HYPHOREIC ZONE HYDRAULIC TESTING:

RIVER TAME, BIRMINGHAM, UK

by

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Abstract

The hyporheic zone is a dynamic ecotone, within which important physical, chemical and biological processes occur. It is believed that the hyporheic zone has the potential to naturally attenuate contaminants and limit their exchange between groundwater and surface waters. Fieldwork was conducted on an urbanised stretch of the River Tame in north Birmingham in order to characterise the hydraulic properties of the unconfined sandstone aquifer and river bed sediments. The hydraulic conductivity of the Kidderminster Sandstone was calculated as 2 m/d. Groundwater head measurements indicated that the Tame was gaining along the entire study reach with a mean hydraulic gradient of 0.1036. The river bed sediments were found to be very heterogeneous with their hydraulic conductivities ranging from 0.61 m/d to 52.96 m/d, the mean hydraulic conductivity is 5.28 m/d. Groundwater flow modeling of a proposed pump test next to the river indicated that pumping at a rate of 200 m^3/d will reduce the observed hydraulic gradients by between 2.07% and 48.51%, this is very much dependent on the hydraulic parameters for the river bed sediments, and to a lesser extent, the superficial drift deposits.

Acknowledgements

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Chapter 1. Introduction

1.1 Project Background

The hyporheic zone represents a critical interface between groundwater and surface water. It is a dynamic ecotone within which important microbial and geochemical processes take place, which are thought to have an effect on contaminant transport due to natural attenuation. The introduction of the EU Water Framework Directive has led to a more integrated management of surface and groundwater resources, increasing the need to understand contaminant movement through the hyporheic zone.

This project forms part of the early stages of the SWITCH project which aims to improve the conceptual understanding of the hyporheic zone and its potential to naturally attenuate pollutants. The focus is on a research site currently being developed on an urbanised stretch of the River Tame in north Birmingham.

Bank-side extraction of groundwater will be carried out at the study site. This will perturb the groundwater-surface water exchanges in a controlled manner to assess the natural attenuation capacity of the hyporheic zone. The transport of injected tracers, solutes and contaminants will be monitored during these tests which are expected to take place over a prolonged period of time.

1.2 Project Aims and Objectives

This project aims to investigate the hydraulic properties of the geological units within the study area. This will then be used to build a basic numerical model of the aquifer/river system with which the effects of hydraulic testing on the hyporheic zone can be assessed. In particular the

model will look at the changes in hydraulic gradient below the river bed due to groundwater abstractions, as this will determine the pump rate used in future experiments at the study site.

To meet these aims the following objectives must be met:

- A desk study should be conducted to gain an understanding of the regional and site geological, hydrogeological and hydrological conditions.
- A detailed characterisation of the hydraulic properties of the site should be conducted in the field to collect data for a numerical groundwater flow model.
- The model should test the effect pump tests will have on the observed hydraulic gradients below the river bed.

1.3 Approach

The objectives for this project were met by carrying out a combination of field studies, desk based study and computer modelling. Previous work on the site by Ellis (2002), Lydon (2006) and Conran (2006) provided a good understanding of the methodologies used for data collection in the field.

A literature review (Chapter 2) provided an insight into the current understanding of groundwater-surface water interactions and the processes occurring within the hyporheic zone which affect the exchange of water. Chapter 3 examines the study area and gives a comprehensive review of the geological, hydrogeological and hydrological knowledge of the regional and local area.

The methods used to collect and analyse data are given in detail in Chapter 4, with the results detailed in Chapter 5. The subsequent modelling of the study area is described in Chapter 6.

Chapter 2. Groundwater – Surface Water Interactions

2.1 Introduction

The importance of understanding groundwater-surface water interactions has increased with the introduction of the Water Framework Directive (WFD) in December 2000. The WFD has the primary goal of ensuring that water bodies achieve the environmental objectives set out in Article 4 of the directive, these include good chemical and quantitative status of groundwater bodies and good chemical and ecological status for surface water bodies (Environment Agency, 2002). Member states are required to ensure that water bodies achieve these objectives by 2015, therefore there is a great necessity to improve our understanding of how groundwater and surface water bodies interact, particularly in terms of the potential for attenuation of pollutants (Smith, 2005). Groundwater and surface water are not isolated components of the hydrologic system, and therefore development or contamination of one will commonly affect the other (Sophocleous, 2002). By combining the disciplines of hydrogeology, hydrology and ecology a more comprehensive understanding of groundwater-surface water interactions is being achieved, which will lead to more effective management of water resources.

This chapter will review the current understanding of groundwater-surface water interactions, in particular looking at the hyporheic zone and its potential for natural attenuation of pollutants. Methods for quantifying groundwater-surface water flow will also be discussed.

2.2 Understanding Groundwater-Surface Water Interactions

To understand the complex nature of groundwater-surface water interactions a number of physical factors must be considered, these include climate, topography, geology and the

position of the surface water body in relation to the groundwater flow system (Winter, 2000; Sophocleous, 2002). Woessner (2000) has stated that the controlling factors for groundwater flow to a river channel are:

- the distribution and magnitude of the hydraulic conductivities within the river channel, the associated fluvial plain sediments and the underlying bedrock
- the relation of stream stage to the adjacent groundwater head
- the geometry and position of the river channel within the fluvial plain

2.2.1 Conceptual Models

There are four basic conceptual models which describe the flow between groundwater and a stream channel. These models, shown in Fig 2.2.1, are based on the relationship between the stream stage and the groundwater head.

A gaining or effluent stream has a groundwater head at the channel interface which is greater than the stream stage (Fig 2.2.1a). This results in flow from the aquifer into the stream. These streams are often sensitive to nearby groundwater abstractions, which alter the groundwater heads and affect the contribution of groundwater discharge to the streams base flow (Griffiths et al, 2006; Winter et al, 1998).

A losing or influent stream has a groundwater head at the channel interface which is lower than the stream stage (Fig 2.2.1b). This results in flow from the stream to the aquifer. These streams c can be at risk from climate change and may dry out during prolonged dry periods (Wheater et al, 2007).

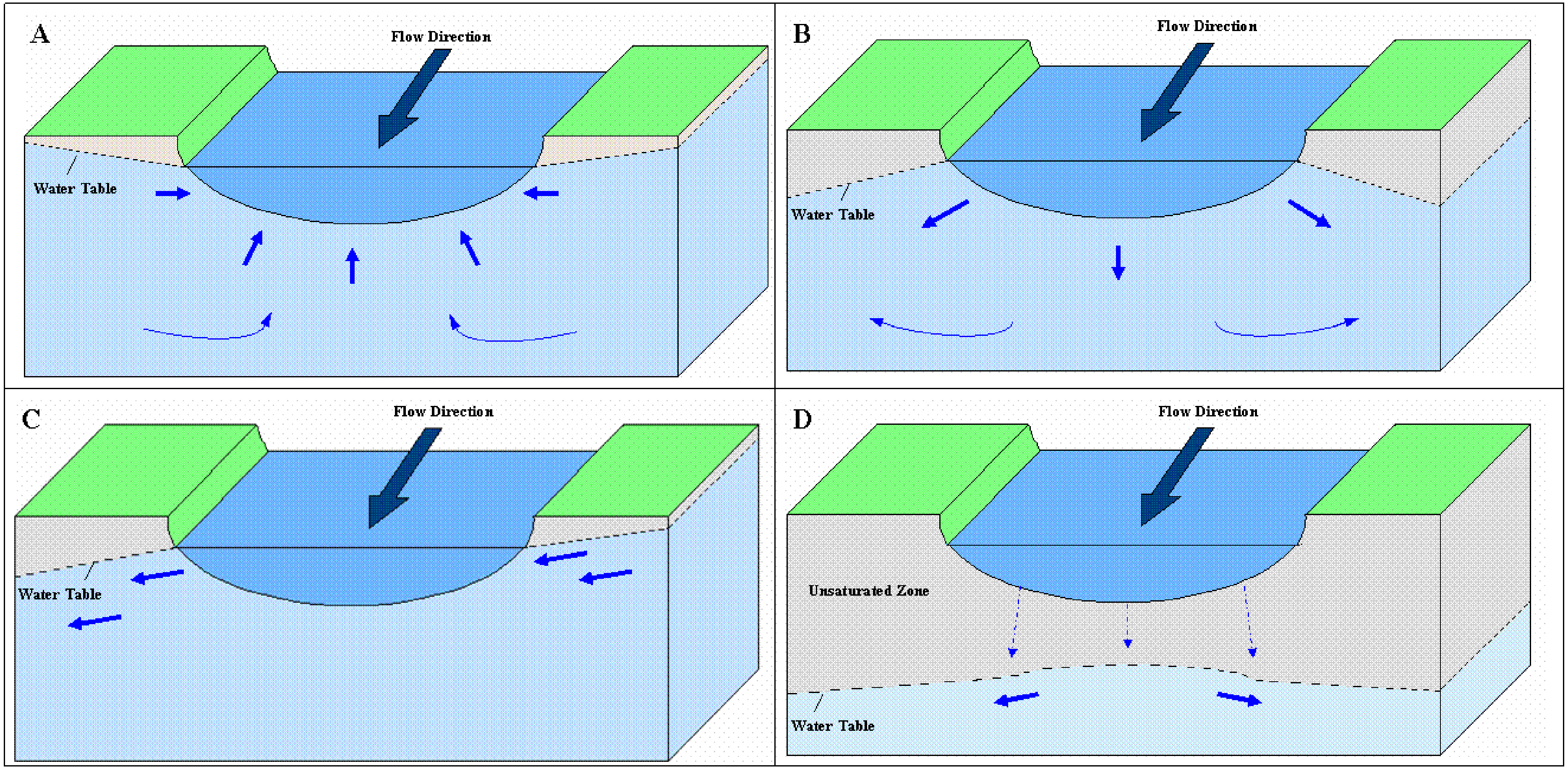

Figure 2.2.1 Four basic conceptual models indicating direction of flow between a stream channel and the underlying aquifer; A, Gaining or effluent stream; B, Losing or influent stream; C, Groundwater through flow; D, Disconnected stream.

Groundwater through flow occurs when the groundwater head is higher than the stream stage on one bank but lower on the other (Fig 2.2.1c). These conditions commonly exist when the stream channel cuts perpendicular to the fluvial-plain groundwater flow field (Sophocleous, 2002).

Under some conditions there may be no hydraulic connectivity of the groundwater with the stream (Fig 2.2.1d). These disconnected streams typically occur in very arid conditions, where low water tables exist. Water may still be exchanged between the two systems by seepage through the unsaturated zone.

These conceptual models define the direction of water flow between the aquifer and stream. The magnitude of the flow is largely controlled by the hydraulic conductivity of the stream bed sediments and the bedrock. Calver (2001) conducted an analysis of the variations in riverbed sediments and found that horizontal and vertical hydraulic conductivities can vary by several orders of magnitude, 1.0×10^{-7} to 1.0×10^{-3} m/s. The heterogeneous nature of a stream bed will result in varying magnitudes of flow between groundwater and surface water.

2.3 The Hyporheic Zone

2.3.1 Definition

Definitions of the hyporheic zone vary according to scientific discipline (ecology, hydrology and hydrogeology). The term was first used by ecologists who identified the zone as a dynamic ecotone, the crossing point between two ecosystems, in which surface water and groundwater mixed, creating unexpectedly high concentrations of dissolved oxygen and a unique habitat for invertebrate fauna known as ‘hyporheos’(Biksey & Gross, 2001; Hancock, 2002; Harvey &

Wagner, 2000). Some common themes, identified by Smith (2005), in the definitions of the hyporheic zone are:

- it is the zone below and adjacent to a streambed in which water from the stream channel exchanges with interstitial water in the bed sediments
- it is the zone around a stream in which characteristic fauna of the hyporheic zone are distributed and live
- it is the zone in which groundwater and surface water mix

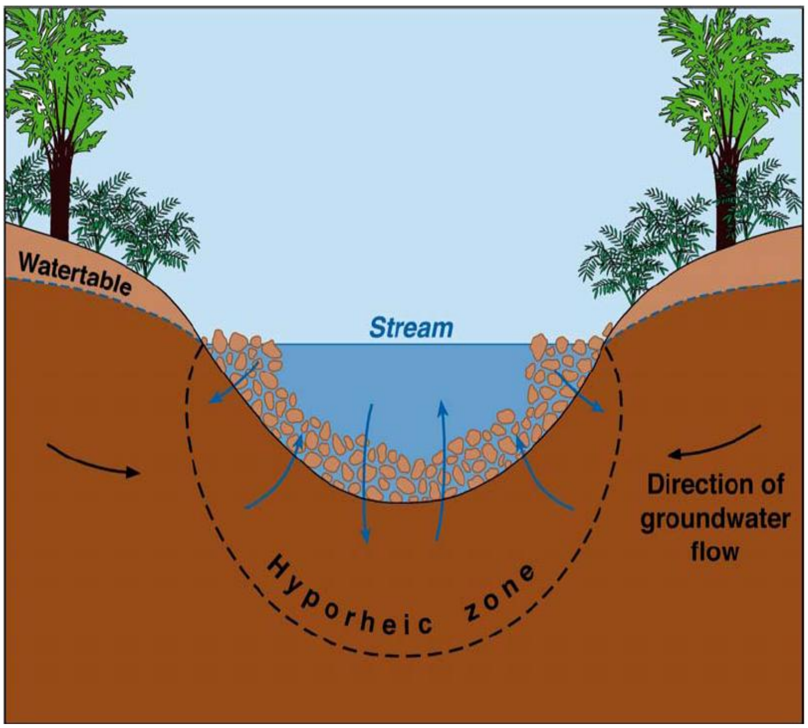

Figure 2.3.1 Schematic diagram of the hyporheic zone indicating the mixing of surface waters and groundwater within the region (*From Smith, 2005*)

Hydrogeologists commonly define the hyporheic zone as the water-saturated, transitional zone between surface water and groundwater (Woessner, 2000), as shown in Fig 2.3.1. The zone is thought of as carbon and microbial-community rich in comparison to the underlying aquifer

sediments, and it is therefore considered to have a greater potential for natural attenuation of pollutants than the aquifer (Smith, 2005).

The hyporheic zone is a dynamic system; its extent will vary according to daily or seasonal fluctuations in river stage and groundwater flow, proportions of groundwater and surface water, levels of oxygen, organic matter, temperature and pH. The hyporheic zone can range in size from centimetres to hundreds of metres, depending on the nature of the sediments in the stream bed and banks, the hydraulic gradients within the adjacent groundwater system and the variability in slope of the stream bed (Biksey & Gross, 2001). Increasing the permeability of stream bed sediments and stream channel width will, in general, increase the extent of the hyporheic zone (Winter, 2000).

2.3.2 Mixing in the hyporheic zone

Flow paths within a streams hyporheic zone are primarily a function of the surface morphology of the bed and its hydrologic features (Yamada et al, 2005). Within an overall gaining or losing stream reach localised flow paths within the bed are present as a result of the stream bed heterogeneities, which cause variations in the aquifer and stream fluid pressures, resulting in a mixing zone (Ellis et al, 2007; Sophocleous, 2002). The flow processes occurring in the hyporheic zone can be studied at two scales; the reach scale and the sediment scale (Fig 2.3.2).

At the reach scale geomorphological features, including the channel shape, the roughness and permeability of the stream bed, discontinuities in slope and depth of riffle-pool sequences, result in upwelling and downwelling regions of hydrological exchange (Fig 2.3.2a). Downwelling

zones in streams tend to occur at the head of riffles or where there is an increase in streambed elevation. Downwelling surface waters entering the stream bed displace the interstitial waters, which may then travel for a considerable distance within the hyporheic zone before upwelling back into the stream (Boulton et al, 1998). Upwelling waters, and groundwater discharge, tend to occur at the upstream edge and base of pools. Anoxic conditions and anaerobic processes, caused by the upward movement of groundwater, typically characterise upwelling zones. In contrast the downward movement of surface waters in downwelling zones results in these regions having high oxygen levels and the occurrence of aerobic processes (Biksey & Gross, 2001).

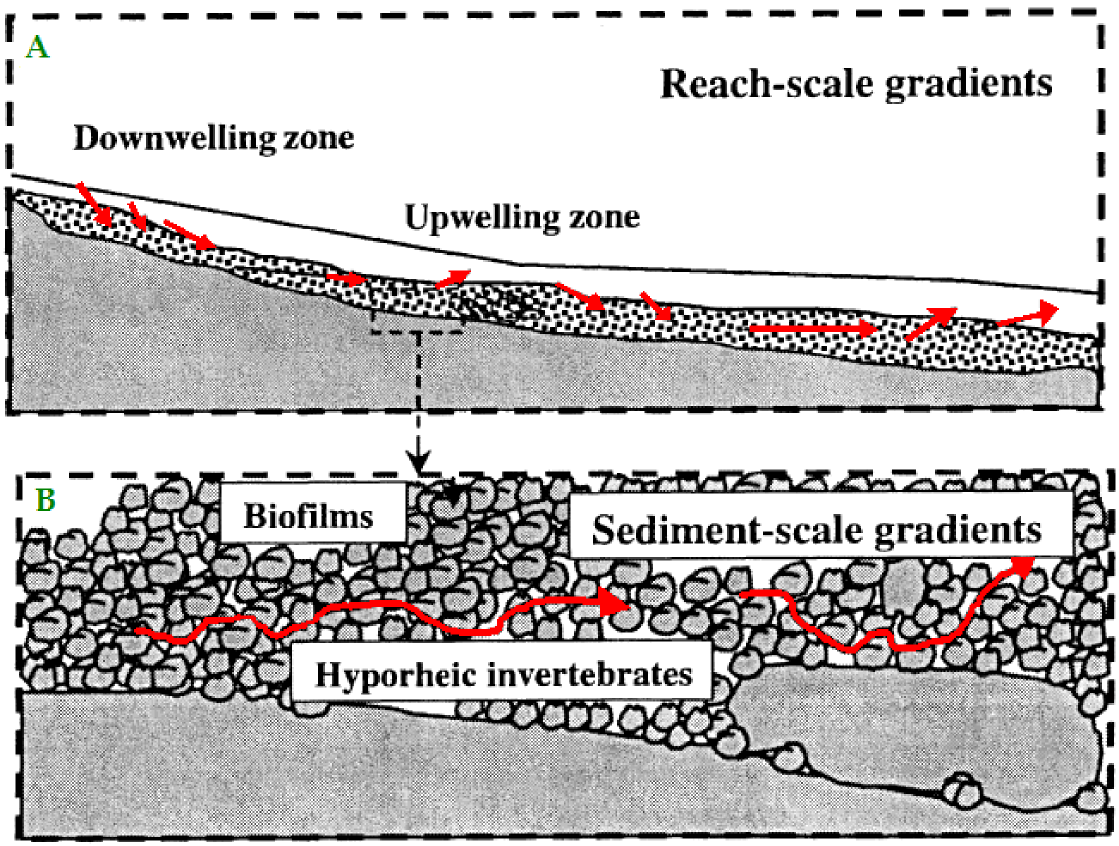

Figure 2.3.2 Hyporheic zone at two spatial scales (*arrows* indicate the water flow paths); A, the reach scale, upwelling and downwelling zones alternate; B, the sediment scale, microbial and chemical processes occur, creating microscale gradients. (*Adapted from Boulton et al, 1998)*

At the sediment scale the hydraulic gradient and stream bed porosity control the interstitial flow paths (Fig 2.3.2). The flows are irregular and turbulent, which results in the creation of areas of slow, rapid and no flow. An apparently well oxygenated hyporheic zone at the reach or catchment scale is found to contain anoxic and hypoxic pockets at the sediment scale; these are associated with irregular sediment surfaces, small pore spaces and deposits of organic matter (Boulton et al, 1998).

The heterogeneous nature of the hydraulic gradients and hydraulic conductivities observed within the hyporheic zone are not solely caused by the parent bedrock and stream bed sediment distributions. There are a number of physical, chemical and biological processes which may occur within the hyporheic zone reducing or increasing permeabilities locally. Biofilms predominantly form on small sediment particles because of their large surface areas. Biofilms have a low porosity and therefore result in localised areas of low permeability (Smith, 2005; Boulton et al 1998). Sediment reworking by macroinvertebrates will increase permeabilities locally. Chemical processes within the hyporheic zone may result in mineral dissolution or precipitation, which will increase and decrease permeabilities respectively (Smith, 2005).

The most significant physical process which has an impact on stream bed permeability is colmation. Colmation is caused by a gradual clogging of fine grained sediments, usually deposited as a result of the filtering of downwelling waters by the porous sediments of the stream bed (Sophocleous, 2002; Smith, 2005). In general high levels of colmation lead to decreased oxygen and nitrate concentrations and an increase in ammonification (Hancock, 2002). Colmation results in a low permeability sediment layer, known as colmatage.

2.3.3 Groundwater-surface water contamination

Contaminants present in a groundwater will be transported to surface water through the hyporheic zone, and vice versa. Transport is predominantly by advective processes from surface waters into the hyporheic zone. From groundwater to surface water, contaminants move in response to concentration gradients or hydraulic head of the adjacent hillslope (Triska et al, 1989). The organisms present in the hyporheic zone are useful for assessing the impact of groundwater contamination, as they will typically show the effects of pollution on discharging groundwaters before surface water organisms (Ellis, 2002). It is necessary to understand how any processes occurring in the hyporheic zone will affect the contaminant fate to ensure the successful management of groundwater and surface water quality.

As discussed above, subsurface conditions can change from anaerobic to aerobic over short distances within the hyporheic zone; it is thought that these processes could be important for natural attenuation and removal of contaminants (Biksey & Gross, 2001). The highly dynamic nature of the hyporheic zone is believed to enable more rapid natural attenuation of pollutants than within the aquifer system, by processes including sorption, microbial biodegradation and redox based reactions (Smith, 2005).

Biofilms provide a site of increased microbial and biological activity within the hyporheic zone and are critical for processing and cycling organic carbon and nutrients. It has been shown that these processes provide attenuation of contaminants in surface water as a result of the continuous exchange between the stream channel and the hyporheic zone (Smith, 2005), reducing the contamination of groundwater as a result of polluted surface waters. Colmation within the stream

bed is also thought to have a significant impact on the transport of contaminants as the low permeability zones act as a barrier, preventing the transfer of contamination (Sophocleous, 2002).

The application of monitored natural attenuation to the hyporheic zone requires a highly detailed study to assess the impact of flow heterogeneities on residence time and potential for attenuation to occur (Smith, 2005).

2.4 Methods to Characterise Groundwater-Surface Water Exchange

There are a number of methods which have commonly been used to characterise the exchange of groundwater and surface water. Woessner (2000) states that characterisation is typically accomplished by:

- Measuring water levels in wells, piezometers and piezometer nests installed within the fluvial plain, on the banks of the channel and within the streambed.
- Performing stream gauging at a number of stream cross sections over a short time period and comparing the discharge measurements.
- Comparing groundwater and stream geochemistry.
- Conducting one-dimensional stream channel tracer studies.

Calculating the flux of water within the hyporheic zone can be done using many methods; however the use of a seepage meter is considered to be the only direct measurement of water flux. The seepage meter consists of a chamber which sits on the stream bed with a tube inserted into the top or the side of the chamber. A small bag is attached to the tube and is used to measure

changes in water volume to determine the flux over time (Biksey & Gross, 2001). The seepage meter can give a good measurement of water flux, however in some situations it can be difficult to install, for example in coarse gravel beds with a strong current (Yamada et al, 2005).

There are numerous other methods which can be used to make indirect measurements of the water flux, they measure hydraulic head, hydraulic conductivity, temperature or electrical conductance from which a water flux can be calculated. Mini-piezometers installed in the stream bed can be used to measure the difference in hydraulic head between the groundwater and surface water to determine the hydraulic gradient. This is used to calculate the hydraulic conductivity and cross-sectional area of the flux, which in turn is used in the calculation of the water flux. Mini-piezometer nests can be installed at varying depths, within the stream bed sediments, to determine the water flux across a larger profile of the hyporheic zone (Biksey & Gross, 2001). Sometimes physical measurements made using piezometers can be difficult to interpret due to the heterogeneous nature of the hyporheic zone resulting in what appears to be conflicting negative, positive and neutral head differences in adjacent sections of the stream channel (Woessner, 2000). Piezometers can also be used to obtain water samples for geochemical analysis and to assess the permeability of the stream bed sediments, by conducting slug tests. Slug tests may not work very well in highly permeable gravel beds however as the water table will return to the original level very quickly after the removal of the water (Yamada et al, 2005).

The localised fine scale flow systems commonly found within the hyporheic zone are difficult to assess using piezometers alone. Tracer tests are also used to define the significance of water

exchange and the extent of the hyporheic zone. Chemical tracers are commonly used in hyporheic zone studies, however interpreting the observations may be problematic if rapid changes in tracer concentration occur (Yamada et al, 2005). Chemical tracers have been used in many field experiments to assess the extent of the hyporheic zone, which Hancock (2002) states is the area where 98-10% of water originates from the surface stream. Heat is another common tracer used. Groundwater temperature is constant in comparison to surface water temperatures which vary seasonally, and from day to night. The difference in temperature between the two waters can be used to develop a temperature profile, which indicates the direction of groundwater flow and the extent of the hyporheic zone (Smith, 2005). The extent of the hyporheic zone can also be assessed using gradients in pH, oxygen concentrations, and the composition of the hyporheos.

2.5 Estimating Groundwater-Surface Water Exchange

The exchange flow of water in hydraulically-connected aquifer-stream systems is a function of hydraulic conductivity of the sediments and the hydraulic gradient. By considering flow between the stream and aquifer to be controlled by a similar mechanism as leakage through a semi-impervious layer in one-dimension, a simple approach to estimating the flow can be used which is based on Darcy's Law (Sophocleous, 2002).

$$q = k\,\Delta h$$

Where:

- $\Delta h = h_a - h_r$ (h_a is aquifer head, h_r is river head)
- k is a constant representing the stream bed leakage coefficient (hydraulic conductivity of the semi-impervious stream bed stratum divided by its thickness)

- q is flow between the river and aquifer (positive for baseflow, or gaining streams, negative for recharge, or losing streams)

The use of the above equation assumes a linear relationship between q and Δh. Subsurface flow is assumed to be laminar, i.e. has a Reynolds number less than or equal to 1, because turbulent flow does not have a linear relationship between q and Δh (Shaw, 1990; Yamada et al, 2005).

The assumption of a linear relationship between hydraulic gradient and flow is, in many cases, too simplistic. The complex heterogeneities found within stream and aquifer systems can induce non-Darcian flow within the hyporheic zone sediments (Smith, 2005). A number of studies have shown that the total baseflow during streamflow recession is largely independent of the leakage coefficient, k, and that during periods of high recharge the leakage calculated is much greater than is observed in practice (Sophocleous, 2002). Rushton (2003) suggests that a non-linear relationship may be more appropriate for flow calculations. It was found that the results from non-linear relationships were very similar to those calculated using a linear relationship.

2.6 Conclusions

The hyporheic zone is a complex and dynamic system, which has ecological, hydrological and hydrogeological significance. The hyporheic zone can be viewed as a subset of small scale interactions between the stream channel and groundwater, which occur within the larger scale groundwater-surface water exchange system. Distinguishing the interactions between the stream and the hyporheic zone, and the stream and groundwater flow is important to understand the nature of water exchange. Hyporheic flowpaths leave and return to the stream many times within

a single study reach, unlike groundwater flowpaths which enter the channel reach only once (Harvey & Wagner, 2000).

Physical, chemical and biological processes occurring within the hyporheic zone may have significant effects on the exchange of water and contaminants. To fully understand the exchanges taking place a highly detailed investigation of the stream reach is required, to assess the impact heterogeneities within the system may have. This is particularly essential when investigating the potential of the system for monitored natural attenuation.

Chapter 3. Site Setting

3.1 Study Setting

The River Tame drains a catchment area of 408 km^2, of which 250 km^2 is highly urbanised as shown in Fig 3.1.1. The river flows eastwards through north Birmingham and forms part of the larger, River Trent drainage system, which eventually discharges into the North Sea. The Tame is a typical example of an urban river and has suffered problems of pollution from surrounding industries in Birmingham (SMURF 2005). The unconfined Triassic sandstone aquifer underlies the study site and part of the city of Birmingham, dominating an approximately 111 km^2 area.

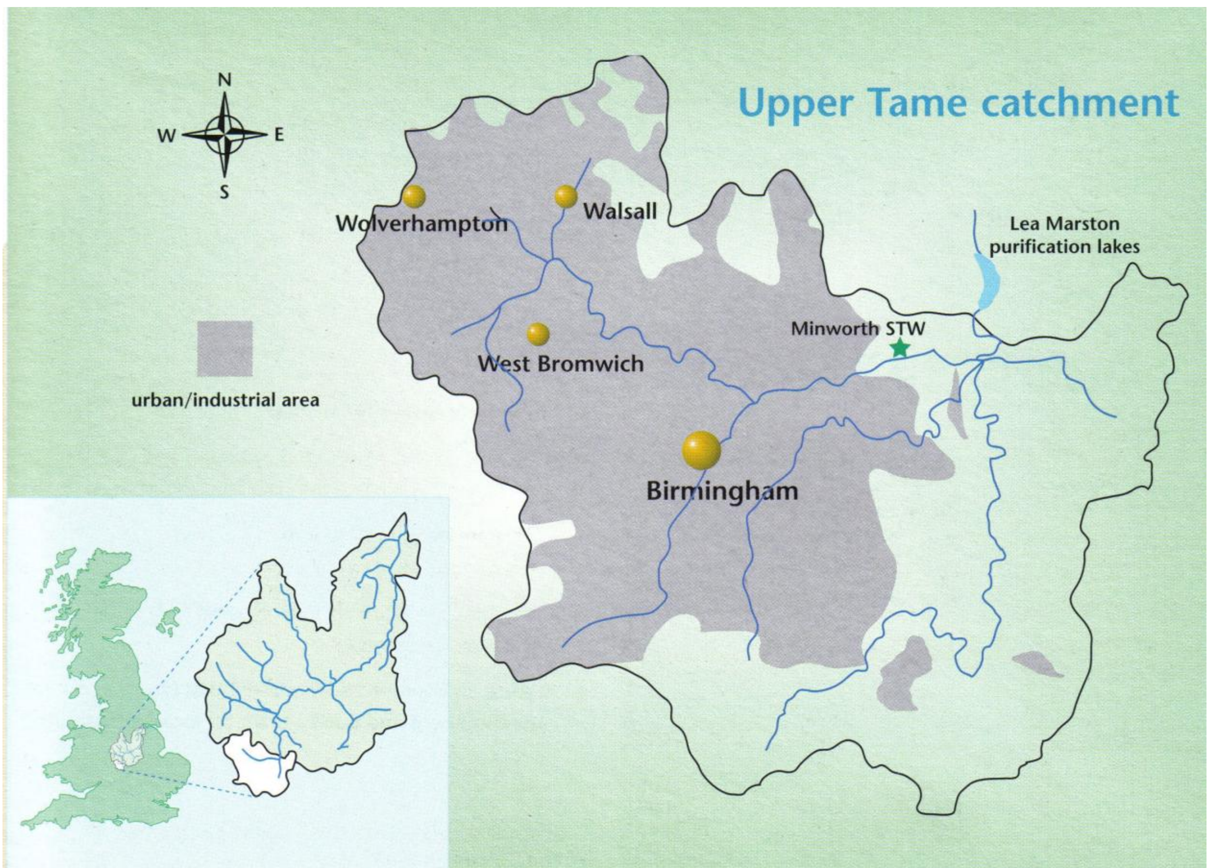

Figure 3.1.1 Map of the River Tame catchment indicating the extent of urban land use

(*From SMURF 2005*)

The Tame has been the subject of a number of case studies in recent years primarily focusing on the known contaminant plumes present in the underlying aquifer which discharge into the river by base flow and seepage. Since ~2000 the SWITCH research group has been studying the hyporheic zone of the Tame (Rivett et al, 2006).

3.2 Study Site

The study site is situated approximately 4 miles north of Birmingham city centre between the Holford Industrial Estate and the M6 motorway. The site can be accessed from Tameside drive (Holford Industrial Estate). The site location is shown in Fig 3.2.1.

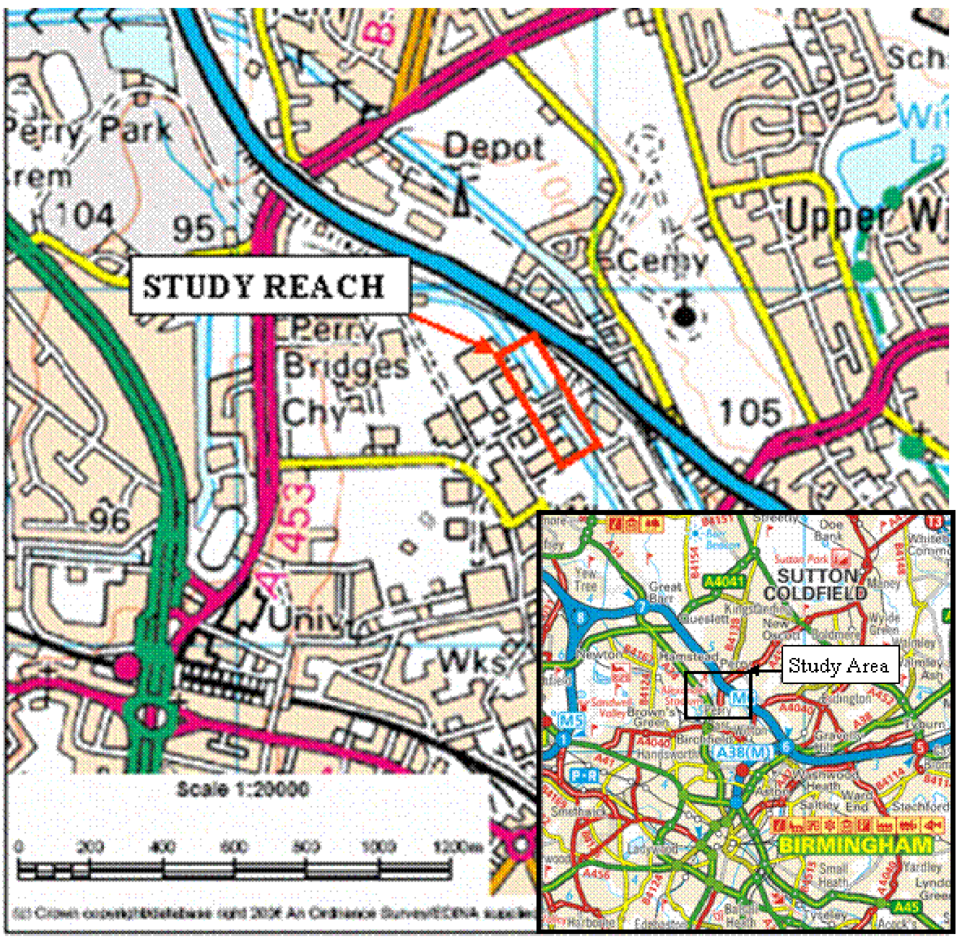

Figure 3.2.1 Location of study site

The site is being developed to conduct a series of hyporheic zone control experiments over the next few years. Selection of a suitable site was constrained by a number of factors including; provision of a power supply, suitable river conditions for monitoring, access for drilling rigs, supportive land owners and a reasonably uniform channel. The site selected was the best available option (Rivett et al, 2006). The current study has been carried out on a 200m stretch of the River Tame, which is shown in Fig 3.2.2. During July/August 2007 a 16.5m borehole was drilled on the north bank of the Tame (see Fig 3.2.2).

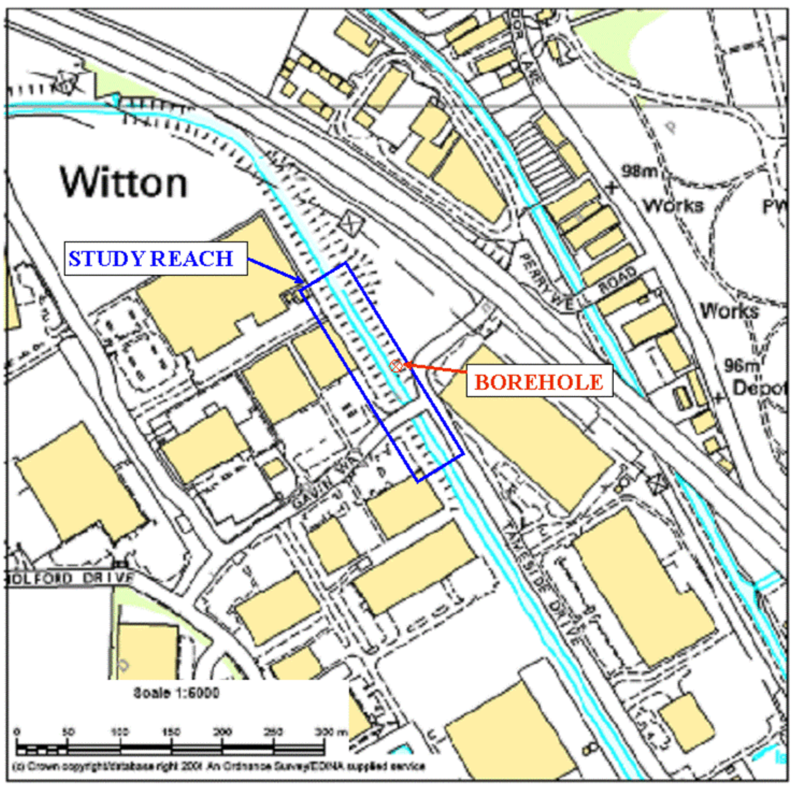

Figure 3.2.2 Location of Borehole within the study reach

3.3 Regional Geology

The geology of the Birmingham area consists of a series of sedimentary sequences deposited in the centre of the Knowle Basin. The Knowle Basin is an east-west extensional rift-basin formed

during the Permian to Triassic (Powell et al, 2000). The boundaries of the Knowle Basin are defined, to the west, by the north-east trending Birmingham Fault and, to the east, by the north trending Western Boundary Fault of the Warwickshire Coalfields. The map in Fig. 3.3.1 shows the regional geology and locations of the major faults.

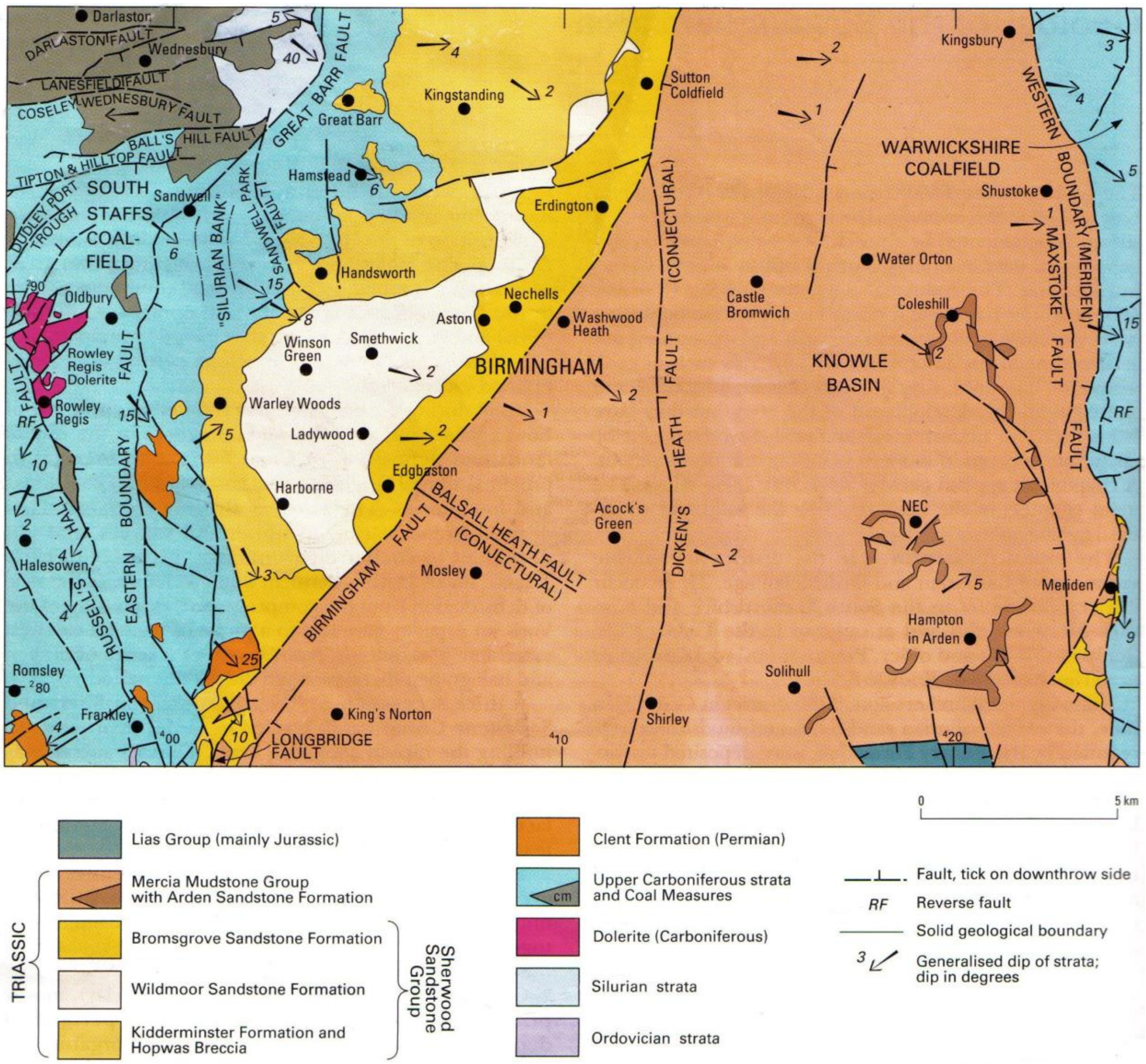

Figure 3.3.1 Sketch map of the solid geology of the Birmingham area

(*From Powell et al, 2000*)

During the Triassic a thick sequence of sedimentary rocks were deposited in the Knowle Basin. This sequence is known as the Sherwood Sandstone group. They are predominantly fluvial,

continental and lacustrine sediments deposited in an arid to semi-arid climate. The Sherwood Sandstone Group is overlain by the Mercia Mudstone Group, a series of red-bed mudstones and siltstones deposited under lacustrine and fluvial conditions (Powell et al, 2000). The stratigraphy of the Triassic sequence is shown in Table 3.3.1. The bedrock in the area is generally covered by fluvial and glacial drift deposits which range in thickness from 1 to 40m.

System/Series		Stage	Lithostratigraphy: *Group*	Lithostratigraphy: *Formation*
Lower Jurassic		Hettangian	Lias Group	Blue Lias Formation
Triassic	Upper	Rhaetian	Penarth Group	Lilstock Formation
				Westbury Formation
			Mercia Mudstone Group	Blue Anchor Formation
		Norian		
		Carnian		Arden Sandstone Formation
	Middle	Ladinian		
		Anisian	Sherwood Sandstone Group	Bromsgrove Sandstone Formation
	Scythian	Induan-Olenekian		Wildmoor Sandstone Formation
				Kidderminster Formation
				Hopwas Breccia

Table 3.3.1 Stratigraphic sequence of the Triassic and Lower Jurassic in the region (*adapted from Powell et al, 2000*)

3.3.1 Sherwood Sandstone Group

The base of the Triassic succession is defined locally by the Hopwas Breccia. This is overlain by the Sherwood Sandstone Group, which comprises, in upward sequence, the Kidderminster

Formation, the Wildmoor Sandstone Formation and the Bromsgrove Sandstone Formation. The Sherwood Sandstone Group varies in thickness across the region; generally it is up to 200m thick although there is geophysical evidence suggesting a minimum thickness of about 625m in the Knowle Basin (Powell et al, 2000).

Kidderminster Formation

The Kidderminster Formation consists of red-brown pebble conglomerate, pebbly sandstone and medium- to coarse-grained, cross-bedded sandstone with intermittent, thin mudstone beds. Generally the formation is weakly cemented and friable, however harder calcite cemented beds are present locally (Powell et al, 2000). The formation crops out sporadically to the north-west of the Birmingham Fault. It is thought that the formation was deposited as part of a major, northward flowing, braided river system (Ellis, 2002). The formation has proved to be up to 119m thick in the central Birmingham area, and appears to thin out to the southeast as it approaches the Birmingham Fault (Powell et al, 2000).

Wildmoor Sandstone Formation

The boundary between the Wildmoor Sandstone Formation and the Kidderminster Formation is gradational. The Wildmoor Sandstone Formation consists of finer grained, micaceous soft sandstone, which is orange-red in colour, and thin, red-brown and grey-green mudstone layers (Powell et al, 2000). The formation is poorly cemented. Low-angle, planar cross-bedding indicates fluvial deposition. It is thought that the formation was deposited by ephemeral streams with some aeolian deposition (Ellis, 2002). There are few outcrops of the Wildmoor Sandstone

Formation in the region. Borehole data suggests that the formation is 30-86m thick to the north-west of the Birmingham Fault and 7-40m thick to the south-east (Powell et al, 2000).

Bromsgrove Sandstone Formation

The Bromsgrove Sandstone Formation lies unconformably on the Wildmoor Sandstone Formation and Kidderminster Formation. The unconformity represents a period of uplift and erosion (Powell et al, 2000).The formation is composed of a series of upward fining sequences from sandstone to mudstone. The sequence is made up of red-brown medium- to coarse-grained, sub-angular arkosic sandstone, which locally contains pebble conglomerate beds, interbedded with thin, red mudstone and siltstone layers. Deposition of the formation occurred in mature, meandering river systems. To the south-east of the Birmingham Fault the formation is 84-180m thick (Powell et al, 2000). The boundary with the overlying Mercia Mudstone Group is typically gradational.

3.3.2 Drift

About 75% of the Birmingham area is covered in Quaternary, unconsolidated superficial deposits (Powell et al, 2000). The strata of the superficial cover is varied and includes glaciofluvial and glaciolacustrine deposits, till, river terrace deposits, alluvium and head (Shepherd et al, 2006).The deposits range in thickness from 1 to 40m (Ellis et al, 2007). Their distribution across the area is varied, as shown in Fig 3.3.2. In addition to the glacial and post-glacial deposits, the Birmingham region has several areas of made ground deposited on top of the natural ground surface. Made ground has been deposited over large areas of the River Tame valley during construction of flood defences (Powell et al, 2000).

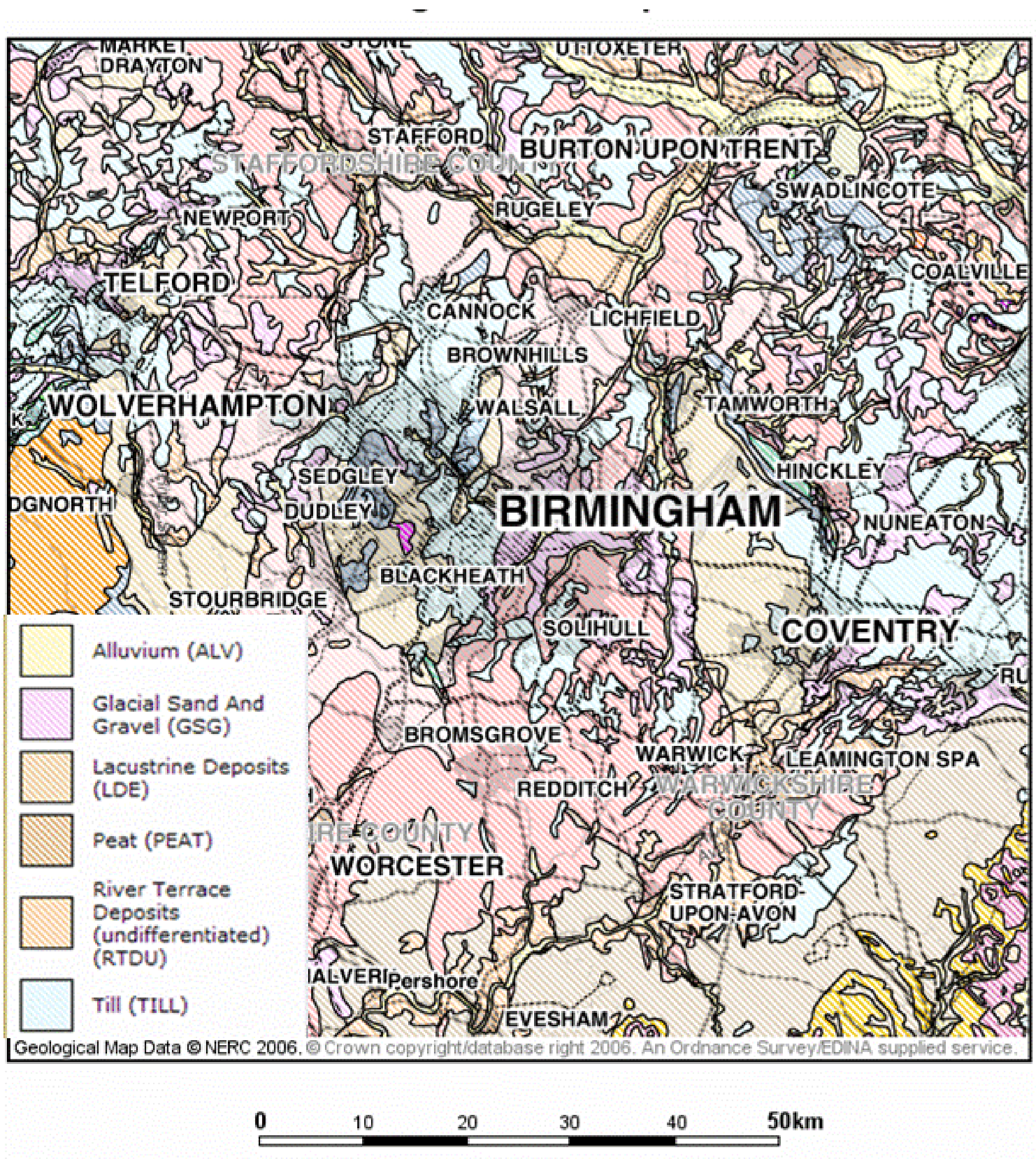

Figure 3.3.2 Regional drift map of Birmingham and surrounding area

3.4 Regional Hydrogeology

The Sherwood Sandstone Group forms the main aquifer system in the region, and is locally known as the Birmingham aquifer. The eroded surface of the Carboniferous below the Sherwood Sandstone is believed to represent the aquifer base. Regionally the aquifer is both confined and unconfined. To the northwest of the Birmingham fault the aquifer is unconfined (Shepherd et al,

2006). To the east of the region the sandstone is overlain by the, effectively, impermeable Mercia Mudstone Group creating confining conditions (Greswell, 1992). The map in Fig 3.4.1 indicates the relationship between the confined and unconfined aquifer.

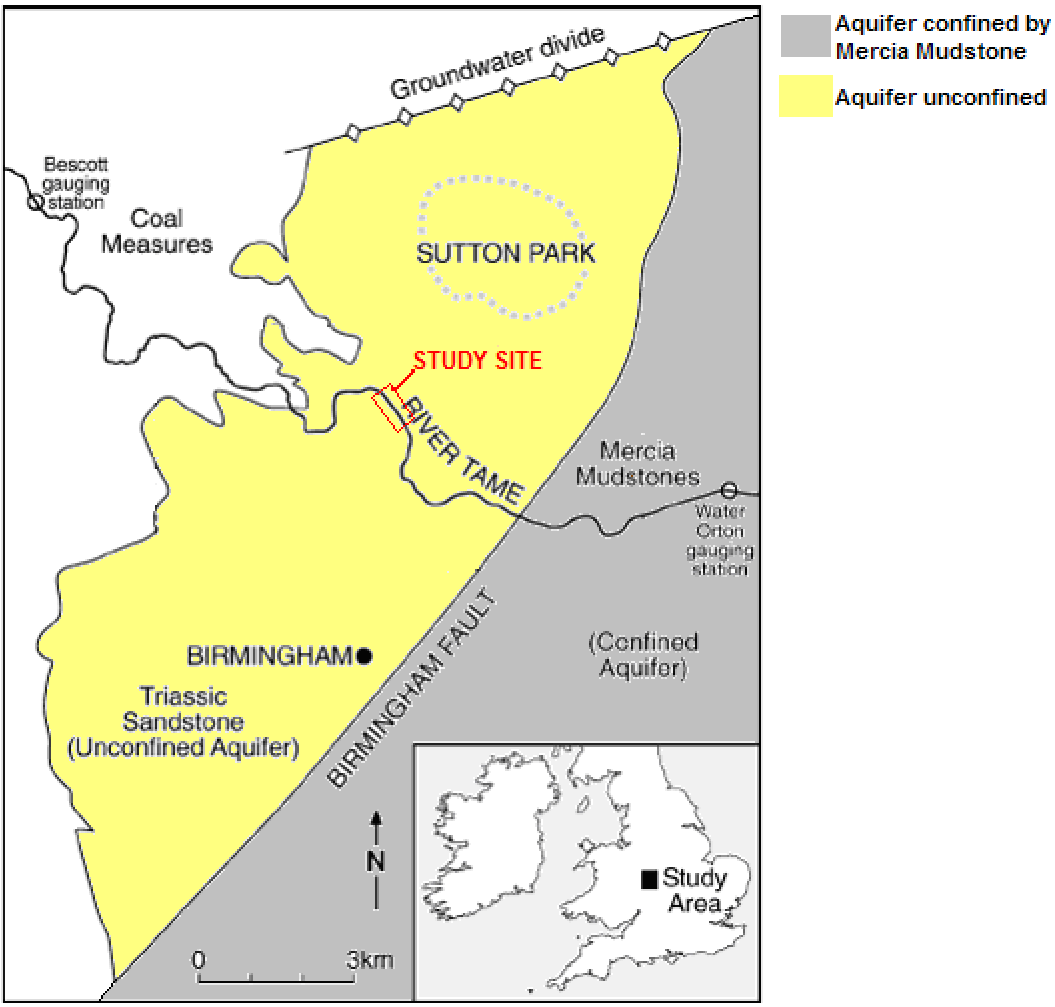

Figure 3.4.1 Map of Birmingham region indicating the location of the main aquifer and the study site (*adapted from Ellis et al 2007*)

Groundwater flow in the region is typically down gradient in an easterly direction where it discharges to water courses in low relief areas, the River Tame and its tributaries are the main focus for natural discharge in the region (Ellis, 2002). The Birmingham Fault has created a low

permeability feature, across which the piezometric head decreases significantly causing a boundary to groundwater flow (Ellis, 2002). A simple conceptual model of groundwater flow in the Birmingham aquifer is shown in Fig 3.4.2.

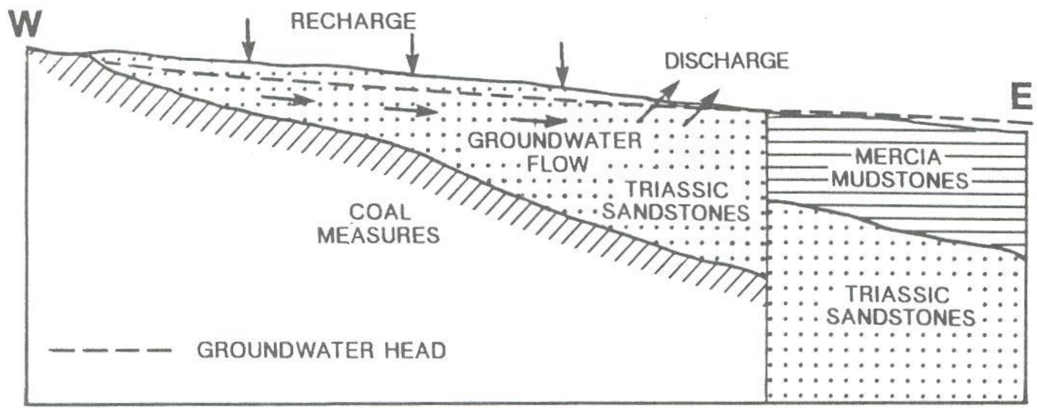

Figure 3.4.2 Conceptual model of groundwater flow in the Birmingham aquifer (*From Greswell, 1992*)

The Birmingham aquifer is typically of low permeability/high storativity. Pumping tests conducted on the sandstone give a hydraulic conductivity of approximately 2m/d, porosity of ~0.28 and Specific yield of ~0.12 (Shepherd et al, 2006). Hydraulic anisotropy is present at a regional scale with the horizontal hydraulic conductivity being greater than the vertical hydraulic conductivity; this is due to low permeability mudstone layers which occur within the formation (Ellis et al, 2007). At a regional scale groundwater flow in the aquifer can be considered as single porosity, however fractures and joints in the sandstone result in dual porosity and permeability being important at the local scale (Ellis et al. 2007). The mean regional hydraulic gradient has recently been estimated as ~0.004, indicating a groundwater pore velocity of ~10m/yr (Shepherd et al, 2006).

Recharge to the aquifer system is greatly influenced by the distribution and thickness of superficial drift deposits that cover the region (Powell et al, 2000). The sands and gravels present in the region potentially have high hydraulic conductivities of 10 – 30 m/d, and high storage (Ellis et al, 2007). The average annual rainfall is non-uniform across the region, with approximately 800mm in the south-west and 650mm in the north-east (Powell et al, 2000). Run-off in the area is high due to widespread urban land use resulting in a low recharge from precipitation, estimated at 130mm/yr (Shepherd et al, 2006). However significant recharge to the aquifer is caused by leakage from sewers and water mains, and seepage from the extensive canal network, estimated as 600mm/yr (Ellis, 2002).

Historically groundwater abstractions in the region have occurred for over 150 years, predominantly for industrial purposes (Shepherd et al, 2006). Abstractions peaked in the 1950s with over 60Ml/d, this has since steadily declined with the reduction in industrial demand to the present day abstractions of approximately 13Ml/d (Ellis et al, 2007). The decline in groundwater abstractions over the past 50 years has resulted in a significant rise in water levels (Powell et al, 2000).

3.5 Local Geology

The study site is located approximately 1km northwest of the Birmingham fault. The geological map shown in Fig 3.5.1 indicates that the site rests on the Kidderminster Formation, which is part of the Sherwood Sandstone group. Various borehole logs from the surrounding area indicate that the sandstone is 71 – 80m thick, with a median thickness of ~76m (Lydon, 2006). The base of the sandstone rests disconformably on the Hopwas Breccia.

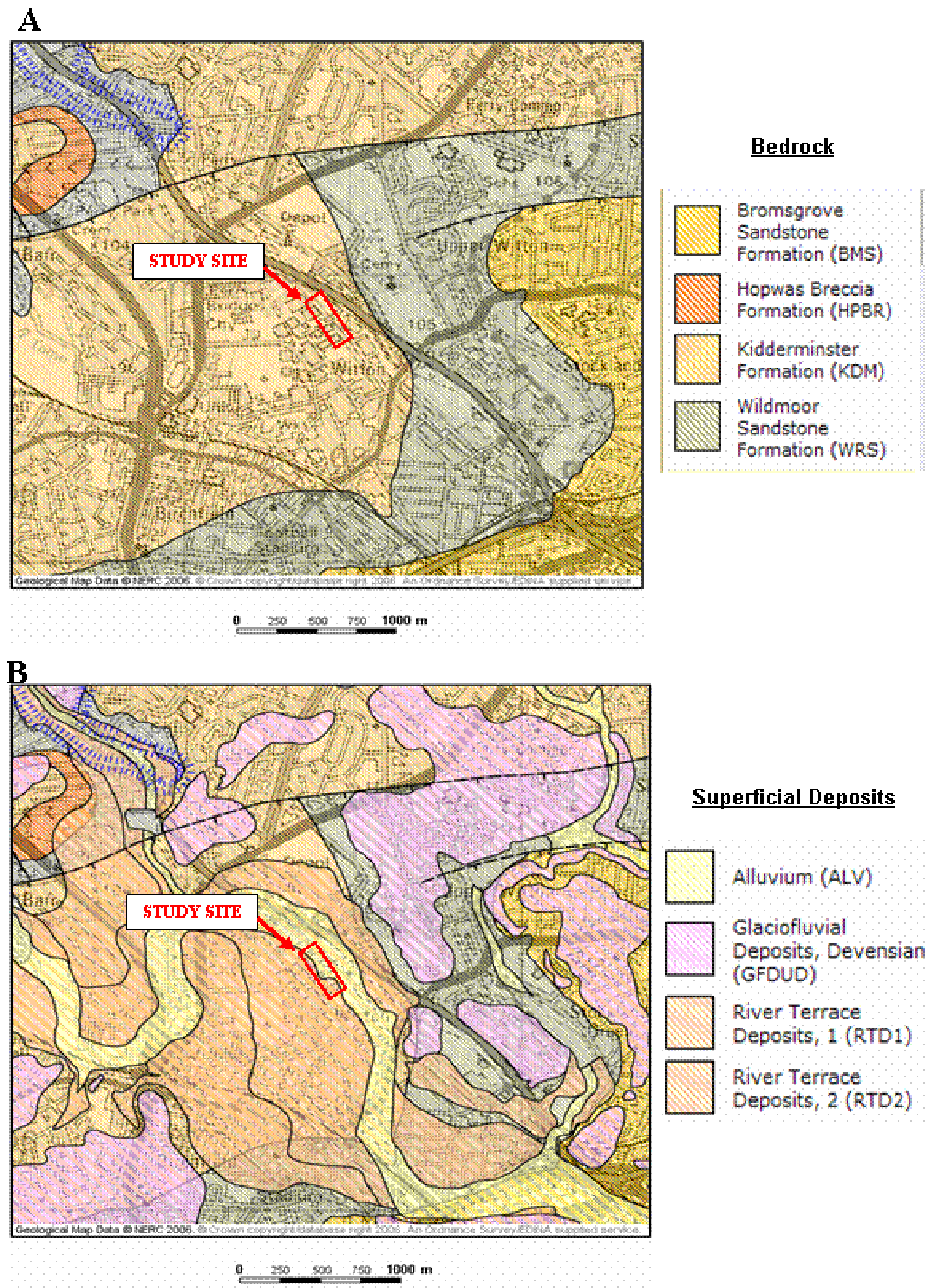

Figure 3.5.1 A: Map of the bedrock at the study site; B: map of drift cover at the study site

Superficial deposits cover the sandstone across the whole site area. The valley bottom of the Tame is composed of alluvial deposits of sands, gravels and clays. There are also two stages of

river terrace deposition, which predate the recent alluvial deposits. Fig 3.5.1 shows the complexity of the drift cover in the area. Glaciofluvial deposits also cover the area. The borehole drilled on site indicated that superficial cover on the north bank of the River Tame was approximately 8.5m thick, with the alluvium composed of sand and gravel. A 2m layer of made ground covers the drift.

3.6 Local Hydrogeology

The study site is situated on an unconfined part of the Birmingham aquifer. Groundwater flow in the area is typically from west to east. A numerical model constructed by Ellis (2002) indicates that in close proximity to the River Tame groundwater flows towards the river, as shown in Fig 3.6.1. Head measurements, from piezometers installed in the riverbed during field work, and from previous studies, indicate that the River Tame is gaining along the entire study reach, with a median hydraulic gradient of 0.0875.

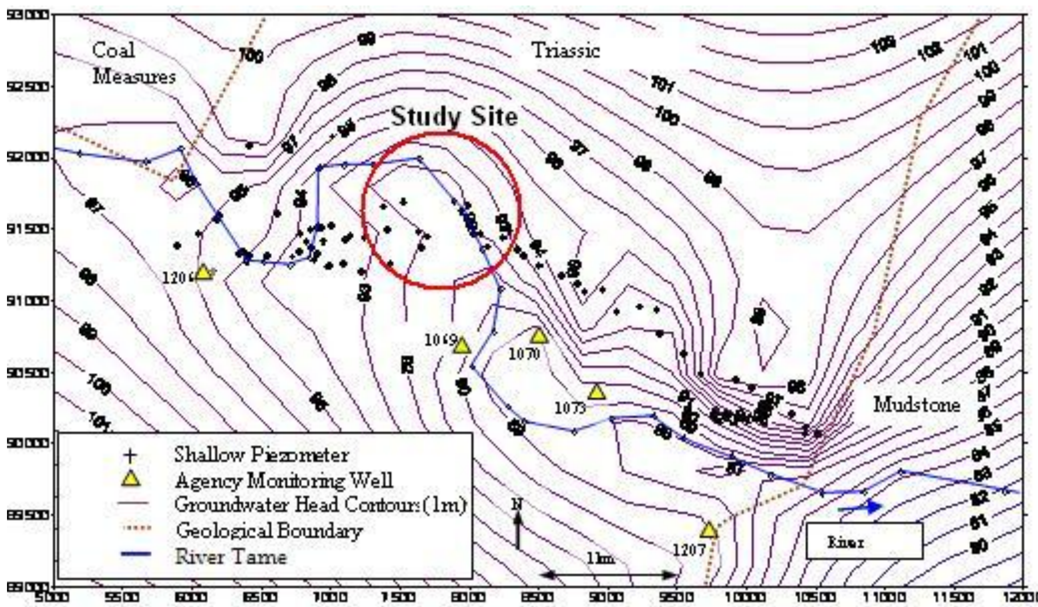

Figure 3.6.1 Distribution of groundwater head in the study area (*from Ellis 2002)*

3.7 Hydrology

The catchment area drained by the Tame is 408 km^2 (Ellis et al, 2007). The river is described as extensively modified; it bares little resemblance to its natural form of a meandering braided river system. Flood defence work carried out along the river has resulted in the river being straightened along many sections and river banks being strengthened, this includes the study site as shown in Fig 3.7.1, the base of the channel has remained natural and unlined.

Figure 3.7.1 Study site looking upstream from the Gavin Way Bridge, the river has been straightened as part of flood defence work carried out in the area.

In the study area the river is typically 8-12m wide, 0.2-2m deep and has an average dry weather flow velocity of 0.1-0.8 m/s (Ellis 2002). Over the study section the river stage maintains a comparatively constant shallow gradient of 0.0013. The riverbed is composed of a range of materials from sub angular cobbles to gravel, sand and silt. During the summer months weed

growth in the river is substantial and is thought to affect the hydrological regime by reducing flow velocities (Ellis, 2002).

Chapter 4. Data Collection

4.1 Introduction

The project area used for this study consisted of a 200m stretch of the River Tame to the north of the Holford industrial estate. This chapter looks at the methods used for collecting data in the field to use for subsequent analysis and modelling. A network of riverbed piezometers were installed for the study and these, along with a small scale pump test conducted on the borehole on site, were used as the main sources for data. The installation of piezometers and data collection was carried out during August 2007, following delays in accessing the river (due to high water levels following a prolonged period of rain) and gaining permissions to access the site.

The aims of the fieldwork applied to the study area were:

- To identify the hydraulic gradients below the river bed.
- Obtain hydraulic conductivity (K) estimates of river bed sediments from falling head piezometer tests.
- Obtain hydraulic conductivity (K) estimates of the sandstone aquifer from a small scale pump test.

4.2 Borehole Installation

During late July/early August 2007 a borehole was installed on site, to be used for future experimental work on the study area (Fig 4.2.1). The 6" borehole was drilled using rotary drilling with an air flush to clear the hole of cuttings. During the drilling process cuttings samples were collected approximately every 1m to accurately log the borehole section (Fig 4.2.1c). The

borehole was drilled to a depth of 16.5m. The entire length of the borehole was cased, with 6m of slotted casing at the base of the borehole to permit water to enter the borehole from the sandstone aquifer.

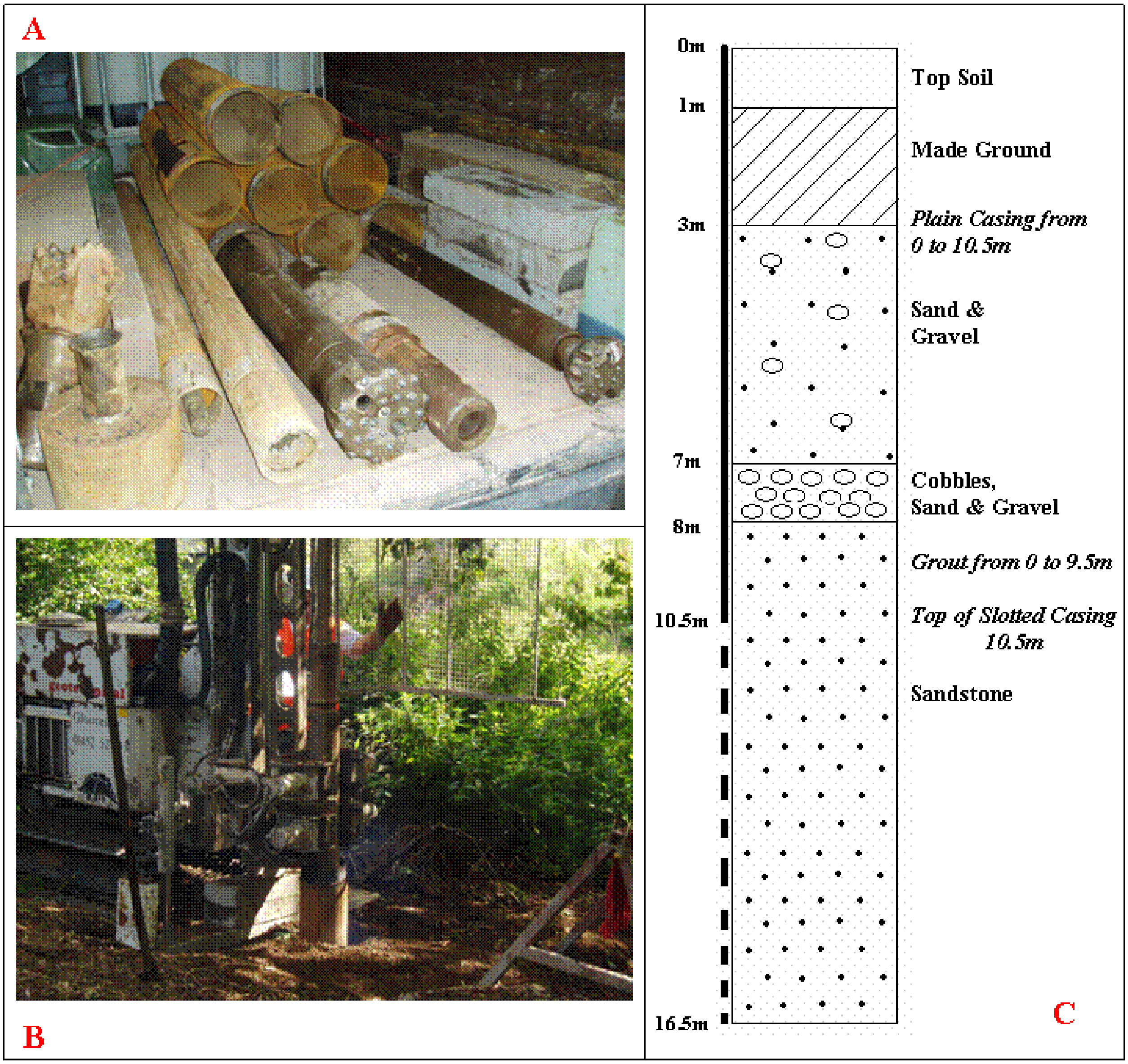

Figure 4.2.1 Installation of borehole on the north bank of the River Tame; A, Drill bits and casing used; B, Drilling the borehole; C, Borehole log indicating depth below ground level of different sediment layers and the depths of plain and slotted casing

4.3 Pump Test

A groundwater sample was required from the borehole for chemical analysis as part of the main research study on the site. A submersible pump was used to remove enough well-bore volumes to retrieve a suitable water sample. The opportunity was taken to measure the drawdown in the well during pumping and recovery to allow an estimate of the sandstone hydraulic conductivity to be calculated.

Collecting accurate drawdown data from a pumping test will depend on the following:

- Maintaining a constant yield during the test
- Measuring the drawdown in the pumping well carefully
- Taking drawdown readings at appropriate intervals
- Recording data from both the pumping and recovery periods

A small generator was used to power the pump. Pumping was at a constant discharge of 0.67 l/s. The discharge was measured by observing the time taken to fill a 10 litre bucket. This is a practical way of measuring discharge at low pumping rates (Driscoll, 1995).

Water level measurements were only made in the pumping borehole. Ideally, measurements from nearby piezometers are needed to get accurate data from a pumping test, however there were none available at the time the test was carried out. The accuracy of data taken from single-well pump tests is usually less reliable because of turbulence created by the pump, however the data can still be used to give a useful estimate of the aquifer transmissivity.

Measurements were made using a dip meter which provided a noise when the probe was immersed in water. Prior to pumping the static water level in the borehole was measured. All measurements were taken relative to the top of the well casing. As early test data provide the most important information when conducting a pumping test, on commencement of pumping measurements of the water level were recorded to the nearest 10 seconds. During the pumping, the pump became jammed (possibly with leaves and debris that were sitting in the borehole), to resolve this issue the pump was switched on and off a couple of times to clear the blockage and continue pumping at a constant rate. This resulted in water level data only being accurately measured for the first 3 minutes of pumping. Following 1 hour of continuous pumping the water level in the borehole was at a relatively steady state. The pump was switched off and the recovery was monitored. During the first four minutes of recovery the water level was recorded every 10 seconds, and then for every 30 seconds until the water level had recovered to its original level in just over 10 minutes.

4.4 Analysis of Pump Test Data

Pump test data was analysed using the Cooper-Jacob (1946) method which can be applied to single-well pump tests to estimate the aquifer transmissivity. The drawdown observed during the pump test was calculated by subtracting the water level measurements from the original static water level measured prior to pumping. The pumping and recovery data were analysed separately.

Data obtained during the discharge period of the pump test were analysed by constructing a semi-log plot of observed values of drawdown in the pumping well (s_w) versus the time (t) on the

logarithmic scale. A best-fit straight line was drawn through the data points, and the slope of the line calculated, i.e. the drawdown difference Δs_w per log cycle of time. Using the Cooper-Jacob method the transmissivity is calculated from:

$$KD = \frac{2.30Q}{4\pi\Delta s_w}$$

Where:

- KD is the transmissivity (K is hydraulic conductivity, D is saturated aquifer thickness)
- Q is well discharge

Using the relationship that the transmissivity is equal to the hydraulic conductivity multiplied by the saturated aquifer thickness, an estimate of the hydraulic conductivity for the sandstone aquifer could be made. As the borehole only partially penetrated the aquifer, the length of the slotted casing (6m) was used to estimate the hydraulic conductivity; this was appropriate because the borehole only penetrated 10% of the aquifer thickness (assuming it is 100m thick) and has a low transmissivity (Halford et al, 2006).

Data obtained from the recovery period of the test was analysed by constructing a semi-log plot of residual drawdown (s'_w) versus the ratio between the time since the start of pumping and time since the end of pumping (t/t') on the logarithmic scale. A best fit straight line was drawn through the data points, and the slope of the line calculated ($\Delta s'_w$). The transmissivity is calculated using:

$$KD = \frac{2.30Q}{4\pi\Delta s' w}$$

The transmissivity calculated for the recovery period was used to estimate the hydraulic conductivity using the same method as described above for the discharge period.

4.5 Construction of Riverbed Mini Drive Point Piezometers (MDP's)

In order to identify the hydraulic gradients between the river bed sediments and the river, mini drive point piezometers (MDP's) were constructed and installed at selected profiles within the river bed. A series of multi-level chemical sampling tubes were attached to the MDP in order to enable chemical water samples to be taken during future experimental work at the site. The MDP's were required to cover a depth of up to 1m below the river bed, in order to full penetrate the hyporheic zone which is thought to be up to 0.6m deep in this stretch of the river (Ellis, 2002).

The MDP's were constructed using flexible, high density polyethylene tubing (HDPE), 100 micron nylon mesh, stainless steel bolts, washers and wire. The HDPE tubing had an internal diameter of 10mm and an outer diameter of 13mm. Chemical sampling tubes were required at 10cm intervals over a 1m depth of the river bed, whereas half of the MDP's were required to have an open section at 1m depth and the other half at 0.5m depth. To allow for this all the MDP's were designed to be installed at a depth of 1m below the river bed, with the position of the open section of the MDP modified accordingly, the basic design is shown in Fig 4.5.1. The HDPE tubing was cut into 2m lengths. For the 1m MDP's open section a 4mm drill bit was used to drill holes through the tubing placed 1cm apart longitudinally to cover a 10cm length at the end of the tube. The drilled section of tube was then covered in 1 layer of 100 micron nylon mesh which was secured with thin, stainless steel wire (Fig 4.5.2). The nylon mesh acts as a

screen to prevent clogging of the open section. For the 0.5m MDP's the open section was constructed in the same way however it was positioned 0.5m from the end of the tubing.

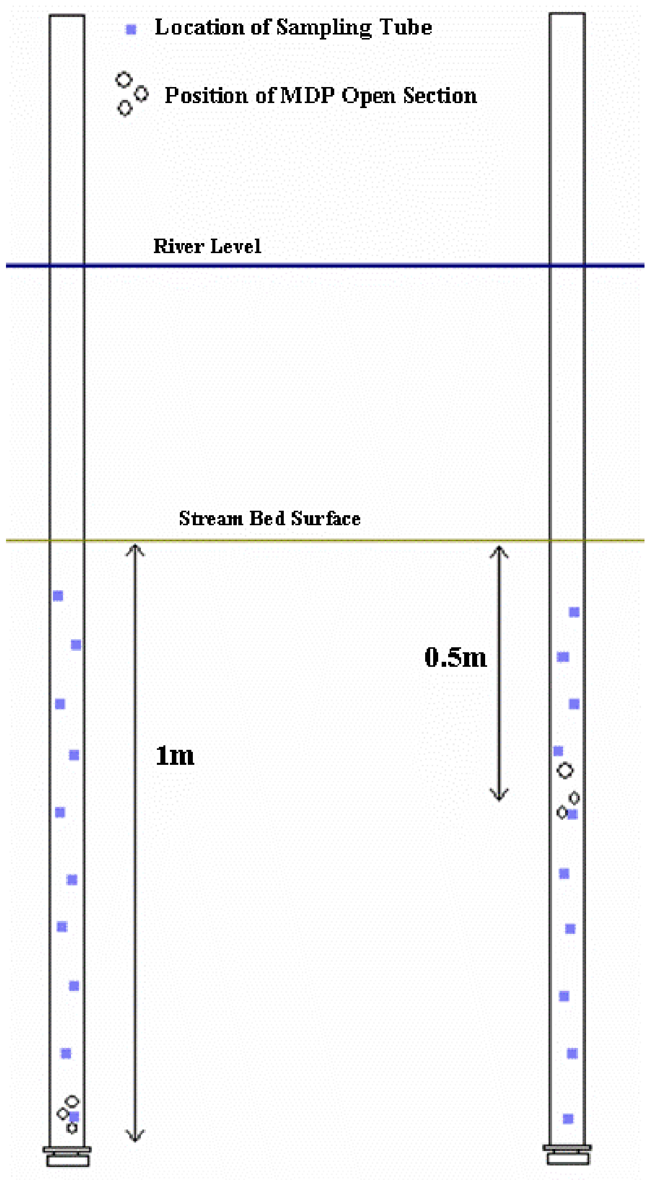

Figure 4.5.1 Schematic diagram of MDP design. The 1m MDP (on the left) is inserted to a depth of 1m below the stream bed, the open section is at a depth of 1m, the position of the multi-level samplers are placed at 10cm intervals over a depth of 1m. The 0.5m MDP (on the right) is inserted to a depth of 1m below the stream bed, the open section is at a depth of 0.5m, multi-level samplers are placed at 10cm intervals over a depth of 1m.

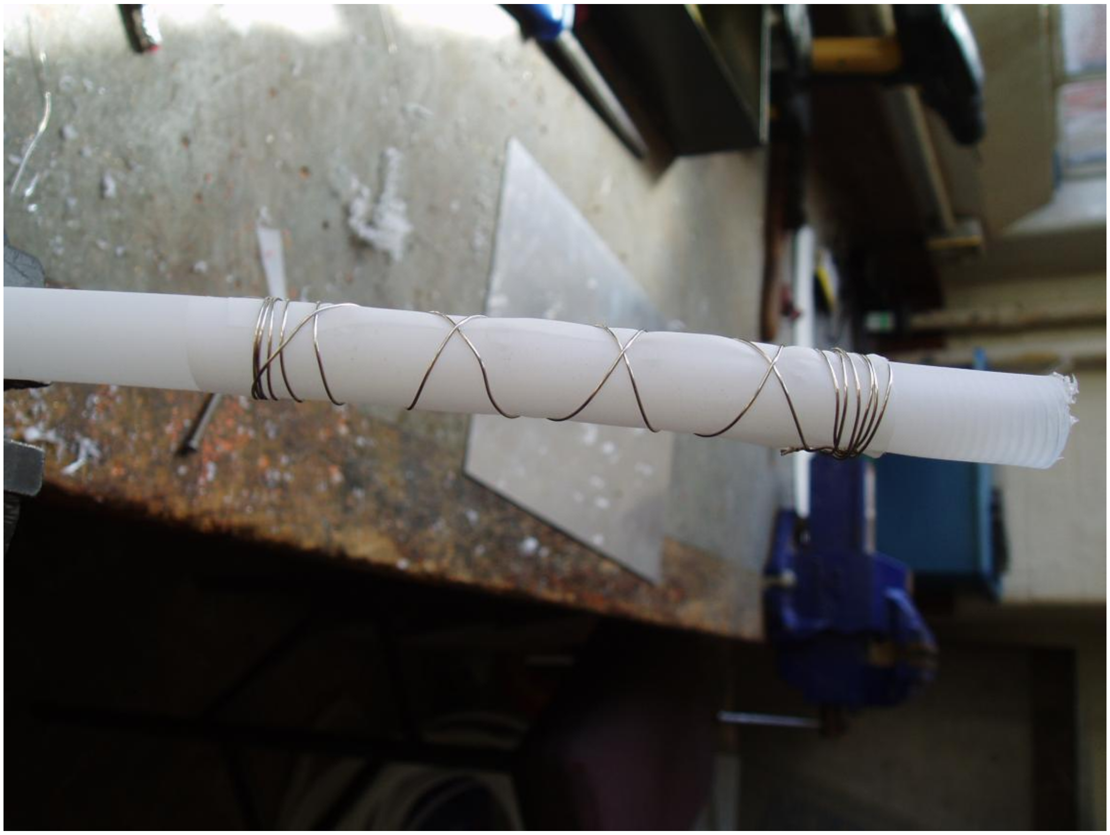

Figure 4.5.2 Open section of MDP. 100 micron nylon mesh is held in place by stainless steel wire to cover the drilled holes.

Following the construction of the MDP, 10 sampling tubes were attached to its outside at 10cm intervals. The sampling tubes were constructed using 3mm outer diameter HDPE tubing, 100 micron nylon mesh and stainless steel wire. 10 lengths of HDPE tubing were needed, the first was cut to a length of 200cm and then subsequent lengths were cut 5cm shorter than the previous one, with the shortest being 155cm. As the separation distance of the sampling tubes was to be 10cm, cutting each tube to a different length allowed easy identification of the tube sample depth in the field. The shortest tube above the river bed would correspond to the deepest sampling point, and the longest tube would represent the shallowest. One end of the sampling tube was wrapped in 100 micron nylon mesh and held in place with stainless steel wire. To attach the

sampling tubes to the MDP, 10cm intervals from the end of the MDP tube were measured; a single length of stainless steel wire was wrapped around the MDP, and then the first sampling tube (1m depth) to attach it securely to the MDP. The wire was continually wrapped around the MDP to ensure that the sampling tube would remain in place, for every 10cm interval another sampling tube was attached until all 10 were securely held in place (Fig 4.5.3). Once the sampling tubes were attached, brass screws were inserted into the open end to prevent algae growth and cross contamination of sample horizons. Fig 4.5.4 shows the completed multi-level, mini drive point piezometer.

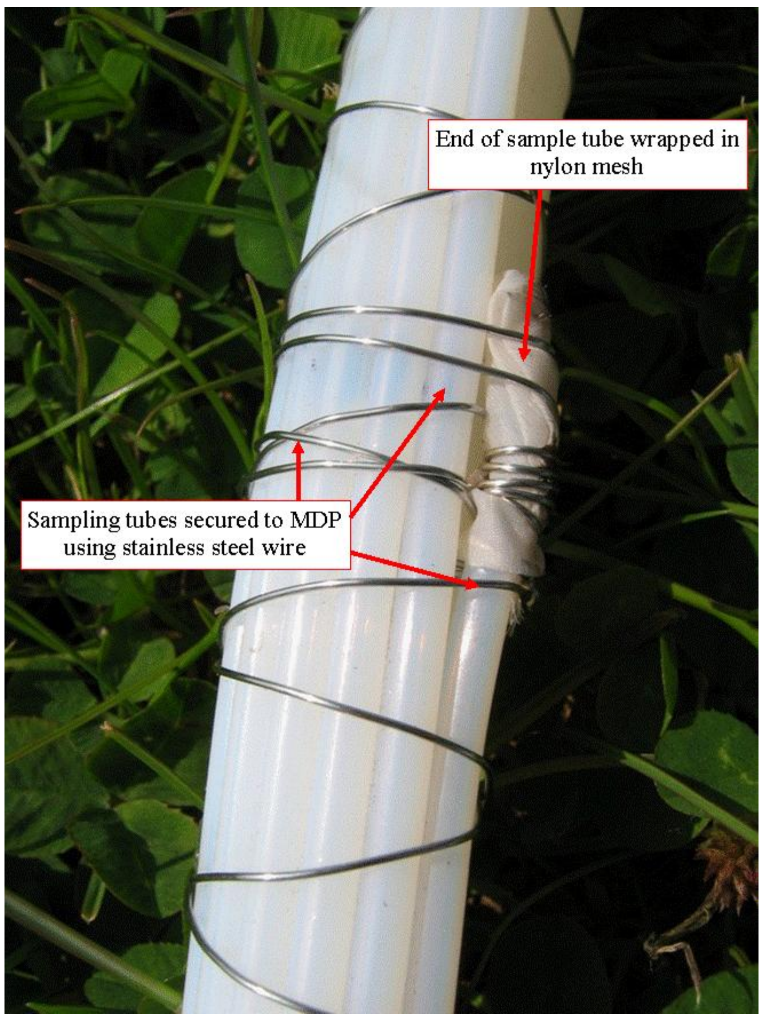

Figure 4.5.3 Design and attachment of sampling tube to the MDP.

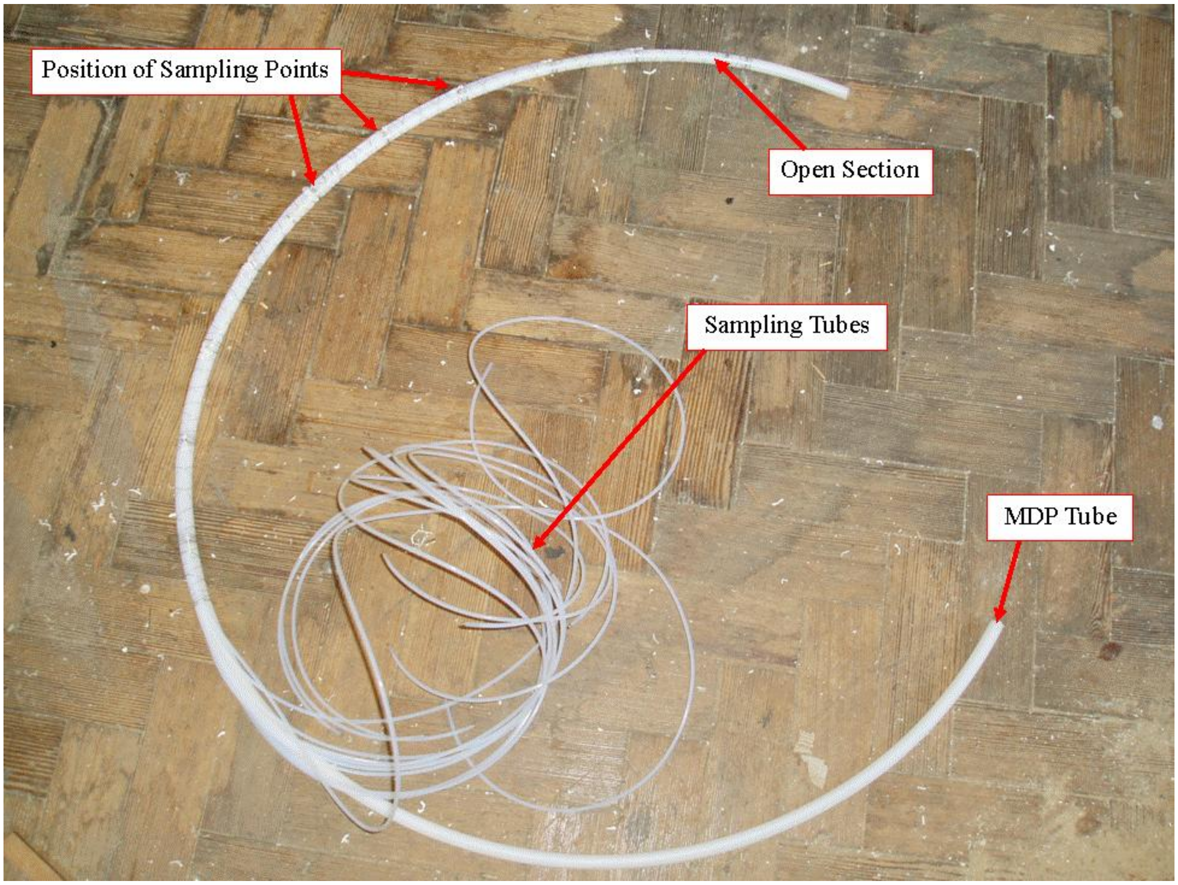

Figure 4.5.4 Completed multi-level, mini drive point piezometer

4.6 Installation of MDP's

The installation of the piezometers required a hollow 3m steel pole and a 15kg fence post driver. The piezometer was inserted inside the steel pole with care taken to ensure the sampling tubes did not slip out of place. Once the piezometer was inside the steel pole two washers and a stainless steel bolt were attached to the end of the MDP tube, as the washers and bolt head had a larger diameter than the steel pole they would prevent slippage of the piezometer during installation and act as the drive point. The steel pole was driven by hand, using the fence post driver, into the river bed to the required depth which was marked on the outside of the steel pole

(Fig4.6.1). Once the steel pole had been driven to the required depth, it was removed allowing the sediments to collapse naturally around the piezometer and hold it in place.

Figure 4.6.1 Installation of piezometer, a fence driver post is used to drive the steel pole to the required depth

The variability of sand and gravel bed sediments in the Tame provided varying levels of resistance to installation of the piezometers. In many cases it was not possible to drive the piezometers down to the required 1m depth. In these cases the piezometer was driven as far as possible. The length of the MDP tubing remaining above the river bed was measured for each piezometer so its exact depth could be calculated.

Particular problems were encountered during installation of the MDP's which had an open section at 0.5m along its length. In some cases these piezometers were driven down to 0.5m or less resulting in the open section protruding above the river bed surface; some of the sampling tubes were below the river bed. For these locations simple MDP's were constructed, with the open section at the end of the HDPE tubing. Their construction used the same method as described in section 4.5 however no chemical sampling tubes were attached to these piezometers, which were then installed next to the previously installed multi-level MDP's.

4.7 Hydraulic Gradient

Calculating the hydraulic gradients between the groundwater and the river has been carried out to establish the direction of water movement below the river bed i.e. gaining or losing. To calculate the hydraulic gradient, groundwater head measurements were taken from the MDP's.

The piezometers were held upright and the water level was given time to equilibrate with atmospheric pressure. Due to the relatively transparent nature of the tubing used in the construction of the MDP's, the groundwater head level could be measured directly using a tape measure. This was possible because the piezometers had been newly installed. However, even after only a week of installation, there was significant algae growth on some of the piezometers which made it difficult to observe the water level. This was resolved by cleaning the algae off the piezometer tubes using a metal scouring pad. The head levels in the piezometers were measured as height above the river water level, which was used as the datum.

The groundwater head level and known depth of the piezometer could then be used to calculate the hydraulic gradient as follows:

$$i = \frac{h_2 - h_1}{L}$$

Where:

- i is the hydraulic gradient
- h_1 is the river water level (datum point = 0m)
- h_2 is the water level in the piezometer
- L is the depth of the open section of the piezometer in the river bed

4.8 Falling Head Tests

Falling head tests were conducted on all the MDP's to calculate estimates of the hydraulic conductivity of the river bed sediments. A falling head test involves the application of a greater hydraulic pressure in the piezometer, and measuring the rate of change in the pressure as it equilibrates to the natural conditions. The change in hydraulic pressure was done, in this case, by introducing water into the piezometer tube. The rate of change in pressure was represented by measuring the rate of change of the water level in the piezometer.

To conduct the test the open end of the piezometer tube was held under the river water level and allowed to fill with river water, making sure there was no air left in the tube. By placing a thumb over the end of the tube, it could be lifted upright out of the water and the hydraulic pressure could be maintained. The piezometer was held next to a wooden board which had 10cm intervals

marked onto it (Fig 4.8.1). The thumb was removed from the end of the tube and the timer was started. As the water level dropped the time taken to reach each 10cm interval was noted until the water level had reached a steady state. This process was repeated 3 times for each piezometer to ensure consistency of the data.

Figure 4.8.1 Falling head test on piezometer. 10cm intervals were marked on the wooden board held next to the piezometer to allow the change in head level to be easily measured.

4.9 Analysis of Falling Head Test Data

The analysis of falling head test data taken from the piezometers was done using the Hvorslev (1951) method. Hvorslev found that water levels will return to their normal static level at an exponential rate, with the time taken dependent on the hydraulic conductivity of the porous

material (Hiscock, 2005). The calculation of hydraulic conductivity can be made using the following equation:

$$K = \frac{r^2 \log_e (L/R)}{2LT_0}$$

Where:

- K is the hydraulic conductivity
- r is the radius of the well casing
- L is the length of the piezometer open section
- R is the radius of the well screen
- T_0 is the time taken for the water level to fall to 37% of the initial change

(see Fig 4.9.1a)

The calculation of hydraulic conductivity can be found provided the length, L, is greater than eight times the radius of the well screen, R.

A semi-log plot of the ratio, h/h_0 (where, h is the height of the water level above the static water level, and h_0 is the initial height of the water level above the static level at the start of the test) versus time, t, was made and a straight line was fitted to the data, from which T_0 could be found (see Fig 4.9.1b). The T_0 value could then be used in the above equation and the hydraulic conductivity calculated.

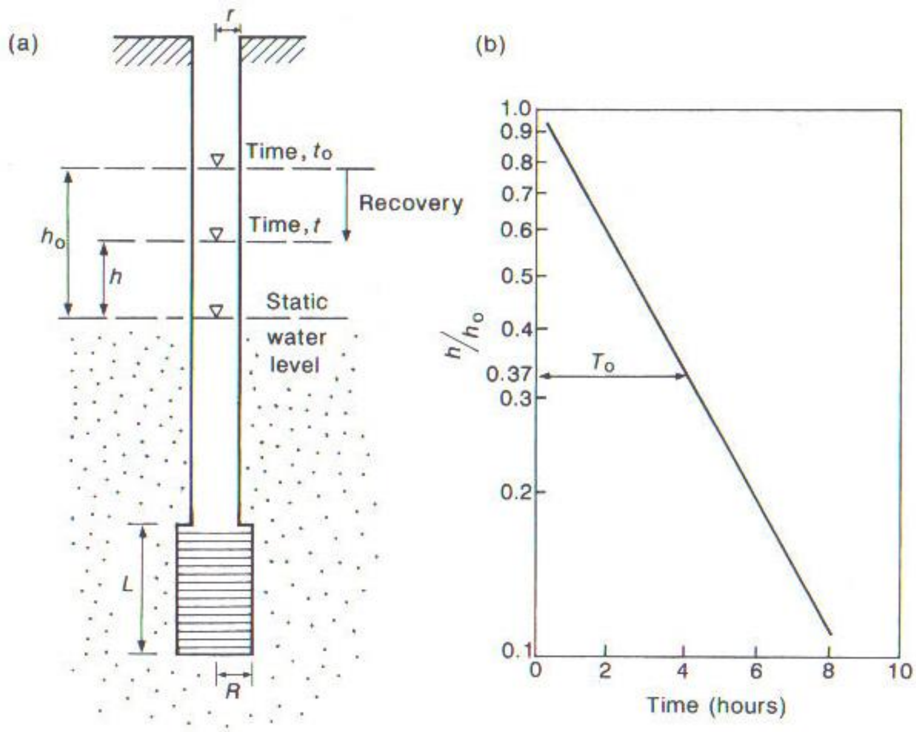

Figure 4.9.1 Hvorslev piezometer test; (a) piezometer geometry; (b) graphical method for analysis of T_0 (*From Hiscock, 2005*)

A hydraulic conductivity value was calculated for each of the three falling head tests carried out on an individual piezometer. The arithmetic mean of the three values was then taken to give a hydraulic conductivity estimate for the river bed sediments surrounding that piezometer.

Chapter 5. Fieldwork Results

5.1 Introduction

The collection and analysis of data collected in the field are described in the previous chapter. This chapter presents the results, which aimed to define the hydraulic properties of the sandstone aquifer and river bed sediments, for use in subsequent modelling.

5.2 Pump Test

The full list of results taken from the pumping test are given in Appendix 1. The graphs of drawdown (s) versus time (t) for the discharge period of the test and for residual drawdown (s') versus t/t' are shown in Fig 5.2.1 and 5.2.2 respectively.

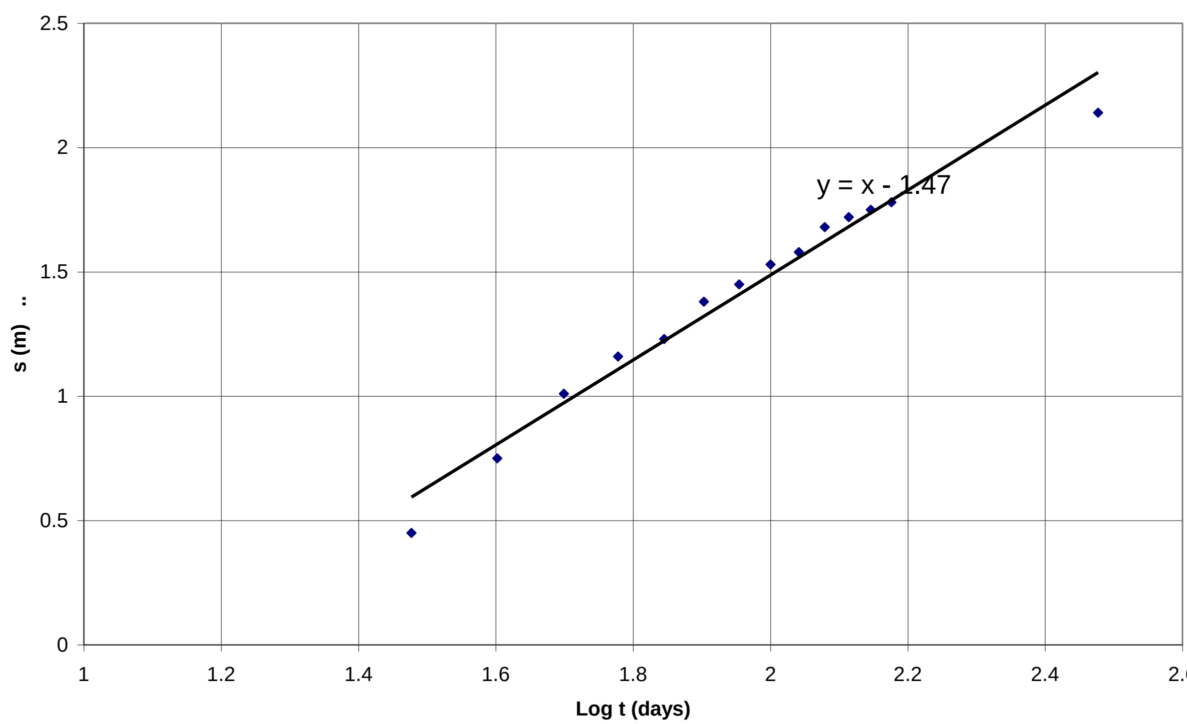

Figure 5.2.1 Discharge period of pump test

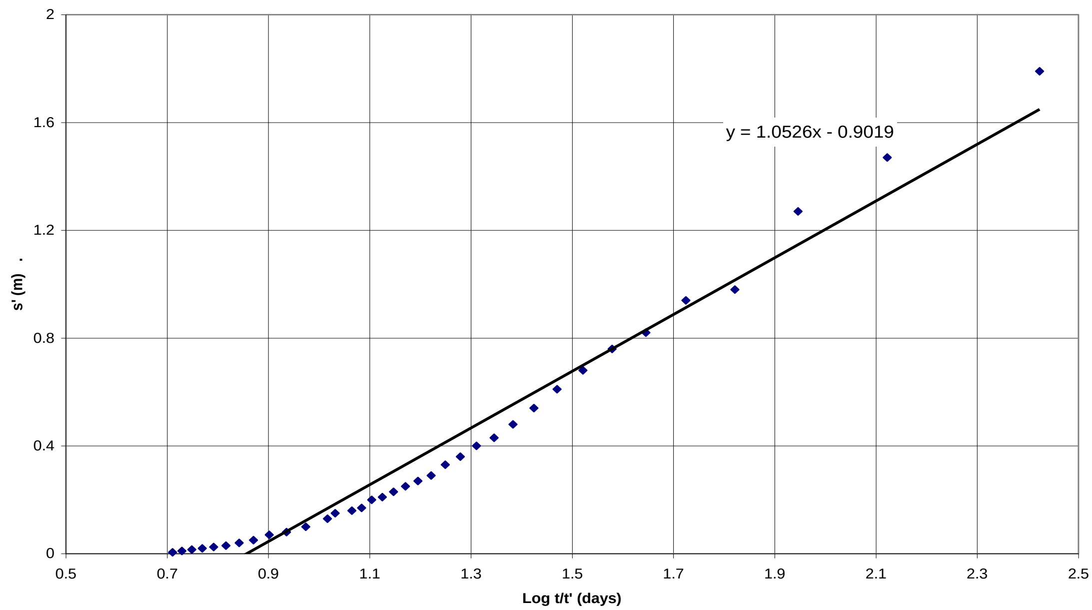

Figure 5.2.2 Recovery period of pump test

The calculated transmissivity and estimates of the hydraulic conductivity are given in Table 5.2.1 below.

Test Period	Calculated T (m^2/d)	Estimated K (m/d)
Discharge	10.61	1.76
Recovery	10.09	1.68

Table 5.2.1 Calculated transmissivity and estimated hydraulic conductivity values from discharge and recovery periods of the pump test.

The two estimates for the hydraulic conductivity of the sandstone aquifer show good consistency. The arithmetic mean hydraulic conductivity is 1.72 m/d. This result is consistent with previous estimates of the hydraulic conductivity of the Kidderminster Sandstone, which is given as 2 m/d by Shepherd et al (2006). As the pump test had only been monitored at the

pumping borehole, transmissivity is the only hydraulic property for the sandstone which could be calculated.

5.3 Piezometer Installation

In total 20 piezometers were installed in the river bed at 8 lateral profiles along the study reach. The locations of the piezometers are shown in Fig 5.3.1.

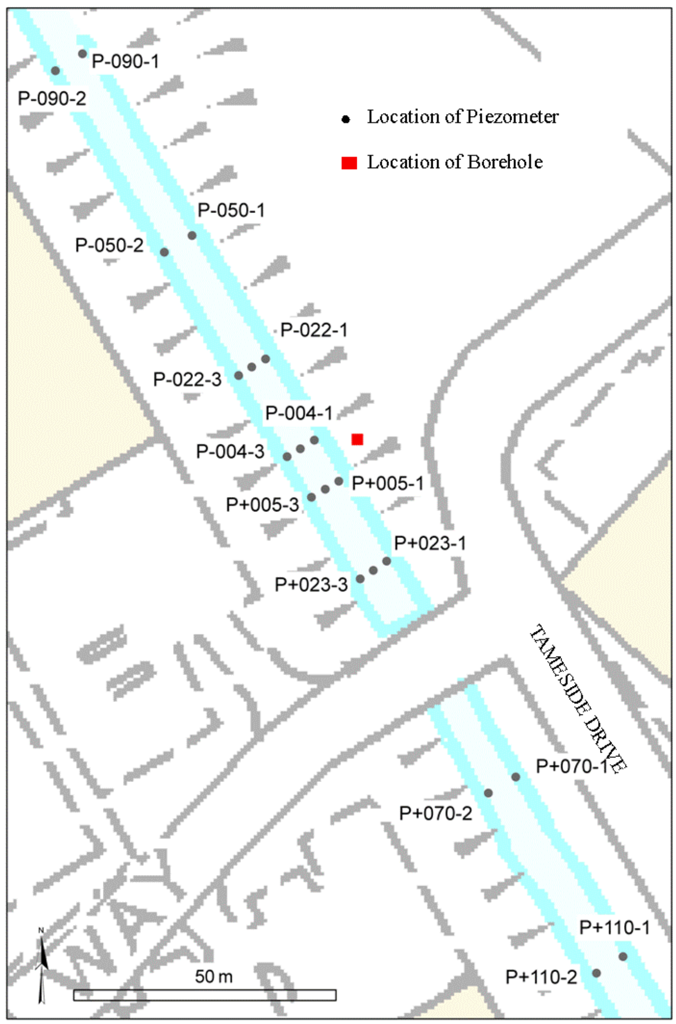

Figure 5.3.1 Map of study site indicating the location of all the piezometers installed on site for this project

The sequence in which the piezometers are referenced is as follows. First the location of the piezometer (P) in relation to the borehole installed on site is referenced; -, if it is upstream of the borehole, +, if it is downstream of the borehole, along with its distance in metres from the borehole, e.g. P-050, refers to a piezometer 50 m upstream from the borehole. Finally the position of the piezometer in the river is noted; 1, if it is closest to the north river bank; 2, if it is in the centre of the river; 3, if it is closest to the south river bank. The locations of the piezometers can be easily found in the field as a yellow of spot, painted onto the rocks on the south river bank, indicates the position of each piezometer profile.

The installation depths of the MDP's ranged from 0.02 m to 1m. They were driven on average to 0.51 m. Penetration of the bed sediment became increasingly difficult below this depth due to the nature of the sediment strata. A full list of the depths of each piezometer is given in Appendix 3.

The positioning of the piezometers across the channel and at varying depths was done to assess the heterogeneities within the river bed sediments. The positioning of an equal number of piezometers (and similar pattern) either side of the borehole was done so that the influence of pump tests, taking place in the future, on the groundwater levels within the river bed could be accurately monitored.

5.4 Groundwater Head Measurements

The groundwater level in each piezometer was measured as height above the river water level. It would be expected that in a homogeneous medium, the groundwater head would increase with depth. However a plot (shown in Fig 5.4.1) of the piezometer groundwater head measurements

versus their depth produced a large scatter of results which did not follow this rule and indicated the heterogeneous nature of the river bed sediments

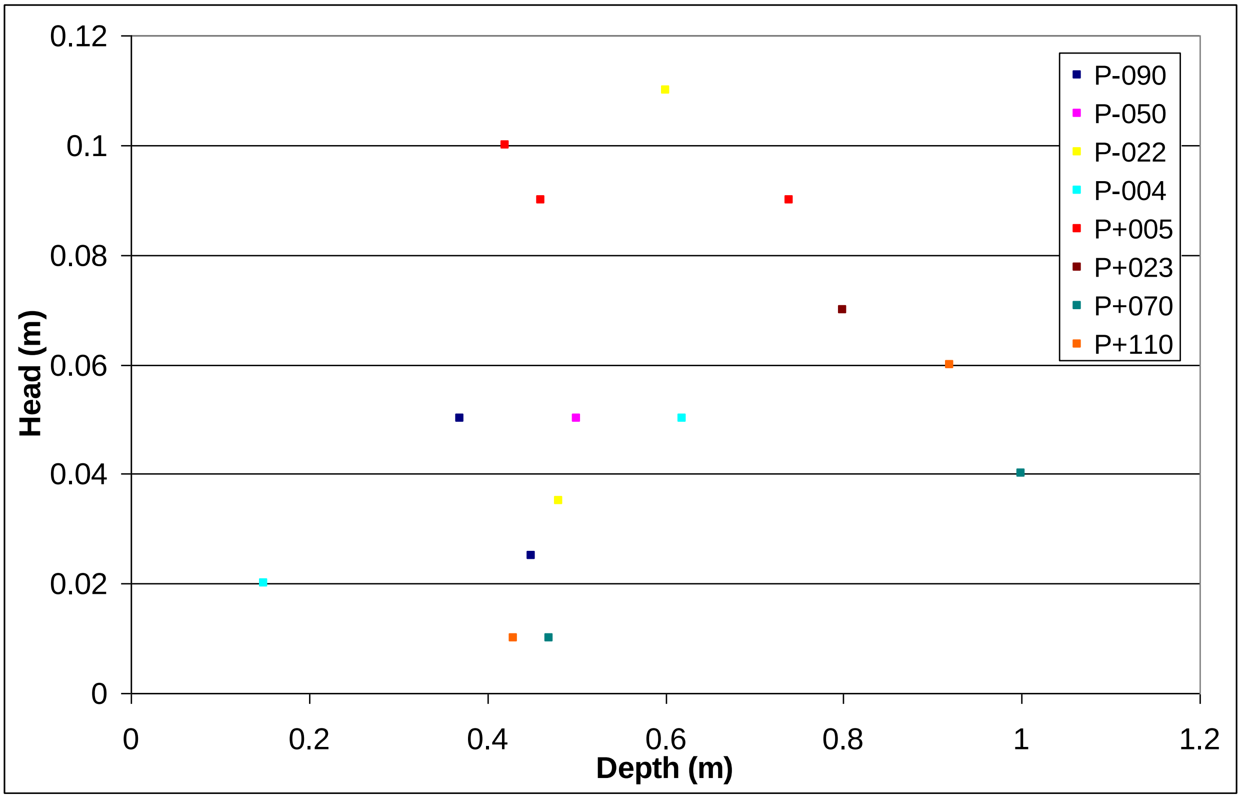

Figure 5.4.1 Plot of groundwater head versus depth, indicates the heterogeneous nature of the river bed sediments. Individual piezometer profiles are picked out.

The plot shown in Fig 5.4.1 does, however, show some correlation between groundwater head and depth if individual piezometer profiles are picked out. Profiles P-004, P+100 and P-022 all indicate that the groundwater head is increasing with depth, suggesting that there may be less vertical heterogeneity in these areas. The profiles for P+005 and P-090 do not show this

relationship, with the groundwater head appearing to decrease with depth, indicating large heterogeneities within the river bed sediments.

5.5 Hydraulic Gradients

Groundwater head measurements were taken for all the piezometers, and this value along with the known depth of the piezometer was used to calculate the hydraulic gradient of the river. The results are given in Table 5.5.1.

Piezometer	Depth (m)	Groundwater Head (m.a.d)	Calculated Hydraulic Gradient
P-090-1	0.45	0.025	0.0556
P-090-2	0.37	0.05	0.1351
P-050-1	0.5	0.05	0.1
P-050-2	0.5	0	0
P-022-1	0.55	unable to measure	?
P-022-2	0.48	0.035	0.0729
P-022-3	0.6	0.11	0.1833
P-004-1	0.17	0	0
P-004-2	0.62	0.05	0.0806
P-004-3	0.15	0.02	0.1333
P+005-1	0.46	0.09	0.1957
P+005-2	0.42	0.1	0.2381
P+005-3	0.74	0.09	0.1216
P+023-1	0.24	0	0
P+023-2	0.8	0.07	0.0875
P+023-3	0.32	0	0
P+070-1	0.47	0.01	0.0213
P+070-2	1	0.04	0.04
P+110-1	0.92	0.06	0.0652
P+110-2	0.43	0.01	0.0233

Table 5.5.1 Calculated hydraulic gradients for each piezometer. Groundwater head measured as height above datum (river water level).

The results give a range of hydraulic gradients from 0 to 0.2381, with an average gradient of 0.1036. This indicates that along much of the study reach the river is gaining. P-050-2, P-004-1, P+023-1 and P+023-3 indicated a head level equal to that of the river indicating neutral conditions, i.e. the river is neither gaining nor losing at these points.

The range of hydraulic gradients observed indicates the heterogeneous nature of the streambed. These heterogeneities are particularly apparent by comparing the results of piezometers from a single profile. For example piezometer 1 in profile P-90 has a hydraulic gradient of 0.0556, which is less than half that observed in piezometer 2 in the same profile which has a gradient of 0.1351. The variations indicate that the stream bed is heterogeneous both vertically and laterally.

No calculation of the hydraulic gradient could be made for P-022-1. This was because a measurement of the groundwater head could not be made in the field; this is discussed further in section 5.6.

5.6 Falling Head Tests

Falling Head tests were conducted in 17 of the 20 piezometers installed during the fieldwork.

Falling head tests could not be carried out in P+070-1, P+070-2 and P+110-2, this was due to the river water level being deeper in these area than expected during installation of the piezometers resulting in the piezometer tube being too short to and sticking out no more than about 15cm above the river surface. This meant that there was not a long enough section of piezometer tube to conduct a falling head test in. Attempts were made to connect a longer piece of tubing onto the

end of the piezometers using plastic connectors and metal clamps, however this method was unsuccessful.

The results of falling head tests in 5 of the riverbed piezometers failed or proved inconclusive when tested. The tests on P-022-1 were recorded as having failed when no percolation of water occurred through the open section. Several reasons may account for this type of failure. It is likely to have been caused by either damage to the screened section during the installation procedure or as a result of encountering very low permeability strata. The tests on P-050-2, P-004-1, P+023-1 and P+023-3 proved inconclusive. As the groundwater level in these piezometers was in equilibrium with the river level, there was no pressure difference between the two, resulting in a falling head test being impossible to carry out. This may suggest that the piezometer was in fact in hydraulic connection with the river rather than the groundwater in the river bed sediments. This could possibly be due to damage caused during installation to the piezometer and/or the removal and reorganisation of sediments within the river bed during installation.

The 12 falling head tests that proved successful were analysed using the Hvorslev method. Values obtained for the bed conductivity ranged between 0.61 m/d and 52.96 m/d, indicating the heterogeneous nature of the sediments. The full set of results is given in Table 5.6.1.

Piezometer	Depth (m)	K (m/d)
P-090-1	0.45	9.31
P-090-2	0.37	6.32
P-050-1	0.5	2.09
P-050-2	0.5	?
P-022-1	0.55	?
P-022-2	0.48	6.54
P-022-3	0.6	0.61
P-004-1	0.17	?
P-004-2	0.62	3.67
P-004-3	0.15	10.45
P+005-1	0.46	52.96
P+005-2	0.42	16.2
P+005-3	0.74	1.85
P+023-1	0.24	?
P+023-2	0.8	12.58
P+023-3	0.32	?
P+070-1	0.47	?
P+070-2	1	?
P+110-1	0.92	1.24
P+110-2	0.43	?

Table 5.6.1 Hydraulic conductivity estimates from falling head test. ? indicates failed, inconclusive or no test.

A cumulative probability plot of the hydraulic conductivity values found that they had a log normal distribution; however this did not include the hydraulic conductivity value for P+005-1 which appeared to be an erroneous point (see Fig 5.6.1). The hydraulic conductivity calculated for this piezometer was 52.96 m/d, which was much greater than all the other calculated values. There are two possible reasons for this result; it could be a genuine result and indicate a very high permeability zone within the river bed sediments, or the sediments could have been disturbed during installation of the piezometers creating an empty space around the open section

of the piezometer. The second reason may be valid as the falling head tests were only carried out after 1-2 weeks of installation, which may not have been enough time for the sediments to have completely collapsed back around all the piezometers. A repeat of all the falling head tests needs to be made in the future which may help indicate if this is the case i.e. if the hydraulic conductivity has significantly changed.

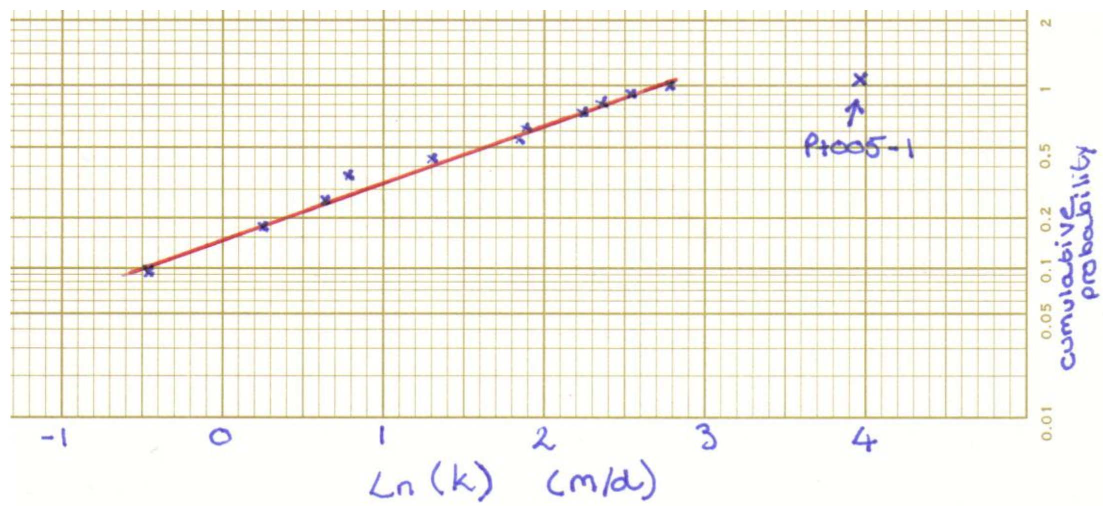

Figure 5.6.1 Cumulative probability plot of river bed hydraulic conductivity values indicating a log normal distribution; however this does not include the value calculated at P+005-1

It was decided that the hydraulic conductivity of P+005-1 would be included in all calculations to deduce the mean hydraulic conductivity of the river bed sediments. Due to the log normal distribution of the results the geometric mean value for the river bed sediments was calculated and found to be 5.28 m/d.

The values obtained for the river bed hydraulic conductivity were then split into two categories, those at 0m to 0.5m depth below the river bed and those at >0.5m to 1m below the river bed. This was to assess the differences in values at varying depths.

Values obtained for the river bed hydraulic conductivity between 0m and 0.5m ranged between 2.09 m/d and 52.96 m/d, giving a geometric mean of 9.54 m/d. Values obtained for the river bed hydraulic conductivity between 0.5m and 1m ranged between 0.61 m/d and 12.58 m/d, giving a geometric mean of 2.30 m/d.

The range of hydraulic conductivity values observed between the individual piezometers indicates the extremely heterogeneous nature of the river bed sediments. Low values of hydraulic conductivity, such as 0.61 m/d calculated in P-022-3, could indicate low permeability zones, e.g. clay lenses, within the sediments.

5.7 Conclusions

Estimates of hydraulic conductivity for the sandstone from pump test data, gave good results which were comparable with literature values.

The geometric mean value for hydraulic conductivity (5.28 m/d) from falling head test analysis is considered to provide a good estimate for the combined vertical and horizontal riverbed hydraulic conductivity. A repeat of all the falling head tests needs to be made in the future to allow the river bed sediments more time to settle out following the piezometer installation. New piezometers may need to be installed to solve the problems encountered when attempting to

carry out some of the falling head tests, i.e. P-022-1 where the open section appears to be blocked or damaged and longer piezometers installed in the locations where the current ones are too short.

Overall the results shown for hydraulic conductivity, groundwater head and hydraulic gradients support the theory that river bed sediments have a very heterogeneous nature. All the data sets described in this chapter are given in the appendices.

Chapter 6. Modelling

6.1 Introduction

To analyse the effects of a proposed pump test on the study site a basic numerical model of the aquifer system was created. The aim of the modelling work was to investigate:

- the effect of pumping on the hydraulic gradients observed within the river bed
- the sensitivity of both known and estimated hydraulic parameters applied to the model

The model could then be used to predict the expected response in the river bed piezometers for future experiments on the site.

The basic model grid had been created by Dr. Veronique Durand; this model was used to apply the hydraulic parameters established in the field and test its sensitivity. MODFLOW was the code used in the modelling. MODFLOW uses a block-centred finite-difference approach to simulating groundwater flow within an aquifer. The code was run using the software interface Processing Modflow Pro (version 7.0.11), which allows for a large quantity of layers to be modelled.

6.2 Basic Conceptual Model

A conceptual model of the groundwater flow system was established, using field and archive data, to describe the aquifer geometry, the local boundary conditions and estimates of the hydraulic parameters. To conceptualise the system some simplifications had to be made.

6.2.1 Aquifer geometry

The aquifer geometry was based upon the geometry used in modelling work on the same site by Lydon (2006). Four distinct lithological units were identified from borehole data taken from the surrounding area. The elevations Lydon based his model on were used here, however for this model the upper layers were considered to be made ground rather than the glacio-fluvial deposits suggested by Lydon, Table 6.2.1 lists the elevations of stratigraphical units used in the model. All layers were assumed to be horizontal and of equal lateral thickness. The basic geometry is shown in Fig 6.2.1.

Model Layer	**Stratigraphical Unit**	**Top Elevation (maod)**	**Base Elevation (maod)**
1 - 3	Made ground / River & River Bed	100	98.65
4 - 17	Glacio-fluvial deposits	98.65	91.65
18 - 21	River terrace deposits	91.65	89.72
22 - 58	Kidderminster Sandstone	89.72	0

Table 6.2.1 Elevations of stratigraphical units used in model (*adapted from Lydon, 2006*)

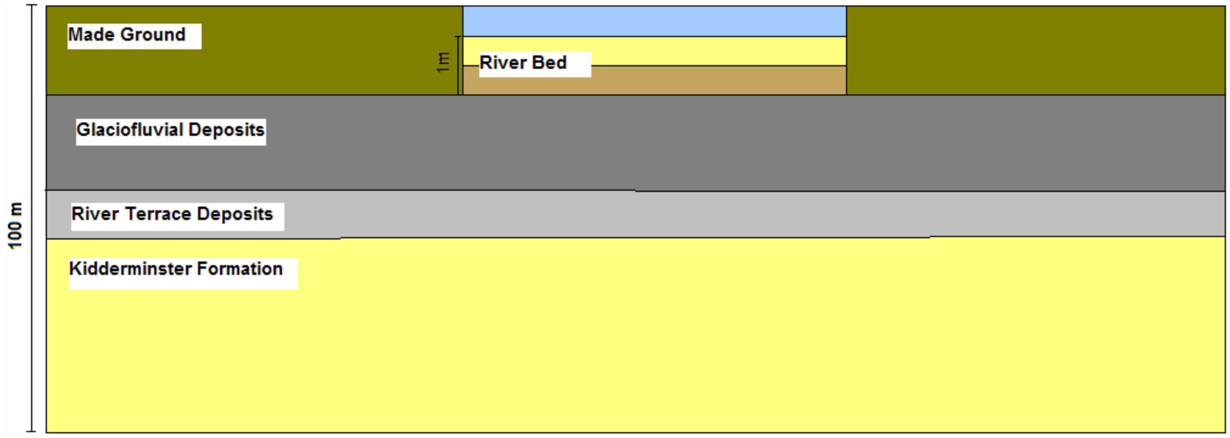

Figure 6.2.1 Basic geometry used in modelling (not to scale)

6.2.2 Boundary conditions/initial heads

Constant head boundaries were set parallel to the River Tame to the north east and south west both had a value of 100 maod along their entire length (Fig 6.2.2). These boundaries were set far enough away from the pumping borehole so that they would not influence it.

The River Tame was also modelled as a constant head boundary of 100 maod. Applying a constant head boundary to act as the river ensured a constant supply of water from the river when running the model. Initial heads across the model were set at 100 maod. Modelling the river and initial head levels in this way was acceptable as the model was to be used to look at the change in groundwater head below the river (i.e. drawdown), not the actual groundwater levels observed.

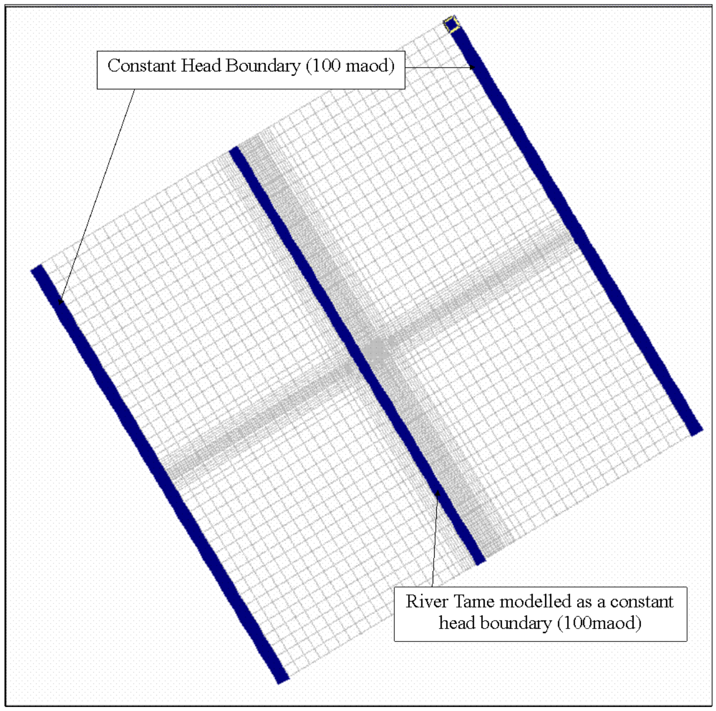

Figure 6.2.2 Position of constant head boundaries on model grid

6.2.3 Aquifer hydraulic parameters

The initial estimates of hydraulic parameters used in the model were taken from the fieldwork results, and combined with literature values for the properties that were not measured in the field. All hydraulic conductivity values were assumed to be isotropic. The initial values used in the model are listed in Table 6.2.2. Values for the made ground were unknown, it was assumed in this case to have a low hydraulic conductivity (1m/d).

The layers of superficial deposits above the sandstone aquifer are not assumed to be confining. A nominal Storativity value of 0.0001 was assigned to all the model layers, although this value is basically not required as the aquifer is unconfined. Specific yield was assigned a value of 0.2 to all model layers.

Stratigraphy	Lithology	Hydraulic Conductivity (m/d)
Made Ground	Various	1
River Bed	Sands, gravels, cobbles	5.3
Glacio-fluvial deposits	Sands, gravels	20
River Terrace Deposits	Sands, gravels	20
Kidderminster Sandstone	Sandstone	2

Table 6.2.2 Initial Hydraulic conductivity values used for model layers

6.2.4 Model Grid

The model grid covers a 410 m by 400 m area. The grid was divided into 78 rows and 59 columns. The model grid was tilted so that the rows would lie parallel to the river (see Fig 6.2.3). The size of the cells in each row and column varied from 1 m to 10m, this was to allow for a finer mesh size around the river, borehole and piezometers where detailed data was needed. The model has a thickness of 100m and was split into 58 layers, with thicknesses ranging from 0.5 m to 5 m. The model contains 266916 cells in total.

The locations of the piezometers at the study site were added to the model as observation points. The river bed is split into two layers both are 0.5m thick; therefore the varying depths of the piezometers could be represented.

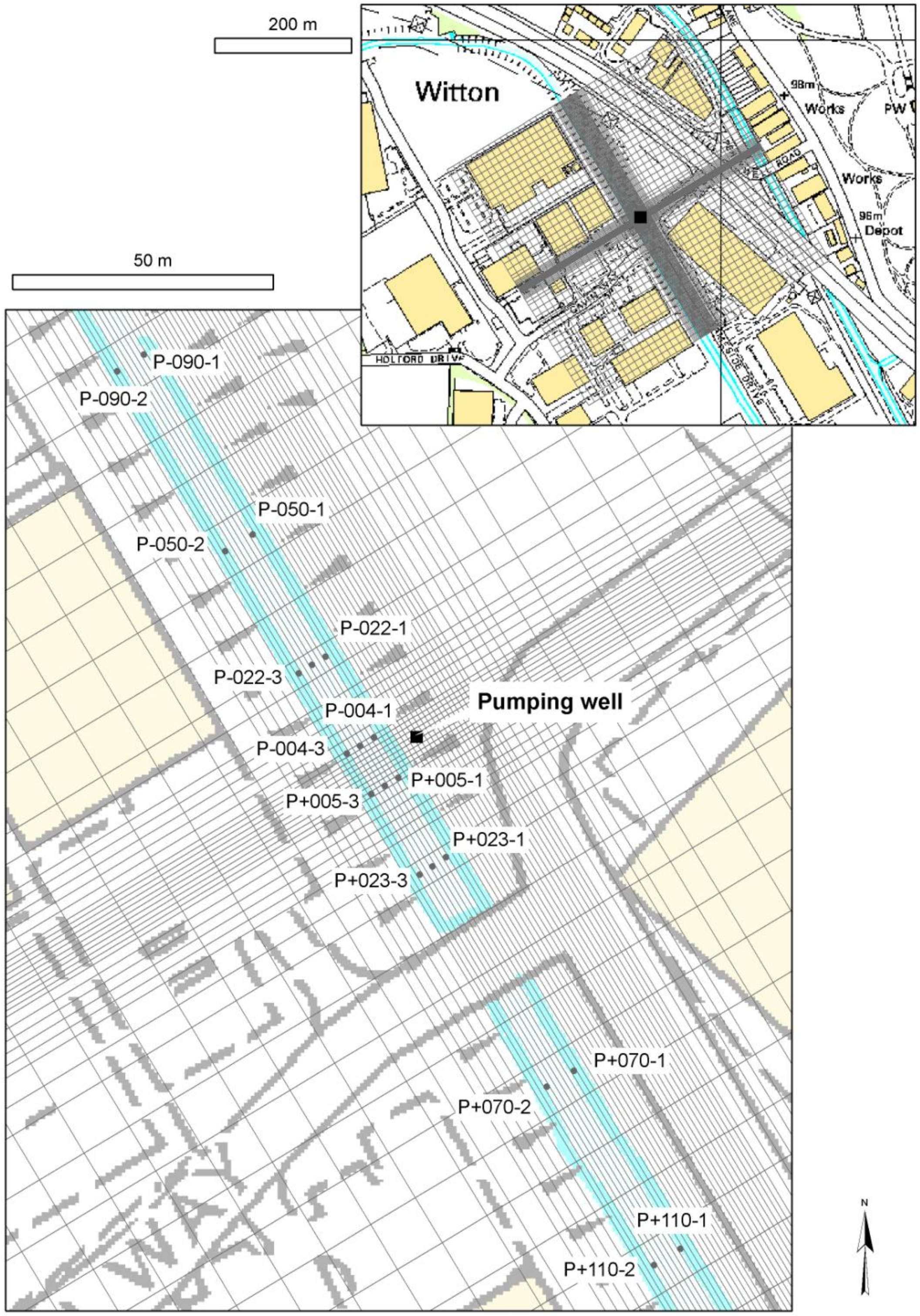

Figure 6.2.3 Location of model grid in relation to the study area (top); detailed grid mesh around borehole, river and piezometers (below)

6.3 Transient Model

With the basic model grid established a transient regime was applied to represent a proposed pump test which is to take place on site in autumn 2007. The pump test is to be carried out over a 4 week period and will involve 4 periods of pumping/ recovery at varying pump rates. By applying the plans for the pumping to this model the expected drawdown, and therefore change in gradient, in the river bed piezometers could be investigated.

The transient model was set up with 9 stress periods. The length of each stress period, number of time steps, and pumping rate are summarised in Table 6.3.1.

Stress Period	Duration (days)	Time Steps	Pumping Rate (m^3/d)
1	1	1	0
2	3	10	50
3	4	20	0
4	3	10	100
5	4	20	0
6	3	10	150
7	4	20	0
8	3	10	200
9	4	20	0

Table 6.3.1 Stress period set up for transient model

6.4 Results

The model was run in transient state using the parameters above. Figure 6.4.1 illustrates the maximum drawdown contours observed 1m below the river bed (layer 3) from stress period 8, time step 10.

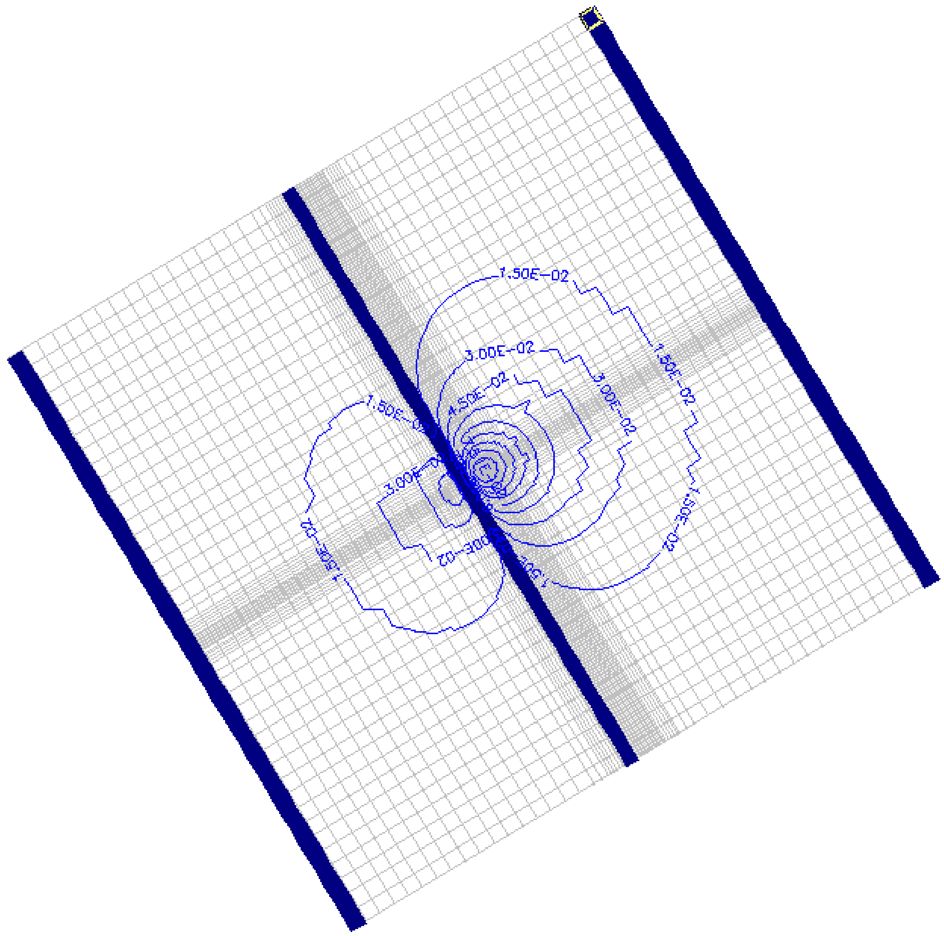

Fig 6.4.1 Drawdown contours (layer3 of model, stress period 8, time step 10)

The maximum drawdown observed for this model was 0.05m, in P-004-1. The drawdown results observed in selected piezometers for this model run are shown in Fig 6.4.2, as expected the drawdown observed increases with an increase in pump rate.

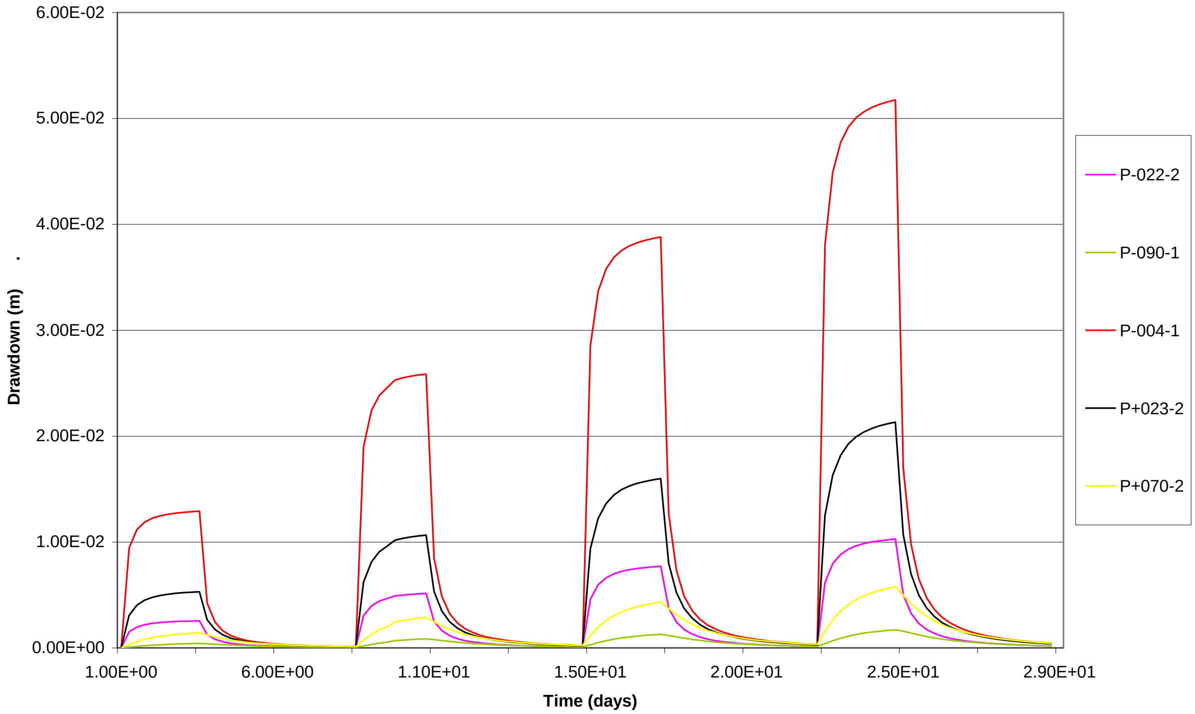

Figure 6.4.2 Drawdown curves for selected piezometers from initial model run, as expected piezometers nearest to the borehole show the highest drawdown.

To calculate the change in hydraulic gradient within the river bed sediments, the drawdown observed in each model piezometer was subtracted from the known groundwater head level measured in the field, to give a new groundwater head level. This was then used in the equation described in section 4.7 to calculate a new hydraulic gradient for that piezometer. A summary of the gradient changes are given in Table 6.4.2.

Pumping at a rate of 200m^3/d will have a significant effect on the hydraulic gradients in the river bed sediments, reducing them by between 2.6% and 48.51%.

Piezometer	Initial Gradient	Gradient when Pump = 200m^3/d	Percentage Change
P+005-1	0.1957	0.148	-24.37%
P+005-2	0.2381	0.2	-16.00%
P+005-3	0.1216	0.0845	-30.51%
P-004-2	0.0806	0.0415	-48.51%
P-004-3	0.1333	0.0955	-28.36%
P-022-2	0.0729	0.0515	-29.36%
P-022-3	0.1833	0.1605	-12.44%
P+023-2	0.0875	0.0675	-22.86%
P-050-1	0.1	0.0903	-9.70%
P+070-1	0.0213	0.0155	-27.23%
P+070-2	0.04	0.034	-15.00%
P-090-2	0.1351	0.1303	-3.55%
P-090-1	0.0566	0.05245	-7.33%
P+110-2	0.0233	0.021	-9.87%
P+110-1	0.0652	0.0635	-2.61%

Table 6.4.2 Summary of hydraulic gradient changes observed in river bed sediments when pumping at 200m^3/d

A second model was run; this time the vertical change in the river bed sediments was considered. The river bed between 0 m and 0.5 m depth was assigned a hydraulic conductivity value of 9.5 m/d; the river bed between 0.5 m and 1m depth was assigned a hydraulic conductivity value of 2.3 m/d. These values were taken from the fieldwork data, as discussed in chapter 5. All other hydraulic parameters remained the same.

Using these values for the river bed sediments reduces the drawdown observed in the piezometers, to 0.048 m in P-004-1, a 4% reduction in comparison to the previous model (see Fig 6.4.3). The most significant change is in the hydraulic gradients. Table 6.4.3 summarises the changes in hydraulic gradient for this model run. The change in hydraulic gradient for this model run is much less than in the initial model run, ranging between 2.07% and 38.59%.

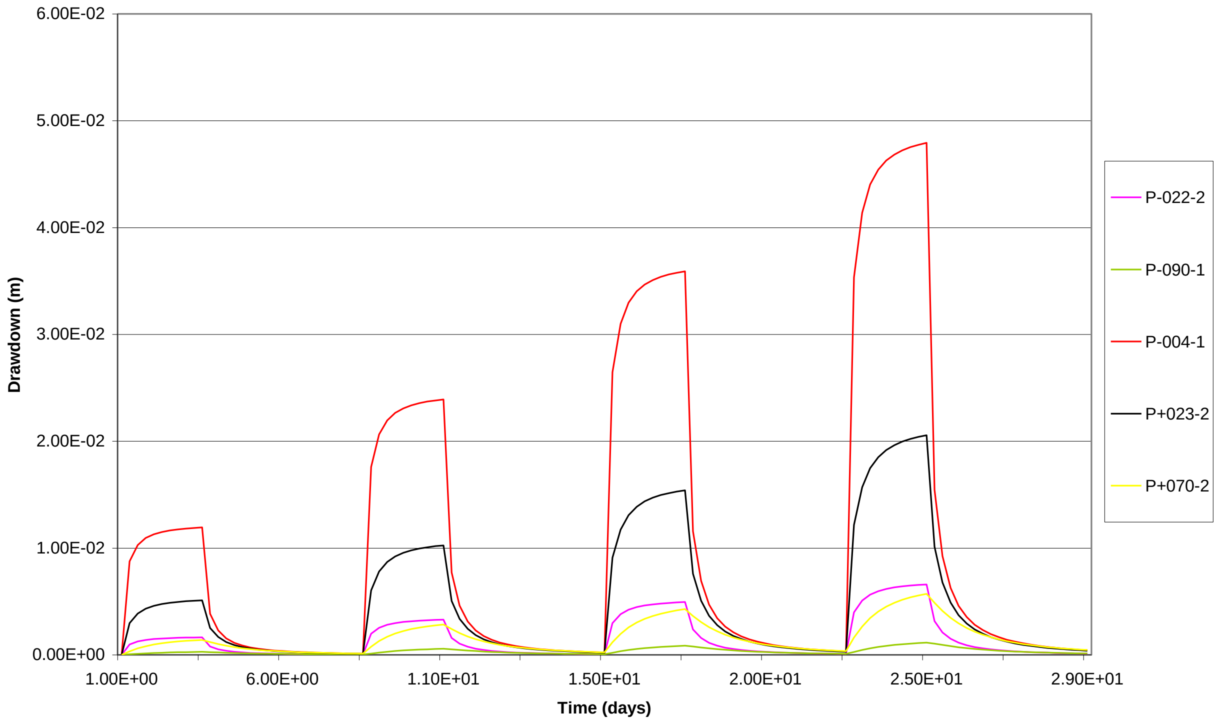

Figure 6.4.3 Drawdown curves for selected piezometers from second model run (the two river bed layers have been assigned different values) as expected piezometers nearest to the borehole show the highest drawdown.

Piezometer	Initial Gradient	Gradient when Pump = 200m^3/d	Percentage Change
P+005-1	0.1957	0.1575	-19.52%
P+005-2	0.2381	0.208	-12.64%
P+005-3	0.1216	0.09	-25.99%
P-004-2	0.0806	0.0495	-38.59%
P-004-3	0.1333	0.101	-24.23%
P-022-2	0.0729	0.0545	-25.24%
P-022-3	0.1833	0.164	-10.53%
P+023-2	0.0875	0.0715	-18.29%
P-050-1	0.1	0.092	-8.00%
P+070-1	0.0213	0.0165	-22.54%
P+070-2	0.04	0.035	-12.50%
P-090-2	0.1351	0.1323	-2.07%
P-090-1	0.0566	0.0531	-6.18%
P+110-2	0.0233	0.02145	-7.94%
P+110-1	0.0652	0.0636	-2.45%

Table 6.4.3 Summary of hydraulic gradient changes observed in river bed sediments when pumping at 200m^3/d during second model run

6.5 Use of Model As A Predictive Tool

The two initial model runs indicate that the model could be used to give an estimate of the hydraulic gradient change that may be observed in the river bed sediments. The change in gradient could be significant, by up to 50% in the piezometers closest to the borehole. However as has already been shown the observed change in hydraulic gradient can vary significantly depending on the value of hydraulic conductivity assigned to the river bed sediments in the model.

It is probably more realistic to use the second model as a predictive indicator of gradient changes as this takes into account some of the variability which exists in the river bed sediments. However, as has already been discussed, there is a significant degree of heterogeneity within the river bed sediments, and this may need to be represented to a greater degree to give a better indication of the expected gradient changes.

This has not been done as part of the present project due to a limited amount of data and time being available. Defining the extent of the heterogeneities within the river bed will be 'easier' to do when there is actual pump test data available to calibrate the model. The extent of the heterogeneities could be defined by comparing observed piezometer drawdown data with the calculated drawdown in this model to assess where differences lie.

6.6 Sensitivity Analysis

A series of tests were carried out to assess the sensitivity of the model to different hydraulic parameters, in particular for the hydraulic conductivity values of the layers of glacio-fluvial and river terrace deposits, the exact values of which are uncertain.

The hydraulic conductivity for each zone was analyzed by changing one parameter at a time and assessing the effect the change had on the drawdown in each piezometer. Some of the zones for hydraulic conductivity were highly sensitive relative to other zones.

The analysis found that the model was very sensitive to changes in conductivity in layers 1 to 3, in particular the river bed sediments. Decreasing the hydraulic conductivity by a factor of two was found to increase the drawdown by a factor of two; the layer 2 river bed sediments in particular had a big effect on the drawdown.

The model was slightly sensitive to changes in the hydraulic conductivity of layers 4 to 21 (glacio-fluvial and river terrace deposits), with the upper layers (4 to17) having the biggest effect. This indicates the need to have a good representative hydraulic conductivity value for the superficial deposits, and is something which may need investigating further.

The model was generally insensitive to change in the specific yield, with only layer 1 showing any sensitivity

6.7 Conclusions

The modelling results have indicated that pumping at a discharge rate of $200m^3/d$ from the borehole at the study site will significantly influence the hydraulic gradients observed within the river bed sediments. A gradient reduction of up to 50% can be expected in the piezometers closest to the borehole. At the present time the second model run described above is thought to best represent the system, as it takes into account the vertical change in hydraulic conductivity of the river bed sediments.

The sensitivity analysis indicates that constraining the hydraulic conductivity values of the river bed sediments and upper layers of the superficial drift deposits will have the most effect on the on the drawdowns observed in the model.

More data is needed to calibrate the model, which will be gained when the pump tests are carried out. This will help make the model more accurate, and in particular help constrain the heterogeneities within the river bed sediments and drift deposits.

Chapter 7. Conclusions

7.1 Summary

The aims of this project were to investigate the hydraulic properties of the aquifer system and examine the effects of a hydraulic test on the hydraulic gradients observed below the river bed. Relevant conclusions from individual areas of investigation have already been presented in chapters 5 and 6. This section concludes generally on the aims and objectives of the study.

A comprehensive network of piezometers has been installed at various depths and locations within the river bed which have given a good set of data results for this project and can continue to be used as part of the future research on site. Falling head tests and groundwater head measurements carried out on the piezometers indicated the varied nature of the river bed sediments hydraulic properties. The hydraulic conductivity (2 m/d) calculated from the pump test on the Kidderminster Sandstone, correlated well with previous pump test data, indicating that a single-well pump test can provide valuable data.

A basic groundwater flow model of the study area was completed. Simulations of hydraulic tests indicated that, significant changes to the natural hydraulic gradients below the river bed could be achieved (up to 50%). It is recommended that a pump rate of 200m^3/d could be used for future tests as this will cause a significant degree of mixing within the hyporheic zone. More data is required to calibrate the model and assess the importance of heterogeneities within the river bed sediments and superficial drift deposits.

7.2 Future Work

A number of recommendations are made for future work on the site, in particular to help improve upon the understanding of the hydraulic properties developed in this project:

- A repeat of all falling head tests on the piezometers is required in order to assess if there is any variability in hydraulic conductivity due to river bed sediments having more time to settle around the piezometers following the disturbance caused during their installation.
- Install new, longer piezometers (or find a better method for extending the tubes of the current piezometers) at P+070-1, P+070-2 and P+110-2, in order to conduct successful falling head tests at these locations.
- Install a new piezometer next to P-022-1 to assess if the difficulty in conducting a successful falling head test was due to a very low permeability horizon or due to damage to the piezometers open section.
- Carry out the pump tests detailed in chapter 6, recording data from all piezometers and the pumping borehole, to give a detailed data set which can be used for more calculations of the systems hydraulic properties.
- Use pump test data to fully calibrate the current groundwater flow model and assess the extent and importance of heterogeneities within the aquifer and river bed sediments.
- Use the newly calibrated model to investigate the changes in hydraulic gradients below the river bed as part of the plan for any future experiments carried out on the site.

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