Complementary use of tracer and pumping tests to characterize a heterogeneous channelized aquifer system in New Zealand

R. L. Dann · M. E. Close · L. Pang · M. J. Flintoft · R. P. Hector

Abstract The combined use of pumping and tracer test data enabled the derivation of equivalent average hydraulic conductivities (K_{avg}) for each test in a heterogeneous channelized alluvial aquifer, whereas K values of the preferential flow paths were two orders of magnitude higher. Greater and earlier drawdown was generally observed along preferential flow lines in a pumping test, within an array of 21 wells. The study aim was to characterize hydraulic properties of a channelized aquifer system in New Zealand by combining tracer and pumping test data. Estimates were able to be made of the percentage of highly permeable channels within the profile (~1.2%), effective porosity that reflected the maximum fraction of highly permeable channels within the aquifer ($\phi_{eff\text{–}pc}$ ~ 0.0038), and flows through highly permeable channels (~98%) and the sandy gravel matrix material (~2%). Using $\phi_{eff\text{–}pc}$, a tracer test K_{avg} value (~93 m/day) was estimated that was equivalent to pumping test values (~100 m/day), but two orders of magnitude smaller than K calculated solely from transport through permeable channels (K_{pc} ~8,400 m/day). Derived K values of permeable and matrix material were similar to values derived from grain size distribution using the Kozeny-Carman equation.

Résumé L'utilisation combinée de données obtenues par pompage d'essai et par traçage a permis de déduire des conductivités hydrauliques moyennes (K_{avg}) pour chacun des tests dans un aquifère alluvial hétérogène parcouru de chenaux, alors que les valeurs de K des lignes d'écoulement préférentiel étaient plus élevées de deux ordres de grandeur. Un rabattement plus important et antérieur a été observé en général le long de lignes d'écoulement préférentiel lors d'un pompage d'essai, au sein d'un dispositif de 21 puits. Le but de l'étude était de caractériser les propriétés hydrauliques d'un système aquifère parcouru de chenaux, en Nouvelle Zélande, en combinant des donnés de traçage et de pompage d'essai. Des estimations ont pu être faites du pourcentage de chenaux très perméables dans la section (~1.2%), de la porosité efficace qui reflétait la fraction maximum de chenaux très perméables au sein de l'aquifère ($\phi_{off\text{–}pc}$ ~ 0.0038), et des écoulements à travers des chenaux très perméables (~98%) et du matériau de la matrice de gravier sableux (~2%). En utilisant $\phi_{off\text{–}pc}$, une valeur de K_{avg} des tests de traçage (~93 m/jour) a été estimée qui était équivalente aux valeurs des pompages d'essai (~100 m/jour), mais deux ordres de grandeur plus petite que K calculé uniquement à partir du transit à travers des chenaux perméables (K_{pc} ~8.400 m/jour). Des valeurs de K déduites du matériau perméable et de la matrice étaient semblables aux valeurs obtenues à partir de la distribution de la taille des grains en utilisant l'équation de Kozeny-Carman.

Resumen El uso combinado de ensayos de bombeo y trazadores permitió la estimación de las conductividades hidráulicas promedio (K_{prom}) para cada ensayo en un acuífero heterogéneo canalizado, en tanto que los valores de K de las vías preferenciales de flujo resultaron dos órdenes de magnitud mayores. En una batería de 21 pozos, se observaron descensos mayores y más rápidos en ensayos de bombeo a lo largo de líneas preferenciales de flujo. El propósito del estudio fue la caracterización de las propiedades hidráulicas de un sistema acuífero canalizado en Nueza Zelandia, mediante la combinación de datos de ensayos de bombeo y trazadores. Fue posible estimar el porcentaje de canales de alta permeabilidad en el perfil (~1.2%), la porosidad efectiva que refleja la fracción máxima de canales de alta permeabilidad en el acuífero ($\phi_{ef\text{–}cp}$ ~0.0038), y el flujo a través de los canales de alta permeabilidad (~98%) y la matriz de grava y arena (~2%). Utilizando $\phi_{ef\text{–}cp}$, con un ensayo de trazador se estimó la K_{prom} (~93 m/día) que resultó equivalente a la obtenida lcon ensayos de bombeo (~100 m/día), aunque dos órdenes de magnitud menor que la K calculada a partir del transporte a través de canales permeables (K_{cp} ~8,400 m/día). Las estimaciones de la K de la matriz y del material permeable

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son similares a los valores derivados de la distribución del tamaño de granos usando la ecuación de Kozeny-Carman.

Keywords New Zealand · Heterogeneity · Pumping test · Tracer tests · Channelized aquifer

Introduction

An understanding of the hydraulic properties and flow geometry of a heterogeneous aquifer system is vital for predicting the transport of contaminants. Information on aquifer characteristics is often scarce and when present, is obtained from a variety of methods including well logs, core sample analysis, outcrop analysis, slug tests, pumping tests and tracer tests. Natural gradient tracer tests are the most effective method to obtain direct information on contaminant transport. However, the high cost associated with well installation, monitoring, and sample analysis limit their wide and frequent use. In the absence of other data, hydraulic properties such as transmissivity (*T*) and storativity (*S*) derived from interpreting pumping test data are commonly used to obtain 'average' values of an aquifer.

Several authors have compared the results of tracer and pumping tests, some showing greater hydraulic conductivity (*K*) values derived from pump tests than tracer tests (Niemann and Rovey 2000; Rovey and Niemann 2005), whilst others have shown the opposite (Thorbjarnarson et al. 1998). In a study of stratified aquifer material, Thorbjarnarson et al. (1998) observed tracers following high conductivity strata and noted that using an average *K* value from an aquifer pump test to predict solute transport would result in considerable underestimation of transport distances for a given time period. In glacial outwash sandy material, Niemann and Rovey (2000) observed that aquifer *K* values measured from conservative tracer tests were approximately 10–20 times smaller than the average pumping test values. The authors suggested that in tracer tests, dispersion may have prevented solutes from flowing exclusively within high conductivity pathways, which strongly affected the *K* value derived from pumping tests, and therefore, using *K* values interpreted from pumping tests in transport studies would over estimate transport distances.

The traditionally used analytical solutions for interpreting pumping test data do not take into account aquifer heterogeneities. Through the use of numerical solutions, Wu et al. (2005) questioned the applicability of traditional analysis of pumping tests in heterogeneous formations. They found that at early times *T* and *S* vary with time, whilst at late times the estimated *T* approached the effective *T* (i.e. T calculated from two-dimensional parallel flow conditions). Sanchez-Vila et al. (1999) used analytical simulations of the Jacob's method (Cooper and Jacob 1946) to assess its value in a simulated heterogeneous field. They reported that even in a heterogeneous field, estimates of *T* and *S* provided useful information about the aquifer. They found that estimates of *T* for different observation points converged to a single value, which corresponds to the effective *T*, and the geometric mean of the estimated *S* value could be used as an estimator of actual *S*. Meier et al. (1998) carried out a similar study of Jacob's method using numerical methods and found similar results with late time *T* values being quite similar across the field and also similar to the effective *T* value.

Investigations in channelized systems have been carried out in some studies. Feehley et al. (2000) used a dual domain mass transfer approach, combined with a fractal geostatistical method to model solute transport in a channelized aquifer. They reported that solutes were being transported along preferential flow pathways smaller than could be practically represented in flow and transport models. In a study of pumping tests in heterogeneous aquifers, Trahan et al. (2002) and Weissmann et al. (2004) observed a cone of depression following coarse-grained deposits (channels) in an alluvial aquifer, suggesting that water was extracted predominately from this unit. Other researchers have used a variety of facies and geostatistical methods to describe and model transport in heterogeneous alluvial formations (Anderson 1989; Fogg et al. 1998; Ritzi et al. 1994; Webb and Anderson 1996; Weissmann and Fogg 1999).

This report focuses on the complementary use of data from pumping and tracer tests to improve understanding of the hydraulic properties of a highly channelized gravel aquifer. The aim was to derive values of hydraulic conductivity and effective porosity, and to assess the areal response of the highly channelized aquifer system to pumping. Fractions of high conductivity channels were estimated in addition to estimates of flow through both high and low conductivity material.

Methods and materials

Field site

The study site is located in Burnham, Canterbury, on the South Island of New Zealand (Fig. 1). The site is part of the Central Plains aquifer system, which is in a geological formation of well-sorted, little-weathered fluvioglacial outwash gravels with varying proportions of sand, silt and clay. The gravels are derived from indurated greywacke sandstone, and they extend up to 125 m in depth according to nearby well logs. The highest permeabilities tend to occur parallel with the depositional surface in a downstream direction, and the lowest permeabilities occur vertically. Groundwater tends to flow through preferred routes representing a multitude of buried, permeable interconnected and meandering river channels (NCCB and RWB 1983). In his assessment of Canterbury Plains aquifers, Davey (2006) indicated that significant flows of groundwater are confined to distinct permeable lenses situated within a matrix-rich gravel of low permeability. Bore log information noted in this study indicates that the profile consisted of sandy gravel; clay bound gravels and gravely sand with brown clay. Unfortunately bore logs do not identify the location or extent of the more permeable

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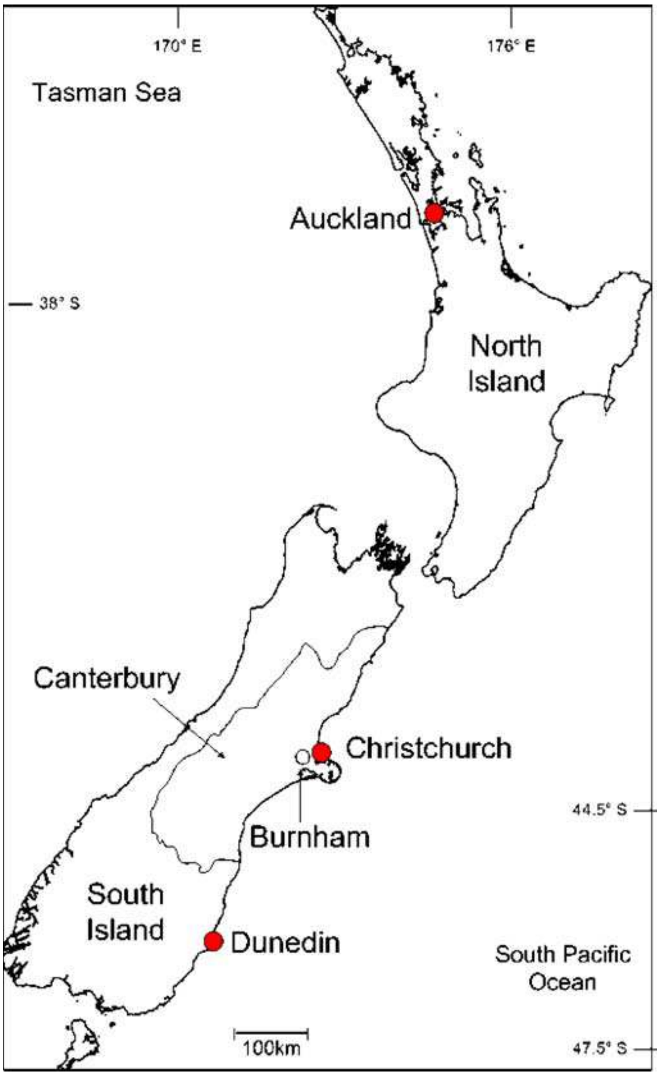

Fig. 1 Location of the Burnham field site, near Christchurch, South Island of New Zealand

channels or ribbons that are noted in the 1983 report and in Davey (2006). This is due to the general narrowness of the highly permeable material and the inability of most drilling methods to be able to distinguish changes in gravels over small intervals.

The highly permeable channels noted above consist of openwork gravels, which are well-sorted rounded pebbles, often coated with iron and manganese oxides and organic matter (Fig. 2). The permeable channels vary in geometry and horizontal and vertical orientation. In their study of the characteristics of the Canterbury gravels, Moreton et al. (2002) indicate that the highest permeability material is associated with secondary channel fill deposits. This facies had a mean grain size of 64 mm and contained significant openwork gravels, with extremely high permeability. Ashworth et al. (1999) described the Canterbury gravel profile using a number of different fine and coarse-grained facies including bar cores, which frequently comprise clean openwork gravels. In their work on gravel formations in Switzerland, Jussel et al. (1994) described openworks gravels as having mean porosities of 34.9%, geometric mean hydraulic conductivity of 8,640 m/day, maximum lens height of 0.25 m and comprising 1.9% of the overall profile. Whilst the work of Ashworth et al. (1999) and Moreton et al. (2002) indicate a variety of different facies (pure sand lenses and mixed sand, silt and gravel facies), most of which is sandy gravel (matrix material), the contrast in permeabilities between the openwork gravel (permeable channels) and the remaining facies is such that the aquifer system works essentially as a dual permeability system. This conceptual model is supported by observations in exposed vertical profiles in Canterbury where water literally flows out of permeable channels whilst slowly seeping out of the surrounding matrix material.

The regional groundwater flow boundaries for the unconfined system are all further than 20 km from the study site (major rivers to the north and south, the sea to the east, and a mountain watershed to the west). The groundwater level generally fluctuates seasonally over about 4–8 m, with highs in September and lows in April. During this study, groundwater levels fluctuated from about 21 m below ground level (bgl) to about 13 m bgl. The groundwater flow direction in the area is generally south-easterly toward the ocean with an average hydraulic gradient of about 0.003 (NCCB and RWB 1983). The Burnham site has been used for a number of contaminant transport studies (Pang et al. 2005; Pang and Close 1999b; Sinton et al. 2000) and has 22 wells with one injection well (well 0) and 21 observation wells (wells 1–21) situated in four arrays at approximately 20, 40, 65, and 90 m down gradient from the injection well (Fig. 3). Each observation well (including the injection well) has a depth of 18 m and a diameter of 100 mm and has slotted PVC screen from 12 to 18 m (screen porosity approximately 20%). An additional larger well (pumping well) was

Fig. 2 Image of highly permeable channel material located within the Canterbury sandy gravels matrix; picture taken at a nearby site, from Davey (2006)

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installed within the array for the purpose of the pumping test; it is 24 m deep with perforations from 12 to 23.5 m (Fig. 4).

Pumping tests

The first pumping test was carried out when water levels were at a historical low for the site, at about 17 m bgl (September 2005), and as such the base of the monitoring wells (screened from 12–18 m bgl) were only approximately 1 m below the water table (Fig. 4). The screened interval of the pumping well was from 23.5 m bgl to the water table (~6.5 m of saturated screen), and as such, most of the saturated pumping well screen length was situated below the base of the observation well screens. Whilst conditions were not optimal, this test was run to enable an initial comparison of hydraulic properties with tracer tests while the wells were still viable. Groundwater was abstracted from the pumping well at a rate of approximately 730 L/min for 41 hrs (test stopped due to pump failure). During this period water levels were recorded in 13 of the monitoring wells. In addition, water levels were

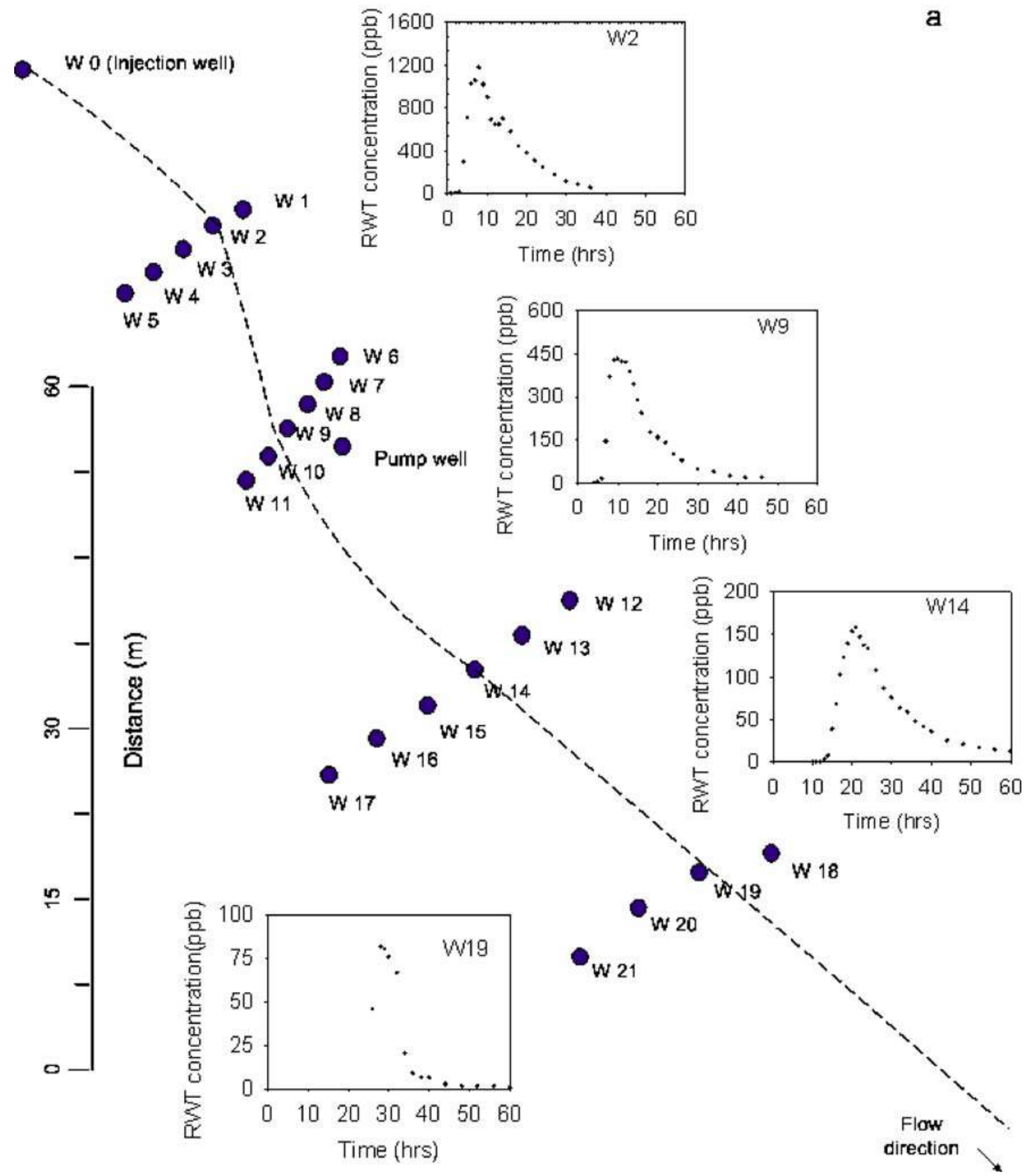

Fig. 3 Plan view of the Burnham experimental site showing the preferential flow paths and rhodamine WT (*RWT*) breakthrough curves derived from **a** the 1995 tracer tests (Pang et al. 1998) and **b** the 2006 tracer tests. *ppb* parts per billion

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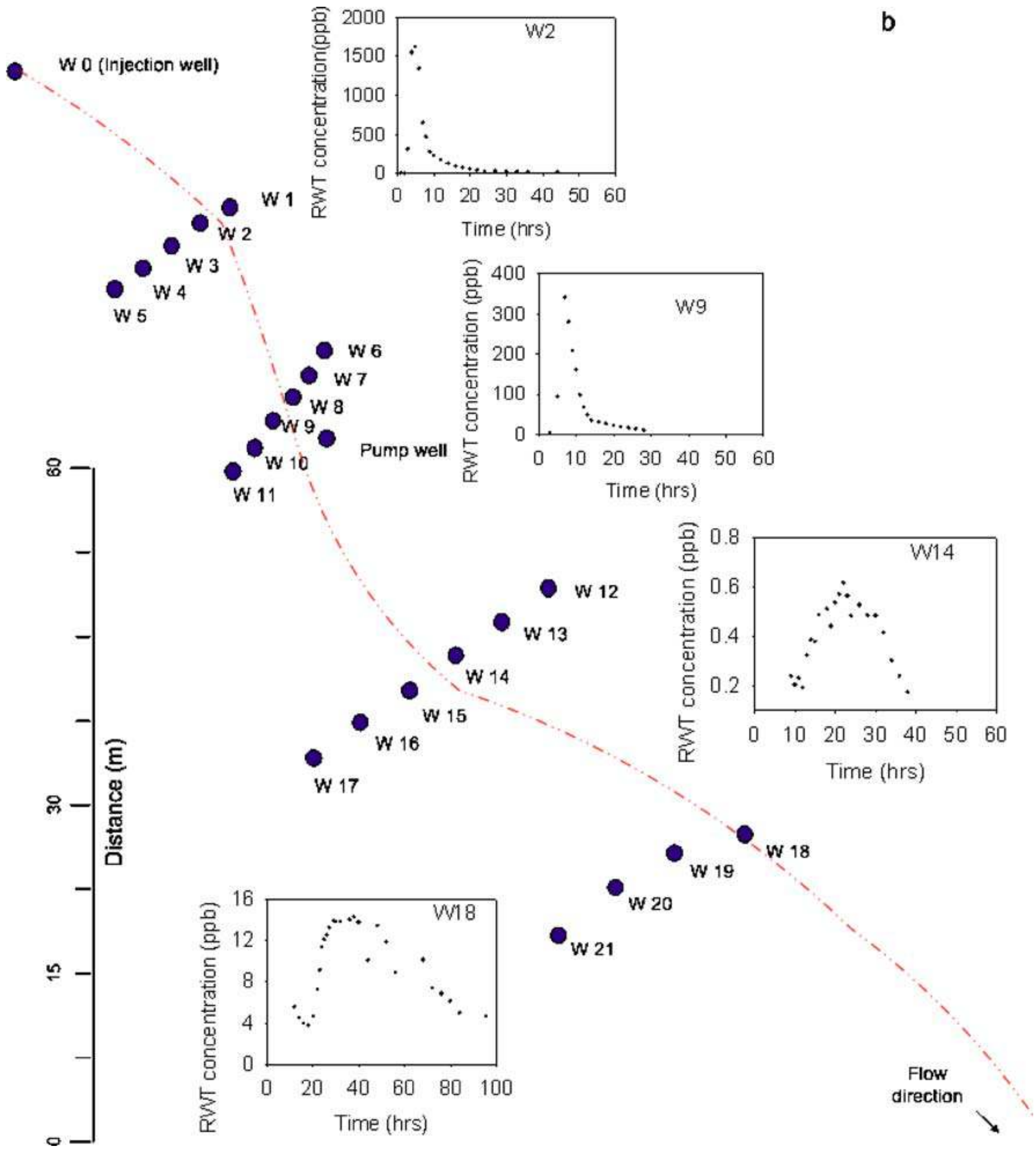

Fig. 3 (continued)

also monitored in the pumping well and at a 'control' well located approximately 900 m to the north-east of the pumping well. Three of the monitoring wells, the pumping well and the 'control' well were instrumented with pressure transducers and data loggers to automatically record water levels. Water levels were recorded in the control well for approximately 24 hrs prior and 36 hrs after the pumping test was carried out. Barometric pressure was also recorded in the nearby 'control' well prior to, during, and after the pump test. Water level in well 0 before the start of the first pumping test was 17.48 m bgl.

The second pumping test was undertaken (October 2006) when water levels had risen to the level of previous tracer tests (approximately 13 m bgl). For this test (48-hr duration at approximately 200 L/min), the pumping well was modified (backfilled with sand and sealed with bentonite at 18 m bgl) to be screened over the same portion of the aquifer as the monitoring wells (Fig. 4). In this test, all 21 wells within the array were monitored. The pumping well was pumped from a 5.25 m saturated section from 12.75 to 18 m bgl. Background water levels and barometric pressure were recorded adjacent to the site prior to, throughout and after the pumping test.

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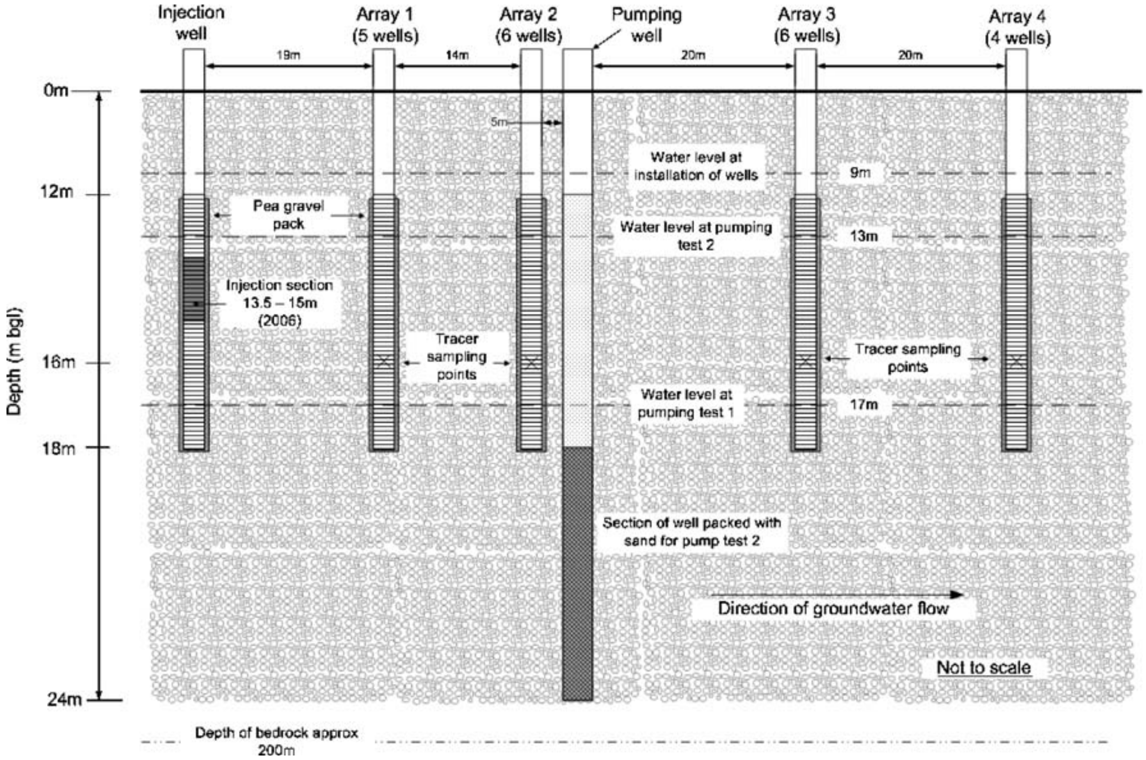

Fig. 4 Cross section schematic of the Burnham experimental site showing the well installation and water levels for each pumping test

Tracer tests

A tracer test was carried out in September 2006, approximately 2 weeks prior to the second pumping test, so as to be able to compare the results of the second pumping test and a tracer test in the same portion of the aquifer. This tracer experiment was conducted when the water level in well 0 was at 10.91 m bgl. Twenty litres of solution containing 1,600 mg/L of rhodamine WT (RWT) dye and 6,200 mg/L of bromide was injected into well 0 over the interval 13.5–15 m bgl. The injection took 2 min to complete and was followed by 4 L of fresh water to flush out residual tracer in the system. Seventeen out of the 21 downgradient wells were sampled for a period of up to 95.5 hrs after the injection. Sampling depths were selected following a preliminary tracer test using RWT and amberlite XAD-7 resin bags (Close et al. 2002) to determine the location of preferential flow paths. Samples were analysed for rhodamine on a Shimadzu RF1501 spectrofluorimeter and for bromide using a bromide selective electrode (Thermo-Orion 9635BN).

Tracer tests carried out in previous studies

A number of natural gradient tracer tests were undertaken at the Burnham site between 1995 and 1997 in conjunction with bacterial and trace metal contaminant transport studies (Close and Pang 1995; Pang et al. 2005; Pang and Close 1999a; Pang et al. 1998). Each experiment provided similar results regarding derived hydraulic conductivity values. Only a short summary of methods and results (in the results section) from one of these experiments is presented here, as a full description of methods and results can be found in these papers.

A tracer experiment was undertaken at the Burnham site in May 1995 (Pang et al. 1998). Water level in well 0 was 13.4 m bgl at the time of the tracer test. Forty litres of a solution containing rhodamine WT, chloride, and 2.9 L of a solution containing *Bacillus subtilis* was injected over an interval of 0.1–1.6 m below the water table (13.5–15 m bgl). Injection was completed in 13 min. Following tracer injection, groundwater was sampled at 20 downgradient observation wells over a period of 100 hrs. Sampling depth at each well was selected following a preliminary flow path tracer test, which used rhodamine WT and resin bags to identify preferential flow paths.

The breakthrough curves (BTCs) from the tracer tests (Fig. 3) were analysed via a curve fitting method using the AT123D model from Yeh (1981) and the PEST optimisa-

Table 1 Parameters derived from the delayed yield (Boulton 1973) analysis of Burnham pumping test 1 (September 2005)

Observation well	Radius (m)	Transmissivity (m^2/day)
0	41.3	331
5	22.9	460
9	5.5	260
10	7	216
16	23.9	432
20	45	547
	Mean	374
	SD	127

SD standard deviation

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Table 2 Parameters derived from Cooper and Jacob (1946) analysis of Burnham pumping test 2 (October 2006)

Observation well	Transmissivity (m^2/day)	t (mins) for u<0.01
0	512	17.2
1	569	5.1
2	603	3.8
3	569	2.3
4	683	2.7
5	603	14.8
6	625	0.1
7	569	0.3
8	569	0.2
9	603	0.1
10	488	16.3
11	512	16.2
12	569	4.5
13	569	9.1
14	512	12.7
15	670	8.2
16	683	16.0
17	641	32.0
18	641	6.8
19	512	35.5
20	666	22.8
Average	589	
SD	61	

SD standard deviation

tion package (Doherty et al. 1994). The input parameters used in the modelling included an effective porosity (ϕ_{eff}) of 0.2; a bulk density of 2.12 g/cm^3; and hydraulic gradient of 0.0012. Hydraulic conductivity and dispersivity were optimised.

Particle size distribution methods

Hydraulic conductivity of sediments can be estimated from a combination of particle size distribution (PSD) and porosity data using the Kozeny-Carman equation (Carman 1937; Carman 1938);

$$K = \left\{\frac{\phi^3}{(1-\phi^2)}\right\}\left\{\left(\frac{g}{\upsilon K_{cc} s_s^2}\right)\right\} \tag{1}$$

where K is hydraulic conductivity, ϕ is the porosity, g is gravity acceleration (9.81 m/s^2), υ is the fluid kinematic viscosity, K_{cc} is the Kozeny-Carman coefficient (value of 5 used for this study) which is a product of tortuosity and a shape factor, s_s is the specific surface area. The specific surface area was calculated from the combined sediment fractions in each sample following Louden (1952):

$$s_s = \Lambda\left\{\sum_{i=1}^{n} X_i s_{si}\right\} \tag{2}$$

$$s_{si} = 6/D_{mi} \tag{3}$$

where Λ is a dimensionless coefficient ranging from 1.1 (for rounded grains) to 1.4 (for angular grains), X_i to X_n are the fraction of sediment within each size grouping, s_{si} is the specific surface area of fraction i, and D_{mi} is the mean diameter of the grains in fraction i.

Porosity values were obtained for the matrix and permeable channel material using volume replacement methods (unpublished data). These methods use either water (for horizontal surfaces) or expanding foam (for vertical surfaces such as cliff faces) to obtain a volume estimate after the extraction of a known mass of sediment material, from which porosity can be calculated (assuming particle density 2.65 g/cm^3). Where measured porosity values were not available, the porosity of matrix material was estimated from particle size empirical equations, as given by Urish (1981), which were optimised using measured porosity. The equations derive maximum and minimum porosity values ϕ_{max} and ϕ_{min}, from which an average value ($\frac{\phi_{max}+\phi_{min}}{2}$) can be calculated:

$$\log \phi_{max} = 1.62563 - 0.08653 \log D_{50} - 0.03636 \log (D_{90}/D_{10}) \tag{4}$$

$$\log \phi_{min} = 1.53902 - 0.18986 \log D_{50} - 0.08201 \log (D_{90}/D_{10}) \tag{5}$$

These equations are derived from repacked unsorted natural glacial sand samples with porosities determined from the loosest state attainable (ϕ_{max}) and the densest state (ϕ_{min}) (Urish 1981). Equation 4 was optimised using the mean measured porosity values and PSD values from

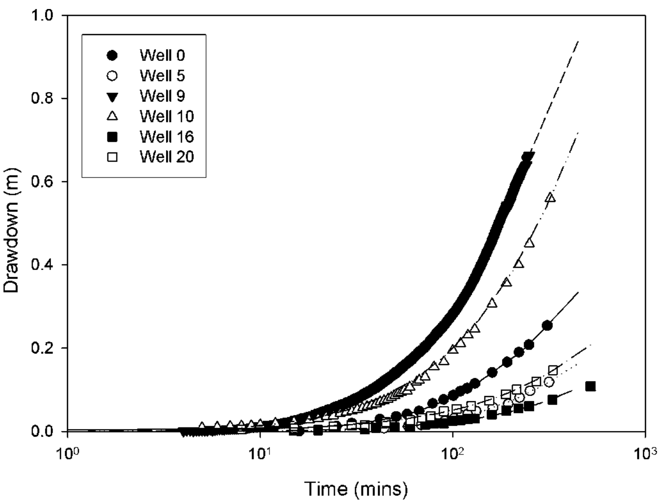

Fig. 5 Pumping test 1 measured drawdown (*symbols*) and calculated drawdown (*lines*) for six observation wells, reflecting a delayed yield response

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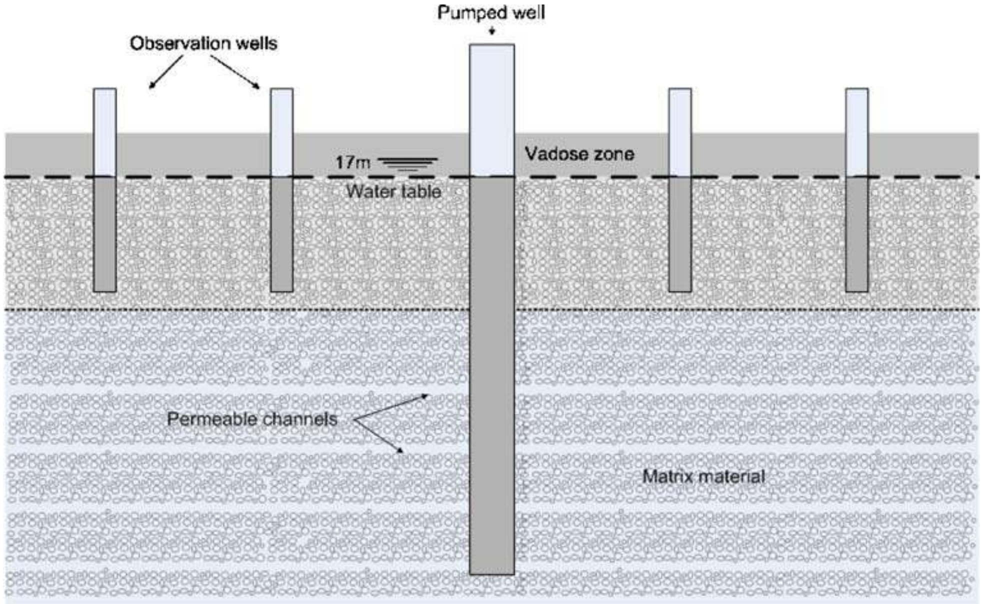

Fig. 6 Cross section schematic of the interpreted geology and well set-up in pumping test 1

the measured samples to provide an optimised maximum porosity value for gravel matrix:

$$\log \phi_{\max-\mathrm{opt}} = 1.62085 - 0.07631 \log D_{50} - 0.02818 \log \left(D_{90}/D_{10}\right) \quad (6)$$

Results

Pumping tests

Results of the pumping test analysis are summarized in Tables 1 and 2. Terms used in the analysis of the drawdown data include transmissivity (T), which is equal to the product of the average hydraulic conductivity (K_{avg}) and saturated aquifer thickness (B); the storage co-efficient (S), the radius (r), time (t), and hydraulic conductivity of the aquifer (K).

Pumping test 1

Of the 13 observation wells monitored during the first pumping test, six wells provided drawdown data consistent with a delayed yield response to pumping (Fig. 5). The remaining wells (drawdown data not shown) either showed little evidence of drawdown suggesting a possible lack of connection with the pumped section of the aquifer, or highly variable drawdown suggesting either interference or clogging around the well and/or variable drainage at the water table. Interpretation of drawdown curves suggest that for pumping test 1, most wells are either screened in a matrix material of lower hydraulic conductivity overlying (or perhaps adjacent to) a portion of the aquifer with channels of significantly higher hydraulic conductivity from which the pump is extracting most of its water (Fig. 6). The Boulton (1973) solution (solution Eta_3 within the Hunt (2007) Function.xls framework) which assumes a pumping well within an aquifer, bounded on top with an aquitard containing a free water table, within which the monitoring well is screened, was used to analyze pumping test 1 drawdown data.

The average of the calculated transmissivities was 374 m^2/day, with a standard deviation of 127 m^2/day. Deriving a K_{avg} value from T is problematic in that B is unknown in this environment. From the solution used, the top of the 'aquifer' must be below the base of the wells, therefore the maximum value of B is 5.5 m. Using this value the K_{avg} value derived from the mean of the transmissivities (K_{avg}=T/B) would be about 68 m/day. No correlation was found between distance and drawdown. Wells further from the pump well often had greater drawdown than wells close to the pump well indicating that these wells had better hydraulic connection to the pump well.

Burnham pumping test 2

The aquifer response to the second pumping test was vastly different to the first pumping test. The drawdown response to pumping was more consistent, while still reflecting the heterogeneous nature of the aquifer. This variable response is shown in semi-logarithmic plots in Fig. 7 where wells along the main flow line (circles) plot separately (due to a trend of earlier and greater drawdown) from the observation wells located off the main flow line (squares). Due to the variable drawdown response, the data were fitted both collectively to a global type curve using an unconfined solution (Moench 1997; Fig. 8) and individually using the single well Cooper and Jacob

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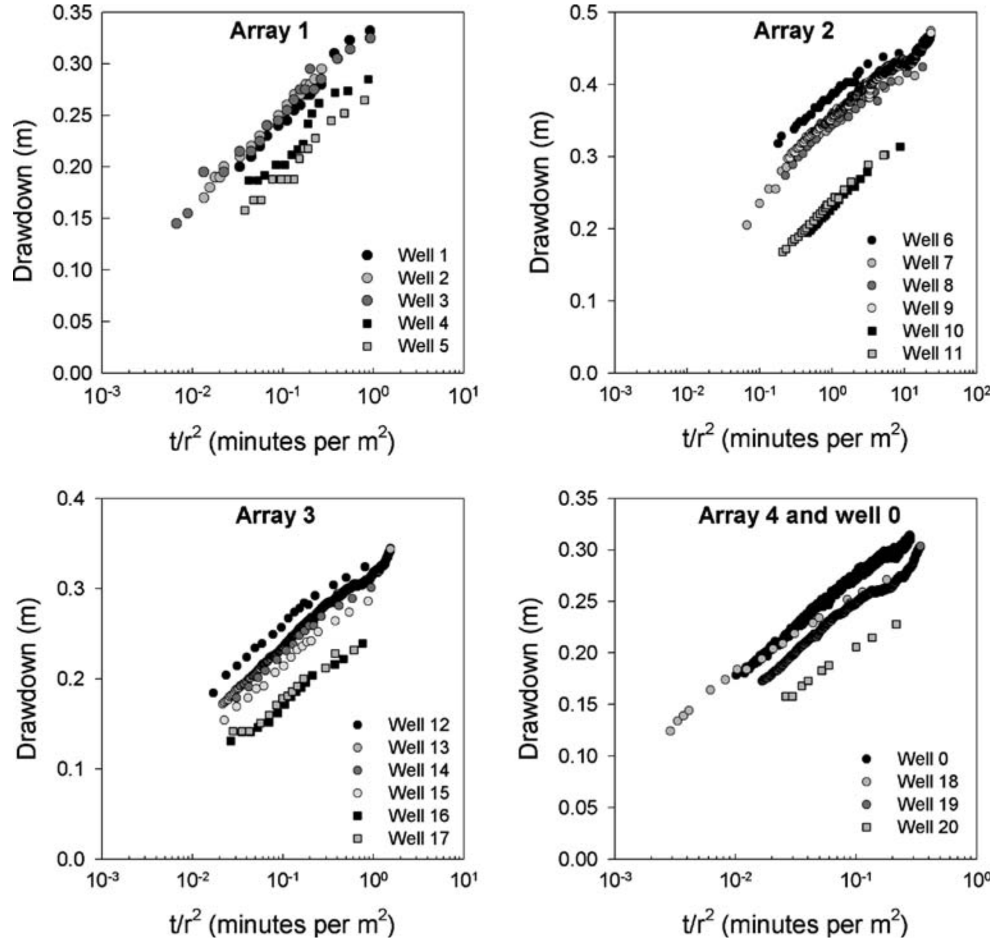

Fig. 7 Semilogarithmic plots for each well array from pumping test 2 showing the different response of wells located along the main flow line (*shaded circles*) and those located off the main flow line (*shaded squares*)

(1946) solution (Table 2). For the Moench type curve solution the transmissivity is calculated from:

$$T = \frac{Q}{4\pi}\left(\frac{h_D}{(h_i - h)}\right) \tag{7}$$

where Q = pumping rate; h_D = dimensionless drawdown at the match point (Fig. 8); and (h_i–h) is the measured drawdown at the match point. Transmissivity is calculated from the Cooper-Jacob method by drawing a straight line through the measured drawdown data points on a semi-logarithmic plot and extending it to the point of zero drawdown on the time axis (t/r_0^2). The drawdown per log cycle (Δh) is then derived from the slope of the line, and transmissivity is calculated from:

$$T = \frac{2.3Q}{4\pi\Delta h} \tag{8}$$

Early-time observation drawdown data points with values of u > 0.01 (a condition of applicability of the Cooper-Jacob method) were omitted from the analysis, where:

$$u = \frac{r^2 S}{4Tt} \tag{9}$$

Unfortunately an unknown external well in the vicinity impacted the late-time data. Due to this, only early (with u < 0.01) and mid-time data are used in this analysis. The T calculated from the collective type curve analysis is 556 m^2/day, calculated with a B value of 6 m. Due to the wells being screened at the water table, as the water table drops, so too will the B value. Therefore B is likely to be approximately 5.5 m, which would result in a K_{avg} value of about 100 m/day. The average T value from the analysis of 21 single well analyses using the Cooper-Jacob method (Table 2) was 584 m^2/day with a standard deviation of 58 m^2/day. If one again assumes a B of 5.5 m this gives a similar approximate K_{avg} value for the aquifer of about 105 m/day.

Tracer tests

In a previous assessment of the site, Pang et al. (1998) used natural gradient conservative tracer tests to derive

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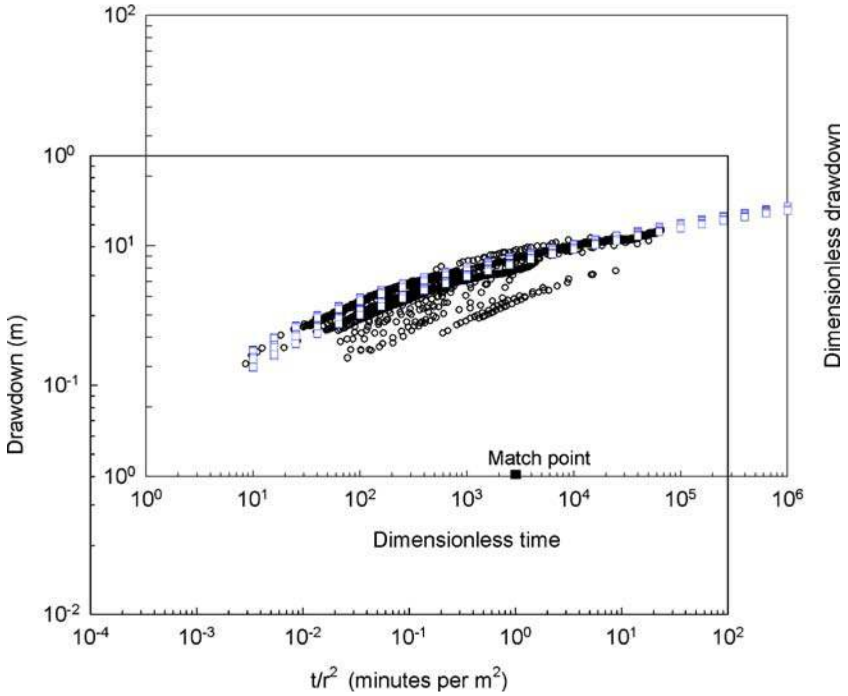

Fig. 8 Composite plot of measured drawdown (*open black circles*) and calculated dimensionless drawdown (*open blue squares*); water table aquifer type curves with B=6 m; XKD (ratio of vertical to horizontal K)=10^{-5}; and σ (ratio of storativity to specific yield)=1.0

transport parameters. They observed a velocity range of 30–85 m/day (median 63 m/day) from 15 observation wells. Though these velocities are relatively high, they are within the range of 0.1–290 m/day reported for the Canterbury Plain (NCCB and RWB 1983). These velocities translate to hydraulic conductivities of the preferential flow paths (i.e. highly permeable channels) at the site ranging from about 5,000 to 14,100 m/day with a mean of 9,720 m/day (geometric mean 9,354 m/day): $V=K*I/n$; with I=0.0012 and n=0.2, based on Pang et al. (1998) results. These K values reflect transport solely through the highly permeable channels, not the bulk of the aquifer; from now on permeable channel K values will be denoted K_{pc}.

The preferential flow path at the site, which was derived from three tracer experiments and presented by Pang et al. (1998), is shown in Fig. 3. It was obtained using a combination of dye tracer experiments using either resin bags or sampling to detect the dye tracer. Pang et al. (1998) noted that preferential flow paths identified from the relative concentrations of dye in resin bags showed significant vertical variations within a given array of wells and between arrays for the three experiments (i.e. not within a given test), though the main flow line derived was similar. Variations were attributed to differences in the injection interval and, perhaps more significantly, changing water levels between experiments. Secondary flow paths were also observed at the site, with a flow path also intersecting well 7. The main flow path is curved, which may reflect the meandering depositional pattern of a braided gravel river (Pang et al. 1998).

The tracer test carried out in this study produced similar results to previous tracer tests, with the main flow path generally following the mid to left-hand side of the array (Fig. 3). The small differences in flow direction may be due to a higher water table in the more recent experiment, providing a greater number of preferential flow paths for the tracers to travel through. Groundwater velocities, generated from peak arrival times along preferential flow line, were similar to the previous test with a range of 54–192 m/day (from ten observation wells), with an average of 90 m/day. This translates, using a mean gradient for this experiment of 0.0024 and an effective porosity of 0.2 as used by Pang et al. (1998), into K_{pc} values ranging from 4,500 to 16,000 m/day with an average of 7,576 m/day (geometric mean 7,018 m/day).

Hydraulic conductivity determined from particle size distribution

Saturated hydraulic conductivity of matrix, sand and highly permeable channel samples was estimated from porosity and particle size distribution data using the Kozeny-Carman equation (Eq. 1) and are presented in Table 3. Porosity values were either measured using volume replacement methods or calculated using the empirical equations of Urish (1981), modified for maximum porosity (see section Particle size distribution methods).

Calculated K values are highly dependant on the porosity value (Table 3). The mean hydraulic conductivity of the measured matrix samples was estimated as 44.6 m/day using volume replacement total porosity values, and as

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Table 3 Particle-size distribution, porosity and hydraulic-conductivity values for matrix, sand and permeable channel material

Parameter	Trench matrix samples (20 samples)			Bucket auger matrix (13 samples)			Permeable channel (10 samples)			Sand lens (3 samples)		
	Average	Min	Max	Average	Min	Max	Average	Min	Max	Average	Min	Max
D_{10} (µm)	361	242	586	1,152	415	2,127	7,674	3,221	12,535	165	156	172
D_{50} (µm)	21,362	13,435	37,500	22,796	16,163	27,903	17,633	8,244	27,959	311	260	347
D_{90} (µm)	56,806	40,685	72,606	72,529	59,412	85,817	36,260	19,000	53,292	538	502	571
(D_{90}/D_{10}) (µm)	171.75	79.44	269.72	82.3	29.0	162.1	5.0	2.6	6.6	3.27	3.22	3.32
Urish average porosity	9.18	8.47	9.79	9.4	9.0	10.0	–	–	–	17.61	17.36	17.98
Porosity (field measured)[c]	0.17	0.13	0.24	17.2	16.8	18.0	33.8[a]	27.0	39.0	26.08[b]	25.84[b]	26.43[b]
Specific surface area (cm^{-1})	60.91	33.85	94.88	77.4	52.0	152.9	19.43	11.99	31.95	243.00	218.58	278.67
K sat (cm/s) - Kozeny-Carman	6.3E-03	1.6E-03	1.7E-02	4.3E-03	8.5E-04	8.3E-03	–	–	–	2.7E-03	2.2E-03	3.1E-03
K sat (m/day)	5.44	1.36	14.66	3.8	0.7	7.2	–	–	–	2.4	1.9	2.7
K sat (cm/s) - Kozeny-Carman[c]	5.2E-02	9.2E-03	1.4E-01	3.2E-02	6.3E-03	5.9E-02	–	–	–	1.1E-02	8.6E-03	1.3E-02
K sat (m/day)[c]	44.60	7.93	117.28	27.3	5.5	50.9	5,406	1,498	10,646	9.6	7.5	11.1

[a] Porosity average measured for four out of ten permeable channel samples

[b] Porosity derived from optimized Urish equation

[c] *K* calculated with field measured or optimized Urish (sand) porosity values

D is the grain diameter; D_{10} is equal to the grain diameter at which 10% of the sample is finer

5.4 m/day using a mean porosity value derived from particle size distribution (Urish 1981). Values obtained from excavated matrix material at a site close by are similar with mean *K* values of 3.8 and 27.3 m/day using unmodified mean porosity and modified maximum porosity respectively. For permeable channel material there is a significant difference between *K* values derived from mean calculated (Urish) porosity (average ϕ=11.02, average *K*=100 m/day) and those derived from measured (volume replacement method) porosities (average ϕ=33.8, average *K*=5,406 m/day). The calculated porosity values for permeable channels are unrealistically low, suggesting that the Urish (1981) method developed with unsorted natural glacial sand samples was not suitable for these large uniform sized sediments. Average calculated hydraulic conductivity values for the pure sand lenses were 2.4 m/day (for ϕ of 17.61) and 9.6 m/day (for ϕ of 26.09) for the average and optimised Urish (1981) methods respectively.

Discussion

The first pumping test, undertaken with approximately 1 m of screen for the observation wells below the water table, observation wells showed highly variable responses to pumping (mainly either delayed or little to no response). Lee and Lee (1999) found similar 'typical' and 'anomalous' responses with various observation wells in fractured rock aquifers, where lower drawdown was related to poor hydraulic connection to the pumping well. The delayed yield solution used in the pumping test analysis is consistent with the conceptual interpretation of the aquifer as a series of high permeability channels or ribbons meandering through a lower permeability matrix. It is likely that the pumping well was withdrawing water mainly from the high permeability channels below the base of the observation wells and responses in the wells, screened in the lower permeability matrix material, were delayed as the matrix water is variably drained at the water table to recharge the permeable channels.

The second pumping test, undertaken with the observation wells screened in the same portion of the aquifer as the pumping well screen, provided a much more consistent drawdown response. Interestingly though, the wells located along the flow line generally recorded more and earlier drawdown than wells located away from the general flow line (Fig. 7). This process is consistent with the findings of Trahan et al. (2002) and Weissmann et al. (2004) who also reported an irregular cone of depression with greater drawdown in zones of coarse-grained deposits in a heterogeneous alluvial aquifer. It appears that after the greater early-time drawdown, the field starts to react more evenly and the gradient of the drawdown curve converges to a similar value, and therefore a similar *T*, for all observation wells. The findings of *T* converging to a similar value across the aquifer are consistent with the observations of Meier et al. (1998) and Sanchez-Vila et al. (1999) in a heterogeneous aquifer. Similarly, using

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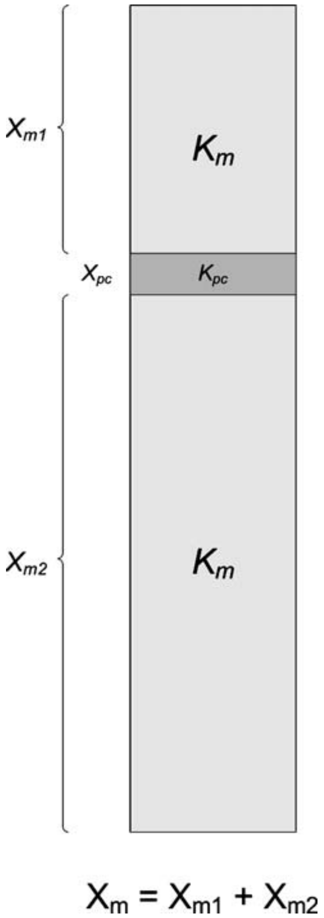

Fig. 9 A schematic of the aquifer profile showing the fraction of matrix material (x_m) with a hydraulic conductivity of K_m and the fraction of permeable channel (x_{pc}) with a hydraulic conductivity of K_{pc}

the Cooper-Jacob solution in simulated heterogeneous fields, Wu et al. (2005) found that T values from later time data are similar throughout the field, whereas the storativity values are more dependant on the location and connectivity of the observation well to the pumping well. The variation in response to pumping that was observed in wells located away from the main flow line, as opposed to those along the flow line, may be reflective of variable storativity values. The combined tracer and pumping test results provide some insight for variable early-time drawdown curve gradients in a channelized aquifer system. It appears that water is preferentially withdrawn from along the high permeability channels connected to the pumping well at early times; at mid- to later times the drawdown is more consistent across the area as the matrix material contributes more to the pumped volume.

The average hydraulic conductivity values ($K_{avg–pt}$) derived from the two pumping tests in this study were substantially lower than the hydraulic conductivity derived for the preferential flow paths from natural gradient conservative tracer tests. Drawdown curves observed from the two pumping tests, undertaken at times of different groundwater levels, were significantly different. This reflects both the heterogeneity at the site and the different pumping well and water-table conditions for each test. Due to the sub-optimal conditions and the associated highly variable observation well drawdown response of the first pumping test, a more reliable estimate of $K_{avg\text{-}pt}$ was obtained from the second pumping test. Focus will be on the results of the second pumping test for the comparison with tracer tests.

A $K_{avg\text{-}pt}$ value from pumping test 2 was derived from dividing the mean estimated transmissivity at the site by an aquifer thickness (B) of 5.5 m. Similar transmissivity values were obtained using both the Moench collective type curve (556 m/day) and the mean of the Cooper-Jacob solution (584 m/day). These result in $K_{avg\text{-}pt}$ values of about 100 and 105 m/day respectively. These values are nearly two orders of magnitude lower than the K_{pc} derived from tracer tests. This observation of K being smaller when derived from pump tests is consistent with the findings of Thorbjarnarson et al. (1998) in their study of a stratified aquifer where the tracer followed higher conductivity layers, but contrasts those of Niemann and Rovey (2000).

Table 4 Fractions of matrix and permeable channel and various estimates of the K of the matrix material and the average K reflecting flow through permeable material

x_{pc}	x_m	K_m	Bulk–$\Phi_{eff\text{-}pc}$	$K_{avg–pc}$
0.008	0.992	35	0.0025	73
0.009	0.991	26	0.0028	82
0.01	0.99	18	0.0032	93
0.011	0.989	10	0.0035	102
0.012	0.988	2	0.0038	111
Parameters used in calculations				
K_{pc}	K permeable channel		8,200	Derived from tracer tests
K_{avg}	K average for profile		100	Calculated from pumping test 2
K_m	K matrix		Variable	
x_m	Fraction of matrix		Variable	
x_p	Fraction of permeable layers = 1–x_m		Variable	
Bulk–$\Phi_{eff–pc}$	Bulk effective porosity		Variable	
$\Phi_{eff–pc}$	Effective porosity permeable channel		0.315	Assumed 90% of total porosity
V	Groundwater velocity		70	Derived from tracer tests
I	Hydraulic gradient		0.0024	Measured from 2006 tracer test

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The K values derived from the 2006 tracer test, ranged from 4,500 to 16,000 m/day (average 7,576 m/day, geometric mean 7,018 m/day), which is very similar to the values reported by Pang et al. (1998) for the preferential flow paths, range 5,064–14,112 m/day (average 9,720 m/day, geometric mean 9,354 m/day). The average of the two geometric mean values is 8,186 m/day, which is similar to the value reported by Jussel et al. (1994) for openwork gravels (geometric mean 8,640 m/day). These K values represent flow through high permeability channels (K_{pc}), not the bulk of the aquifer (K_{avg}).

Used in isolation, it was not possible to derive K values solely from pumping test data that were similar to those observed during tracer tests. Pumping test derived K values are greatly reduced by the lower conductivity matrix material, whereas the tracer tests derived K values are from flow through the preferential flow paths only. By combining tracer and pumping test data, it is possible to estimate the fraction of permeable channels within the overall aquifer and the average hydraulic conductivity of the aquifer. Pumping test data suggest a $K_{avg\text{-}pt}$ value of approximately 100 m/day (average of Moench and Cooper-Jacob methods for pumping test 2) for the screened portion of the aquifer. The tracer test data indicate that there are connected high permeability channels or ribbons with K_{pc} values of approximately 8,200 m/day (rounded average of the two tracer test geometric mean values). However, the K value of the lower permeability matrix and the fraction of matrix and permeable channel material in the profile are unknown (Fig. 9). In an attempt to estimate these, a weighted average approach was used, as below:

$$K_{avg} = K_m x_m + K_{pc} x_{pc} \tag{10}$$

where K_m and K_{pc} are the hydraulic conductivities of the matrix and the permeable channels respectively, and x_m and x_{pc} are the fractions of the matrix and channel material respectively within the profile. Using $K_{avg\text{-}pt}$ derived from the pumping tests of 100 m/day and a K_{pc} value of 8,200 m/day from the tracer tests, the fraction of channel material cannot exceed 0.012 or 1.2% of the profile, otherwise Eq. 10 leads to negative K_m values (Table 4). For a given section of aquifer, a value for x_{pc} 1.2% dictates a K_m value of 2 m/day. At 1% of permeable channels in the profile K_m is 18 m/day. These calculations indicate that only a small portion (about 1%) of the aquifer consists of connected highly permeable channels, with the remainder comprising the lower permeability sandy gravel matrix. This figure is consistent with the findings of Jussel et al. (1994) who reported 1.9% of single openwork gravels in similar alluvial gravel media.

The derived K_m value range above for the matrix (2–18 m/day) is similar to that derived from the Kozeny-Carman equation (averages 3.8–44.6 m/day, dependant on porosity; Table 3). In addition, the K_{pc} value derived for permeable channels from tracer tests (about 8,200 m/day) is also relatively similar to the value derived from Kozeny-Carman equation (about 5,400 m/day) using measured porosity values. Though only a small sample size (three) was available, estimates of average K values for pure sand lens (2.4–9.6 m/day) indicate K values of the same order as the matrix material, confirming suggesting that sand and sandy gravel matrix material may be grouped together for the purposes of flow and transport modelling.

If the effective porosity for openwork gravels at the facies scale ($\phi_{eff\text{–}pc}$) is assumed to be 90% of the total porosity (0.35), as measured in a study of Canterbury gravel unsaturated zone (unpublished data) and by Jussel et al. (1994), then with the fraction of channel material (x_{pc}), a bulk effective porosity for the permeable material (bulk–$\phi_{eff\text{–}pc}$) can be estimated:

$$\text{bulk} - \phi_{eff-pc} = x_{pc} \times \phi_{eff-pc} \tag{11}$$

Thus an average hydraulic conductivity related to the water being transported through permeable channels ($K_{avg\text{-}pc}$) for the aquifer from tracer test velocity data would be:

$$K_{avg-pc} = \frac{V \times \text{bulk} - \phi_{eff-pc}}{I} \tag{12}$$

where V is the groundwater velocity and I is the hydraulic gradient. Calculated values for $\phi_{eff\text{–}pc}$ and $K_{avg\text{–}pc}$ are listed in Table 4 and show that the bulk–$\phi_{eff\text{–}pc}$ of the aquifer material (i.e. the whole profile) is about 0.0032 (assuming 1% permeable channels; i.e. x_{pc}=0.01) and $K_{avg\text{-}pc}$ is 93 m/day (assuming I=0.0024 and V=70 m/day, average of tracer test geomean velocities). These $K_{avg\text{–}pc}$ values are then similar to the $K_{avg\text{–}pt}$ values derived from the pumping tests (about 100 m/day). To test the assumption of the effective porosity being 90% of total porosity, the effective porosity of the permeable channels at the facies scale was reduced to 70% of total porosity. This reduced the $K_{avg\text{–}pc}$ value to about 70 m/day, indicating little sensitivity to this assumption. This finding of similar average K values from pumping and tracer tests was also observed by Vandenbohede and Lebbe (2003), who argued that conductivity values derived from pumping tests and tracer tests, in layered aquifers, could be complementary, if analysed with a realistic aquifer model incorporating geological data of aquifer geometry.

It is possible to approximately estimate the groundwater velocity through the matrix material and flow through the matrix and permeable channels, using the assumptions listed in the above calculations. Measurement of the matrix material within the unsaturated zone has shown the average total porosity of the matrix material ϕ_m to be 0.17 (unpublished data). Assuming that effective porosity at the facies scale is 90% of the total porosity, a facies effective porosity of about 0.15 was derived. Assuming that 1% of the profile is permeable channel material

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(therefore, 99% is matrix material, x_m), the profile or bulk effective porosity for the matrix is bulk–ϕ_{eff-m}; 0.99×0.15 ≅ 0.15, and K_m is 18 m/day, the groundwater velocity through the matrix material (at the same gradient, 0.0024) is estimated to be about 0.29 m/day. Lowering the effective porosity of the matrix material at the facies scale to 70% of total porosity (about 0.12) increases the estimated groundwater velocity through the matrix material to 0.36 m/day.

Overall flow and flow through the matrix and permeable channels can also be estimated from:

$$Q = K_{avg} I A \tag{13}$$

where A is the area, I=0.0024 and assuming K_{avg}=100 m/day (from the second pumping test), the estimated mean flow through a square metre section is 0.24 m^3/day. Of this, the flow through the matrix is estimated to be 0.043 m^3/day or about 18% of total flow (assuming 99% matrix, and K_m 18 m/day) leaving an average flow through the 1% of permeable material of 0.197 m^3/day (about 82% of total flow) through a square metre section of the aquifer. If the percentage of permeable material goes to 1.2% (K_m=2 m/day) then approximately 98% of flow is through the permeable material.

These calculations indicate that whilst there is only a small percentage of the aquifer that is highly permeable, it is enough to transport most of the water (and solute) through the aquifer at high velocities, as observed in tracer tests. The matrix material transport velocity (approximately 0.4 m/day) is estimated to be two orders of magnitude slower than the preferential flow lines (approximately 70 m/day), which are comprised essentially of permeable channels.

Conclusions

Average hydraulic conductivity values derived from pumping tests in an alluvial gravel aquifer were ~100 m/day. Variable drawdown responses were observed in an array of 21 observation wells surrounding the pumping well, with generally greater and earlier responses observed in observation wells located along preferential flow channels. Hydraulic conductivity values derived for the permeable channels K_{pc} using tracer test analysis (~8,400 m/day) were similar to those derived from particle size analysis (~5,400 m/day) but were approximately 2 orders of magnitude greater than the average K_{avg} values derived from the pumping tests. Using the complementary tracer and pumping test data and assuming effective porosity at the facies scale is 90% of the total porosity, a bulk or profile effective porosity (ϕ_{eff-pc}=0.0032) was calculated that reflected the percentage of permeable channels throughout the whole profile. An average K_{avg} value from the tracer tests (~93 m/day) was estimated that is similar to pumping test $K_{avg\text{-}pt}$ value (~100 m/day), which implies a maximum 1.2% of permeable channel material within the profile. This equates to fractions of flow through the matrix (~2%) and permeable channels (~98%), though if the estimate of permeable material is changed slightly to 1% the flow fractions of the matrix and the permeable channels change to 18 and 82% respectively.

This work indicates that while permeable channels may make up only a small proportion of the alluvial aquifer system, they can have a major influence on flow and transport of contaminants. Further work is ongoing to better characterize the sediments and to model transport of solutes and colloids through these environments.

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