduction in intense farming.

Fisher and Acreman (2004) conducted a review of 57 studies from around the world that sought to determine the nutrient filtering capacity of wetlands that included many floodplains. They found that the majority of wetlands reduced nitrogen and phosphorus loading (Figure 2.12). Oxygen content and waterlogging of the sediment were the main factors that were attributed to the retention of nitrogen. These factors are often associated with periods of flooding that alter the oxidation state of groundwater in the floodplain (e.g. Clilverd *et al*. 2008). Oxygen concentration was noted as the most important factor controlling the retention of inorganic P, attributed to the binding capacity of iron and aluminium. In contrast to nitrogen, inorganic P retention in soils requires oxic conditions. The onset of anoxia results in lower redox potential which is coupled with pH, and causes phosphorus to desorb from iron and aluminium solids and return to solution (Hedricks and White 2000). However, in alkaline soils, the availability of phosphorus is controlled by the solubility of calcium compounds. Calcium-bound phosphorus is relatively stable and typically unavailable to plants (Dunne and Reddy 2005). Hence, the management of floodplains for nutrient retention is different for phosphorus and nitrogen, and different soil-types.



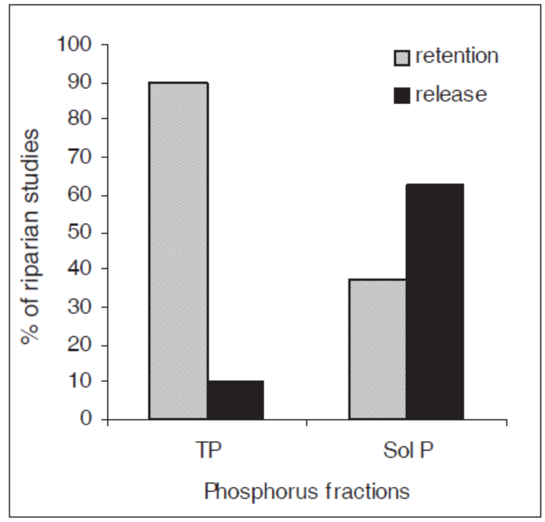


Figure 2.12: The percentage of wetlands studied which exhibited reduction and an increase in N and P loading in riparian zones from Fisher and Acreman (2004). TN/TKN = total nitrogen/Kjeldahl nitrogen; NO3 = nitrate; NH4 = ammonium; TP = total phosphorus; Sol P = soluble phosphorus.

Seasonal variation in the extent and duration of soil saturation affects the demand for oxygen and consequently the redox potential, which controls nitrogen loss via denitrification (Duff and Trska 2000), and P mobility (Dunne and Reddy 2005). As described above, high river stage and flooding can substantially enhance river-water intrusion into the hyporheic zone, supplying DOC, N and P to plant roots and subsurface microbes, which regulate the loss of nutrients (Schade *et al*. 2002; Baker and Vervier 2004; Forshay and Stanley, 2005). However, while floodplains may act as a sink for nutrients during periods of high river stage (Findlay 1995; Bartley and Croome 1999; Adair *et al*. 2004), in contrast at low river stages when groundwater flow may be directed away from the floodplain and towards the river, riparian zones can be a source of nutrients to the river. If, for example, low river stage occurs in the autumn and/or winter, N and P released from plant senescence and root turnover may be a source of nutrients to the river (Kröger *et al*. 2007).

Water table fluctuations, and associated cycles of drying and re-wetting in floodplain soils, can significantly affect soil fertility by causing bursts of microbial activity and soil organic matter turnover (Chepkwony *et al*. 2001). Nitrogen-retention is maximised by varying water table heights. Where there is a close juxtaposition of aerobic and anaerobic zones in sediments, coupled nitrification –denitrification, i.e. diffusion of nitrate from oxidized surface layers to denitrifying bacteria in adjacent anaerobic zones, can result in a strong potential for nitrogen removal (Hedin *et al*. 1998; Strauss *et al*. 2006). In contrast, Pant and Reddy (2000) and Aldous *et al*. (2005) suggest that frequent wetting and drying cycles that alternate between anoxic and oxic conditions maximise the release of mineral-bound and organic P (via mineralisation under drying conditions), whereas stable, moist soil conditions minimise phosphorus release from floodplain soils. A stable hydrological management regime that maximises phosphorus retention is not appropriate however for supporting diverse riparian floral and faunal communities, which typically flourish in environments with naturally fluctuating hydrological conditions (Grevilliot *et al*. 1998; Ward 1998; Gowing *et al*. 2002a; Woodcock *et al*. 2005). This highlights the importance of identifying the goals of riparian management projects and the plans for implementation.

2.4 Riparian zone community composition

2.4.1 The effects of waterlogging on plant community-composition

A wetland's hydrological regime is a fundamental environmental factor that determines the plant community-composition (Gowing *et al*. 1998; Grevilliot *et al*. 1998). The Flood-Pulse Concept (FPC) (Junk *et al*.1989) and its more recent extensions (Tockner *et al.* 2000, Junk and Wantzen 2004) concerns the key role of pulsing river discharge on supply of flood water, sediments and nutrients onto the floodplain. The FPC predicts that recurring overbank inundation is important for the formation of a dynamic physical environment, which drives plant species composition, and high biodiversity in floodplain ecosystems (Ward 1998;

Opperman *et al*. 2010). The FPC has also drawn attention to the importance of the intensity, frequency and timing of flood pulse events for floodplain biota to benefit optimally from flood-deposited resources (Tockner *et al*. 2010). For example, in floodplain meadows the vegetation composition of grassland communities is sensitive to water table fluctuations, fertility, and management (Wheeler *et al*. 2004). While flood disturbances of an intermediate level, in terms of frequency and duration, is expected to enhance environmental heterogeneity on the floodplain and increase species richness (Ward *et al*. 1999), excessive flooding that results in prolonged soil saturation in the rooting zone during the growing season can limit plant growth and cause mortality in plants (Michalcová *et al.* 2011). This can lead to the development of one of a few dominant species that are adapted to these conditions (e.g. *Holcus lanatus, Juncus* spp.) (Grevilliot *et al*. 1998). On the other hand, prolonged flooding that occurs during the winter when plants are dormant is less likely to negatively impact floodplain plant communities (e.g. Beltman *et al*. 2007). Wantzen and Junk (2000) highlight that disturbance sensitivities vary among species and developmental stages of species, and point out that flood pulses should not be considered as disturbance alone, but also a resource of riverine nutrients and organic matter.

Many studies worldwide have reported the zonation of vegetation in response to water table depth below the ground surface (Sánchez *et al*. 1998; Castelli *et al*. 2000; Dwire *et al*. 2006; Toogood *et al*. 2008; Jung *et al*. 2009). Silvertown *et al.* (1999) showed that plant species in European wet meadows have individual tolerance ranges to aeration stress in the root zone that results in niche-segregation along fine-scale hydrological gradients (Figure 2.13). The generality of this mechanism was tested by Araya *et al.* (2011) by quantifying the hydrological niches of plants in fynbos communities in the Cape Floristic Region, South Africa. Despite the vast floristic, functional and phylogenetic differences between fynbos and wet meadow communities, they found that the same trade-offs occurred in response to aeration/or drying stress resulting specialisation of species into distinct hydrological niches.

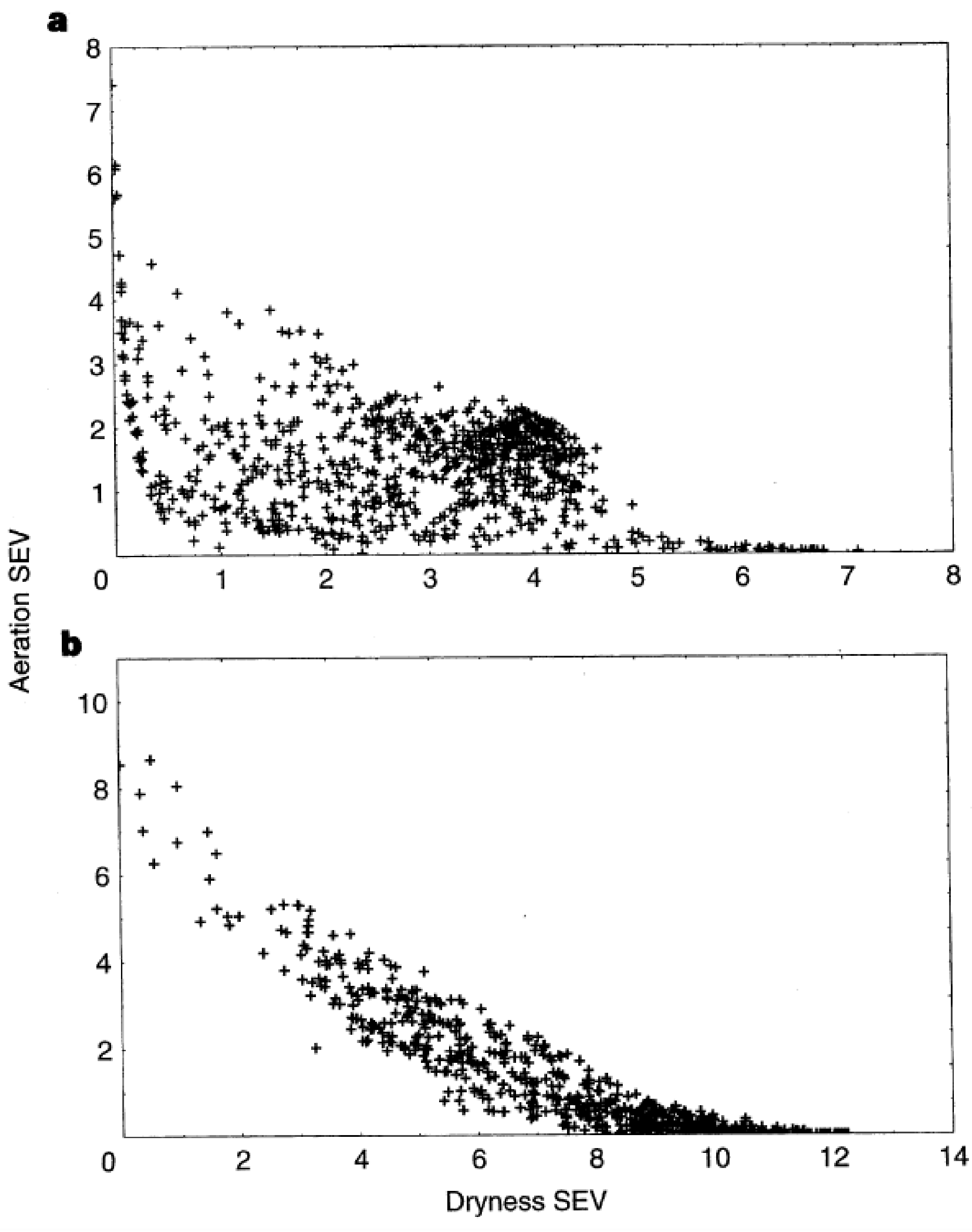


Figure 2.13: Niche space at (a) Tadham and (b) Cricklade. High aeration SEVs indicate waterlogged conditions; high soil-dryness SEVs indicate droughted conditions. Each point is a sampling location for plant community composition. From Silvertown *et al*. (1999).

In wet and waterlogged conditions, root respiration and thus plant growth is limited in the majority of vascular plants. Wetland species that tolerate such conditions are adapted to retain high root-oxygen levels or to provide adequate oxygen supply to their roots (Jackson and Colmer 2005). Where low oxygen concentrations persist plants may also need to cope with the accumulation of metal cations (e.g. Fe^{2+}, Mn^{2+}) and organic phytotoxins (e.g. ethanol and acetaldehyde) released under reducing conditions due to redox-sensitive reactions (Jackson and Colmer 2005; Shabala 2011). Furthermore, nutrient availability changes with soil moisture content, with a maximum availability in mesic soils and minimum availability occurring in waterlogged and very dry conditions (Araya 2005). As discussed by Peñuelas *et al*. (2011), plants are forced to specialise in order to successfully compete for varying demands on resources, and thus hydrological gradients are strongly linked to the biogeochemical niche.

The soil-water regime tolerances of wet meadow vegetation to aeration stress have been investigated using both qualitative (e.g. Ellenberg 1974) and quantitative (e.g. Gowing *et al.* 1998) methods. The above-mentioned studies (i.e. Gowing *et al.* 1998; Silvertown *et al.* 1999; Araya *et al.* 2011) quantified hydrological niche segregation using a cumulative stress index, ‘sum of exceedance value’, based on the position of the water table. Water table depth is used as a proxy of soil water content and air-filled porosity, which in turn is used to determine aeration stress in the rooting zone of plants (Gowing *et al.* 1998; Silvertown *et al.* 1999). This is discussed further in Section 8.3.5. This model was also used by Gowing *et al.* (1997) to predict the aeration and drought stress in the rooting zone of plants and to account for spatial patterns in wet meadow plant community composition in over 2,000 species-rich sites throughout England. The different water regimes were used to provide a quantitative description of favoured water regimes of particular wetland species. This study suggested that stable water table conditions during the growing season are linked with low levels of aeration and drought stress, that can result in high diversity swards (>22 species per m^2). Conversely, extreme fluctuations in soil water conditions result in high levels of combined aeration and drought stresses and thus the lowest species richness (<18 species per m^2). High levels of aeration stress or drought stress typically resulted in low species richness, which is probably due to fewer available species adapted to these conditions (e.g. Silvertown *et al*. 1999).

Further work on the water regime requirements of wet grasslands was conducted by Gowing *et al.* (2002). This research encompassed the analysis of extensive botanical data from 18 wet grassland sites across England. Using sum exceedence values, the preferred water regime of different community types were quantified based on their tolerance to aeration and drying stress (Figure 2.14). The water regime tolerances of wetland communities in Figure 2.14 range from high aeration stress/low drying stress in fens (S25, S24), to low aeration stress/high drying stress in hay meadows (MG5a, MG3). In general, Gowing *et al*. (2002) show that species-rich grassland communities (MG3, MG4, MG5, MG8) are intolerant of waterlogging and were predominantly located on the drier end of the hydrological spectrum, whereas more species-poor communities (MG13, OV28, *Agrostis*-*Carex*) can tolerate waterlogging. The relationship between water level (and thus aeration and drying stress) and vegetation type presented in this study, provide a quantitative method that can be applied to other UK wet meadow sites to predict the possible botanical outcomes of changing water regime.

However, predictions of botanical responses to modified hydrological conditions that allow an optimization of wet grassland management for biodiversity require an understanding of the relationship between the soil moisture and oxygen status of the root environment. As stated above, the cumulative stress index for aeration stress uses water table depth, which is relatively easy to monitor (e.g. Gilman 1994) and represents a useful descriptor of soil water content and air-filled porosity that in turn influence the rate of oxygen diffusion in soil (Hillel 1998). However, air-filled porosity and soil oxygen status are not always strongly correlated. The oxygen concentration in soil is a function of oxygen supply (i.e. diffusion from the soil surface) and consumption (i.e. respiration) rates in the soil profile. Thus, in soils where respiration exceeds diffusion from the surface, soil pores may be filled primarily with respiratory products (e.g. CO_2, CH_4, H_2S) rather than oxygen (e.g. Lloyd 2006). Barber *et al*. (2004) used water content and redox potential measurements to characterize the aeration status of peat soil. Although redox potential was related to water table depth at shallow depths, no significant relationship was found for data from 0.4 m depth (Figure 2.15). This was attributed to rates of oxygen diffusion being less than aerobic respiration. The relationship between aeration status and water table depth needs further study, and more direct measures of oxygen status in response to changing hydrological conditions could help to establish the degree to which the water table position can be used as a proxy for aeration stress in plants.

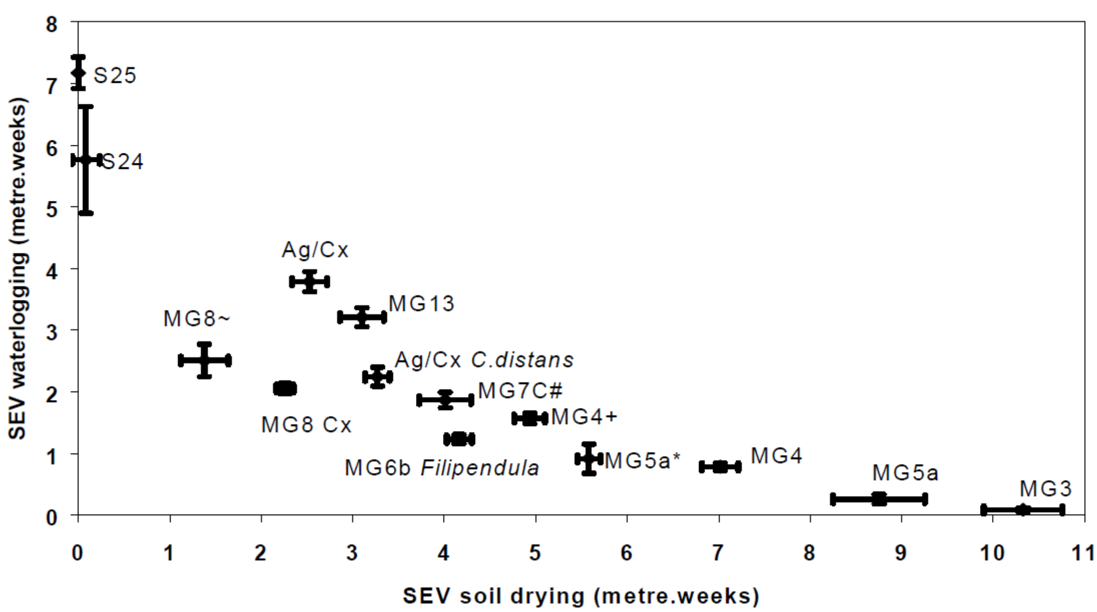


Figure 2.14: Water-regime of each community type (mean and 95% confidence interval) from Gowing *et al*. (2002).

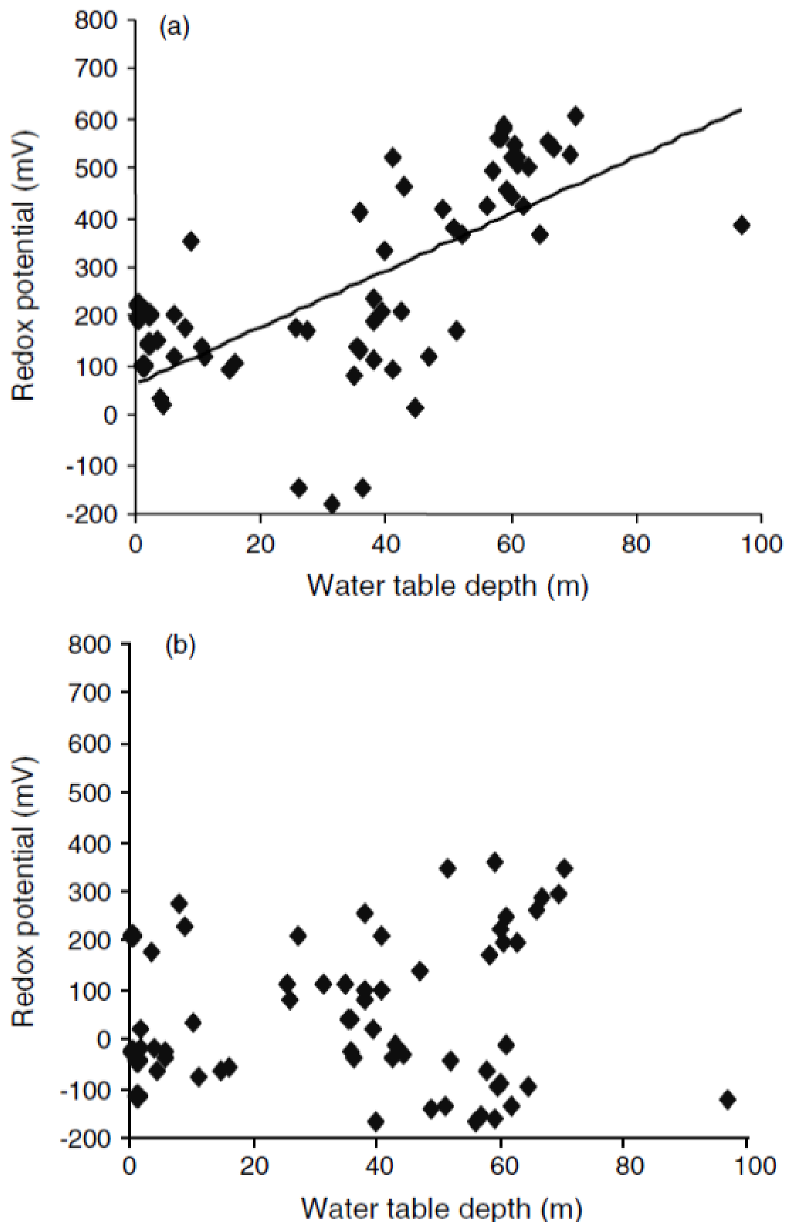


Figure 2.15: The relationship between redox potential and water table depth (a) at 0.1 m depth and (b) at 0.4 m depth. From Barber *et al*. (2004).

2.4.2 Fertilisation of riparian zones

In addition to soil water regime, nutrient availability is an important factor controlling floodplain plant assemblages at the community level (Willby *et al*. 2001), with nitrogen and phosphorus primarily limiting plant growth (Keddy

2000). In oligotrophic environments, the supply of nutrients from aquatic to riparian environments can be essential for floodplain primary productivity (Shade *et al*. 2002). For instance, Lisuzzo *et al*. (2008) report that the nutrient requirements of productive early successional plant communities that colonise nitrogen limited Taiga floodplains cannot be accounted for by nitrogen fixation, mineralisation, and deposition alone, which collectively only account for approximately 26% of community nitrogen requirements. Using injections of enriched $^{15}NO_3^-$ into buried flowboxes (perforated pvc boxes installed in floodplain sediments at a depth of 1.3 m), they found substantial uptake of hyporheic nitrogen downstream in floodplain willow stands, and concluded that during high river stage, nitrogen supply from mass flow of water through the hyporheic zone to the roots of plants equalled or exceeded total nitrogen supply from nitrogen mineralisation and nitrogen fixation (also see Tockner *et al*. 1999; Schade *et al*. 2002).

As discussed above, river-derived nutrients can enter riparian areas via two main transfer mechanisms: (1) through hyporheic flow; and (2) via overbank inundation. Both mechanisms lead to enhanced river water intrusion in floodplain soils; however overbank inundation also has the potential to deposit sediment with its associated nutrients on the floodplain. The rate of nutrient acquisition in plants is generally controlled by diffusion to the root surface (Lambers *et al*. 1998). Therefore in riparian zones, infiltration of nutrient-rich floodwater into the rooting zone can substantially accelerate the transport of nutrients to plant roots. Such fertilisation favours the growth of competitive species at the cost of slower growing species, which can lead to a loss of species richness. Therefore, while floodplains may function to protect adjacent ecosystems from nutrient loading, flood-deposited sediment and enhanced hydrological exchange with nutrient rich river water may pose a risk to floodplain plant communities, particularly where there is a fine balance between species-richness and biomass (Vermeer and Berendse 1983; Waide *et al*. 1999). If the river has become eutrophic, then the removal of river embankments and hydrological reconnection between rivers and historical floodplains (as discussed in Section 2.5) may not be the most appropriate way to restore floodplain biodiversity.

For example, a N fertilizer experiment on a species-rich flood-meadow under hay-cutting management in the Czech Republic by Joyce (2001) showed that species diversity was highly sensitive to nitrogen enrichment, reducing significantly in 8 weeks, with forbs and moss most affected. The study concluded that flood-meadows are susceptible to fertilization inputs from intensive agricultural practice or other human activities (which could include nutrient-rich flood water). Similarly, Beltman *et al*. (2007) found that on a species-rich (average 24 species per m^2) *Arrhenatherion*-dominated grassland floodplain, flooding caused a general reduction in species richness and an increase in biomass production that was similar in magnitude to the effects of fertilisation. As flooding occurred in winter when plants were dormant, inundation was better tolerated by sensitive species than it would have been in summer. Hence, rather than aeration stress, the decrease in species richness after the flood events was attributed to the germination of highly competitive species, initially tall forbs, and later graminoids (e.g. *Carex hirta* - hairy sedge), *Elymus repens* - couch grass). Beltman *et al*. (2007) concluded that increased inundation of floodplain grasslands that already have relatively high biomass production (above-ground production >500 g m^{-2} yr^{-1}) is likely to lead to a reduction in species richness, which is unlikely to recover without sufficient time between flooding (>10 years).

In UK mesotrophic environments, Michalcová *et al*. (2011) report that species richness is best managed for low levels of waterlogging and low soil phosphorus concentration (<10 μg P g soil $^{-1}$). As discussed in Section 2.3.3, under anoxic conditions mineral-bound phosphorus is released into bioavailable forms. This could in part explain the positive relationship between waterlogging and soil available phosphorus, and the resulting negative impact on plant biodiversity in wetlands reported in this study. Similarly, Snow *et al*. (1997), Gowing *et al.* (2002) and Critchley *et al*. (2002) found that species-rich wet grasslands in the UK require low extractable P levels, within the range of 5 – 10 mg P kg^{-1}.

The resource balance hypothesis states that multiple nutrient limitation favours plant species richness (Braakhekke and Hooftman 1999). Multiple resource limitation is thought to be beneficial for plant communities as it results in a competitive balance that permits plant species to coexist. This concept is from the competitive exclusion principle, which states that species that compete for the same resources cannot permanently coexist (Hardin 1960). Braakhekke and Hooftman (1999) propose that species coexistence is described by parabolic 'humped back' curves, with highest diversity when nutrient supply is balanced and lowest diversity at the extremes of nutrient ratios (Figure 2.16), which they investigated in 25 natural grasslands.

This model was also tested by Aerts *et al*. (2003) in a long-term (11 years) study by adding commonly limiting nutrients, N and/or P to two mesotrophic grasslands. They hypothesised that adding nutrients to grassland that was in shortage of N (N:P <10) or P (N:P >14) would increase biodiversity. However instead, nutrient addition resulted in a reduction in species diversity, which was attributed to seed and colonisation limitation. Seemingly, although the correct balance of N:P play a role in grassland diversity, other factors are also important e.g. disturbance in forms of flooding (Helfield *et al*. 2007) and herbivory (Olff and Ritchie 1998; Grace 1999). Furthermore, Braakhekke and Hooftman (1999) recommend that nutrient addition as a 'quick fix' is not prescribed in grasslands to attempt optimal N:P ratios, as this can disturb the balance between nutrient and light limitation. Instead, long-term solutions are suggested, such as biomass removal, which removes non-limiting nutrients more rapidly than limiting nutrients, with an over-arching emphasis on management that considers as many of the regulating factors as possible.

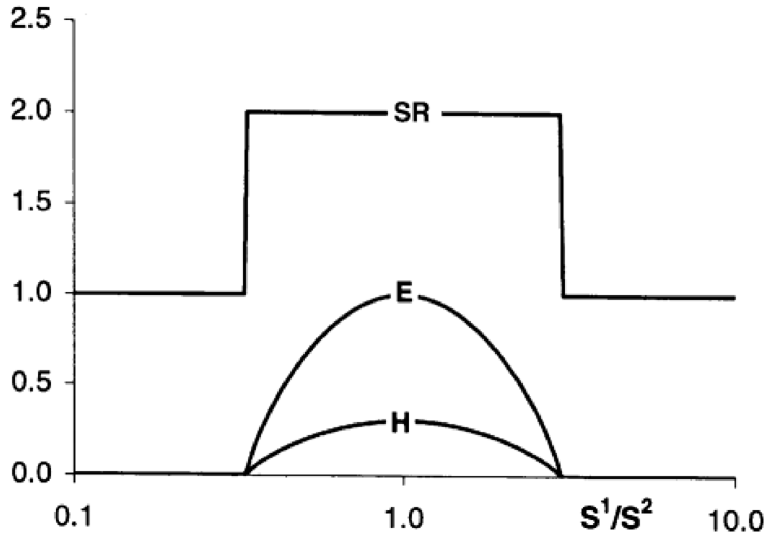


Figure 2.16: Relationship between nutrient supply ratio (S1/S2) and equilibrium species richness (SR), evenness (E) and Shannon index (H) in a theoretical two plant species community competing for two essential nutrients from Braakhekke and Hooftman (1999).

2.4.3 Grazing impacts on plant diversity

Grazing can have both positive and negative effects on species richness, but in floodplains the effect is more likely to be positive due to the high productivity of the vegetation (Proulx and Mazumder 1998). This can be explained by the intermediate disturbance hypothesis (Grime 1979), which predicts that productivity and competitive exclusion is high under conditions of low environmental stress, whereas under high environmental stress few species are adapted to survive. However, when in balance, i.e. following a similar ‘humped-back’ relation to the resource balance hypothesis (see Section 2.4.2), these stresses contribute importantly to the maintenance of biodiversity by reducing the abundance of dominant species and increasing the number of less competitive species.

The impact of grazing pressure on plant species richness in nutrient-rich and nutrient-poor ecosystems was evaluated by Proulx and Mazumder (1998). They compared data from 30 published studies, and found that plant species richness typically increases with grazing in nutrient-rich ecosystems and decreases with grazing in nutrient-poor ecosystems. Under nutrient-poor conditions, declines in species richness are attributed to a limitation of available resources required for regrowth after grazing. In contrast, in nutrient-rich environments, regrowth is less likely to be impacted by nutrient limitation.

Intensive grazing reduces botanical diversity by favouring a few species that can either tolerate repeated defoliation (e.g. *Lolium perenne* (English ryegrass), and *Trifolium repens L*. (white clover)), or are extremely competitive (e.g. *Poa trivialis L* (rough meadow-grass)), or have strong defenses against herbivory (e.g. *Cirsium vulgare (Savi) Ten* (common thistle). Intense grazing regimes also reduce the chance for flowering and seeding (Tallowin *et al*. 1995). Moderate grazing, on the other hand, can decrease competition by dominant plant species, and encourage sward heterogeneity. However, the timing of grazing, in terms of allowing species to set seed, and type of grazing e.g. sheep, cattle, or hay cutting, or a combination, have varying effects on meadow plant assemblages (Table 2.2). Long-term studies on the use of grazing during the re-creation of a floodplain meadow in Oxfordshire, UK undertaken by Woodcock *et al*. (2006; 2011), indicate that under sheep and cattle grazing, plant assemblages are closer to the target species-rich floodplain meadows, compared with un-grazed meadows. Hence grazing is thought to be central to management of plant diversity of lowland grasslands.

Table 2.2: Characteristics of sheep grazing, cattle grazing and hay cutting in agriculturally unimproved grasslands (from Crofts and Jefferson 1999, cited in Vickery *et al*. 2001).

Sheep	Cattle	Cutting
Bite the vegetation, graze close to ground level and produce very short swards of min. height 3 cm.	Bite, pull and tear the vegetation, cannot graze as close to ground level, and maintain longer swards with min. height 5-6 cm.	
Able to manipulate vegetation and select items from very low in the grassland profile.	Coarse level manipulation of vegetation, and relatively unselective grazers.	Completely unselective.
Avoid tall plants in the sward, leave grass stems and often select flowers.	Take tall plants and grass stems, and occasionally select flowers (orchids).	
Dead material and litter left.	Some dead material taken.	
Often avoid rough, tall swards and tussocky areas.	Utilize rough, tall swards and tussocky areas.	
Graze preferentially in small patches, selecting the most palatable patches available.	As sheep. Cattle swards are often particularly patchy.	Leaves swards extremely homogenous in height.
Returns some organic matter in dung and urine.	Returns organic matter in dung and urine. Large dung pats promote sward heterogeneity.	Usually returns little or no organic matter. If cuttings are left as a dense mat they cannot be utilized by decomposers.

2.4.4 Flooding disturbance and propagule dispersal

Pristine floodplains are complex, heterogeneous systems, in part down to regular overbank inundation that increases disturbance. Flooding can create micro-habitats that are important for the coexistence of different vegetation types and thus maintenance of biodiversity (Pollock *et al*. 1998; Ward *et al*. 1999). As discussed in Sections 2.2 and 2.3, hydrological connectivity plays a major part in river and floodplain health and functioning. In addition to the supply of water and nutrients, floods open space for colonisation and aid seed recruitment on the floodplain (Grime 1979; Silvertown *et al*. 1999; Helfield *et al*.

2007; Auble and Scott 1998; Nilsson *et al*. 2010). The local species pool can affect the relative distribution and richness of wetland plant species, and may determine the effectiveness of restoration schemes (Bischoff *et al*. 2009). Transport of riparian plant propagules into the river, and dispersal downstream onto newly flood-created patches for colonization, is dependent on sufficient hydrological exchange between the river and floodplain during flooding (Figure 2.17), and thus is likely to be impeded by river embankments (Auble and Scott 1998). Indeed, Nilsson *et al*. (2010) state that wetland plant communities may often be recruitment-limited. Recent studies that have considered the role of flood pulses for propagule dispersal suggest that flooding is important for providing bare and wet soils, gaps in the vegetation, and the necessary dispersal vector required for the maintenance or restoration of species-rich floodplain vegetation (Leyer 2006; Gurnell *et al*. 2006; Ozinga *et al*. 2009; Merritt *et al*. 2010).

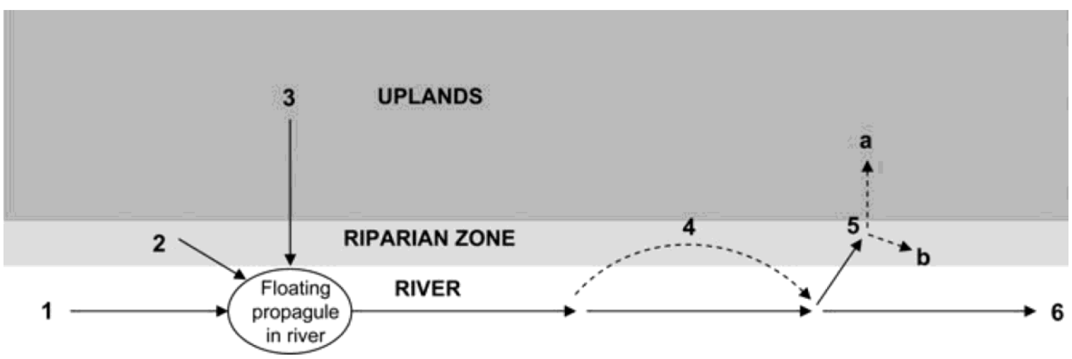


Figure 2.17: Conceptual model showing pathways for water-dispersed propagules in rivers. Floating propagules can be derived from (1) upstream, (2) riparian zones, and (3) uplands. During frequent flooding, (4) some propagules are temporarily stranded and later dispersed before germination takes place. During large infrequent floods, (5) stranded propagules can germinate where stranded, (5a) disperse to uplands or (5b) across the riparian zone via wind and animals. During large floods, (6) some propagules can disperse over very long distances. From Nilsson *et al*. (2010).

Habitat and river restoration efforts (detailed in Section 2.5) that attempt to optimise the water and nutrient status of grasslands in order to enhance grassland diversity often report reduced success when there is a lack of viable seed banks and/or dispersal opportunities for target species are limited (Donath *et al*. 2003). To overcome dispersal limitation along the Upper Rhine in Germany, Hölzel and Otte (2003) conducted a diaspore transfer experiment using freshly mown plant material, in conjunction with topsoil removal to reduce nutrient concentrations and create bare soil for seedling recruitment. After four years, substantial increases in species richness were observed, and rare and endangered plants were introduced. These findings are in agreement with other studies that have found plant species diversity in grasslands to be strongly governed by recruitment limitation (e.g. Tilman 1997; Pywell *et al*. 2002; Fraser and Madson 2008; Hellström *et al*. 2009; Zeiter *et al*. 2013).

2.4.5 Human pressures on lowland wet grasslands

Lowland wet grasslands are mostly semi-natural habitats, subject to periodic freshwater flooding or waterlogging (Jefferson and Grice 1998), and are often characterised by high plant and animal biodiversity. They can support numerous plant species and vegetation types such as grasses, broad-leaved herbs, sedges and rushes (Wheeler *et al*. 2004), and provide habitat for invertebrates, and breeding wading birds such as *Vanellus vanellus* (lapwing), *Tringa totanus* (redshank), *Gallinago gallinago* (snipe), *Limosa limosa* (black-tailed godwit), *Philomachus pugnax* (ruff), *Haematopus ostralegus* (oystercatcher), and *Numenius arquata* (curlew) that are attracted to winter flooding (Ausden *et al*. 2001).

The extent and botanical nature conservation value of lowland wet grassland in Britain declined substantially in the 20th century. Between 1930 and 1984 semi-natural lowland grassland decreased by an estimated 97% to approximately 0.2 million ha (JNCC 1995). These losses have continued, such that it is now estimated that the total area of species-rich wetland meadow in the UK comprises <1,500 ha (Figure 2.18) (JNCC 1995). This is primarily attributed to agricultural intensification, including: channelization and embankment of rivers; substantial increases in the application of inorganic fertilisers (nitrogen and phosphorus); a switch from hay to silage, where more frequent cutting reduces seeding opportunities for plants; and increased stocking densities on grazed meadows, which have altered the ecohydrology of these habitats (Vickery *et al*. 2001; Tilman *et al*. 2002; Benton *et al*. 2003; Kleijn *et al*. 2009). Land development, land drainage, and encroachment by invasive species have also contributed to the recent declines in biodiversity of floodplain meadows (Joyce and Wade 1998). Species-rich vegetation communities have largely been out-competed by relatively dense, fast-growing uniform swards (Joyce 1998; Vickery *et al*. 2001).

Grassland that has not been subjected to artificial fertiliser or excessive hay cutting or grazing is a valued resource of high nature conservation value (Bignal and McCracken 1996; Eriksson *et al*. 2002; Wheeler *et al*. 2004). The continued decline of many lowland wet grassland has led to increased efforts to conserve and restore many of the UK's wetlands. This includes protection at a number of meadows at Sites of Special Scientific Interest (SSSI) and Natural Nature Reserves (NNRs) e.g. North Meadow Cricklade, Wiltshire, and Upwood Meadows, Cambridgeshire (JNCC 1995). In addition, lowland hay meadows (NVC type MG4: *Alopecurus pratensis, Sanguisorba officinalis)* are listed as a priority habitat on Annex I of the EU Habitats Directive (92/43/EEC). In the UK, this community is characterised by species-rich swards containing *Festuca rubra (red fescue)*, *Cynosurus cristatus (*crested dog's-tail), *Alopecurus pratensis* (meadow foxtail), S*anguisorba officinalis (*great burnet), *Filipendula ulmaria (*meadow sweet*)* and *Ranunculus acris (*meadow buttercup*)*, and provides the main habitat for *Fritillaria meleagris (*snake's head fritillary*)* (Wheeler *et al*. 2004). Some of these sites, designated as Special Areas of Conservation (SACs), are of European importance and are protected legally under the Habitats Directive (92/43/EEC). In the UK they form part of the European Natura 2000 network (JNCC 2016).

Figure 2.18: The distribution of species-rich lowland wet grasslands (MG4 under the UK National Vegetation Communities system) in England from Wheeler *et al*. (2004). The size of the circle reflects the relative area of each site on an arbitrary scale.

2.4.6 Management of lowland hay meadows

Traditional management of floodplain meadows involved a midsummer (July) hay harvest, followed by low-density cattle and then sheep grazing of regrowth in the autumn and winter, respectively (Jefferson and Grice 1998). This maintained a low sward during most of the year, promoted seed dispersal and created niches for seed germination (e.g. Linusson *et al*. 1998; McDonald

2001), and balanced the input of river-derived nutrients to the floodplain with the removal of nutrients in plant biomass by the use of hay cutting and grazing (Wheeler *et al*. 2004). The conversion of semi-natural grassland to intensive agriculture imposed a number of abiotic and biotic constraints on floodplain plant diversity, largely linked to the drainage, fertilisation, and impoverishment of seed-pools through habitat loss and fragmentation, in short, an intensification of the low-impact farming outlined above.

Effective management strategies for improving biodiversity and favouring target plant communities also include the removal of topsoil, which in addition to reducing soil fertility creates bare soil and poor competition (Tallowin and Smith 2001; Verhagen *et al*. 2001); diaspore transfer (Kiehl and Wagner 2006; Hedberg and Kotowski 2010), or a combination of these methods (Hölzel and Otte 2003). Those strategies that are most effective are thought to include as many controlling factors as possible, and avoid situations where a single factor can dominate over the others (Braakhekke and Hooftman 1999).

2.5 River regulation, and restoration

2.5.1 Channel modification

In their natural form, rivers are heterogeneous and complex, free to change their form and flow in relation to the natural properties of the river bed, banks, and climate (Poff *et al*. 1997). Strong interactions between the river and floodplain are characteristic of natural river systems and contribute to a state of dynamic balance due to the regular floods that continuously reshape river channels and their banks, and transport water, sediment and nutrients onto the floodplain (Ward 1998). This leads to a patchwork of habitats within the river channel and on the floodplain that can support a variety of plant and animal assemblages (Décamps *et al*. 2010). Despite occupying a relatively small proportion of the total land area, riparian zones support more species than any other terrestrial habitat (Nilsson 1992).

Aquatic ecosystems, however, have been disturbed by humans for thousands of years through the diversion of rivers and streams for harnessing waterpower, deforestation, and drainage and conversion of wetlands for agriculture (Tockner and Stanford 2002). In contrast to natural rivers, modified channels are simplified and diversity and changes to channel form are reduced, connectivity with the floodplain is restricted, and the role of riparian vegetation is diminished (Figure 2.19). The most severe impacts have occurred globally since the late 19th century. Intensified agriculture and industrial pressures has increased the channelization, drainage, deforestation, and pollution of rivers, streams and the riparian zone; in addition the construction of dams, water extraction, and flood control, have had a lasting impact on the hydrological characteristics of river-floodplain ecosystems (Tockner *et al*. 2009; Poff and Zimmerman 2010; Cheng *et al*. 2012; Elosegi and Sabater 2013; Perfect *et al*. 2013; Roni and Beechie 2013). These disturbances have led to a rapid degradation of water quality and habitat, which has impacted humans through impaired water supply and flood control, and affected biota resulting in species loss (Goudie 2006). It is reported that 50 – 60% of wetlands worldwide have been lost (Davidson 2014), and estimates suggest that globally more than 75% of riverine habitats are degraded (Vörösmarty *et al*. 2010).

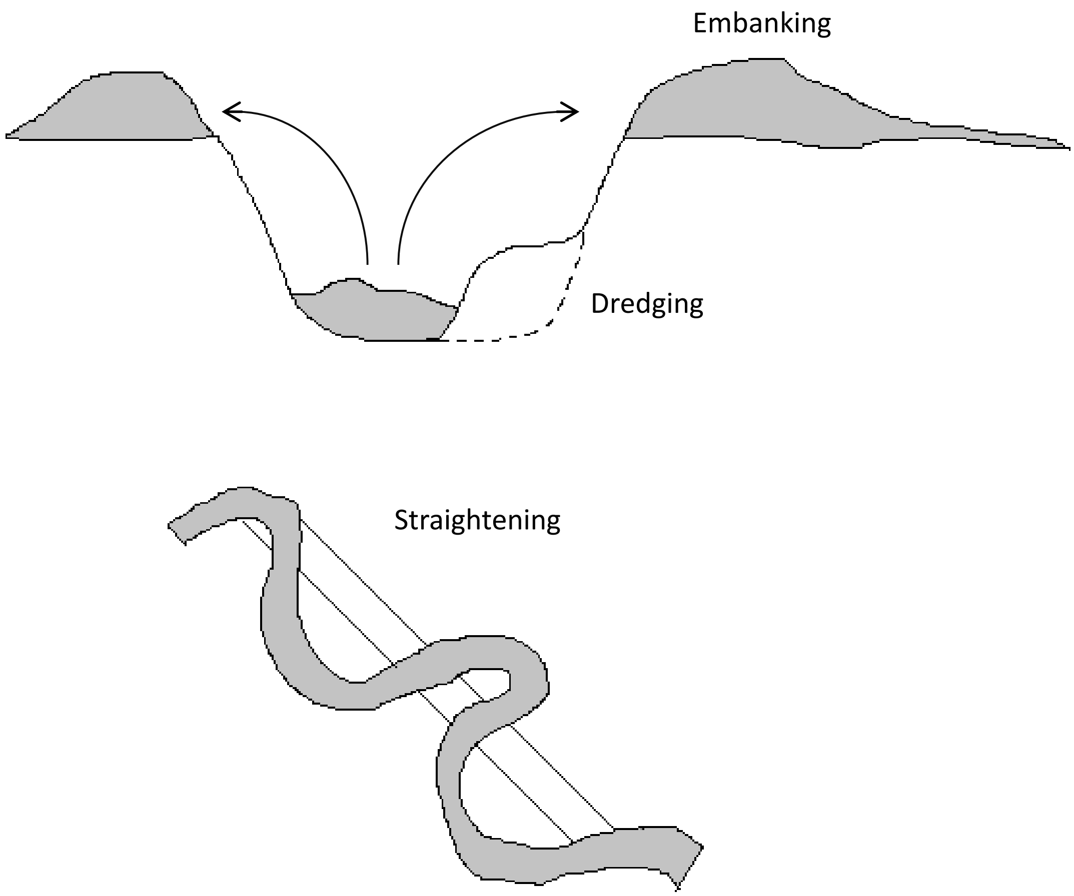


Figure 2.19: Examples of modifications to river channels, amended from Sear *et al*. (2000).

Regulation of rivers and streams over past centuries has had a pervasive impact on the hydrological characteristics of floodplain ecosystems. Many floodplains, where overbank flow was historically a regular occurrence, no longer flood frequently due to alteration of the natural flow regime (Poff and Zimmerman 2010). River embankments limit overbank flows onto the floodplain in order to protect adjacent land from flooding so that it can be employed for other uses including agriculture and urban development (Darby and Simon 1999; Sear *et al*. 2000). However, local river embankment can severely impact flood defence downstream. Embankments lead to increased channel volume and flow depth and reduced resistance to flow, which in turn results in higher flow velocities, decreased contact time of water with sediments (important for the nutrient filtering capacity of aquatic environments) and increased downstream transport of water (Darby and Simon 1999; Gilvear 1999). Other modifications that have altered the flow-regime in riverine ecosystems include channel straightening, simplifying, widening, damming, deepening and dredging, culverting, bank stabilisation, and vegetation removal (Beechie *et al*. 2013). These alterations have a major impact on natural patterns of longitudinal and lateral connectivity, homogenising the river corridor (Poff *et al*. 1997; Tockner and Stanford 2002), which in turn affects in-stream (e.g. Jungwirth *et al*. 1995) and riparian biotic composition (Nilsson *et al*. 1997) and water quality (e.g. Lefebvre *et al*. 2004).

2.5.2. Restoration techniques

River restoration spans a variety of activities, and can range from the creation of new habitats, the partial restoration of lost habitats and habitat improvement, to full restoration of ecosystem processes and function (Box 1996; Beechie *et al*. 2013; RRC 2014). Habitat restoration is intended to assist the recovery of a habitat that has been degraded in order to return the system to its original or near-natural, undisturbed state (Bond and Lake 2003; Darby and Sear 2008). Whilst passive restoration involves the removal of human disturbance to facilitate recovery, active restoration physically alters the habitat (e.g. via the removal of river embankments, re-meandering previously straightened rivers reaches, addition of spawning gravel, introduction of woody debris) and typically focuses on the recovery of natural components and processes e.g. overbank inundation (Hammersmark *et al*. 2010; Barlaup *et al*. 2008; Miller *et al*. 2010). By addressing the root causes of habitat degradation the assumption is that if the physical conditions and processes that control the habitat characteristics are adequately restored then target flora and fauna will respond positively without further intervention (Rohde *et al*. 2006; Katz *et al*. 2009). However, the link between habitat and biotic restoration can break down and inhibit the recovery of historic conditions where there are: (1) barriers to colonisation; (2) new species present (native or non-native); and/or (3) physical and chemical substrate changes (Bond and Lake 2003; Katz *et al*. 2009). In these instances, further intervention and management may be required to attain habitat restoration goals.

For the reasons discussed above, restoration efforts that start from a less disturbed point are in the end more likely to succeed due to a greater species reservoir, and physical conditions more similar to the natural state. They are also likely to be more cost effective (e.g. Wheeler *et al*. 2004). Protection of near-pristine aquatic ecosystems that are already functioning naturally have an important role to play in restoration ecology, in part because they provide a baseline and control point for restoration efforts, but also because continued human pressures such as growing population and climate change are likely to disturb aquatic habitats at rates much greater than restoration efforts.

Several habitat restoration projects, that have physically altered the morphology of over-regulated rivers to restore the natural flow regime, have reported positive effects on river and floodplain plant and animal diversity (e.g. Buijse *et al*. 2002; Nilsson and Svedmark 2002; Nienhuis *et al*. 2002; Poff and Zimmerman 2010; Arlettaz *et al*. 2011). However there are few published studies on small rivers in which the effects of river restoration have been monitored. Small-scale river restoration works have increased in recent years, but they are often initiated by local farmers and landowners that understand the benefits of restoring natural flow regimes for floodplain biodiversity and water management, and are not widely reported in the literature.

Floodplain restoration, through embankment removal (the focus of this thesis) is now being increasingly employed to re-establish river–floodplain connections (Bernhardt *et al*. 2005). The aims of these restoration works are often multifaceted and include enhanced floodplain biodiversity, improved nutrient-attenuation capacity and the provision of temporary storage of flood water (Muhar *et al*. 1995; Buijse *et al*. 2002; Bernhardt *et al*. 2005). The River Restoration Centre (RRC 2016a) lists best-case studies of river restoration from across the UK. Of the 72 projects listed, most were related to in-stream restoration works to improve habitat quality for fish (e.g. weir removal, creation of riffles, pools, and backwaters), however 15 studies involved floodplain reconnection schemes. Four of these were located in urban environments, where concrete channels were replaced with new natural bed profiles in order to enhance flood protection, channel heterogeneity and aesthetic values of the streams (i.e. River Quaggy, Chinbrook Meadows, London; River Ravensbourne, Cornmill Gardens, London; Yardley Brook, Birmingham; Braid Burn, Edinburgh). The remaining floodplain reconnection projects which were documented involved breaching of flood banks (e.g Burn of Mosset, Forres, Elgin), bank regrading (e.g. River Cam, Trumpington Meadows, South of Cambridge), the creation of new sinuous channels (e.g. River Wensum, Bintree), floodplain mosaics (e.g. River Thames, Farmoor reservoir, Oxford) and increased variation in channel depth, width and flow (e.g. River Glaven, Lettteringsett).

An example of stream restoration, of a similar scale to the River Glaven (the study site in this thesis), is reported by Baattrup-Pedersen *et al*. (2000). A 1.4 km section of stream channel in the headwaters of the River Gudena, Denmark was excavated with the aim of re-establishing hydrological contact between the stream and its valley. This included the addition of meanders, and spawning gravel to the stream bottom, shallower banks and the raising of the streambed. The preliminary findings after 2-years indicated that groundwater levels increased by up to 30 cm after the restoration, and that plant communities on the banks changed from a dominance of non-riparian species to more diverse communities containing greater numbers of water-tolerant riparian grasses. Similarly, a study of five lowland and 13 mountainous 200 m section river restoration sites (also of a similar scale to the River Glaven) in western Germany reported increased riparian habitat diversity following the removal of river embankments. Increases in habitat diversity were attributed to an increase in mesohabitats generated by the creation of sand and gravel bars, islands and floodplain ponds, and diversified microhabitat conditions as a result of more frequent inundations (Januschke *et al*. 2011).

Restoration successes have been reported on large river systems. For example, removal of river embankments and widening of river channels on two sections of the Moesa River and one section of the Emme River in Switzerland designed to restore hydrogeomorphic processes was assessed by Rohde *et al*. (2004). These restoration measures led to the restoration of near-natural physical habitat conditions, in comparison with undisturbed reference sites. This improved patch richness and the number of riparian habitats. These changes mainly promoted early successional plant species, however, due to the limited size of the river widenings. Rohde *et al*. (2004) conclude that the success of such restoration projects could be increased by extending the length and width of the widenings. In another example, an ambitious project to restore inundation regimes altered by channelization along 70 km of the Kissimmee River in Florida, USA, is expected to improve dissolved oxygen conditions, result in the establishment of wetland vegetation, and improve habitat for rheophilic (fast water-loving) benthic invertebrate, fish populations, and water birds (Dahm *et al*. 1995). Initial results suggested that signature marshland plant communities, *Sagittaria lancifolia (*bulltongue arrowhead) and *Pontederia cordata (*pickerel weed), have responded positively to increased water table levels. However in this case the elimination of an exotic species, *Lygodium microphyllum (*climbing maidenhair fern), with a herbicide treatment was required to aid the hydrological restoration trajectory of the wetland community (Toth 2009; 2010).

While there has been an exponential increase in the implementation of river restoration projects over the past few years (Bernhardt *et al*. 2005), largely due to an increased appreciation of the socioeconomic environmental services that natural floodplains and other wetlands provide (e.g. flood defence, improvements to water quality, sediment management) (Acreman *et al*. 2007), river restoration ecology is still in its infancy and hence information on the effectiveness of different river restoration methods are still being gathered. Many projects have been implemented without clear aims, and sufficient monitoring before and after the restoration works to adequately determine their success or failure (Bernhardt *et al*. 2005). Hence, a number of nation-wide databases of river restoration schemes have been established to make information about their design and success more readily available. For instance, The River Restoration Centre provides information about more than 2,640 river restoration schemes that have been initiated in the UK, with the aim of informing future restoration projects (RRC 2016b).

Floodplain restoration strategies are often based on the maintenance and recreation of wet grassland areas (e.g. via river-floodplain reconnection) related to specific soil moisture requirements of target floodplain species (e.g Duranel *et al*. 2007; Hammersmark *et al*. 2010). Niche and habitat-suitability models of plant sensitivity to the soil moisture regime, such as the cumulative aeration stress index used by Gowing *et al.* (1997) (discussed in Section 2.4.1), can be used to guide adaptive management of wetland vegetation, especially when the models are linked with predictive tools such as physically-based hydrological models (e.g. Thompson *et al*. 2009; Hammersmark *et al.* 2010; Booth and Loheide 2012) (discussed in Section 2.6). The ability to predict changes in soil moisture conditions and associated impacts on the plant community composition is promising in river restoration schemes that comprise floodplain reconnection. Here, the length of time required to observe post-restoration effects on plant community composition can be substantial. For instance, Woodcock *et al.* (2006) report that re-creation of target (NVC MG4) plant communities on a floodplain meadow was >18 years due to limited seed-dispersal. Often, there is also insufficient time within monitoring programmes to capture the full range of hydrological conditions (i.e. wet versus dry years) that can drive community composition changes. Hydrological/hydraulic models therefore provide opportunities to investigate the expected long-term hydroecological effects of river restoration on plant communities through the simulation of extended periods that include this range of conditions, an approach that is used in this thesis, and addressed in Chapters 8 and 9.

2.6 Hydrological modelling

2.6.1 Model classification and representation of hydrological processes

The maintenance and management of wetlands and their influence on biological communities and processes is primarily dependent on the hydrological regime (Silvertown *et al*. 1999; Robertson *et al*. 2001; Bonn *et al*. 2002; Baker and Vervier 2004; Mitsch and Gosselink 2007). A thorough understanding of hydrological processes in wetlands, and the ability to predict changes in hydrological conditions associated with human modifications or climate is important for improving river health and services, as stipulated by the Water Framework (Directive 2000/60/EC) and Floods Directives (Directive 2007/60/EC).

Hydrological and hydraulic models are employed to simulate subsurface and surface movements of water. These models can be used to quantify key components of the hydrological regimes of wetlands, such as water table elevation; frequency of overbank flows; depth and extent of surface flooding; and flood-peak attenuation (e.g. Gilvear *et al*. 1993; Refsgaard *et al*. 1998; Bradley 2002; Thompson 2004; Thompson *et al*. 2009; Frei *et al*. 2010; House *et al*. 2016a). Models that adequately capture wetland hydrological processes are therefore valuable tools for the management of water resources and the assessment of potential ecological impacts of anthropogenic modifications (e.g. Zhang and Mitsch 2005; Woldeamlak 2007; Thompson 2012).

There are a range of ways in which mathematical models used to represent hydrological systems can be classified. One approach reflects a model's representation of the inherent uncertainty within hydrological systems with two categories being identified: 1) deterministic (mechanistic) and, 2) stochastic (probabilistic) (Figure 2.20) (Shaw 2011). Deterministic models are process-based, and simulate various fluxes in the hydrological system (e.g. groundwater and channel flow) using uniquely defined physical properties. This approach is non-random and thus deterministic models perform in a reproducible manner for a given set of input parameters. Examples of deterministic models include HBV (Bergström 1976, Bergström *et al*. 1992), TOPMODEL (Beven and Kirkby 1979), and MIKE SHE (Refsgaard and Storm 1995). Stochastic models, conversely, simulate hydrological processes using a probabilistic approach. This allows for random variation in one or more of the input values, which can account for uncertainty that is inherent to measurements of earth systems. Hence despite having set initial conditions, stochastic models produce many possible solutions that are influenced by frequency-magnitude distribution of the variable inputs.

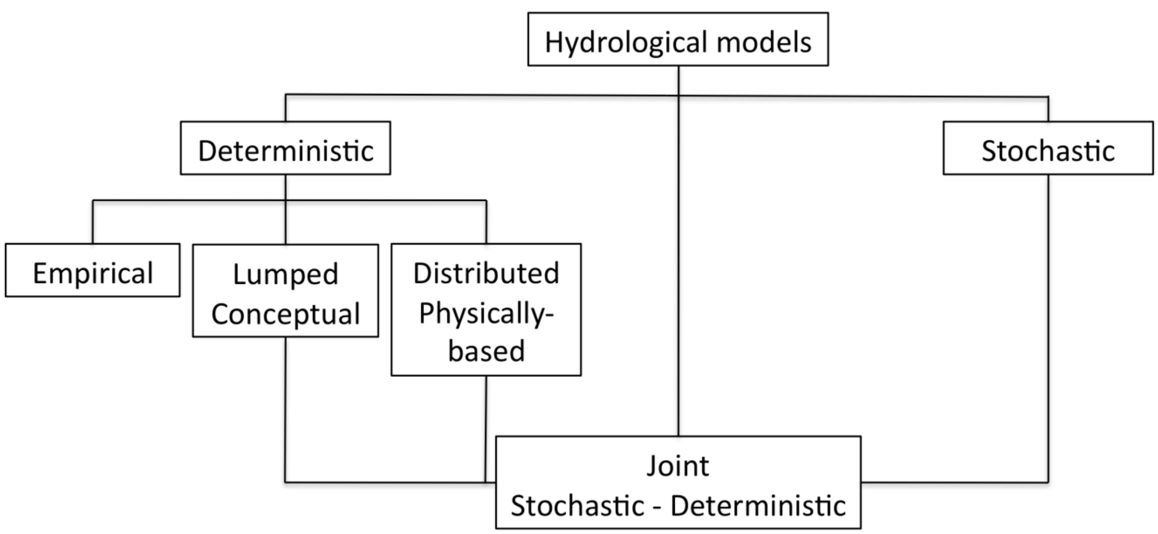


Figure 2.20: Model classification modified from Abbott and Refsgaard (1996).

Deterministic models can be classified further into the following three categories: empirical, conceptual, and physically based (Figure 2.20). Empirical models use mathematical equations from analyses of input and output time series rather than from the physical processes in the catchment. A well-known example of an empirical hydrological model is the Unit Hydrograph Model (Sherman 1932), which relates rainfall excess to direct runoff assuming a linear relationship between the two for a given catchment (Kokkonen and Jakeman 2001). Empirical models are data-dependent; hence they are specific to a given geographical region or time-period and cannot be used accurately outside the observed conditions. Some more process-based models may also employ empirical approaches. For example, Thompson *et al*. (2015) used an empirically established relationship between river inflows and outflows to the Inner Niger Delta, a major floodplain wetland in West Africa, within a conceptual catchment model used to simulate river flow and floodplain inundation. This enabled the simulation of flood extent under a range of climate change scenarios. However it is important to keep in mind that if conditions change, which is possible given the dynamic nature of many environments especially when simulation periods extend over a decade, these relationships will change and potential errors will be introduced into model results.

In a conceptual hydrological model, physical processes are mostly represented schematically, such as a series of interconnected stores and with the use of simple flow equations (Davie 2008; Shaw 2011). For example, the storage of water in soil may be estimated using a simple ‘bucket’ model, where predicted soil water is analogous to that determined for the rise (when precipitation > evapotranspiration) and drop (evapotranspiration > precipitation) of water in a bucket (e.g. Robock *et al*. 1995; Thompson and Hollis 1995; Gasca-Tucker and Acreman 1999). Such models can be used to capture the key processes in a catchment, sub-catchment, or an individual wetland whilst remaining parsimonious. However, the parameters that drive these models are not directly measured and need to be calibrated for a given area or catchment, hence the conceptual model may not be transferable to other catchments that were not used in the calibration process (Thompson and Hollis 1995; Davie 2008). They are also subject to change if the catchment characteristics that control hydrological processes are modified (e.g. due land cover change) and given their lack of physical-basis such changes are not readily incorporated within revised parameter values (Thompson *et al*. 2015).

Physically-based models, such as those used in this thesis, are based on our understanding of the mechanisms and physics that control the hydrological fluxes within a catchment or part thereof. In these models, the transport processes within the hydrological cycle are represented by the governing differential equations (Feyen *et al.* 2000). For example, the Richards equation for unsaturated zone flow, the Saint Venant equations for surface flow, and the Darcy/Boussinesq equations for groundwater flow. The physical characteristics of a catchment/sub-catchment/wetland, such as the soil properties, topography, land cover, are naturally spatially variable, hence physically-based models often use a distributed approach (see further details below) (Abbott and Refsgaard 1996). Physically-based distributed models, such as MIKE-SHE (Refsgaard and Storm 1995), IHBM (Beven *et al*. 1987), THALES (Grayson *et al*. 1992), and MODFLOW (Harbaugh 2005) can be used to represent spatial variability within the model domain. Since model parameters, which can be spatially distributed, have physical meaning, distributed models have the advantage over conceptual models in which parameters are obtained through calibration and cannot be measured directly. They enable the simulation of changes in catchment configuration such as land use modifications.

Hydrological models can be further classified as lumped, semi-distributed or distributed. Lumped models are spatially homogeneous and treat the model domain as a single unit. They do not allow for spatial variation in physical input parameters or the hydrological outputs, and instead represent an average response across the model domain (Sutcliffe and Parks 1989; Thompson and Hollis 1995; Beven 2000). The advantage of using these simplified models is that fewer parameters or data need to be defined or calibrated. However, a clear and inherent drawback is the lack of information about spatial variability in hydrological conditions so that, for example, maps of flooding depth and water table levels throughout a wetland area cannot be produced. Consequently, the utility of these models to characterise spatial differences in wetland hydrology is limited.

Conversely, distributed models account for spatial variation (potentially both horizontally and vertically) in hydrological properties within a modelled area including topography, soil properties, land use as well as meteorological inputs used to drive the model (e.g. precipitation and evapotranspiration). Distributed hydrological models may organise the catchment into a large network of grid cells, each of which are characterized by a set of variables, and parameter equations that are solved independently for each grid cell (Refsgaard 1997, Beven 2000). These models are therefore data intensive and are dependent on good quality input data, and computational power (Feyen *et al*. 2000). Semi-distributed models combine the principles of both lumped and distributed models and discretize the catchment into homogeneous zones based on factors that may include topography, drainage area, common combinations of soil and land use (e.g. the Hydrological Response Unit approach employed within SWAT (Arnold and Fohrer 2005).

The classic example of a fully distributed, physically-based (although conceptual, semi-distributed approaches are available for some processes – e.g. Thompson *et al*. 2013) model is MIKE SHE (DHI 2007a) (Figure 2.21), developed from the Système Hydrologique Européen (SHE) (Abbott *et al*. 1986). As a distributed model, MIKE SHE allows for spatial variations in the physical environment (e.g. topography, geology, soil properties) through discretisation of the study area into a regular spatial grid. For each grid cell, the finite difference approach is used to solve the differential equations that describe (1) saturated flow (three-dimensional Boussinesq equation), (2) unsaturated flow (one-dimensional Richards' equation/2 Layer UZ), and (3) overland flow (two-dimensional Saint Venant equation). Channel flow (one-dimensional Saint Venant equation) is simulated using the one-dimensional hydraulic modelling system, MIKE 11 (Havnø *et al*. 1995).

For three-dimenstional saturated flow, the general equation is:

$$\frac{\partial}{\partial_x}\left(K_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial_y}\left(K_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial_z}\left(K_z\frac{\partial h}{\partial z}\right)=-S_S\frac{\partial h}{\partial t} \tag{2.2}$$

where K_x, K_y, K_z describe the hydraulic conductivity along the x, y and z axes where material properties are different (i.e. anisotropic), *h* is the hydraulic head, and *Ss* is the specific storage coefficient (DHI 2007b; Fitts 2013).

The unsaturated flow equation is similar to the saturated flow equation. However, in unsaturated flow, changes in water content occur. Thus the one-dimensional Richards' equation can be written as:

$$C\frac{\partial \Psi}{\partial t}=\frac{\partial}{\partial z}\left(K(\theta)\frac{\partial \Psi}{\partial z}\right)+\frac{\partial K(\theta)}{\partial z}-S \tag{2.3}$$

Where C is soil water capacity (which is the slope on the soil moisture retention curve), ψ is the pressure head, *t* is the time, *K(*θ*)* is the unsaturated hydraulic conductivity*, z* is the vertical coordinate (positive upwards), and S is the root extraction sink (DHI 2007b; Fitts 2013).

The Saint Venant equations include the continuity equation and the dynamic/ momentum equation, and when solved yield a fully dynamic description of shallow, two-dimensional free surface flow (DHI, 2007b). The conversion of mass gives:

$$\frac{\partial h}{\partial t}+\frac{\partial}{\partial x}(uh)+\frac{\partial}{\partial y}(vh)-i \tag{2.4}$$

and the dynamic/ momentum equation gives:

$$s_{fx} + s_{Ox} - \frac{\partial h}{\partial x} - \frac{u \partial u}{g \partial x} - \frac{1 \partial u}{g \partial t} - \frac{qu}{gh} \tag{2.5}$$

$$s_{fy} + s_{Oy} - \frac{\partial h}{\partial y} - \frac{v \partial v}{g \partial y} - \frac{1 \partial v}{g \partial t} - \frac{qv}{gh} \tag{2.6}$$

Where *h (x, y)* is the flow depth (above the ground surface), g is acceleration due to gravity, *u* (*x, y*) and *v* (*x, y*) and the flow velocities in the x- and y-directions, respectively, *I* (*x, y*) is the net input into overland flow (net rainfall less infiltration), *Sf* is the friction slopes in the x- and y-directions, S_O is the slope of the ground surface. The dynamic solution of the two-dimenstional Saint Venant equations is numerically challenging, therefore the diffusive wave approximation implemented in MIKE SHE uses a simplification of the momentum equation (see derivation of the equations in DHI 20007b).

The one-dimensional Saint Venant equation used to simulate river flows and water levels is a simplification of the two-dimensional Saint Venant equation. The derivation of the equations of continuity and momentum, employed by MIKE 11, is given in DHI (2007c). The resulting equations are:

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = q \tag{2.7}$$

$$\frac{\partial Q}{\partial t} + \frac{\partial\left(\alpha \frac{Q^2}{A}\right)}{\partial x} + gA\frac{\partial h}{\partial x} + \frac{gQ|Q|}{C^2 AR} = 0 \qquad 2.8)$$

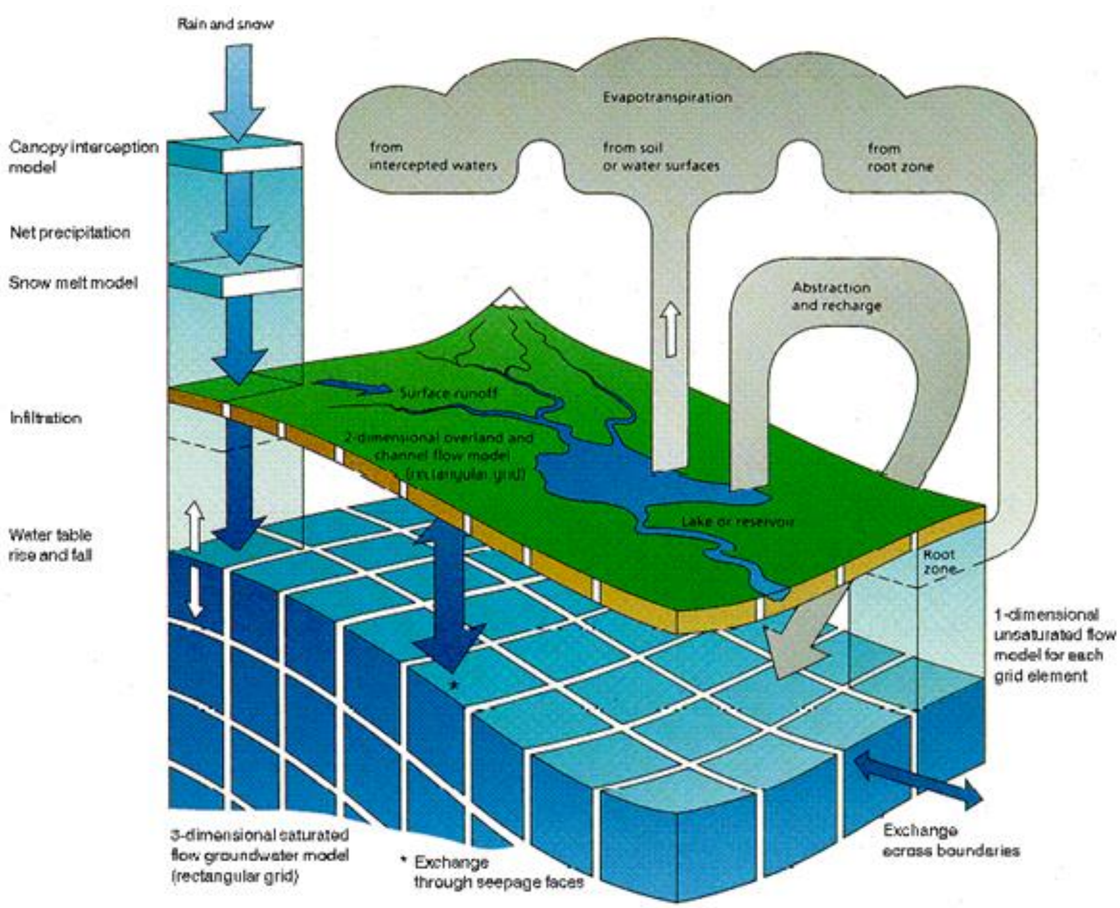


Figure 2.21: Schematic of MIKE SHE model (after Refsgaard and Knudsen, 1996, cited in Thompson *et al*. 2004).

Analytical solutions are used to describe evapotranspiration and interception (Thompson 2004). DHI (2007a) and Refsgaard and Storm (1995) provide further detail and description of the MIKE SHE equations and the parameterisation of the MIKE SHE and MIKE 11 models. Dynamic coupling of the MIKE 11 river model and the MIKE SHE hydrological model enables the simulation of river-aquifer exchange, inundation from the river onto the floodplain and the return of overland flow to the river (Thompson *et al.* 2004; DHI 2007a). This type of physical model has become the main approach for assessing the response relationship between ecosystem properties and hydroecological processes (Ma *et al*. 2016).

Generally, physically-based models are data-intensive; therefore when selecting appropriate hydrological models it is important to consider the complexity of the modelling system in relation to the availability of input data needed to parameterise and drive these models (Shiklomanov *et al*. 2002; Rawlins *et al*. 2003). Even where field data are plentiful, a number of unknown variables are likely to exist with regard to parameterisation of the models, necessitating calibration and careful model validation in which model results are compared against adequate field data is essential (Feyen *et al*. 2000).

The River Glaven, studied in this thesis (see Chapter 3), is a dynamic hydrological system characterised by interaction of unsaturated and saturated zones, and surface and subsurface flows between the river and groundwater. In this thesis many different components, and physical properties of the hydrological system (i.e. precipitation, potential evapotranspiration, stream flow, groundwater elevation, topography, hydraulic conductivity, soil texture) have been surveyed, measured and monitored. In addition, a floodplain restoration project has modified the physical properties of the site that control the interactions between hydrological components. MIKE SHE (DHI 2007a) was therefore selected as a suitable model with the complexity and computational capability to represent surface-subsurface exchange at small spatial scales, and the flexibility to simulate hydrologic processes using a combination of distributed and semi-distributed methods, in line with process understanding at the site. In addition, the applicability of MIKE SHE to simulate surface-subsurface hydrological exchange and quantify the impacts of embankment removal on river-floodplain hydrology is documented in the literature, and as the following section demonstrates, this model system has been successfully employed in the simulation of wetlands with similar characteristics to the site which is central to this thesis.

2.6.2 Surface water-groundwater modelling and river restoration

Hydrological and hydraulic models are employed to simulate subsurface and surface movements of water. These models can be used to quantify the hydrological effects of river restoration efforts, such as water table manipulations, embankment removal, reintroduction of meanders, and river widening (Thompson 2004; Hammersmark *et al*. 2008). The effects of river restoration on hydraulic and hydrological processes are complex, and are often difficult to determine if there is insufficient monitoring (Kondolf 1995; Darby and Sear 2008). Direct comparisons of measured pre- and post-restoration hydrological conditions, while informative, are often not possible (e.g. Hammersmark *et al*. 2008). Even in cases where hydrological monitoring was initiated before a reach was restored, there is often insufficient time to capture the range of hydrological conditions that occur in response to climate extremes (i.e. wet versus dry years). Understanding the long-term impacts of river restoration is important for predicting changes in wetland function and subsequent response patterns of the floodplain biota. As a substitute for long-term data, hydrological/hydraulic modelling is increasingly used to better understand the effects of river restoration activities under a variety of possible hydrometeorological conditions, such as extreme rainfall and river-flow events, and thus form an important component of wetland restoration research (Thompson 2004; Thompson *et al*. 2009; Hammersmark *et al.* 2010).

Coupled wetland MIKE SHE/MIKE 11 hydrologic/hydraulic models were initially developed for floodplains on the Danube (Sørensen *et al*. 1996; Refsgaard and Sørensen 1997), and have been used to study a range of wetland and other hydrological settings at vastly different scales from major international river basins of hundreds or thousands of km^2 (Huang *et al*. 2010; Singh *et al*. 2010; Thompson *et al*. 2013), to small individual wetlands (<10 km^2) (Thompson 2004, Hammersmark *et al*. 2008). For instance, a coupled hydrological/hydraulic model of the Elmley Marshes, North Kent, UK, was developed by Thompson *et al*. (2004) using the MIKE SHE/MIKE 11 system. The model domain comprised of 9,271 grid squares of 30 m × 30 m and covered an area of 870 ha of lowland wet grassland. This study provides a detailed account of the modelling system and demonstrates the successful application of coupled MIKE SHE/MIKE 11 models to simulate seasonal water table and ditch water level fluctuations (e.g. Figure 2.22) as well as shallow flooding in a wetland. The Elmley Marshes model also shows the utility of MIKE SHE/MIKE 11 for the simulation of water-level management scenarios, specifically the impact of changing the height of drop-board sluices upon groundwater and ditch water levels as well as flood extent (Thompson 2004, Figure 2.23).

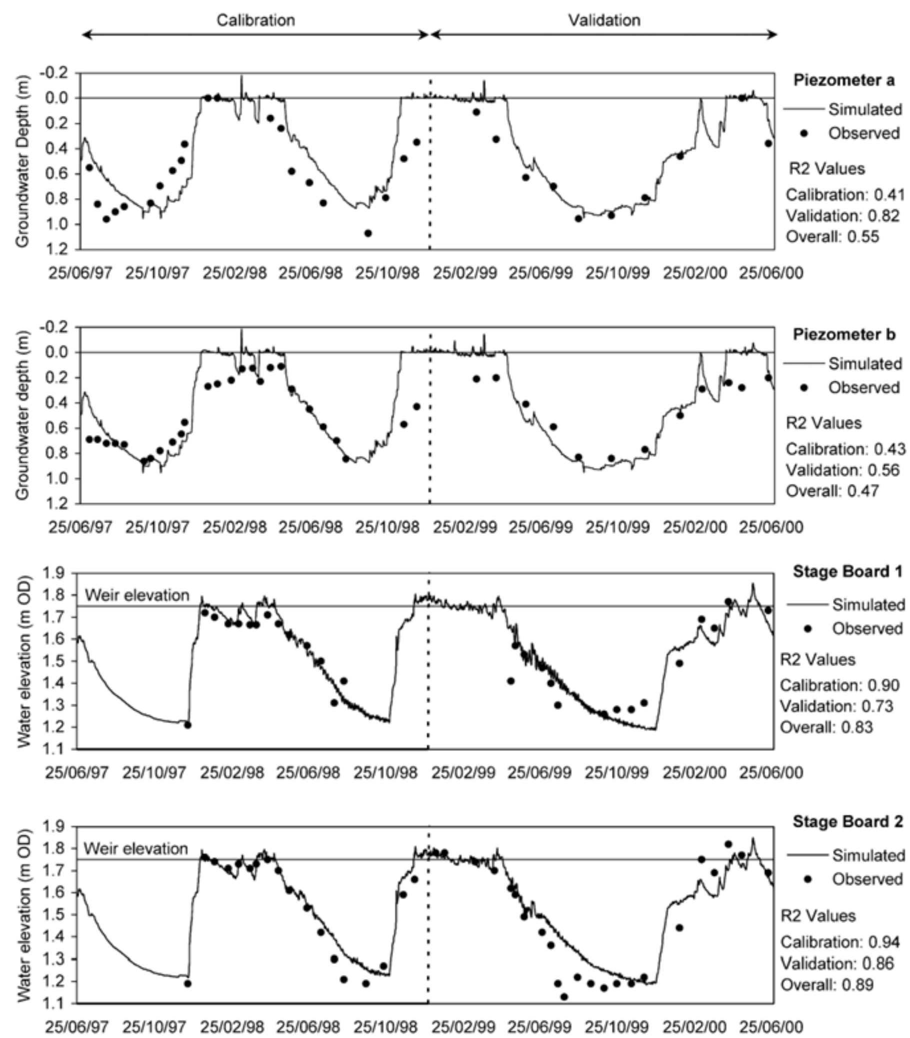


Figure 2.22: Simulated groundwater depth (top two panels) and ditch water level (bottom two pannels) at four locations within the Elmley Marshes (amended from Thompson *et al.* 2004).

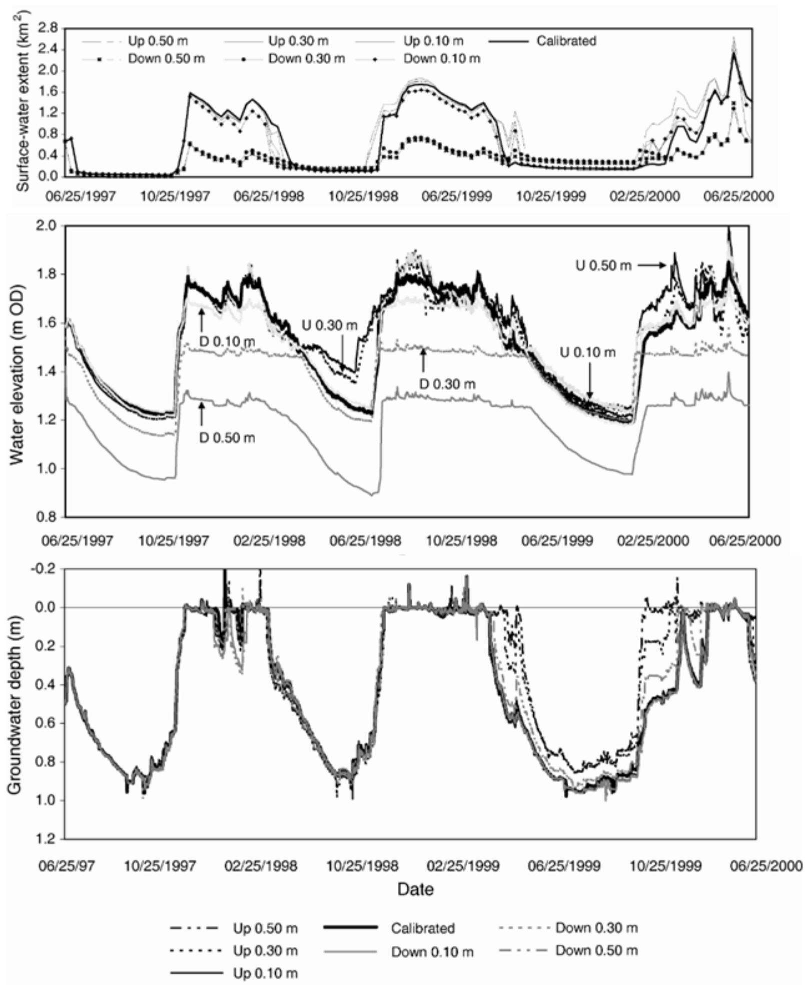


Figure 2.23: Impact of changing weir height upon simulated surface flooding (top panel), ditch levels (middle panel), and groundwater depth (bottom panel) within the Elmley Marshes. Amended from Thompson (2004).

The study conducted by Hammersmark *et al*. (2008) at Bear Creek, Northern California is of particular relevance to this thesis because MIKE SHE/MIKE 11 was used to develop coupled hydrological/hydraulic models of an incised and restored river and its floodplain. The models were comprised of 2,898 30 m × 30 m grid squares, representing a total area of 261 ha. Due to limited pre-restoration data, the restored channel was used during the calibration and validation processes. Channel and floodplain topography were then altered to represent pre-restoration geomorphological conditions. Good agreement was obtained between observed groundwater depths and surface water elevations from the calibrated post-restoration MIKE SHE/MIKE11 model (Figure 2.24).

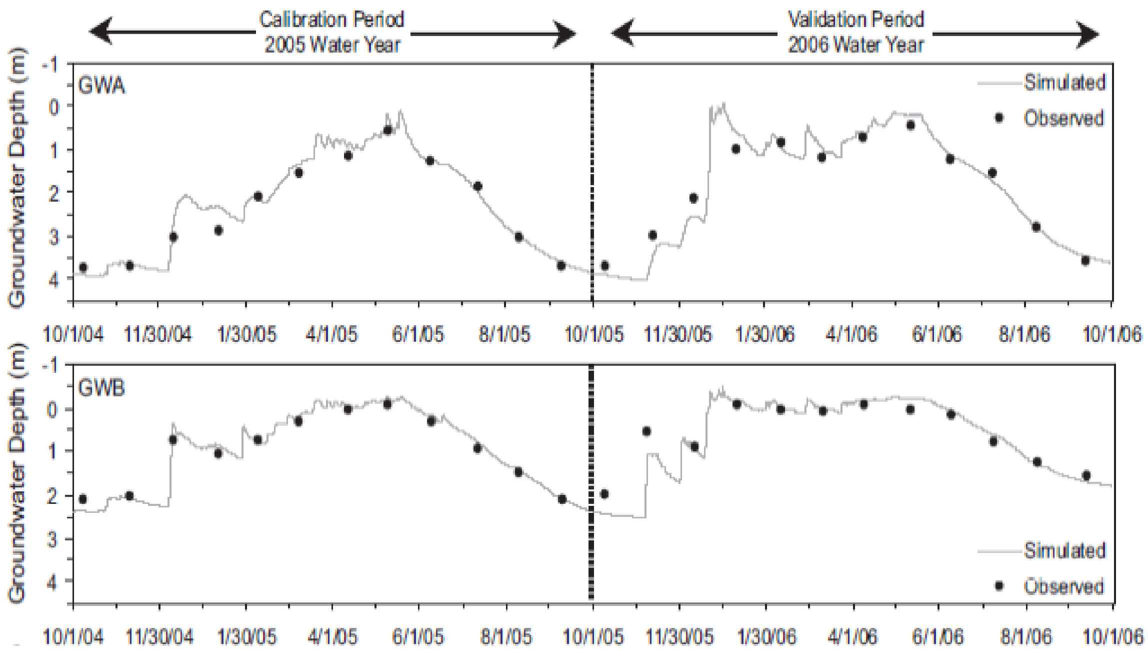


Figure 2.24: Comparison of simulated and observed groundwater depth at two piezometer locations within a wet meadow. The 2005 water year (left side) was used for model calibration and the 2006 water year (right side) was used for model validation. From Hammersmark *et al*. (2008).

Hydrological/hydraulic modelling of Bear Creek provided a quantitative assessment of the effects of river restoration. Specifically, the study highlighted three important hydrological consequences of embankment removal: (1) increased groundwater levels and subsurface storage; (2) increased frequency of floodplain inundation and a reduction in the magnitude of flood peaks; and (3) decreased baseflow and annual runoff, and demonstrated the use of MIKE SHE/MIKE 11 for the direct assessment of the effects of embankment removal on wetland hydrology (e.g. Figure 2.25).

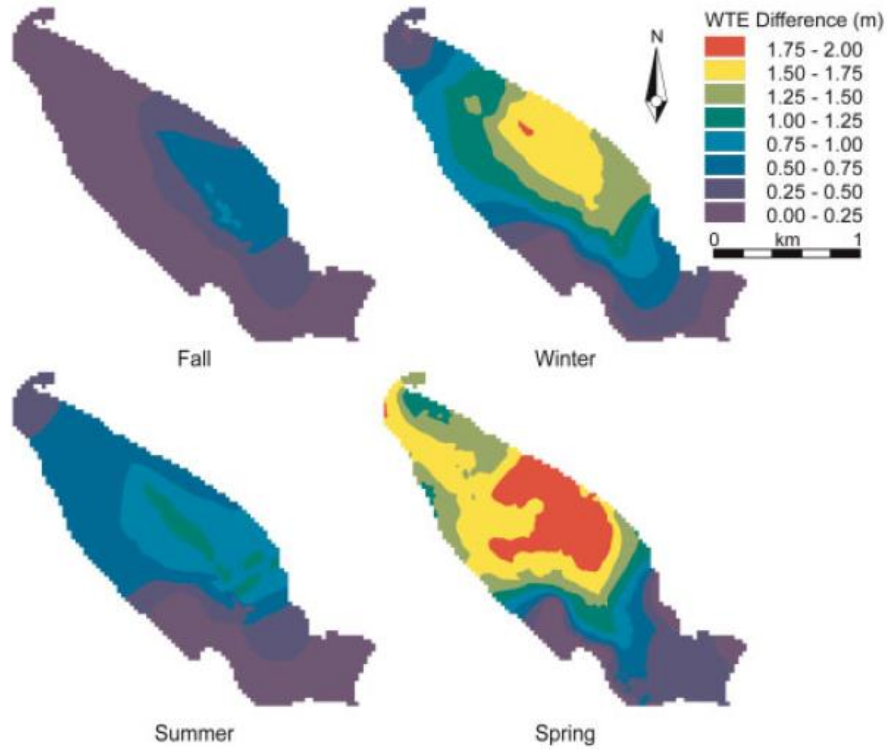


Figure 2.25: Seasonal water table elevation (WTE) differences between the incised and restored simulations. Positive difference indicates the restored water table is higher than the incised water table. From Hammersmark *et al*. (2008).

Coupled groundwater/surface water models of a lowland chalk wetland along the River Lambourn, similar to the wetland studied in this thesis, were successfully developed by House *et al*. (2016a) using the MIKE SHE/MIKE 11 system. A 10 ha model area was discretised using a 1 m × 1 m grid, comprising 101,689 grid cells within the model domain. Surface water and groundwater interactions and wetland functioning were assessed (Figure 2.26) to inform management practices at the site. The study demonstrates the complex and significant role of groundwater-surface water interactions in wetlands, and highlights the importance of fully considering both contributions to flooding. The study also identified the acute impact of seasonal in-stream vegetation growth on channel bed roughness and flow resistance, which in turn affects river stage and flow rates, and should be considered during model development and calibration, particularly in small streams (Figure 2.26).

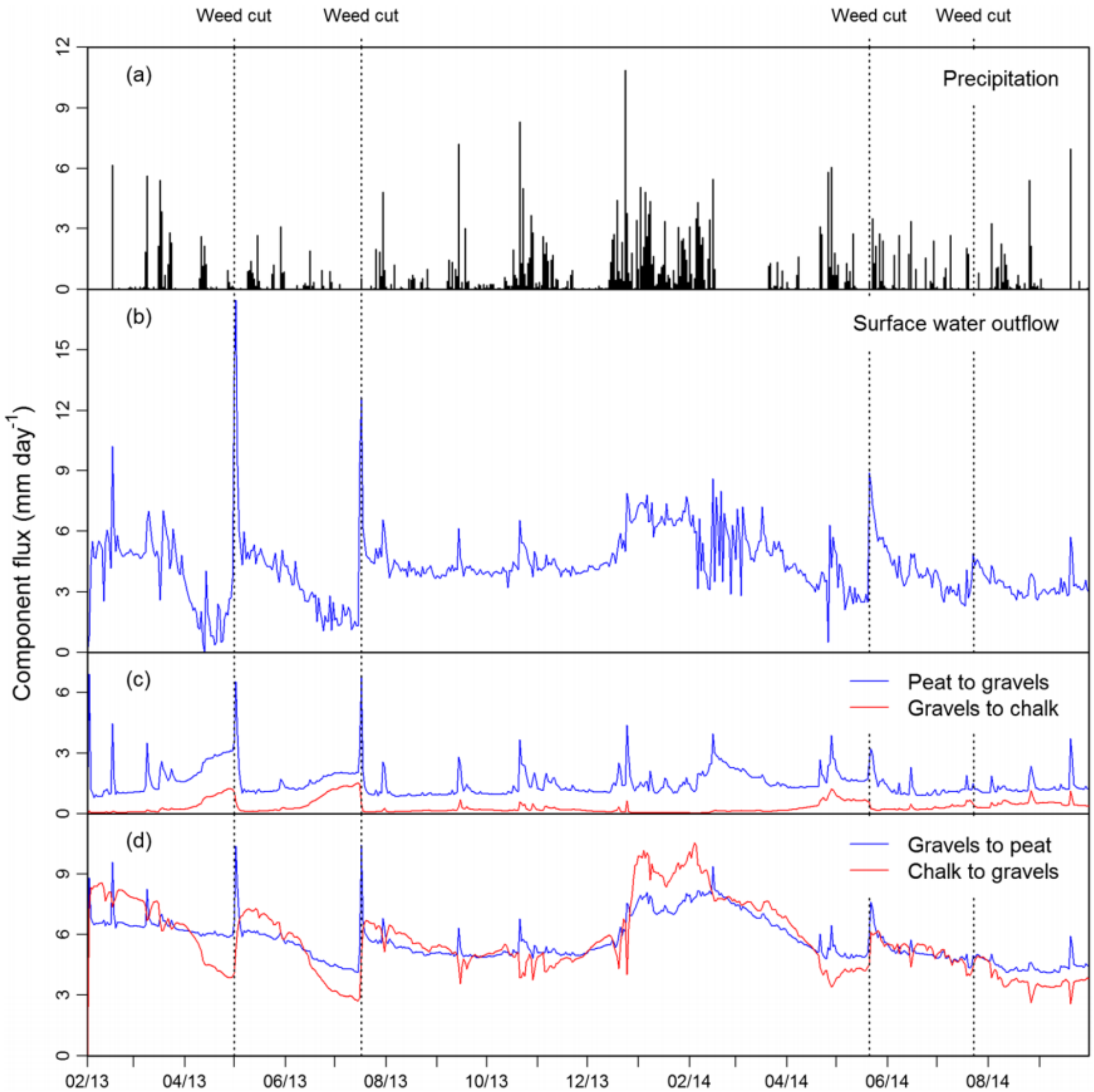


Figure 2.26: (a) Precipitation, (b) surface water outflow, (c) downward groundwater flow between geological layers and (d) upward groundwater flow between geological layers. From House *et al*. (2016a).

2.7 Conclusions

The existing literature is in general agreement that fluctuations in water levels (flood pulses) are an integral component of natural floodplains, necessary for the maintenance of biodiversity, and ecosystem functioning. Whilst hydrology, in terms of water table depth or inundation frequency/extent, is the primary factor that determines plant community composition, other factors such as nutrient availability, and past and current management regimes can interact to affect different aspects of plant community composition, and thus should not be viewed in isolation. As a consequence, there is a need for integrated, process-driven wetland restoration research. Literature on the ecological response of river-flow alteration is still quite fragmented, in part due to the length of time required to study post-restoration effects. Hence, there are opportunities to investigate the long-term ecohydrological effects of river restoration using coupled hydrological/hydraulic models. These issues are evaluated using the river restoration project undertaken on the River Glaven as an example of an integrated, ecohydrological assessment of the impacts of embankment removal on river-floodplain functioning.

Chapter 3: The River Glaven and Hunworth Meadow

3.1 Introduction

Hunworth Meadow, a floodplain site on the River Glaven in Norfolk, UK, was selected as the study site due to the rare opportunity to monitor a river restoration project before and after the works were conducted. A complete data set was essential for assessing the measured hydrological response of embankment removal, and for calibration and validation of the hydraulic-hydrological models. This study also provides an opportunity to investigate the implications of river embankment removal on river–floodplain hydrological connectivity in a chalk setting. The River Glaven is a low altitude, calcareous river that has been impacted by farming and land management practices. However it still supports many important habitats of conservation value along its length and thus contains a reservoir of species that can benefit from the restoration of more natural hydrological conditions. Restoration studies of chalk rivers are also limited, so Hunworth Meadow provides an opportunity to study the effects of river restoration in a chalk setting. This chapter details the physical, chemical and biological characteristics of the River Glaven and Hunworth Meadow. A description of the river restoration work conducted at Hunworth Meadow is also given.

3.2 Location and climatology

The study was conducted at Hunworth Meadow on the River Glaven, in North Norfolk, approximately 34 km north-west of Norwich (52° 52' 55''.53 N, 01° 03' 55''.45 E; elevation ca. 21 m above Ordnance Datum Newlyn) (Figure 3.1). The climate of the study region is temperate, with average annual air temperature of 9.8 °C, ranging on average from 4.7 °C in January and 17.3 °C in July (East Anglia, UK 1985-2015 average; Met Office 2016) (Figure 3.2). Mean annual rainfall for the East Anglia region (for the period 1985 – 2015) is approximately

623 mm and is characterised by higher rainfall during the autumn and winter months (Figure 3.3). On average, the annual potential evapotranspiration, (evaluated using the Hargreaves-Samani method (Hargreaves and Samani 1985)) reaches 600 mm, and exceeds precipitation in the summer.

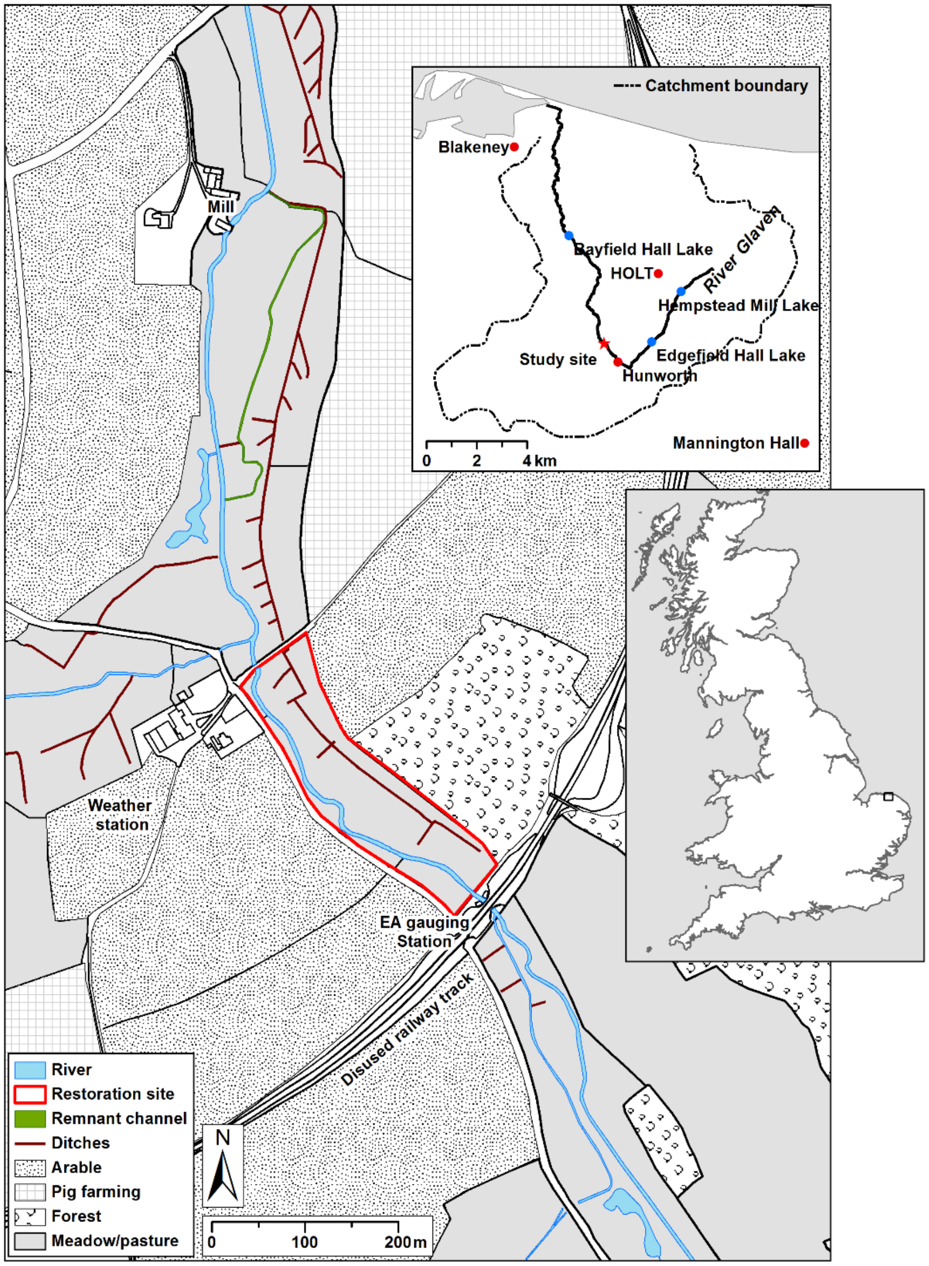


Figure 3.1: The River Glaven restoration site at Hunworth, North Norfolk. The woodland and arable border along the northeast of the meadow delineates the base of a hillslope. The River Glaven is shown inset, with the location of the study site at Hunworth. River flow direction is northward.

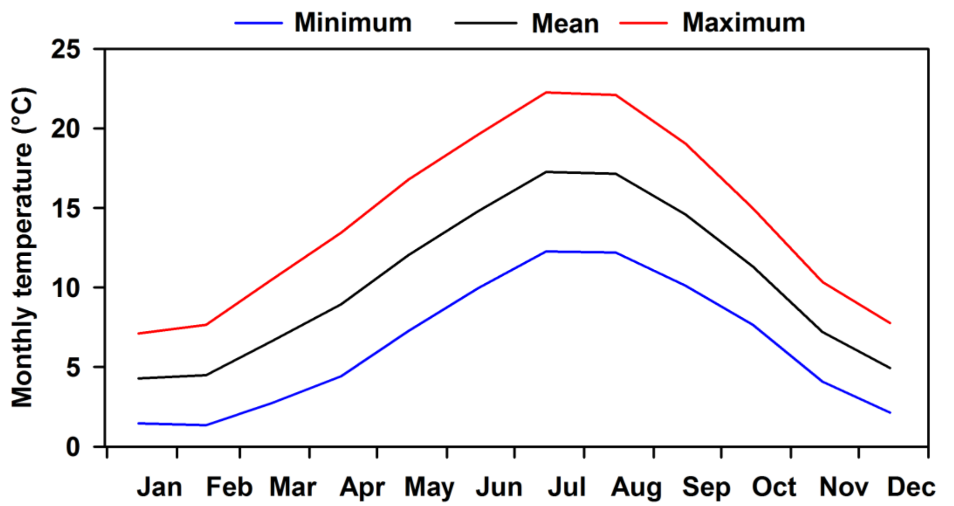


Figure 3.2: Monthly minimum, mean and maximum air temperature for East Anglia for the period 1985 – 2015.

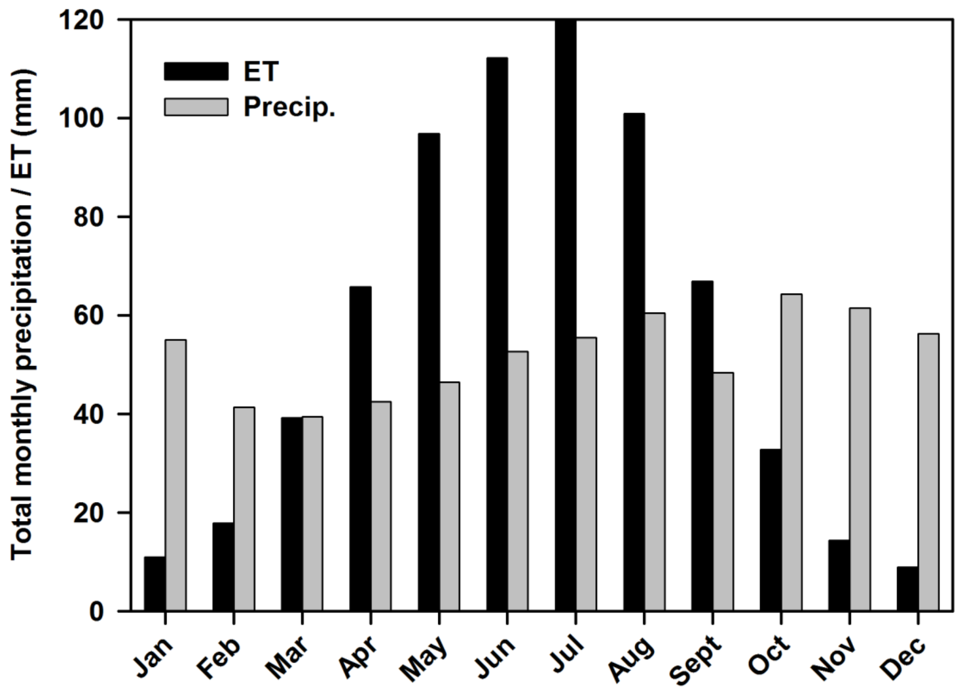


Figure 3.3: Mean total monthly precipitation and potential evapotranspiration (1985 – 2015) for East Anglia, UK. Climatology data are from UK Met Office regional climate summaries (Met Office 2016). Potential evapotranspiration was estimated using the Hargreaves-Samani method (Hargreaves and Samani 1985).

The River Glaven is a small (17 km in length), lowland, calcareous river, has a catchment area of 115 km^2 (Figure 3.1). The river flows southwest from headwaters in the Lower Bodham and Baconsthorpe area before taking an acute turn at Hunworth to continue northwards to Blakeney Point, where it discharges into the North Sea (Figures 3.1 and 3.4). Previous glaciation of this area has resulted in the formation of glacial hill features throughout the catchment in an otherwise flat landscape (Moorlock *et al*. 2002). The elevation change from the headwaters to the mouth of the River Glaven is 50 m, which equates to a valley slope of 2.9% (Figure 3.4). The river has two main tributaries, the Stody Beck which meets the Glaven approximately 500 m upstream of the study site and the Thornage Beck which joins immediately downstream of the study site (Figure 3.1).

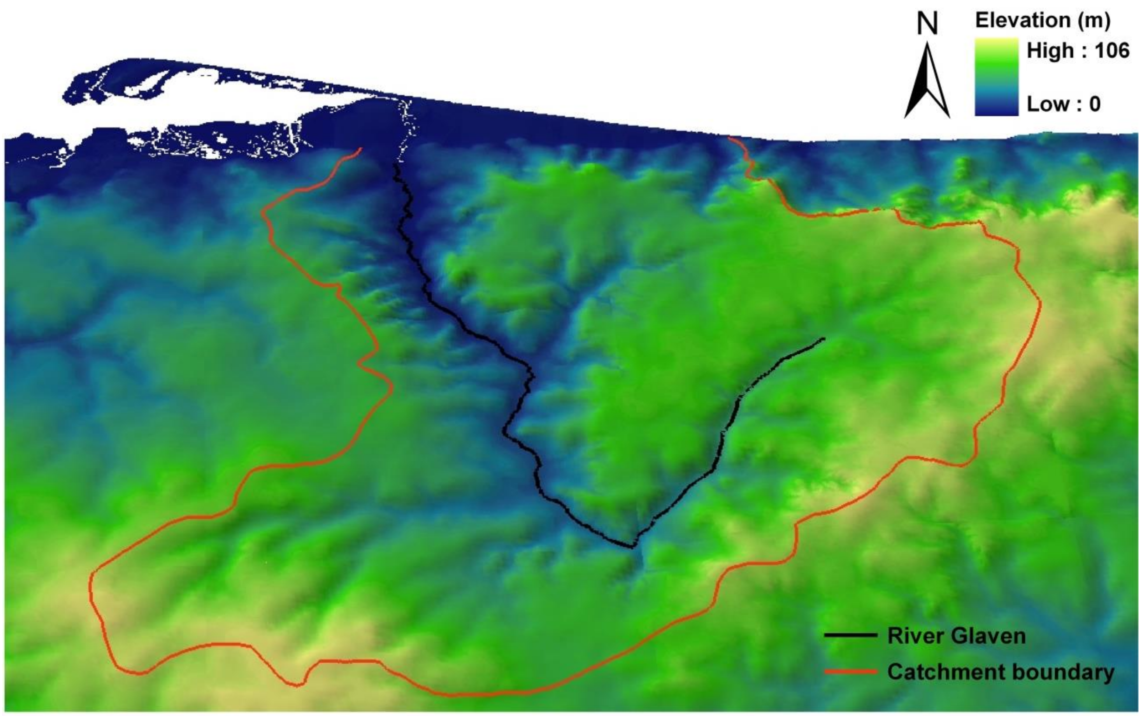


Figure 3.4: Topography of the River Glaven catchment, based on Ordnance Survey 1:50k elevation data. The three-dimensional DEM was produced in ArcScene, and was exaggerated 5×.

The River Glaven rises on Cromer Ridge (see Section 3.3) and flows northward to the North Sea (Figure 3.4), with the groundwater divide generally following the surface water divide (Figure 3.5). Similarly, other North Norfolk rivers that drain north include the Burn, and the Stiffkey (Figure 3.6). A major groundwater divide separates the east and west Norfolk regions and follows the surface water divide which trends south-west for the Nar and the Wissey, and generally east towards Great Yarmouth for the Bure, the Yare, and the Wensum (Figures 3.5 and 3.6). Chalk bedrock, which underlies the whole area, has a general easterly dip of less than 1° (Figure 3.5). The Upper Chalk forms the main aquifer unit. Aquifer tramissivity values vary from 1 $m^2 d^{-1}$ to >10 000 $m^2 d^{-1}$. A strong pattern of the high transmissivity occurs in river valleys, with the lower values typically found in interfluves, for instance high transmissivity is found at Glandford in the Glaven valley (Allen *et al*. 1997). The physical hydrogeology of the Chalk aquifer in North Norfolk is given in detail by Allen *et al*. (1997).

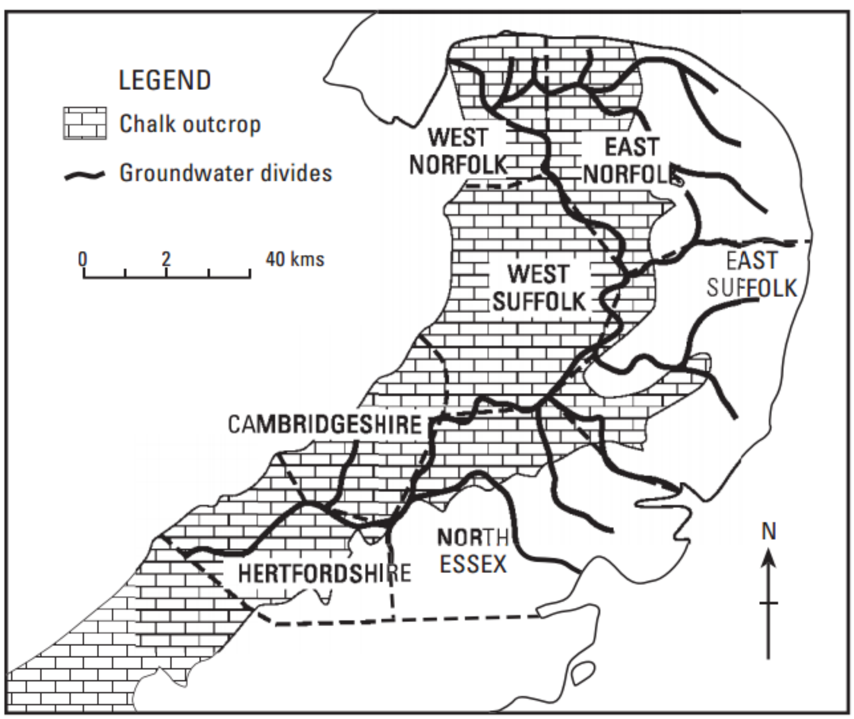


Figure 3.5: Chalk outcrops and groundwater divides for regions around South-East England (from Allen *et al*. 1997)

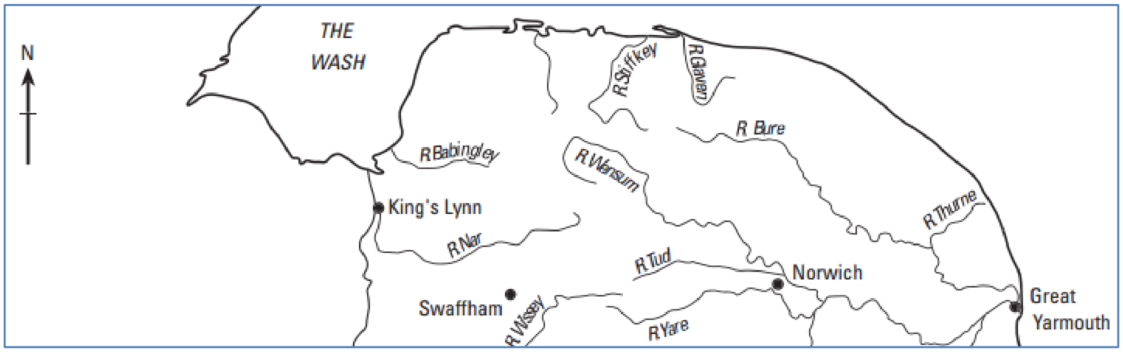


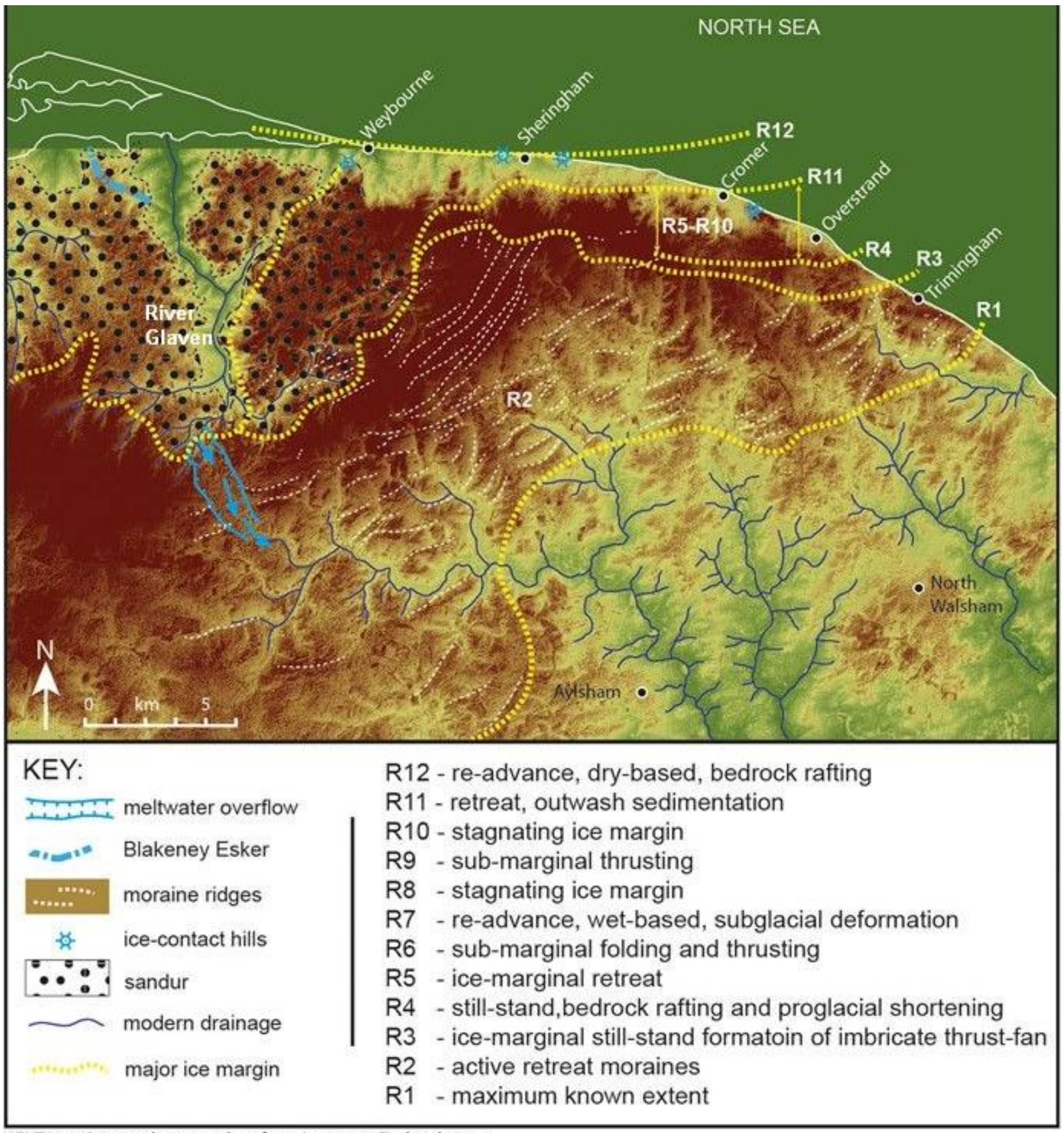
Figure 3.6: Major rivers in North Norfolk (amended from Allen *et al*.1997).

3.3 Geology of the River Glaven catchment

North Norfolk exhibits a low-lying relief associated with Quaternary sediments overlying Cretaceous chalk bedrock (BGS 2016). The region has been affected by glaciation on a number of occasions during the Mid - Late Pleistocene, with the most extensive episode during the Anglian Glaciation, approximately 480,000 years ago (e.g. Lee *et al*. 2012). While the most recent glaciation, approximately 35,000 – 16,000 years ago only reached the northern-most Norfolk Coast (e.g. Carr *et al*. 2006), this contributed to the formation of the Wash (the estuary along the northwest margin of East Anglia) and the establishment of periglacial conditions impacting the landscape of North Norfolk (Bateman *et al*. 2014).

The shallow geology of North Norfolk is strongly influenced by the dynamic subglacial and ice marginal processes that acted during the Anglian glaciation (Lee *et al*. 2016). This was due to the oscillation (retreat and advance) of the ice-margin that resulted in a complex sedimentary stratigraphy and associated glacial landforms. These glacial deposits, and structures composed of sand, gravel, and clasts of chalk, form a broad moraine complex known as the Cromer Ridge (Figure 3.7), which runs from Syderstone to Cromer, accounts for all topography in North Norfolk (Moorlock *et al*. 2002; Allen *et al*. 1997; Lee *et al*.

2016), and controls the direction of river flow in the upper reaches of the River Glaven (i.e. due to moraine ridge orientation of approximately SW-NE; Figure 3.7).



NEXTMap Britain elevation data from Intermap Technologies.

Figure 3.7: The glacial structures of the Cromer Ridge (in brown) associated with ice marginal retreat. A summary of the processes operating during each ice margin retreat stage is indicated in the table key (amended from Lee *et al*. 2016).

The River Glaven's catchment is characterised by Upper Cretaceous chalk bedrock that is overlain by chalk-rich stony, sandy, silty, boulder clay (Lowestoft Formation) up to 40 m thick, and estimated to be less than 20 m thick at the study site (British Geoglogical Survey borehole log Brinton Hall (TG 0378 3580) approximately 2.5 km west of the study site; note the borehole at Brinton Hall was drilled on the interfluve and thus the recorded till thickness of 20 m is likely to be greater than the thickness of the till in the valley at Hunworth Meadow) (Figure 3.8). The Lowestoft Formation outcrops extensively throughout the Glaven catchment but is overlain by Quaternary glaciogenic sand and gravel deposits (Briton's Lane Sand Gravel Member) at the study site. Hillwash (also known as Head), a poorly sorted mixture of clay, sand, silt and gravel, typically occurs as a veneer less than one metre thick on the valley slopes, and up to several metres thick at the base of steep slopes. Alluvium along the floodplains of the River Glaven is estimated to be a maximum of 2 m thick, and consists predominantly of unconsolidated layers of sand and silt, but also includes sediments that range from clay to coarse gravel (Moorlock *et al*. 2002) (Figure 3.8).

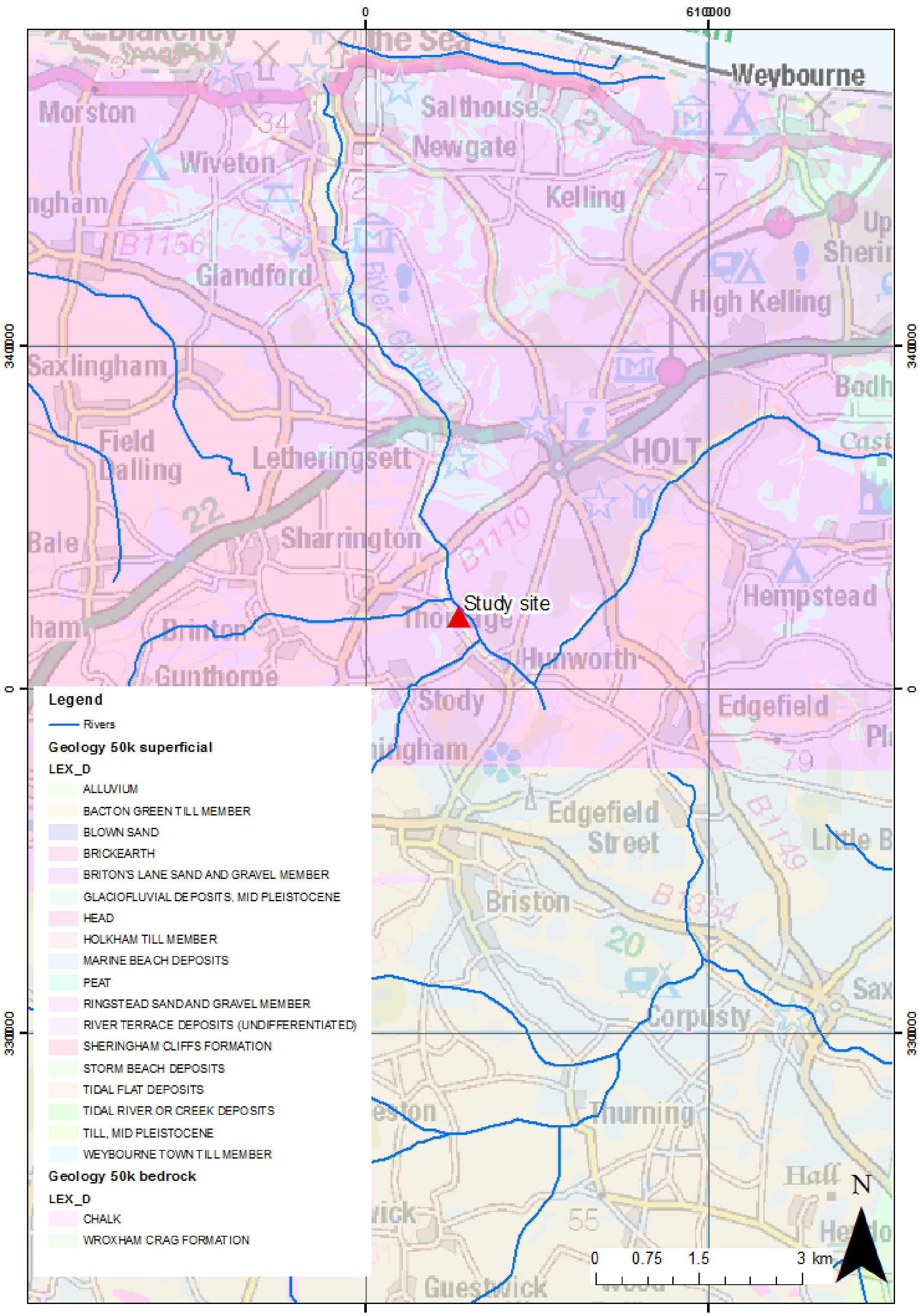


Figure 3.8: Superficial geology of the River Glaven catchment, and regional bedrock geology. British Geological Survey data.

For much of its length, the River Glaven is classified as a chalk stream (Candy *et al*. 2008) (Figure 3.8). Chalk is predominantly found in the south and east of England, which accounts for the major part of the chalk river resource of Europe (UK BAP 2011). There are approximately 35 chalk rivers and major tributaries in the UK that range from 20 to 90 km in length (UK BAP 2011). Chalk aquifers are a major groundwater resource in northwestern Europe, and are the principle aquifer in the UK (Allen *et al*. 2010). They form an important contribution to river flow, maintaining stable flows of clear cool water in chalk rivers, even during extended periods of low rainfall (Sear *et al*. 1999). Rain falling in a chalk catchment mainly infiltrates into the soil and underlying bedrock, with very little overland flow. Indeed, it is estimated that up to 90% of discharge of chalk streams can be from groundwater (Ladle and Westlake 2006; Sear *et al*. 1999). Consequently, chalk streams have a characteristic annual hydrograph due to the slow release of water from the porous aquifer, which attenuates rainfall events, and provides a fairly stable hydrological regime. Typically, stream flow increases from December until March/April, when it begins to decline steadily until the next winter (Berrie 1992).

Chalk rivers typically flow through mixed geologies for some of their length, and although local hydraulic features are important, they often maintain some of the characteristics of a groundwater dominated chalk system, e.g. high base flow index, steady thermal regimes, dampened discharge fluctuations, stable substratum dominated by gravel, high water clarity, and relatively high alkalinity (Berrie 1992; Sear *et al.* 1999). For example, in some sections of the upper reaches, the River Glaven has characteristics of a fast flowing upland gravel-bed river, with good riffle-pool structure, whereas in parts of the lower reaches the river flows over a chalk bed. Impermeable clay layers (see Figure 3.8) can have some influence on flows in chalk rivers, resulting in more rapid response to rainfall in some parts of the catchment than would otherwise be expected for a groundwater dominated system (UK BAP 2011). In addition human acitivities (e.g. water extractions, sewage treatment dicharges, and land use can affect the groundwater dominance signature in chalk streams (Sear *et al*. 1999). The element chemistry of the Chalk aquifer in North Norfolk is presented in Chapter 5 and is given in detail by Ander *et al*. (2006).

3.4 Surface water quality

As with flow, the chemical (e.g. pH) and physical (e.g. water temperature) properties of surface water are relatively stable in un-impacted chalk streams (Berrie 1992; UK BAP 2011). The River Glaven is slightly alkaline, averaging between pH 7.7 – 8.0 (Table 3.1), which is indicative of a chalk river, mainly due to calcium carbonate (Berrie 1992). Long-term averages of river chemistry upstream of the study site indicate that nutrient concentrations have been fairly constant (Table 3.1). River Glaven nutrient concentrations are moderate, with nitrate concentrations ranging from 5.8 – 7.5 (mg N L^{-1}) along the river, approximately four-fold lower than the Environment Agency surface water threshold of 30 mg NO_3^--N L^{-1} (Environment Agency 2013). Phosphate concentrations are low and average less than 0.05 mg P L^{-1}; these concentrations are well below the Environment Agency threshold of 0.1 mg P L^{-1} (Environment Agency 2013) and meet the Water Framework Directive ‘High’ water quality standard of ≤ 0.05 mg L^{-1} (Table 3.2).

Dissolved oxygen concentrations are high, ranging from 95 – 98% saturation (10.7 – 10.8 mg L^{-1}) on average in the river. Slightly lower oxygen concentrations occur in Selbrigg Lake, averaging 89% saturation (10.1 mg L^{-1}), which is to be expected due to less turbulence and mixing in the lake. Treated sewage discharges occur into the River Glaven downstream of the study site near Letheringsett, and downstream of Glandford. Although measurements of river nutrients upstream of the water treatment works indicate that the water quality in these sections is good (Table 3.1), river nutrients are likely to be higher in the immediate vicinity of the water treatment works.

Table 3.1: Water chemistry (pH, nitrate, phosphate and dissolved oxygen) of the River Glaven measured by the Environment Agency. The Selbrigg Lake and Letheringsett Mill mean values are from samples collected from 2007 - 2010 (n=35). The Hempstead Mill and Edgefiled Bridge mean values are from samples collected from 1983 – 2007 (n=283).

Sample Location (upstream/ downstream of study site)	pH	NO_3^- (mg N L^{-1})	PO_4^{3-} (mg P L^{-1})	DO (% sat.)	DO (mg L^{-1})
Selbrigg Lake (upstream)	7.73 ± 0.10	ND	0.06 ± 0.01	89.34 ± 2.04	10.10 ± 0.33
Hempstead Mill (upstream)	7.93 ± 0.03	7.49 ± 0.31	0.034 ± 0.004	98.47 ± 1.87	10.83 ± 0.53
Edgefield Bridge (upstream)	8.03 ± 0.03	7.12 ± 0.24	0.032 ± 0.005	95.72 ± 1.17	10.73 ± 0.60
Letheringsett Mill (downstream)	7.77 ± 0.30	5.84 ± 0.32	0.036 ± 0.004	94.50 ± 2.40	10.69 ± 0.30

3.5 Flora, fauna, and conservation value of the River Glaven

In pristine chalk rivers, stable flows of clear, cool water provide favourable conditions for the development of diverse river macrophyte and faunal communities, making them an important ecological resource (Berrie 1992). As such, they have received considerable conservation attention. A number of chalk rivers have been designated as Sites of Special Scientific Interest (SSSIs) (Mainstone 1999), and they are a priority habitat under the EU Habitats Directive (92/43/EEC). However, chalk rivers are under increasing management pressures, including low-flow issues, due to enhanced groundwater abstraction. This has led to the drying out of upper reaches, as well as the accumulation of silt and changes in the aquatic macrophyte structure (UK

BAP 2011). Like most lowland rivers, chalk rivers are generally highly modified systems. Many have been dredged, lowered and channelized for flood defence and land drainage purposes (UK BAP 2011).

At Hunworth Meadow, a seasonal pattern of macrophyte growth and recession occurs in the river channel, which is characteristic of chalk stream flora (Ham *et al*. 1982). Dominant macrophytes are *Apium nodiflorum (*fool's watercress*), Phalaris arundinaceae (*reed canary grass*) and Rorippa nasturtium-aquaticum* (watercress); these species characteristically dominate chalk stream flora in summer (e.g Holmes *et al*. 1998). The large surface area provided by macrophytes is important for stream invertebrates, providing refugia and habitat diversity (Berrie 1992).

Although chalk streams and rivers are an important ecological resource, they are scarce and declining, having suffered from human pressures. The River Glaven has been modified by agricultural and flood management practices, such as river channelization, construction of artificial embankments, soil drainage, and substantial increases in the application of inorganic fertilisers; nevertheless the river flows through numerous habitat types that are of high conservation value (e.g. wet meadows, fen meadows, riparian woodlands, shallow lakes, and coastal marshes), which support a diversity of aquatic and terrestrial invertebrates and several important and protected bird species such as heron (*Ardea cinerea*), barn owls (*Tyto alba*), kingfisher (*Alcedo atthis*), lapwing (Vanellus vanellus), and osprey (*Pandion haliaetus*) (Sayer and Lewin, 2002).

At the Hunworth Meadow river restoration site (detailed in Section 3.7), many aquatic species of conservation interest were already present before the restoration work commenced, including otters (*Lutra lutra*), the native white-clawed crayfish (*Austropotamobius pallipes*), brook lampreys (*Lampetra planeri*), and bullheads (*Cottus gobio),* which are all listed Special Area of Conservation (SAC) species under the European Habitats Directive (92/43/EEC of 21 May 1992) (JNCC 2013a) as well as wild brown trout (*Salmo trutta)*, and water voles (*Arvicola amphibious)* which are listed as a UK Biodiversity Action Plan (UK BAP) priority species (JNCC 2013b).

The River Glaven is protected under the Freshwater Fish Directive, Natura 2000 (Habitats and/or Birds Directive), Nitrates Directive, and Shellfish Water Directive; hence, respectively, the River Glaven is deemed a suitable water body for sustaining fish populations, contains valuable and threatened species and habitats, is necessary of protection against agricultural runoff, and warrants protection of habitat (in terms of water quality) for edible shellfish (Table 3.1). These management directives form integral components of the EU Water Framework Directive (WFD) environmental targets for surface waters to achieve ‘good’ ecological and chemical status.

The WFD ecological status of a water body is recorded on a 5 tier scale (high, good, moderate, poor and bad), and chemical status is recorded as good or fail (Environment Agency 2009). The current overall WFD classification for the River Glaven is ‘Moderate’ based on moderate fish status and good invertebrate status, good status for surface water chemistry (e.g. ammonia, dissolved oxygen, pH, phosphate), and poor flow dynamics (Table 3.2). A goal of “Good” ecological status is to be achieved by 2027 for the river rather than 2015 due to the expected high cost of achieving the WFD target.

River restoration is one method being employed to improve ecohydrological conditions within rivers and along the riparian zone. In this study the removal of river embankments is being evaluated as a means to improve flood storage, and the hydrological regime on the floodplain for diversifying wet meadow vegetation. The River Glaven is currently failing to meet the WFD goals for river flow, and fish habitat. Hence it is important to determine the success of restoration projects that aim to reinstate more natural river morphology, and whether these methods can be used to help achieve the WFD ecological, chemical and hydrological goals.

Table 3.2: Water Framework Directive Classifications for the River Glaven (Environment Agency 2009).

Waterbody Category and Map Code.:	River - R3	**Surveillance site:** No
Waterbody ID and Name:	GB105034055780 Glaven	
National Grid Reference:	TG 09121 37493	
Current Overall Status	Moderate	
Status Objective (Overall):	Good by 2027	(For Protected Area Objectives see Annex D)
Status Objective(s):	Good Ecological Status by 2027	
Justification if overall objective is not good status by 2015:	Disproportionately expensive	
Protected Area Designation:	Freshwater Fish Directive, Natura 2000 (Habitats and/or Birds Directive), Nitrates Directive, Shellfish Water Directive	
SSSI (Non-N2K) related:	No	
Hydromorphological Designation:	Not Designated A/HMWB	
Reason for Designation:		
Downstream Waterbody ID:	GB520503403600	

Ecological Status	
Current Status (and certainty that status is less than good)	Moderate (Very Certain)

Biological elements

Element	**Current status (and certainty of less than good)**	**Predicted Status by 2015**	**Justification for not achieving good status by 2015**
Fish	Moderate (Very Certain)	Moderate	Disproportionately expensive (M5a)
Invertebrates	Good	Good	

Supporting elements

Element	**Current status (and certainty of less than good)**	**Predicted Status by 2015**	**Justification for not achieving good status by 2015**
Ammonia (Phys-Chem)	High	High	
Dissolved Oxygen	Good	Good	
pH	High	High	
Phosphate	High	High	
Temperature	High	High	
Copper	High	High	
Zinc	High	High	
Ammonia (Annex 8)	High	High	

Supporting conditions

Element	**Current status (and certainty of less than good)**	**Predicted Status by 2015**	**Justification for not achieving good status by 2015**
Quantity and Dynamics of Flow	Does not Support Good (Uncertain)	Does not Support Good	Disproportionately expensive (HR2a)
Morphology	Supports Good	Supports Good	

3.6 Modification of the River Glaven and land management at Hunworth Meadow

The river flows through arable land, deciduous and coniferous woodland in the upper reaches, while grazing meadows dominate the middle reaches and former floodplain in the lower reaches, with most of the former floodplain environments currently disconnected from the river by embankments. Many reaches have been subject to extensive alterations, which have involved the deepening and straightening of the channel and the construction of embankments. In addition, some floodplain areas have been drained and the natural vegetation has been widely cleared and transformed for agriculture. The natural flow of the river has also been interrupted or diverted by numerous weirs and mills (five mills in total: Hempstead, Hunworth, Thornage, Letheringsett and Glandford). The river Glaven flows through three manmade lakes, two upstream of the study site at Edgefield Hall and Hempstead Mill, and one downstream at Bayfield Hall (Oddy 2010). The mill at Thornage is located approximately 500 m downstream of the study site and is shown in Figure 3.1.

At Hunworth, the River Glaven was constrained by embankments along the entire length of the meadow study site (Figure 3.9). Hunworth Meadow slopes gently in an approximate SW to NE direction, is approximately 400 m long, 40 – 80 m wide and has an area of approximately 3 ha. It is bounded to the north-east by an arable and woodland hillslope (Figure 3.1). An agricultural drainage-ditch on the meadow runs parallel to the river close to the base of this hillslope, and at the time of the study was blocked towards the downstream end of the meadow, impairing the site’s drainage and leading to near-permanent surface water within a ponded area adjacent to the ditch.

Waterlogging at the site was evident during much of the year, particularly in the northern, downstream region of the meadow where a pond (Figure 3.10) had developed adjacent to a blocked drainage ditch (Figure 3.1). These waterlogged soil conditions had affected the vegetation community composition on the meadow; site walks across the meadow indicated that *Holcus lanatus (*Yorkshire fog) and *Juncus effusus* (Common rush) were the dominant plant species. Meadows that are dominated by these species are classified under the National Vegetation Classification (NVC) as mesotrophic rush-pasture communities (MG10), and are indicative of wet soils with low plant diversity (Rodwell, 1992). In the ponded area, true aquatic plants are found such as *Potamogeton natans (*broad-leaved pondweed), *Typha latifolia (*broadleaf cattail)*, Sparganium erectum* (branched bur-reed), which provided a nesting and feeding habitat for wildfowl.

The management history of Hunworth Meadow is known from 1992 onwards. The meadow was intensively grazed by cattle until 2000, after which a less intense grazing regime, using mainly cattle with some sheep, has been established. Low levels of inorganic fertiliser were used until 1997, but since then Hunworth Meadow has not received any fertiliser application (Ross Haddow, Stody Estate, personal communication, 9 October 2012).

Figure 3.9: The River Glaven at Hunworth Meadow, showing the river embankments along the reach in 2007.

Figure 3.10: Pond at the downstream end of Hunworth Meadow. Note the visually different vegetation around the pond, and extensive patches of *Juncus effusus* in the background along the ditch.

The River Glaven has been straightened and its channel relocated at various times in the past. At Hunworth Meadow, the river was most recently moved around 1800 during the reconstruction of Thornage Mill located approximately 100 m downstream from the study site (Figure 3.1) (Oddy 2010). A disused brick railway bridge, which was once part the old Midland and Great Northern Joint railway line from Cromer to Melton Constable, is located in the south east corner of the study site (Figure 3.1). It is likely that the river was diverted to flow under the bridge, which was completed in 1884 (Science Museum Group 2007).

The historical Ordnance Survey map of Hunworth Meadow from 1890 shows the remnants of an old channel adjacent to the current location of the river channel (Figure 3.9b), which has significantly reduced in size and separated from the channel in later Ordnance Survey maps of the meadow (Figure 3.11b-d).

Historical Ordnance Survey maps confirm that the stretch of river at the study site has not been moved since records began at Hunworth in 1890 (Figure 3.11a). The river channel was subsequently deepened and embanked for flood defence purposes during the 1960s and 1970s. The agricultural drainage-ditch, which runs parallel to the river along the floodplain, first appears on maps in 1980, and was possibly installed sometime between 1950 and 1980 (Figure 3.11d).

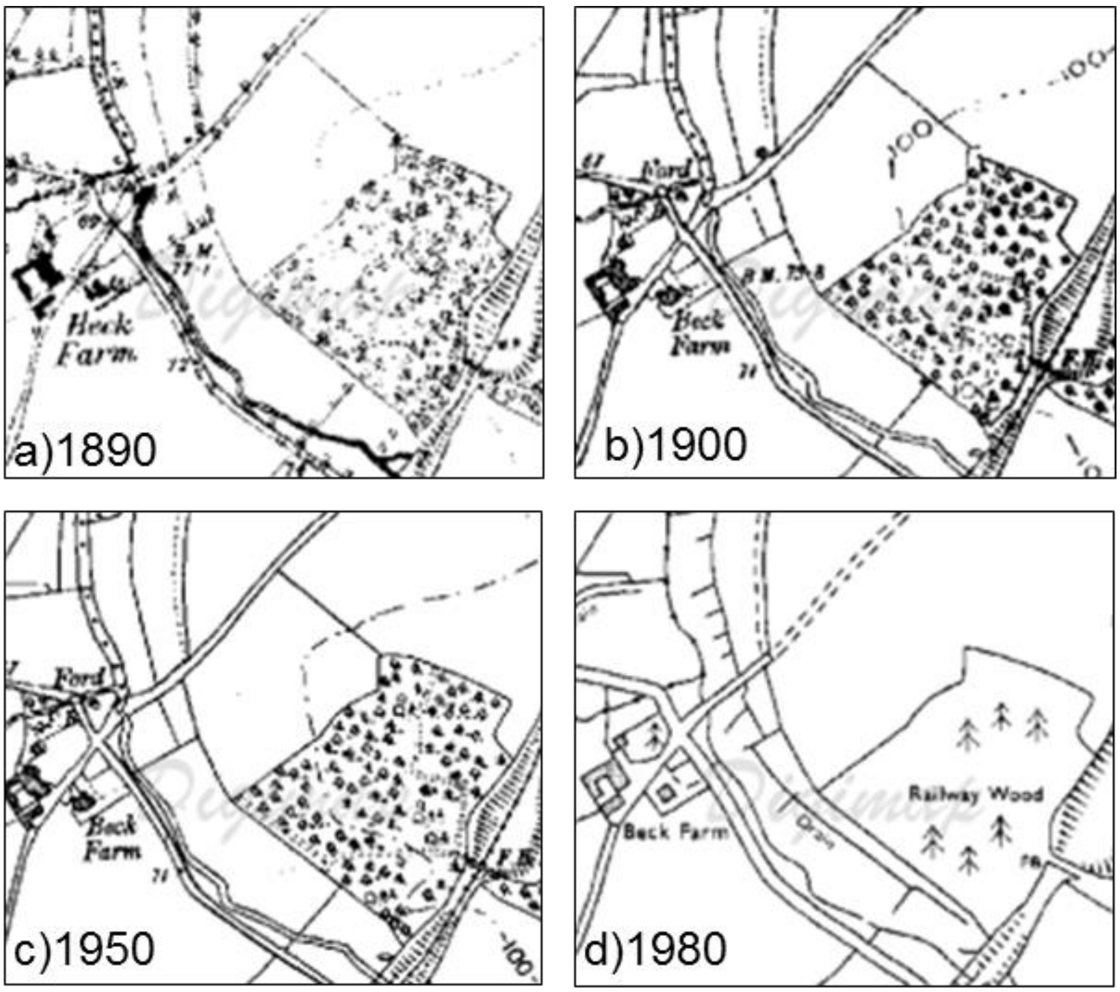


Figure 3.11: Historical maps of the study site at Hunworth Meadow from Historic Digimap.

3.7 The River Glaven restoration project

Chalk rivers such as the Glaven are low energy systems poorly suited to autonomously reinstate their natural channel structure once it has been disturbed by engineering works (Sear *et al*. 2000). Hence, river restoration through the reconfiguration of river embankments and the channel bed form an integral part of returning the natural state and functioning of many chalk rivers. At Hunworth, restoration of the 400 m reach of river was undertaken between 18 - 27 March 2009 by the Environment Agency in collaboration with the River Glaven Conservation Group, the Wild Trout Trust and Natural England. The aim of this restoration project was to increase hydrological connectivity between the over-deepened, embanked river, and its long abandoned floodplain to improve flood storage capacity, site drainage, and ecological diversity within the floodplain (e.g. Leyer 2005; Acreman *et al*. 2007; Hammersmark *et al*. 2008).

Removal of the embankments and the re-profiling work extended up to 10 m into the meadow and involved the use of heavy machinery, which resulted in the removal of approximately 1,400 tonnes of soil from the embankments (Figure 3.12). Removal of the embankments lowered the surface elevation of the riverbanks to the level of the adjacent meadow (Figure 3.13a and b); the river margin was subsequently profiled so that elevation gradually increased from the river towards the meadow, with the hope of introducing a wet to damp moisture gradient from the river.

The restoration work was restricted to the removal of the river embankments and did not involve modification of river channel geomorphology (Figure 3.13a and b). However, during the excavation of the river banks, some sediment was inadvertently deposited in the river and likely accumulated in downstream reaches. One section along the embankment was not removed during the restoration works in order to protect water vole burrows that were found during a pre-restoration flora and fauna impact survey. This section was, by chance, a low point along the embankments, and was located on the river bend midway along the meadow. It is easily identified in Figure 3.13b as the undisturbed, vegetated area surrounded by exposed soil. Relocation of a few young alder trees that lined the river in the southern part of the meadow was required in order for the embankment to be reprofiled in this region. The alders were replanted at other points along the lowered embankment (Figure 3.13b).

Figure 3.12: Embankment removal work in progress (March 2009).

The spoil, which was removed from the site, was largely composed of peaty soil, however within this there were sand and gravel horizons that likely originated from the river. These river materials were probably placed on the meadow when the embankments were constructed and the river was deepened in the 1960–70s. Approximately half of the embankment spoil was removed to an arable field, and the remaining half was stored on the meadow for use during a second phase in-stream restoration project. This second phase, which is discussed in Section 10.2.2, was conducted on the same stretch of the River Glaven in August 2010, one year after the embankment removal and after the main period of fieldwork reported in this study. This involved the creation of a new, narrower and more geomorphologically diverse, meandering river channel, aimed at enhancing the in-stream habitat for macroinvertebrates and spawning fish. This thesis focuses on the ecohydrological research undertaken in the monitoring and modelling of the embankment removal at Hunworth Meadow, which is presented in the succeeding chapters.

Figure 3.13: Photographs of the River Glaven at Hunworth Meadow showing (a) the river embankments in January 2009 prior to the restoration and (b) the completed restoration work with embankments removed in March 2009.

Chapter 4: Methods Part I – hydrological and chemical monitoring

4.1 Introduction

In order to investigate the ecohydrological implications of river embankment removal, a three year hydrological monitoring program was conducted at Hunworth Meadow. It included pre- and post-restoration conditions, fine scale analyses of floodplain topography, soil and water biogeochemistry, and vegetation biodiversity. Results from this programme were subsequently employed within simulations of key hydrological surface water and groundwater processes.

This chapter describes the study design, and details the monitoring regime used to study river-floodplain hydrological connectivity prior to and following the river restoration. This includes information on the collection of hydrological data from the river and floodplain, sampling and analysis of surface and groundwater biogeochemistry, monitoring of local meteorological conditions, and topographic surveys of the river, embankments, and floodplain.

4.2 Study design

Continuous observations of groundwater depth and river stage, measurements of groundwater chemistry, and surveys of topography were collectively used to determine the hydrological impacts of river restoration. Groundwater wells were installed across the meadow in February 2007 in three transects approximately 33 – 39 m in length, each consisting of four or five wells. Transects extended from the base of the arable and woodland hillslope to the river-embankments and were aligned perpendicular to the river, i.e. parallel to the assumed main groundwater flow direction (Figures 4.1 and 4.2). In addition to the Environment

Agency gauging station located immediately upstream of the study site (Figure 3.1), a stage board was installed in the river adjacent to the downstream well transect (Figure 4.2).

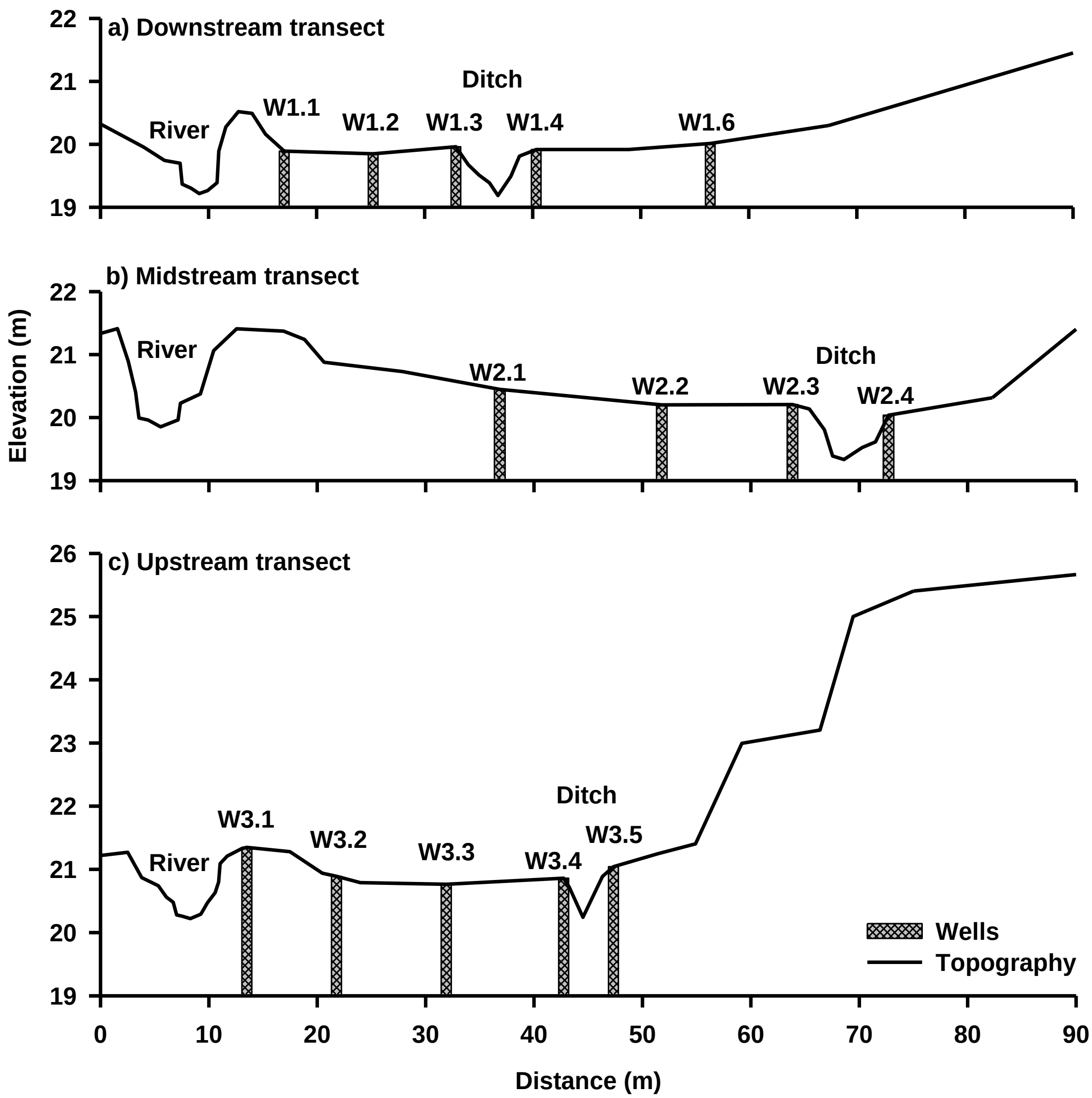


Figure 4.1: Cross-sections of the embanked river and floodplain showing the location of groundwater wells along the three well transects.

The wells were installed at varying depths between 1.3 and 2.0 m due to the presence of alluvial gravels which proved difficult to penetrate with a hand-auger. The wells were constructed from polypropylene pipe (inside diameter =

3.0 cm), screened with 3 mm diameter holes, and wrapped in geotextile cloth to prevent blockage by fine silts. The tops were covered with rubber caps between sampling dates. To prevent cattle trampling and damaging the wells, the top of the wells were approximately 1 – 2 cm below the soil surface, and covered with a concrete slab (ca. 30 cm × 30 cm) (Figure 4.3).

An enclosure was installed at the upstream well transect in 2009 to protect *in-situ* monitoring equipment from damage from grazing livestock. The enclosure contained solar panels and dataloggers, which were wired to oxygen optodes, tensiometers, and theta probes buried in the soil profile (Figure 4.4a and 4.4b) to monitor soil and well oxygen concentrations, soil water potential and soil volumetric moisture content, respectively (see Section 4.4.3). Well 3.3 was also located within the enclosure (Figure 4.2).

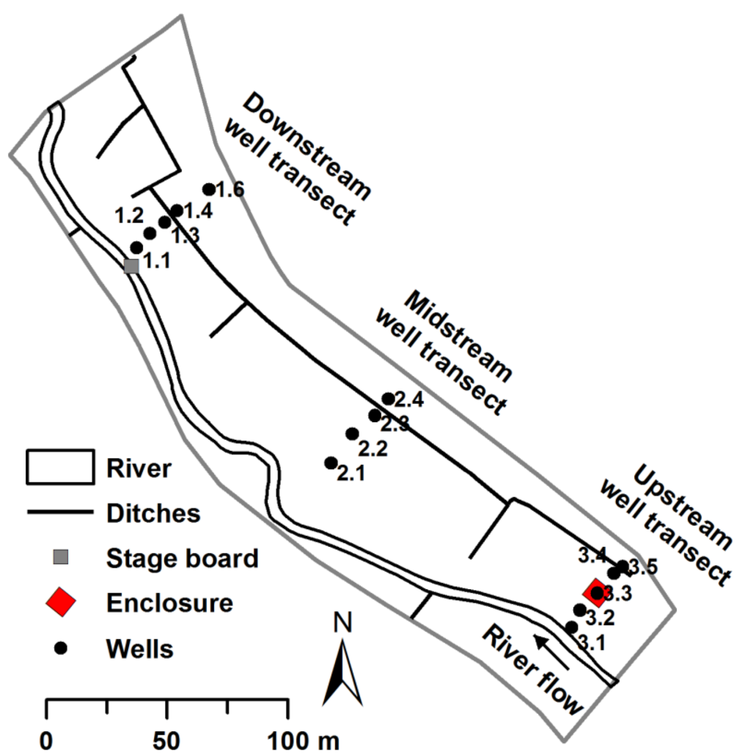


Figure 4.2: Sampling design at Hunworth Meadow showing the location of the groundwater wells.

Figure 4.3: Photograph of the concrete slab covering Well 2.1.

Floodplain vegetation surveys and sampling of floodplain soil chemistry were conducted prior to the restoration to determine the physicochemical controls on floodplain plant community composition. Vegetation surveys were conducted in late June 2008 across the entire meadow on a regular 10 × 10 m sampling grid using 1 m^2 quadrats, which resulted in 214 survey points across the meadow (Figure 4.4a). A detailed description of the vegetation survey methods are given in Section 8.2. Floodplain soils samples were collected on 29/04/2008 across the meadow at depths of 0.1 – 0.2 m. Soil samples were collected from 113 points at regular intervals across the vegetation sampling grid (Figure 4.4b). Soil collection and analytical methods are given in Sections 4.4 and 8.3.

The response of groundwater elevation relative to precipitation and river stage was determined using data from an automatic weather station (MiniMet SDL 5400, Skye, Powys, UK) that was located approximately 200 m from the study site, and an Environment Agency (EA) gauging station (#034052) located immediately upstream of Hunworth Meadow (Figure 3.1; Figure 4.5). The weather station was installed in 2007 and stored precipitation, air and soil temperature, net radiation, relative humidity, windspeed and wind direction data at 30 minute intervals. The weather station was programmed to average measurements at 30 second sample periods, with the exception of total precipitation which was measured at 30 minute intervals.

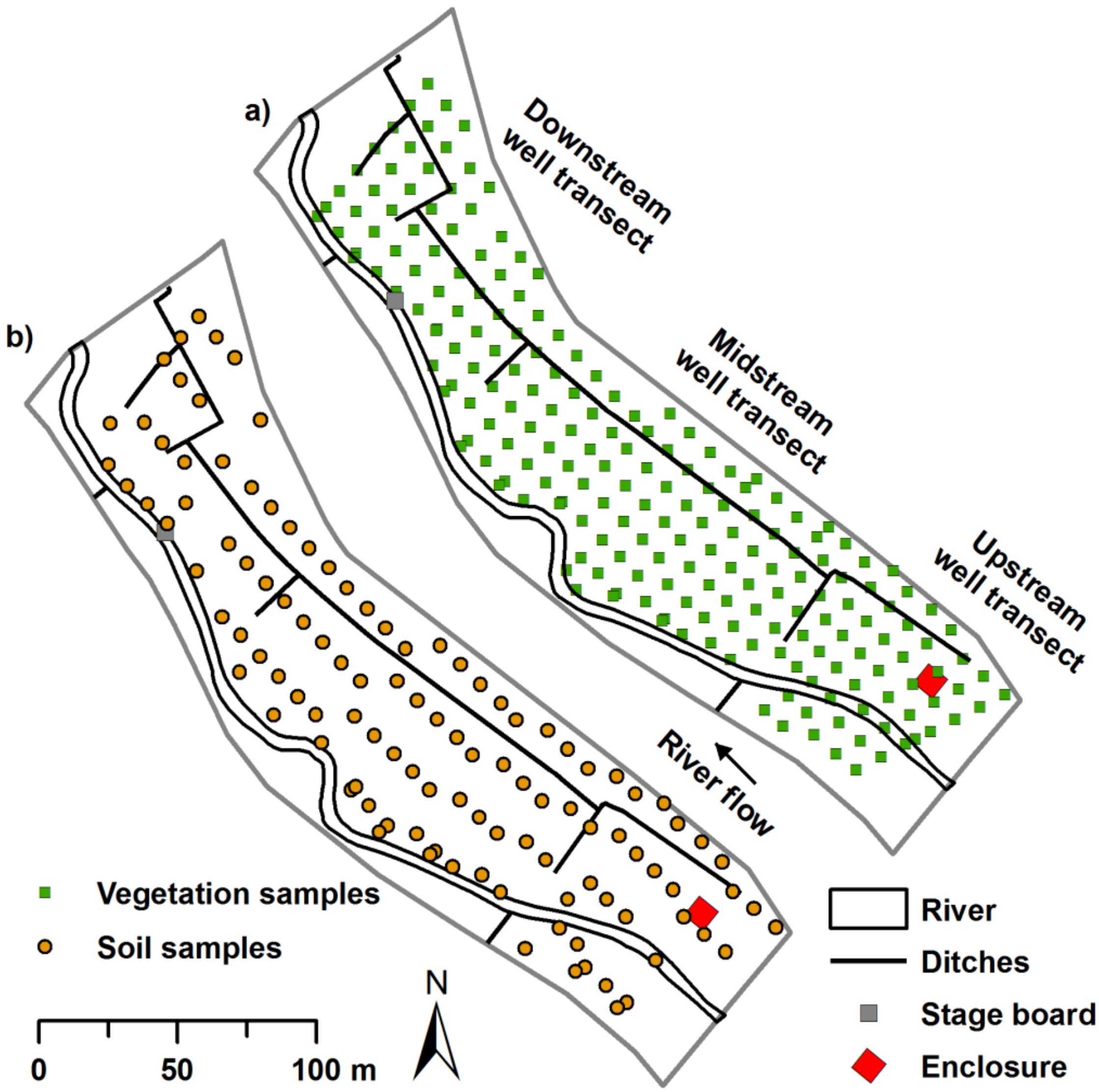


Figure 4.4: Sampling design at Hunworth Meadow showing the location of the (a) botanical sampling and, (b) soil sampling points. The enclosure at the upstream end of the meadow protected *in-situ* sampling equipment within from grazing cattle.

Figure 4.5: Photographs of (a) the automatic MiniMet weather station and (b) the Environment Agency gauging station located at Hunworth (#034052).

The EA gauging site at Hunworth Meadow is a relatively wide, shallow channel with a bed width of 4 m and maximum depth of 1 m. River discharge is measured using a v notch weir design (Figure 4.5b). Mean daily river stage and discharge data were available for the Hunworth gauging station for the period 2001 – 2010. However, there were some gaps in the record, because seasonal macrophyte growth in the channel downstream of the gauging station caused water to backup over the weir that impacted on the rating curve and reduced the accuracy of the data during these periods. This was manifested in a slow increase in baseflow through summer, despite low or no rainfall, upon which individual peaks associated with rainfall events were superimposed. Subsequently, this apparent elevated baseflow would decline during the autumn due to macrophyte dieback. The influence of the vegetation on discharge measurements was, in some cases, removed abruptly during flood events, possibly due to a devegetation of the river channel and relocation of sediment downstream. The influence of macrophyte growth on river discharge was easily identified in the discharge record when compared with precipitation data. The affected data were excluded from quantitative analyses (baseflow index, flow duration) (see Figures 5.5, 5.10, and 6.26). Another EA gauging station is located on the River Glaven approximately 5 km downstream of the study site at Bayfield (Station Number: 034016). Mean river discharge at Bayfield was recorded from 1970 – present, however, unfortunately the river discharge at Bayfield is not rated above 0.3 m (ca. 0.65 $m^3 s^{-1}$) due to the influence of Letteringsett mill, located immediately upstream. Data quality and the methods used to substitute missing data in the discharge record at Hunworth are described in Sections 6.3 and 6.5.

4.3. Hydrological monitoring

4.3.1 Groundwater and surface water levels

To characterize fluctuations in groundwater elevation, Solinst combined pressure transducer-dataloggers (Levelogger Gold 3.0, Ontario, Canada) were installed in four wells at the upstream transect, one well at the midstream transect, and in five wells at the downstream transect. Groundwater elevation was recorded hourly from February 2007 – August 2010. A barologger was positioned on a fence post on the boundary of Hunworth Meadow above the water level and in air in order to compensate for barometric pressure. The Solinst level loggers record temperature compensated water level and operate accurately within a thermal range of -20 to +80 ˚C. The loggers are completely autonomous, and can remain *in situ* recording for long periods of time (battery life = 10yrs; max. 40,000 readings) (Solinst 2006). However, for this study the leveloggers were downloaded and the readings checked with manual measurements of groundwater elevation during regular field visits.

4.3.2 Base flow index and flow exceedance values

To assess the permeability of the surrounding geology at the study site, and the potential for subsurface river-floodplain hydrological interactions, the river base flow index (BFI) was calculated for each full year of EA river discharge data at the site using a base flow separation program (BFI version 4.15) (Wahl and Wahl 2007). The BFI program follows the Institute of Hydrology (1980) base flow separation procedure, where the water year is divided into 5-day increments to identify minimum flow, called a turning point. The turning points are then connected to obtain the baseflow hydrograph. The volume of base flow for the period is estimated by the area beneath the hydrograph. The base-flow index is calculated as the ratio of baseflow volume to the total volume of streamflow. High BFI values indicate groundwater dominance, which is broadly represented by a stable flow regime (Sear *et al*. 1999). A detailed description of the BFI calculation is given by Gustard *et al*. (1992).

The contribution of groundwater to total river flow was also analysed using flow exceedance values for Q10 (a high flow threshold that is equalled or exceeded for 10% of the flow record) and Q95 (a low flow threshold that is equalled or exceeded for 95% of the flow record), which were determined from a river flow duration curve. Q95, expressed as the percentage of mean annual river flow, and comparisons of Q10 and Q95 provide a measure of the variability (i.e. flashiness) of the flow regime (Gustard *et al*. 1992; Marsh and Hannaford 2008).

4.3.3 Evapotranspiration

Daily values of Penman-Monteith potential evapotranspiration (Monteith 1965) were computed from meteorological data (temperature, net radiation, humidity, wind speed) provided by the on-site weather station (Figure 4.6). These data were supplemented by a nearby (<9 km distance) British Atmospheric Data Centre weather station (Source ID: 24219, Mannington Hall) (Figure 3.1), however, there were periods of time where data were not available. In instances where wind speed data was unavailable an average wind speed value of 1.6 m s^{-1} was used, computed from the Hunworth Meadow data.

The reference formula for grasslands described by Allen (2000) was used to calculate potential evapotranspiration. This equation assumes a constant grass height of 0.12 m throughout the year (see also Hough and Jones, 1997), and a fixed surface resistance of 70 s m^{-1} with an albedo of 0.23 (Allen 2000), and was calculated as:

$$ET_o = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T + 273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)} \qquad (4.1)$$

where, ET_o is reference evapotranspiration (mm day^{-1}), R_n is net radiation at the crop surface (MJ m^{-2} day^{-1}), G is the soil heat flux (MJ m^{-2} day^{-1}) which is small compared to R_n, especially when the surface is covered by vegetation and calculation time steps are >24 hours and thus was set to 0 in this instance, T is the mean daily air temperature at 2 m height (°C), u_2 is the wind speed at 2 m height (m s^{-1}), e_s is the saturation vapour pressure (kPa), e_a is the actual vapour pressure (kPa), Δ represents the slope of the saturation vapour pressure temperature relationship (kPa $°C^{-1}$), and λ is the psychrometric constant (0.067 kPa $°C^{-1}$ for 0 m altitude) (Allen 1998).

Actual vapour pressure, e_a (kPa), the vapour pressure exerted by the water in the air, was calculated using relative humidity data as:

$$ET_o = \frac{e^o(T_{min})\frac{RH_{max}}{100} + e^o(T_{max})\frac{RH_{min}}{100}}{2} \qquad (4.2)$$

where e°(Tmin) is the saturation vapour pressure at daily minimum temperature (kPa), e°(Tmax) is the saturation vapour pressure at daily maximum temperature (kPa), RH_{max} is the maximum relative humidity (%), and RH_{min} is the minimum relative humidity (%).There were some short periods of time where the maximum humidity data was either missing or of questionable quality (Figure 4.6). In the absence of maximum humidity data an alternative method of calculating actual vapour pressure for the potential evapotranspiration equation was used. The actual vapour pressure (e_a) which equals the saturation vapour pressure at the dewpoint temperature (T_{dew}), was estimated by assuming that $T_{dew} \approx T_{min}$ in the following equation:

$$e_a = e^o(T_{dew}) = 0.6108 \exp\left[\frac{17.27 T_{dew}}{T_{dew} + 237.3}\right] \tag{4.3}$$

The relationship $T_{dew} \approx T_{min}$ holds in wet environments where the air is nearly saturated with water vapour at sunrise when the air temperature is close to the daily minimum temperature (T_{min}) (Allen 1998).

Saturation vapour pressure, e_s, was calculated as:

$$e_s = \frac{e^o(T_{max}) + e^o(T_{min})}{2} \tag{4.4}$$

where, e°, is the saturation vapour pressure (kPa) at a given temperature, T:

$$e^o(T) = 0.6108 \exp\left[\frac{17.27 T}{T + 237.3}\right] \tag{4.5}$$

The difference between the saturation and actual vapour pressure, ($e_s - e_a$), is the saturation vapour pressure deficit and indicates the actual evaporative capacity of the air. The slope of the relationship between saturation vapour pressure and temperature, Δ, at a given temperature was calculated as:

$$\Delta = \frac{4098\left[0.6108 \exp\right]\frac{17.27T}{T + 237.3}}{(T + 237.3)^2} \tag{4.6}$$

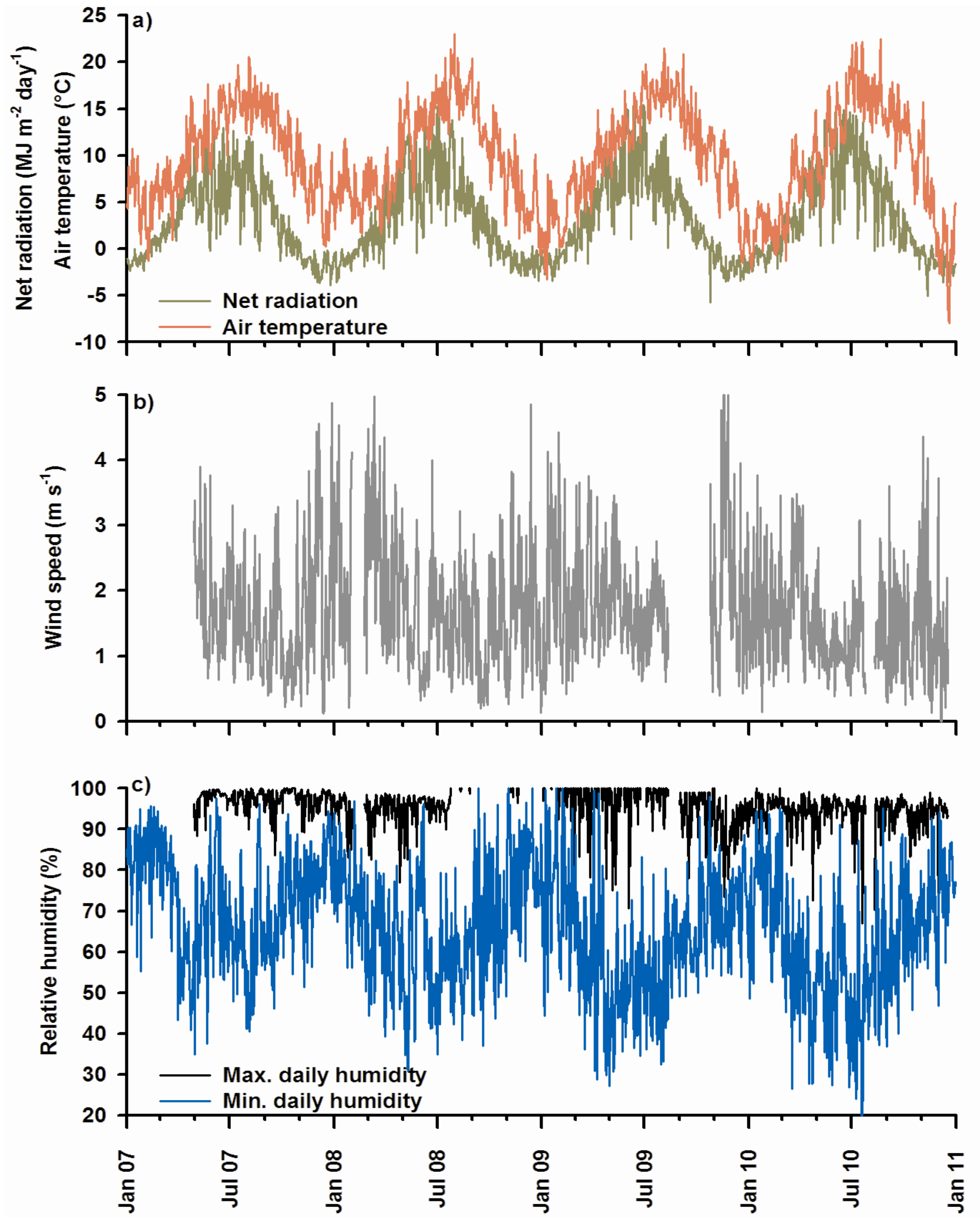


Figure 4.6: Time series of (a) mean daily air temperature and net solar radiation, (b) mean daily wind speed, and (c) maximum and minimum relative humidity.

4.3.4 Hydraulic conductivity

Hydraulic conductivity values of the top 2 m of soil were determined using piezometer slug tests (n = 25) following the approach of Surridge *et al.* (2005). This included six slug tests in the geotextile-screened monitoring wells (inside diameter = 3.0 cm) already installed at the site, and 19 slug tests using piezometers with larger intakes (Figure 4.7) specifically designed to reduce flow resistance through the screens and more accurately measure flow through porous organic layers. After inserting a continuous-logging Solinst 3.0 pressure transducer into the base of the piezometer, programmed to take readings every 10 seconds, a slug, which consisted of a sand ballast-filled PVC tube (outside diameter = 2.2 cm) sealed at both ends with rubber bungs, was lowered into the piezometer (Figure 4.8).

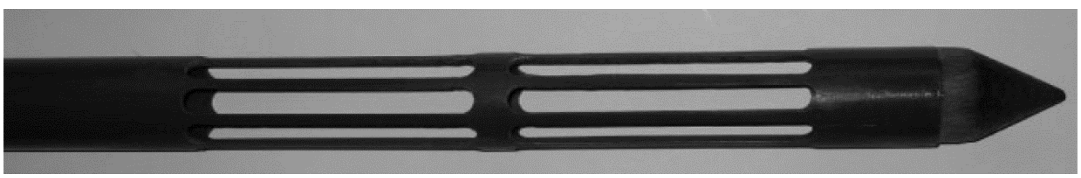


Figure 4.7: Photograph of the piezometer intakes used in the slug tests.

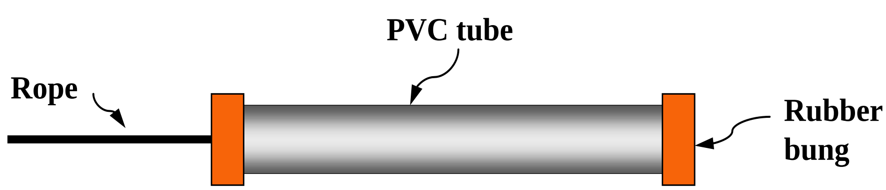


Figure 4.8: Diagram of the sand-filled slug.

The water level (head) in the piezometer was then left to recover to pre-test levels, at which point the slug was removed smoothly and swiftly from the piezometer. The pressure transducer was removed and downloaded once the water level in the piezometer had again returned to pre-test levels. A typical response in water table height during the slug tests at Hunworth Meadow to determine hydraulic conductivity is shown in Figure 4.9. The water level recovery in Figure 4.9 is from a test piezometer installed adjacent to the upstream observation well 3.1. The water level readings show the recovery of the water table from 0 – 2,000 secs after the insertion of the pressure transducer and the slug during the set-up of the test, and the response of water levels during the test from 2,150 – 5,030 secs after the removal of the slug at 2140 secs.

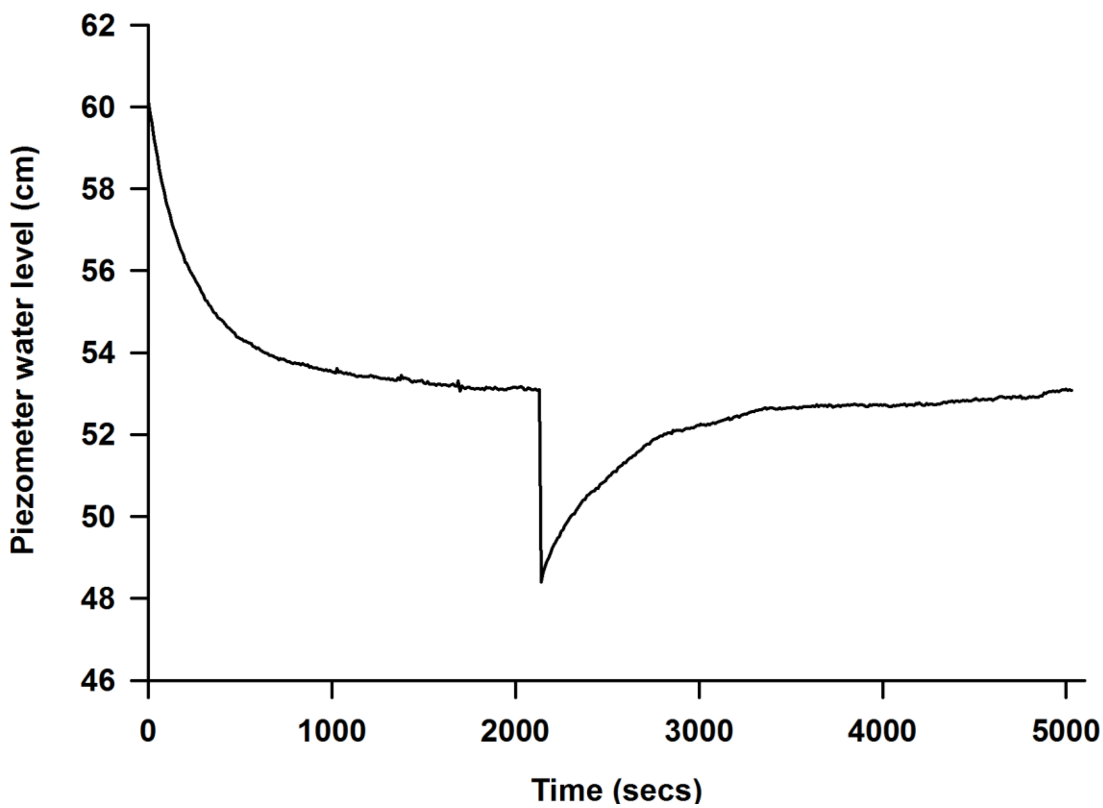


Figure 4.9: Typical response in water table height during slug tests at Hunworth Meadow to determine hydraulic conductivity.

Hydraulic conductivity was calculated from the recovery in hydraulic head assuming Darcian flow as:

$$K = \frac{A}{FT} \tag{4.7}$$

where K is hydraulic conductivity (cm s^{-1}), T is the basic hydrostatic time lag (secs), A is the inside cross-sectional area of the piezometer (cm), and F is the shape factor of the piezometer intake (cm) (Hvorslev 1951; Baird *et al*. 2004; Surridge *et al*. 2005), calculated as:

$$F = \frac{2.4\pi l}{\log_e\left[\frac{1.2l}{d} + \sqrt{1 + \left(\frac{1.2l}{d}\right)^2}\right]} \tag{4.8}$$

where d is the outside of the intake (cm) and l is length of the intake (cm) (Brand and Premchitt 1980; Baird *et al*. 2004). The hydrostatic time lag (T) was solved by fitting equation 4.9 to the measured head recovery data using a least-squares minimization with T as the fitted parameter:

$$\frac{h}{h_0} = e^{-t/T} \tag{4.9}$$

where h is the head difference at t (cm), h_0 is the initial head difference (cm), and t is time from the start of the test (sec) (i.e. slug withdrawal) (Hvorslev 1951; Baird *et al*. 2004).

Assuming Darcian flow, head recovery during slug tests should be log-linear (Hvorslev 1951; Baird *et al*. 2004). Comparisons of the decline in normalised head (h/h_0) following slug-removal (t_0) and the best-fit Hvorslev equation indicate that head recoveries were adequately log–linear (Figure 4.10). Eight additional slug tests were undertaken at the site; however these tests were excluded because the water level recovery was too slow to be captured adequately during the observation period, and the early water level response showed substantial departures from log-linearity. As only a small part of the recovery could be used in the calculation of K, the accuracy of the test was likely affected and these tests were removed from the analysis.

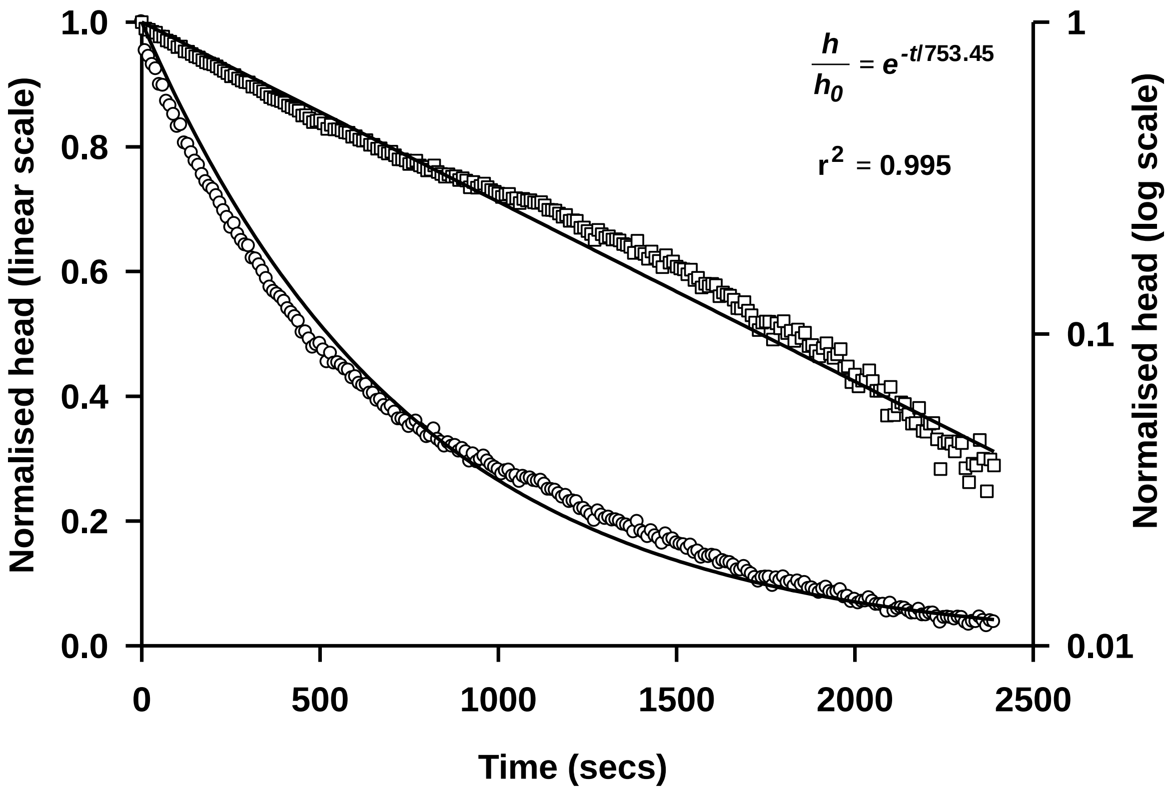


Figure 4.10: Typical example of the log-linear change in normalized head (h_0/h) following slug removal at time 0. Normalised heads are plotted on a linear scale (open circles) and logarithmic scale (open squares). Line indicates best-fit Hvorslev equation; the basic hydrostatic time lag = 753.45 secs, r^2 = 0.995.

Subsurface flow rates were calculated assuming Darcian flow as:

$$u = \frac{K \frac{\Delta h}{\Delta l}}{n_e} \tag{4.10}$$

where u is flow rate (m day^{-1}), n_e is effective porosity, $\Delta h/\Delta l$ is water table slope obtained from hydraulic head measurements taken at each well along the transect, and K is hydraulic conductivity in m day^{-1} (Domenico and Schwartz 1998). Mean values of K and $\Delta h/\Delta l$ for each well transect were used in this calculation.

4.4 River-floodplain biogeochemical monitoring

4.4.1 Soil structural and physical properties

Measurements of soil variables were confined to the upper 0.1 – 0.2 m of floodplain soils. In April 2008, 113 soil samples were collected across the floodplain (Figure 4.4b). Duplicate field and analytical samples were taken randomly, were checked for sampling bias and analytical accuracy and precision, and were averaged to give final values.

Bulk density and porosity were calculated from the difference in the volume of saturated and dry soil (Elliot *et al*. 1999) collected using bulk density rings (I.D = 53 mm, height = 51 mm, volume = 100 cm^3) from the top 0 – 0.2 m of soil. Organic matter content was subsequently analysed by Loss on Ignition (Heiri *et al*. 2001). Total porosity was calculated assuming a particle density of 2.65 g cm^{-3} (Jarrell *et al*. 1999). For the analysis of soil pH, duplicate 15 g field-moist sieved (4 mm) soil subsamples were mixed with 15 ml of dionised water. The samples were left to equilibrate for 30 min and subsequently measured with a HANNA HI8424 pH probe (HANNA Instruments Ltd , Bedfordshire, UK) (Robertson *et al*. 1999).

Soil particle size (soil texture) was determined with optical laser diffraction using an LS 13320 Coulter Counter particle size analyzer (Beckman Coulter Corp., Hialeah, Florida, USA). Prior to analysis, the soil samples were treated with hydrogen peroxide to remove organic matter, and subsequently with sodium hexametaphosphate (Calgon) to disaggregate the soil particles (Chappell 1998). Soil textural class was also recorded using the “feel” method in the field during the collection of soil samples (Brady and Weil 2002). Soil samples were also analysed for exchangeable ions in order to determine nutrient availability for plant and microbial uptake. The salt solutions used to extract the exchangeable ions from the soil, and the analytical methods employed to measure nutrient concentrations in the extractants are described in Chapter 8.

4.4.2 Water chemistry sampling and analysis

River water, shallow groundwater (top 1 − 2 m of soil), and rooting zone pore water samples were taken bimonthly from 2007 − 2008 to examine the spatial variation in subsurface and rooting zone chemistry. Groundwater samples were collected using a point-source bailer and stored in pre-washed 250 mL polyethylene bottles. Before acquiring samples, the wells were purged to introduce fresh groundwater and the collection bottles were rinsed with well water. Pore water in the rooting zone was collected using 0.1 μm rhizon samplers, 0.1 m in length (Figure 4.11a). The rhizon samplers were attached to evacuated test tubes with syringe needles, and inserted into the top 0.1 m soil (Figure 4.11b). The water samples were stored in a cooler until return to the laboratory, refrigerated, and then filtered through 0.45 μm filter paper. The rhizon samples were not filtered in the laboratory as the water had already passed through the 0.1 μm rhizon sampler. Any samples that could not be analyzed within two days were frozen.

Figure 4.11: Photographs of rhizon soil moisture samplers taken in the laboratory (a), and in the field attached to test-tubes collecting pore-water from the rooting zone.

Dissolved oxygen, pH and electrical conductivity were measured in the field at the time of water sampling using a handheld YSI-555A dissolved oxygen meter (YSI Hydrodata Ltd, Letchworth, UK), a Mettler Toledo MP120 pH meter and a Mettler Toledo MC126 conductivity meter (Ohio, USA), respectively. Prior to use in the field, the pH meter was calibrated with pH 4 and pH 7 BDH Prolabo calibration solutions (buffer tablets, VWR, Leicestershire, UK), and the conductivity meter was calibrated using a 1413 μs calibration solution (0.01M KCl @25°C).

In the laboratory, cations (Ca^{2+}, Mg^{2+}, K^+, Na^+, NH_4^+) and anions (SO_4^{2-}, Cl^-, NO_3^-, NO_2^-, PO_4^{3-}) were analysed by ion exchange chromatography (ICS-2500, Dionex Corp., California, USA). Dissolved organic carbon (DOC) and total dissolved nitrogen (TDN) were determined using a HiPerTOC carbon analyser plumbed to a HiPer5000 total nitrogen chemoluminescent detector (Thermo Electron Corp., Delft, The Netherlands). Deep groundwater chemistry data were obtained from an EA borehole within the Glaven catchment, located at Edgefield (52° 52′ 49.36″ N, 01° 05′ 52.91″E), approximately 2 km from Hunworth study site (Figure 3.6). The borehole is used to monitor water levels and chemistry in the chalk strata, and had a response zone of 38 – 41 m below ground surface. The chemistry data obtained from the EA borehole are assumed to be representative of the groundwater chemistry of the chalk underlying the study site. This is reasonable considering the close proximity of the EA borehole to the study site, and that the chalk is laterally continuous in the region (Figure 3.6). It is unknown whether the chalk aquifer is in hydraulic continuity with the river at the study site; however as a point of comparison the chalk groundwater samples were included in the chemical analyses.

4.4.3 Oxygen concentration in soil pores

Measurements of dissolved oxygen in the soil profile were obtained at 30 minute intervals from two Aanderaa 4175 oxygen optodes (Bergen, Norway) attached to Campbell Scientific CR1000 dataloggers (Loughborough, UK; Figure 4.12). The Aanderaa Oxygen Optodes are based on the ability of certain molecules, in this case oxygen, to influence the florescence of other molecules, in a process called dynamic luminescence quenching (lifetime based). The Aanderaa Oxygen Optodes were chosen for this study because, unlike electro-chemical sensors, they do not consume oxygen and are less affected by fouling, making them suitable for systems with limited exchange of water such as in groundwater (Aanderaa 2006). To evaluate temporal fluctuations in oxygen concentration within the rooting zone, the oxygen optodes were installed at the upstream well transect (Well location 3.3) from January 2009 to August 2010 (see Figure 3.1).

The optodes, dataloggers and solar pannels were installed in an enclosure to protect them from livestock (Figure 4.12). Initially, the optodes were installed in wells at 0.1 and 0.3 m depths. However, in January 2010 the 0.3 m oxygen optode was moved from the well and buried directly in the soil at 0.1 m below the ground surface. Although the optodes can remain *in-situ* for more than one year without repeated calibration (Aanderaa 2006), the calibration was checked periodically using a zero-oxygen solution (sodium sulphite saturated in deionised water) and 100% saturated solution (deionised water bubbled with air).

Figure 4.12: Clockwise from top of the photograph, dataloggers, position of the oxygen optodes (protected in plastic containers), and varying depths of the tensiometers.

Soil water potential (matric potential) was measured at 30 minute intervals with five water-filled ceramic cup Delta-T SWT4 tensiometers, three were buried in the enclosure at depths of 0.1, 0.3, and 0.5 m, and two were buried outside of the enclosure: one approximately 13 m from the river at a depth of 0.5 m, and one approximately 9 m from the ditch at a depth of 0.5 m. Two Delta-T ML2X theta probes, attached to Campbell Scientific CR1000 dataloggers (Loughborough, UK), were installed in the enclosure at the site to measure volumetric soil water content at 30 minute intervals (Figure 4.12). The optimal operating range for these probes covered 0 − 0.5 m^3 m^{-3} (accuracy within 1% for soil specific calibration), which corresponds to a voltage output 0 − 1.0 V (Delta-T Devices 1998). All of the readings were at or above 1.0 V, which suggests that these probes were not suitable for soils that are so wet. Due to a narrow range of soil moisture conditions measured at the site, an *in-situ* soil moisture release curve (soil tension versus soil water content) could not be measured, and laboratory analyses of soil moisture release were conducted (see Section 8.3.3). Nevertheless, the theta probes remained *in-situ* for the study period in order to measure any decrease in soil moisture that might suggest periods of soil drying stress. A manual hand-held Delta-T ML2X theta probe was used to cross-check the buried probe readings periodically; furthermore, the response of the buried theta probes in drier soils (collected from the hillslope in a bucket) was checked midway through the study period and compared with the measurements from the hand-held meter.

4.4.4 River and floodplain topography

Surface elevation of the meadow, river channel and before their removal, the embankments were surveyed using a differential Global Positioning System (dGPS) (Leica Geosystems SR530 base station receiver and Series 1200 rover receiver, Milton Keynes, UK; Figure 4.13) in April 2008 prior to the restoration, and in July 2009 after the restoration. Each survey was conducted using the survey pole in static mode, which according to the output data resulted in a 3D coordinate quality of 1 – 2 cm. The meadow was surveyed at intervals of approximately 10 m, whereas the river embankments and cross-sections were surveyed at a higher resolution using intervals of approximately 0.25 – 0.5 m (Figure 4.14). River channel cross-sections were surveyed at 32 transects along the study reach prior to the restoration (Figure 4.14a) and at 23 transects after the restoration (Figure 4.14b). Digital elevation models (DEMs) were created in Arc-GIS using the Kriging interpolation function, which estimates values from a statistically weighted average of nearby sample points (de Smith *et al*. 2007).

Figure 4.13: Photographs of the differential Global Positioning System base station (a) and the rover in-use at Hunworth Meadow.

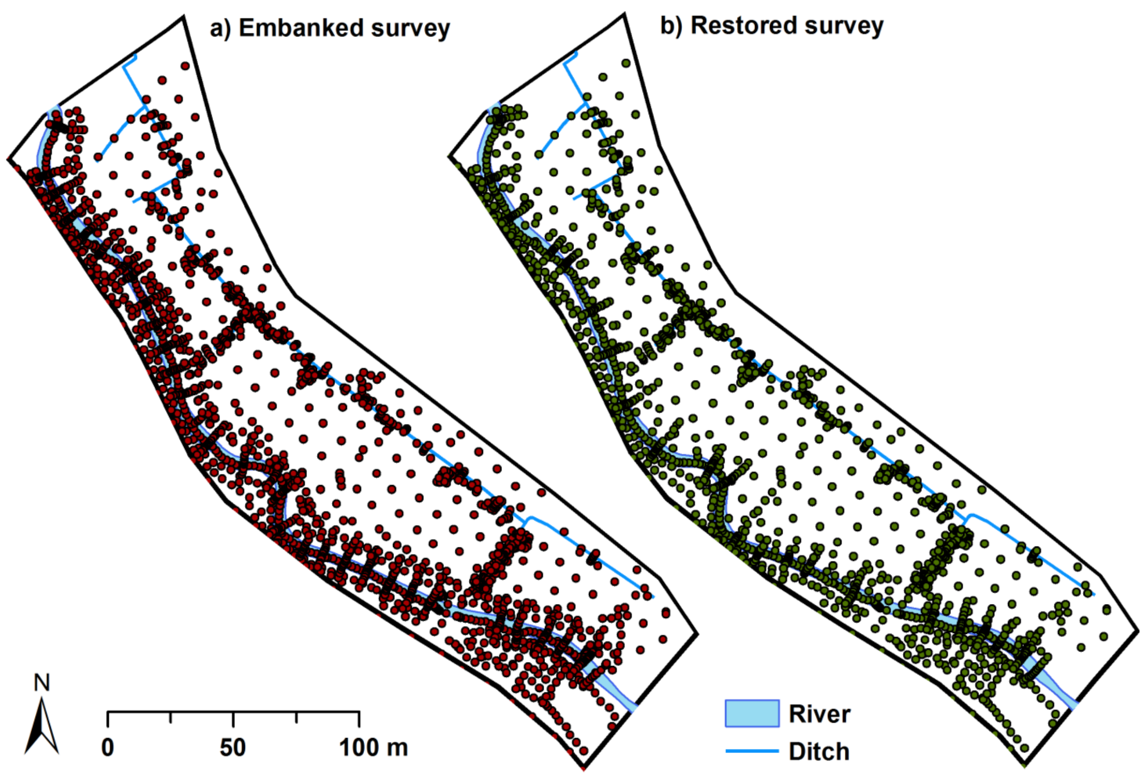


Figure 4.14: Location of dGPS sample points for the embanked (a) and restored (b) topographic surveys.

4.5 Flood prediction

4.5.1 Bankfull capacity

The effects of embankment removal on the frequency of overbank inundation were evaluated by estimating the change in bankfull capacity. A field-based method was used to relate bankfull elevation, measured before and after embankment removal using dGPS, to a stage board installed adjacent to the downstream well transect (Figure 4.15). Bankfull discharge was then predicted using a regression relationship between river stage in the Hunworth Meadow reach and discharge at the EA gauging station upstream of the Meadow, which was located approximately 0.3 km above the stage board (Figure 3.1). This method assumes that groundwater exchanges and runoff inputs do not significantly affect river discharge between the gauging station and stage board, which is reasonable considering their close proximity.

The regression relationship between river stage and discharge for spring/summer (April – September) follows a slightly lower trajectory than for autumn/winter (October – March) ($p<0.05$, F value = 35.41) (Figure 4.15). This can be attributed to the extensive growth of aquatic macrophytes, such as watercress (*Rorippa nasturtium-aquaticum*), fool's watercress (*Apium nodiflorum*), and reed canary grass (*Phalaris arundinaceae*), within the channel during the growing season (Figure 4.16). Seasonal in-stream vegetation growth can have a significant effect on the fluvial dynamics of streams by modifying flow velocities and sedimentation rates (Champion and Tanner 2000; Clarke 2002), resulting in a reduction in channel capacity during spring and summer.

For example, river stage at the downstream well transect was higher during the summer compared with the winter (Table 4.1). However, the higher river stage during the summer was actually associated with lower river discharge compared with the winter (Table 4.1). The effects are often diminished, however, at high flows due to compression of the macrophyte stems, or devegetation, which can substantially reduce flooding potential (Chambers *et al*. 1991; Wilcock *et al*. 1999; Champion and Tanner 2000). For example, Chambers *et al*. (1991) reported that, in slow-flowing rivers studied in western Canada, macrophyte biomass decreased with increasing flow velocities over a mean range of 0.2 – 0.7 m s^{-1}, with aquatic macrophytes typically absent at velocities above 1 m s^{-1}.

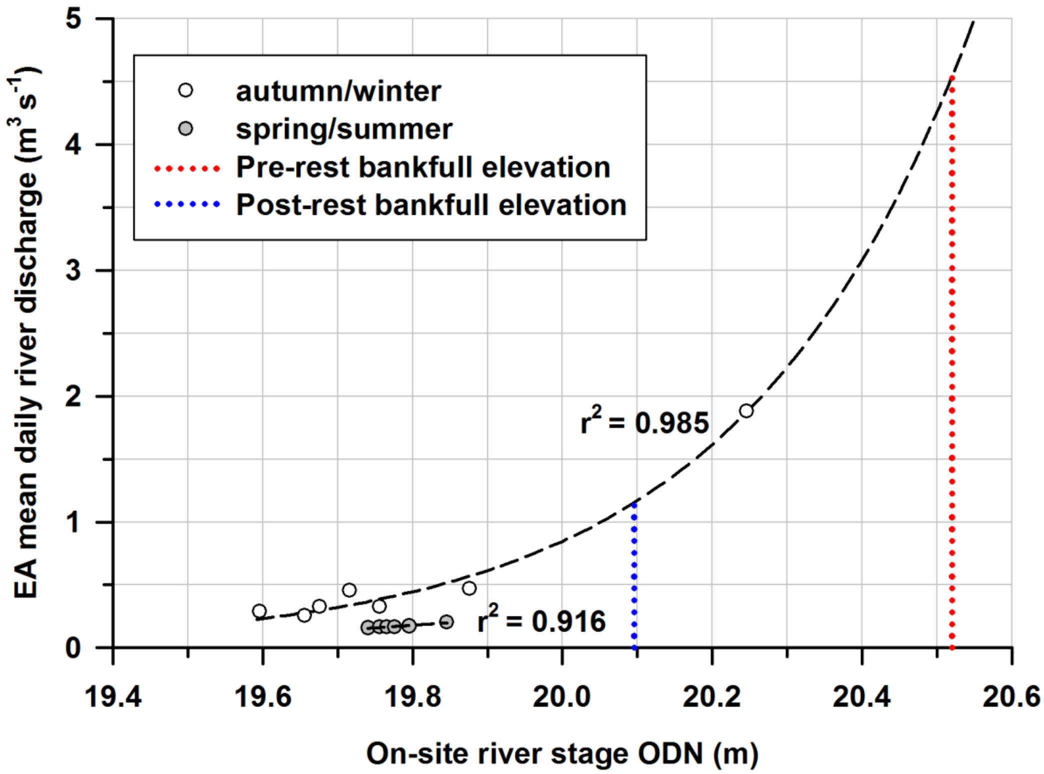


Figure 4.15: Relationship between river stage and mean daily river discharge used to determine bankfull capacity. Lines indicate statistically significant regression at $p<0.05$, $y_{autumn/winter} = 7.6795E^{-29}e^{3.2285x}$, $y_{spring/summer} = 0.415x - 8.0379$. River Glaven discharge data are from the EA gauging station (#034052) located immediately upstream of the Hunworth Meadow.

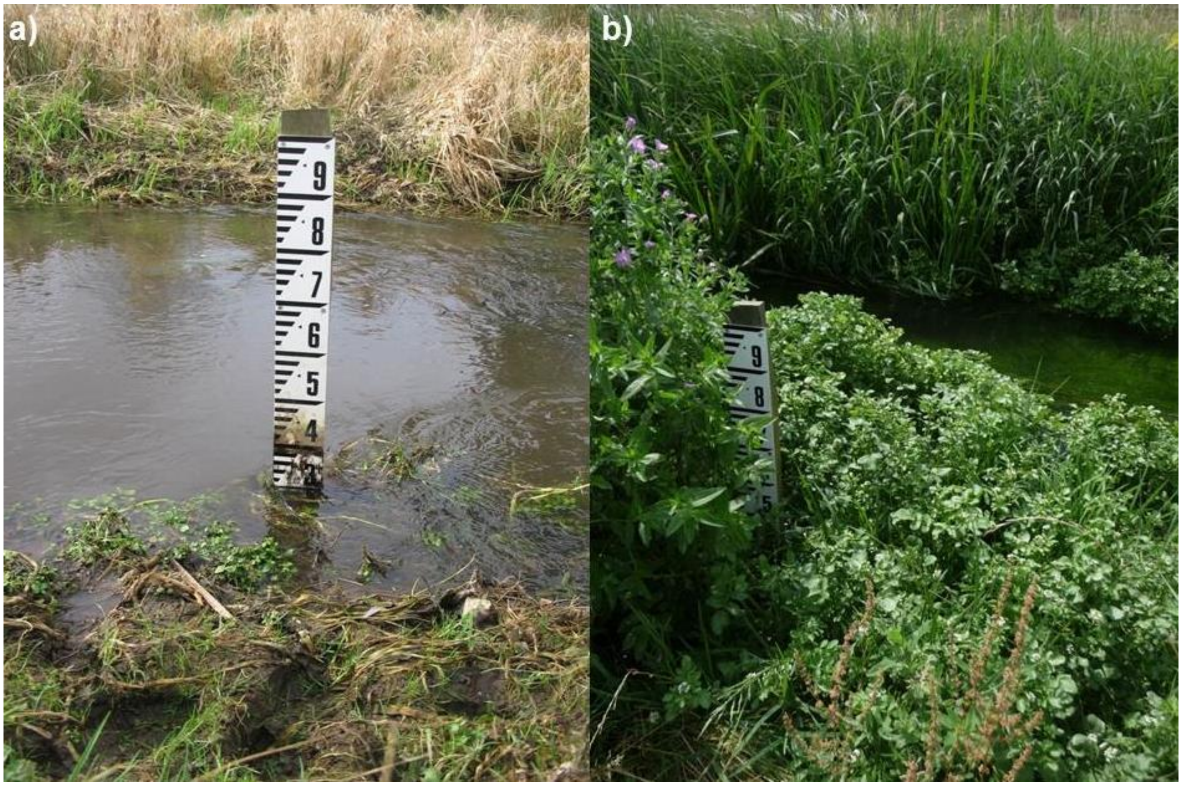


Figure 4.16: Photographs reflecting the different in-river macrophyte abundance during the winter (a) and summer (b) months at the river stage located at the DS transect. Photograph (a) was taken on 3rd March 2009, prior to the restoration. Photograph (b) was taken on 25th July 2010 following embankment removal.

Table 4.1. Corresponding Environment Agency river discharge and stage, and downstream river stage for the low (a) and high (b) levels of in-river macrophyte abundance shown in Figure 4.17.

Date	EA River Discharge (m^3 s^{-1})	EA river stage (m)	Downstream transect river stage (m)
3rd March 2009 (a)	0.325	0.212	0.30
25th July 2010 (b)	0.143	0.154	0.39

Indeed, macrophyte abundance on the River Glaven was observed to be substantially lower during high-flow summers (mean flow: 0.38 $m^3 s^{-1}$) (Figure 4.17a), compared with low flow summers (mean flow: 0.19 $m^3 s^{-1}$) (Figure 4.17b), probably due to devegetation at higher flows. As spring/summer river stage was only measured during periods of low river flows (0.15 – 0.24 $m^3 s^{-1}$) (Figure 4.15), and the effects of macrophyte abundance is likely reduced during high flow conditions, the spring/summer regression equation was not used to determine bankfull capacity. Instead, bankfull capacity was calculated with the autumn/winter regression equation (Figure 4.15), which encompasses river stage measurements for a range of wider flows (0.25 – 1.9 $m^3 s^{-1}$) during low macrophyte abundance.

Figure 4.17: Photographs reflecting the different in-river macrophyte abundance during a wet (a) and dry (b) summer. Photograph (a) was taken on 17th July 2007, prior to the restoration. Photograph (b) was taken on 25th July 2010, following embankment removal.

Bankfull discharge was also determined by a semi-empirical method using the Manning's equation for uniform flow as:

$$Q = \frac{1}{n} A R^{2/3} S^{1/3} \tag{4.11}$$

where Q is discharge ($m^3\ s^{-1}$), n is Manning's roughness coefficient, A is bankfull cross-sectional area (m^2), R is hydraulic radius (m), and S is water surface slope (Gordon *et al*. 2004). The value of Manning's roughness coefficient (n value), describes the channel bed resistance to flow, and was estimated using the Rosgen (1996, 2007) stream classification method. This required the following river morphology input parameters: bankfull width-to-depth ratio, entrenchment ratio (flood-prone width at 2 × bankfull depth/bankfull width), water surface slope and channel sinuosity, which were obtained from river channel cross-sections, surveyed using dGPS, and measurements of river channel length in Arc-GIS.

The River Glaven at Hunworth has a low gradient of <0.004 (Figure 4.18), a moderate sinuousy of 1.4 (Table 4.2), and low width to depth ratio of <9 (Table 4.2). Collectively, these characteristics most closely fit a Rosgen type C stream, which is defined as slightly entrenched, with a riffle-pool bed form and channel width > depth (Table 4.3). A type E stream may also be applicable, but is typically assigned to streams with high sinuosity, which does not fit the River Glaven. The removal of the river embankments and the associated reduction in river incision increased entrenchment ratio and increased the width of the river relative to depth; however these changes did not alter the classification of the stream within the Rosgen scheme. Bed slope and sinuosity were unchanged.

Table 4.2: Stream morphology of the River Glaven. Entrenchment ratio, width to depth ratio, sinuosity and bedslope parameters required for Rosgen's stream classification system.

	Pre restoration	**Post restoration**	**Applicable stream type**
Entrenchment ratio *<1.4 = entrenched* *1.4-2.2 = moderate* *>2.2 = slightly entrenched*	2.79	15.51	C, E
Width to depth ratio *<12 = low* *12-40 = moderate* *>40 = high*	7.32	8.80	E
Sinuosity *<1.2 = low* *1.2-1.5 = moderate* *>1.5 = high*	1.40	1.40	C
Bed slope	0.00353	0.00353	C, E

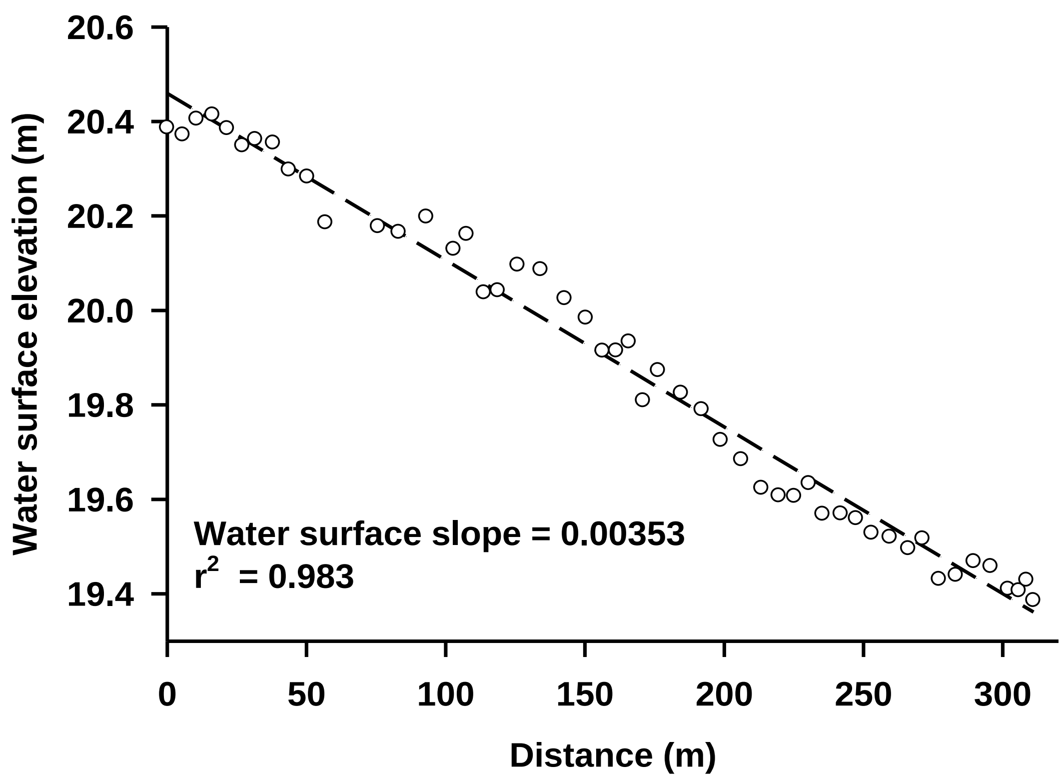


Figure 4.18: Water surface elevation of the River Glaven at Hunworth.

Table 4.3: General descriptions and physical criteria for Rosgen's stream classification system. Modified from Rosgen (2007).

Stream type	**Description**	**Entrench -ment ratio**	**W/D ratio**	**Sinuosity**	**Slope**
Aa+	Very high relief. Deeply entrenched, debris transport, torrent streams. Vertical steps with deep scour pools; waterfalls	<1.4	<12	1.0 – 1.1	>0.10
A	High relief. Entrenched, cascading, step-pool streams. High energy/ debris transport associated with depositional soils. Very stable if bedrock or boulder-dominated channel. Frequently spaced, deep pools in associated step-pool bed morphology	<1.5	<13	1.0 – 1.2	0.04 – 0.10
B	Moderate relief. Moderately entrenched, moderate gradient, riffle dominated channel with infrequently spaced pools. Rapids predominate with scour pools. Very stable plan and profile. Stable banks.	1.4 – 2.2	>12	>1.2	0.02 – 0.039
C	**Broad valleys with terraces, in association with floodplains, alluvial soils. Low gradient, meandering, point bar, riffle/pool, alluvial channels with broad, well-defined floodplains.**	**>2.2**	**>12**	**>1.2**	**<0.02**
D	Broad valleys. Braided channel with longitudinal and transverse bars. Very wide channel with eroding banks.	n/a	>40	n/a	<0.04
DA	Broad, low-gradient valleys with fine alluvium and/or lacustrine soils. Anastomizing (multiple channels) narrow and deep with extensive, well-vegetated floodplains and associated wetlands. Very gentle relief with highly variable sinuosities and width-to-depth ratios. Very stable streambanks.	>2.2	Highly variable	Highly variable	<.005
E	Broad valley/meadows. Low gradient, meandering riffle/pool stream with low width-to-depth ratio and little deposition. Very efficient and stable. High meander width ratio.	>2.2	<12	>1.5	<0.02
F	Entrenched meandering riffle/pool channel on low gradients with high width-to-depth ratio.	<1.4	>12	>1.2	<0.02
G	Entrenched gully step-pool and low width-to-depth ratio on moderate gradients	<1.4	<12	>1.2	0.02 – 0.039

The river substrate (percentage composition of sand, gravel, cobble) was surveyed at 53 regular points along the river on 03/03/2009. The river bed at the site was composed of a mixed sand and gravel bed, averaging 60% sand, 35% gravel, and 5% cobble (Figure 4.19). The substrate composition was fairly variable along the study reach, with sandbar and gravelbar sections. The river can be further defined within the Rosgen scheme as either a C4 (gravel bed) or C5 (sand bed) stream. Figure 4.20 shows the typical Manning's roughness coefficients for different stream types. Using the values for small streams with a vegetation influence, a C4 gravel bed stream and C5 sand bed stream correspond with average bankfull manning's n values of approximately 0.04 and 0.056, respectively (Figure 4.19) (Rosgen 2007). A range of bankfull discharge estimates were calculated by holding all other terms constant and varying Manning's n only between the two extremes of 0.04 and 0.056. The Manning's bankfull estimates were subsequently compared with the bankfull stage–discharge estimates.

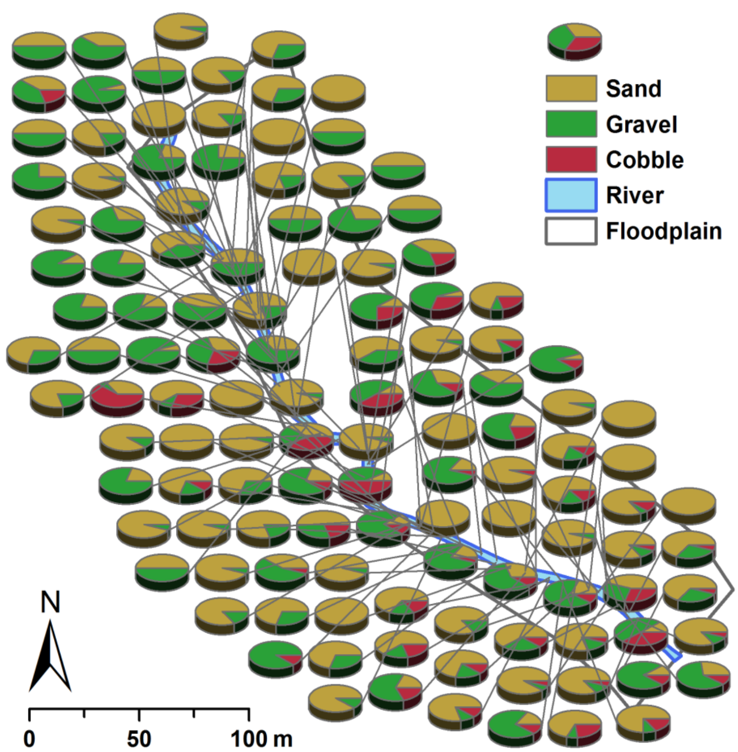


Figure 4.19: River bed substrate composition of the River Glaven at Hunworth (n=154 along 53 points).

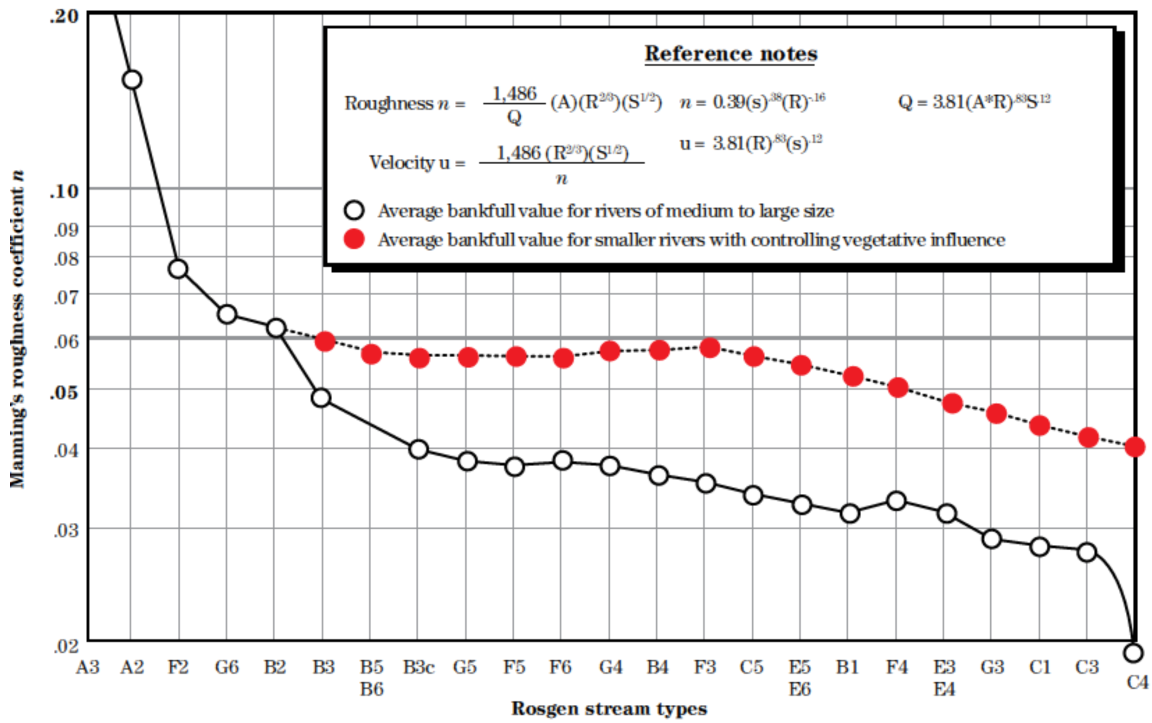


Figure 4.20: Bankfull roughness coefficients by stream type for 140 streams in the USA and New Zealand (Rosgen 2007).

4.4.2 Recurrence intervals

Recurrence interval (return period in years) of bankfull discharge was computed using flood peaks over threshold (POT) data, which were extracted from discharge data measured at the on-site EA gauging station from 2001 to 2010. Flood POT was used rather than annual maximum floods because it provides more information on flood regime and is more suitable for records of <13yrs (Institute of Hydrology 1999). Discharge data were arranged into water years beginning on 1 October, the threshold value was set at 0.6 $m^3\ s^{-1}$ so that, on average, five independent peaks per year were included in the series (Institute of Hydrology 1999, Cunderlik and Burn 2001). Aquatic macrophytes caused breaks in the river discharge record during some years; however, these years were included in analysis where: (a) the gaps did not constitute a major portion of the record, and (b) precipitation data available to confirm that large flood peaks were unlikely to have been missed. The recurrence interval for bankfull discharge was determined as:

$$Q_T = q_0 + \beta \ln \lambda + \beta \ln T \tag{4.12}$$

where Q_T is the magnitude of flood ($m^3\ s^{-1}$) with a return period of T years, q_0 = POT threshold discharge (=0.6 $m^3\ s^{-1}$), β is mean discharge (=1.01 $m^3\ s^{-1}$) minus threshold discharge, λ is the number of peaks (=37) divided by the number of full years of data (=8 years) (Wilson 1983).

4.6 Statistical analyses

4.6.1 Linear regression models and diagnostics tests

Simple linear regression was used to determine correlations between river stage and mean daily river discharge for summer and winter periods. To test whether the summer and winter regression functions differed, a full and reduced F test comparison was used. The full model was a regression of river discharge (y) on river stage (X_1), the indicator variable for winter and summer (X_2), and the interaction term for differences in slope (X_1*X_2). The reduced model was a regression of river discharge on river stage. Differences in hyporheic chemistry along well transects were tested using one-way Analysis of Variance (ANOVA). If a significant difference was found ($p<0.05$), a Tukey's test was used to determine which wells were significantly different ($p<0.05$). To meet the required assumptions for statistical inference, plots of residuals versus X were used to assess linearity, constant variance, and identify any outliers. The Brown-Forsythe t-test was used to assess constancy of variance. Boxplots of residuals and normal probability plots of residuals were used to identify nonnormality of the error terms. Significance tests for normality using the Pearson Correlation Coefficient were also completed to test that the normality assumption was reasonable. Regressions and diagnostics were computed using SAS 9.2 statistical software for Windows (SAS Institute Inc., North Carolina), and Sigma Plot 10.0 (Systat Software Inc., London).

Chapter 5: The hydrology and biogeochemistry of the embanked and restored floodplain meadow

5.1. Introduction

This chapter presents the hydrological and biogeochemical data collected prior to and after the removal of river embankments. These data are used to characterise river-floodplain hydrology and assess the initial changes in the hydrological regime to answer the first set of research questions outlined in Chapter 1:

(i) What is the hydrological and biogeochemical regime of an embanked-river floodplain?

(ii) What is the measured response to embankment removal?

5.2. Results

5.2.1 River embankments

Prior to the restoration, the River Glaven was constrained by embankments that ranged from 0.4 to 1.1 m (mean = 0.6 m) above the meadow surface (Figures 5.1 and 5.2). The embankments are evident in Figure 5.2 as a band of higher elevation along the river compared with the adjacent floodplain. The width and depth of the channel was fairly uniform along its length (Figures 5.1 and 5.2), indicative of a channelized and deepened river. Prior to the restoration, channel depth (river bed to bank top) averaged 1.4 ± 0.1 m along the study reach. After the removal of the river embankments, channel depth was reduced by approximately 44%, averaging 0.8 ± 0.1 m along the study reach (Figure 5.1), and riverside elevation was aligned with that of the floodplain (Figure 5.1b). Embankment removal reduced channel cross-sectional area by approximately 51%, from a mean of 6.5 ± 0.6 m^2 to 3.2 ±

0.7 m^2 along the study reach. Surface elevation on Hunworth Meadow is below river bankfull elevation, and decreases with distance from the river, with the exception of a few local highs (Figure 5.2 a and b). A section along the embankment that was not restored due to the presence of water vole burrows (see Section 3.8) is evident in Figure 5.2c as the section of zero elevation change (in white) on the main bend in the river channel. A slight increase (ca. 0.1 – 0.2 m) in river bed elevation at some locations along the reach occurred after the embankment removal (Figure 5.1). Although the restoration work did not involve mechanical work in the channel, some sediment fell into the river during the excavation of the river banks, which is the most probable cause of modified river bed geomorphology observed after the restoration (see also Figure 5.6).

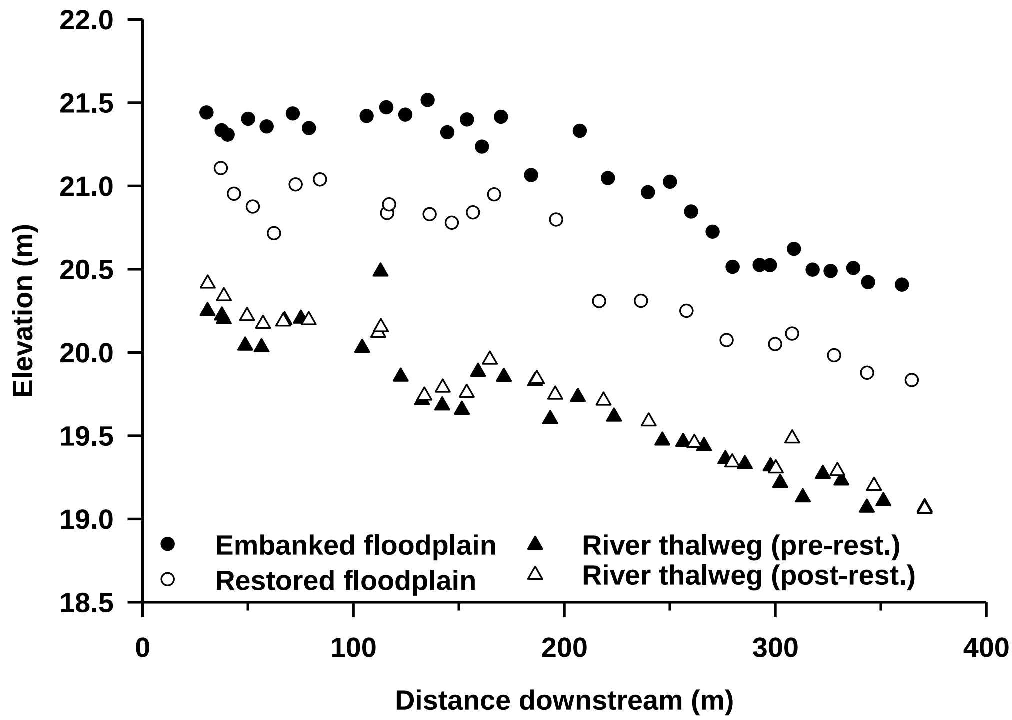


Figure 5.1: Comparison of floodplain elevation adjacent to the river channel and thalweg (lowest point along the river bed) along the study reach before (embanked) and after (restored) embankment removal. The embanked data represent the highest point on the embankments, and the restored data represent the closest corresponding sample point after embankment removal.

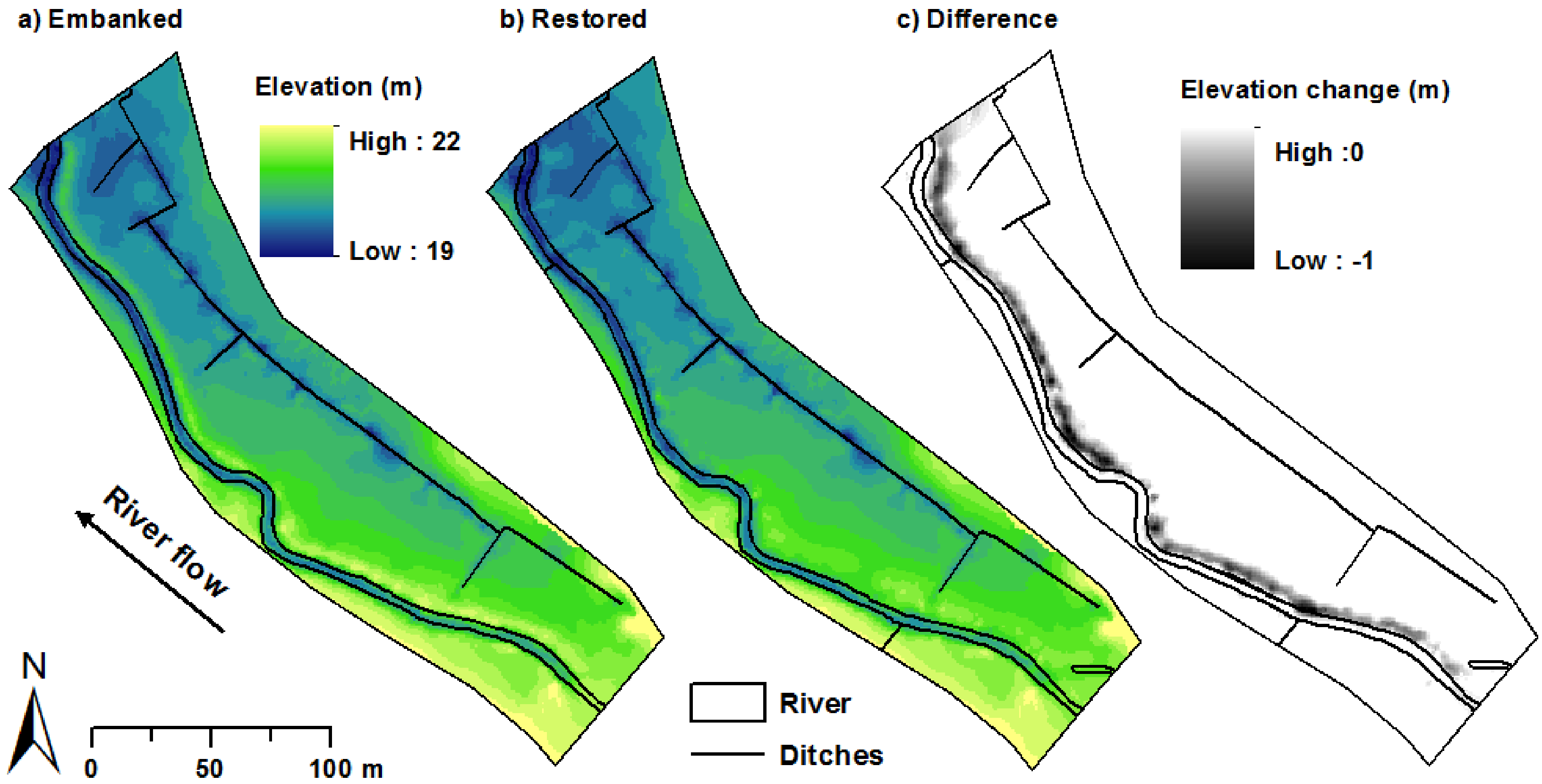


Figure 5.2: Elevation of Hunworth Meadow study site showing (a) before and (b) after embankment removal, and (c) the difference in elevation. The DEMs were created using dGPS survey data collected in (a) June 2008 and (b) July 2009.

5.2.2 Climate and hydrology

The River Glaven base flow index (BFI) averaged 0.83 ± 0.04 (range = 0.75 – 0.88) at the Hunworth gauging station, and Q95 (expressed as % of mean annual flow) was 51%, indicating high groundwater contributions to discharge (Figure 5.3; Table 5.1). This is typical of permeable chalk aquifers, which average 0.83 BFI (range = 0.53 − 0.99) and approximately 41% Q95 (permeable catchments >30%, impermeable <15%) (Gustard *et al*. 1992; Sear *et al*. 1999). Similarly, the flat nature of the river flow duration curve from Q10-Q95, i.e. the small difference between the low flow (Q95) and high flow (Q10) parameters, is indicative of a steady flow regime (Figure 5.4a). River flow, in general, followed the characteristic annual hydrograph of a chalk stream, with increased discharge over the winter from early December until March, when river levels begin to decline. However, some of the highest recorded river flows occurred during the summer (Figure 5.4b).

Precipitation had a distinct effect on river discharge, with peaks in river discharge coinciding with precipitation events (Figure 5.5a). River response to precipitation was rapid, typically within one day, although some events characterised by low intensity rainfall, prolonged over a few days, resulted in a muted stream flow response (Figure 5.5a). Shallow groundwater elevation within the Hunworth Meadow responded rapidly (<1 day) to changing river levels, resulting in prolonged saturation of surface soils during winter/spring (November – April), with periodic saturation in summer/autumn (May – October) (Figure 5.5b).

Year	2002	2003	2004	2005	2006	2007	2008	2009	2010
Base flow Index (BFI)	0.88	0.81	0.79	0.83	0.85	0.75	0.83	0.87	0.84

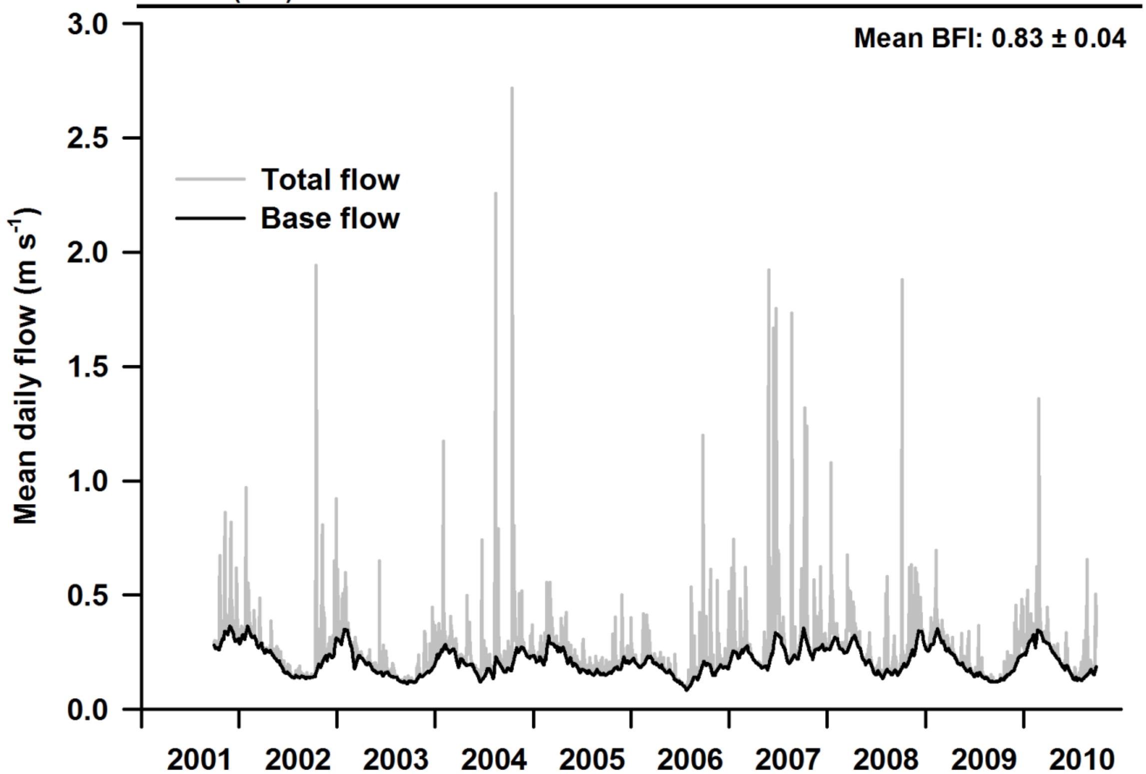


Figure 5.3: Mean daily total river flow and base flow from 2002 to 2010. The table inset shows base flow index (BFI) values calculated for each full year of record using a computerised base flow separation program (Wahl and Wahl 2007).

Table 5.1: Mean annual river flow (range), Q10, Q95, and Q95 (as % of mean annual flow) using continuous river discharge data from 2002 − 2010.

Mean annual river flow (m^3 s^{-1}) (range)	Q10 (m^3 s^{-1})	Q95 (m^3 s^{-1})	Q95 (as % mean annual flow)
0.26 (0.08 – 2.72)	0.37	0.13	51.39

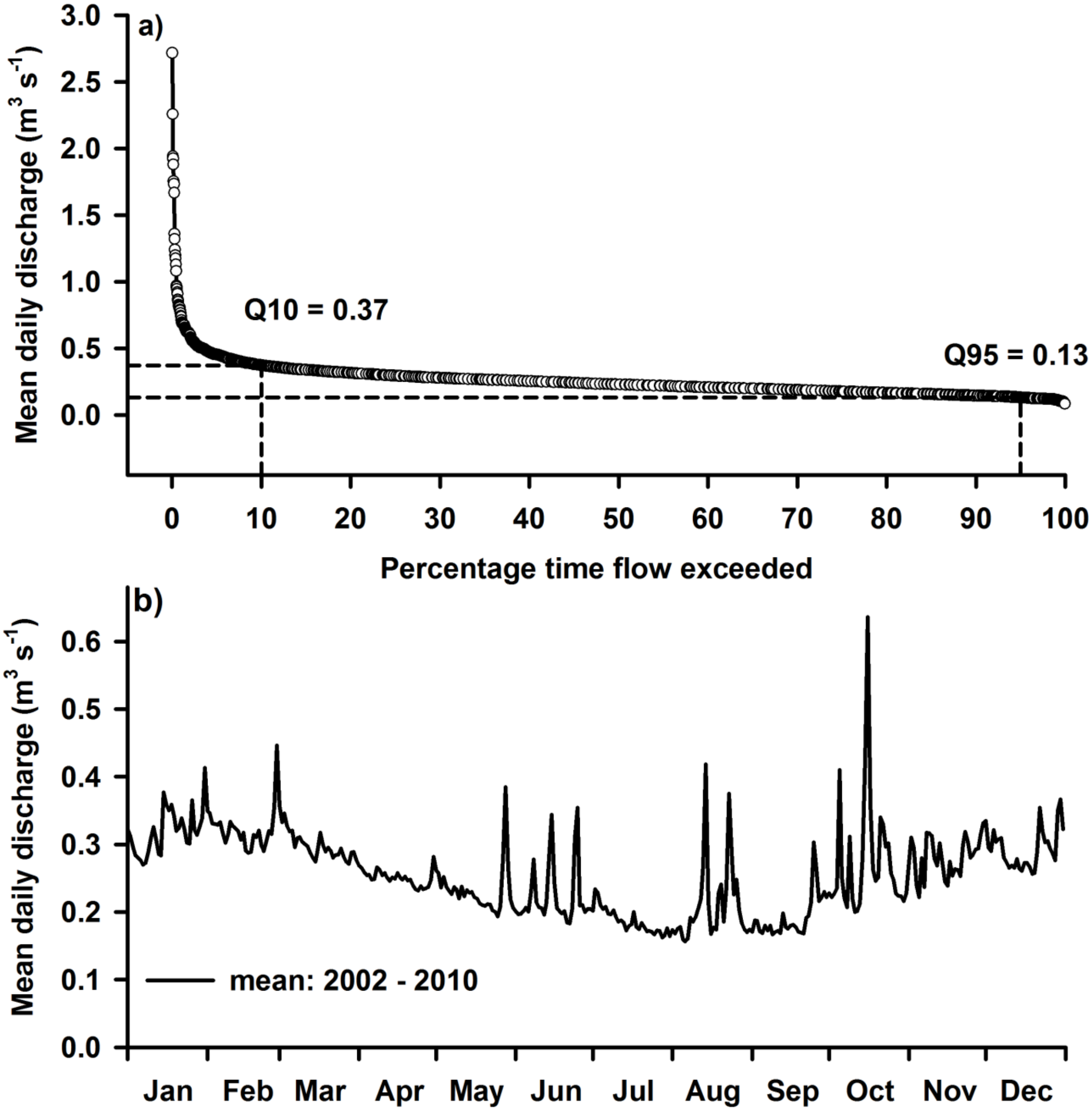


Figure 5.4. Flow duration curve (a), and mean daily river discharge (b) for the River Glaven from October 2001 – September 2010.

The period of observation prior to the restoration (2007 – 2008) was characterised by cooler, wetter summers (Table 5.2). Summer (June − September) precipitation in 2007 and 2008 totalled 393 and 281 mm, respectively and exceeded potential evapotranspiration, which totalled 248 and 262 mm, respectively (Table 5.2). Average total summer precipitation for East Anglia from 1985-2015 is169 mm (Met Office 2016), hence summer precipitation in 2007 and 2008 was approximately double the long-term average. Summer precipitation in 2007 was highest of the four-year study period, which resulted in a shallow water table within 0.10 m of the ground surface for much of the growing season between March and September (Figure 5.5b). In comparison, summer water table elevations in 2008 were typically within 0.10 m of the ground surface between March and May, due to a period of prolonged precipitation and high river discharge, and were slightly lower during the summer, averaging 0.33 m below the ground surface (Figure 5.5a and b) Contrary to 2007 and 2008, the summers after the restoration were warm and dry, with total precipitation of 178 and 261 mm in 2009 and 2010, respectively. Furthermore, potential evapotranspiration was 10 − 19% higher, and exceeded total precipitation (Table 5.2). Collectively, this led to average summer water table depths of 0.57 and 0.59 m below the ground surface in 2009 and 2010, respectively (Figure 5.5b). Summer river hydrographs were very similar in 2008, 2009 and 2010, all with a steady decline in discharge from May to August (Figure 5.5a), and mean summer discharge of approx. 0.2 $m^3 s^{-1}$ (Table 5.2). However, whereas summer water table elevation differed by less than 0.4 m on average between 2009 and 2010, water table elevation was on average 0.13 – 0.17 m higher in 2008 compared to 2009 and 2010 (Figure 5.5b).

Table 5.2: Summer (June − September) mean (± 95% confidence interval) air temperature, total precipitation, total potential evapotranspiration, and mean annual river discharge (±95% confidence interval)

Year	**Temperature (°C)**	**Precipitation (mm)**	**Evapotranspiration (mm)**	**River discharge ($m^3 s^{-1}$)**
2007	15.1 ± 0.3	393	248	0.38 ± 0.05
2008	15.3 ± 0.5	281	262	0.19 ± 0.01
2009	15.6 ± 0.4	178	286	0.24 ± 0.01*
2010	15.7 ± 0.5	261	295	0.19 ± 0.01

Note: ∗2009 mean river discharge is for 303 days.

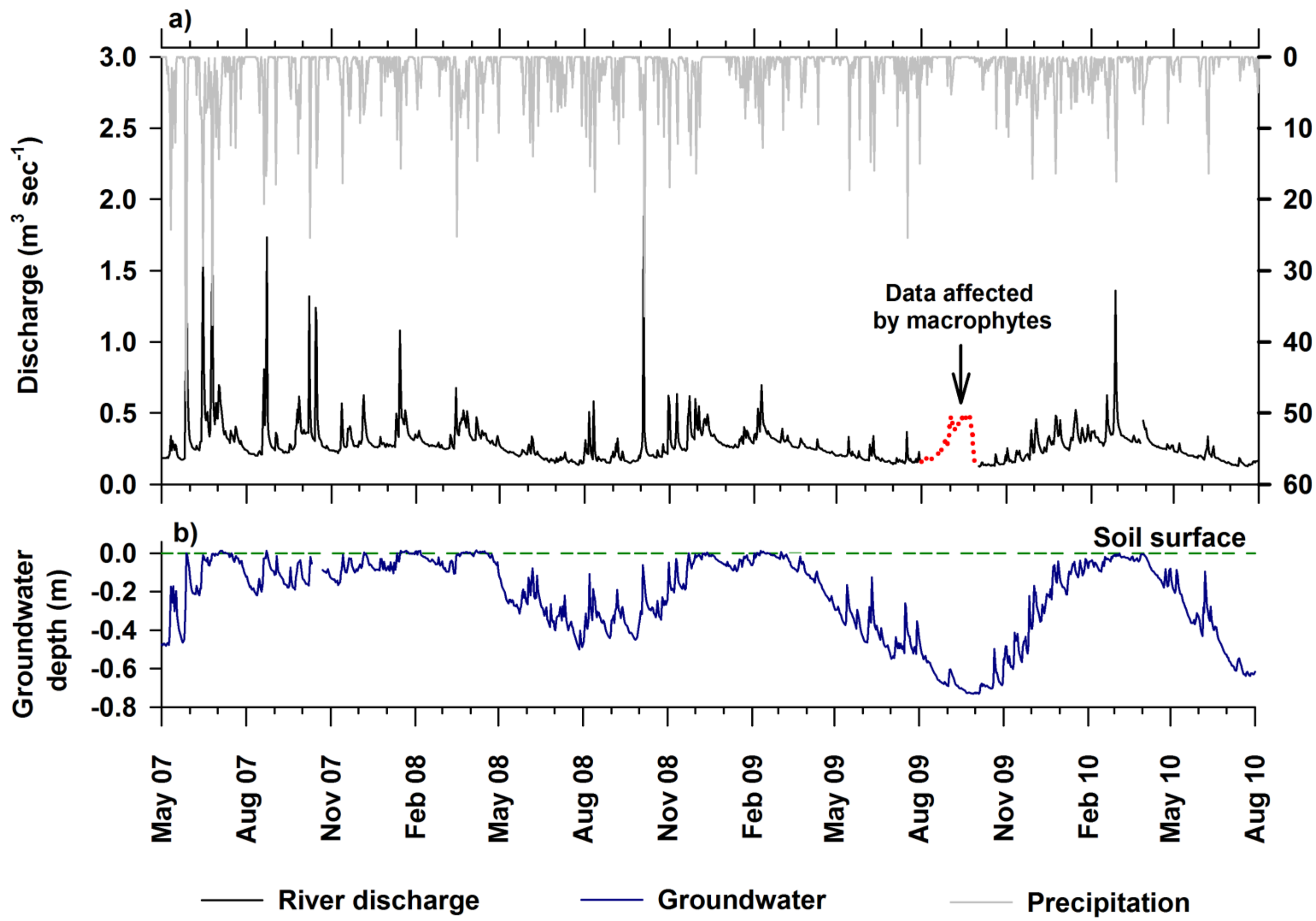


Figure 5.5: Temporal variation in (a) mean daily river discharge and total daily precipitation, and (b) representative mean daily groundwater depth (downstream Well 1.6) for the four study years (2007 – 2010). The river discharge data affected by aquatic macrophyte growth (August – October 2009) are highlighted in (a).

Figure 5.6 shows groundwater levels on the floodplain during three different river-flow conditions; high flow in autumn (October 2008), low flow in spring (March 2009), and low flow in summer (June 2008). The hydraulic gradient across the floodplain is relatively flat, averaging 6.1 – 9.4 mm m^{-1} (Table 5.3). Groundwater levels at the upstream transect indicate complex movements of groundwater, with shifts in hydraulic gradient observed between periods of different river flow. In general, groundwater levels at the upstream and midstream transect decreased from the river towards the ditch, which was located at the lowest point of the floodplain (Figure 5.6a and b). This was most apparent at the midstream well transect, where the river thalweg (lowest point along the river bed) was above the ditch thalweg (Figure 5.6b). Conversely, at the downstream transect, where the topography flattens, groundwater levels tended to flow from the base of the hillslope towards the river (Figure 5.6c).

Groundwater levels measured in wells next to the river (3.1 and 1.1) were not always lower than river stage, indicating temporal changes in the hydraulic gradient (Figure 5.3a and c). During peak discharge conditions, river stage was above the water table in the floodplain and flow was directed away from the river and into the floodplain, creating a bank storage ridge (Figures 5.6 (c) and 5.13). A streamward hydraulic gradient was re-established after the flood peak had passed (see Figure 5.13c). During low river stage in winter, groundwater levels on the floodplain were above that of the river (Figure 5.6a and b). Conversely, during dry summers, when groundwater levels were typically low in the soil profile, river stage was often slightly above groundwater levels (Figure 5.6).

Table 5.3: Hydraulic gradient, hydraulic conductivity and groundwater flow rate (mean ± 95 % confidence interval) for the well transects.

Transect	**Hydraulic gradient ($\Delta h/\Delta l$)**	**K(cm day^{-1})**	**Groundwater flow rate (cm day^{-1})**
Upstream	0.0094 ± 0.0033	17.87 ± 28.17	0.29 ± 0.06
Midstream	0.0067 ± 0.0042	3.04	0.03 ± 0.02
Downstream	0.0061 ± 0.0057	14.23 ± 20.93	0.13 ± 0.15

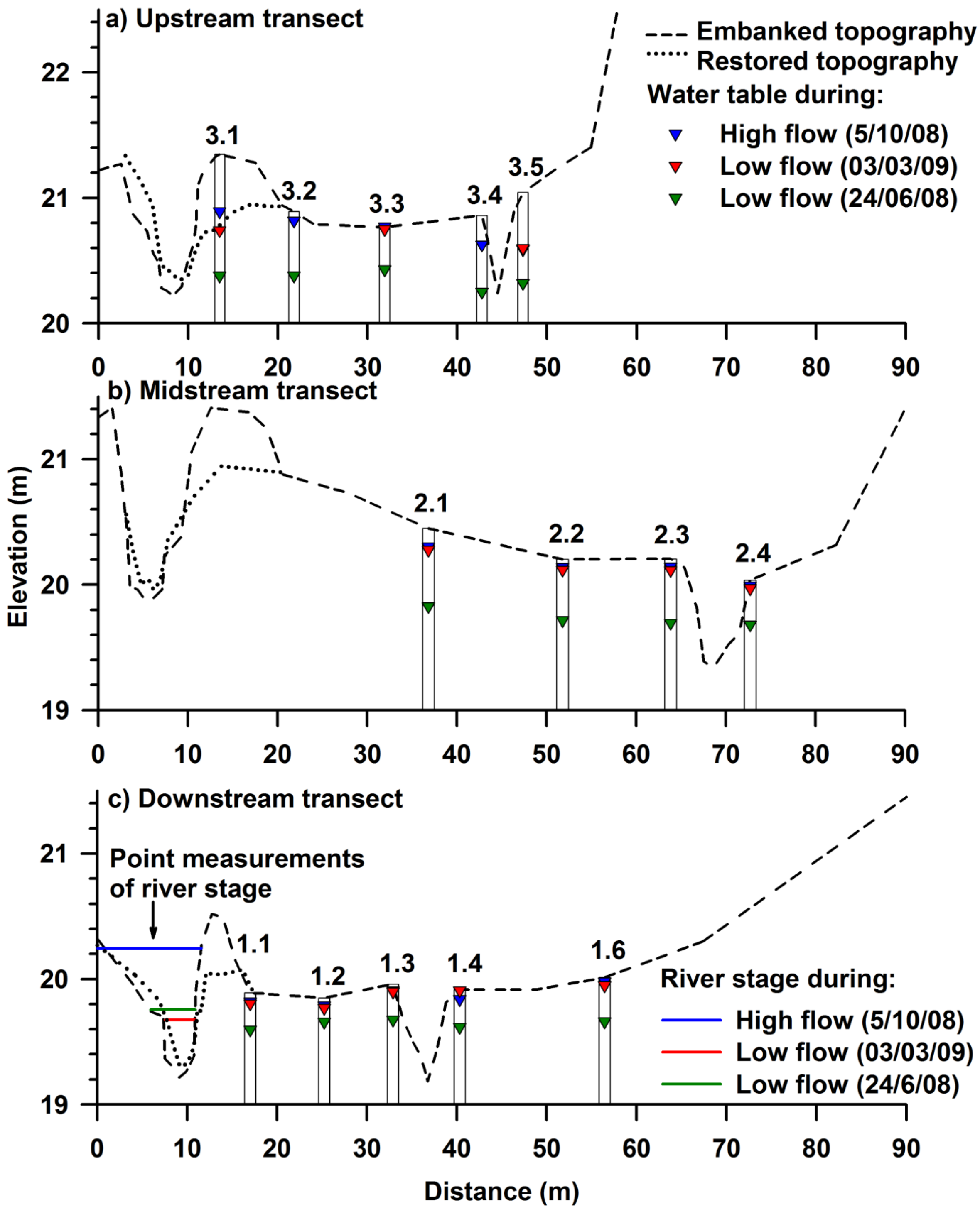


Figure 5.6: Cross-sections of the meadow and river channel before (embanked) and after (restored) embankment removal. Typical mean well water levels are shown along the three well transects in relation to low and high river stage. River stage measurements at the downstream transect are single point measurements.

5.2.3 Soil physical and chemical properties

Hunworth Meadow soils are predominantly composed of sand and silt, which collectively composed 76 – 97% of the sample volume. Soil texture on the meadow is predominantly sandy loam, and does not differ substantially from the river embankments ($p<0.05$) (Figure 5.7). Soil texture in the ponded area of the meadow however was more variable, and had higher clay content (Figure 5.7). Soils were high in calcium concentration, with an average of 1.7 – 2.7 mg Ca^{2+} g^{-1} (Table 5.4), and moderately fertile, with Olsen P concentrations of 6.2 – 9.5 mg P kg^{-1} on average, and mean plant available potassium concentrations of 1.0 – 2.8 mg K^{+} g^{-1} (Table 5.4). Plant available ammonium concentration (average: 12.8 – 32.3 mg NH_4^{+}-N kg^{-1}) was 11 – 26 times greater than nitrate concentration (average: 0.5 – 3.0 mg NO_3^{-}-N kg^{-1}) (Table 5.4).

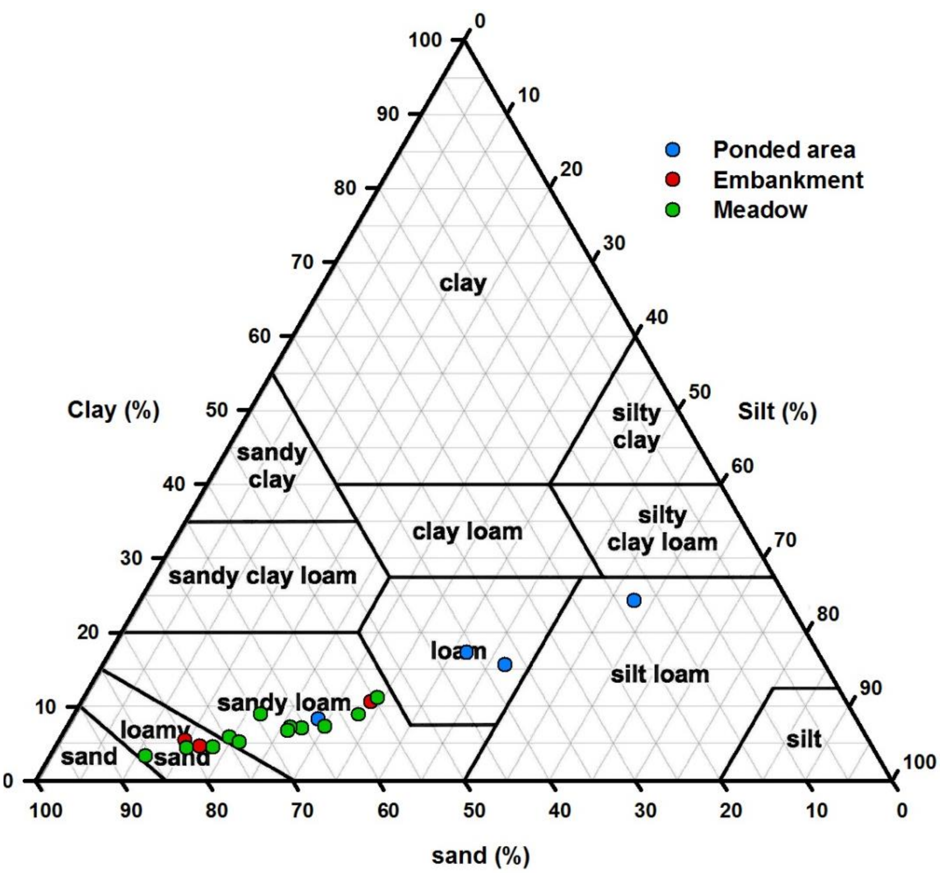


Figure 5.7: Textural triangle of soils sampled along the three well transects.

Table 5.4: Soil chemistry of Hunworth Meadow along the three well transects (mean ± 95% confidence interval). Soils were sampled on 29th April 2008.

	Upstream Transect	**Midstream Transect**	**Downstream Transect**
pH	6.34 ± 0.45	6.49 ± 0.74	6.51 ± 0.13
Organic matter content (%)	12.16 ± 2.62	10.98 ± 2.14	15.38 ± 10.37
Bulk density	0.71 ± 0.07	0.75 ± 0.00	0.50 ± 0.23
Ca^{2+} (mg g dry soil^{-1})	2.70 ± 0.73	1.87 ± 1.08	1.66 ± 0.93
Na^{+} (mg g dry soil^{-1})	0.12 ± 0.09	0.08 ± 0.04	0.04 ± 0.02
Mg^{+} (mg g dry soil^{-1})	0.09 ± 0.06	0.14 ± 0.09	0.10 ± 0.07
K^{+} (mg g dry soil^{-1})	0.96 ± 0.39	2.81 ± 1.36	1.25 ± 1.19
Total iron (mg kg dry soil^{-1})	16.50 ± 1.91	38.84 ± 54.78	66.90 ± 69.95
Al^{3+} (mg kg dry soil^{-1})	6.49 ± 2.02	9.99 ± 8.94	6.14 ± 0.85
NH_4^{+} (mg N kg dry soil^{-1})	25.31 ± 9.99	12.81 ± 16.48	32.33 ± 13.21
NO_3^{-} (mg N kg dry soil^{-1})	2.96 ± 3.41	0.47 ± 0.92	2.22 ± 22.36
Olson P (mg P kg dry soil^{-1})	7.12 ± 4.97	6.18 ± 3.76	9.49 ± 3.58
TOC (mg kg dry soil^{-1})	0.57 ± 0.14	0.58 ± 0.50	0.92 ± 1.14
TN (%)	0.51 ± 0.19	0.44 ± 0.16	0.58 ± 0.44
TC (%)	5.89 ± 2.33	5.56 ± 2.38	6.94 ± 4.62

Topsoil (ca. 0-0.3m) on the meadow was slightly acidic (mean pH = 6.3 – 6.5), moderately organic (range: 13 – 35% organic matter content; Table 3.4), with the highest organic matter content recorded in the wetter parts of the meadow, and had low bulk density (mean: 0.69 g cm^{-3}) (Table 5.4). An organic-rich topsoil was present at the site to depths of approximately 0.3 m, below which the sandy loam topsoil was observed (Figure 5.8). Topsoil layers were underlain by alluvial gravels at depths from approximately 0.8 m (Figure 5.8).

The majority of hydraulic conductivity measurements, obtained through well and piezometer slug tests (see Section 4.3.4), ranged between 0.98 and 7.72 cm day^{-1} (Table 5.3). However, due to some higher conductivity measurements, the average was 16.3 ± 17.2 cm day^{-1} (Table 5.3; Figure 5.9). Hydraulic conductivity measurements were relatively low, with measured rates of the order expected for silt/loess soils (see Table 5.5). Such low values suggest slow hydrological exchange between the floodplain soils and river water, which is likely responsible for the poor on-site drainage and ponding of water that occurred at the downstream end of the meadow. Hydraulic conductivity (K) values tended to be higher in the unscreened piezometers, which measured K at a specific depth (see Section 4.3.4), in comparison to the slug tests conducted in the geotextile-screened monitoring wells, which gave an average soil K value for the entire length of the piezometer. However, no clear patterns were observed across the floodplain (Figure 5.9). The hydraulic conductivity of the underlying alluvial gravels on the floodplain could not be measured, but are likely to be substantially higher than the values measured in the top 2 m of sandy loam floodplain soil, for example values of 0.2 – 844 m day^{-1} are reported by Miller *et al.* (2014) for gravel-dominated alluvial floodplains (also see Table 5.3).

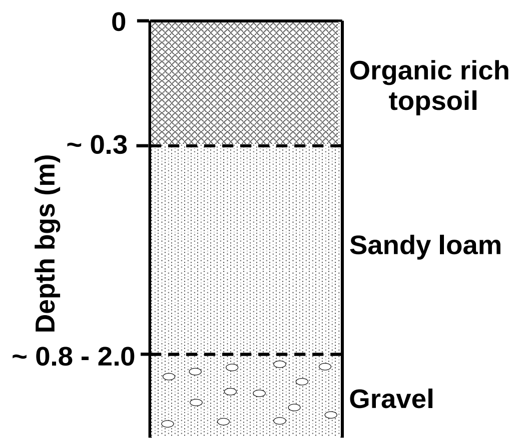


Figure 5.8: A portrayal of the soil horizons on Hunworth meadow. Note: bgs = below ground surface.

Table 5.5. Hydraulic conductivity values for various sediment, modified from Domenico and Schwartz (1998).

Substrate	Hydraulic conductivity (cm day^{-1})
Gravel	$2.6 \times 10^{3} – 2.6 \times 10^{5}$
Coarse sand	$7.8 \times 10^{0} – 5.2 \times 10^{4}$
Medium sand	$7.8 \times 10^{0} – 4.3 \times 10^{3}$
Fine sand	$1.7 \times 10^{0} – 1.7 \times 10^{3}$
Silt loess	$8.6 \times 10^{-3} – 1.7 \times 10^{2}$
Till	$8.6 \times 10^{-6} – 1.7 \times 10^{1}$
Clay	$8.6 \times 10^{-5} – 4.1 \times 10^{-2}$

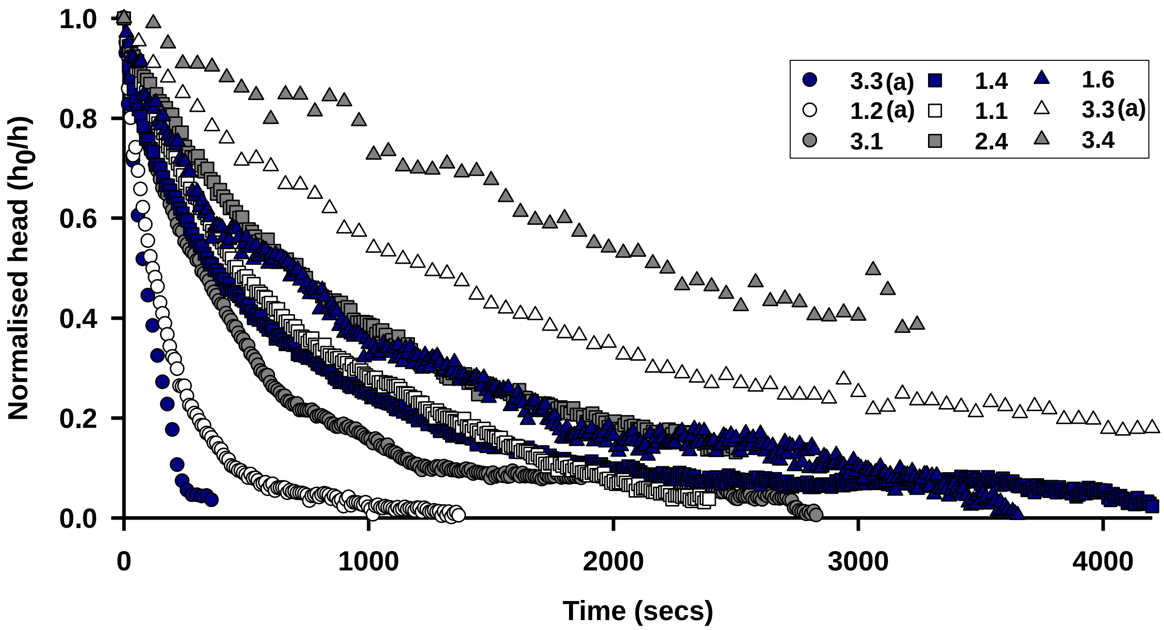


Figure 5.9: Decline in normalised head (head ratio) for all hydraulic conductivity tests that showed adequate log-linear behaviour. The unscreened peizometer intakes are labelled (a).

5.2.4 Bankfull capacity

Generally good agreement was obtained between the bankfull stage-discharge estimates and the Manning's equations estimates for pre and post- restoration bankfull discharge (Table 5.6). Bankfull capacity of the embanked river channel using the first of these methods was predicted to be 4.5 $m^3 s^{-1}$ (Table 5.6 and Figure 5.10). Flows above this threshold did not occur at all during the period of the record of the EA gauging station (2001 – 2010). Similarly, the lowest of the bankfull discharge estimates using the Manning's equation (4.09 $m^3 s^{-1}$) was not exceeded within the discharge record, confirming that overbank flow onto the floodplain was infrequent (at least >10-year intervals) (Figure 5.10).

Table 5.6: Bankfull height above ODN, bankfull river discharge from the river stage–discharge relationship, and calculated using Manning's equation, and bankfull recurrence interval (not estimated for the pre-restoration bankfull discharge due to the high uncertainty associated with extrapolating beyond the range of data).

Transect	**Bankfull elevation ODN (m)**	**Stage-discharge relation of bankfull discharge (m^3 s^{-1})**	**Manning's bankfull discharge (m^3 s^{-1})**
Pre-restoration	20.520	4.53	4.09 – 5.73
Post-restoration	20.096	1.15	1.33 – 1.87

Note: ODN: Ordnance Datum Newlyn.

Following the removal of the embankments, bankfull capacity (evaluated using the stage–discharge method) was reduced by 75% to 1.15 m^3 s^{-1}. River discharge exceeded this bankfull threshold during one short (1-day) high-flow event in February 2010 that averaged 1.36 m^3 s^{-1} (Figure 5.10). Historical river discharge data from before the restoration indicate that flows of this magnitude are regular. For instance, from 2001 to 2010, river discharge was above the 1.15 m^3 s^{-1} post-restoration bankfull capacity during a minimum of 14 high flow events, the majority of which (11 out of the 14 recorded) occurred during the summer and autumn (Figure 5.10). These were of short duration (1 day), with the exception of a high river discharge period in May – June 2007, successive high flows were above bankfull capacity within 10 – 18 days (Fig. 5.10). It is important to note that the river discharges presented in Figure 5.10 are mean values over 24 hours. Hence during precipitation events, peak discharge may be significantly greater than the mean daily discharge value, particularly on a small river such as the River Glaven, where flood events may last only a few hours.

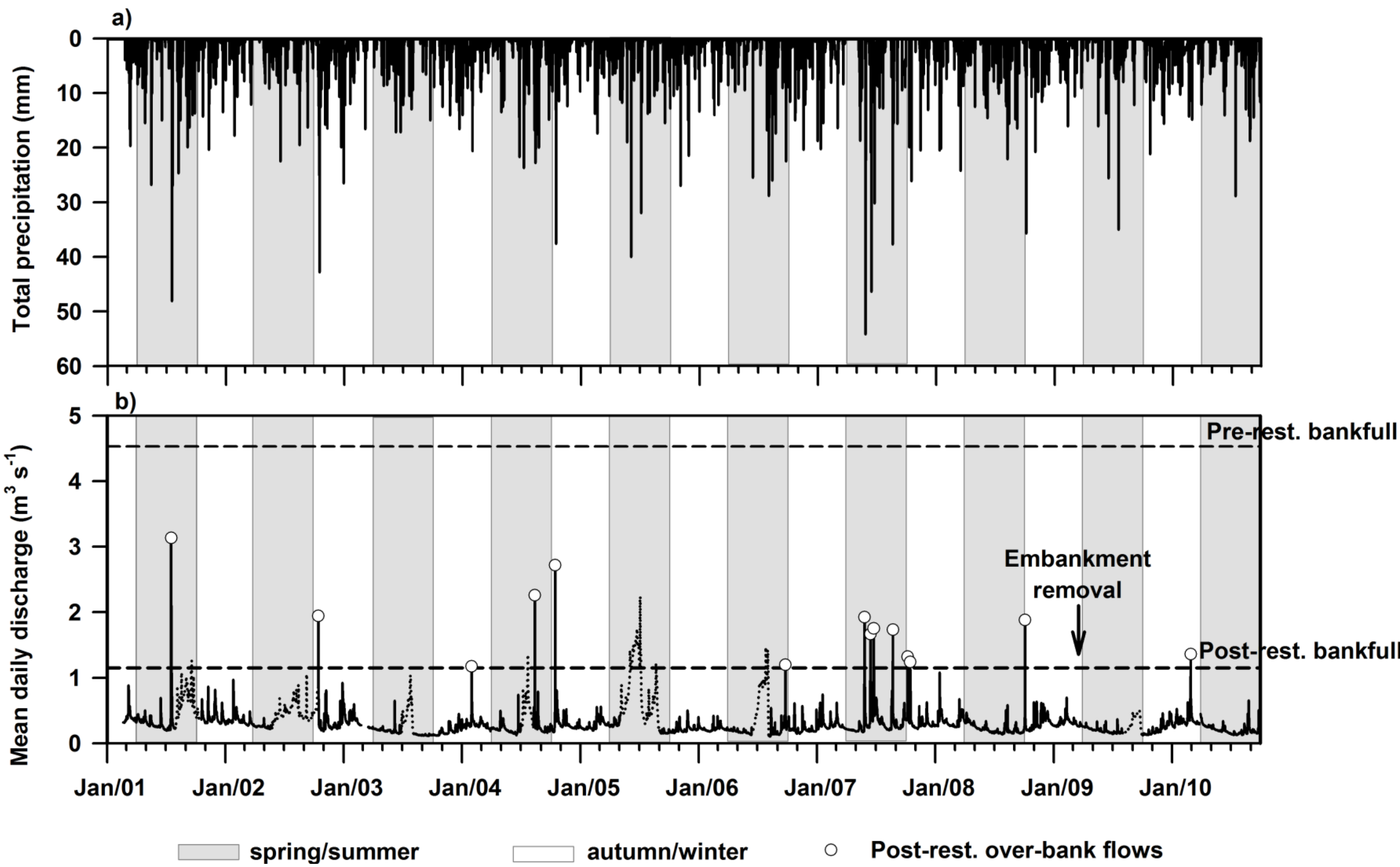


Figure 5.10: Time series of (a) total precipitation, and (b) mean daily river discharge from 2001 to 2010. Pre- and post-restoration bankfull capacity at the downstream river stage is shown in (b), above which inundation of the floodplain would have occurred. River discharge data affected by aquatic macrophyte growth have been highlighted with dotted lines and were not included in the analysis of bankfull discharge.

As bankfull flows prior to embankment removal (i.e. 4.5 $m^3 s^{-1}$; Table 5.6) did not occur during the period of record (Figure 5.10), recurrence interval was not estimated due to the high uncertainty associated with extrapolating beyond the range of data. It can be stated, however, that the return period for these flows was at least >14 years (Figure 5.11). Following embankment removal, the recurrence interval for overbank flows (i.e. 1.15 $m^3 s^{-1}$; Table 5.6) was substantially reduced to every 0.83 years (Figure 5.11).

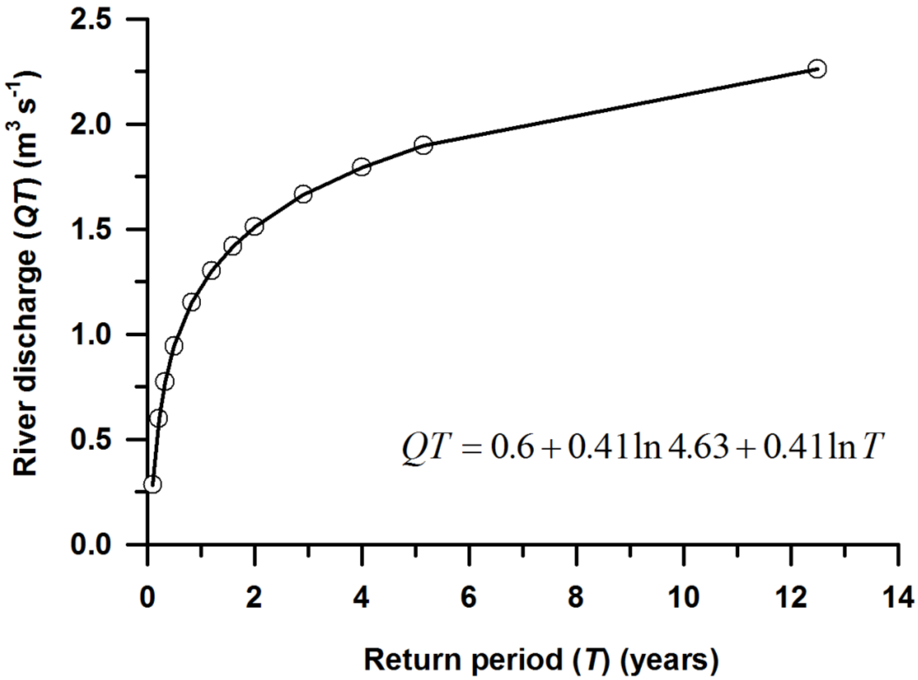


Figure 5.11: Recurrence interval (return period in years) of various river discharges on the River Glaven at Hunworth, computed using flood POT data.

A further indication that embankment removal will result in reconnection of the river and floodplain via overbank flows is given in Figure 5.12b which shows river stage in autumn 2008 above the restored bankfull elevation of 20.096 m (Table 5.6). This is also apparent in photographs taken of the same location on the River Glaven during high river flow (1.9 $m^3 s^{-1}$) in October 2008 before embankment removal (Figure 5.12a) and during low river flow (0.45 $m^3 s^{-1}$) in January 2010 after embankment removal (Figure 5.12b). The October 2008 stage board level is highlighted by the white circles in Figure 5.12, and indicates that river flows of a similar magnitude to October 2008 are likely to inundate the post-restoration floodplain.

Figure 5.12: Photographs of the same location on the River Glaven during differing river flows (a) before and (b) after embankment removal . Photograph (a) was during a high river stage (0.87 m), and (b) was during low river stage (0.34 m).

5.2.5 Groundwater response to embankment removal

After the removal of the embankments, groundwater levels at wells 3.1 and 1.1, which were located at the river–floodplain interface, generally remained higher than those within wells further from the river (Figure 5.13 a and c) for much of the post-removal period. This was particularly evident during the dry summer of 2009, possibly due to enhanced river water intrusion at the river-floodplain interface (Figure 5.13). In contrast to the observations close to the river, water levels in all other wells were approximately 0.1 – 0.6 m lower in the summers following the restoration (Figure 5.13a–c). This could be a result of the prevailing low summer precipitation levels in these years (see Section 2.3.2). Winter water table elevations remained unchanged following the restoration (Figure 5.13a–c), with water levels within 0.1 m of the soil surface in all wells. Furthermore, embankment removal lowered the soil elevation to within 0 – 0.5 m of the water table at Well 3.1. This resulted in increased saturation of surface soils on the restored river banks for much of the summer in 2009 and 2010 (winter 2009 data are not available) (Figure

5.13a), which is likely to have important effects on soil physicochemistry in this part of the floodplain.

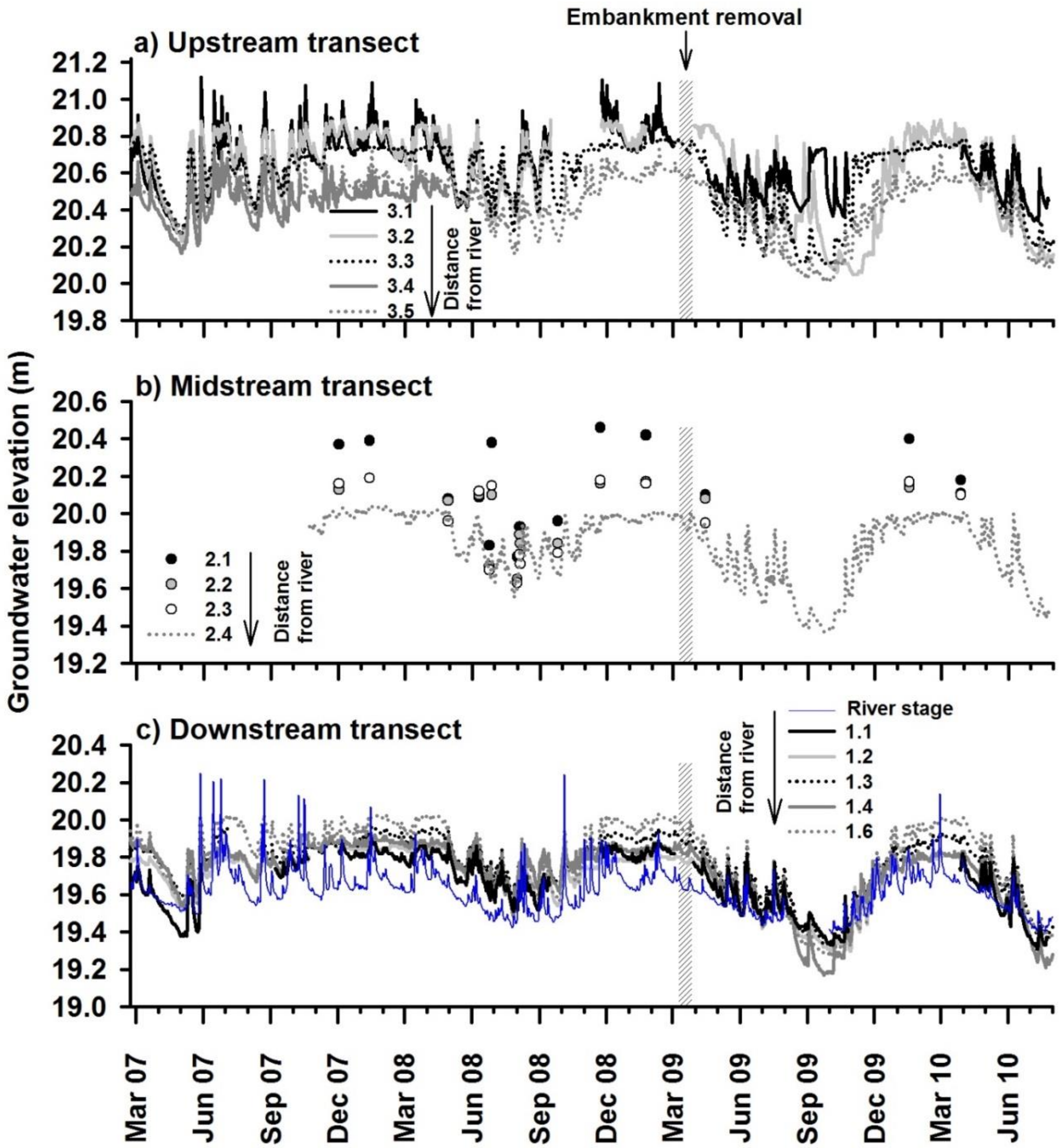


Figure 5.13: Temporal variation in mean daily groundwater height above Ordnance Datum Newlyn (ODN). Well 3.1 (a) was located on the river embankment; Wells 3.5 (a), 2.4 (b) and 1.6 (c) were located at the base of the hillslope (see Figure 4.2). Continuous measurements of groundwater height were not available for all of the wells at the midstream transect, hence hand measurements are also plotted (circles). River stage is plotted at the downstream transect.

5.2.6 Hydrological controls on chemistry

River water and regional groundwater (sampled in the chalk borehole) was slightly alkaline (mean pH: 7.3 and 7.4, respectively) (Table 5.7). In contrast, the floodplain well water was slightly acidic (mean pH: 6.5–6.7), potentially due to the organic-rich soils on the meadow (see Section 5.2.3). Groundwater in the floodplain wells and chalk borehole, and river water cation chemistry were dominated by calcium (Table 5.7). Cation concentrations followed the pattern of Ca>Na>Mg>K, whereas anion chemistry followed the pattern $Cl^->SO_4^->NO_3^-$ (Table 5.7).

The relative percentage composition of base cations in the floodplain well samples were mostly aligned in a linear fashion along the calcium and sodium axes (Figure 5.14) between the two main potential water sources: groundwater that was strongly dominated by calcium ions and river water that was characterised by higher levels of sodium ions (Table 5.7). The spatial variation in calcium and sodium cations in well water indicate that the direction of groundwater flow is not strictly from upslope through the floodplain to the stream but that floodplain well water is a mixture of hillslope and river water. However, differences in concentration between end members were quite small, particularly during baseflow conditions when a groundwater signature was evident in the river water. There is also an indication of evapotranspiration in well samples enriched in calcium. Hence, further analysis using an end-member mixing model was not considered appropriate (e.g. Genereux *et al*. 1993).

Conservative solutes such as sodium, calcium and chloride, varied spatially along each well transect. However, there was no obvious pattern with distance from the river (Figure 5.15a-c). One exception was the chemistry in Well 2.4 and Well 3.5, located at the base of the woodland hillslope (Figure 5.15a and b), which was markedly different from other wells on the meadow. Chloride and sodium concentrations in these two wells were, respectively, on average between 1.7 − 4.2 and 1.8 − 3.8 times greater than average concentrations in the floodplain wells (Figure 5.15a and c). This could be due to anthropogenic inputs; alternatively the source may be geological, possibly due to the weathering of hillwash at the base of the hillslope (Moorlock *et al*. 2002). Chloride and sodium chemistry of wells located close to the river (within 10 m) was closer in concentration to regional groundwater chemistry than that of the river, possibly due to limited hydrological exchange between river water and groundwater on the floodplain.

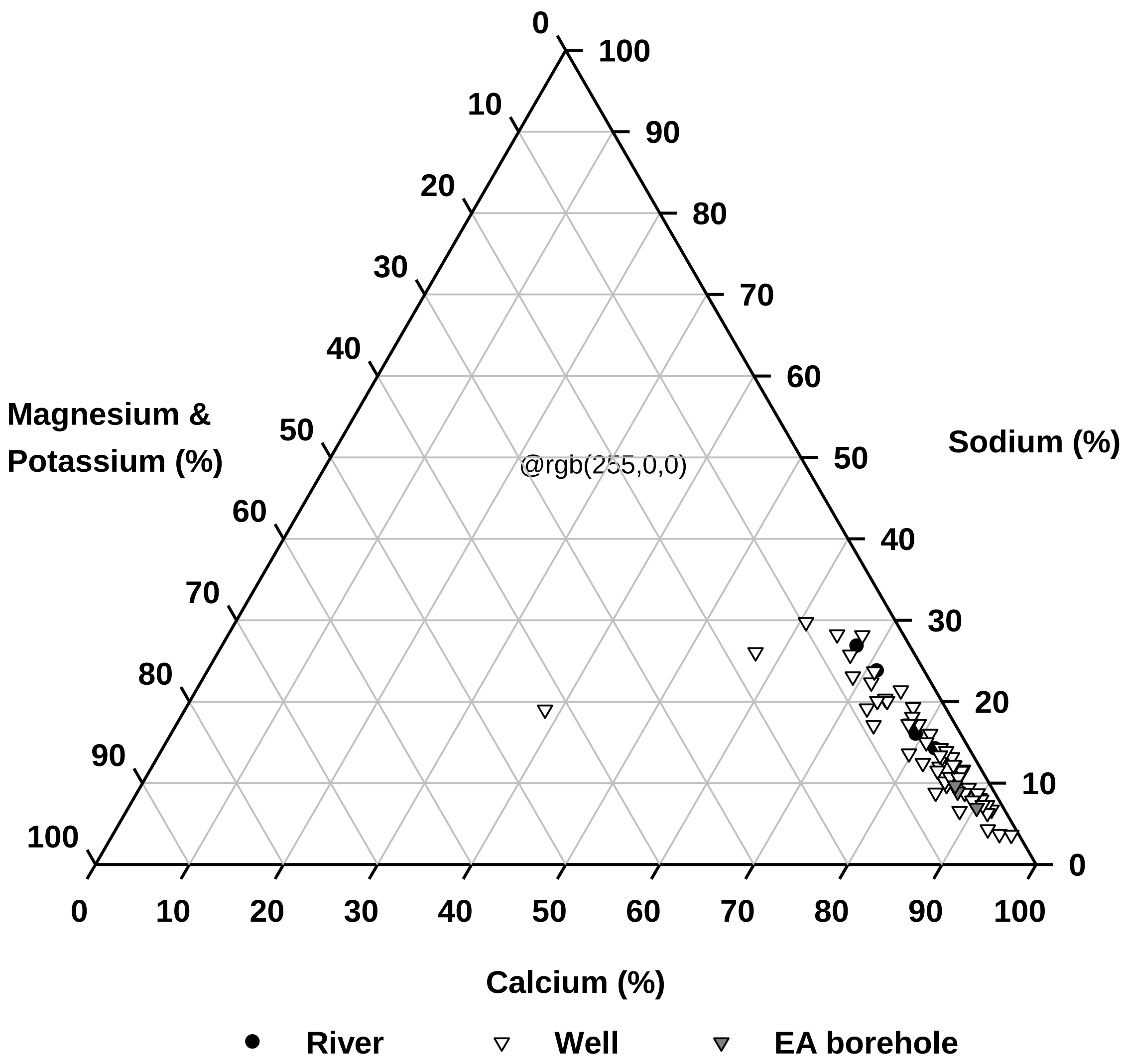


Figure 5.14: Ternary plot of base cation chemistry in river water, wells on the floodplain, and the EA borehole (instrumented in the chalk).

Table 5.7: Chemistry of the River Glaven and Hunworth Meadow groundwater wells(Mean ± 95% confidence interval).

	River	Upstream Transect	Midstream Transect	Downstream Transect	EA borehole
pH in-situ	7.3 ± 0.5	6.6 ± 0.1	6.7 ± 0.2	6.5 ± 0.1	7.4 ± 0.2
Conductivity (μs cm^{-1})	953 ± 213	676 ± 130	775 ± 767	812 ± 111	519 ± 50
Ca^{2+} (mg L^{-1})	95.8 ±12.4	113 ± 81	71.2 ± 28.2	102.6 ± 11.5	89.7 ± 4.8
Na^{+} (mg L^{-1})	25.6 ± 5.1	13.7 ± 3.2	24.8 ± 10.5	17.6 ± 1.3	16.1 ± 0.2
Mg^{+} (mg L^{-1})	5.0 ± 0.8	3.5 ± 0.6	5.2 ± 2.2	4.3 ± 0.5	5.5 ± 0.3
K^{+} (mg L^{-1})	2.6 ± 0.4	1.0 ± 0.7	5.5 ± 3.0	1.9 ± 1.0	1.8 ± 0.1
NH_4^{+} (mg N L^{-1})	0.6 ± 0.1	0.8 ± 0.3	0.6 ± 0.3	0.5 ± 0.1	0.0 ± 0.0
Cl^{-} (mg L^{-1})	45.4 ± 1.0	33.9 ± 3.7	48.6 ± 23.0	27.5 ± 5.3	31.7 ± 2.5
SO_4^{-} (mg L^{-1})	48.4 ± 4.1	30.7 ± 12.7	33.6 ± 17.7	14.5 ± 4.4	43.7 ± 10.1
NO_3^{-} (mg N L^{-1})	6.2 ± 0.5	0.2 ± 0.1	0.4 ± 0.4	0.1 ± 0.03	0.2 ± 0.0
$PO4^{3-}$ (mg P L^{-1})	0.1 ± 0.1	0.8 ± 3.3	0.1 ± 0.1	0.1 ± 0.03	No data
TDN (mg L^{-1})	5.5 ± 4.5	3.3 ± 0.6	4.7 ± 1.4	2.6 ± 0.6	No data
DOC (mg L^{-1})	10.8 ± 2.3	33.8 ± 6.0	46.1 ± 7.6	32.7 ± 6.8	0.8 ± 0.2
DO (mg L^{-1})	10.8 ± 0.4	0.4 ± 0.2	1.0 ± 0.8	0.5 ± 0.3	1.4 ± 0.8

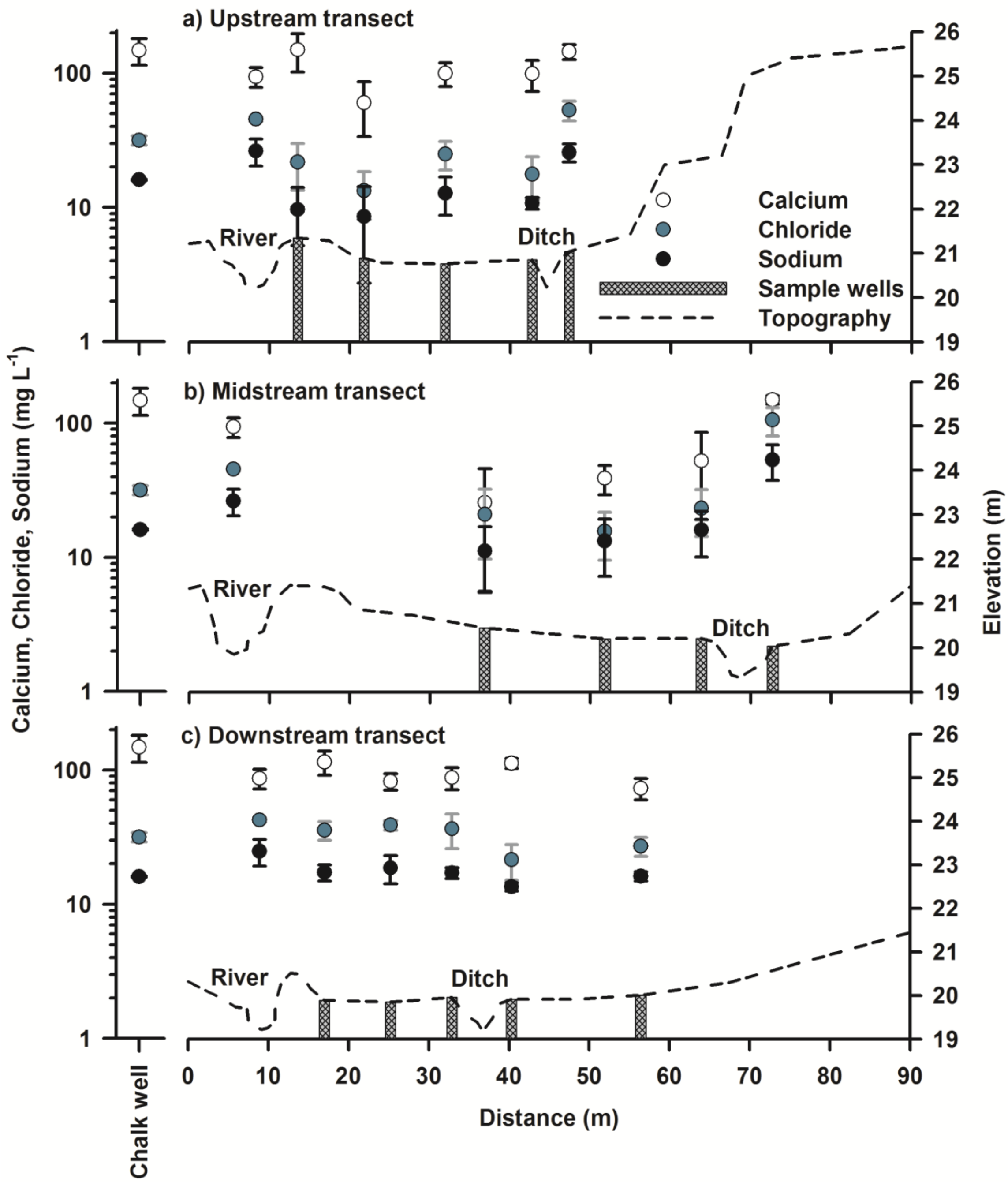


Figure 5.15: Spatial variation of selected ions (mean ± 95% confidence interval; log scale) along subsurface flowpaths at the upstream, midstream and downstream well transects for 2007 − 2008. Concentrations in the EA borehole (instrumented in the chalk) are also shown. Pre-restoration topography is plotted to identify the sample locations in relation to the river and hillslope.

Dissolved oxygen and nitrate concentrations in river water were, respectively, approximately 18- and 14-fold greater than the concentrations in shallow groundwater wells on the meadow ($p<0.05$) (Figure 5.16). Groundwater in these wells was consistently depleted in DO (0.6 mg O_2 L^{-1}) and nitrate (mean = 0.21 mg NO_3^--N L^{-1}) relative to river water (mean: 10.8 mg O_2 L^{-1} and 6.2 mg NO_3^--N L^{-1}, respectively) ($p<0.05$), and exhibited little change with distance from the river. However, nitrate concentrations at Well 3.1, which was located on the embankment, showed greater variation than the other wells on the floodplain, possibly indicating greater connectivity with river water (Figure 5.16a). Dissolved organic carbon concentration, in contrast, was significantly ($p<0.05$) higher in the floodplain wells (mean at different wells between 33 and 46 mg L^{-1}) than in river water (mean = 11 mg L^{-1}) (Table 5.7), likely due to the presence of organic matter on the floodplain.

Dissolved oxygen concentration in surface soils (top 0.1 – 0.3 m) was strongly coupled with water table height (Figure 5.17b and c). As groundwater rose vertically through the soil profile and surface soils became saturated, DO concentration decreased rapidly within a day to 0 – 2 mg L^{-1}, indicating reduced conditions (Figure 17b). As the water table height fell once again, DO concentration increased at a rate of about 0.8 – 1.4 mg L^{-1} day^{-1} to atmospheric saturation (Figure 5.17b). Groundwater DO did not increase at any point during periods of high river flow and elevated water table, suggesting that oxygen-rich river water did not inundate (via overbank flow) the upstream area of the floodplain where the DO probes were located during the study period (Figure 5.17b). This is further supported by site observations during high-flow events. Although Figure 5.10 suggests that one overbank flow event occurred following the embankment removal, the event was of short duration (≤1 day) and likely only inundated the downstream, relatively lower-lying section of the floodplain.

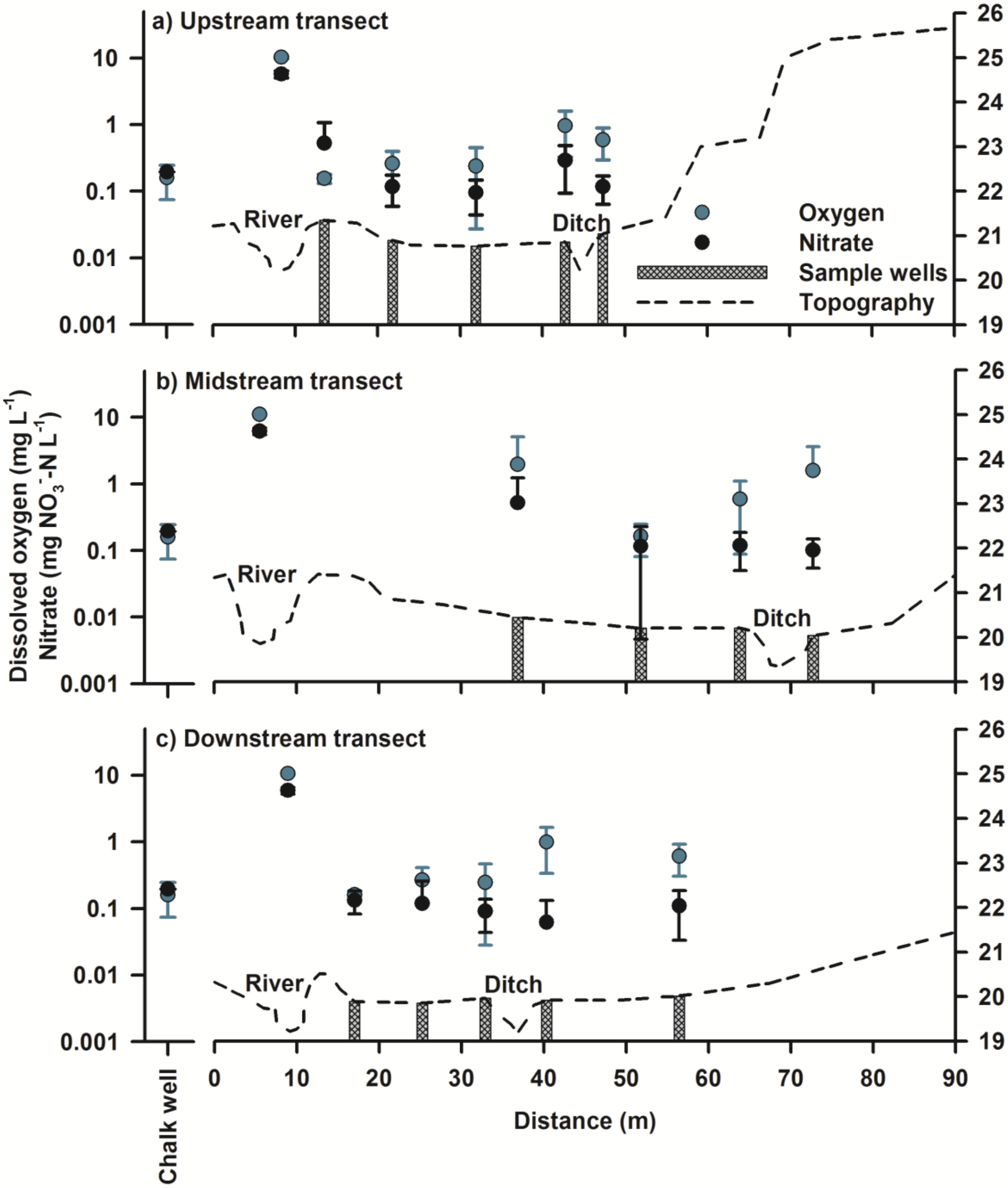


Figure 5.16: Nitrate and DO concentrations (mean ± 95% confidence interval; log scale) in river water and groundwater along subsurface flowpaths at the upstream, midstream and downstream well transects for 2007 – 2008. Concentrations in the EA borehole (instrumented in the chalk) are also shown. Pre-restoration topography is plotted to identify the sample locations in relation to the river and hillslope.

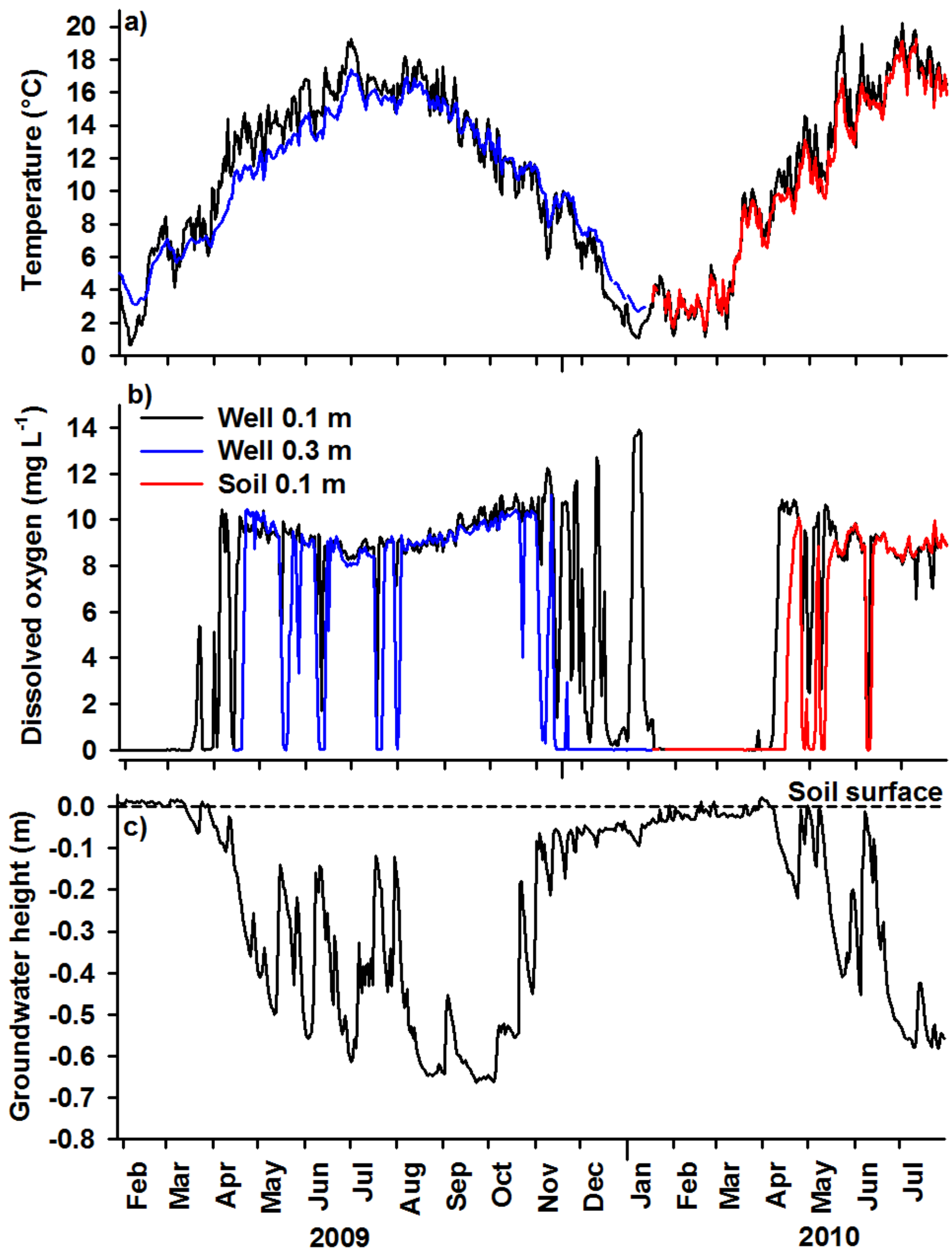


Figure 5.17: Temporal variation in (a) temperature and (b) DO concentration in well water and soil in relation to (c) changes in groundwater height. DO concentrations in well water were measured continuously at 0.1 and 0.3 m below the soil surface. In January 2010, the 0.3 m oxygen optode was moved from the well and buried directly in the soil at 0.1 m below the soil surface. When groundwater height is below the respective DO optode installation height, optode data represent atmospheric temperature and saturated DO conditions.

5.3 Discussion

5.3.1 River-floodplain hydrological linkages

As ecosystems strongly influenced by disturbances linked to flooding, floodplains are widely thought to be important for mediating the flow of water, energy, matter, and organisms between aquatic and terrestrial environments (Junk *et al*. 1989; Tockner and Stanford 2002; Naiman *et al*. 2005; Mitsch and Gosselink 2007). Functioning hydrological links are key for the numerous interstitial foodwebs (Stanford and Ward 1993) and biogeochemical processes, such as aerobic respiration, nitrification, denitrification and methanogenesis, that occur in the saturated sediments beneath and adjacent to rivers and streams (Jones and Holmes 1996). However, the reduction or absence of overbank inundation and the associated flood-related natural disturbance regime have drastic effects on biological and chemical conditions on floodplains. For example, overbank flooding is important for the deposition of nutrients and sediments, the control of dominant species and the transport of propagules that leads to increased species richness, and flood water storage (Brunet *et al*. 1994; Ward and Stanford 1995; Silvertown *et al*. 1999; Bullock and Acreman 2003; Nilsson *et al*. 2010).

The embankments along the River Glaven represented a substantial barrier to river-floodplain interactions. Without overbank flow, slow horizontal subsurface flow (<0.4 cm day^{-1}) was the primary mechanism for the exchange of water, DO and nutrients between the river and floodplain. With such low-flow velocities, it would take approximately 1 year for a molecule of water to travel a distance of 1.5 m from the river towards the floodplain. During high river flows when river stage was above groundwater height on the floodplain, there was a greater potential for lateral subsurface mixing of river and groundwater. However, considering the low subsurface flow rates of these soils, groundwater movement on the floodplain is likely to be dominated by the rapid (<1 day) vertical transfer of deeper oxygen and nitrate-depleted groundwater through the soil profile that was observed in response to precipitation. This prompt groundwater response to precipitation input, despite low hydraulic conductivity, is a phenomenon that has been reported in many small catchments and is not well understood (Kirchner 2003; Cloke *et al*. 2006).

A similar study on a chalk river in the Berkshire Downs, UK, found that the majority of surface-subsurface hydrological exchange occurred a few tens of centimetres from the river bed, and was limited to the gravel aquifer. The underlying chalk at the site was found to be hydraulically separate from the river (Allen *et al*. 2010). Throughout the River Glaven catchment, a chalk-rich boulder clay (Lowestoft Formation) underlies the alluvium and gravels (Moorlock *et al*. 2002); this formation is reported to be variably permeable, containing groundwater only when weathered, fractured or interspersed with sand and gravel horizons (BGS 2007). As the local permeability of this layer is unknown, further investigation is required. This could involve the instrumentation of boreholes into the different geological layers at the site. The presence of this less permeable layer at Hunworth could restrict hydrological contact between the river and chalk bedrock. However, the chemical similarity between the floodplain wells and chalk well samples, and the high baseflow index and flow exceedence values for Q95, indicate interaction with the chalk aquifer (see Sections 3.3 and 4.3.2).

The alluvial and glaciogenic gravels that overlay the Lowestoft Formation are likely to have substantially higher hydraulic conductivity than the overlying alluvium. While this could provide a route for more substantial mixing between river and groundwater at depth, conservative ion chemistry in the wells has a groundwater signature. The higher nitrate concentration measured in Well 3.1, which is located next to the river, suggests some connectivity between the river and groundwater on the floodplain. In general though, there was a lack of spatial and temporal variation in the conservative ion chemistry of shallow groundwater across the floodplain, indicating that even in a chalk setting there are regions of limited hyporheic extent. Overbank inundation therefore, represents the only potential mechanism for substantial surface water-groundwater connectivity.

Soils at Hunworth Meadow were of intermediate fertility (i.e. within the range of 5 – 15 mg kg^{-1} Olsen extractable phosphorus specified by Gowing *et al*. 2002a), likely due to the cessation of fertilizer application at the site in 1997 and the absence of flood water and river sediment inputs. Water flowing within the River Glaven is substantially richer in nitrate and DO relative to floodplain groundwater, and thus represents a potential source of nutrients to floodplain vegetation and microbes. The large difference in nitrate and DO concentrations between river water and groundwater <2 m away from the channel indicate that a strong redox gradient is present at the river–floodplain interface. This part of the floodplain is likely to be an important zone for reducing nitrate concentrations (e.g. Dahm *et al*. 1998; Hedin *et al*. 1998; Clilverd *et al*. 2008).

The removal of the river embankments adjacent to Hunworth Meadow sufficiently reduced the channel cross-sectional area, and thus bankfull capacity, to initiate overbank inundation and reconnect the river with its floodplain. One overbank event was observed during the period of study, and long-term river discharge data from before the restoration indicate that river flows will regularly exceed the restored bankfull capacity. Flood events will be of short duration, as bankfull capacity was typically exceeded for only one day, and they will often occur during the summer months when surface soils are dry and have a greater capacity for water storage, which is likely to maximise floodpeak attenuation (e.g. Burt *et al*. 2002). Flooding may also persist for even longer periods, depending on the infiltration and evapotranspiration rates, and the influence of in-river macrophyte growth during the summer. Increased frequency and duration of floodplain inundation due to embankment removal is consistent with other river restoration studies (e.g. Acreman *et al*. 2003; Helfield *et al*. 2007; Hammersmark *et al*. 2008) and is seen as one of the main aims of river restoration projects.

So far, an increase in the frequency of overbank flooding is suggested to be the most dramatic hydrological effect following the restoration of the floodplain. Increased groundwater levels at the river–floodplain interface have also been observed, possibly due to enhanced river water intrusion. The most noticeable change in soil saturation in this part of the floodplain, however, was due to a lowering of the surface elevation relative to the water table height. This occurred along a 1 – 2 m strip where the embankments were previously located, and is likely to promote re-colonisation by wetland plant species that can tolerate periodic waterlogging and aeration stress in the rooting zone, particularly during the growing season (e.g. Silvertown *et al.* 1999; Barber *et al*. 2004; Wheeler *et al*. 2004). However, water levels in the rest of the floodplain were lower during the summers following the restoration. This could be a result of the prevailing low summer precipitation and high evapotranspiration rates in these years. The areal extent of inundation on the floodplain could potentially be far-reaching due to the low-lying elevation of the meadow in relation to the river; however, further analysis including the application of hydrological/hydraulic modelling (e.g. Thompson *et al*. 2004) is required to determine flood inundation extents for a range of flows (see Sections 6 and 7).

5.3.2 Floodplain ecohydrology

Prior to embankment removal, initial site assessments indicated that Hunworth Meadow was comprised of a degraded *Holcus lanatus* – *Juncus effusus* rush pasture community classified as MG10 under the UK National Vegetation Classification system (Rodwell 1992) (see Section 3.6), which is typically associated with waterlogged soils. This is congruent with prolonged saturation of surface soils observed pre-restoration during winter and spring, and with the periodic saturation occurring during summer and autumn months, all of which occurred during within-bank river flows. Flooding, particularly during the growing season, can cause aeration stress in plants, with prolonged waterlogging leading to species-poor plant communities (Jackson and Colmer 2005). This stands in stark contrast to the effects related to frequent low-intensity summer flooding of the floodplain with oxygen-rich river water, which are predicted for the site following the removal of the embankments along the River Glaven. Recurrent overbank inundation increases environmental heterogeneity and is believed to have a positive effect on floodplain plant diversity, firstly by limiting competition by dominant plant species (e.g. Silvertown *et al*. 1999; Helfield *et al*. 2007), and secondly by opening new patches for colonization by hydrochorically deposited propagules (e.g. Auble and Scott 1998; Nilsson *et al*. 2010).

In contrast to brief summer inundation events, infrequent floods of lengthy duration during the growing season can be expected to negatively affect floodplain diversity, either by burying plants with river sediment or by exceeding tolerance limits for anoxia in the rooting zone of sensitive species (e.g. Gowing and Youngs 1997; Friedman and Auble 1999); however, floods of this type were not observed during the 10 years for which river discharge data are available. Furthermore, at Hunworth Meadow there is initial evidence to suggest that reconnection will improve drainage and create drier conditions between flood events due to the easier drainage of the floodplain following the removal of the embankments, which partly moderates the effects of large floods (this is further explored in Section 7.2.2).

Re-establishment of overbank flooding may result in nutrient enrichment of floodplain soils from flood-deposited sediment and nutrient-rich river water (Gowing *et al*. 2002b). While this may function to protect adjacent ecosystems from nutrient loading, increased nutrient supply may pose a risk to plant species richness at the restoration site (Vermeer and Berendse 1983; Verhoeven *et al.* 1996; Janssens *et al.* 1998; Michalcová *et al*. 2011) and over-ride the ecological benefits of improved river–floodplain connections. In such instances, further management may be required. In the case of Hunworth Meadow, which is a mesotrophic wet grassland, the additional supply of nutrients could be managed with traditional hay cutting. This would help balance the input of river-derived nutrients to the floodplain with the removal of nutrients in plant biomass (Linusson *et al*. 1998; Wheeler *et al*. 2004).

Chapter 6: Methods – Part II: Hydrological/hydraulic modelling

6.1 Introduction

In this chapter the methods employed to simulate surface water and groundwater elevation, groundwater recharge, surface runoff, and floodplain storage are outlined. It describes the development (input data and parameterisation), calibration, and validation of two coupled surface water/groundwater MIKE SHE-MIKE 11 models used to simulate floodplain hydrology before and after the removal of the river embankments along the River Glaven at Hunworth Meadow.

6.2 MIKE SHE model development

6.2.1 Model domain and topography

Coupled MIKE SHE/MIKE 11 hydrological/hydraulic models were developed for the pre-restoration (embanked) and post-restoration (no embankment) scenarios, which differed in embankment and riverbed elevation resulting from the embankment removal. Figure 6.1 provides a representative example of the change in river and embankment topography along the study reach. The riverbed was slightly higher after the embankment removal works due to the accidental loss of bank material into the river (see also Figure 5.1). This raised the riverbed in relation to the agricultural drainage ditch, such that the river thalweg (lowest point along the river bed) was above the ditch thalweg along the length of the meadow. In both models, the model domain included Hunworth Meadow and extended up to the tops of the adjacent hillsides on either side of the river. The upstream limit of the modelled area coincided with the disused railway embankment whilst the smaller embankment carrying an agricultural track crossing the floodplain defined its downstream limit. The model was extended downstream of the study meadow in order to prevent interference around the model boundary affecting groundwater simulations on the meadow (Figure 6.2).

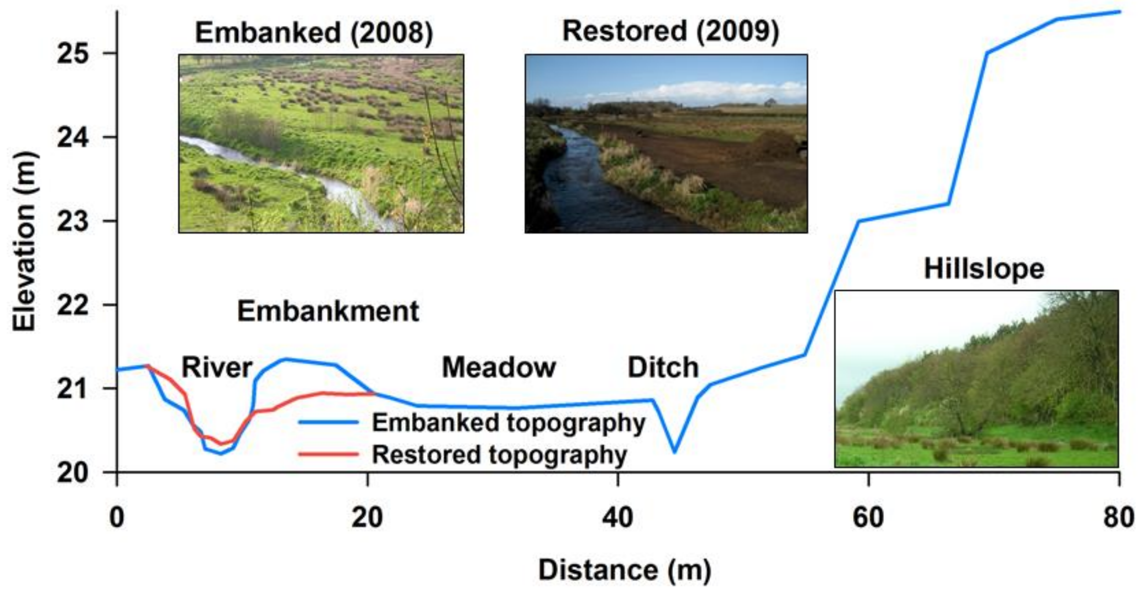


Figure 6.1: Cross-sections and photographs of the river and floodplain topography before (embanked) and after (restored) embankment removal.

Two digital elevation models, one representing the embanked river, and the other the restored river, were derived from dGPS surveys conducted before and after embankment removal (see detailed description in Section 4.4.4) (Figures 6.2 and 6.3). The surface topography delineates the surface boundary of the model for overland flow, and acts as a reference point for the depth of the saturated and unsaturated zone layers and groundwater level observations. The DEMs were created in Arc-GIS using the Kriging interpolation function. A comparison of three Arc-GIS interpolation functions (Kriging, Nearest Neighbour, and IDW) indicated that the Kriging method resulted in the least absolute error (mean error = 0.001 m) between observed and interpolated data points, and was best able to represent changes in topography over small spatial scales (Figure 6.4).

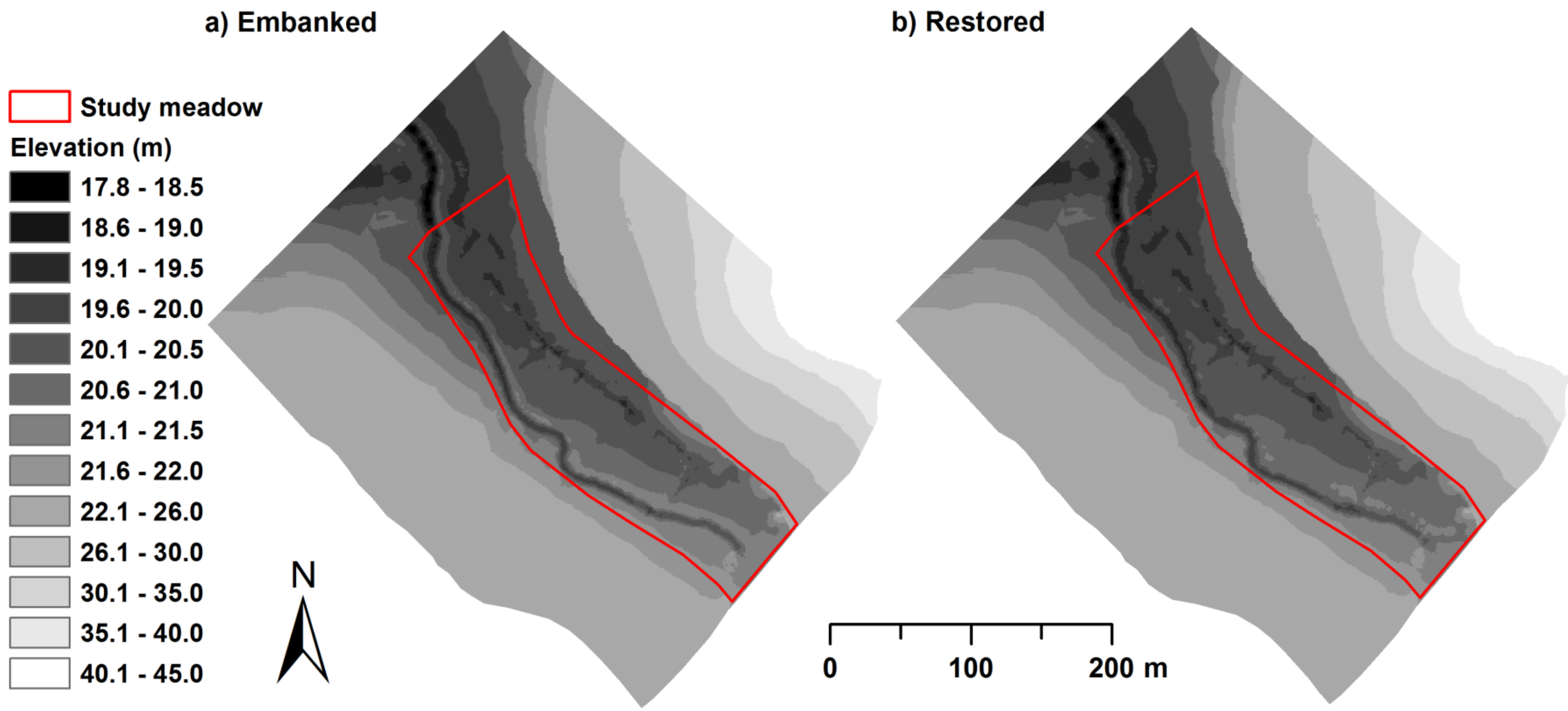


Figure 6.2: Digital elevation models used in MIKE SHEfor the (a) embanked and (b) restored river embankment models. The extent of each the DEM delineates each respective model domain. An enlarged map of the study meadow is shown below in Figure 6.3.

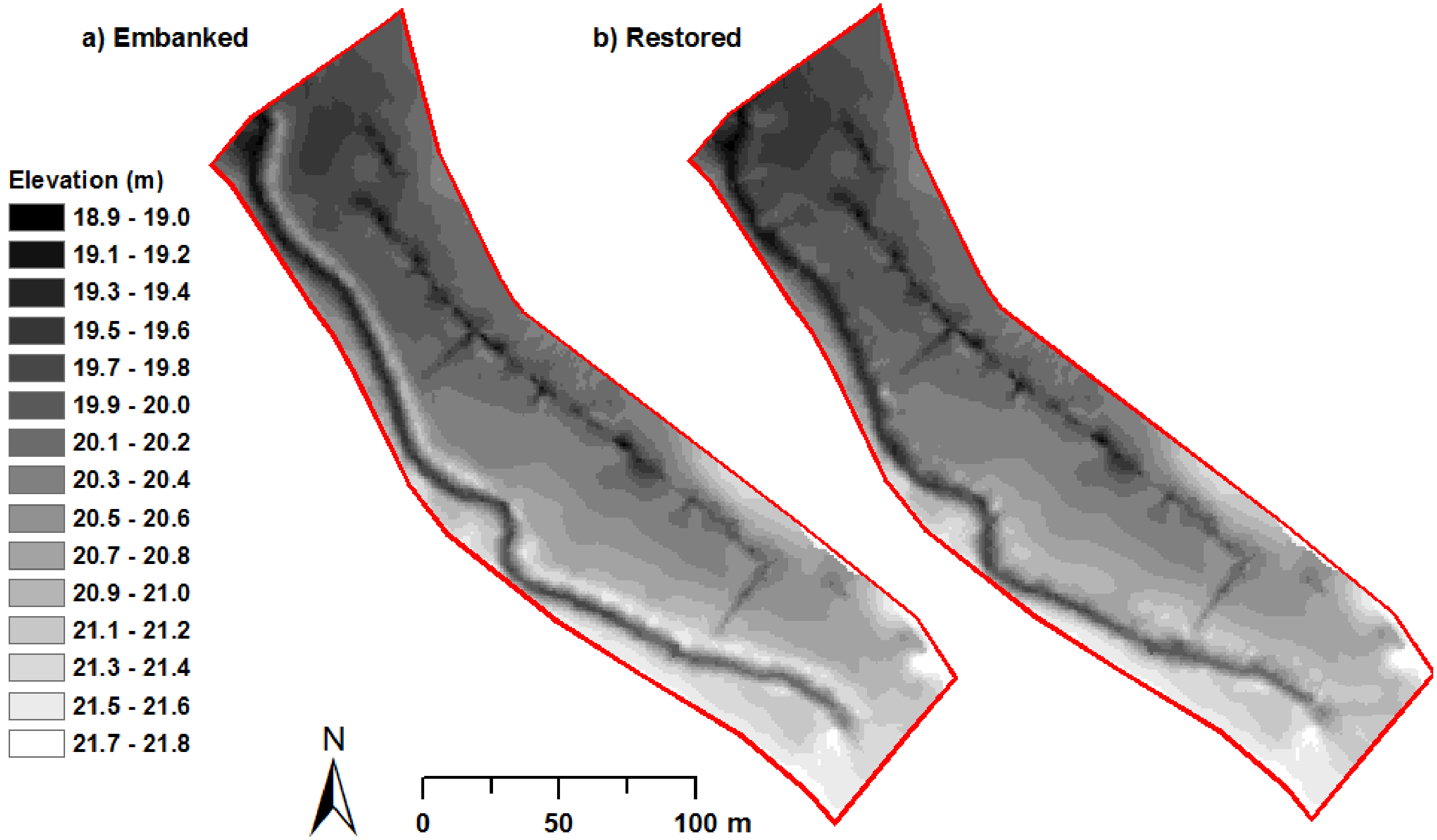


Figure 6.3: Digital elevation models of the (a) embanked and (b) restored sections of the study meadow. Surface elevation of the river embankments was reduced on average by 0.8 m (see Section 5.2 for detailed results).

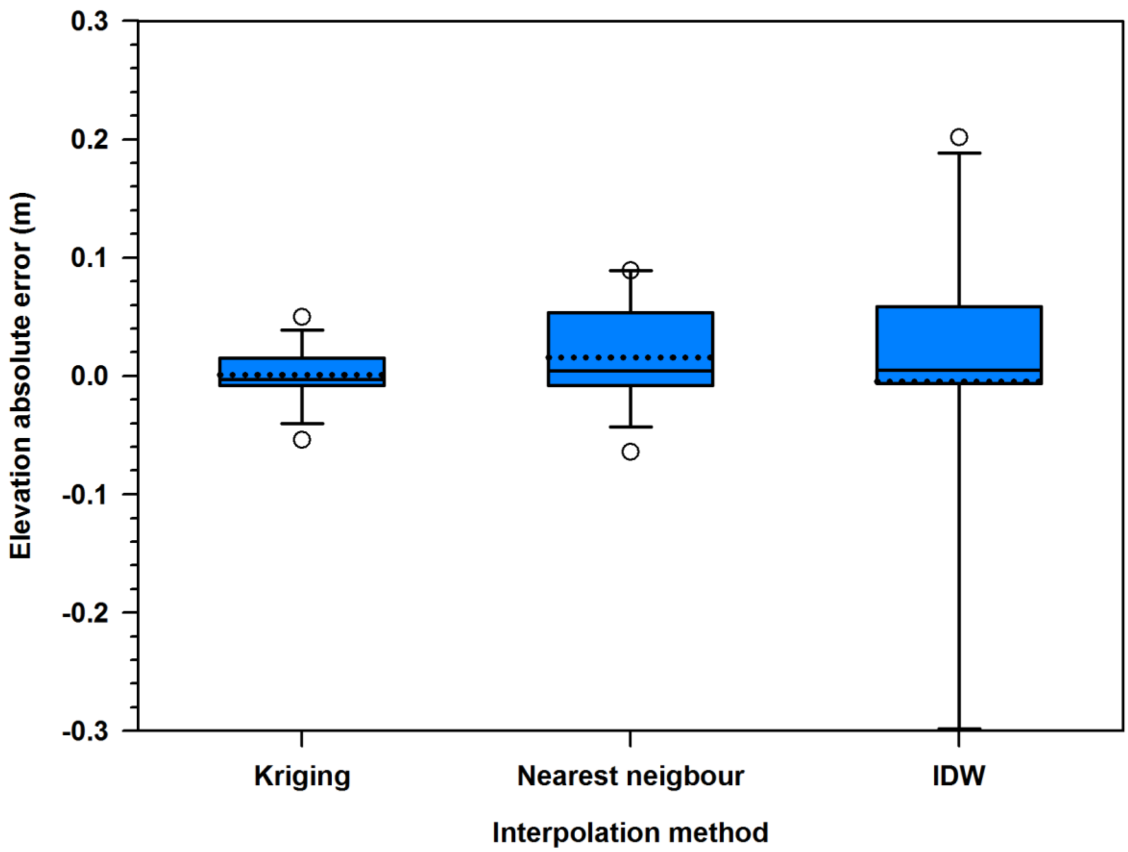


Figure 6.4: Comparison of absolute error for different interpolation methods at 1 m grid spacing. The solid centre line and broken line within the boxplot indicate the median and mean, respectively. The box extent and error bars denote the 25th and 75th percentiles, and 10th and 90th percentiles, respectively. The circles indicate the presence of outliers.

Both DEMs in Figure 6.2 were re-sampled to the MIKE SHE model grid. The model domain was divided into 5,038 grid cells of 5 m × 5 m, which represented a total area of 12.6 hectares. As discussed below in Section 6.4.2, initial calibration steps employed a larger grid size of 15 m × 15 m grid (610 grid cells), which resulted in appropriate computational times for the multiple model runs during the autocalibration process. Subsequently, the model grid size was reduced during manual fine-tuning of the model (see section 6.4.2). Experiments during model calibration showed little difference in simulated groundwater elevations between 1 and 5 m. In order to balance between grid resolution and computational speed, common with distributed hydrological models (McMichael *et al*. 2006; Thompson *et al.* 2013), 5 m x 5 m was selected as the optimal grid size relative to the high resolution of topographic data available, whilst also ensuring appropriate computational speed.

The relatively fine discretisation of the final model was needed to accurately characterise topographic variations across the floodplain including the blocked ditch and small-scale features such as shallow depressions and raised hummocks that can provide microhabitats of differing soil water content that are important for fostering high species diversity (Wheeler *et al*. 2004). At higher grid spacing, the river and ditch topography were significantly dampened in the interpolation, and were lost altogether from the 20 m grid DEMs (Figure 6.5). As the dGPS data are discontinuous, with higher sampling densities (1 – 2 m) in key areas of topographic change i.e. in the river, on the embankments, and in the ditch, 10 m spacing on the meadow where the topography was relatively flat, and substantially lower sampling densities (ca. 20 – 40 m) outside of the study meadow i.e. on the hillslopes, it is acknowledged that the DEMs are spatially aliased across the model domain. However, in order to capture the key study area on the floodplain in sufficient detail and to simulate subtle differences in the water regime across the site, a 5 m grid resolution was necessary. Ideally, the model would be organised into a network of smaller grid sizes on the meadow (e.g. 1 m × 1 m), where it was important to represent small changes in topography, and larger grid sizes on the hillslope (e.g. 20 m × 20 m). However spatially variable grid sizes were not an option in the version of MIKE SHE used (Release 2009) and is also not available in the latest software release.

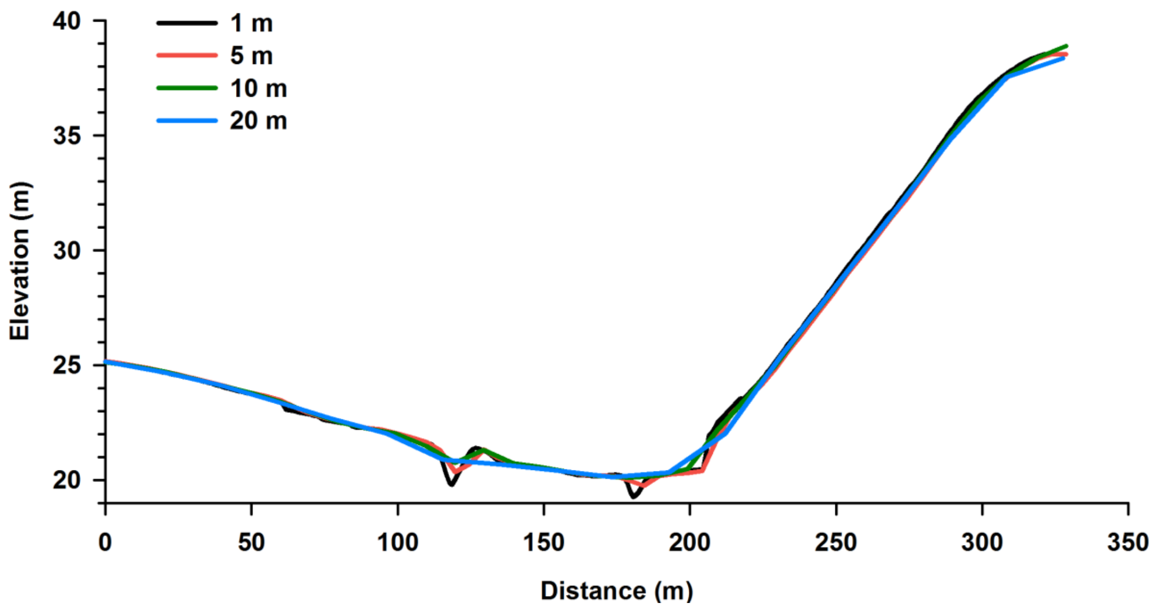


Figure 6.5: Loss of topographic resolution with increased interpolation grid size. Cross-sections depict the embanked river, floodplain and hillslopes from a Kriging interpolation in ARC-GIS.

6.2.2 Hydrological and climate data

A detailed description of the well locations, instrumentation, and evapotranspiration calculations are given in Sections 4.2 and 4.3. Over three years of river discharge and meteorological data, and observed groundwater elevations were used to, respectively, parameterise and calibrate/validate the models. Water level data used in the calibration and validation process (discussed below) were available for a 24 month period from 22/02/2007 to 15/03/2009 prior to the embankment removal and 16 month period from 29/03/2009 to 25/07/2010 after the embankment removal. Spatially uniform precipitation and potential evapotranspiration were specified, an approach justified by the small size of the model domain (Thompson 2004). Daily river discharge available from 2001 – 2010 from the onsite Environment Agency gauging station (see Section 4.2) is discussed below in the MIKE 11 section).

Daily precipitation and potential evapotranspiration (see Section 4.3) inputs required to simulate water fluxes in the unsaturated zone and recharge to the saturated zone, were based on records from an automatic weather station (Skye MiniMet SDL 5400) installed 100 m from the meadow (see Figure 3.1) supplemented during periods of instrumental failure with data from a nearby (<10 km) UK Met Office meteorological station (source ID: 24219, Mannington Hall). Daily Penman-Monteith potential evapotranspiration was computed from air temperature, net radiation, relative humidity, and wind speed observations from the on-site weather station (see Section 4.3.3).

Pressure transducers were installed in 10 wells at Hunworth Meadow in February 2007 to measure changes in groundwater height (as described in Section 4.3.1). The level logger in Well 3.4 was removed in June 2007 due to its close proximity to the level logger in Well 3.5 and the need for a logger in the ditch at Transect 1. Hand measurements of water table depth recorded during site visits, which were subsequently converted to water table elevation using dGPS measurements of the top of the well casings, show good agreement with the level logger data, which were generally within a few cm of each other (Figures 6.6 - 6.8). These data underpin all of the groundwater modelling, and thus great care was taken to check the quality of the water level data. The exception was Well 1.2, where a disagreement of approximately 7 cm occurred between the hand measurements and level loggers after the restoration; this potential error was taken into account during the modelling analysis (Figure 6.8).

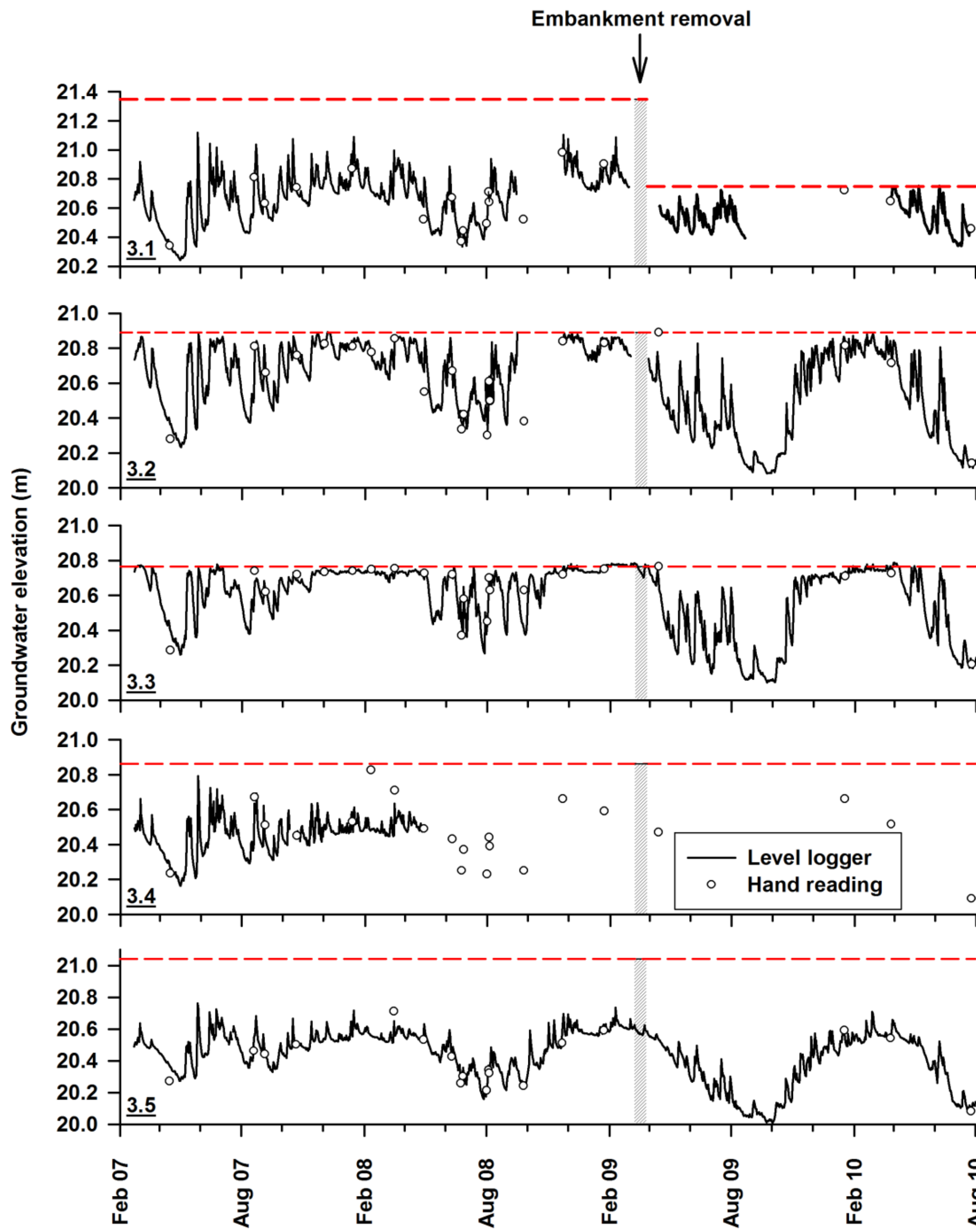


Figure 6.6**:** Comparison of level logger and hand measurements of groundwater elevation at the Upstream Well Transect. The embankment removal in March 2009 is highlighted by the vertical hashed bar. The soil surface is shown by the broken red line. Note the change in soil surface elevation at Well 3.1, which was located on the embankment prior to the restoration.

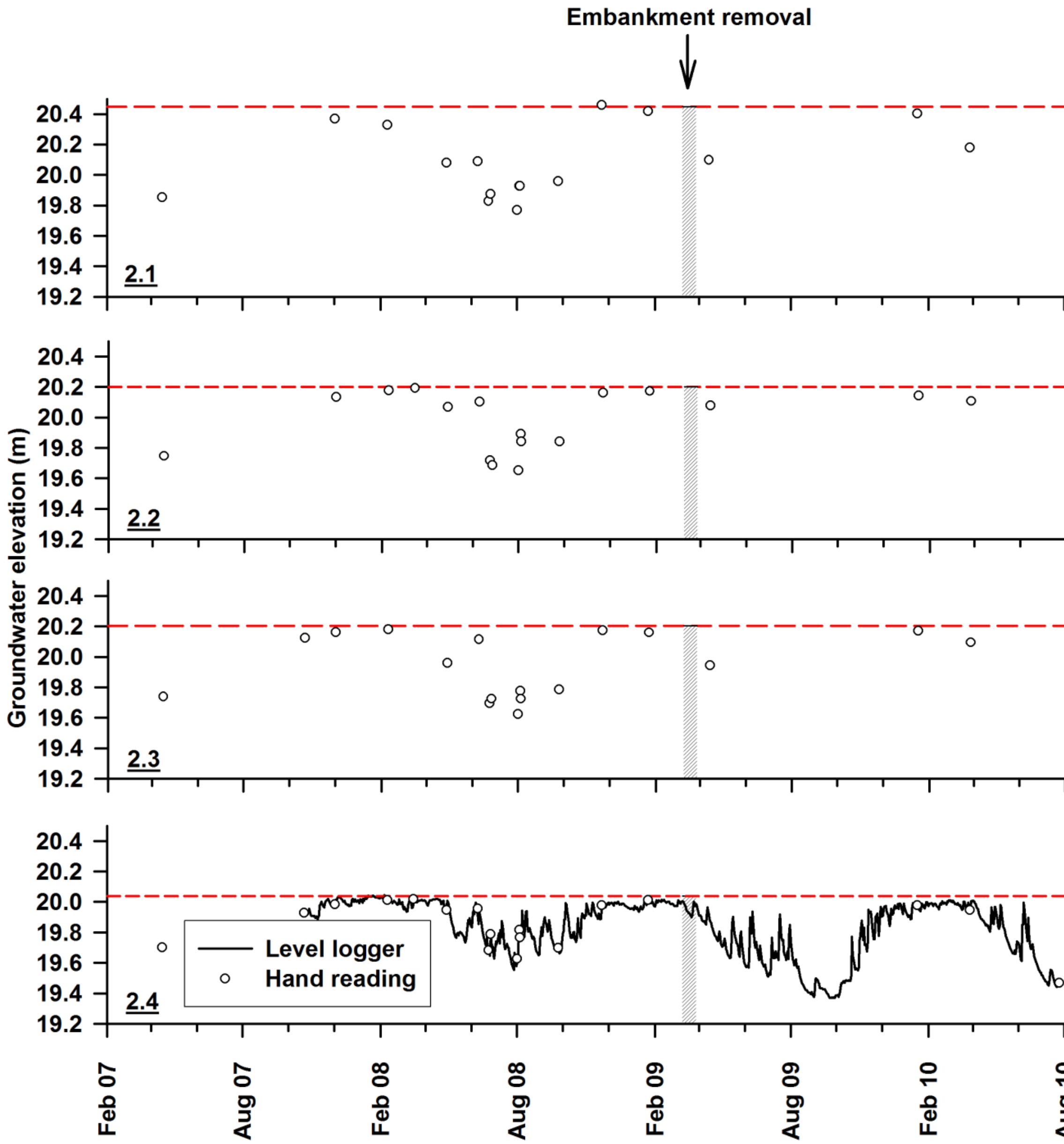


Figure 6.7: Comparison of level logger and hand measurements of groundwater elevation at the Midstream Well Transect. The embankment removal in March 2009 is highlighted by the vertical hashed bar. The soil surface is shown by the broken red line.

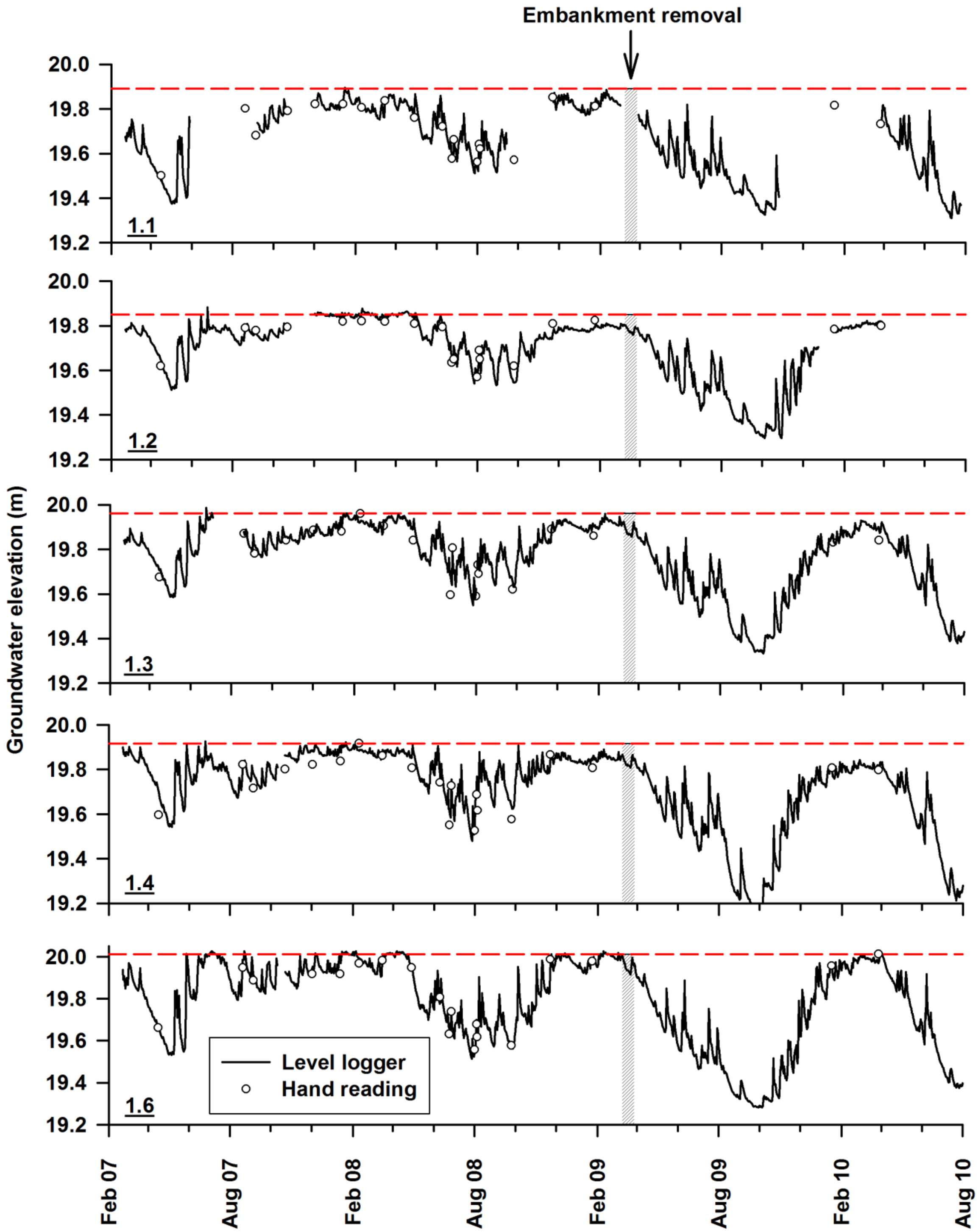


Figure 6.8: Comparison of level logger and hand measurements of groundwater elevation at the Downstream Well Transect. The embankment removal in March 2009 is highlighted by the vertical hashed bar. The soil surface is shown by the broken red line.

Differences in land use (e.g. grassland versus woodland), which affect evapotranspiration, runoff and thus groundwater recharge, are defined within the MIKE SHE model domain by the allocation of different vegetation types. Each land use category is assigned a Leaf Area Index (LAI) and rooting depth specific to the vegetation type that it represents. Four different land use classes were defined in the Hunworth models: roads and buildings, arable land, riparian grassland, and mixed deciduous/coniferous woodland (Figure 6.9).

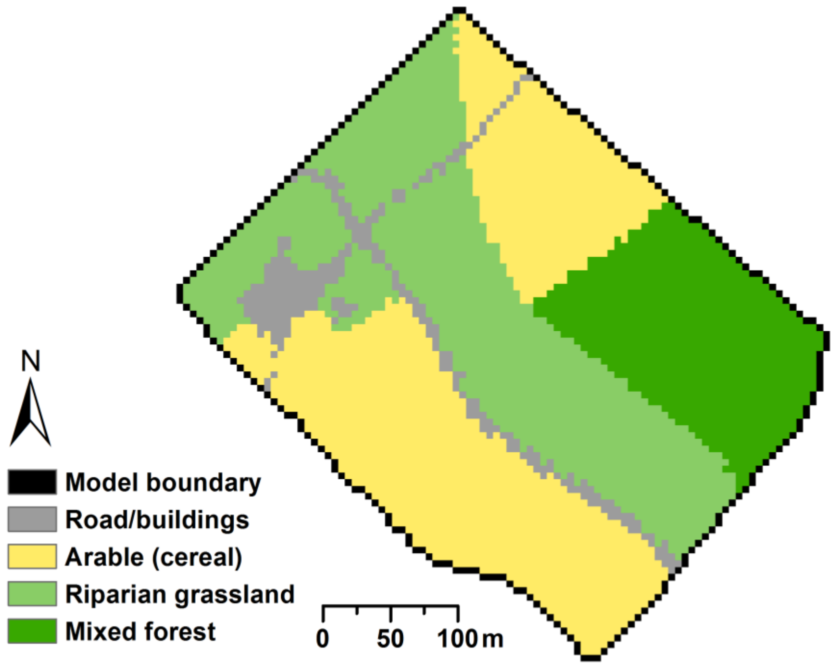


Figure 6.9: Spatial classification of land use in the model for delineating differences in LAI and root depth.

Constant rooting depths were applied to most land use classes, with the exception of the arable class, which was varied seasonally (range: 0 – 1.8 m) (Figure 6.10). Root depth was set at 0.3 m on the meadow and 2.7 m was used for the mixed deciduous/coniferous woodland (Figure 6.10c and d). Root depth values for the woodland and arable crop (classified as winter wheat) were taken from the literature (Canadell *et al*. 1996; Thorup-Kristensen *et al*. 2009; FAO 2013), whereas the rooting depth for the meadow was based on investigations at the site and measurements of water table depth, which showed that a shallow region of topsoil was aerated during the growing season. Root depth was defined as 0 for the roads and buildings land cover class (Figure 6.10).

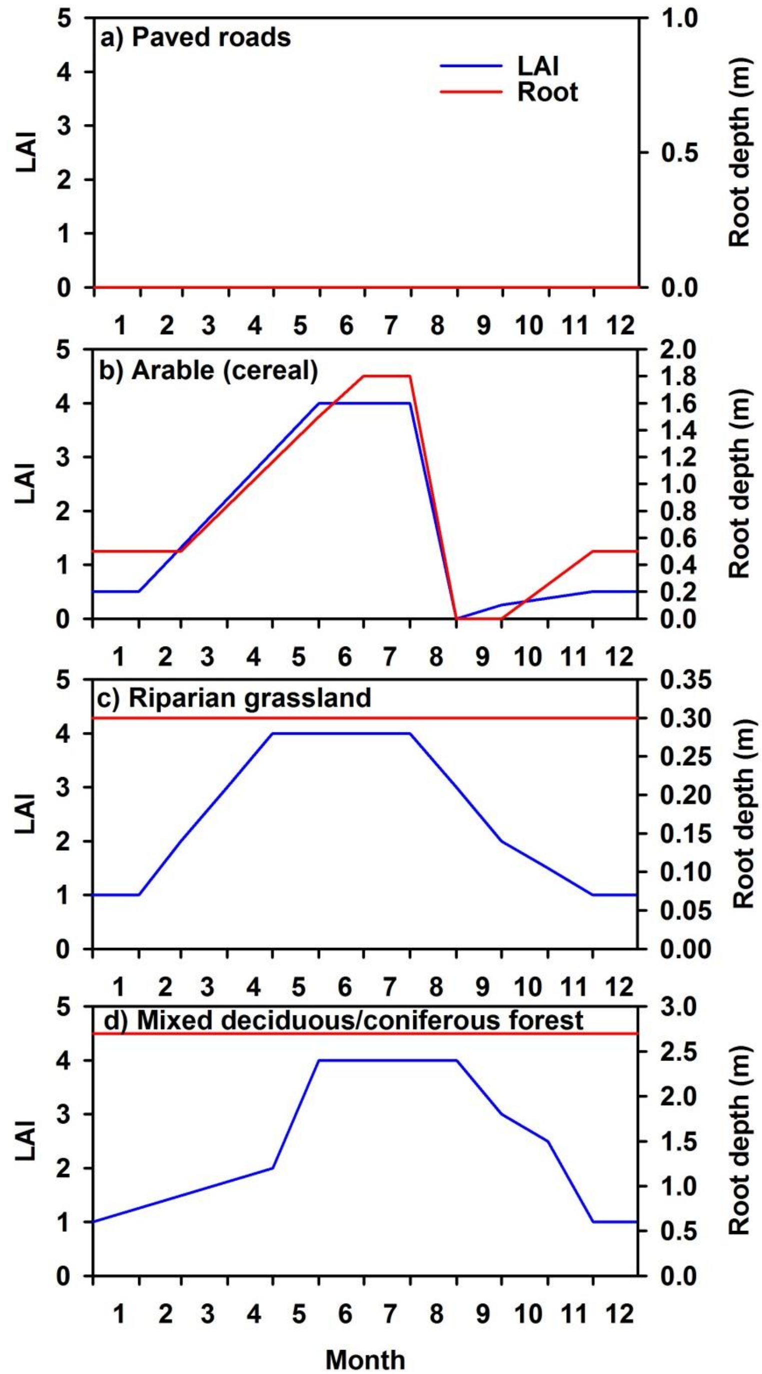


Figure 6.10: Leaf area index and rooting depth values assigned to the four land use classes.

Seasonal changes in Leaf Area Index (LAI) were applied to the arable (range: 0 – 4), meadow (range: 1 – 4) and mixed woodland (Range: 1 – 4) classes to account for increased LAI values during the growing season (Herbst *et al*. 2008; Hough and Jones, 1997). LAI was defined as 0 for the roads and buildings (Figure 6.10). LAI values for the mixed deciduous/coniferous woodland were based on data from Herbst *et al*. (2008) for two heterogeneous broad-leaved woodlands in southern England and maximum LAI values reported by Hough and Jones (1997) for conifer woodlands. Deciduous woodlands have a bare soil characteristic (LAI = 0) in the winter. However, LAI values above zero were applied all year round in the model to account for the presence of some evergreen trees and understory in the woodland (Herbst *et al*. 2008).

6.2.3 Overland flow

A finite difference solution of the diffusive wave approximation was used to simulate overland flow. The direction and amount of overland flow is controlled by the topography, resistance to flow (described by the roughness coefficient, Manning M for overland flow), the loss of water via evapotranspiration and the infiltration capacity of the soil. Manning M is the inverse of the more often used Manning’s n roughness coefficient. The higher the value of Manning’s M, the faster the water is routed overland towards the river. Manning M was initially set at a uniform value of 3.0 $m^{1/3}s^{-1}$ for the automatic calibration process (described in Section 6.4), which corresponds with values given for grassland in the literature (USDA 1986; Liu and Smedt 2004; Thompson 2004; Hammersmark *et al*. 2008). Subsequently, a distributed option of Manning M was applied to the MIKE SHE models to account for differences in overland resistance between the woodland hillslope, patches of rushes (*Juncus effusus)* in the vicinity of the ditch, and the grass meadow. The initial water depth for the overland flow calculations is usually set as zero (DHI 2007b). An initial water depth greater than 0 mm resulted in continuously flooded conditions across the meadow, which was not realistic; hence the initial water depth was set at a uniform value of 0 mm and was not included as a calibration factor. In order to simulate the ponded conditions that are present at the downstream end of the meadow, an area of lower soil permeability was specified for the spatial extent of the pond to account for the accumulation of fine sediment in this region (Figure 6.11). A subsurface leakage coefficient of $1{\times}10^{-9}$ s^{-1} was used for the pond area, and detention storage (the amount of water stored in local depressions which must be filled before water can flow laterally to an adjacent cell) and initial water depth were both set at 0.05 m. The MIKE SHE models were started using these initial water values, and run for a while to get some reality of the starting conditions (see Section 6.2.7).

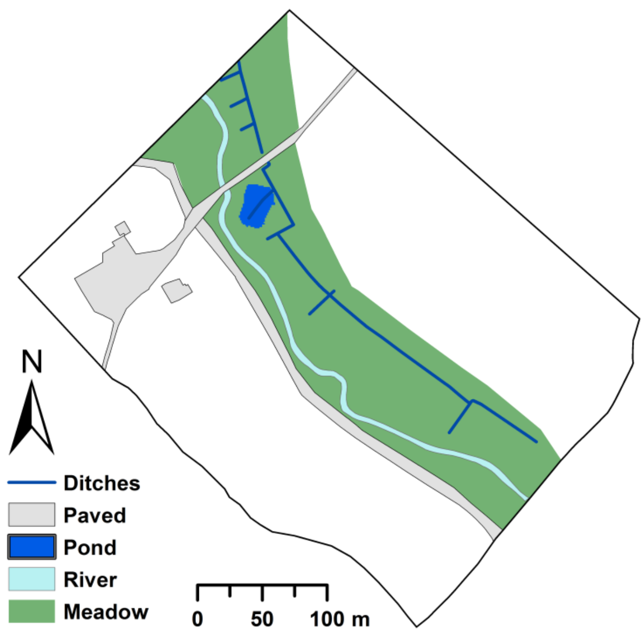


Figure 6.11: Spatial extent of water bodies in MIKE SHE domain.

6.2.4 Unsaturated zone

Water that enters the unsaturated zone can be removed by plants and transpired, stored in soil pores, or percolate downwards to recharge groundwater. Flow in the unsaturated zone can be simulated in MIKE SHE using the full Richard's equation, which is the most accurate approach but also the most computationally intensive (Graham and Butts 2005). MIKE SHE also includes a simplified two-layer water balance method, which uses a uniform soil for the entire depth, is less computationally demanding compared to the full Richard's equation (Graham and

Butts 2005; DHI 2007b), and is suitable for conditions that include the high water tables and a rapid groundwater response to precipitation that characterize Hunworth Meadow (e.g. Thompson 2012).

Soil properties were defined for a spatially uniform (global) unsaturated zone that was represented using the two-layer water balance approach. The infiltration rate of the unsaturated zone (mean: 1.88×10^{-6} m s^{-1}) was obtained from piezometer slug tests (n=25) conducted at the site (see Hydrological Conductivity methods in Section 4.3.4) and was varied during calibration (see below). Soil water content at saturation (approximately equal to the porosity) (DHI 2007b), and the water content at field capacity were additional calibration terms. Initial values were based on measurements of the water release characteristic (pF-curve) using a manual 08.01 sandbox (Eijkelkamp, Giesbeek, The Netherlands) (described in detail in Section 8.2.5), and averaged 0.7 (volumetric basis) and 0.2 (volumetric basis), respectively (Table 6.1). Water content at wilting point was also varied during calibration but could not be obtained from the water release curve as the sand box tension was not decreased below 10 kPa (field capacity) and the wilting point in soils is typically 1,520 kPa (Brady and Weil 2002). Therefore a range of wilting point values for sandy loam soils were obtained from the literature (mean = 0.07, volumetric basis) to guide the initial value in the calibration (Table 6.1) (Zotarelli *et al*. 2010).

Table 6.1 Soil properties of Hunworth Meadow topsoil (top 10 cm)

Unsaturated zone soil properties	**Mean**	**Range**
Water content at saturation (volumetric):	0.7	0.6 – 0.8
Water content at field capacity (volumetric):	0.2	0.1 – 0.4
Water content at wilting point (volumetric):	0.07	0.04 – 0.1
Saturated hydraulic conductivity (m s^{-1}):	2×10^{-6}	1×10^{-7} – 9×10^{-6}

The two-layer water balance method divides the unsaturated zone into a root zone that is subject to vertical draw from evapotranspiration, and lower zone that is not (Graham and Butts 2005). The volume of water available for groundwater recharge and actual evapotranspiration is calculated using the input precipitation and evapotranspiration data (that employs the LAI and RD as described in Section 6.2.2 and physical soil properties detailed below - Table 6.1). The rate of water loss via evapotranspiration from the unsaturated zone is dependent on the thickness of the capillary fringe, the zone of upward capillary movement of water from the water table. A capillary fringe can occur some distance above the water table, in particular in fine soils (DHI 2007b; Clilverd *et al*. 2008). If the capillary fringe reaches the soil surface, capillary action draws water directly from the water table, saturating surface soils, and resulting in maximum rates of evapotranspiration. Similarly, plant roots can draw water directly from the saturated zone if the roots reach the capillary zone. Hence the height of the capillary zone is used as the water table depth at which the influence of evapotranspiration starts to decrease. Fine soils have the most influence on capillary action. The maximum height of capillary rise for sandy loam soils at Hunworth Meadow was calculated as a function of soil pore size using Hazen's formula (Das 2002) of capillary rise as:

$$h_1 = \frac{C}{\mathrm{e}D_{10}} \qquad (6.1)$$

where the maximum height of capillary rise is h_1 (mm), C is a constant that varies from 10 to 50 mm^2, *e* is the void ratio and D_{10} is the effective particle size, the diameter (mm) of the smallest size fraction that accounts for less than 10% of total soil mass. The void ratio was obtained from soil moisture measurements using 100 cm^3 bulk density tins and D_{10} was measured with optical laser diffraction using an LS 13320 Coulter Counter particle size analyzer (Beckman Coulter Corp., Hialeah, Florida, USA). The low effective particle size (D_{10}) values in Table 6.2 are indicative of the low hydraulic conductivity values (mean: 1×10^{-5} m s^{-1}) measured at Hunworth meadow, and are in the order expected for soils that are composed of fine sands (Murthy 2002; Figure 6.11). The capillary fringe values for Hunworth Meadow sandy loam range from 0.4 m – 1.9 m (Table 6.2), which is consistent with capillary rise values of >0.5 m in fine sands and silts reported by DHI (2007b), and measurements in the range of 1.0 –1.5 m for weakly compacted alluvial sandy loams and 1.5 – 2.0 m for alluvial loams (Chubarova 1972).

Table 6.2: Range of height of capillary rise for Hunworth Meadow.

Constant (mm^2)	**Soil void ratio *e***	**D_{10} (mm)**	**Capillary height (m)**
10	3.08	0.008	0.39
50	3.08	0.008	1.93

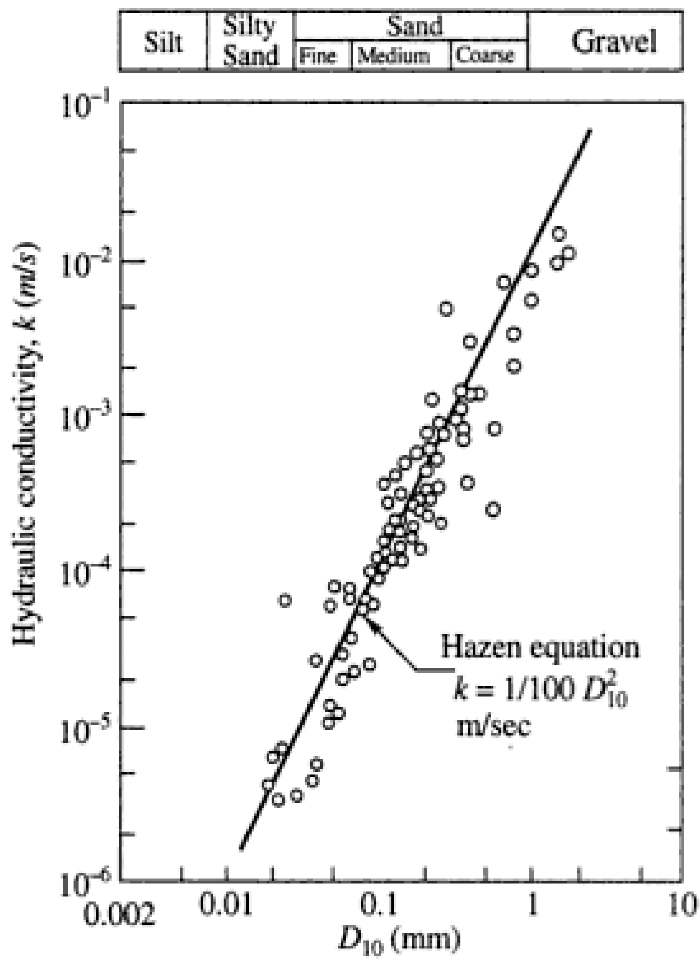


Figure 6.12: Relationship between hydraulic conductivity and D_{10} of granular soils (Louden 1952, cited in Murthy 2002).

Preferential flow through macropores in unsaturated soil can provide an important bypass mechanism for flow to the saturated zone in certain soil types (DHI 2007b).

However, the low clay content (see Section 5.2.3) and high water table conditions (see Section 5.2.2) observed across the floodplain throughout the study period suggest that the influence of macropores are limited and therefore macropore flow was not included in the model.

6.2.5 Saturated zone

A finite difference solution to the three dimensional groundwater flow equation was used to simulate flow in the saturated zone. This method allows for the vertical discretisation of geological layers and lenses in the saturated zone, each with specific hydraulic properties (DHI 2007b). Four geological layers are likely present at the site based on a British Geological Survey of superficial and regional bedrock geology in the catchment (Figure 3.6), geological strata information recorded during the installation of a British Geological Survey groundwater borehole at Brinton Hall (see Section 3.3), and onsite soil texture data recorded during the installation of groundwater wells on the floodplain. These layers consist of (1) hillwash (a poorly sorted mixture of clay, sand, silt and gravel), followed by (2) glaciogenic sand and gravel, (3) boulder clay, and (4) chalk. In addition an alluvium lens is present on the floodplain.

In order to minimise the computational demand, and because the physical properties of these layers are relatively unknown at the study site, the model was composed of a relatively simple saturated zone that represented average geological conditions in the upper alluvial and glacial soils that were considered separated from the chalk aquifer at the site by a layer of low permeability boulder clay. This consisted of one geological layer with a lower elevation of 10 m above Ordnance Datum Newlyn (ODN), and uniform hydrodynamic properties (i.e. hydraulic conductivity and storage coefficients) which were varied during the calibration process within a range of values determined from onsite investigations and typical values reported in the literature (Domenico and Schwartz 1998). Hydraulic conductivity values of the top 2 m of floodplain alluvium were determined using piezometer slug tests (n=25), described in detail in Section 4.1. The results of these tests indicate that hydraulic conductivity of the alluvium is relatively low, ranging from 1.13×10^{-7} to 5.35×10^{-6} m s^{-1} in the order expected for silt soils (see Table 5.5). The initial horizontal and vertical hydraulic conductivity values used in the calibration process were guided by results from piezometer slug tests (mean = 1.88×10^{-6} m s^{-1}) conducted on the floodplain and values reported by Domenico and Schwartz (1998) for various geological materials (Table 6.3), both were subject to adjustment during model calibration.

Table 6.3: Representative hydraulic conductivity values for various geological materials (Domenico and Schwartz 1998).

Material	**Hydraulic Conductivity (m s^{-1})**
Gravel	3×10^{-4} – 0.03
Coarse sand	9×10^{-7} – 0.006
Medium sand	9×10^{-7} – 5×10^{-4}
Fine sand	2×10^{-7} – 2×10^{-4}
Silt	1×10^{-9} – 2×10^{-5}
Till	1×10^{-12} – 2×10^{-6}
Clay	1×10^{-11} – 4.7×10^{-9}
Limestone, dolomite	1×10^{-9} – 6×10^{-6}

Specific yield, defined as the volume of water that drains per unit surface area of aquifer per unit decline of water table, primarily represents the release of water under gravity at the phreatic surface (Rushton 2004; DHI 2007b). Specific yield was included in a sensitivity analysis prior to the calibration of the model. An initial value of 0.2 was used in the sensitivity analysis based on typical ranges within the literature for sand (Table 6.4). Specific storage, which is slightly different than specific yield, relates to a unit volume of aquifer, and is defined as the volume of water that drains per volume of aquifer per unit decline of water table under saturated conditions. Thus specific storage represents the water released from storage from the entire column of water in the aquifer, not just at the phreatic surface (Rushton 2004; DHI 2007b). Specific storage was also used in the sensitivity analysis; an initial value of 1.0×10^{-4} m^{-1} was used based on typical values in the literature for sand (Table 6.5). A value of 19 m (ODN) was used for the initial potential head, which was based on average water level observations across the floodplain equivalent to between -0.2 m to -0.7 m below the ground surface.

Table 6.4: Specific yield for various geological materials (Johnson 1967 as reported in Domenico and Schwartz 1998).

Material	**Specific Yield (fraction)**
Gravel, coarse	0.23
Gravel, medium	0.24
Gravel, fine	0.25
Sand, coarse	0.27
Sand, medium	0.28
Sand, medium	0.23
Silt	0.08
Clay	0.03
Limestone	0.14
Peat	0.44
Till, predominantly silt	0.06
Till, predominantly sand	0.16
Till, predominantly gravel	0.16

Table 6.5: Representative specific storage values for various geological materials (Domenico and Mifflin 1965, as reported in Batu 1998).

Material	**Specific storage (m^{-1})**
Plastic clay	2.6×10^{-3} - 2.0×10^{-2}
Stiff clay	1.3×10^{-3} - 2.6×10^{-3}
Medium hard clay	9.2×10^{-4} - 1.3×10^{-3}
Loose sand	4.9×10^{-4} - 1.0×10^{-3}
Dense sand	1.3×10^{-4} - 2.0×10^{-4}
Dense sandy gravel	4.9×10^{-5} - 1.0×10^{-4}
Rock, fissured	3.3×10^{-6} - 6.9×10^{-5}

6.2.6 Boundary conditions

A combination of zero-flux (no-flow) and specified head subsurface boundary conditions were applied around the model domain (Figure 6.13) (e.g. Hammersmark *et al*. 2008). A zero-flow boundary is the default condition and is realistic for watershed boundaries. The zero-flow boundaries are a simplification of the system, but were justified for the summits of the hillsides on either side of the meadows following the assumption that the groundwater divide followed the topographic divide and provided a hydraulic boundary (e.g. Thompson 2012). Similarly, the foundations of the railway embankment defined a physical boundary at the upstream end of the meadows that was assumed to restrict flow into the site. Some subsurface flow perpendicular to the river is, however, possible across the downstream boundary of the floodplain. To facilitate this exchange a constant head boundary of 18.8 m ODN was specified at this location using mean groundwater height of -0.2 m ± 0.17 m from a well transect at the downstream end of the meadow (Figure 6.13). Specified-head and constant-head boundaries can supply an inexhaustible source of water no matter how much water is removed from a system model (e.g. Franke *et al*. 1985). This is unlikely to cause a problem at the downstream boundary of the Hunworth model as the constant head value is based on mean groundwater elevation that fluctuated very little in this region of the floodplain. A manual sensitivity analysis of alternative boundary options (specified head, flux, zero-flow) was performed and demonstrated negligible effects on simulated groundwater elevations across the floodplain beyond the immediate location of the boundary conditions.

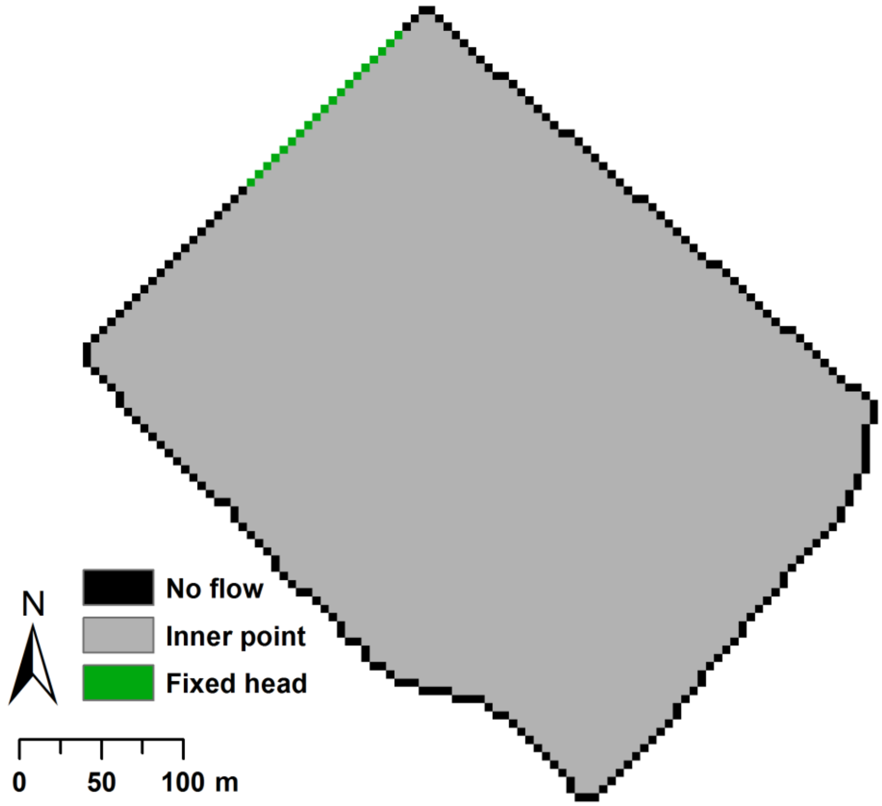


Figure 6.13: Boundary conditions of the MIKE SHE model (grid size = 5 m).

The MIKE SHE drainage option was used to represent relatively small-scale, fast run-off along the base of the hillslope and to route drainage into topographical lows along the agricultural ditch (e.g. Thompson *et al*. 2009). A drainage level and time constant were applied along the base of the hillslope and the ditch, and were altered in the sensitivity analysis and model calibration (Figure 6.14). A drainage level of -1.6 m, and time constant of 6×10^{-8} s^{-1} along the base of the hillslope, with a higher time constant of 2.6×10^{-7} s^{-1} closer to the model boundary, provided the best overall fit.

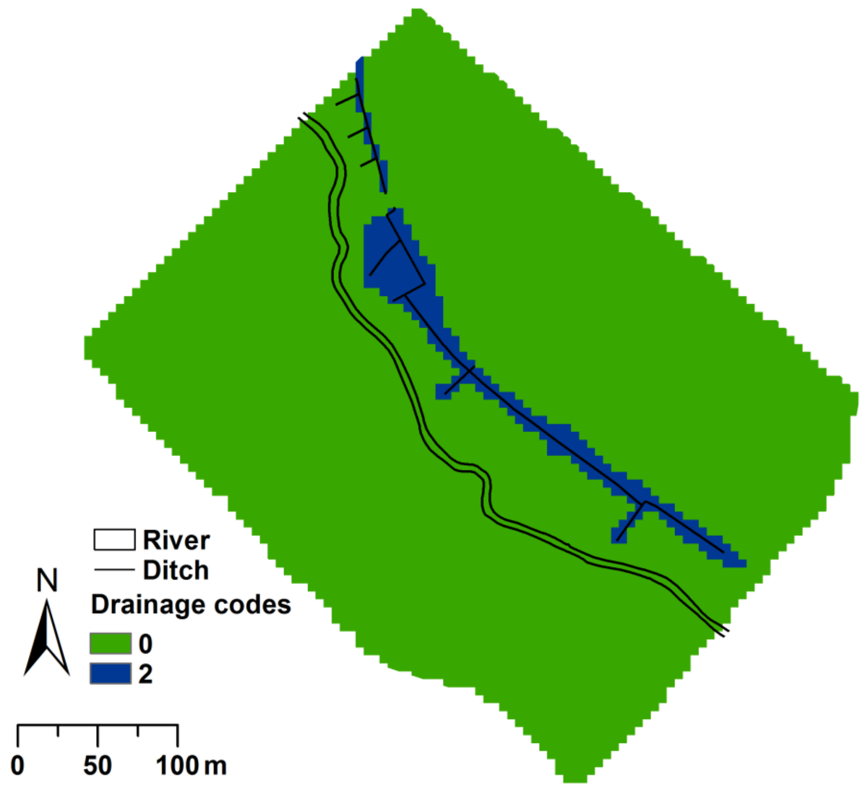


Figure 6.14: Spatial distribution of drainage codes used in the MIKE SHE models along the ditch and pond. Grid cells with code = 2 received drainage flow. Grids cells with code = 0 did not receive drainage flows. Preference for drainage occurred in the following order: river > boundary > depression. Grid cells not connected to river or boundary cells drained to cells with the lowest drain level (to create a pond).

6.2.7 Simulation specification

The MIKE SHE simulations were started using initial conditions (e.g. initial surface water depth) stored from a point in a previous simulation run of the same model in a 'hot start' file. The hot start model was run for a year and the initial conditions for groundwater head were obtained from a representative time period, selected after the model had stabilised. This provided more realistic initial conditions, and prevented the need for a 'warm-up period' at the start of each model. The maximum allowed time steps for the unsaturated flow (using the two-layer water balance method), saturated flow (finite difference) and evapotranspiration components were set at 24 hours. A shorter time step of 0.25 hours was specified for the overland flow (finite difference) component to ensure model stability. However, in flat areas with ponded water, such as on floodplains, the difference in water depth between grid cells is close to zero, which requires very small overland flow time steps. To allow the simulation to run with longer time steps and further reduce numerical instability the calculated overland flows between cells were multiplied by a damping factor to reduce flow between cells when the flow gradient is close to zero. Rather than the default damping function in MIKE SHE, an alternative single parabolic function was specified, which goes to zero more quickly and is consistent with the approach used in MIKE FLOOD (DHI 2007b). This alternative damping function was applied below a specified gradient of 0.001. All model results were stored at 24-hour intervals to coincide with the temporal frequency of observations.

6.3 MIKE 11 model development

6.3.1 River channel and ditch network

Channel flow was simulated with one-dimensional hydraulic MIKE 11 models (Havnø *et al*. 1995). Two MIKE 11 models were developed, one for the embanked river scenario and another for the restored scenario. Dynamic coupling of each MIKE 11 river model and the appropriate (embanked / restored) MIKE SHE model through the exchange of simulated water level at MIKE 11 h-points (points where water level data are calculated along the river branch) and MIKE SHE river links enabled the simulation of river-aquifer exchange, overland flow to channels and inundation from the river onto the floodplain (Thompson *et al.* 2004; DHI 2007a). The MIKE 11 river channels were linked to the MIKE SHE grid via river links at the edge of each grid cell, which simplified the river geometry (e.g. Figure 6.15). However the high resolution of the MIKE SHE grid meant that the river network was represented well by the MIKE SHE river links.

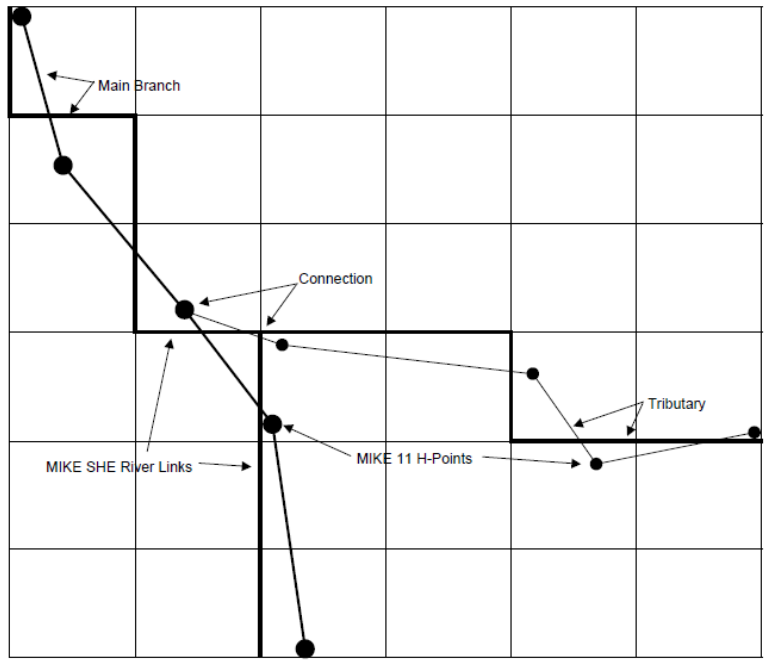


Figure 6.15: An example of MIKE 11 river branches with H points and the corresponding river links in a MIKE SHE model from DHI (2007b).

River-aquifer exchange can be calculated using different river-bed permeability options, which account for differences in riverbed and aquifer conductance. River-aquifer exchange of the River Glaven was simulated using the aquifer only formulation, where the river is assumed to be in full contact with the aquifer material. This was a suitable method given the similarity between river and groundwater chemistry along the river-banks, and the high baseflow index (0.81) and flow exceedance values for Q95 (51%) which indicate high groundwater contributions to discharge at the site (see Section 5.2.2).

A 576 m section of the River Glaven beginning immediately upstream and ending just downstream of Hunworth Meadow was digitised in MIKE 11 from 1:10,000 Ordnance Survey digital data (Land-Line.Plus) (Figure 6.16). The river was divided into three connected sections (i.e. branches): a main river section that was within the MIKE SHE model domain, and an upstream and downstream section that were both outside the MIKE SHE model domain (Figure 6.16). River cross-sections for the two MIKE 11 models representing pre- and post-restoration channel configurations were specified using the results from the dGPS surveys conducted before and after embankment removal (as discussed in Section 4.4.4). A total of 32 river transects were surveyed prior to the restoration, at intervals along the length of the river of approximately 10 m, and 23 transects after the river restoration, at intervals of approximately 15 m (Figure 6.16). The river channel was fairly uniform in width along the reach, nevertheless some variation in the channel bed topography occurred due to patchy deposits of sand and silt (Figure 6.17a).

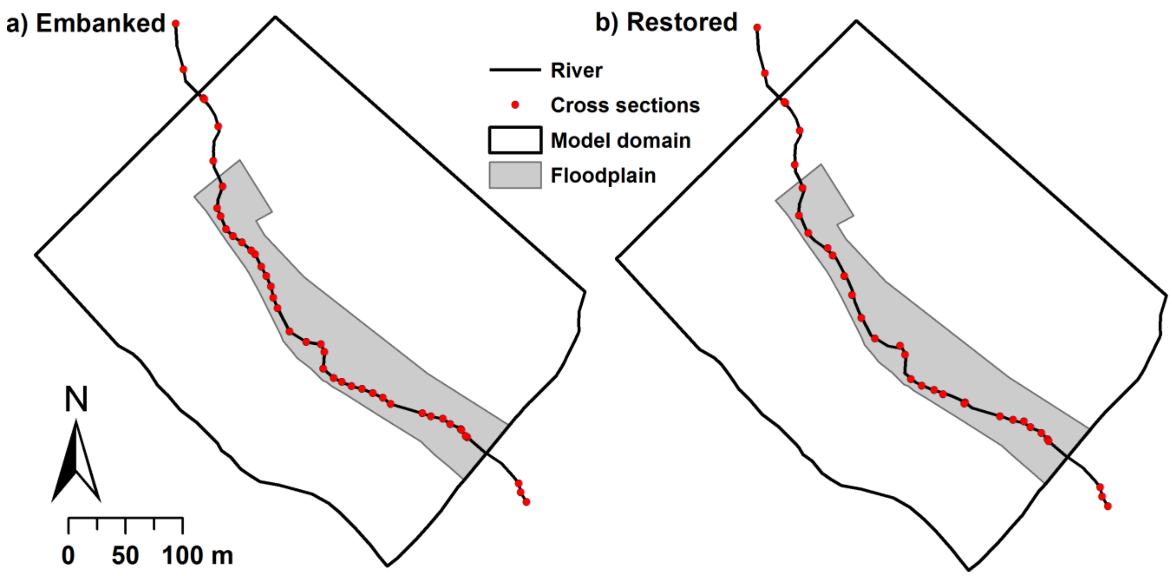


Figure 6.16: Delineation of the MIKE 11 River Glaven channel and location of cross-sections for the (a) embanked and (b) restored models.

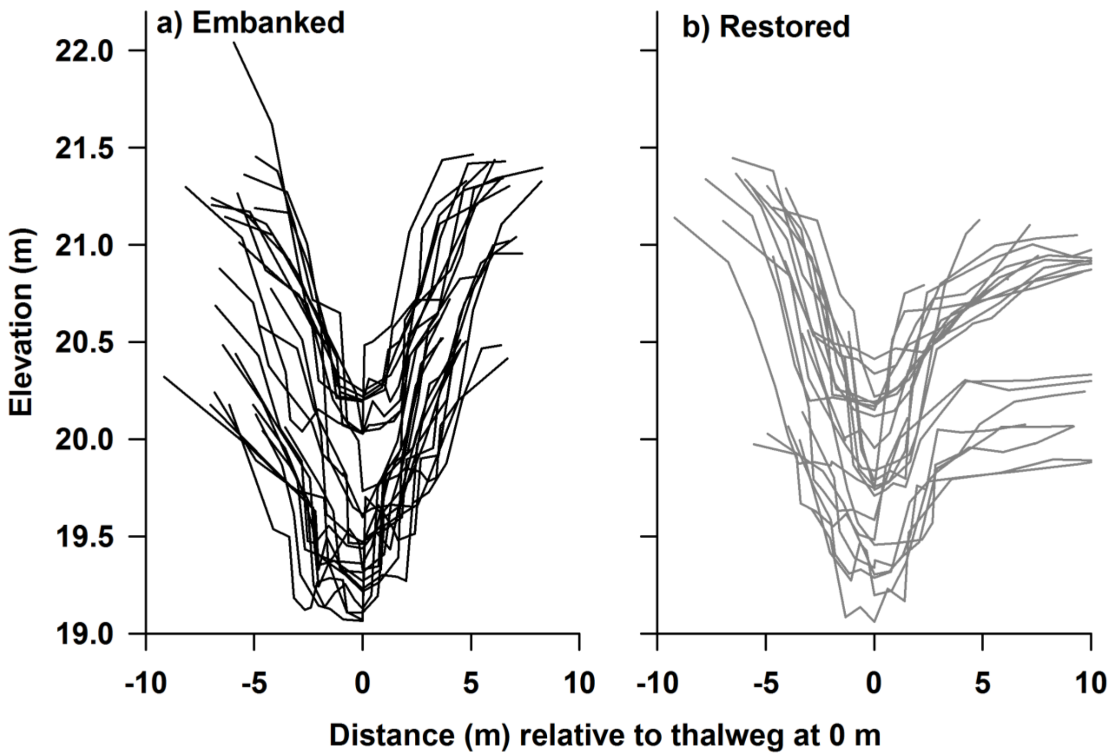


Figure 6.17: MIKE 11 cross-sections for the (a) embanked and (b) restored river-banks.

A discharge boundary condition was specified at the upstream end of the MIKE 11 model using daily discharge records from the Environment Agency gauging station (#034052) located immediately upstream of Hunworth Meadow. A constant water-level boundary condition of 18.6 m AOD was applied at the downstream end of the MIKE 11 model. This level was just above the bed of the river at the lowest cross-section and prevented the river drying out whilst permitting discharge of water from the end of the model (Thompson *et al*. 2004). An initial water depth of 0.2 m throughout the MIKE 11 model at the start of the simulation period was obtained from the records of a stage board installed in river towards the downstream end of the reach.

A blocked agricultural drainage-ditch that runs parallel to the river was also digitised in MIKE 11, and was initially added as a separate MIKE 11 branch, which totalled 415 m in length. The ditch channel, which was not subject to alteration during the river restoration, was surveyed once resulting in 19 cross-sections. The ditch was a uniform width of approximately 4 m, but varied in depth due to varying amounts of organic matter and silt that had accumulated along the bottom (Figure 6.18).

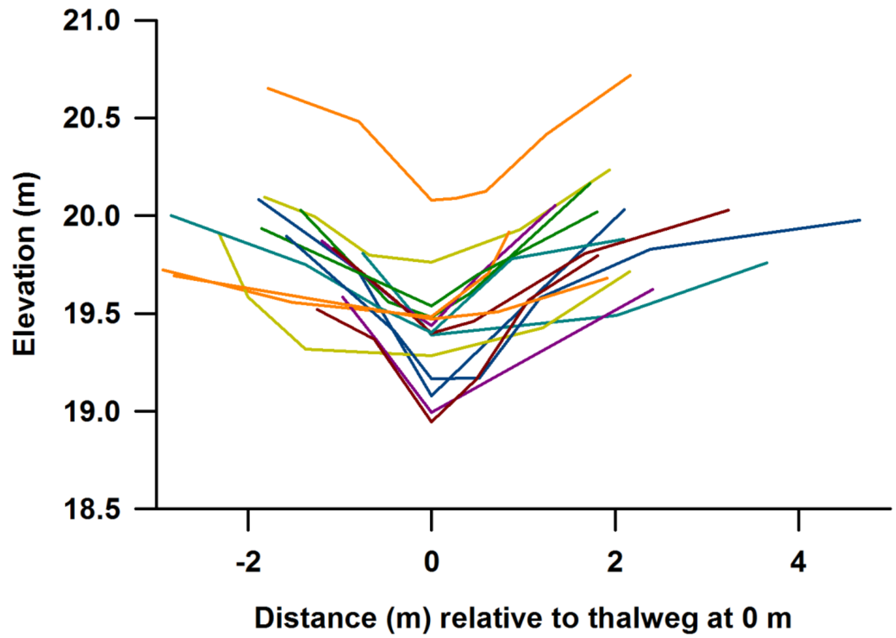


Figure 6.18: MIKE 11 ditch cross-sections.

A zero flow boundary condition was set at upstream end of the ditch channel (as the ditch did not exhibit surface inflow) and MIKE SHE provided flows into the ditch. However, MIKE 11 does not simulate evaporation, and thus periods of wetting and drying in the ditch could not be modelled by MIKE 11. Instead of fluctuating in tandem with groundwater levels (Figure 6.19), the modelled ditch water levels remained constant along the MIKE 11 branch. The MIKE 11 branch of the ditch was therefore removed. The final grid size, as discussed above in Section 6.2.1, was sufficiently small to be able to include the ditch topography in MIKE SHE and simulate fluctuations in the ditch surface water associated with evaporation.
A solinst pressure transducer (Levelogger Gold 3.0, Ontario, Canada) installed in the ditch recorded hourly water levels. The ditch water level observations were

closely coupled with fluctuations in groundwater elevation (Figure 6.19) and were used as an added tool in the calibration and validation process.

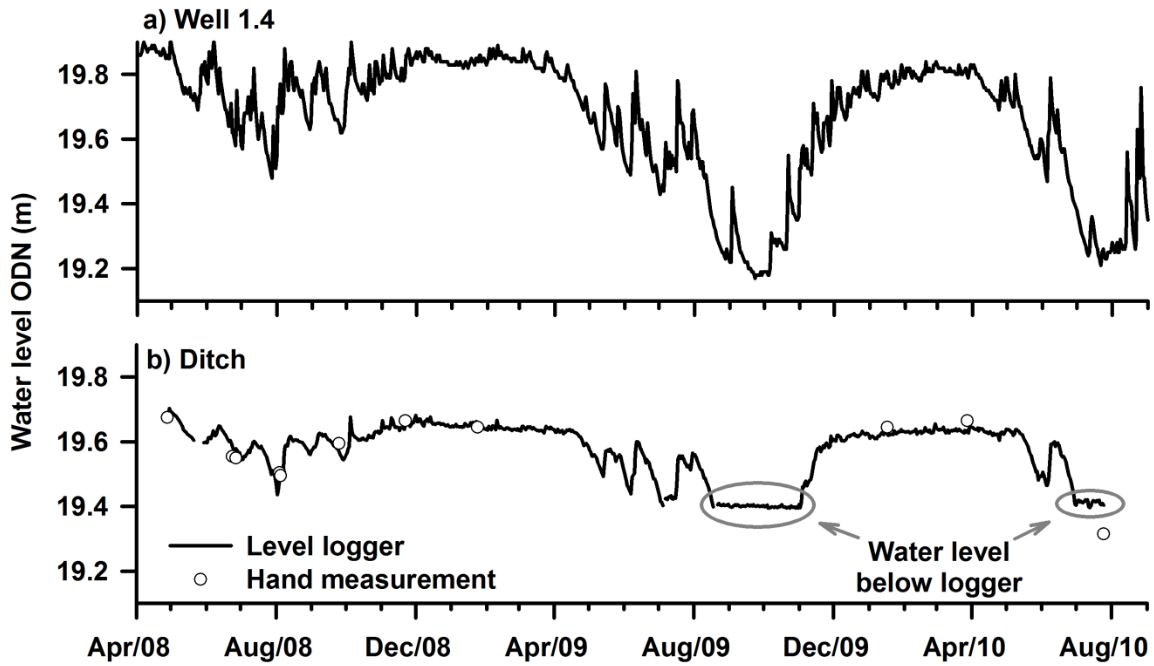


Figure 6.19: Water levels in (a) Well 1.4 and (b) the adjacent ditch.

6.3.2 Specification of hydrodynamic parameters

A dynamic wave approximation, which uses the full Saint Venant momentum equation, was applied to the river to calculate channel flows. The Manning's n coefficient for bed resistance of the river channel usually ranges from 0.01 sec $m^{-1/3}$ for smooth channels (small flow resistance) to 0.10 sec $m^{-1/3}$ for thickly vegetated channels (large flow resistance) (DHI 2007c). A constant Manning's n coefficient for bed resistance of 0.08 sec $m^{-1/3}$ was initially applied to the model, however this value resulted in river levels being too high in the winter and too low in the summer. Instead, a time varying Manning's n coefficient was specified throughout the MIKE 11 model for the entire discharge record from 2001 – 2010 (Figure 6.20) based on the approach used by House *et al*. (2016a), to account for seasonal differences in bed resistance associated with in-stream macrophyte growth. Seasonal macrophyte growth in the river was easily identified in the river discharge record (see Section 4.5, and Figure 5.10), as it impacted the rating curve and resulted in a slow increase in baseflow through the summer, despite low or no rainfall. This effect declined during the autumn due to macrophyte dieback, or more abruptly during flood events due to devegetation of the river channel (e.g. Chambers *et al*. 1991). Two general summer conditions were identified for varying Manning's n values among years: 1) high flow summers where macrophyte growth was restricted, and 2) low flow summers where stable conditions resulted in substantial vegetation growth. A Manning's n coefficient of 0.058 sec $m^{-1/3}$ was applied in the winter, and maximum values of 0.08 sec $m^{-1/3}$ and 0.15 sec $m^{-1/3}$ were applied in June during high flow and low flow summers, respectively (Figure 6.20). These values were guided by the fit between observed and simulated river stage, and were within the range of 0.045 to 0.353 $m^{1/3}$ s^{-1} reported for a UK chalk stream by House *et al*. (2016a). The growth period was defined as April – September and Manning's n values during this period were interpolated between the winter and summer values, which was guided by macrophyte growth measurements in a UK chalk stream reported by Flynn *et al*. (2002).

It is worth noting that the modelled river levels were simulated from average daily stage measurements recorded at the EA gauging station, whereas the observed data were from point measurements taken at a stage board adjacent to the downstream well transect (see Figures 3.1 and 4.2). Hence an exact fit between observed and modelled data was not expected, particularly during high flow events where river stage was likely to change substantially over a 24hr period. Selection of the Manning's n parameter was therefore based on the value that provided a good fit between modelled and measured river levels during periods of relatively stable river flows when point measurements were likely to be closest to mean daily values.

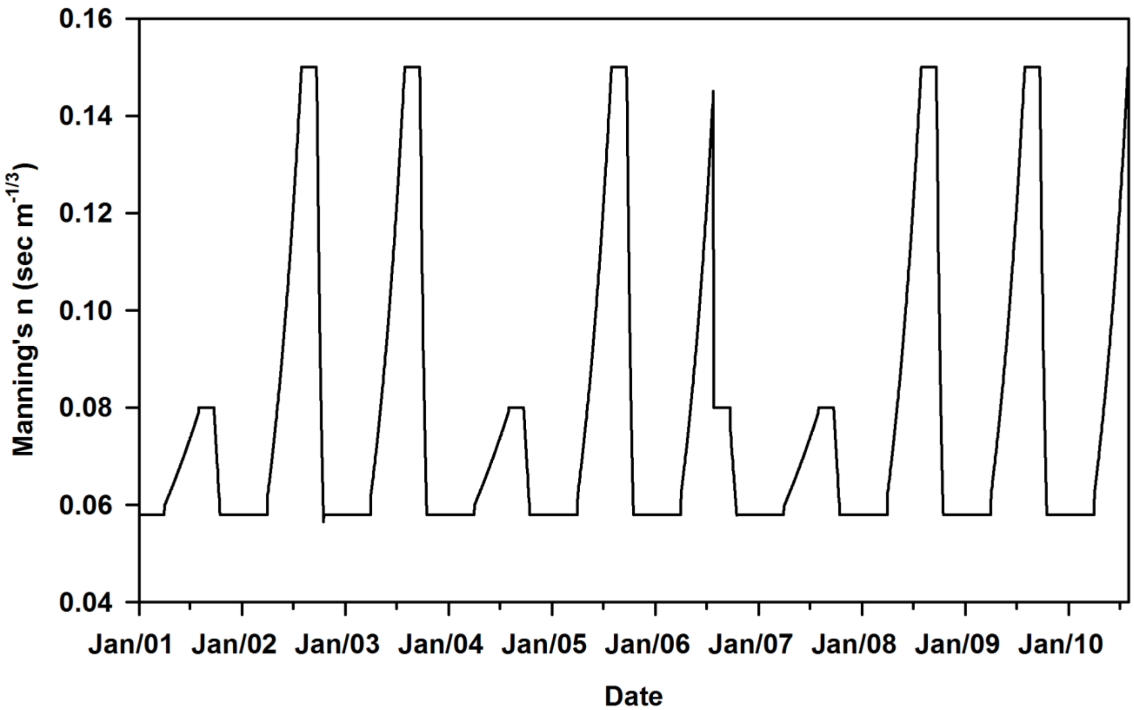


Figure 6.20: Time series of variable Manning's n values used in the MIKE 11 model to account for seasonal (summer) and intrannual (wet versus dry summer) variation in-stream macrophyte growth.

Using the approach adopted by Thompson *et al.* (2004) flood codes were used to specify MIKE SHE model grid cells which could be directly inundated from the MIKE 11 model. A delete value of 2 was given to grid squares in MIKE SHE that would not flood, i.e. beyond the ditch, and the hillslope (Figure 6.21). Potentially flooded cells (flood code value = 10) comprised the immediate riparian area, which included the grid cells through which the river ran, those coincident with embankments (if present) and the zone up to 10 m (two grid cells) onto the meadow (Figure 6.21). These MIKE SHE grid cells were flooded by the river if water levels simulated by MIKE 11 were higher than the corresponding MIKE SHE grid surface level. Once a grid cell was flooded, the overland flow component of MIKE SHE could simulate surface water movement onto adjacent model grid cells further away from the river. Infiltration and evapotranspiration from flooded cells would also be simulated in the same way as if flooding occurred from precipitation and surface runoff or the water table reaching the ground surface (Thompson, 2004).

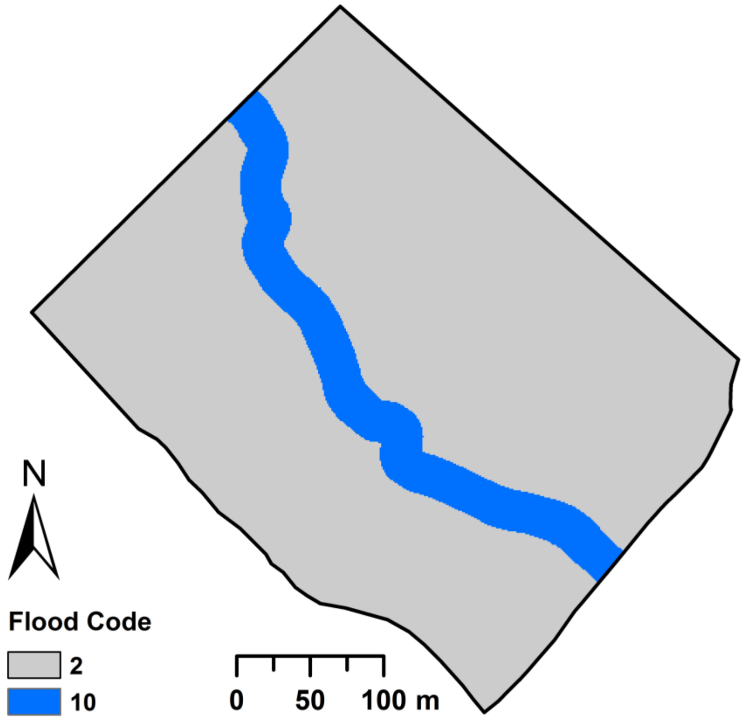


Figure 6.21: Spatial distribution of floodcodes in the pre-restoration MIKE SHE model. A value of 10 indicates the area assigned a flood code value.

The MIKE 11 models were set-up to run at 1 min time steps from 22/02/2007 to 15/03/2009 prior to the embankment removal and from 29/03/2009 to 25/07/2010 after the embankment removal. Once coupled to the MIKE SHE model, the specified MIKE SHE time-step over-rode the MIKE 11 simulation time settings to store river flow and water levels at hourly intervals.

6.4 Model calibration and parameter optimisation

6.4.1 Sensitivity analysis

A sensitivity analysis was performed as an initial step in the calibration procedure to identify the most important model parameters to be included in the calibration of the model. This important step is necessary to reduce to the computational time required to undertake the calibration process. The sensitivity analysis was performed using the MIKE ZERO autocalibration program, AUTOCAL. A total of 18 parameters were included in the sensitivity analysis. Each parameter was successively varied in AUTOCAL around an initial value within a lower and upper range of physically possible values, which are outlined above and summarised in Table 6.6.

Table 6.6: List of model parameters and their initial, lower and upper limit values used in the AUTOCAL sensitivity analysis.

Parameter	**Initial**	**Lower limit**	**Upper limit**
Overland Manning M ($m^{1/3}\ s^{-1}$)	3	1	4
Detention storage (mm)	5.0×10^{-3}	5.0×10^{-3}	0.3
Water content at saturation (volumetric)	0.3	0.3	0.8
Water content at field capacity (volumetric)	0.2	0.1	0.41
Water content at wilting (volumetric)	0.07	0.04	0.2
Infiltration ($m\ s^{-1}$)	2.0×10^{-6}	1.0×10^{-11}	0.03
Evapotranspiration surface depth (m)	0.5	0.39	1.9
Geological layer lower level (m)	-10	-100	-1
Horizontal hydraulic conductivity ($m\ s^{-1}$)	1.9×10^{-6}	1.0×10^{-11}	0.03
Vertical hydraulic conductivity ($m\ s^{-1}$)	3.8×10^{-7}	2.0×10^{-12}	4.0×10^{-3}
Specific yield (fraction)	0.2	0.03	0.3
Specific storage (m^{-1})	1.0×10^{-4}	4.9×10^{-5}	0.25
Initial head (m ODN)	20	15	21
Fixed head (m ODN)	19	17	21
Drainage level (m)	-0.4	-2	0
Drainage time ($1\ s^{-1}$)	1.0×10^{-7}	1.0×10^{-7}	1.0×10^{-6}
River water depth (m)	0.17	0.1	0.6
Riverbed resistance ($m^{1/3}\ s^{-1}$)	0.08	0.01	0.1

A forward difference approximation was used to evaluate the sensitivity coefficients, as described in DHI (2007d), in which values for the model parameters are individually perturbed and the resulting root mean square error (RMSE) of the simulation are compared to a control simulation that contains the initial parameter values that were not perturbed. RMSE was calculated in the MIKE ZERO autocalibration program as:

$$RMSE = \sqrt{\frac{1}{N}\sum_{i=1}^{N} w_i (TARGET_i - SIM_i)^2} \qquad (6.2)$$

where RSME is the root mean square error for model$_i$, $TARGET_i$ is the target value, in this case observed groundwater levels, and SIM_i, is the simulated value (i.e. water level), and w_i is the weight assigned in the calculation of the output measure when SIM_i, ≤ $TARGET_i$ or when SIM_i, > $TARGET_i$ (DHI 2007d).

AUTOCAL calculates local sensitivity coefficients for each parameter with respect to the measured and modelled water table elevation and objective functions i.e. RMSE. In order to compare the local sensitivity coefficients for each parameter, AUTOCAL computes scaled sensitivity coefficients for each parameter of the specified output measures (i.e. water levels) and objective functions (i.e. RMSE). The scaled sensitivity values ($S_{i,scale}$) were calculated in AUTOCAL as:

$$S_{i,scale} = S_i(\theta_{i,upper} - \theta_{i,lower}) \qquad (6.3)$$

where S_i is the calculated un-scaled sensitivity coefficient, and $\theta_{i,upper}$ and $\theta_{i,lower}$ are the upper and lower values of the parameter. The scaled sensitivity coefficients are ranked with respect to the importance of the parameter. Higher sensitivity values, i.e. the distance from zero, indicate more sensitive parameters. Parameters are considered insensitive if their scaled sensitivity value is <0.01 – 0.02 times the maximum scaled sensitivity value (DHI 2007d).

The most sensitive parameters in descending order were horizontal hydraulic conductivity, drainage level, drainage time, Manning M, vertical hydraulic conductivity, and soil water content at saturation (Table 6.7). A manual manipulation of the most sensitive parameters selected by AUTOCAL was conducted in the MIKE SHE model in order to confirm the results of the AUTOCAL sensitivity analysis. Subsequently, the insensitive parameters were set to constant values using the initial values in Table 6.6, which as discussed throughout this section were established from the field observations and the literature, whereas the most important parameters outlined above were adjusted during the following parameter optimisation.

Table 6.7: Scaled sensitivity coefficients for parameter used in the AUTOCAL sensitivity analysis. Greater RMSE (absolute values), i.e. the distance from zero, indicate more sensitive parameters. Parameters are considered insensitive if their scaled sensitivity value is <0.01 – 0.02 times the maximum scaled sensitivity value (absolute values).

Parameter	RMSE Aggregate
Drainage level	1.14
Water content at saturation	0.11
Infiltration	0.09
Specific storage	0.02
Initial head	0.01
Fixed head	0.01
Detention storage	0.00
Specific yield	0.00
River water depth	0.00
Riverbed resistance	0.00
Water content at wilting point	-0.03
Evapotranspiration surface depth	-0.06
Geological layer lower level	-0.10
Vertical hydraulic conductivity	-0.14
Overland Manning M	-0.14
Drainage time	-0.96
Horizontal hydraulic conductivity	-3.06

6.4.2 Model *calibration and validation*

A detailed account of the MIKE SHE model results and comparisons of the observed and simulated groundwater elevations is provided in Section 7. Calibration and initial model validation was undertaken using the embanked model using a split sample approach. The 13-month period 22 February 2007 – 14 March 2008 was used for calibration and the following 12 months (15/03/2008 – 15/03/2009) for validation. The end of this period coincided with embankment removal so that calibrated parameter values were specified within the model representing restored conditions with the subsequent 16 months (29/03/2009 – 25/07/2010) providing a second validation period. As described above, a number of model parameters were varied during model calibration.

Initial calibration was undertaken using an automatic calibration procedure that was based on the shuffled complex evolution method with the optimal parameter set being selected according to the lowest aggregate RMSE (a measure of the average magnitude of error) for the comparisons between observed and simulated groundwater and river water levels (Duan *et al.* 1992; Madsen 2000 2003; DHI 2007c). This approach was undertaken for the coarser 15 m × 15 m model grid to reduce the computational time due to the number individual model runs (n=480) required for the automatic calibration routine to determine an optimal parameter set.

Interpolation of surface elevation over a coarse grid size of 15 × 15 m in MIKE SHE resulted on average in a 0.25 m over-estimate of surface elevation at the well locations, with greater absolute error at the base of the hillslope (absolute error = 0.53 to 1.16 m) and on the river embankments (absolute error = 0.53 m), which is represented by the outliers in Figure 6.22. At a grid spacing of 15 × 15 m multiple wells were represented by the same grid cell, which did not allow for variation in water levels between these wells. For this reason, three wells (one from each well transect) located in the middle of the meadow and away from the embankments and hillslope were used in the autocalibration process; the interpolated surface elevation at these grid cells had an absolute error <1 cm. The models were run with the selected parameter sets using the Shuffled Complex Evolution optimization method in the MIKE ZERO AUTOCAL program (DHI 2007d). The optimal parameter values in the autocalibration after 420 runs were selected according to the lowest aggregate RMSE scores, which equaled 0.027. These final MIKE SHE and MIKE 11 parameter values for the automatically calibrated model are summarised in Table 6.8.

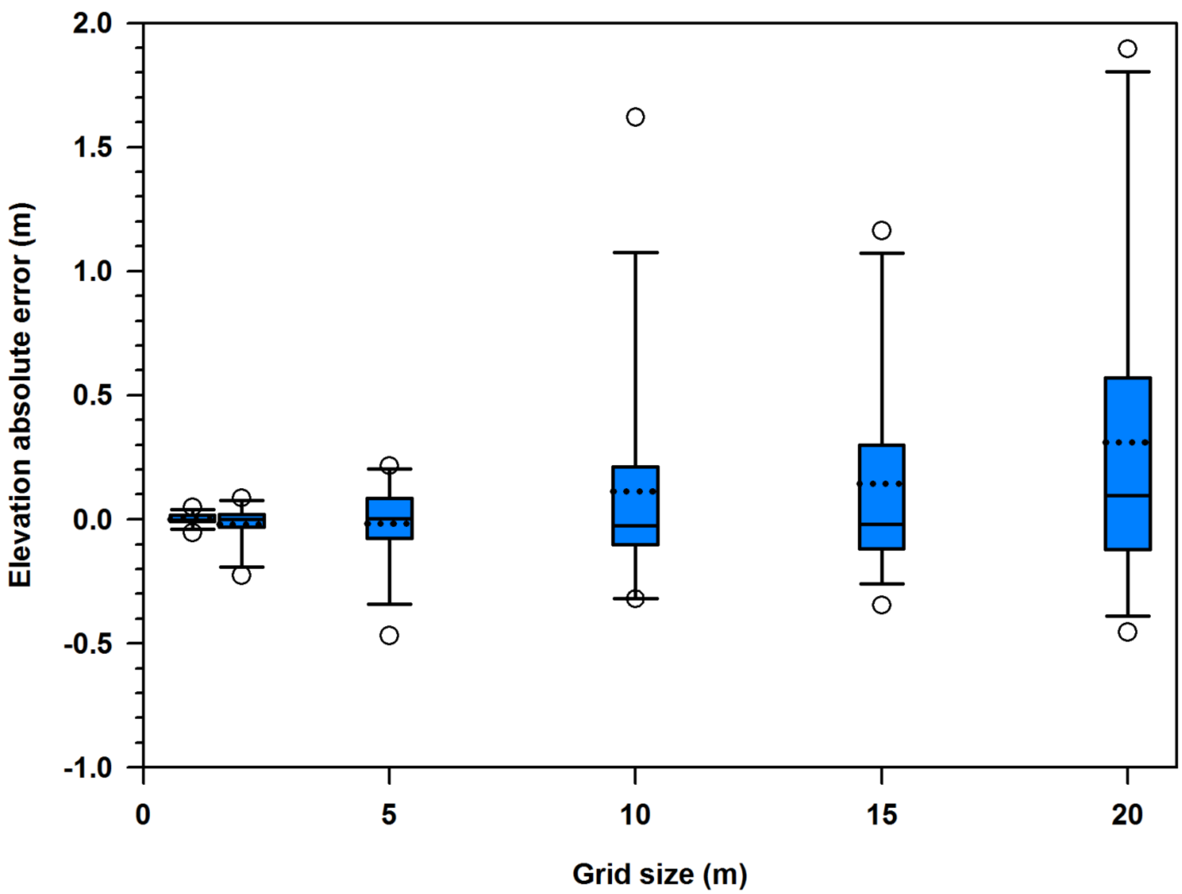


Figure 6.22: Increase in elevation absolute error with grid size. The solid centre line and broken line within the boxplot indicate the median and mean, respectively. The box extent and error bars denote the 25th and 75th percentiles, and 10th and 90th percentiles, respectively. The circles indicate the presence of outliers.

Table 6.8: MIKE SHE and MIKE 11 parameter values for the automatically calibrated 15 x 15 m grid model.

Parameter	**AUTOCAL final values (15 × *15 grid)***
<u>MIKE SHE</u>	
Overland Manning coefficient ($m^{1/3}\ s^{-1}$)	3.0
Water content at saturation (volumetric)	0.31
Water content at field capacity (volumetric)	0.26
Water content at wilting (volumetric)	0.05
Infiltration ($m\ s^{-1}$)	1.0×10^{-5}
Evapotranspiration surface depth (m)	1.10
Horizontal hydraulic conductivity ($m\ s^{-1}$)	9.0×10^{-7}
Vertical hydraulic conductivity ($m\ s^{-1}$)	3.8×10^{-7}
Drainage level (m)	-0.1
Drainage time ($1\ s^{-1}$)	5.9×10^{-7}
<u>MIKE 11</u>	
Riverbed resistance ($m^{1/3}\ s^{-1}$)	0.08

Following the autocalibration, the model grid size was reduced to 5 × 5 m and the calibration was checked manually with model performance being assessed statistically using the RMSE, the correlation coefficient (R), and the Nash-Sutcliffe efficiency coefficient (NSE) (Nash and Sutcliffe 1970) with data from all 14 wells. These key statistics assess different aspects of the model performance (bias, correlation, goodness of fit), and have been widely used in similar studies including those where optimised parameter values from auto-calibration routines are refined manually (Rochester 2010; Thompson 2012; Thompson *et al.* 2014; House *et al*. 2015). In general, the 5 × 5 m grid provided good agreement between the measured and modelled surface elevation, with an average absolute error of -0.02 m (Figure 6.21), each well located within a separate grid cell, and clearly defined river and ditch topography. However, the surface elevation of wells located at the base of the woodland hillslope (i.e. Wells 3.3 and 2.4) and some of the wells positioned next to the ditch (i.e. Wells 2.3 and 1.3) still exhibited greater elevation absolute error than the rest of the wells, with respective errors of approximately +0.2 m and -0.5 m. This was taken into account during the manual fine-tuning of the model, and a less accurate fit between the modelled and measured water surface elevations was expected at these wells. The final values of the calibration terms defined at the end of this process are summarized in Table 6.9. The final calibrated distributed manning's M values for overland flow are sensible for grassland (3.0 $m^{1/3} s^{-1}$) and light underbrush (2.5 $m^{1/3} s^{-1}$) given in the literature (USDA, 1986, Thompson, 2004). The volumetric soil water content for saturation, field capacity and wilting point are within an appropriate range for a sandy loam soil based on the laboratory results from the water release characteristic (see Section 6.2.4) and values reported for sandy loam soils (e.g. Zotarelli *et al*. 2010).

The calibrated evapotranspiration surface depth (i.e. the height of the capillary zone) of 1.1 m is between the values of 0.4 m – 1.9 m calculated for Hunworth Meadow using Hazen's formula (Das 2002), and comparable to values of capillary rise between 0.5 – 2.0 m for sandy loams (Chubarova 1972; Brouwer *et al*. 1985). Furthermore, the final hydraulic conductivity values are in accordance with values provided in the literature for medium sand soils (ie. 9×10^{-7} – 5×10^{-4} m s^{-1}) (Domenico and Schwartz 1998), and field tests at the River Glaven restoration site (see section 6.2.5). Lastly, the time varying river bed resistance approach, described in Section 6.3.2, was essential to account for variation in riverbed resistance to flow associated with seasonal and interannual differences in macrophyte abundance observed at the River Glaven. The final values of between 0.058 sec $m^{-1/3}$ and 0.15 sec $m^{-1/3}$ (see Figure 6.20) are sensible for a small stream with high macrophyte growth in the summer (e.g. House *et al*. 2016a).The same statistical measures were subsequently employed to assess model performance for both of the validation periods.

Once the model parameters were selected, the performance of the model was assessed for the two validation periods from statistical comparisons of simulated and measured water using RMSE, the correlation coefficient (R), and the Nash-Sutcliffe efficiency coefficient (R2) (Nash and Sutcliffe 1970). Subsequently, key components of surface water and groundwater hydrology, such as groundwater recharge, surface runoff, and floodplain storage were simulated at daily time steps and compared between the embankment and restored floodplain scenarios. A detailed account of the MIKE SHE/MIKE 11 model results are presented in Chapter 7.

Table 6.9: Final calibrated MIKE SHE and MIKE 11 parameter values.

Parameter	***Value***
MIKE SHE	
Overland Manning M ($m^{1/3}\ s^{-1}$)	3.0 (grass) 2.5 (light underbrush)
Water content at saturation (volumetric)	0.40
Water content at field capacity (volumetric)	0.20
Water content at wilting (volumetric)	0.07
Saturated hydraulic conductivity ($m\ s^{-1}$)	1×10^{-5}
Evapotranspiration surface depth (m)	1.0
Horizontal hydraulic conductivity ($m\ s^{-1}$)	9×10^{-7}
Vertical hydraulic conductivity ($m\ s^{-1}$)	1×10^{-7}
MIKE 11	
Riverbed resistance ($m^{1/3}s^{-1}$) (Time varying)	0.058 – 0.15*

**(Time varying)*

6.5 Impact assessment of embankment removal

The hydrological effects of removing the embankments along the River Glaven were investigated by running the two MIKE SHE / MIKE 11 models representing pre- and post-restoration conditions for the same extended period with identical climatic and river-flow conditions. This method avoids the differences in simulated hydrological conditions that are due to interannual climate variability within the pre- and post-restoration periods used for model calibration and validation. For example, 2007 and 2008 (pre-restoration) were characteristically wetter than 2009 and 2010 following restoration, with total annual precipitation of 880, 784, 684, and 606 mm respectively (Table 6.10). Total annual river discharge for 2007, the period of record used in the model calibration, was one of the highest on record (Table

6.10). Simulating pre- and post-restoration for the same period therefore enables the effects of embankment removal alone to be assessed.

Table 6.10: Total annual precipitation at Mannington Hall (<10 km from the study site) and river discharge at Hunworth from 2001 – 2010.

Hydrological year	Precipitation (mm)	Discharge ($m^3\ s^{-1}$)
2001	614	101
2002	766	91
2003	822	87
2004	721	94
2005	703	73
2006	954	110
2007	880	103
2008	784	92
2009	684	91
2010	606	N/A*

**Discharge data not available from August 2010 onwards (see Section 4.2).*

The simulation period for this assessment was the decade 2001 – 2010. As for the calibration and validation periods, the upstream boundary condition of the MIKE 11 model was specified as mean daily discharge at the Hunworth gauging station. In the absence of data from the local automatic weather station, daily precipitation and Penman-Monteith potential evapotranspiration were derived from records from the Mannington Hall meteorological station. Comparisons of the meteorological data collected on site at the Hunworth weather station and the UK Met Office weather station at Mannington Hall (ca. 10 km from the study site) were conducted to validate the use of off-site weather data, which covered a longer period of record, and were required to extend the period of simulation in the MIKE SHE/11 models. Hunworth and Mannington Hall weather stations showed excellent agreement in temperature ($r^2 = 0.91$) (Figure 6.23), but more variation among the precipitation data ($r^2 = 0.41$) (Figure 6.24). This was largely due to a slight offset in the timing of precipitation events, which may be expected when using an off-site weather station. However, the magnitude of rainfall measured at the Mannington Hall station compared very well with the on-site station and was considered representative of the upstream meteorological conditions affecting discharge dynamics at Hunworth study site.

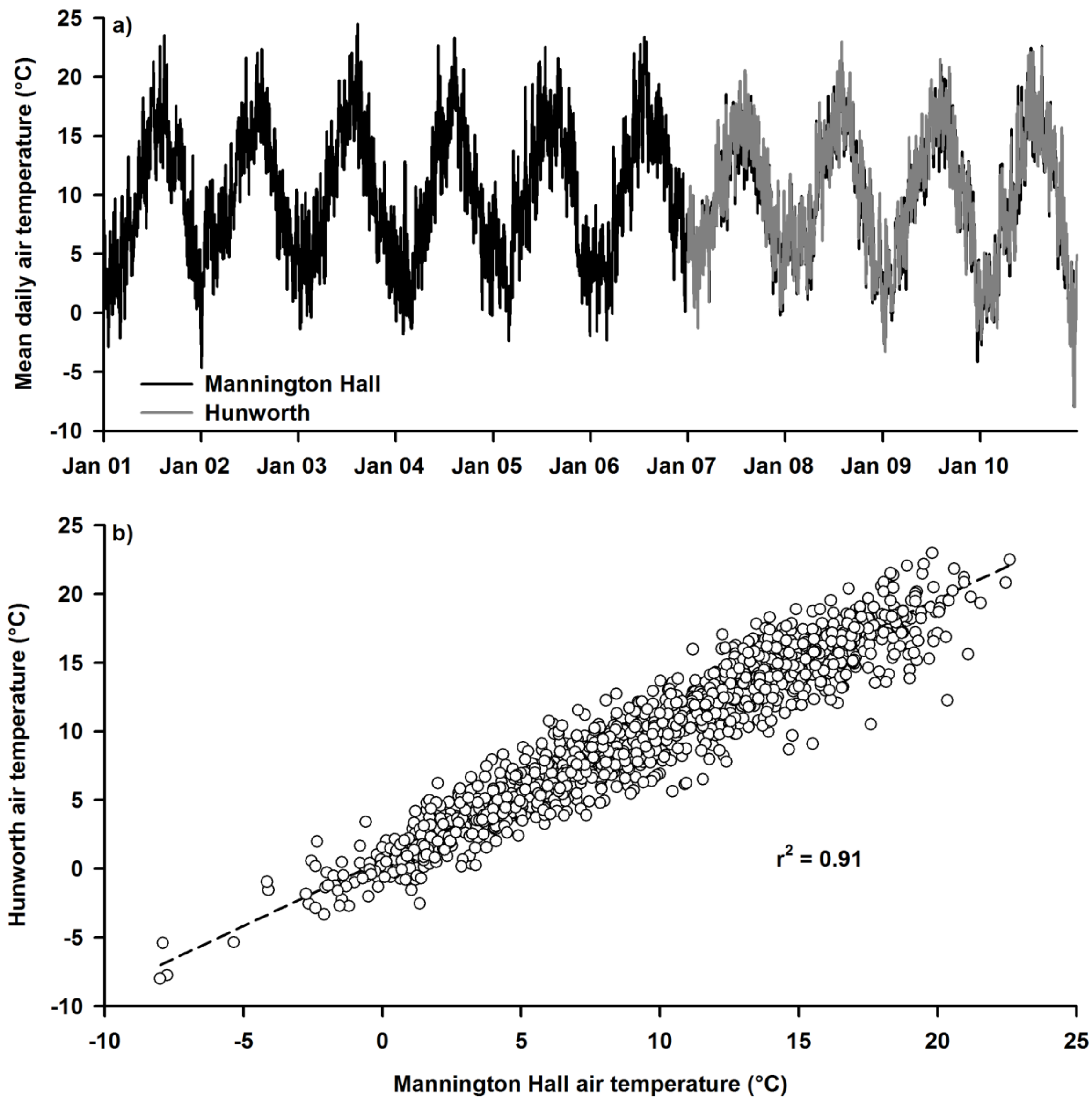


Figure 6.23 (a) Time series of mean daily air temperature at Hunworth Meadow and Mannington Hall, and (b) relationship between air temperature at Mannington Hall and Hunworth.

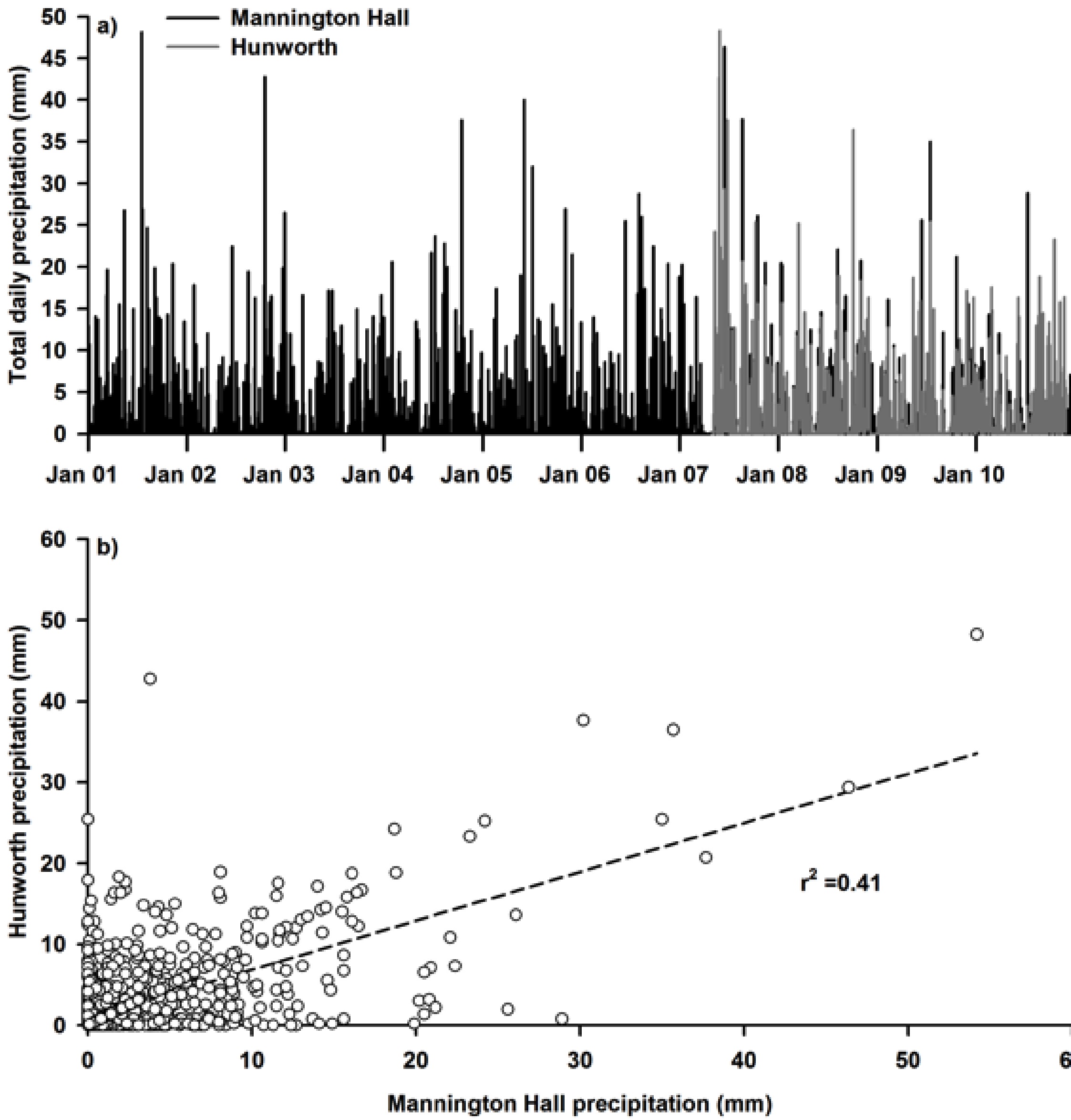


Figure 6.24: (a) Time series of total daily precipitation at Hunworth Meadow and Mannington Hall, and (b) relationship between precipitation at Mannington Hall and Hunworth.

River discharge data at Hunworth was affected by macrophyte growth for short periods during the summer in some years (see Section 5.2.4). In order to run the MIKE SHE/11 models continuously from 2001 to 2010, gaps in the Hunworth discharge data resulting from macrophyte growth was plugged using a regression established between river discharge at the EA gauging station at Bayfield (station

#:3034016) and the onsite EA gauging station at Hunworth (located ca. 5 km apart) during low flow conditions ($y = 0.4087x + 0.0396$; $r^2 = 0.86$) (Figures 6.25 and 6.26). As discussed in Section 3.5, river discharge at Bayfield was not rated above 0.3 m (ca. 0.65 $m^3\ s^{-1}$). However this did not affect the data used as the missing data for Hunworth only occurred during low flow conditions in the summer when macrophyste growth was at its greatest, and accordingly only baseflow data was used in the regression (Figure 6.25).

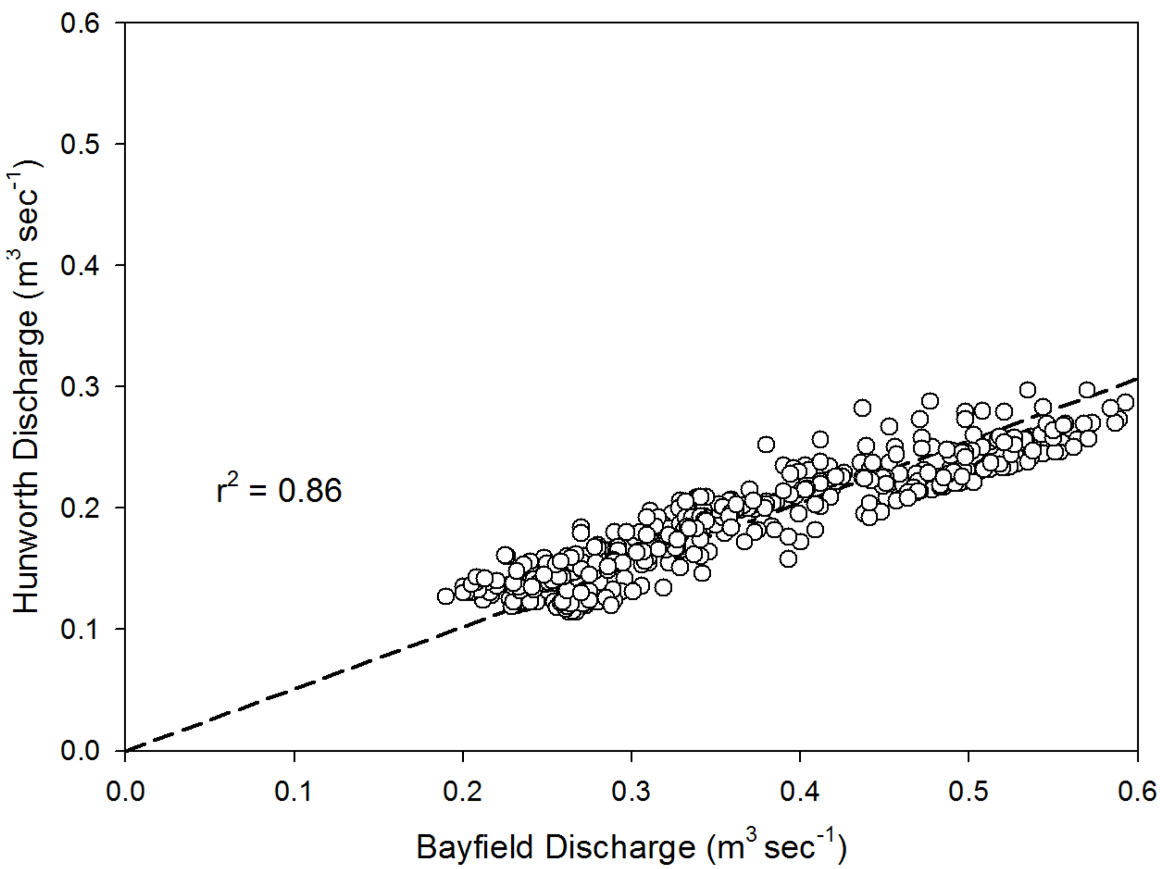


Figure 6.25: Simple linear regression between river discharge at Bayfield and Hunworth gauging stations ($y = 0.4087x + 0.0396$; $r^2 = 0.86$).

River discharge data at Hunworth was not usable from August 2010. A second-phase in-stream restoration project conducted in August 2010, which involved extensive geomorphological changes to the river channel (narrowing, depth diversification, and re-meandering), inadvertently caused water to back-up over the weir which affected the rating curve at the gauging station. This resulted in inaccurate discharge calculations, which can clearly been seen in the sudden increase in discharge after the restoration in August 2010 (Figure 6.26). Unfortunately Bayfield is the only other gauging station on the River Glaven, and as this section of the river was not rated above discharges of approximately 0.65 $m^3 s^{-1}$, the River Glaven does not currently have a fully operational gauging station. The available meteorological and river discharge data from 2001 – 2010 represented a range of climate and river-flow conditions, including extreme high and low river-flow years, which enabled the simulation of a spectrum of probable flow conditions expected on the floodplain under both pre- and post-restoration conditions. For example, during this period, some of the wettest years in the UK occurred since records began in 1910 (Figure 6.27a). In East Anglia, the region of the study site, total annual rainfall in 2001 averaged 780 mm, and was second highest to the 2012 record of 810 mm. Wet years also occurred in 2004, 2007, and 2008, and were substantially above the baseline average for the region (Figure 6.28). The driest contemporary years occurred in 2003, 2005, and 2011 and were within 3 – 18% of the driest five years on record, which averaged 346 mm (Figure 6.28a). In the UK, five of the sixth wettest years have occurred since 2000, and eight of the warmest years have all occurred since 2002. Indeed, 2014 was the wettest winter and warmest year in the UK for over 100 years (Figure 6.27a and b).

Seasonal rainfall in East Anglia is highly variable, but appears to have increased in winter and summer months since 2000. Less obvious trends were evident for spring and autumn (Figure 6.28). Contemporary river discharge and climate data for the River Glaven provided a realistic range of expected hydrological conditions, including extreme heavy precipitation events (e.g. summer 2007), and periods of drought (e.g. summer/autumn 2005) that were used to predict the effects of high and low flow scenarios on the floodplain hydrology following the removal of the river embankments.

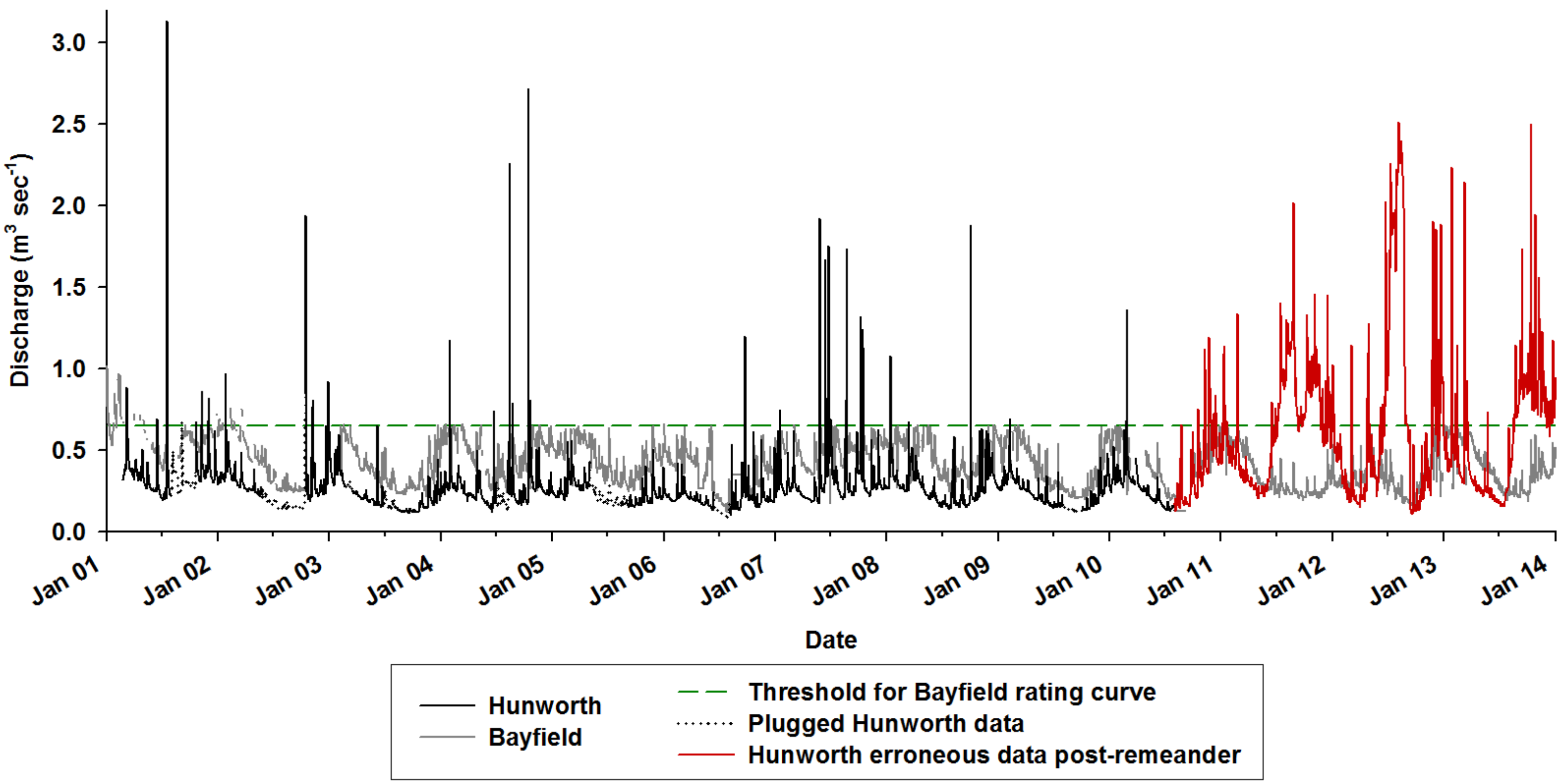


Figure 6.26: Time series of mean daily discharge on the River Glavenat Hunworth (Station #: 034052) and Bayfield (Station #: 034016) Environment Agency gauging stations from 2001 to 2014. River discharge at Bayfield was not rated above 0.3 m (ca. 0.65 m^3 s^{-1}). Measurements of river discharge at Hunworth were affected by a restoration project (remeandering of the river channel) immediately downstream of the gauging station in August 2010.

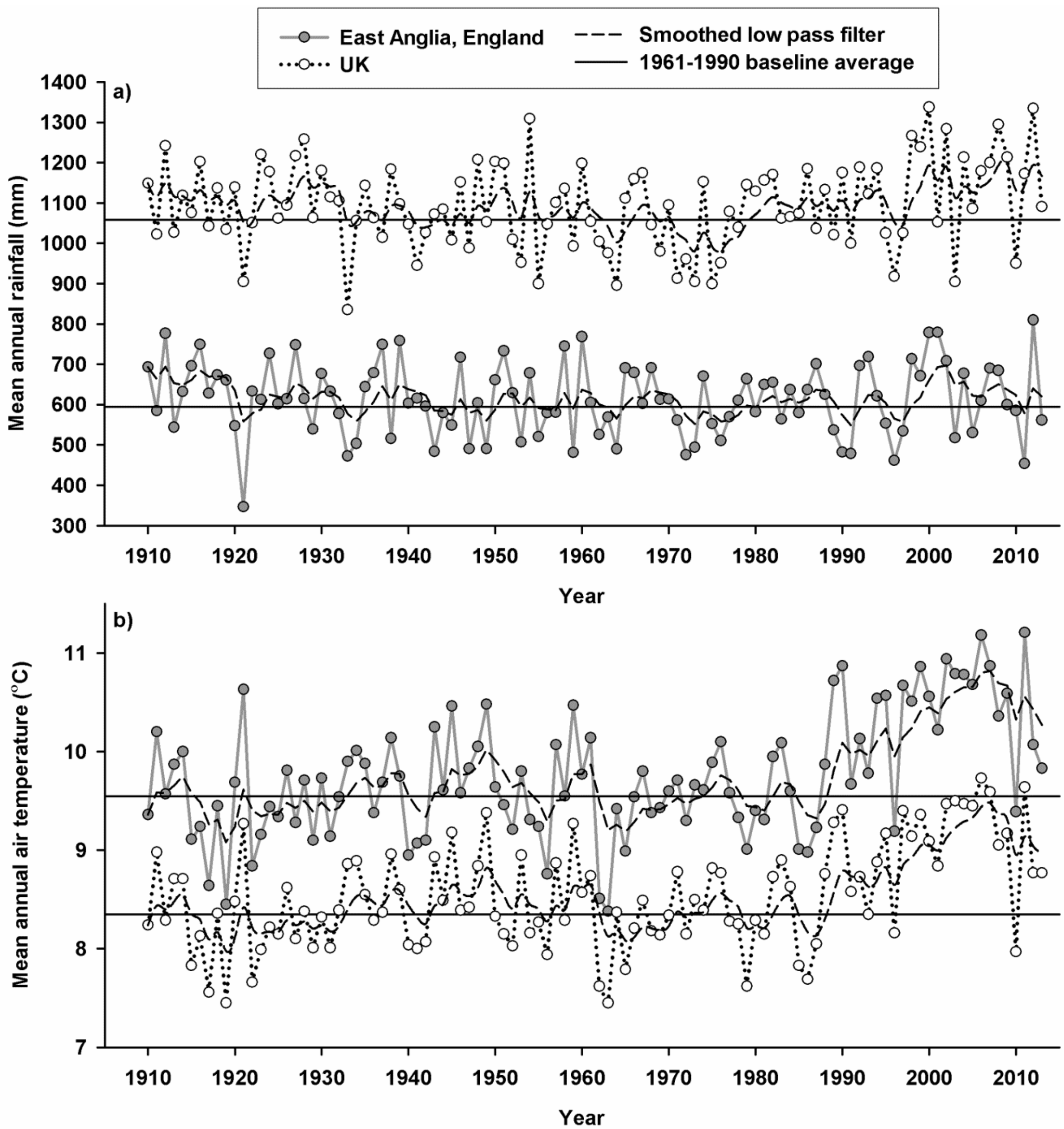


Figure 6.27: Time series of mean annual rainfall (a) and air temperature (b) for East Anglia, England (location of the study site) and the United Kingdom. Data are plotted relative to the average of the 1961 – 1990 baseline period (Jones *et al*. 1999). Data were smoothed with a first order low pass recursive filter to highlight trends in the data relative to the 1961 – 1990 average. Data are from the Met Office regional climate summaries (Met office 2015).

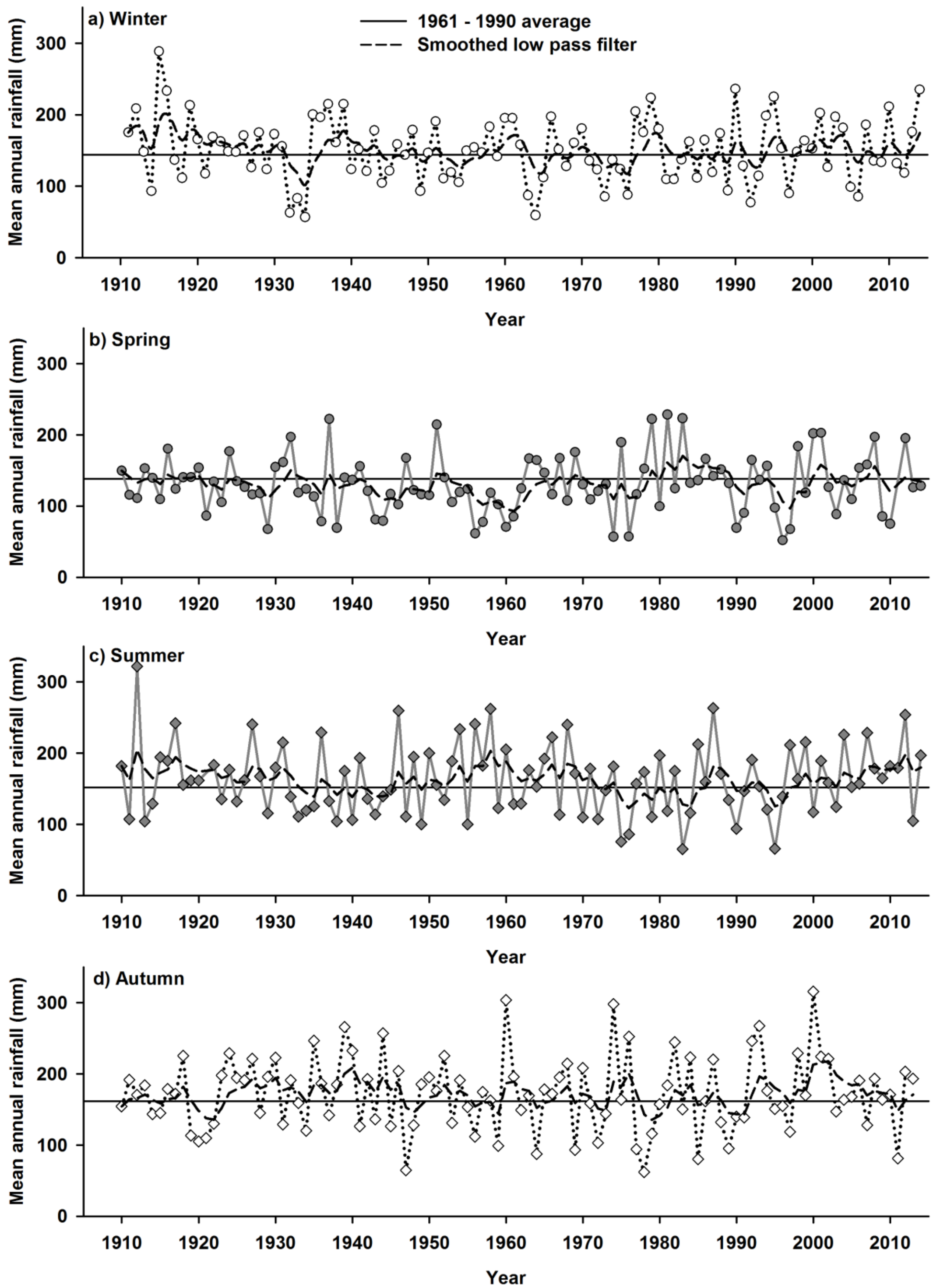


Figure 6.28: Time series of mean seasonal rainfall for the region of East Anglia, England (location of the study site), plotted relative to the average of the 1961 – 1990 baseline period (Jones *et al*. 1999). Data are from the Met Office regional climate summaries (Met office, 2015).

Chapter 7: Coupled hydrological/hydraulic modelling of river restoration and floodplain hydrodynamics

7.1 Introduction

This chapter presents a combined analysis of the monitoring and simulation results of pre-embankment and post-embankment hydrological conditions. Simulations from the calibrated and validated MIKE SHE / MIKE 11 models are used to predict changes in river-floodplain hydrology due to embankment removal for an extended 10-year period, and to further enhance understanding of floodplain functioning. This chapter therefore addresses the second set of research questions outlined in Chapter 1:

(iii) What are the effects of embankment removal on key components of river-floodplain hydrology (water table elevation, frequency and extent of floodplain inundation, flood-peak attenuation)?

(iv) How will embankment removal impact river-floodplain hydrology under a range of expected river-flow conditions?

7.2 Results

7.2.1 Model calibration and validation

For the majority of wells, there is very good agreement between the observed and simulated groundwater levels throughout the calibration and validation periods (Figures 7.1-7.3). The timing of simulated groundwater fluctuations fit well with the observed data (Figure 7.1-7.3). In particular, the rapid response of groundwater during high magnitude rainfall and river-flow events is captured well by the model (Figures 7.1-7.3). The observed and simulated rates of groundwater decline following periods of elevated water tables (typically March – May) also show generally good agreement each year. During some periods of low rainfall, e.g. August to mid-October 2009 (see Section 5.2.2), simulated groundwater levels close to the river are higher than the observed levels, possibly due to over-estimated instream macrophyte growth, however this difference is ≤0.2 m (Figures 7.1-7.3).

Groundwater levels on the floodplain were controlled by river stage and responses to rainfall. The model reproduces the close connection between groundwater and river water levels, and captures the recession of groundwater levels in response to decreasing river levels (Figures 7.1-7.4). Seasonal changes in groundwater levels are reproduced well by the model. Levels at each of the well locations exhibit similar temporal patterns, with distinct seasonal fluctuation in groundwater levels in the range of 0.4 – 0.6 m. Across the floodplain, greater fluctuations in groundwater levels are simulated during the summer when drier conditions result in water levels that are typically lower in the soil profile, compared with the winter when surface soils are predominantly saturated (Figures 7.1-7.3). Consequently, greater variability in groundwater levels occurs between summers than between winters. The model clearly reproduces the lower groundwater levels observed during the dry summers of 2009 and 2010 following embankment removal, compared with the wet summers in 2007 and 2008 in which both observed and simulated groundwater levels are higher (Figures 7.1-7.3).

The ability of the model to represent observed conditions within Hunworth Meadow is further demonstrated in Table 7.1 that summarises the model performance statistics for each well for the calibration period and each of the validation periods (pre- and post-restoration). The mean error for groundwater levels was typically less than ± 0.05 m and the correlation coefficient averaged 0.85, 0.80 and 0.85 for the calibration and pre- and post-validation periods, respectively. Values of the Nash-Sutcliffe efficiency coefficient were between 0.5 – 0.8 for most of the wells, indicating fair to good model performance. In particular, excellent performance is indicated for wells 3.1 and 3.2. The first of these was the only well located on the embankment and as a result necessitated re-installation of monitoring equipment after restoration (note the change in soil surface elevation in Figure 7.1a).

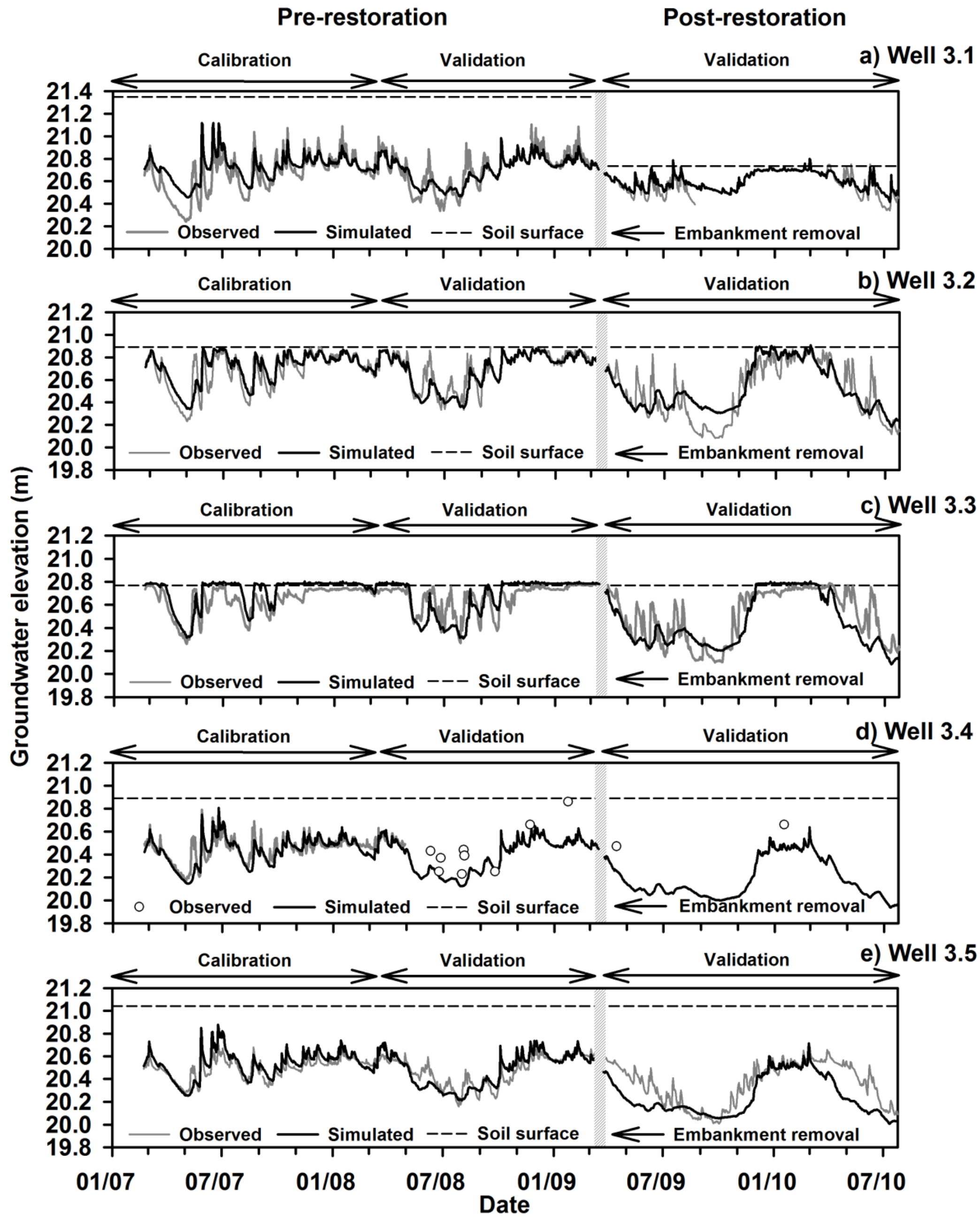


Figure 7.1: Comparison of observed and modelled groundwater depths for the calibration and validation periods at the upstream well transect. Note: soil surface elevation change at Well 3.1 (a), which was located on the embankment prior to the restoration.

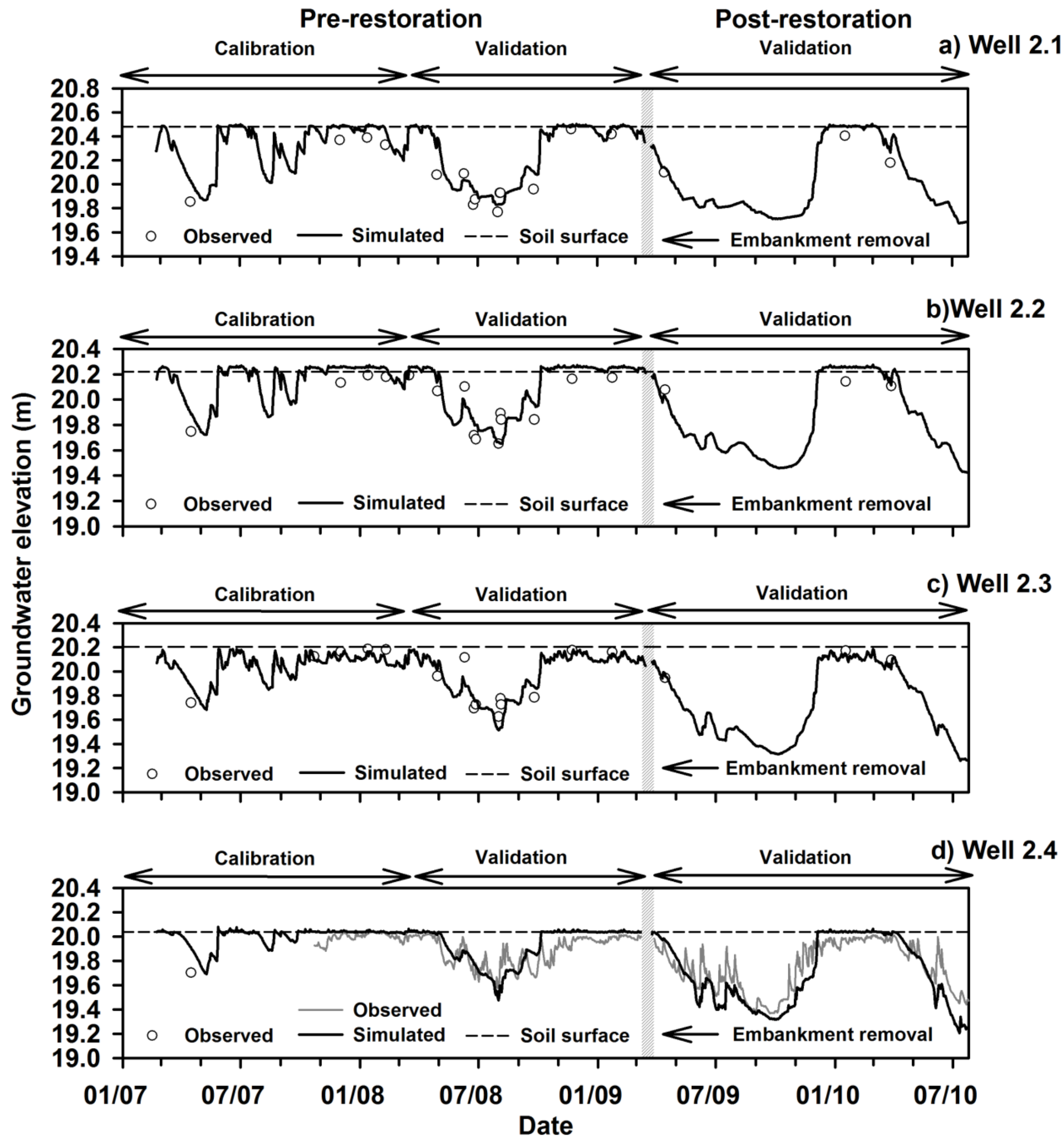


Figure 7.2: Comparison of observed and modelled groundwater depths for the calibration and validation periods at the midstream well transect. The embankment removal in March 2009 is highlighted by the vertical hashed bar.

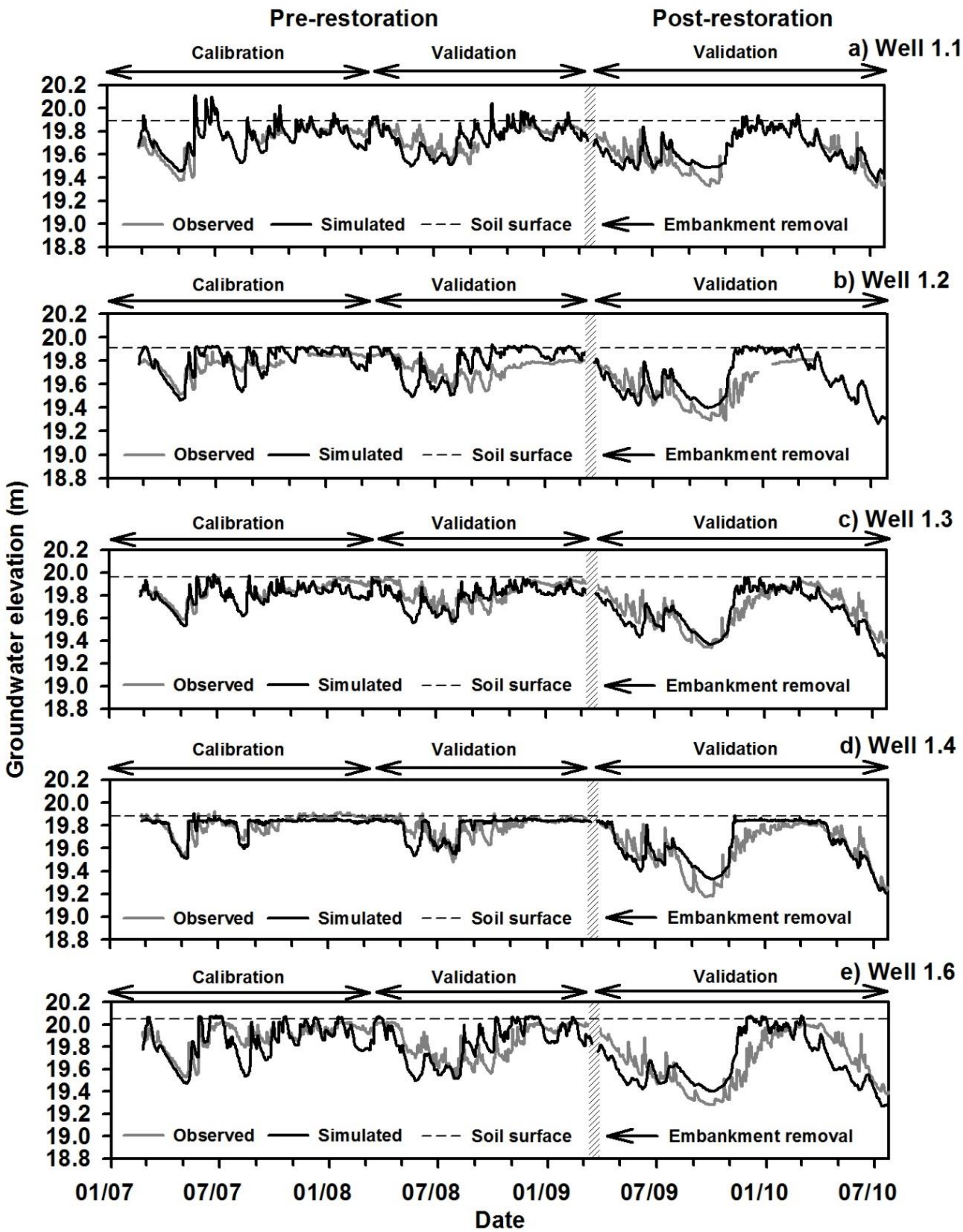


Figure 7.3: Comparison of observed and modelled groundwater depths for the calibration and validation periods at the midstream well transect. The embankment removal in March 2009 is highlighted by the vertical hashed bar. Note: there were problems at times with the level logger at Well 1.2, hence this well was discounted during the calibration and validation of the model.

Water levels simulated by the model also provide a good fit at Well 1.1, which was located next to the embankment at the downstream end of the meadow, and at wells spanning the middle section of the meadow (Wells 2.1-2.3). The model performs less well at the edge of the ditch, i.e. at those wells that were in many cases within 1 m of this channel, and at the floodplain-hillslope margin. Model performance statistics indicate a poorer fit in this narrow section of the floodplain, with simulated groundwater levels being periodically slightly higher than observed at Well 2.4 and lower than observed at Wells 3.4 and 3.5 (Figures 7.2 and 7.3). These wells proved the most difficult to calibrate. In many of the calibration simulations too much water was present at these locations.

The drainage option in MIKE SHE was employed to simulate the transfer of water along the base of the hillslope to the drainage ditch, the topography of which was not fully captured at the 5 m model grid resolution (see Section 6.2.6). These effects were largely removed at smaller grid spacing (i.e. 2 m), however the computational requirements for this model resolution was too intensive. Model performance in some of the lower meadow wells (e.g. Wells 1.1 and 1.3) was poor during the pre-restoration validation due to lower than observed groundwater levels during a period of low rainfall from April-May 2008 (see Section 5.2.2). Groundwater levels in the ditch were simulated and guided the calibration process; however there was a slight offset in the simulated and observed water levels of approximately +0.1 – 0.2 m (Figure 7.4). This was probably due to smoothing of the ditch topography at the 5 m grid resolution (see Figure 6.5). Ultimately, however, the magnitude of daily groundwater changes in the ditch were captured very well by the model (Figure 7.4).

Collectively the comparisons between observed and simulated groundwater levels and the associated model performance statistics indicate a good ability of the model to reproduce groundwater levels across most of the meadow for periods both before and after the removal of river embankments. These results suggest that the model is an appropriate tool to assess the impacts of embankment removal upon hydrological conditions across the floodplain.

Table 7.1: Mean error (ME - m), correlation coefficient (R), and Nash-Sutcliffe model efficiency coefficient (NSE) for the calibration (22/02/07 – 14/03/2008) and validation (pre-restoration:15/03/2008 – 15/03/2009; post-restoration: 29/03/2009 – 25/07/2010) periods.

	Calibration (pre-restoration)			**Validation (pre-restoration)**			**Validation (post-restoration)**		
Well	**ME**	**R**	**NSE**	**ME**	**R**	**NSE**	**ME**	**R**	**NSE**
1.1	-0.02	0.79	0.60	0.02	0.71	0.14	-0.02	0.70	0.47
1.2*	-0.02	0.74	-0.23	-0.05	0.42	-1.71	-0.07	0.75	0.29
1.3	0.03	0.81	0.49	0.03	0.65	0.16	0.03	0.83	0.62
1.4	0.00	0.79	0.60	0.00	0.66	0.27	-0.03	0.86	0.70
1.6	0.04	0.74	0.12	0.00	0.65	0.17	0.01	0.68	0.37
2.1	-0.13	1.00	0.62	-0.05	0.91	0.75	-0.07	0.99	0.67
2.2	-0.10	0.99	0.67	-0.01	0.85	0.58	-0.02	0.99	-6.56
2.3	0.01	0.96	0.77	0.05	0.85	0.56	0.02	0.89	0.72
2.4	-0.05	0.62	-0.80	-0.04	0.80	0.31	0.05	0.91	0.54
3.1	-0.06	0.85	0.56	0.01	0.90	0.76	-0.04	0.84	0.52
3.2	-0.04	0.89	0.73	0.03	0.89	0.75	-0.03	0.84	0.70
3.3	-0.07	0.86	0.37	0.00	0.75	0.20	0.05	0.81	0.57
3.4	0.01	0.82	0.55	0.03	0.85	0.18	n/a	n/a	n/a
3.5	-0.04	0.89	0.45	0.00	0.88	0.69	0.11	0.81	0.26
River stage	*n/a*	*n/a*	*n/a*	0.08	0.97	0.65	-0.03	0.65	0.18

**Note that there were problems at times with the level logger at Well 1.2 so this well was discounted during model calibration and validation.*

n/a: data not available for the period.

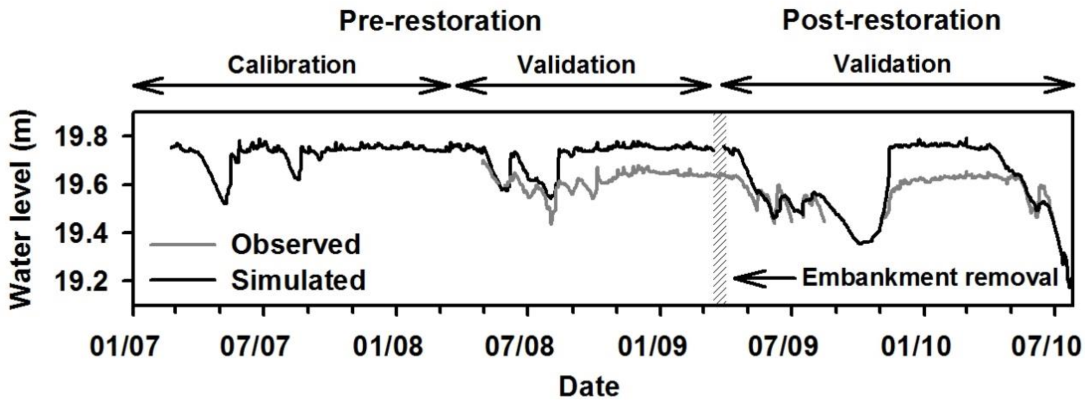


Figure 7.4: Comparison of observed ditch water levels and MIKE SHE-simulated groundwater levels in the ditch at well transect 1.

7.2.2 Impacts of embankment removal on overbank flows and floodplain inundation

Bankfull capacity for the embanked river channel was estimated using a cubic regression between river stage and discharge (Figure 7.5), a similar approach to that used in Section 4.5.1, however in this instance river stage and discharge were obtained from MIKE 11 outputs, and therefore for multiple locations along the study reach. The winter Manning's value (0.058 $m^{1/3}s^{-1}$) was used to determine the embanked bankfull capacity in order to remove the effect of vegetation that occurred at the lower river flows and simplify the regression relationship between river stage and discharge (see Sections 4.5.1 and 6.3.2). Figure 7.5 presents bankfull discharge values for two locations along the river, which are representative of values for the study reach. The embanked bankfull threshold for the lower meadow section was 5.1 m^3 s^{-1}, whereas a slightly lower bankfull capacity of 4.5 m^3 s^{-1} applies to the upper meadow (Figure 7.5).

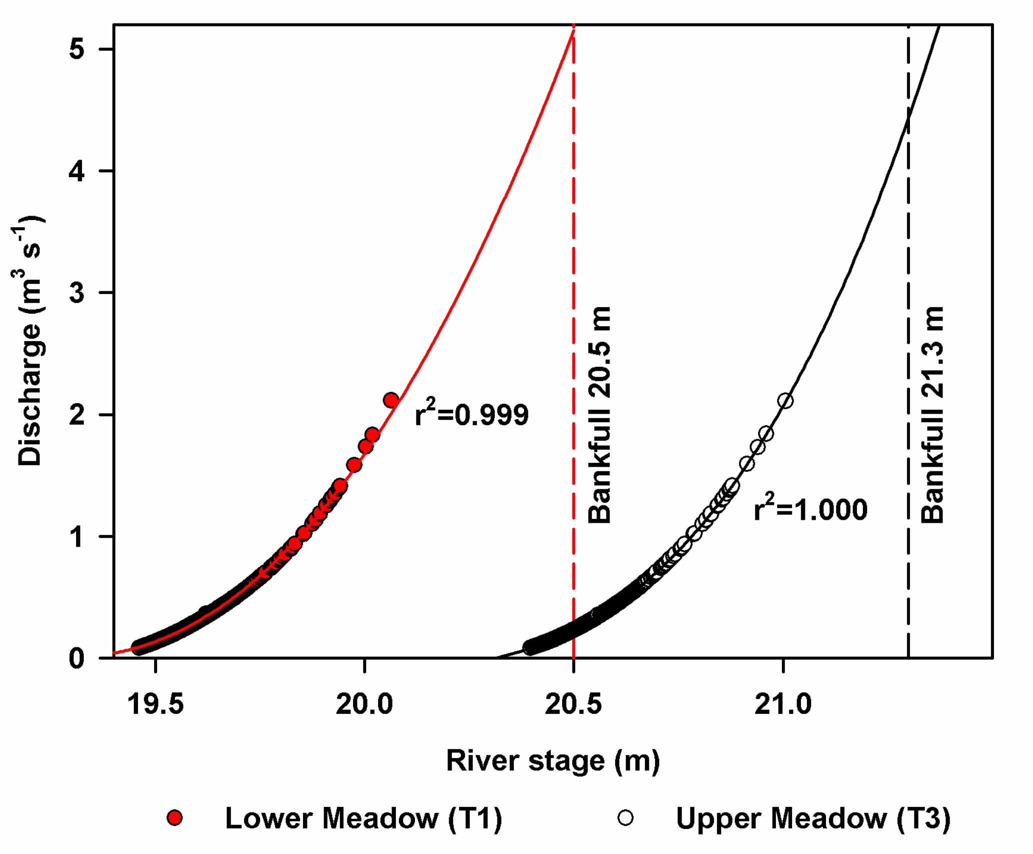


Figure 7.5: River stage-discharge relationships from MIKE 11 outputs for the embanked scenario at the Lower Meadow (r^2 = 0.999; y = -635.0860 + 162.9633 * x + -11.7679 * x^2 + 0.2606 * x^3) and Upper Meadow (r^2 = 1.000; y = -13843.7297 + 2083.8580* x + -104.5615* x^2 + 1.7489* x^3) areas. Bankfull discharges for the embanked scenario were not observed during the 10-year period of the discharge record; hence bankfull capacity was extrapolated beyond the available data and thus should be treated with caution.

Bankfull capacity for the restored river was derived from MIKE 11 stage-discharge relationships and bankfull measurements from the river cross-sections (Figure 7.6). In addition, bankfull capacity was evaluated using MIKE SHE results depicting the depth of overland water, which enabled the identification of two thresholds for overland flow: a high discharge threshold above which widespread inundation occurred, and a lower threshold above which localised flooding (up to one grid cell - i.e 5 m - from the river) occurred (Figure 7.6). The high flow threshold was selected based on values for bankfull capacity at the Lower Meadow section, flows above which were required to fully connect the river and floodplain (widespread inundation threshold) (Figure 7.6). There is some variation in the bankfull discharge for localised flooding, which is due to interannual differences in summer in-stream macrophyte growth, i.e. periods of high macrophyte growth raised river levels (due to increased flow resistance) which reduced bankfall capacity for a given discharge (Figure 7.6).

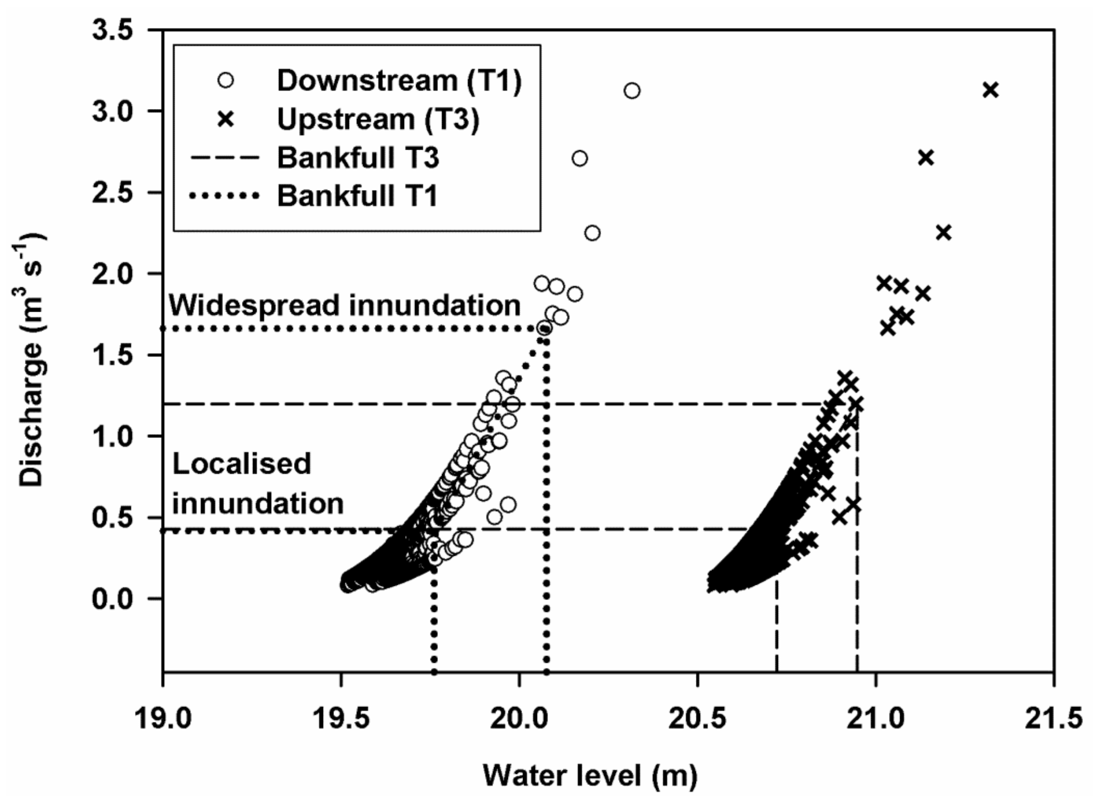


Figure 7.6: River stage-discharge relationships from MIKE 11 outputs for the restored scenario at the Lower Meadow and Upper Meadow areas. River discharge and stage thresholds for widespread and localised inundation on the floodplain are shown.

The impact of embankment removal upon the potential for overbank flows is summarised in Figure 7.7. This shows the daily discharge at the Hunworth gauging station for the period 2001 – 2010 upon which are superimposed the estimated bankfull channel capacities under both embanked and restored conditions.

Throughout the whole 10-year period no overbank flows were simulated since the bankfull channel capacity (5.1 m^3 s^{-1}) was greater than the maximum observed flow of 3.1 m^3 s^{-1}. This is in agreement with the observation-based bankfull results detailed in Section 5.2 from the river stage-discharge relationship at well transect 1 (T1) and bankfull river discharge calculated using Manning's equation.

In contrast, river flows frequently exceeded bankfull capacity in the restored model, where two thresholds for inundation on the floodplain were identified: the high flow channel capacity (1.67 m^3 s^{-1}) above which widespread floodplain inundation occurred; and the low flow channel capacity (0.59 m^3 s^{-1}) which resulted in localised inundation at the river edge in an area corresponding to the former location of the embankments (Figures 7.6 and 7.7). Such flooding did not occur in pre-restoration conditions due to the steep sides of the embankments. Throughout the 10-year period, discharge exceeded the high flow channel capacity for widespread flooding in the restored model on nine occasions, albeit only for short periods (1 day). Three large overbank events occurred over a month-long period from late May – June 2007, interspersed with eight smaller localised flooding events at the river-edge. Localised flooding was much more frequent (61 occasions) and of longer duration (2 – 3 days), and is likely to result in a more dynamic and natural transitional zone between the river and the floodplain.

Surface flooding on the floodplain is simulated within the MIKE SHE models when groundwater levels intercept the ground surface (in which case precipitation cannot be infiltrated) or when the river over-tops the channel banks. In the embanked model groundwater was the only source of flooding on the floodplain, whereas under restored conditions inundation also occurred due to overbank flows. Restoration of these overbank flows and the reconnection of the river and its floodplain therefore had a marked effect on simulated floodplain hydrology. This is clearly demonstrated in Figure 7.8, which shows the simulated extent and depth of surface water for the pre- and post-restoration models for two high river-flow events. The first (which occurred on 28/05/2007) is associated with a mean daily discharge of 1.9 $m^3 s^{-1}$, just above the threshold channel capacity associated with widespread inundation under restored conditions (Figure 7.8a), whilst the second (18/07/2001) is the largest event (3.1 $m^3 s^{-1}$) during the 10-year simulation period (Figure 7.8c).

Results for the embanked, pre-restoration model show that river water was constrained within the river channel by the embankments, which were not flooded in both events shown in Figure 7.8 and indeed throughout the 10-year simulation period. During the smaller flood event (Figure 7.8a), flooding was limited to the margins of the floodplain ditch and the downstream ponded area and was driven by rising groundwater tables. During the larger river-flow event (Figure 7.8c), there was limited groundwater flooding behind the embankments, with surface water depth ranging between 0.0 – 0.02 m across much of the meadow, and up to 0.4 m in topographic depressions along the ditch and ponded area in the lower meadow. This was attributed to an extended period of low rainfall, high evapotranspiration and low water table depths that preceded the high flow event.

Under post-restoration conditions overbank flows resulted in widespread inundation on the floodplain that would supplement groundwater-fed surface water. During the smaller flood event (Figure 7.8b) much of the floodplain was subject to shallow (<0.3 m depth) inundation. Embankment removal enabled some overbank flows at the top end of the floodplain although a relatively high section of the riverbank and adjacent floodplain in the upper-middle part of the site was not flooded. Further downstream the lower half of the floodplain was directly connected with the river and the previously embanked area was inundated. During the largest flood event (Figure 7.8d) nearly the whole floodplain was directly connected to the river, and extensive and much deeper flooding (0.2 – 0.6 m) occurred.

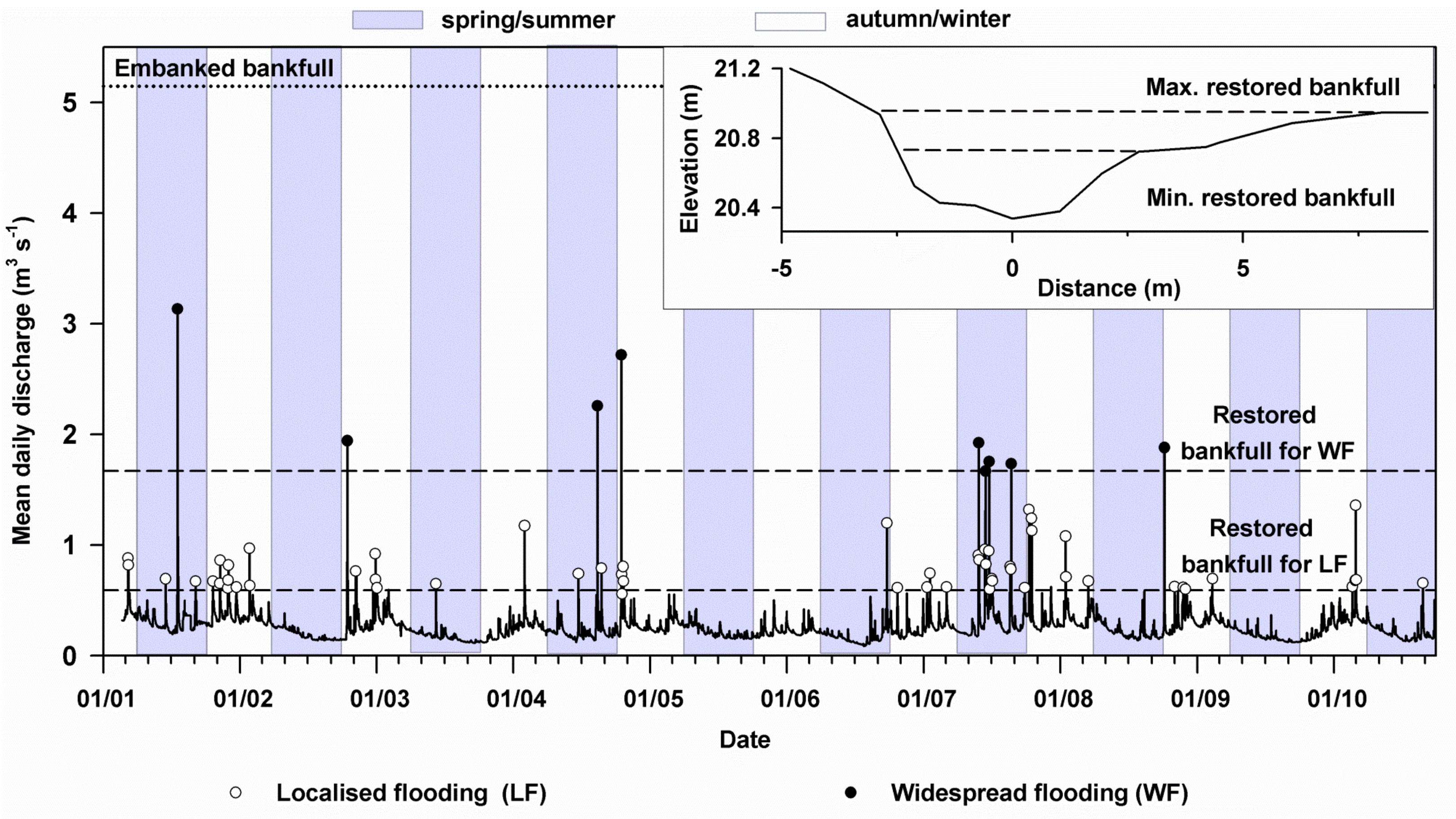


Figure 7.7: Mean daily river discharge from 2001 – 2010. The embanked and restored bankfull capacity is shown, above which widespread inundation of the floodplain would have occurred. Two bankfull thresholds, a minimum and maximum, are shown for the restored river, which correspond to the cross-section inset; flows above these thresholds result in localised and widespread flooding, respectively.

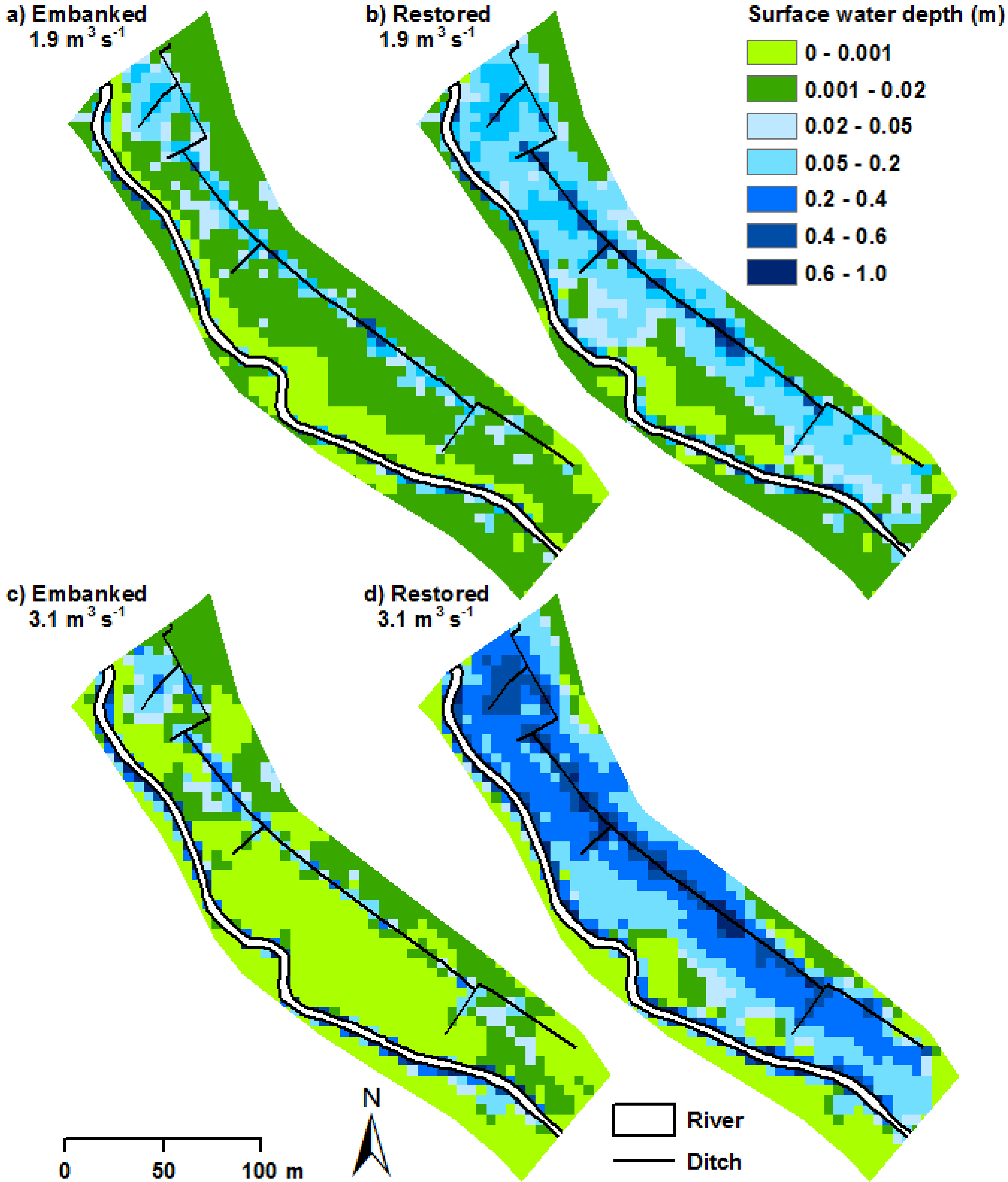


Figure 7.8: Comparison of simulated surface water extent and depth for the embanked and restored scenarios during a small overbank (post-restoration) event (28/05/07; flow = 1.9 $m^3 s^{-1}$) (a-b) and a larger overbank (post-restoration) event (18/07/01; flow = 3.1 $m^3 s^{-1}$) (c-d).

Simulation results show that the ditch running parallel, but to the north of the river, played an important role in distributing flood water. Surface water resulting from high water tables or overbanking of the river was channelled across and down the floodplain into the ditch, which then filled and contributed to flooding along the ditch marginal areas, ponding in topographic depressions, and subsequent groundwater recharge leading to higher water table elevations. Surface water accumulated in the lower section of the meadow in the region of the pond (Figure 7.8a and 7.8c). Prior to the restoration, the ponded area that was subject to groundwater flooding as well as being fed by the ditch, was saturated for much of the year. In this state, the embankment acted as a barrier for water that had accumulated in this part of the floodplain, preventing its return to the river. However, after the removal of the embankments, drainage of surface water from the floodplain to the river was restored. Water stored in this low-lying area of the meadow during flood events subsequently acted as a source of return flow to the river (Figure 7.8).

7.2.3 Impacts of embankment removal on groundwater

Throughout the 10-year simulation period, groundwater levels close to the river (i.e. within 30 m) were on average 0.01 m higher under restored conditions, whereas groundwater levels in low lying areas of the meadow that were previously flooded were on average 0.01 m lower in the restored scenario. This is reflected in Figure 7.9 showing the differences in groundwater levels at the 14 wells simulated by the embanked and restored models. During periods of the highest river flows, groundwater levels were up to 0.8 m higher under restored conditions. The largest increases in water table elevation occurred along the river banks (e.g. wells 3.1 and 1.1), in the region of the ditch (e.g. wells 2.3 and 2.4), and on the relatively low-lying downstream end of the floodplain. The smallest effects were seen at well 2.1 adjacent to the section of riverbank that was not restored, where increases in water table elevation during high river-flow periods were typically less than 0.3 m (Figure 7.9b). This location corresponds to the relatively high part of the floodplain, where the embankments were not removed (see Section 3.8), that consequently was not flooded under post-restoration conditions (Figure 7.8b). Some short periods of slightly lower groundwater levels (up to -0.18 m) were simulated under restored conditions immediately after periods of groundwater and overbank flooding. These are most noticeable at Well 1.1 (Figure 7.9c) that was located close to the river and the ponded area in this part of the floodplain and at Well 1.4 that was located in a low-lying area next to the ditch. These changes are most likely due to the previously discussed improved drainage at the river-floodplain margin.

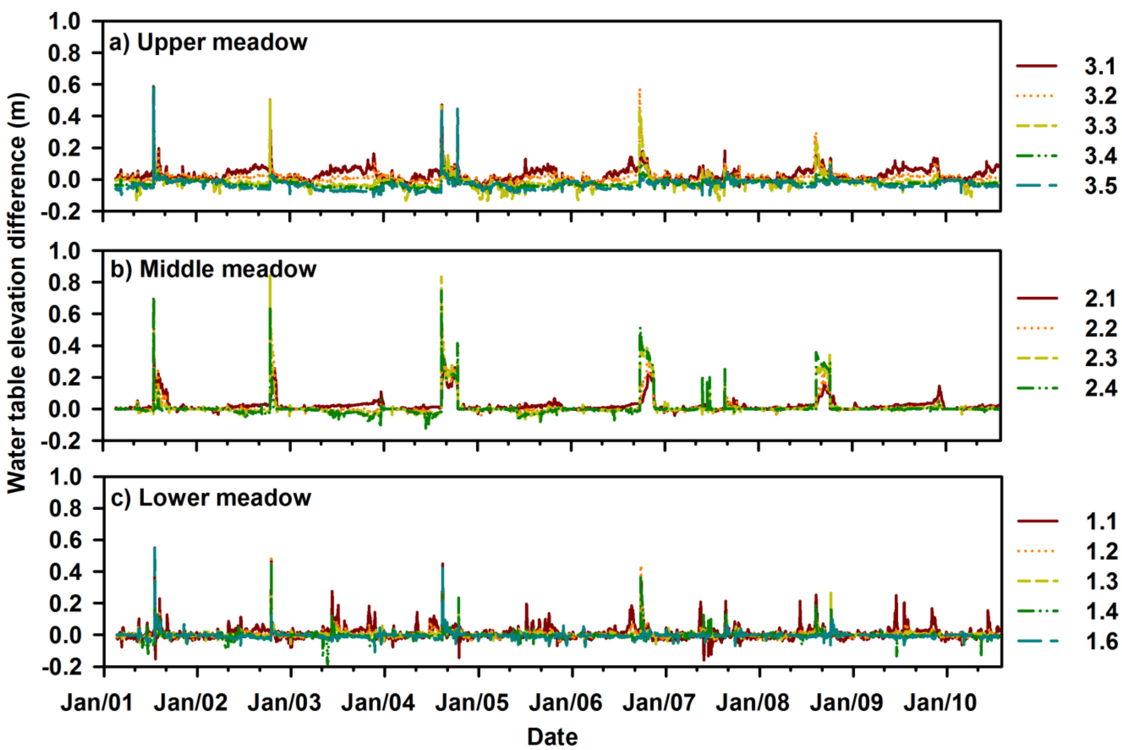


Figure 7.9: Simulated time series of water table elevation (WTE) differences between the restored and embanked scenarios from 2001 – 2010 (a period which encompassed a range of wet and dry conditions). Positive differences indicate restored WTE > embanked WTE. Differences in WTE are shown at well locations across the floodplain at the (a) upper, (b) middle, and (c) lower well transects (see well locations in Figure 3.1 and 7.12).

The greatest differences in water table elevation between the embanked and restored model results occurred in spring/summer during periods of low river flows. Simulated groundwater levels along the river (i.e. within 30 m) for the restored model were on average 0.03 m higher ($p<0.05$) than those for the embanked model in the spring/summer. No significant differences were found in the autumn/winter ($p=0.754$) (Figure 7.9). This can be attributed to increased surface flooding and floodplain storage during a number of inundation events that occurred in the summer months. Increased floodplain storage before the beginning of the spring/summer drawdown combined with periodic additions from summer flooding reduces the summer groundwater head recession within the meadow (Figure 7.10). The higher simulated groundwater levels after embankment removal causes some differences in the hydraulic gradient between the river and floodplain during the summer. In comparison to the embanked model, results for the restored model show that summer groundwater levels at the river margin are closer to river water levels (Figure 7.10), resulting in more frequent reversals in the hydraulic gradient and consequentially a more dynamic subsurface exchange.

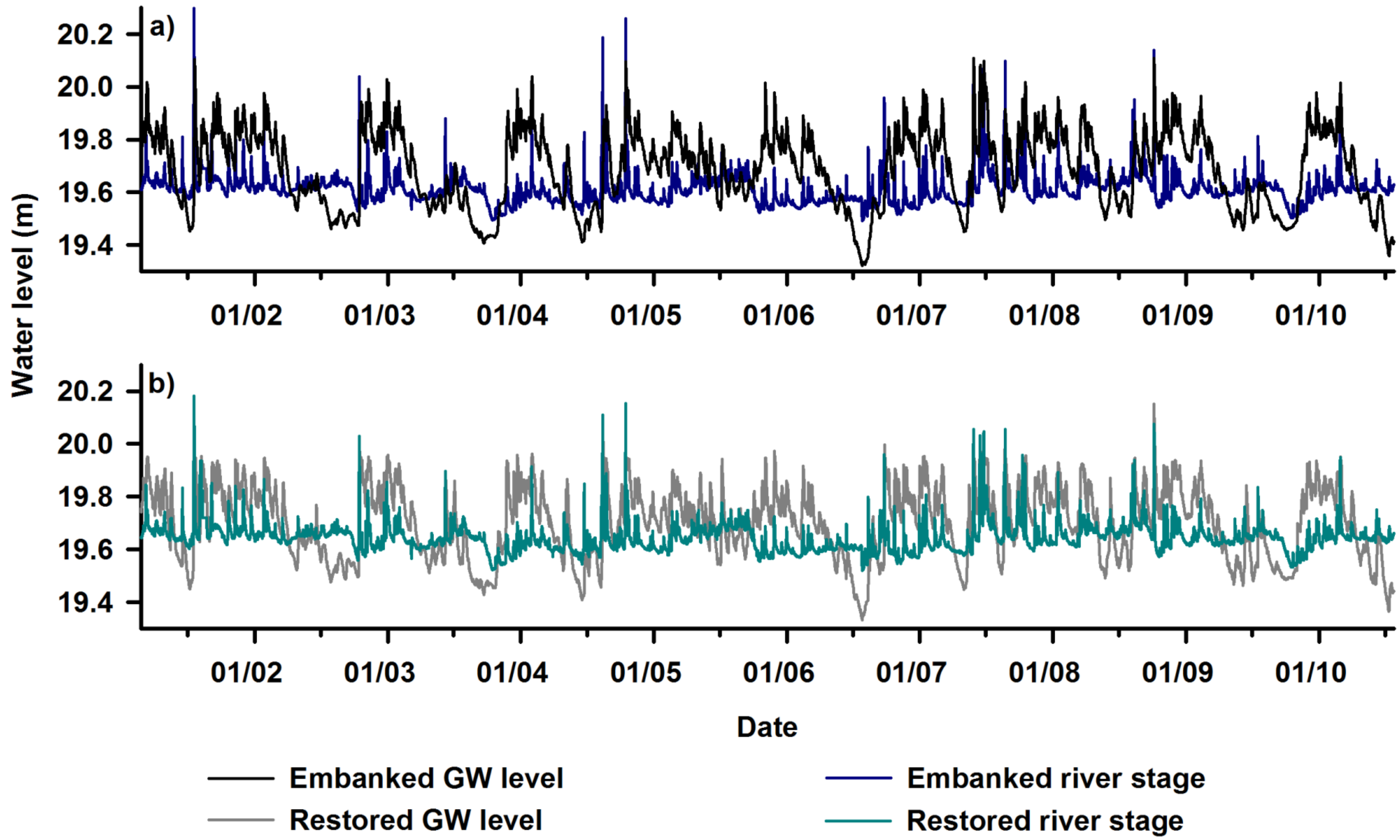


Figure 7.10: Comparison of simulated groundwater elevation (ODN) at well 1.1 and river stage (ODN) for the (a) embanked and (b) restored scenarios from 2001 – 2010.

7.2.4 Impacts of embankment removal on groundwater flowpaths

As discussed in Section 5.2.2, observed groundwater heads generally decreased from the direction of the river to the floodplain at the upstream and midstream transects groundwater (Figure 7.11a-b). In contrast, at the downstream transect, the usual hydraulic gradient was from the hillslope towards the river (Figure 7.11c). Hillslope water entered the floodplain in the region of the ditch (a possible remnant river channel of the River Glaven) that is positioned at the lowest point on the floodplain. At the upstream well transect, hillslope water intersected with groundwater on the floodplain flowing from the direction of the river. The confluence of these two water sources is supported by the analysis of groundwater chemistry in Section 5.2.6. From this point on the floodplain, simulated groundwater flow is in a down-valley direction, with the ditch providing storage. Towards the downstream transect, where the topography flattens, groundwater flow within the floodplain is perpendicular to the river and hillslope (Figures 7.11 and 7.12).

During the autumn and winter, simulated groundwater movements across the floodplain are complex. A groundwater divide is simulated at the upstream and midstream parts of the meadow with sub-surface flow simulated from both the river and ditch to the central part of the floodplain (e.g. Figure 7.11a). At this time of year, groundwater levels on the floodplain are close to or above river water levels. In the lower part of the meadow, the high water table acts as a source of water to the river, with some groundwater exchange back to the river being simulated (Figure 7.12). During dry summer conditions, simulated river levels are above groundwater levels in all wells (Figure 7.13). The hydraulic gradient from the floodplain to the river is reversed and instead simulated subsurface flows are predominantly directed from the river to the floodplain (Figure 12d). Short-term (1 – 2 day) groundwater ridging and increases in floodplain storage are simulated during periods of peak river flows (Figure 12b and d, Figure 7.13). However, longer-term (2 – 3 months) reversal of the hydraulic gradient, and the consequent loss of river water to groundwater storage are simulated during dry periods in the summer, possibly due to a dominant down-valley hydraulic gradient (Figure 7.12c and Figure 7.13).

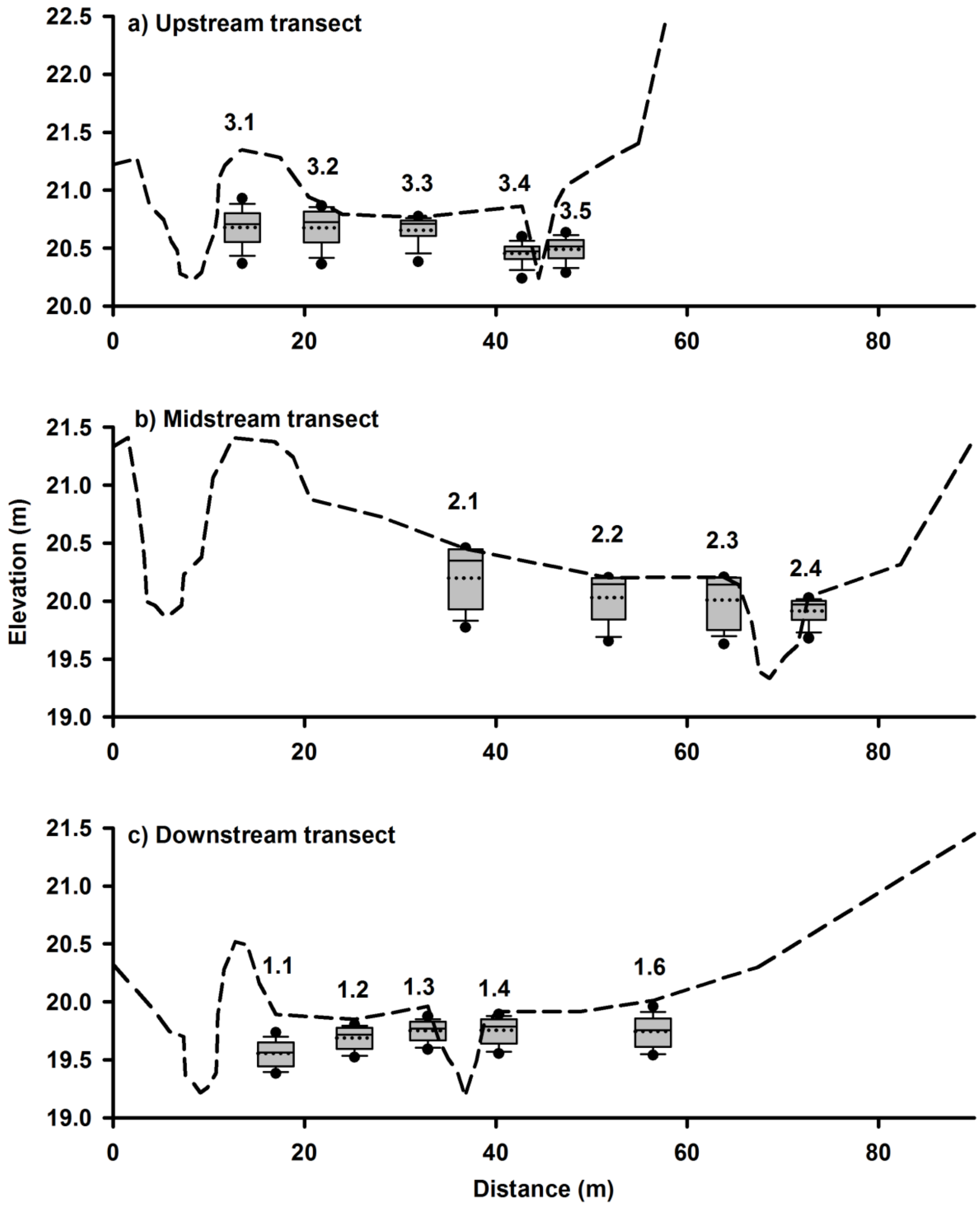


Figure 7.11: Boxplots of simulated groundwater elevation in relation to surface topography along the three well transects prior to the restoration.

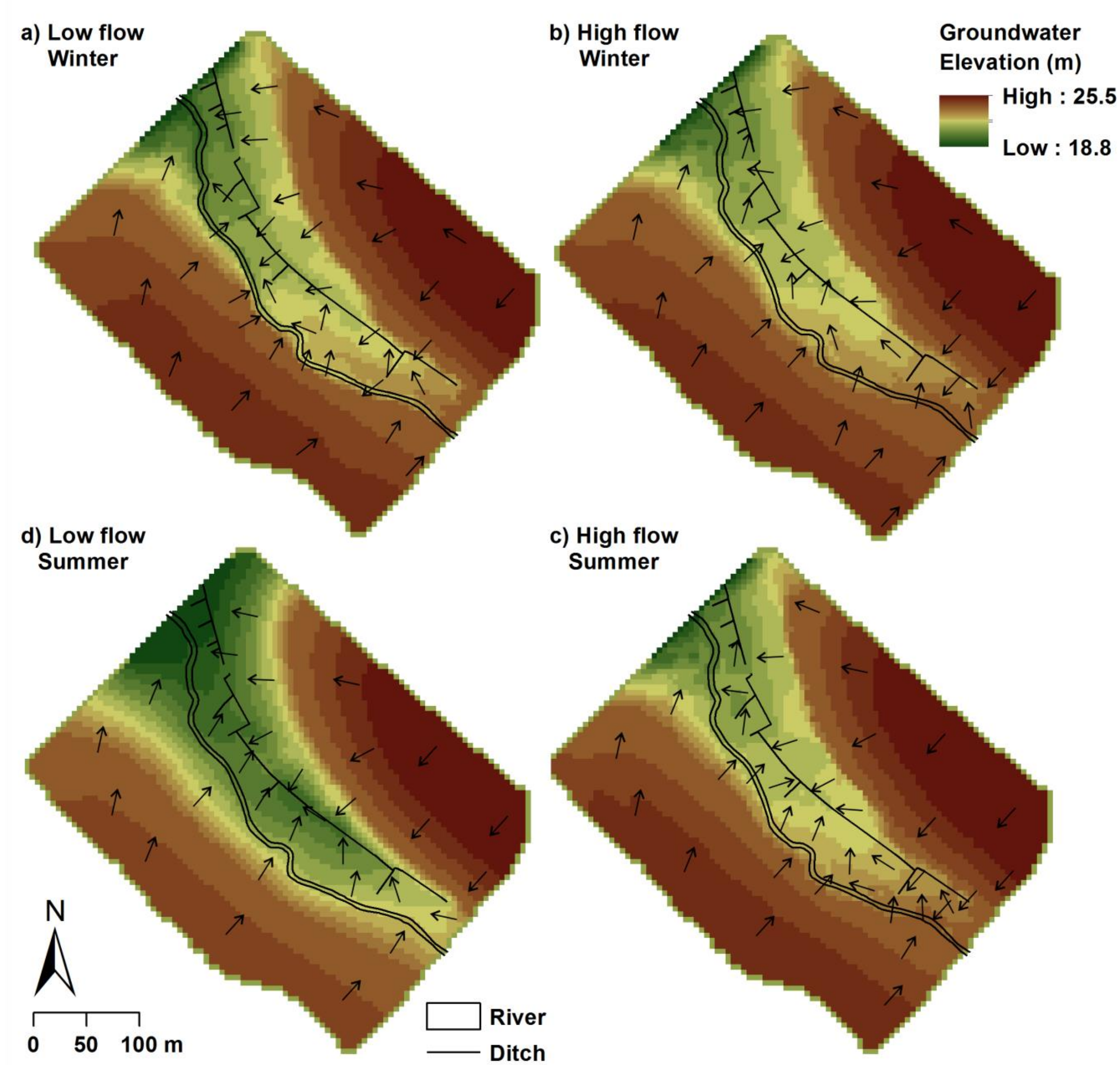


Figure 7.12: Simulated groundwater elevation and flow direction (arrows) during (a) low (01/01/07) and (b) high (15/10/04) river flow winter conditions, and (c) low (01/09/09) and (d) high (28/05/07) river flow summer conditions simulated using the restored MIKE SHE scenario.

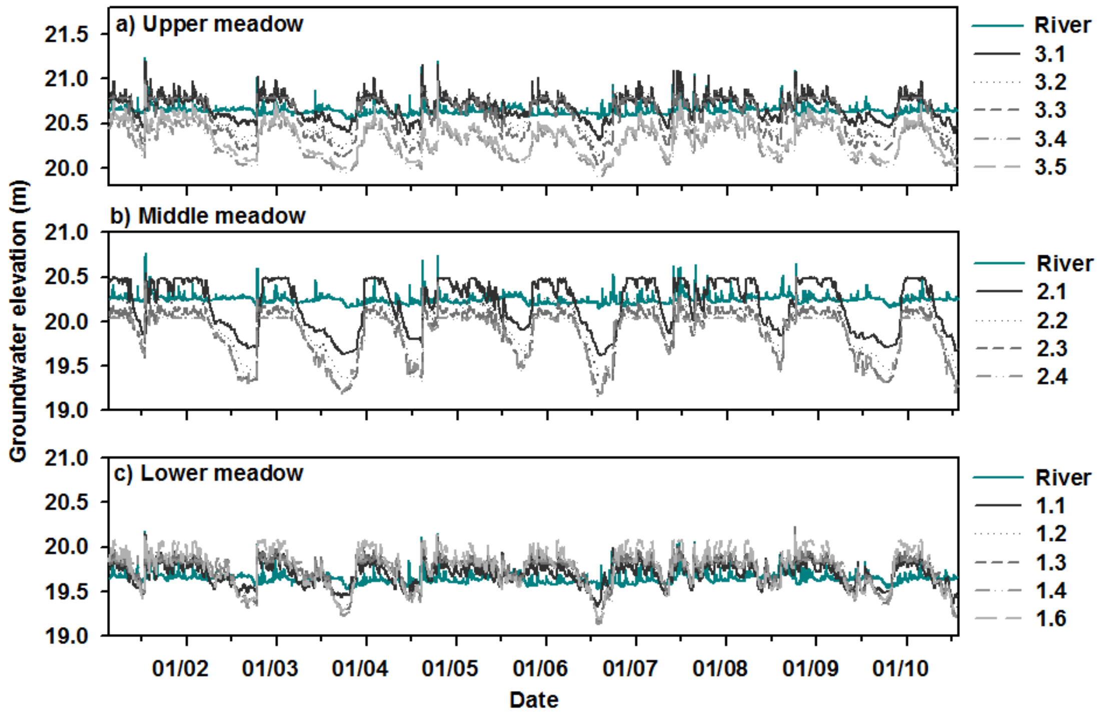


Figure 7.13: Time series of simulated post-restoration groundwater levels relative to the simulated river levels at each well transect from 2001 – 2010.

Seasonal fluctuations in exchange flow between the river and saturated zone are shown in Figure 7.14, and support the analyses above of Figures 7.12 and 7.13. During the winter, the positive exchange values in Figure 7.14b-d indicate that the saturated zone acted as a source of water to the river. During the summer, the saturated zone was, at times, a source of water to the river, but to a lesser extent compared to the winter. During large river-flow events, particularly when they occurred in the summer (e.g. July 2001), and/or after a period of low flow (e.g. October 2002) and thus coincided with lower groundwater levels on the floodplain, exchange reversed between the saturated zone and river, with the saturated zone briefly becoming a sink for river water. Overall, however, this section of the river was a gaining reach. The reversal of flow direction simulated during flood events is consistent with the bank ridge model presented by Burt *et al*. (2002) (see Figure 2.6). In the restored scenario the exchange flow from the saturated zone to the river was reduced, resulting in greater retention of groundwater. Furthermore, during high river flow, subsurface exchange with the river was a less important exchange of water in the restored scenario, likely due to the precedence of surface (overbank) exchange on the floodplain.

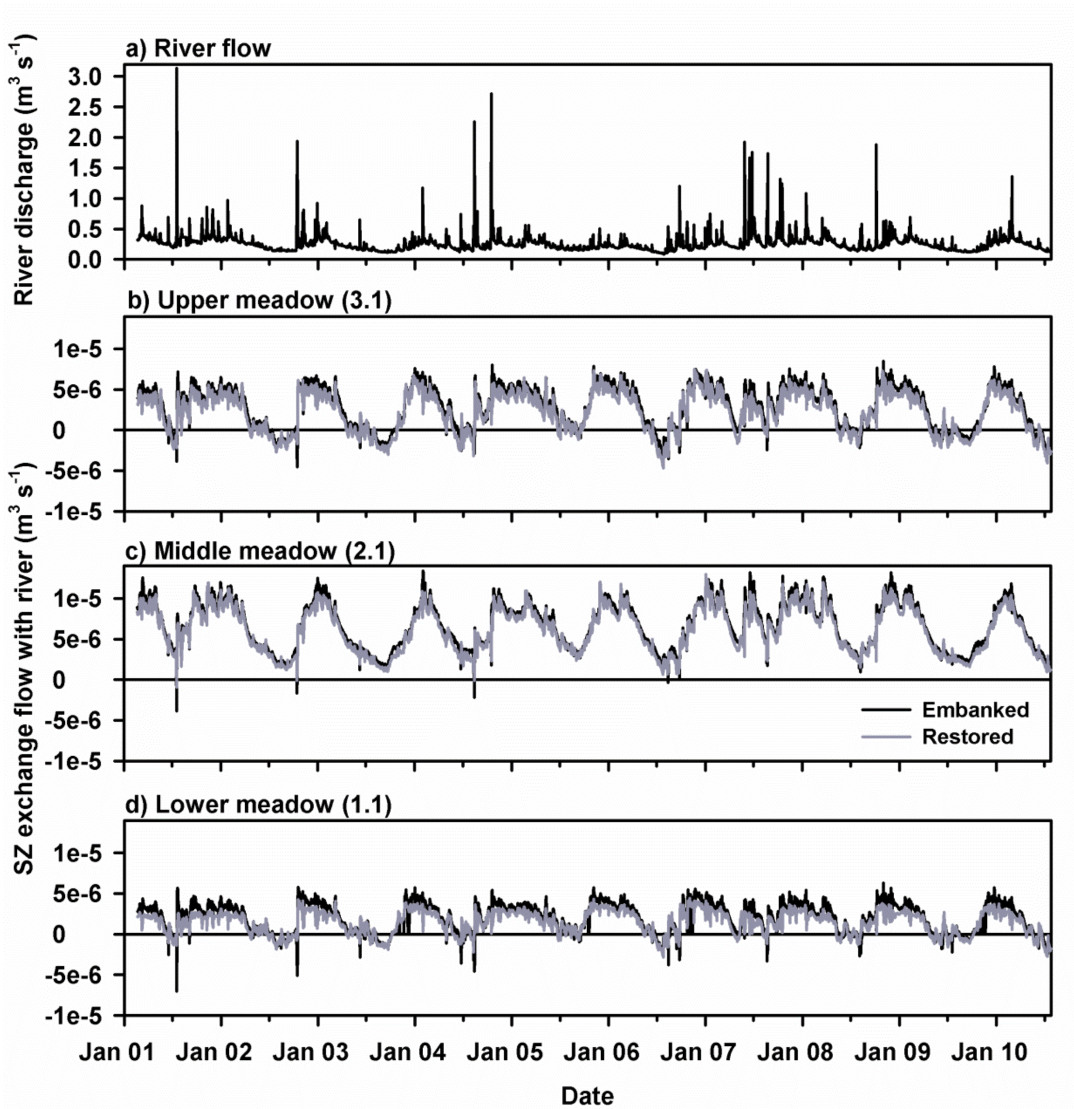


Figure 7.14: Simulated exchange flow between the saturated zone and river. Values > zero indicate exchange from the SZ to river, whereas negative values indicate exchange from the river to the saturated zone.

7.2.5 Impacts of embankment removal on floodplain storage and flood-peak attenuation

The impacts of embankment removal upon both overland and subsurface water storage within Hunworth Meadow are summarised for the 10-year simulation period in Figure 7.15. The volume of simulated surface water stored on the floodplain is greater in the restored model (Figure 7.15a). Particularly large differences between the results of the two models are associated with periods when major overbank flood events are simulated under restored conditions. For example, the overland storage volume increases approximately six-fold during the highest flow event (18/07/2001) after simulated restoration (maximum storage increase of 2159 m^3 compared to 373 m^3 for the embanked model). As discussed above, although surface water is stored on the floodplain in the embanked scenario during these periods, groundwater rather than river water overtopping the riverbanks is the source of flooding (i.e. Figure 7.8). Overbank flows substantially enhanced surface storage, which increased 600% from an average of 144 m^3 in the embanked model to an average of 841 m^3 in the restored model over the 14 peaks in overland storage shown in Figure 7.15a.

Differences in the simulated volume of subsurface storage between the embanked and restored models are much less pronounced (Figure 7.15b). During winter months, groundwater storage is very similar for both models as soils were typically at or near saturation and had limited available storage capacity. However, during the drier floodplain conditions that characterised summer months, subsurface storage is greater under restored conditions. The largest difference in subsurface storage occurred during a period of higher river flow at the end of the dry summer in 2004. At this time, storage change for the original embanked model was -1099 m^3 compared to -401 m^3 for the restored model (Figure 7.15b), equivalent to storage volumes in the floodplain of 38,022 m^3 and 38,675 m^3, respectively.

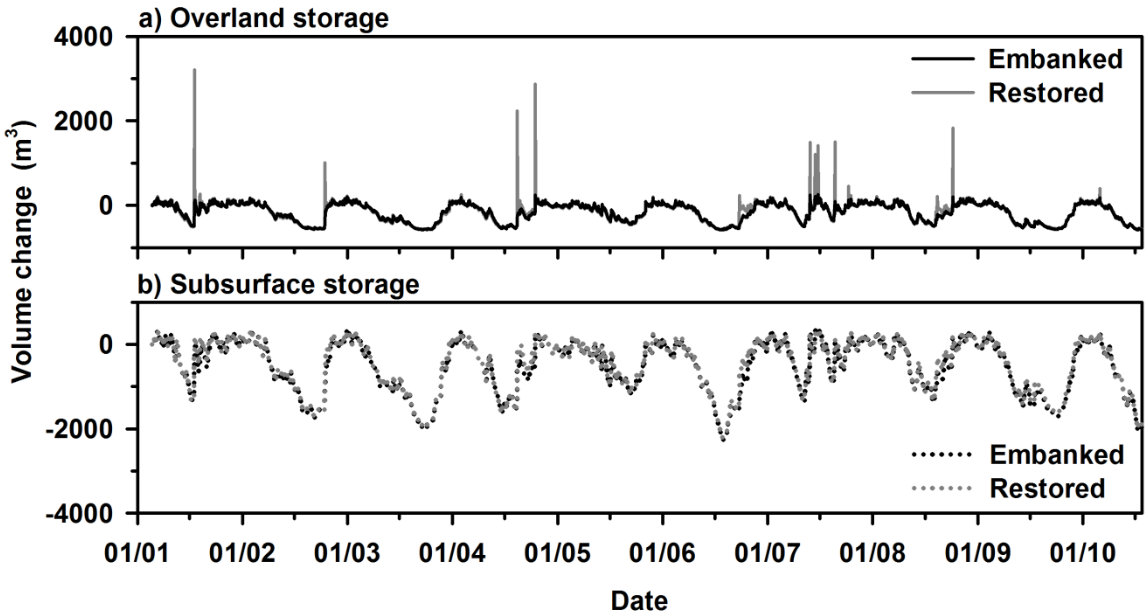


Figure 7.15: Times series of change in (a) overland and (b) subsurface storage for the embanked and restored scenarios. Volume change is set at 0 m^3 at the beginning of the simulations (i.e. 20/02/2001).

Although the annual actual evapotranspiration (ET) totals did not differ between embanked and restored models (Figure 7.16), the different components of total ET were significantly different between the two models. Annual ET from the unsaturated zone was on average 7% larger in the embanked model compared with the restored model ($p<0.05$). This is the result of the higher water tables under restored conditions that limits the depth of the unsaturated zone and the duration of unsaturated conditions at the surface. Conversely ET from the saturated zone and evaporation from ponded overland water were on average 10% and 12% larger for the restored model ($p<0.05$), respectively.

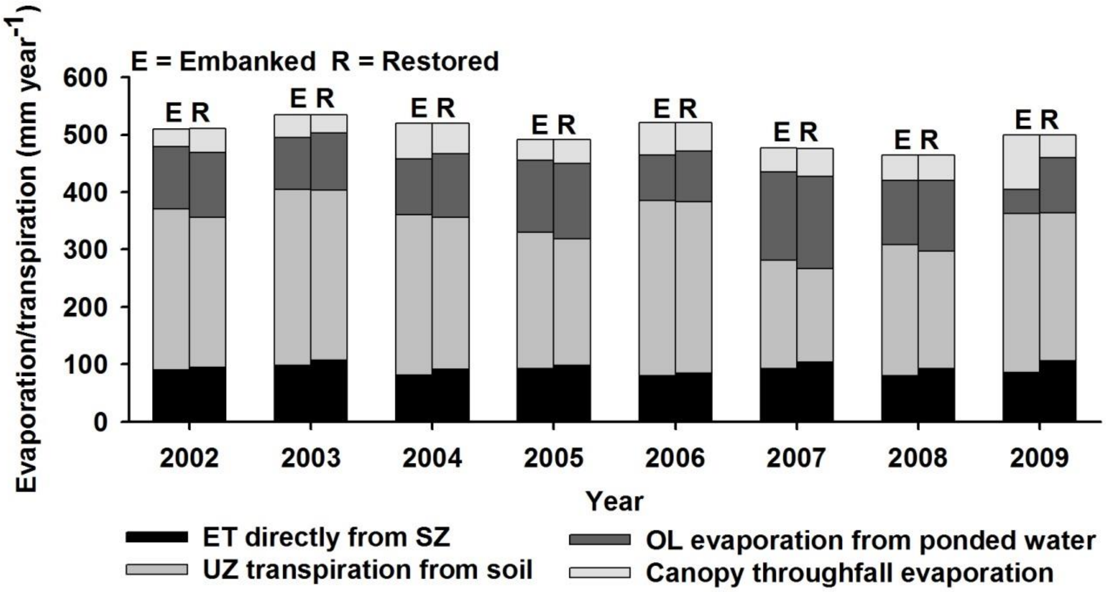


Figure 7.16: Annual evapotranspiration (ET) rates for the embanked (E) and restored (R) scenarios from 2002 – 2009. Four components of ET are shown: transpiration from the unsaturated zone (UZ), ET from the saturated zone (SZ), overland (OL) evaporation from ponded water, and canopy throughfall evaporation.

For the embanked model, river discharges were almost identical at the upstream and downstream ends of the modelled reach demonstrating that most flows are retained within the river channel (Figure 7.17). However, a slight reduction in outflows (of between 1 and 3%) is evident during the highest river-flow events (flows >1.2 $m^3 s^{-1}$), likely associated with loss of flow to bank storage given the absence of simulated overbank flooding. For the restored model, differences between river inflows and outflows began at lower flows (around 1.0 $m^3 s^{-1}$) compared to the embanked model (Figure 7.18). The largest overall reductions in river flow, however, occurred during the largest overbank events (>1.5 $m^3 s^{-1}$) when inundation and recharge to the water table occurred across the floodplain (Figure 7.18). Embankment removal and restoration of overbank flows onto the floodplain had a moderate effect on flood-peak attenuation. The peak discharge of the largest flood (18/07/2001) was reduced by 24% from 2.94 $m^3 s^{-1}$ at the top of the restored reach to 2.31 $m^3 s^{-1}$ at the downstream end (Figures 7.17 and 18). Following the highest river flows, outflow was marginally greater than inflow (max.

2% and 3% in the embanked and restored scenarios, respectively), due to some return flow from the floodplain to the river. However, these differences are barely noticeable in Figure 7.18.

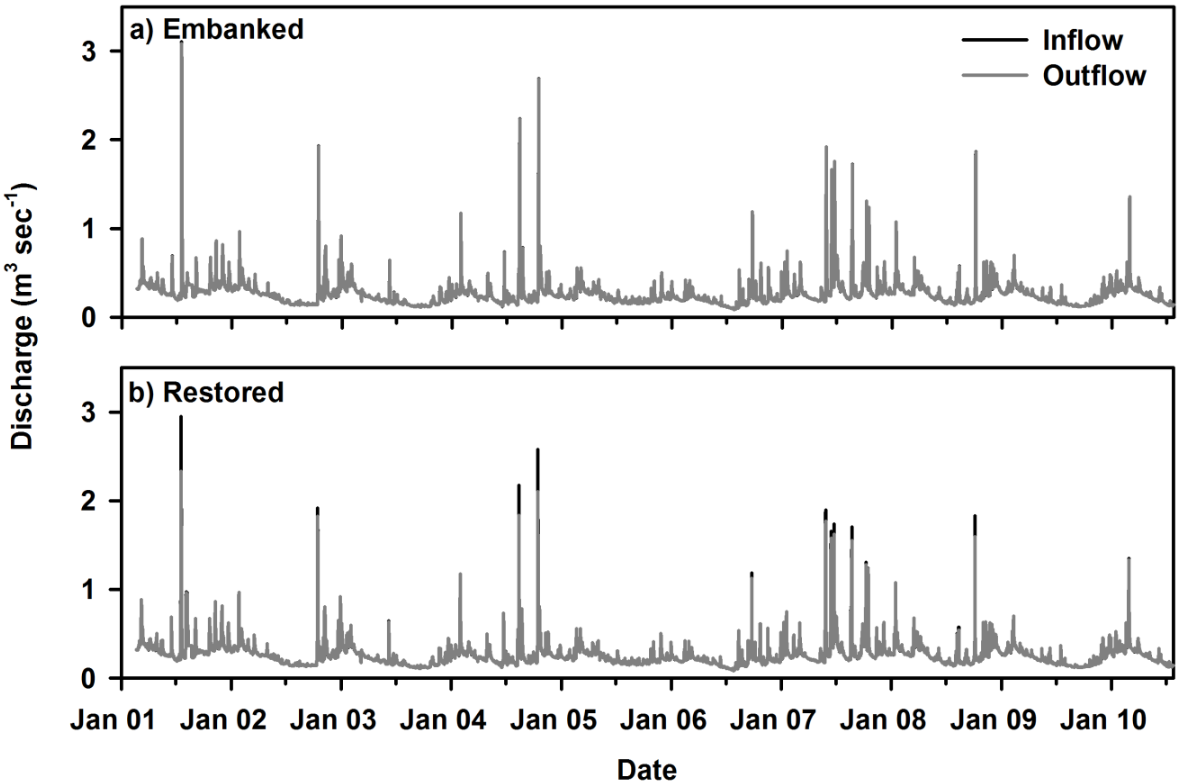


Figure 7.17: Comparison of modelled mean daily river inflow and outflow for the (a) embanked and (b) restored scenarios.

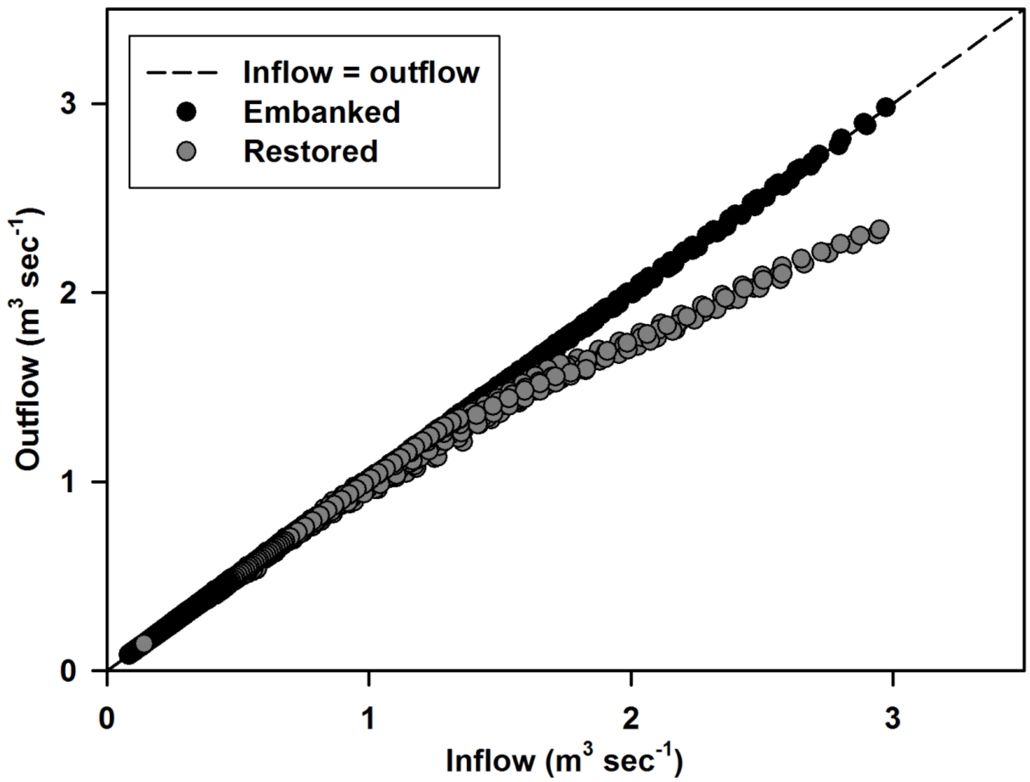


Figure 7.18: Comparison of modelled hourly river inflow versus outflow before and after the river restoration. Values above the solid line would indicate water loss (outflow > inflow) whereas values below the line indicate net retention (outflow < inflow) within the reach.

7.3 Discussion

7.3.1 Channel modification

River channelization and embankments constrain river flows within deeper, narrower cross-sections to reduce overbank flows and thus restrict hydrological connectivity between the river and floodplain. In contrast, the bankfull discharge of more natural river channels is generally thought to be in the range of the 1 – 2 year recurrence interval flood event (Darby and Simon 1999). Floodplain inundation is a major hydrological event that can attenuate downstream flood peaks through surface water storage and recharge of the floodplain aquifer, and create a more heterogeneous riparian habitat through flood disturbance and deposition of nutrient rich sediments (Tockner *et al*. 2000; Amoros and Bornette 2002; Naiman *et al*. 2010; Shrestha *et al*. 2014). Restoration of rivers to a more natural form is an increasingly accepted long-term solution for improving river health and functioning, and is likely to increase in practice, encouraged through legislative requirements of the Water Framework (Directive 2000/60/EC) and Flood Directives (Directive 2007/60/EC) and the interests of local groups (Richter *et al*. 2003; Perfect *et al*. 2013). Understanding how restoration affects river-flow dynamics and connections with the floodplain is necessary to be able to predict and evaluate the success of restoration schemes and guide future practices.

7.3.2 Simulation of floodplain hydrological processes

Observed groundwater levels on the floodplain of Hunworth Meadow before and after the embankment removal were simulated well by the two coupled hydrological/hydraulic MIKE-SHE/ MIKE 11 models. The models successfully reproduced groundwater responses to high magnitude flood events although they overestimated groundwater levels at the base of the hillslope (e.g. Well 2.4). This may be because either the model grid resolution was unable to sufficiently represent the topography of the ditch and its immediate surroundings, or the MIKE SHE drainage function did not adequately simulate drainage towards the topographic lows in the region of the ditch. Nonetheless, the coupled MIKE SHE/MIKE 11 models were able to adequately predict temporal changes in groundwater levels across the floodplain, capturing intra-annual variations in these levels associated with climate as well as changes in hydrological fluxes related to the restoration. Sensitivity analyses during model calibration revealed that the models were responsive to the overland Manning’s coefficient. Greater resistance to flow on the floodplain (e.g. applying Manning’s n values for woodland versus grassland) reduced overbank flow depth, and after flooding, increased flood retention on the meadow. This demonstrates the importance of vegetation type for the management of riparian lands for reducing flood risk downstream (e.g. Piegay 1997; Tabacchi *et al*. 2000).

The results from the two models developed using identical hydrometeorological conditions but with different topographical characteristics to reflect pre- and post-restoration conditions indicate four main hydrological responses to embankment removal on the River Glaven: (1) an increase in the frequency at which bankfull discharges are exceeded and in turn overbank inundation of the floodplain which was not simulated under embanked conditions, (2) increased groundwater levels and subsurface storage within the floodplain, (3) increased overland storage on the floodplain surface, especially during winter, and (4) moderate declines in downstream flood peaks. These responses are consistent with those reported following embankment removal and ‘pond and plug’ meadow restorations (where floodplain alluvium is excavated to plug incised channels) on, for example, the River Cherwell, Southeast England (Acreman *et al.* 2003), the headwaters of the Feather River, Northern California (Loheide and Gorelick 2007), and Bear Creek, Northern California (Hammersmark *et al*. 2008).

7.3.3 River-floodplain connectivity

A major aim of the river restoration was the re-establishment of hydrological linkages between the river channel and floodplain. Model results suggest that prior to restoration the embankments restricted river flows to the channel, which limited river-floodplain hydrological exchange to slow lateral subsurface flow (See Chapter 5). Removing the embankments has restored overbank water transfers onto the floodplain, modifying the floodplain’s hydrological regime, to form a more natural and dynamic wetland ecotone driven by flood disturbance. Widespread inundation occurred across the floodplain during high river flows (>1.7 $m^3 s^{-1}$), and reached as far as the hillslope (ca. 50 m from the river). Large overbank flows were of short duration (around a day) and were separated by large time intervals (2.9 year return period). Localised inundation of the immediate riparian area (within 5 m of the channel) was a much more frequent event (0.22 year return period). Increased river water incursions on to the floodplain is likely to improve continuity with groundwater, and enhance the supply of river nutrients to soil microbes and plant roots, an important influence on species composition, richness, primary productivity, and nutrient cycling (e.g. nitrification, denitrification, methanogenesis) within wetland environments (Hedin *et al*. 1998; Pinay *et al*. 2002; Amoros and Bornette 2002; Clilverd *et al*. 2008).

The groundwater regime is one of the most important factors determining the plant communities that are present on floodplains (Silvertown *et al*. 1999; Castelli *et al*. 2000). Hydrological models such as MIKE SHE therefore provide useful tools for evaluating the effects of river restoration on water table depths, which can in turn be used to predict shifts in vegetation communities and guide floodplain management (e.g. Thompson *et al.* 2009). At Hunworth Meadow, groundwater levels responded differently across the floodplain to embankment removal. Substantial increases in groundwater levels (0.4 – 0.6 m) occurred at the river-floodplain margin, where connectivity with the river was greatest and frequent localised overbank flooding occurred. This resulted in increased surface soil saturation throughout the year, which is likely to promote colonisation by wetland plant species that can tolerate waterlogging (e.g. Wheeler *et al*. 2004). Restoration also improved drainage between flood events, which could reduce flooding stress and lessen the impact of large floods on plant communities during the growing season. Smaller increases in water table elevation occurred as distance from the river increased, with the exception of the ditch area which received flood waters during large overbank events. As a result, the effects of restoration on floodplain biota are expected to vary spatially across the floodplain. Surface flooding and consequent surface water storage increased the volume of subsurface storage and reduced aquifer head recession over the summer. This was due to increased surface water inundation at the river-floodplain margin, and ponding of flood water in topographic depressions on the floodplain. The simulated increases in groundwater levels and subsurface storage in this study are consistent with modelled increases in groundwater levels simulated by Hammersmark *et al*. (2008) using a MIKE SHE model of floodplain restoration in Northern California.

7.3.4 Floodplain storage and flood-peak attenuation

Prior to restoration, model results suggest that bank storage contributed to a slight (maximum 3%) decrease in downstream flood peaks. River water intrusion increased during periods of elevated river stage, which reversed the hydraulic gradient on the floodplain and directed some subsurface flow away from the river. However removal of the embankments resulted in a substantially more marked response in flood-peak attenuation. Most of the overbank water was stored temporarily on the floodplain surface and in the ditch. Most flood water returned to the channel downstream with improved drainage being facilitated by embankment removal whereas prior to restoration embankments acted as a barrier for surface water exchange from the floodplain to the river. While some overbank water was infiltrated no noticeable changes in baseflow due to return flows occurred following inundation events.

Before embankment removal, the floodplain at Hunworth Meadow was a groundwater-dominated system. Rapid groundwater recharge occurred in response to precipitation and rising river levels, likely associated with pressure differences across the floodplain (e.g. MacDonald *et al*. 2014). During high winter river flows, groundwater was typically close to the soil surface, which limited the capacity for subsurface storage. Increased storage was available in soils in the summer. Therefore after restoration the greatest attenuation of flood-peaks occurred when floods followed a period of low rainfall (in particular during warm and dry summers). Although restoration increased surface water inundation and surface water storage, total evapotranspiration was unchanged. This was attributed to the rapid response of groundwater to river levels and subsequent groundwater flooding that resulted in saturated surface soils in both pre- and post restoration conditions. This response may vary in different hydrogeological settings, where evapotranspiration from inundated areas may act to reduce overland runoff and further attenuate flood peaks.

Expansive inundation and storage of flood waters on Hunworth Meadow resulted in a maximum reduction in peak river flows of 6 – 24%, along the length of restored reach (ca. 400 m). This is a similar contribution to flood-peak attenuation reported by other modelling studies. For instance, reductions in peak flows of 10 – 15% were simulated along a 5 km reach of the River Cherwell, UK (Acreman *et al*. 2003), and 13 – 25% reductions in river discharge were reported along 3.6 km of restored channel at Bear Creek, Northern California (Hammersmark *et al*. 2008). Logically, providing increased room for flood water storage on floodplains favours greater reductions in flood peaks, which is an appreciable benefit of river restoration.

Many recent reviews have identified the need for larger-scale restorations that include an environmental management plan for the catchment as a whole, particularly where problems persist throughout the catchment, e.g. agricultural fertiliser runoff, habitat fragmentation, urbanisation (Harper *et al.* 1999; Wharton and Gilvear 2007; Bernhardt and Palmer 2011). Indeed, the restoration project at Hunworth Meadow is part of a wider landscape approach to restoration being implemented along the River Glaven to reconnect and buffer an array of aquatic habitats of varying sizes (e.g. rivers, streams, ponds and ditches), with the aim of repairing autonomous river processes and associated ecosystem services (e.g. biodiversity, water quality) within the catchment (Sayer 2014). The removal of embankments along other reaches of the river that is proposed as part of this project could therefore be expected to have a cumulative impact of flood-peak recession.

7.3.5 Climate

River restoration, and the associated improvements to river-floodplain functioning (e.g. enhanced hydrological connectivity, groundwater retention, and flood-peak attenuation), may provide an important tool for buffering the hydrological regime of wetlands and other aquatic ecosystems against some of the extreme climate variability predicted over the next century (IPCC 2014). In the UK, five of the six wettest years have occurred since 2000, and eight of the warmest years have all occurred since 2002 (Met Office 2015a). The wettest May to July on record since 1766 occurred in 2007 during the observational period of this restoration study (IPCC 2014). Indeed, 2014 was the wettest winter and warmest year in the UK for over 100 years, suggesting a trend towards warmer and wetter weather (Met Office 2015b). The majority of climate change scenarios for the UK predict that the frequency and magnitude of floods will increase due to increased winter precipitation (Wilby *et al*. 2008; Thompson 2012). Increases in air temperature will also likely alter evapotranspiration rates and groundwater recharge, which is likely to affect wetland species that are sensitive to changes in hydrological regime (e.g. Gowing *et al*. 1998, Araya *et al*. 2011).

For example, a climate impacts study conducted by Thompson *et al*. (2009) using MIKE SHE/MIKE 11 and UK Climate Impacts Programme (UKCIP) projections for the 2050s simulated lower water table depths and reduced magnitude and duration of surface water inundation within the Elmley Marshes, Southeast England. It was suggested that these hydrological changes would lead to a loss of specialist wetland plants adapted to the current high water tables. Similarly, House *et al.* (2016a) used a MIKE SHE/MIKE 11 model of a riparian wetland on a tributary of the River Thames to demonstrate spatially varying hydrological impacts due to climate change that would have implications for both wetland flora and fauna.

Chapter 8: Methods – Part III: botanical and soil chemistry data collection and analysis

8.1 Introduction

This chapter describes the methods used to assess vegetation response to river restoration and embankment removal. It includes details of the fine scale botanical surveys, analyses of plant available nutrients, estimation of plant aeration stress using a cumulative stress index (e.g. Gowing *et al.* 1998), and measurements of dissolved oxygen concentration in the rooting zone using a novel oxygen optode technique. In addition, a description is given of the statistical analyses that were used to assess the importance of soil moisture and nutrient status on plant community composition.

8.2 Floodplain plant community composition

8.2.1 Vegetation composition

Floodplain vegetation surveys prior to the restoration were conducted in late June 2008 across the entire meadow on a regular 10 × 10 m sampling grid using 1 m^2 quadrats (n=206) (see Figure 4.4a). Percentage composition of all plant species was estimated visually and typically exceeded 100% due to vegetation layering (Figure 8.1). Soil moisture content was measured on 30/06/08 at a subset of survey points (n = 138) aligned to the botanical grid using a ML-2X soil moisture ThetaProbe (Delta-T Devices, Cambridge, UK) (see Figure 4.4b).

The main vegetation types for Hunworth Meadow were classified according to the National Vegetation Classification (NVC) (Rodwell 1992) using CEH TABLEFIT (Hill 1996b). TABLEFIT measures the degree of agreement (goodness-of-fit) between vegetation samples and the expected species composition of each NVC vegetation type (Hill 1996b). NVC communities were assigned for the middle meadow, ditch and river embankments using the average goodness of fit value produced in TABLEFIT.

Figure 8.1: Photograph of Hunworth Meadow showing the multiple layers of vegetation, taken in a NW direction, perpendicular to the River Glaven, in July 2007.

Diversity patterns of the floodplain vegetation were analysed using the Shannon Index, which is the most commonly used measure of α-diversity (Jost 2006; Reddy *et al*. 2009; Stein *et al*. 2010). This index describes both species richness and the abundance distribution of a sample or site (Magurran 2004). Species richness was calculated as the total number of species per 1 m^2 quadrat. The Shannon Index (H') expresses heterogeneity of an assemblage, and was calculated as:

$$H' = -\Sigma \, p_i \ln p_i \qquad (8.1)$$

where p_i is the proportional abundance (% cover) of the *ith* species. The Shannon index value is usually between 1.5 − 3.5 and rarely exceeds 4 (Magurran 2004).

8.2.2 Floodplain topography

Surface elevation of the meadow and river embankments (see methods in Section 4.4.4) was surveyed at each point on the botanical survey grid at intervals of 10 m, whereas the river embankments were surveyed at a higher resolution using intervals of approximately 0.25 – 0.5 m (Figure 4.14.). Digital elevation models (resolution 1 × 1 m), were created in ArcGIS using the kriging interpolation method which (as discussed in Sections 4.4.4 and 6.2.1) estimates values from a statistically weighted average of nearby sample points (de Smith et al. 2007). The same approach was used to create a soil moisture map (resolution 10 × 10 m), from the ThetaProbe data described above in Section 8.2.1.

8.3 Soil physicochemistry

8.3.1 Soil extractable ions

Soil extractable ions were measured to examine the links between plant available nutrients and the spatial variations in plant community composition. Floodplain soils were collected on 29/04/2008 across the meadow at shallow rooting depths of 0.1 – 0.3 m (n=113), which were observed for grasses on the meadow during soil sampling. This method does not allow for a detailed analysis of overall plant-available nutrients throughout the different soil horizons, however, since this would require an even more extensive sampling and analytical approach.

Soil samples were stored in a cooler with ice until returned to the laboratory, where they were refrigerated. Any samples that could not be analysed within two days were frozen. In the laboratory, plant available nutrients were determined using the following extraction methods: for analysis of nitrate, nitrite, and ammonium, 100 ml of 1M potassium chloride was added to 10 g of soil (Robertson *et al.* 1999); for determination of potassium, calcium, magnesium, sodium, aluminium, and iron, 100 ml of 1M ammonium acetate solution was added to 10 g of soil (Hendershot *et* *al.* 2008); for analysis of dissolved organic carbon, 100 ml of deionised water was added to 10 g of soil (method amended from Robertson *et al.* 1999); and phosphate was analysed using the Olsen *et al.* (1954) sodium bicarbonate extraction method (Schoenau and O'Halloran 2008). Prior to extraction, roots were removed from the field-moist soil, and the soil was mixed and passed through a 2 mm sieve. Following extraction, the samples were placed on a mixing table for 24 h, after which 50 ml subsamples were centrifuged at 3,000 rpm for 20 minutes (method amended from Robertson *et al.* 1999). The supernatant was filtered using 0.45 µm Supor® hydrophilic polyethersulphone membrane filter paper and frozen pending analysis.

Table 8.1: Soil chemical extractants

Soil extract	**Extraction chemicals**	**Soil weight (g)**	**Extraction volume (ml)**	**Analytical method**
NO_3^-, NO_2^-, NH_4^+	1M KCl[a]	10	100	Skalar San++
K^+, Ca^{2+}, Mg^{2+}, Na^+, Al^-, Fe^{3+}	1M CH_3COONH_4[b]	10	100	ICP[2]
TOC	Dionised water[a]	10	100	TOC analyzer
Olsen P	0.5M $NaHCO_3$, 0.0013M NaOH, and Polyacrylamide solution[c]	2.5	50.25	Skalar San++

[a]*Robertson et al., (1999);* [b]*Hendershot et al., 2008;* [c]*Schoenau and O'Halloran, 2008.*

The percentage moisture content was determined for each soil sample by drying triplicate 10 g subsamples of sieved field-moist soil at 105 °C overnight (Robertson *et al.* 1999). This allowed the respective ion concentration for each extract to be corrected for dilution. Inorganic nitrogen species (NO_3^-, NO_2^- and NH_4^+) and phosphorus were analysed colorimetrically using an automated continuous flow analyser (SAN++, SKALAR, Delft, The Netherlands) following the standard San++ methods for preparation of reagents. Base cation analysis (K^+, Ca^{2+}, Mg^{2+}, Na^+, Al^-, Fe^{3+}) was conducted using a Vista-PRO inductively coupled plasma optical emission spectrometer (ICP-OES) with a SPS3 autosampler (Varian, The

Netherlands). All soil extractable ions are reported with respect to dry weight of soil.

Dissolved organic carbon (DOC) and total dissolved nitrogen (TDN) were determined using a standard high temperature (1000°C) combustion method (Thermo 2006), on a HiPerTOC carbon analyser plumbed to an HiPer5000 total nitrogen chemoluminescent detector (Thermo Electron Corp., Delft, The Netherlands). Prior to analysis of DOC, inorganic carbon was removed with the addition of 1M Hydrochloric acid. For all analytes, calibration standards were prepared fresh each week. Laboratory control samples of a known concentration, blank samples, and duplicates (sampling and analytical), were interspersed throughout the sample runs approximately every 10 samples.

8.3.2 Soil pore water chemistry

As discussed in Section 4.4.2, pore water in the rooting zone was collected bimonthly from April 2007 to June 2008 using 0.1 μm rhizon samplers (see Figure 4.12b). The pore water samples were stored in a cooler until return to the laboratory, and then refrigerated. Any samples that could not be analyzed within two days were frozen. In the laboratory, cations (Ca^{2+}, Mg^{2+}, K^+, Na^+,NH_4^+) and anions (SO_4^{2-}, NO_3^-, PO_4^{3-}) were analysed by ion exchange chromatography (ICS-2500, Dionex Corp., CA, USA).

8.3.3 Analysis of total carbon and nitrogen content in soil

Total carbon (TC) and total nitrogen (TN) soil contents (%) were determined by elemental analysis (FLASH 1112 elemental analyser (EA) Thermo-Finnigan, Bremen, Germany) using an aspartic acid standard (C: 36.09 %; H: 5.30 %; N: 10.52 %; O: 48.08 %) (Skjemstad and Baldock 2008). Prior to analysis, soil samples were dried at 50 °C overnight, and ground and homogenised using a agate mortar and pestle. All apparatus was cleaned thoroughly between samples with acetone. Approximately 5 − 10 mg of soil was inserted into tin caps (8 × 5 mm, Elemental Microanalysis Ltd, Okehampton, UK) on a 6 decimal point precision microbalance (Supermicro, Sartorius UK Ltd).

8.3.4 Air-filled porosity

To determine temporal variation in soil aeration status with changes in groundwater elevation, the water release characteristic (pF-curve) of topsoil along Well Transect 3 (which was located near the oxygen optodes) (see Figure 4.2) were measured (n=15). This gave the relationship between air-filled porosity and soil pore water tension (pressure head). As pressure head and water-filled pore space decrease air defuses into draining soil pores. Air-filled porosity can therefore be calculated from the difference in water content of soil at a particular pressure head and the water content at saturation (Barber *et al.* 2004). Soil samples were collected in bulk density tins from the top 0 − 0.1 m of soil at three locations at Hunworth Meadow (n=15). In the laboratory, the tins were sealed with nylon filter cloth, held tightly in place with two rubber bands. The samples were saturated in water for five days, after which a small amount of swelling (1 − 2 mm) was noticed in some samples which were trimmed so that sample volume was the same as tin volume.

Using a manual 08.01 sandbox (Eijkelkamp, Giesbeek, The Netherlands) at the Centre for Ecology and Hydrology (Wallingford), soil water content was determined at a range of pressures from pF 0 (saturation) to pF 2.0 (-100 hPa), equivalent to a water table elevation (i.e. pressure heads) of 0 to -1 m. Between measurements, samples were left to equilibrate for a minimum of two weeks at each pressure. Differences in sample volume due to shrinkage was estimated by measurement of vertical and horizontal shrinkage space between the soil sample and the bulk density ring, which ranged from 0 − 5 mm vertically, and 0 − 4 mm horizontally. Some samples exhibited bubbling and swelling as the soil was dried, presumably due to biological activity in the samples, hence these samples were excluded from the analysis. After the final sand table measurement at pF 2.0, samples were oven dried at 105 °C for 48 hrs to obtain the dry sample weight. Volumetric water content was calculated for each tension accounting for any change in sample volume due to shrinkage (Elliot *et al*. 1999) (Figure 8.2).

Soil water content decreased fairly quickly at low tensions, indicating the presence of large pores that readily emptied of water as the soil began to dry (Figure 8.2). Greater tension was required to remove water from the smaller soil pores, indicated by the flattening of the water release curve between tensions of 0.1 – 0.6 m, followed by the sharp decline in water content at higher tension (i.e. between 0.6 – 1.0 m). Saturated soil water content was estimated as 0.72 ± 0.04 (volumetric basis). Soil shrinkage was most pronounced at tensions greater than 0.2m. In some samples, despite the loss of water as the sample dried, the decrease in sample volume due to shrinkage kept the soil close to saturation and limited the aerated pore space (Figure 8.2).

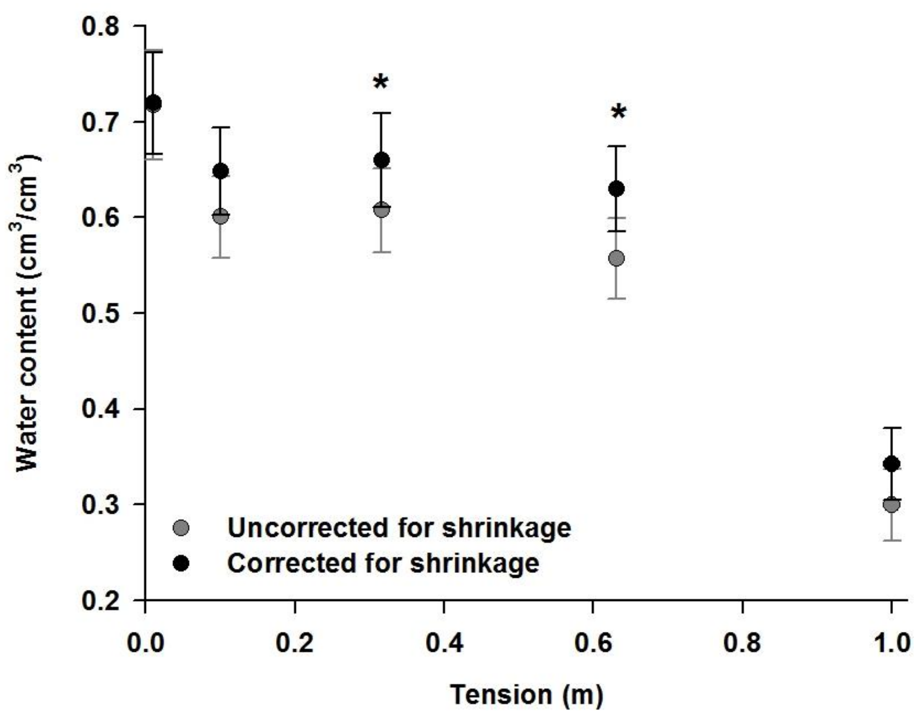


Figure 8.2: Soil water release characteristic (from 0.1m depth). Errors bar depict 95% confidence interval. * denotes significant differences ($p<0.05$) between the data corrected and uncorrected for shrinkage at each tension.

The effect of shrinkage and swelling on air-filled porosity should be a consideration in organic soils (Barber *et al*., 2004). Hence, differences in sample volume due to shrinkage (i.e. the shrinkage space between the soil sample and the bulk density tin), were measured and change in water content due to this shrinkage was calculated. Air-filled porosity accounting for shrinking and swelling soil was calculated as:

$$e_a = \left(\frac{\theta_{sat}}{\beta} - \theta\right) - \left(\frac{1}{\beta} - 1\right) \qquad (8.2)$$

where ea is air-filled porosity, θ_{sat} is saturated water content, θ is the water content of the sample, and β is the relative sample volume as a function of θ (defined below in Figure 8.3) (Barber *et al*. 2004). Sample shrinkage did not occur in all of the samples and, although the relationship between relative shrinkage of the soil and soil water content was significant ($p<0.05$), the correlation was fairly weak (r^2 = 0.33) (Figure 8.3). Thus while it is worth noting that soils are heterogeneous, and air-filled porosity may be slightly lower (and thus aeration stress higher) in areas on the meadow with more organic soils, which are more likely to swell and shrink, in order to use the cumulative aeration stress index described by Gowing *et al*. (1998), and standardise the results for comparison to other studies, air-filled porosity was calculated as the difference between the water content of the soil and saturated water content, assuming no shrinkage or swelling of the soil (i.e. a rigid soil).

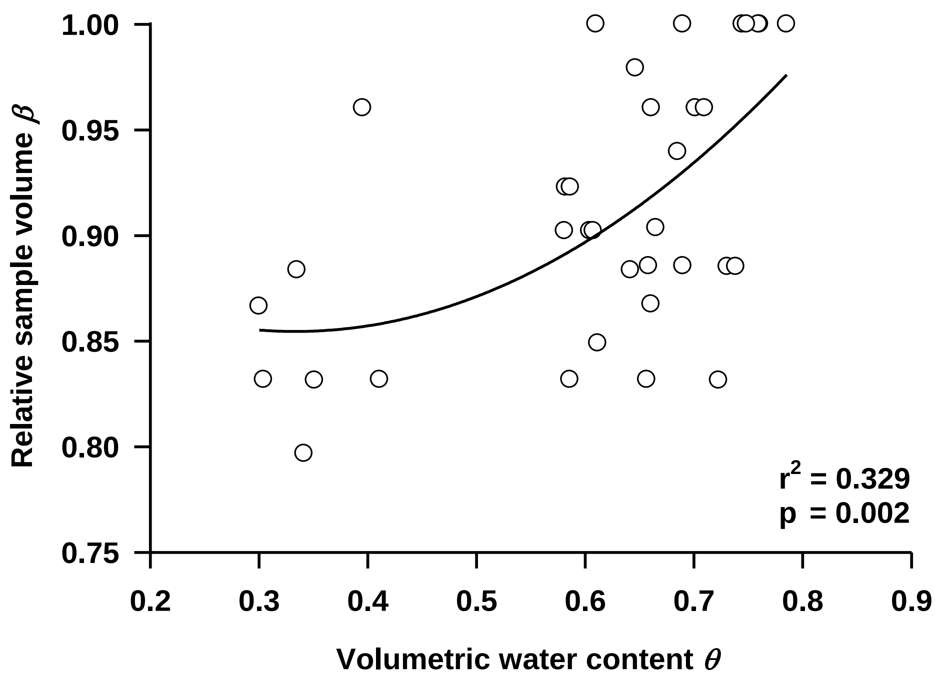


Figure 8.3: Relative shrinkage of the soil versus soil water content. A value of 1 indicates no shrinkage. $\beta = 0.5948\theta^2 - 0.3966\theta + 0.9207$.

Air-filled porosity increased exponentially ($p<0.05$) as tension increased as the water table dropped, and increased rapidly when the water table (i.e. pressure head) was below ~0.60 m (Figure 8.4). Assuming soil rigidity, the 0.1 (10%) threshold porosity expected for aeration stress in plants (Wesseling and van Wijk, 1975; Gowing *et al.* 1998), occurred at the surface when the water table was on average less than 0.34 m below the ground surface (Figure 8.4). This reference water table depth is comparable to a threshold depth of 0.35 m used by Gowing *et al*. (1998) to calculate sum exceedance values for aeration stress (SEV_{as}) within a peat-based wet grassland.

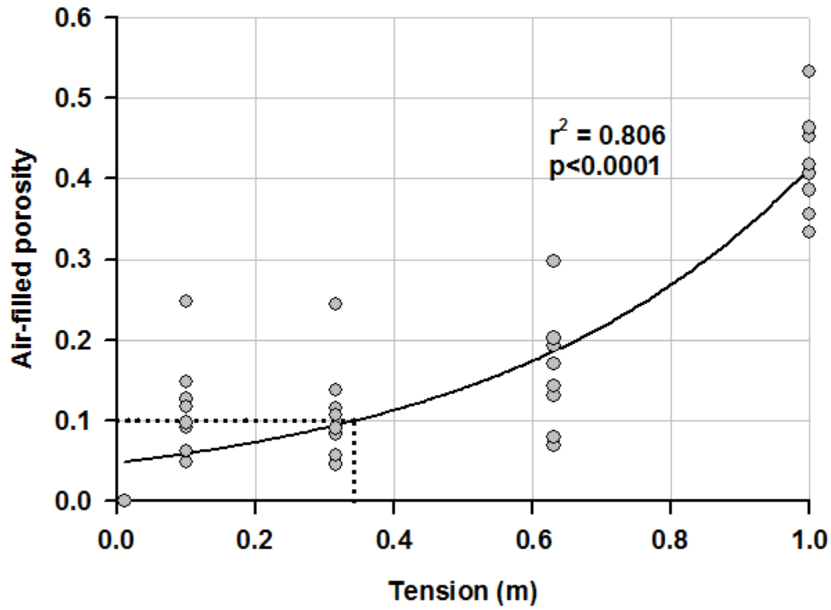


Figure 8.4: Relationship between water table depth below the soil surface (tension) and air-filled porosity for Hunworth Meadow soil. The solid line represents a statistically significant exponential model (y=0.0478exp(0.0216x)). The dotted drop lines indicate mean water table depth at 0.1 porosity.

8.3.5 Aeration stress index

An aeration stress index for each well location (see Figure 4.2) was estimated by determining sum exceedence values (SEV_{as}). This index uses water table position as a proxy for aeration stress under shallow water table conditions, i.e. where the water table is <1 m below the soil surface (Gowing *et al*. 1998; Silvertown *et al*. 1999). Aeration stress was calculated as the integral of the difference between the water table depth and a reference water table depth:

$$SEV_{as} = \int_{1}^{N} \left(D_{ref} - D_{W}\right) dt \qquad (8.3)$$

where SEV_{as} is the sum exceedence value in metre weeks, which increases in value with aeration stress; N is the number of weeks during the active growing season for grasses (March – September inclusive; Gowing *et al*. 2002b); D_{ref} is the reference water table depth (m) where air-filled porosity at the surface = 0.1 which, as described above, was taken to be 0.34 m; and D_w is the average depth to the water table (m). The integral was solved numerically for one week intervals and only positive values were included in the integration.

As described in chapters 6 and 7, pre-restoration and post-restoration water table elevations were simulated for a variety of possible hydrological conditions on the meadow using coupled surface water-groundwater models from which average water table depths during the growing season (March – September) and SEVs were calculated for the meadow. The pre-restoration (embanked) and post-restoration (no embankment) conditions on the meadow were simulated with two MIKE SHE/MIKE 11 models, which differed only in embankment and riverbed elevation as a result of the embankment removal (see Chapter 6.2.1). Fine model discretisation (5 m grid, see Section 6.2.1) was used to accurately characterise small-scale topographic variations in surface topography on the floodplain (e.g. the river banks, shallow depressions, and raised hummocks), which control soil water content, water table depth and influence habitat suitability for different plant species (Wheeler *et al*. 2004). In order to predict compositional change of plant communities on the meadow, in response to changing hydrological regimes associated with the removal of river embankments, sum exceedence values for the meadow were compared with known tolerances of meadow plant communities established by Gowing *et al*. (1998) and Wheeler *et al*. (2004). SEVs were also calculated from observed water table elevations from 2007 – 2009 to validate simulated SEVs.

8.3.6 Oxygen concentration in soil pores

To evaluate temporal fluctuations in oxygen concentration within the rooting zone and to test the use of water table depth as a useful proxy for aeration stress in plants, 4175 Aanderra oxygen optodes (Bergen, Norway), described in Section 4.4.3, were used to measure DO concentrations in saturated and unsaturated soil pore spaces in the rooting zone (0 – 0.10 m below the soils surface) at the upstream well transect from January to December 2010. Aanderra oxygen optodes measure oxygen partial pressure of dissolved oxygen and respond equally when measuring in air to when measuring in air-saturated fresh water (Aanderaa 2006; ADDI 2009) Hence, in addition to the measurements of oxygen concentration in groundwater, the optodes provided unique data on oxygen concentrations in soil air. The amount of oxygen dissolved in water at equilibrium with air is given by Henry's Law:

$$[O_2water] = P_{O2}air \bullet K_H \tag{8.4}$$

where O_2water is the concentration of DO in water (mol L^{-1}), $P_{O2}air$ is oxygen partial pressure in atmospheres (atm) and K_H is Henry's constant (mol L^{-1} atm^{-1}). Henry's law constants are dependent on temperature and salinity; in terms of the Bunsen solubility coefficients (β) (ml gas/l water) given in Weiss (1970), K_H is equal to β/V, where V is the volume of the gas at STP (O_2 = 22.4 l) (Weiss and Price 1980). The oxygen optodes were configured to record DO concentration in µmol L^{-1}, therefore equation 8.4 can be rearranged to give the oxygen partial pressure in the soil pores. Oxygen saturation values in air are calculated assuming 100% relative humidity, a pressure of 1 atmosphere, and zero salinity (Aanderaa 2006). Dry air can result in slightly higher oxygen measurements (up to 3%) than air that is 100% saturated with water vapour as dry air can hold more oxygen. This effect is expected to be minimal in soil, particularly at Hunworth Meadow where the water table was in close proximity to the buried optodes. Wetting and drying of the sensors occurred at the study site as the water table rose above and fell below the optodes, which can lead to a maximum error of 2% (Aanderaa 2006).

8.4 Statistical analyses

8.4.1 Linear regression models and diagnostics tests

Multiple linear regression was used to determine correlations between surface elevation, distance from the ditch and soil organic matter content on the meadow (predictor variables) and soil moisture content (response variable). Full and reduced models were compared in SAS 9.2 statistical software for Windows (SAS Institute Inc., North Carolina) using the PROC REG command. Boxplots for each of the predictor variables and the response variable provided basic range of validity, outlier, and influence information. Residual plots identified outliers and departures from the model assumptions. Normal probability plots of the residuals were used to identify non-normality of the error terms. Organic matter content data were log transformed to linearize the relationships and meet the assumptions of constant variance and normality. A correlation matrix of all the variables using PROC CORR computed Pearson correlation coefficients and was used to identify multicolinearity between the predictor variables.

8.4.2 Multivariate analysis

Spatial variation of plant community composition in relation to soil environmental conditions was analysed using ordination techniques in CANOCO 4.5 for Windows (Informer Technologies, Inc.). Since elevation and soil moisture contents were significantly correlated ($y = -23.301x + 523.45$, $r^2 = 0.52$, $p<0.05$), and soil moisture was not available for all plots (see Section 8.2.1), elevation was used in the subsequent analysis as a strong proxy of soil moisture content. Given the substantial turnover in the plant species data (>4 SD units), correspondance analysis (CA) was deemed most suitable for analysis of the spatial relationship in plant species composition. Unconstrained CA biplots of the relative position of species and samples along the two main explanatory ordination axes were used to visualize patterns in plant composition across the meadow (Lepš and Šmilauer 2003).

Spatial patterns of plant species in relation to soil environmental parameters were investigated using Canonical Correspondence Analysis (CCA) (Lepš and Šmilauer 2003). Prior to CCA, the data were tested for normality using normal probability plots and the Kolmogorov-Smirnov test. Where necessary, the data were log transformed to achieve normality. Multicollinearity among the environmental variables was tested using Pearson and Spearman correlation coefficients. Where independence was not found among environmental parameters, one parameter was removed from the analysis. After conducting the CCA, constrained bi-plots were used to investigate patterns in plant community composition in relation to soil environmental conditions.

Chapter 9: Simulation of the effects of river restoration on plant community composition

In this chapter, fine scale botanical and chemical data are used to assess the spatial pattern of plant communities in relation to soil physicochemical conditions, and water table simulations from coupled MIKE SHE-MIKE 11 hydrological-hydraulic models presented in Chapter 7. These data are used to predict aeration stresses in the rooting zone of floodplain plants prior to, and following the removal of river embankments from the River Glaven at Hunworth Meadow. The final set of research questions are addressed:

(v) What are the importance of soil moisture and fertility on plant communities on a disconnected floodplain?

(vi) What is the relationship between water table depth and oxygen content in the root zone?

(vii) What are the likely long-term impacts of floodplain restoration on the vegetation?

9.1. Results

9.1.1 *Groundwater hydrology and soil moisture gradients*

As demonstated in Section 5.2.2, groundwater levels within Hunworth Meadow were typically close to the soil surface. Using 50% time exceedence as a common reference value, mean water table depth during the wettest summer of the study period (2007) was between 0 and 0.4 m below the ground surface (bgs) at the upstream well transect (excluding Well 3.1 on the embankment) and between 0 and 0.2 m bgs at the downstream transect (Wells 1.1 – 1.6) for 50% of the time

(Figure 9.1). Hence for prolonged periods during the 2007 growing season the rooting zone (ca. 0 – 0.30 m bgs) was saturated, particularly near Well Transect 1 (Figure 9.1). Even during the driest summer (2009), the water table was frequently above the rooting zone. Elevated regions on the meadow (see Figures 5.2 and 6.2) such as the river embankments (Well 3.1) and areas near the hillslope (Well 3.5) remained dry for the majority of the summer (>95% of the time) (Figure 9.1).

Soil moisture and elevation were closely linked, showing a negative linear relationship ($r^2 = 0.52$, $p<0.05$; Figure 9.2). An average 23% decrease in soil water content resulted from an average 1 m increase in surface elevation on the meadow (Figure 9.2a). Organic matter content was strongly correlated with soil water content ($r^2 = 0.75$, $p<0.05$; Figure 9.2b). Organic matter content was weakly, though significantly, correlated with surface elevation (0.31; $p<0.05$) (Figure 9.2c) Thus, elevation, organic matter content and distance from the ditches accounted for 85% of the variation in soil water content (<0.05; Table 9.1). However, due to some multicolinearity between the predictor variables, surface elevation and organic matter content provided the most parsimonious model of soil water content ($r^2 = 0.846$; <0.05; Table 9.2).

Soil moisture increased in a northeasterly (downstream) direction in conjunction with a decrease in elevation (Figure 9.3). Soil water content was lowest (11 − 30%) along the river embankments, where surface elevation was between 21.3 (upstream) − 20.4 m ODN (downstream), and approximately 0.4 – 1.1 m above the main meadow (Figure 9.3a). Soil moisture content was near saturation in the region of the ditches, and at the downstream section where surface elevation was less than 20 m above Ordnance Datum Newlyn (ODN) (Figure 9.3a), and where surface water ponded due to the blocked drainage ditch (see Figure 7.8). Across the remainder of the meadow, soil moisture content ranged between 20 and 80% (Figure 9.3b).

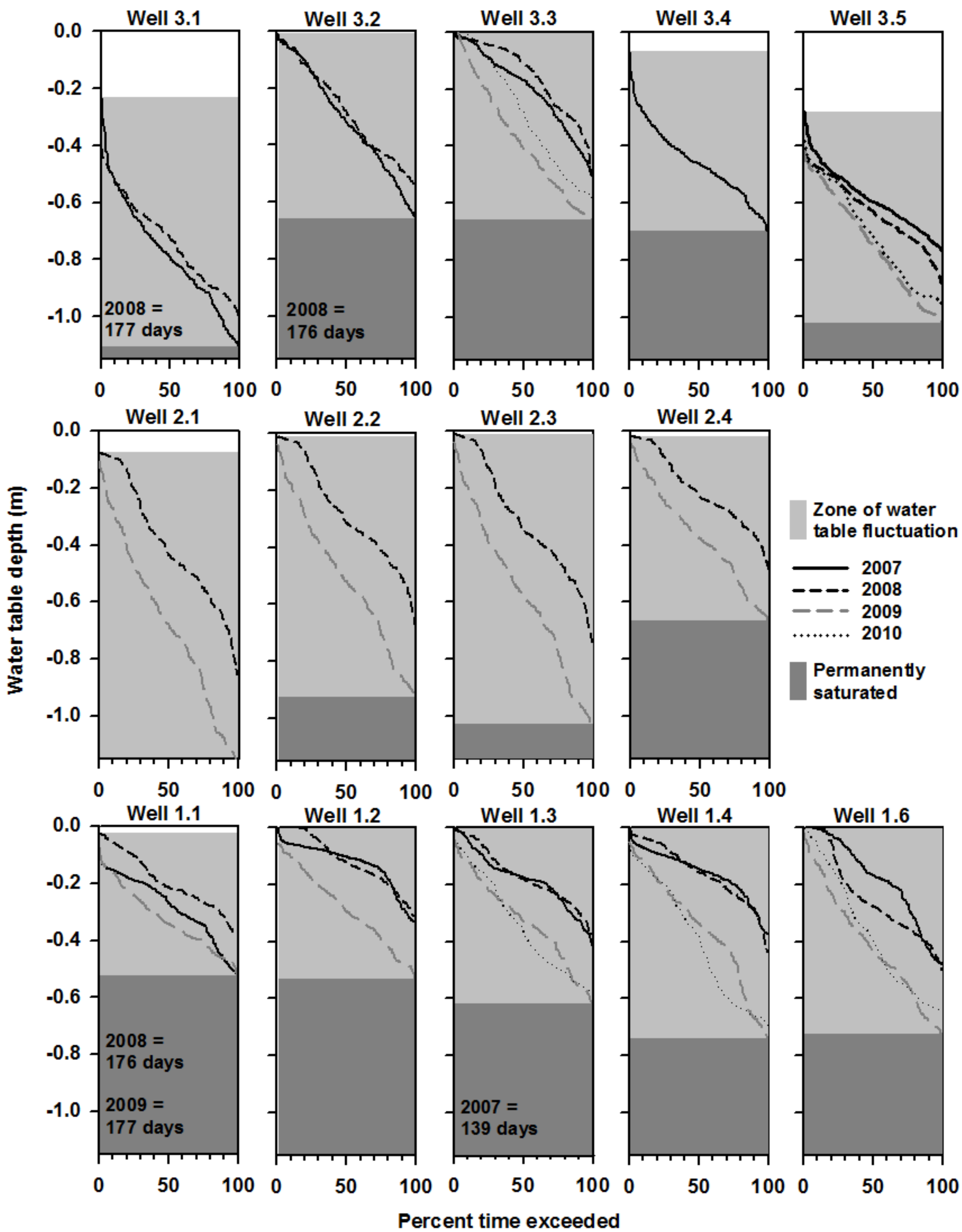


Figure 9.1: Water table duration curves derived from mean daily water table depth below the soil surface (0.0 m) during the growing season (March to September inclusive). The top, middle and bottom panels represent wells along the upstream (Wells 3.1 – 3.5), midstream (Wells 2.1 – 2.4) and downstream (Wells 1.1 – 1.6) well transects, respectively (see Figure 4.2).

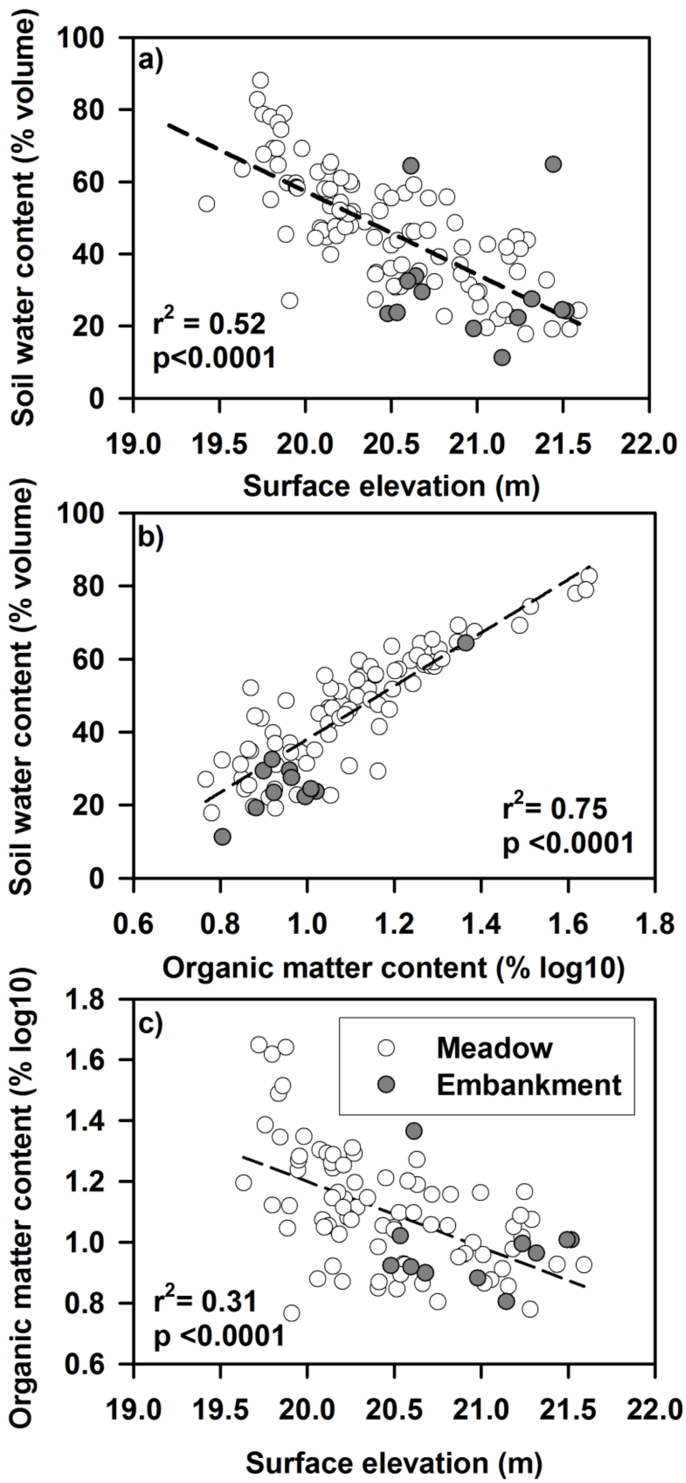


Figure 9.2: Correlations between (a) surface elevation on the meadow and water content of the soil (y = -23.301x + 523.45); (b) organic matter content (log transformed) and soil water content (y = 72.8504x + -34.79), and (c) surface elevation and organic matter content (log transformed) (y = -0.217 + 5.533).

Table 9.1: ANOVA and parameter estimates for predicting soil moisture content with the following parameter inputs: elevation, organic matter content, and distance from the river.

ANOVA				
R-Square	**Adj. R-Sq**		**F value**	**Pr > F**
0.8518	0.8468		168.65	**<.0001**

Parameter Estimates

Variable	**DF**	**Estimate**	**Error**	**t value**	**Pr > F**
Intercept	1	197.84	40.53	4.88	**<.0001**
Elevation (m)	1	-10.60	1.87	-5.66	**<.0001**
Organic matter (%)	1	59.02	4.47	13.22	**<.0001**
Distance to ditch (m)	1	-0.11	0.055	-1.93	**0.0571**

Table 9.2: ANOVA and parameter estimates for the reduced soil moisture model, using the following parameter inputs: elevation, and organic matter content.

ANOVA				
R-Square	**Adj. R-Sq**		**F value**	**Pr > F**
0.8456	0.8421		243.68	**<.0001**

Parameter Estimates

Variable	**DF**	**Estimate**	**Error**	**t value**	**Pr > F**
Intercept	1	232.51	36.88	6.31	**<.0001**
Elevation (m)	1	-12.41	1.65	-7.53	**<.0001**
Organic matter (%)	1	59.10	4.53	13.04	**<.0001**

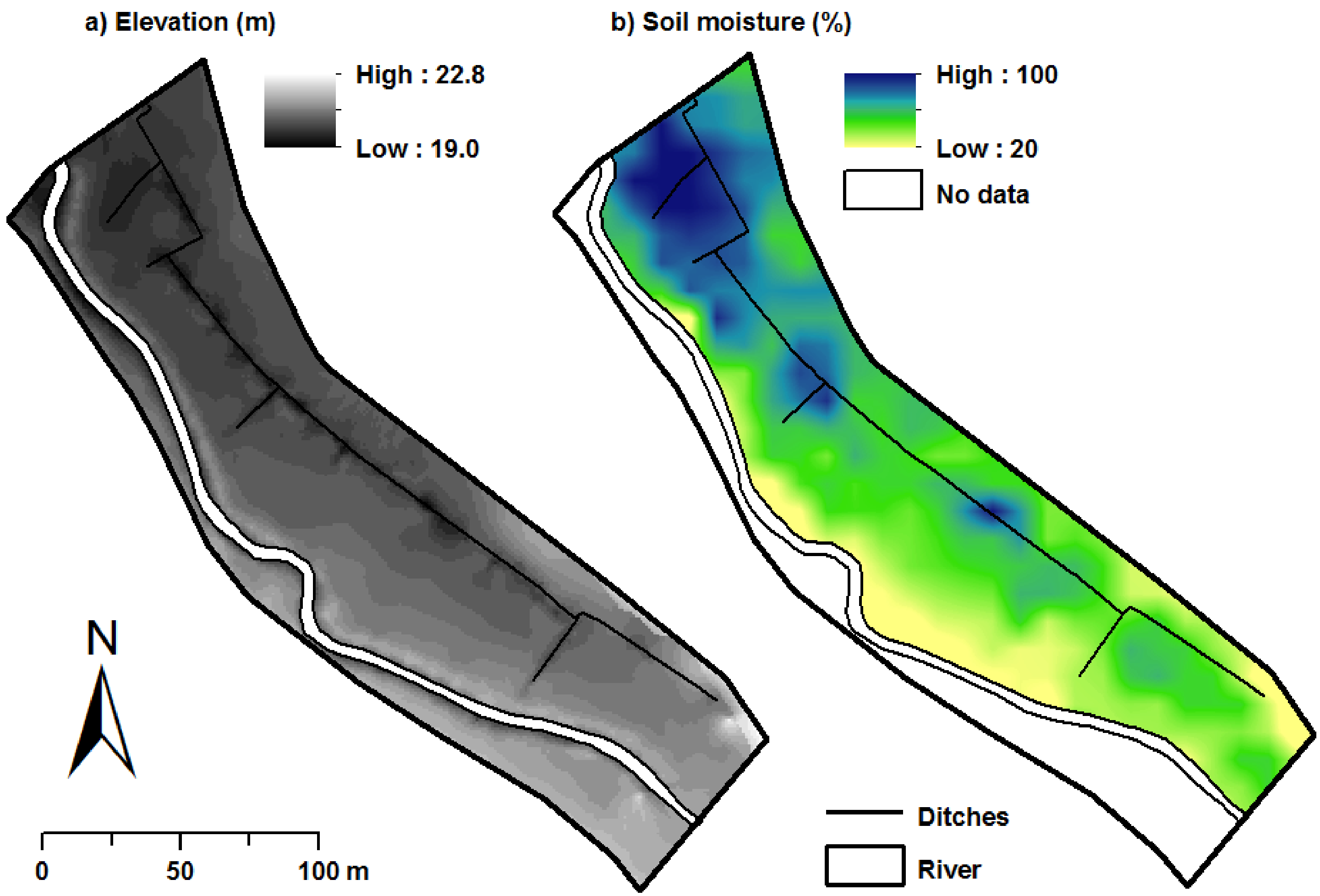


Figure 9.3: Spatial variation in (a) elevation and (b) soil water content in June 2008 on the embanked Hunworth Meadow. The digital elevation model was created using dGPS survey data collected in June 2008.

9.1.2 Soil Nutrient Status

Topsoils (ca. 0 – 0.3m) contained moderate levels of organic material (OM; range: 13 – 35% OM content), had low bulk density (mean: 0.69 g cm^{-3}), and were slightly acidic (mean *in-situ*-pH: 6.1) (Table 9.3). Analysis of plant available nutrients in the topsoil revealed moderately fertile conditions, with Olsen P and plant-available potassium concentrations in the range of 0.1 – 34 mg P kg^{-1} (mean: 9.2 mg P kg^{-1}) and 0.4 – 4.8 mg K g^{-1} (mean: 1.6 mg K g^{-1}), respectively. Plant available ammonium and nitrate concentrations ranged from 5.7 – 249.9 mg NH_4^+-N kg^{-1} (mean: 32.5 mg NH_4^+-N kg^{-1}) and below detection limit (i.e. 0.1) to 25.2 mg NO_3^--N kg^{-1} (mean: 2.9 mg NO_3^--N kg^{-1}), respectively. Soils were high in calcium, with an average of 2.6 mg Ca^{2+} g^{-1}, which far exceeded concentrations of magnesium (mean: 0.11 mg Mg^+ g^{-1}) and sodium (mean: 0.08 mg Na^+ g^{-1}). Plant available N:P ratios were low, averaging <5:1 (Table 9.3).

Plant available Olson P and ammonium concentrations were highest (>16 mg P kg^{-1} and >75 mg N kg^{-1}, respectively) in water-logged soils at the downstream end of the meadow (Figure 9.4). Plant available ammonium concentrations were also higher (>51 mg N kg^{-1}) in wet soils along the ditch margin (Figure 9.4b). In contrast, plant available nitrate was typically highest (>5 mg N kg^{-1}) in the drier, upstream soils (Figure 9.4c). Olsen P and nitrate concentrations were also raised (i.e. above 11 mg P kg^{-1} and 5 mg N kg^{-1}, respectively) in soils on the river embankments, and along the north-east boundary of the meadow (Figure 9.4) which borders the woodland and arable hillslope (see Figure 3.1). These higher values are in contrast to those in soils in the middle of the meadow, where concentrations of less than 5 mg P kg^{-1} and 1 mg N kg^{-1}, respectively, were more typical (Figure 9.4). Plant available potassium concentrations did not appear to show a clear spatial pattern across the meadow (Figure 9.5a). Soil extractable TOC and TON were positively correlated (r^2 0.64; $p<0.05$), and occurred in the highest concentrations at the downstream region of the meadow, and along the ditches (Figure 9.5b-c).

Table 9.3: Soil (n = 113) and pore water (n = 53) chemistry of Hunworth Meadow along the vegetation sample transects (mean ± 95% confidence interval).

Analyte	**Mean ± 95% confidence**
Soil:	
pH (laboratory)	6.53 ± 0.24
pH (*in-situ*)	6.09 ± 0.23
Organic matter content (%)	13.64 ± 1.50
Bulk density (g m^{-3})	0.69 ± 0.07
Ca^{2+} (mg g dry $soil^{-1}$)	2.64 ± 0.36
Na^{+} (mg g dry $soil^{-1}$)	0.08 ± 0.01
Mg^{+} (mg g dry $soil^{-1}$)	0.11 ± 0.01
K^{+} (mg g dry $soil^{-1}$)	1.64 ± 0.16
Total iron (mg kg dry $soil^{-1}$)	80.96 ± 35.18
Al^{3+} (mg kg dry $soil^{-1}$)	10.27 ± 1.54
NH_4^{+} (mg N kg dry $soil^{-1}$)	32.46 ± 5.02
NO_2^{-} (mg N kg dry $soil^{-1}$)	0.28 ± 0.09
NO_3^{-} (mg N kg dry $soil^{-1}$)	2.86 ± 0.75
Olsen P (mg P kg dry $soil^{-1}$)	9.20 ± 1.23
N:P ratio (NH_4^{+} + NO_3^{-} /Olsen P)	4.89 ± 0.61
TOC (mg kg dry $soil^{-1}$)	0.67 ± 0.11
TON (mg kg dry $soil^{-1}$)	0.10 ± 0.09
Pore water:	
Ca^{2+} (mg L^{-1})	119 ± 19
Na^{+} (mg L^{-1})	21.5 ± 5.7
Mg^{+} (mg L^{-1})	8.2 ± 1.5
K^{+} (mg L^{-1})	6.0 ± 2.8
NH_4^{+} (mg N L^{-1})	1.5 ± 1.2
SO_4^{2-}(mg S L^{-1})	3.1 ± 2.9
NO_3^{-} (mg N L^{-1})	0.5 ± 0.5
PO_4^{3-} (mg P L^{-1})	0.1 ± 0.02

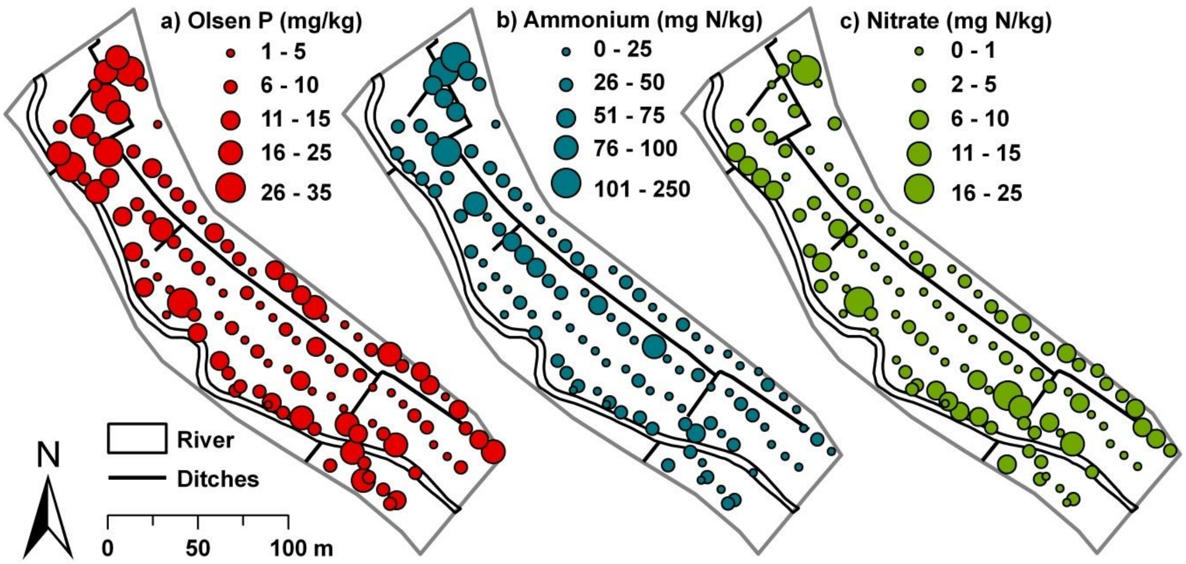


Figure 9.4: Spatial differences in plant available Olson P, ammonium and nitrate across Hunworth Meadow.

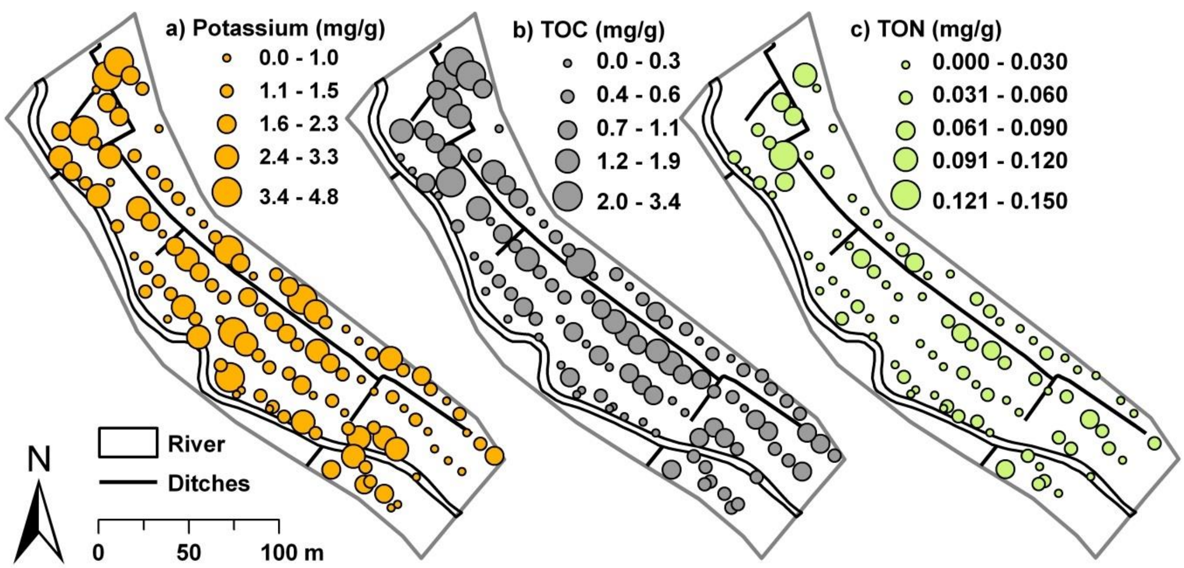


Figure 9.5: Spatial differences in plant available potassium, and TOC and TON across Hunworth Meadow.

The nutrient composition of root-zone pore-water was depleted in phosphate, nitrate, and ammonium, which averaged 0.06 mg PO_4^{3-}-P L^{-1}, 0.5 mg NO_3^--N L^{-1} and 1.5 mg NH_4^+-N L^{-1}, respectively (Table 9.3). Mean ammonium concentration was approximately seven times greater than the median concentration (0.21 mg NH_4^+-N L^{-1}) due to a number of outliers. These occurred during a period of sheep grazing and were possibly related to high concentrations of nitrogen in urine patches. Average soil pore-water nutrient concentrations were not significantly different to concentrations measured in groundwater wells on the meadow, reported in Section 5.2.6 (p values: phosphate = 0.36; nitrate = 0.16, ammonium = 0.07 including outliers, and 0.78 excluding 3 outliers).

9.1.3 Community composition

The vegetation on Hunworth Meadow was dominated by *Holcus lanatus*, *Ranunculus repens* and *Agrostis stolonifera* which were present in 85, 83 and 67% of the surveyed quadrats (Table 9.4), respectively. These three species, which had a mean relative percentage cover of 21, 23 and 7% across the meadow (Table 9.4), are indicative of damp, mesotrophic to richly fertile soils, as indicated by their respective moisture and nitrogen Ellenberg's values (Table 9.4). *Arrhenatherum elatius* and *Festuca rubra* were also fairly common on the meadow, present in 27 and 43% of the samples, respectively, with a similar mean relative percentage cover to *Agrostis stolonifera*. Reflecting the variable hydrological conditions encountered on the meadow, a variety of wet- (e.g. *Juncus effusus, Glyceria fluitans*, *Equisetum palustre*) and dry-loving (*A. elatius*, *Urtica dioica*, and *Dactylis glomerata*) plants were recorded (Table 9.4).

Table 9.4: Frequency of presence and mean relative percent cover of plant species on Hunworth Meadow, and their typical Ellenberg's indicator values for tolerance of moisture and nitrogen (from Hill *et al.* 1999). Rare species (<5% presence) are not included in the table.

Species	**Frequency of presence (%)**	**Mean relative cover (%)**	**Ellenberg's indicator**	
			Moisture (1 – 12)*	**Nitrogen (1 – 9)****
Holcus lanatus	*85*	20.9	*6*	*5*
Ranunculus repens	*83*	22.8	*7*	*7*
Agrostis stolonifera	*67*	7.3	*6*	*6*
Festuca rubra	*43*	5.8	*5*	*5*
Rumex acetosa	*39*	1.6	*5*	*4*
Glyceria fluitans	*35*	3.9	10	6
x Festulolium loliaceum	*35*	1.4	N/A	N/A
Cerastium fontanum	*33*	0.8	*5*	*4*
Arrhenatherum elatius	*27*	7.0	5	7
Urtica dioica	*24*	3.1	*6*	*8*
Cirsium arvense	*23*	1.4	6	6
Poa trivialis	*23*	0.3	*6*	*6*
Equisetum palustre	*21*	0.8	8	3
Juncus effusus	*20*	7.1	*7*	*4*
Dactylis glomerata	16	2.3	*5*	*6*
Stellaria alsine	13	0.3	*8*	*5*
Rumex obtusifolius	12	0.6	*5*	*9*
Taraxacum officinale	11	0.4	N/A	N/A
Equisetum arvense	10	0.79	*6*	*7*
Cardamine pratensis	9	0.14	*8*	*4*
Anthoxanthum odoratum	8	0.18	*6*	*3*
Carex hirta	8	0.41	*7*	*6*
Ranunculus acris	7	0.21	*6*	*4*
Equisetum fluviatile	7	0.27	*10*	*4*
Phalaris arundinacea	7	2.96	*9*	*7*
Veronica chamaedrys	6	0.21	5	5
Gallium aparine	6	0.15	*6*	*8*
Carex acutiformis	5	1.34	*9*	*6*

**1 = extremely dry; 12 = submerged; **1 = extremely infertile; 9 = extremely rich; N/A = not available.*

A total of 88 plant species were encountered within the 206 quadrats. Average species richness was low at 8 species per m^2 (Figure 9.6a), but varied substantially between quadrats (range: 1 – 16 species per m^2) (Figure 9.7). Shannon index of diversity averaged 1.4 (Figure 9.6b), suggesting low heterogeneity of the plant assemblages (i.e. a high degree of dominance in communities) on the meadow. Plant diversity was generally low in the centre and in lower lying regions of the meadow, while patches of higher species richness were encountered on the embankments and along the ditch margins (Figure 9.7).

Correspondence Analysis (CA) identified three distinct vegetation groups or clusters: 1) vegetation on the river embankments, 2) vegetation along the ditch margins, and 3) the vegetation on the central sections of the meadow between the river and ditch (Figure 9.5). In the CA plot, the first axis explained 16% of the total species variability (Table 9.5). This main axis appears aligned to a gradient in soil wetness, with samples from the wet ditch margins located at one end of the axis and samples from the dry river embankments at the other. The CA plot furthermore showed an arch-effect, indicative of the quadratic dependence of the second ordination axis on the first ordination axis (Lepš and Šmilauer 2003). However, the DCA's exhibited a large amount of scattering, and as interpretation of DCAs is restricted to the first axis only, interpretation was not improved. Initial CA analysis identified some sample outliers that were vastly different from all other samples. These samples were located at the downstream ponded area of the meadow and were primarily composed of aquatic vegetation (*Lemna minor*, *Lemna trisulca* and *Potamogeton natans*). As this area was an aquatic habitat (see Figure 7.8) rather than a wet meadow these samples were excluded from further ordination analyses.

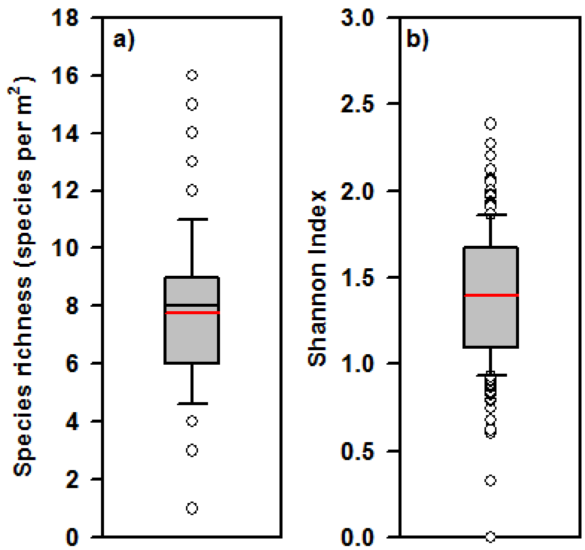


Figure 9.6: Species richness and Shannon Diversity Index on the embanked Hunworth Meadow. The black centre line, red centre line and box extent denote the median, mean, and 25th and 75th percentiles, respectively. Whiskers indicate the 10th and 90th percentiles, and the circles signify outliers outside of the 10th and 90th percentiles.

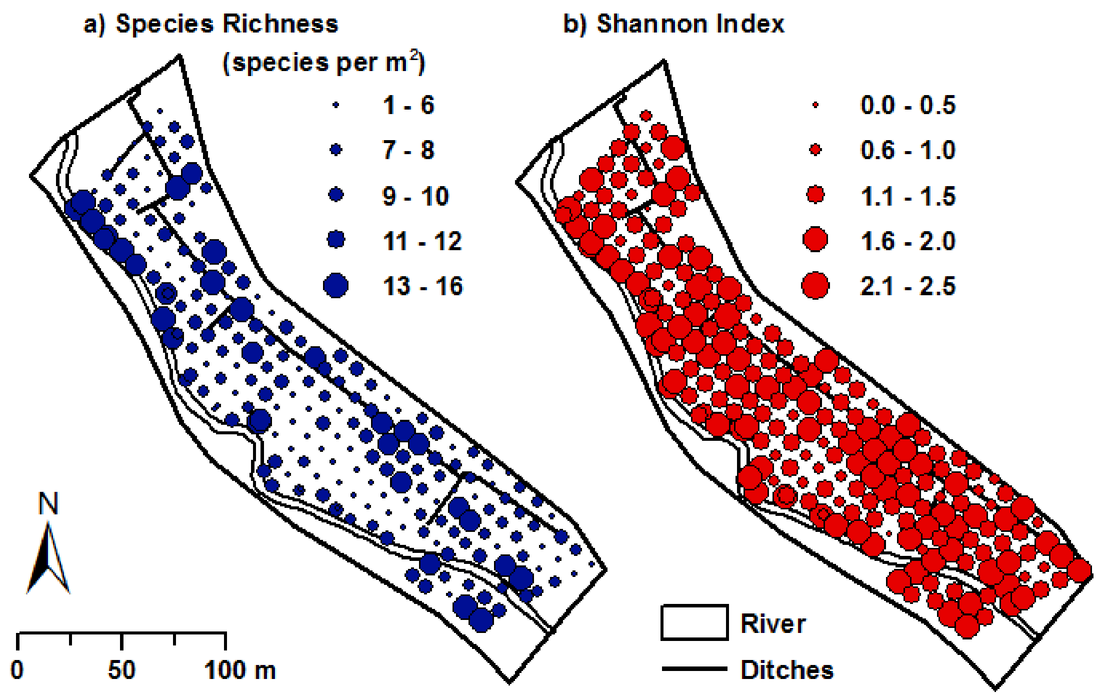


Figure 9.7: Spatial patterns of species richness and Shannon Diversity Index on the embanked Hunworth Meadow.

Vegetation assemblages along the river embankments were characterised by drought-tolerant species typically thriving on mesotrophic soils, such as *A. elatius*, *U. dioica*, *D. glomerata, Plantago lanceolata* and *Geranium molle* (Figure 9.8). Accordingly, the community was classified as MG1 *A. elatius* grassland (mean goodness of fit: 88). In contrast, the ditch margins were characterised by water-tolerant species often found on less fertile soils, such as *Carex acutiformis, Equisetum fluviatile*, *J. effusus* and *Juncus inflexus* (Figure 9.6) as represented by MG10 *H. lanatus* – *J. effusus* rush – pasture (Table 9.6). Unlike the plants encountered on the embankments, plants found along the ditches were also present in some areas of the meadow. Consequently the TABLEFIT results place the meadow community in the MG10 flood-sward, along with the OV28 *A. stolonifera* – *R. repens* grassland community (mean goodness of fit: 60 – 68; Table 9.6).

As the embankment habitat represented dry grassland rather than wet meadow, a second CA was performed where embankment samples were removed so that changes in the wet meadow plant communities could be examined in detail (Figure 9.7). Axis 1 explained 15% of the total species variability (Table 9.5) and again appeared to follow a gradient in plant moisture-tolerance, with water-loving plants (e.g. *E. fluviatile*, *J. effusus* and *J. inflexus*) negatively associated with axis 1, and less waterlogging tolerant species (e.g. *A. elatius*, *U. dioica*, *Lolium perenne* and *D. glomerata*) found at the other end of the gradient. Many of the meadow samples, however, were clustered together on the centre of the axis. Plant species closely associated with these samples were *H. lanatus*, *R. repens* and *Festuca pratensis*, with the dense clustering indicating little variation in community composition amongst these samples (Figure 9.7).

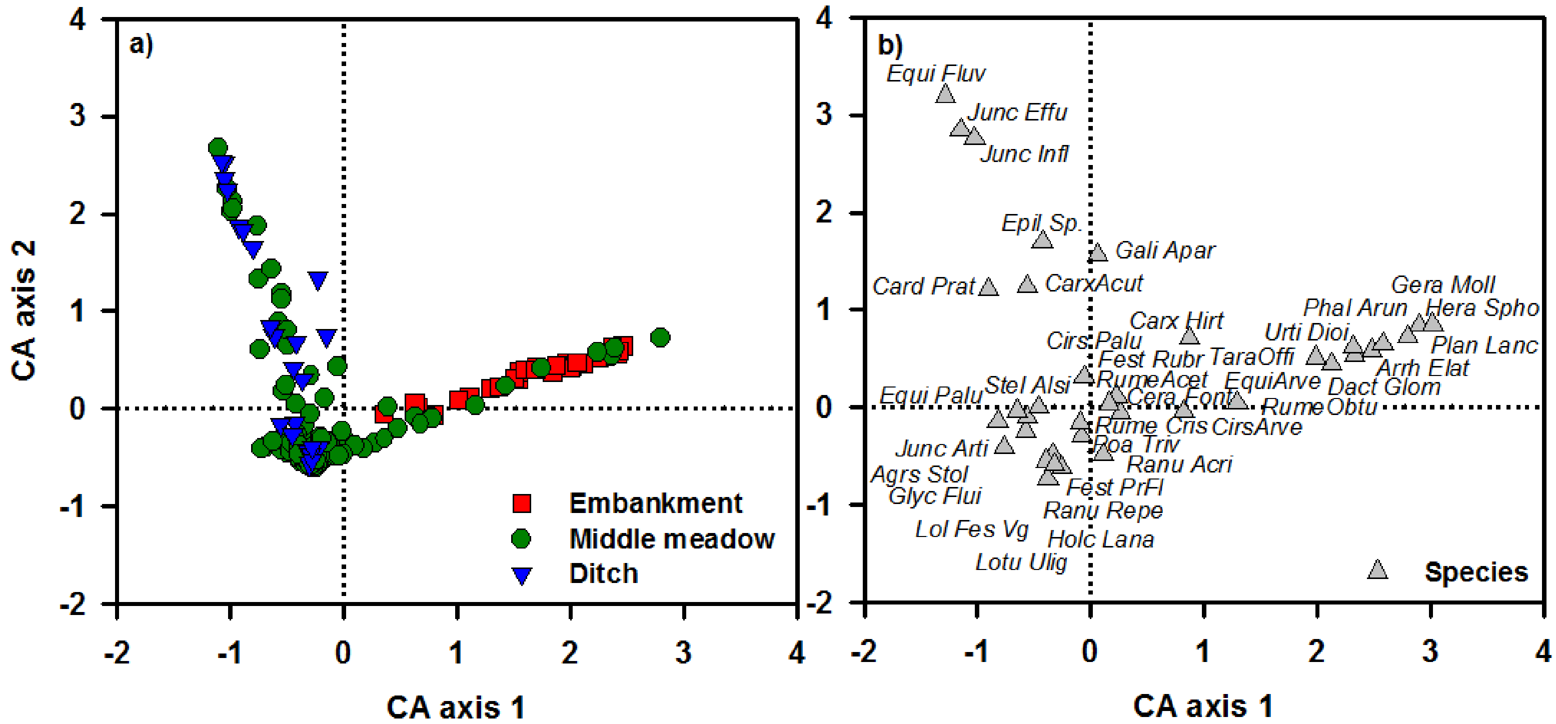


Figure 9.8: (a) Correspondence Analysis (CA) of the vegetation data showing (a) the embankment, middle meadow and ditch sample points (n=195) and (b) the associated species (n=80). The relative distance between the species points represents the similarity or dissimilarity of species relative abundance across the samples. To improve the visibility of species labels, rare species (<5% presence) were removed from the plot after the CA analysis.

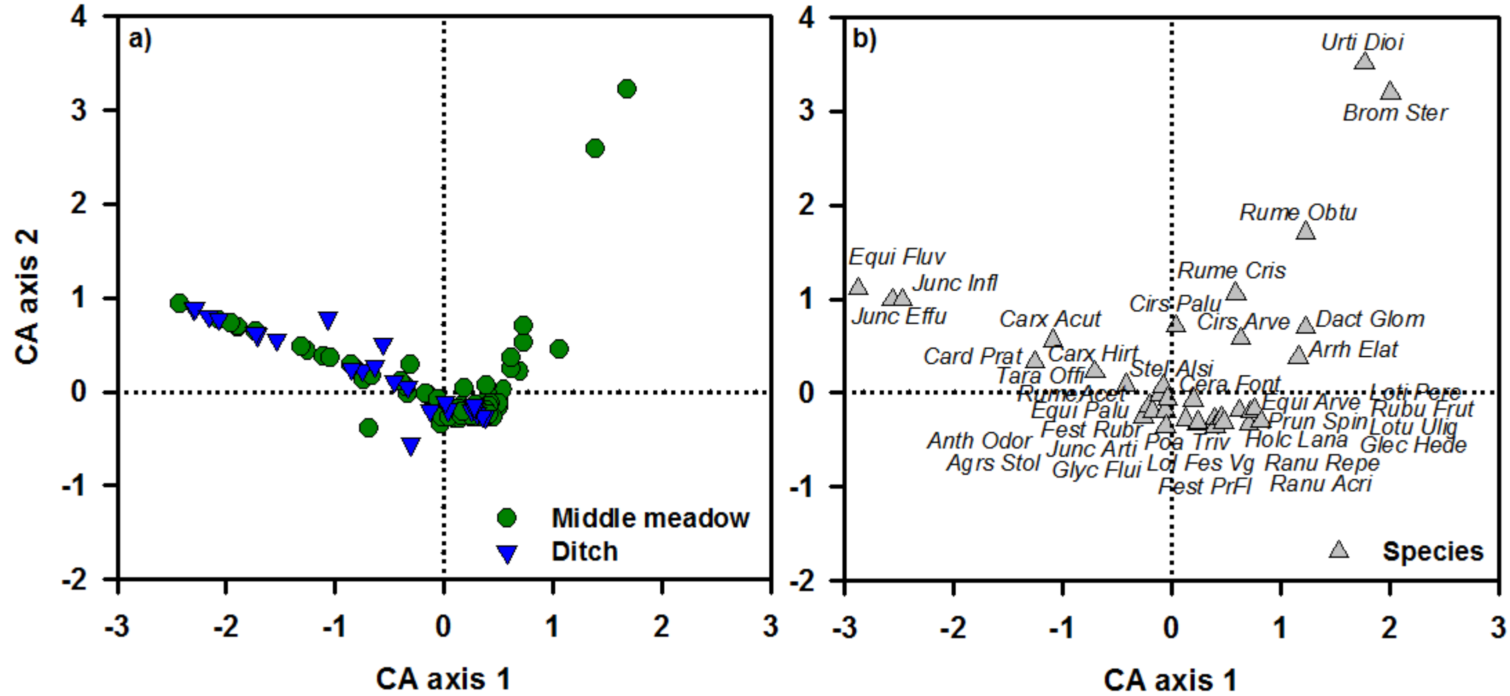


Figure 9.9: Correspondence Analysis (CA) of the vegetation data showing (a) showing the middle meadow and ditch sample points only (n=163) and (b) the associated species (n=67). The relative distance between the species points represents the similarity or dissimilarity of species relative abundance across the samples. To improve the visibility of species labels, rare species (<5% presence) were removed from the plot after the CA analysis.

Table 9.5: Eigenvalues and cumulative percentage variance for each CA axis of the vegetation data.

Axes		**1**	**2**	**3**	**4**	
Embankment, meadow, and ditch:	Eigenvalues:	0.674	0.554	0.447	0.373	
	Cumulative percentage variance of species data:	15.7	28.7	39.1	47.8	
	Sum of all eigenvalues:					4.283
Meadow and ditch:	Eigenvalues:	*0.572*	*0.542*	*0.391*	*0.321*	
	Cumulative percentage variance of species data:	15.4	29.9	40.4	49.1	
	Sum of all eigenvalues:					*3.721*

Since elevation and soil moisture content were significantly correlated (see Figures 9.2a and 9.3), and soil moisture was not available for all plots (see Section 8.2.1), elevation was used in the subsequent analysis as a strong proxy of soil moisture content. Changes in the micro-topography of the meadow that represented a proxy for variations in the soil moisture content were closely linked to vegetation changes in the constrained Canonical Correspondence Analysis (CCA) ordination plot ($r = 0.79$; $p<0.05$; Figure 9.8; Table 9.7). Soil fertility was a significant secondary predictor of plant species composition changes; with plant available ammonium, phosphorus and potassium all closely correlated with CCA Axis 2 (Figure 9.8). Similar to the first CA plot (Figure 9.6), the plant assemblage encountered on the river embankments was clearly separated from the remaining vegetation, with this area being characterised by low soil moisture and higher phosphorus availability (Figure 9.8). The source of phosphate on the embankment is unclear, as these elevated areas were not flooded with river water or sediment prior to the restoration, although river water might provide a P source via hyporheic flowpaths (see Chapter 5). Due to the reduced sample size of the CCA (108 samples), environmental data was only available for four ditch samples. As a consequence, changes in vegetation community composition in relation to environmental data were not discernible among the ditch and middle meadow sample groups.

Table 9.6: British National Vegetation Classification (NVC) communities for the middle meadow, ditch and river embankments. NVC type was assigned using the average goodness of fit value produced in TABLEFIT (Hill 1996b).

NVC Type	Mean Goodness of fit	Community (Sub-community)
Middle Meadow		
OV28a	68	*Agrostis stolonifera – Ranunculus repens* grassland (*Pol hyd – Ror syl)*
OV28	67	*Agrostis stolonifera – Ranunculus repens* grassland
MG10b	60	*Holcus lanatus – Juncus effusus* rush-pasture (*Junc inflexus)*
MG10a	59	*Holcus lanatus-Juncus effusus* rush-pasture (Typical)
MG10c	58	*Holcus lanatus – Juncus effusus* rush – pasture (*Iris pseudacor)*
Ditch		
MG10a	82	*Holcus lanatus-Juncus effusus* rush-pasture (Typical)
M23b	75	*Juncus effusus/acutiflorus – Galium palustre* rush – pasture (*Juncus effuses)*
OV28a	71	*Agrostis stolonifera – Ranunculus repens* grassland *(Pol hyd – Ror syl)*
M23	70	*Juncus effusus/acutiflorus – Galium palustre* rush-pasture
MG10c	68	*Holcus lanatus – Juncus effusus* rush – pasture (*Iris pseudacor)*
Bank		
MG1	88	*Arrhenatherum elatius coarse grassland*
MG1a	82	*Arrhenatherum elatius coarse grassland(Festuca rubra)*
W24b	80	*Rubus fruticosus – Holcus lanatus* underscrub *(Arr ela – Her sph)*
MG1b	76	*Arrhenatherum elatius coarse grassland (Urtica dioica)*
MG1c	71	*Arrhenatherum elatius coarse grassland (Filip ulmaria)*

**Goodness of fit: 80 – 100 = very good, 70 – 79 = good, 60 – 69 = fair, 50 – 59 = poor, 0-49 = very poor.*

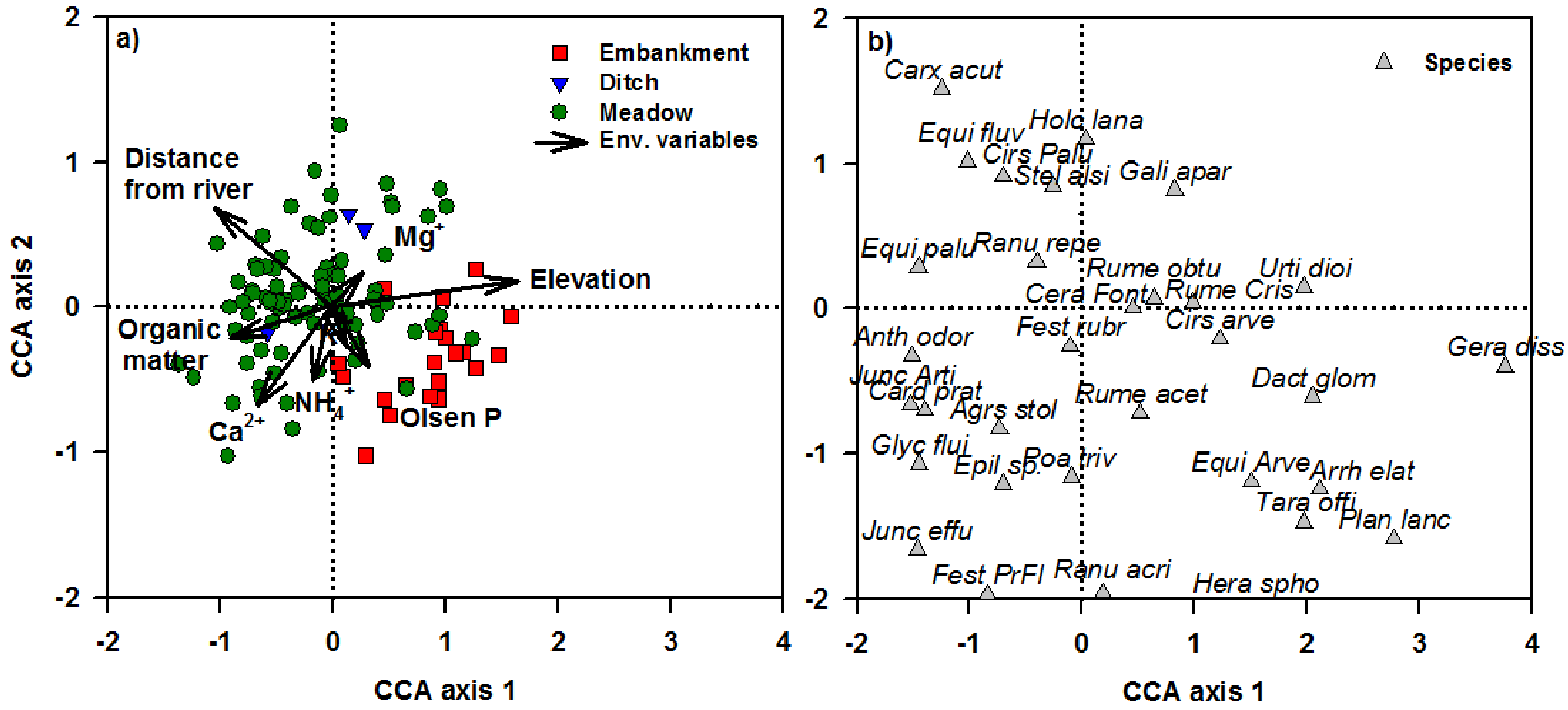


Figure 9.10: Constrained Canonical Correspondence Analysis (CCA) of species composition on environmental variables, showing (a) embankment, middle meadow and ditch sample points, and (b) the associated species. Correlations of the soil variables (elevation, distance from the river, organic matter content, calcium, ammonium, Olsen P, magnesium, and potassium concentrations) with the two main axes are shown by the arrows. In total, 67 species were analysed from 108 samples that spanned 31 transects across the meadow study site.

Table 9.7: Eigenvalues and cumulative percentage variance for each CCA axis of the vegetation and environmental data, and summary of the Monte Carlo test.

Axes	**1**	**2**	**3**	**4**	
Eigenvalues:	0.419	0.169	0.077	0.058	
Species-environment correlations:	0.794	0.631	0.558	0.448	
Cumulative percentage variance					
of species data:	9.9	13.9	15.7	17	
of species-environment relation:	50.3	70.5	79.8	86.8	
Sum of all eigenvalues					4.241
Sum of all canonical eigenvalues					0.833
Test of significance:	First axis				All axes
F-ratio	10.634				2.635
P-value	0.002				0.002

Average Ellenberg moisture values on the meadow ranged from 5 (damp) to 11 (saturated roots) (Figure 9.11a), and highlight a wet soil-water regime experienced by plants at the site. High values, in the range of 9 (wet) to 11 (saturated roots), occurred at the downstream ponded area of the meadow and along the ditch margins, whereas lower values occurred along the river embankments, typically in the range of 5 – 6 (Figure 9.11a). Ellenberg nitrogen values indicate high soil fertility (nitrogen values: 6 − 7) along the river embankments and at the base of the hillslope (Figure 9.11b). Variations in Ellenberg scores across the meadow indicate a spatial arrangement of vegetation in response to differing soil moisture and nutrient status and support the CCA results in Figure 9.10.

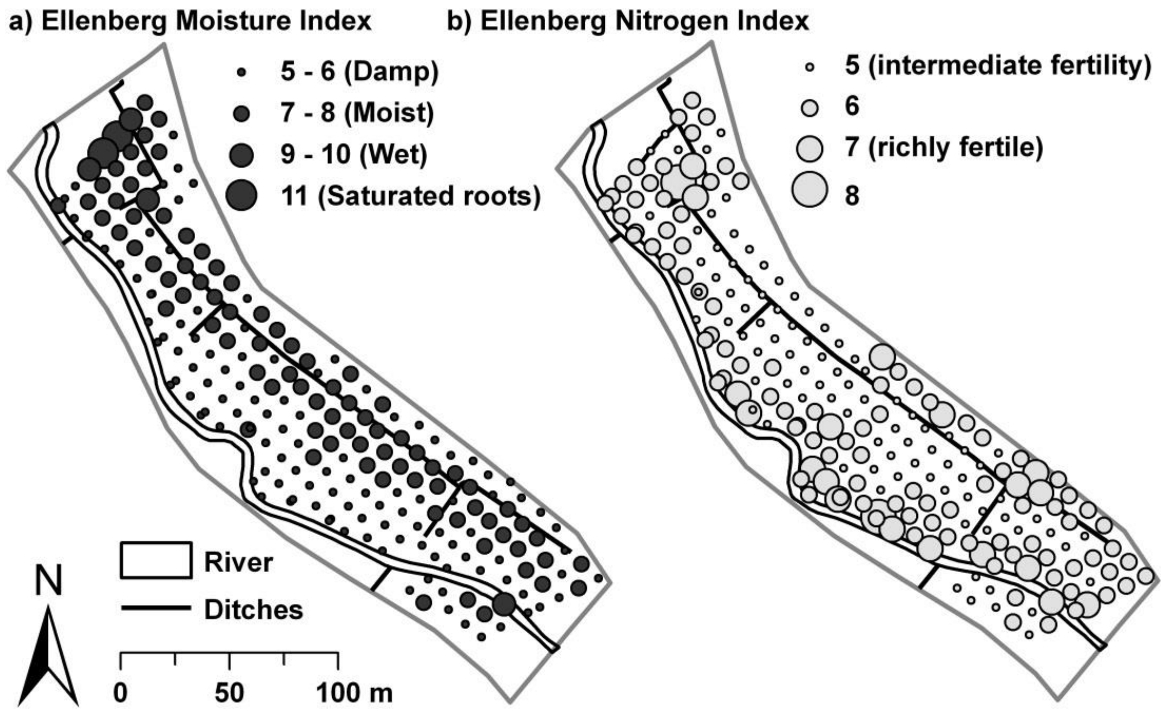


Figure 9.11: Spatial variation in Ellenberg's indicator values for plant species tolerance of moisture (a) and nitrogen (b). Mean Ellenberg's values are presented for each sample quadrat. Moisture indicator values range from 5 (damp soils) to 11 (plant rooting under water for part of the time), and nitrogen indicator values range from 5 (intermediate soil fertility) to 8 (between richly fertile and extremely rich soil).

9.1.4 Hydrological modelling outputs

To predict changes in soil aeration stresses and plant community composition following embankment-removal, groundwater levels were simulated for pre-restoration (embanked) and post-restoration (no embankment) conditions on the meadow under identical climatic conditions for a 10-year period using coupled MIKE SHE/MIKE 11 models, as described in Chapters 6 and 7. The simulated MIKE SHE groundwater levels provided a good fit to water table observations from most wells (see Figures 7.1-7.3). The mean error was typically less than ± 0.05 m, the correlation coefficient averaged 0.85 and 0.84 for the calibration and validation periods, respectively, and the Nash-Sutcliffe model efficiency coefficient ranged between 0.5 and 0.8 for most of the wells (see Table 7.1). Seasonal changes in groundwater levels for the baseline conditions were reproduced well by the models. In particular, they replicated timing and magnitude of groundwater response to high rainfall and river-flow events with great accuracy (see Figures 7.1-7.3). The models also captured the groundwater recession in response to decreasing river levels although at times simulated groundwater levels were higher than observed levels. This difference was however typically <0.1 m.

Surface flooding on the meadow was simulated in both the embanked and restored models. However, in the embanked model, flooding occurred due to elevated groundwater levels following periods of high precipitation. This flooding was mainly limited to the lower-lying areas of the meadow, while the embankments remained dry (see Figure 7.8). Under restored conditions, flooding also occurred due to overbank inundation (see Figure 7.8). Embankment removal increased the extent and depth of surface water, especially close to the river, along the ditch, and in the lower part of the meadow (see Figure 7.8).

Simulated groundwater depths below the soil surface averaged -0.31 m and -0.25 m during the growing season for the embanked and restored scenarios, respectively (Figure 9.12a-b). Differences in water table depth after embankment removal resulted from both topographic changes (lower surface elevation and increased groundwater flooding e.g. Macdonald *et al*. (2012) and hydrological changes (increased flood water storage and subsequent seepage) following inundation events. The largest increases in water table elevation (i.e. declines in depth) following restoration occurred adjacent to the river, and at the relatively low-lying downstream end of the meadow (Figures 9.12 and 9.13a). The smallest effects were seen adjacent to the un-restored section of riverbank (Figure 9.12).

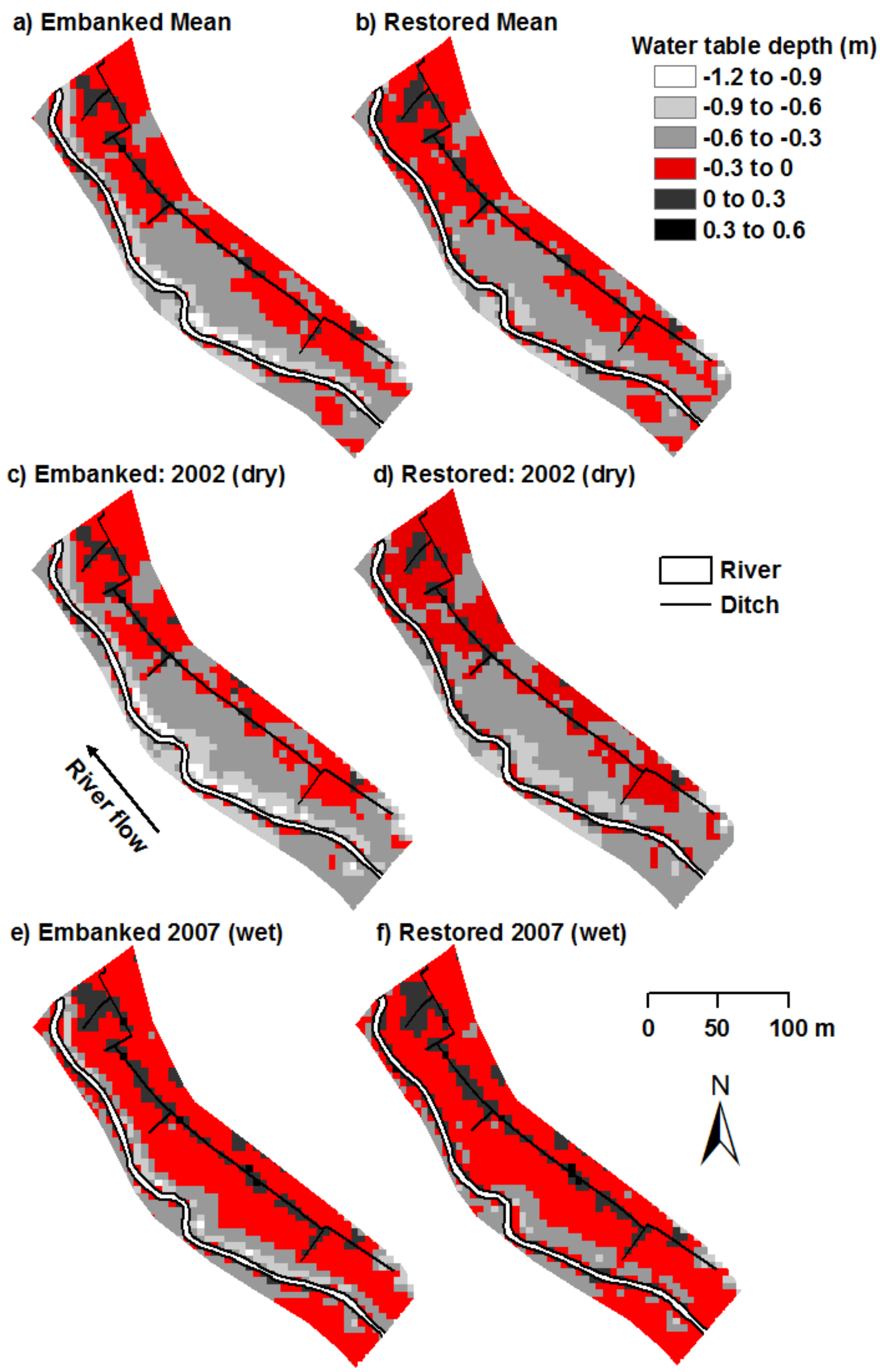


Figure 9.12: Comparison of average water table depth for the embanked and restored scenarios during the active grass growing season (March – September inclusive) (a-b) from 2001 – 2010, and during (c-d) a dry and (e-f) wet spring/summer. Positive values indicate the meadow surface is inundated. Values between -0.3 to 0 m (in red) indicate that water table depth is above the rooting zone, and plants are likely to experience aeration stress.

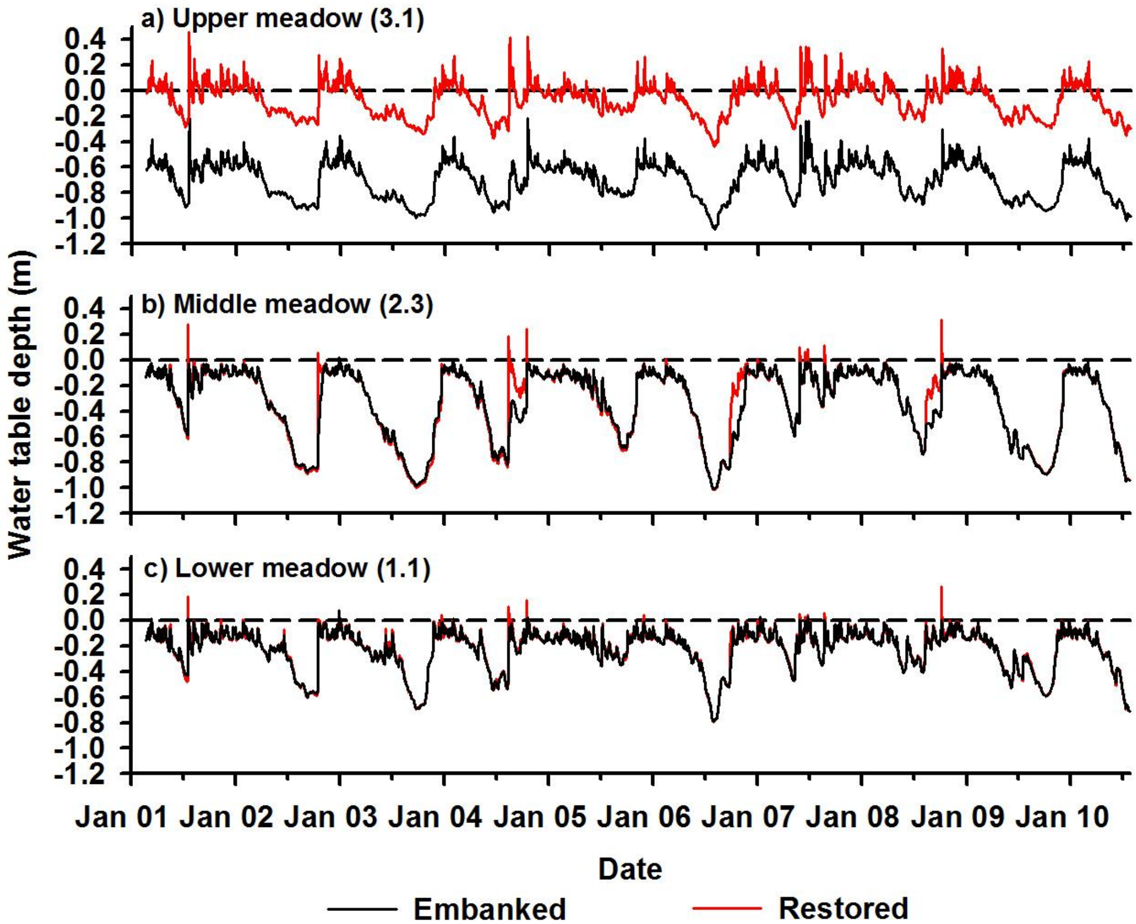


Figure 9.13: Comparison of simulated water table depth relative to the soil surface for the restored and embanked scenarios from 2001 – 2010 (a period which encompassed a range of wet and dry conditions). Water table heights are shown for three representative well locations across the floodplain at the (a) upper, (b) middle, and (c) lower well transects (see well locations in Figure 1).

During dry growing seasons with no overbank flows such as in 2002, groundwater depth averaged -0.36 m and -0.31 m below ground in the embanked and restored scenarios, respectively (Figure 9.12c-d). In the embanked scenario, groundwater depth remained below the rooting zone (0 to 0.3 m) along the embankments and across much of the upper meadow (Figure 9.12c). Under restored conditions, increases in water table elevation largely occurred adjacent to the river in the previously embanked area (Figures 9.12d and 9.13). Surface flooding was restricted in both scenarios to the lower meadow and to areas immediately adjacent to the ditch. For a wet growing season, with multiple high river-flow events and overbank flows in the restored scenario, average groundwater depths for the embanked and restored models were -0.20 m and -0.14 m, respectively (Figure 9.12e-f). Shallow groundwater depths occurred across the meadow in both scenarios, indicating a high degree of aeration stress, with the exception of the near-channel areas under embanked conditions that remained much drier.

9.1.5 Habitat suitability

Oxygen status of surface soil pores (10 cm bgs) was closely linked with water table depth (Figure 9.14). During waterlogged conditions, dissolved oxygen (DO) concentrations responded rapidly (within one day), and soil pores in the rooting zone were filled with anoxic groundwater (mean DO = 0.85 ± 0.26 µM). As the water table fell, DO concentration increased close to atmospheric saturation (mean DO = 295 ± 5 µM) typically within 3 – 6 days. Oxygen levels responded faster, typically within 1 – 2 days, if the duration of flooding was <1 week. During the growing season, vadose DO averaged 295 ± 5 µM (88 ± 2% air saturation), which was close to atmospheric levels (Figure 9.14). DO concentration of groundwater measured at 0.3 m below the ground surface followed a similar trend (see Figure 5.17), however the optode which provided these data was installed in a well and thus when groundwater levels were below the optode, measurements were of DO concentration in air (rather than DO in soil shown in Figure 9.14). Consequently, DO levels in the well responded more quickly (within 1 day) to drying conditions compared to DO levels in the soil.

Large seasonal variations in soil oxygen concentration occurred during the study period (Figure 9.14b). Surface soils were saturated from mid September – mid April resulting in anoxic soil conditions (<1 µM) during autumn, winter, and early spring (Figure 9.14b). For the majority of the active grass growing season, oxygen concentrations were typically close to atmospheric concentrations (Table 9.8). For example DO in the soil pores measured 255 μM at the peak summer temperature (19.4 °C), which is equivalent to 181,829 ppmv (Table 9.8). However five high water table events occurred during the summer that resulted in anoxic surface soils for 3 – 8 consecutive days.

The high oxygen concentrations measured during unsaturated soil conditions in the summer (Figure 9.14b) indicate that diffusion of oxygen from the soil surface exceeded the rate of consumption within the rooting zone. These results are consistent with low sum exceedance values for aeration stress (SEVas) for the same period (i.e. 2010) (see below). The ability to relate water table depth to soil aeration status validates the use of water table position as a proxy of aeration stress in wetland soils, which underpins the sum exceedance values for aeration stress (SEVas) index (Gowing *et al.* (1998). However, the delayed recovery of DO concentrations to lower water tables after saturation may lead to longer periods of aeration stress than those predicted by water table depth alone, particularly if prolonged flooding (>1 week) were to occur during the growing season.

Table 9.8: Temperature and dissolved oxygen concentration measured in soil pores during 2010. Bunsen coefficients (from Weiss 1970) and Henry's coefficients (Bunsen/1 mole of pure gas at STP) used to determine oxygen concentration in parts per million by volume.

Date	**O_2 (μM L^{-1})**	**O_2 (mg L^{-1})**	**Temp (°C)**	**Bunsen coefficient (ml L^{-1})**	**Henry coefficient (moles L^{-1} atm^{-1})**	**O_2 (ppmv)**
12/26/10	0.82	0.03	1.2	0.047	0.0021	387.63
02/25/10	0.45	0.01	5.1	0.043	0.0019	232.41
04/23/10	306	9.53	10.2	0.038	0.0017	180,618
06/16/10	287	8.91	15.0	0.034	0.0015	187,567
07/02/10	255	7.93	19.4	0.031	0.0014	181,829

Oxygen concentration of dry air = 210,000 (ppmv) (Jacob 2009).

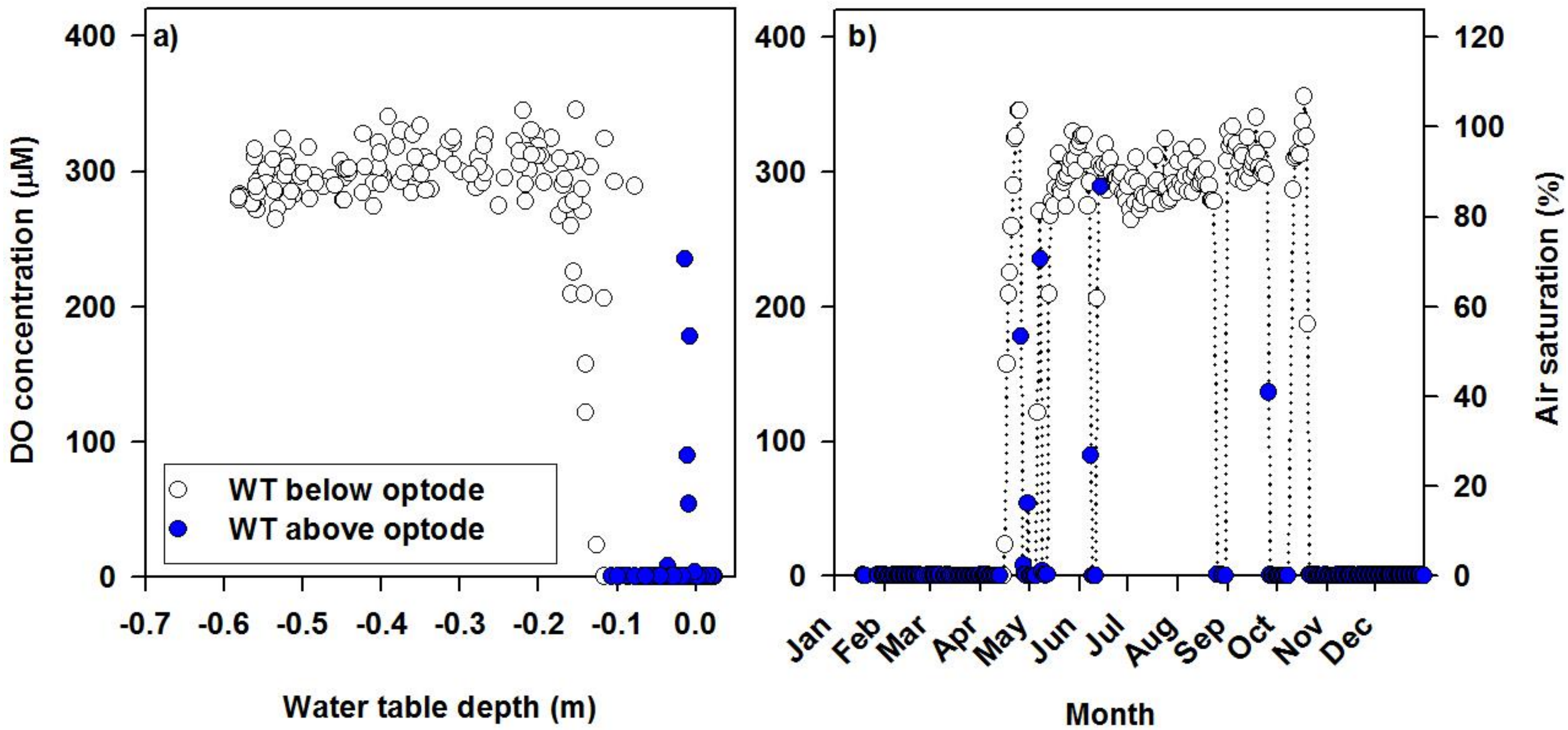


Figure 9.14: Relationship between mean daily dissolved oxygen (DO) concentration in soil and mean daily water table (WT) depth (a), and time series of DO concentration for 2010 (b). The DO optode was buried in soil 0.10 m below the ground surface.

Observed sum exceedance values for aeration stress (SEV_{as}) averaged 2.8 ± 1.0 m weeks (range: 0 – 6.9 m weeks), suggesting that plants experienced a high degree of aeration stress (Table 9.9). The highest parts of the meadow, i.e. on the embankment (Well 3.1) or on the hillslope (Wells 3.4, 3.5) had the lowest values of SEV_{as} (<0.2 m weeks), and thus plants in these areas were expected to experience little to no aeration stress (Table 9.9). SEVas for the embanked and restored modelled scenarios fitted well with the observed values (Figure 9.15). The observed data indicated that SEVas were typically high, but were lower after the restoration (in 2009 and 2010). The modelled SEVas indicate that lower water table heights in 2009 and 2010 were associated with prevailing low rainfall during this period, rather than with embankment removal (Figure 9.15). Modelled SEVas did not differ significantly, averaging 2.9 and 3.0 m weeks ($p = 0.7611$) before and after the restoration, respectively. However, a large increase in aeration stress occurred immediately next to the river, where SEVas increased from 0 m weeks before embankment removal to 6.9 m weeks ($p<0.05$), on average, after embankment removal (Figure 9.15a).

Results for the restored (no embankment) MIKE SHE model show an increase in groundwater levels during the summer for the simulation period 2001 – 2010. These are attributed to a number of inundation events that occurred in summer months after embankment removal (Figure 9.16; see also Figure 7.7). The marked increases in simulated groundwater levels along the river-margin are evident. The modelled water table results indicate that on average, a shallow rooting zone on the restored meadow is aerated during the growing season (Figure 9.16). A shallow water table depth (-0.05 to -0.3 m) is simulated near the river on the previously embanked area, whereas some deeper, but still relatively shallow water tables (0.02 to -0.5 m) occur on the meadow. The restored water table regime in some parts of the meadow satisfies the target hydrological conditions required for an MG8 species-rich floodplain meadow community (Figure 9.16b), or a MG13 inundation grassland community (Figure 9.16d). During wetter than average years (e.g. 2007), overbank inundation and surface flooding on the meadow results in groundwater levels which are above the limits believed to be tolerable for these grasslands. However, since these communities can endure occasional overbank flows (indicated by the grey areas in Figure 9.16), and wet conditions do not occur year on year or across the entire meadow, there is the potential for the sustained establishment of these diverse plant communities at the site. Restored conditions on the meadow are nonetheless too wet for an MG4 species-rich hay meadow, since the respective plant species cannot tolerate soil waterlogging at any time during the year (Figure 9.16f).

Table 9.9: Cumulative aeration stress index for plants, also referred to as sum exceedance values for aeration stress (SEVas), on Hunworth Meadow during the grass growing season from March – September inclusive (31 weeks total).

	Aeration stress index (SEV_{as}) (metre weeks)			
	2007	**2008**	**2009**	**2010**
Well 1.1	2.64	4.35[a]	1.64[b]	
Well 1.2	6.70	6.94	3.09	
Well 1.3		5.45	2.61	2.78
Well 1.4	5.86	5.78	2.51	2.32
Well 1.6	5.49	3.78	2.16	2.96
Well 2.1		2.21	0.91	
Well 2.2		3.39	1.56	
Well 2.3		3.23	1.45	
Well 2.4		4.50	2.12	
Well 3.1	0.00	0.00[c]		
Well 3.2	2.86	2.84[a]	2.49[b]	1.60
Well 3.3	5.20	5.92	2.54	4.19
Well 3.4	0.19			
Well 3.5	0.00	0.00	0.00	0.00

3 weeks of data missing[a]; 4 weeks of data missing[b]; 6 weeks of data missing[c]; blank cell = too much data missing to calculate SEV_{as}.

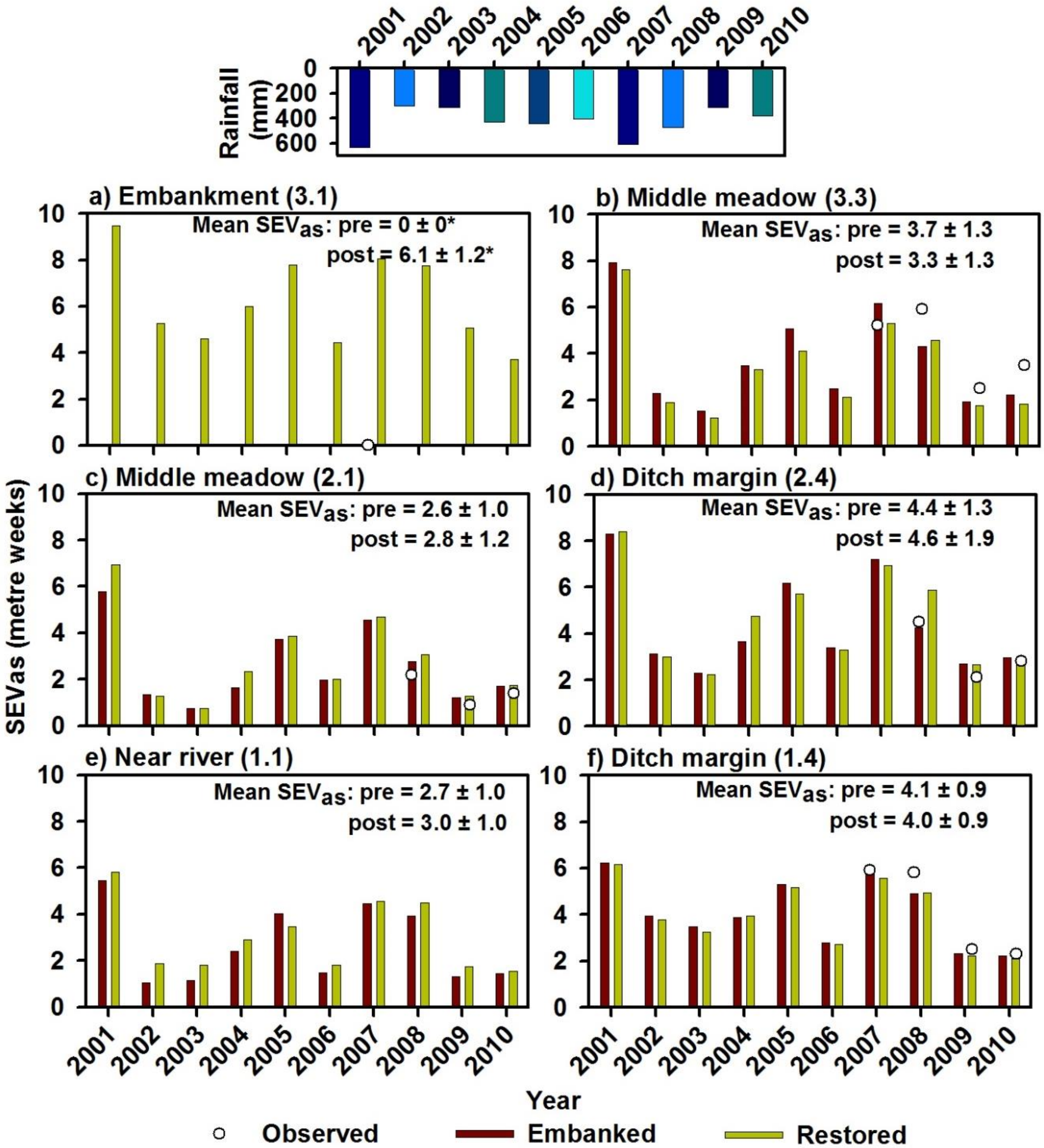


Figure 9.15: Comparison of sum exceedance values for aeration stress (SEVas) for the embanked and restored scenarios across the meadow during the active grass growing season (March – September inclusive). Modelled SEVas is validated with observed SEVas, where continuous water table depth data were available. Top panel shows total rainfall (March – September). ***Note that SEVas for the embanked scenario on the embankment at (a) 3.1 was 0 m weeks from 2001 – 2010, which is not visible in the figure.*** The river embankment (3.1) was the only location where mean SEVas's were significantly different ($p<0.05$).

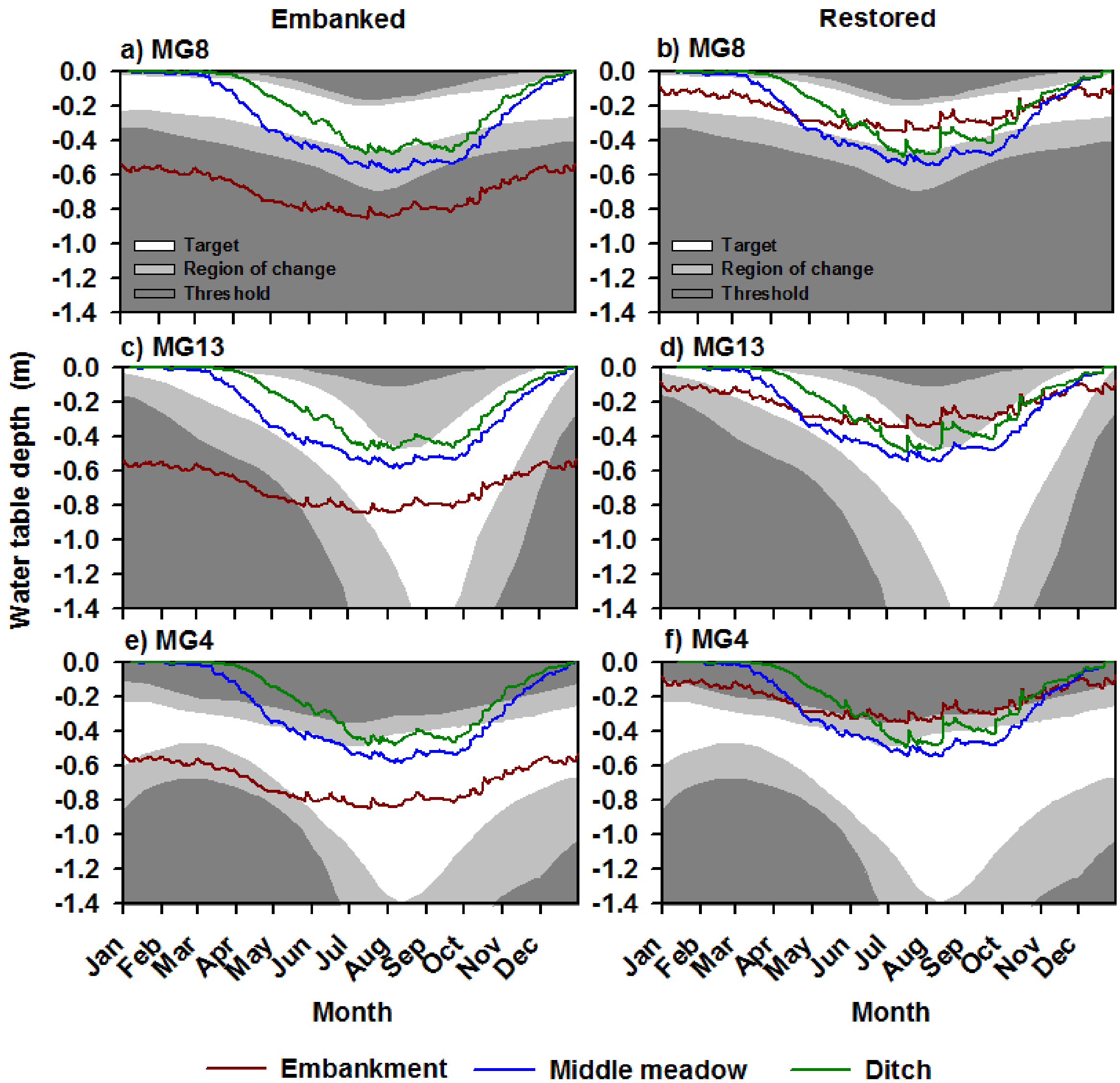


Figure 9.16: Post-restoration mean groundwater height for three representative locations across the meadow: the embankment/river margin (Well 3.1), middle meadow (Well 2.1), and ditch margin (Well 2.4) superimposed upon ecohydrological guidelines for (a-b) MG8, (c-d) MG13 and (e-f) MG4 British National Vegetation Classification (NVC) grasslands from Wheeler *et al.* (2004). The target area (white) are the conditions required for the community; conditions that fall outside of this target area regularly result in community change (light grey); the threshold areas (dark grey) indicate the more extreme wet or dry conditions that if experienced in one year only will result in community change (Wheeler *et al.* 2004).

9.2 Discussion

9.2.1 Hydrological controls on floodplain processes and plant diversity

River regulation and modification of the natural flow and flood regimes are major threats to biodiversity in floodplain ecosystems (Ward *et al*. 1999). Plant species richness, in particular, is reported to be sensitive to flow alteration (Poff and Zimmerman 2010; Kuiper *et al*. 2014). The Flood-Pulse Concept predicts that floodplain ecological functioning (e.g. riparian production and nutrient retention) is flow-dependent, governed by the lateral transport of flood water, sediments and nutrients onto the floodplain (Junk *et al*. 1989; Tockner and Stanford 2002). Recurrent overbank inundation and flood-based disturbance of an intermediate level (i.e. in terms of flooding magnitude and duration) increases environmental heterogeneity within the riparian zone (Ward *et al*. 1999). This is thought to positively affect plant community dynamics, composition, and promote maximal diversity, firstly by opening space for colonisation by less competitive plant species and thus allowing many species to coexist, and secondly by aiding seed recruitment on the floodplain (Grime 1979; Silvertown *et al*. 1999; Helfield *et al*. 2007; Auble and Scott 1998; Nilsson *et al*. 2010).

At Hunworth Meadow, overbank flows were infrequent (>10 year recurrence interval) under embanked conditions (see Sections 5.2.4 and 7.2.2). This severely impeded the exchange of water and nutrients between the river and floodplain meadow. In certain hydrogeological settings, hyporheic exchange is an important pathway for linking biological and chemical processes (i.e. nutrient uptake and cycling) that occur in rivers with their floodplains and vice versa (Stanford and Ward 1993; Hedin *et al*. 1998). However, subsurface flow rates on the meadow were only in the order of cm day^{-1} (see Section 5.2.2), and without regular overbank events, the floodplain was essentially disconnected from the river. Despite this, groundwater levels were generally close to the soil surface, and during high river flows, the rooting zone often became waterlogged due to groundwater flooding (e.g. Macdonald *et al*. 2012).

The seasonal timing of flooding and accompanying aeration stresses plays an important role in structuring wetland plant communities (Gowing and Youngs 1997; Robertson *et al*. 2001), with summer flood duration being particularly important. Joyce (1998) found that a less dynamic soil water regime tends to promote a community characterised by stress-tolerant, competitive perennials. Indeed, at Hunworth Meadow, the relatively wet and stable hydrological conditions of the embanked site supported a species- poor plant community dominated by a degraded *H. lanatus – J. effusus* (NVC: MG10 type) flood-sward.

Ordination of the botanical data suggested that soil moisture, predicted by surface elevation, was the dominant environmental influence on the spatial distribution of meadow vegetation. This is in agreement with earlier studies (e.g. Gowing *et al.* 1998; Silvertown *et al*. 1999; Castelli *et al.* 2000; Dwire *et al.* 2006). The degree of waterlogging and resultant aeration stresses in the root environment control plant functioning, productivity and survival (Armstrong *et al*. 1994; Visser *et al*. 2003; Jackson and Colmer 2005). Results from Hunworth Meadow show that wet, anoxic soil conditions persisted in the winter and intermittently in the growing season in both the embanked and restored meadow. This is likely to be a major factor controlling plant assemblages at the site. In addition to oxygen deprivation, wetland plants are affected by the accompanying changes in soil chemistry (Pezeshki and DeLaune 2012). In particular, anaerobic conditions lead to higher availability of phosphorus related to the reduction of iron-complexes and phosphorus desorption in low redox soils (Baldwin and Mitchell 2000; Zeitz and Velty 2002), and low amounts of available nitrogen for plants, firstly by limiting nitrification and secondly by promoting the removal of nitrate via denitrification (Pinay *et al*. 2007).

Soil fertility was found to be a secondary driver of plant species composition and likely has a combined affect with hydrology. Topsoils were of intermediate phosphorus fertility, probably due to historic farming practices (fertiliser additions) on the meadow. As is the case in many nutrient-rich habitats, Hunworth Meadow exhibited a high dominance of a few plant species. Biodiversity of mesotrophic plant communities is generally found to decrease in response to increased nutrient supply (Willems *et al*. 1993; Grime 2001; Loreau *et al*. 2001; Aerts *et al*. 2003). More diverse grassland communities have been found to require plant-available phosphorus concentrations within the range of 5 – 10 mg P kg^{-1} (Snow *et al*. 1997; Gowing *et al.* 2002a; Michalcová *et al.* 2011). Average topsoil concentration at Hunworth was at the upper limit of this target range, and in numerous areas of the meadow exceeded 15 mg P kg^{-1}. In contrast, nitrate concentrations measured in soil pore water and soil were typically low, likely associated with prolonged waterlogging and anoxia during the winter and high demand from soil microbes and plants. The resource balance hypothesis suggests that plant species diversity is highest when nutrient supply ratios (N:P) are balanced, i.e. following a ‘humped-back’ relation, which is thought to favour plant species coexistence (Braakhekke and Hooftman, 1999; Aerts *et al*. 2003). In contrast, the observed low plant species richness on Hunworth Meadow is likely due to the combined effects of high aeration stress due to waterlogging, high available phosphorus, and nitrogen limitation.

Flood waters and sediments are major sources of plant nutrients (P and N) to floodplains. In turn floodplains can remove river nutrients from through-flowing water via plant assimilation, denitrification and phosphorus adsorption, which can reduce nutrient loading downstream. While additional nitrogen inputs could improve diversity at the site, drier soil conditions are likely required to increase plant availability of added nitrogen (i.e. to increase mineralisation and nitrification) and limit phosphorus availability (i.e. increase P-adsorption) (Zeitz and Velty 2002). For instance, Van Oorschot *et al*. (1997) found that in fertiliser experiments on English floodplain meadows, added phosphorus resulted in a higher P-pool, whereas a limited affect was noted for N-pools (likely due to losses via denitrification). Hence, allochthonous phosphorus supplied by floods following reconnection of the river and floodplain may reduce species richness on the meadow (e.g. Beltman *et al*. 2007; Michalcová *et al*. 2011). In this case, further management of the restored site may be required; for instance via hay-cutting that can effectively balance additional inputs of flood-derived nutrients to the floodplain (Linusson *et al*. 1998; Wheeler *et al*. 2004).

9.2.2 Water regime of the restored floodplain meadow

Removal of the river embankments has naturalised the river form, and restored Hunworth Meadow to an active floodplain. Hydrological/hydraulic modelling reported in Chapter 7 indicates that under restored conditions, large floods were fairly infrequent (flood return period of ~3 years) and typically short-lived (<1 day). Indeed, observations of overbank flows onto the meadow confirm that flood waters recede quickly, normally within 1 day. This is likely to minimise the negative impacts of widespread inundation (i.e. due to prolonged waterlogging and aeration stress) on plant diversity (e.g. Baldwin *et al*. 2001), particularly if inundation occurs in the winter. In contrast, following restoration smaller-scale flooding at the river-floodplain margin (i.e. within 5 m of the river) is simulated as occurring regularly (return period = 0.2 years) and persisting for longer (2 – 3 days) (see Sections 7.2.2 and 7.3.3). This results in a more hydrologically connected river-floodplain ecotone. Simulated water table elevation was higher after the embankment removal due to lower surface elevation and increased flood water storage on the meadow. However, results in Chapter 7 also showed periods of drier conditions on the meadow, particularly in the ditch region and in topographic depressions, as a result of improved drainage along the river that allows surface water to drain more freely following floods, and so reducing the effects of large floods. These findings may reduce the concerns of land managers who do not wish to lose productive grazing land to flood waters, but are interested in the ecosystem benefits of river restoration.

Low soil aeration under water-logged conditions was demonstrated using a novel approach based on oxygen optodes, which provided direct measurements of vadose DO concentrations that can used to identify when plants will begin to experience oxygen stress. High vadose DO concentrations were consistent with low aeration stress values during the same period, confirming that air-filled porosity and soil oxygen status were directly linked at the site. Oxygen optodes have been successfully used in a range of environments, such as in the open ocean and in the air above the water surface (e.g. Bittig and Körtzinger 2015), in benthic chambers (e.g. Sommer *et al*. 2008; Almroth-Rosell *et al*. 2012), and in river sediments (Tengberg 2004), but to our knowledge, this study is the first to use DO optodes to measure continuous oxygen concentration in floodplain soils. Redox potential is commonly used to characterise the oxygen status of wetland soils (e.g. Barber *et al*. 2004), however the accuracy of redox measurements are constrained by the difficulty of calibration and the time taken to obtain stable readings (Strawn *et al*. 2015). Further work is needed to examine the change in soil oxygen-moisture relationship with depth, during overbank inundation, and the relationship between oxygen concentration and air-filled porosity in different soil types. However the method employed in this study offers huge potential for assessing the impacts of waterlogging and the accompanying aeration stresses on wetland plants.

9.2.3 Predicting plant community composition change

Modelled water table results for the 10-year simulation period permitted the quantification of the hydrological effects of embankment removal for a range of probable climate and river-flow conditions, including extreme high and low flow years. Although both pre- and post-restoration hydrological data were monitored in this study, and were essential for the initial site assessment as well as model calibration and validation, these data alone were insufficient to clearly determine water-table response to embankment removal (see Chapter 5). This was because climate and river conditions were so different between the pre-restoration (i.e. dry) and post-restoration (i.e. wet) observational periods that it was difficult tease apart hydrological effects resulting from inter-annual climate variability from those due to the restoration. As a result, hydrological/hydraulic model results from simulations of pre- and post-restoration conditions with identical climate inputs significantly improved our ability to predict the potential response of plant communities to the altered water regime.

Reinstatement of overbank flows did not substantially affect root saturation and aeration stress on the meadow, largely because prior to embankment removal, the meadow was already very wet and groundwater flooding was typical during high river flows. However, an exception to this is along the river embankments, where simulated sum exceedance values for aeration stress (SEVas) increased dramatically from 0 m weeks (i.e. dry grassland) to approximately 7 m weeks (i.e. fen) ($p<0.05$). It can be predicted, therefore, that this region is likely to undergo the greatest plant community change, from species intolerant of flooding (e.g. *A. elatius* and *D. glomerata*) to plants that tolerate waterlogged soils throughout most of the growing season (e.g. *Phalaris arundinacea, Ranunculus acris* and *Cardamine pratensis*) which were already present in wetter regions of the meadow and could colonise along the restored river banks.

Species-rich MG4 meadows require lower levels of waterlogging than observed across the entire meadow. Interestingly, prior to the restoration the hydrological regime of the river embankments was suited to dry grassland communities. Evidently, other factors (e.g. high fertility, seed dispersal) limited colonisation of MG4 grassland in this part of the meadow. Michalcová *et al*. (2011) reported that increases in plant diversity for wet mesotrophic grasslands in the UK are most likely to occur when SEVas values are between 0 – 1 m weeks. Gowing *et al*. (2002b) presented average SEVas values of ca. 2.5 m weeks (upper limit: ca. 4.5 m weeks) as the favoured water-regime of species-rich MG8 *Cynosurus cristatus* – *Caltha palustris* grazing marsh communities, and average SEVas values of ca. 3.2 m weeks (upper limit: ca. 6 m weeks) for MG13 *Agrostis stolonifera* – *Alopecurus geniculatus* inundation grassland communities. The restored water table regime may be suitable for MG8 or MG13 grassland communities, both of which are of importance for wading birds (Wheeler *et al*. 2004). Indeed, some particularly characteristic species of MG8 (e.g. *C. cristatus, Cirsium palustre, Carex hirta*) and MG13 (e.g. *A. stolonifera, Rumex crispus, Alopecurus geniculatus*) communities were already present on the meadow. However SEVas at Hunworth are near the upper limits for these more diverse plant assemblages, and although there is a gradient of hydrological change across the meadow that could support a range of communities, results suggest that in order to achieve the greatest increases in plant species diversity at the site, drier conditions would be required during parts of the growing season.

Changes in the quality of flood water on the meadow may act to reduce aeration stress during waterlogged conditions. Removal of the river embankments has altered the flooding regime on Hunworth Meadow to include overbank flows of oxygen-rich river water (mean: 10.8 mg O_2 L^{-1}) (see Section 5.2.6). In contrast, prior to the restoration, the meadow was dominated by oxygen-depleted groundwater (mean: 0.6 mg O_2 L^{-1}). Mommer *et al*. (2004) demonstrated that passive diffusion of DO from the water column into submerged terrestrial plants, i.e. *Rumex palustris*, is an important source of oxygen. Thus, submergence with oxygen-rich river water may reduce shortages of oxygen and lessen stress on plant functioning, and thus significantly affect which plant assemblages can survive the respective conditions.

A botanical study of wet meadow sites along the River Glaven conducted by Wotherspoon (2008) identified a number of meadows of higher botanical value in terms of species richness (mean species richness: 15 – 20 species per m^2) than Hunworth Meadow (mean species richness: 8 species per m^2). These local species pools may provide a source of hydrochorically deposited propagules during overbank flow events (e.g. Merritt *et al*. 2010), providing that river embankments along the river upstream do not limit seed dispersal. The river embankments at Hunworth were a substantial barrier to propagule dispersal. This was likely to be an important factor limiting colonisation and plant diversity on the meadow, particularly along the river embankments, given the low levels of waterlogging prior to the reconnection that could have supported more species-rich assemblages.

9.2.4 Management implications

Conservation efforts for UK wetland environments are particularly focused on MG4, MG5, and MG8 species-rich wet meadows (Manchester *et al*. 1999). Based on hydrological and soil physicochemical conditions of the restored floodplain meadow, three main management options at Hunworth Meadow are proposed to maximise the botanical value of the sward following the embankment removal: 1) restoration and maintenance of the ditch network to lessen waterlogging during the growing season. This would promote more favourable conditions for species-rich MG8 communities; 2) reinstatement of traditional grazing and hay cutting regimes to reduce nutrient loading from flood-deposited sediments, and reduce the dominance of competitive grasses and tall species such as rushes (*Juncus* spp.) (e.g. Proulx and Mazumder 1998; Crofts and Jefferson 1999; Woodcock *et al.* 2006); and 3) reintroduction of species (e.g. hay spreading from local meadow sources) to supplement the local species pool, which is likely to be seed-poor due to previous river regulation and habitat fragmentation at the site (e.g. Walker *et al*. 2004).

A number of other restoration and habitat improvement schemes are being implemented in the Glaven catchment to reinstate interconnectivity among aquatic habits, with focus on landscape-wide improvements to biodiversity (Sayer 2014). Arable reversion in the upper reaches of the River Glaven, which includes a management regime of cutting and baling, has successfully reinstated wet meadow and enhanced flora from one of arable weeds (*Raphanus raphanistrum* and *Equisetum arvense*) to grassland dominated *H. lanatus*, *Poa trivialis*, and *Potentilla anserina*. In addition, the restored meadow now supports abundant orchid populations (particularly *Dactylorhiza fuchsii* and also *Dactylorhiza praetermissa*) (Sayer 2014). These species are of conservation interest, and are present locally in other meadows along the River Glaven (Wotherspoon 2009), especially those managed with hay cut and rake regime. Reinstatement of overbank flooding and improved connectivity with meadows along the river corridor could result in spontaneous regeneration of these species, although more targeted management that involves the introduction techniques discussed above could be implemented to encourage and hasten colonisation at the site.

Chapter 10: Conclusions and recommendations for future research

10.1 Conclusions

In a natural or near-natural state river-floodplain ecosystems are characterised by dynamic hydrological connections and complex environmental gradients (Triska *et al*. 1989; Ward *et al*. 2001; Pringle 2003). Flood-pulsed disturbance regimes maintain and promote spatio-temporal heterogeneity, which in turn is responsible for high levels of biological diversity on floodplains (Poff *et al*. 1997; Grevilliot *et al*. 1998). This thesis has demonstrated that the removal of river embankments at Hunworth Meadow has enhanced river-floodplain interactions via overbank flows thereby restoring more natural, flood-pulse dominated hydrological conditions. Using data from a comprehensive pre- and post-restoration hydrological monitoring campaign, hydro-chemical monitoring, botanical surveys, hydrological/hydraulic modelling, and the simulation of pre- and post-restoration floodplain groundwater levels, the primary objectives of this thesis were as follows: to determine baseline pre-restoration hydrological and chemical conditions, and botanical composition on the floodplain and assess the measured hydrological response to river restoration; to model the effects of embankment removal on key components of river-floodplain hydrology (water table elevation, frequency and extent of floodplain inundation, flood-peak attenuation) under a range of expected river-flow conditions; and to predict plant community changes to altered soil moisture and chemistry resulting from river-floodplain reconnection.

Results from hydrological and chemical monitoring at Hunworth Meadow presented in Chapter 5 indicate that prior to the embankment removal the river and floodplain were linked primarily via slow subsurface flowpaths, resulting in limited hyporheic extent (Figure 10.1). Consequently, the soil water regime on the floodplain was controlled by anoxic, nutrient-poor groundwater. Removal of the embankments reduced the channel capacity by an average of 60%. This re-established opportunities for regular overbank flow allowing for bidirectional surface-subsurface flow, and exchange of water, sediment, nutrients and matter between the river and its floodplain. Accordingly, the floodplain is likely to shift to a more disturbance-based environment controlled by oxygen-rich river water as well as anoxic floodplain groundwater (Figure 10.1). Restoration of river-floodplain connectivity is likely to cause more frequent, short duration inundation of the floodplain, resulting in a more favourable soil water regime that may enhance floodplain plant diversity. This will be associated with a change in the quality of flooding, i.e. from long-term, stagnated inundation with oxygen-poor groundwater prior to the restoration, to short-term pulses in oxygen-rich river water following restoration. These changes should in turn reduce aeration stress during submergence, and create flood conditions that are much more easily tolerated by a variety of wet meadow plant species (e.g. Mommer *et al*. 2004). Furthermore, regular overbank flows and the supply of nutrient-rich river water to the floodplain during the summer months when microbial and plant activity is high will favour conditions for removal of river nutrients by floodplain sediments, particularly at the river-floodplain interface where a strong redox gradient is present.

Despite a fairly lengthy observational period (1.5 years) after the restoration, only one overbank event was observed. As this flood was relatively small (resulting from a river flood with a mean daily discharge of 1.36 $m^3 s^{-1}$), and was below the bankfull threshold for widespread inundation (1.67 $m^3 s^{-1}$), flooding was limited to the river-margin, and thus changes in floodplain hydrology (other than the frequency of overbank flooding) could not be determined based on the observational data alone (see Section 5.3.1). Interannual climate variability complicated direct comparisons of pre- and post-restoration hydrological conditions. For instance, the two very wet summers prior to the restoration, and the significantly drier summers after the restoration, rendered it difficult to clearly determine the effects of embankment removal on the floodplain soil-water regime -

with the possible exception of the near-river environment. This highlights the importance of long-term monitoring (before and after restoration works) that is required to fully evaluate the impacts of river restoration works on floodplain hydro-dynamics. The need for such long-term monitoring should be recongised by regulatory bodies interested in using river restoration to meet legislative requirements for river water quality and ecology.

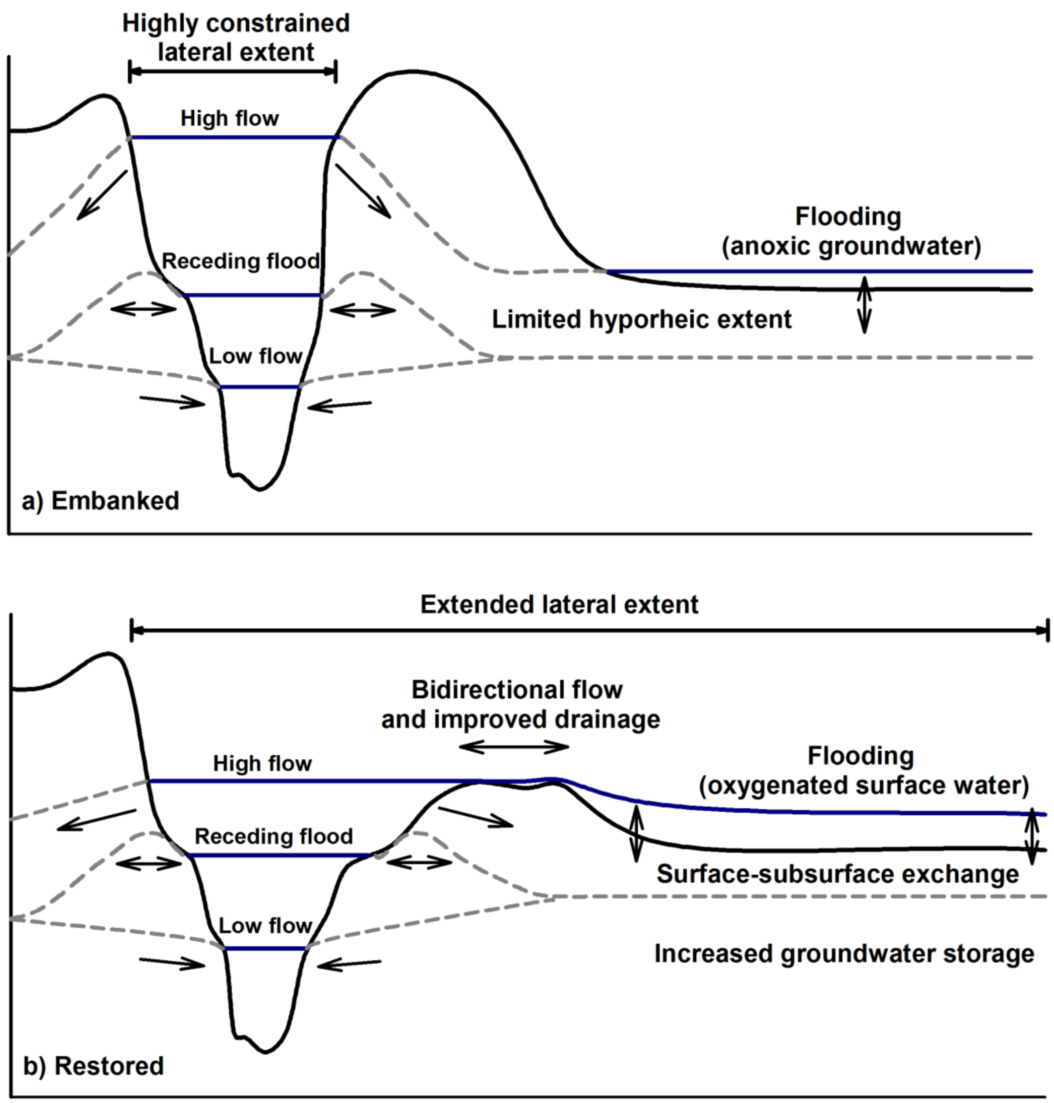


Figure 10.1: A schematic representation of the hydrological regime of Hunworth Meadow in the (a) embanked and (b) restored scenarios during three river flow conditions.

In the absence of long-term datasets, hydrological/hydraulic monitoring can be used to assess the response to restoration under a range of hydrological conditions (i.e. for high flow and low flow events). Coupled MIKE SHE-MIKE 11 hydrological-hydraulic models detailed in Chapters 6 and 7 were successfully employed to simulate the significant and complex hydrological impacts of river restoration. Model results indicate that the removal of the river embankments provided the physical geomorphological conditions to allow regular overbank flows that resulted in widespread floodplain inundation at high river flows (>1.7 $m^3 s^{-1}$), and frequent localised flooding along the river edge during smaller events (>0.6 $m^3 s^{-1}$), which is seen as one of the main aims of river restoration projects (e.g. Acreman *et al*. 2003, Helfield *et al*. 2007, Hammersmark *et al*. 2008). Subsequently, groundwater levels were higher, particularly along the river-floodplain margin where connectivity with the river was substantial resulting in increases between 0.4 – 0.6 m (see Section 7.2.3), and subsurface storage was greater (see Section 7.2.5). The restoration provided space for water to spill out onto the adjacent floodplain. This had a moderate effect on flood-peak attenuation, resulting in a maximum reduction in peak river flows of 6 – 24%, along the length of restored reach of the River Glaven (see Section 7.2.5). In addition, removal of the river embankments improved free drainage to the river following periods of inundation. The restoration increased river-floodplain hydrological connectivity, creating a more disturbance-based riparian zone that extended laterally from the river towards the edge of the floodplain (see Figure 10.1).

The process-based methods used in this thesis to quantify the hydrological impacts of river restoration provide powerful and practical scientific and management tools to predict changes in habitat suitability for target biological communities, in this instance due to changes in water table depth and duration of floodplain inundation resulting from embankment removal. They can be used to inform policy decisions and conservation management strategies. This study is based on rarely available pre- and post-restoration hydrological data. However, the hydrological/hydraulic modelling approaches employed for Hunworth Meadow can equally be used at other sites when either pre- or post-restoration data are not available (e.g. Hammersmark *et al*. 2010; see also Section 10.2), or when inter-annual climate variability and relatively short observational periods prevent direct pre- and post-restoration comparisons. This approach may also be applied in the planning stage of restoration projects to assess the feasibility of restoration at a site, whether optimal water regime requirements are likely to be achieved, or to assist the design of the restoration works.

In order to restore wet meadow biodiversity, it is important to consider the joint effects of hydrology and soil nutrients on plant community composition. Hydrology was identified as primary driver of plant community composition, while soil fertility was also important (see Sections 9.1.3 and 9.2.1). Unique continuous measurements of vadose DO concentrations using oxygen optodes indicated a strong coupling between water table depth and DO concentrations in the rooting zone and showed that the relationship between oxygen status and water table fluctuations in floodplain soils can assist in the understanding of spatial and temporal distribution patterns of lowland wet meadow vegetation. Reinstatement of overbank flows did not substantially affect the degree of aeration stress on the meadow because of pre-existing, very wet conditions. An exception to this was seen along the river embankments where sum exceedance values for aeration stress increased from 0 m weeks (dry grassland) to 7 m weeks (fen) due to a lowering of the surface elevation relative to the water table height.

There is a gradient of hydrological conditions and post-restoration change across the meadow that could support a range of communities. Although conservation focus is on more species-rich meadows (MG5, MG4, MG8), there is conservation value in restoring other British National Vegetation Classification grassland types, particularly in terms of improving habitat for other biological species (e.g. wading birds, dragonflies, amphibians). Embankment removal alongside the River Glaven created a more natural flood-pulsed hydrological regime, characterised by regular, short-duration, oxygen-rich inundation of the floodplain meadow that will likely result in improvements to river-floodplain ecosystem functioning (e.g. enhanced habitat connectivity, patch heterogeneity) and improvements in ecosystem services (plant biodiversity, water quality). This has important implications for the rehabilitation, maintenance and resilience of floodplain plant communities at the site and elsewhere.

Embankment removal as a measure for improving ecosystem functioning in degraded wet meadows is effective. However, reduction of nutrient levels and waterlogging are also important for restoration efforts to succeed in promoting species-rich plant communities (Critchley *et al.* 2002; Michalcová *et al.* 2011), particularly where floodwaters are enriched in nutrients (see Section 5.2.6). In addition, propagule availability of target species, i.e. due to dispersal limitations along the river or paucity in the seed-bank, is likely to be a limiting factor in the recovery of biodiversity (Bischoff *et al*. 2009; Nilsson *et al*. 2010), and thus seed transfer could be an effective additional restoration measure. River restoration should focus on reinstating self-regulating, dynamic river processes important for creating morphological heterogeneity that can benefit riparian communities. Ideally, this would be initiated at the landscape-scale (e.g. Tockner *et al*. 2000, Ward *et al*. 2001), such that benefits in ecosystems services can be transmitted along the river corridor.

10.2 Further research directions

10.2.1 Limitations of the study

The soil and water chemistry analyses conducted in this study for the determination of baseline biogeochemical conditions were comprehensive. However it would clearly have been beneficial to collect further measurements after the restoration so that changes in soil fertility due to the restoration could be assessed, particularly in terms of developing a management strategy for the meadow after the restoration. This was not undertaken largely due to time and laboratory constraints associated with such an extensive chemical sampling campaign. However, the pre-restoration data provide an essential baseline reference against which further chemical analyses at the site can be compared, and changes due to the restoration can be assessed. In addition, the potential supply of phosphorus from riverine sediments, an important driver of grassland productivity (Vitousek 2015), to the reconnected meadow was not measured, and could be followed-up in further studies.

For the development of the MIKE SHE/MIKE 11 hydrological/hydraulic models of Hunworth Meadow, it would have been useful to have conducted more detailed hydrogeological surveys of the meadow using, for example, Ground Penetrating Radar or three-dimensional electrical resistivity tomography (see House *et al*. 2016a). Although the geology of the region was available from British Geological Survey maps, more detailed representation of the subsurface geology could have assisted the model calibration process. Targeted analysis of hydraulic conductivities of identified geological layers could then have been conducted, and the use of distributed values for hydraulic conductivity in MIKE SHE would have been an option.

10.2.2 Further restoration and habitat enhancement

In addition to the embankment removal, a second-phase in-stream restoration project was conducted on the same stretch of the River Glaven in August 2010, one year after the embankment removal and after the main period of fieldwork reported in this thesis (Figure 10.2). This involved the creation of a new, narrower and more geomorphically diverse, meandering river channel, with the aim of improving in-stream habitat and further increasing hydrological connectivity between the river and floodplain (Figures 10.2 and 10.3). Excavation of the drainage ditch was carried out in conjunction with the re-meandering works in order to promote drainage and a more dynamic soil water regime.

Re-meandering of the river channel has had a substantial effect on in-stream geomorphology, most notably on river sinuosity, which increased by 16% from 1.1 to 1.3. The downstream river length was also increased by 61 m (Table 10.1). A further 58 m of channel length was added in the form of backwaters, which were created from the remnant channel (Figure 10.3). Variation in river depth increased slightly along the reach, with average river depths of 0.76 ± 0.15 m and 0.76 ± 0.23 m (measured as height from the bank top to river thalweg; Figure 10.4) before and after the re-meandering, respectively. Continued monitoring of hydrological conditions (based on the methods detailed in Chapter 4) on the floodplain alongside regular vegetation surveys (using the methods described in Chapter 8) will be used to evaluate changes in hydrological regime following the two differing stages of restoration, and measure the long-term effects on floodplain hydrodynamics and plant community composition.

Figure 10.2: Photographs of the River Glaven at Hunworth Meadow showing (a) the river embankments in January 2009 prior to embankment removal, (b) the completed Stage I restoration work with embankments removed in March 2009, and (c) completed Stage II restoration work with embankments removed and remeandered river channel in December 2010.

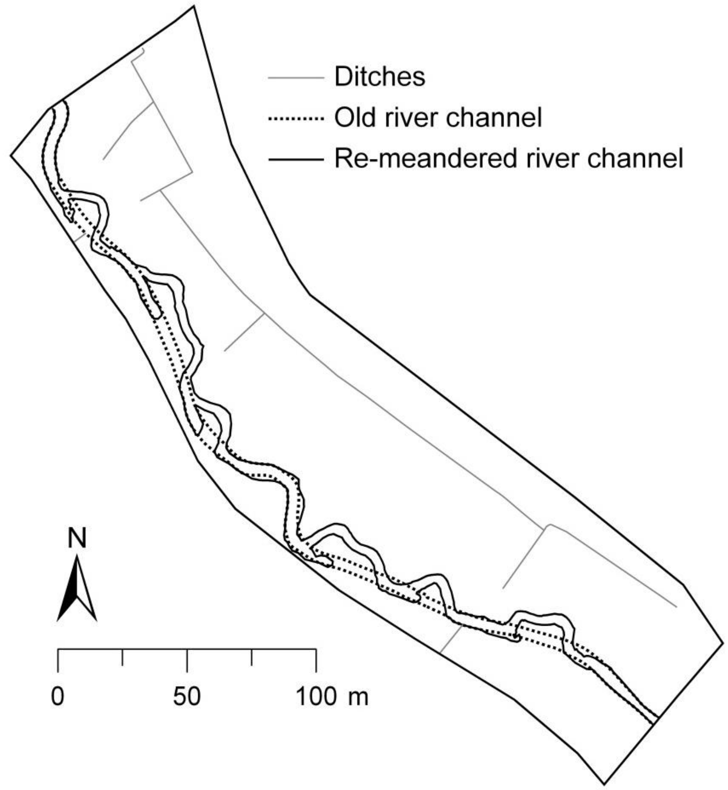


Figure 10.3: Re-meandered river channel at Hunworth Meadow.

Table 10.1: Comparison of river length and sinuosity before and after remeandering of the river channel.

	River length (m)	Sinuosity (ratio of channel to valley length)
Pre-meander	370	1.1
Post-meander	430	1.3

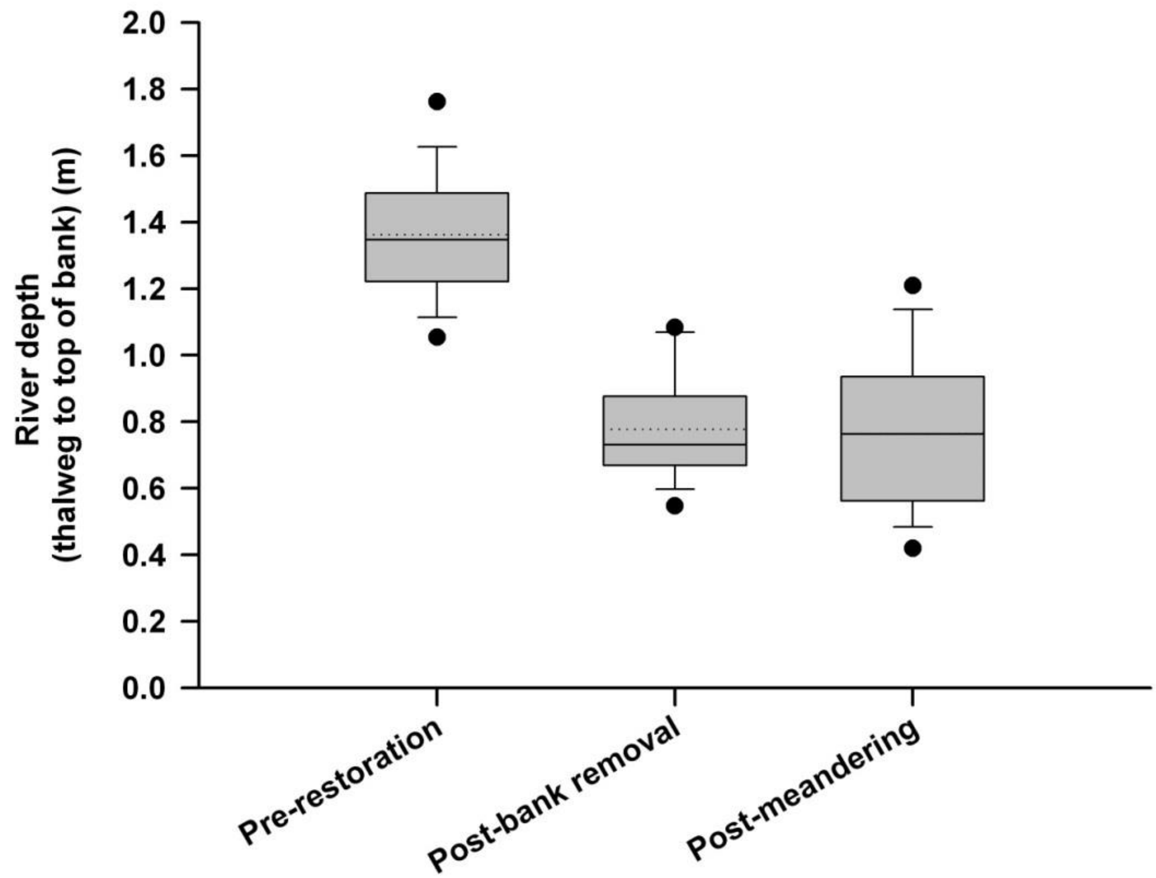


Figure 10.4: Comparison of river depth before and after the two stages of restoration (embankment removal and re-meandering). The black centre line, dotted centre line and box extent denote the median, mean, and 25th and 75th percentiles, respectively. Whiskers indicate the 10th and 90th percentiles, and the circles signify outliers outside of the 10th and 90th percentiles.

The hydrological-hydraulic models developed in this thesis could be applied to the second-phase restoration. This would require the creation of a further revised digital elevation model of the meadow for the MIKE SHE model and river cross-sections representing the new channel configurations for the MIKE 11 model. River discharge data at Hunworth for the upstream MIKE 11 boundary condition are not usable from August 2010 after the second-phase in-stream restoration. Extensive geomorphological changes to the river channel (narrowing, depth diversification, and re-meandering) have inadvertently caused water to back-up over the weir which has affected the rating curve at the gauging station, and resulted in inaccurate discharge calculations (see Section 6.5). Thus, a model of the re-meandered river channel could be run for the same period of analysis presented in this thesis, i.e. 2001 – 2010, to determine the relative effects of embankment removal and re-meandering on floodplain hydrological processes.

In addition to the restoration projects at Hunworth Meadow, a number of other restoration activities are ongoing in the Glaven catchment involving a mosaic of aquatic habitats of varying sizes (e.g. rivers, streams, ponds and ditches), with the aim of repairing autonomous river processes and enhancing landscape heterogeneity and biodiversity. Based on the hydrological changes quantified at Hunworth Meadow due to embankment removal (i.e. increased river-floodplain connectivity, reinstatement of flood-pulsed hydrology, increased subsurface storage, and flood-peak attenuation) removal of embankments is suggested for other reaches of the river in order to have a cumulative impact on river health and associated ecosystem services (e.g. flood protection, biodiversity, water quality) within the catchment.

10.2.2 Climate impact studies

Hydrological models are also a powerful tool when used to link hydrological impact assessments and climate-change studies (Thompson *et al*. 2009; Thompson *et al*. 2013; House *et al*. 2016b). The majority of climate change scenarios for the UK predict that the frequency and magnitude of floods will increase due to increased winter precipitation (Wilby *et al*. 2008; Thompson 2012). Increases in air temperature will also likely alter evapotranspiration rates and groundwater recharge, which is likely to affect wetland species that are sensitive to changes in hydrological regime (e.g. Gowing *et al*. 1998; Araya *et al*. 2011). Hydrological models can be combined with climate scenarios by using the latter to perturb meteorological inputs to the former. This can produce potential scenarios of climate-change effects on hydrological processes at a watershed scale (e.g. Ficklin *et al*. 2009; van Roosmalen *et al*. 2009; Clilverd *et al*. 2011). Several case studies of climate change impacts on wetland hydrology have used MIKE SHE (Thompson *et al*. 2009; Singh *et al*. 2010; Stoll *et al*. 2011; Thompson 2012; Vansteenkiste *et al*. 2013). A variety of carbon dioxide emissions scenarios are typically used (e.g. low, medium, high) from relevant GCMs. The predicted changes in hydrological regime (e.g. water table depth) under different climate scenarios can then be used to assess the ecological responses to climate change and guide future management for important flora and fauna conservation species accordingly (e.g. House *et al.* 2016b).

For example, a climate impacts study conducted by Thompson *et al.* (2009) using MIKE SHE/MIKE 11 and UK Climate Impacts Programme (UKCIP) projections for the 2050s simulated lower water table depths and reduced magnitude and duration of surface water inundation within the Elmley Marshes, Southeast England. It was suggested that these hydrological changes would lead to a loss of specialist wetland plants adapted to the current high water tables. Similarly, House *et al*. (2016b) used a MIKE SHE/MIKE 11 model of a riparian wetland on a tributary of the River Thames to demonstrate spatially varying hydrological impacts due to climate change that would have implications for both wetland flora and fauna. These studies point to the potential for further analysis using the hydrological/hydraulic models of the Hunworth Meadow to assess the capacity of river restoration to proof wetlands from the hydrological impacts of climate change. Since hydrological conditions at the site are strongly controlled by river flow, such a study would require the development of an additional model component that could simulate baseline discharge at the Hunworth gauging station. This model would in turn be forced with peturbed meteorological inputs (precipitation and PET) to assess future river flow changes for application to the MIKE SHE/MIKE 11 models. The latter would themselves be forced with the corresponding scenario precipitation and PET (see House *et al.* 2016b). In addition, other factors that may be impacted by climate changes and indirectly affect floodplain hydrology include vegetation (e.g. type, cover, and length of the growing season) and soil (e.g. water holding capacity) properties (see Holman 2006), which could be altered for specific climate scenarios

In conclusion, this thesis contributes a process-based, quantitative, hydroecological framework to river restoration. The modelling approach described can be used to assess the complex hydrological effects of restoration, and predict the possible response in floodplain biota. The restoration at Hunworth Meadow highlights the importance of dynamic hydrological linkages and interactions between river-floodplain ecosystems, and the potential for embankment removal to restore form and function to floodplain meadows. This study has important implications for the planning and management of river restoration projects that aim to enhance floodwater storage, floodplain plant biodiversity and biogeochemical cycling of nutrients, and can serve as context for understanding and predicting the hydroecological response to restoration in other ecosystems.

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