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UNDERSTANDING THE GROUNDWATER SYSTEM OF A HEAVILY DRAINED COASTAL CATCHMENT AND THE IMPLICATIONS FOR SALINITY MANAGEMENT

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ABSTRACT

The Thurne catchment in north-east Norfolk, UK, is an extremely important part of the Broads National Park, an internationally important wetland environment. Extensive engineered land drainage of the marshes of this low-lying coastal catchment over the past two centuries has led to land subsidence and the need for drainage pumps to control water levels sufficiently below sea-level to maintain agricultural productivity. Consequently, seawater from the North Sea has intruded into the underlying Pleistocene Crag (sand) aquifer and brackish groundwater enters into land drainage channels, thereby raising their salinity. Powerful pumps discharge these brackish drainage waters into a Special Area of Conservation (SAC) and RAMSAR site, leading to adverse ecological impacts on salt-sensitive species.

Chloride concentrations within drainage channels throughout the network have been found to significantly vary, with several influential factors affecting channel salinity such as proximity to the sea and connectivity to the underlying aquifer. A thorough understanding of the surface-water/groundwater system and a subsequent quantification of the various processes has been necessary for the development for the drain/aquifer interactions and a numerical groundwater model. These models are used to estimate the long-term distribution of the salinity within the drainage system under current conditions. The model credibility is justified by comparable aquifer-drain water balance, a comparable coast water inflow/ total groundwater ratio and the particle tracking from the coastal reaches trace to previously-measured saline-vulnerable locations. The numerical groundwater model has demonstrated that the average daily inflow of saline groundwater into the Crag aquifer of the Thurne catchment is 3,081 m^3/day, of which the Hempstead Marshes main drain is one of the main conduits for saline inflow into the Brograve system, which discharges directly into the SAC.

Various changes to the engineering design or operation of the drainage system have been proposed to minimise the saline inflow to the SAC, but the implementation of any proposals must be considered in conjunction with the current dynamics of the system. Three separate management or engineering remedial measures have been modelled: (i) raising the water levels in the drains of the Hempstead Marshes in the north east of the catchment (ii) lining the main drain of the Hempstead Marshes with low permeability material, and (iii) The construction of a new coastal open ditch drain which is intended to 'intercept' the saline intrusion and prevent ingress into inland drains of the Brograve system. The results suggest that raising the water levels in the Hempstead Marshes will reduce the saline inflow into the Brograve sub-catchment substantially, and decrease the overall saline inflow into the Thurne catchment from 3081 m^3/day to 2822 m^3/day). The lining of the main drain in Hempstead produces a less than 10% decrease in saline inflow into the catchment from 3,081 m^3/day to 2,958 m^3/day. The simulated coastal interceptor drain could *in theory* through maintaining a low groundwater head near the coast, prevent the inflow of saline groundwater into the Brograve system. However, such a drain would increase the saline inflow across the coastal boundary by around six times (from 3,081 m^3/day to 19,750 m^3/day), remove large quantities of fresh groundwater from the Pleistocene Crag aquifer and lead to high energy and pumping costs.

The research has shown that there are partial solutions to reducing the saline inflow into the drainage systems in this lowland coastal catchment. However, any intended alterations must first consider other potential impacts, such as changes to flood risk, land management restrictions or hydrodynamic effects on the receiving watercourse through changed discharge volumes.

DEDICATION

To my darling wife Elaine,

To God be the glory, great things he hath done

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1 INTRODUCTION

1.1 Background

The Broads National Park in East Anglia is one of Britain's most important wetland areas for ecology, recreation (boating, bird watching and angling) and agriculture. The Broads or shallow lakes are generally freshwater systems, with the exception of the Broads of the Upper Thurne catchment that are brackish. The Upper Thurne catchment is situated in northeast Norfolk with the villages of Happisburgh immediately to the north, Winterton in the southeast and Thurne in the south.

The unusual brackish nature of the Thurne River system in north-east Norfolk has been reported since the beginning of the twentieth century when the source of the salinity was thought to be "probably due to salt springs" within Hickling Broad and Horsey Mere (Gurney, 1904). Due to the flat, low-lying nature of the area and the close proximity to the North Sea, the region is susceptible to flooding. Several major sea breaches are recorded as far back as 1287, when the sea swept inland as far as Potter Heigham and Ingham, drowning 108 people in Hickling alone. The floodwaters reached a level of 30 cm higher than the altar of Hickling Priory (Sainty et al. 1938). Other floods caused by breaches of the (sand) dunes occurred in 1608, 1665, 1791 and 1792, 1805, 1897, 1938 and 1953 (Holman, 1994).

In 1938 and 1953 the Upper Thurne catchment experienced more severe flooding as the sea breached the dunes, with the latter event leaving the people of Horsey village isolated for four consecutive months (Mr. J.J. Buxton, *pers.comm*.). This continual flooding was thought to be an alternative source of the salinity within the Upper Thurne system (Barry and Jermy, 1953) but this was shown not to be the case as the annual winter rainfall gradually diluted the saline groundwater. In the light of the work completed by Pallis (1911), it is generally recognised that the source of the salinity is by direct aquifer communication between the sea and the groundwater of the Upper Thurne catchment.

The high water table marshland that is found within the catchment has historically been used for arable farming with the aid of a complex interconnected network of open ditch drains to lower the water table. This land drainage has major adverse effects such as significant land subsidence, due to rapid initial shrinkage resulting from the removal of the large amounts of water, which subsequently causes a compression of the peat under its own weight and an increase in its bulk density. Under such aerobic conditions, decomposition (biochemical oxidation) becomes the dominant process significantly affecting the profile above the water table (Hutchinson, 1980). The overall effect is that the peat and in particular the open ditch drains that run throughout it are now several metres below sea level.

The drains that are cut throughout the marshes are often cut to offer assistance with runoff from the neighbouring elevated land units and are therefore are cut along the thinnest parts of the alluvium and directly into the underlying Pleistocene Crag. The source of salinity identified by Pallis (1911) as being the crag (sand) aquifer beneath the marshes is in direct communication with specific drains that are below sea level, thereby encouraging saline inflow into specific surface water locations.

The network of drains is subdivided into sub-catchments in which the drains are connected to a main drain that leads to a specific electrical discharge pump. The electrical drainage pumps are strategically located within the coastal marshes to remove water from the drains and transfer it into a nearby broad such as Hickling Broad, Horsey Mere or Martham Broad, which are in turn connected to the River Thurne. In some of the sub-catchments, the electrical discharge pump is located so that the outfall is directly into the river.

The largest area covered by a single pumping station within the catchment is the Brograve sub-catchment located to the immediate north of the channel that leads to the broad known as Horsey Mere. Horsey Mere lies within an area that has certain species of wildlife that are protected by law. The quality of water being received by this channel is contaminated with Iron (iii) Hydroxide and Iron (iii) Oxide commonly known as *pyritic ochre.* In anthropogenically-drained Ferruginous soils near low-lying coastal regions, the resulting encrustation of aquatic vegetation by Iron (iii) Hydroxide and Iron (iii) Oxide (*Ochre*) is significantly detrimental to local ecological systems (Harding and Smith, 2002). Effectively, the sulphate-rich seawater intrudes via the underlying crag aquifer and reacts with the Iron (ii) sulphide within the acidic soil to produce the orange-brown Ferric hydroxides and oxides. It is widely accepted that a reduction in the saline intrusion would reduce the amount of ochre being produced and thereby bring about a level of ecological stability. However, the view taken by English Nature and Professor Brian Moss (*pers. comm.*) is that the salinity is more ecologically significant than the ochre.

There is increasing interest in changing the engineering, operation and management of the drainage systems within the catchment away from a focus on supporting intensive agriculture towards fulfilling multiple functions for extensive agriculture, biodiversity and fluvial and coastal flood defence. The suggested changes are intended to reduce the amount of salt and ochre discharged by the drainage pumps into the river and Broad network, which should enhance the water quality of the Broads and the River Thurne.

The ultimate desire of the interested parties, such as the Broad Authority and English Nature, is to have an improvement in the water quality within the Broads and a progressive move towards more clear waters (i.e. low in nutrients, low in suspended solids). The clear waters would increase the assemblage of diverse aquatic plant species and thus increase the stability of the ecological environment. Consequently, the ecosystem would be more resilient against any natural changes or human intervention.

One of the proposed solutions (Harding and Smith, 2002) to reduce the amount of brackish and ochre-rich water entering the Horsey Mere is the re-siting of the Brograve electrical discharge pump. The hydrological and ecological ramifications of re-siting the Brograve pumping station are not yet fully known. The Brograve outfall is pumped directly into the channel that leads to the Horsey Mere and as such reduces the effect of the tidal backflow of the Upper Thurne (Prof. Sue White *pers. comm.*). Other proposed measures include raising water levels within drains or lining ditches with low permeability materials. The proposed measured will have impacts, which are not well understood. Therefore, it is necessary to identify and subsequently quantify the significant factors that affect the drainage network in terms of saline inflow into the surface water system via the connectivity of specific drains, as this will assist with the future management of the Upper Thurne catchment. The development of a numerical simulation model is therefore necessary to quantitatively understand the various groundwater and drain–aquifer interactions that may be occurring within the catchment and be able to predict the possible outcomes of various re-engineering proposals.

1.2 Aim and Objectives

The aim of this research project is to predict the long-term effects on the salt load discharged at specific electrical pumps under proposed changes to the current engineering, operation and management of the land drainage systems of the Upper Thurne catchment. In accomplishing this aim, major objectives must be undertaken:

- To develop quantified conceptual models of the groundwater behaviour and the groundwater/surface water interactions within the catchment.
- To develop and refine a numerical groundwater/surface water simulation model of the flow patterns against historical data.
- To use the model to simulate the movement of the groundwater throughout the catchment under current and proposed conditions.
- Analyse model results to assess the impacts upon pump discharge and coastal inflow.

1.3 Research Area

The research area covers the groundwater catchment of the River Thurne in north east Norfolk (Figure 1.1), an area of approximately 110 square kilometres centred around 1^{o} 35'E and 52^{o} 45'N. The catchment area forms the northeastern part of Broadland and contains several important Broads including Hickling Broad, Horsey Mere and Martham Broad. Significantly, the catchment area is of low-lying terrain to the north of the River Thurne with land subsidence lowering the marshland to below sea level, and elevated land (above 20 mODN) to the south of the River. The towns and villages through out the catchment exist on the higher land as the lower land is historically susceptible to coastal flooding. The drains that are depicted within the Figure 1.1 are cut into the relatively impermeable marshland.

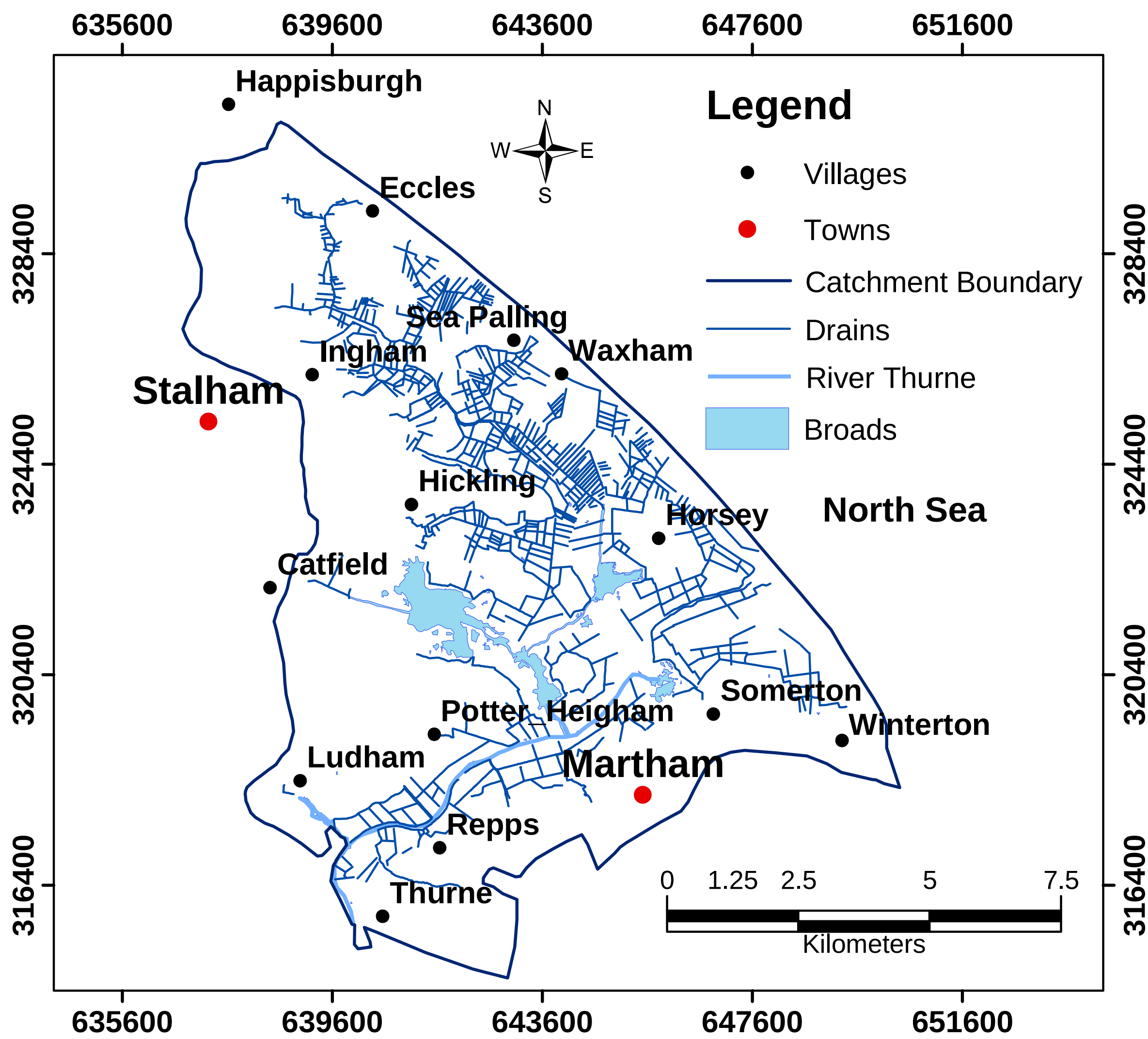

Figure 1.1 Locations and surface water drainage map of the River Thurne catchment

1.4 Structure of the Thesis

This thesis is divided into three parts. Part A, comprising of Chapters 1 to 3, brings to focus the level of importance of the catchment from a national to international perspective. Chapter 2 presents a collection of the methods and approaches used within the thesis. Chapter 3 describes the conceptualisation of the catchment from various hydrogeological and hydrological perspectives and introduces some important concepts that are used in later chapters to understand the complexity of the groundwater/surface water behaviour

Part B, Chapters 4 to 7, introduces and describes certain aspects of theoretical groundwater behaviour and describes the modelling process from theory to actual implementation and sensitivity analyses. Chapter 4 introduces the background to the numerical modelling of groundwater. Chapter 5 describes the development of computational models and the estimation of input parameters. Chapter 6 describes the implementation of a preliminary model with a discussion on model outputs. Chapter 7 describes the modelling refinement procedure and the various sensitivity analyses used to access the relationship between model inputs and model outputs.

Part C Chapters 8 to 10, describes the implementation of the proposed management changes within the catchment from a modelling and an impact perspective. Chapter 8 describes the actual management changes from a quantitative perspective. Chapter 9 presents a discussion on the important and valuable details that can be extracted from the entire body of research and in particular, the ramifications this may have on managing low-lying coastal aquifers in temperate climates. Chapter 10 provides conclusions, recommendations and possible further areas of research in the light of this project.

1.5 Literature review

Throughout the world coastal aquifers constitute an important source for water, as many of these catchments are also heavily urbanized (Bear & Verruijt, 1987),(Post, 2005) and given that the most common pollutant in fresh groundwater is saline groundwater (Todd & Mays, 2005) the need for an in-depth understanding of fresh water/saline water interaction is therefore crucial. Badon Ghyben (1888) and Herzberg (1901) independently investigated this interaction and found that salt water occurs underground not at sea level but at a depth of about forty times the height of fresh water above sea level. This relationship, known as the Ghyben-Herzberg relationship, or analytical solutions derived from it (Rushton, 2003) were the only means available to estimate quantitatively the position of an assumed fresh water-saltwater interface until the development of the computer that enabled the solving of complicated sets of equations for variable density groundwater flow. The relationship is often used as an initial estimate in approximating the depth of saline groundwater (Guswa and LeBlanc, 1985; LeBlanc et al. 1986; Masterson 2004, Masterson and Garabedian, 2007) although the extraction of the fresh water complicates the distribution of the saline water (Misut et al., 2004).

Typically, the main method of extracting fresh water from a coastal aquifer involves the sinking of a vertical borehole (Walter and Whelan, 2004), which has the effect of encouraging to a shallower depth more saline water and altering the estimated distribution of the coastal waters within the aquifer. The existence of an extensive drainage system is another mechanism whereby the inflow of saline water is encouraged into a coastal aquifer system (Genereux and Guardiario, 1998; Zechner and Frielingsdorf, 2004; Giambastiani, 2007). In these cases, the shallow horizontal drains are used to control the water table within the less permeable covering deposits and the drain-aquifer interactions that subsequently occur are often varied throughout the catchment. The locations where the drains are in direct contact with the underlying aquifer can be identified by the high concentrations of saline water.

The sinking of vertical boreholes and shallow horizontal drains therefore represent two different type of surface water/groundwater interactions that occur within a coastal aquifer system. The conceptualisation of the drain-aquifer interactions is an important component in a heavily drained coastal aquifer system, such as the Upper Thurne, as this gives valuable insight into the understanding of the behaviour of the aquifer (Bredehoeft, 2007). There are three approaches to quantifying the drain-aquifer interactions for gaining drains- analytical solutions, field investigations and numerical modelling.

Analytical solutions, such as the three that are fully described by Rushton (2006), can be used to obtain order of magnitude estimates of the flow from the aquifer to the drain. The first solution described by Rushton (2006) is derived from the method of additional seepage resistances (Streltsova, 1974). Due to a channel extending into an aquifer, an additional seepage resistance can be calculated to represent the deformed flow path into the channel and the resulting head losses. These head losses can be superimposed onto the head distribution. A second approach, proposed by Herbert (1970), is that when the river channel is small compared to the saturated thickness of the aquifer, the flow from the aquifer to the river perimeter is therefore approximately radial. The third approach is based on analytical solutions for water loss from trapezoidal channels through an aquifer to an underlying high permeability zone, which are reported by Harr (1962).

Field investigations methods such as hydrochemical analysis of drain waters (Heathcote and Lloyd, 1984; Malcolm and Soulsby, 2000; Capaccioni et al., 2004), or the measurement of the contribution of specific drains (canals) using Acoustic Velocity Meters placed within the aquifer nearby (Genereux and Slater, 1999), allow for the direct measurement of quantifiable variables. These quantifiable variables such as Electrical Conductivity or the rates of water exchange with canals are then used to identify the saline distribution within the surface water network. These investigations can also used with numerical models of the drain configuration in order to quantify the drain-aquifer interactions.

Another important interaction that occurs within a coastal aquifer is the outflow of groundwater at the coastal interface known as Submarine Groundwater Discharge (SGD). The accurate measurement of SGD remains very difficult (Burnett et al., 2001) and although several methods of estimating these discharges are currently in use by both oceanographers and hydrogeologists, the volume of discharge estimated is often dependent upon the chosen method (Mulligan and Charette, 2005). These methods include Seepage meters, Salinity/Tidal Budget and a Water Budget (Giblin & Gaines, 1990), radioactive tracers (Cable et al., 1996), Darcy's Law and a Soil Moisture Budget (Oberdorfer et al., 1990), and Numerical Modelling (Rasmussen, 1998). Corbett et al. (2001) use three categories to generalize the commonly used methods for measuring rates of submarine ground water discharge: (1) calculations using Darcy's law, (2) direct measurements with seepage meters, and (3) studies using natural or artificial tracers. Currently none of these methods are available for use within the Upper Thurne, and so therefore the measurement of the SGD is not possible. However, the identification of locations where SGD occurs must be verifiable within any groundwater model developed as the loss of fresh groundwater from the catchment will affect the total groundwater balance.

One example of a coastal catchment with an extensive drainage system is the Biscayne Aquifer of southeast Florida where management of drain water levels meets flood control, water supply, and environmental objectives. Observations lasting nearly 20 years from 1940 to 1960 (Kahout 1960) indicate that the salt front in the Biscayne aquifer is dynamically stabilised seaward as much as 8 miles from the position computed by the Ghyben-Herzberg principle. The reason for this discrepancy is that the salt water in the aquifer is not static but flows in a cycle from the floor of the sea into the zone of diffusion and back out to the sea (Kahout 1964). This cycle acts to lessen the extent to which the salt water occupies the aquifer. Further investigations show that the position of the zone of diffusion is dependent upon the amount of fresh water recharge entering the aquifer. Representation of such a zone of diffusion, by a groundwater density distribution, within a numerical groundwater model is complicated by the sensitivity of the saline inflow process in the hydrologic system to the initial conditions within the model. The initial density or chloride concentration distribution should be known accurately, especially in three-dimensional density-dependent groundwater modelling (Oude Essink, 2001).

An alternative approach used by Zheng (1990) and Zheng and Bennett (2002) has been to ignore the density differences of the groundwater and instead implement a single density groundwater regime with a post-processor that estimates the flow direction.

2 METHODOLOGIES USED IN DEVELOPING A NUMERICAL GROUNDWATER MODEL OF THE CRAG AQUIFER SYSTEM

2.1 Introduction

This chapter opens with a brief introductory description of the hydrogeological background to the current study, which includes the geology of the main aquifer, its covering deposits and the historical investigations of the Upper Thurne catchment. These details give the context to the need for additional information and the justification for the various methodologies implemented in the development of the numerical groundwater model. The key stages in the development of the numerical groundwater model for the Thurne catchment are presented and analysed. The identification and presentation of the various conceptual themes that describe the behaviour of both the surface water and groundwater regimes is the first task to be completed. The quantification of the groundwater and surface water contributions to the drainage system using a hydrograph separation technique is developed. This technique, in conjunction with the quantification of the drain-aquifer interactions using drain coefficients for gaining drains, assists with the development of a numerical groundwater model. The approach of representing variable density groundwater flow within a heavily drained aquifer using a single density groundwater model with particle tracking is introduced.

2.2 The hydrogeological framework of the catchment

2.2.1 An introduction to the Pleistocene Crag aquifer

The Pleistocene Crag aquifer comprises of sand and gravels, and is underlain by the impermeable Eocene London clay. The crag increases in thickness eastwards, being shallowest in the west where the basal material is found at -30 ODN (Figure 3.1) and slopes gently to a depth of approximately -65 ODN in the east at the coast. The body of the crag is varied in texture with a 'clayish layer' found at around -25m ODN (Downing, 1959) in the west which extends laterally to the coastline in the east. The major superficial materials overlying the Crag are the Norwich Brickearth, peat and estuarine alluvium (clay), which all vary both in thickness and in hydraulic conductivity, influencing the hydrology and hydrogeology of the catchment.

2.2.2 An introduction to the superficial deposits

The Norwich Brickearth is the major superficial deposit found at outcrop within the catchment, with extensive areas in the northwest of the catchment (Figure 3.2), to the south of the River Thurne and in patches of isolated higher land in the east. The texture of the Norwich Brickearth is that of sandy clay to sandy loam, thus only offering a partial barrier to infiltration to the underlying aquifer. In the southwestern area of the catchment, the crag aquifer is covered by highly permeable sandy soils allowing this region to be classified as exposed crag (Figure 3.2).

The alluvial deposits covering the aquifer within the catchment comprise of peat (Section 3.2.3) in the northern marshland from Hempstead Village to the Brograve Pump and the estuarine alluvium (clay) in the southern marshlands from the Brograve pump to the mouth of the River Thurne in the southwest and Winterton in the southeast. The thickness of the peat is varied, as it is less than a metre at the margins with the Norwich Brickearth and deepest beneath the line of the main drain from Lessingham Village to the Brograve pump (Figure 3.2). The estuarine alluvium has a similar varied profile with thicknesses of up to 11.5 m beneath the River Thurne, the thickness reducing to less than 0.5m at the margins such as at Potter Heigham to the north of the River or at Thurne village to the south. However, in the Horsey–Martham–Somerton region, the thickness is a uniform 4m and in either case the peat or the clay is heavily drained suggesting reduced vertical flow to the aquifer.

2.3 Hydrogeological Investigations of the Upper Thurne catchment

2.3.1 Previous studies

Pallis (1911) first investigated the origins of the saline waters within the catchment, and suggested that saline water enters the surface water system via the underlying aquifer. Subsequently, she concluded that the saline water within the aquifer at and below the low water elevation of the North Sea has concentrations of chloride similar to seawater. Other investigators such as Driscoll (1984) added weight to this theory by suggesting that deepening of the drains within the West Somerton Level coincided with general increases in the salinities in the drains. Lambert et al (1960) and Watson (1981) both suggested that lake leakage from Martham and Hickling Broads also increased the aquifer salinity. Other sources of brackish water entering the aquifer include the possible leakage from the River Thurne, which now stands several metres above the marsh levels, after years of drainage that has reduced the land elevation to several metres below sea level. No direct measurements of the saline distribution within the aquifer has ever been recorded owing to the lack of borehole data in much of the catchment, although surface electrical resistivity soundings (a measure of saltwater content) were found by Holman (1994) to have very low readings below what was interpreted as a fresh groundwater lens.

2.3.2 The current available information

The available information about the movement of groundwater and surface water within the catchment is limited. The Upper Thurne catchment is subdivided into sub-catchments that are each individually drained by electrical discharge pumps. Holman (1994) estimated the percentage of seawater for the individual catchments by measuring the salinities of drains entering the mouths of the electrical discharge pumps. These percentage estimates are used to approximate the saline loads discharged from each of the individual sub-catchments. The individual sub-catchment volumetric flow rates are estimated using electricity consumption readings, the rated capacity and an electrical conversion factor for the individual pumps. The electrical consumption is easily deduced from consecutive weekly readings and the manufacturer gives the rated capacity for the individual pumps. The estimation of the dynamics of the groundwater and surface water therefore rely heavily upon an accurate estimation of the electrical conversion factor for each of the individual pumps. Electrical conversion factors are estimates that are used to determine the volume of water discharge by the electrical pumps for each unit of electricity used. The estimation conversion factors was first carried out by Watson (1981) using a conversion factor calculated from engineering data. However, the values deduced by Watson (1981) may be erroneous as calculations from an aquifer test completed by the LWRC (1992a) predicted discharge volumes far different to those estimated using Watson's values. In a research project undertaken from April 1991 to March 1993, Holman (1994) produced revised conversion factors and took various direct and indirect measurements of the behaviour of both the surface and groundwater. These measurements are used in conjunction with recently acquired field data to assist with the parameterisation of the groundwater model within the current study.

The five datasets collected by Holman (1994) are briefly reviewed.

1. *Weekly electrical usage of the discharge pumps*: these measurements require an electrical conversion factor to convert the values into the estimated volume of water discharged. The electrical conversion factors listed for the Brograve pump vary from 142 m^3kWhr^{-1} to 31 m^3kWhr^{-1}. This creates a window of uncertainty of 215% to 40% when compared to the value of 75 m^3kWhr^{-1} quoted by Holman (1994). The final values listed by Holman (1994) of all the electrical conversion factors for the individual pumps are used to estimate the catchment groundwater balance.
2. *Salinities of the water in the drains near the electrical pumps*: these measurements were taken at the same time as the electricity readings and assist with estimation of the saline water percentage discharged from the individual pumps. It is necessary to have an estimate of the total amount of coastal water entering the groundwater system as this aids the parameterisation of the surface water/groundwater interaction occurring along the length of the coast.
3. *Electrical conductivity readings of water from various open ditch drains*: these measurements assist with the identification of the drain locations within the catchment where the drains are encouraging the groundwater inflow of saline water.
4. *Weekly water levels within observation wells*: In the study of the Upper Thurne catchment, Holman (1994) measured water levels within 41 observation wells for the two-year period ending April 1993. These weekly water levels within the observation wells will be used assist with the understanding of the groundwater head distribution within the aquifer. It is not known how the actual structure of the observation well influences the readings taken as substantial horizontal flows may occur within the more permeable deposit such as the Norwich Brickearth thus attenuating the water level fluctuations. Local conditions such as proximity of the wells to drains that are in good contact with the underlying crag or the thickness and structure of the overlying deposits may also affect the fluctuation of the measured water level.
5. *Electrical resistivity soundings of the aquifer*: duplicate measurements were made by Holman (1994) using a Rhometer and a Terrameter who conducted about 120 electrical resistivity soundings in the study area in order to map quantitatively the depth and extent of the seawater within the aquifer. These readings were conducted at ground elevation and the investigator’s conclusions were that the quality of the data was poorest at depth and near to the coast thus reducing the consistency of the readings. The combination of the very conductive groundwater and the high porosity of the Crag matrix resulted in very low

resistances, which the instruments either had difficulty in measuring (Rhometer) or in measuring consistently (Terrameter). Once a significant part of the current path intersected the more saline groundwater, both meters appeared incapable of measuring the resistance, which made quantitative depth interpretation difficult. These reading were therefore **not** used as they were found to make a quantitative depth interpretation difficult

2.3.3 The required additional information needed to develop numerical models of the groundwater behaviour

In order to determine the additional fieldwork required to support the development of a numerical groundwater model within the current study there must be an understanding of certain conceptual themes. Once these themes have been identified and understood the quantification process can be undertaken and numerical models can be developed. The three conceptual themes identified are surface water flows along the drains (SWF), drain-aquifer interaction (DAI), and the movement of the groundwater throughout the crag aquifer (DRGF). The first conceptual theme concerns the surface water movement and as this research is concerned with groundwater and the influences of drains upon groundwater movement, it is not addressed. The drain-aquifer interaction (DAI) requires information concerning the various types of drain configurations based on thickness and the hydraulic conductivity of the underlying deposit. The last theme (DRGF) requires a knowledge of the structure of the crag aquifer, a selection of representative hydraulic conductivities of the various deposits and a quantified relationship for the influence that the drains have upon the groundwater flow regime.

2.4 The development of a hydrograph separation technique

It is necessary to identify the groundwater components of the drainage pump discharges. The hydrograph of the discharge volumes from the electrical pumps is plotted and the period of time in which the surface water (runoff) contribution is estimated to be negligible is identified. This critical period within the summer months is user-determined and it is found that using different length of times for this representative period influences the groundwater-to-drain flow equation (Section 6.6.2). The principal assumption is that the head difference between the aquifer and the drain is proportional to the quantity of groundwater entering into the drain. The purpose of a hydrograph separation technique is to estimate the relative fractions of groundwater to surface water contributions within the drain-aquifer hydrological system.

2.5 The quantification of drain-aquifer interactions for gaining reaches using drain coefficients

Drain–aquifer interactions (similar to stream-aquifer interactions) are normally included in regional groundwater models using drain coefficients (equivalent to streambed conductances) because the dimensions of the drain are usually significantly smaller than the mesh spacing of the regional model. Initial studies of drain–aquifer interactions such as that of Prickett and Lonnquist (1971) relate to *losses* from drains to the underlying aquifer, whereas in the case of the Upper Thurne the flow is typically from the aquifer into the drain. The numerical values of drain coefficients deduced by using the linear relationship proposed by the Prickett and Lonnquist (1971) ignore the aquifer properties and rely pre-dominantly on the estimated hydraulic conductivity of the drain bed deposits, and so previous investigations into drain-interactions may not be suitable for investigating gaining drains. The development of equations for drain-aquifer interactions must therefore include an analysis of how gaining drains behave with different hydraulic conductivity values for the underlying aquifer. The representation of drain-aquifer interactions (Section 6.6.2) involves the construction of a 200m wide and 6m thickness vertical-section finite–difference numerical model. This method of independently deducing the approximate magnitude of drain coefficients by use of fine-grid solutions has been found to be adequate when quantifying drain aquifer interaction in large scale numerical groundwater models (Rushton, 2006).

2.6 The use of ARCGIS as an aid in the preparation of a conceptual model in plan

The use of the geographical information management system ARCGIS to assist in conceptualising the modelling domain in plan view involves the construction of one regional database and several smaller databases. The regional database contains geological and hydrogeological input information for each cell within each layer of the numerical groundwater model. This regional database is comprised of three datasets, which are assigned to the model; the recharge and the hydraulic conductivity assigned to each of the upper layer cells; and the elevations of each of the layers. The smaller databases contain spatial information on specific types of cells such as drains, lakes or coastal cells and their individual attribute values. Within individual cells, it is important to ensure that the values assigned are consistent and that any values such as drain water level are at a higher elevation than the drain base elevation. This method of pre-processing allows for the easy identification, manipulation and error checking of input information and the collation of output data. Once the databases are constructed, the individual cell details are exported into Groundwater Vistas, the MODFLOW graphical user interface.

2.7 The representation of a variable density groundwater flow

2.7.1 Multiple density approaches to modelling saline inflow into coastal aquifers

The inclusion of the difference in groundwater densities between fresh and saline groundwater in numerical models usually takes two general approaches: the dispersed interface approach (Hukayorn et al., 1987) and the sharp interface approach (Reilly and Goodman, 1985; Essaid, 1990). The dispersed interface approach explicitly represents the presence of the transition zone, where there is mixing of fresh water and salt water due to the effects of hydrodynamic dispersion. The increased computational effort required to solve density-dependent transport problem has limited most solutions to two-dimensional vertical cross sections (SUTRA-2D). There are solute transport codes available that model three-dimensional flow but they are limited in their application to regional coastal systems by computational constraints (Essaid, 1990). This approach requires various parameters to be estimated in order to solve the advection-dispersion equation, including the magnitude of the dispersion coefficient which is usually very difficult to assess for a real aquifer (Volker and Rushton, 1982).. This information is not available and the saline water within the crag aquifer may have taken hundreds or thousands of years to obtain its current distribution.

The second approach, (the sharp interface approach) represents the distribution of fresh water and salt water within the aquifer by estimating the advection and dispersion equations and reduces the analysis by assuming that the transition zone is thin relative to the dimensions of the aquifer. A number of methods have obtained solutions for the sharp interface problem. One approximate method, known as the Ghyben-Herzberg lens (Todd and Mays, 2005), lacks accuracy in regions of high vertical velocities and is incapable of handling anisotropy and heterogeneity of the aquifer (Volker and Rushton, 1982). This approach is therefore considered inappropriate for the Upper Thurne catchment as the open ditch drainage system is, in places, in direct contact with the underlying crag aquifer and as such exerts significant vertical velocities and also the aquifer itself is both anisotropic (Erskine, 1991) and heterogeneous (Downing, 1959).

2.7.2 Single density groundwater modelling with particle tracking

Single density groundwater modelling with particle tracking is a useful method of estimating the flow of groundwater within a coastal aquifer. The simulation software MODFLOW (McDonald and Harbaugh, 1988) is a single density finite-difference model of groundwater flow. The particle tracking method traces flow paths by locating the movement of small imaginary particles placed in the field of flow (Anderson and Woessner, 1992) and uses the head distribution from groundwater simulation outputs to compute the velocity distribution (Goode, 1990). Particle tracking programs, such as MODPATH are used as postprocessors to MODFLOW (McDonald and Harbaugh, 1988) and workers such as Todd and Mays (2005) outline the benefits of using this method in model verification and identifying conceptual errors that are not easily detectable by the sole examination of head distribution. In this current study, the single density method is used in conjunction with a total groundwater balance and coastal groundwater balance to quantify the groundwater flow processes.

2.8 Synopsis

The aim of this study is to predict the long-term effects on the salt load that is being discharged at specific electrical pumps under proposed changes to the land drainage systems of the Upper Thurne catchment, but this understanding is dependent upon the available information. The available information is limited in part because of the lack of drain flow data and borehole data. However, the development of conceptual and computational models of the drain-aquifer interactions and the groundwater flow will assist in the understanding of the saline surface water movement within the catchment.

The previous understanding of this distribution of surface water salinity in the drains is based primarily on the theory that the salinity of the main drains is a result of direct inflow from saline groundwater from the immediate underlying crag aquifer. In reviewing the existing data and obtaining new data of the salinities of the various drains, it is now recognised that the salinities of any particular drain is dependent upon the salinities of both the upstream surface water and the upstream groundwater contributions. These groundwater contributions are quantified using proxy drain coefficients developed by constructing numerical models of vertical cross sections of typical drain configurations. The total catchment-wide groundwater contribution is estimated using a hydrograph separation technique of the discharge waters of the electrical pumps. The estimated values of drain coefficients are assigned to individual cells in the numerical model to quantify the drain-aquifer interactions throughout the drained area within the catchment. Since there are no actual measurements of groundwater salinities, since the usual method for obtaining deep-groundwater samples require borehole sinking, the representation of variable density groundwater using a single density groundwater model with particle tracking and groundwater and coastal water constraints is an alternative and suitable approach.

3 CHARACTERISTICS OF THE CATCHMENT AREA

3.1 Introduction

In this chapter, a review of the historical drainage techniques used within the research area is presented. An introduction to the geometrical and geological properties of both the crag (sand) aquifer and the superficial deposits that cover it is presented. An introduction to the geological units and the various soils found within the catchment are detailed along with the history of drainage windmills within the area. The use of groundwater as a water resource identified by the proliferation of wells throughout the catchment is detailed. The major surface water bodies such as the broads and the River Thurne are introduced along with the notion of sub-catchments within this drained landscape and their relationship to the modern Water Level Management Plans is identified.

3.2 Geology of the Upper Thurne

3.2.1 The Pleistocene (sand) aquifer

The Pleistocene Crag and the overlying glacial sands and gravels comprise the crag aquifer of northeast Norfolk and represent a very complex and heterogeneous semi-confined aquifer with the alluvium behaving as the confining layer. This assertion of the alluvium behaving as a confining layer is made with the understanding that only the alluvial covering within the catchment requires open ditch drainage, which implies that the natural drainage rates are insufficient to avoid surface water logging. Hence, the alluvial materials are hydrogelogically behaving in a manner similar to an aquitard. The crag is generally considered to be a layered aquifer with minor thin clay horizons (Downing, 1959) and a significant laterally continuous 'clayish' horizon at around -25m Ordinance Datum (Figure 3.1) that at least extends as far east as the coast (Holman et al. 1999).

A profile of the crag at the Ludham Borehole revealed 50.44 m of Pleistocene sediments resting on the relatively impermeable Eocene London Clay (Bristow, 1983), which is essentially non-water bearing and forms the basal platform that separates the crag aquifer from the Cretaceous Upper Chalk aquifer. Water movement within the crag is inter-granular, though the layering of the aforementioned clay horizons may reduce vertical permeability and thus promote horizontal flow. Typically the crag is reddish or brown in colour above the water table due to Iron (iii) oxide staining, but is greyish green to greenish black below the water-table due to the presence of glauconite (Holman, 1994).

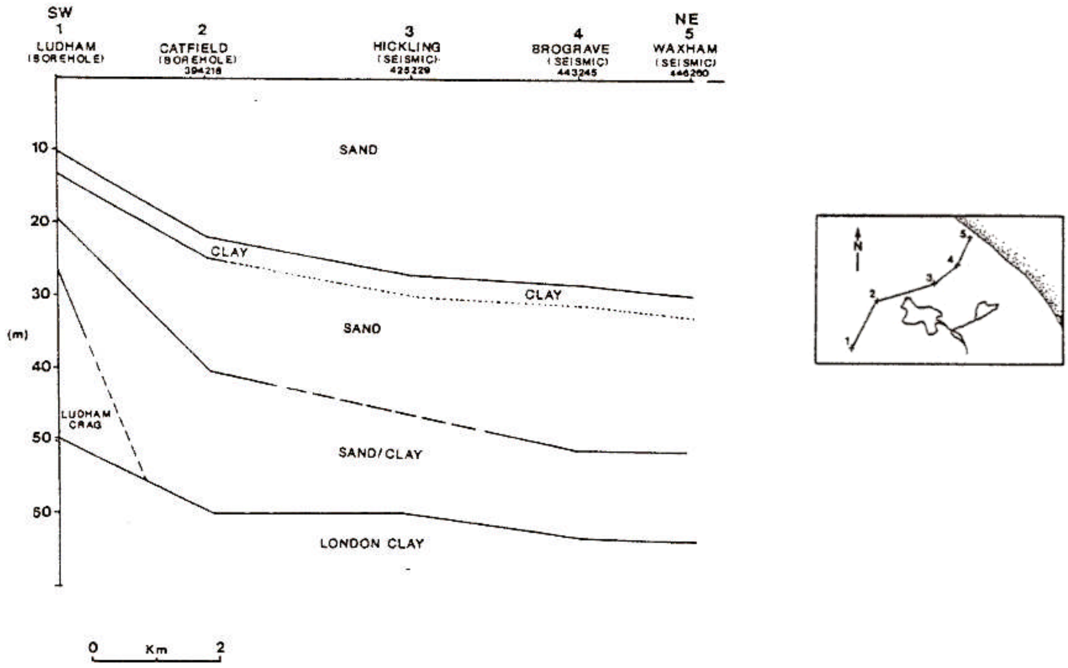

Figure 3.1 Cross-section of the Norwich Crag (Holman, 1994)

The problems associated with exploiting the crag aquifer for its groundwater often arise from its fine grained and high iron stained characteristics. The finer sands within the aquifer often significantly reduce the abstraction rates by clogging the screen, whilst iron-oxidizing bacteria such as *Gallionella Ferruginea* oxidize the dissolved iron, therefore removing it from the water and producing an insoluble precipitate of ferric hydroxides which blocks the borehole filter (Clarke and Phillips, 1984).

3.2.2 Superficial deposits

The Quaternary drift deposits that overlay the crag aquifer significantly determine the quantity of actual recharge to it. The most extensive drift covering the crag is the *Norwich Brick-earth [also known as Contorted Drift]*, which has been described as a sandy clay (Cox et al. 1989) and thought to only constitute a partial barrier to infiltration. Boswell (1916) classifies it as a sandy loam, which covers an area of 906.5 sq. km^2 and is remarkably uniform in mechanical composition. Apparently, no other geological deposit in this country possesses exactly this grade composition over its whole outcrop, the distribution of which within the Upper Thurne catchment can be found in Figure 3.2.

The Norwich Brickearth is found (Figure 3.2) in places within the catchment, south of the river and just south of Ingham to be covered with The *Corton Sands [Glacial Sands and Gravels]*. These sands are made up well sorted, medium- and fine-grained quartz sands (Cox et al. 1989) in strata up to 8 m thick and yield moderate supplies of water. A spring at Cranworth (outside the catchment) yielded 8 litres /second (Chatwin, 1961). However they are rarely used for obtaining water supplies as the unconsolidated sediments make it difficult to sink and complete the abstraction boreholes (Cox et al. 1989).

The *Boulder Clay [Lowestoft Till Group]* that is seen on the surface in patches along the southern boundary of the catchment was formed because of the last glacial shift (estimated to have occurred 40,000 years ago). A borehole at Corton (4.5 km south of the catchment) revealed Pleasure Till as the uppermost sub-layer underlain by the Oulton Beds which is in turn underlain by the Lowestoft Till (Arthurton et al. 1994). The Boulder Clay is found (Figure 3.2) to have a thickness of no more than 10 m within the catchment and due to its high clay contents has the effect that it reduces the amount of recharge to the underlying aquifer. In Table 3.1 are preliminary values of hydraulic conductivities and thicknesses for the various geological deposits found within the catchment.

Table 3. 1 Preliminary values of hydraulic conductivities

Geology	Saturated hydraulic conductivity	Typical thickness
Marine alluvium	Kv=0.00055 m/day*	1-5m
River alluvium	Kv=0.00055 m/day*	1-5m
Lowestoft Till (Boulder Clay)	Kv = 10^{-6}-10^{-4} m/day (Lloyd et al.1981)	3-4m
Corton Sands (Glacial Sands)	K_H = 10-100 m/day (Holman et al. 1999)	7-8m
Norwich Brick-earth (North Sea Drift)	Kv = 0.1-1 m/day Van Genutchen (sandy loam)	0-9m
Norwich Crag (Pleistocene Crag)	Kz = 10-100 m/day (Holman et al. 1999)	65m

* for further information see Section 6.2.3

The *Alluvium* found in the marshes is of two varieties; river alluvium, which is peaty and is found in the area around the Hickling marshes and northern marshes (Holman, 1994). Whereas the other variety, found in the main river valley, is of a clayey material transported and deposited by the River Thurne and is known as Marine alluvium.

3.2.3 Geological landscape units

The Upper Thurne catchment can be represented by four distinct ‘geological landscape’ units (Loam Uplands, Holmes, Flegg Hundreds, and the Marshes) that each have different hydrogeological settings due their different surface geologies and their lithological profiles. The northern and western coverage of Norwich Brick-earth and the area of exposed crag forms what is known as the *Loam Uplands.* The surface elevation varies from 1 mODN just outside Ludham in the south to 10mODN just east of Stalham. Patches of glacial sands and gravel have been found at elevations of +5 mODN approximately 8 km to the east of Ingham and groundwater levels have been found to be as high as 7 mODN in Happisburgh in the far north of the catchment (Holman, 1994) suggesting a significant coverage of unmapped glacial sands and gravel. The lithologcal profile of this geological unit is Norwich Brick-earth and then the underlying crag to the north and exposed crag in the southeast area around Ludham.

The *Holmes* is represented by the isolated elevated patches of Norwich Brick-earth found along the coast and within the areas of alluvium. The surface elevation within the Holmes is notably higher than the surrounding alluvium, and is typically 1 to 6 mODN. This elevation removes the need for drainage control of groundwater levels within the Holmes. Drains are cut around the margins of the patches of Norwich Brick-earth to intercept groundwater flowing from the Holmes into the marshes. The lithographical profile of this geological unit is Norwich Brickearth over the underlying crag.

The region south of the River Thurne marshland is locally known as the *Flegg Hundreds* (Figure 3.2) and typically is relatively hilly terrain with altitudes exceeding 20mOD. The geological profile is Lowestoft Till, which sits on the Corton Sands beneath which is the more extensive Norwich Brickearth. The Wick soil (Section 3.2.4) is typically a well-drained soil of wetness class I (Hodge et al. 1984) so that there is therefore no need for artificial land drainage within this part of the catchment, thus allowing a significant proportion of the precipitation to infiltrate through to the underlying till.

The marshes of the Brograve catchment (Figure 3.3) are covered with the Altcar 2 soil Association which is peaty, underlain at a shallow depth by marine alluvium (Hodge et al. 1984). The formation of the peat occurs when plant material, such as grasses and sedges, in marshy areas are inhibited from decaying fully by acidic, waterlogged conditions. The southern marshes are covered with the Newchurch soil association, which is a pelo-calcareous alluvial gley (clay) soil (Hodge et al. 1984), and developed, in estuarine alluvium.

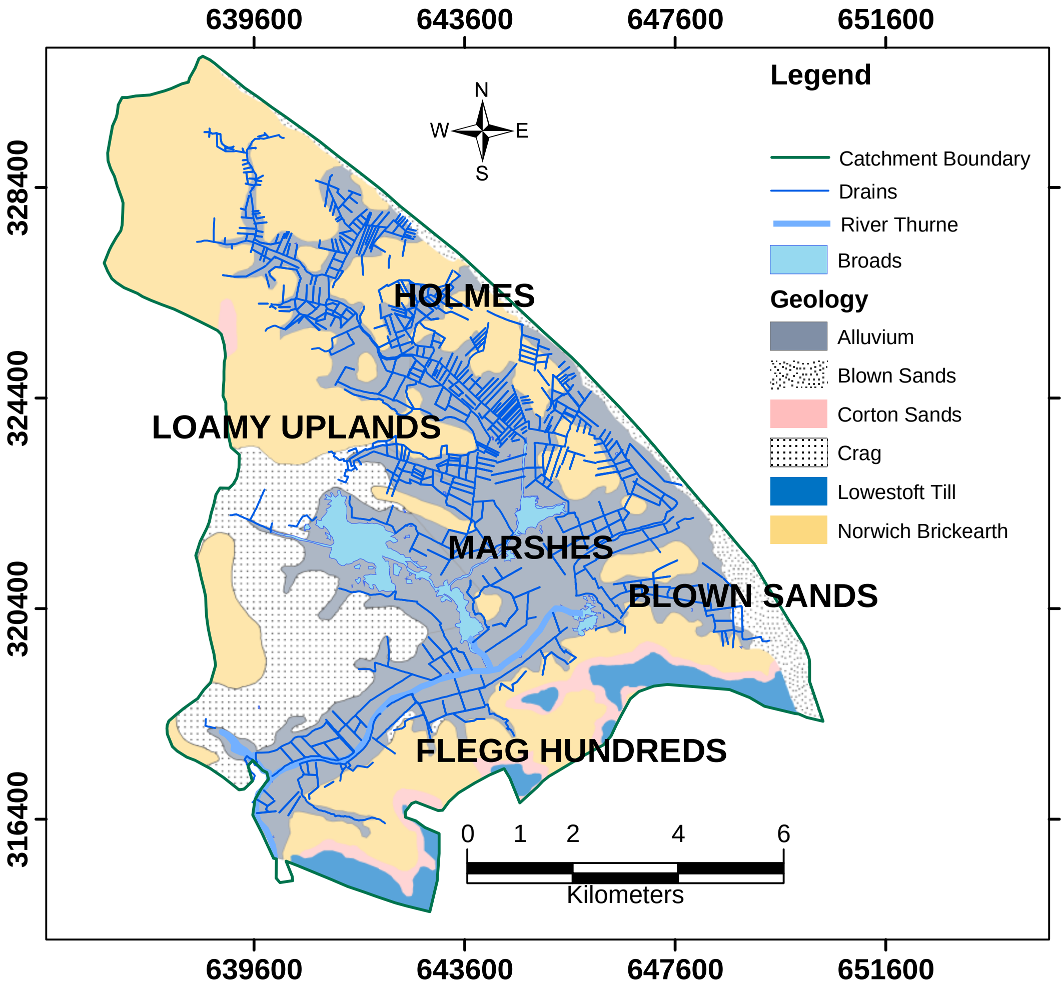

Figure 3.2 The various geologies and geological units of the Upper Thurne catchment

3.2.4 Soils of the catchment

The soil properties affect runoff generation and the degree of water retention and therefore actual evapo-transpiration by vegetation and the potential rates of recharge. The slightly elevated land (Loam Uplands, Holmes and Flegg Hundreds) away from the marshes are covered with soils of the Gresham and Wick 2 and 3 Associations (Hodge et al. 1984). The main soil type of the Gresham Association, the Gresham Series, is a poorly drained soil and is silty in its upper horizons but is overall coarse-loamy. When found on slightly elevated terrain, the Gresham Series soils are well drained, being moderately permeable. Gresham association (Figure 3.3) is found within the catchment area covering the underlying Norwich Brickearth in parts of the western uplands and around Horsey, Waxham and Sea Palling.

The Wick 2 and 3 Association covers the remaining western uplands and all of the uplands south of the river (known as the Flegg Hundreds). The main soil of the Wick Associations is the Wick series, which is typically coarse loamy over sandy or coarse loamy Cover loam. The Wick series have highly permeable surface and subsurface layers, which are unaffected by groundwater and are well drained. (Hodge et al., 1984)

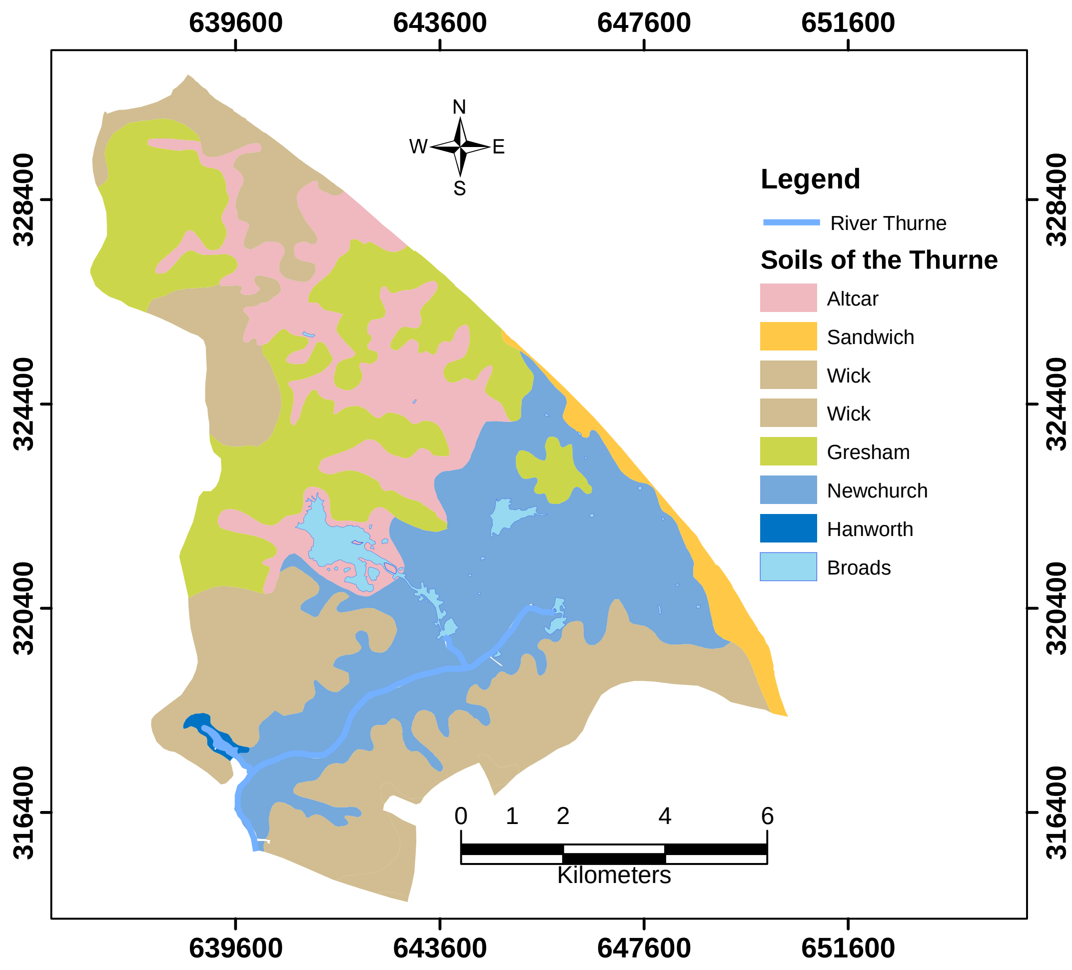

Figure 3.3 Soils found within the catchment (Hodges et al., 1984)

The upper marshes of the river Thurne contain the Altcar 2 soil association, which are developed in various thicknesses of peat over mineral substrates (Holman, 1994). The acidic nature of the Altcar Association (Hodge et al. 1984) and hence the ochre production, is a distinguishing feature of the soil. The other major association that is found within the catchment is the Newchurch 2 Association, which covers the majority of the remainder of the marshes along the river Thurne. It is described as a silty clay, with approximately 50% clay (Hodge et al. 1984) and requires efficient field drainage systems to control groundwater levels.

A relatively small patch of the marshes south of Ludham is covered with Hanworth Association soils. These are typically deep non-calcareous coarse loamy and peaty soils with high groundwater (Hodge et al., 1984) and thus often require artificial drainage. The Sandwich Association which borders the dunes are deep calcareous and non-calcareous sands (Hodge et al., 1984).

3.3 Broads and river Thurne

Typically, the Broads are characterised as shallow lakes (maximum depth 2 metres) with uniformly soft bottoms of gravel, peat or clay covered by a thin layer of mud (Lambert et al. 1960). There is evidence that they are in hydraulic continuity with the crag aquifer, and therefore a potential source of recharge or a point of discharge. Horsey Mere, which is the major receptor for the ochre-rich water of the Brograve level, lies to the northeast of Heigham Sound, which is connected to Heigham Sound via Meadow Dyke that runs south-westerly (Figure 3.4). Hickling Broad drains into Whiteslea Broad and then subsequently into Heigham Sound, which in turn is connected to the River Thurne via the south-easterly flowing Candle Dyke.

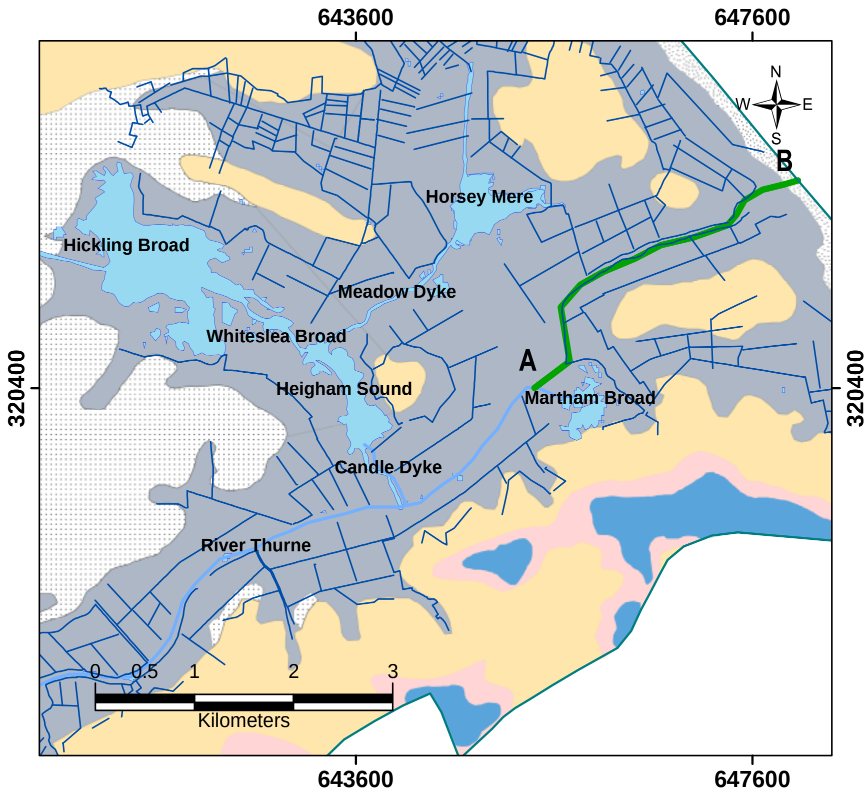

Figure 3.4 The major surface water bodies and the path of the Hundred Stream

The river's course in medieval times ran from just north of Martham Broad directly out to sea, the path of which is depicted by the line A-B in figure 2.4 and is called the Hundred Stream, so called because it formed the boundary between the hundred of Happing and West Flegg (Williamson, 1997). Due to the substantial wastage of the Altcar peat deposit caused by drainage and cultivation, the River Thurne is several metres above the level of the marshes and thus drainage pumps are needed to raise the water several metres from the marshes to the rivers or Broads.

3.4 The drainage of groundwater within the Broad lands

3.4.1 History of wind mill drainage in north east Norfolk

The modern history of land drainage within the catchment dates back to the early 18th century when the land was 1 metre higher than today (Ellis, 1965) and the process of land reclamation had began to mature and thus the gradual proliferation of mills (Moss, 2001). Mills in northeast Norfolk were built to grind corn or to drain land to allow the farmers cultivate the land, which had a high water table. The land suffered sea breaches in 1717, 1718 and 1720 and at this time, the Brograve marshes were still un-drained and were only protected from sea flooding by the 3 metre-high sand dunes (George, 1992).

Table 3.2 The approximate construction dates for drainage wind pumps within the catchment (Williamson, 1997

Drainage Windmill	In Existence by
Brograve	1771
Lessingham	1797
Morse's Thurne	1820
St Margaret's (Fleggburgh)	1840
Eel Fleet (Potter Heigham)	1860
Lambrigg	1880
Calthorpe	1880
Ludham Bridge North	1880
West Somerton	1898
High's (Potter Heigham)	1806
Eastfield	1862
Martham	1908
Stubb Mill	1795-1825
Repps Level	1795-1825
Catfield (fen)	1816-1840
Ludham Womack water	1824-1840
Catfield (swim coots)	1825-1840
Hickling Broad	1825-1840
Horsey	1712*

*rebuilt to replace a 200year old mill that had become derelict

In 1771, Sir G. B. Brograve built a series of drainage channels (ditches) through Brograve Level, and the Brograve mill. Survey maps by Faden in 1797 reveal that the Brograve catchment was more extensive than now and included several areas now designated as Hickling level (Williamson, 1997) and showed the existence of Waxham (Brograve) Mill. The Enclosures Act of 1801, which restricted the usage of lands by common people to graze animals on areas when planted by crops, now required that the General Commissioners construct flood banks and drainage mills.

In the early nineteenth century the Eastfield and Stubb Mills had been constructed and water from Stubb mill was now being discharged into Meadow dyke via newly constructed drains including the Commissioners' drain, which resulted in the loss of three small broads: - Wigg's, Gage's and Hare Park Broads. It was at this time that steam drainage began to assist, and in some cases replace, wind power (such as at Brograve in 1850). The replacement with fewer but larger steam-driven mills made the management of the Marshes cheaper. The more localised windmills which were operated by mill men who lived in very close proximity and could operate them at very short notice was under threat, although the progression from wind to steam was not simple or steady. It was the spread of electricity through the national grid, which saw the demise of wind pumps at the hands of electric pumps (Williamson, 1997).

3.4.2 The use of electrical discharge pumps throughout the catchment

The Brograve outfall is pumped directly into a channel that discharges into the Horsey Mere and as such reduces the effect of the tidal backflow up the Upper Thurne (Williamson, 1997). The electrical drainage pumps are strategically located within the coastal marshes to remove water from the drains and transfer it into a nearby broad (such as Hickling Broad, Horsey Mere or Martham Broad) or the River Thurne. Holman (1994) carefully documented the electrical usage of the pumps (Horsefen pump not included) covering a period of two years from April 1991 to March 1993, including the salinities of the waters entering the individual mouths of the discharge pumps and showed that some pumps (e.g. Brograve) were discharging large amounts of saline water. The effects of the increased salinity within specific drains associated with improved drainage and thus the recipient surface water bodies had been investigated a decade earlier by Driscoll (1984). The investigator found that the increase in salinity could significantly jeopardise the security of the habitats of certain protected species. The contributing drains within the surface water catchments of the different drainage systems are represented by the various coloured regimes in Figure 3.5.

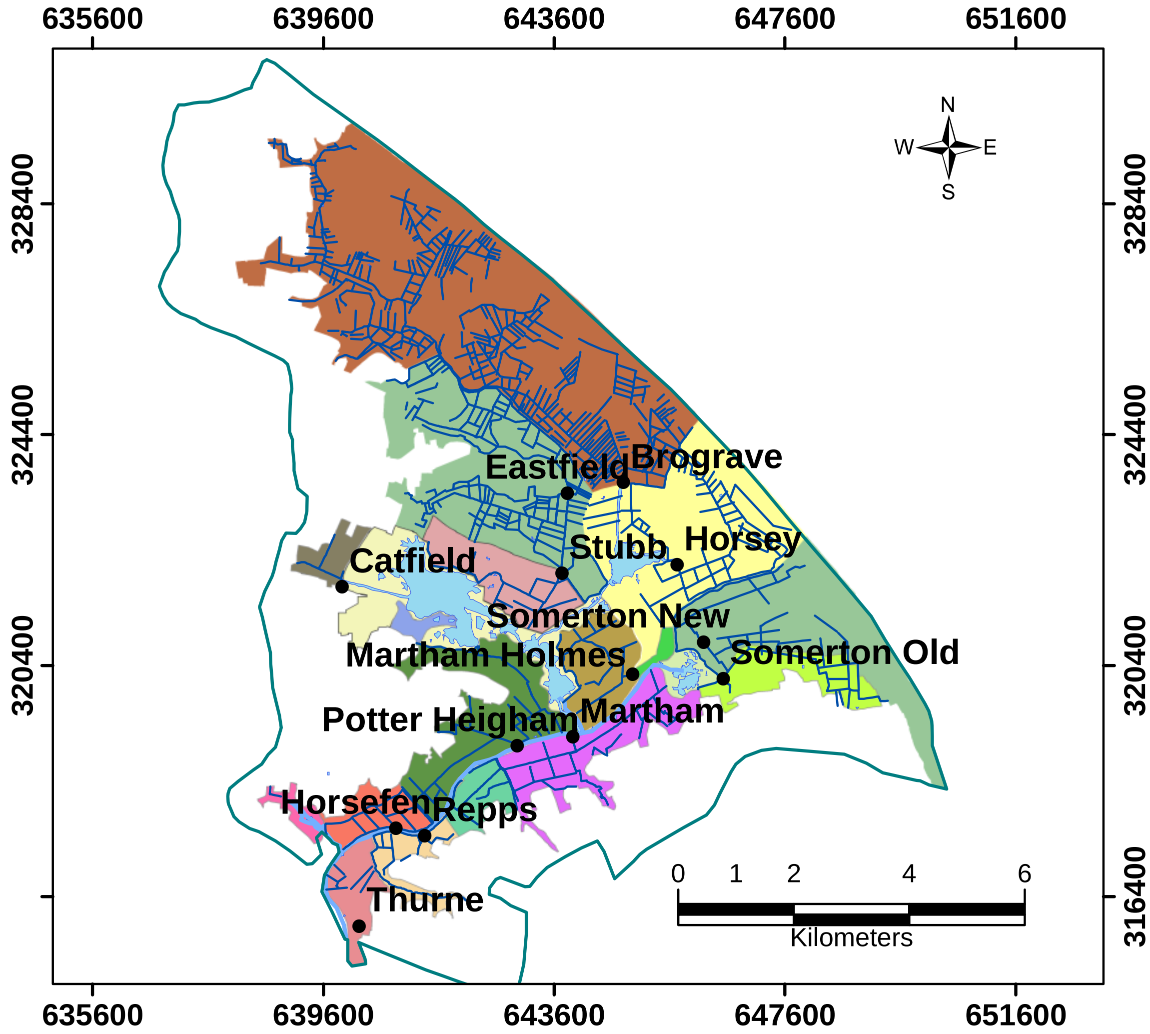

Figure 3.5 The various electrical drainage pumps and their *surface* water drainage sub-catchments

3.5 An introduction to the distribution of observation wells in the catchment

In the aforementioned study conducted by Holman (1994) a range of wells and boreholes that penetrated into the upper crag were monitored on a weekly basis to observe the water level fluctuations within the local groundwater levels. The locations and the subsequent hydrographs are analysed in detail in the following chapter, in Section 4.3.5 along with recently collected water level data.

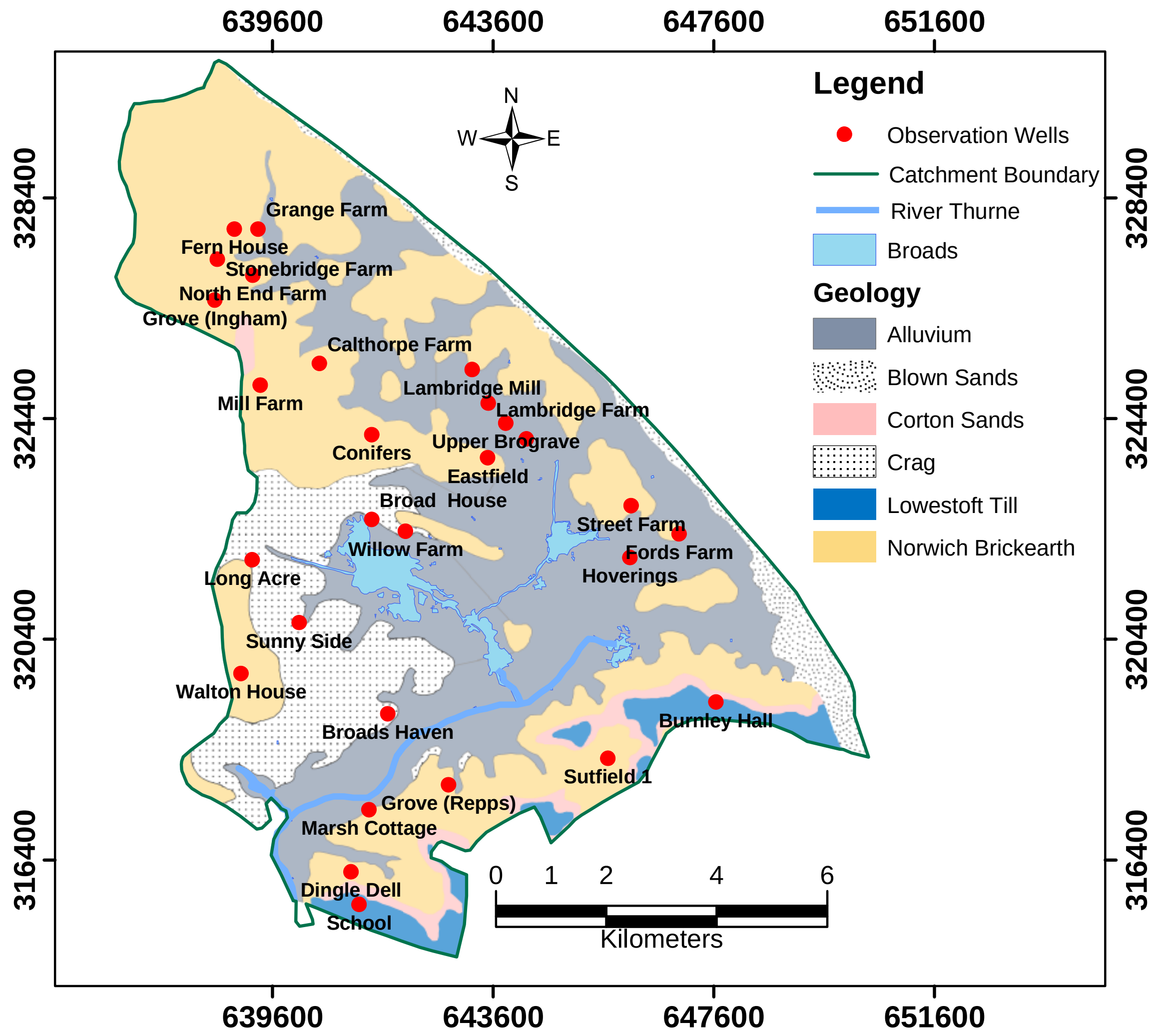

Figure 3.6 The locations of the crag observation wells (after Holman.1994)

3.6 Summary

The Upper Thurne catchment is a heavily drained coastal catchment with a semi-confined aquifer with two confining layers namely the peat and the clay. These confining layers restrict recharge and influence the behaviour of the horizontal groundwater flow. The unconfined regions covered with Norwich Brickearth are the main areas for groundwater recharge, and thus influence the vertical flows within the aquifer. The identification of four geological landscape units is made and their relationship with the surface deposits described.

The locations of the original distribution of wind drainage mills are more localised than the current distribution of electrical pumps. These wind drainage mills had substantially lower capacities than the modern pumps and therefore had to be located where they could be manually respond to localised changes in drain levels.

The raw data collected by Holman (1994) at the electrical pumps included electricity used and equivalent seawater amounts entering the mouth of the pump and water levels within the observation wells throughout the catchment. In Chapter 4, other field information collected will be used to help assist with the input information within the modelling process.

4 QUANTIFIED CONCEPTUAL MODELS OF THE UPPER THURNE CATCHMENT

4.1 Introduction

In the previous chapter the notion of landscape units, the structure of the crag aquifer, the existence of observation wells and the major surface water bodies such as the broads and the River Thurne were established and the drainage regimes described. To further understand and quantify the various hydrological and hydrogeological processes in the catchment, three separate conceptual themes have been developed in this chapter, namely, Surface Water Flow along the drains (SWF), Deep Regional Groundwater Flow (DRGF) and Drain/Aquifer Interactions (DAI) and quantified conceptual cross-sections are produced using available data. The three themes are intended to cover all the relevant aspects of the catchment hydrology and hydrogeology pertaining to the drainage network.

In Section 4.2, the flow along the network of drains, the Surface Water Flow (SWF), is conceptualised and investigated using data collected from the drainage pumps and from spatial sampling of drain water salinities throughout several of the drainage systems.

Within the conceptual theme of the Deep Regional Groundwater Flow (DRGF), detailed in Section 4.3, the influences on the movement of the groundwater in unconfined and confined regions of the aquifer are discussed. The historical observation well hydrographs are reviewed, with explanations given of the groupings and synchronicity of responses to recharge and water released from storage. Recently obtained well hydrograph data are analysed and compared to the historical data.

The configuration and location of the drains determine the level of interaction occurring between the drain and the aquifer, thus the Drain/Aquifer Interactions (DAI) is introduced and discussed in Section 4.4. Finally, cross sectional diagrams of the lithographical profiles are given of specific locations in Section 4.5, which describe the individual pathways of rainfall recharge through the various landscape units.

4.2 Conceptual Model 1-Surface Water Flow in open drains (SWF)

4.2.1 Introduction

This section focuses on the surface waters collected within the open ditch drains and subsequently discharged by the electrical pumps into Horsey Mere, Hickling Broad or the River Thurne, which are termed *high-level carriers*. This is important because the quantity and quality of the water within the drains provides an indication of where saline groundwater inflow to the surface water systems is occurring. Available field information includes spot readings of salinity of water samples collected from the drains (includes Electrical Conductivity readings) and the weekly readings at the drainage pumps of electrical consumption and salinity from Holman (1994). From this limited field information attempts are made to establish the possible locations of saline groundwater inflow into the drainage network and the equivalent volumes of seawater entering the deep drainage system. The different components of water in the deep drains are:

- rainfall runoff and water from the pipe drains in the cultivated layer above the peat/clay which are fed into the riparian collector open ditch drains;
- the perimeter drains along the boundary with the Norwich Brickearth which collect both excess runoff and provide an outlet for groundwater in the Loam Uplands;
- groundwater from the underlying aquifer.

The individual salinities and volumes discharged by the various pumps differ, but the process by which the surface water is removed from the field to the sea is essentially the same.

4.2.2 The use of under soil pipe drains within the drainage process

When the rainfall hits the relatively impermeable Altcar or Newchurch (collectively referred to as the alluvial marsh deposits), the complex interconnected system of drains combine to remove the excess waters and reduce the risk of surface water flooding. The undersoil 12 cm circular perforated pipes drain the marshland waters away firstly. These are buried beneath the surface topsoil at a depth of 60-70cm (see Plate 1) and typically cover the length of the individual fields, being positioned between 25-30 m apart.

Plate 1 The outlet of an under soil perforated pipe

These perforated pipes accelerate the drainage response time for the individual fields and subsequently allow water-sensitive crops to be farmed on fields within the marshes. They extend the length of the individual field and have their respective outlets in the riparian open ditch drains (Plate 2) which run perpendicular to them. The riparian drain waters all flow into the main drains, which are owned and maintained by the Internal Drainage Board (IDB).

Plate 2 Riparian Drain at Brograve

Plate 3 IDB main drain at Hempstead

These IDB drains (Plate 3) lead to the individual drainage pumps that then discharge into the high-level carriers such as the Horsey Mere or the River Thurne.

The conceptual diagram of a soil layer is represented in Figure 4.1. The relatively impermeable underlying peat forms the base upon which the cultivated layer sits. The net effect of the under soil pipe regime is that the water captured within the agitated soil (cultivated) layer is efficiently removed and the water table is lowered thereby providing the roots of the water sensitive crops with a sufficiently deep unsaturated zone.

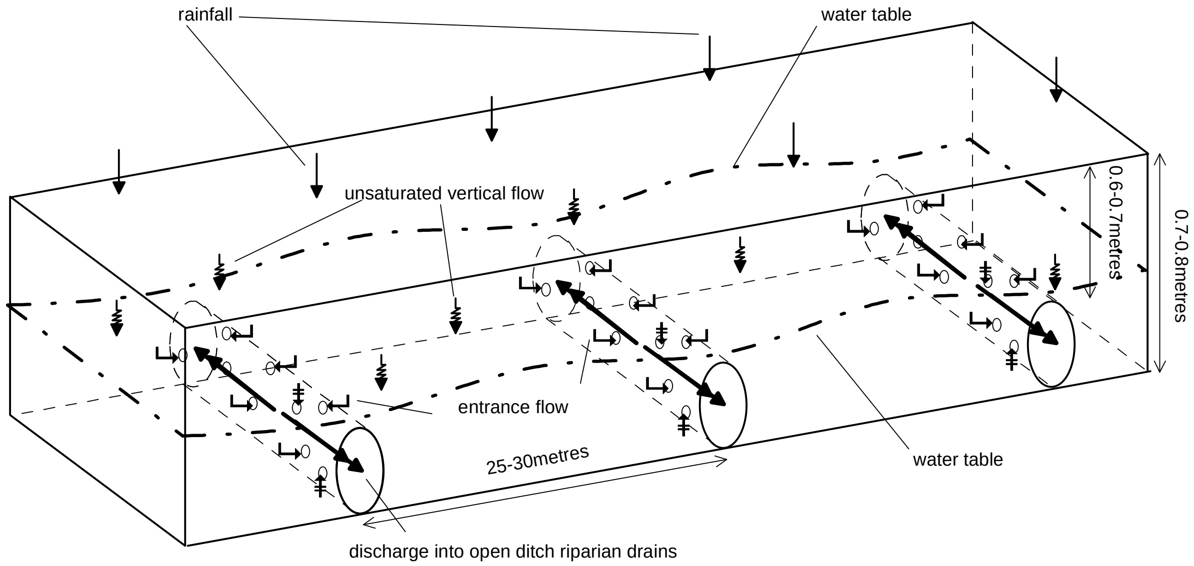

Figure 4.1 A conceptualised cross section of the soil layer within the marshland

The volumetric output of the individual components of this complex drainage system is shown in Figure 4.2. The water levels within the first three components are all below sea level and therefore the waters within the main IDB drains have to be lifted by the various electrical pumps to be discharged into the high-level carriers.

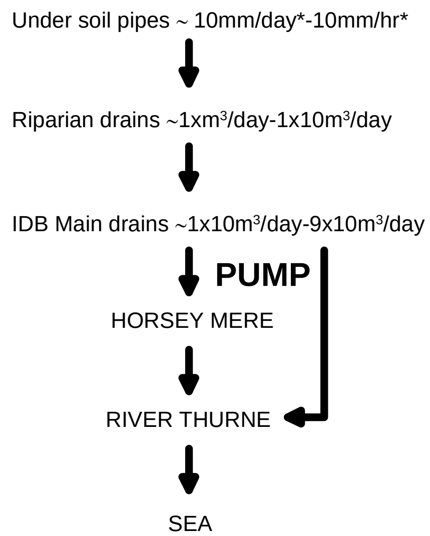

Figure 4.2 Schematic drain-water removal chart

(* N.B. The volumetric outflow of the under soil pipes are listed in heights reflecting the typical amount of rainfall recharge that can be removed using a collection of them at an optimum separation.)

4.2.3 The use of open ditch drains to capture runoff

The riparian drains are maintained by the landowners and in some cases are overgrown and poorly maintained, thus allowing water to enter into it freely but subsequent flow along it is often hindered by overgrowth. These drains often extend from the thickest areas of the alluvial layer to the margins and in some cases inadvertently draw in water from the underlying crag aquifer. In certain locations, as the examples depicted in Figure 4.3, there is a need to collect rainfall runoff from higher land before it reaches the relatively impermeable alluvium.

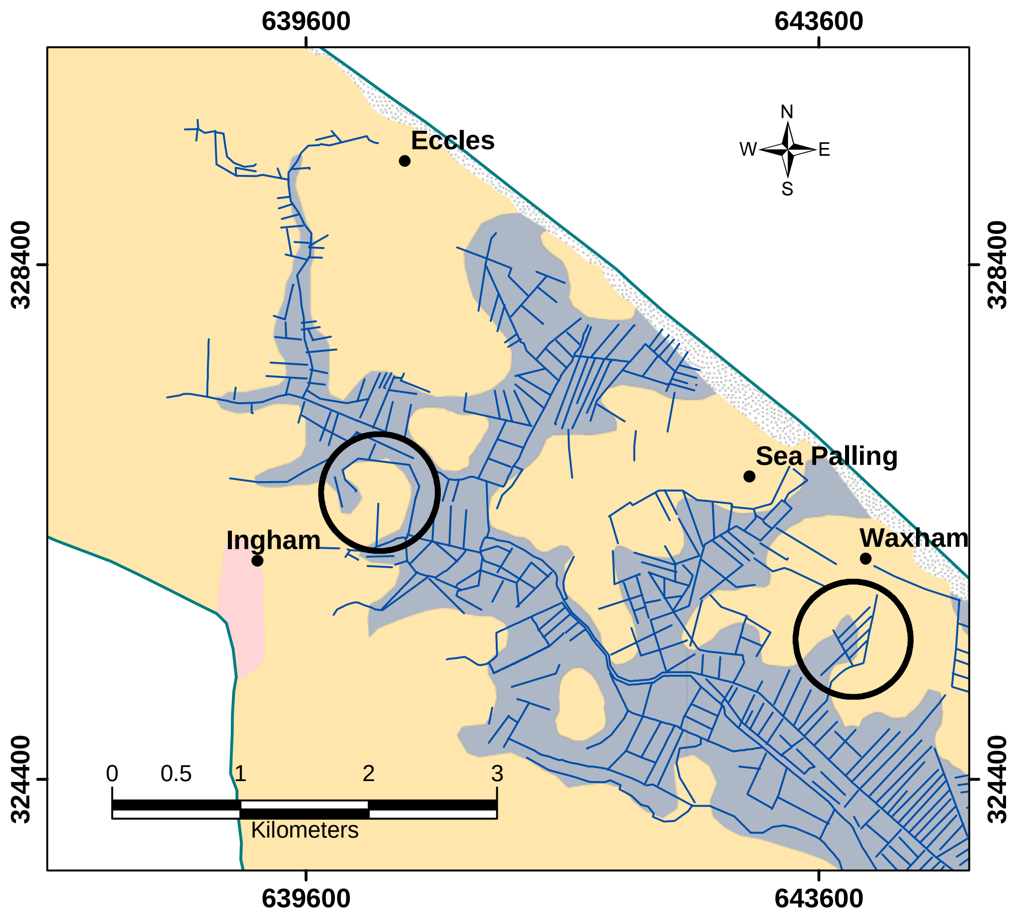

Figure 4.3 Examples of locations where the 'runoff' drains possibly cut into the underlying crag

These drains are cut along (perimeter) or perpendicular (spike) to the margins of the Norwich Brickearth (only two examples are given). Due to their location, they are often cut deep into the Norwich Brickearth or into the underlying crag aquifer and thus draw in significant amounts of groundwater. These drains are therefore in direct communication with the aquifer and may have an influence on the head fluctuations through out the catchment.

4.2.4 Previous measurements of salinities within the network of drains

In the spring of 1991 Holman, (1994) conducted a salinity survey of the dykes of Brograve, Horsey, Somerton Estate and Eastfield sub-catchments. These locations were chosen specifically to coincide with the sampling points of Driscoll, (1984), so that comparisons could be made. There were two sampling methods used to estimate the salinity; the first involved collecting samples of dyke water in 150 ml polythene bottles and storing them at 4°C until analysis of

Chloride ion concentration in the laboratory by titration using silver nitrate solution against a potassium iodide indicator with an accuracy of ±50 mg Cl^{-} l^{-1}. Seawater has a chloride ion concentration of around 19,500 mg/l or 19.5g/l. Table 3.1 converts specific values of Chloride content into equivalent seawater with the assumption that fresh water has a negligible amount of Chloride content.

The second method involved measuring the field samples on location using a calibrated digital electrical conductivity meter. Increasing concentration of salt gives higher electrical conductivity. Typical seawater has an electrical conductivity of 53,000 $\mu S\ cm^{-1}$ whereas deionised freshwater has a conductivity of 5.5 x 10^{-6} $\mu S\ cm^{-1}$. However, it must be noted that the electrical conductivity method is providing a measure of the Total Dissolved Solids (TDS), rather than the Chloride ion concentration.

Table 4.3 A linear conversion table that equates Chloride content to percentage of seawater (deduced from collected data)

Chloride (g/l)	*% seawater*
0.5	2.6
1.0	5.1
1.5	7.7
2.0	10.3
2.5	12.8
3.0	15.4
3.5	17.9
4.0	20.5
4.5	23.1
5.0	25.6

The samples collected in the various locations represent the spot salinity along the various branches; however, these individual spot salinities cannot reveal the extent of any surface water mixing. The measurements of spot samples at Brograve, Horsey and Somerton show that in some drains the percentage of seawater is 25% seawater or higher. (Figure 4.4)

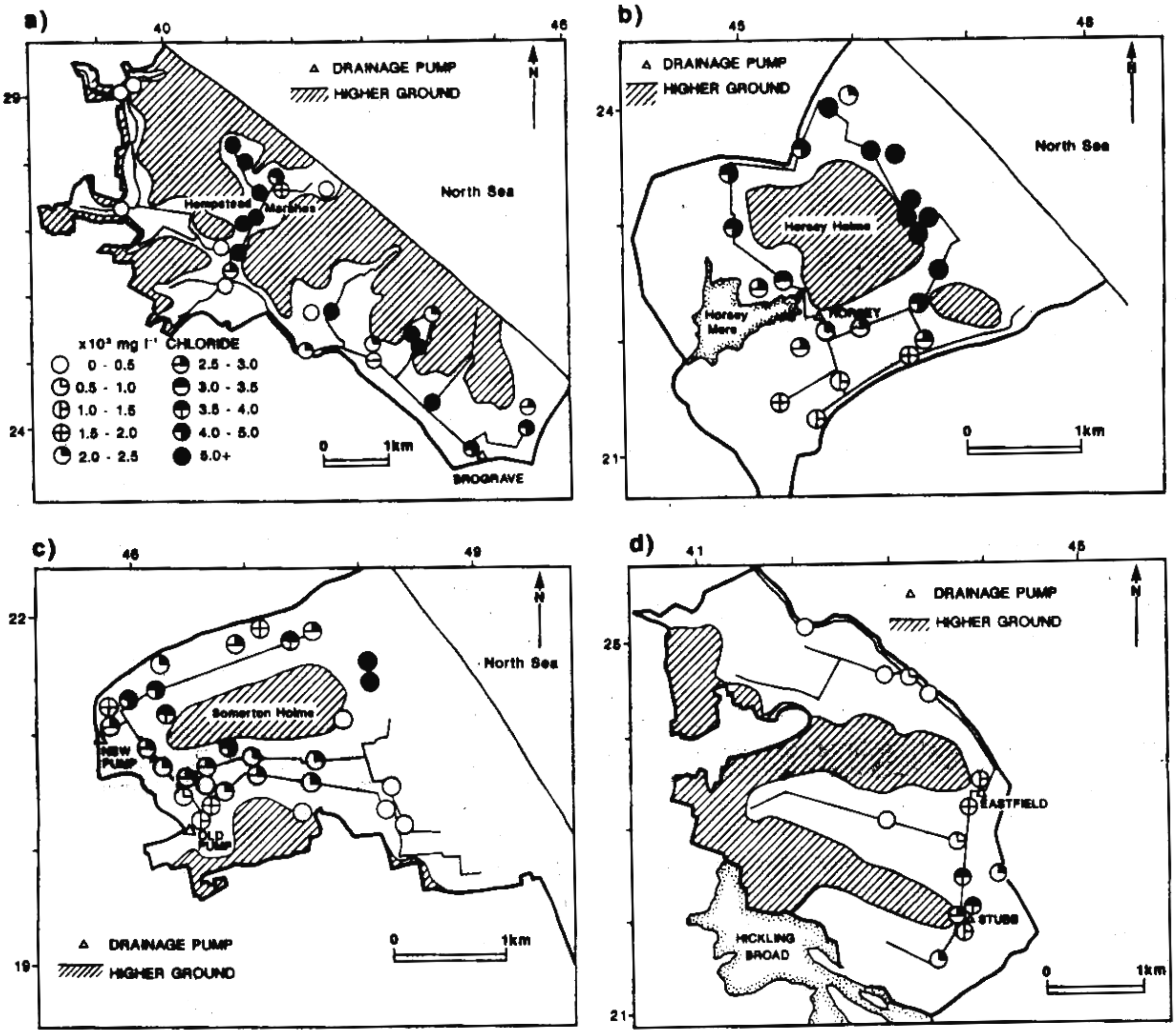

Figure 4.4 Distribution of surface drain water chloride concentration in the Happisburgh to Winterton IDB (map (a), (b) and (c) and the Smallburgh IDB (map (d) measured in spring 1991 (Holman and Hiscock, 1998)

The Eastfield and Stubbs sub-catchments (Fig. 4.4d) are the only non-coastal sub-catchments to be sampled and the values obtained show that the maximum contribution to the more saline drains is 20% seawater. This is a reduction of at least 5% seawater compared to the samples obtained at Brograve, Horsey and Somerton, indicating that the fresh water contributions from recharge and rainfall runoff are proportionally slightly more than that of the coastal sub-catchments. The Hempstead marshes results show that the percentage of seawater contribution along the main drain at Hempstead is approximately 50% with the electrical conductivities of the nearby riparian drains being substantially lower indicating a lower seawater contribution.

The collection of these high-saline sampling points and series of electrical conductivity readings assisted Holman, (1994) in his development of a map of the catchment with the locations where it was suggested the surface water would be prone to saline inflow (Figure 4.5).

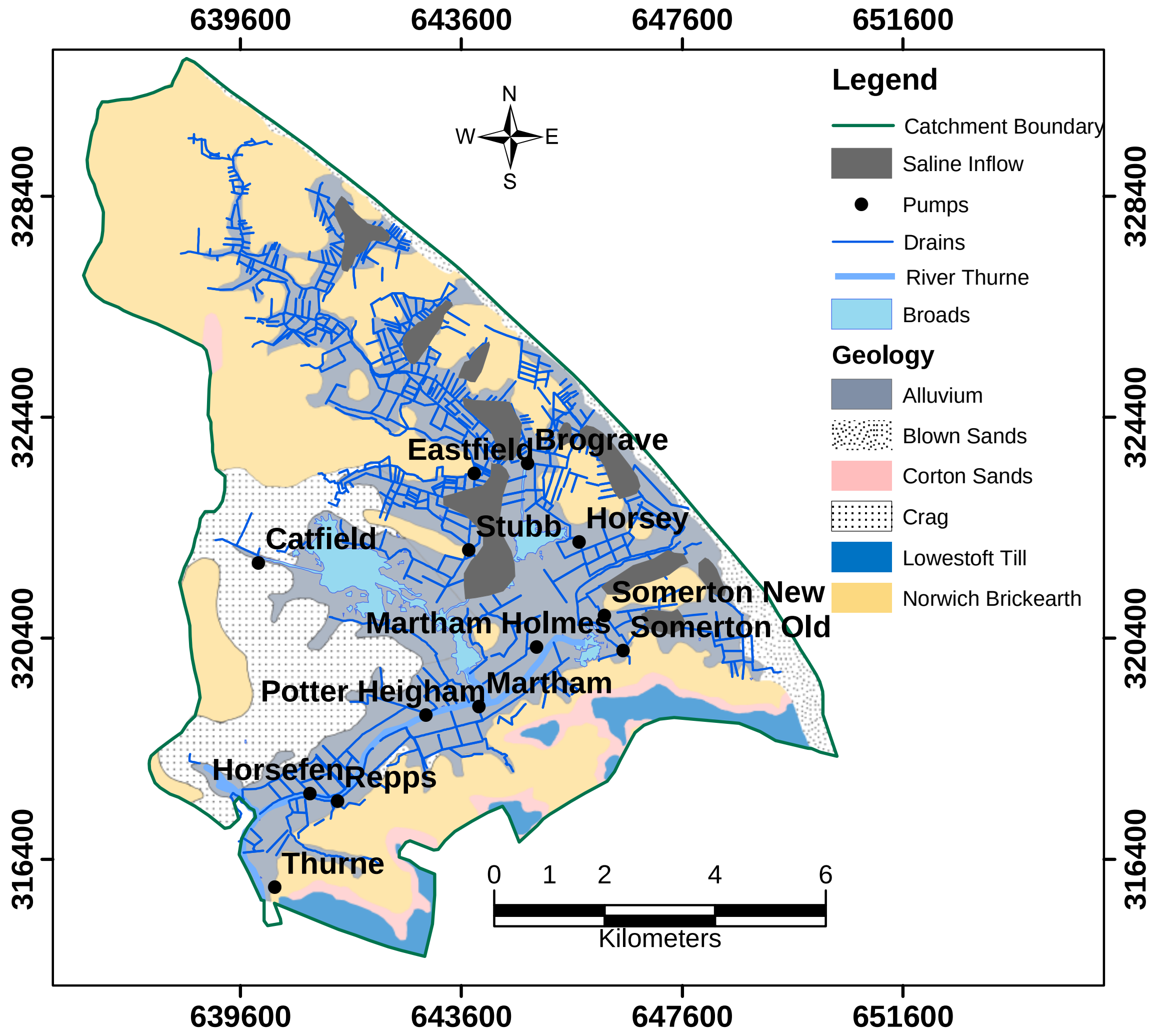

Figure 4.5 The areas Holman (1994) considered to have substantial saline groundwater inflow

4.2.5 Current fieldwork

Whilst drilling to determine the thickness of the Newchurch clay (Section 4.4.2) certain salinity and electrical conductivity readings were taken on nearby drain water samples. The water sampling of the Upper Thurne catchment was completed in the months of August and September 2005. The purpose of collecting the salinities for specific locations is threefold:

- To identify the current distribution of salinities within the drainage network (i.e. general increases since previous studies).
- Provide insights into the possible contribution of the riparian drain to the surface water quality of the main drain.

- Provide insights into the level of surface water mixing that occurs between the branches at Lessingham and Hempstead.

Figure 4.6 shows the locations where the salinities of the drain waters were tested.

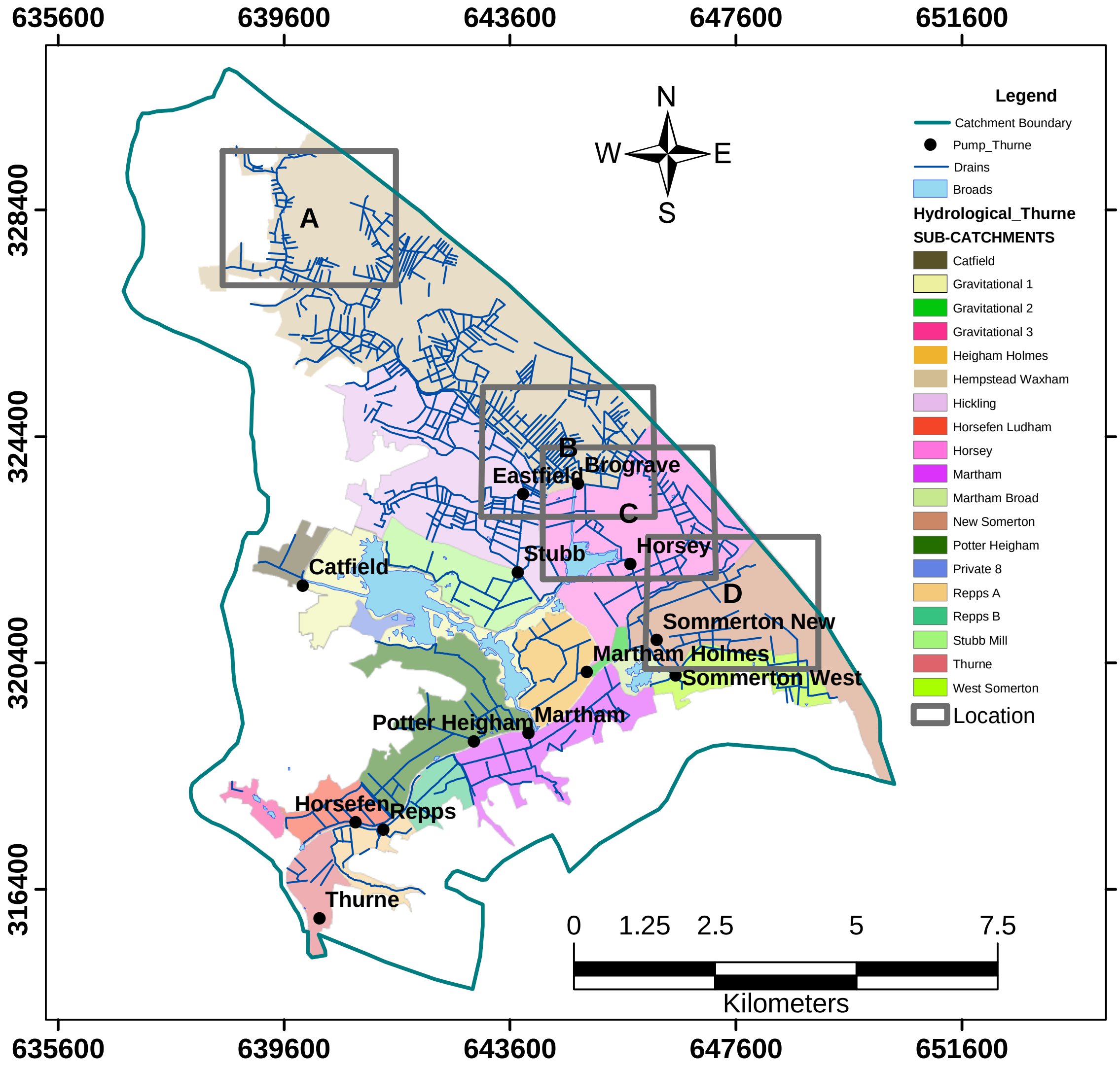

Figure 4.6 Locations where the salinities of the drain waters were tested

Spot readings of water samples from Hempstead main drain with an electrical conductivity (E.C.) of 45.08 mS/cm and 38.78 mS/cm and from Lessingham main drain 2.31 mS/cm, 2.44 mS/cm and 2.39 mS/cm demonstrate that the Lessingham waters are substantially fresher than the Hempstead waters (see Figure 4.7). However, the electrical conductivity of a sample (11.02 mS/cm) taken from the main drain after the confluence of these branches is more saline than the Lessingham sample but is fresher than the Hempstead sample suggesting a level of mixing. The spot readings of water samples taken along the stretch of the Brograve main drain all lie between the range of 8-11 mS/cm with the final reading being 10.25 mS/cm (Figure 4.8) suggesting a level of constant salinity in the waters along the main drain.

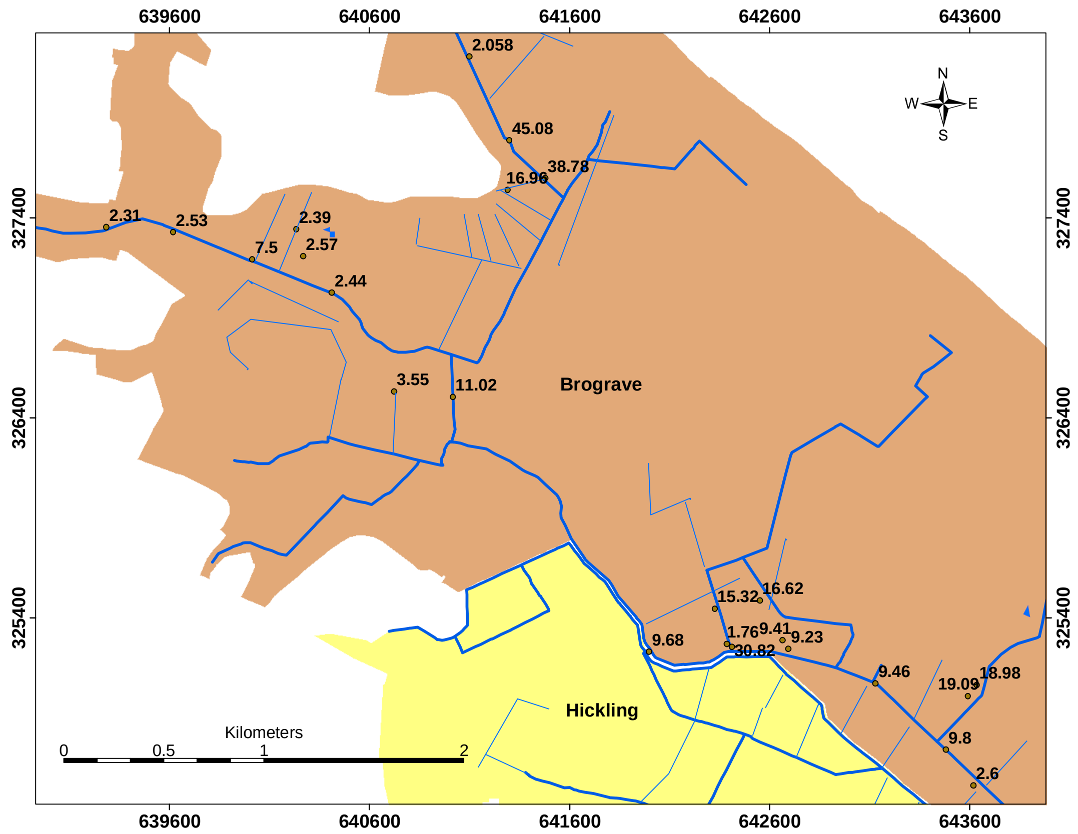

Figure 4.7 Ditch water conductivity (mS/cm) in the Lessingham, Hempstead and Brograve area

The readings of samples taken from seaward riparian (Figure 4.8) drains are substantially higher than the values taken prior (10.49 mS/cm) and after (10.51 mS/cm) the confluence suggesting the contributions from these riparian drains are not large enough to change significantly the saline proportion. If there were very substantial vertical up-flow into the main drain in the Brograve Level then the spot EC readings would demonstrate this saline inflow. Some locations may have vertical flows and therefore inflows exist, however these contributions are small as the spot readings do not reveal any substantial increase in conductivity (salinity). The electrical conductivity values shown within Figures 4.7-4.9, when compared to the spot readings (Figure 4.4) collected by Holman (1994) show a level of consistency in the locations of substantial amounts of coastal saline groundwater inflow.

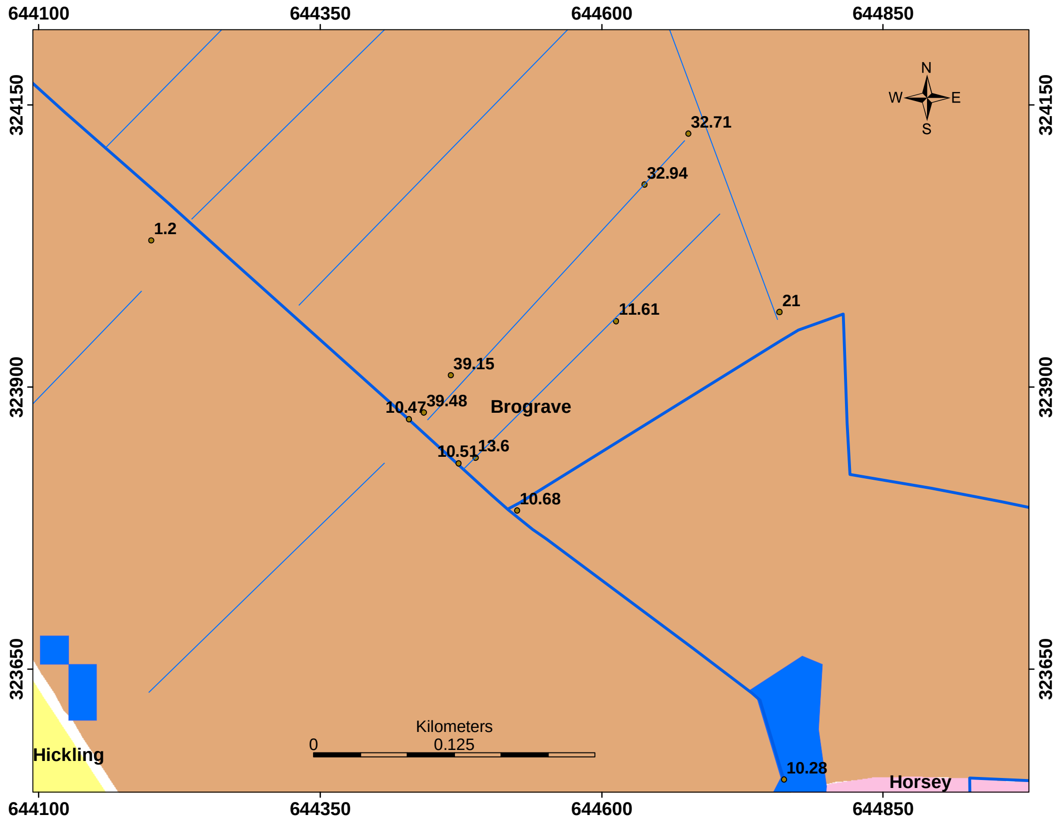

Figure 4.8 Ditch water conductivity (mS/cm) in the Lower Brograve area

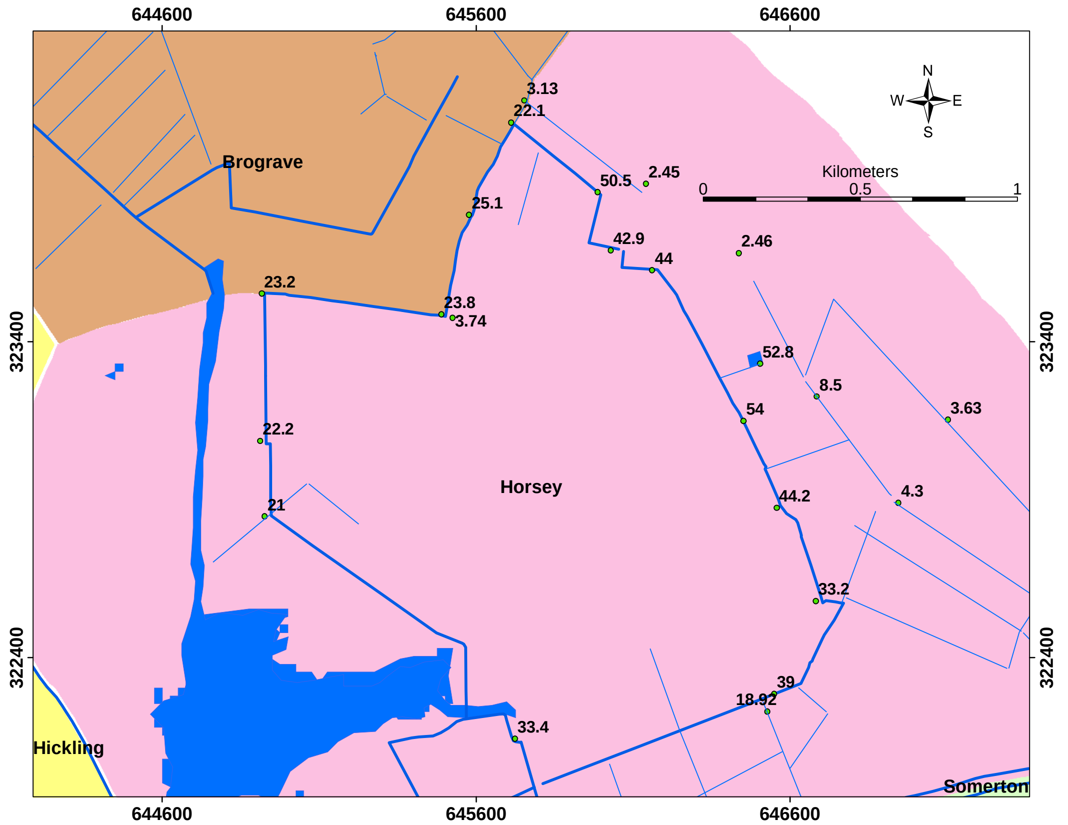

Figure 4.9 Ditch water conductivity (mS/cm) in the Horsey area

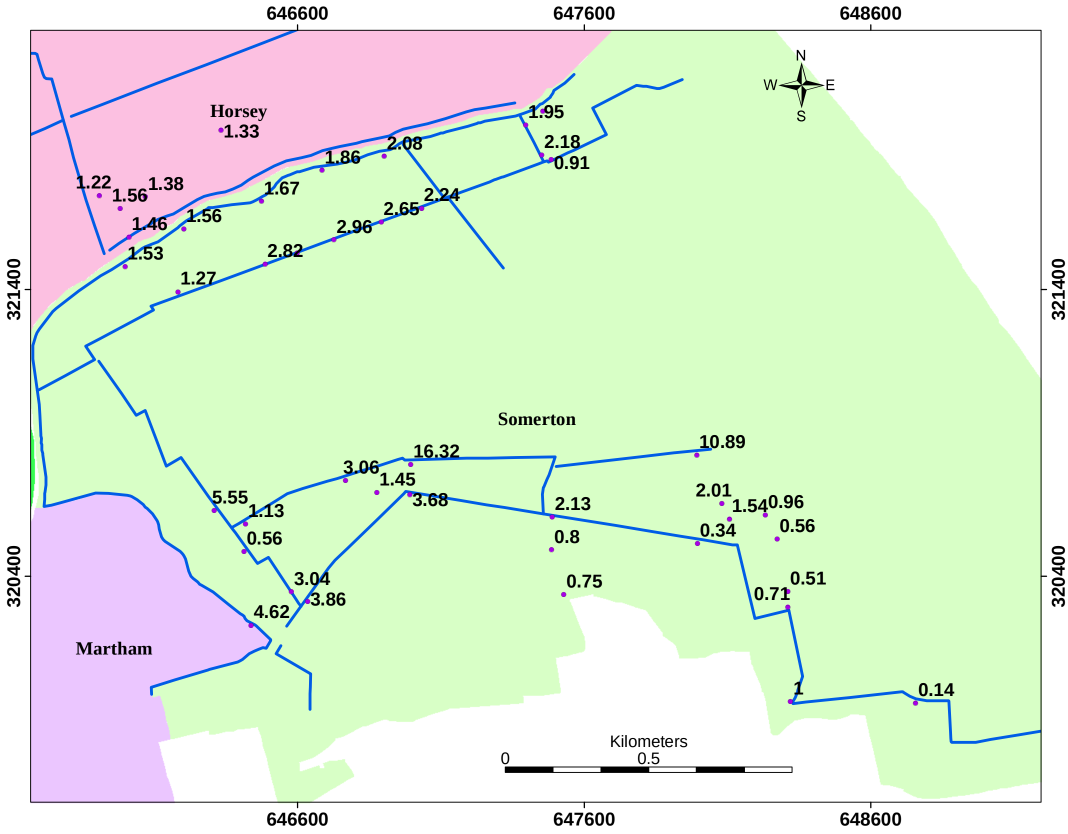

Figure 4.10 Ditch water salinities (mg Cl^{-1}/l) in the Somerton IDB area

4.2.6 Conceptualisation of seasonal fresh water/saline water interactions

The spot reading of electrical conductivities within various drains throughout the catchment reveal a level of surface water mixing. A conceptualisation of this seasonal variation is presented in Figure 4.11. The fresh water proportion within the drains often increases within the winter months whereas the inflows from the coastal region represent a higher proportion within the summer months.

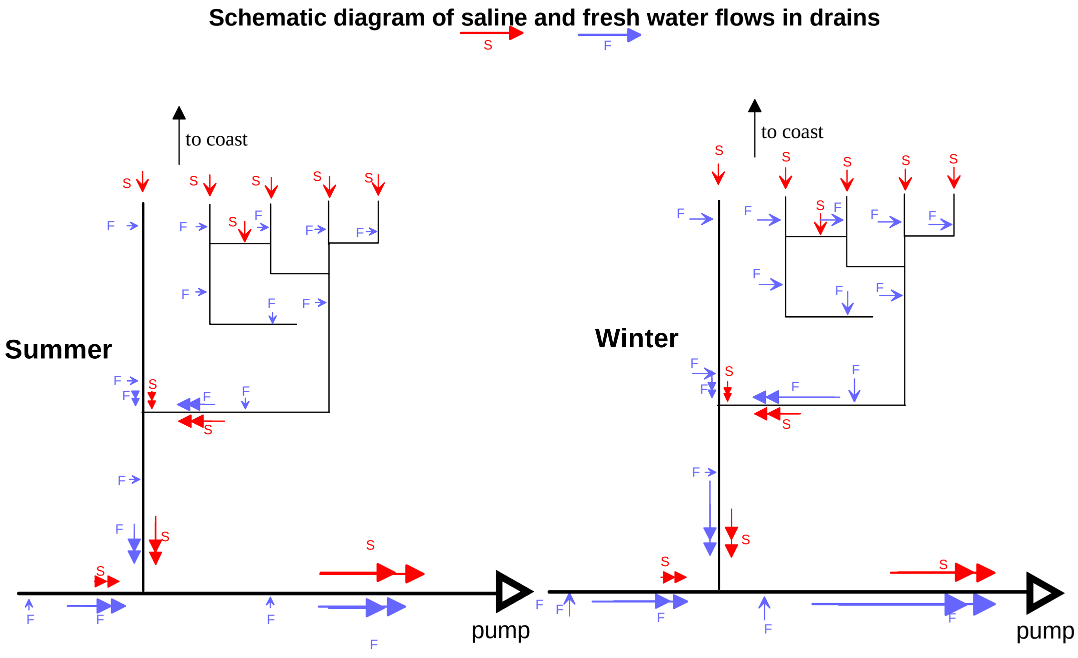

Figure 4.11 Schematic diagram saline and fresh water flow in drains

4.2.7 The estimation of drainage pump discharges from electrical consumption values

Throughout the catchment, there are electrical discharge pumps that lift the drain water into the high-level carriers. (e.g. the River Thurne or the channel leading to the Horsey Mere). Since there is no actual discharge data collected from the drainage pumps, the amount of electricity consumed since the previous meter reading and an electrical conversion factor are used to determine the discharge volumes. The actual values of these electrical conversion factors are a source of uncertainty within the hydrology of the catchment. Several workers have used various mechanically based theories to estimate these factors.

Watson, (1981) was the first to determine conversion factors for the various pumps within the Thurne catchment (Table 4.2), based upon typical pump efficiencies at the average head then operating during the winter. The method used and the parameters incorporated within these calculations are unfortunately unrecorded. Holman (1994) found that although the discharge rates and the drain water levels (i.e 'lifts') for the various drainage pumps were significantly different, the Watson (1981) conversion factors were all very similar. During the pumping test on the National Rivers Authority Fish Refuge Borehole at Catfield during January and February 1992 approximately 39,000 m^3 of water were discharged into the drainage system (LWRC, 1992), yet using Watson's conversion factor only about 23,000 m^3 were predicted to have been discharged by the drainage pump (Holman, 1994).

The Land &Water Research Centre Ltd (1992) method of estimating the volume of water pumped during the test was determined by applying the running time of a unit of electricity and so calculated the discharge from the rated capacity of the drainage pump:

$$\text{Volume Pumped} = \text{Units of electricity} \times \text{Running time per unit} \times \text{Rated Capacity} \quad (4.1)$$

Unfortunately when this approach was applied to the aforementioned investigation the estimated volume of water pumped rose to 76,000 m^3, indicating the need for an improvement in the pump conversion rates rendered by both Watson (1981) and LWRC (1992). The penultimate method of estimating the discharge volumes of the electrical drainage pumps is based upon the conservation of energy theory. The energy required lifting the volume of water and the energy dispensed due in discharging is estimated by considering the electricity consumed with minor corrections for frictional losses. Holman (1994) used the values derived from drainage pump theory for the Catfield, Eastfield, Potter Heigham, Repps, Stubbs and Old and New Somerton but derived another method for the remaining drainage pumps. The problem of determining which of the factors are to be used thus arises and therefore an estimation of the adequacy of the Holman (1994) figures needs to be determined.

Table 4.4 The various conversion rates for the electrical pumps of the Upper Thurne catchment.

Pump	Discharge Rate (m^3/s)	Watson (1981) (m^3/kWhr)	LWRC (1992a) (m^3/kWhr)	Holman (1994) Pump Theory (m^3/kWhr)	Holman(1994) (m^3/kWhr)
Brograve	0.72*	31.15	142.00	44	75.00
Catfield	0.17	30.58	100.00	153	153.47
Eastfield	0.42	37.10	57.20	66	66.35
Heigham Holmes	0.42	33.41	-	-	48.28
Horsey	0.78*	30.58	55.00	48	92.37
Ludham	0.25	34.00	-	-	56.88
Martham	0.70	N/A	35.00	57	80.53
Potter Heigham	0.76	37.10	-	54	53.91
Repps	0.34	34.00	-	82	82.52
Stubb	0.28	-	-	83	83.36
Thurne	0.32*	34.00	-	-	80.53
Somerton New	0.45	N/A	50.80	51	51.30
Somerton Old	0.20	41.63	56.60	63	63.20

The electricity values represent the amount of electricity used to pump the total volume of water since the last reading. This total amount is the sum of rainfall runoff (RO) and groundwater contributions (GWC). For each sub-catchment the estimated proportions of RO and GWC dependent on various factors such as percentage of recharge areas (exposed crag and/or Norwich Brickearth), percentage of sloped land (e.g. Loam Upland) and are therefore different. Once this ratio is estimated the amount of groundwater discharged at the pump can be estimated. The algorithm for estimating groundwater contributions discharged at an individual pump is as follows:

1. Total **annual** electricity used × electrical conversion factor = Total **annual** discharge volume.
2. Total **annual** discharge volume / number of days in the year (366 yr1 and 365 yr2) = Average **daily** discharge volume.
3. Average **daily** discharge volume × GWC/(RO+GWC) = GWC daily total target (fresh and saline)
4. Total annual amount had *X*% saline (from Holman (1994) salinity readings at the entrance of the pump) therefore Saline water target = GWC daily target × *X*%

4.3 Conceptual Model 2-Deep Regional Groundwater Flow (DRGF)

4.3.1 Introductory to Darcy's law in its simplest form

The fundamental theory of groundwater movement is based upon the simple Darcian equation, which relates the volumetric flow through a porous medium to the product of the hydraulic conductivity of that medium and the cross-section hydraulic gradient. To understand the behaviour of hydraulic flow the types of velocities encountered in groundwater must be investigated.

The velocity at which groundwater moves can be estimated by considering a pipe of cross-sectional area **A** and flow rate **v**. The volume of water moving per unit time is the flow **Q** and is given by the equation **Q = vA**. In an aquifer the velocity is given by the rearrangement of the Darcian equation

$$v = \frac{Q}{A} = -K\frac{dh}{dl} \tag{4.2}$$

where K is the hydraulic conductivity of the medium [m/d] and the dh/dl is the head gradient inducing the flow [m/m]. This however is the *Darcian* or *apparent* velocity and is the speed that water moves through an open pipe given the rate of flow measured. However, an aquifer is not an open pipe and the cross sectional area of flow is not open space in an aquifer. The actual distance travelled by the water is often far lengthier than the lengthwise displacement.

The pathways available for flow is much smaller equal to the *effective pore space*. To convert apparent velocity measured from flow through an aquifer into the actual velocity that the water is moving (i.e. the *seepage velocity*) one must divide specific discharge by the effective porosity N.

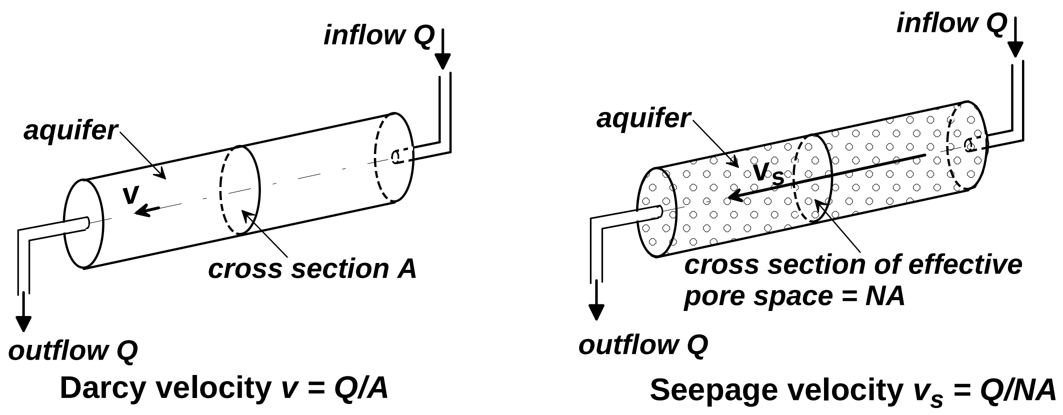

Figure 4.12 A diagram showing the difference between the Darcian velocity and Seepage velocity

4.3.2 The geometry and geology of the aquifer, inflows, outflows, and boundaries

The geometry and geology of the crag aquifer is depicted in Figure 4.13 with possible inflow and outflow lines and depicts the unconfined and confined areas of the aquifer. The unconfined areas of the aquifer occur in the west, from the groundwater divide to Eastfield (E) and from the Hempstead Marshes (HM) to the coast in the east with the confined area of the aquifer found between Eastfield and the Hempstead Marshes.

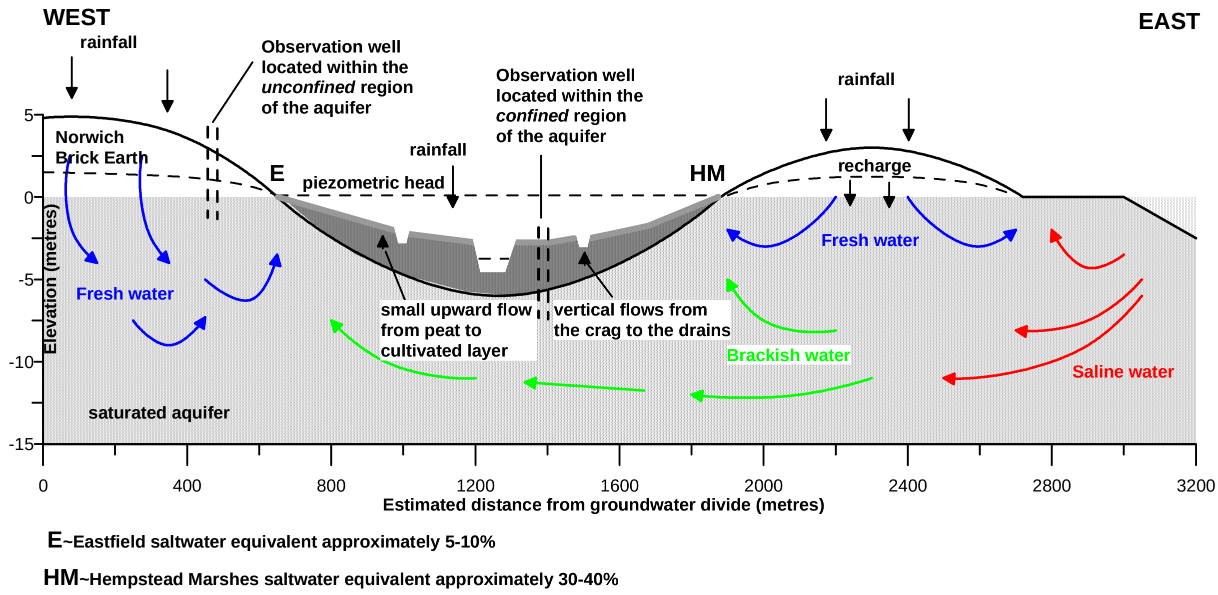

Figure 4.13 A cross-section of the Pleistocene Crag showing the mechanism for mixing

The freshwater flows *into* the aquifer are predominantly vertical and are a result of rainfall recharge or water released from storage (depicted by the blue arrows in Figure 4.13), whereas the coastal waters (saline) flow mainly horizontally into the aquifer from the sea (depicted by the red arrows in Figure 4.13). The groundwater which then enters into the drains that are well connected to the underlying aquifer is composed of a mixture of varying amounts of fresh and saline water (depicted by green arrows),.

4.3.3 The spatial variation in groundwater head responses throughout the catchment

The variation of groundwater heads with time throughout the study area depends on a number processes; these can be identified in the schematic cross-section of Figure 4.13. The landscape units of the Loam Uplands and the Holmes (Figure 3.2) are covered with the relatively permeable Norwich Brickearth and exposed crag, which allows the underlying crag in these areas to behave as an unconfined aquifer. In these regions, the groundwater level fluctuations can be due to recharge; abstraction; or the lateral movement of water through the aquifer towards discharge zones, especially drains. For a typical unconfined storage coefficient for the crag (Section 5.3.2), a recovery in water level of 0.3m in a week requires a vertical flow of 0.03m in a week assuming a specific yield of 0.1. Beneath the substantially less permeable alluvium the behaviour is typical similar to a confined aquifer. Here changes in groundwater head do not mean that substantial flows of water have occurred through the confining layer, as a confined storage coefficient is typically very small e.g. 0.0001. These changes occur mainly due to lateral flows within the confined aquifer such as flows to drains at the margins of the Norwich Brickearth and the alluvium.

4.3.4 The temporal influence of prominent rainfall events

In order to identify the hydrogeological influence of prominent rainfall events it is necessary to understand how evapotranspiration, runoff, and infiltration combine to determine the likelihood of recharge occurring (the effective precipitation is the actual precipitation minus the actual evapotranspiration, whereas infiltration is the actual precipitation minus the runoff). It is assumed that recharge can only occur when the soil is at or above field capacity (Rushton 2003) which is equivalent to zero soil moisture deficit. Soil moisture contents are estimated by use of *daily* approximations, with the start-of-day Soil Moisture Deficit influencing the likelihood of recharge. An examination of the rainfall totals for the study period (Figure 4.14), measured daily by the Hickling rain gauge of The British Atmospheric Data Centre (BADC), gives an indication of when significant actual recharge is likely to occur. Three prominent weekly rainfall totals are briefly examined. A weekly rainfall total of (0+0+0.7+0+5.2+8.4+31.8 =) 46.1mm (Figure 4.14) occurred towards the end of September 1991 (Week 26). During this period, the evapotranspiration would have been moderate, the 5.2mm and 8.4mm rainfall events possibly having negligible runoff (Table3.5, Rushton 2003), making the infiltration fraction substantial, and thus reducing the start-of-day SMD for the last day of the week. The subsequent 31.8mm rainfall event, even with a substantial proportion (20-30%) contributing to runoff, would still mean a substantial infiltration fraction available to reduce the SMD and thus making recharge likely to occur. The weekly rainfall total of (0+0+0+57.2+0+0+3.2 =) 60.4 mm occurred towards the middle of July 1992 (Week 69). The evapotranspiration would have been at its highest, the runoff proportion substantial, leading to a minimal infiltration rate making the likelihood of recharge minimal. The final prominent weekly rainfall total of (0+0+0+32.1+13.4+0 =) 45.5mm occurred towards the beginning of September 1992 (Week 85). The evapotranspiration would have been moderate, the runoff proportion for the 32.1mm being 20-30%, making the infiltration fraction large enough to reduce the start-of-day SMD for the sixth day. The subsequent 13.4mm rainfall event with a negligible fraction contributing to runoff would still mean a substantial infiltration fraction available thus reducing the SMD making recharge likely to occur.

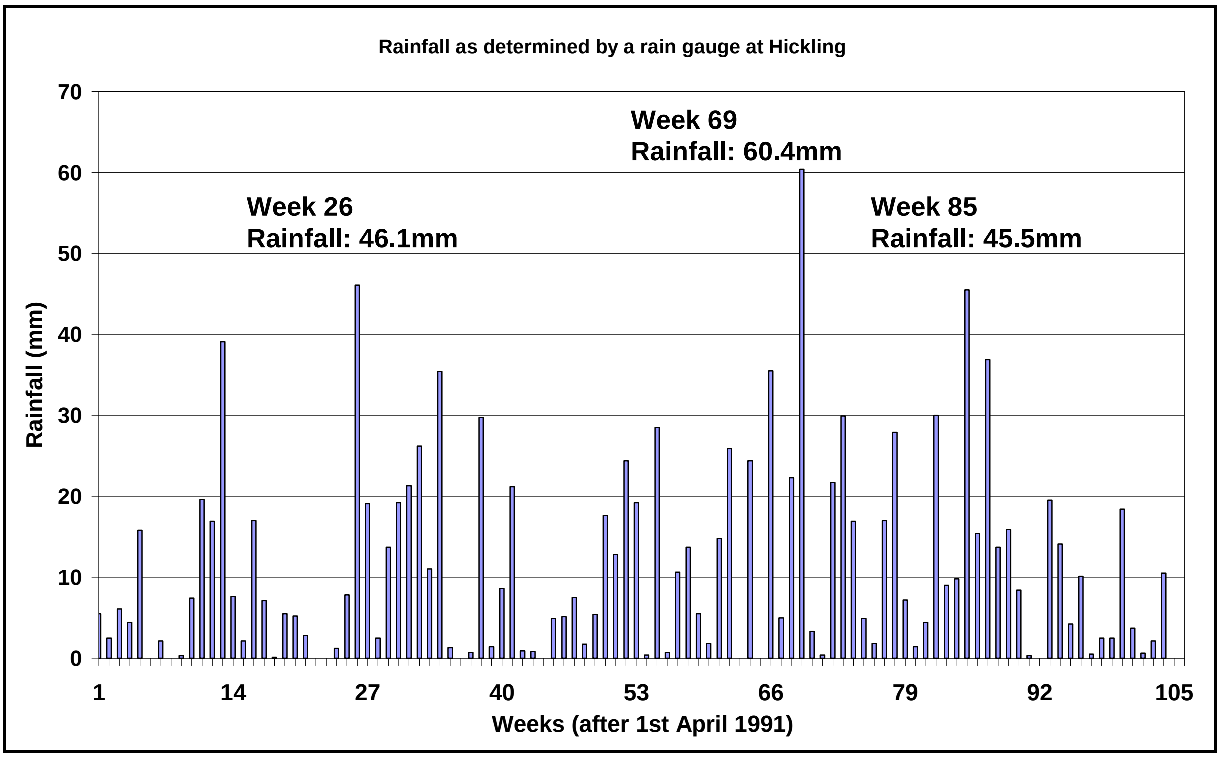

Figure 4. 14 Rainfall as determined by a rain gauge at Hickling

One important input to an aquifer system is a quantity known as *runoff recharge*. Runoff recharge is precipitation that has flowed from less permeable areas of the catchment, such as the Lowestoft Till, to more permeable sandy deposits and subsequently becomes recharge to the aquifer system. In the observation well hydrographs groups, substantial runoff recharge is identified by the uniform attenuated response to prominent rainfall events, such as the rainfall totals of Week 85, as shown in the hydrograph of the Flegg Hundreds (Figure 4.34).

4.3.5 An introduction to theoretical hydrographical groups

In identifying the major factors that influence the fluctuations that occur within the observation wells, monitored by Holman (1994), it is now possible to plot the distribution of different hydrographical groups (Figure 4.15 and Table 4.3).

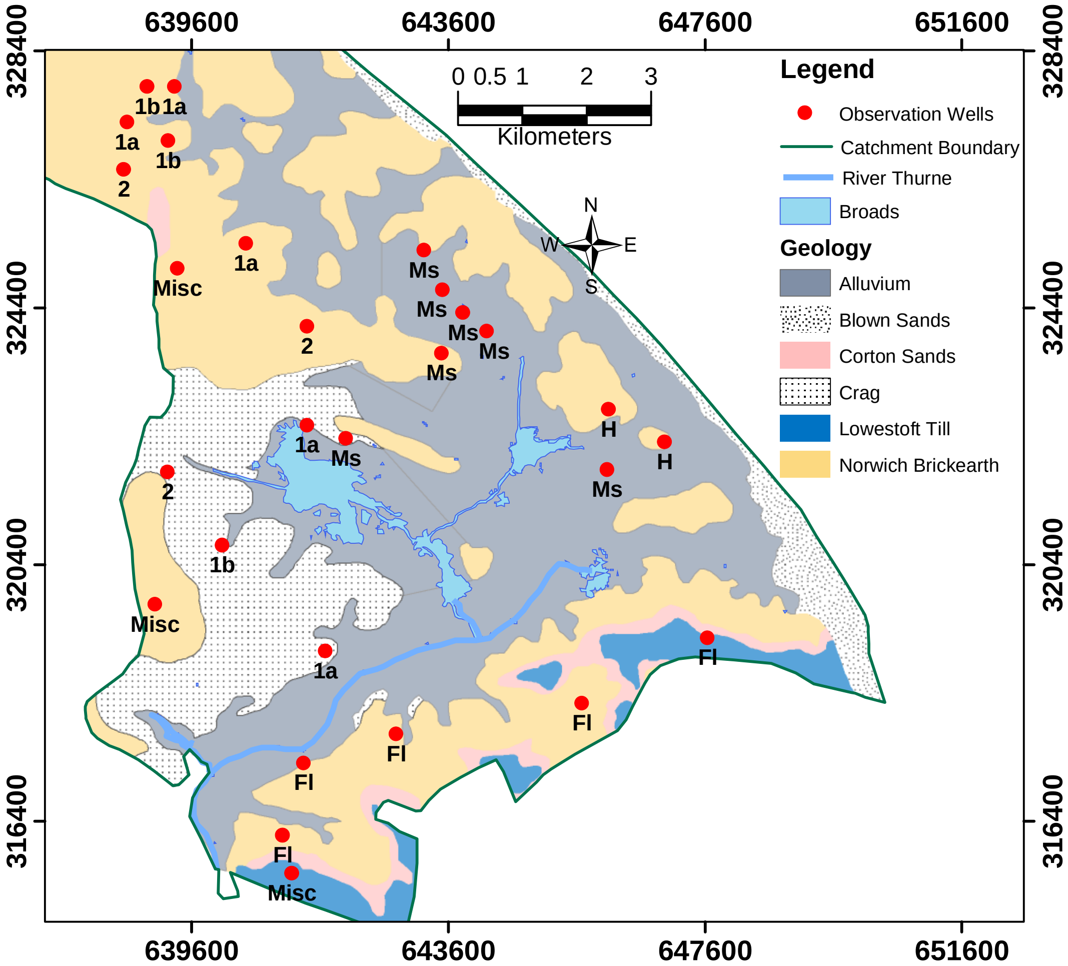

Figure 4.15 The distribution of the observation wells in the classification of hydrographical groups

The *theoretical* hydrographical groups can be analysed individually and collectively to render an improved understanding to the aquifer behaviour of the monitored period.

Table 4.5 The abbreviations used to denote the theoretical hydrographical groups

Geological Unit	*Hydrographical Groupings*
Flegg Hundreds	Fl
Holmes	H
Loam Uplands	1a & 1b, 2
Marshes	Ms
Miscellaneous	Misc.

4.3.6 Observations of the groundwater level fluctuations of the hydrographical groups

In an attempt to explain the general fluctuations of the water levels within individual wells located across the catchment, it is important to compare groups of observations wells that are located close to each other; this will indicate how localized conditions influence observation well fluctuations.

Within the hydrographical group of *the Marshes*, there are two distinct sets of profiles containing Hoverings, Willow Farm, and Lambridge Mill and containing Lower Brograve, Upper Brograve and Lambridge Farm.

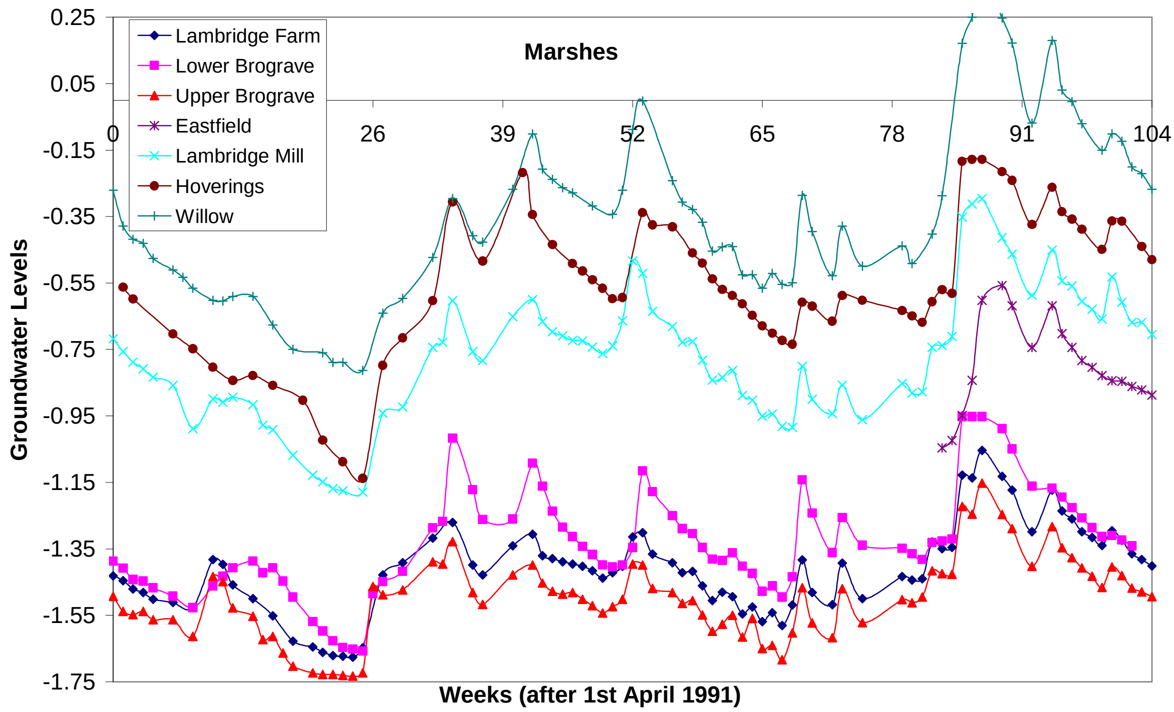

Figure 4.16 The hydrograph of the observation wells grouped as the Marshes

However, focusing on the hydrographs from Hoverings and Lower Brograve (Figure 4.16), they are very similar in general profile except for an approximate 0.6 m difference in water levels. One possible reason for this difference is that the high water levels maintained in the drains within the Horsey estate are influencing the groundwater levels at Hoverings.

The hydrographical profile of the observation well at Willow Farm (Figure 4.16) is similar to the profiles of these wells within this confined area except that the profile is generally higher. The observation wells of Lower Brograve, Upper Brograve and Lambridge Farm with fluctuations of about 0.15m are all close to the main Brograve Drain in which there is normally no significant variation in the water level. This strongly implies that the drain-aquifer interaction of the main drain and the aquifer immediately below it is limited.

In considering the upper metre of the alluvium, which is cultivated and contains under (or pipe) drains, the water level in this upper zone rises due to rainfall and falls due to drainage and evapotranspiration; with very heavy rain, it is likely that the water table will approach the ground surface. The water levels of the well at Eastfield Farm, which is on the Norwich Brickearth close to the edge of the peat, has fluctuations that are similar to the water levels of the well at Lambridge Mill, which are partly due to fluctuations in the cultivated zone of the surrounding land. In the case of the minimum water level obtained at Willow Farm, the well is influenced by inflows in the unconfined regions and so when the water level is low there is a slight release of groundwater from a nearby portion of aquifer that has a higher groundwater head. The maximum groundwater level in this marginal area of the aquifer is not restricted by a confining layer and is susceptible to inflows from the surface water body of Hickling Broad.

The well at Hoverings (Figure 4.16) provides important insights; the minimum head is about -1.1 mOD and the maximum about -0.2 mOD. The elevation of the land surface at Hoverings is around -0.1mOD, while the Somerton Pumps maintain water levels in the drains of the neighbouring drainage level at around -2.0 mOD. The groundwater head in the confined aquifer beneath Somerton is likely to be similar to that in the confined region below the Brograve marshes (due to similar Main Drain water levels) at about -1.6 mOD. Both of these regions will therefore lower the water levels at Hoverings. The water level of the observation well at Lambridge Mill is typically about 0.6 m above the water level in the Lower and Upper Brograve wells- this is because the heads in the cultivated layer are higher possibly due to a direct connection to watercourses (Waxham Cut).

The observation wells at Fords Farm and Street Farm are located (Figure 4.17) within the landscape units known as the *Holmes* (Figure 3.2). These Holmes constitute an *unconfined* region of the aquifer and therefore a source of recharge. The Week 69 rainfall event increases the groundwater levels within both the Street Farm and Fords Farm wells by approximately 700 mm within 4 weeks. When compared to the fluctuations measured at the observation wells at Willow Farm and Lambridge Farm, which penetrate the confined region of the aquifer, the fluctuations are similar. This suggests that the drain-aquifer interaction on the margins of the Marshes-Norwich Brickearth is the dominant effect as opposed to the main-drain aquifer interaction.

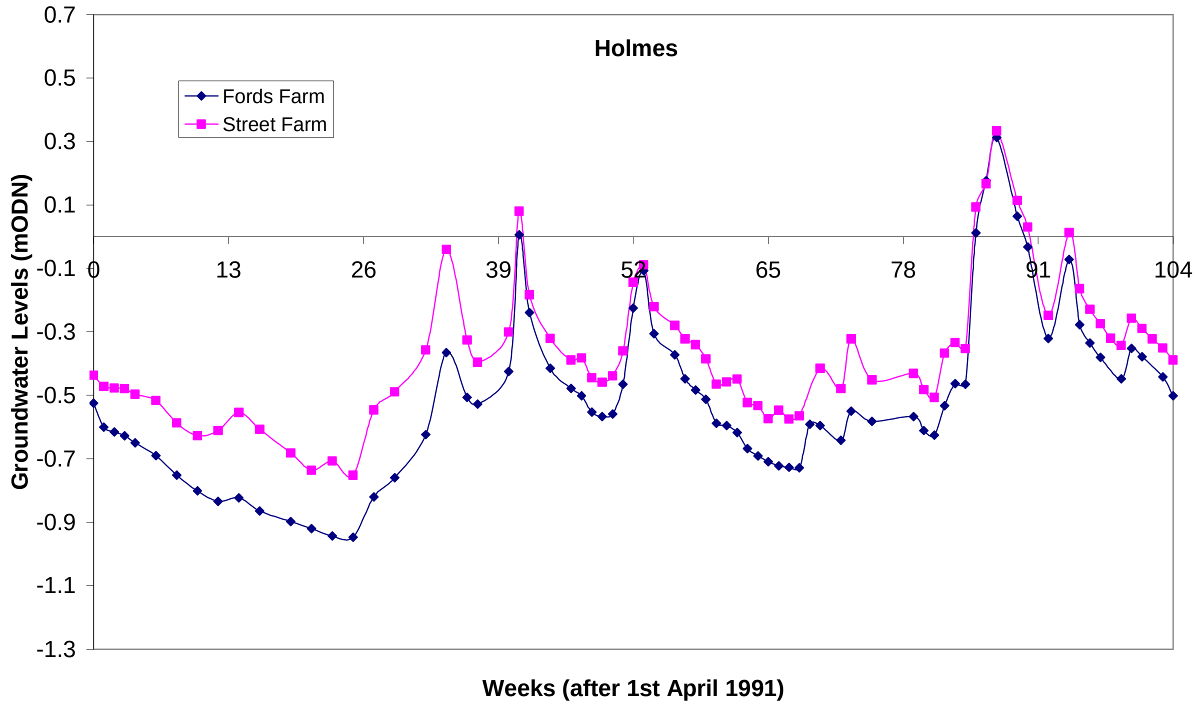

Figure 4.17 The hydrograph of the observation wells grouped as the Holmes

Loam Uplands 1a (Figure 4.18) the large recessions shown in the hydrograph of the well at Stonebridge Farm during Weeks 0 to 25 and Weeks 53 to 66 is possibly due to the lateral flows that are occurring through the Norwich Brickearth towards the spring drains. These drains are cut on the margins of the Norwich Brickearth and therefore are well connected to the underlying aquifer. The observation well at Broads Haven maintained water levels below sea level, possibly due its proximity to a drain within the Potter Heigham drainage sub-catchment. The observation well at Calthorpe Farm responded to the major rainfall event of Week 85 at about Week 88 with a water level change of approximately 2.3m, which is an unusually large increase, suggesting the possibility of flooding. The observation wells within this hydrograph group have a tendency to respond noticeably to the prominent rainfall events (Figure 4.14) and have substantial recession rates, possibly due to high rates of infiltration.

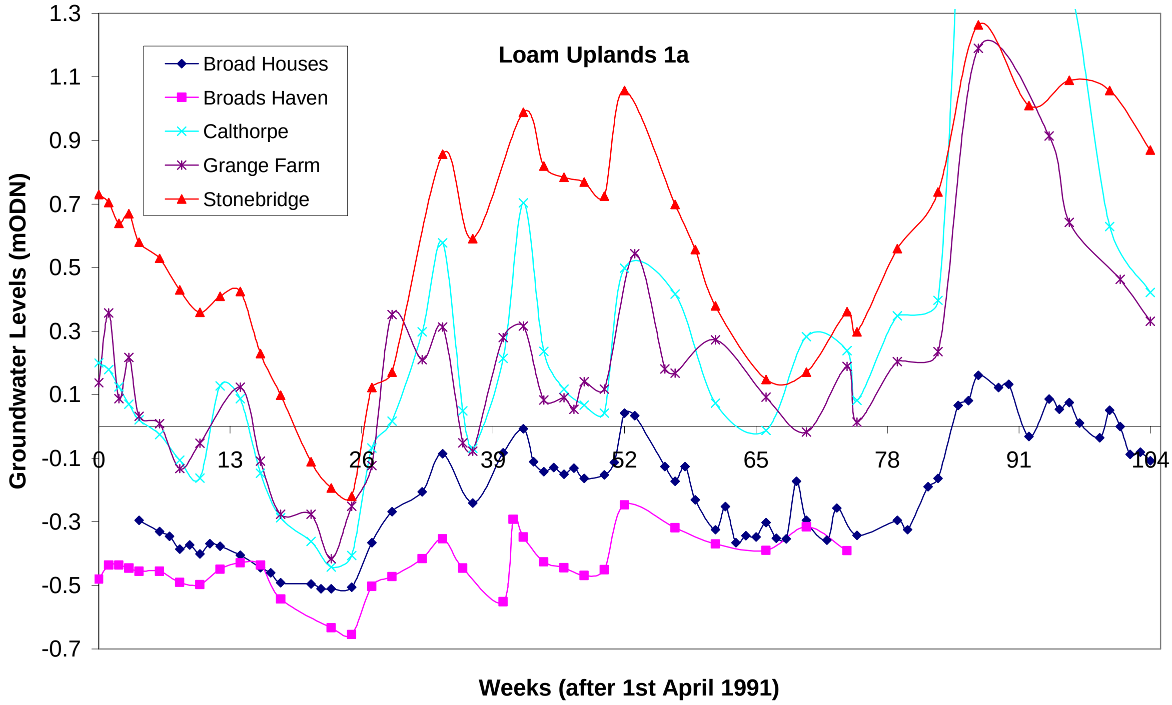

Figure 4.18 The hydrograph of the observation wells grouped as the Loam Uplands 1a

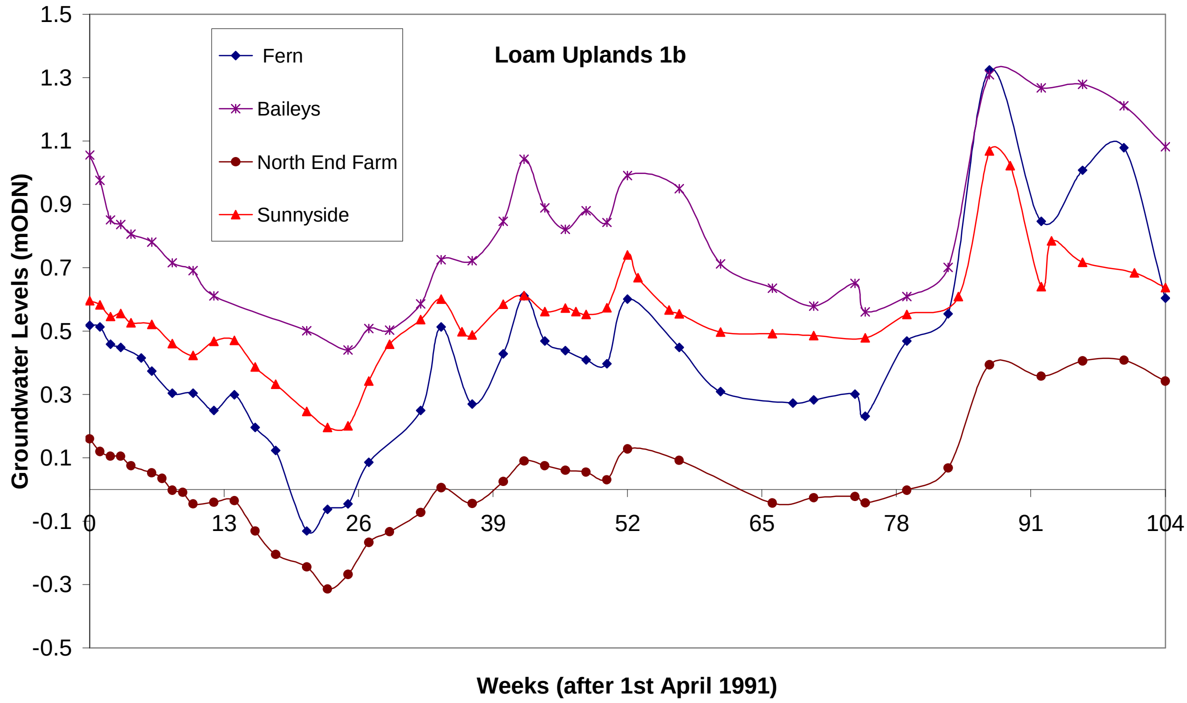

Figure 4.19 The hydrograph of the observation wells grouped as the Loam Uplands 1b

Loam Uplands 1b: The groundwater fluctuations in this group are generally less than in the wells within Loam Uplands 1a (Figure 4.18) although Fern House does show a very rapid recovery. One possible cause could be localized flooding although this is not verifiable. An alternative reason for this type of response could be the influence of significant horizontal flows occurring in areas of unmapped Corton Sands (Section 3.2.3). Notably the wells within this group are further away from the margins of the Norwich Brickearth and alluvium than the wells within the Loam Uplands 2 group.

Loam Uplands 2; the hydrographs of the Brumstead Hall well and Brumstead Hall borehole (Figure 4.20) are of very different levels but are at the same location; the difference in water levels occurs because the Brumstead Hall borehole penetrates into the underlying chalk. The other three wells (Long Acre, Grove and Conifers) show substantial fluctuations with rises in response to high effective rainfall (Section 4.3) that occurred at Week 85 (Figure 4.14).

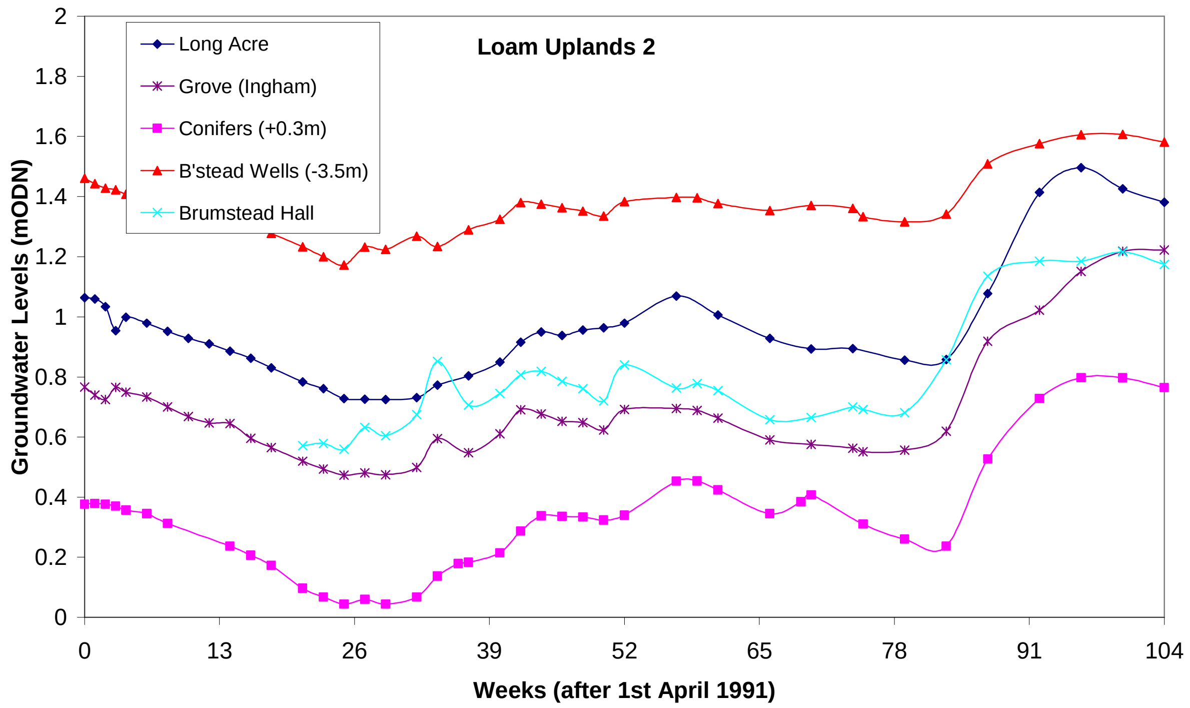

Figure 4.20 The hydrograph of the observation wells grouped as the Loam Uplands 2

The only hydrographical groups to show major responses to the rainfall event of Week 86 are the Marshes (Figure 4.15) and the Holmes (Figures 4.16). The wells located within the Marshes and within the Holmes both respond to the drain-aquifer interactions occurring along the margins of the alluvium–Norwich Brickearth. This is possibly because the soils within unconfined regions of the Holmes reached field capacity so that the aquifer in the local region subsequently received a substantial amount of recharge.

In the *Flegg Hundreds* the combination of the clayey Lowestoft Till and the highly permeable Corton Sands provide the infiltration with multiple routes to the extensively underlying Norwich Brickearth and in turn to the underlying crag (Figure 4.34). This results in the observation wells within the Flegg Hundreds having very similar attenuated profiles with delayed responses to prominent rainfall events (Figure 4.14). For an in-depth study of the quantification of runoff recharge, Bradbury and Rushton (1998) give an informative example by way of the investigation into the South Lincolnshire Limestone.

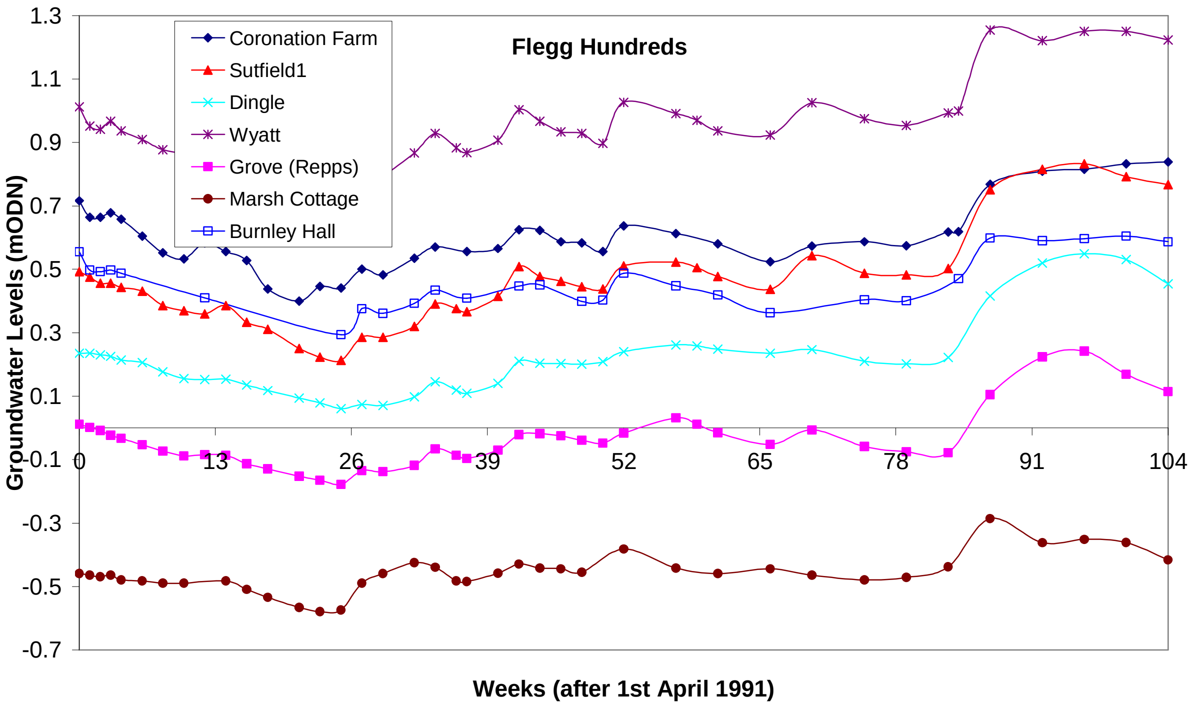

Figure 4.21 The hydrograph of the observation wells grouped as the Flegg Hundreds

4.3.7 Additional Hydrographs

Additional hydrographical data was collected from the observation well at Lower Brograve to reveal the current behaviour of the groundwater beneath the peat. The data was collected from the 8th September 2005 to 4th September 2006. The average water level is a little lower than that collected by Holman (1994) possibly the respective datum are different. The initial rise in groundwater level is caused by the aquifer receiving recharge from the less permeable deposits such as the Norwich Brickearth and then moving laterally through the crag. The recession that occurs from Week 26 through Week 45 represents the drier months of March through June 2006.

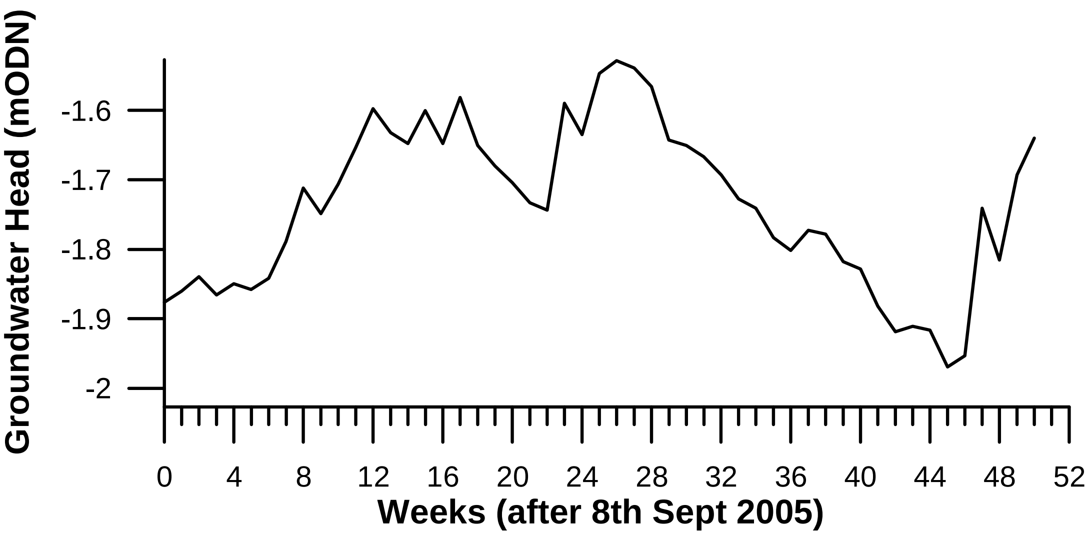

Figure 4.22 The hydrograph of the Lower Brograve observation well fluctuations weekly levels

4.4 Conceptual Model 3-Drain/Aquifer Interactions (DAI)

4.4.1 Introduction

The description of the method of estimating the thickness of the alluvial layer is presented, along with five cross sections of drains that are found within the peat alluvial layer. The cross sections of the various geological units are also presented with possible lines of flow.

4.4.2 Estimating the thickness of the alluvial layer

The thickness of the alluvial deposits significantly affects the upward vertical flows that may enter the various drains. These alluvial deposits of Altcar (peat) and Newchurch (clay) series differ significantly in texture and in thickness and because of the high importance of the Brograve level and marshes which lie immediately to the north, the data available for the peat is far more extensive than that for the clay. Harding (2002) has surveyed the Brograve level for surface elevation and Burton and Hodgson (1987) have surveyed the peat thickness. The elevation and thickness data collected within this research study therefore focussed primarily on the much less surveyed clay, which covers the valley flow of the River Thurne and extends as far east as Winterton village and north to Horsey corner (Figure 4.23).

The mechanical method used to determine the thickness of the clay layer involved the usage of a ¾ inch hollow pipe connected via flexible rubber tubing to a portable pump which in turn is connected to the nearest surface water body. The location that is to be profiled is first augured to approximately 15-25 cm depth to allow the water to be efficiently pumped through the bottom of the pipe. The pipe is held vertically and then water is pumped through it while pounding the leading edge through the alluvial deposit. The depth at which the pipe first encounters significant resistance was deemed as the top of the sands and gravels although there was found to be a thin sandy layer was approximately 10-30cm thick lying on a gravel layer at approximately 4.5-4.9m depth. The gravel hindered any further vertical movement of the metal tubing.

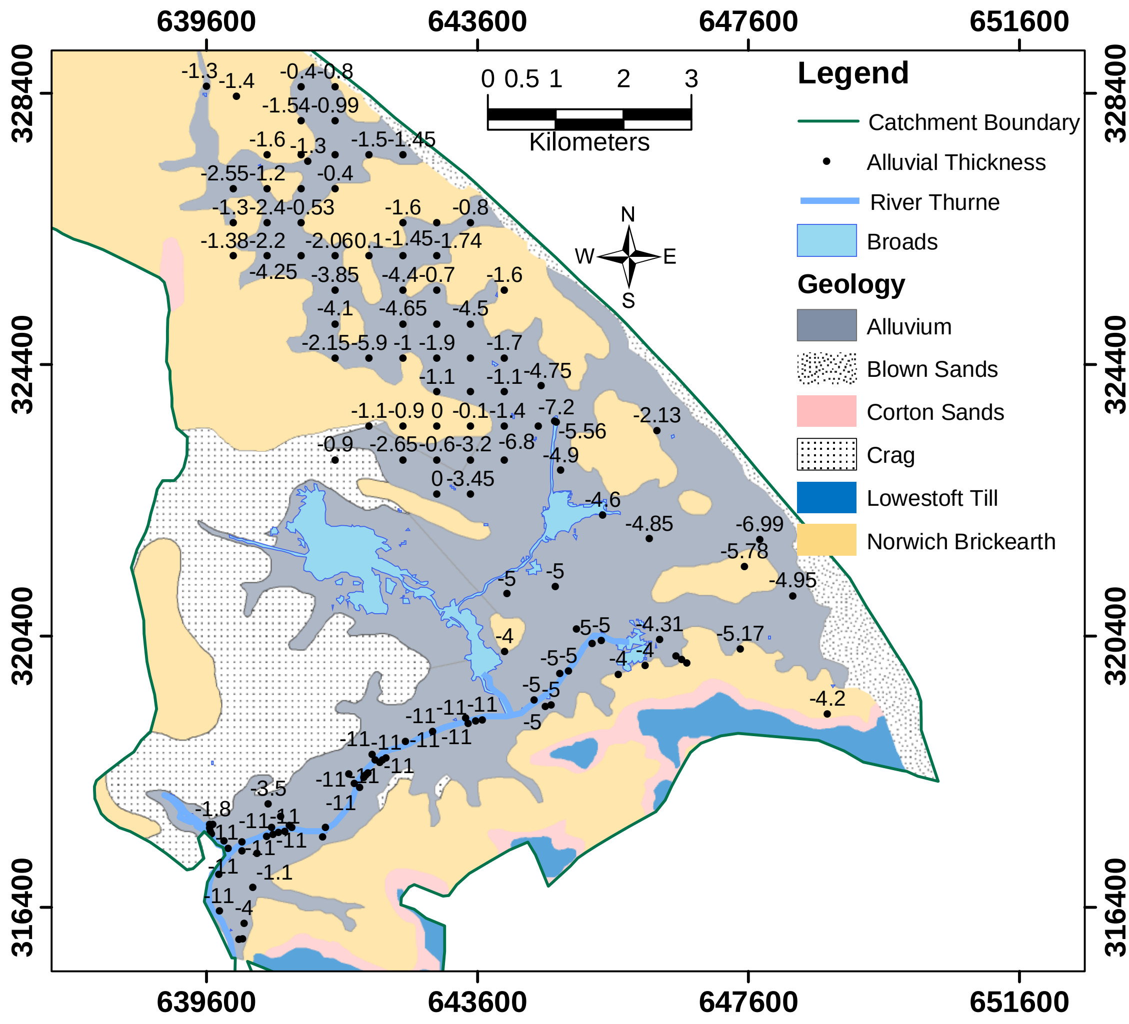

Figure 4.23 Surveyed thickness of the alluvial layer (Edmund Nuttall, Burton 1987, project field work)

Engineers from the engineering company Edmund Nuttall using the Cone Penetration Test (CPT) have measured the depth of the clay along which the River Thurne sits. The method in brief consists of a vertically penetrating cone attached to a resistance-sensitive sensor.

The variation in resistance encountered by the cone is interpreted as being dependent on the variation in the geological material. The results of which, along the bank of the River Thurne, confirm measurements made within this research project, that this thickness is approximately 12 metres, although allowance is made in Figure 4.23 for the 1.2 m height of the banks of the River Thurne.

4.4.3 Drain elevation and aquifer contact considerations

At various locations throughout the catchment, the drain beds have differing elevation and varying degree of contact with the underlying aquifers. In chapter 3, the notion of spring dykes was introduced and their locations along the margins of the Norwich Brickearth emphasized.

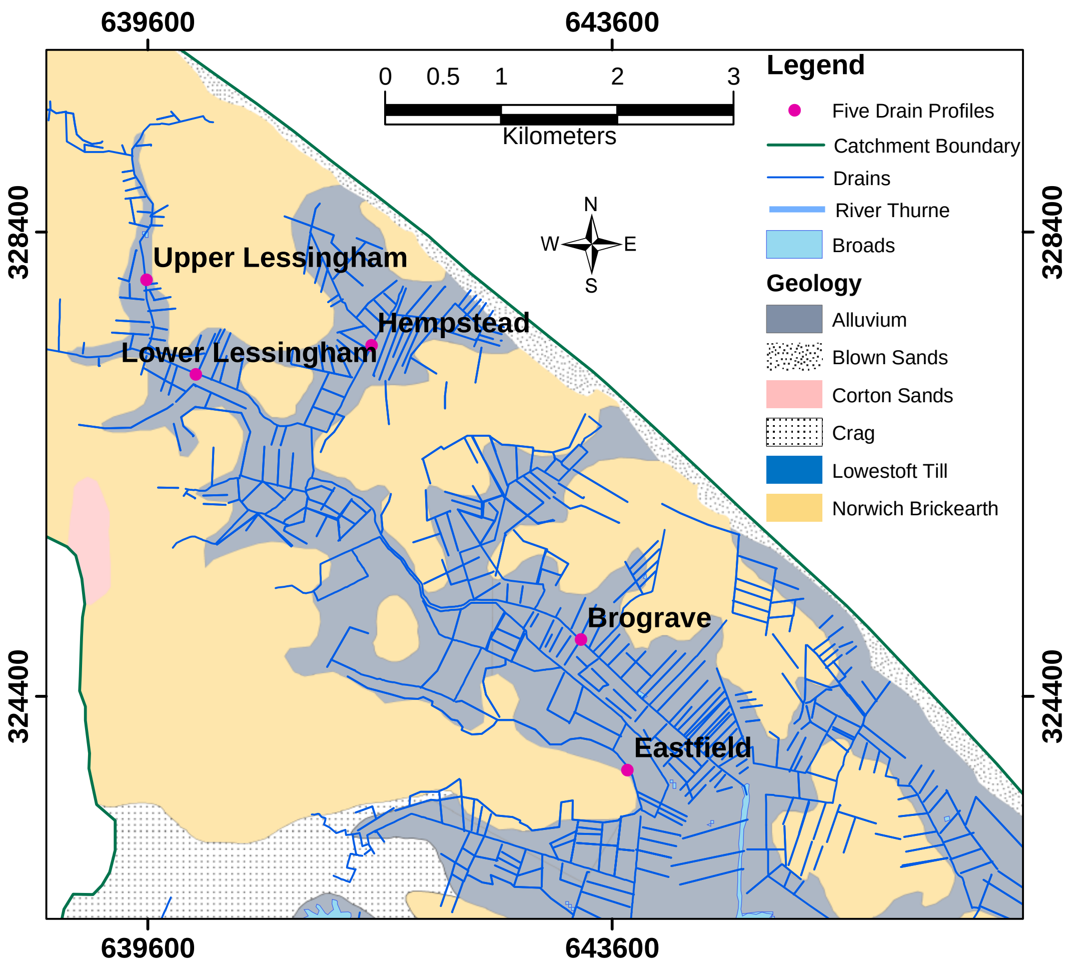

Figure 4.24 The location map for the five drain profiles under discussion

These drains were cut to capture runoff from the much hillier Loam Uplands and in some locations have penetrated to depths that allow water from the aquifer beneath to seep in (e.g. around Eastfield in Figure 4.24). These drains will behave differently to the drains that were cut to remove waters collected by riparian drains, which were cut for the most part in the thickest part of the Altcar peat (e.g. Brograve main drain). There are cases where the main drains have penetrated through the alluvial layer within the catchment.

In the case of the Hempstead drain in the north of the catchment, the purpose of its existence is akin to that of the main drain below the confluence but due to deepening of the base beyond a depth that offered resistance to basal seepage from the aquifer, the drain now not only carries riparian water but also saline groundwater. The Lessingham branch is immediately down gradient from the hillier Loam Uplands and the fresh groundwater contributions at Lessingham are significantly larger than those in the Eastfield sub-catchment to allow permanently fresh water flow. The cross-sectional profiles represented in Figures 4.25-27 represent drain configurations estimated using both drain and ground elevation information from Harding & Smith (2002) and aquifer elevation information from Burton and Hodgson, (1987) and data received from Edmund Nuttall for locations shown in Figure 4.23.

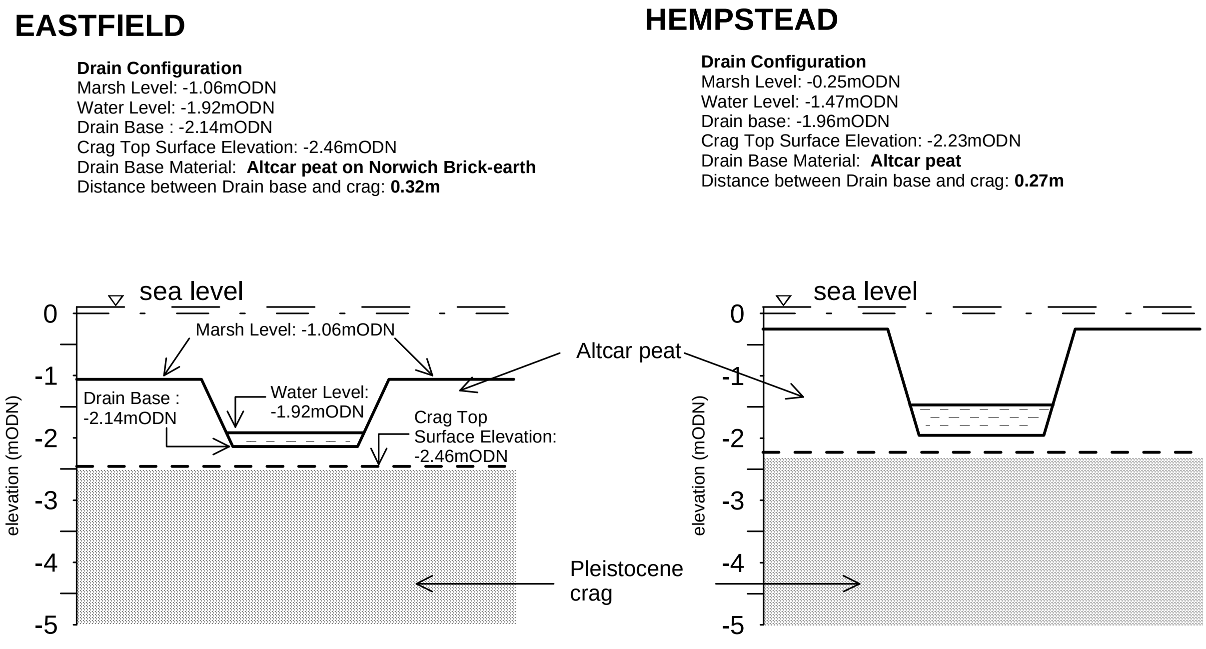

Figure 4.25 The cross-sectional profile of the Eastfield and Hempstead locations

The configurations of drains within the Eastfield and Hempstead areas are presented in Figure 4.25. The water level within the drain, the elevation of the drain base and the marsh level are from Harding & Smith (2002). The marsh level is defined as the surface elevation of the ground immediately next to the measured drain. The elevation of the crag is an *interpolated approximation* of the peat base elevations (Burton and Hodgson, 1987) taken at 500 metre intervals. It is noticeable that the Eastfield and Hempstead cross sections have similar thicknesses of peat below the drain bed of 0.32 m and 0.27 m respectively. These drains both have water levels that are below sea level but have been reported by Holman, (1994) and in this research project to have very different saline proportions in the drain water (Figures 4.4 and 4.7 respectively). One possible reason for this difference in saline proportion is the fresh water contribution to these drains is substantially different. In both cases, the drains are have a basal thickness of less than 0.3 m but in the case of the Eastfield drain the distance from the coast allows for more fresh water and saline groundwater mixing to occur, thus reducing the saline contribution to the composition of the drain water.

The cross sections for drains located at lower and upper Lessingham is presented in Figure 4.26, with the drain elevation of the lower Lessingham drain giving a distance of 2.05m between the base and the surface of the crag. These drains are located down gradient from the nearby Loam Upland and likely to receive substantial amounts of surface runoff and lateral flow through the permeable Norwich Brick earth.

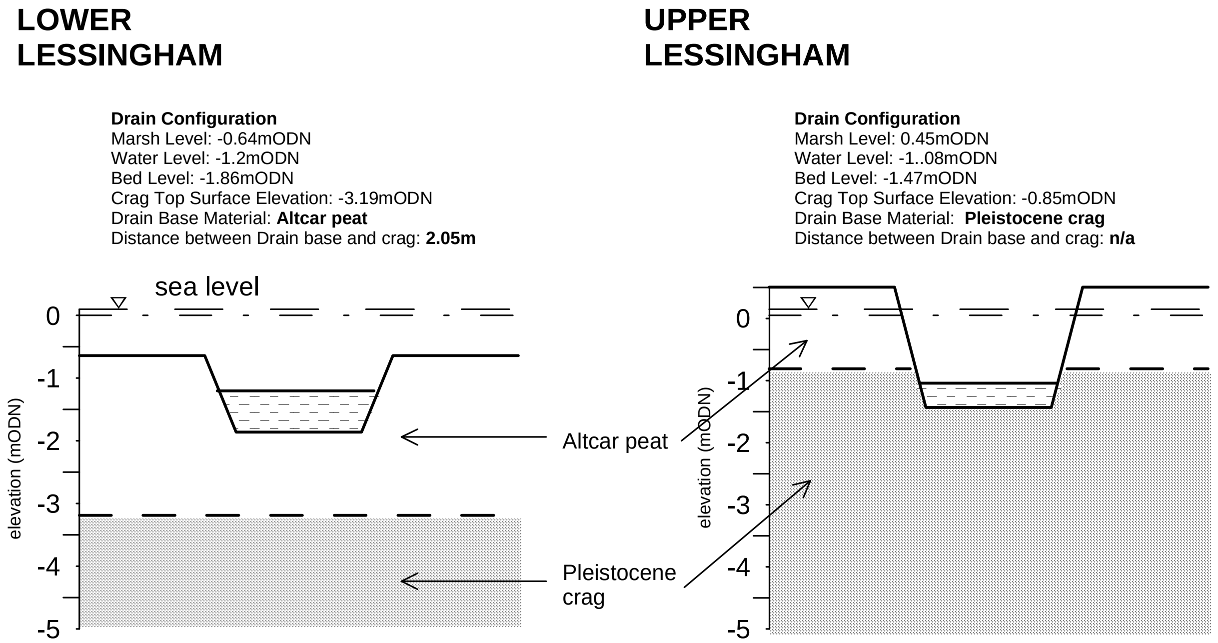

Figure 4.26 The cross-sectional profile of the Upper and Lower Lessingham locations

The influence of the Lower Lessingham drain on the groundwater levels is limited as the thickness of the peat beneath the base of the drain is substantial. In the case of the Upper Lessingham drain the penetration through to the crag is evident. The influence of this drain upon the groundwater heads would be substantial.

The vertical profile of the Brograve main varies along its length, with the thickness of the peat between its base and the surface of the crag varying. In Figure 4.27, the vertical profile of the

Brograve main drain around 2km upstream from the pump is presented. The thickness of the peat between the drain base and the surface of the crag is 1.33 m, which might reduce the magnitude of vertical flows entering the main drain from the underlying aquifer.

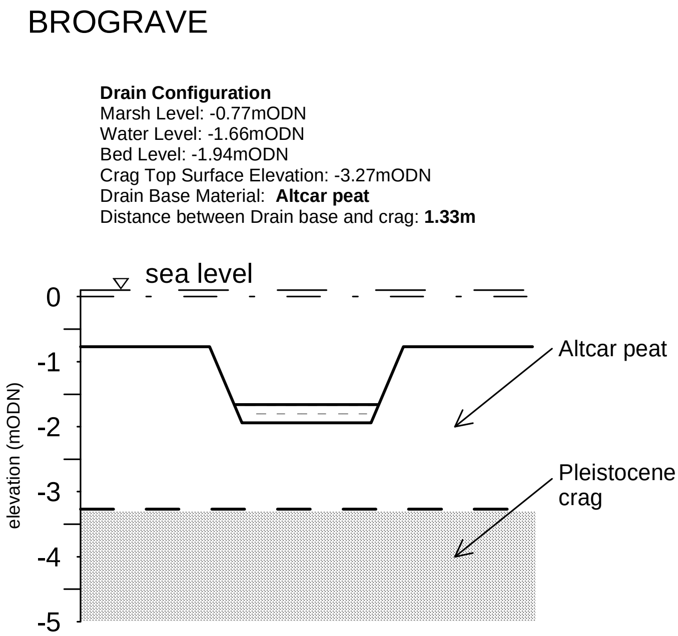

Figure 4.27 The cross-sectional profile of the Brograve drain at the chosen location

The five drain locations previously represented have differing compositions of the four contributing flows shown in Figure 4.28. The contributing proportions to the drain water provided by these various flows are dependent on the geology, geometry and locality of the individual drain. An example of how these contributions vary between drains will easily illustrate how these flows determine the salinity of any sampled drain.

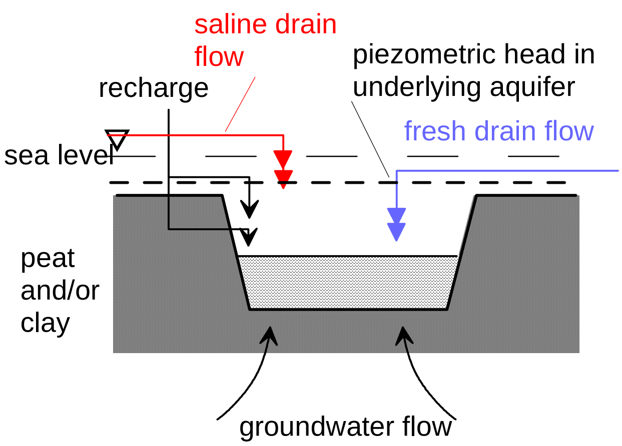

Figure 4.28 A Schematic diagram of saline and fresh water drain/aquifer interactions

The Hempstead drain has been shown to be highly saline in Figure 4.7, the drain basal thickness is approximately only 0.3 m (therefore a high groundwater contribution is likely from the underlying aquifer) and drain locations measured further upstream of the location are also substantially saline. The 'spike' drains that run between the main drain and the more permeable Norwich Brickearth (Figure 4.8) capture rainfall runoff and relatively fresh groundwater contributions, it is likely therefore that the saline contribution to these spike drains is less than the levels found within the actual Hempstead main drain. A qualitative assessment of the typical contributing flows into the five chosen locations is given in Table 4.4.

Table 4.6 A qualitative assessment of the flows entering five different locations

	Recharge	Saline dr. flow	Fresh dr. flow	GW flow	Composition
Upper Lessingham	high	low	high	high	fresh
Lower Lessingham	high	low	high	medium	fresh
Eastfield	medium	medium	medium	high	brackish
Brograve	low	high	high	low	brackish
Hempstead	low	high	medium	high	saline

4.5 Quantified conceptual cross-sections

4.5.1 Introduction

In the previous sections, the fluctuations within the measured groundwater levels and the individual configurations of the various drains were introduced and discussed. In this section the conceptualisation of various cross-sections throughout the catchment are presented in an attempt to combine the two conceptual themes of the deep regional groundwater flow with the drain-aquifer interactions. In Figure 4.28 the locations of the cross sections are presented.

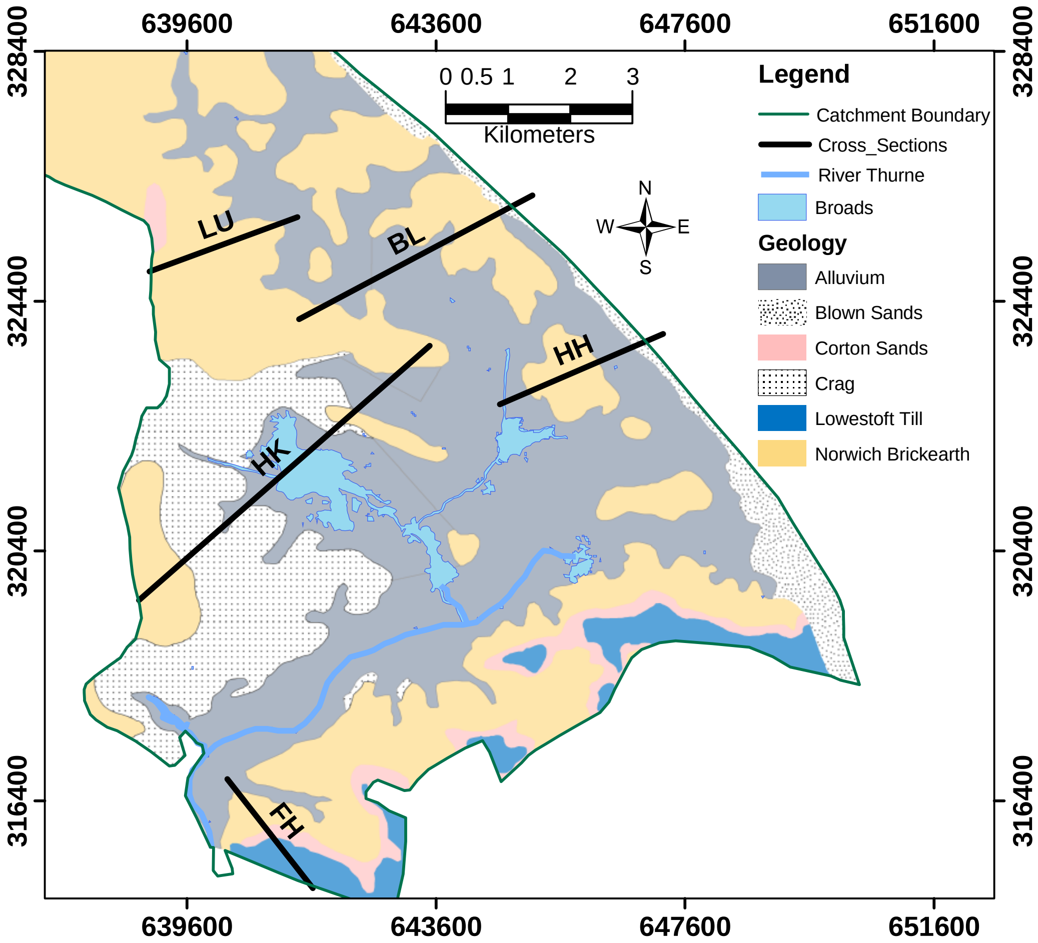

Figure 4.29 The locations of the cross sections

The Eastfield drainage sub-catchment is one of the few non-coastal areas that have coastal saline water within its drains (Figure 4.4). Figure 4.30 is the 'BL' cross section of Figure 4.29 that incorporates the groundwater flow lines that give a possible mechanism as to how the Eastfield drains might be receiving coastal saline water.

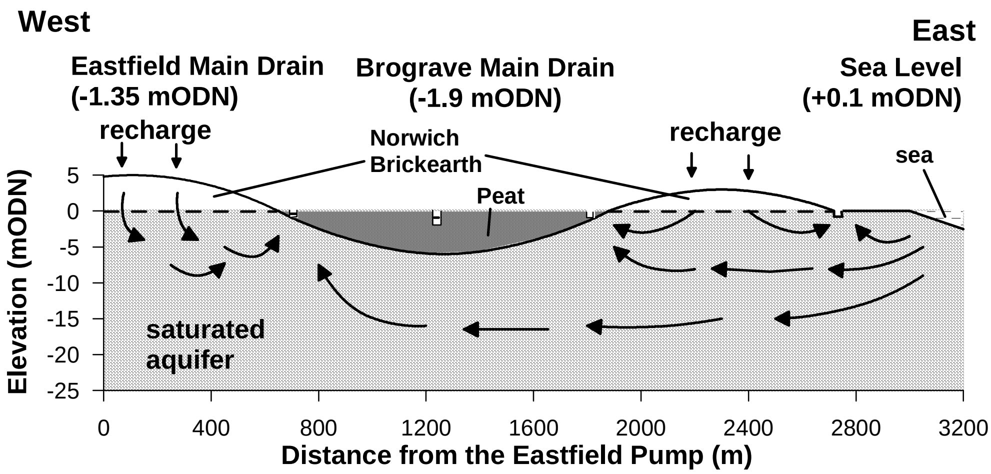

Figure 4.30 The regional groundwater flow through the layered mediums of the peat and upper 25m of the crag

4.5.2 Hickling (HK)

The cross section from the Walton House to Eastfield observation wells is presented below in Figure 4.31. The mean groundwater level deduced from the well at Walton House to the well at Eastfield, are possibly affected by the drains that are cut on the margins of the Norwich Brickearth and the Peat/clay.

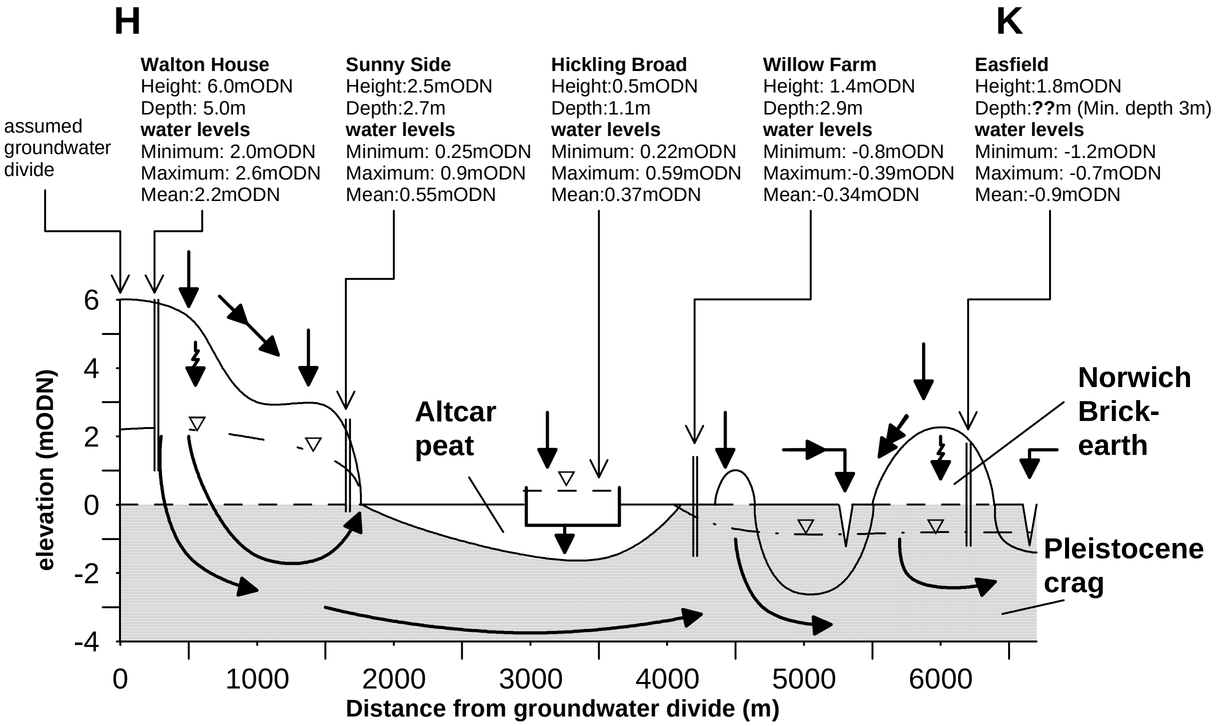

Figure 4.31 The possible flow mechanisms near the Hickling Broad

4.5.3 Horsey Holmes (HH)

The Horsey estate drain that surrounds the Horsey Holme shown in Figure 4.32 has brackish drain water on the landward portion of the circular drainage system (Figure 4.4) and a more saline composition along the seawards portion. Figure 4.32 is a cross section representation of the possible mechanisms that influence the drainage composition of the various sections of the Horsey drainage system.

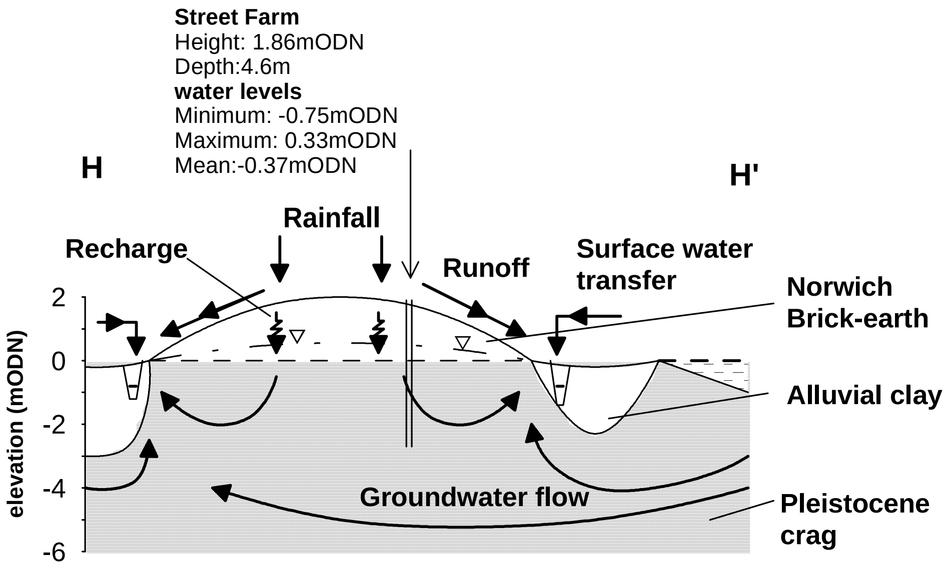

Figure 4.32 The possible flow mechanisms within Horsey Holme

4.5.4 Loam Uplands (LU)

The surface water flows and groundwater flows towards the drains cut on the margins of the peat/clay of the runoff /spring drain are illustrated in Figure 4.33. These spring drains assist in the drainage of the marshes and prevent excess runoff from arriving in the lower lying areas.

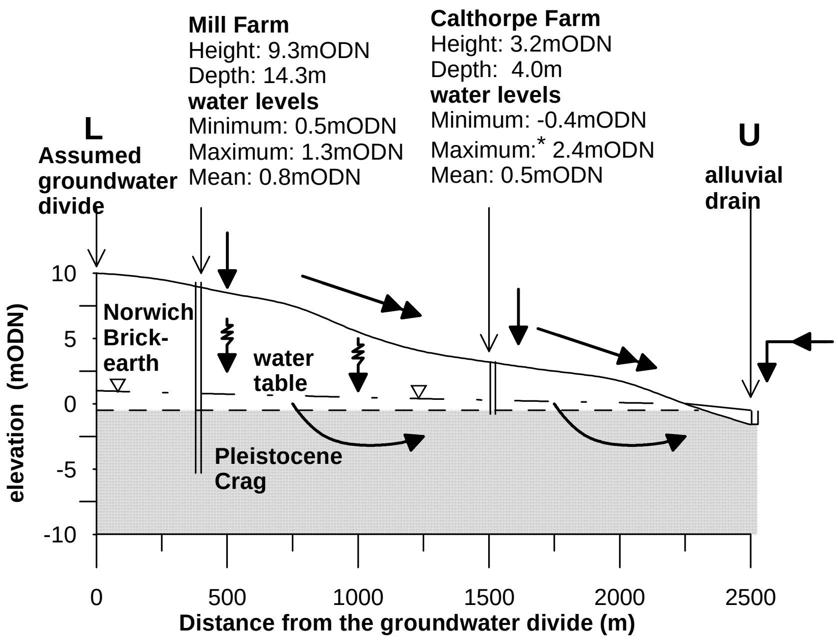

Figure 4.33 The possible flow mechanisms within the Loam Uplands

4.5.5 Flegg Hundreds (FH)

The southern Flegg Hundreds landscape unit (Figure 4.34) is the steepest portion of land within the Upper Thurne catchment. The existence of a more permeable stratum within the Flegg Hundreds (Corton Sands) allows for the construction of wells such as the Moregrove observation well to extract water from this perched aquifer.

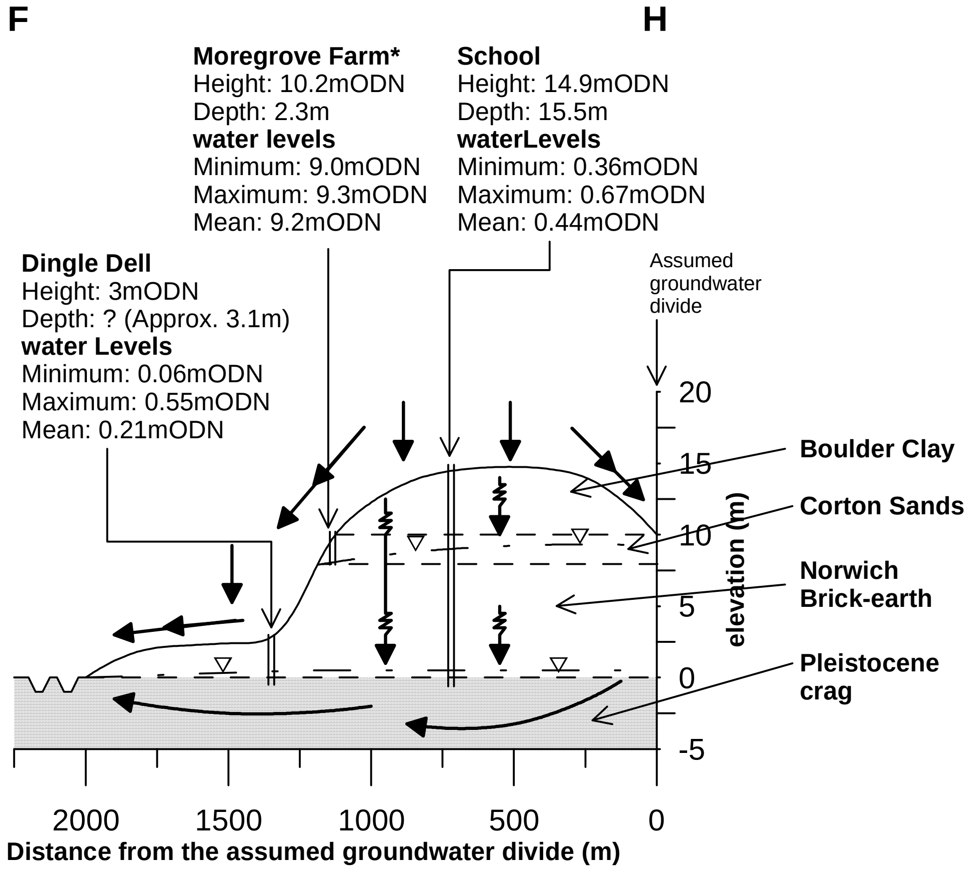

Figure 4.34 The possible flow mechanisms within the Flegg Hundreds

4.6 Summary

The introduction of the three types of drainage mechanisms (e.g. undersoil pipes, riparian drains and conduit drains) used to drain the marshes is presented. Three conceptual themes are also identified within this chapter, Surface Water Flow along the drains (SWF), Deep Regional Groundwater Flow (DRGF) and Drain/Aquifer Interactions (DAI). The combination of the DRGF and DAI conceptual themes in the penultimate section of the chapter is intended to give insights into the complex nature of the groundwater flow.

The distribution of salinities within the Lessingham and Hempstead branches of the main drain appear to dominate the composition of the water along the main drain up to the pump at Brograve. Inflows contributed by the feeder drains along the lower reaches of the main drain seem not to affect the surface water salinity.

The grouping and analysis of the fluctuations within the observation well water levels measured by Holman (1994) give an indication of how various areas of the underlying aquifer respond to prominent rainfall events. Three different types of responses are recognised, namely (i) the confined response beneath the alluvial layer (Marshes) (ii) The unconfined direct response (Loam Uplands 1a, 1b, and Holmes) and (iii) the unconfined attenuated response (Loam Uplands 2 and Flegg Hundreds).

5 AN INTRODUCTION TO THE NUMERICAL MODELLING OF GROUNDWATER FLOW

5.1 **Introduction**

In chapter 4, three separate conceptual themes that cover the surface water and groundwater systems, namely; Surface Water Flow Along the drains (SWF), Deep Regional Groundwater Flow (DRGF) and Drain/Aquifer Interactions (DAI), were developed along with the production of quantified conceptual cross-sections. However, in order to adequately understand the *groundwater system* and thereby appropriately implement the concepts developed within the DRGF and DAI into a numerical groundwater flow model, the techniques and problems associated with groundwater flow modelling need to be discussed. Therefore in this chapter, the first chapter in Part B (which provides a background to numerical modelling and model preparation), the governing equations of groundwater flow are reviewed and the numerical methods for solving partial differential equations are investigated.

The introduction of idealisations of a modelling situation is often used as a means to obtain a useful solution to a complex problem. The idealisations in this case might include simplification of layers within a layered aquifer or applying a dual density approach or a single density approach with particle tracking to a variable density situation. The single density equation and numerical techniques are introduced, and the advantages and disadvantages of the numerical techniques of the Finite-Difference Method (FDM) and Finite Element Method (FEM) are reviewed. The choice of a well-tested appropriate software package to simulate the groundwater behaviour completes the chapter.

5.2 A background to groundwater flow

5.2.1 The physical problem to be solved

The physical problem, which is to be solved within the Upper Thurne, is the case of a coastal aquifer that is exposed to coastal saline inflow from the eastward coastline boundary. Superficial deposits that have very different hydraulic conductivities, which affect the rate of vertical recharge, cover most of the aquifer. Those superficial deposits that have been identified as having particularly low hydraulic conductivities are collectively known as the alluvial deposits. They cover the lower-lying land within the catchment and as such receive both direct (rainfall) and indirect (runoff) fresh waters. The rainfall rates for the catchment are such that the drainage capabilities of these alluvial deposits are alone insufficient to remove excess surface waters. Therefore, a system of dendritic open-ditch drains has been cut into the upper 1-2 metres of alluvial deposits to assist with the drainage. These open ditch drains, however, have in some cases penetrated through to the underlying aquifer and thus there is a direct continuum of saline inflow from the sea into specific surface water locations. The problem is compounded by the land being several metres below sea level thus providing a hydraulic gradient for seawater to enter into the underlying aquifer. The problem therefore is determining the mechanisms that influence the quality of water from various fresh or coastal (saline) sources that enter into the drainage systems. In order to determine these mechanisms the problem of simulating variable density flows need to be investigated.

5.2.2 The problem of modelling variable-density flows

The problem of seawater intrusion in coastal aquifers can be formulated in terms of two partial differential equations (Huyakorn et al. 1987). The first equation (5.1) is used to describe the flow of the variable–density fluid (mixture of fresh water and seawater), and the second equation (5.2) is used to describe the transport of dissolved salt. The three-dimensional flow equation may be written as

$$\frac{\partial}{\partial x_i}\left[K_{ij}\left(\frac{\partial h}{\partial x_j}+\eta c e_j\right)\right]=S_s\frac{\partial h}{\partial t}+\phi\eta\frac{\partial c}{\partial t}-\frac{\rho}{\rho_0}q \qquad (5.1)$$

i,j =1,2,3

where K_{ij} is the hydraulic conductivity tensor [LT^{-1}], *h* is the reference hydraulic head referred to as the freshwater head [L], x_j (j = 1, 2, 3) are the Cartesian coordinates, η is the density coupling coefficient, *c* is the solute concentration [L^3/L^3], e_j is the *j*th component of the gravitational unit vector. S_s is the specific storage [L^{-1}], *t* is time [T], ϕ is porosity [L^3/L^3], q is the volumetric flow rate of sources or sinks per unit volume of the porous medium [T^{-1}], and ρ is the density of the mixed fluid [ML^{-3}] (fresh water and seawater) and the ρ_0 is the density of the reference (freshwater) density. To describe salt transport, the following form of the *advection-dispersion* equation is used:

$$\frac{\partial}{\partial x_i}\left(D_{ij}\frac{\partial c}{\partial x_j}\right)-V_i\frac{\partial c}{\partial x_i}=\phi\frac{\partial c}{\partial t}+q(c-c^*) \qquad (5.2)$$

where $D_{ij} = \phi \widetilde{D}_{ij}$, with $\widetilde{D}_{ij}$ being the dispersion tensor defined by (Bear, 1979, p234), V_i is the Darcy velocity vector [LT^{-1}], and c* is the solute concentration in the injected (or withdrawn) fluid [ML^{-3}]. An analytical solution to the equation 4.2 is not possible and so numerical solutions are required. The governing flow equations 5.1 & 5.2 are both difficult to solve numerically and cause discrete solutions to be affected by a phenomenon known as *numerical dispersion*, (pp 327, Anderson and Woessner, 1992) which refers to the artificial dispersion caused by errors associated with the spatio-temporal discretization of the model domain. An alternative approach to solve the variable-density flow problem is to approximate the modelling domain by use of the initial concentrations, initial pressure and density distributions (pp 446, Zheng and Bennett, 2002) to approximate the advection-dispersion equation.

Modelling advection-dispersion transport through a porous medium involves (a) flow simulation for each time step, (b) a transport simulation for the said time step, (c) a check for pressure and concentration convergence, and (d) a check to see if the density distribution is consistent with the pressure and concentration distributions. In Figure 5.1 the solution sequence in an *implicit* (an implicit approach is one where all the unknowns must be obtained by means of a simultaneous solution of the modelling domain), procedure for variable–density flow and transport modelling is given in the form of a flowchart.

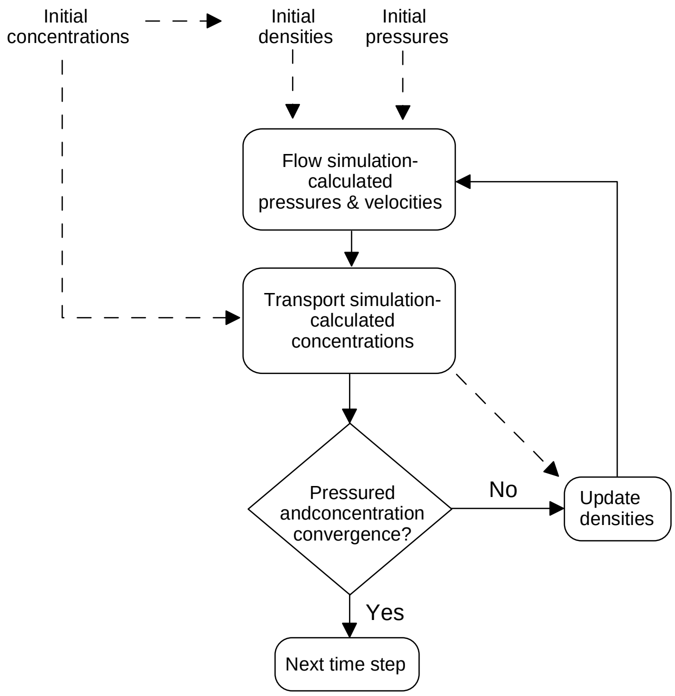

Figure 5.1 The solution sequence in an *implicit* procedure for variable–density flow and transport modelling (p446, Zheng & Bennett, 2002)

It is important to note that the three variables, namely the concentrations, densities and pressures must render values simultaneously which allow the advection-dispersion transport equation to be solved numerically. This level of complexity is impractical for such a complicated study area as the Thurne catchment. Therefore, to avoid such numerical problems a number of investigators (Salmon, 1989; Rushton and Tomlinson, 1977) prefer to idealise the modelling situation. These idealisations can take one of two forms; (i) dual density or (ii) single density with particle tracking. The following sub-sections address both of these approaches.

5.2.3 The use of dual-density hydrostatic theory to model a hydrodynamic environment

The use of dual-density hydrostatic theory to model a hydrodynamic environment within coastal aquifers was first investigated using a simple model developed independently by Badon Ghyben in 1888 and by Herzberg in 1901. Their model (known as the Ghyben-Herzberg model) is based on the *hydrostatic* balance between fresh and saline water in a coastal aquifer. Figure 4.2(a) shows the coastal situation where h_f is the elevation of the water table above sea and z is the depth of the fresh-saline interface below sea level.

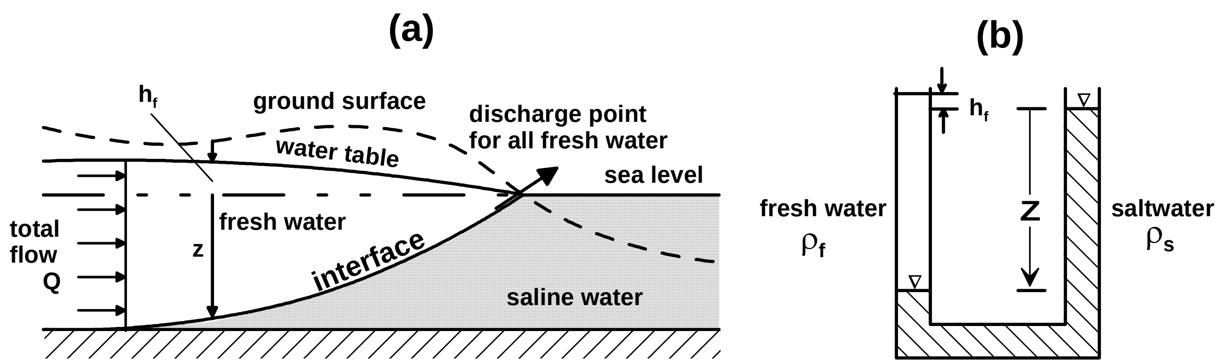

Figure 5.2 The hydrostatic balance between fresh water and saline water (a) at the coast (b) in a U-tube

The hydrostatic balance between fresh water and saline water is illustrated by the U-tube shown in Figure 5.2 (b)

$$\rho_s g z = \rho_f g(z + h_f) \tag{5.3}$$

where ρ_f = density of fresh water [ML^{-3}], ρ_s = density of salt water [ML^{-3}],

h_f = head of the fresh water above level [L], z = depth to the interface [L],

They showed that the depth below sea level to saltwater, z, in an unconfined, homogenous isotropic aquifer is

$$z = [\rho_s / (\rho_s - \rho_f)] \times h_f \tag{5.4}$$

The typical densities of freshwater and seawater are 1.000 g/cm^3 and 1.025 g/cm^3 respectively and if substituted into the above Equation 4.4 we find that the hydrostatic balance between fresh water and saline water is found below sea level at a depth of forty times the head of the fresh water above sea level.

$$z = 40 \times h_f \tag{5.5}$$

which is referred to as Ghyben-Herzberg approximation or relationship. However, the assumptions that form the basis for this relationship are *primarily* hydrostatic assumptions that generally do not necessarily apply to hydrodynamics. The hydrostatic assumptions are:

i. Vertical flows are assumed negligible, also known as the Dupuit assumption.(Bear and Dagan, 1964).
ii. The aquifer is assumed homogeneous. (Chandler and McWhorter, 1975)
iii. The dispersion effects within the saline water are ignored. (Bear, 1972).
iv. The interface is assumed thin in comparison to the thickness of the fresh water layer. (Volker and Rushton, 1982)

The flow regime immediately below the open ditch drains (Figure 3.10) suggest that condition vertical flows are considerable and essential to the groundwater movement in the Thurne and thus (i) is not adhered to. Within the upper layer (< 25m) the aquifer was found to have thin (< 0.5m) of sand above thicker layers of gravels thus making the crag heterogeneous (Rose and Allen, 1977). A more in-depth discussion on the limitations of applying the Ghyben–Herzberg approximation is given by Bear and Dagan, (1964).

5.2.4 The use of single density with particle tracking

A pragmatic approach to simulating the effects of fresh water recharge and the saline inflow at the coast within the groundwater system is to consider a convection-only solution based on mass balance and Darcy's law. The United State Geological Survey (USGS) MOC model (Prickett et al. 1981) combines particle tracking for convection with a finite-difference solution of the dispersion portion of the problem. MT3D (Zheng, 1990) is a particle tracking code with dispersion that is compatible with the finite-difference software MODFLOW (McDonald and Harbaugh, 1988).

In modelling convection only through a layered aquifer (see Figure 3.4), the problems associated with the variable density approach are avoided. The difference in density between fresh water and seawater of 0.025 g/cm^3 may not be important at a distance from the coast where significant mixing produces large volumes of intermediate density brackish groundwater. Typically, over hundreds of years, mixing will have occurred with the distribution of fresh and saline waters being highly dependent on the recharge pattern of the last few hundred years. This level of detail of historical information is not available thus making the initial conditions for a variable density simulation difficult to identify. The mass movement of water from the coastal region can be simulated using a single density approach and the aid of particle tracking. (Prickett et al. 1981) When investigating the mass movement of groundwater within an aquifer, important aspects of convection transport, such as the groundwater flow equation and storage effects, need to be introduced and discussed.

5.3 **Considering single density transport only (convection)**

5.3.1 Three-dimensional time variant groundwater flow for convection transport

The simple derivation of the three-dimensional groundwater flow equation can be found by considering Darcy's Law in the direction of the principal axes (x,y,z,) :

$$v_x = \frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) \qquad v_y = \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) \qquad v_z = \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) \tag{5.6}$$

and by considering the conservation of mass in a three dimensional element (Figure 5.3)

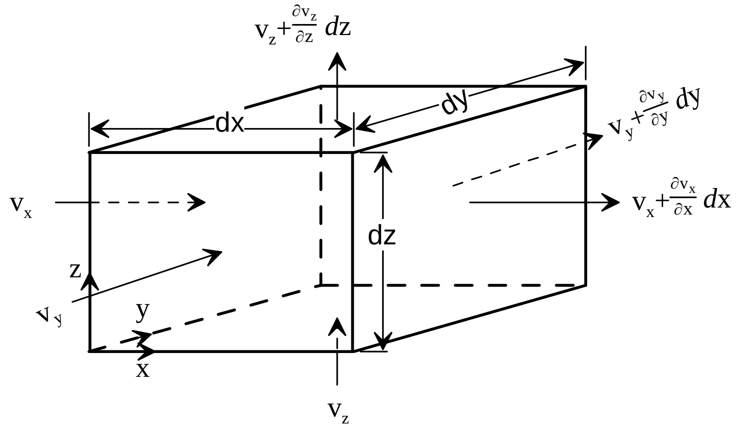

Figure 5.3 A unit three-dimensional element

The flow entering in the *x* direction in time $\delta t = v_x dydz\delta t$ and the flow leaving in the *x* direction in time $\delta t = v_x + \frac{\partial v_x}{\partial x} dydz\delta t$ it can be deduced that the net flow entering in time δ t in the x direction $= -(\partial v_x / \partial x)dxdydz\delta t$ and the net flow entering in time δ t in the y direction $= -(\partial v_y / \partial y)dxdydz\delta t$. Likewise the net flow entering in time δ t in the z direction

$= -(\partial v_z / \partial z) dxdydz\delta t$ and likewise the net flow entering in time δ t due to increase in head $\delta h = -\delta h S_S \ dxdydz$. These flows must sum to the zero; in the limit, δ h/δ t tends to ∂ h/∂ t,

$$-\frac{\partial v_x}{\partial x} - \frac{\partial v_y}{\partial y} - \frac{\partial v_z}{\partial z} = S_S \frac{\partial h}{\partial t} \tag{5.7}$$

thus combining equations 4.6 and 4.7

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_S(x,y,z)\frac{\partial h}{\partial t} \tag{5.8}$$

This is the governing equation for three-dimensional flow as defined by Jacob (1950) and is represented in Figure 4.4 (a) where K_x, K_y, K_z are values of hydraulic conductivity [LT^{-1}] along the x, y and z co-ordinate axes. S_S the specific storage (volume of water released from a unit volume of aquifer for a unit fall in head) of the porous material and is described fully in section 4.2.3, h the piezometric head [L], and t is the time [T]. At the top of the element, two physical processes provide the vertical flux, the recharge –q and the flux component due to the specific yield S_Y $\partial h/\partial t$. Therefore flux difference in the z direction (vertically upwards is positive) is $S_Y \frac{\partial h}{\partial t} - q$. Equation (5.8) is known as the three dimensional time-variant differential equation that describes groundwater flow.

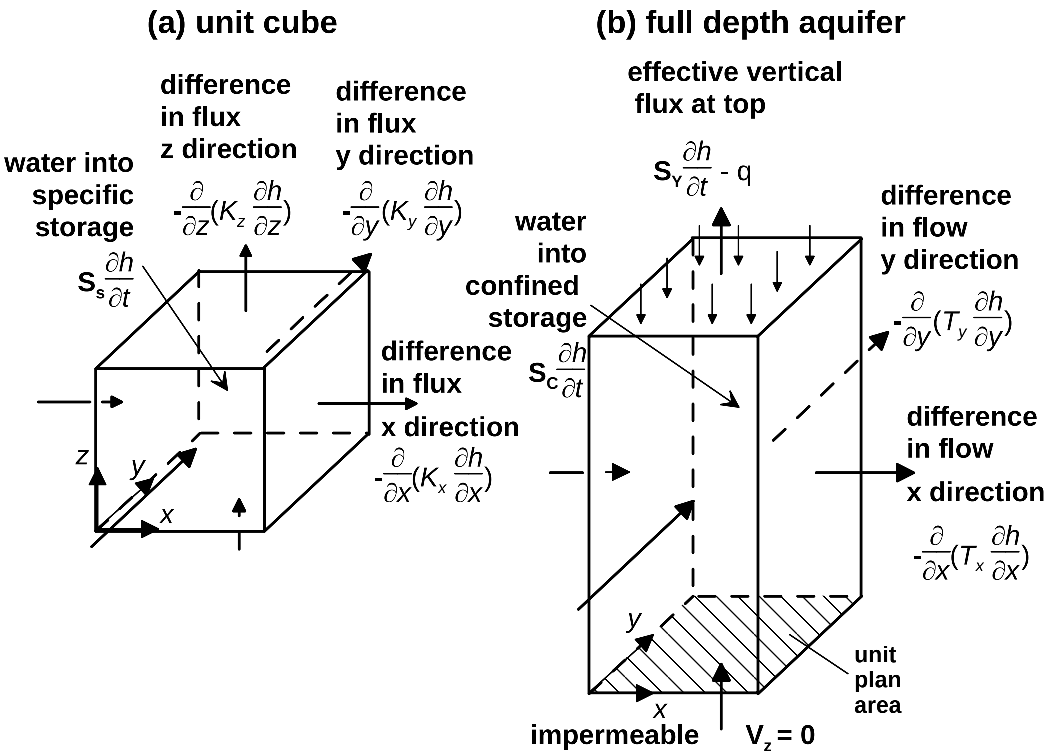

Figure 5.4 A diagram illustrating how vertical flow components are implicitly contained in the two-dimensional regional groundwater flow formation (Rushton, 2003)

However, in most practical situations, there is no need to work with the unit cube in three dimensions plus the time dimension, but instead the more workable Equation 5.9, which represents the groundwater flow in a full depth aquifer.

$$\frac{\partial}{\partial x}\left(T_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_y \frac{\partial h}{\partial y}\right) = S_C \frac{\partial h}{\partial t} + S_Y \frac{\partial h}{\partial t} - q \tag{5.9}$$

This represented in Figure 4.4 (b), S_C is the confined storage and S_Y is the specific yield and are both dimensionless (see sub-section 4.3.2) and T_x and T_y are the transmissivities in the x and y direction and defined as the product of the directional hydraulic conductivity and the unit perpendicular saturated thickness. This equation is known as the Boussinesq equation and Connorton, (1985) derived the proof that it includes the necessary vertical flows (i.e. flows in the z direction) implicitly. There are no known analytical solutions to the three-dimensional time variant problem and only on the condition of axial symmetry are analytic solutions available (Rushton and Redshaw, 1979). Only if the dominant effect on the flow is the pumping from a borehole does the condition of axial symmetry hold; it follows then that alternative, namely numerical, solutions usually have to be employed. One approach to solve the case of a layered aquifer which has layers *i*, *j*, and *k*, such as the crag, is the approach (see sub-section 4.4.2) used in MODFLOW (McDonald and Harbaugh, 1988) and shown in Figure 4.5.

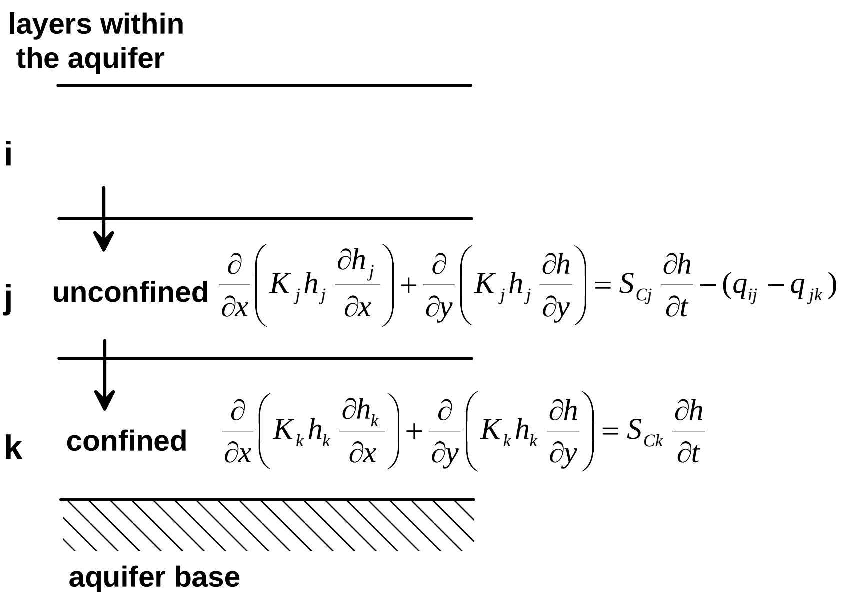

Figure 5.5. The application of the groundwater equation to a layered aquifer (McDonald and Harbaugh, 1988)

5.3.2 An introduction to storage coefficients

In the representation of groundwater flow, the notion of water being released or taken into storage is an importance concept in the dynamics of the groundwater system and the behaviour of the water table. An important property of the aquifer is its ability to store (or release groundwater) and is known as the aquifer's storage coefficient. There are three variations of the storage coefficients namely the Specific Yield, Specific Storage and Confined Storage (Figure 5.6).

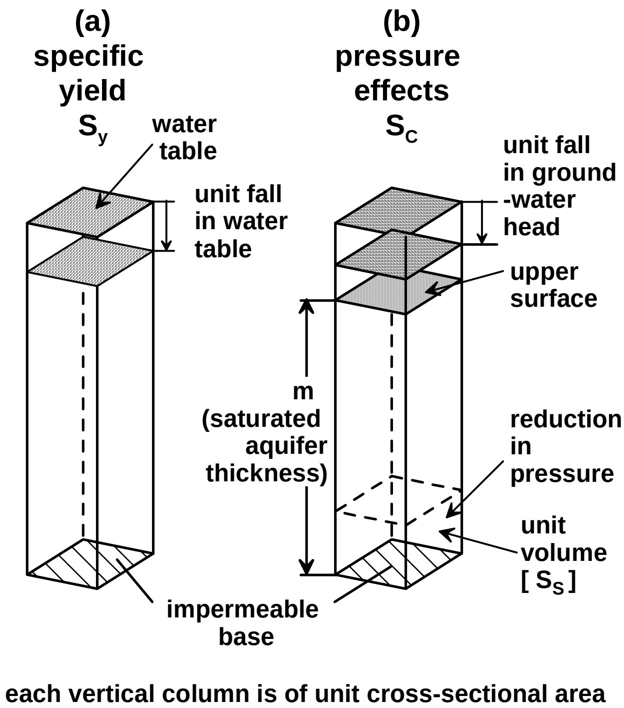

Figure 5.6 A diagrammatic explanation of the three storage coefficients (Rushton, 2003)

The Specific yield (the volume of water drained from a unit plan area of aquifer for a unit fall in head) is $\mathbf{S_Y}$ and is dimensionless and has a typical value of between 10^{-3} and 0.25 depending on the constituent material (p13, Rushton, 2003). Whereas the Specific storage ($\mathbf{S_S}$) is the volume of water released from a unit volume of aquifer for a unit falling head and has dimensions $[L]^{-1}$ with values of between 10^{-6} m^{-1} and $10^{-4}m^{-1}$ (p13, Rushton, 2003). The confined storage ($\mathbf{S_C}$) is the volume of water released per unit plan area of aquifer for a unit fall in head and is dimensionless with typical values of between 10^{-5} and 10^{-3} (p13, Rushton, 2003).

5.3.3 Numerical solutions to partial differential equations

There are two major numerical techniques for solving the partial differential equations that describe fluid flow through any medium, namely the Finite-Difference Method (FDM) and Finite-Element Method (FEM) (Gray, 1984). Several workers such as (Faust and Mercer, 1980a; (Wang and Anderson, 1982; Spitz and Moreno, 1996) have reported experiences with both methods in application to groundwater flow and have concluded that each method has its distinct advantages and disadvantages.

The FDM is the older of the two methods and is very simple to implement and understand because it requires approximating Equation 4.8 with a *discrete* difference equation. The resulting difference equation is solved for each nodal point along the modelling domain boundary to form a system of algebraic equations that must be solved at each internal node and for each time-step. There are two variants to the FDM, namely Block-centered and Nodal centered, but both systems start by dividing the aquifer with two sets of parallel lines that are orthogonal as shown in Figure 5.7.

In the block-centered formulation, the blocks formed by the sets of parallel lines are the cells; the nodes are at the centre of the cells Figure 5.7(a). Whereas in the nodal-centered formulation, the nodes are at the intersection points of the sets of parallel lines, and cells are drawn around the nodes with faces halfway between nodes (Figure 5.7(b)). In either case, spacing of nodes should be chosen so that the hydraulic properties of the system are generally uniform over the extent of a cell.

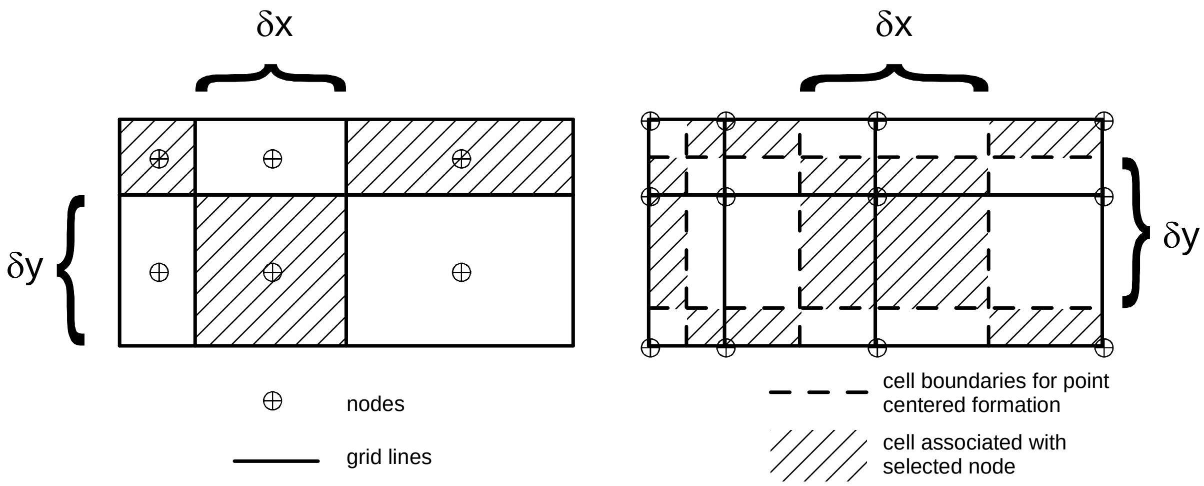

Figure 5.7(a) Block-Centered Grid system (b) Node-Centered Grid System

(McDonald & Harbaugh, 1988)

The Finite Element Method (FEM) approach is more sophisticated mathematically, more flexible in structure and thus is typically used where the flow near an irregular boundary requires simulation (p21, Anderson and Woessner, 1992; p202, Zheng and Bennett, 2002; Townley and Wilson, 1980). In this method, three nodes, one at each corner, define the typically triangular elements. These nodes are the points at which the heads within the problem domain are computed, whereas the head *within* each element is defined according to an interpolation function incorporating the values of these three nodal points. The computed heads at the nodal points are evaluated by applying a weighted residual principle expressed directly in terms of Equation 5.8 (Bear, 1979).

Since its inception into groundwater modelling the FEM codes have been used to simulate variable density flows e.g. SUTRA (Voss, 1984), The Princeton Transport Code (PTC) (Babu and Pinder, 1984; Huyakorn et al. 1987). However, while the early motivation for applying the FEM was to overcome the numerical difficulties associated with the finite–difference solution of the advection-dispersion equation (Huyakorn and Pinder, 1983), the finite element method itself is subject to these same numerical problems (p202, Zheng and Bennett, 2002) limiting its widespread use.

The example shown below is of the Nile delta of Egypt, North Africa (Townley and Wilson, 1980) where the irregular structure of the aquifer is discretised efficiently with the use of triangular elements. The coarse discretisation in the northern regions of the catchment reflects the expected low variations in speed and direction of flows, whereas the finer mesh in the southern tip of the catchment reflects a more complex flow regime.

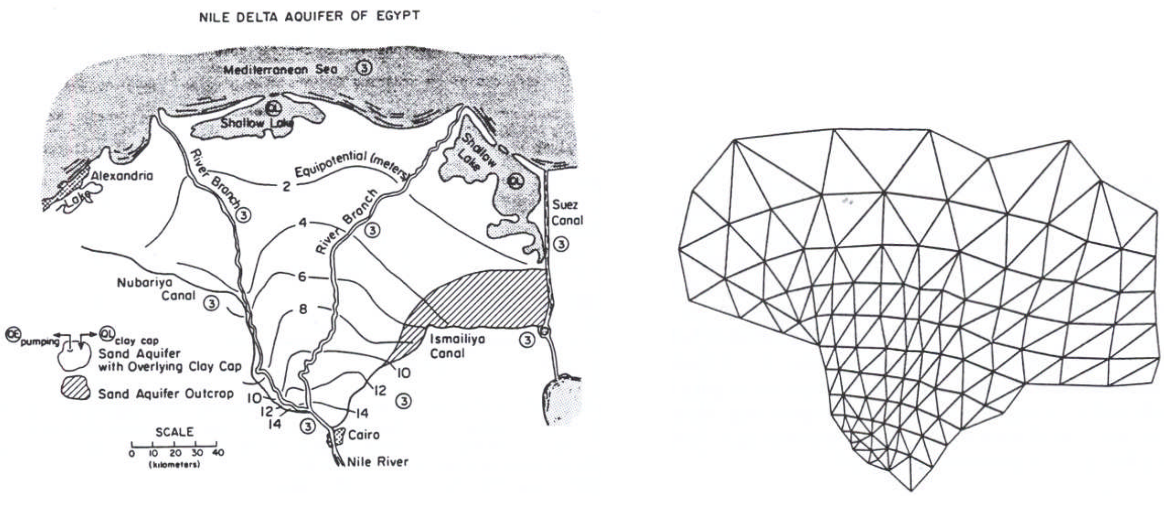

Figure 5.8 (a) Map of the Nile delta (b) Triangular elements are shaped to fit the irregular boundary (Townley and Wilson, 1980)

5.3.4 The choice of an appropriate numerical method for the Upper Thurne

In the case where the modelling domain is complicated with a network of drains/rivers and the aquifer is both confined and unconfined in various locations, the chosen method must be able to reliably incorporate this complexity (Anderson and Woessner, 1992; (Kolm, 1993; (San Juan and Kolm, 1996). Although within the FEM the elements can be made flexible to adjust to the geometry of the problem, the necessary computational models that are developed within the intermediary stage of the modelling process would be beyond the scope of this study. This is because the mathematical method of approximation for the nodal head values uses a weighting function which is derived using Galerkin's method (Wang and Anderson, 1982; p198, Zheng and Bennett, 2002) which requires detailed mathematical manipulations. One serious difficulty with FEM is the node numbering; (p67 Anderson and Woessner, 1992) suggest a method for numbering the mesh given in Figure 5.6 (b) but this system of numbering nodes is only applicable where the mesh is ordered in one direction. In many cases where the grid is not so well ordered, the numbering system is almost arbitrary (p275, Zheng and Bennett, 2002).

Since nodal identification is essential for the creation of intermediate computational models the FDM holds one distinctive advantage above the FEM in that the necessary computational models can easily be developed to assist with the understanding of the head and subsequent flow variation within the modelling situation.

The fundamental difference between the two methods is the calculation of head at a node (p22, Anderson and Woessner, 1992). The finite difference method computes a value for the head at the node, which is the average head for the cell that surrounds the node. There are no assumptions made about the form of variation of head from one node to the next. The finite elements, however, precisely define the variation of head within an element by means of an interpolation function (Gray, 1984).

The FDM is the method that will be used to model the important drain-aquifer interactions in the Thurne catchment and so in this case is the preferred method of discretisation. In deciding to use the Finite Difference Method to simulate the modelling domain the mesh spacing used must be able to allow for the capture of any important variations within the modelling domain, such as each grid cell must essentially represent only one drain.

5.4 The use of software packages within the modelling environment

5.4.1 A review of previous Finite–Difference simulation codes

There are a number of available FDM models such as PLASM (Prickett and Lonnquist, 1971), FLOWPATH (Franz and Guiguer, 1990), MT3D (Zheng, 1990), SEAWAT (Guo et al. 2001) and MODFLOW (McDonald and Harbaugh, 1988) which have all been developed over the years to simulate groundwater flow in aquifers. FLOWPATH (Franz and Guiguer, 1990) includes a solution for a two dimensional, steady-state problems with particle tracking. The flow code is similar to PLASM (Prickett and Lonnquist, 1971) but modified to simulate mesh-centered boundary conditions. Both MT3D (Zheng, 1990) and SEAWAT (Guo et al. 2001) are finite difference codes that solve the simulation of variable density flow by use of an *explicit* approach and particle tracking. (An explicit approach is one where all the unknowns within the same time step are used to solve the unknown within the next steps). Explicit methods are often unstable unless great care is taken in limiting the size of the time step (p, Anderson 1995) (p176, Zheng and Bennett, 2002). MODFLOW (McDonald and Harbaugh, 1988) is a quasi-three-dimensional finite difference code that uses an implicit method (SIP: strongly implicit procedure, SSOR: slice successive over-relaxation) for solving the groundwater flow equation. The software has a modular structure that allows compatible packages to be added to the original software to solve specific boundary conditions such as drains or rivers.

5.4.2 The choice of an appropriate computer software code

The quasi-modular three-dimensional finite difference code, MODFLOW, developed by (McDonald and Harbaugh, 1988) was the computer code used to build and solve the finite difference model for this study. MODFLOW is selected because the computer code is in the public domain; the code is widely accepted and has been validated. The finite-difference structure is compatible with *raster* output from ArcView (See Section 4.4.3) and allows the graphical user interface (GUI) of Groundwater Vistas™ to view imported and exported files directly. MODFLOW uses the block-centered Finite-Difference Method and can simulate steady and transient flow in an irregularly shaped flow system in which aquifer layers can be confined, unconfined, or a combination of confined and unconfined. Flow from external stresses, such as flow to wells, aerial recharge, evapotranspiration, flow to drains, and flow through riverbeds, can also be simulated. Hydraulic conductivities or transmissivities for any layer may differ spatially and be anisotropic (restricted to having the principal direction aligned with the grid axes and the anisotropy ratio between horizontal coordinate directions is fixed in any one layer), and the storage coefficient may be heterogeneous (McDonald and Harbaugh, 1988).

5.4.3 The geometrical configuration of the modelling domain using Arc VIEW

Arc View edition 8.3 is a geographic information system (GIS) data visualization, query, and map creation solution designed for the Windows desktop. It is based on the same technology on which the other Arc GIS Desktop products are based. Arc View 8.3 has two important components namely, Arc Map and Arc Catalog:

Arc Map is the application for map making and analysis in two different forms; Data view or layout view. Data view is an all-purpose view for exploring, displaying, and querying the data on the map created by the user. This view hides all the map elements on the layout, such as titles, North arrows, and scale bars, and allows the user work on the data in a single data frame (i.e. editing or analysis). In layout view, the user can see a virtual page upon which map elements are arranged and the map can be designed.

Arc Catalog is a tool for accessing and managing data. In Arc Catalog the user can assemble individual folders, connections, coverages, and maps are in a catalogue of geographic data sources.

Obtaining the correct geometry is fundamental to the structured modelling of complex hydro-geological systems (Anderson and Woessner, 1992), (Kolm, 1993), (San Juan and Kolm, 1996) (Rushton, 2003). In using Arc View to view the modelling domain in plan the worker can methodically assign boundary condition or general values to identified locations in Arc View and export these values into Groundwater Vistas.

The procedure for conceptualising the modelling domain in Arc View is as follows:

1. The number of layers to be modelled is first identified.
2. The modelling domain in plan for each layer is identified in Arc View (this includes the landward and seaward boundaries).
3. The modelling domain is then mapped over with a grid of pre-determined spacing, (e.g. 250/√2m×250/√2m).

4. Each cell centre is then assigned a node identifying spatially the centre of the individual cell using HAWTHTOOLS. This allows a British National Grid projection X-Y co-ordinate to be assigned to each cell centre and effectively each cell.
5. A database is opened using Microsoft excel assigning the individual properties of each cell (e.g. X co-ordinate, Y coordinate, layer base elevation, layer material etc)
6. The layers 1 & 2 are layers which have cells that host unique attributes (i.e. boundary conditions) such drain nodes require that the locations of these nodes are first identified then another database is created with only these nodes having entries within the subset database. An example of this is the General Head Boundary Condition (GHB) for layer 1. The GHB database is given columns that identify the cell properties required by Groundwater Vistas™ to attribute a cell as a GHB cell (i.e. conductance, GHB head, width of boundary, length of boundary, Reach number). The reach number is a way of grouping cells which have identical boundary conditions.
7. The properties unique to layer 1 are also assigned using the database for layer 1 created in Arc View. An example of this is the distribution of recharge across the layer, where each cell layer 1 cell is assigned a recharge value.

The export of results from the Groundwater Vistas output database tables to the easily accessible Arc View database tables allow for the necessary post audit investigations.

5.5 Summary

Within this chapter, the problem of modelling of a *variable* density fluid flowing through a porous medium of *variable* hydraulic conductivity is investigated and appropriate idealisations are made. The understanding that the variation in fluid density distribution is dependent on several hundreds of years of vertical fresh water recharge, discharges and horizontal saline water inflow is made. The variation in hydraulic conductivity is dependent on geological changes that have occurred over even longer timescales (Banham, 1968); these variations make the simulation of the initial concentrations of the groundwater very difficult. Applying hydrostatic solutions such as the Ghyben-Herzberg (G-H) condition to complex hydrodynamic systems would require the over simplification of the modelling domain properties (G-H assumption of homogeneous medium) and the oversimplification of fluid movement within it (G-H assumption of negligible vertical flows). These assumptions do not hold in a heterogeneous aquifer such as the Pleistocene Crag (Rose and Allen, 1977) and below a substantially drained environment such as the Upper Thurne marshes, which have significant vertical flows (Rushton, 2007) induced by the dendritic drainage system.

The groundwater flow equation for advection flow through a porous media was derived and the important concepts related to it such as storage coefficients were explained. The available discrete solutions for the groundwater flow equation such as the finite-difference method (FDM) and the finite-element method (FEM) were introduced and the advantages and disadvantages of each method were investigated. The chosen software MODFLOW (McDonald and Harbaugh, 1988) and the model boundary restrictions, discretisation and general configurations were explained. The modelling domain configuration is now complete with boundaries, allocation of the layering, the mesh spacing and domain orientation. In the following chapter, the estimated input parameters are deduced and the development of quantified conceptual models is introduced.

6 PARAMETERISATION OF THE UPPER THURNE GROUNDWATER SYSTEM

6.1 Introduction

The previous chapter introduced the background to the numerical modelling of regional groundwater of variable density using a single density approach with particle tracking. The method of finite–differences was reviewed and the justification of the choice of modelling software (MODFLOW) was given along with the Groundwater Vistas^{TM} (Rumbaugh and Rumbaugh, 2000) Graphical User Interface (GUI). There are two main computational models within the current study, the regional groundwater model (DRGF) and the drain-aquifer interaction model (DAI), which were introduced in Sections 4.3 and 4.4 and are used to understand the possible distribution of saline water within the catchment. In this chapter, the hydraulic conductivities of the geological deposits are estimated and the modelling domain boundary is defined, along with the mesh spacing and alignment. The estimation of input parameters such as the recharge and the drainage coefficients, which take into account the aquifer's geological and geometrical properties, are also presented.

Extensive weekly data (Holman, 1994) for electricity usage for the individual electrical discharge pumps and observation well water levels for the Upper Thurne catchment are only available from April 1st 1991 through March 31st 1993. No other groundwater monitoring data are available. This period incorporate two very different meteorological years and with very little knowledge of the initial conditions, they are almost impossible to replicate in a time–variant manner. Therefore, a pragmatic approach has been adopted to make the best use of the data available and to attempt to replicate the average conditions for this time-period. In this time-invariant analysis, the storage coefficients are ignored and an average daily value for the recharge in each of the individual landscape units is required.

6.2 Estimating the hydraulic conductivities of the various geological deposits

6.2.1 Characteristics of the alluvial deposits

As explained in Chapter 3, within the Upper Thurne catchment the low-lying land is covered with two types of soil deposits. To the north of the Brograve pump (Figure 3.5) the Altcar 2 Association which is peat underlain at a shallow depth by marine alluvium (Holman, 1994), is the predominant coverage. The southern marshes consists of the Newchurch Association which is a pelo-calcareous alluvial gley soil (Hodge et al. 1984) developed in marine alluvium, which is found in the river valleys (p25, Cox et al. 1989) and their tributaries and comprising mainly of silt and clay. Within the catchment thicknesses of up to 4 metres have been measured (Fig. 4.21).

6.2.2 Estimating the hydraulic conductivity of the superficial deposits

The hydraulic properties of overlying deposits determine how readily recharge is transmitted to the saturated zone (p127, Younger, 2006), although there is an assumption that the hydraulic conductivity within the peat decreases (a form of heterogeneity) gradually with depth or degree of decomposition (Beckwith et al., 2003b; Holden and Burt., 2003). The horizontal hydraulic conductivity (K_h) of a peat formation is often greater than its vertical hydraulic conducticity (K_v), usually refered to as anistropy (Chason and Siegel, 1986; Beckwith et al. 2003b). Field investigations have been conducted to provide information about the hydraulic conductivities of the actual deposits within the catchment area.

The method of collecting soil specimens and then measuring the saturated hydraulic conductivity first requires a metal cylinder (internal diameter 10 cm) of known internal volume and with a sharp leading edge to be hammered or pressed into the soil. All samples were taken from between 0.8-1.0 m below ground level. After the sample has been taken, it is saturated overnight with water to remove all the air. The saturated sample is connected to the falling head permeameter, which is represented schematically in Figure 5.1.

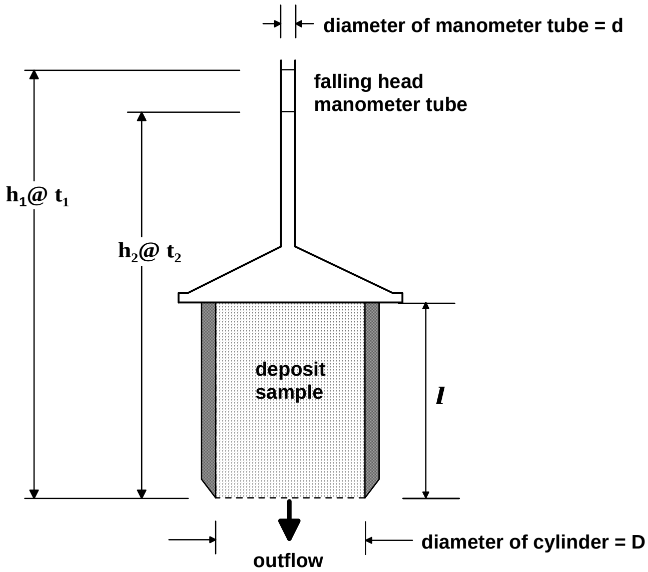

Figure 6. 1 Schematic diagram of the falling head permeameter

The typical dimensions for the equipment is as follows; the length of the cylinder, ***l***, is 0.1 m with an internal diameter, ***D***, of 0.08 m, the diameter of the manometer, ***d***, will vary from between 6 mm to12 mm depending on whether the deposit being investigated is expected to have a low or high hydraulic conductivity respectively. The speed of the fall in head within the permeameter is then measured from which the hydraulic conductivity can be deduced (Johnson et al., 2005).

The inflow from the manometer will be equal to the flow through the soil sample. The water level in the manometer falls Δh in time Δt. The volume of water lost from the standpipe unit is Q in m^3/day. Using Darcy's law for a fixed head in the cylinder

$$Q = KA\frac{\Delta h}{l} \tag{6.1}$$

K is the hydraulic conductivity (m/d) A is the cross sectional area of the sample (m^2) and l is the cylinder length (m) (see Figure 6.1) and for a falling head (a is the cross sectional area of the manometer tube).

For the manometer tube

$$Q_1 = -a\frac{dh}{dt} \tag{6.2}$$

For the specimen using

$$Q_2 = KA\frac{H}{l} \tag{6.3}$$

where H is the head difference across the sample. Since $Q_1 = Q_2$ from equation 5.2 and 5.3

$$KA\frac{H}{l} = -a\frac{dh}{dt} \tag{6.4}$$

separating the variables and integrating with $H = h_2$ at $t = t_2$ and $H = h_1$ at $t = t_1$

$$K = \left[\frac{al}{A(t_2 - t_1)}\right]\ln\left(\frac{h_1}{h_2}\right) \tag{6.5}$$

The cross sectional area of the manometer is given by $a = \frac{\pi d^2}{4}$ and the area for the cylinder is given by $A = \frac{\pi D^2}{4}$ the hydraulic conductivity K can be estimated from

$$K = \left[\frac{d^2 l}{D^2(t_2 - t_1)}\right]\ln\left(\frac{h_1}{h_2}\right) \tag{6.6}$$

To reduce the effects of variance, several samples were taken for each geological material with the results of 20 samples taken through out the catchment given in Table 6.1.

Table 6. 7 Hydraulic conductivity for individual superficial deposits

Material	No of samples	Range m/day
Clay	4	0.0089-0.001
Peat	4	0.001-0.024
Norwich Brickearth	5	0.1-2.00
Sand	3	4.7-5.0

The highest hydraulic conductivity value of clay obtained was 8.9×10^{-3} m/d, which is greater than expected, typical values for clay lie within the range of 10^{-3} and 10^{-6} m/d (Gerber and Howard , 2000). The values of the more permeable material showed more consistency, whereas the values obtained for the less permeable deposits were slightly nearer to the expected values.

6.2.3 The estimation of the hydraulic conductivity of the crag

The transmissivity values derived from pumping tests completed for the National Rivers Authority range from 150 m^2/day to 1100 m^2/day, which for an aquifer of 50m thickness, produces a range of hydraulic conductivity values of 3-22 m/day (Holman, 1994) which is consistent with representative values for fine to coarse sands (Domenico and Schwartz, 1997). Also Erskine (1991), reported that pumping tests conducted in the crag at Sizewell in Suffolk, gave horizontal conductivity value of 20 m/day and a vertical hydraulic conductivity value of 2 m/day. The lower vertical hydraulic conductivities is probably due to the presence of extensive horizontal zones of lower hydraulic conductivity (Bliss and Rushton, 1984). Therefore, the crag has been represented as homogeneous and isotropic, but with a low conductivity layer (Fig. 3.1 & Fig. 6.5) consistent with Holman (1994).

6.2.4 Input hydraulic conductivity values for the modelling domain

Choosing a value of hydraulic conductivity for both the peat and the clay is a difficult task as both deposits have hydraulic conductivities that may be anisotropic (Holden & Burt, 2003; Beckwith et al., 2003b) and/or heterogeneous. In practice hydraulic conductivity values vary over the study area but there is insufficient field information to justify areal variations. Therefore a pragmatic approach has been taken to choose values of the likely correct order of magnitude, which have then been tested in a sensitivity analysis to see whether the model is sensitive to the chosen parameter values of the alluvium. Table 6.2 presents the listing of the various hydraulic conductivities that are to be used in the preliminary model.

Table 6. 8 Aquifer parameter values to be used in the model

Geology	Saturated hydraulic conductivity	Location	Typical thickness
Overlying strata			
clay (Newchurch Association)	*K*v=0.0008 m/day* (Approximated)	Southern Marshes	1-5m Fieldwork
peat (Altcar 2 association)	*K*v=0.0016 m/day (Approximated)	Northern Marshes	1-5m (Burton et al., 1987)
Exposed Crag	*K*v =5 m/day (Fieldwork)	Between Catfield and Ludham	1-2m
Drift for Flegg Hundreds	*K*v=0.3 m/day (Approximated)	Flegg Hundreds	10-23m (OS map)
Norwich Brick-earth (North Sea Drift)	*K*v = 0.1-1 m/day Boswell (1916) (sandy loam)	Loamy Uplands, Holmes & Flegg Hundreds	0-9m (BGS map)
Main aquifer			
Norwich Crag (Sands and Gravels)	*K*h = 20 m/day Erskine (1991)	Catchment wide	25-65m (Holman et al. 1999)
'Clayish' layer within Crag (consolidated material)	*K*v = 0.1 m/day (Approximated)	Catchment wide	3 m (Holman et al. 1999)

The deposits of the Flegg Hundreds were grouped together as one superficial deposit with the knowledge that the river divide would reduce its influence on the groundwater flows in the northern half of the catchment. The value given of 0.3 m/day is intended to reflect the combination of the low hydraulic conductivity Boulder Clay covering the more permeable Corton Sands and Norwich Brickearth. The Norwich Brick-earth was assigned the hydraulic conductivity value of 1m/day for a sandy loam given by Boswell (1916), which also corresponds to that obtained from the laboratory experiments. The Norwich Crag is assigned a value of 20 m/day for both the vertical and horizontal hydraulic conductivities. The exposed crag is covered with sand that was tested using the falling head permeameter described in Section 6.2.2 and thus was given a hydraulic conductivity of 5 m/day. Model input screen of the distribution of the hydraulic conductivity of the uppermost layer (Layer 1) is presented in Appendix A.

6.3 Configuring the modelling domain in plan

6.3.1 Defining the modelling domain

The external boundaries for the groundwater system in the Thurne catchment are difficult to identify, but an approximation can be found in the surface water boundaries. This is based upon the surface water boundaries of ESNRA, (1971) and modified in the vicinity of the surface water drainage systems which are outside the Smallburgh IDB and Martham, Repps and Thurne IDB drainage sub-catchments (Fig. 3.7).

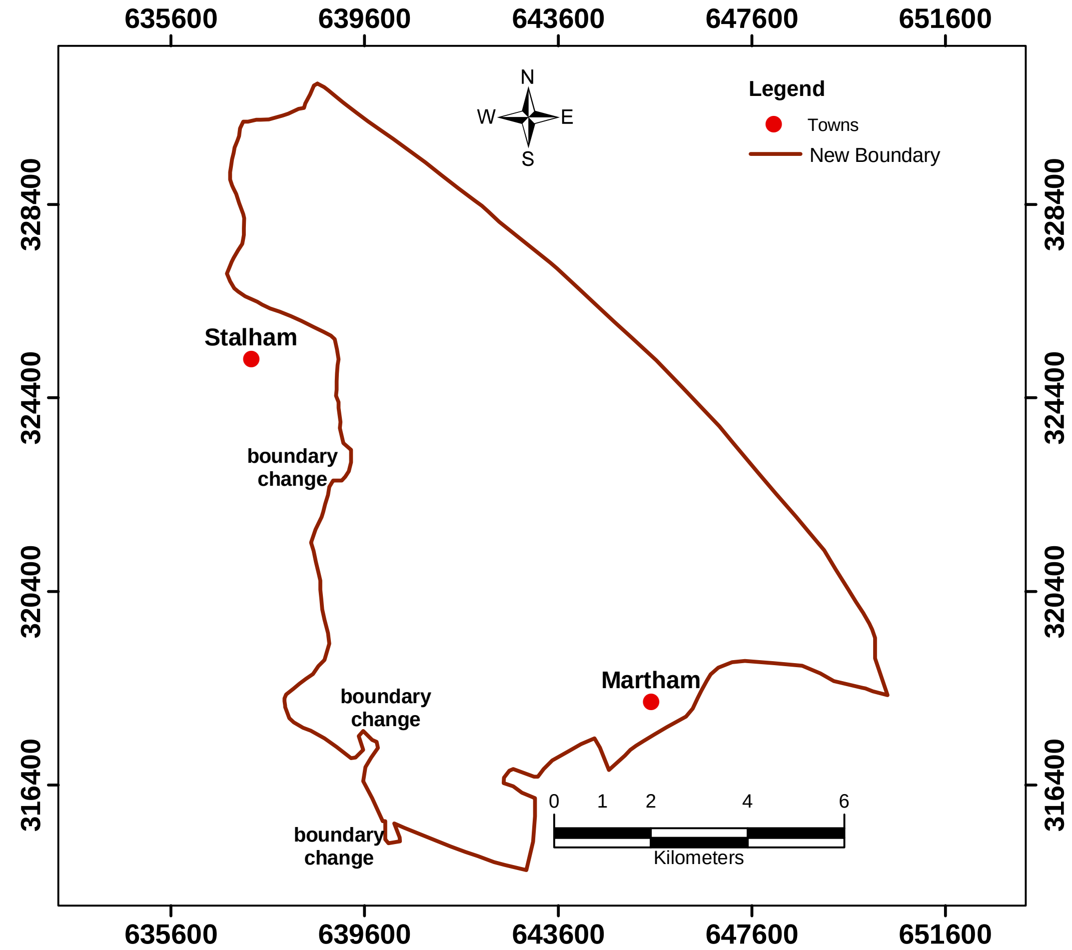

Figure 6. 2 Map of groundwater boundary (based on ESNRA, 1972)

6.3.2 The choice of an appropriate axes alignment

It is advisable to align the cells in the direction of flow so that the major groundwater flow direction is along one of the principle axes (Section 5.2.2). The mesh alignment was also chosen so that the catchment grid, when rotated 45°clockwise, would still coincide with the kilometre intersections of the national grid. This is an important feature of the grid spacing, as it allows future finite-difference investigations that may cover a larger area of Norfolk to coincide with nodal points on the kilometre intersections.

6.3.3 The choice of an appropriate mesh spacing

The choice of the mesh size is usually a matter of expediency (Rushton and Tomlinson, 1977) and is often dependent upon the individual situation being modelled. In the case of the Upper Thurne, the main boundary condition on the internal cells is the drains' connectivity to the underlying aquifer. The mesh spacing should be such that each cell is representative of the deepest drain that is encompassed within it. In the case where several drains are within one cell a drainage coefficient must represent the total connectivity. The mesh spacing should fit with any modelling *ideals* placed upon the discretised domain.

The two ideals are:

a) The mesh alignment of 45^o to the national grid
b) The modelling should be discretised with between 500 and 2000 cells in total, so that the task of preparing the data is not too time consuming.

The three alternatives considered are 125/√2m spacing, which would render the current modelling domain to be discretised into 80,000 cells per layer, 250/√2m giving 5,000 cells per layer or 1,000/√2 with 300 cells per layer. The 125/√2m spacing gives the model 400,000 cells in total since there are five layers the total number of cells increase is five fold within the model (Section 6.4). The 1,000/√2m spacing leads to a cell length of approximately 707m, which would not capture the variety of drain connectivity that can be found within the complexity of the network of drains. The chosen discretisation is therefore 250/√2m, which renders the landward portion of the modelling domain to be approximated with 3,000 cells per layer.

6.4 Representing the layered aquifer using MODFLOW

6.4.1 The concept of conductance in the MODFLOW model

When the aquifer has sources or sinks within the groundwater system then an extra term Q is added to the left hand side of Equation 4.9

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right)+Q(x,y,z,t)=S_S(x,y,z)\frac{\partial h}{\partial t} \qquad (6.7)$$

This is the partial-differential equation of groundwater flow used in MODFLOW (McDonald & Harbaugh, 1988). K_x, K_y and K_z are values of hydraulic conductivity along the x, y, and z

coordinate axes which are assumed to be parallel to the major axes of hydraulic conductivity [L/T]; h is the potentiometric head [L]. Q is the volumetric flux per unit volume representing sources and/or sinks of water, with $Q < 0.0$ for flow out of the groundwater system, $Q > 0.0$ for flow in [T^{-1}]. S_s is the specific storage of the porous material [L^{-1}] and t is time [T]. Equation 6.7, when combined with boundary and initial conditions, describes transient three-dimensional groundwater flow in a heterogeneous and anisotropic medium, if the principle axes of hydraulic conductivity are aligned with the co-ordinate directions.

MODFLOW solves the governing Equation 6.7 using a finite difference method in which the groundwater flow system is divided into grid cells. For each cell, there is a single point, referred to as a node, at the which head is calculated. The finite-difference equation for a cell is

$$
\begin{aligned}
& CR_{i,j-\frac{1}{2},k}\,(h^m_{i,j-1,k} - h^m_{i,j,k}) + CR_{i,j+\frac{1}{2},k}\,(h^m_{i,j+1,k} - h^m_{i,j,k}) \\
+ \quad & CC_{i,j-\frac{1}{2},k}\,(h^m_{i,j-1,k} - h^m_{i,j,k}) + CC_{i,j+\frac{1}{2},k}\,(h^m_{i,j+1,k} - h^m_{i,j,k}) \\
+ \quad & CV_{i,j-\frac{1}{2},k}\,(h^m_{i,j-1,k} - h^m_{i,j,k}) + CV_{i,j+\frac{1}{2},k}\,(h^m_{i,j+1,k} - h^m_{i,j,k}) \\
+ \quad & P_{i,j,k}\, h^m_{i,j-1,k} + Q_{i,j,k} = SS_{i,j,k}(\mathrm{DELR_j}\times\mathrm{DELC_i}\times\mathrm{THICK_{i,j,k}})\,\frac{h^m_{i,j,k} - h^{m-1}_{i,j,k}}{t^m - t^{m-1}}
\end{aligned}
\qquad (6.8)
$$

where $h^m_{i,j,k}$ is head at cell i, j, k at time step [L]; CV, CR, and CC are hydraulic conductances, between node i, j, k and a neighbouring node [L^2T^{-1}]. $P_{i,\,j,\,k}$ is the sum of the coefficients of head from source and sink terms [L^2T^{-1}]; $SS_{i,j,k}$ is the specific storage [L^{-1}]; ($\mathrm{DELR_j}$) is the cell width of column j in all rows [L]. ($\mathrm{DELC_i}$) is the cell width of row i in all rows [L]; ($\mathrm{THICK_{i,\,j,\,k}}$) is the vertical thickness of cell i, j, k [L]; and t^m is the time step m [T]. For steady-state stress periods, the storage term and therefore, the right-hand side of the Equation 6.8 is set to zero.

6.4.2 Allocation of layers to represent the Upper Thurne aquifer system

Figure 6.3 (a) shows the actual layering within the Upper Thurne aquifer system according to Holman (1994) and Downing (1959), which is used to represent the conceptual models of Figure 4.1 and Figures 4.17 to 4.22.

Layer 1 can represent a variety of covering deposits. In the low-lying marshes, Layer 1 represents the peat or the clay; the top of the peat or the clay is the *cultivated layer* in which the water table is usually controlled by undersoil pipe drains. In the higher–lying land, such as the Loam Uplands or the Holmes (Figure 3.2) it represents the more permeable Norwich Brickearth or exposed crag.

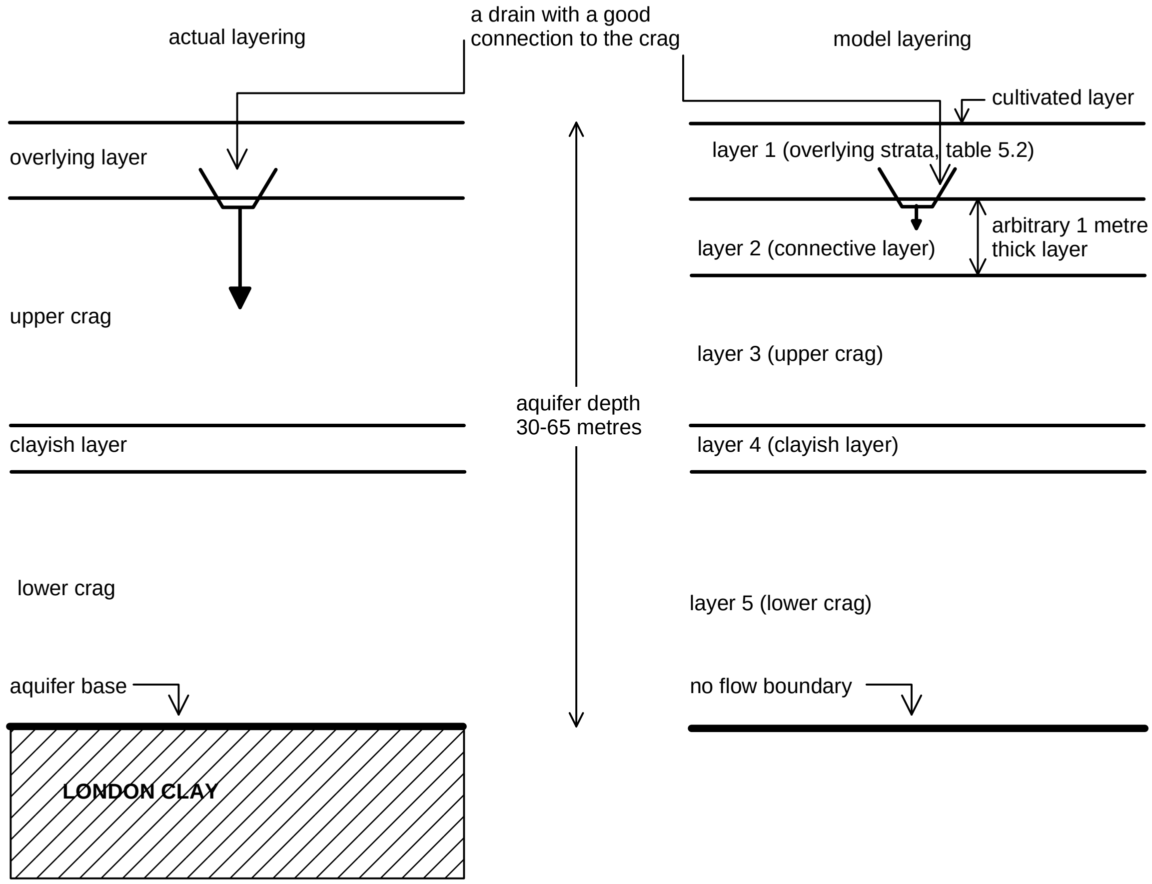

Figure 6. 3(a) the actual layering and (b) model layering of the crag aquifer (not to scale)

In the MODFLOW model, an additional layer (Layer 2) is introduced in Figure 6.3 (b) so that the vertical hydraulic conductivity of the overlying layer can be represented adequately. Layer 2 ensures that drains are connected to the top of the upper aquifer. If this additional layer is not provided the influence of the drain would be simulated by MODFLOW as if it penetrated to a depth of 11.5 m depth into Layer 3 (0.5× 23m = 11.5 m), Figure 6.4 (a). The nodal distance m in Figure 6.4 (a) is approximately [(1+23)/2] = 12 metres, whereas the nodal distance n in Figure 6.4 (b) is approximately [(1+1)/2] = 1 metre.

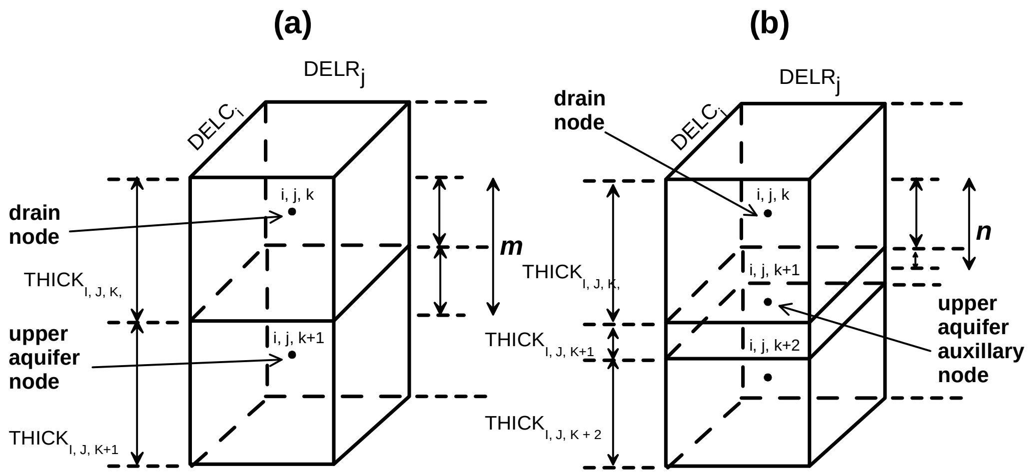

Figure 6. 4 (a) drain-aquifer nodes representing actual layering (b) drain-aquifer nodes with connective layer

6.4.3 The current MODFLOW approach to modelling rivers/drains

In the development of a regional groundwater model in which the drain-aquifer interaction is an influential feature, the geometry and properties of a drain channel cannot be represented in detail because the mesh spacing is usually greater than the dimensions of the drain. The drain-aquifer interaction is represented by individual drainage coefficients and associated equations relating drain-aquifer flows to the drain-aquifer head gradient. The MODFLOW approach to stream-aquifer relations refers to stream and rivers as opposed to drains. The current MODFLOW approach uses a drainage coefficient (DC) in m^3/d/m, which represents the drain's ability to 'attract' water due to the localised gradient in groundwater head according to the equation:

$$Q = DC \times (H_{riv} - h) \qquad \text{for} \qquad H_{riv} \le h \tag{6.7}$$

where
$$DC = L \times W \times K_{bed} / M \tag{6.8}$$

and L = length (m), W = width (m), K_{bed} = hydraulic conductivity of the drain bed (m/d), M is the thickness of the drain bed material in metres, H_{riv} the head in the river and h the head in the nearby aquifer (m). (Prickett and Lonnquist, 1971) were two of the early investigators to use this form of linear relationship. A typical river will have a bed (Figure 6.5a) which is both varied in thickness and in geology, along with an aquifer that has a varying hydraulic conductivity making any estimation of the passage of water through the riverbed difficult. The current drainage coefficient formulation used by MODFLOW (Figure 6.5b) assumes that the drainage coefficient is inversely proportional to the thickness of the drain bed material and ignores the underlying aquifer properties. In order for MODFLOW to represent the Thurne system, the river-aquifer modelling approach in MODFLOW has needed to be adapted:

- For conditions in the coastal zone (Section 6.4.3)
- To represent the interaction between the deep drains (Section 6.4) and the crag aquifer, this has required the drainage coefficient to be calculated in a different way. (Section 6.6)

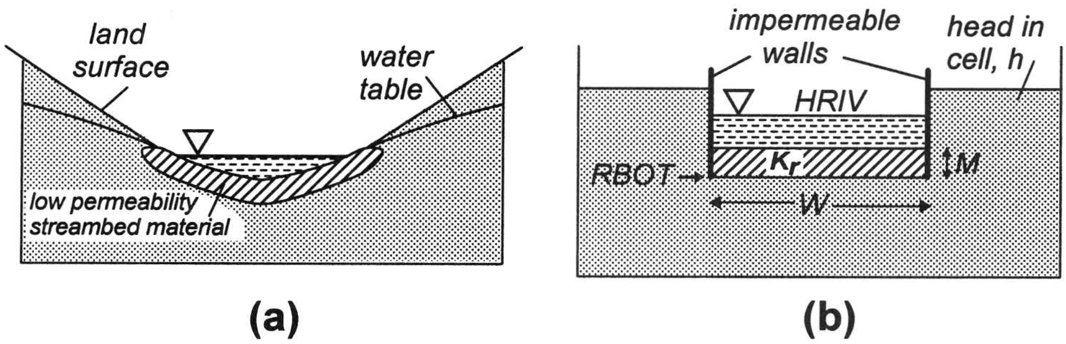

Figure 6. 5 Representation of river-aquifer interaction in MODFLOW:
(a) actual geometry of river section, (b) idealisation used in MODFLOW (Anderson and Woessner, 1992)

6.4.4 The coastal boundary

A conceptual model of conditions at the coastal boundary is illustrated in Figure 6.6. The slope of the land beneath the sea is approximately 1:200. This means that the upper boundary of the aquifer representing the contact with the sea occurs in Layer 1, Layer 2 and Layer 3.

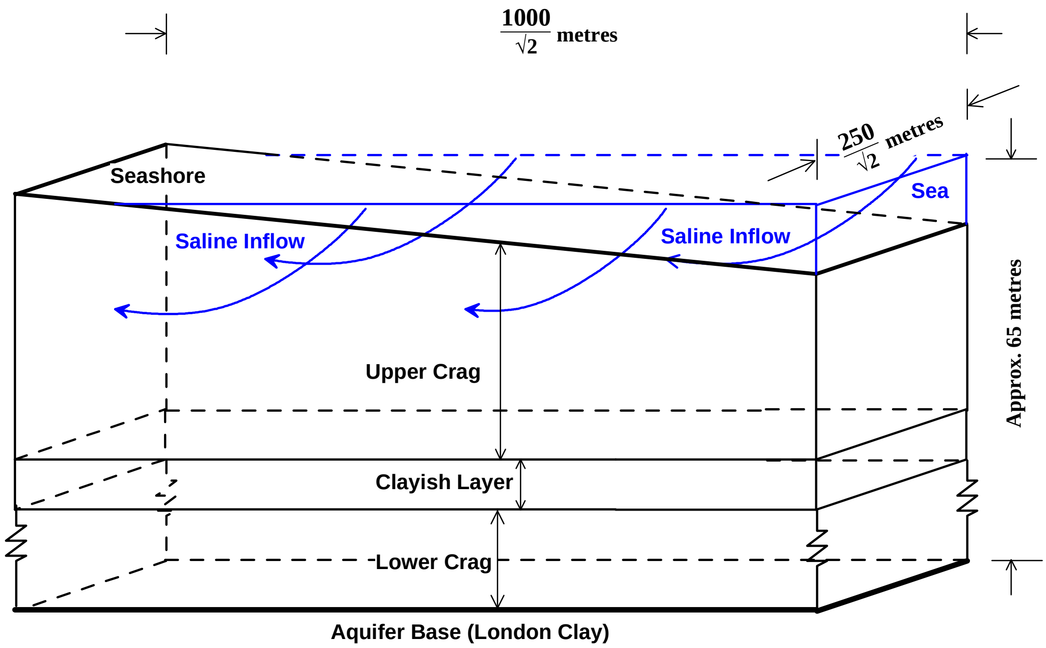

Figure 6. 6 A 250/√2m-width cross-section of the Coastal Boundary Zone

This contact is illustrated in Figure 6.7 for a 250/√2m width cross-section of the coastal boundary. The coastal offshore area is modelled in MODFLOW, using the river boundary condition. This allows for a calculation of the amount of water entering or leaving the coastal grid cell.

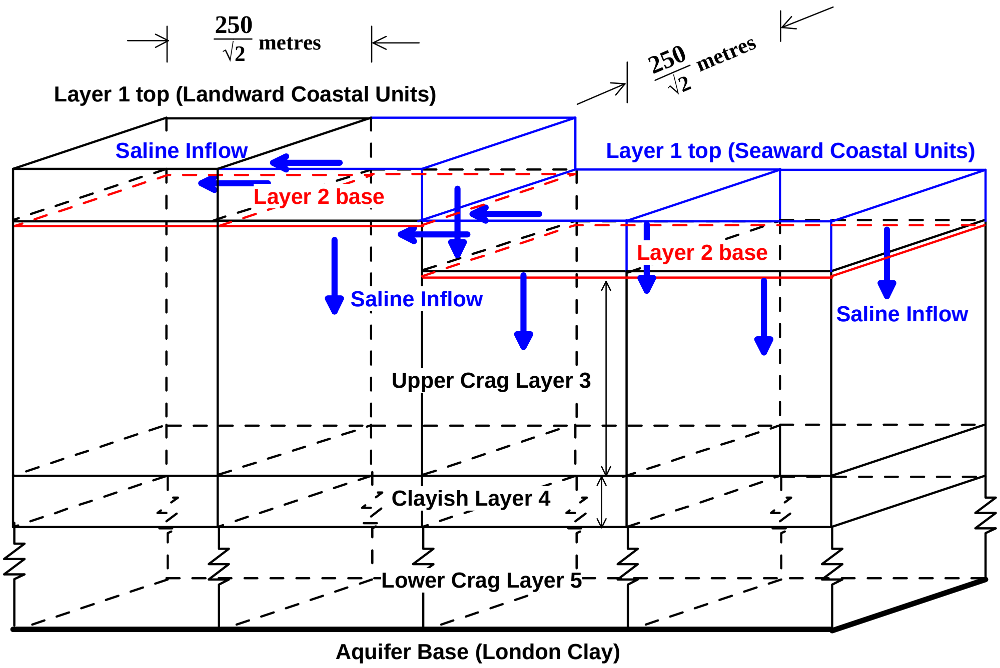

Figure 6. 7: Discretisation of the Coastal Boundary Zone

Figure 6.7 represents the conceptualised coastal shelf with Layer 1 *landward* coastal units in black and Layer 1 coastal *seaward* units in blue. The 1 metre thick Layer 2 base is shown in red. This 'square in plan' river has a river bottom (RBOT) which has a very high conductance (of > 10^7 m^2/d) with a negligible thickness (of $<10^{-7}$ m); this allows the coastal shelf to be modelled as a 'river' as shown in Figure 6.8 (b).

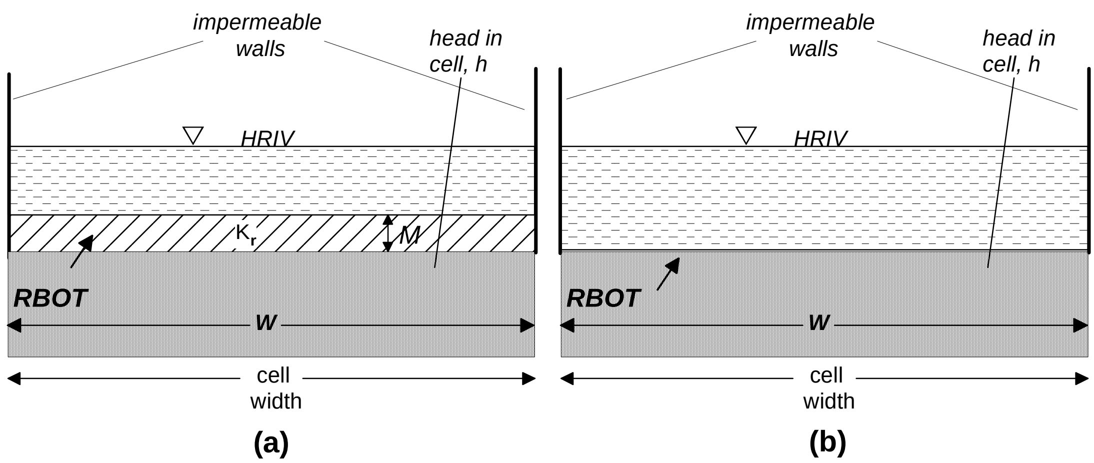

Figure 6. 8 Representation of coastal shelf -aquifer interaction in MODFLOW using a 'square' river

(a) actual geometry of river section, (b) effective geometry of the 'square

6.4.5 The introduction of the MODFLOW General Head Boundary Condition

When representing features such as the water table elevation in the cultivated layer and the interaction between lakes and the aquifer, the General Head Boundary (GHB) MODFLOW subroutine is used. The GHB can be configured to allow the representation of a specified head; MODFLOW provides the flow resulting from this specified head. It is important to emphasize that the use of the GHB is a convenient technique for determining the nodal flows from the specified heads.

The General Head Boundary condition is based on the Darcian concept of flows through a porous medium being proportional to the head gradient along the length of the medium (i.e.$Q \propto \Delta h/\Delta l$). Flow through the boundary (Q_b) is calculated as the product of the conductance (C_b) of the boundary (i.e. the boundary's ability to accept flow) and the difference between the head at the boundary (h_b) and the head in the aquifer (h) measured in metres:

$$Q_b = C_b \times (h_b - h) \tag{6.9}$$

The conductance term is analogous to representing the resistance to flow between the cultivated layer and the aquifer. When using the GHB condition to simulate a specified head boundary condition, C_b is set to a large value.

Within the modelling process, there is a need to ensure that boundary conditions such as GHB do not generate unrealistically high horizontal or vertical flows, therefore as a modelling precaution the GHB distribution is limited to cells that are vertically more than one metre thick and are not adjacent to any cells allocated as exposed crag or Norwich Brick-earth.

6.4.6 The use of the MODFLOW GHB condition

The MODFLOW GHB is used to represent two important elements of the behaviour of the Upper Thurne catchment. Firstly, an under soil pipe drainage system (Plate 4a) is needed to assist water table control throughout the coverage of the seasonally waterlogged peaty (Altcar 2 Association) and clayey (Newchurch Association) alluvial soils, which requires effective simulation in the modelling process. The method chosen to replicate this field condition in the model is the General Head Boundary condition. Details of the GHB condition used in Layer 1 is presented in Appendix C.

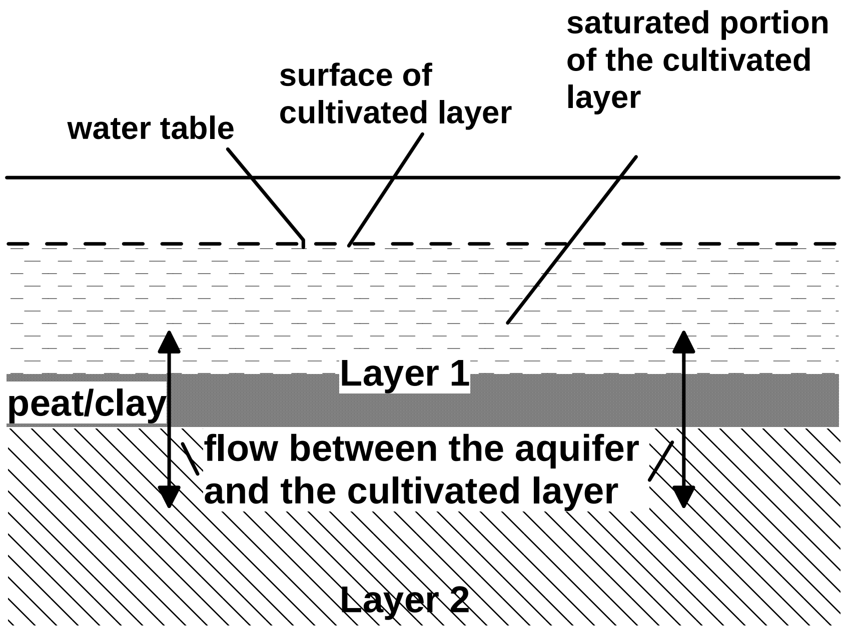

Figure 6. 9 A simplified schematic profile of the interaction between the cultivated layer and the underlying aquifer

In the case of the representation of the flows between the Broads and the underlying aquifer the beds of the broads are generally composed of soft unconsolidated sediments (Lambert et al. 1960) so that flow is occurring between specific broads (Martham Broad) and the underlying aquifer. The General Head Boundary Condition is used to represent the low fluctuating heads in the Broads.

6.5 Estimating potential recharge

6.5.1 Introduction

The definition of potential recharge is the amount of infiltrating water which passes the root suction zone (p90, Rushton, 2003). There are several direct ways of estimating potential recharge to British aquifers such as Water balance Simulation (WaSim) model of Hess et al. (2000), or by evaluating the soil moisture balance (SMB) (Rushton et al. 2006). These methods require the estimation of actual evapotranspiration and runoff. The evaluation of the former can found using techniques developed by Penman (1948), whereas the evaluation of the percentage of runoff by can be found by use of a runoff soil curve number (SCN) of Ponce and Hawkins (1996) or the Standard Percentage Runoff (SPR) of Boorman et al. (1995). The SCN is an American development, created to estimate immediate runoff quantities from soils after storm events. Boorman et al. (1995) is a hydrologic based classification of the soils of the UK and was developed from existing datasets that described both the soils and their distributions, and the hydrological response of catchments (Boorman et al. 1995). The document has classification listings for all soil types, according to their parent substrates and the physical properties of the soils, and estimates the SPR that causes the short-term increase in flows seen at the catchment outlet and which is used to estimate the fraction attributed to runoff after a storm event.

Figure 6.10 is a simplified flow diagram showing the principal stores and thresholds which need to be considered to evaluate the quantity of “potential recharge”. The diagram introduces the relevance of important soil conditions such as “field capacity” [which is defined as the amount of water that a well-drained soil can hold against gravitational forces] and “wilting point” [which is the soil moisture content below which plant roots cannot extract moisture].

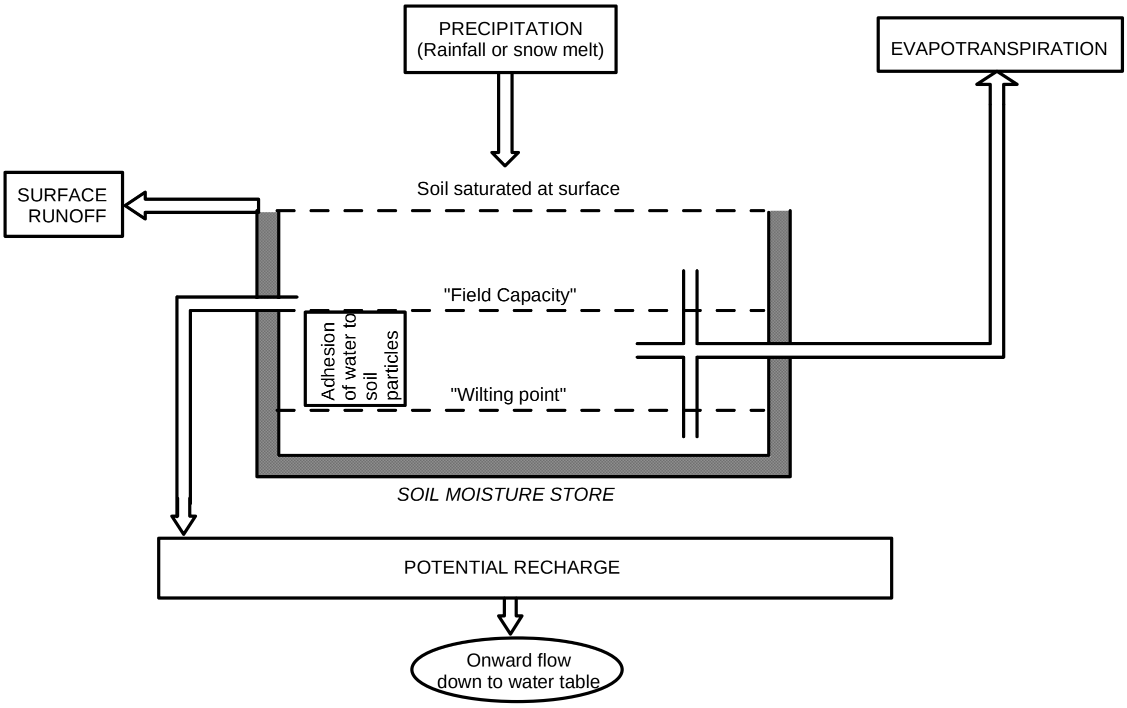

Figure 6. 10 A simplified flow diagram showing the principal stores and thresholds, which need to be evaluated to evaluate the quantity of “potential recharge”. (p33, Younger, 2006)

Simple estimates of potential recharge (Pot. Rech) are found by evaluating the storage components for the precipitation (PPT), evapotranspiration (ET) and the runoff (RO) at field capacity.

Precipitation (PPT)- Evapotranspiration (ET)-Runoff (RO) = Potential Recharge (Pot. Rech) (6.10)

6.5.2 Estimating potential recharge for the Upper Thurne catchment

For the time-invariant modelling of the Upper Thurne catchment using MODFLOW, a single potential recharge estimation for the drift covering the individual landscape units is required. The recharge estimation for the marshland will take the form of specified heads within the GHB. In the other landscape units, annual average estimates are adequate.

Various superficial deposits cover the Upper Thurne catchment that will vary in their individual abilities to allow rainfall recharge to pass through. In order to represent the impact of the drift, one approach is to consider a *recharge factor* for each of the constituent portions of drift; for further discussion see pp 91, Rushton (2003). The advantage of this approach is that the worker has full control of the recharge estimates. The estimated potential recharge entering via the more permeable (sandy) deposits is reflected by use of a recharge factor that is equal to unity and likewise a less permeable deposit or terrain with greater slopes will have a recharge factor less than unity. A simple calculation of the potential recharge is made using the weekly MORECS precipitation and actual evapotranspiration data used by Holman (1994)

Year 1 (1991/2)

PPT Precipitation = 539.4 mm/year ET Evapo-transpiration = 413.5 mm/year

PR Potential recharge (PPT-ET) = 539.4 - 413.5 = 125.9 mm/year = 0.00034 m/day.

Year 2 (1992/93)

PPT Precipitation = 657.1 mm/year ET Evapo-transpiration = 522.6 mm/year

PR Potential recharge (PPT-ET) = 657.1 – 522.6 =134.5 mm/year = 0.00037 m/day.

The calculated mean recharge value for the two years is then approximately 0.00036 m/day, which represents the mean potential recharge per unit area for the entire catchment assuming no runoff or rapid lateral flow. This value is assigned to the area that has been designated exposed crag. The areas of recharge that will require a recharge factor are the areas covered by the Norwich Brickearth and the unit known as the Flegg Hundreds. To determine an appropriate recharge factor for these areas the composition of the covering materials must be considered. A sample of Norwich Brickearth collected at Corton 3 miles north of Lowestoft, just outside the catchment was sieve analysed by Boswell (1916), and described as a sandy loam (clayey sand) with 53.95% sand grade (1mm and 0.1mm) and the remainder evenly divided between silt (20.4%) and clay (25.65%).

The Norwich Brickearth coverage in the Holmes and in the Loam Uplands is between 3 and 10 metres thick (Table 3.2) with the Ordinance Survey maps suggesting that the majority of this land is closer to 3 metres. It follows from Rushton (2003) that an approximate recharge factor for the Holmes and the Loam Uplands would be between 0.9 and 0.8 depending on the estimated mean thickness of the coverage (Table 6.2).

Table 6. 9 Typical recharge factors for a representative location (p.91, Rushton, 2003)

Thickness/Drift type	Sand	Clayey sand	Sandy clay	Clay
0 to 3m	1.00	0.90	0.20	0.02
3 to 10m	1.00	0.80	0.10	0.01
>10m	1.00	0.60	0.03	0.00

A similar reasoning can be used to estimate the recharge factor for the Flegg Hundreds. This estimation however must include the fact that a fraction of the Flegg Hundreds is covered with the sand and gravels of the Corton Woods deposit and it in turn is covered with the less permeable Boulder Clay of the Lowestoft Till Group (see Section 3.2). The factor of 0.73 also reflects the increased estimated mean thickness and the increase slope of the terrain. The potential recharge distribution adjusted with the individual recharge factors is given in Table 6.3.

Table 6.10 The estimated potential recharge values for the various deposits (m/day)

Deposit	Recharge Factor	Factored Potential Recharge
Flegg Hundreds	0.73	0.00029
Norwich Brickearth	0.84	0.00034
Exposed crag	1.00	0.00040

(Note that the values factored potential recharge is based upon an initial potential recharge estimate of 0.0004 m/d.)

An alternative method of calculating recharge for the various geological units is to consider the soils that cover the catchment and use the aforementioned HOST SPR (Boorman et al., 1995) to estimate runoff. In the case of the Flegg Hundreds, the predominant coverage is the Wick 2 Association, which has a HOST class of 5 and a Standard Percentage Runoff of 14.5%. This suggests that surface infiltration is 85.5%. However, there is a lateral outflow from the intermediate more permeable layer of the Corton Sands, which may thus reduce the recharge to the underlying aquifer and a recharge factor closer to 0.73 is used in Table 8.3. A similar calculation can be made for the landscape units covered by the Norwich Brickearth, whereby the SPR is approximately 20%, which is consistent with the recharge factor of 0.84 in Table 8.3. A particle-tracking plot for the recharge distribution is presented in Appendix B.

6.6 The estimation of drainage coefficients using a 2-D finite-difference mesh

6.6.1 A non-linear relationship between Q and head difference

Flow between an aquifer and a water conduit such as a river or drain is usually assumed to be controlled by the same sort of mechanism as leakage through a semi-pervious stratum into a confined aquifer (Rushton and Tomlinson, 1979) with the assumption of a linear relationship between Q and head difference. This relationship was first proposed by Prickett and Lonnquist (1971) and is often too simplistic because it ignores the hydraulic and geometrical properties of the aquifer. One of the earliest investigations of Rushton & Tomlinson (1979) suggested that a combination of a linear and non-linear coefficient might best represent the actual relationship,

$$Q = K_3 HDIFF + K_1\left[1 - \exp(-K_2 * HDIFF)\right] \text{ for } HDIFF \geq 0 \qquad (6.9)$$

Where K_1, and K_3 are constants ($L^3/T/L$) and K_2 is a constant (L^{-1}) and depend on the individual river configuration and aquifer properties.

6.6.2 Estimating drainage coefficients using a finite-difference mesh

In order to derive realistic drainage coefficients for the current study, a series of finite-difference solutions of vertical section models have been created. A problem is specified in Figure 6.11(a), which refers to flow from the underlying crag, through the peat, and into the drain; the drain extends to - 2.6 mOD, and the drain water level is -2.0 mOD. Using a vertical section x-z numerical model for the problem of Fig. 6.11 (a), the drainage coefficient can be estimated from solutions with different values of the head in the crag. Note that the head on the wetted perimeter of the drain is -2.0 mOD, there is a seepage face on the sides of the drain and the base of the cultivated layer is treated as a no- flow boundary.

Solutions for the drainage coefficients are obtained with different values of base to aquifer thickness (m_b) and are quoted in Table 6.4, making it possible to gain an insight into the drain-aquifer interactions further along the main drain where the thickness of the basal material varies (Figure 6.21).

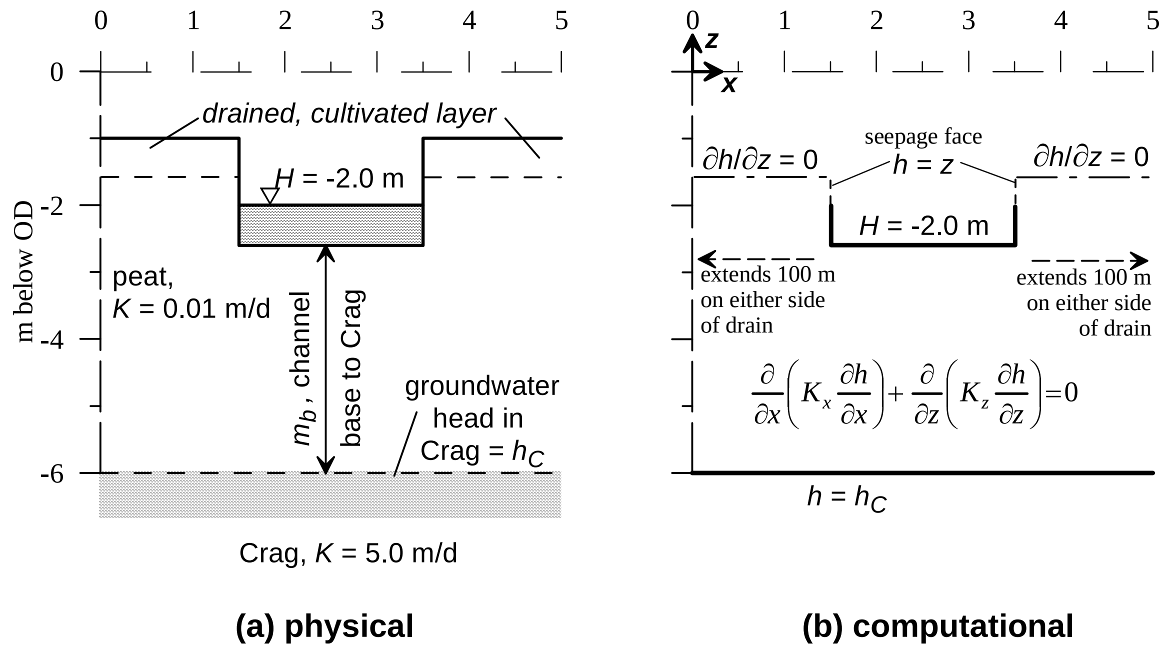

Figure 6. 11 2-D representation of the main drain 2km from the pump

Modelling the actual drain configuration (Case 1) which is representative of the main drain 2km upstream from the Brograve pump can assist with the understanding of the magnitude of the vertical flows into the drain from the underlying aquifer. With subscripts denoting the following: *a* = aquifer, *b* = material above the crag; three scenarios (cases) were tested: **Case 1**: Peat on sand (crag); K_b = 0.01 (m/d) / K_a = 5.0 (m/d) thus a hydraulic conductivity ratio of 2×10^{-3}. **Case 2**: Peat on N.B.E; K_b = 0.01 (m/d) / K_a = 1.0 (m/d) thus a hydraulic conductivity ratio of 1×10^{-2}. **Case 3**: N.B.E on sand; K_b = 1 (m/d) / K_a = 5.0 (m/d) thus a hydraulic conductivity ratio of 2×10^{-1}

Table 6.11 Numerical simulation of various drainage scenarios

	m_b(m)		**DC** (m^3/d per m length)	
		Case 1 K_b/K_a =0.002	**Case 2** K_b/K_a =0.01	**Case 3** K_b/K_a =0.2
1	3.40	0.0204	0.0204	2.01
2	1.10	0.0390	0.0387	3.30
3	0.40	0.0800	0.0775	4.64
4	0.20	0.1380	0.1300	5.67
5	0.10	0.2460	0.2190	6.56
6	0.05	0.4480	0.3640	7.21
7	0.00	7.7700	1.4500	7.85
8	-0.40	8.5800	1.7200	8.96

A value of the hydraulic conductivity of peat of 0.01m/d was chosen so that the conductivity ratios would be three consecutive orders of magnitude (i.e. 10^{-1}, 10^{-2} and 10^{-3}). However, the results of Table 6.4 suggests that even if the peat has a hydraulic conductivity of < 0.01 m/d the effect on the drainage coefficient would not decrease significantly until the thickness of the basal material was less than 0.5m (Table 6.4).

The following interesting numerical details can be noted from such simulations:-

1. For m_b of 0.05m or more, the K_b (basal material) has the dominant effect, such that Case 1 and Case 2 have almost identical DC values.
2. At zero thickness (i.e. fully penetrating drains), the Drainage Coefficient is dependent on the chosen value of the hydraulic conductivity of the aquifer. (i.e. DC per unit length is approximately $1.5 \times K_a$, at zero thickness)
3. The relationship between the variables m_b and DC is either an exponent function or composite function, being influenced by the magnitudes of the two hydraulic conductivities K_b (basal material) and K_a (aquifer).

The first six points for each of the cases 1, 2 and 3 are plotted in Figures 6.12, 6.13 & 6.14

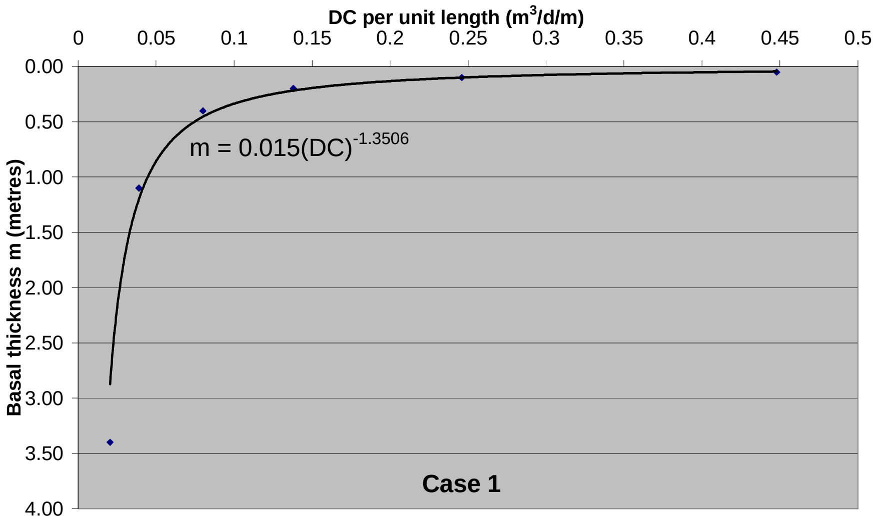

Figure 6. 12 Relationship between basal thickness and drainage coefficient for Case 1 (Hydraulic ratio 0.002)

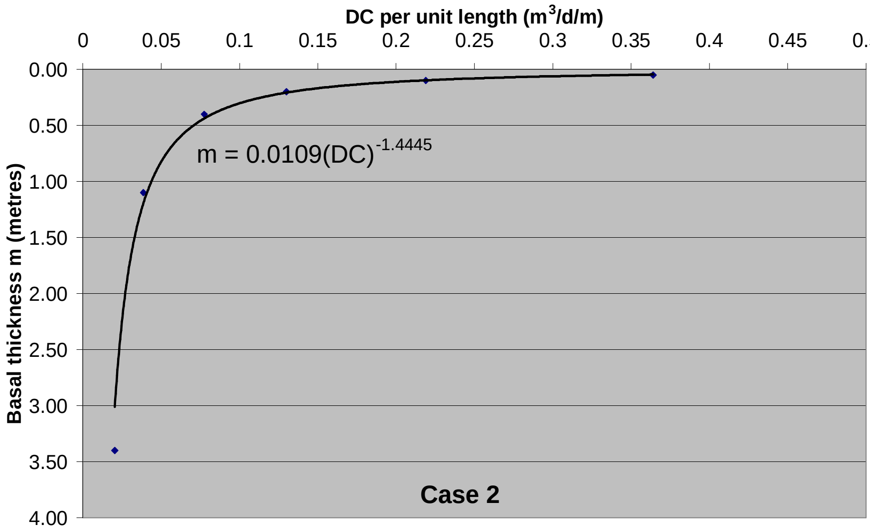

Figure 6. 13 Relationship between basal thickness and drainage coefficient for Case 2 (Hydraulic ratio 0.01)

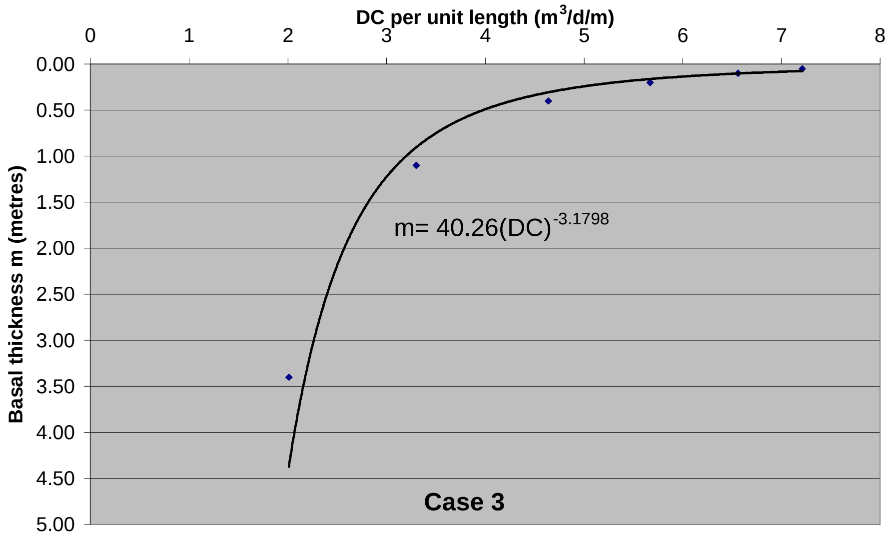

Figure 6.14 Relationship between basal thickness and drainage coefficient for Case 3 (Hydraulic ratio 0.2)

The allocation of drainage coefficients for individual cells must incorporate the length of the cell, which in this case is 177 metres. The actual thicknesses of the basal material for the drains throughout the catchment are not actually known but can however be approximated from data collected from various sources such as (Harding and Smith, 2002), (Burton and Hodgson, 1987) and fieldwork measurements taken in the process of this research project. In a pragmatic approach to drainage coefficient allocation, the basal thicknesses can be categorised to represent all the different levels of drain-aquifer interactions. From Table 6.4 Cases 1 & 2 it possible to define the drains as individually in the second column of Table 6.5, whereas the third column is the actual drainage coefficient for a cell of length 250/√2m.

Table 6.12 The approximation of drainage coefficients

m_b (m)	**DC** (m^3/d/m)	**DC** (m^3/d)	**DC*** (m^3/d)
> 3.4	-	-	1
3.4	0.02	3.6	1 or 5
1.1	0.04	6.9	5 or 10
0.4	0.08	14.1	10
0.2	0.14	24.4	10 or 50
0.1	0.25	43.5	50
0.05	0.5	79.2	50 or 100
0.00	$\approx 1.5 \times K_a$	1373.0	1000
-0.4		1516.1	1000

DC*=drainage coefficient used in the modelling domain

6.7 Summary

Within this chapter, the various hydraulic conductivities of the geological deposits are either deduced or approximated. The configuration of the modelling domain in plan and in cross section is complete along with mesh alignment and mesh spacing. The configuration of the coastal shelf reaches is presented and the estimation of potential recharge using a recharge factor approach is presented.

The previous research by Rushton and Tomlinson, (1979) concerning drainage coefficients showed that the combination of linear and non-linear functions of head difference provided the best representation of a drainage coefficient relationship with the flows into the drain. The current research reveals that the flows into the drain are dependent on the ration between the hydraulic conductivity of the aquifer material K_a and the hydraulic conductivity of the material between the base of the drain and the aquifer surface K_b. It showed also that the hydraulic conductivity of the basal material K_b is dominant for less penetrative drains, whereas the aquifer material K_a is dominant for drains that are more penetrative. It is possible, therefore, to represent a substantial variety of drain-aquifer interactions by considering values of only 1, 5, 10, 50, 100 and 1000 m^3/d/cell depending upon the level of drain connectivity.

7 THE IMPLEMENTATION OF A PRELIMINARY TIME-INVARIANT NUMERICAL MODEL

7.1 Introduction

The development of a time-invariant numerical model is described in this chapter from interpreting raw collected data to the performance of a first simulation with four specified target parameters (aquifer-drain water balance, coastal inflow water balance, particle tracking and the distribution of groundwater heads) and respective tolerance levels. The justification of the prioritisation of these four target parameters is also given. The theoretical basis of setting target parameters with tolerance levels is developed along with actual target values deduced from collected data. A series of output data analyses are made concerning the various sub-catchments and specified locations. These preliminary analyses will give guidance to the level of refinements required to improve the quality of subsequent simulations. In the following Chapter, after a sensitivity analysis has been made, refinements to the appropriate parameters can be made so that the final model better represent the data collected.

7.2 The introduction, prioritisation and estimation of target parameters and tolerances

7.2.1 Introduction

In modelling groundwater behaviour the two parameters that are often used to determine model performance are *groundwater heads* and *solute concentrations* (Zechner and Frielingsdorf, 2004), the latter in a coastal aquifer system, such as the upper Thurne, being considered as locations with high levels of saline concentrations (which depend on seepage fluxes). Zechner & Frielingsdorf (2004) found that in modelling the Biscayne Aquifer in south east Florida (i) the models calibrated solely on groundwater head measurements predict poorly regardless of the number of estimated input parameters. (ii) the models calibrated solely on seepage fluxes (i.e. the equivalent in the Thurne catchment are estimated drain flow volumes) predict accurately if the number of estimated parameters is restricted; and (iii) the models calibrated on at least two types of data (e.g. hydraulic head and seepage fluxes) exhibit good predictive capabilities.

In the upper Thurne catchment, readings are available of groundwater heads for 1991-93 (Holman, 1994). The solute concentrations (which depend on seepage fluxes) can be identified in the field by the groundwater component of the pump discharge, salinities of the pumped water and spot quality readings. Thus, four parameters have been identified as tools to assist with model refinement:

(i) groundwater component in the deep drains;

(ii) total coastal water inflow;

(iii) particle tracking and;

(iv) groundwater heads.

The modelling instrument known as *particle tracking* is a useful tool found in modelling packages such as PLASM (Prickett and Lonnquist, 1971) and MODFLOW (McDonald and Harbaugh, 1988). The groundwater pathlines leading to (forward) or from (reverse) a specific location can be determined.

7.2.2 The prioritization of target parameters

The concept of prioritising the various target values lays mainly in the philosophy that there are specific parameters that have less associated error attached to them under measurement. Typically *total water balances* are catchment–wide integral approximations and are generally robust and less affected by individual measurement error, whereas *head* readings are typically individual point readings that reflect a very local measurement. In the case of the upper Thurne the groundwater heads are water levels measured at a variety of observation wells. The four parameters under analysis in decreasing priority is:

1) Aquifer-drain water balance
2) Coastal water-drain water balance
3) Forward/Reverse particle tracking
4) Well water levels

The following section explains how the magnitude of the aquifer-drain water balance is estimated.

7.2.3 The estimation of a target value for the groundwater contributions (for Brograve) using a hydrograph separation technique

The volume of water in any of the catchments' drains consists of two components; namely, a rainfall-runoff component and (baseflow) groundwater contributions. The drainage pump discharge volumes calculated from the weekly electricity usage are the only volumetric estimates available of the total discharge (Section 4.2.6). A hydrograph separation technique for the sub-catchments of the Upper Thurne is required so that an adequate estimate of the groundwater contributions from the underlying crag aquifer to the drainage pump discharge can be made.

For most hydrographs, it is assumed that some of the stream discharge would have occurred even without the rainfall, this is usually called the baseflow (Beven, 2004). The method of hydrograph separation is intended to differentiate the rainfall-runoff component from the baseflow. Many workers carry out hydrograph separation but Beven's negative and pessimistic view, "One of the best ways to deal with baseflow separation is to not bother with it at all"(Beven, 2004) may give an insight as to how difficult a hydrograph separation can be. The problem is that there are no satisfactory techniques for hydrograph separation (Beven, 2004) for stream flows. In the case of the Upper Thurne, where there are drains as opposed to streams, springs and seepages, the baseflow component is really a groundwater contribution to the drainage pump discharge. The rainfall storm runoff is now equivalent to all the flows that are not groundwater contributions, thus making it necessary to identify the origins of the waters within the pump discharge waters:

1. Groundwater up flowing into the deep drains;
2. The coastal groundwater inflow, that enters into the deep drains;
3. The rainfall runoff that enters from the cultivated layer;
4. The runoff water collected by the spring drains that trace along the perimeter of the marshes.

The groundwater up flowing into the deep drains and the coastal groundwater inflow into the drains constitute the groundwater contributions. The inflow into the drain is expected to be proportional to the difference between the water surface level in the drain and the underlying groundwater head (Equation 6.3). For a deep drainage node within the model, the flow from crag aquifer to the drain is $Q_D = DC \times (H - h)$ where DC = is the drain coefficient, H = the head at the base of the drain and h = the head with the aquifer. For the drains upstream of the Brograve pump, the total groundwater inflow is $\Sigma Q_D = \Sigma DC \times (H - h)$. The groundwater head hydrographs are similar throughout the Brograve catchment (Figure 3.7) so the equation can be expressed as

$$Q_{gwc} = P \times (H - h_{gwc}) \tag{7.1}$$

where P is a constant of proportionality m^2/week. Since the value of P is not known, it can be found by considering a time period within the two years worth of readings where the flow is at a minimum and almost the entire pump discharge contribution is from the groundwater. The length of the time period chosen will affect the value of P.

Two alternatives exist:

(i) use one week's reading to estimate minimum flow (**Case A**)

$$Q_{\min} = P_1 \times (H - h_{\min}) \tag{7.2}$$

or (ii) use several weeks' readings to estimate minimum flow (**Case B**).

$$\overline{Q_{\min}} = P_2 \times \overline{(H - h_{\min})} \tag{7.3}$$

Note that h_{min} is the lowest groundwater head but it is always higher that H

P_1 and P_2 are constants of proportionality for each equation respectively. Any subsequent increases in head (due to recharge) throughout any given time period will result in a proportional increase in groundwater flow into the reach, which can be written as

$$Q_{gwc} = P^* \times (H - h_{gwc}) \tag{7.4}$$

where P^* is either P_1 or P_2 depending on which method of approximation was chosen. The minimum flow calculated (using the drainage pump conversion factor for the Brograve pump of 75 (kWhr/m^3)

Case A

H (is the water level in the drain) -1.9 mOD,

$Q_{\min}$ = 9367.5 m^3/week (week 22)

$h_{\min}$ = -1.73 m, (head within the local aquifer, based on reading at Lower Brograve in week 22)

P_1 = 9367.5/(-1.9-(-1.73)) ≈ 55,100 m^2/week

Case B

H =-1.9 mOD,

$\overline{Q_{\min}}$ ≈ 11,000 m^3/week (average flow volume for the lowest **three** weeks, (weeks 22, 23, and 24)

$\overline{(H - h_{\min})}$ = -0.17m, (average head difference for the lowest **three** weeks, based on reading at Lower Brograve)

P_2 =11,000/(-1.9-(-1.73)) ≈ 64,705 m^2/week

In **Case A**, where the flow volume from week 22 was chosen to estimate the groundwater contributions to the drain, it was found that the total groundwater component over the period April 1991 to March 1993 was equal to 2,332,273 m^3 . (This is given by the area under the dashed line in Figure 6.1) and the total volume of water discharged from the Brograve pump for the entire two years was 4,653,743 m^3. Therefore, the groundwater component constituted $\frac{2,332,273}{4,653,743} \approx 0.50$ of the total discharge volume.

In **Case B**, where the mean flow volumes from weeks 22, 23, and 24 were chosen to estimate the groundwater contribution, it was found that the groundwater component was equal to 2,738,832 m^3 over the two years (as given by the area under the solid red line in Figure 6.1) $\frac{2,738,832}{4,653,743} \approx 0.60$ of the total discharge volume.

It is also important to consider the effect that the given value of H has upon the estimation of groundwater contribution to the discharge volumes. **Case B** (Figure 6.1, encircled) matches well with the total volume discharge only between the 14th and the 26th week.

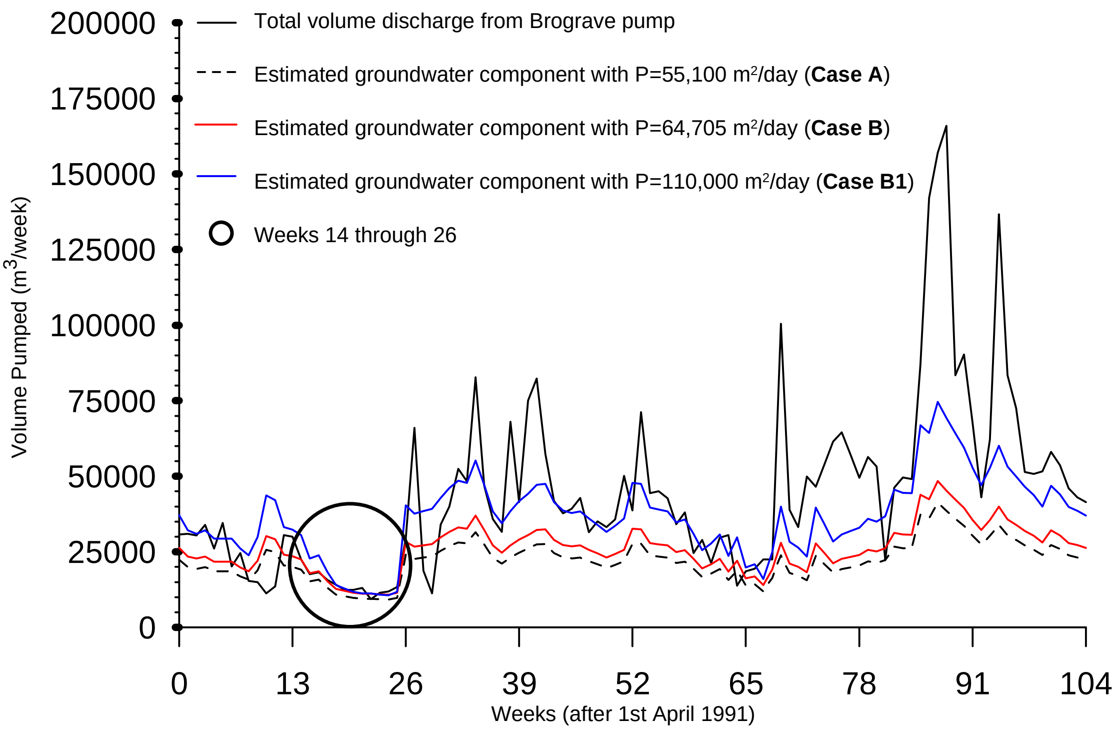

Figure 7.1 The hydrograph separation of the discharge waters of the Brograve pump

Case B1

For Case B, the mean head difference used, $\overline{(H - h_{\min})}$ = -0.17 m, is based on the measured head differences in the lower reaches of the Brograve sub-catchment. However, in the extensive upper reaches of the sub-catchment the head difference is likely to be smaller due to the higher water levels, and consequently a smaller average head difference for the catchment of $\overline{(H - h_{\min})}$ = –0.01 m might be appropriate.

Such a head difference gives P = 110,000 m^2/week and subsequently the groundwater contributions are estimated to $\frac{3{,}847{,}580}{4{,}653{,}743} \approx 0.83$. However, this gives a substantial departure from the total volume hydrograph in the low flow period (weeks 14 through 26 in Fig. 7.1) suggesting a more accurate profile would render a lower overall groundwater contribution. A pragmatic estimation is to assume a groundwater contribution of 70% (which is roughly halfway between **Case B** and **Case B1**), which equate to an average daily discharge of groundwater from the Brograve pump of 4,439 m^3/day (Table 7.1). The other sub-catchments were attributed to have 40% groundwater contributions (see Table 7.1), as this is an approximate equivalent for baseflow contributions in regular stream–aquifer interactions.

The level of tolerance attached to the average daily discharge of groundwater from the pumps is taken as a percentage of its value; in this case a tolerance level of 20% is given. Thus the target equation for water entering the aquifer is (Q = Quantity).

$$Q_{\text{(modelled)}} = Q_{\text{(measured)}} \pm 20\% \times Q_{\text{(measured)}} \tag{7.5}$$

Table 7.1 was derived from the weekly electricity and salinity reading collected by Holman (1994).

7.2.4 Estimating the saline contribution in the groundwater component of the pump discharges

The direct coastal inflow into the groundwater system is very complex and there is no physical way of measuring it. The only measurements collected are the salinities of the waters entering the individual mouths of the various discharge pumps.

Holman (1994) estimated the seawater-equivalent proportion of pump discharge by measuring the chloride concentration, and carrying out a weekly *End-Member-Mixing-Analysis*, with chloride end-members of drain water and seawater. These percentages are listed in Table 7.1, column 5. At pumps where salinity within drainage water samples was considered to be due to leakage through the flood banks from the River Thurne, the leakage values are not considered within the coastal water balance.

The task of estimating the amount of coastal waters entering the aquifer is similar to the aforementioned estimation of the total daily quantity of water entering the aquifer. The coastal daily quantity target equation can be written as (*C.I.Q.* = Coastal Inflow Quantity)

$$C\,I\,Q_{\text{(modelled)}} = C\,I\,Q_{\text{(measured)}} \pm 20\% \times C\,I\,Q_{\text{(measured)}} \qquad (7.6)$$

Although in this case the tolerance condition is only valid in the same direction as the total daily water inflow discrepancy. (i.e. if the total daily inflow is underestimated by 7% then the coastal inflow will only be acceptable for 0-10% underestimation). This additional constraint is to minimise the variation of the relative proportions of coastal water to total water.

Table 7.13 Target values for average daily total groundwater flow and coastal inflow waters into the sub-catchment (m^3/day)

Pump	Average daily discharge*	Estimate groundwater contribution (%)	Average daily groundwater inflow*	Seawater-equivalent % of daily discharge	Average daily Seawater-equivalent inflow (m^3/d)
Brograve	6342	70	4439	17	1070
Catfield	1063	40	425	5	51
Eastfield	4814	40	1929	5	253
Horsey	2285	40	914	10	230
Ludham	1555	40	622	-	-
Martham	5468	40	2187	-	-
Martham Holmes	1455	40	582	-	-
Potter Heigham	2822	40	1129	-	-
Repps	2488	40	995	-	-
Stubb	2965	40	1186	-	-
Thurne	1972	40	789	-	-
Somerton New	3913	40	1585	10	403
Somerton Old	3935	40	574	4	168
Gravity_3	-			-	-
Total			**18333**		**2175**

- these values were not incorporated into the total amount as they were considered to be leakage volumes by Holman (1994).

7.2.5 Estimating model performance with the aid of particle tracking

The final parameter used to estimate model performance is particle tracking. The use of particle tracking needs to be quantified within the modelling procedure so that the movement of coastal and non-coastal source waters can be simulated and the model performance can be measured. The procedure for estimating the locations of high saline inflow is as follows:

(i) The particle tracking source line is placed along the coast.

(ii) Location are considered 'fresh' if particles from the coast do not reach the inland monitored location.

(iii) If the particle reaches the inland monitored location, then location is considered saline.

(iv) If for the separate summer and winter conditions (Section 8.7), the particle reaches in summer but not in the winter then the location can be considered brackish.

It is important to note that the coast is not the only source of eastward inflow to the groundwater system, there are recharge areas in locations quite close to the coast which will mix with any saline inflow.

7.2.6 Setting groundwater head target values and tolerance levels

The groundwater heads were estimated using a target of the *average* readings from observation wells (Figure 3.11) over the period April 1991 – March 1993. Thus, the target equation for the groundwater heads reads as follows:

$$H_{\text{(modelled)}} = \text{Average } H_{\text{(measured)}} \pm 0.8\text{m} \qquad (7.7)$$

7.3 Input details for the initial model

7.3.1 Auxiliary input details

The model configuration details are developed in Chapters 4 and 5. The following details are recalled to remind the reader of the characteristics of the modelling domain. The mesh spacing is 250/√2 m×250/√2 m with a mesh alignment of 45^{o} (Sections 6.3.2 & 6.3.3.). The model has five layers including an auxiliary Layer 2 of 1 metre thickness (Section 6.4.2.). The use of the General Head Boundary for Layer 1 is restricted to the cells in Layer 1 above the alluvium but not to cells adjacent to the Norwich Brickearth (Section 6.4.5.). The Broads are also modelled using the General Head Boundary condition (Section 6.4.6.). The coastal shelf is modelled using drain cells that are cell-wide and have no coverage of drain bed sediment (Section 6.4.4.).

The individual recharge rates for the various landscape units are presented in Table 7.2. The table includes the number of cells attributed to the individual landscape units.

Table 7.14 The initial recharge rates for the catchment

Landscape Units	Recharge rate (m/day)	cell area (m^2)	No. of cells	Total m^3/day
Flegg Hundreds	0.00029	31250	545	4939
Norwich Brickearth	0.00034	31250	1100	11688
Exposed Crag*	0.000405	31250	558	7062
Total				**23689**

The alluvium is not assigned any recharge as the existence of General Head Boundary conditions in Layer 1 above both the peat and the clay provides the alluvium with the residual recharge. The initial input values of hydraulic conductivities for the preliminary model are given in Table 7.3.

Table 7.15 The initial values of hydraulic conductivity

Geology	Hydraulic Conductivity (m/day)
Altcar	0.0016
Newchurch	0.0008
Norwich Brickearth	1.000
Flegg Hundreds	0.3
Exposed Crag	20.0
Clayish Layer	5.0*

* Amended under model refinement.

7.4 Analysis of output information for the Preliminary Model

7.4.1 Introduction

In this section, output tables and plots for the *preliminary* model are presented and analysed with regard to the parameter targets. The four output tables represent the four previously identified model targets; namely the aquifer-drain water balance, the groundwater coastal inflows, the areas vulnerable to saline inflow and the distribution of groundwater heads. Once the appropriate refinements have been made to the preliminary model, further analysis is made into the adequacy of the groundwater model (this will be done in the following chapter).

7.4.2 Analysis of aquifer-drain flow balance

The aquifer-drain water balance combines Inflows, which include the rainfall recharge at the outcrop, the coastal inflows, and the residual vertical flows through the peat and clay, whereas the Outflows include the fresh groundwater outflows to sea.

The results presented in Table 7.4 show that the Preliminary model has output values for each of the individual drainage sub-catchments that lie within the 20% tolerance level; the only exception to this is the Brograve catchment (Table 7.4). The combined total volume of water estimated to have entered the Crag aquifer by the simulation is 27,007 m^3/day, which is more than the targeted 18,333 m^3/day (Table 7.4). The difference in flows between the modelled value and the target value is mostly caused by the discrepancy of the Brograve values. The modelled estimate of the average Brograve drainage inflows is 11,147 m^3/day, which is approximately 7,800 m^3/day over the target value of 4,463 m^3/day. This over-estimate of the volume of groundwater flowing out of the aquifer into the drains of the Brograve sub-catchment can be rectified with an adjustment of the total sub-catchment drainage coefficient from 18,900m^2/day (sum of column 3) to approximately 6,600 m^2/day.

Table 7.16 Output from the preliminary model showing the various sub-catchment groundwater balances

Catchment	No. of cells	DC per cell [m^2/day]	Simulated Flux into drains [m^3/day]	Simulated Flux out of drains [m^3/day]	TARGET flux [m^3/day]
	7	140		250.91	
	3	0.25	0.47		
	228	1	165.84		
	2	3	6.67		
	3	35	74.52		
Brograve	1	50	43.34		
	2	62	97.11		
	1	68	89.05		
	89	140	7697.46		
	16	300	2464.28		
	3	355	508.44		
Total			**11147.18**	**250.91**	**4439**
Catfield	5	10	55.05		
	3	100	347.12		
Total			**402.17**		**425**
Martham	9	1	4.17		
Holmes	20	100	697.8		
Total			**701.97**		**582**
	94	1	62.96		
Eastfield	30	100	1865.48		
Total			**1928.44**		**1926**
	16	1	16.81		
Ludham	7	100	761.47		
Total			**778.28**		**622**
	84	1	22.92		
Horsey	30	50	977.67		
Total			**1000.59**		**914**
	33	1	28.95		
Martham	25	100	2128.06		
Total			**2157.01**		**2187**
Somerton	36	1	74.88		
New	11	100	2002.51		
Total			**2077.39**		**1565**
Potter	41	1	31.54		
Heigham	13	100	1308.71		
Total			**1340.25**		**1129**
Repps	27	1	19.85		
	17	100	1245.2		
Total			**1265.05**		**995**
	38	10	205.79		
Stubbs	22	100	1507.92		
Total			**1713.71**		**1186**
Thurne	18	1	10.68		
	14	100	804.76		
Total			**815.44**		**786**
Somerton	23	1	42.95		
Old	9	100	1584.6		
Total			**1627.55**		**1574**
Gravity_3	16	10	52.89		
TOTALS	**996**	**2344.25**	**27007.92**	**250.91**	**18,333**

7.4.3 Analysis of coastal inflow

Due to the complex nature of the coastal flows, their contributions to the individual sub-catchments are very difficult to replicate. The total target value of 2,175 m^3/day is a composition of coastal waters from the full length of the coast. A time-invariant simulation should at least be able to approximate the total amount of coastal inflows entering the sub-catchments that are vulnerable to saline inflow. (i.e. all the coastal sub-catchments together with the Eastfield and Stubbs sub-catchments).

The coastal reaches within the model have been labelled to reflect the landward sub-catchment that they individually border – this labelling does not assume that all of the coastal inflow from the reach enters the drains of those bordering landward sub-catchment. The modelled 6,042 m^3/day of coastal inflow waters are much higher than the total target of 2,175 m^3/day. However, the simulated average daily coastal saline inflow should decrease with the lowering of the Brograve sub-catchment drainage coefficients suggested in the previous section .

Table 7.17 The simulated average daily coastal inflows

Coastal Region	Inflow (m^3/day)	Outflow (m^3/day)	Target (m^3/day)
Happisburgh	0	-1097	0
Hempstead	1196	0	0
Brograve Level	3066	0	0
Horsey	1228	0	0
Somerton New	511	0	0
Somerton Old	41	0	0
Grand Total	**6042**	**-1097**	**2175***

* includes coastal waters that are pumped from Eastfield and Stubbs pumps

7.4.4 Analysis of particle tracking (pathlines)

The ability to replicate the consistently fresh and consistently saline (coastal-fed) drainage reaches is the most realistic criteria that a time-invariant model can be expected to replicate. Figure 7.2 show that particle tracking indicates that the Upper Lessingham drains are fed by fresh groundwater, which is consistent with fieldwork observations (Section 4.2.4). The Horsey subcatchment (brown) is shown to have direct coastal waters entering its drains and itself to be providing recharge to the aquifer beneath the peat layer in the Brograve Level. Coastal waters are entering beneath the peat towards the brackish Eastfield drains). However, although the particle tracking demonstrates that the model is replicating the observed understanding of groundwater flowpaths, the model needs some local modifications. In particular, the current time-invariant model needs modifying to capture the more saline drains within the Hempstead marshes (cells **40, 60** and **40, 61**) which are currently simulating only fresh water sources.

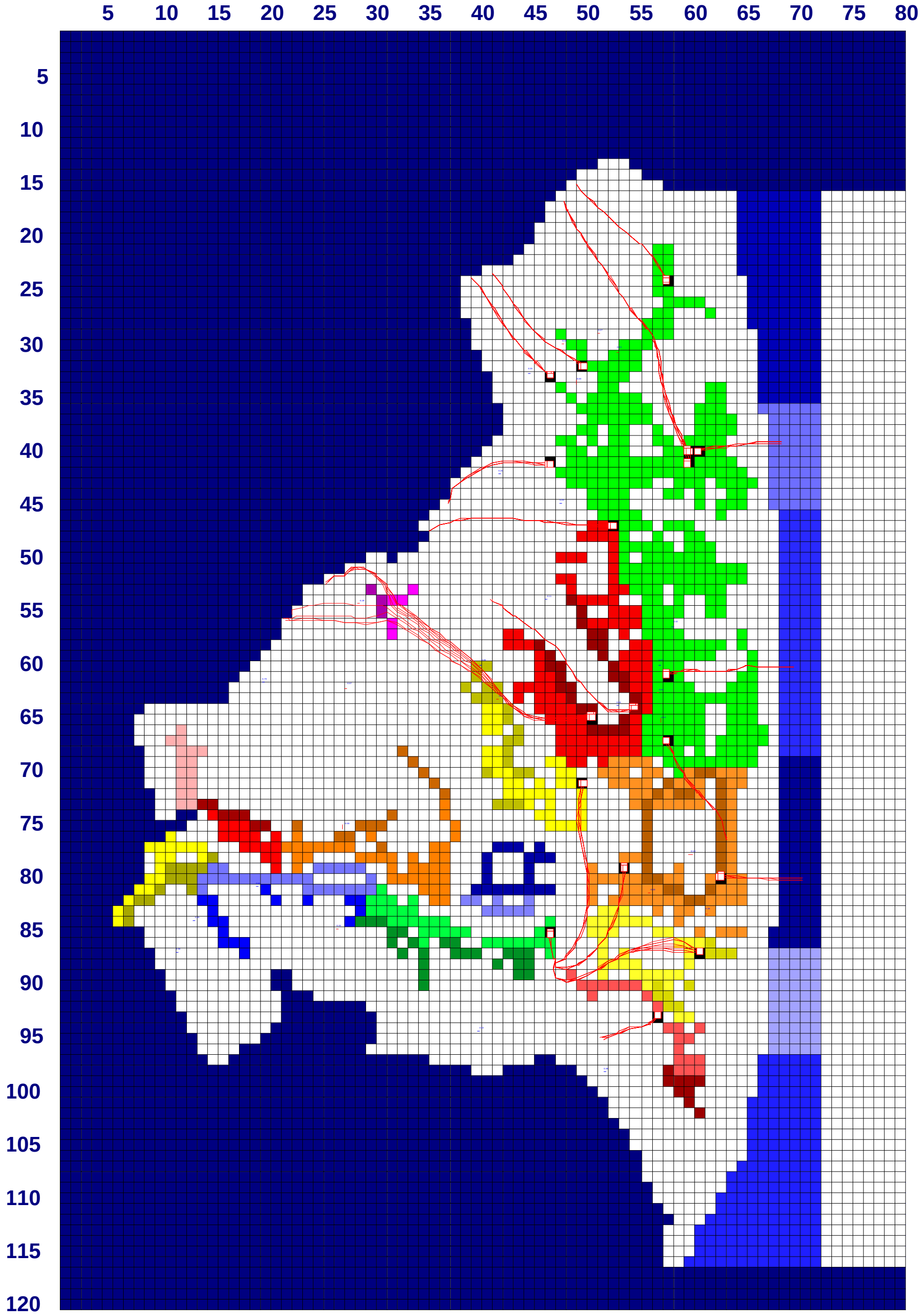

Figure 7.2 The output screen depicting the various particle tracks for Layer 2

The particle tracking destination locations in Figure 7.2 are given in Table 7.6 along with the source direction and the location of their individual sources. It is noticeable that the individual source locations all have heads greater than their respective destination location.

Table 7.18 A table of specified locations detailing reverse particle tracking

Particle destination		*Particle Source*		*From source*	*Model*
Node	**Head**	**Node**	**Head**	**Direction**	**SOURCE**
24,58	0.72	**15,50**	1.77	NW	fresh
32,50	0.47	**24,42**	1.58	NW	fresh
33,47	0.57	**24,40**	1.6	NW	fresh
40,60	-0.83	**16,48**	1.77	NW	fresh
40,61	-0.86	**39,69**	0.1	E	saline
41,47	0.03	**45,37**	0.84	W	fresh
41,60	-0.92	**16,48**	1.77	NNW	fresh
47,53	-0.54	**47,36**	0.91	W	fresh
61,58	-0.68	**60,70**	0.1	E	fresh
65,51	-0.46	**52,26**	1.37	W	fresh
67,58	-0.60	**76,63**	-0.21	SE	saline
71,50	-0.48	**88,43**	0.19	S	fresh
79,54	-0.34	**88,43**	0.19	S	fresh
80,63	-0.18	**80,71**	0.1	E	saline
85,47	-0.10	**88,43**	0.19	S	fresh
87,61	-0.39	**88,43**	0.19	W	fresh
93,57	-0.24	**95,52**	0.06	SW	fresh

7.4.5 Analysis of water levels

The distribution of the observation wells that were shown in Section 4.4 varied in their individual response to recharge and reveal a local distribution of heads. Table 7.7 lists observed average water level measurements at a selection of the wells for both the confined (alluvial) and unconfined (Norwich Brickearth and Flegg Hundreds) aquifer areas and the simulated average head for the node in which the well is located. Although the model may require adjustments, the head distribution does reflect the major highs (Walton House) and lows (Lower Brograve) within the catchment.

Table 7.19 Observed average groundwater heads at selected wells and the modelled head value for the node

Node	**Target groundwater head (mOD)**	**Modelled groundwater head (mOD)**	**Residual (m)**	**Name**	**Confinement**
60,40	**0.08**	0.03	0.05	Broad House	Unconfined
45,48	0.22	-0.03	0.25	Calthorpe Farm	Unconfined
54,47	**0.00**	-0.11	0.11	Conifers	Unconfined
64,53	**-0.79**	-0.83	0.04	Eastfield	Unconfined
31,53	0.15	0.38	-0.23	Grange Farm	Unconfined
60,57	**-0.77**	-0.84	0.07	Lambridge Farm	Confined
65,57	-1.34	-0.82	-0.52	Lower Brograve	Confined
42,42	0.80	0.56	0.24	Mill Farm	Unconfined
87,12	0.44	0.08	0.36	School	Unconfined
78,60	**-0.40**	-0.27	-0.13	Street Farm	Unconfined
62,28	0.55	1.22	-0.67	Sunny Side	Unconfined
61,20	2.17	1.38	0.79	Walton House	Unconfined

The head distribution in Layer 2 is presented in Figure 7.3; the overall distribution of heads is not unrealistic. The lowest simulated head values are found in the Marshland (<-0.5mODN), whereas the higher simulated heads were found in the western Loam Uplands (>1.50mODN).

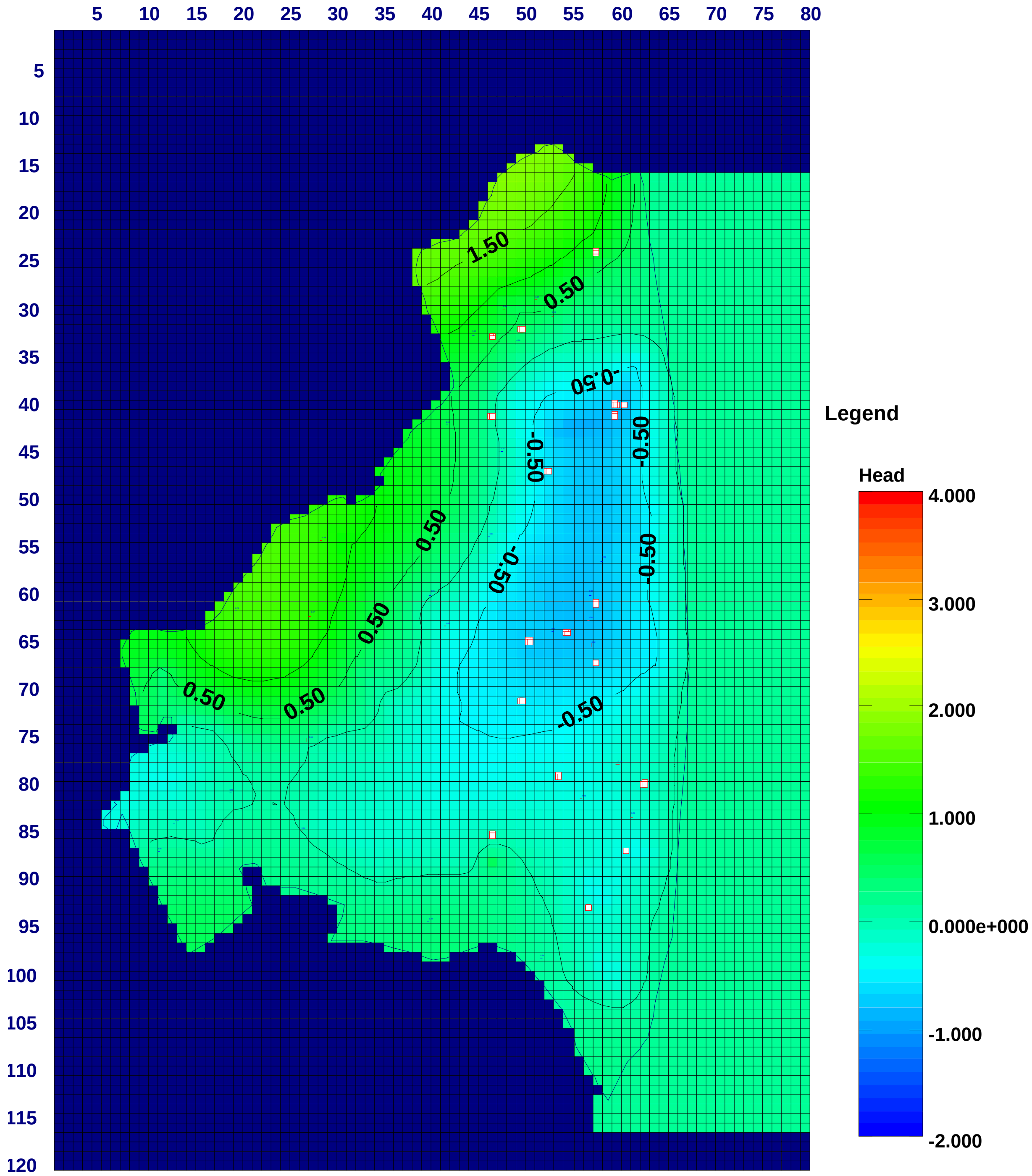

Figure 7.3 The output screen depicting the head distribution

7.5 Modelling Conclusions

The most obvious conclusion stemming from the current simulation is that there is an over estimation of the groundwater flows within the Brograve sub-catchment with modelled value of 11,147 m^3/day compared with the estimated measured value of 4,439 m^3/day. Reductions of specific drainage coefficients is clearly required to reflect a more realistic volume of groundwater entering it and this must be the priority in the further investigations. The spatial distribution of coastal water inflows can only be correctly investigated by varying the rainfall recharge to determine whether the brackish locations obtain coastal waters under reduced rainfall recharge circumstances. Once these refinements have been implemented and the water balances (fresh and saline) are representative then the distribution of head values can be re-examined.

The modelled values even for this initial simulation suggest that some sub-catchments have the correct magnitude of drainage coefficients. Catfield, Horsey, Ludham, Martham, Martham Holmes, Potter Heigham, Repps, Thurne, Somerton Old and Somerton New sub-catchments all have modelled values groundwater-to-drain inflow that fall within the 20% tolerance level criteria. This level of agreement for the initial simulation is promising.

7.6 Summary of the modelling process thus far

The four stages of the modelling process (Table 7.8) are the collection of *field measurements*, assessment of the usefulness or *scope of measurements*, the development of *conceptual models* using theory and collected data, and finally, the development of *computational models*. The four target parameters (groundwater flow into drains, the coastal inflow, particle tracking, groundwater heads) are analysed to understand better the possible sources of model inadequacies.

The first stage is the collection of Field Measurements, which allows such modelling goals as the water balance to be attempted. In the case of the Upper Thurne catchment, the discharge volumes are not directly measured but are deduced using various electrical pump conversion factors, for which (Table 4.2) quotes four different values for the Brograve pumping station. The total coastal saline volumes are also not directly measured but are deduced from drain water samples collected from near the various electrical pumps. The inaccuracy of these measurements is masked by the inaccuracy of the total discharge volume calculation. The saline deep drains are easier to identify using spot electrical conductivity or chlorinity readings although the actual locations of saline inflow cannot easily be determined due to mixing within the drains. Head measurements are accurately measured but reflect the time-variant behaviour of the aquifer and are each individually strongly influenced by local groundwater movement. Drain coefficients are impossible to determine accurately but with the introduction of the *magnitude-determined* DC developed in Section 6.6 allows for a reduction in the inaccuracies as the level of contact the drain has with the underlying aquifer is classified into four or five single-valued categories. The conceptual and computational model assumes a level of uniformity in recharge behaviour and geological structure, which is usually non-uniform.

Table 7.20 The process of modelling from Field Measurements to Computational Models

Parameter	Field Measurements	Scope of Measurements	Limit of Conceptual Model	Limit of Computational model
Discharge volume (surface water flows)	Only electricity readings from discharge pumps.	Deduced volumes very dependent on conversion factors and the estimated groundwater contributions fractions.	Scope of aquifer and drain contributions assumed to be constant.	Only g/w proportion considered. The hydraulic conductivity of the outcrop assumed to be constant for each individual deposit
Saline volume (coastal inflow)	Nil direct measurements	Saline waters and leakage waters are indistinguishable	Actual (hydro)geology along the coast is not exactly known.	The geology along the coast is assumed to be uniform.
Saline-prone drains (locations of saline inflow into drains)	Nil direct measurements Occasional spot salinity measurements within drains.	Assumes spot reading locations are the actual locations of saline inflow.	Limited mixing. The actual influx of saline waters are not known	The saline inflow into drains is not modelled, estimations are made using varying fresh/saline water proportions.
Heads (groundwater levels)	Accurately measured, but reflect the time variant nature of the g/w fluctuations.	Very dependent upon LOCAL aquifer properties. Quality of measuring instrument.	Replicating time-invariant behaviour. Simplification of aquifer properties and geometry.	Aquifer variability not captured. Mesh spacing assumes aquifer uniformity throughout
Drain Coefficient (drain-aquifer interactions)	Based solely upon drain-aquifer Theory, difficult to determine Accuracy.	The amount of water discharged not exactly known due to imprecise conversion factors. Groundwater proportion estimated using theory.	Replicating time-invariant behaviour. Assumes g/w to s/w proportions are constant. No actual flow measurements	Only g/w proportion considered.

8 MODEL REFINEMENT, SENSITIVITY ANALYSES AND A SIMULATION OF SUMMER AND WINTER CONDITIONS

8.1 Introduction

In the previous chapter, a Preliminary time-invariant simulation of the groundwater flow in the Thurne catchment was implemented and the output results presented. The volume of groundwater entering into the drains within the Brograve sub-catchment was over-estimated by the model by around 6,708 m^3/d. Therefore, within the first part of this chapter various refinements are made to the parameterisation of the model to better represent the estimations deduced from field information. These refinements include the reduction in the overall drainage coefficient (Scenario 1) for the Brograve sub-catchment and the repositioning of the drains with the larger drainage coefficients (Scenario 2).

In the second part of this chapter, various sensitivity analyses are completed on the refined groundwater model (Scenario 1) to gain insights into the relationships between model input parameters and model output results. In order to distinguish these adjustments, the conditions and variations that are made to this groundwater model are listed in Table 8.1. The *scenarios* are the sensitivity analyses involving the uncertainty of specific input parameters or a representation of the average winter and average summer, whereas the *predictions* in Chapter 9 are intended to give insight into the behaviour of the groundwater system after the implementation of a salinity management proposal. All scenarios and predictions (except for Scenario 2) are simulated using the Brograve drain configuration of Scenario 1.

Table 8.21 The Output data analysis from various sensitivity analyses (where X denotes completion)

Variation	Name of run	Described in	Heads	Groundwater balance	Coastal Inflow	Particle Tracking
Large DCs	Preliminary	Section 7.4.2		Brograve	X	
Reduced DCs	Scenario 1	Section 8.2.3	X	All Sub-catchments	X	
Rearranged DCs	Scenario 2	Section 8.3	X	Brograve		
K-Crag	Scenarios 3,4	Section 8.5.1			X	
K-Clayish Layer	Scenarios 5, 6	Section 8.5.2			X	
K-alluvium	Scenarios 7,	Section 8.6	X			
Doubled DCs	Scenario 8	Section 8.6		All Sub-catchments	X	
Summer Conditions	Scenario 9	Section 8.7		All Sub-catchments	X	X
Winter Conditions	Scenario 10	Section 8.7		All Sub-catchments	X	X
Water Levels	Prediction 1	Section 9.2		Brograve	X	
Interceptor Drains 1,2	Predictions 2,3	Section 9.3		Brograve	X	
Lining Hempstead	Predictions 4	Section 9.4	X	Brograve	X	X

8.2 Model refinement by adjustments to the initial Brograve drainage coefficients

8.2.1 Introduction

In the previous chapter, a comparison of simulated (11,147 m^3/day) and estimated (4,439 m^3/day) average daily pump discharge volumes revealed a discrepancy, which is likely to have been related to the estimations of the drainage coefficients for the Brograve catchment. Table 8.1 represents the Brograve drain-aquifer water balance for the initial model. The values of drain coefficients were derived using estimated drainage coefficients obtained from the drain coefficient basal thickness equations developed in the Section 6.4. The drain coefficients must now be refined to produce an average aquifer outflow total that is comparable to the estimated 4,439 m^3/day.

Table 8.22 The Brograve sub-catchment aquifer-drain water balance for the Preliminary Model

Sub-catchment	No. of cells	DC [m^2/day]	Out of aquifer [m^3/day]	Into aquifer [m^3/day]	Target [m^3/day]
Brograve	3	0.25	0.47	0	0
	228	1	165.84	0	0
	2	3	6.67	0	0
	3	35	74.52	0	0
	1	50	43.34	0	0
	2	62	97.11	0	0
	1	68	89.05	0	0
	89	140	7697.46	0	0
	16	300	2464.28	0	0
	3	355	508.44	0	0
Total		**1014***	**11147**	**251**	**4439**

*does not include the inflowing drain

8.2.2 Method of model refinement

During the development of the Preliminary Model, it was necessary to select preliminary parameter values at certain locations, which were selected with only an elementary understanding of the properties and flow mechanisms of the aquifer system at those locations. Therefore, it is necessary to review some of these parameter values to refine the model. Results in Chapter 7 (Table 7.4) show that aquifer-drain flows are represented adequately in all sub-catchments apart from in the Brograve sub-catchment. Before proceeding with any parameter adjustments to try to improve the simulation of the groundwater component of the Brograve pump discharge, it is necessary to review the likely accuracy of the target value. The groundwater component of the Brograve pump discharge is based on:

(i) An electrical conversion factor to estimate the actual discharge for the pump from the electricity consumption (Section 4.2.7)

(ii) A procedure for estimating the proportion of the Brograve pump discharge which is due to groundwater flow into the drains. (Section 7.2.3)

There are uncertainties in the electrical conversion factor as Table 4.2 quotes three alternative factors. Making the assumption that the flows deduced from the current conversion are only 70% of the correct value, the target value would then become (100/70) × 4439 ≈ 6300 m^3/d. Since the area of the Brograve catchment plus the outcrop that drains into it is at least three times the area associated with Eastfield pump, it is realistic to assume that the groundwater component will be about three times the Eastfield value (i.e. about 6000 m^3/d).

Considering point (ii) above, it is unlikely that the proportion of groundwater discharge will be more than 70%. To achieve a reduction in the groundwater flows into the Brograve catchment from 11,150 m^3/d, a reduction in the drainage coefficients is necessary: this is described as Scenario I. For the preliminary Model, Table 8.1 shows that the sum of the drainage coefficients is 18,908 m^2/d. In Scenario I, the cells with the original drainage coefficients of 355 m^2/d, 300m^2/d and 140m^2/d are reduced to 50 m^2/d, and those with coefficients originally 50 m^2/d are reduced to 10 or 5 m^2/d, giving a summed drainage coefficient of 6,200 m^2/d (Table 8.7). The groundwater flow into the drains of the Brograve sub-catchment then equals 5863 m^3/d. These changes in drainage coefficients are possibly marginally too severe given the inherent uncertainty in the electric conversion factor for the Brograve pump.

8.2.3 Results

The aquifer-drain water balance for the entire Thurne catchment is presented in Table 8.3. The model estimates for both the inland drains and the Coastal inflow falls within the defined tolerance levels of 20%. However, the volumetric flows into the deep drains are over-estimated, whereas the flows into the coastal drains are under estimated. The model ratios for the coastal inflow/ (deep drains + Cultivated layer) is 12.4% (=3,081/ [20,326+4458]) (Table 8.3) whereas the estimated ratio from Table 7.1 is 12% (= 2,175/18,333), which is satisfactory.

Table 8.23 The aquifer drain water balance table (m^3/day)

Groundwater	Into Aquifer	Out of Aquifer
Inland drains	36	20326
Coastal inflow	3081	2086
Lakes	1098	1035
Cultivated layer	1	4458
Recharge	23689	0
Total	27905	27905

The deep-drain water balance in Table 8.4 shows that the modelled values for individual sub-catchments are mostly within the target tolerance levels. The only sub-catchment that falls outside the tolerance levels is Brograve. The existence of drains in Brograve that have flows *entering* the aquifer may not be verifiable, but due to its relatively small magnitude, their contribution to the total aquifer-drain water balance can be ignored.

Table 8.24 The average daily flow deep inland drain water-balance table (m^3/day)

Inland drains	Into Aquifer	Out of Aquifer	Within target tolerance range*
Brograve	36	5863	no
Catfield		454	**yes**
Martham Holmes		474	**yes**
Eastfield		2158	**yes**
Ludham		708	**yes**
Horsey		1006	**yes**
Martham		1947	**yes**
Somerton New		1647	**yes**
Potter Heigham		1144	**yes**
Repps		1196	**yes**
Stubbs		1231	**yes**
Thurne		841	**yes**
Somerton old		1597	**yes**
Gravity		60	n/a
Total	**36**	**20326**	**yes**

*Within +20% tolerance range

The outflow to the sea as predicted by the model (Table 8.4) is not verifiable in terms of volume but has been verified in existence by persons familiar with the local beach when they observed canines drinking from small puddles. The coastal inflow predicted by the model is mostly within the coastal region of the Brograve level and Horsey, which correlates well with previous studies such as Driscoll, (1984) and Holman (1994).

Table 8.25 The average Coastal inflow water balance table (m^3/day)

Coastal	Into Aquifer	Out of Aquifer
Happisburgh	0	1873
Hempstead	272	0
Brograve Level	1437	0
Horsey	921	0
Somerton New	270	0
Somerton Old	181	213
Total	3081	2086

The individual water balances of the cultivated layer and the lakes are presented below in Tables 8.6 and 8.7. The conditions within the cultivated layer are simulated using the General Head Boundary condition. Nearly all of the flow generated by the GHB condition are out of the aquifer and therefore vertically upwards into the cultivated layer. This is indicating that the modelled heads within the cultivated layer are lower than the modelled heads within the underlying aquifer.

Table 8.26 The average water balance for the cultivated layer (m^3/day)

Name	Into Aquifer	Out of Aquifer
Hempstead	0	761
Brograve Level	0	379
Gravity 1	0	187
Lessingham	0	1286
Heigham Holmes	0	6
Eastfield	0	978
Ludham	1	0
Horsey	0	5
Martham	0	8
Gravity 2	0	2
Somerton New	0	22
Potter Heigham	0	329
Private 8	0	95
Repps	0	1
Stubbs	0	392
Thurne	0	2
Somerton Old	0	5
Total	**1**	**4458**

The noticeable values within Table 8.7 are the volumes of groundwater exiting the aquifer into Womack Water and the volumes of surface water entering the aquifer at Hickling Broad. The modelled area of the Womack Waters is 437,500 m^2 (= 14 × 31,250 m^2) and so a loss of 960 m^3/day equates to an aquifer outflow of 2.2 mm/day per unit area, which whilst unverifiable, does equate to a reasonable seepage rate. For Hickling Broad; a similar calculation can be completed for this body of surface water, with a loss of 978 m^3/day equating to a seepage rate into the aquifer of approximately 0.7 mm/day.

Table 8.27 The average lakes water balance table (m^3/day)

Lakes	Into Aquifer	Out of Aquifer
Womack Waters	0	960
Heigham Sound & Candle Dyke	15	0
Hickling Broad	978	21
Horsey Mere & Breydon Marshes	15	0
Martham Broad	86	0
River Thurne	4	54
Total	**1098**	**1035**

The replication of groundwater heads through out the catchment is a very difficult task to accomplish, because the average water levels measured within the observation wells are often influenced by local conditions. An example of this can be seen in the group of observation wells shown in Figure 8.1. The observation wells located at Grove Farm (1.17 mODN) and North End Farm (0.03 mODN) are separated by distance of approximately 1 kilometre and yet the average difference in mean water levels is over a metre, giving a hydraulic gradient of 1.1 x 10^{-3}.

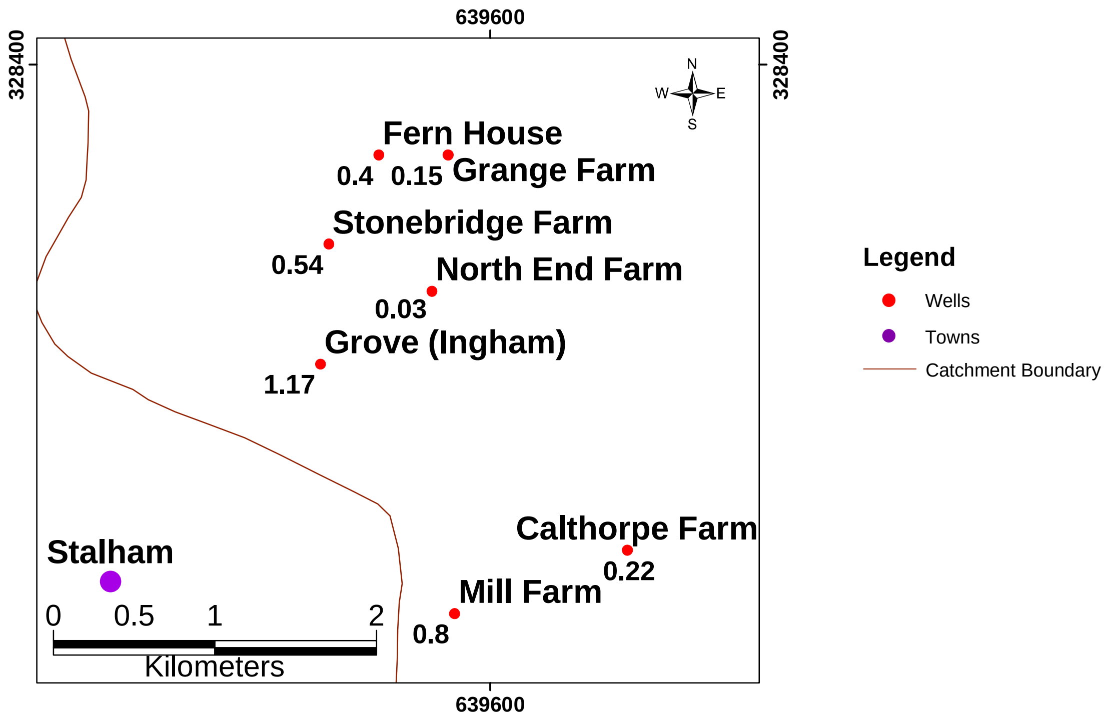

Figure 8.1 The observation wells located immediately east and northeast of the town of Stalham

Figure 8.2 represents the simulated head distribution of Layer 2. The contour line of -0.5 mODN traces a path around the Brograve Level area and is the lowest contour line presented. This demonstrates that the lowest *modelled* heads are replicated in the area with the lowest *measured* groundwater heads within the catchment. The groundwater levels near Catfield, where groundwater levels are high enough to cause significant road damage (*pers. comm.* Andrew Alston), are simulated as high as 1.5 mODN within the head distribution values and it is known that location is approximately 1.0 mODN.

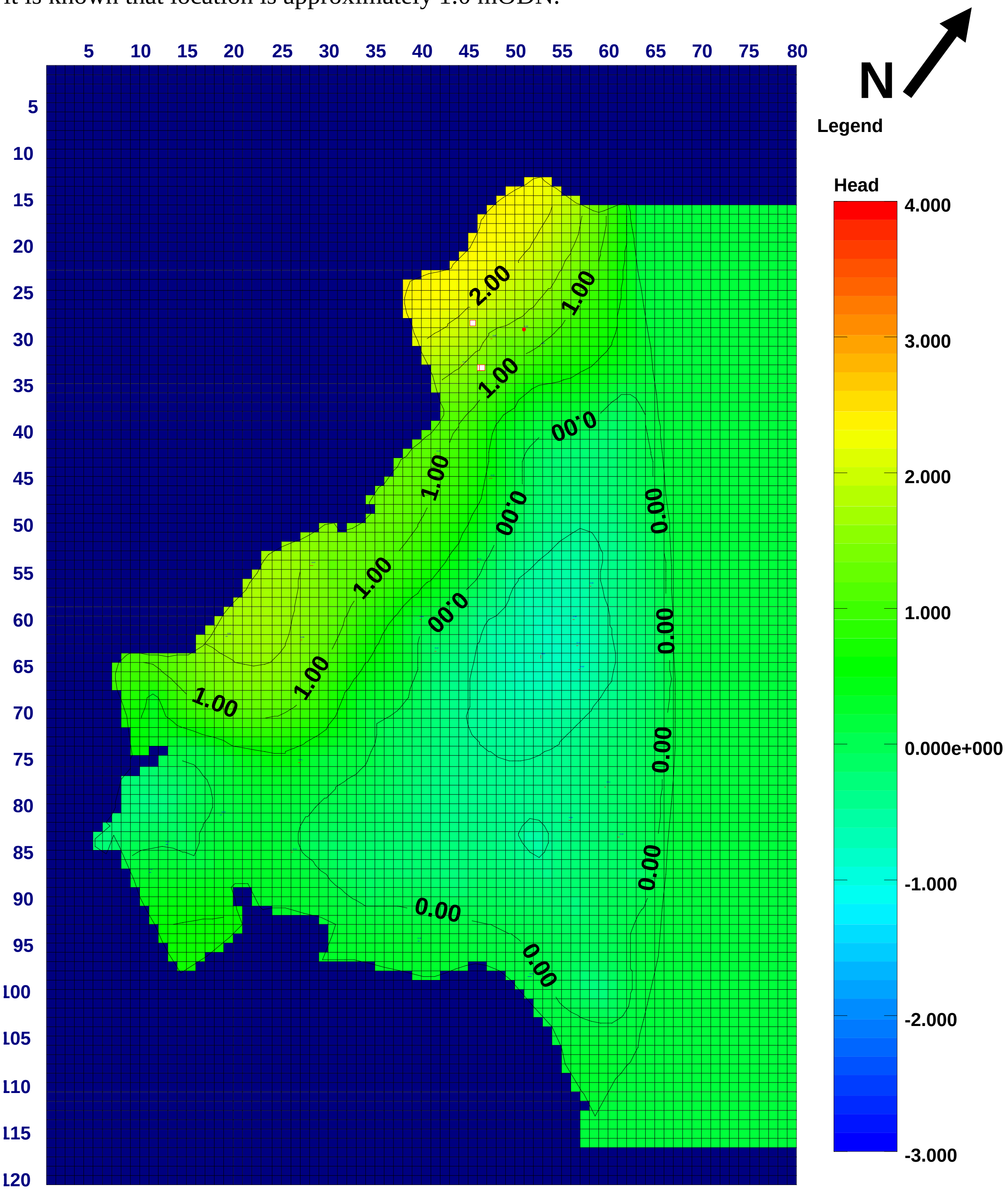

Figure 8.2 Simulated time-invariant head distribution (mODN) in Layer 2

8.3 The analysis of adjusted conditions near the Brograve main drain (Scenario 2)

8.3.1 Introduction

The replication of the spatial distribution of groundwater heads within the catchment has proven a difficult task. This was shown in the previous section where the observed variation in groundwater heads over a very short distance is substantial. The adjustment to the drainage coefficients within the Brograve sub-catchment is intended to lower the groundwater heads in the aquifer beneath the main drain to below -1m ODN in an attempt to replicate the low groundwater heads found along the Brograve main drain.

8.3.2 The adjustment to the model parameters

In the Scenario 1 model, the Brograve sub-catchment had a distribution of cells with drainage coefficients, which were mostly 50 m^2/day and 1 m^2/day, whereas the redistributed version has a significant number of cells that have drainage coefficients of 100 m^2/day.

Table 8.28 The table of re-distributed drain coefficient values for the Brograve catchment (m^2/day)

Standard				Re-arranged			
No. of cells	**DC**	**Out of aquifer**	**Into Aquifer**	**No. of cells**	**DC**	**Out of aquifer**	**Into Aquifer**
none	100	0	0	35	100	3500	0
119	50	5950	0	48	50	2400	0
5	5	25	0	4	5	20	0
232	1	232	0	269	1	269	0
Total		**6207**	**0**	**Total**		**6189**	**0**

The new distribution of drainage coefficients within the Brograve sub-catchment is presented in Figure 8.4, where the dark blue cells within the Brograve catchment represent the cells with drainage coefficients of 100 m^2/d, whereas the light blue cells represent the cells with 50 m^2/d and the green cells represent those with 1m^2/d.

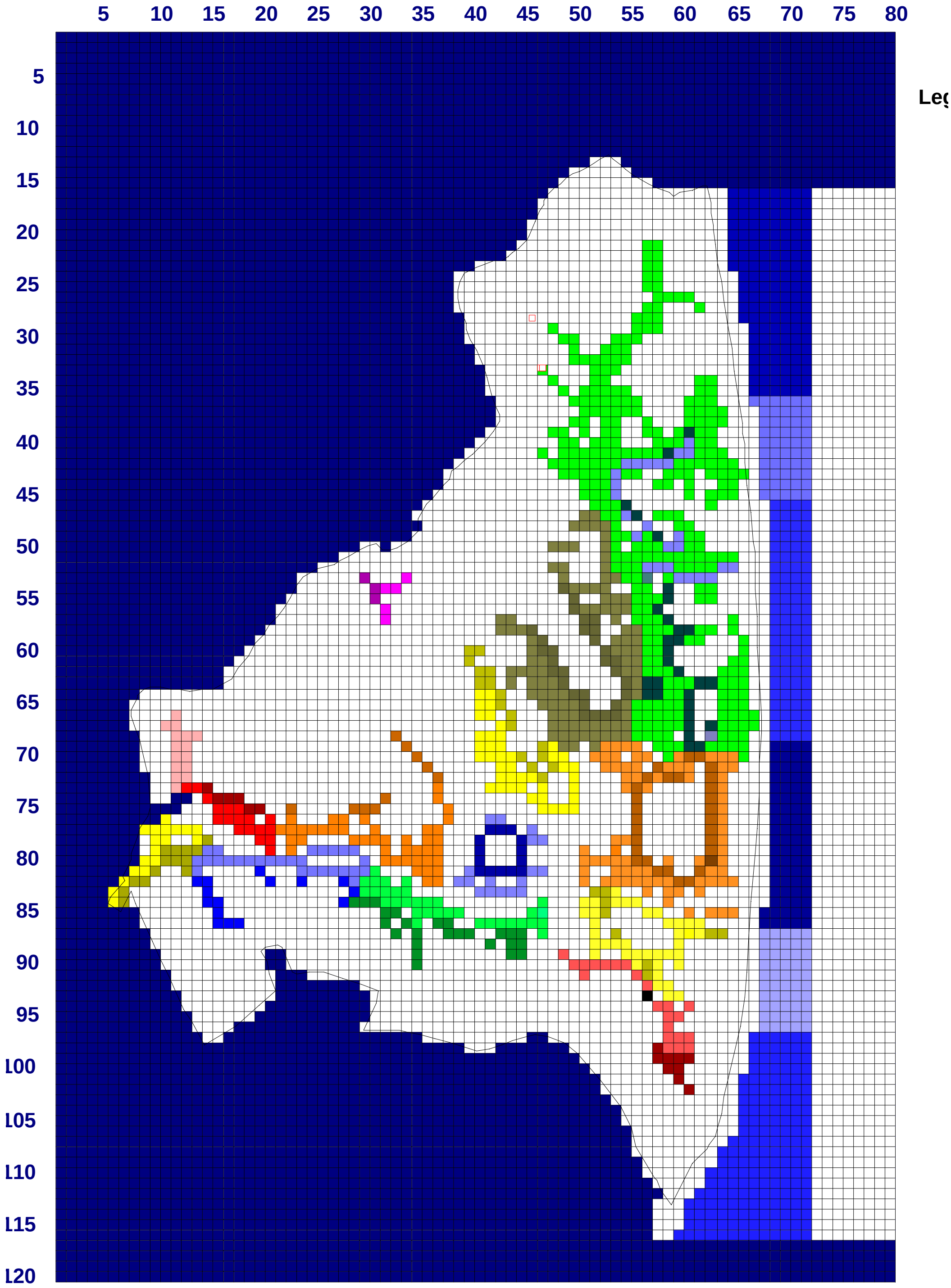

Figure 8.3 The re-distributed drainage coefficients within the Brograve catchment

8.3.3 The results

The re-arrangement of drain coefficients near the thickest parts of the peat produced a simulated lowering in the groundwater heads beneath the main drain. The four well-connected cells (darkened green) in the centre of the Brograve sub-catchment Figure 8.3, were designated along the main Brograve drain are each given a drain coefficient of 100 m^2/day. The previous distribution of DC's included several well-connected drains within the Lessingham area which were receiving fresher water from the eastern higher outcrop. The redistribution of the value of the Drain coefficient of these cells has allowed the groundwater heads within the northwestern outcrop to increase. The head distribution for the Brograve level is similar to the distribution before the re-arrangement of Drainage coefficients. The groundwater heads of approximately 1.0 mODN was achieved in the aquifer beneath the main drain.

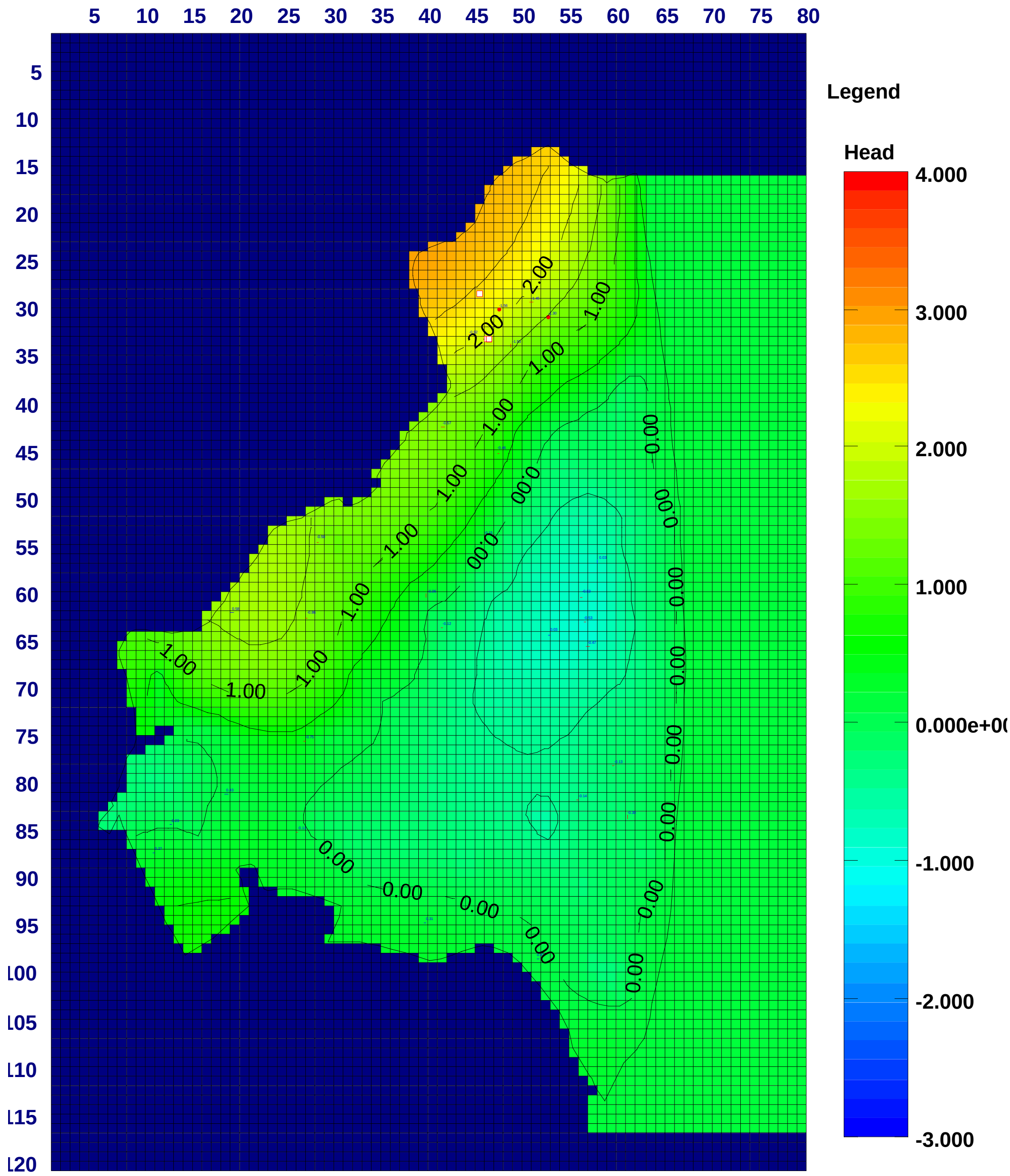

Figure 8.4 The head distribution in Layer 2 throughout the catchment after the re-arrangement of Brograve drainage coefficients

8.4 Method and scope of sensitivity analysis

It is generally regarded that a carefully planned sensitivity analysis is a prerequisite to model understanding. The benefits gained include identification of inadequacies in the conceptualisation and subsequent model performance, the ability to identify the areas in which data are sparse and whether further fieldwork may be necessary or additional research required. The analysis often allows the modeller to identify whether a specific output parameter is relatively insensitive, sensitive or hyper-sensitive to change and thus a possible source of model discrepancy. The purpose of a sensitivity analysis is to quantify the uncertainty in the *calibrated* model caused by uncertainty in the estimates of aquifer parameters, stresses, and boundary conditions (p.246, Anderson and Woessner, 1992).

There are several techniques of sensitivity analysis which are used in modern model development procedures (Beven, 2004), including one-at-a-time sensitivity measures (Dubus et al. 2003)), the sensitivity index (Bauer and Hamby, 1991) and standard regression techniques (Iman and Helton, 1988). The one-at-a-time (*ceteris paribus* approach) sensitivity analysis method ((Hamby, 1994) has been chosen to investigate the effects of parameter in the Thurne groundwater model. The method of analysis used must be based on realistic assumptions of the uncertainties in the model conditions (Bergström, 1991)). The maxima and minima values therefore used within the sensitivity analysis should reflect the realistic extremities that might be encountered.

Within the Thurne catchment the complex nature of the hydrological and hydrogeological processes make a full sensitivity analysis of all the input variables very difficult. For example, the analysis of the hydraulic conductivity of the Pleistocene Crag would have the subsequent effect of changing the real values of the individual drainage coefficients. The sensitivity analyses that are covered in this chapter will include changes in the various parameters that have modelled values which may be influential to the four output categories i.e. Head distribution, groundwater contributions the coastal inflows and the particle tracking outputs. Certain sensitivity analyses such as the analysis of the hydraulic conductivity of the Norwich Brick-earth would be of very little effect. This is due to the fact that the realistic limits of variation with this deposit is 0.1 m/day to 5 m/day (Boswell, 1916) which would not hinder the recharge from entering the aquifer. The parameters that are to be analysed for catchment-wide variations are listed in Table 8.1 with further specialised analyses concerning specific times (average and average winter conditions) analysed in Section 8.7.

8.5 The analysis of variations in hydraulic conductivities

8.5.1 The hydraulic conductivity of the crag (Scenarios 3 & 4)

The lowering of the hydraulic conductivity of the crag by 50% of its original value to 10 m/day saw a decrease in the coastal inflow of from 3,081 m^3/day to 2062 m^3/d (Table 8.9). The decrease in hydraulic conductivity also brought about a decrease in the Happisburgh outflow from 1873 m^3/d to 1660 m^3/d. Increasing the hydraulic conductivity of the crag by 150% (from 20 m/day to 50 m/day) produces an increase in flows leaving the aquifer, and an increase in the coastal inflow from 3,081 m^3/d to 5,289 m^3/d - a percentage increase of 66% . The coastal inflow is said to be sensitive to the hydraulic conductivity of the crag.

Table 8.29 The variation of coastal inflows influenced by the variation K-crag

	K-crag 10 m/day		K-crag 20 m/day		K-crag 50 m/day	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1660	0	1873	5	2065
Hempstead	191	0	272	0	460	0
Brograve	979	0	1437	0	2379	0
Horsey	629	0	921	0	1556	0
Somerton New	165	1	270	0	491	0
Somerton Old	98	280	181	213	398	136
Total	**2062**	**1941**	**3081**	**2086**	**5289**	**2201**

The results from the three simulations suggest that the mean water levels for a selection of wells, covering the extent of the catchment and the various local deposits, would be improved by the use of 10 m/day for the hydraulic conductivity for the crag (Table 8.14). However, the wells at Conifers (in the north of the catchment) and Broads Haven (on the periphery of the Hickling Broad) would have closer values of mean water levels if the value of hydraulic conductivity for the crag was *increased* to 50 m/day.

Table 8. 30 The effect on modelled water levels at various wells of varying K-crag

Well	Observed mean water level (mODN)	Simulated water level [K-crag=10 *m/day]*	Simulated water level [K-crag=20 *m/day]*	Simulated water level [K-crag=50 *m/day]*	Deposit
Conifers	0.0	0.3	0.1	**0.0**	Loamy
Broads Haven	-0.4	0.5	0.3	**0.1**	Exposed crag
Dingle Dell	0.2	**0.0**	0.0	-0.1	Flegg
Eastfield	-0.8	-0.9	**-0.8**	-0.5	Loamy
Hoverings	-0.6	**-0.6**	-0.4	-0.2	Clay
Lambridge Mill	-0.8	**-0.7**	-0.6	-0.3	Altcar peat
Street Farm	-0.4	**-0.3**	-0.3	-0.2	Holmes
Upper Brograve	-1.5	**-1.0**	-0.8	-0.4	Altcar peat

8.5.2 The hydraulic conductivity of the 'clayish' layer (Scenarios 5 & 6)

The default hydraulic conductivity of the 'clayish' layer within the Crag aquifer was changed from 1 m/day to 0.1 m/day and 20 m/day. When rounded to one decimal place the simulated groundwater heads at the chosen well locations are insignificantly different and therefore the groundwater heads are said to be *insensitive* to change in the hydraulic conductivity of the 'clayish layer'.

Table 8.31 The effect on modelled water levels at various wells of varying the K-clay layer

well	Observed mean water level (mODN)	Simulated water level [K-Clay=0.1 *m/day]*	Simulated water level [K-Clay=1 *m/day*]	Simulated water level [K-Clay=20 *m/day]*	Deposit
Conifers	0.0	0.1	0.1	0.1	Loamy
Broads Haven	-0.4	0.3	0.3	0.3	Exposed crag
Dingle Dell	0.2	0.0	-0.1	-0.1	Flegg
Eastfield	-0.8	-0.8	-0.8	-0.8	Loamy
Hoverings	-0.6	-0.4	-0.4	-0.4	Clay
Lambridge Mill	-0.8	-0.6	-0.6	-0.6	Altcar peat
Street Farm	-0.4	-0.3	-0.3	-0.3	Holmes
Upper Brograve	-1.5	-0.8	-0.8	-0.7	Altcar peat

8.6 The effect on coastal inflow of variations in the hydraulic conductivity of the alluvium and the value of the catchment-wide drain coefficients (Scenarios 7 & 8)

The hydraulic conductivity of the alluvial deposits of the peat (northern marshes) and the clay (southern marshes) were reduced each by 50% (the peat reduced to 4×10^{-4} m/day, clay to 8×10^{-4} m/day) to measure the effect upon the amount of coastal inflow. The modelled average daily coastal inflow volume was reduced by 12% from 3,081 m^3/day to 2,601 m^3/day. The coastal inflow is therefore said to be relatively *insensitive* to the hydraulic conductivity of the alluvium. The doubling of the drain coefficients within the entire catchment saw an increase in the coastal inflow from 3,081 m^3/day to 6,792 m^3/day. The Hempstead, Somerton

New and Somerton Old coastal sub-catchments seem to be particularly *sensitive* to the increase in drain coefficients. A particle-tracking plot for the numerical groundwater model with the drainage coefficients doubled is presented in Appendix B.

Table 8. 32 Effects of reducing the hydraulic conductivity of the alluvium and doubling of drainage coefficients on coastal inflows and outflows (m^3/day)

	Scenario 1		Reduced k-alluvium		DC-Doubled	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1873	0	2158	16	1501
Hempstead	272	0	92	73	741	0
Brograve Level	1437	0	1123	0	2589	0
Horsey	921	0	885	0	1753	0
Somerton New	270	0	285	0	785	0
Somerton Old	181	213	216	213	908	35
Total	**3081**	**2086**	**2601**	**2444**	**6792**	**1536**

8.7 A simulation representing average summer and average winter conditions (Scenarios 9 & 10)

8.7.1 The required modelling adjustments

The average winter conditions were simulated by assuming that the average winter daily recharge is 150% of the time-invariant average daily recharge value. Whereas the average summer conditions were simulated using 66% of the time-invariant daily recharge value and assume that an *equivalent recharge* (Rushton, 2003) amount is released from storage.

8.7.2 Result

The standard time-invariant, ‘winter’ and ‘summer’ groundwater balances are tabulated in Table 7.13. The net inflow into the inland drains reduces from 25,063 m^3/day in the ‘winter’ to 16,985 m^3/day in the ‘summer, a reduction of 17% whereas the recharge reduction was 44%. The net result is that the loss in fresh water recharge from rainfall allows more fresh water from the cultivated layer and saline water from the coast to enter into the aquifer. The change in modelled coastal inflow caused by the reduction in ‘summer’ rainfall recharge is tabulated in Table 8.14. The total coastal inflow is at its highest in the summer (4,761 m^3/d) and at its lowest in the winter (1,730 m^3/d). The pathlines that are presented in Figures 8.5 and 8.6 represent the movement of the groundwater within the catchment in the two seasons. The olive/mustard cells represent the drain cells for the Eastfield sub-catchment.

Table 8.33 A seasonal comparison of groundwater balances (m^3/day)

	Scenario 1		winter		summer	
Inland drains	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Brograve	36	5863	0	7405	95	4946
Catfield	0	454	0	678	0	269
Heigham	0	474	0	650	0	332
Eastfield	0	2158	0	2746	0	1868
Horsefen	0	708	0	876	0	574
Horsey	0	1006	0	1163	0	928
Martham	0	1947	0	2389	0	1465
Somerton New	0	1647	0	1786	0	1601
Potter H.	0	1144	0	1432	0	918
Repps	0	1196	0	1532	0	817
Stubbs	0	1231	0	1543	0	1052
Thurne	0	841	0	1068	0	625
Somerton Old	0	1597	0	1722	0	1549
Gravity	0	60	0	73	0	41
Total	**36**	**20326**		**25063**	**95**	**16985**

The typical ‘summer’ and ‘winter’ conditions simulate different locations along the coast from which the Eastfield sub-catchment receives its coastal waters. The ‘summer’ simulation shows that the length of the coastal waters pathlines are increased, suggesting that the reduction of recharge in the Eastfield sub-catchment allows more coastal waters to travel further west and enter the drains of the Eastfield sub-catchment.

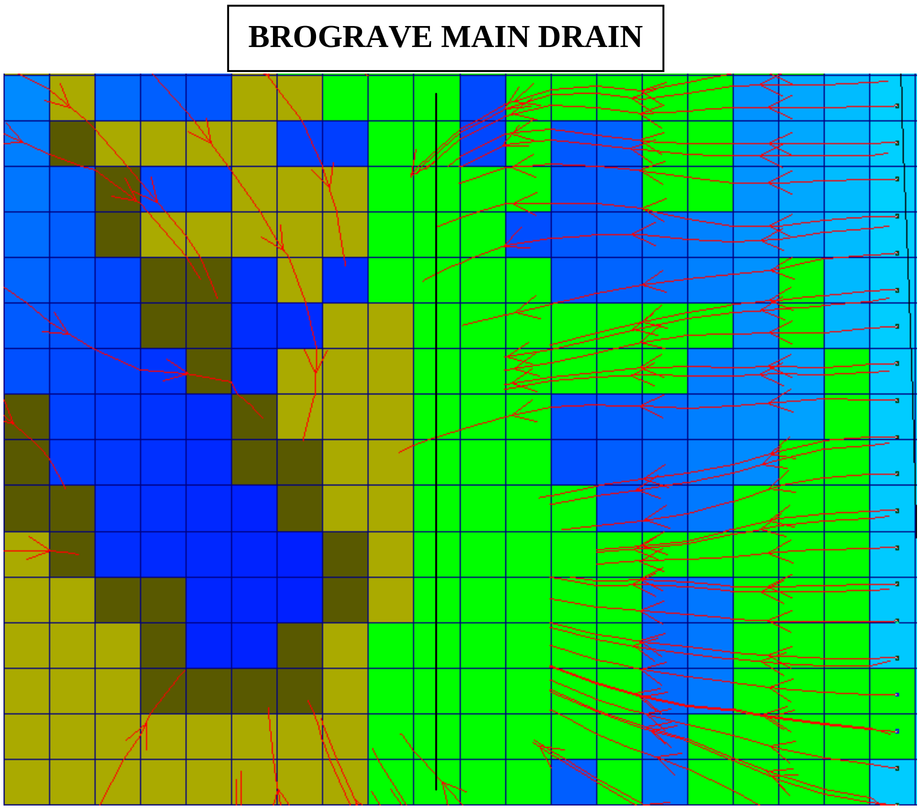

Figure 8.5 Groundwater flow pathlines in the Eastfield and Brograve sub-catchment under ‘winter’ conditions

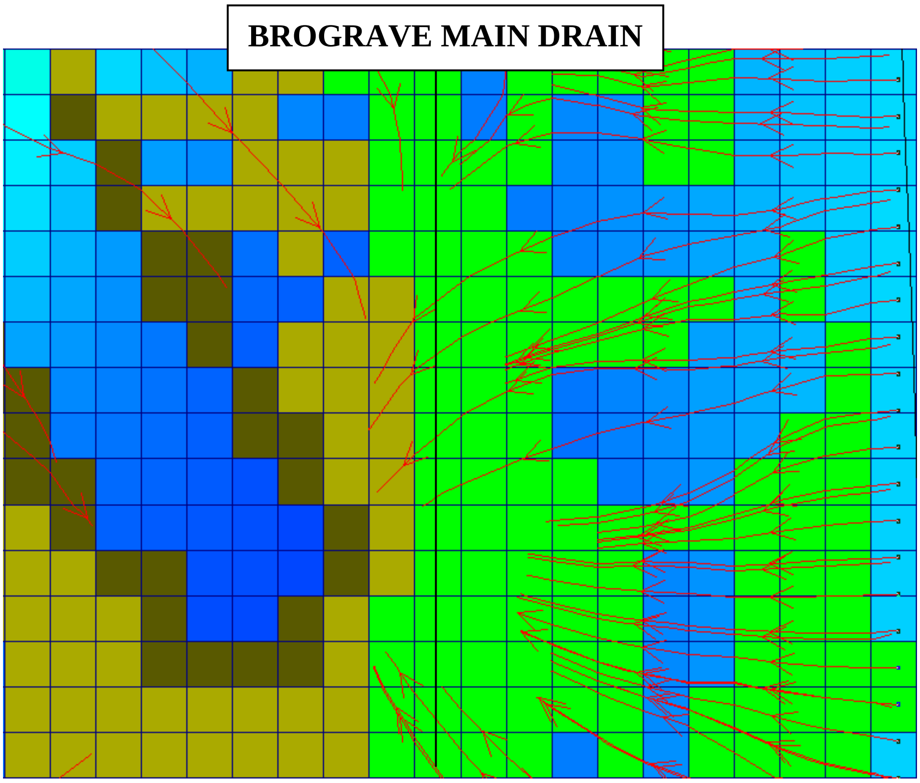

Figure 8.6 Groundwater flow pathlines in the Eastfield and Brograve sub-catchment under 'summer' conditions

Figures 8.5 and 8.6 are plots for two very different sets of pathlines; to interpret them it is important to appreciate very approximately how far a particle of water will move in a season of 100 days. By considering the seepage velocity defined by Equation 3.1 in Section 4.3.1. Distance moved, *s*, is given by:

$$s = \frac{K}{N}\frac{dh}{dl} \times t \tag{8.1}$$

where *K* is the hydraulic conductivity of the crag *dl* is the distance between the contour lines *N* is the effective porosity and *dh* is the head gradient between the two contour lines and *t* is the time in days.. With $K = 20$ m/d, $N = 0.15$ m^3/m^3, $dl = 4 \times 177$ m, $dh = 0.5$ m, $t = 100$ days

$$s = \frac{20}{0.15} \times \frac{0.5}{4 \times 177} \times 100 \approx 10 \text{ metres} \tag{8.2}$$

This means that although the pathlines of the coastal inflows may approach internal sub catchments such as Eastfield and Stubb, the estimated time taken for water to arrive within these drains is of the order of hundreds of years. Two particle-tracking plots for the numerical groundwater model for both 'summer' and 'winter' scenarios are presented in Appendix B.

Table 8.34 A seasonal comparison of coastal inflows (m^3/day)

	Scenario 1		Winter		Summer	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1873	0	3126	41	1107
Hempstead	272	0	42	96	532	0
Brograve	1437	0	945	2	1932	0
Horsey	921	0	626	0	1249	0
Somerton New	270	0	117	57	488	0
Somerton Old	181	213	0	544	519	35
Total	**3081**	**2086**	**1730**	**3825**	**4761**	**1142**

8.7.3 Summary and conclusions

The *Total volume of groundwater* estimated from data to be entering the deep (inland) drains from the aquifer (18,333 m^3/day) is less than the amount suggested by the model (20,318 + 4,458 m^3/day). The amount of *coastal waters* entering the catchment is also over-estimated, but the fraction of coastal water to total groundwater for both the estimated and the model is around 12%.

The *particle tracking* traces for both winter and summer conditions suggest that the model is correctly simulating the flows from the coastal region to the Eastfield and Stubb sub-catchments (Figures 8.6 & 8.7). The rearrangements of the drain coefficients within the Brograve sub-catchment suggest that feasible pathlines can be obtained from various re-arrangements of the drain coefficients within the sub-catchment. The modelled head distribution is on average higher than the estimated average groundwater level. The source of this discrepancy could be (a) the aquifer hydraulic conductivity is modelled as homogeneous when several aquifer tests throughout the catchment give varying values for the horizontal hydraulic conductivity, or (b) the local groundwater level variation within the point well measurements are large and possibly reflect very local influences, which are almost impossible to reproduce within a model.

The coastal inflow is proportional to the chosen value of hydraulic conductivity of the aquifer and increases or decreases depending the amount of recharge entering the aquifer. The distribution of groundwater heads are influenced by the chosen value of hydraulic conductivity.

The halving of the hydraulic conductivity of the alluvium saw a small response for water entering the catchment from the coast. This can be interpreted as the coastal inflows being insensitive to the hydraulic conductivity of the alluvium. Doubling the drain coefficient catchment wide doubled the amount of coastal water entering the catchment, indicating the potential importance of remedial engineering measures, such as ditch lining, which have the effect of decreasing what is represented by the drainage coefficient in the model.

9 PREDICTION TESTING AND RESULTS

9.1 Introduction

In Chapter 8, sensitivity analyses were completed to estimate how certain input variables affect conditions in the aquifer system. The groundwater model was considered adequate as from the Brograve pump and secondly whether the proposed solutions are practicable. The suggested re-engineering predictions are

- Altering the water levels within the Hempstead Marshes and Lessingham area in the north of the Brograve sub-catchment;
- Excavating of a new interceptor drain along the Brograve section of the coast;
- The lining of the main drain at the Hempstead Marshes with low permeability material.

9.2 Prediction 1- The adjustment of drain water levels in Hempstead Marshes and Lessingham areas

9.2.1 Introduction

The raising of water levels within the Brograve sub-catchment was first recognised as a possible remedial measure by Harding and Smith (2002) and subsequent refinements of water levels rises were discussed by Harding et al. (2005). In the latter report, it was suggested that a strategy be put in place whereby average water tables would be 0.35 m below ground level throughout the problematic drainage locations of the sub-catchment during the summer months. During the course of this research project, it was confirmed that there is a substantial proportion of saline coastal water entering the Brograve system via the Hempstead Marshes branch of drains (see Figure 4.8). The ground level elevation throughout the Hempstead Marshes was surveyed by Harding and Smith (2002) and found to be very uneven making any raising of water level relative to the ground elevation difficult. An improved course of action proposed by the author of this text is to raise the water level to a specific elevation (-0.4 mODN) throughout the Hempstead marshes.

9.2.2 The scientific rationale for the adjustment of the drain water levels

The movement of water from the aquifer into a drain is governed by the Darcian-based drainage Equation 6.8, which relates the amount of water entering a specific drain to the local aquifer-drain head differential. Raising water levels in the Hempstead marshes would reduce the head differential and therefore lead to a reduction in the total flow into the drains. The proposed water level maintained along the Hempstead branch of drains at -0.4 mODN would be significantly higher than that maintained along the main Brograve drain at -1.9 mODN. To maintain these relative water levels requires the installation of a water control device, preferably a weir, at the proposed location (Figure 9.1) at the confluence of the Lessingham and Hempstead branch of drains within the Brograve sub-catchment.

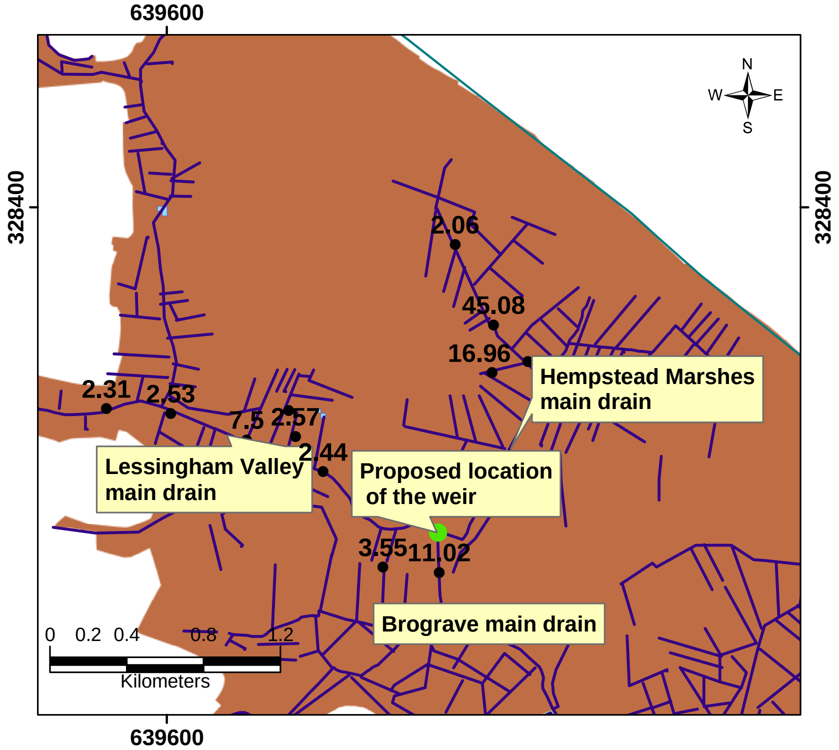

Figure 9.2 The proposed location of the weir (spot electrical conductivity readings in mS/cm)

The raising of water levels to -0.4 mODN within the drains of the Hempstead Marshes is intended to reduce the inflow of saline waters entering the Brograve main drain and the subsequent salt load into the Horsey Mere. Effectively the Brograve main drain would receive a reduced saline contribution flowing from the Hempstead drains, although with a local level of -0.4 mODN in the Hempstead marshes, there would continue to be saline inflow because the drain water levels are still below sea level. The raising of the water levels in the Hempstead Marshes has been combined with a lowering of the currently high drain water levels in Lessingham Valley to cause more fresh water to enter the Brograve main drain ensuring a similar total flow at Brograve pump with less saline waters.

9.2.3 The required adjustments to the numerical groundwater model

The model adjustments require two separate changes to the modelling domain; the first is to raise the water levels throughout the entire Hempstead marshes to -0.4 mODN. The current water levels are between -1.1 mODN and -1.8 mODN. The second is to lower the water levels within specific drains in the Lessingham Valley, which put at them at levels between -1 mODN and -1.8mODN. Both of these adjustments are represented in Figure 9.2.

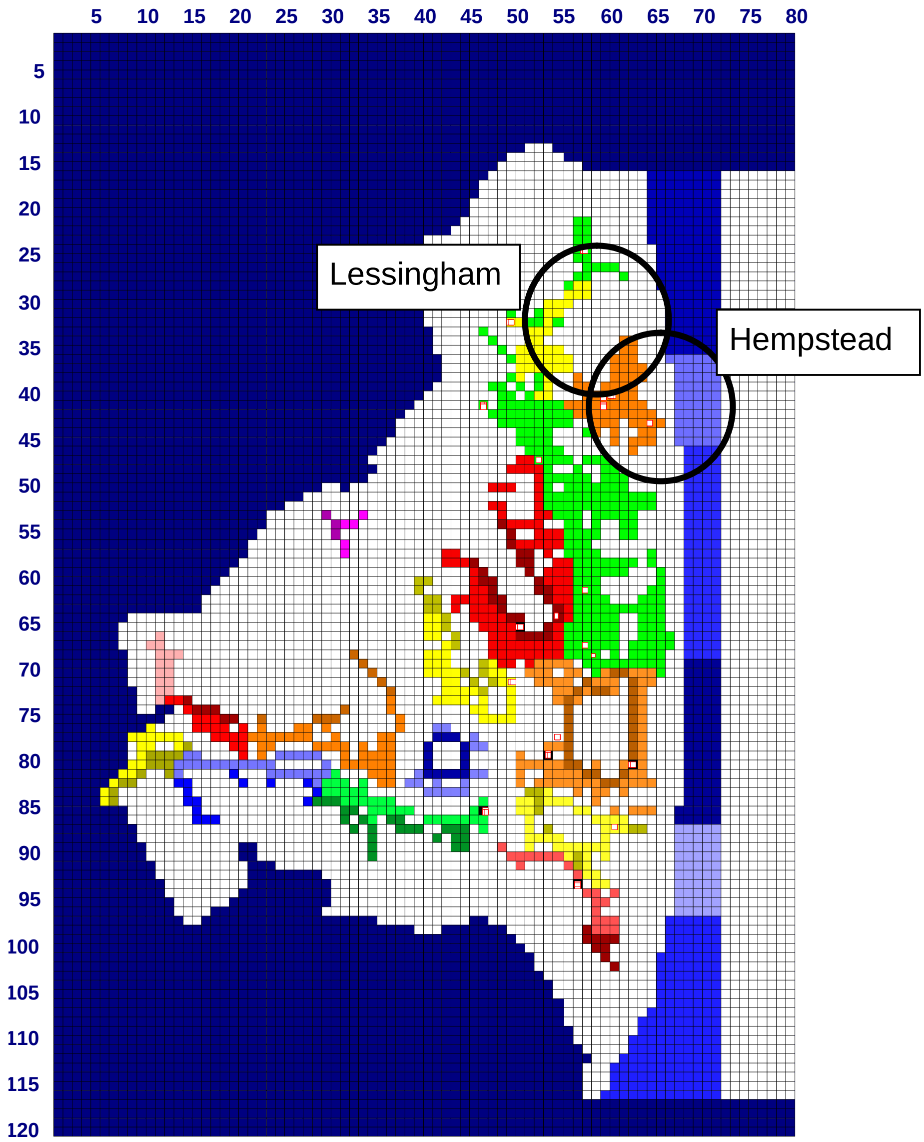

Figure 9.3 The locations in the modelling domain where the water level adjustment are made

9.2.4 Results

The simulated coastal inflow from the Hempstead and Brograve coastal reaches both show decreases from 272 m^3/d to 84 m^3/d and 1,437 m^3/d to 1367 m^3/d respectively (Table 9.1).

The total coastal inflow from the Hempstead and Brograve reaches *before* the adjustments of 1,709 m^3/d, is reduced to 1,451 m^3/d *after* the adjustments, which represents a 15% decrease in the coastal water inflow from the coastal reaches immediately east of the Brograve sub-catchment.

Table 9. 35 The effect on the coastal inflow by adjusting the Lessingham and Hempstead water levels (m^3/d**)**

	Scenario 1		Prediction 1	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1873	0	1964
Hempstead	272	0	84	28
Brograve	1437	0	1367	0
Horsey	921	0	920	0
Somerton New	270	0	270	0
Somerton Old	181	213	181	218
Total	**3081**	**2086**	**2822**	**2210**

The standard model estimated the total (fresh and saline) groundwater contribution to the Brograve pump discharge total as being 5,863 m^3/d (Table 8.4) whereas the adjustments to the water levels in the Lessingham valley and Hempstead Marshes lead to an estimated decrease to 5,332 m^3/d. A particle-tracking plot for the numerical groundwater model with the adjustment to the water levels along the Lessingham and Hempstead branches is presented in Appendix B.

9.3 Predictions 2 & 3- The construction of an interceptor drain

9.3.1 Introduction

The utilisation of an interceptor drain for the removal of excess groundwater from a seeping canal has been adopted for irrigation schemes in USA and Pakistan (Willardson et al., 1971), Bhutta and Wolters, 2000; Jansen et al., 2006). Harding and Smith, (2002) proposed the construction of a coastal interceptor drain in the Thurne catchment as a means of preventing the inflow of saline groundwater into the Brograve sub-catchment.

The path of the interceptor drain, as proposed by Harding et al. (2005) Figure 9.3 shows that it would run approximately from the seaside village of Eccles to the coastal area parallel to the Brograve pump covering a distance of just over 6 km .

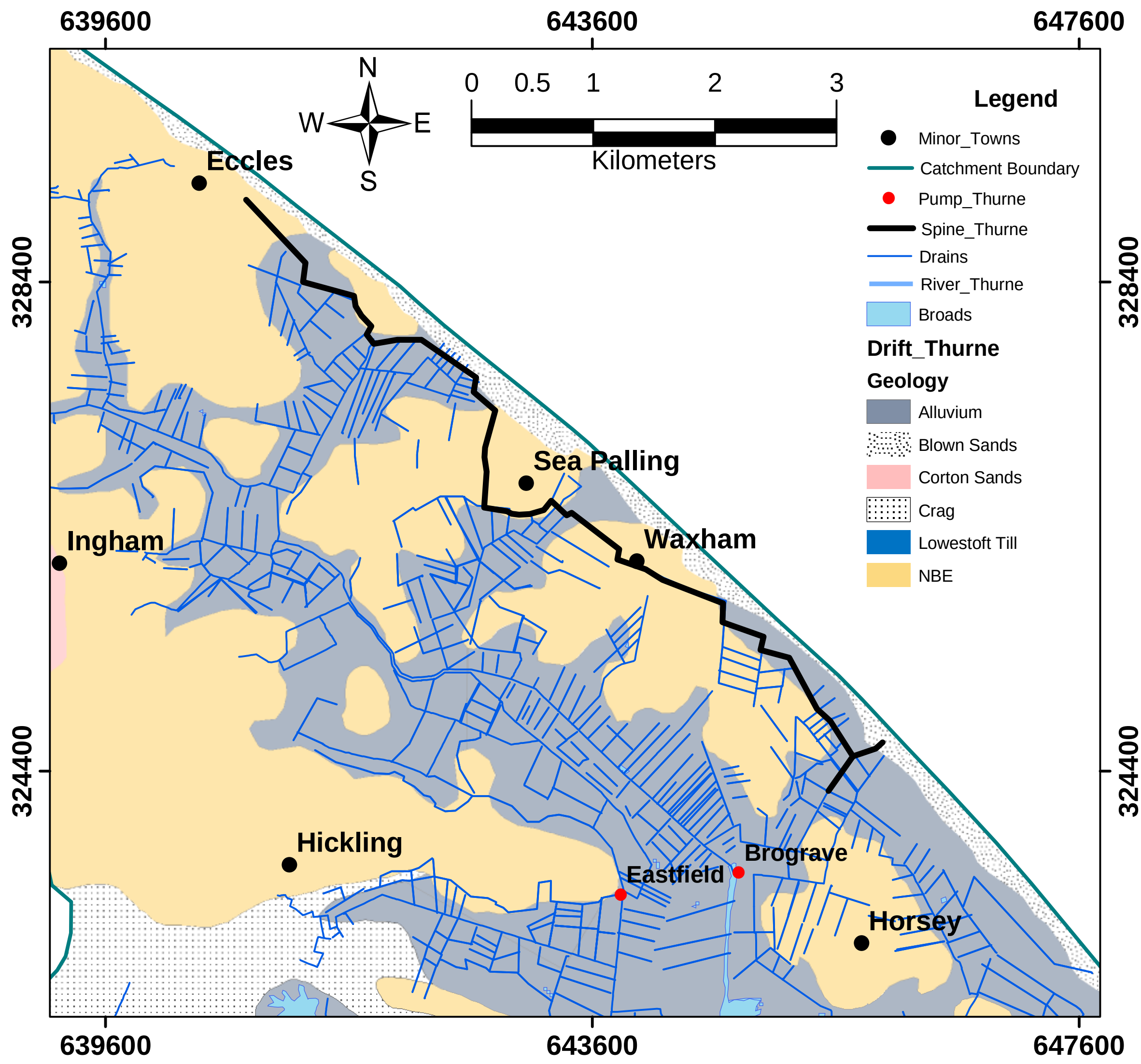

Figure 9.4 The path of the proposed interceptor drain from Harding & Smith (2002**)**

9.3.2 The influence of the interceptor drain on the individual conceptual theme components

In Section 4.1, the introduction of three conceptual themes allowed for the complexity of the drain-aquifer system to be understood in separate components. Similar components are found in the interceptor drain–aquifer system.

The first component that is considered is the *flow within the aquifer system*; the numerical groundwater model represents this. In Figure 9.4, a typical block section of the aquifer from the western margin of the peat to the coast is presented. The interceptor drain will not only intercept coastal waters but will receive fresh waters from the west and from recharge through the overlying Norwich Brickearth.

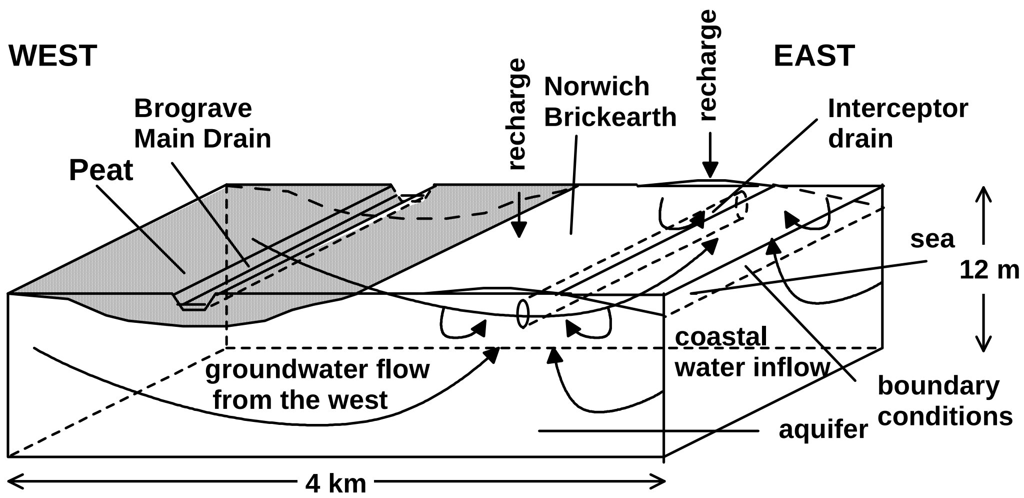

Figure 9.5 A typical block section of the proposed interceptor drain

The second component that is considered is the *flow from the aquifer to the drain*; the drain coefficient *Cd* for the interceptor pipe drain represents this.

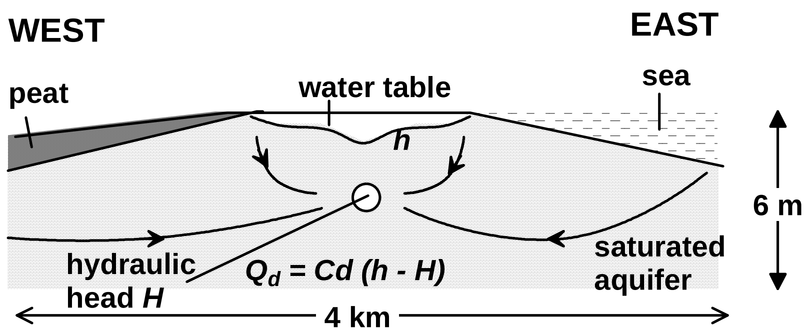

Figure 9.6 A typical cross section of the proposed interceptor drain

The third and last component that is considered is the *hydraulic flows within the drain*: Within the drain, water enters from the sides through the holes causing turbulent flow.

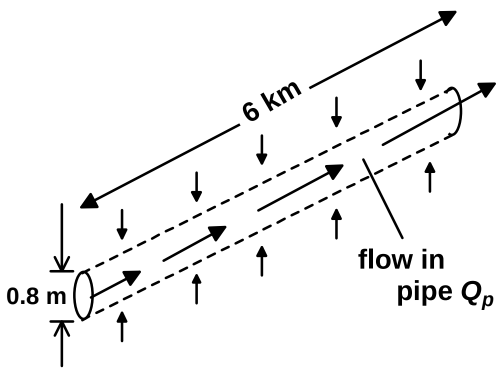

Figure 9.7 Hydraulic flows within the drain

9.3.3 The required adjustment to the numerical groundwater model and the analysis of results

The actual adjustment implemented in the numerical groundwater model required the introduction of drainage coefficients to Layer 3. The 42 cells along the proposed line of the coastal interceptor drain were each given a water level of -2 mODN and a conductance of 133.3 m^2d/cell or 1000 m^2/d/cell reflecting the two predictions of the drain performance. Table 9.2 shows the comparison of the coastal inflows with an interceptor drain with both *practicable* (133.3 m^2/d/cell) and the *required* (1,000 m^2/d/cell) drain coefficients, with the coastal inflows without the interceptor drain.

Table 9.36 Comparison of the coastal inflows with the implementation of an interceptor drain (m^3/d)

	Scenario 1 (with no interceptor drain)		Model with interceptor drain with Cd = 133.3 m^2/d/cell		Model with interceptor drain with Cd = 1000 m^2/d/cell	
Coastal section	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1873	597	1435	2451	1153
Hempstead	272	0	1866	0	4350	0
Brograve	1437	0	4197	0	9890	0
Horsey	921	0	1405	0	2572	0
Somerton New	270	0	278	0	298	0
Somerton Old	181	213	183	215	190	214
Total	3081	2086	8526	1650	19751	1367

The comparison of the results for the two differently parameterized interceptor drains reveals some valuable details concerning the effects of increasing the drainage coefficients:

1) The drainage coefficient ratio for the two interceptor drains is 15:2, whereas the ratio between the resulting total coastal inflow is only 2:1 (19,751 m^3/d for *Cd* = 1,000 m^2/d/cell compared to the 8,526 m^3/d for *Cd* =133.3 m^2/d/cell).
2) Both types of interceptor drains induce coastal inflow from the Happisburgh section, which previously was exclusively a zone of coastal outflow, and reduce coastal outflow. The simulation with drain coefficients of 1,000 m^2/d/cell increases the Happisburgh coastal inflow by approximately 1,800 m^3/d compared with the 133.3 m^2/d/cell simulation and the fresh water outflow is reduced from 1,453 m^3/d to 1,153 m^3/d.
3) In the case of the simulation with drain coefficients of 1,000 m^2/d/cell, the total coastal inflow is almost 20,000 m^3/d.

4) Both types of interceptor drains induce increased groundwater inflow from the Horsey coastal section. From a baseline of 921 m^3/d, the 133 m^2/d/cell interceptor drains induces 1,405 m^3/d whereas the 1,000 m^2/d/cell induces 2,572 m^2/d/cell.
5) The induced seepage, percentage wise is drastically reduced for the Somerton New coastal section.

The ratios of drainage coefficient and total coastal inflow suggest that the volume of coastal inflow is *dependent* on the value of the drainage coefficient of the interceptor drain, but is not directly *proportional* to it. There is simulated coastal outflow at Happisburgh, which occurs even with an interceptor drain with a large drainage coefficient, suggesting that the influence of the interceptor drain might be greater to the south along the coast.

Table 9.3 gives the estimated groundwater flows into the Brograve sub-catchment and the flows into the interceptor drain itself. The flows into the Brograve drainage systems are 4,396 m^3/d with the interceptor drain with drainage coefficients of 133.3 m^2/d/cell in place, whilst flows into the interceptor drain itself are 8,160 m^3/d. With the increased 1,000 m^3/d/cell drainage coefficients, the interceptor drain is collecting 23,773 m^3/d, which is approximately a threefold change. This suggests that volume of flows into the drain itself is not proportional to the drainage coefficient of the interceptor drain.

Table 9.37 The comparison of the inflows with the implementation of an interceptor drain (m^3/d)

	standard		Cd = 133.3 m^2/d/cell		Cd = 1000 m^2/d/cell	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Brograve drains	36	5862	49	4396	160	2278
Interceptor drain	n/a	n/a	0	8160	0	23773

One of the important observations to be made regarding the effects of the interceptor drains on the inflows into the Brograve drains is the volumetric pumping of adding an interceptor drain. However, the forward particle tracking reveals that only a small proportion of coastal waters passes the interceptor drain and continues to flow further inland, as shown in Figure 9.7 suggesting that an interceptor drain could *in theory* significantly reduce the salinity of the water being discharged by the Brograve pump. However, Figure 9.7 also demonstrates that such an interceptor drain induces increased flows of fresh groundwater from the higher land to the north west.

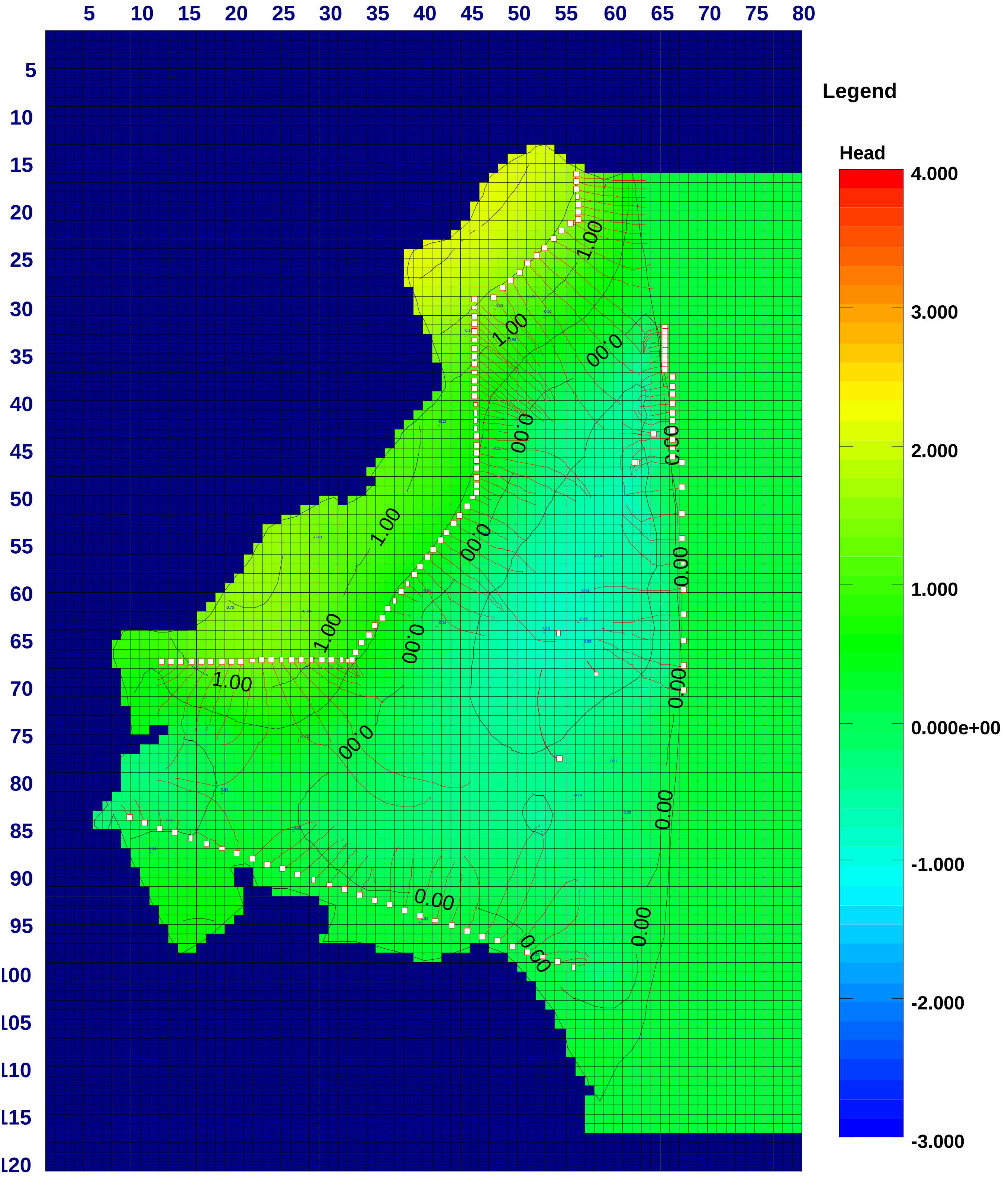

Figure 9.8 The catchment head distribution for an interceptor drain with Drain Coefficient=133m^2/d/cell and resultant particle tracks (Layer 2)

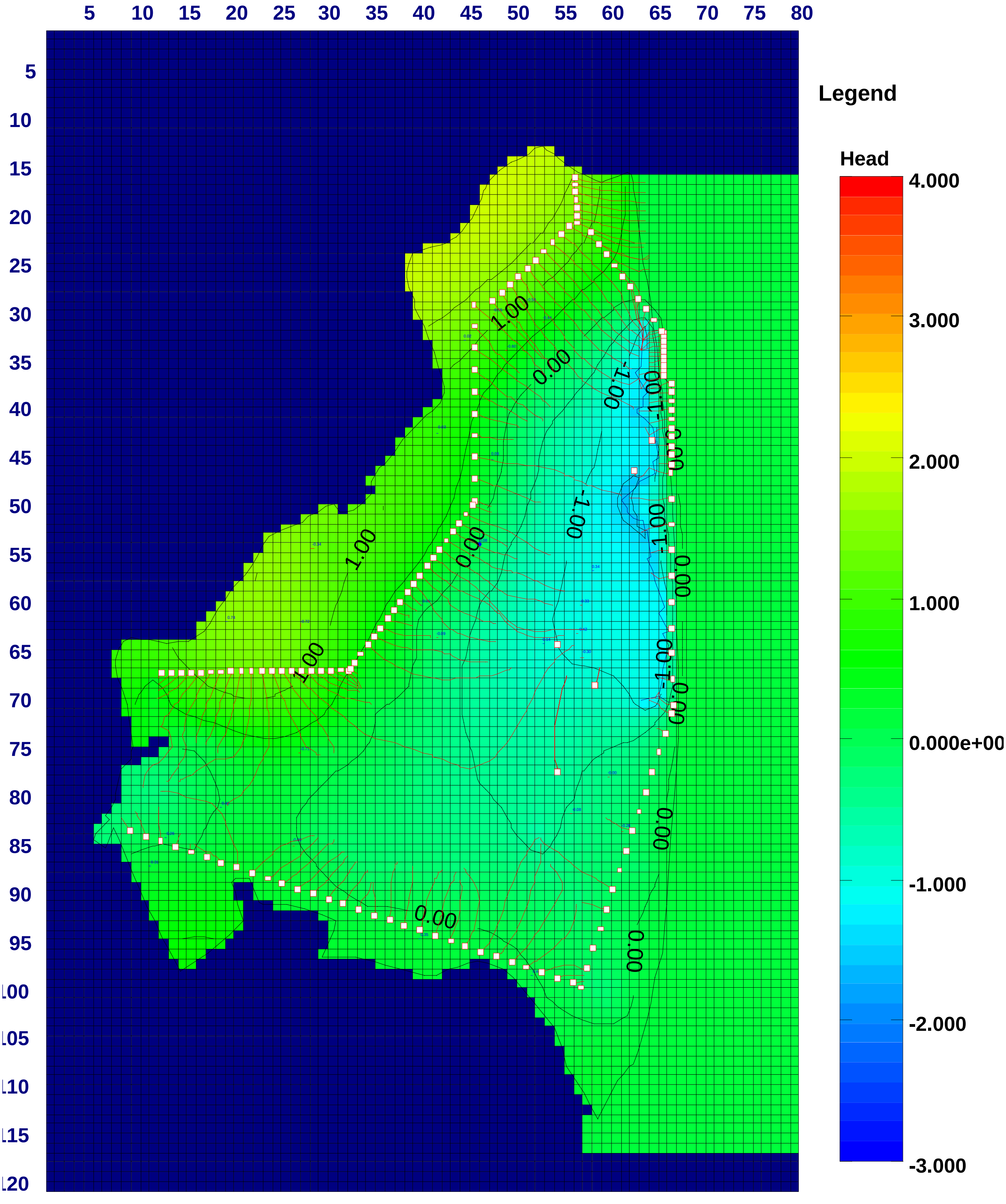

Figure 9.9 The catchment head distribution for the interceptor drain with Drain Coefficient =1000 m^2/d/cell and resultant particle tracks (Layer 2)

9.4 Prediction 4-The lining of the main drain in the Hempstead Marshes

9.4.1 The scientific rationale for the lining of the main drain at Hempstead

In a study conducted by Bhutta and Wolters (2000) on canal lining in the Fordwah Eastern Sadiqia South (FESS) project, the effect of lining canals was studied using two separate *methods* of lining (i) a standard lining or 'production' lining and (ii) an 'experimental' lining. The production lining consisted of 0.75 mm thick very-low density-polyethylene, protected with 75 mm thick in-situ concrete, with an intermediate layer of 200 or 350 g/m^2 geo-textile. The construction was executed in dry conditions, which added considerably to the cost.

The other method, which used various experimental materials, comprised an installation of nine broad types of lining systems with a 0.5 mm thick polypropylene alloy. The other experimental materials included the usage of rapid lining methodologies (to reduce costs) such as concrete with sealed joints or an impermeable membrane with concrete or soil covering. Both lining methodologies included impermeable geo-membranes and both resulted in an average reduction of about 70% of the initial seepage (Bhutta and Wolters, 2000). However, the investigators found that even though the linings were installed in an unconfined aquifer of medium to high hydraulic conductivity, the initial seepage from the canals was not high enough to justify the expenses, and that they would not reduce the drainage requirements enough.

In the current study, it is crucially important that the proposed linings be made of semi-permeable material due to the possibility of reverse pressure exerted by the crag aquifer beneath the lining. This will allow a long-term construction to respond effectively to the aquifer in a fashion similar to a drain that has a thick low-permeability base. The reduction percentage given by Bhutta and Wolters (2000) will be used as a guide to reduce the standard model drainage coefficients.

9.4.2 The required adjustment to the numerical groundwater model

The lining of the Hempstead main drain was simulated by reducing the drainage coefficient for the 12 drainage cells that represented the Hempstead main drain within the model (Figure 8.8). The original drainage coefficients were 50 m^2/d/cell, which were subsequently reduced to 15 m^2/d/cell. .

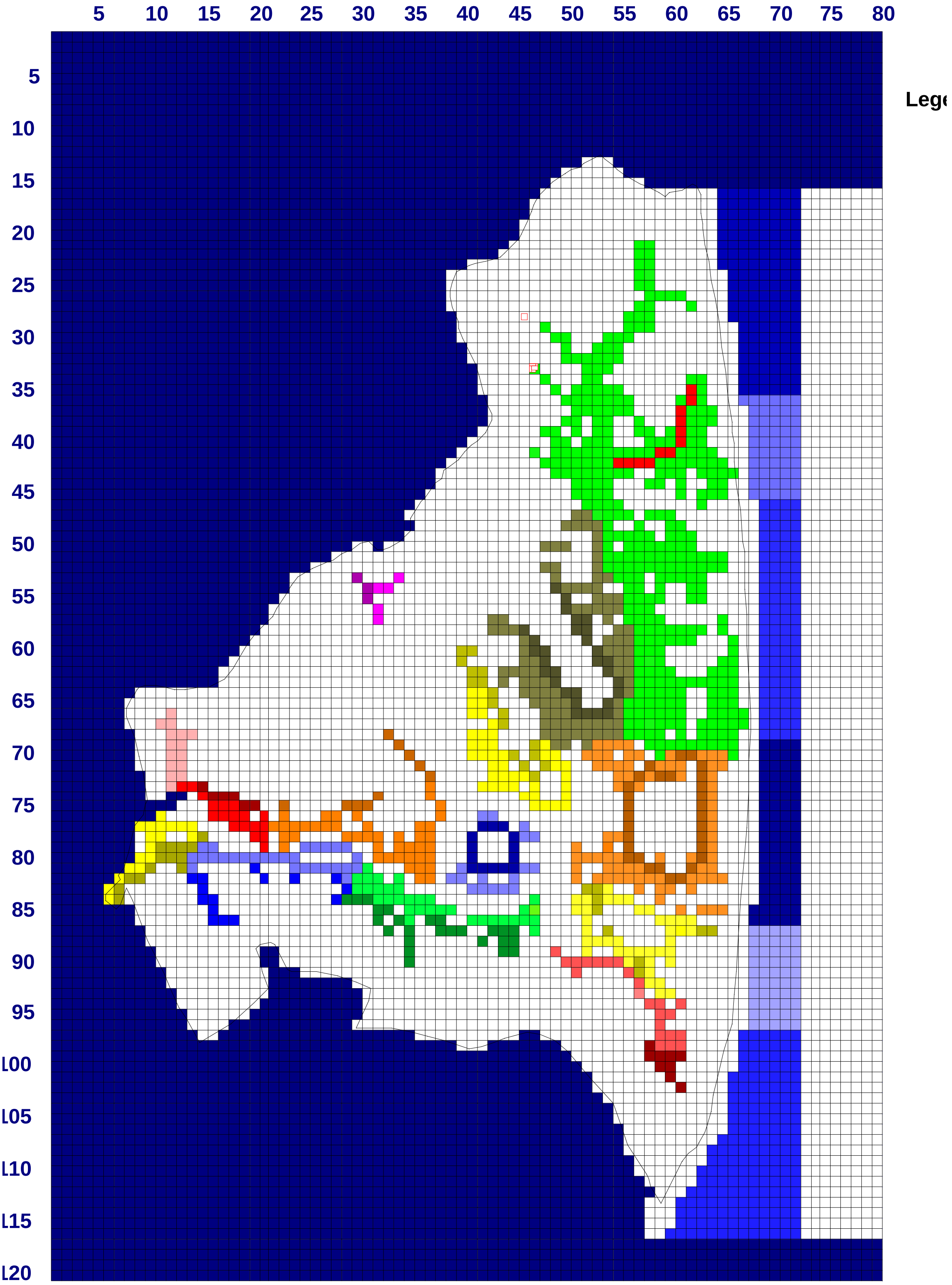

Figure 9.10 The locations in the modelling domain (red squares within the northern green area) where the adjustments in drainage coefficient to reflect the proposed lining of Hempstead Marsh main drain is made

9.5 Results

Although the drain coefficient was reduced by 70%, the responding coastal inflows in the Hempstead coastal reach reduced by only 35%, which is fifty percent efficiency rating. In addition, the simulated lining of the main drain at Hempstead has increased the coastal outflow in the Happisburgh coastal reach, produced new coastal outflows in the Hempstead coastal reach, and reduced the saline inflows in the Brograve coastal reach. The output particle-tracking plot for the numerical groundwater model with the Hempstead main drain lined is presented in Appendix B.

Table 9.38 The effects of the lining of the main drain at Hempstead (m^3/d) on average daily coastal flows

	Scenario 1		Lining the Hempstead main drain	
Coastal	**Into Aquifer**	**Out of Aquifer**	**Into Aquifer**	**Out of Aquifer**
Happisburgh	0	1873	0	1962
Hempstead	272	0	175	16
Brograve	1437	0	1412	0
Horsey	921	0	920	0
Somerton New	270	0	270	0
Somerton Old	181	213	181	218
Total	**3081**	**2086**	**2958**	**2196**

9.6 Summary

The prediction that combines lowering of the water level within the Lessingham drains and the raising of the water level within the Hempstead drains seems the most plausible approach to reduce long-term saline inflows into the Brograve system. This course of action allows for the hydrology of the major water bodies to be maintained as the Brograve pump discharges similar volumes of water as currently.

The realistic (133.3 m^2/d/cell) interceptor drain does not lower the heads in the aquifer enough to intercept the required amount of coastal inflow. The required interceptor drain of 1,000 m^2/d/cell is almost certainly impossible to maintain hydraulically for any period. This is because levels of energy loss incurred by the turbulent flow within the pipe generated by the mixing of the entrance flow and the longitudinal pipe flow would be substantial.

The lining of the main drain at Hempstead by itself is relatively ineffective, given that the actual reduction in the drainage coefficient is 70% whereas the reduction in groundwater inflow is only around 35%. The reason for this is the low water levels still maintained within the Hempstead drains.

10 DISCUSSION AND WIDER IMPLICATIONS

10.1 Introduction

This chapter considers the major issues concerning the development of the understanding of the hydrogeology of the Upper Thurne catchment and the recommendations for salinity management. This includes the important steps taken in the development of the numerical model and the furtherance of knowledge obtained by its implementation. The last section discusses the wider implications of the research completed, the general contributions to hydrogeology and the areas in which further research can be undertaken.

10.2 The development in the understanding of the hydrogeology of the Upper Thurne catchment

10.2.1 Developing a numerical groundwater model of the Upper Thurne catchment

The problems associated with saline coastal groundwater inflowing into the upper Thurne catchment is well-documented (Pallis, 1911; Driscoll, 1985; Holman, 1994; Harding and Smith, 2002). However, little work had been completed in terms of groundwater modelling and the quantification of the drain-aquifer processes. Therefore, one of the objectives of this research project was to further the understanding of the groundwater system and investigate the influence that individual drain configurations have on the saline coastal water distribution within the drainage network.

To further the understanding of the groundwater behaviour the collection of new data or the re-interpretation of historical information is needed. The historical information available concerning groundwater behaviour includes the measurements of water levels taken within observation wells distributed throughout the catchment over a two-year period from April 1st 1991 to March 31st 1993. The other quantifiable historical data available for the two-year period are (i) the electricity usage of the discharge pumps and (ii) spot readings of drain-water both near and at a distance from these discharge pumps. These measurements were taken by Holman (1994) and provide useful insights into the influence of the various drains and the distribution of salinity within the various sub-catchments. The available data are interpreted and additional supplementary data collected to inform the identification of target parameters so that a process of manual trial-and error adjustment could be undertaken to ‘calibrate’ the model.

10.2.2 Conceptual themes

Several worker emphasise that the correct conceptualisation of the modelling domain is a pre-requisite to the modelling process.(Anderson and Woessner; 1992; San Juan and Kolm, 1996; Rushton, 2003) Therefore, there is a need to conceptualise the modelling domain, identify the usefulness of the historical data and determine whether more data are needed. If more data are needed, the data should be collected and the extent of the advancement of understanding determined. In analysing the usefulness of the available historical information, three important conceptual themes are identified:

(i) Deep regional groundwater flow

(ii) Drain aquifer interactions

(iii) Surface water flow within drains

Detailed information on the structure and geometry (Downing, 1959; Holman, 1994) and on the hydraulic conductivity (Erskine, 1991) of the crag aquifer are available, whilst limited data are available for the drain-aquifer interactions within the catchment. These drain-aquifer (or stream-aquifer) interactions influence the nature of groundwater behaviour. Several investigators have attempted to quantify them (Prickett and Lonnquist, 1971; Rushton and Tomlinson, 1979) because they are normally included in regional groundwater models in the form of drain coefficients. The current MODFLOW (McDonald and Harbaugh, 1988) approach is based on the thickness of drain bed deposits and was developed for gaining drains (Prickett and Lonnquist, 1971) and ignored the properties of the aquifer. In this study, result have been presented that corroborate with the findings of Fleckenstein et al. (2006) that drain-aquifer interactions are governed by the hydraulic properties of both aquifer and aquitard materials beneath the drain.

The previous understanding (Holman, 1994) of the Brograve main drain was that it was a major source of saline intrusion by upward flow from the underlying aquifer immediately below that section of the drain. However, several indicators from collected data suggest that the hydrological influence of the main drain may not be this dominant. The salinities of the drains before and after the Hempstead-Lessingham confluence and just before the Brograve pump, indicates that the saline contributions from the groundwater seepage is not enough to alter substantially the main drain salinity (Figures 4.7 and 4.8). Other collected information such as the fluctuations within the observation wells within the Brograve sub-catchment indicate that the groundwater heads are influenced by drain-aquifer interactions other than the Brograve main drain. The observation wells' water levels fluctuate whereas the water level within the Brograve main drain is maintained effectively at a constant level.

It assumed that any error incurred in the measured water levels within the observation wells is minimal and that the fluctuations are genuine increases in groundwater head. This allows for the grouping of the observation wells according to their individual hydrographical fluctuations. In grouping the observation well hydrographs, it is recognised that the groundwater response to the effective recharge varies both spatially and temporally. Several influences are recognised as affecting the local fluctuation in water level, (i) the magnitude of the local drain-aquifer interaction (ii) the local recharge mechanism (iii) the local aquifer conditions. In developing the drain-coefficients, by use of vertical cross-section finite element model??????(Section 6.6.2), it is noticeable that the drains with a significant contact with the underlying aquifer induce a substantially larger flow and so disproportionately influence the groundwater head fluctuations. Therefore, this research has indicated that within any heavily drained aquifer, the drain-aquifer interaction is one of the major influences of the groundwater flow.

10.2.3 The development and usefulness of target parameters

The purpose of identifying target parameters is to enable the *quantitative* determination of a model's performance (Anderson and Woessner, 1992; Rushton, 2003; Hassan, 2004). The simulation of the groundwater behaviour within an aquifer requires data concerning a groundwater balance (Gerber and Howard, 2000; Rushton, 2003) without which it is difficult to pursue adequately the modelling exercise. So the use of groundwater head distributions and target volumes of saline coastal groundwater inflows (Huyakorn et al. 1987) in the current study assist in the improvement of the numerical groundwater model. Four targets are identified and quantified (the total groundwater balance, the coastal water contributions, the drain locations identified as having high salinities and the groundwater head distribution), which when identified can assist with understanding the usefulness of the numerical model.

10.2.4 The development and analysis of the numerical groundwater model

The existence of the three interlinked conceptual themes (Chapter 4) makes possible the estimation of the distribution of the surface water salinities by the analysis of the drain-aquifer interactions and the deep regional groundwater flow. The drain-aquifer interactions are quantified by drainage coefficients that are assigned to individual cells within the groundwater model. The initial numerical groundwater model developed required further adjustments to the total drainage coefficient for the Brograve sub-catchment because the *modelled* groundwater contribution was substantially greater than the *estimated* groundwater contribution. In refining the numerical groundwater model, the modelled coastal inflow volume was found to be comparable to the estimated coastal volume (Section 6.4.2) indicating that the simulation replicated the primary and secondary target parameters (Section 8.2.3) adequately.

The development of the numerical model also required the introduction several new techniques, which represent the unique characteristics of a heavily drained coastal catchment. The coastal shelf (Section 6.4.4) is represented by 'square in plan' river cells that have a river bottom (RBOT) with a very high conductance (of > 10^7 m^2/d) and a negligible thickness (of <10^{-7} m). This allows for a calculation of the volume of water entering or leaving the coastal grid cells. The introduction of a pseudo layer (Section 7.4.2) beneath the cultivated top layer is another novel technique, which is necessary because within MODFLOW (McDonald & Harbaugh, 1988) the computation of the vertical flow into a drain is partially determined by the thickness of the layer immediately beneath it. In the case of the Upper Thurne, this would mean a drain would have a theoretical depth of influence of approximately 23.5m. So in order to represent realistically the vertical flow entering the drain a pseudo layer of Crag of 1 metre thickness is introduced.

In verifying the consistency of input data and analysing output data, the use of the geographical information management system ARCVIEW proved to be an efficient tool. The regional database contains geological and hydrogeological input information for each cell within each layer of the numerical groundwater model and allows the collation of flow data of coastal cells, which are subsequently used to quantify the inflow within the numerical model. The development of the smaller databases of the drain properties made the easy identification of the required model adjustments under each proposed scenario possible.

10.2.5 Summary of the improved understanding of the Upper Thurne system

This research has shown that within the Upper Thurne catchment, there are at least three interlinking conceptual themes; namely, the Deep Regional Groundwater Flow (DRGF), the

Drain-Aquifer Interactions (DAI) and the Surface Water Flow along the drains (SWF). These themes show that the understanding of the salinity within the surface water network is possible by the use of fine-grid modelling of the drain-aquifer interactions and the numerical simulation of the groundwater flow. This numerical simulation requires target parameters that are necessary to assess the model performance. These target parameters ((i) Total groundwater balance, (ii) coastal water balance, (iii) drains locations identified as having high salinities and the (iv) groundwater head distribution) each have a level of uncertainty in their individual estimations. The actual total groundwater balance is obtained by use of estimated electrical conversion factors of the discharge pump and a new hydrograph separation technique. Both of these estimations incorporate a level of uncertainty, although the effects of this composite uncertainty can be reduced by comparing the relative changes within the groundwater behaviour after each salinity management proposal. The coastal water fraction in each sub-catchment is estimated as a fraction of the total groundwater volume thereby masking the error within the initial total groundwater estimation. The effects upon the groundwater flow regime caused by the error in the estimation of the total groundwater balance and the coastal fraction is an area where further research is possible.

10.3 Salinity management recommendations for the Upper Thurne catchment

10.3.1 Introduction

At the beginning of this research project, it was recognised that the Upper Thurne catchment was important from a national and international perspective. There had been several proposals to reduce the saline outfall from the Brograve pump, but no real quantification of any improvements. In this body of research, an adequate groundwater model is developed and the consequences of the three options on groundwater flow that were originally suggested can now be quantified. The recommendations for the remedial measures will consider the individual proposals in the light of this research and previous investigations on wetlands management.

10.3.2 The adjustment of water levels of drains within the Hempstead marshes and Lessingham

The Hempstead and Lessingham upstream (Section 9.2.2) branches contribute to the volumetric output of the Brograve pump and given that the River Thurne is tidal, any reduction in input into the River might allow the increased upstream incursion of saline water up the river Thurne. It is suggested that the water levels within the Lessingham area can be lowered to compensate for reduced flow from the Hempstead marshes. The raising of the water level within the Hempstead Marshes will reduce the hydraulic gradient between the drains and the coastal water subsequently reducing the saline inflow into the local drains. The long-term land use in around the Hempstead marshes must also be considered and realised before any adjustments are made to the water levels. If the local land users decide that they no longer require the draining of the Hempstead marshes for agricultural purposes then the raising of water levels to -0.4 mOD can be implemented immediately. Further research is required to determine the actual contributions of the individual drainage branches of Lessingham and Hempstead, but the simulation of raising the water levels within the Hempstead Marshes would reduced the fraction of saline waters discharged by the Brograve pump.

10.3.3 The construction of an interceptor drain along the coast

The simulation of two different interceptor drains with substantially different drainage coefficients is presented in Section 9.3.3. The *physically maintainable* interceptor drain of 133.3 m^2/d/cell would not sufficiently reduce the saline inflow from the coast, whereas the required drain coefficient of 1,000 m^2/d/cell would be almost certainly impossible to maintain for any period. The construction of *any* interceptor drain is *not* viable as the net effect of the drain would be to dewater the aquifer and encourage into the catchment *more* saline water and therefore *must not* be implemented under any circumstances.

10.3.4 Lining of the main drain at Hempstead

The final prediction investigated involves the re-lining of the main drain at Hempstead with geo textile material. This approach to salinity management is only 50% efficient as the linings reduce the drainage coefficient of the main drain by 70% but only reduce the saline inflow into the catchment by 35%. This can only be implemented if the long-term need to drain the Hempstead Marshes is determined by the local land users. A long-term financial commitment to the upkeep of the drain therefore must be in place before any work is commenced. It is possible that the relining of the drain is not financially viable as was found by Bhutta and Wolters (2000).

10.4 General contributions to hydrogeology

10.4.1 Introduction

This section presents the more general insights and major contributions that can be used in groundwater studies and recognises that in developing a groundwater model of the coastal catchment several procedures and methods were implemented to obtain estimates of input parameters. These procedures and methods provided the opportunity for the development of an understanding of the Upper Thurne catchment and modelling techniques in general. The construction of numerical models of vertical cross-sections of typical drain configurations furthered the knowledge of drainage coefficients for gaining drains. A discussion on the feasibility of linear drainage coefficients given the prior approximations of pump discharge volumes and the groundwater fraction is presented. The theory of hydrographical groupings and the identification of the various factors that influence the measurement of groundwater heads in observation wells are investigated. The development of a hydrograph separation technique and the validation of the use of a single density groundwater model to represent a variable density groundwater situation are also described.

10.4.2 Drainage coefficients for gaining drains

The previous approach for quantifying the drain-aquifer interactions, which focused on *losing* drains was suggested by Prickett and Lonnquist (1971) and required the use of a linear combination of the drain's width and the vertical hydraulic conductivity and thickness of the drain's basal material. This current research shows this approach to be inadequate and an improved estimation of quantifying drain-aquifer interactions for gaining drains is possible by use of non-linear drainage coefficients (Section 6.6.2). These drain coefficients are deduced using numerical vertical cross-section models that include the typical dimensions of the drain with a reference water level, the magnitude of the hydraulic conductivity of the aquifer material K_a and the hydraulic conductivity of the material between the base of the drain and the aquifer surface K_b. This work reveals the importance of recognising that the hydraulic conductivity of the aquifer material K_a is dominant for drains in which the base is closer to the underlying aquifer, whereas the hydraulic conductivity of the aquitard material K_b is dominant for drains that are less penetrative (Section 6.6.2).

10.4.3 The importance of drainage coefficient magnitudes

The estimation of the total drainage coefficient for any sub-catchment is dependent upon two previous estimations. These two estimations are (i) the total discharge volume of the electrical pump serving the sub-catchment using unverified electrical conversion factors (Section 4.2.7) and (ii) the groundwater fraction entering the drainage network using a new hydrograph separation technique (Section 6.6.2). In accommodating these estimations in the process of determining the values of drainage coefficients for the individual sub-catchments, a reduction in precision is accepted. Therefore, the individual drainage coefficient allocated to a cell within the numerical groundwater model is an approximation of the *magnitude* of the local drain-aquifer interaction and so there is little justification in allocating a value any more precise than a ½ order of magnitude (i.e. 1, 5, 10, 50, 100, 500 and 1000 m^3/d/cell). This is possibly because in this situation the aquifer-drain flow is not very sensitive to the magnitude of the drainage coefficient.

10.4.4 The development of a hydrograph separation technique for flows into drains

The separation of a flow hydrograph from a pumped drainage system is possible using the flow equation (7.1) provided the required data are available. The data required are (i) one year's flow data from the drainage pump (weekly), (ii) one year's head data (weekly) for both the aquifer and the gaining drain (stream). This estimation of groundwater contribution to the total flow assumes that the groundwater contributions are directly proportional to the head difference between the head in the local aquifer and the head at the base of the gaining drain. In estimating the constant of proportionality, two approaches are investigated resulting in three different values, with the eventual approach implemented rendering a groundwater contribution for the Brograve sub-catchment of 70%. The estimation is sensitive to (i) the period used for the approximation of 'groundwater only' contributions (dry period) and (ii) the estimated head difference between the aquifer and the gaining drain (Section 7.2.3). The research has also shown that the estimated profile of the groundwater contribution when superimposed on the flow hydrograph should (i) coincide with the 'groundwater only' contributions (dry period) and (ii) reveal the extended peaks that represent large volume of rainfall runoff contributions. The overall use of this method can be extended to estimate the baseflow percentage entering rivers.

10.4.5 The use of a single density groundwater model to represent an aquifer containing water of different saline concentrations

By considering only the convection process, this body of research has confirmed that the use of a single density groundwater model to represent an aquifer containing water of different saline concentrations is a valid approach. This approach requires additional constraints such as (i) Total groundwater balance (ii) the coastal flow estimation (iii) that the flow regime within the aquifer is physically described (i.e. particle tracking). The estimation of the total groundwater balance and coastal inflow fraction provides the worker with a quantitative understanding of the groundwater behaviour. The use of particle tracking allows for the identification of the source and destination of the important flow paths.

10.4.6 Conclusions and further research

Generally, the recommendations of the predictive simulations emphasise the necessity of an integrated approach that combines land-use management as well as the influence of the drain-aquifer interactions. The adjustment of any hydrological influence must be carefully examined before implementation. It is also of the utmost important to consider the financial, social and hydrological impact any change may have before any hydrological change is implemented.

The recommendations arising from this study concern both the Upper Thurne catchment and groundwater studies in general. Within the Upper Thurne catchment, further fieldwork is necessary so that areas of uncertainty within the current understanding can be removed. Of particular concern is the calibration of the flows from the electrical discharge pumps, which would reduce the uncertainty of the total groundwater volume within the drainage system. Once the total groundwater volume is estimated, the volumetric contributions of individual drains could be justifiably approximated using flow meters. The contributions of individual drains are important because current field evidence suggests that specific drains are disproportionately contributing to the saline water discharge of the various pumps, in particular the Brograve pump. Another concern is the salinity of groundwater. Whereas the salinity of the drain water has been measured using electrical conductivity meters and chloride concentration tests, the salinity of groundwater is unknown.

In addition to the fieldwork, further desk studies concerning the Upper Thurne catchment are necessary and include the development of a time-variant groundwater model by implementing dynamic recharge calculation using the Soil Moisture Balance technique (Eilers, 2002). Other investigations are needed to determine the influence that the variable mesh spacing has on the simulated groundwater flow regime.

The general contributions to groundwater studies include the development of a new hydrograph separation technique for drains, which can be extended to estimating the baseflow contributions to rivers. The sensitivity of drainage coefficients suggests that the magnitude is more important than the actual value attributed to the groundwater cell and that a precision of ½ order adequately represent the full range of drain-aquifer interactions. The final contribution to groundwater studies presented by this study is the verification that a groundwater flow model can adequately represent the distribution of flows within an aquifer with water of different salinities.

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APPENDIX A: MODEL INPUTSCREENS

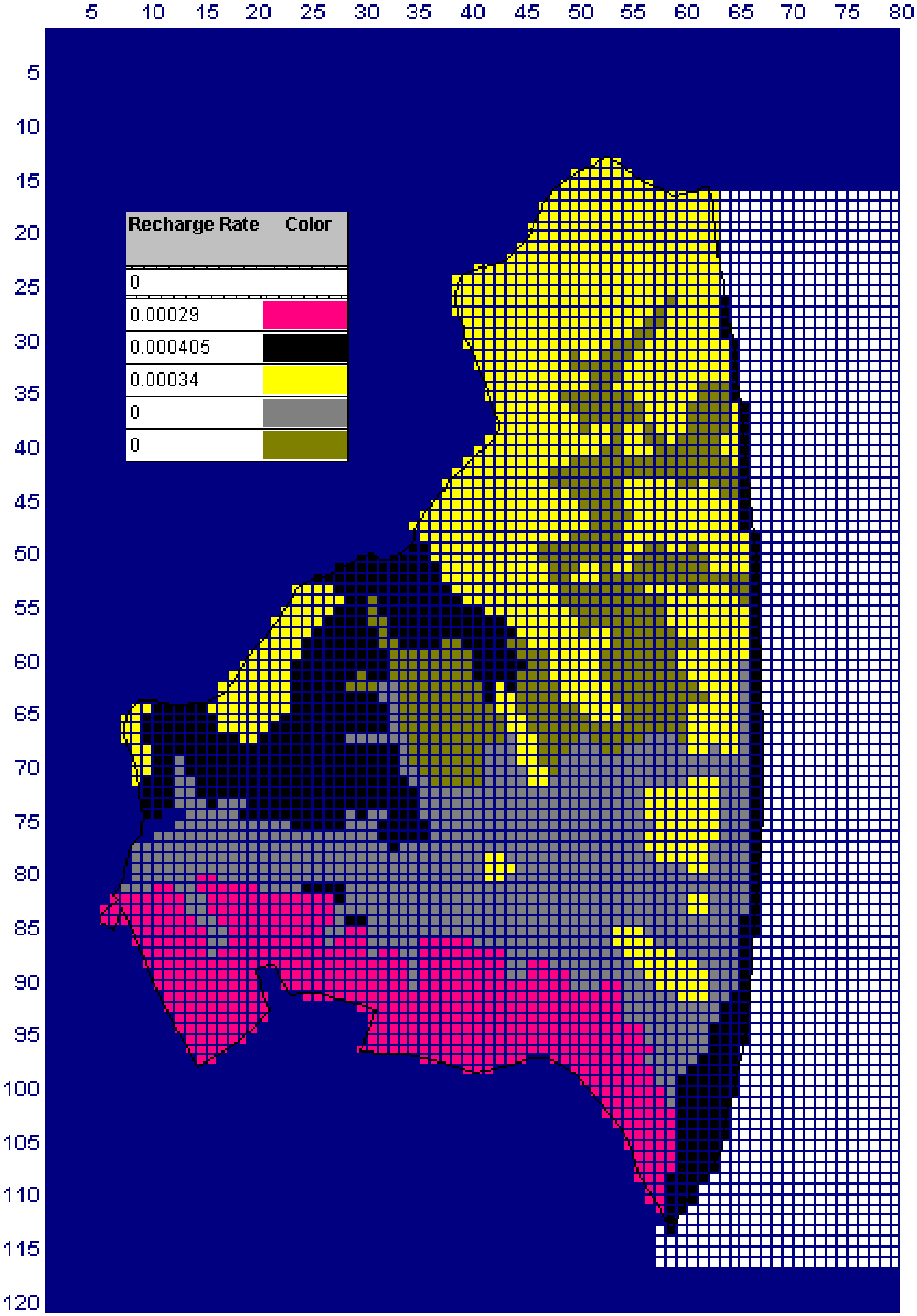

Figure A1 Recharge distribution (m/day)

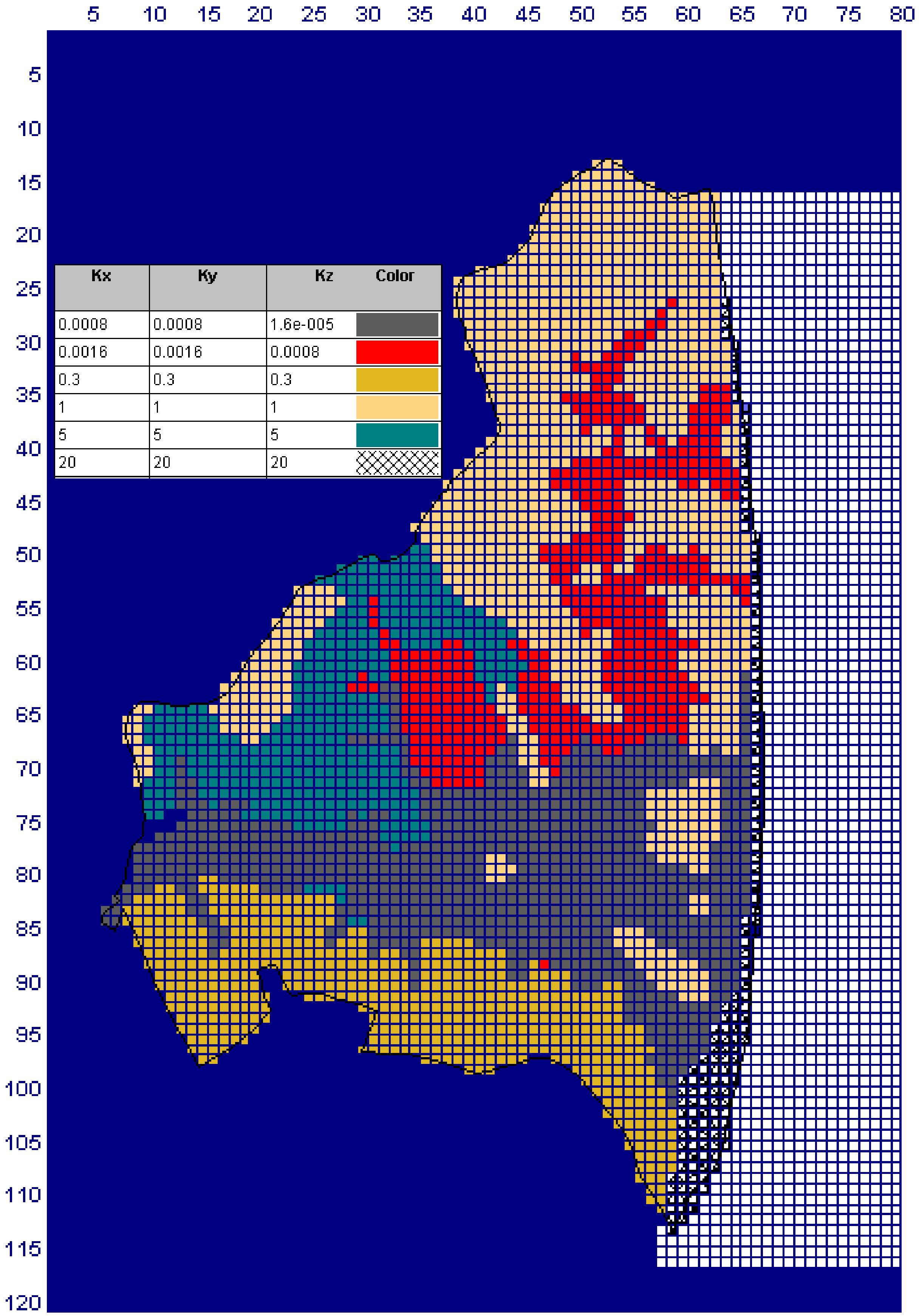

Figure A 2: Hydraulic Conductivity distribution (m/day)

APPENDIX B: MODEL OUTPUT SCREENS

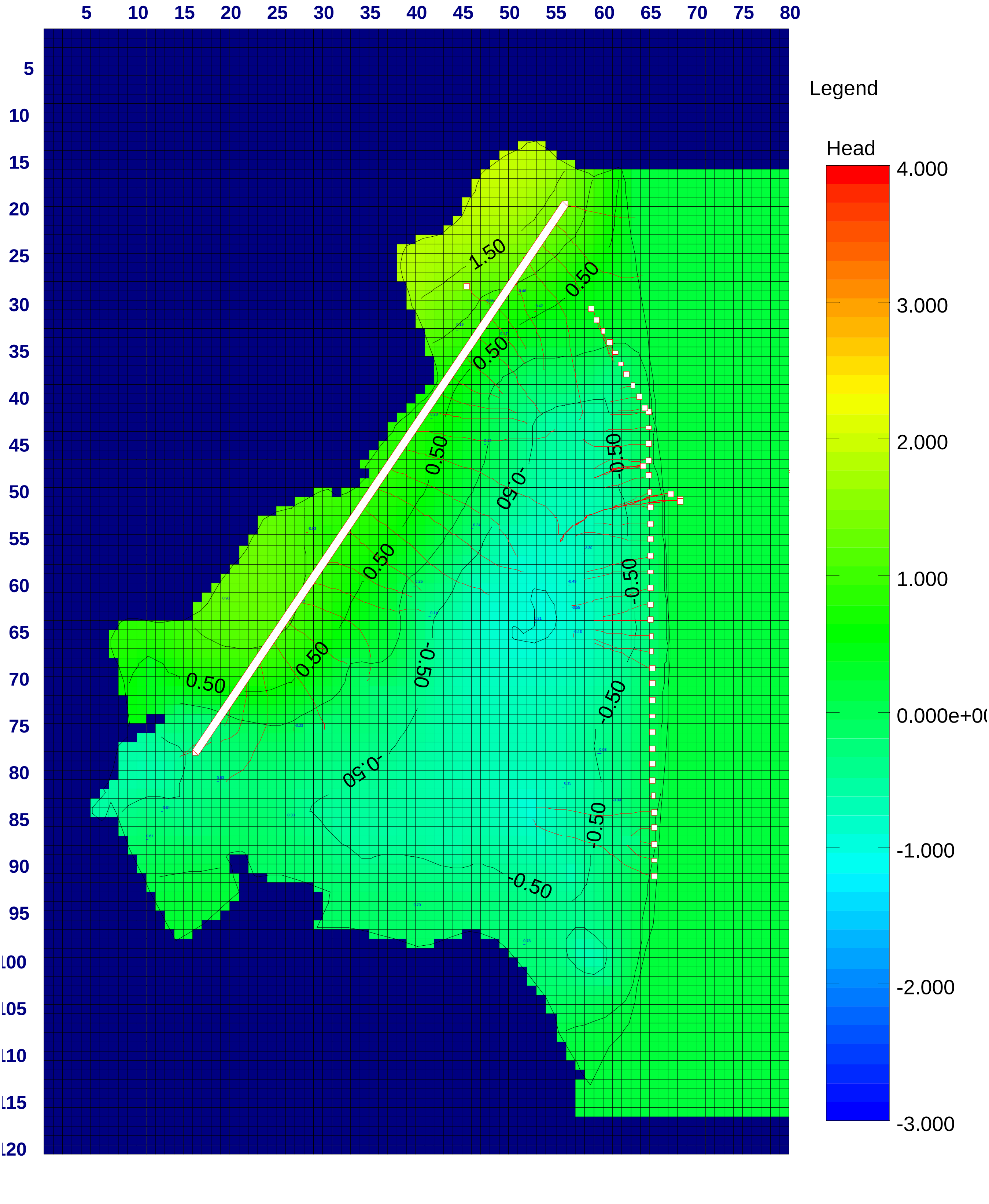

Figure B 1 Twice × Drain Coefficients

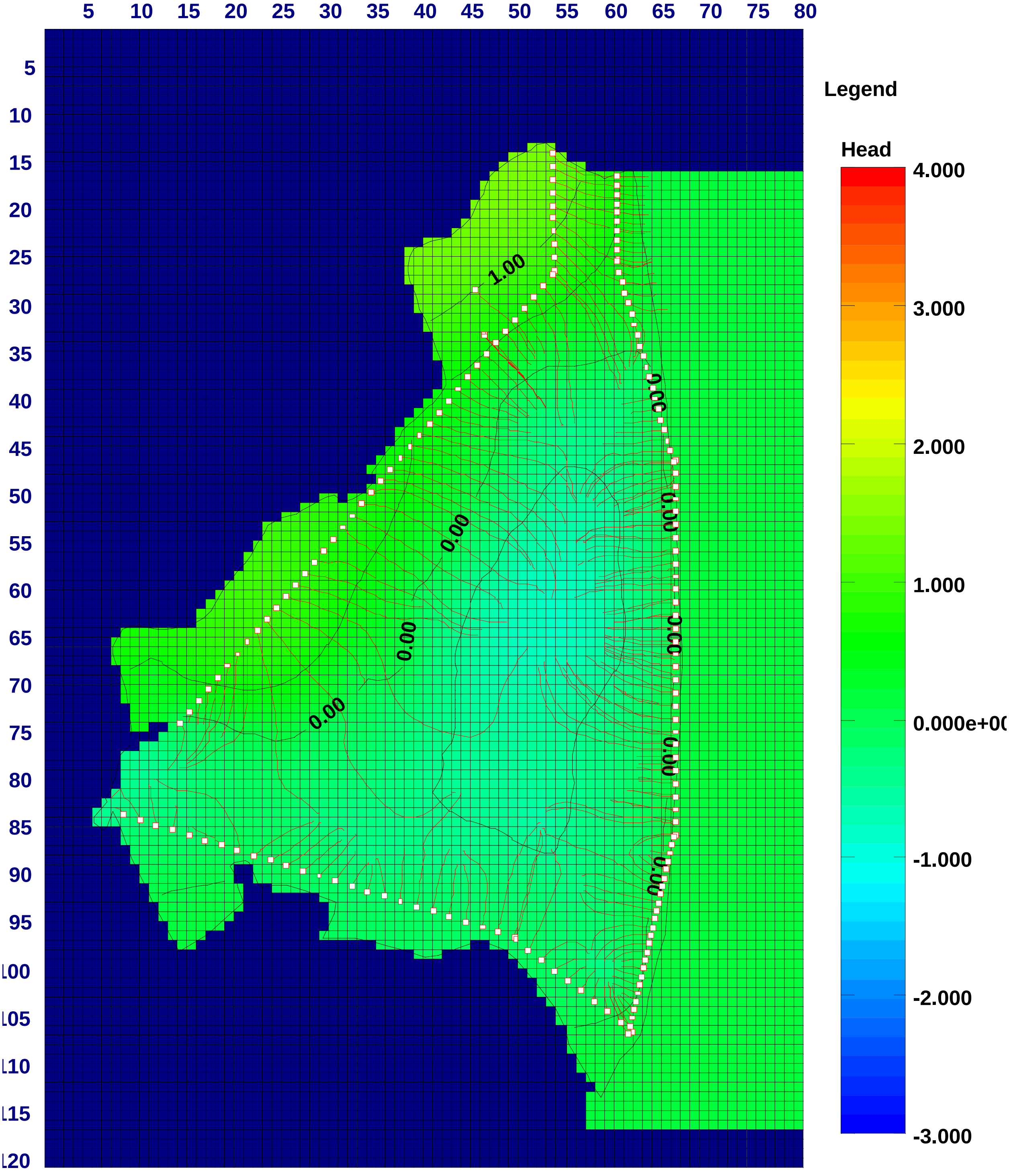

Figure B 2 Summer conditions (66% Average recharge)

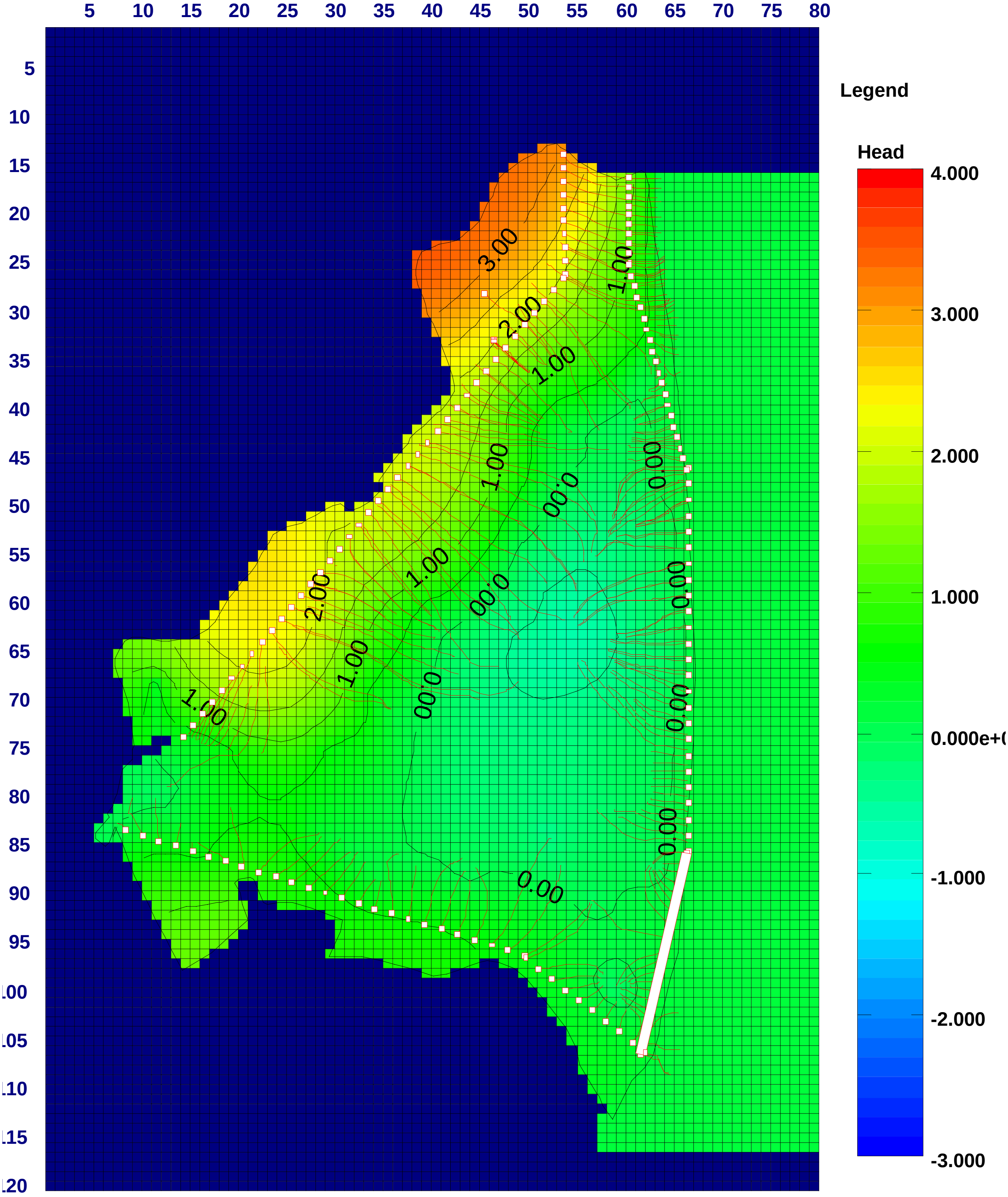

Figure B 3 Winter conditions (150% Average Recharge**)**

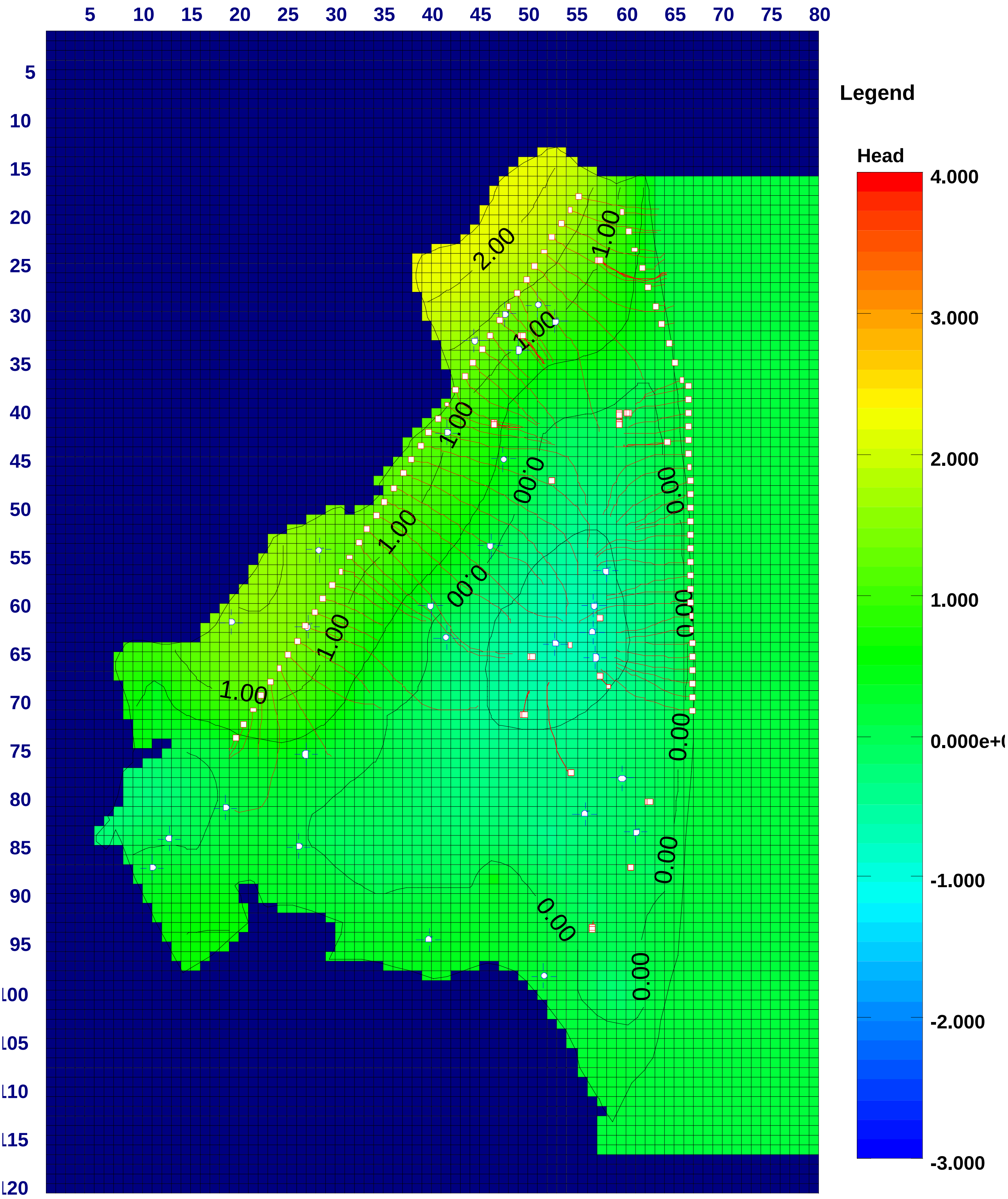

Figure B 4 Adjustment of Water levels in Lessingham and Hempstead

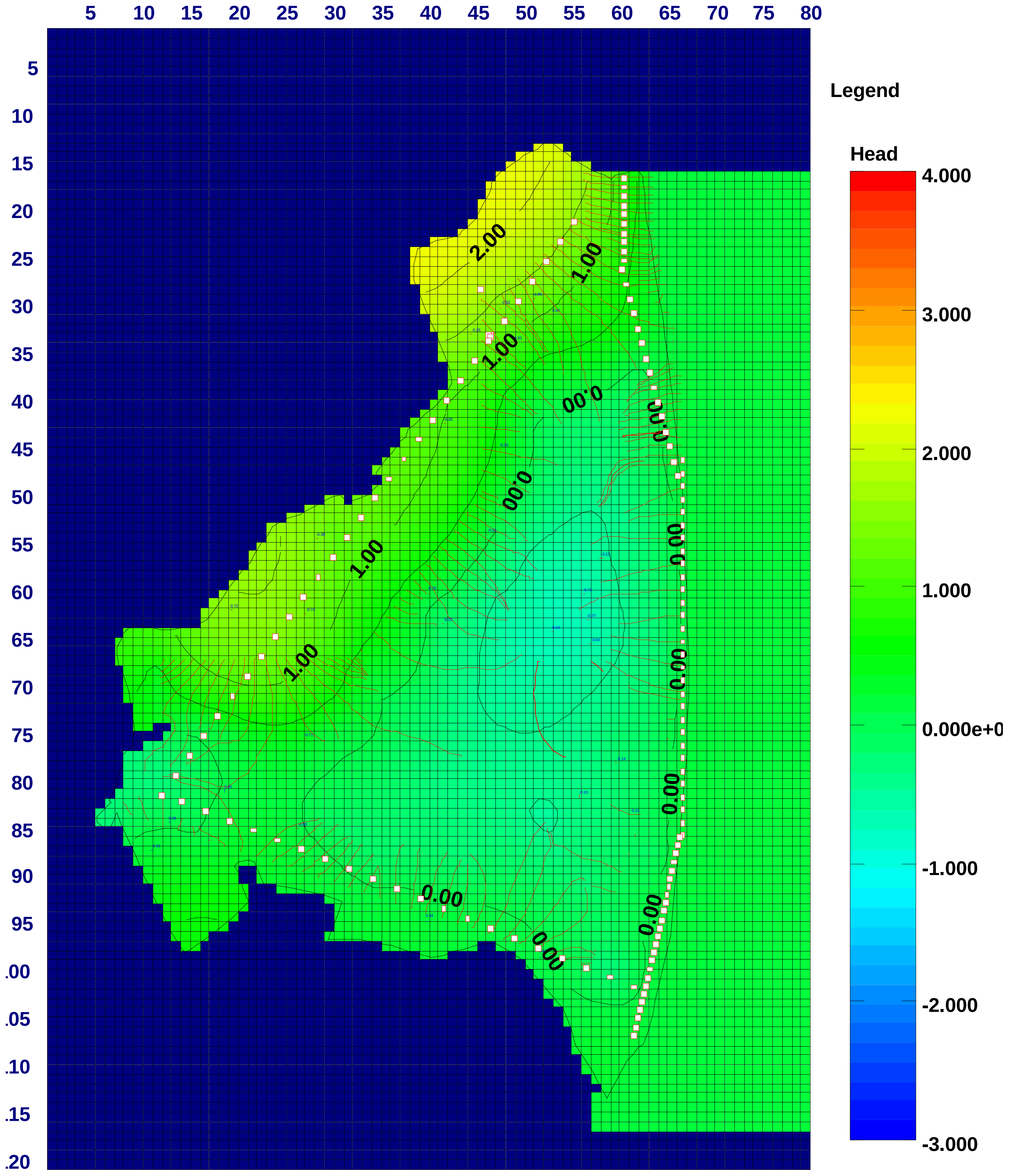

Figure B 5 Lining the Hempstead main drain