Technical Report

Investigations and Monitoring Group

Vertical flow in Canterbury groundwater systems and its significance for groundwater management

Environment Canterbury

Your regional council

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Report No. R09/45
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August 2009

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Executive summary

This technical report has been prepared for a wide-ranging audience including technical experts, water resource managers and decision makers. The report investigates the degree of interaction and connection between sedimentary deposits located at different depth intervals in Canterbury.

Sections of the report:

- describe the general geological structure of the aquifers of the Canterbury Plains;
- illustrate the significance of leakage in groundwater flow dynamics by providing details on the propagation of drawdown effects in groundwater systems such as those in Canterbury and explain what both natural vertical flow between aquifers and pumping-induced leakage is;
- describe in detail how the magnitude, distribution and timing of drawdown and storage release in leaky aquifer systems can be calculated, based on appropriate aquifer tests;
- recommend techniques for aquifer testing and analysis in leaky aquifer systems;
- provide recommendations on the management of groundwater abstractions to ensure that pumping-induced leakage is addressed appropriately.

Modelling of groundwater abstraction has illustrated how the effects on piezometric heads are different in magnitude and timing depending on the depth of the strata from which abstraction occurs.

Modelling has also shown that abstraction from any depth induces vertical flow in a leaky aquifer system. It has shown that the volume of water removed from the uppermost strata is eventually equivalent to the pumped volume. Where there are other sources of accessible water, such as a hydraulically connected river, the storage depletion will be apportioned between drainage of the water table and the other sources.

Measuring effects at the water table caused by individual abstractions at depth is difficult because such effects are generally diffuse over a substantial area, delayed in time, and commonly obscured by water level fluctuations caused by other abstraction and weather-related changes over the period of measurement. Despite this, where practicable, it is useful to monitor these changes during a pumping test to calibrate the analysis.

Abstraction from deep wells means that leakage develops over a wide area. Consequently the areal extent of water table effects is more widespread but smaller in magnitude than the direct drawdown effects that would occur during pumping from the shallowest aquifer.

The significance of leakage to managers of groundwater and associated ecological systems is that ultimately, any take, from any depth, can impact on the level of the water table and modify interaction with surface waterways, although the magnitude and location of the effect will differ depending on the location and depth of the abstraction.

Key conclusions and recommendations of this report are:

- Shallow and deep strata should not be managed as disconnected entities as the flow of water between them can be substantial.
- Effective management of a groundwater resource requires recognition that the effects on piezometric heads will be different in magnitude and timing depending on the depth of the strata from which groundwater abstraction is occurring.
- Management of the groundwater resource should take into account that, although individual consents pumping from deep layers may cause very small changes in the level of the water table, those deep abstractions can eventually affect the water table level and change recharge or discharge components of the system, such as stream and spring flows. The magnitude and location of these changes (i.e. which streams could be affected and to what degree) is dependent on the location of the groundwater takes, their depth and the hydrogeological properties of the system.

- For each of the recognised groundwater systems within Canterbury, leakage between strata is only one of a number of issues that needs to be incorporated into management decisions and this should be done using an integrated approach. The modelling of leakage presented here looks at a simplistic setting where the aquifers are of infinite extent and there are no recharge sources present. The complexities of each system need to be considered when management systems are being developed.
- Whether a sedimentary sequence is characterised as being layered or as simply displaying a degree of anisotropy is not usually of great relevance in the assessment of leakage effects.
- The terms used to label natural sediments (e.g aquitards, aquifers) should not influence decisions made on the management of groundwater or assessments of groundwater flow. Those decisions should be made based on actual or potential effects that arise from the abstraction of groundwater from the resource.
- The use of the Hantush and Jacob (1955) solution to analyse aquifer tests, whilst valid as an indicator of pumping-induced leakage, cannot estimate realistic long-term drawdowns in a groundwater system and a more complete and relevant method such as the Hunt and Scott (2007) or Zhan and Zlotnik (2002) solutions should be used for the assessments. Alternatively, where necessary, a method that allows for changes in the recharge and discharge components of the system should be used.
- Careful analysis of appropriate aquifer tests undertaken outside the irrigation season, provides useful information for the assessment of drawdown effects and of long-term effects on the hydrogeological system. Such tests, although not precise, are the best and most practical tool for assessing the magnitude and timing of pumping-induced leakage effects related to an individual abstraction.

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Unit Abbreviations

Length
m metre
km kilometre

Volume
m^3 cubic metre

Glossary[1]

Term	Meaning
Leakage and vertical flow	Leakage is the flow of water from one hydrogeological unit to another. In this report, for clarity and consistency with existing literature on flow induced between sedimentary deposits resulting from groundwater abstraction, the term 'leakage' has been adopted. The term 'vertical flow' is also used interchangeably with leakage, to emphasise for the reader the vertical component of flow. The term leakage has previously been used to describe vertical flow between deposits under natural flow conditions, as well as when an aquifer is influenced by abstraction. Where it is necessary to distinguish between these two conditions, flow between deposits caused by the alteration of groundwater pressure is referred to as pumping-induced leakage.
Alluvial	Deposited by flowing water, usually rivers.
Anisotropic	Where physical properties, such as hydraulic conductivity, vary with direction.
Aquiclude	Low permeability geologic medium that, although porous and able to absorb water, is incapable of transmitting significant quantities of water.
Aquifer	Saturated, permeable geologic medium that is capable of yielding economically useful quantities of water to wells.
Aquifer system	A series of multiple connected aquifers in which flow can occur between aquifers.
Aquitard	Low permeability geologic medium that retards groundwater flow through it. It may not yield its stored water in significant quantities to wells.
Aquitard conductance (K'/B')	Vertical hydraulic conductivity of a confining unit divided by its thickness.
Avulsion	Sudden change in the course of a body of a river. Where a river ceases to be confined by terraces it may change its course down-gradient of the avulsion point.
Braided river	A river that contains many inter-connecting channels or braids; in Canterbury braided rivers commonly transport gravel-sized detritus.
Confined aquifer	An aquifer confined above and below by lower hydraulic conductivity deposits. Water in a confined aquifer is under greater pressure than atmospheric pressure.
Confinement	Confining material is saturated material bounding an aquifer that has a lower permeability than the aquifer and therefore restricts the rate of flow into and out of the aquifer. Where information such as bore-logs or aquifer testing indicates this material is continuous over some distance, it may be classed as an aquitard or a series of aquitards.
Darcy's Law	Discharge through a porous medium is directly proportional to the hydraulic gradient, hydraulic conductivity, and cross-sectional area through which the flow is occurring.
Fluvial	Produced by the action of a river or stream.
Fluvio-glacial	Deposited by a mixture of fluvial and glacial processes (detritus glacial in origin, deposited by rivers).
Flux velocity (v)	Velocity of flow averaged across a cross-sectional area of a porous medium.

[1] Many of these definitions are based on the National Ground Water Association Compendium of Hydrogeology: Porges and Hammer (2001).

Term	Meaning
Head (h)	Total hydraulic head is the sum of elevation head, the velocity head, and the pressure head at a given point in an aquifer.
Heterogeneous	Where physical properties, such as hydraulic conductivity and porosity, vary with location.
Hydraulic conductivity (K)	Coefficient of proportionality describing the rate of fluid flow for an isotropic porous medium and homogeneous fluid. It is the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow.
Leakage factor or leakance (L)	Represents leakage through an aquitard(s) into a leaky aquifer If L is high, leakage occurs over a wider area.
Perched aquifer	Locally saturated zone overlying a low-permeability unit in an otherwise unsaturated zone; this is a type of unconfined aquifer.
Permeability	A term commonly used as a qualitative synonym for hydraulic conductivity of a porous medium. Intrinsic permeability is a measure of the ease with which a fluid moves through a porous medium that is dependent upon the physical properties of the medium itself, and not upon the fluid being transmitted.
Piezometer	Non-pumping well, generally of a small diameter, that is used to measure elevation of the water table or piezometric surface.
Piezometric surface	An imaginary surface representing the head of the groundwater within a hydrogeologic unit; it may be contoured to indicate direction of groundwater flow.
Saturated zone	Subsurface zone in which the voids in a porous material are filled with water. The water table is the top of the saturated zone in an unconfined aquifer.
Sedimentary reworking	Continued transport and sorting of sedimentary detritus after original deposition (common in rivers).
Semi-confined aquifer / leaky confined aquifer/leaky aquifer	Aquifer confined by lower permeability deposits that allow leakage of water and through which recharge to the aquifer can occur both naturally and during abstraction.
Sorting	Concentration of sedimentary detritus into a predominance of similar grain sizes (e.g. gravel), by removal of finer material.
Specific discharge	Apparent groundwater velocity calculated from Darcy's Law: flux per unit area of voids and solids; also called darcy velocity
Specific storage (S_s)	Volume of water that a unit volume of aquifer releases from or takes into storage under a unit change in hydraulic head.
Specific yield (S_y)	Proportion of drainable pore space of a material.
Storativity (S)	Volume of water released from or taken into storage in an aquifer per unit surface area per unit change in the component of hydraulic head normal to that surface. In an unconfined aquifer this is approximately equal to the specific yield.
Transmissivity (T)	The rate at which water is transmitted through a unit width of an aquifer under a unit hydraulic gradient. It is equal to the hydraulic conductivity multiplied by the saturated thickness of the aquifer and is a function of properties of the liquid and the porous media.
Unconfined aquifer	Aquifer with no confining material between the saturated zone and the land surface such that water is free to fluctuate under atmospheric pressure. Top of the saturated zone is known as the water table.

1 Introduction

1.1 Contents and purpose of this report

This report describes the occurrence of natural and pumping-induced vertical flow of groundwater, and evaluates its importance for groundwater resource management. Vertical movement of groundwater can occur through the alluvial deposits of Canterbury when there is a vertical component to the hydraulic gradient. For consistency with existing literature on flow induced between sedimentary deposits with groundwater pumping, the term “leakage” is used in this report. Leakage and vertical flow have been used interchangeably in this report.

Leakage has previously been used to describe vertical flow between deposits under natural flow conditions, as well as when an aquifer is influenced by abstraction. To distinguish between these two conditions, flow between deposits caused by the alteration of groundwater pressure is referred to as pumping-induced leakage.

This report has been prepared in part to increase the understanding of the degree of interaction and connection between sedimentary deposits located at different depth intervals.

This study focuses on areas of fully saturated flow and has been designed to answer three important management questions:

- How does pumping-induced leakage from units other than the pumped aquifer control the drawdown effects in wells neighbouring a groundwater abstraction?
- Can pumping from any depth modify the surface water discharge from the system?
- How does the flow of water between deposits located at different depths impact on the way(s) in which the groundwater resource can be effectively managed?

The purpose of this report is to provide the following:

- An overview of Canterbury’s groundwater resources and identification of the potential for natural vertical flow and pumping-induced leakage between sedimentary deposits located at different depth intervals to assist in the understanding of flow dynamics of the groundwater system and the measurement of hydrogeological parameters controlling groundwater flow;
- A detailed explanation of the parameters controlling natural flow between sedimentary deposits and pumping-induced leakage within Canterbury and a summary of information on their magnitude and distribution; and
- A summary of the key conclusions and recommendations associated with the management of the hydraulic connection of sedimentary deposits and pumping-induced leakage at different depths.

1.2 Structure

The report is structured as follows:

- An overview of Canterbury's groundwater resources is provided along with identification of the potential for natural flow and pumping-induced leakage between sedimentary deposits (Section 2);
- The hydrogeological terms used within the context of this report are introduced along with a summary of the groundwater hydraulics related to aquifer leakage (Section 3);
- The significance of natural groundwater flow in directions other than parallel to the land surface within the Canterbury Plains groundwater systems is outlined (Section 4);
- The concept of natural flow between sedimentary deposits and pumping-induced leakage is described (Section 5);
- Recommendations on the optimum design of aquifer tests to determine the hydrogeological parameters controlling leakage are made (Section 6);
- Available methods for the assessment of pumping-induced leakage effects and the estimation of hydrogeological parameters using aquifer tests are presented (Section 7);
- Data from tests on wells located on the Canterbury Plains where pumping-induced leakage effects were observed are presented (Section 8);
- Simulations of pumping-induced leakage in different hydrogeological settings are presented and the appropriateness of the use of an analytical method to model pumping-induced leakage in these different hydrogeological settings is investigated (Section 9);
- Conclusions based on the information in this report are presented (Section 10);
- Recommendations on the management of pumping induced-leakage effects are made (Section 11).

1.3 Target audience

This report has been written for a wide-ranging audience including technical experts, water resource managers and decision makers.

Regional water resource managers and decision makers on RMA consent applications will, it is hoped, obtain a better understanding of the concept of natural vertical flow and pumping-induced leakage by reading this report.

2 Geology and groundwater of the Canterbury Plains

This section of the report outlines the geological formation of and the occurrence of groundwater beneath the Canterbury Plains and includes brief descriptions of other environments such as fault-bounded basins. Understanding the local geology is a pre-requisite for characterising the hydrogeology, and following that, selection of the correct models to assess and manage hydraulic changes in the aquifer system.

2.1 Geology

The most permeable water-bearing sediments in Canterbury have been formed by fluvio-glacial and alluvial processes that have distributed gravel-dominated sediment across the Canterbury Plains, inland basins and river valleys.

The Canterbury Plains comprise a series of large coalescing fluvio-glacial fans built by the main stem rivers (e.g the Rangitata, Rakaia and Waimakariri). During successive glaciations when glaciers partly occupied the inland valleys and extended to the eastern foothills, great quantities of detritus eroded from rapidly rising mountains. Gravel with sand and silt material was transported eastwards and deposited to form the fans of gravel-dominated sediments that extend beyond the present day coastline. During these glacial periods, some re-sorting of the gravel deposits occurred due to alluvial processes (Brown 2001). However, in contrast to interglacial periods the gravels are predominantly of lower permeability and poorly-sorted. The significance of the illustration in Figure 2-1 is that the gravel thickness is very thin in relation to its length and breadth of approximately 200 km by 60 km. On the scale of the plains there is almost no evidence for mappable units of gravel other than broad layering associated with successive glacial periods (Shulmeister 2007).

In the Canterbury Plains, the total thickness of gravels is variable, ranging up to greater than 600 m thick, as illustrated in Figure 2-1. In the fault-bounded basins, total thicknesses are much less.

During the warmer interglacial periods, the glaciers retreated up the valleys and less new gravel material was transported out onto the plains. However, alluvial processes continued to re-work the gravels. During these interglacial periods, while there was a greater proportion of re-working of gravels relative to the deposition of new poorly sorted materials, the areas where this re-working occurred were spatially constrained by the tendency for the alpine rivers to cut down through their fans. Laterally extensive reworking of the previously deposited gravel occurs only down-gradient of the avulsion point, where the river can burst its banks and migrate freely across the fan.

Superimposed on the regional pattern of the coalescing fans created by the major rivers, are inter-fan depressions occupied by smaller-scale fluvial fans of reworked gravels. For example as shown in Figure 2-2, in the current environment the inter-fan Selwyn River occupies the depression between the major overlapping fans of the Rakaia and Waimakariri rivers and consists of reworked gravels exhibiting a more open texture than much of the fluvio-glacial gravels associated with the adjacent fans (Anderson 1994).

Similar reworking of the major fans by rivers such as the Ashburton, Orari, Eyre and Ashley has produced linear strips of high-yielding gravel containing lesser amounts of silt and sand than the original fluvio-glacial fans. Such inter-fan alluvial processes would also have been active in earlier geologic times.

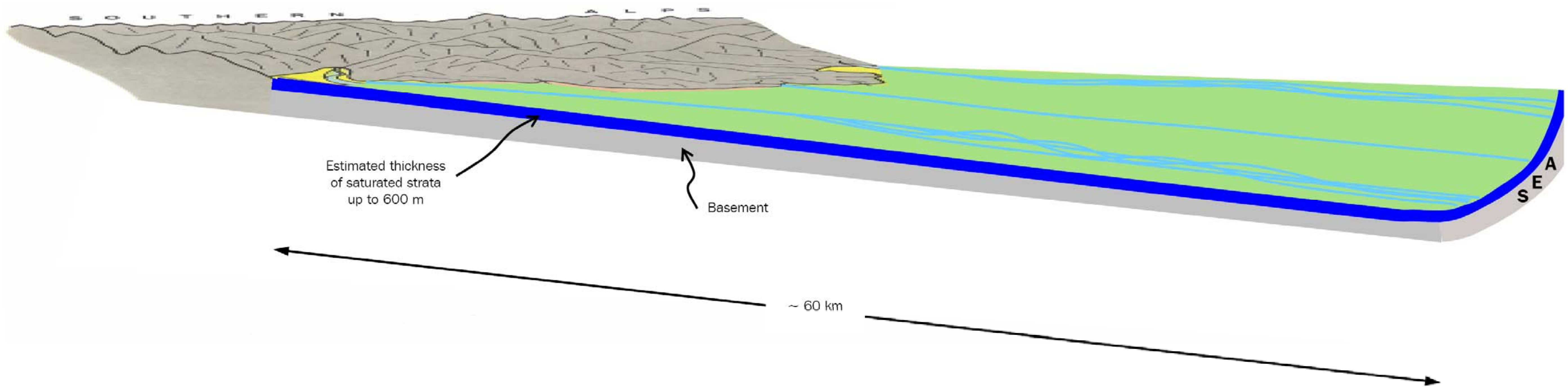


Figure 2-1: Conceptual oblique view cross-section of Canterbury Plains aquifer system with no vertical exaggeration (NB: underlying modified image sourced from NCCB (1983))

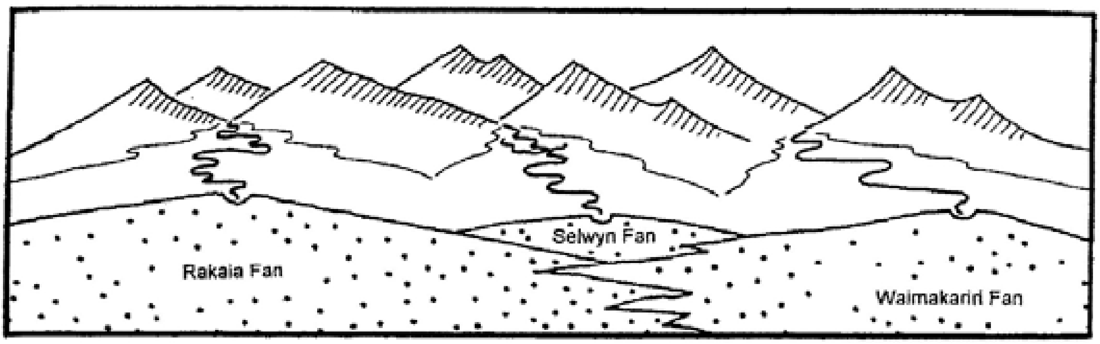


Figure 2-2: Schematic oblique view cross section, looking northwest, showing alluvial fan structure of the Selwyn plains (Anderson 1994)

The Canterbury Plains are characterised by dominant zones where the gravels and finer matrix sediments are poorly sorted and exhibit a relatively lower permeability, and other zones of re-working characterised by more permeable gravels with a coarse matrix.

In contrast to the gravel-dominated sediments that stretch across the plains, there is an area near the coast where marine intrusions during inter-glacial periods have led to the deposition of fine-grained deposits dominated by sand, silt and clay formed in a marine or estuarine environment. These deposits occur as wedge-shaped layers that thicken in a seaward direction. They provide clear geological separations between the gravel-dominated alluvial sediment layers deposited by rivers that extended beyond the current coastline during glacial periods when sea level was lower than at present.

Both inland and in the coastal zone the various gravel sediments have traditionally been classified into aquifers on the basis of the depth frequency of well screens (Davey 2006). At the coast, these zones correlate well with the gravel-dominated sediments that exist between the deposits of fine-grained sediment, as visible in well logs. Brown and Weeber (1992) describe the aquifer system referred to as the Christchurch Artesian aquifer system, extending along the coastal margin from north of the Ashley River to the lower Rakaia River, as a system consisting of a succession of artesian gravel aquifers and inter-bedded confining layers.

Further inland, the aquifers are not discrete entities as at the coast. Rather they represent zones of better sorted, more permeable gravels created by alluvial re-working processes, bounded by zones of poorly-sorted gravel. Over most of the Canterbury Plains, away from the fine grained marine and estuarine coastal sediments, the sedimentary deposits cannot be divided up into laterally extensive aquifers with a measurable thickness and extent because they are not separated by laterally extensive discrete lithological zones of contrasting material. Well logs do not indicate long distance correlation of lithological layering from well to well.

These processes of sediment formation result in a high degree of anisotropy in the deposits, where the horizontal hydraulic conductivity is higher than in the vertical direction. The resultant large vertical anisotropy created by fluvial processes is described in Chapter 4.2 of Freeze and Cherry (1979). Furthermore, the hydraulic conductivity of the sediments will be higher in the more open zones of re-worked gravels than in the more poorly-sorted material creating a large degree of heterogeneity both horizontally and vertically.

The alluvial processes responsible for the gravel deposits of the Canterbury Plains have also created gravel deposits in other Canterbury environments that contain groundwater (e.g. the Hanmer and Culverden basins). Despite their differences to the Canterbury Plains in scale and boundary conditions, there is the potential for groundwater interaction and connection between sedimentary deposits located at different depth intervals within these other systems.

2.2 Groundwater

The gravels of the Canterbury Plains contain a finite though renewable volume of groundwater accessed by wells, with a maximum well depth currently of 332 m near Methven. In some of the fault-bounded basins, total gravel thicknesses are much less. The cross-section of the Canterbury Plains provided in Figure 2-1 illustrates the approximate maximum width and depth of the conceptual model of the Canterbury Plains (the maximum width of the plains is approximately 60 km).

Depth to the groundwater table is highly variable, being close to the surface near the coast and up to 150 m deep in some inland areas closer to the foothills. The predominant sources of groundwater are seepage from the major alpine and foothill rivers, and rainfall infiltration on the plains.

Monitoring of piezometric heads across the Canterbury Plains indicates that variability in rainfall recharge can explain much of the seasonal variability in levels. Abstraction of groundwater, combined with natural discharge to rivers and through the coast, account for the seasonal decline of levels over the summer season when recharge is lower than total discharge (Bidwell 2003).

Monitoring of piezometric heads over upper and mid-plains indicates that heads decline with depth. Near the coast, they commonly increase with depth. Variation in magnitude and direction of gradients is consistent with that expected in a gently sloping prism of permeable sediment receiving recharge at the water table and bordered on one side by a constant head boundary (the sea).

In the areas where heads decline with depth, groundwater is moving from the shallower to deeper gravels as it also travels (predominantly) horizontally, while at the coast where heads are greatest at depth, water is also migrating upwards. The rate at which this groundwater moves is controlled by the hydraulic conductivity of the material through which it passes and the magnitude of the driving hydraulic gradient.

It is probable that groundwater in the uppermost gravels of the Canterbury Plains discharges into the marine environment, with the deeper gravel sediments probably pinched out laterally by marine silt and clay, which would greatly restrict lateral flow. This restriction to seaward flow in the deeper gravel layers forces a significant component of upward vertical flow, especially in the coastal zone both north and south of Banks Peninsula. The significance of this is that the predominant discharge from the groundwater system direct to the marine environment is likely to be through the uppermost sediments.

The general conceptual model of the Canterbury Plains is that of a system with full saturation of all sediment between the water table and the underlying basement. Preliminary investigation indicates localised areas where there is the potential for variable saturation with depth. This is based on observations of piezometric heads which suggest the presence of perched aquifers. This report is limited in scope to fully saturated flow.

The geological structure of the Canterbury Plains is such that the average vertical hydraulic conductivity across the sedimentary deposits is less than the average horizontal or bedding-parallel hydraulic conductivity. Consequently, when groundwater is abstracted from shallower levels there is a localised, direct effect on the water table. Conversely, abstraction from deeper aquifers creates a more diffuse drawdown effect in shallower layers that will ultimately result in a reduction in storage at the water table. This reduction in storage may in turn result in either a decrease in discharge to, and/or increase in recharge from, hydraulically-connected surface water bodies.

The observed change in drawdown with time in observation wells during many aquifer tests in the Canterbury Plains, even those tests of poor quality, has indicated that water that is abstracted from a well is drawn both from within the permeable deposits across which the bore is screened, as well as from leakage of water into the screened zone from less permeable surrounding material.

A recent review by White (2009) concluded that the potential for vertical flow within the Central Plains was high over most of the plains, reducing as the coast was approached.

In summary, the geology of much of the upper and mid Canterbury Plains provides a degree of lithological impediment to vertical groundwater flow - there is greater resistance to vertical flow than horizontal flow over the system. However, groundwater flow can occur in any direction, including both parallel with the major sedimentary bedding structure of the basin and across this structure.

3 Groundwater hydraulics and terminology relevant to this report

This section presents and explains in some detail the terminology and processes referred to throughout this report. An understanding of these terms and processes is necessary to understand natural groundwater processes, the modelling presented in this report and the significance of vertical flow, particularly that induced by pumping, to the management of groundwater.

The changes in head, associated hydraulic gradients, and volume that occur within a groundwater system in response to a recharge event (such as rainfall) or discharge (such as groundwater abstraction) are governed by two key controls: the storage capacity of the groundwater system and the rate at which the sediment allows water to move within it.

The *storativity* of the medium describes how much water can be *stored* within it. Groundwater is stored within the pores of a saturated medium, for example, in the voids between the individual particles of gravel within an aquifer. Depending on the aquifer type, changes in the volume of stored water occur by changes in the saturation of the pore spaces and, to a very much lesser degree, compression or expansion of the water and aquifer medium.

The *hydraulic conductivity* of the medium controls the velocity at which groundwater moves under a specific hydraulic gradient. Under the same hydraulic gradient, water will move through sands and gravels much more rapidly than it could move through materials such as silt or clay, as the sands and gravels have larger connected pores i.e a higher hydraulic conductivity.

These parameters, hydraulic conductivity and storativity, are discussed in more detail in the following sections as they are the fundamental controls on natural and pumping-induced vertical flow. For explanations of other technical terms related to groundwater, refer to the glossary at the beginning of this report.

3.1 Hydrogeological definitions for geologic units

Geologic units from which groundwater is abstracted via wells are referred to as aquifers. Other terms, such as aquitard, are used to describe geologic units that do not allow water to flow through them as easily as an aquifer.

The practice of labelling strata with names such as aquifers and aquitards can lead to debates between groundwater practitioners, where these terms are interpreted differently.

For example, for one individual the word *aquitard* may imply a layer of clay with a discrete thickness, as evidenced by bore logs, separating two permeable gravel layers (*aquifers*) that provides great resistance to flow between them. For another, the word *aquitard* may imply a lens of gravel with a higher silt content than the strata surrounding it, perhaps not even distinguishable in a bore log, which permits slightly lower groundwater velocities than the surrounding strata under the same hydraulic gradients.

In some areas, the word *aquifer* might be used to describe a hydrogeological unit, from which groundwater is abstracted, with a hydraulic conductivity of 5 m/day. However, in other areas with more permeable water-bearing deposits, a recognised *aquifer* may have a hydraulic conductivity of 500 m/day, while what is referred to as an *aquitard* overlying this may have a hydraulic conductivity of 5 m/day. Describing an aquifer as a geological formation from which stored water can be drawn at a useful rate eliminates the dependence of the definition on the actual properties of the medium.

Kruseman and de Ridder (1991) state that their definitions of aquifers, aquitards and aquicludes are purposefully imprecise with respect to permeability, as their definitions are relative ones. This relativity in definitions is significant and it should be appreciated that the classification of these geological units is not based on hydraulic conductivity alone.

The definitions used to describe the hydrogeology in this report are outlined below. However, it should be stressed that the actual words chosen to label natural sediments should not influence decisions made on the management of groundwater or assessments of groundwater flow. Those decisions should be made based on actual or potential effects that arise from the abstraction of groundwater from the resource.

Aquifer

An aquifer is considered to be a geologic unit, or a group of units, which contains water and yields useful quantities of water to wells.

Kruseman and de Ridder (1991) share the view of many groundwater practitioners that an aquifer is something permeable enough to yield economic quantities of water to wells.

In the Canterbury Plains, a concentration of well screens within a depth interval is often used to delineate the thickness of an aquifer (as outlined in Davey (2006)). This is consistent with the definitions above. However, given that the potential exists for wells at depths outside the specified interval to yield economic quantities of water, the interpreted thicknesses should not be viewed as absolute.

Aquitard

An aquitard is considered to be a geologic unit that transmits water, but at a lower rate than aquifers. The definition therefore is a relative one, as with that for an aquifer.

While an aquitard might conduct water at a lower rate than an aquifer, over a very large area, an aquitard can still permit the passage of large volumes of water. An aquitard can be viewed as a geologic unit that retards the flow of water.

The absence of well screens over a certain depth interval should not, in isolation, be used to make an assessment of the thickness of an aquitard as it is possible that there are zones within that depth interval that may yield economic quantities of water, but have not yet been targeted by a well driller.

Aquiclude

An aquiclude is a geologic unit which may contain quantities of water, but does not transmit water; it precludes the flow of water.

As per the definition in this report, and in most groundwater text books and technical papers, any formation that is impermeable should be classed as an *aquiclude* because anything classed as an aquitard is permeable, i.e. water can travel through it. Within the Canterbury Plains groundwater system, there are no known aquicludes.

Kruseman and de Ridder (1991) provide examples of dense un-fractured igneous or metamorphic rocks as typical aquicludes but point out, that in nature, truly impermeable geological units seldom occur; all of them can transmit water to some extent, and must therefore be classified as aquitards.

Use of these terms

Water levels, well screen information, aquifer testing and bore logs are all helpful in delineating zones of varying hydraulic conductivity within a groundwater system, and for clarity in communication. It is appropriate to label these zones as aquifers and aquitards or, rarely, aquicludes. However, as discussed above, it is the physical properties of the strata controlling the release of stored water, the rate of groundwater flow and the transmission of groundwater head changes, not the names or labels, which need to be considered in groundwater assessments and groundwater management decisions. The names or labels are relative terms only.

3.2 Confinement, heterogeneity and anisotropy

Confinement

Confining material is saturated material bounding an aquifer that has a lower permeability than the aquifer and therefore restricts the rate of flow into and out of the aquifer. Where information such as bore logs or aquifer testing indicates this material is continuous over some distance, this "layer" may be classed as an aquitard or a series of aquitards. The presence of confining material usually results in differing water pressures between wells screened in deeper and shallower aquifers. Aquifer confinement is usually thought of as vertical, i.e. above and below, but a rock valley wall or clay bound gravels laterally constraining a buried gravel river channel can also be considered to be confining.

Where an aquifer is not overlain by any confining material, it is classed as an *unconfined aquifer*. At the water table in an unconfined aquifer the water is everywhere at atmospheric pressure (Figure 3-1).

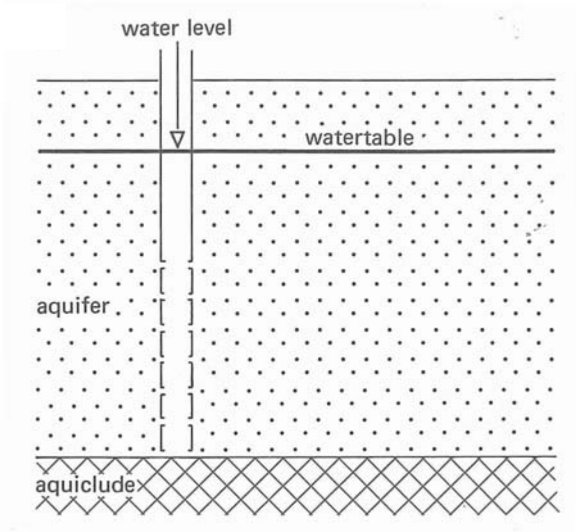


Figure 3-1: Unconfined aquifer (from Kruseman and de Ridder (1991))

Where confining strata overlie and/or underlie an aquifer, the rate of vertical flow into and out of the aquifer through the less permeable material is controlled by the hydraulic conductivity of this confining material together with the hydraulic gradient. This type of aquifer is referred to as a *leaky confined aquifer* or, alternatively, a leaky aquifer or semi-confined aquifer (Figure 3-2). This restriction on flow that the confining material provides has the following implications:

1. groundwater abstraction will cause head changes in the pumped aquifer to propagate further laterally than would be the case in an unconfined aquifer (comparing magnitudes after the same period of pumping) due to different mechanisms of storage release (elastic storage in the leaky confined aquifer versus physical drainage of the pore spaces in the unconfined aquifer).
2. pumping a leaky confined aquifer results in smaller changes in the level of the overlying water table than would occur from pumping an unconfined aquifer directly, but these changes occur over a wider area. Eventually the rate at which stored water is released from the unconfined aquifer equals the rate of pumping from the leaky confined aquifer. The exception is where there are other sources of accessible storage, such as from a hydraulically connected river, in which case the storage depletion will be apportioned between drainage of the water table and the other sources.

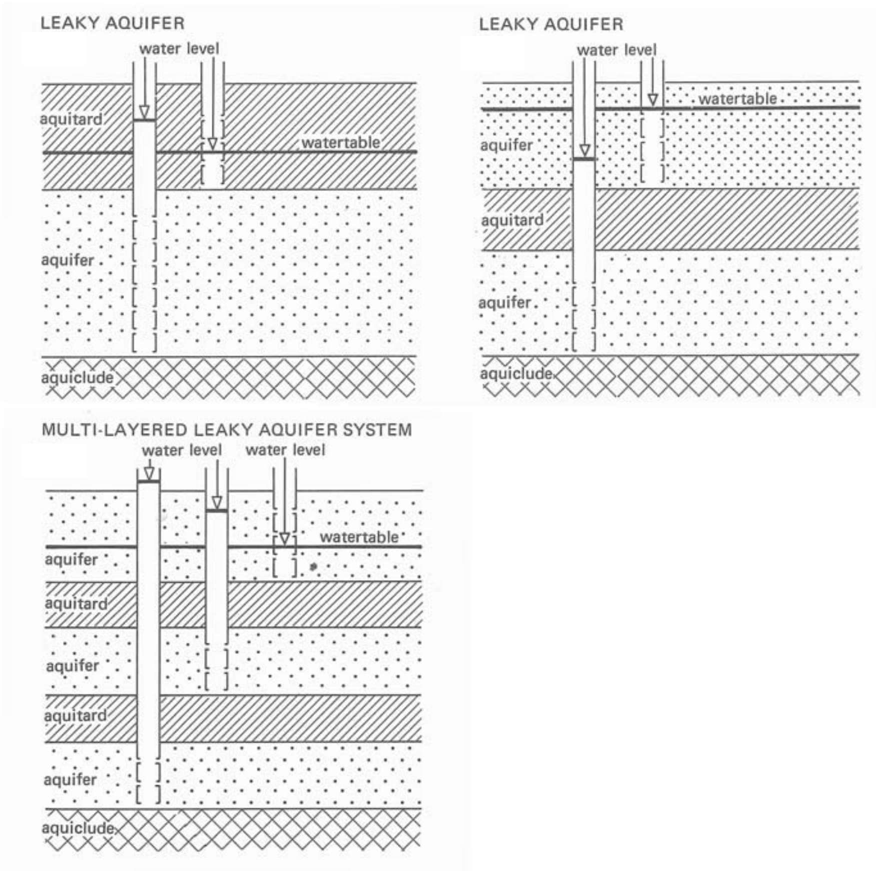


Figure 3-2: Leaky confined aquifers (from Kruseman and de Ridder (1991))

An aquifer bounded above and below by an aquiclude, which means that no water can flow vertically into or out of the aquifer, is classified as a *fully confined aquifer*. In reality, there are unlikely to be any aquifers fitting this description within Canterbury, because even the less permeable deposits are likely to transmit water. Nevertheless, most aquifers where the confining unit is of very low permeability are referred to as confined aquifers within Canterbury.

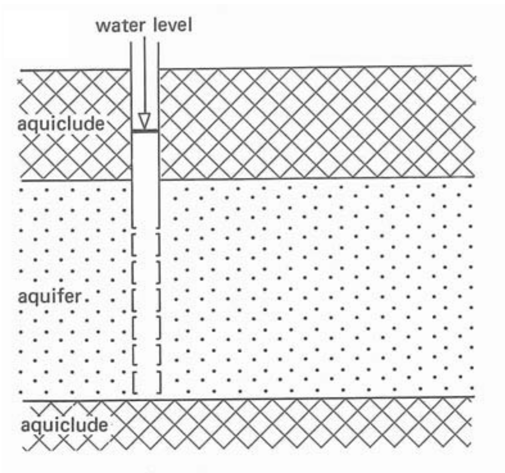


Figure 3-3: Confined aquifer (from Kruseman and de Ridder (1991))

Heterogeneity and anisotropy

Where all physical properties are the same throughout a geologic unit it is described as *homogeneous*. Conversely, if the physical properties vary at different locations within a geologic unit it is described as *heterogeneous*.

The homogeneity of a material is judged by comparing the length scale of the feature of interest (as described by Bear (1979)). A poorly sorted gravel aquifer that has a range of particle sizes may be considered as being heterogeneous over the scale of 10 cm to 1 m, but over a distance of 10 m, there may be sufficient repetition such that the material can be considered to be homogenous. In this aquifer, the drawdown in two wells located the same distance and depth (but more than 10 m) in different directions from a pumped well would be the same, even though the aquifer is heterogeneous over smaller scales.

Where the hydraulic conductivity of an aquifer is the same in all directions at a single point, the aquifer is described as *isotropic.* Conversely, where the hydraulic conductivity changes with direction, it is *anisotropic*. Both the nature of alluvial depositional processes and subsequent pressure of overlying strata tend to result in material that has a much higher hydraulic conductivity in the horizontal direction than vertical as flat particles tend to be oriented with their longest dimension parallel to the plane on which they settle (Bear, 1979). In addition, as described by Dann *et al.* (2008), on a large scale, alluvial depositional processes tend to result in material that has a higher hydraulic conductivity in the flow direction of the rivers responsible for the deposition than in the direction perpendicular to that flow.

Homogeneity and isotropy are both scale-dependent. For example, where there are a number of sedimentary layers, each being homogenous and isotropic but each possessing a different hydraulic conductivity, on a sufficiently large scale, the number of layers could be grouped together and classed as a single anisotropic unit. Consider an alluvial deposited system consisting of alternating deposits of silt-bound gravels and clean gravels each of the order of 0.1 to 1 m thick, and with each deposit of a different average hydraulic conductivity. Over a total thickness of 10 to 100 m, the system could be considered to be a single homogeneous anisotropic aquifer as illustrated in Figure 3-4. In Figure 3-4, K_1 and K_2 refer to hydraulic conductivities of the contrasting strata, L_1 and L_2 are layer thicknesses. K^v and K^p are the effective vertical and horizontal conductivities of the entire system. The *anisotropy ratio,* which will be referred to later in this report, is the ratio of the hydraulic conductivity in one plane to the hydraulic conductivity in another plane, typically expressed as the ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity. [2]

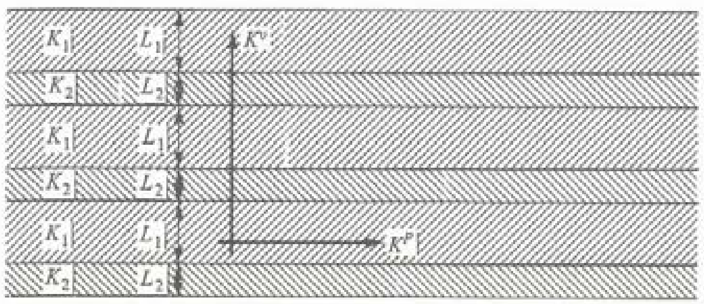


Figure 3-4: A multi-layered system exhibiting anisotropy (from Bear (1979))

In a layered sequence with contrasting hydraulic conductivities between layers, groundwater flow has a larger horizontal component in the layers with higher hydraulic conductivities than in the layers with lower hydraulic conductivities. This refraction of groundwater flow lines is outlined in Chapter 5 of Freeze and Cherry (1979), from which Figure 3-5 is reproduced.

In effect, in strata of contrasting conductivities, in highly conductive layers (K_1), flow is nearly parallel to the layering, while in less conductive layers (K_2), flow is more nearly normal to the layering.

[2] More details of this theoretical explanation of the relationship between layer conductivities, layer thicknesses and the overall anisotropy of hydraulic conductivity are described by equations 2.31 and 2.32 in Freeze and Cherry (1979).

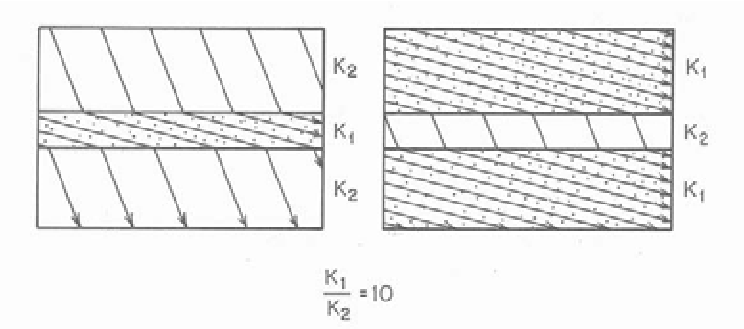


Figure 3-5: Refraction of flow lines in layered systems (from Freeze and Cherry (1979))

3.3 Physical properties of strata controlling groundwater flow rates

This section outlines the controls on the rate of flow through a groundwater system: the hydraulic gradient and the hydraulic conductivities of the saturated strata. Other equations that simply combine the hydraulic conductivity with dimensions are also provided. Units are provided in terms of metres (m) and days.

The **Hydraulic head (symbol h; m**) is the total energy of the water per unit weight of the water. This is also referred to as the piezometric head. For groundwater, it is the height of water in a well above a certain datum.

The **Hydraulic gradient (symbol i; m/m)** is a measure of the change in hydraulic head (**h**) within saturated strata over a given distance (**l**). As the change in hydraulic gradient increases within a groundwater system, so do the groundwater velocities.

The **Hydraulic conductivity (symbol K; m/day)** of a medium is a constant of proportionality between the specific discharge of a fluid through the saturated medium and the hydraulic gradient. It is also referred to as the coefficient of permeability. For a given hydraulic gradient, the velocity through a material with a large hydraulic conductivity value, such as sand and gravel, is much larger than through material with a low value (such as silt or clay).

Darcy's equation (Equation 1), is an empirical relationship originally derived from experimental data by Henri Darcy (1856), and subsequently derived from first principles. It describes how the velocity (**v**) of groundwater through a saturated medium is controlled by the hydraulic gradient (driving force) and the hydraulic conductivity of the strata:

$$v = -K\frac{\Delta h}{\Delta l} = -Ki$$ **Equation 1**

The negative sign is a mathematical convention indicating that the head (h) is decreasing in the direction of flow.

A simplifying assumption in the derivation of this equation is that the flow is non-turbulent. For most groundwater applications, this assumption is appropriate.

While the resulting velocity has units of length/time, this equation assumes flow through the full cross-sectional area and makes no allowance for the fact that most of the cross-sectional area is occupied by solid material (sediment). This calculated Darcy velocity is often referred to as the flux velocity to distinguish it from the pore velocity which is the actual velocity at which flow occurs through the total cross-sectional area of the pores. As the particles occupy much of the space, the pore velocity is always larger than the Darcy velocity. The pore velocity is relevant in assessing the rate of movement of a particle through an aquifer. Only the Darcy velocity is relevant to this report.

The total Darcy flow (**symbol Q, m^3/day**) through a specified cross-section area of saturated strata (**A**) is calculated as follows (Equation 2):

$$Q = vA = -K\frac{\Delta h}{\Delta l}A = -KiA$$ **Equation 2**

The following terms commonly used in equations to describe groundwater flow, are simply the hydraulic conductivity combined in different ways with a dimension:

The term **Transmissivity (symbol T; m^2/day)** is defined as the product of the horizontal hydraulic conductivity of an aquifer (**K**) and its saturated thickness (**B**). Transmissivity is one of the parameters typically interpreted through the analysis of pumping tests.

The term **Aquitard conductance (symbol K'/B'; day^{-1})** is defined as the ratio of an aquitard's vertical hydraulic conductivity (symbol **K'**) to the saturated thickness of the aquitard (symbol **B'**). This term indicates how rapidly groundwater can flow vertically through an aquitard, under a unit vertical hydraulic gradient across the aquitard. It is also referred to as the Vertical Drainage Term or Leakage Coefficient. The reciprocal of this term (B'/K') is referred to as the aquitard resistance, for which the symbol C is sometimes used, e.g. Kruseman and de Ridder (1991). However the symbol C is also used for the aquitard conductance (K'/B') in other literature such as Trinchero *et al.* (2008). To avoid ambiguity the complete expression (K'/B') is used here.

The term **Effective aquitard conductance (symbol $(K'/B')_{effective}$; day^{-1})** is a widely applicable term. In most geologic systems, there is more than a single geologic unit with a contrasting permeability to the pumped aquifer between the pumped aquifer and the water table.

The rate at which water can flow downwards through these units and into the pumped aquifer is dependent on the different vertical hydraulic conductivities of each unit. The thickness of each unit affects the magnitude of hydraulic gradient created and therefore also has a resultant effect on the velocities. Use of the term effective aquitard conductance does not require K' and B' for each layer to be known because it combines them into one term.

Where there are discrete layers overlying a pumped aquifer, each with a known thickness and hydraulic conductivity, the effective aquitard conductance can be calculated as follows (Equation 3):

$$\left(K'/B'\right)_{effective} = \frac{1}{B_{Layer1}/K_{Layer1} + B_{Layer2}/K_{Layer2} + B_{Layer3}/K_{Layer3}}$$ **Equation 3**

In reality, there are seldom discrete geologic layers of uniform thickness present that can be simply classed as aquitards and aquifers. Even if there were, detailed knowledge of the hydrogeological properties of each layer is seldom present. Conveniently, aquifer test data can be analysed to determine a value for the effective aquitard conductance.

Where numerous geologic units exist between the pumped well and the water table, the value obtained from aquifer test data for the effective aquitard conductance provides a measure of the conductance of all strata overlying the pumped well. Inclusion of the word "aquitard" should be viewed as no more than a label for there are many instances in nature where there is no evidence from, for example, well logs, of discrete aquitards. Depositional processes commonly result in strata that provide greater resistance to vertical flow than horizontal flow; that is the vertical hydraulic conductivity is lower than the horizontal hydraulic conductivity.

Therefore, an effective aquitard conductance term derived from drawdown data analysis of an aquifer test is a measure of the vertical hydraulic conductivity and thickness of all overlying strata, regardless of whether the strata consist of discrete aquitards or a complex arrangement of deposits with variable hydraulic conductivity.

A variation on these aquitard conductance terms, is a term called the **Leakage factor (symbol L; m)**, which is the ratio of the aquitard conductance and aquifer transmissivity through the following expression (Equation 4):

$$L = \sqrt{\frac{KB}{K'/B'}} = \sqrt{\frac{T}{K'/B'}} = \sqrt{BB'}\sqrt{\frac{K}{K'}}$$ **Equation 4**

This term provides a combined measure of how much resistance to vertical flow is provided by overlying strata and how permeable the pumped strata are. Large values for a leakage factor indicate that there is a significant contrast between the transmissivity of the pumped strata and the vertical conductance of the overlying aquitard(s) represented by lower permeability strata. A result of this greater resistance to vertical flow is that any resultant drawdown of the water table, induced by abstraction from deeper strata, will be smaller than if the abstraction had been from the strata containing the water table, and will occur over a wider area.

Low values of aquitard conductance and large values of the leakage factor indicate that the drawdown effects of deep abstractions on the water table will be dispersed over larger areas and be smaller in magnitude than abstractions from the shallowest aquifer.

3.4 Properties controlling the volume of stored groundwater

The total volume of groundwater contained within a groundwater system is the amount of water contained within all the pore spaces of all the saturated strata.

The following terms describe how much water is released from storage when there is a reduction in actual water level or groundwater pressure.

The **Specific yield (symbol S_y, dimensionless)** is the proportion of drainable pore space of a material. In a geologic formation that contains a water table, such as an unconfined aquifer or aquitard, water is released from storage as a result of physical drainage of the pore spaces at the water table. The volume of water that is released from storage per unit surface area under a unit decline in head in an unconfined aquifer is equivalent to the specific yield. The specific yield is sometimes referred to as unconfined storativity.

The specific yield is typically less than the total porosity of a material, because when piezometric heads decline, not all of the water stored within the pore spaces at the water table will be released. Some water is retained due to capillary forces and disconnected pores.

Kruseman and de Ridder (1991) report a range of values from 0.01 to 0.3 for the specific yield of an aquifer. High values correspond to gravels with no fines, while small values correspond to fine-grained sediment.

The **Specific storage (symbol S_s; m^{-1})** of confined and leaky confined aquifers is the volume of water that is released from elastic storage per unit volume of the geologic formation under a unit decline in head.

In a leaky confined aquifer, if the head in the aquifer is always above the base of the overlying aquitard there can be no physical drainage of the aquifer pore spaces. Therefore, the volume of water released from the pumped aquifer itself is only that which is released via the expansion of the water and compression of the aquifer strata as heads fall (elastic storage).

The associated **Elastic storage coefficient (symbol S, dimensionless),** or storativity, is the volume of water that is released from storage per unit surface area under a unit decline in head. For a particular confined or leaky confined aquifer, this is equivalent to the specific storage multiplied by the thickness of the formation.

Bear (1979) suggests that typical values for the elastic storage coefficient within a confined aquifer are of the order 10^{-6} to 10^{-4}.

4 Natural groundwater flows within Canterbury

This section presents some water level monitoring data from wells within the Canterbury Plains. Water level monitoring indicates that the flow direction is generally horizontal and towards the coast. At any location in the aquifer system there are likely to be components of both vertical and horizontal flow directions.

Monitoring data from wells of differing depths located in close proximity (piezometer clusters) within Canterbury allow the calculation of the vertical component of hydraulic gradient at those locations. The locations of most piezometer clusters identified by Environment Canterbury are shown in Figure 4-1.

Figure 4-2 depicts the average horizontal gradients at the locations of these piezometer clusters, as interpreted from the piezometric contours shown on the Environment Canterbury GIS system. These contours have been interpolated from the measured piezometric heads at different wells in the area. Although some contours may relate to water measurements in wells with differing screen depths, most contours reflect wells of similar screen depth. Most of the horizontal gradients are fairly similar in Figure 4-2 (in the order of 10^{-3} m/m), with the steepest gradient of $1.1*10^{-2}$ (11 m decline over 1 km) in the vicinity of the Hororata cluster (located near the foothills) and the lowest gradients of $5*10^{-4}$ (0.5 m decline over 1 km) in the vicinity of the Surf Club, Scruttons Road and Plover St clusters near the coast of Christchurch.

Figure 4-3 shows the average vertical gradients between each of the well screens within the piezometer clusters. A comparison between the values for these vertical gradients and the horizontal gradients shown in Figure 4-2 shows that in some areas, vertical gradients are typically much larger than the corresponding horizontal gradients.

These figures have resolved the gradient into horizontal and vertical components. The actual direction of the gradient will usually be at a shallow angle to the land surface, although there are parts of the groundwater system where the net direction of the gradient within the groundwater system is almost entirely vertical (at the coast for example).

Note that a large gradient does not imply a high groundwater velocity. The actual groundwater velocity is controlled by the hydraulic conductivity of the material and, as is illustrated in this report, the vertical hydraulic conductivity is typically lower than the horizontal hydraulic conductivity in alluvial gravel systems.

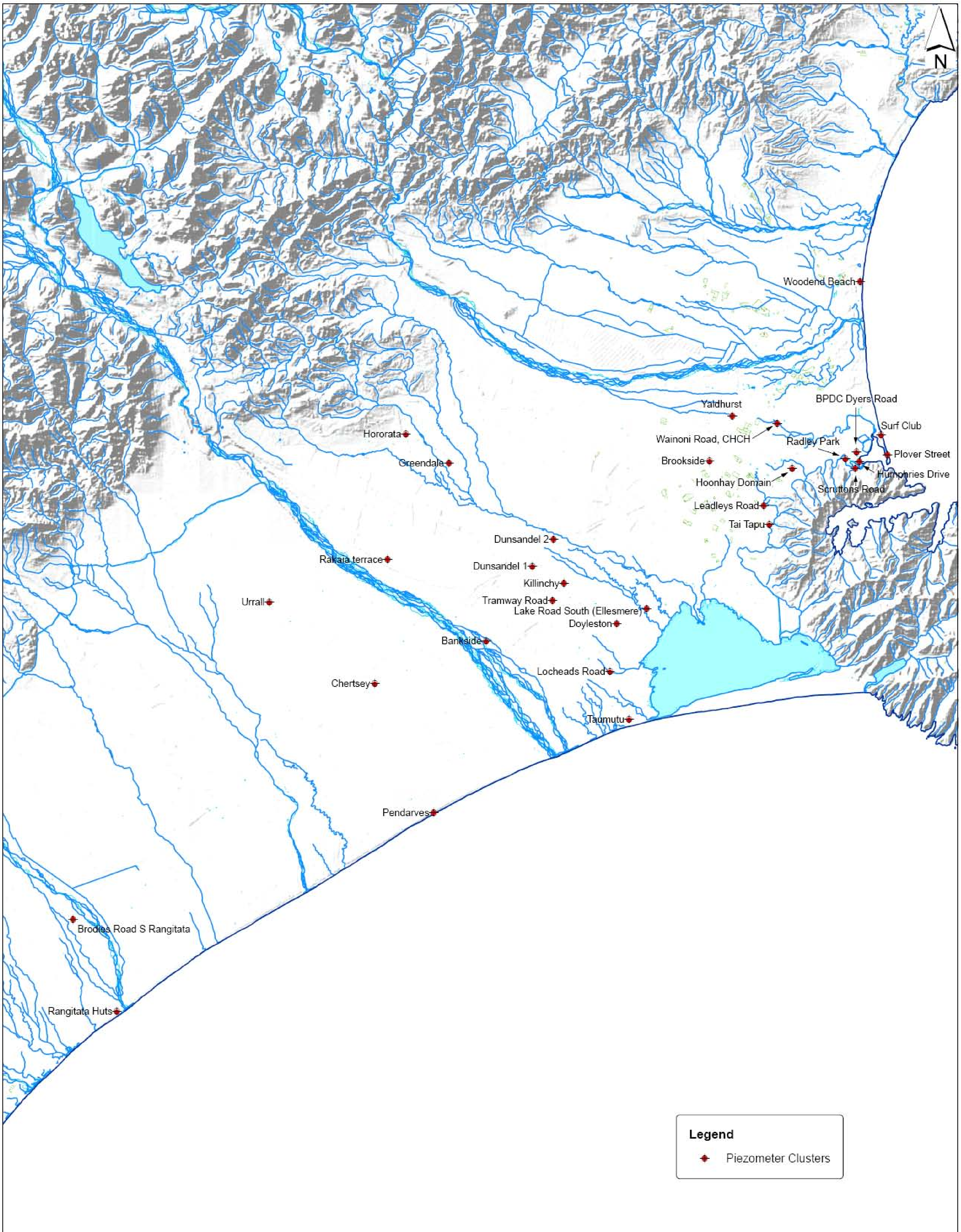


Figure 4-1: Location of piezometer clusters from Environment Canterbury records

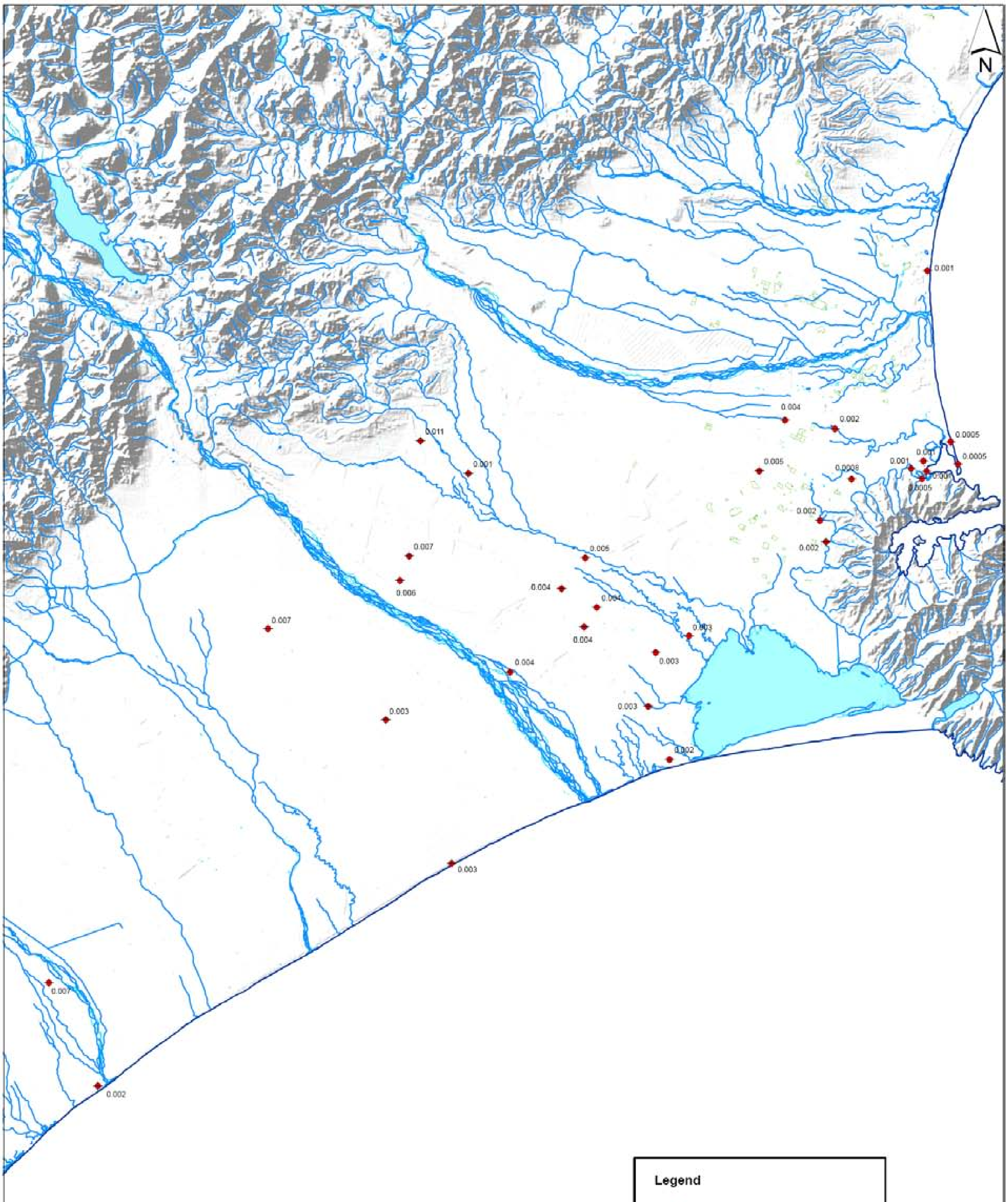


Figure 4-2: Estimated horizontal gradients (as obtained from Environment Canterbury database)

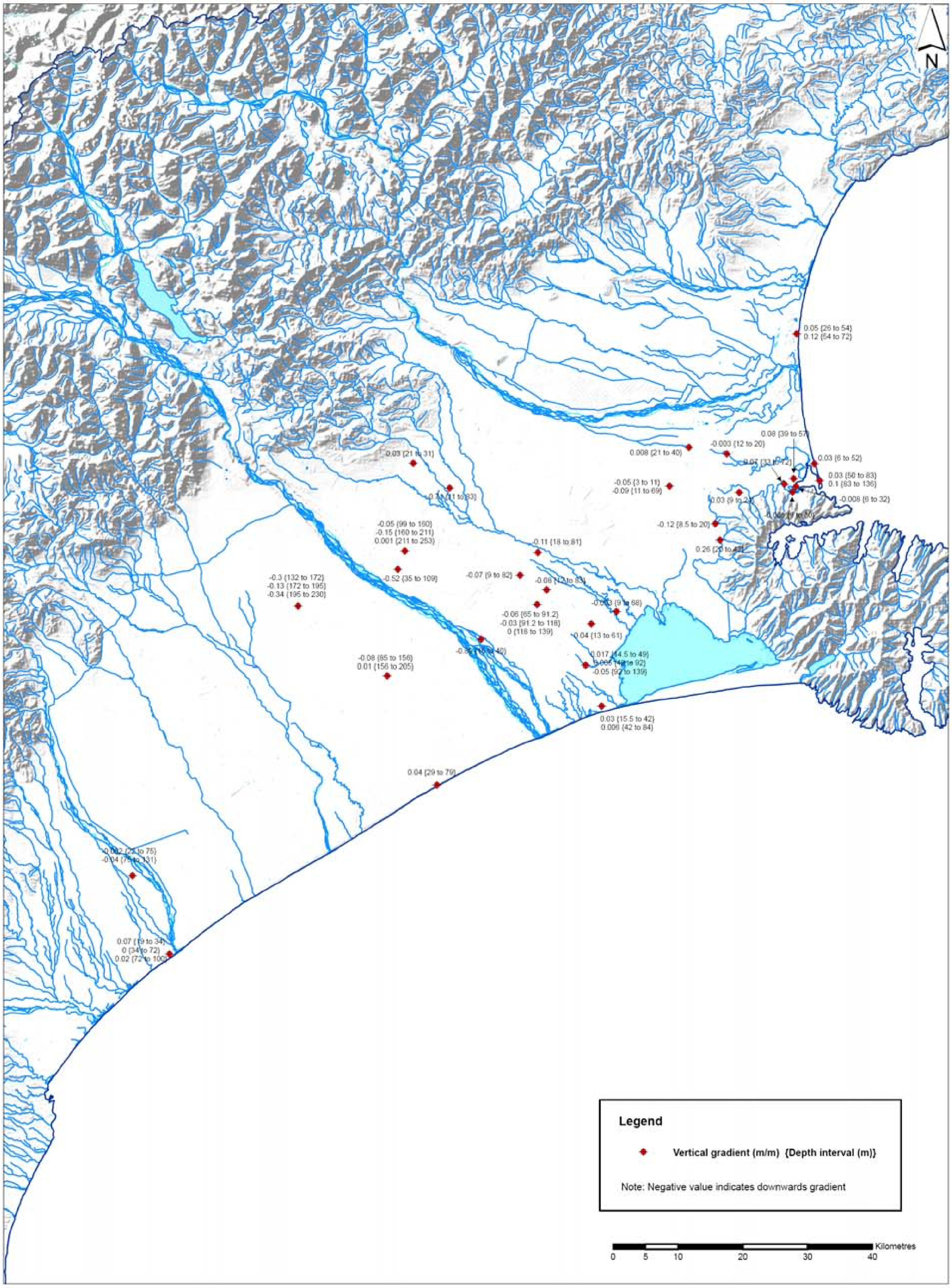


Figure 4-3: Estimated vertical gradients (as obtained from Environment Canterbury database)

An example of the difference in water levels at different depths, representing a vertical gradient, is shown for the Taumutu piezometer cluster in which the bases of the three screens are at 15 m, 42 m and 84 m depth. Taumutu is very close to the coast and hence the effect of the constant head boundary will be large. The water levels are plotted over time in Figure 4-4, and the maximum, minimum and average water levels are plotted versus depth in Figure 4-5.

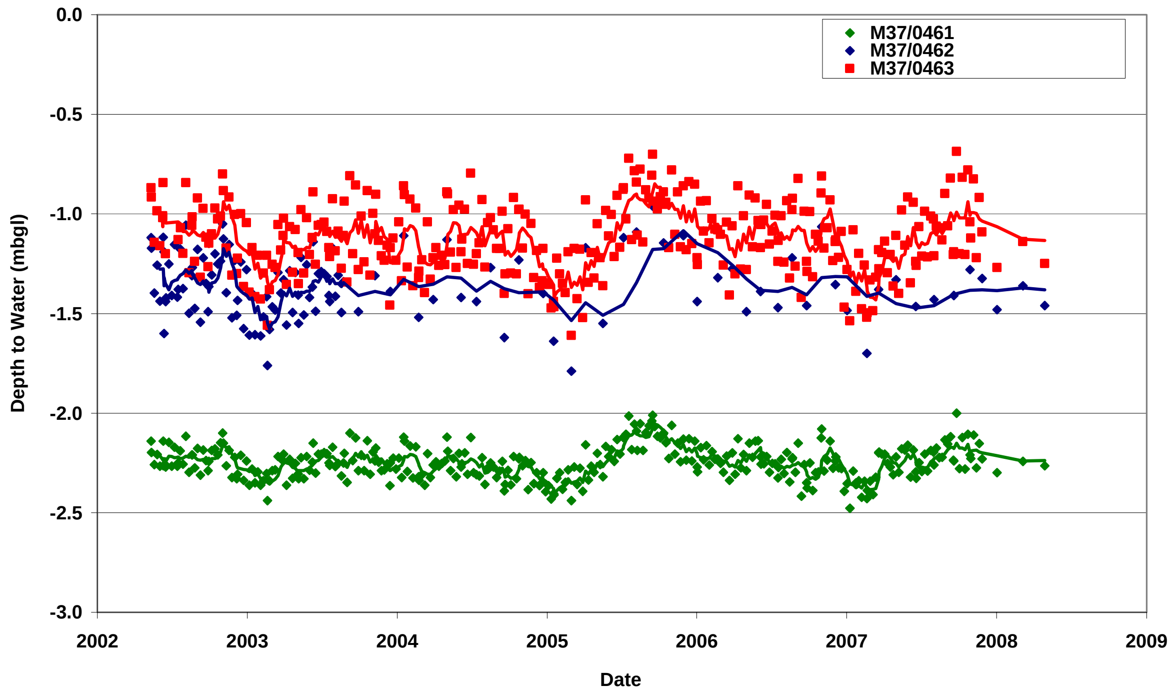


Figure 4-4: Plot of water levels for the three piezometers of the Taumutu piezometer cluster (with 5 point moving averages shown)

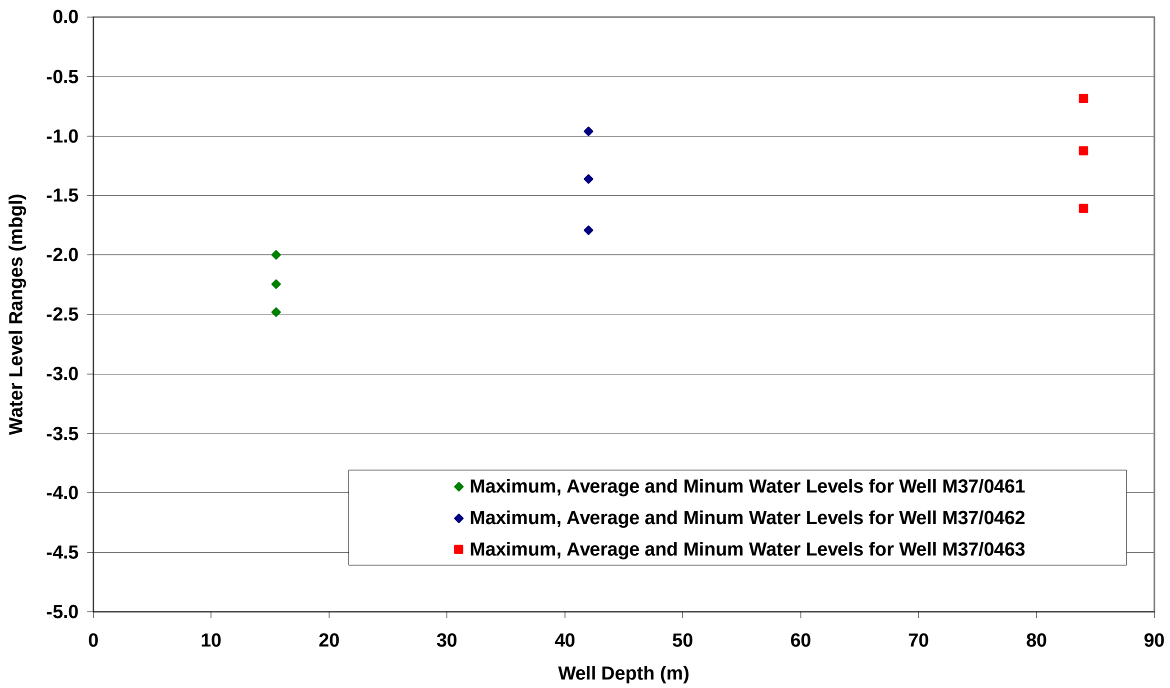


Figure 4-5: Plot of maximum, minimum and average water levels vs. depth for the three piezometers of the Taumutu piezometer cluster

Figure 4-4 and Figure 4-5 demonstrate that the vertical gradient is upward at this location and as the vertical hydraulic conductivity will not be zero there will be movement of water from deeper to shallower levels. The average vertical gradient between the two shallower wells is 0.03 m/m, which is high in contrast to the horizontal gradient in this area of around 0.002 m/m (refer Figure 4-2 and Figure 4-3 for these values). This shows that there is vertical component to the flow in this area, although a calculation of the magnitude of this flow component would require an estimate of the vertical hydraulic conductivity for the deposits between the screens of the piezometers.

While interesting, it is important to understand that background flow patterns within an aquifer have little bearing on pumping-induced leakage effects. These pumping effects are essentially superimposed on the background setting. For example, in a groundwater system that is fully saturated and where water is naturally flowing from a deeper aquifer to a shallow aquifer (upward gradient), pumping from the deeper aquifer will reduce the rate of flow to the shallow aquifer. Where water is naturally flowing from a shallow aquifer to a deeper aquifer (downward gradient), pumping from the deeper aquifer will increase the rate of vertical flow from the shallow aquifer. Where there is no natural vertical flow between permeable strata and the natural flow is only horizontal, pumping from any depth within the system will create vertical head gradients and therefore induce vertical flow to occur (predominantly from the shallow strata).

5 Natural flow between strata and pumping-induced leakage

This section outlines both the concept of natural flow between sedimentary deposits and pumping-induced leakage and clarifies the differences between the two concepts.

5.1 Flow direction

Flow within a saturated medium, such as the sedimentary deposits of the Canterbury Plains, is as a result of either natural or induced head differences in the water occurring within that medium. The direction of flow is determined by the pattern of head differences created by water either entering or leaving the system at some location, for example via rainfall, flow into or out of surface water bodies, or groundwater abstraction and, in some instances, from atmospheric pressure changes.

In most settings, the overall flow direction is almost horizontal. Examples where horizontal flow is dominant include:

- In an unconfined homogeneous aquifer underlain by an aquiclude such as un-fractured bedrock as a result of groundwater abstraction where the well is screened over the entire saturated thickness. Note that in Figure 5-1, the flow is not completely horizontal close to the abstraction well as the well is only screened in the lower half of the original saturated thickness.

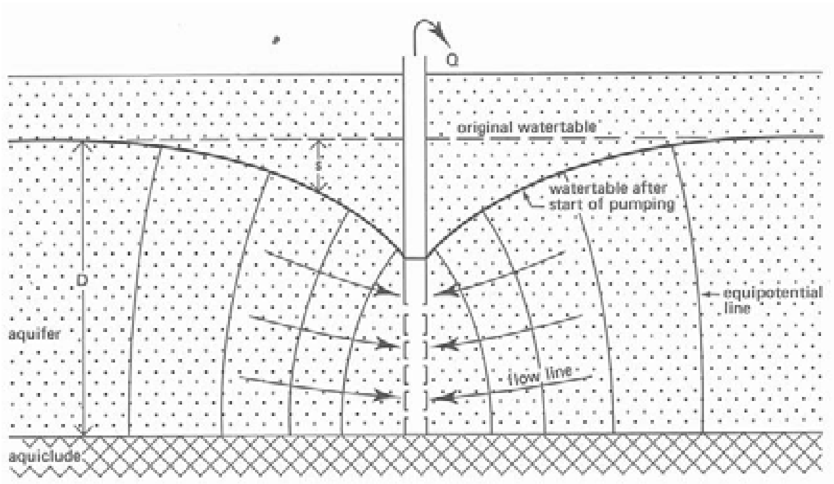


Figure 5-1: Horizontal flow within a pumped unconfined aquifer (beyond a small distance from the well screen) (from Kruseman and de Ridder (1991))

- Towards a pumped well within a deep leaky confined aquifer where either the well is screened over the entire aquifer thickness or at a sufficient distance from a partially penetrating well.
- Flow within a layered groundwater system where the distances between a recharge source and discharge are large, which results in the pressure gradients being predominantly horizontal.

In other hydrogeological settings, the overall flow direction can be predominantly vertical in parts of the system. Examples of where vertical flow is dominant include:

- Within a saturated aquitard overlying a pumped aquifer, that has a much lower hydraulic conductivity than the hydraulic conductivity of the pumped aquifer (due to refraction principles as introduced in Figure 3-5).

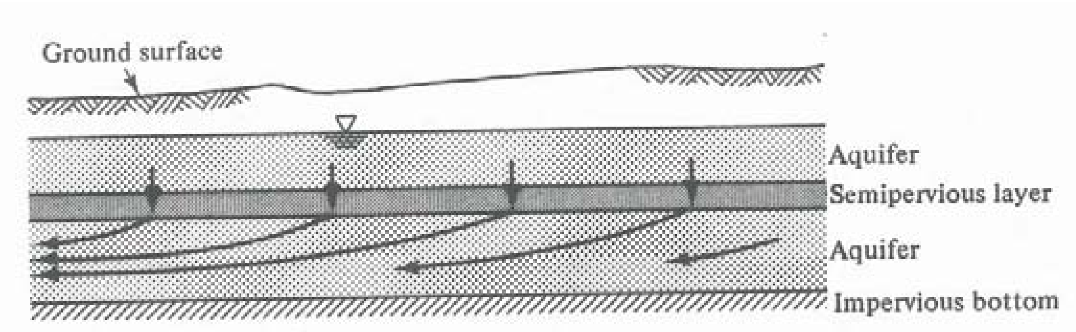


Figure 5-2: Vertical flow within an aquitard (shown as the semi-pervious layer) (from Bear (1979))

- Where a well is partially penetrating (screened only over a short interval at the base of an unconfined aquifer or a leaky confined aquifer), water will flow near vertically from the water table in the near vicinity of the well downwards to the well screen.

In most hydrogeological settings, there will be significant variation in flow directions within different parts of an aquifer due to variation in hydraulic conductivity and storage properties. For example:

- Where there are zones of higher permeability within an aquifer, flow will occur preferentially along these. When pumping from an aquifer that includes these zones, pressure changes will propagate more rapidly within these (Figure 5-2), which will induce flow from adjacent lower permeability areas of the aquifer. This flow from the adjacent areas will flow into the channels obliquely to the net flow direction. This situation is demonstrated in data from field investigations e.g. Dann *et al.* (2008).
- When pumping from a leaky confined aquifer, the predominant groundwater flow direction in the pumped aquifer will be horizontal but will be predominantly vertical in overlying and underlying aquitards (Figure 5-2).

5.2 The source of pumping-induced leakage

Drillers' logs, natural exposures and gravel pits all indicate that there is stratification in the gravel-dominated sediments of Canterbury, although exposures and logs also indicate that each individual layer is laterally discontinuous (Ashworth *et al*. 1999; Shulmeister 2007). The thickness of these layers is generally between 0.1 and 10 m. The properties of these individual strata affect the rate at which water is transmitted and released as a result of groundwater abstraction.

Pumping-induced leakage as opposed to natural vertical flow is flow that enters the pumped aquifer as a result of the release of storage from strata above, below or laterally adjacent to pumped strata. The release of water from strata below and adjacent to the pumped aquifer occurs from the release of elastic storage. The release of water from strata above the pumped aquifer occurs in part due to elastic storage, but also from physical drainage of the pores in the shallowest strata by lowering of the water table.

Many aquifer tests in Canterbury including those of poor quality, indicate the occurrence of pumping-induced leakage (Sinclair Knight Merz, 2008 and Sinclair Knight Merz, 2009).

As the available storage at the water table (within the pore spaces) is typically much larger than the available storage within deeper leaky confined strata (elastic storage), physical drainage (release of storage) from the water table will ultimately become the predominant source of water that is derived from storage when abstraction from leaky confined strata occurs (although groundwater-surface water interaction can modify this balance).

This mechanism of storage release is illustrated in the following hypothetical examples:

Example 1: Consider a large unconfined aquifer with a well screened only in the lower half of the aquifer, located a great distance from recharge and discharge sources, underlain by an aquiclude with an initially horizontal free surface (water table). Withdrawing water from the well will result first in a reduction in pressure around the well screen, which will in turn, by creating a head difference, induce vertical and horizontal flow to the well through the surrounding material. As water is removed from the aquifer, the water level in the aquifer will drop accordingly (e.g. Figure 5-1).

Example 2: Now consider the configuration in Example 1 is changed by separating the single aquifer into two with an aquitard (e.g. Figure 5-2). Pumping is from only the lower aquifer. The volume of water in the system is identical to Example 1. There is now a resistance to the rate at which water can be drawn to the well vertically from drainage of the pore spaces at the water table in the upper (unconfined) aquifer. Water will now preferentially be drawn through the pores within the pumped aquifer with the result that the equivalent head changes (drawdown) will propagate over a much wider area of the aquifer than in Example 1.

In turn, this head drop over a large area beneath the aquitard will induce vertical gradients (decreasing with distance from the well) between the pumped aquifer and the water table, across the aquitard. This gradient induces vertical flow from the shallow aquifer, and this removal of water translates to a drop in the level of the water table.

The difference between Example 2 and Example 1 is that the drawdown of the water table is larger and more localised in Example 1, while in Example 2, water table drawdown is more widespread but smaller in magnitude.

If the shape of the depression in the water table created by the pumping in both examples, “the drawdown cone”, was inspected and the volume of this depression calculated after a reasonable length of pumping time, the volume of the cone would be seen to approach the volume of water abstracted from the well after taking account of specific yield, provided that the pumping had not caused any changes in the recharge and discharge components of the groundwater system.

While the terms aquitard and aquifer are used in this example, what is of relevance is that the greater resistance there is to vertical flow relative to horizontal flow (the larger the net anisotropy) the more widespread the drawdown will be in both the pumped strata and at the water table when compared to an equivalent isotropic aquifer at the same pumping times, although the magnitude of that drawdown will be smaller.

In a fully saturated groundwater system, layering and anisotropy control the way pressures propagate through the system, and therefore the resultant piezometric head changes, and the rate of storage released from various units. The total amount of water released from the different strata within a system is controlled by its storage properties.

The amount of water that can be stored and released via elastic storage within most aquifers is very small in comparison to the amount that can be stored and released via changes in the water table. This means that water taken from any depth within the system will ultimately result in a change in level at the water table. The net changes that occur will depend on the other features of the groundwater system, such as rainfall recharge and the degree of interaction with surface waterways.

The time taken to reach the point where this release of storage due to the physical drainage of the water table begins, and the rate at which water table drainage occurs, depend on how rapidly the head changes reach the water table. This timing is dependent on both the storage properties of all strata and the effective vertical hydraulic conductivity and thickness of the strata between the water table and the pumped aquifer, relative to the lateral hydraulic conductivity.

5.3 Drawdown sensitivity to hydrogeological parameters

Figure 5-3 shows the different influence that aquifer parameters have on the drawdown response to pumping. Note that this graph also applies to both layered systems and equivalent anisotropic systems, where distinct layering is absent, but where the vertical hydraulic conductivity is smaller than the horizontal hydraulic conductivity, as explained in Section 3.2. Note: In Figure 5-3 and following plots in this section, the illustrated responses do not take into account any pumping-induced recharge or discharge.

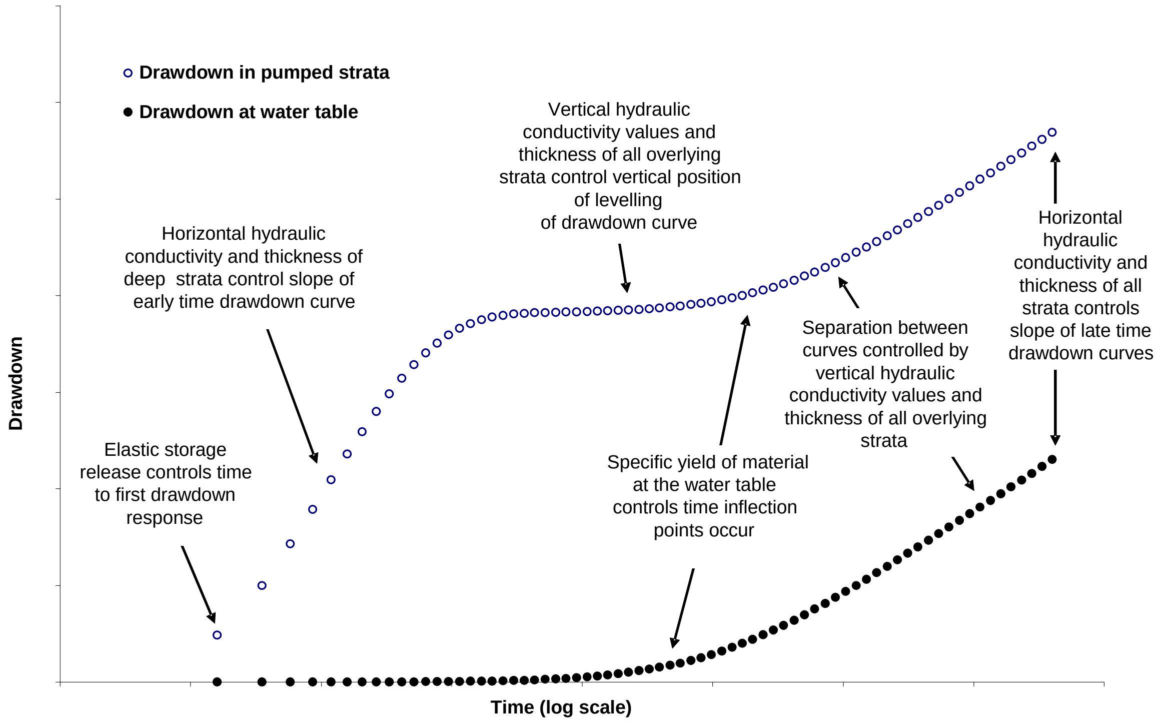


Figure 5-3: Drawdown sensitivity to hydrogeological parameters of a layered system with pumping from a leaky confined aquifer

5.4 Storage release sensitivity to hydrogeological parameters

Figure 5-4 illustrates the general effects of storage parameters of a layered system on the volume of water released from elastic storage and from physical drainage of layers of the water table. Note Figure 5-4, like Figure 5-3, also applies to an equivalent anisotropic system, where there is no distinct layering, but the vertical hydraulic conductivity is smaller than the horizontal hydraulic conductivity. Note that the elastic response of an aquifer is shown to be almost instantaneous whereas physical drainage is a slower process.

There are some points illustrated in Figure 5-3 and Figure 5-4 that relate to monitoring of effects and the management of the resource:

- The time at which release of storage at the water table begins will generally increase with the depth of an abstraction, although it is generally quite rapid due to the difference in the mechanisms of storage release. Conversely, with increased depths of abstraction the drawdowns will typically be smaller in magnitude but occur over a wider area.
- In practical terms, this delay and small response means that it is difficult to track changes in the decline of storage at the water table via drawdown measurements during short-term pumping tests in deep wells.
- A small response in the level of the water table resulting from pumping can be obscured by other water level fluctuations in this environment, caused by rainfall recharge, pumping from the water table aquifer, and other climatic and seasonal effects.
- Figure 5-4 shows that physical drainage or release of storage, by lowering of the water table, will ultimately become the predominant source of water that is derived from storage when abstraction from deeper, confined strata occurs in areas where there are no pumping-induced changes in recharge or discharge. However, in many instances the effects of abstraction are distributed across changes in recharge, discharge and aquifer storage.

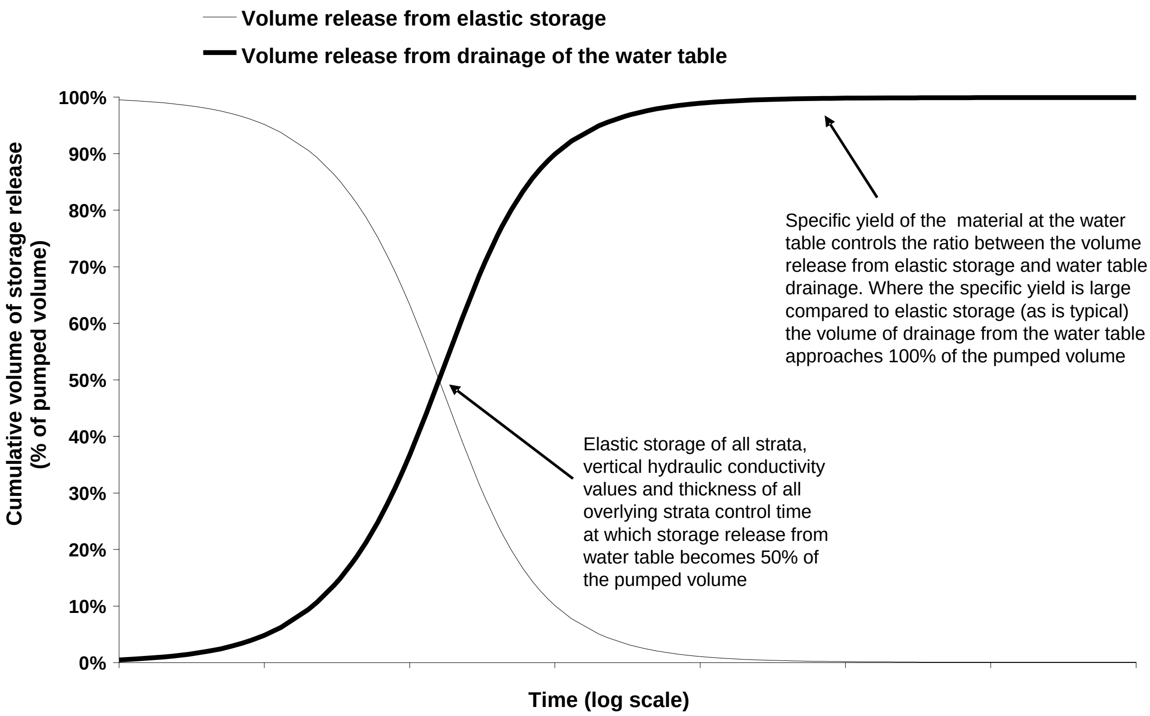


Figure 5-4: Storage release sensitivity to hydrogeological parameters of a layered system

6 Field tests to assess pumping-induced leakage

It is not yet practical to directly measure flow between different strata. Piezometric heads are relatively simple to measure and from these and a consideration of the physics of groundwater flow, inferences about groundwater movement may be made. This section outlines recommendations to optimise the quality of the data and analysis for aquifer tests that are designed to assess leakage.

Constant-rate pumping tests are a convenient method by which head changes can be induced in a groundwater system and monitored via observation wells. Maintaining a constant rate of abstraction simplifies the analysis and reduces the uncertainties associated with long-term measurement and background corrections.

Changes in piezometric heads can be measured in observation wells during a constant-rate pumping test, and with this information and the measured rate of abstraction from the well, the drawdown can be analysed with a model, appropriate to the hydrogeological setting and observed responses, to infer the changes in groundwater flow that were induced over the course of the test. Longer term effects of the pumping can also be extrapolated from the measured drawdown response using a model.

There are other field techniques available to estimate some hydrogeological parameters such as transmissivity, for example step-drawdown tests and other single well pumping tests. A well designed constant-rate pumping test, involving the measurement of drawdown data in surrounding observation wells will provide the most complete information about the response in the media surrounding the groundwater abstraction and, hence, the best information on the properties of the groundwater system.

Models available to analyse drawdown data obtained from a constant-rate pumping test and to predict longer term abstraction effects are outlined in Section 7.

6.1 Recommendations for constant-rate pumping tests conducted to assess pumping-induced leakage

Aquifer tests remain the most reliable method to assess leakage. This section outlines recommendations for tests that are designed to assess leakage to optimise the quality of the data and analysis. These are largely consistent with the updated aquifer test guidelines (Aitchison-Earl and Smith 2008) and, as a result, there is some overlap between the guidelines and this section of the report. It might not be practical to implement all these recommendations at every test site or it might not be necessary for the particular objectives of the test.

6.1.1 Observation wells

The use of observation wells often depends on the wells that are available at any particular location. This section describes an ideal distribution of wells, although it is recognised that the location of such wells may not always be feasible at every test site.

At least one observation well, more where practical, screened at the same depth as the pumped well should be monitored prior to, during and after the course of the test. The well needs to be close enough to provide a reasonable drawdown signal, but far enough from the pumped well not to violate the assumption in many of the analytical models that the pumped well is screened over the entire aquifer thickness.

At least one observation well screened in shallow strata containing the water table should be monitored to determine the absence or presence of drawdown effects at the water table over the pumping duration. This should be as close to the pumped well as possible. The reason for this is that experience and modelling have shown that it is very unlikely for drawdowns to develop at the water table as a result of pumping from a deep well at a magnitude that is discernible from background fluctuations. However, even if no drawdown is discernible in a shallow well, this provides useful information on the system and is important in calibrating the analysis of drawdown data from other wells.

It is useful to monitor at least one background monitoring well at a large distance from the pumped well that experiences either a very small or no drawdown signal to better assess background changes of heads. This can be difficult because if the well is too close it may experience drawdown as a result of the pumping; conversely, if it is too far away the background changes may not be representative of the background changes in the wells used for drawdown analysis.

6.1.2 Monitoring period

Monitoring of water levels in the wells and barometric pressure changes should be carried out for a sufficient period before the test (between 1 and 3 days is generally sufficient). Monitoring of recovery should be for at least as long as the duration of pumping and ideally for an additional one to two days beyond this.

6.1.3 Data corrections

Appropriate corrections are vital to the successful analysis of drawdown data to estimate leakage parameters. Methods for corrections are:

<u>Barometric head changes</u>
A barometric correction is required for changes in heads caused by atmospheric pressure changes. If the barometric change was transmitted to the water in the monitoring well with 100% efficiency, a 1hPa change in atmospheric pressure would cause a change in water head of 0.0102 m. In a confined or semi-confined aquifer, the transmission of atmospheric pressure changes to the groundwater in the aquifer is less than 100% efficient because some of the change is transmitted to the solid aquifer media and some to the water that fills the pore space. It is only the change transmitted to the water for which the monitoring data need to be corrected. The ratio of the observed water level fluctuation to the change that would occur under 100% efficiency is referred to as the barometric efficiency (BE).

<u>Tidal corrections</u>
Where the monitoring record shows a tidal effect, the test data will need to be corrected prior to analysis. This type of correction needs to account for the change in frequency and magnitude of each tidal cycle. Ideally monitoring of water levels in the tidal surface water body should be carried out to allow correction of background fluctuations. If testing is carried out within several kilometres of a tidal body this effect should be looked for in the pre-test monitoring data. Tidal effects can propagate inland through a confined aquifer for surprisingly large distances.

<u>Other background water level fluctuations</u>
Natural background trends that require correction include steadily increasing heads prior to, and following a pumping test caused by a rainfall event prior to the testing or a natural recession following cessation of a natural recharge event. Background trend corrections commonly assume a linear trend over the course of the test, but it should be recognised that linear correction is not always appropriate.

Where water level data still exhibit fluctuations following correction for barometric pressure changes and any tidal effects, further correction is required to remove effects that may be caused by the pumping of nearby wells. Testing outside the irrigation season is best for this reason.

Data corrections are probably the largest source of uncertainty in an aquifer test analysis and they must be made with care as they can have a significant bearing on the ability to determine aquifer parameters with any accuracy, including leakage.

6.1.4 Ratio of magnitude of corrections

Ideally, the magnitude of corrections compared to the observed drawdown signal should be small to minimise the uncertainty in the interpreted drawdown response. However, provided that corrections for barometric pressure effects, tidal effects and background trends are made accurately, it is still possible to reliably identify a drawdown response even where the magnitude of the corrections to the drawdown signal are comparable.

In practice, corrections for water level fluctuations caused by stresses other than the pumping from the well are often difficult to assess with certainty. The unexplainable/uncertain fluctuations prior to, during and after the pumping test need to be as small as possible relative to the drawdown signal, ideally, less than 10% of the drawdown signal.

6.1.5 Duration of a test to assess pumping-induced leakage

To assess the vertical hydraulic conductivity, K', of the overlying strata (incorporated into the term K'/B') the duration of a pumping test in a leaky aquifer needs to be sufficiently long to observe the levelling of the drawdown curve shown in Figure 5-3. Ideally, the test should be carried out for long enough to also determine a value for the specific yield of the material (S_y) at the water table. This parameter controls the departure of the drawdown curve from the pseudo-steady state phase as shown in Figure 5-3. It is also ideal to carry out the test for long enough to assess the horizontal hydraulic conductivity of the strata other than the pumped aquifer which affects the slope of the drawdown curve at later time as shown in Figure 5-3.

However, it is often not practical to carry out a test for long enough to determine a specific yield or to assess the horizontal hydraulic conductivity of the strata other than the pumped aquifer as, depending on hydrogeological properties and the pumped well-observation well separation distance, a test of a duration in excess of 10 days could be required. Over such a period, the magnitude and variability of fluctuations due to other causes typically makes it difficult to isolate the drawdown response precisely. It may also be difficult to maintain a constant pumping rate over such a period, and therefore the data would need to be corrected for this prior to analysis. The benefit of obtaining a full data set needs to be weighed up against practicalities such as the cost of pumping and potentially adverse environmental effects arising from the pumping test itself.

To determine an optimal test duration, it is useful to first record, download and evaluate background water levels from the proposed observation well(s). The magnitude of the background fluctuations can then be compared to the expected drawdown response. Alternatively, a short period of trial pumping from the abstraction well can be carried out to determine the actual drawdown response in the observation well. Where the fluctuations are large in comparison to the estimated or measured drawdown response during the trial pumping, and it is not possible to increase the well pumping rate for the test or to use a closer observation well, a long duration test may be impractical.

Prior to terminating a test, it is good practice to review manual readings from a well (or download electronic data), plot the data over time and compare the response to that expected (see Figure 5-3). If the full expected response for a leaky aquifer system has been observed, then pumping could cease (with water level monitoring continuing for and beyond the recovery phase). If the complete expected response was not present, but unexplainable/uncertain fluctuations appeared in the data or it is not practical to continue the test for some other reason, the test may need to be terminated.

Accordingly, the duration of a test is specific to that test; there is no recommended duration for all tests in Canterbury. The duration must simply be long enough to allow assessment of the aquifer parameters required at a precision sufficient to enable them to be used justifiably in assessments. Consideration of all the corrections required should allow planning of an effective test. Test results show that in some cases, levelling of the drawdown curve (indicating leakage and permitting the estimate of a value for K'/B') may be observable within hours in an observation well, while in other tests, it may take three days. This timing depends on the relative locations of the well screens and hydrogeological parameters. Where the specific yield at the water table and the horizontal hydraulic conductivity of strata other than the pumped aquifer need to be determined, a longer test duration is likely to be required, unless those parameters are already known from previous hydrogeological investigations or can be reliably estimated for the purpose of the assessment. Alternatively, these parameters could be established from a pumping test at a shallower level.

6.1.6 Analysis

Once the data have been corrected to isolate the drawdown response to the pumping from other sources of head changes, the data are ready for analysis with an appropriate solution.

The drawdown and recovery data from all observation wells should be analysed simultaneously using the method of superposition to combine the effects of both the pumping and recovery periods. As outlined in Lough (2004) the principle of superposition and time translation used by Jenkins (1968) enable analysis of recovery data. This is based on the principle that the recovery period can be modelled as continuous abstraction from the pumped well, with a recharge well superimposed in the same location which commences at the instant of pump shutdown and recharges continuously at the abstraction rate. The net abstraction from this time is zero but the effects of the two hypothetical pumps can be superimposed to give the recovery behaviour.

The reason that the analysis of drawdown and recovery data should be carried out simultaneously is that the recovery data provide useful information on how the system would have responded with continued pumping (and therefore potentially allows an interpretation of parameters not determinable from the response during pumping such as the specific yield and transmissivity of other strata).

An appropriate analysis tool should be used with carefully selected parameters to give a realistic match to the measured drawdown response. Section 7 presents available analytical tools. In general, the Hunt and Scott (2007) solution is considered the most widely applicable and useful analytical solution for analysing tests in leaky aquifers in Canterbury, although care must be taken not to “derive” parameters which cannot be justified based on the length of the drawdown record.

Even when there is no drawdown response at the water table, this is useful information as it guides the choice of parameters used to fit the drawdowns at the pumped aquifer. When a solution that allows for drawdown of the water table (e.g. the Hunt and Scott (2007) solution) is used to analyse drawdowns from a well in the pumped aquifer, the hydrogeological parameters chosen for the analysis must also predict zero drawdowns at the water table at the location of the observation well. This confirms the validity of those parameters. When using the Hunt and Scott (2007) solution, it is useful to plot on the same graph the drawdown and recovery data from an observation well in the pumped aquifer and also in a well in the water table aquifer. The theoretical drawdown and recovery data at both well locations should then be added to the graph.

A curve-fitting approach for the analysis, which involves manually changing the parameters to achieve the best match, allows the analyst to obtain insight into the parameter sensitivity. Alternatively, automated processes and alternative methods to estimate hydrogeological parameters from drawdown data are available for some models (such as those outlined in Johns *et al*. (1992), Johnson *et al*. (2001) and Trinchero *et al.* (2008)). Where there is sufficient time available to do so, it may be useful to compare the results obtained from a curve-fitting approach with some of these other analytical techniques.

If a good fit between the corrected data and modelled drawdowns cannot be achieved with parameter optimisation, then corrections may need to be applied or the potential for an alternate model to better describe the system investigated.

6.1.7 Recognition of parameter variability

Where significant uncertainty remains in the values of aquifer parameters, ranges in possible sets of hydrogeological parameters should be developed, leading to corresponding ranges in any longer term predictions.

7 Methods to assess pumping-induced leakage

This section of the report describes some of the analytical methods available to estimate hydrogeological parameters in leaky aquifer systems through the analysis of aquifer test data.

There are a number of models available to analyse transient drawdowns and predict longer term drawdown and groundwater flows within both layered systems and anisotropic systems. This section summarises analytical techniques currently available and briefly outlines situations where numerical modelling is required. A description of these models can also be found in Hunt (2008a) along with the mathematical equations. The assumptions and the appropriateness of each of the analytical techniques are presented together with a discussion on how the shape of the field-measured drawdown curves guides the choice of appropriate technique.

The models discussed in this section are those used to describe transient flow.

Assumptions common to the techniques discussed in the following sections, unless otherwise stated, are outlined here, with brief comments about the appropriateness of these assumptions in the gravel-dominated aquifers in Canterbury follows.

1. *The aquifers and aquitards are isotropic, homogeneous and uniform in thickness*

   Strictly speaking, this assumption is usually violated as most strata in Canterbury are anisotropic, heterogeneous and of variable thickness. However, the scale of the heterogeneity may be small relative to the volume affected by the test. This is demonstrated where analysis of data from a number of observation wells results in a similar set of hydrogeological parameters. For most constant-rate test analyses, the parameters derived are average parameters for the volume of aquifer affected, and, using these average parameters for longer term assessments will, where there is reliable drawdown data and analysis, provide a reasonable approximation of head and flow changes. Trinchero *et al*. (2008) illustrate the relationship between the estimated parameters and the volume of aquifer affected. Therefore this simplifying assumption will, in general, not significantly affect the quality of the analysis and predictions.

2. *The aquifers and aquitards are of infinite lateral extent*

   Aquifers and aquitards cannot be of infinite lateral extent. As described in Section 2, aquifers in Canterbury are bounded laterally by lower permeability material or the sea (a constant head boundary). Some aquifers are also partially bounded by surface water features such as rivers, which reduce the propagation of effects. Where the distances between the area of test influence and any boundaries are large over the course of a constant-rate test, this assumption is reasonable. Where the distances to boundaries are small, or longer term effects are being simulated, boundaries may have an influence and their effects must be accounted for by either using a model that incorporates these or using a technique such as the method of image wells.

3. *The base of the pumped aquifer is impervious (an aquiclude*)

   Again, this assumption will be violated over most of Canterbury. In the Canterbury Plains violation occurs when pumping from any aquifer, except the rare situations where aquifers overly an impervious layer such as volcanic or greywacke bedrock. By assuming the underlying layers are impervious, the release of storage from pervious layers and flow of this released water into the pumped aquifer is not accounted for. The effects of using a model for drawdown analysis where the underlying layer is in fact pervious, is that the derived elastic storage coefficient (S) will be larger than that for the pumped aquifer alone, as it will incorporate the effect of stored water released from below. In addition, the transmissivity (T) will be larger than that for the pumped aquifer alone, as it will incorporate the effects of horizontal flow within these underlying layers. Section 9.2 of this report will demonstrate through modelling that the violation of this assumption does not prevent useful drawdown analyses or predictions.

4. *All layers, other than the pumped aquifer, are incompressible (zero elastic storage)*

   This assumption is similar to the previous. For all realistic values of the elastic storages of the other layers, the effect will be a small delay in the time taken for pressure changes to reach an observation well, which will result in a small increase in the elastic storage coefficient of the pumped aquifer (as shown in the illustrations in Hunt and Scott (2007)). As will also be demonstrated in Section 9.2 of this report, this violation does not prevent useful drawdown analyses or predictions.

5. *Storage is released instantaneously with a decline in head*

Boulton (1955) showed that the release of stored water at the water table is not always instantaneous and the time delay between a decline in the water table level and the release of stored water via drainage from the material can be significant. It was for this “delayed yield” phenomenon that he developed his mathematical solution containing terms for both the elastic storage coefficient and the delayed yield (the specific yield term). Where the reduction in the level of the water table occurs very slowly, this assumption is valid. For example, when abstraction occurs from a leaky aquifer, the shallow water table falls slowly over a large area, so this assumption is likely to be reasonable for most leaky aquifer assessments.

6. *The diameter of the well is infinitesimally small such that storage of water within the well casing can be neglected*

The validity of this assumption can be assessed by calculating the change in the volume of water within the well casing during pumping and comparing this with the amount of water abstracted from groundwater before there is a response in the nearest observation well. Provided the change in the volume of water in the well is comparatively small, this is a reasonable assumption.

7. *The well is screened over the entire thickness of the pumped aquifer*

This assumption is appropriate provided the nearest observation well is located where flow within the pumped aquifer is predominantly horizontal. Monitoring wells should be located outside this zone of convergent flow if the model to be used for the drawdown analysis contains this assumption. Kruseman and de Ridder (1991) suggest that the effect of convergent flow is negligible at a distance greater than 1.5 to 2 times the saturated aquifer thicknesses, but show how this is greater where there is anisotropy on the vertical plane.

8. *A large conductivity contrast exists between the pumped aquifer and aquitard, which implies that flows in the pumped aquifer and aquitard are horizontal and vertical, respectively.*

This assumption is appropriate near the coast where the aquifers and aquitards are discrete entities with a high hydraulic conductivity contrast. This assumption is also appropriate for most other areas, as there generally is a large contrast between the horizontal hydraulic conductivity of Canterbury’s permeable gravel-dominated sediments and the vertical hydraulic conductivity of the interspersed gravels in a fine-grained matrix. In situations where the hydraulic conductivity contrast is small, it may be more appropriate to model the system as an anisotropic aquifer with a solution that accounts for both vertical and horizontal flow. An example of such a solution is the Zhan and Zlotnik (2002) solution, described in Section 7.4 of this report.

9. *Prior to pumping, the piezometric surface in the pumped aquifer and at the water table are horizontal*

Throughout most of the Canterbury Plains, the water table slope is so small that this assumption will not introduce significant errors in applications of the analytical equations described in this report.

7.1 The Hantush and Jacob (1955) solution

Following on from the Theis (1935) solution for confined aquifers, Hantush and Jacob (1955) developed a solution to account for leakage from an aquitard overlying a pumped aquifer. Their conceptual model is one of a pumped aquifer bounded on top by a low permeability aquitard, which lies beneath a more permeable aquifer containing a standing water table as shown in Figure 7-1.

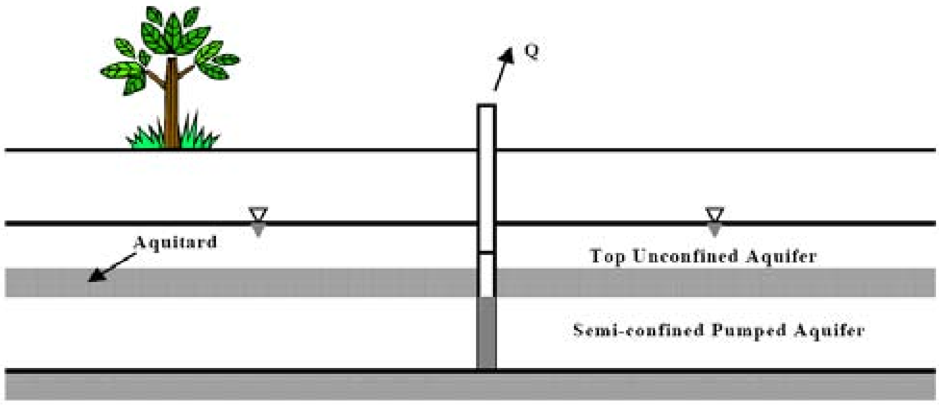


Figure 7-1: Conceptual model for the Hantush and Jacob (1955) solution (from Hunt (2008a))

The solution allows for a reduction in piezometric levels in the bottom aquifer and the resulting vertical piezometric gradient to create downward flow through the aquitard. The main limitation of the Hantush and Jacob solution is that it assumes that the level of the water table in the top unconfined aquifer remains constant over the course of the pumping, and therefore that there is an infinite amount of storage in the unconfined aquifer. This assumption may be appropriate where drawdown and recovery data from a pumping test are available only for the period before the water table is affected by the pumping but will not be appropriate for longer term analysis or predictions. Figure 7-2 illustrates the period of validity for the Hantush and Jacob solution (i.e. left of the vertical red dotted line). Note that if recovery data are available, one of the more advanced analytical solutions described here should be used for the simultaneous drawdown and recovery data and analysis.

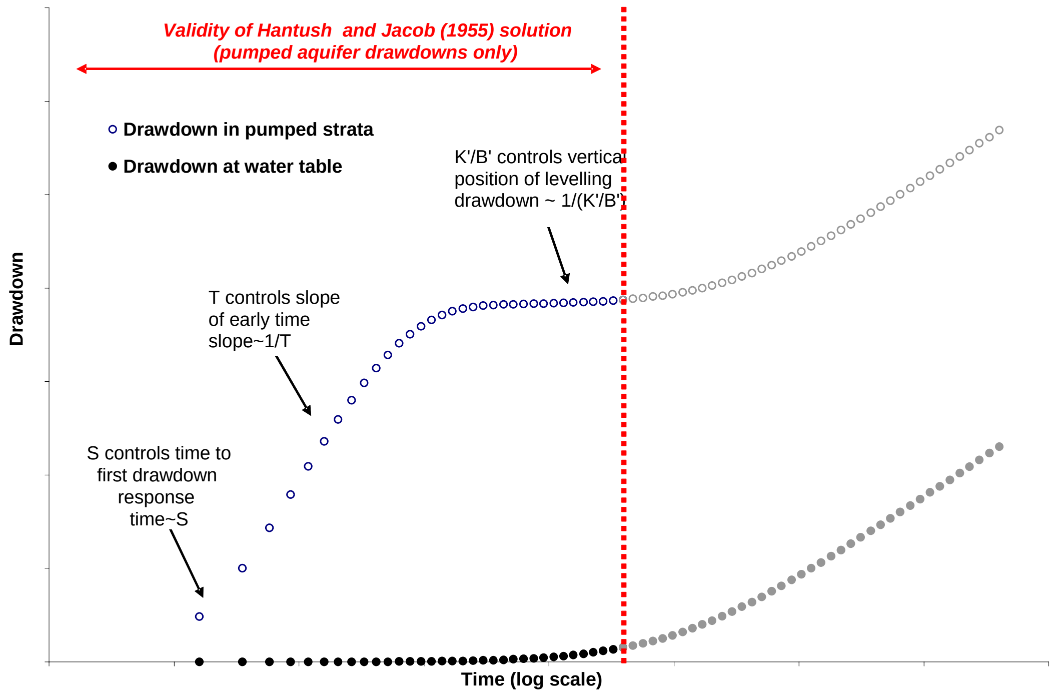


Figure 7-2: Period of validity for analysis of drawdown data with the Hantush and Jacob solution

Neuman and Witherspoon (1968) proposed a similar solution to the Hantush and Jacob solution, but also accounted for elastic storage in the aquitard. Analysis with this solution is outlined in Kruseman and de Ridder (1991) and is described as being applicable at small values of pumping time when the drawdown in the overlying unconfined aquifer is negligible.

The Hantush and Jacob (1955) and Neuman and Witherspoon (1968) solutions are not appropriate to use to assess long-term drawdowns in leaky aquifer systems because the assumption of an infinite amount of storage in the unconfined aquifer is not realistic when assessing long-term effects. However, other methods to assess long-term effects are only appropriate when values for all the parameters that control the effects can be reasonably defined.

7.2 The Boulton (1973) solution

As described in Hunt and Scott (2005), the delayed-yield solution for unconfined flow to a well was originally obtained from an equation proposed by Boulton (1954, 1963). A number of years later, Boulton (1973) and Cooley and Case (1973) showed that this solution can be interpreted as horizontal flow to a well in an aquifer underlain by an aquiclude and overlain by an aquitard containing a free surface.

The Boulton (1973) solution differs from the Hantush solution in that it allows for drawdown at the water table. The conceptual model for this solution is shown below:

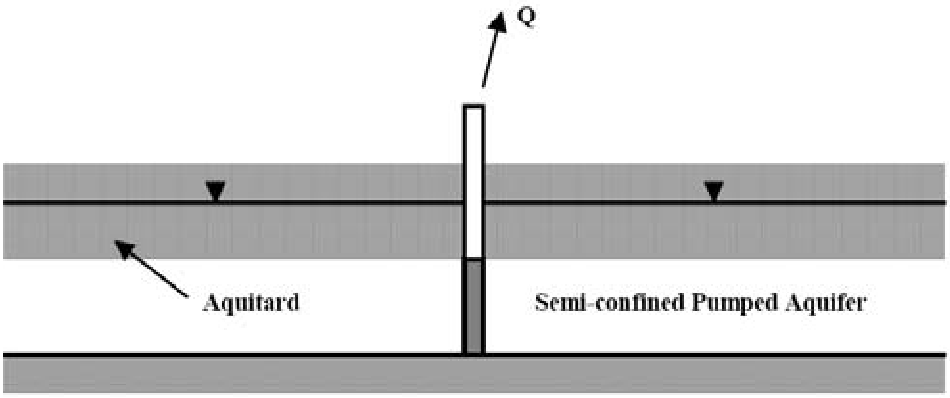


Figure 7-3: Conceptual model for the Boulton (1973) solution (from Hunt (2008a))

Hunt and Scott (2005) demonstrated that the Boulton solution also applies when the pumped aquifer is overlain by any number of layers provided that:

1. the top layer contains a free surface;
2. the elastic storage coefficient of the pumped aquifer is much less than the specific yield of the layer containing the free surface; and
3. none of the layers has a transmissivity that exceeds about 5% of the pumped aquifer transmissivity.

The Boulton (1973) solution is an improvement on the Hantush and Jacob (1955) solution as it accounts for drawdown of the water table. There is no difference between the two solutions in the period before the upward inflection of the drawdown curve for the pumped aquifer shown in Figure 7-2. The time at which the lowering of the water table occurs and corresponding drawdowns in the pumped aquifer depart from their pseudo-steady state values is controlled by the specific yield. Although most pumping tests in Canterbury have not been carried out for long enough to see the drawdown at the water table or the later increase in drawdown in deeper wells, a lower limit for this parameter can still be determined usefully from drawdown analysis.

The Boulton solution is much more appropriate than the Hantush and Jacob (1955) solution for long-term simulations because of the inclusion of water table drawdown; however, specific yield values will need to be estimated based on the type of material containing the water table if a constant-rate test has not been carried out for a sufficient length of time.

The Boulton solution describes the behaviour of an aquifer system consisting of a pumped aquifer overlain by either a single aquitard or a series of aquitards containing a standing water table. The various phases of this behaviour are as follows:

- For a small period after pumping from the well commences, the leaky confined pumped aquifer behaves as a fully confined aquifer as shown in the straight line section of the early drawdown response in Figure 7-2 (for which the drawdown response is described by Theis (1935)).
- Following this initial phase, a period follows where flow is induced through the overlying aquitard (or aquitards) to recharge the pumped aquifer as shown in the pseudo-steady state section of the drawdown response in Figure 7-2. This phase is incorporated into the Hantush and Jacob (1955) solution.
- As pumping from the well continues, a third phase reached in the system response, in which free surface drawdowns in the top layer begin to increase and, ultimately, approach the same values that occur in the pumped aquifer. During this third phase, vertical leakage decreases and drawdowns in the pumped aquifer again approach values predicted from the Theis (1935) solution when the elastic storage coefficient is replaced with the specific yield of the material containing the water table (Hunt, 2008a). Note that for a system where horizontal flow occurs in the overlying layers the time-drawdown curves will not come together, and vertical flow will therefore continue (as shown in Figure 7-2).

The main limitation with the Boulton solution is that it is not applicable where other layers have a transmissivity of more than 5% of the pumped aquifer. This is because the solution cannot model the occurrence of horizontal flow in overlying strata. In a groundwater system such as the Canterbury Plains, there are a number of water bearing zones (aquifers) that are likely to invalidate the assumption of zero horizontal flow in the overlying layers. The effects of flow within these can be accounted for in the Hunt and Scott (2007) solution, described below.

7.3 The Hunt and Scott (2007) solution

Hunt and Scott (2007) introduced a new analytical solution allowing for horizontal flow within an unconfined aquifer overlying a leaky confined aquifer. The conceptual setting is equivalent to that shown for the Hantush and Jacob (1955) solution (Figure 7-1). The solution differs in that it allows for both drawdown and horizontal flow within the unconfined aquifer. The solution reduces to the Boulton (1973) solution where the horizontal flow in the overlying aquifer is negligible.

The following terms appear in this solution and are illustrated in Figure 7-4, which shows an underlying leaky confined pumped aquifer separated by an aquitard from the overlying unconfined aquifer.

T = transmissivity of pumped aquifer

S = elastic storage coefficient

K'/B' = aquitard conductance (ratio of hydraulic conductivity of overlying aquitard, to thickness of the aquitard)

S_y = specific yield of the unconfined aquifer

T_0 = transmissivity of overlying aquifer

s = drawdown in pumped aquifer

η = drawdown in overlying unconfined aquifer

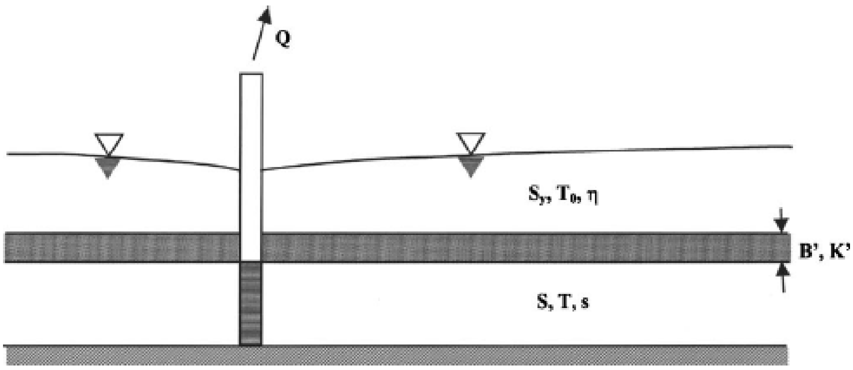


Figure 7-4: Terms used in the Hunt and Scott (2007) solution (from Hunt and Scott (2007))

Figure 7-5 illustrates the effects these aquitard and aquifer parameters have on the drawdown in each of two aquifers when the leaky confined aquifer shown in Figure 7-4 is pumped.

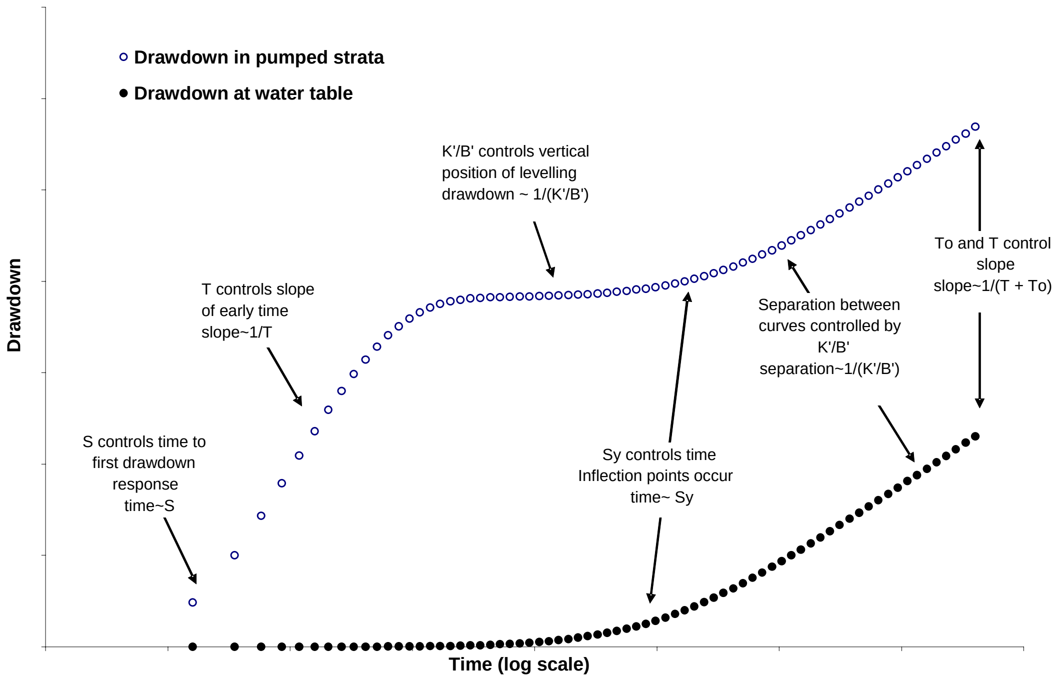


Figure 7-5: Drawdown sensitivity to parameters of a system with two aquifers separated by an aquitard

As described in Hunt (2008a), programs have also been written to calculate both the volume of water sourced from the pumped aquifer and the corresponding volume of water depleted from storage at the water table. Scott and Hunt (2007) illustrate this process.

Figure 7-6 - 7-8 demonstrate the influence parameters have on the calculated loss of storage caused by a decline in the water table.

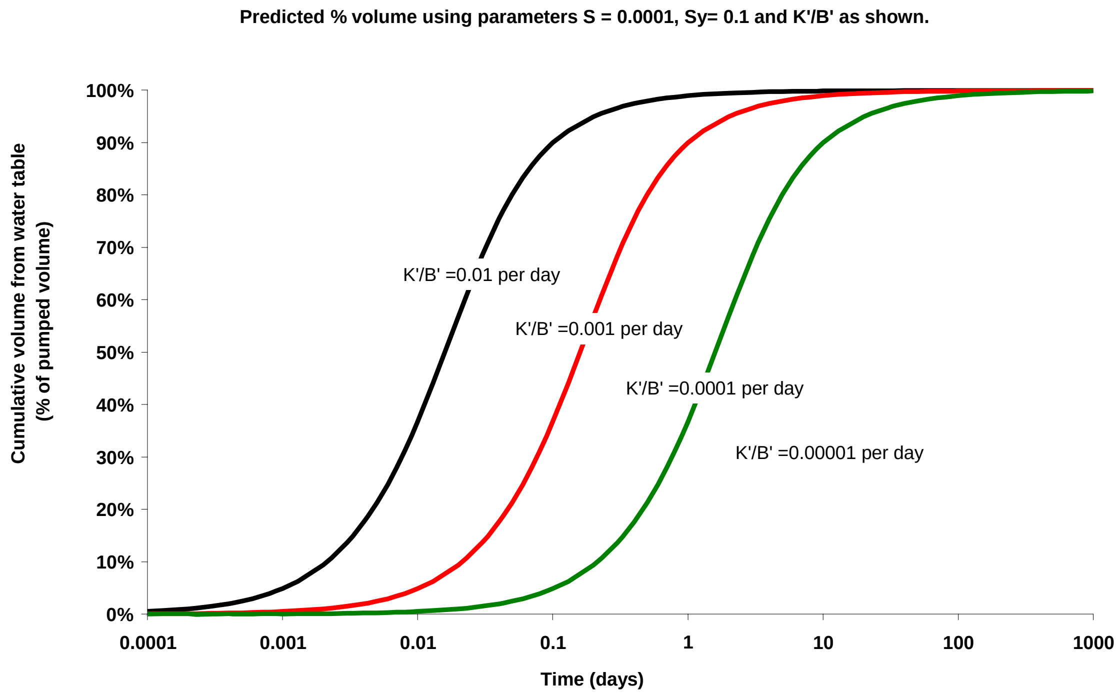


Figure 7-6: Influence of aquitard conductance on volume released from water table drainage (using volumetric extension of the Hunt and Scott (2007) solution)

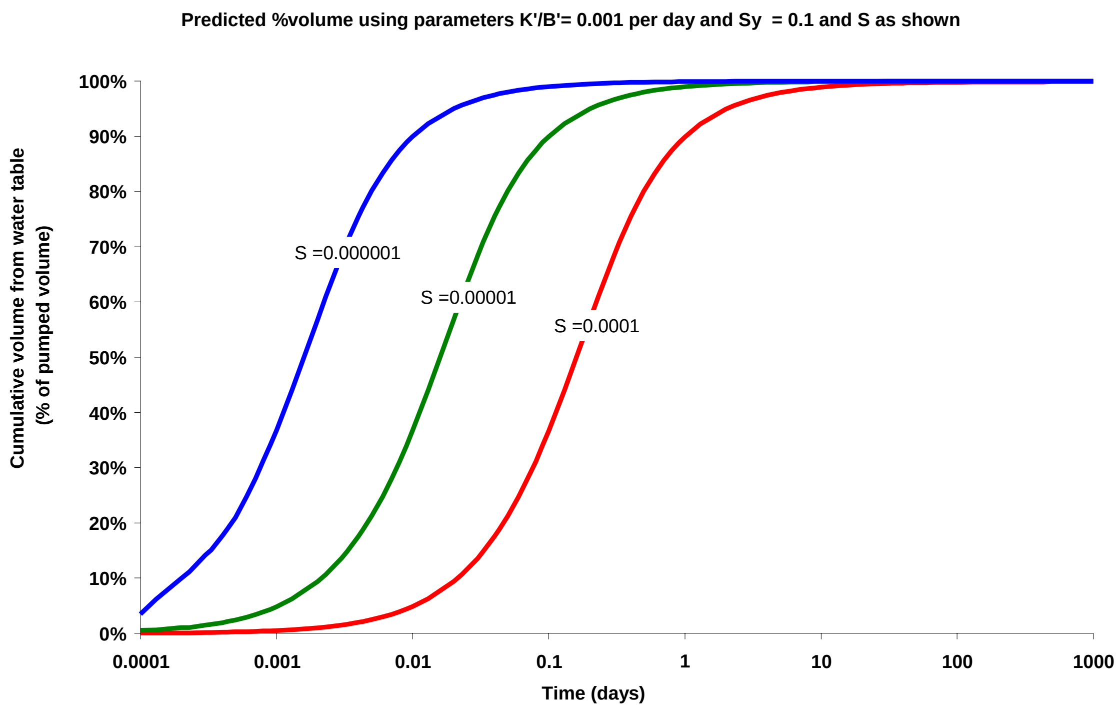


Figure 7-7: Influence of elastic storage coefficient on volume released from water table drainage (using volumetric extension of the Hunt and Scott (2007) solution)

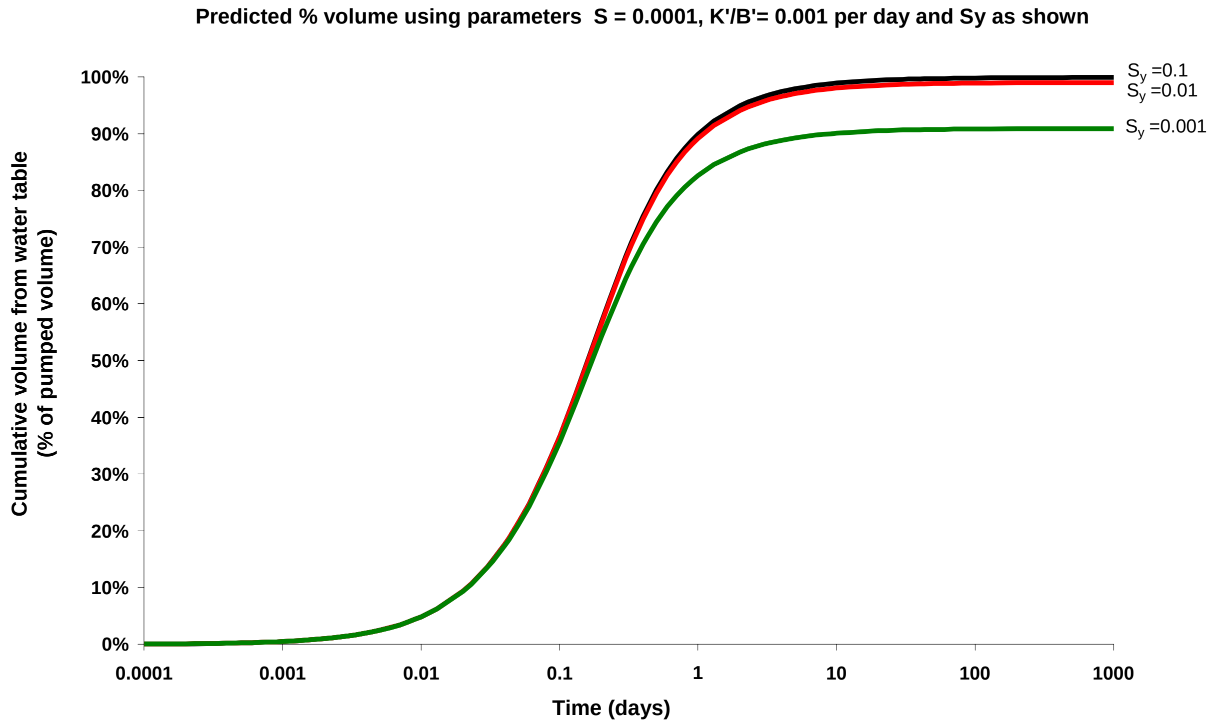


Figure 7-8: Influence of specific yield on volume released from water table drainage (using volumetric extension of the Hunt and Scott (2007) solution)

Figure 7-6 – 7-8 indicate that, regardless of the hydrogeological parameters used, after a sufficient period of groundwater abstraction the volume of drainage at the water table predicted with this solution approaches the pumped volume, and, at later times, becomes insensitive to the parameter values (within realistic ranges). The influence of transmissivity on the volume released from water table drainage has not been illustrated because, in this solution, the volume released is in fact entirely independent of transmissivity.

While these examples are presented for a two-aquifer system, Figures 7-6 - 7-8 also describe the predicted volume release in any multi-aquifer system with an effective aquitard conductance equal to those shown, and with the same contrast between the elastic storage coefficient of the groundwater system and the specific yield at the water table. Therefore, the solution provides a reasonable representation of the Canterbury Plains groundwater system where there is greater resistance to vertical flow than horizontal flow (even where there are no clearly discernible aquitards).

As will be shown later, the Hunt and Scott (2007) solution, while derived for a two-aquifer system, can also provide realistic estimates of pumping-induced leakage for multi-layered and anisotropic systems.

The Hunt and Scott (2007) solution is considered to be the most realistic current analytical solution for the analysis of drawdown data and predictions of long-term effects for most locations within Canterbury (unless the connection to surface water bodies is significant or the setting requires numerical modelling) because it allows for horizontal flow in overlying strata. However, in tests where there are insufficient data to see the complete drawdown curve shown in Figure 7-5, and therefore to determine all the parameters contained in this solution, this should be noted in the analysis. Values for these parameters should not be reported as determinable from the test, although it would be of use to assess a minimum value for the specific yield.

For long-term predictions using this solution, values for parameters beyond those determinable from a pumping test need to be assigned with care. In some instances, tests would need to be run for longer than 10 days under stable background water level conditions to determine values of the specific yield and transmissivity of overlying strata. This is generally not practicable.

7.4 The Zhan and Zlotnik (2002) solution

The characteristic drawdown response shown in Figure 5-3 can also occur when pumping occurs from a well screened at depth within an unconfined, anisotropic aquifer. While the Hunt and Scott (2007) solution can be used in this setting, there are alternate models for which the equations more closely match a conceptual model of an unconfined, anisotropic aquifer.

As outlined in Section 3.2, an aquifer/aquitard system consisting of alternating layers of different properties, is equivalent in its behaviour to an anisotropic aquifer over a large scale. Where there is very fine layering, the depth interval over which a layered aquifer may be considered as a single anisotropic aquifer can be quite small. In addition to layering, anisotropy may be caused by a preferred orientation of solid particles in the aquifer matrix. For either case, vertical and horizontal bulk hydraulic conductivities do not change with either depth or horizontal position within the aquifer.

Hunt (2008a) outlines the equations that govern drawdowns at any particular location (with any x, y, z co-ordinates) within a single, unconfined, homogeneous anisotropic aquifer in response to pumping from a well screened over a set distance. These equations account for partial penetration of wells. A solution obtained by Zhan and Zlotnik (2002) for an unconfined, homogeneous anisotropic aquifer is used by Hunt in his Excel spreadsheet software Function.xls [3].

The Zhan and Zlotnik (2002) solution is, strictly speaking, more appropriate than the Hunt and Scott (2007) solution for pumping test analysis and long-term predictions of drawdown responses in a system that can be considered as a single anisotropic aquifer, as it was developed specifically to model drawdown in this setting.

7.5 The Hunt (2003) stream and Hunt (2004) spring depletion solutions

The solutions presented in Sections 7.1 though to 7.4 do not account for groundwater-surface water interaction. Where there is both pumping-induced leakage from bounding layers and changes in groundwater interaction with a surface waterway over the interval of assessment, a solution that incorporates these is usually more appropriate to use for drawdown analysis or for longer-term predictions.

The Hunt (2003) solution allows for the depletion that results from a stream when water is abstracted from a well in a leaky confined aquifer. This is essentially equivalent to the Boulton (1973) solution, but includes the effects of stream depletion. Stream depletion is either a reduction in groundwater flow to a stream or an increase in flow from a stream to groundwater.

Lough and Hunt (2006) demonstrated that the Hunt (2003) solution is applicable to a setting where the pumped aquifer is overlain by multiple layers, rather than a single aquitard, although this solution does not incorporate horizontal flow in overlying layers. Hunt (2008b) provides a solution for a semi-confined aquifer of finite width.

Where a leaky aquifer is hydraulically connected to a spring, both drawdowns at the water table and flow to the spring will be reduced by pumping. The decrease in flow to the spring is a process known as spring depletion. Where there is both pumping-induced leakage and spring depletion over the interval of assessment, these processes need to be accounted for. The Hunt (2004) spring depletion solution allows for both processes to occur.

This solution is essentially equivalent to the Boulton (1973) solution, but includes the effects of depletion from the spring. Like the Boulton solution, the Hunt (2004) spring depletion solution is applicable to a setting with any number of aquitard layers overlying a pumped aquifer, provided that the transmissivity of each layer does not exceed about 5% of the pumped aquifer transmissivity, and the elastic storage coefficient of the pumped aquifer is much less than the specific yield near the free surface (this principle is outlined in Hunt and Scott (2005)).

Improvements to the Hunt (2004) spring depletion solution are outlined in Hunt and Smith (2008).

[3] Function is a spreadsheet downloadable from the website: http://www.civil.canterbury.ac.nz/staff/bhunt.asp

7.6 Other available analytical models incorporating pumping-induced leakage effects

Amin (2005) describes a method to calculate the rate of leakage into a pumped aquifer. Consent conditions set by the commissioners on groundwater takes within the Rakaia-Selwyn and Selwyn-Waimakariri allocation zones (Environment Canterbury 2008a, 2008b) require assessments of leakage using a method that is essentially the same.

This method appears to be erroneous in that it essentially uses well superposition principles to calculate a net flow rate from a well in a leaky confined aquifer (the pumping rate minus the leakage rate). This is an incorrect approximation as leakage into an aquifer does not occur at a single point and therefore the effects of this cannot be modelled with a radial flow solution for a point source.

The equation also does not account for time-varying leakage, and therefore, should not be used. The preferred method for leakage volume calculations is the extension of the Hunt and Scott (2007) solution, as described in Hunt (2008a) and programmed in Function.xls using the function Eta_Volume_11.

7.7 Numerical models

Where the analytical solutions discussed in the preceding sections can not reasonably describe pumping-induced flow and level changes within a groundwater system, due to complexities such as variable distances to boundaries or multiple locations of groundwater-water surface water interaction, a numerical model should be used.

There is a wide range of numerical models available for simulating groundwater flow. It is most important that a properly verified modelling code is used and that any modelling simulation is presented with appropriate checks to confirm its validity (Pattle Delamore Partners Ltd and Environment Canterbury, 2000).

It is beyond the scope of this report to give detailed instructions on how to set up a numerical model to assess pumping-induced flow between strata at different depths. Guidelines to assess the validity and uncertainty of predictions using numerical modelling are outlined in Pattle Delamore Partners Ltd (2002).

7.8 Regional scale modelling

The analytical models discussed above are most useful when either analysing drawdown data or assessing the effects from a single groundwater abstraction.

Other techniques are more appropriate for assessing leakage effects on a regional scale. In some instances numerical modelling is appropriate.

For modelling as part of water allocation assessments, tools such as the eigen-value modelling described by Bidwell (2003) and net water budgeting as used by Aitchison-Earl *et al.* (2004) may be more appropriate than superimposing the volumetric effects of each abstraction using an analytical solution such as Hunt and Scott (2007).

Information from the analytical analysis of pumping test data within the area of interest can provide useful information on how connected the system is, and therefore, whether it is appropriate to manage aquifers as being interconnected across a range of depths.

7.9 Recommended models

Analytical equations are useful tools to process data from aquifer tests. They can provide relevant information on aquifer behaviour, provided the data are reliable and an appropriate model is used for the analysis. The model results are only as good as the test data used in them.

Analytical equations can also be useful for longer-term simulations provided their assumptions remain appropriate for the longer time period and there is sufficient information to assign values to the parameters contained in the equations.

For most assessments of drawdown and volumetric changes in the leaky aquifers of Canterbury, the Hunt and Scott (2007) solution, and its subsequent advancements, is the most appropriate analytical solution currently available, although it does not allow for changes in the recharge and discharge components of the system. The Zhan and Zlotnik (2002) solution may be more appropriate for drawdown assessments where modelling a particular system as a single anisotropic aquifer.

More advanced analytical models that incorporate both pumping-induced leakage and groundwater-surface water interaction, such as the Hunt (2003) and Hunt (2004) solutions, are appropriate where both these processes occur over the period of the assessments. Numerical modelling may be required for more complex situations.

8 Pumping-induced leakage effects in Canterbury

This section outlines the parameters interpreted from pumping tests carried out in Canterbury in which leaky aquifer behaviour was observed.

8.1 Leakage data

Figure 8-1 shows the location of 265 constant-rate aquifer tests for which Environment Canterbury holds records. Of these tests, 141 have reported aquitard conductance values (K'/B') or leakage factors (L), indicating that leakage effects were observable over the course of the testing.

A sub-set of these tests classed by Environment Canterbury as having reliable data and analysis are shown on a separate map (Figure 8-2) along with their reported aquitard conductance values (of which there are 65). Also included on this map are five "maximum possible aquitard conductance values". These values were derived from re-analysing test data where previous analysis had indicated a confined drawdown response (no significant leakage) over the test duration, and where the tests were of more than one day in duration.

None of the "maximum aquitard conductance values" for these five re-analysed tests was outside the range of derived aquitard conductance values for the tests in which leakage was noted. This does not demonstrate that values of aquitard conductance lower than those shown on the map do not occur in Canterbury. Rather it demonstrates that the available tests on Environment Canterbury's database have not been carried out in locations where lower values occur, or have not been carried out for a sufficient period to determine those lower values (if only the initial "Theis response" was observed).

It is difficult to determine any spatial patterns of aquitard conductance values within Canterbury. In inland Canterbury between the Waimakariri and Rakaia rivers, there are a number of tests with aquitard conductance values in the order of 10^{-4} per day. These wells are located at differing depths; data from similar depths would be required to draw conclusions over any patterns in aquitard conductance values.

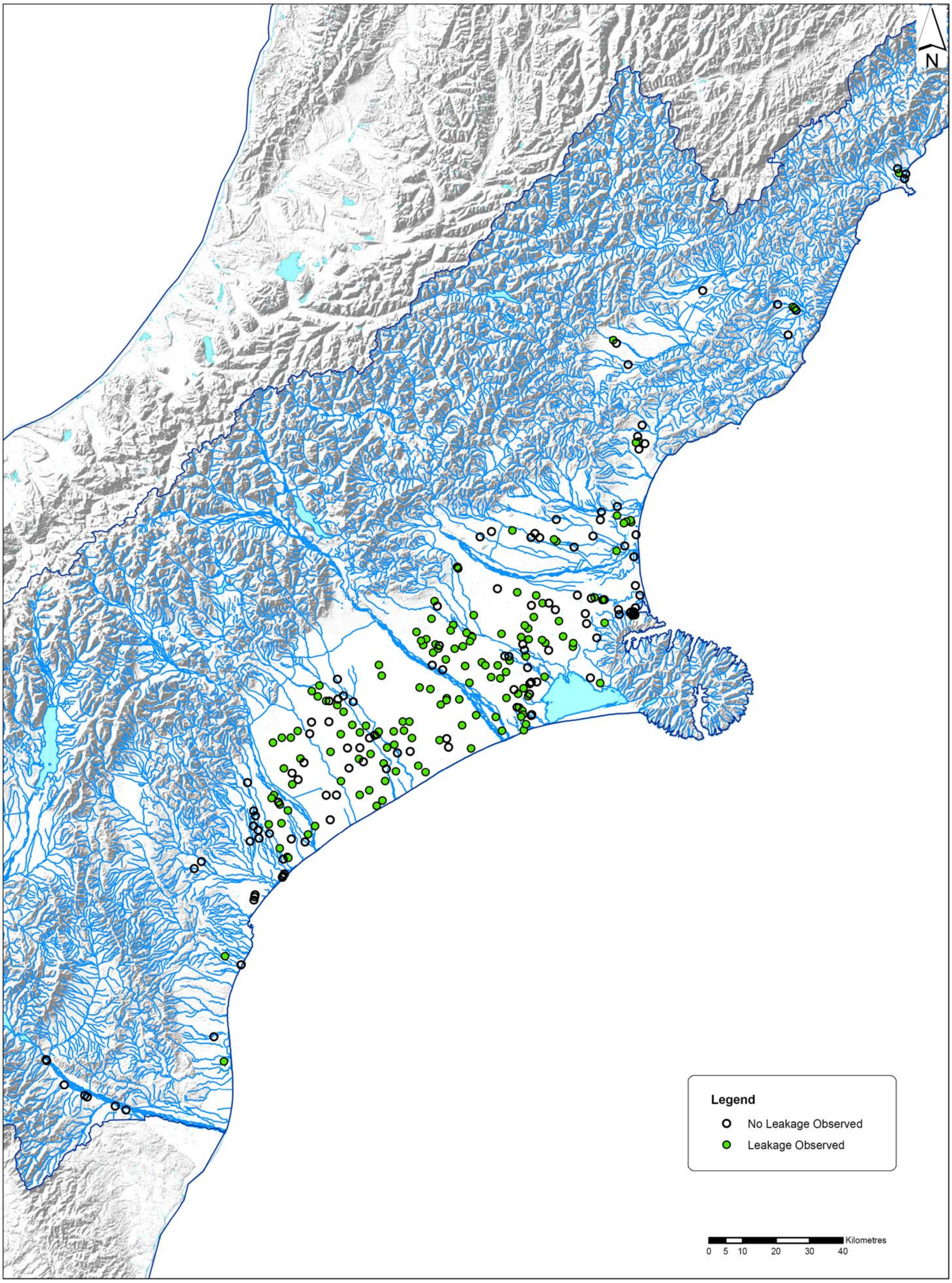


Figure 8-1: Constant rate tests recorded on Environment Canterbury database both with and without observed leakage

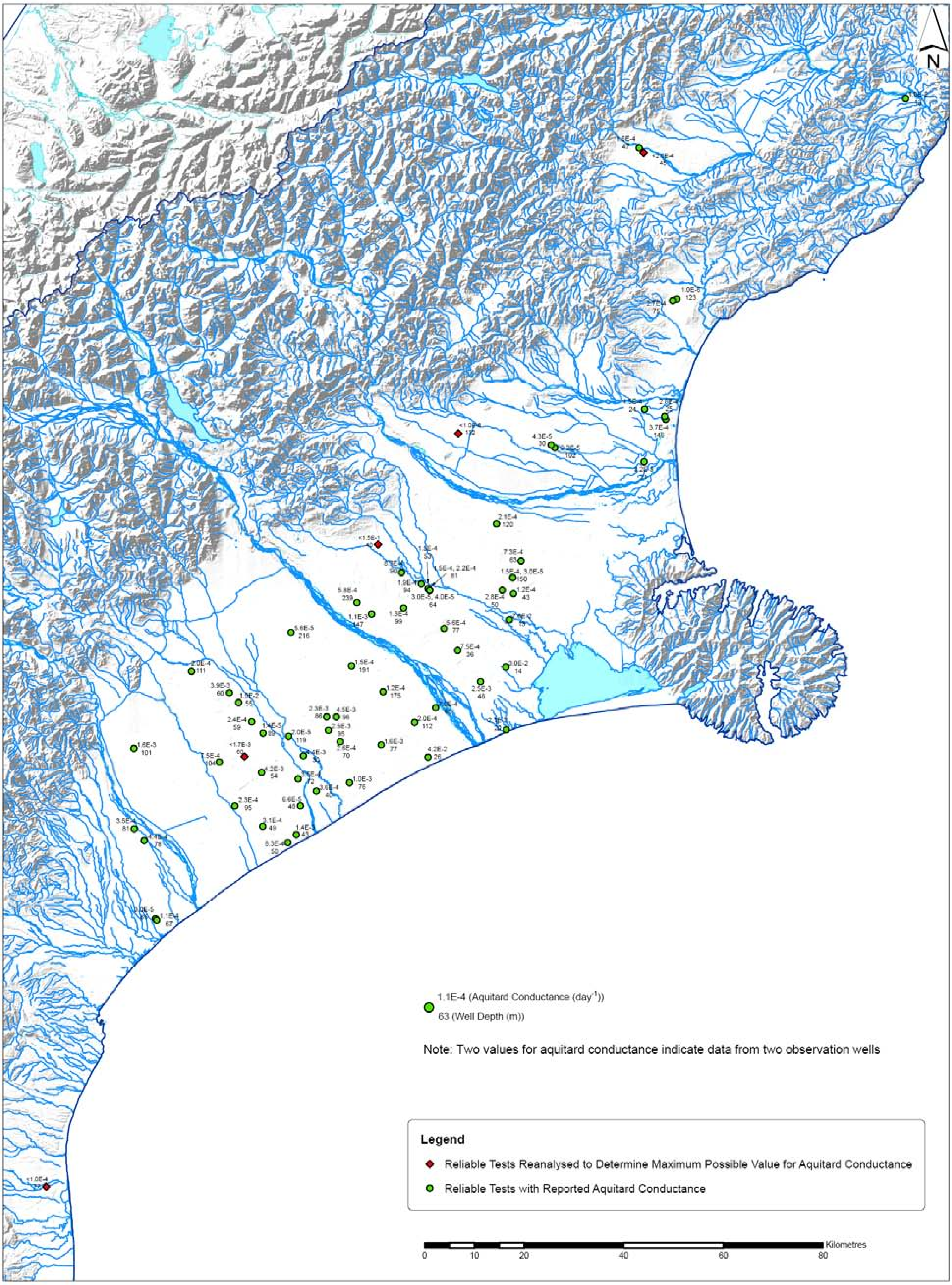


Figure 8-2: Reliable constant rate tests from Environment Canterbury records

Figure 8-3 is a plot of reported aquitard conductance values versus the total saturated thickness above the well screens for tests classed as reliable by Environment Canterbury.

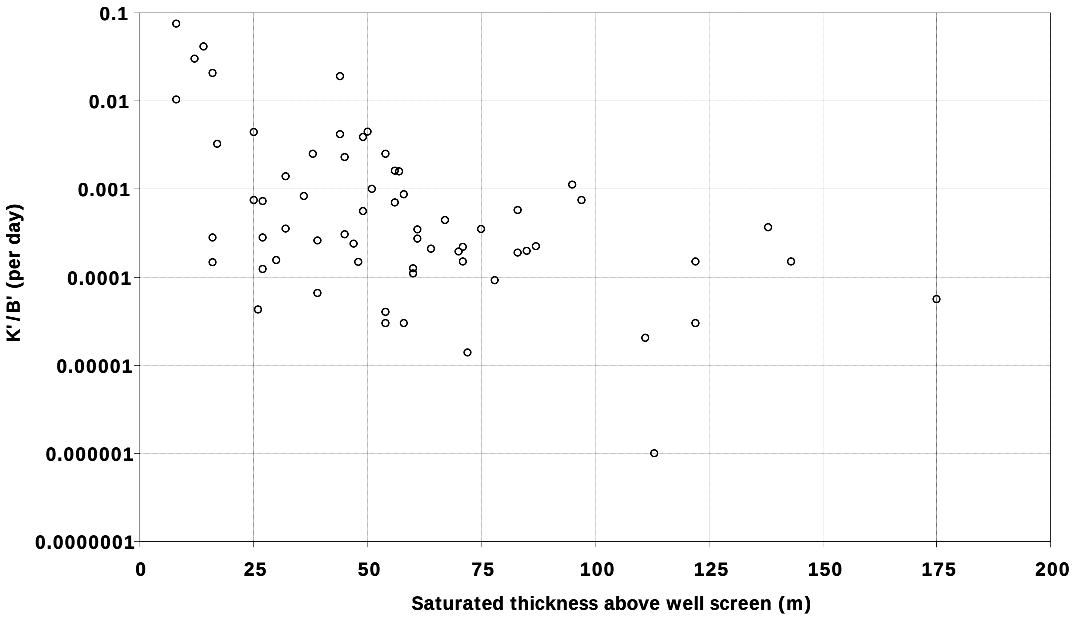


Figure 8-3: Reported aquitard conductance values for tests classed by Environment Canterbury as being reliable (reliability ≤ 2, parameter accuracy ≤3, single screen)

Figure 8-3 shows a general decrease in reported aquitard conductance values with the depth of the well screen below the water table (note that this is plotted on a semi-log scale and the scatter is over two orders of magnitude). This relationship is logical, because as the depth of saturated strata between the water table and the pumped aquifer increases, the value for aquitard conductance (K'/B') decreases, and there is an increase in the resistance to vertical flow from the shallowest strata. With the exception of the value from one test, none of these reported aquitard conductance values are lower than 10^{-6} per day. These are reported values so it can not be concluded that lower values do not exist in Canterbury. Based on the data collected so far, aquitard conductance values in Canterbury will likely be larger than 10^{-6} m/day. In Figure 8-4 the aquitard conductance values shown in Figure 8-3 have been converted to effective vertical hydraulic conductivity values, which provide a measure of the net vertical hydraulic conductivity of all the deposits that restrict vertical flow to the pumped aquifer. This conversion has been made by estimating the saturated thickness of the overlying deposits in the vicinity of the pumped well. The aquitard conductance values (Figure 8-3) were then multiplied by the saturated thickness values to estimate the effective vertical hydraulic conductivity at that location. Appendix A records the well details and aquifer properties included in the assessment.

The resulting effective vertical hydraulic conductivity values plotted in Figure 8-4 are much lower than typical horizontal hydraulic conductivities for gravel aquifers. Kruseman and de Ridder (1991) suggest a range of 100 to 1000 m/day for gravels, similar to Canterbury values estimated in NCCB (1983).

This comparison between the effective vertical hydraulic conductivity estimates and horizontal hydraulic conductivity estimates for the Canterbury Plains groundwater system suggests that this system, over its entire saturated thickness, can be considered to behave as an anisotropic system in which there is greater resistance to vertical flow than horizontal flow.

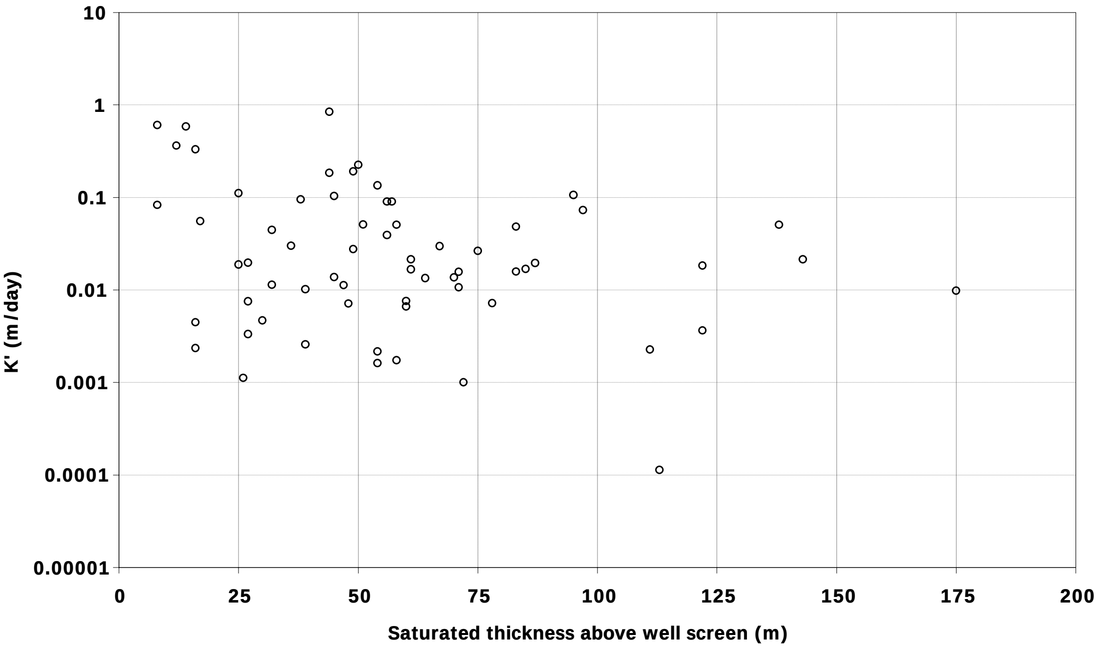


Figure 8-4: Effective vertical hydraulic conductivity values as calculated from aquitard conductance values and saturated thickness estimates for tests classed by Environment Canterbury as being reliable (reliability ≤2, parameter accuracy ≤3, single screen)

8.2 Predictions of volumetric changes

The parameters from each of the wells with reported aquitard conductance values from pumping test analyses (classed as reliable), have been used to assess the effects of pumping from these wells on the volume of water contained in the shallowest strata (in an overlying shallow unconfined aquifer for example). The predictions have been made using the volumetric extension of the Hunt and Scott (2007) solution, which can be used to calculate the volume of water depleted from storage at the water table over a defined period of groundwater abstraction.

Sections 5.4 and 7.3 show that the volume of storage released from water table drawdown is not sensitive to transmissivity values but is sensitive to values of the elastic storage coefficient, aquitard conductance and specific yield. The aquifer tests shown in Figure 8-2 have reported elastic storage coefficient and aquitard conductance values. Values for the specific yield of the material containing the water table are not contained in the database, which suggests that the tests may not have been carried out for a sufficient period of time for one to be determined. NCCB (1983) suggests a range of 0.1 to 0.3 in Canterbury, although values lower than these have been reported for some tests e.g. Lough and Hunt (2004). A value of 0.1 was assigned to the specific yield for this modelling exercise. The calculated change in the volume of water within the shallowest strata is shown at 7 and 150 days in Figure 8-5 (for those tests shown in Figure 8-2).

The predictions using the volumetric extension of the Hunt and Scott (2007) solution show that pumping from any of these locations and depths across Canterbury will create a depletion of the volume of water stored in the shallowest aquifers that is a very significant percentage of the pumped volume, if there is no change to the recharge and discharge components of the aquifer system.

Given the large available volume of stored water per volume of saturated strata at the water table and the potentially large extent of the drawdown cone caused by a deep pumping well, a large proportion of the well pumping rate can often be released from water table drainage with only a very small change in shallow water levels.

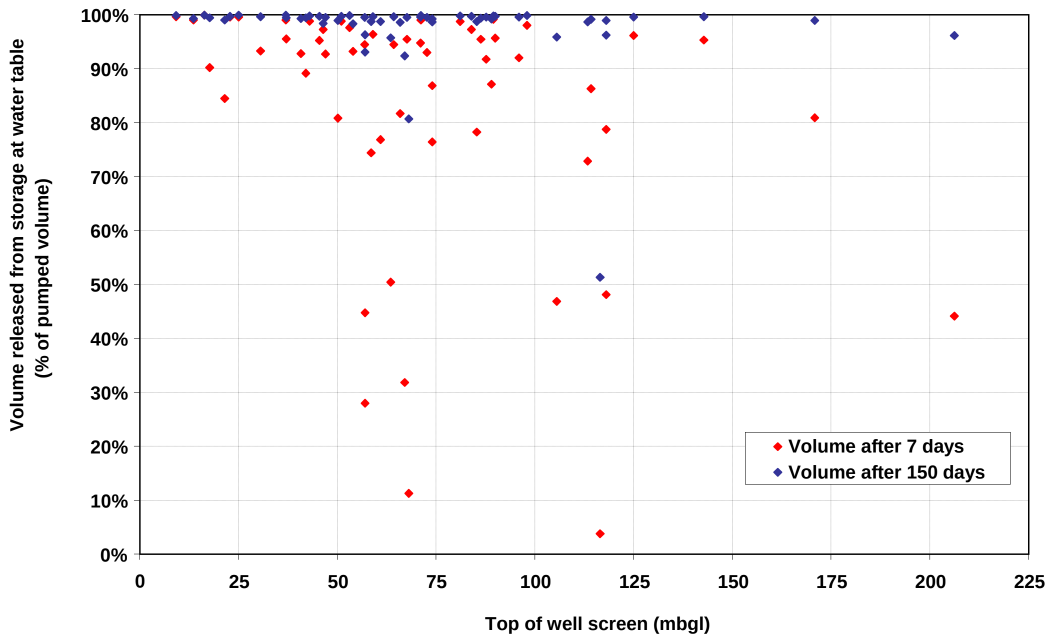


Figure 8-5: Predicted volume of storage release from water table drawdown (as % of pumped volume) for the available Canterbury aquifer test analyses

While the above figures show that the volume of pumped water is sourced from drainage of the water table, the changes in piezometric heads at different depths within the groundwater system will differ significantly depending on the depth of the pumped well.

The reduction in heads, rather than the reduction in volume, is important when assessing drawdown effects in neighbouring wells or stream depletion, as these effects are dependent on the magnitude of the change in heads.

Groundwater abstraction from anywhere will reduce the volume of water at the water table and this will ultimately transpire either as a reduction in discharge from the system or an increase in recharge to the system (Theis 1957). The location of these effects is determined by the hydrogeological setting and the spatial location of the abstraction.

9 Simulation and evaluation of pumping-induced leakage

This report has outlined the use of analytical solutions to interpret aquifer test data to assess the presence of leakage from strata other than the pumped aquifer.

This section describes numerical modelling that has been carried out to investigate the effects of pumping from different depths within a multi-layered system. It also provides details of an investigation used to assess whether the Hunt and Scott (2007) solution can be applied to settings different to those of the conceptual model on which it was based and, therefore, whether it is a useful tool to use in the assessment of drawdown and flow effects within Canterbury's groundwater systems.

9.1 Numerical simulations of pumping-induced leakage in different settings

Numerical modelling has been carried out to investigate the effects on drawdown and flow of pumping from different depths within a multi-layered system. This modelling has also been used to assess whether pumping from layered and non-layered systems can produce similar results.

The MODFLOW model simulations are not intended to represent the Canterbury Plains groundwater system in particular, or any other groundwater system in Canterbury. Rather the intention is to provide an insight into the behaviour of leaky aquifers, which are present throughout the Canterbury Plains, and to investigate the applicability of analytical models to such systems.

The model simulations consist of homogeneous, laterally extensive, uniformly thick layers. This is a simplification of those encountered within the complex heterogeneous Canterbury Plains groundwater system, where strata are discontinuous, indistinct and of very variable thickness. However, as outlined in Section 2, the net effect of the fluvial processes responsible for the sedimentary deposits on the Canterbury Plains is a system that contains stratification and has a greater resistance to vertical flow than horizontal. Therefore, this simple modelling exercise provides insight into the effects that will occur within the real world.

Modelling results presented here illustrate both drawdown and flow distributions in a leaky aquifer system.

9.1.1 Model set-up

The USGS MODFLOW (McDonald and Harbaugh, 1988) model used for this investigation was based on the eight layer model used to assess the validity of the analytical solution derived in Hunt and Scott (2007) with an additional six layers to accommodate a third aquifer.

The model is constructed to allow compressible flow in aquitards together with vertical and horizontal flow in aquitards and in aquifers.

The model comprises a 185 by 185 grid with general head boundaries specified in Columns 1 and 185 of the 7th and 8th layers. The central 81 rows by 81 columns are spaced at 10 m, with the grid spacing increasing geometrically by a factor of 1.1119 beyond this central zone to produce a maximum spacing of approximately 2.5 km. Figure 9-1 shows a plan view of the finite-difference grid.

Transient flow simulations were run with the initial starting head for all cells set to 20.0 m. The pumping rate at a centrally located well was set at a constant rate of 4320 m^3/day.

While hydrogeological parameter values are not intended to represent any particular area of the Canterbury Plains, they are comparable with values estimated from tests in the Canterbury Plains presented in NCCB (1983) and Sinclair Knight Merz (2008 and 2009) and shown in Figure 8-4 of this report.

9.1.2 Simulations

The first four simulations investigated leakage at different depths in a three aquifer/two aquitard system. The units represented in this system and their respective properties are shown in Table 9.1.

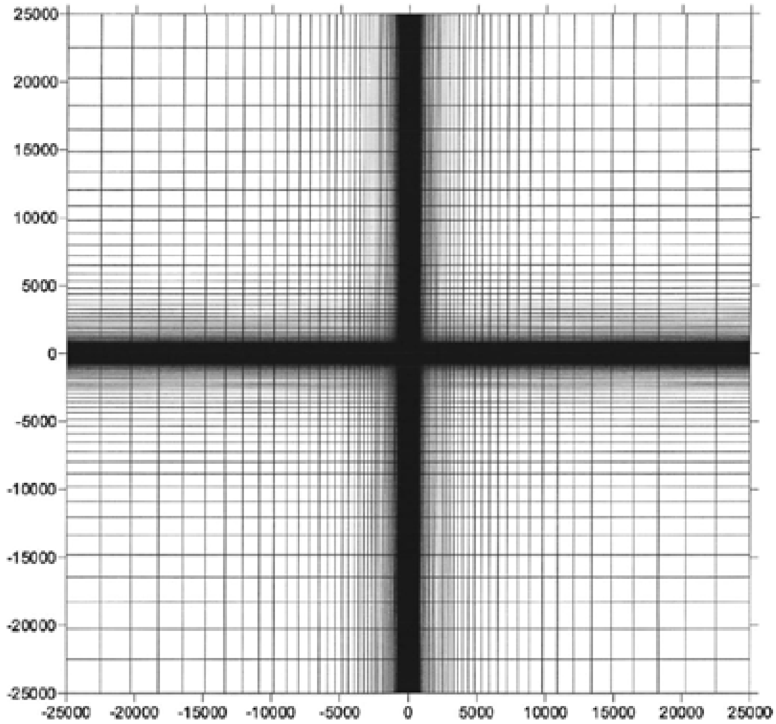


Figure 9-1: Plan view of the finite-difference grid for the MODFLOW model with coordinates in metres from the central pumped well (from Hunt and Scott 2007)

The fifth simulation involved modelling the entire system as a single anisotropic, homogeneous unconfined aquifer that has effective values equivalent to those for a layered system, as shown in Table 9.2.

The simulations are summarised as follows:

Simulation 1: 3 aquifers, 2 aquitards, top aquifer unconfined, pumping from deep leaky confined aquifer, each layer isotropic, homogeneous. No aquitard storage.

Simulation 2: 3 aquifers, 2 aquitards, top aquifer unconfined, pumping from deep leaky confined aquifer, each layer isotropic, homogeneous. Aquitard storage.

Simulation 3: 3 aquifers, 2 aquitards, top aquifer unconfined, pumping from shallow leaky confined aquifer, each layer isotropic, homogeneous. Aquitard storage.

Simulation 4: 3 aquifers, 2 aquitards, top aquifer unconfined, pumping from unconfined aquifer, each layer isotropic, homogeneous. Aquitard storage.

Simulation 5: 1 anisotropic, homogeneous unconfined aquifer, pumping from lower model layers, each model layer anisotropic, homogeneous, storage for all layers.

9.1.3 Model results

The model results are summarised in the figures on the following pages. The difference in the model results between simulations 1 and 2 were not significant except at initial time steps where a slight difference occurred in the timing of drawdown response due to an overall increase in the storage of the system. Therefore only the results from Simulation 2 are shown.

Table 9.1: Layer properties for MODFLOW modelling for first four simulations

Model layers	Represented unit	Total thickness (m)	Horizontal hydraulic conductivity (m/day)	Anisotropy ratio (Kv/Kh) (-)	Specific storage (m^{-1})	Specific yield (-)
1 -2	Unconfined aquifer	5	144	1	0.00001	0.1
3 -6	Aquitard	5	0.0144	1	0.00001	0.1
7-8	Shallow leaky confined aquifer	10	144	1	0.00001	0.1
9-12	Aquitard	5	0.0144	1	0.00001	0.1
13-14	Deep leaky confined aquifer	10	144	1	0.00001	0.1

Table 9.2 Layer properties for MODFLOW modelling for fifth simulation

Model layer	Represented unit	Total thickness (m)	Horizontal hydraulic conductivity (m/day)	Anisotropy ratio (Kv/Kh) (-)	Specific storage (m^{-1})	Specific yield (-)
1 - 14	Unconfined anisotropic aquifer	35 m	103	0.0005	0.00001	0.1

In the cross-sections that follow, the plots of drawdown and aquifer thickness are not drawn to the same scale as it is the relative drawdowns in each layer that are of the most interest, rather than drawdown magnitudes themselves.

The results of simulation 2, where the pumping occurs from the deep leaky confined aquifer, are illustrated in Figure 9-2. The plot of flow versus time shows that within one day, most of the pumped volume is sourced from the unconfined aquifer. Note that the use of the word ‘sourced’ in this context does not mean that the actual water from the unconfined aquifer ends up in the pumped well, but that that there is an equivalent volume of water depleted from the unconfined aquifer.

The results of simulation 3, where the pumping occurs from the shallow leaky confined aquifer, are illustrated in Figure 9-3. The plot of flow versus time shows that within one day, most of the pumped volume is sourced from the unconfined aquifer and that the flow contribution from the underlying strata due to a release of elastic storage is short-lived, reducing to zero within one day.

The results of simulation 4, where the pumping occurs from the unconfined aquifer, are illustrated in Figure 9-4. As expected, the plot of flow versus time shows that from the onset of pumping, the unconfined aquifer provides virtually all the pumped volume. Figure 9-4b also shows that the flow contribution from the underlying strata due to a release of elastic storage never reaches more than 1 % of the pumping rate over 1000 days.

The results of simulation 5, where the pumping occurs from one unconfined, anisotropic homogeneous aquifer, are illustrated in Figure 9-5. The plot of flow versus time shows that within one day, despite the vertical conductivity being much lower than the horizontal conductivity, much of the pumped volume is being sourced from the top of the unconfined aquifer. This occurs as a result of a small but widespread drawdown at the water table.

The modelling for all simulations shows that after pumping for around one day, the rate of pumping is matched by an equivalent rate of reduction in storage at the water table through drainage of the pore spaces. The reason for this is that the amount of water stored and released via elastic storage is very small in comparison to the amount that is stored and released via changes in the water table.

The drawdown plots show that for the three aquifer case, pumping from any of the three aquifers will create a drawdown effect in all of the other layers. Pumping from the deeper confined aquifer creates the largest drawdown effect within that aquifer, with lesser effects in each of the other aquifers. Pumping from the shallow unconfined aquifer creates a large localised effect on the water table and a smaller effect in the underlying aquifers. As expected, pumping from the deeper leaky confined aquifer has less of an effect on the magnitude of water table drawdown than pumping from the shallow leaky confined aquifer.

The drawdown plot for simulation 5 shows that pumping from a single anisotropic aquifer has very similar effects to pumping from an equivalent three-layered system. For this set of parameters, the water table drawdown in both the layered and anisotropic models after 1 day were the same at 1 km from the pumped well, while after 150 days, the difference in the water table drawdown between the two models was only 6.4% at a distance of 1 km. Simulation 5 had the greater drawdown than simulation 2. This difference is so small that it will be within the error margin of any realistic assessment.

The modelling has been undertaken with a relatively small total thickness (35 m) for simplicity but, this can be scaled up. For example, if the model used a total thickness of 350 m, but the hydraulic conductivity and specific storage values scaled down by the same factor, the effects would be the same.

This modelling exercise is merely a schematic of a real stratified groundwater system such as the Canterbury Plains groundwater system but the results have important general implications.

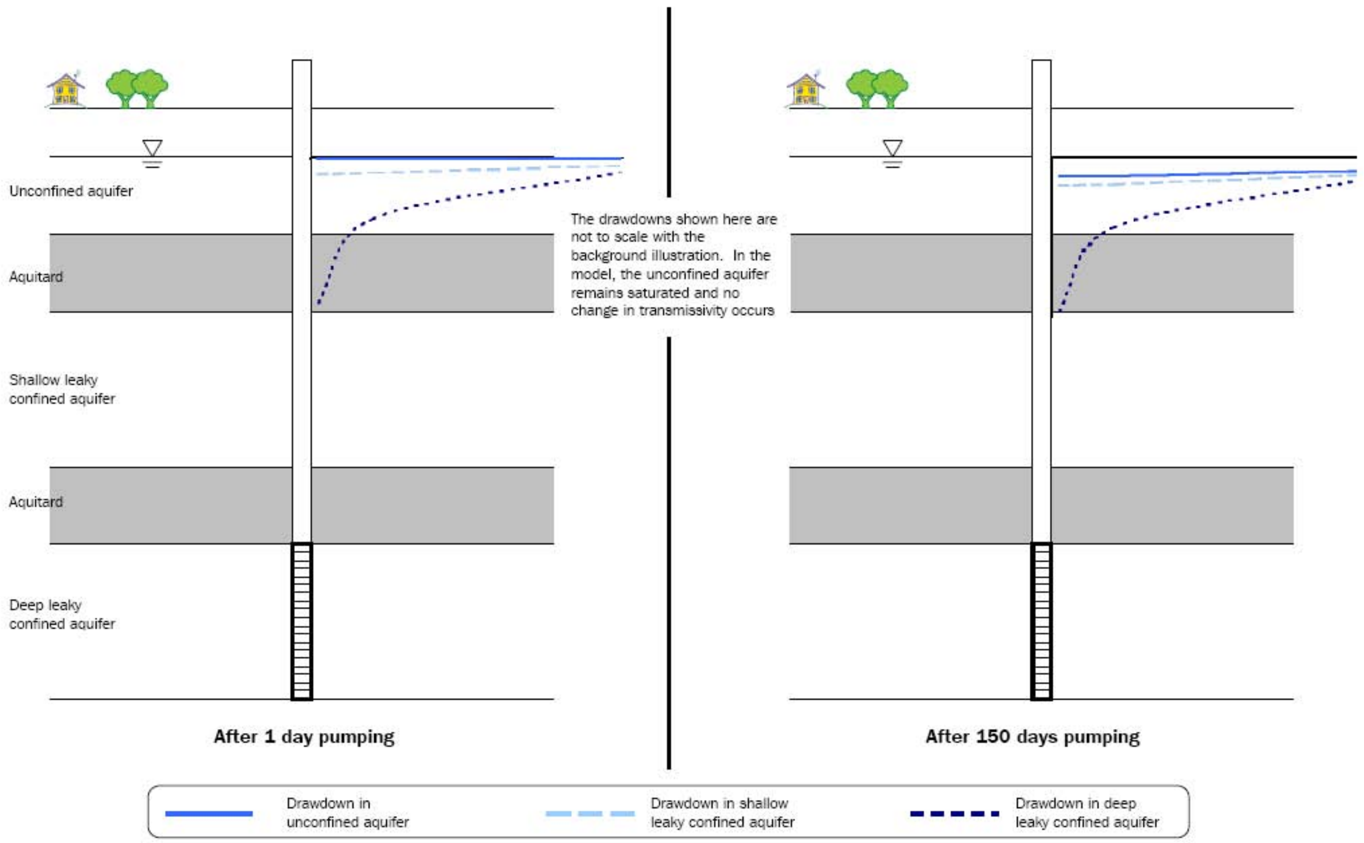
Pumping from any depth within a fully saturated groundwater system, where the distance to recharge and discharge boundaries is large, will create changes in groundwater pressures across the entire depth of the groundwater system. The magnitude of these changes is dependent on the hydraulic properties of the system and the pumping rate.

The drop in the level of the water table is smaller in magnitude when pumping from greater depths. However, in the absence of changes in recharge and discharge components of the system, irrespective of pumping depth there will ultimately be a reduction in storage at the water table that is virtually equivalent to the volume of water pumped from the well.

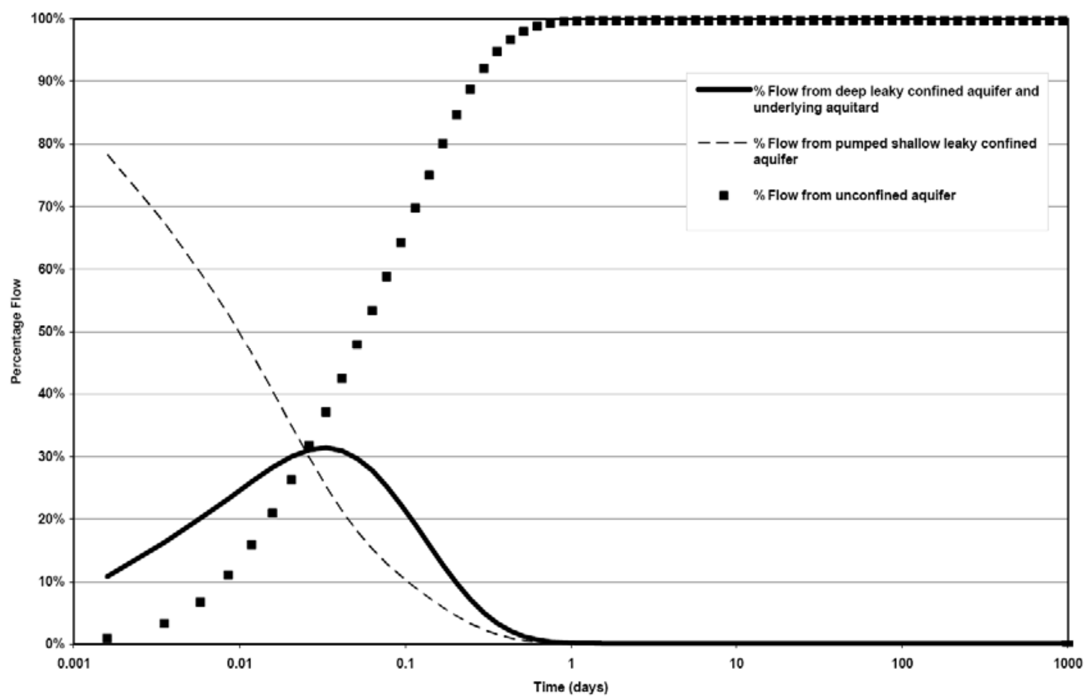
9.2 Investigation of the reliability of the Hunt and Scott (2007) solution for analysis and predictions

This section investigates the applicability of the Hunt and Scott (2007) solution to settings that are different from the conceptual model on which it was based and therefore, its usefulness as a tool in the assessment of effects within Canterbury's groundwater systems.

The drawdowns generated by the MODFLOW model described in section 9.1 were treated as if they were real pumping test data, and drawdowns generated with the Hunt and Scott (2007) solution were matched to them through a curve-fitting process. The resultant parameter values derived from this process were then compared with the original assigned parameters. This process was designed to illustrate any disparities between the two models. These models have been applied without consideration of any changes in aquifer recharge and discharge. Whilst real groundwater systems are much more complex, the simplification of the MODFLOW model is helpful in understanding what the parameters obtained using the Hunt and Scott (2007) solution represent in settings where more than two aquifers are present.

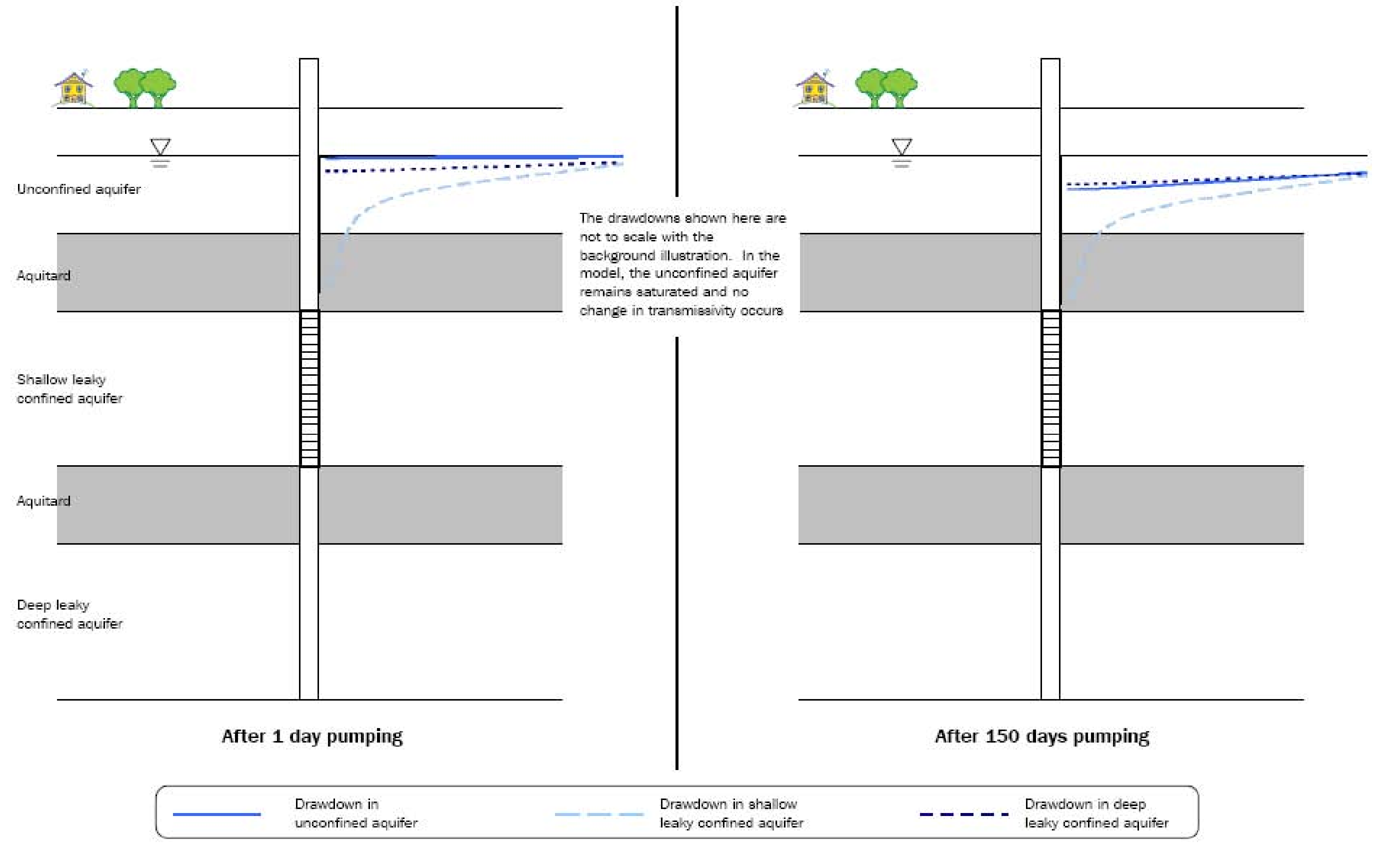


(a) Drawdowns versus distance

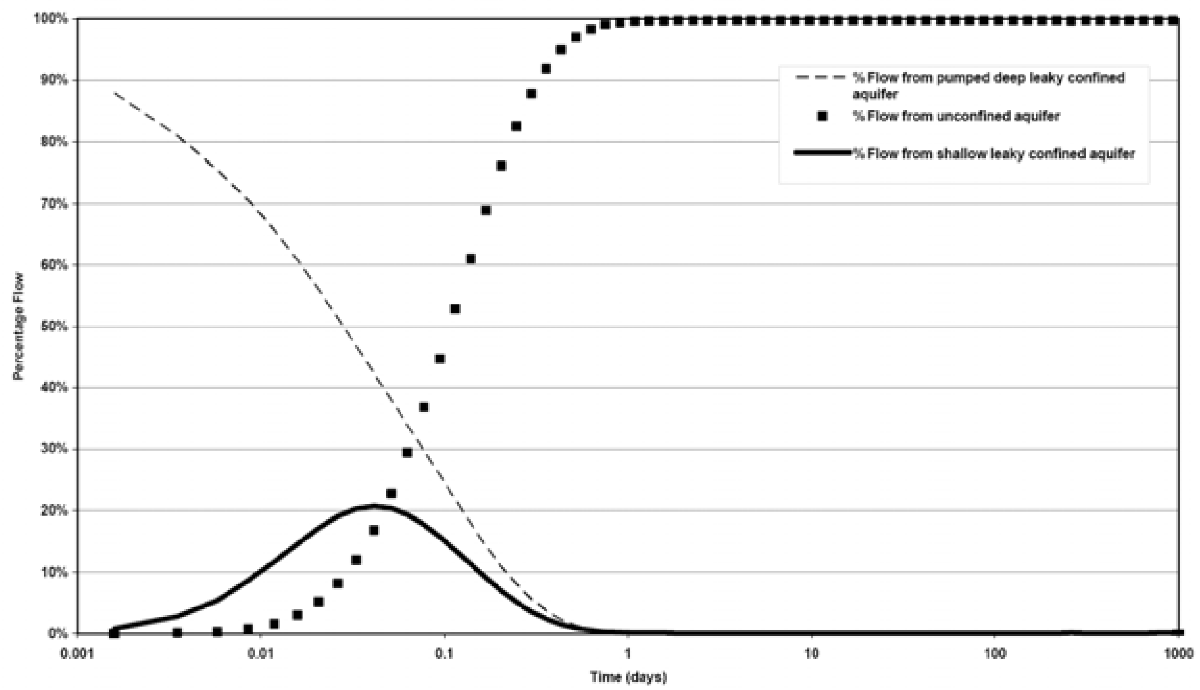


(b) Flow versus time

Figure 9-2: Simulated drawdown and flow for 3 aquifer system, with each layer isotropic, homogeneous, and including aquitard storage. Pumping from deep leaky confined aquifer. (Simulation 2)

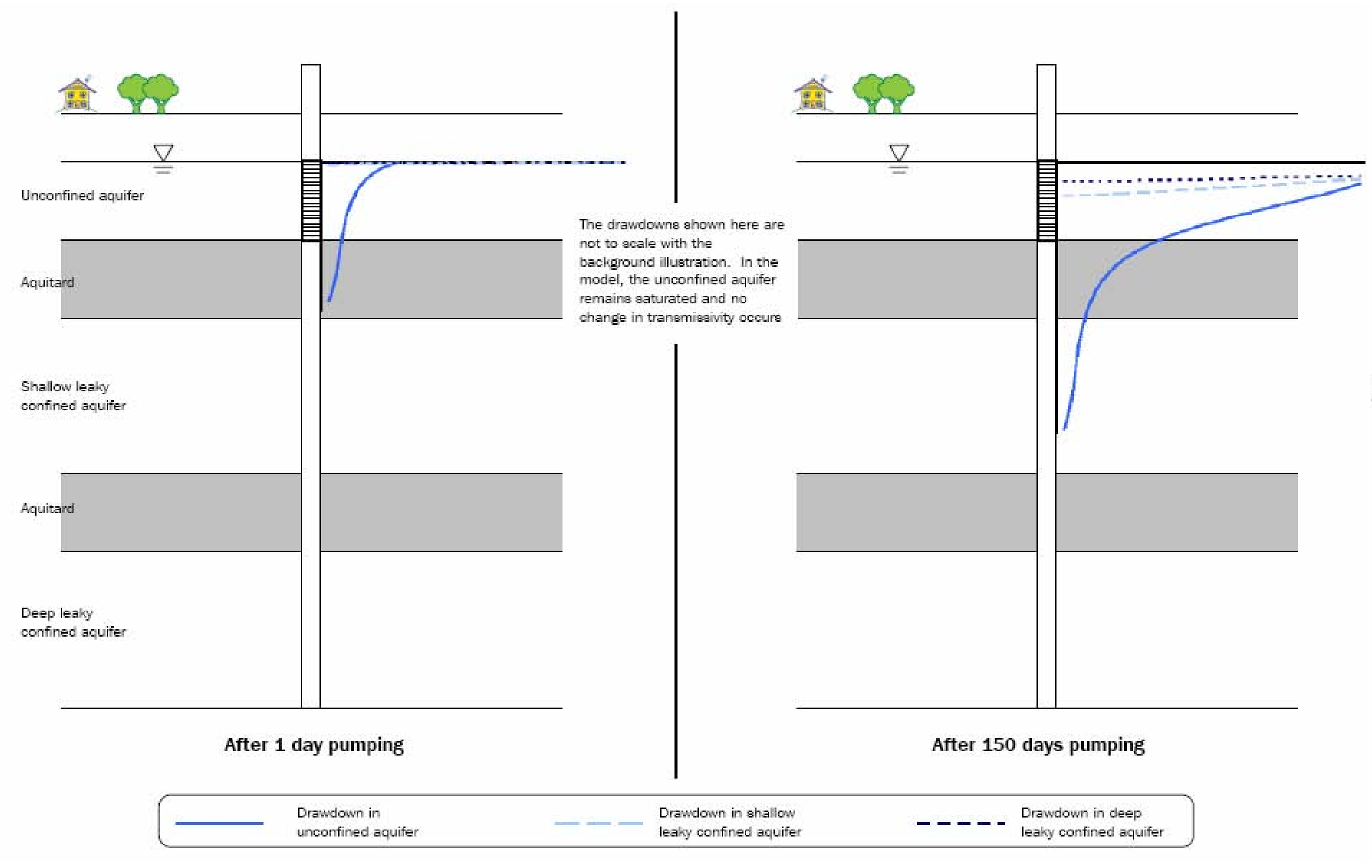


(a) Drawdowns versus distance

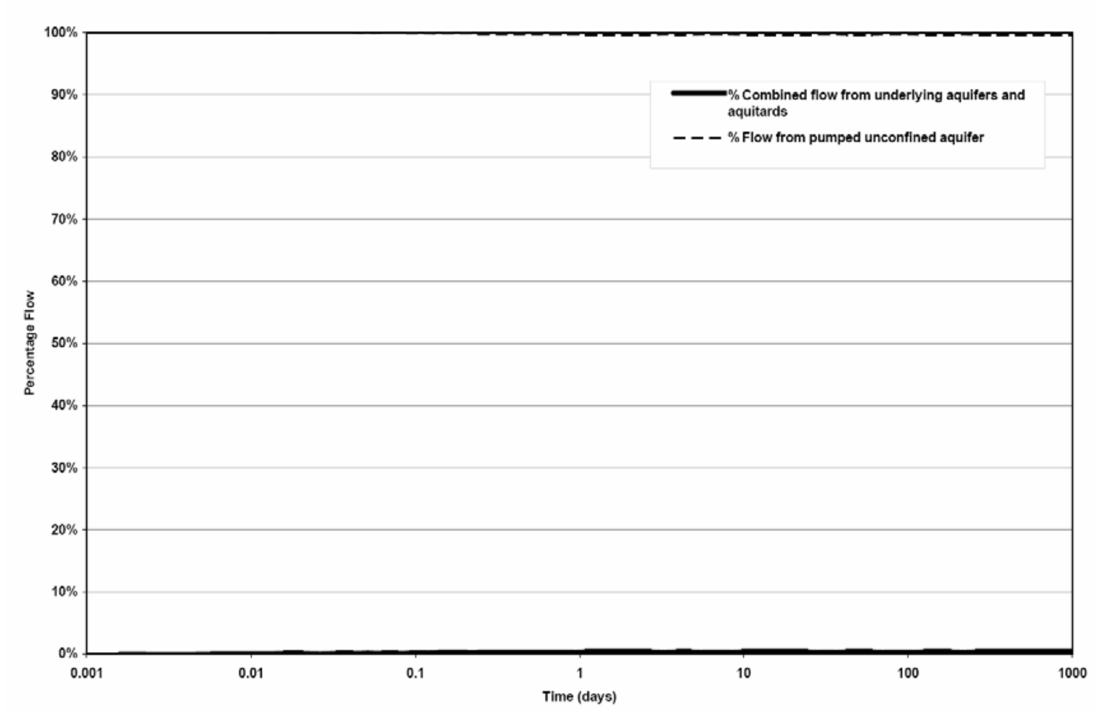


(b) Flow versus time

Figure 9-3: Simulated drawdown and flow for 3 aquifer system, with each layer isotropic, homogeneous, and including aquitard storage. Pumping from shallow leaky confined aquifer. (Simulation 3)

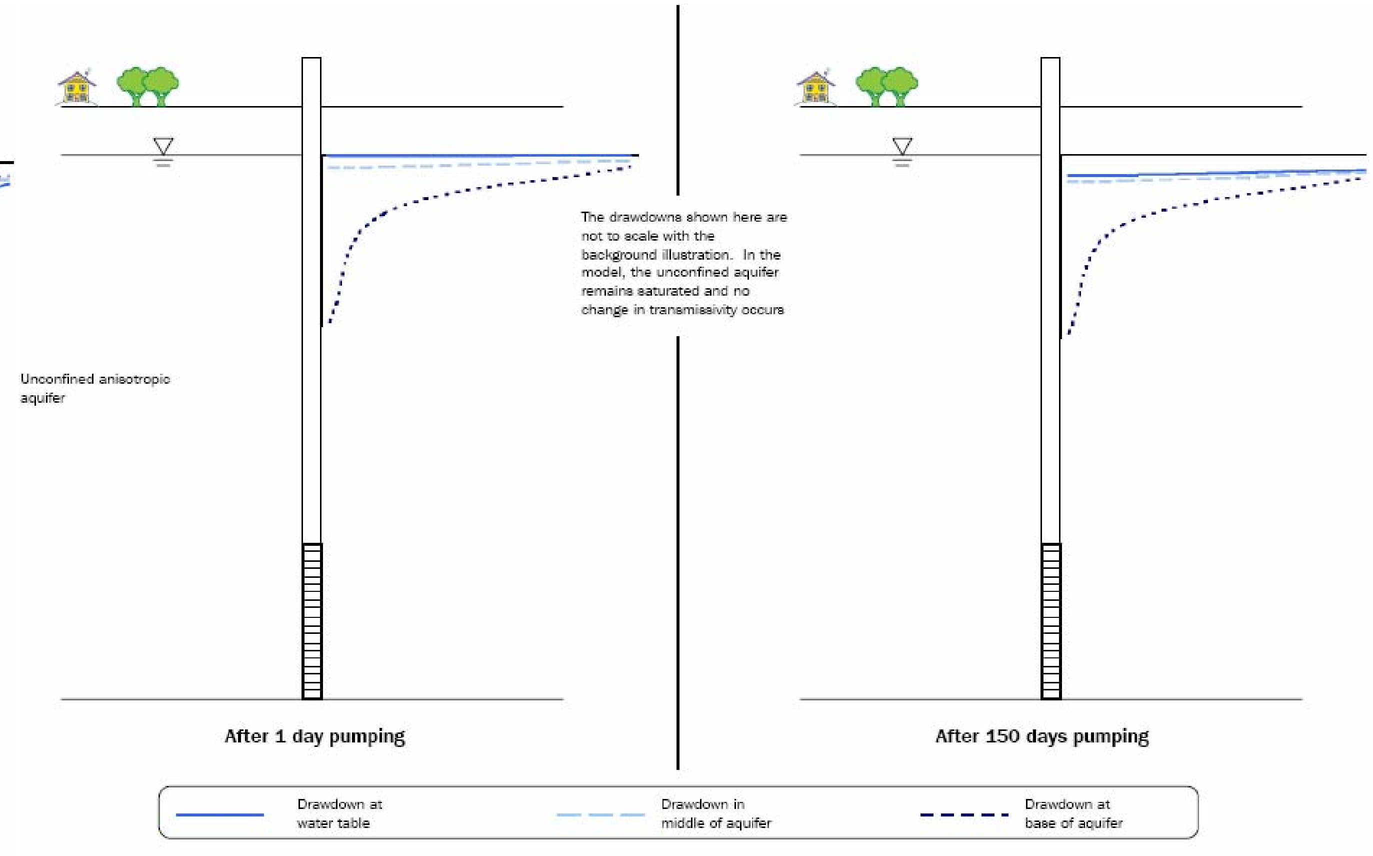


(a) Drawdowns versus distance

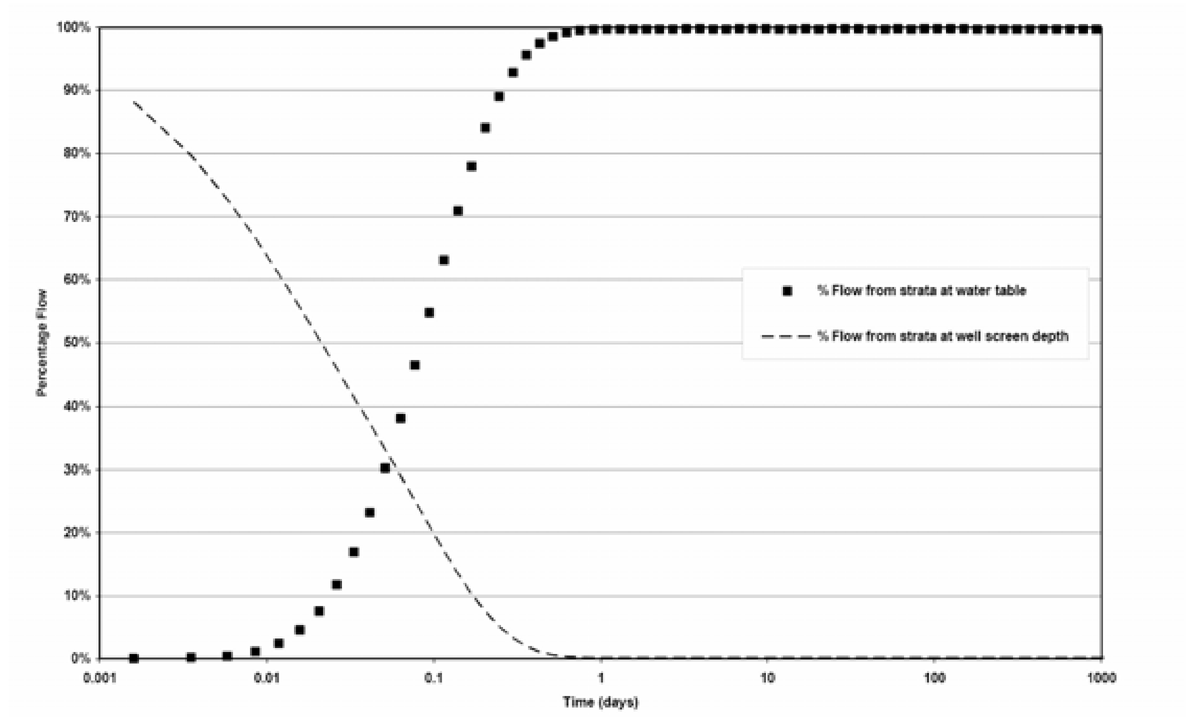


(b) Flow versus time

Figure 9-4: Simulated drawdown and flow for 3 aquifer system, with each layer isotropic, homogeneous, and including aquitard storage. Pumping from unconfined aquifer. (simulation 4)



(a) Drawdowns versus distance



(b) Flow versus time

Figure 9-5: Simulated drawdown and flow within one unconfined, anisotropic, homogeneous aquifer. Pumping from base of aquifer. (simulation 5)

9.2.1 Methodology

The Hunt and Scott (2007) solution was developed to model flows in a setting with two aquifers separated by an aquitard (Section 7.3). The lower aquifer overlies an aquiclude. In their paper, Hunt and Scott (2007) demonstrated the validity of the analytical solution in this conceptual setting by comparison with a 2-aquifer MODFLOW model.

To investigate the applicability of the Hunt and Scott (2007) solution to other settings, the drawdowns generated in the MODFLOW simulations 2, 3 and 5, as described in the previous section, were analysed using Hunt and Scott (2007) and the results of the analysis compared with those originally assigned in the numerical model.

Simulations 2, 3 and 5 were chosen for this investigation as they involved pumping from deeper strata, overlain by strata and an uppermost layer containing a water table. This is consistent with the conceptual model of the Hunt and Scott (2007) solution. Simulation 4 was excluded because the analytical solution does not describe pumping from an unconfined aquifer. As described in the previous section, the difference in the model results between simulation 1 and simulation 2 were not significant; therefore the results of the analysis for simulation 1 are not presented here.

To minimise repetition, the drawdown figures presented in the report are limited to those obtained at a distance of 200 m from the pumped well, which is consistent with the distance used in the Hunt and Scott (2007) paper. The parameters presented are those required to achieve a fit with the drawdowns generated at the same depth as the pumped well. Water table drawdowns and flows were not used in the interpretation.

9.2.2 Analysis of simulation 2 (pumping from deep leaky confined aquifer) using the Hunt and Scott (2007) solution

The results of this investigation for simulation 2 (pumping from the deep leaky confined aquifer) are shown in Figure 9-6. This figure illustrates the correspondence between the drawdown calculated with the Hunt and Scott (2007) solution and the MODFLOW generated drawdown in the pumped aquifer, 200 m from the pumped well.

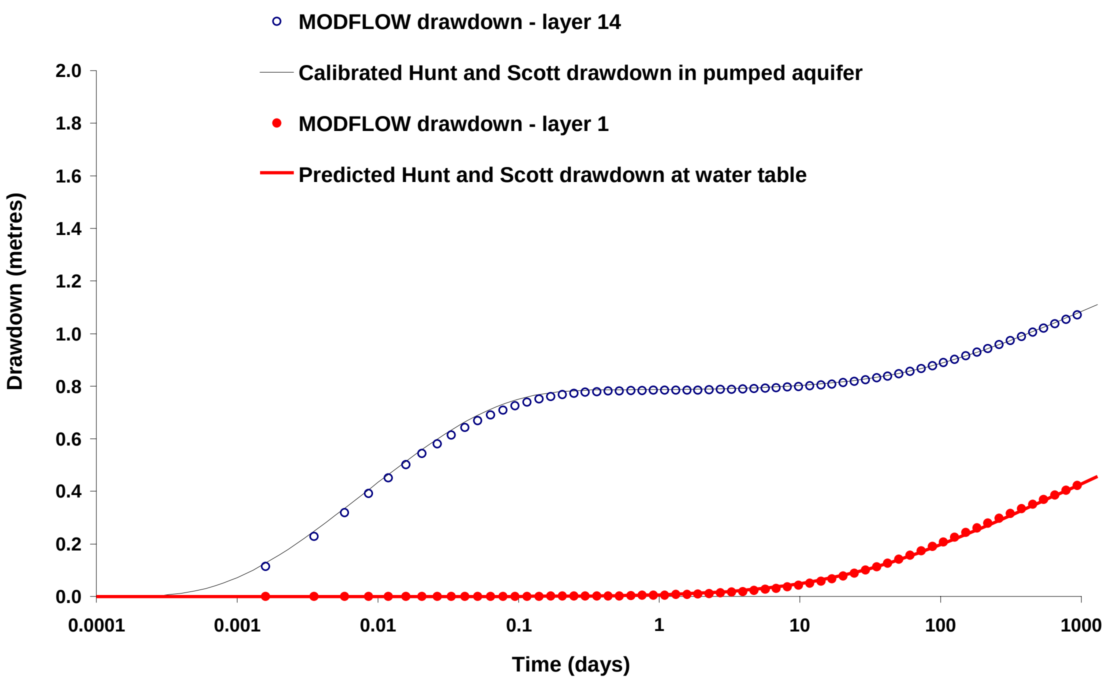


Figure 9-6: Results of curve-fitting for simulation 2

The parameters required to achieve this fit between the drawdown predicted with the Hunt and Scott (2007) solution and the drawdown generated in the MODFLOW model in layer 14 are shown in Table 9.3.

Comparison of the parameters in Table 9.3 shows that the actual (MODFLOW) and interpreted (Hunt and Scott (2007)) pumped aquifer transmissivity, total transmissivity and storage coefficient are similar, and the effective aquitard conductance value is identical. The interpreted specific yield value is higher than that assigned for the MODFLOW simulations.

Table 9.3: Actual and interpreted model parameters

Simulation No.	Case[1]	Transmissivity (m^2/day)		Pumped aquifer storage coefficient (-)	Effective aquitard conductance (1/day)	Specific yield (-)
		Pumped aquifer	Total[2]			
2	Modflow inputs	1440	3600	0.00010	0.0014	0.10
	Interpreted values	1570	3270	0.00012	0.0014	0.16
3	Modflow inputs	1440	3600	0.00010	0.0029	0.10
	Interpreted values	1780	3300	0.00014	0.0029	0.10
5	Modflow inputs	3600	3600	0.00035[3]	0.0017[4]	0.10
	Interpreted values	1234	3600	0.00012	0.0017	0.15

Notes:
1) Parameters specified as Modflow inputs reflect the parameter values assigned for each simulation (refer Tables 1 and 2)
Interpreted values show the effective parameter values interpreted by fitting the Hunt and Scott (2007) solution to the simulated drawdowns in the pumped aquifer.
2) For the Modflow inputs the total transmissivity reflects the combined transmissivity of all 14 model layers. For the interpreted values the total transmissivity is the sum of the transmissivities of the pumped aquifer and the overlying unconfined aquifer.
3) For the unconfined anisotropic aquifer case the storage coefficient has been determined for the total saturated thickness.
4) For the unconfined anisotropic aquifer case this parameter has been reported as the vertical hydraulic conductivity divided by the saturated thickness of layers 1 to 13 (drawdowns are reported for layer 14 of the model).

The differences in each of these parameters are discussed below:

- **Pumped aquifer transmissivity** – the Hunt and Scott (2007) value is larger than the MODFLOW value for the pumped aquifer alone because the Hunt and Scott (2007) solution accounts for some horizontal flow within the overlying aquifers.
- **Total transmissivity –** the sum of the interpreted Hunt and Scott (2007) transmissivities is approximately 10% less than the actual total MODFLOW transmissivity.
- **Pumped aquifer storage coefficient** – the Hunt and Scott (2007) value is larger than the MODFLOW pumped aquifer value, as it accounts for some elastic storage release from overlying layers.
- **Effective aquitard conductance** – the Hunt and Scott (2007) value is identical to the MODFLOW value as the analytical solution accounts for the resistance provided to flow by all layers between the pumped aquifer and the water table.

- **Specific yield -** the Hunt and Scott (2007) value is larger than for the MODFLOW unconfined aquifer value, which may be partly due to some elastic storage release from the intervening layers in the latter part of the drawdown response.

These differences in the estimated parameters are much smaller than the typical uncertainties involved in estimating parameters from aquifer test data.

The Hunt and Scott (2007) predicted drawdowns at the water table using these parameters correspond very well with the MODFLOW modelled drawdowns at the water table. The slope of the drawdown data in the final phase is such that longer-term predictions using the MODFLOW model and the Hunt and Scott (2007) analytical solution will be virtually identical. From a water budgeting perspective, the modelled difference between the predicted flows from the different aquifers will be of interest to those involved in the management of the resource. Comparison of the Hunt and Scott (2007) and MODFLOW flow predictions are shown in Figure 9-7.

Figure 9-7 shows that, although there is a reasonable difference between the predictions during the first day of pumping, beyond that the predicted flows are essentially identical. The early time difference is caused by the differences in Hunt and Scott (2007) derived storage values and those used in the MODFLOW model.

For water budgeting purposes, the later flow effects are those of relevance. Both models show that virtually all water is derived from a release of storage due to a declining water table. The reason for this is that the elastic storage that is released with the pumping is much smaller than that available for release due to a declining water table (refer to the two examples presented in Section 5.2 for a simple explanation of this).

Based on this comparison, the Hunt and Scott (2007) solution, when used to analyse drawdown recorded in a pumped aquifer with more than one overlying aquifer and underlain by an aquiclude, can provide realistic short and long-term predictions of drawdown and volume changes, both within the pumped aquifer and at the water table, where no significant changes in aquifer recharge or discharge occur.

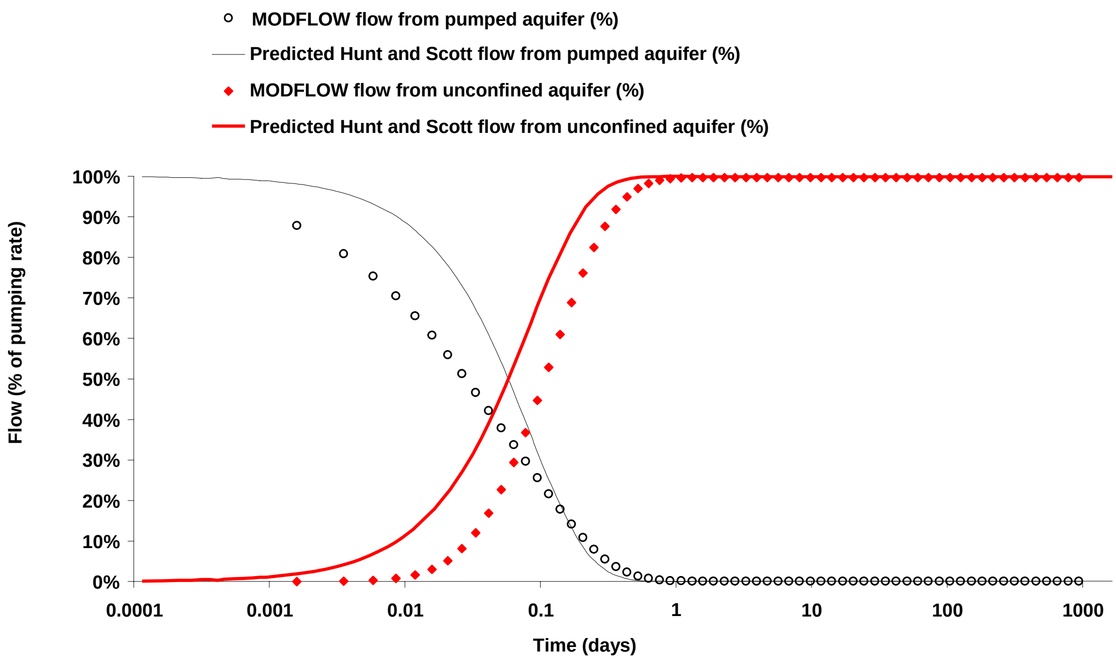


Figure 9-7: Comparison between MODFLOW flows and those predicted using the parameters from the drawdown analysis with the Hunt and Scott (2007) solution for simulation 2

9.2.3 Analysis of simulation 3 (pumping from shallow leaky confined aquifer) using the Hunt and Scott (2007) solution

The results of a similar investigation for simulation 3 (pumping from the shallow leaky confined aquifer) that involved comparing the fit of drawdown calculated with the Hunt and Scott (2007) solution to the MODFLOW generated drawdown in the pumped aquifer at a distance of 200 m is shown in Figure 9-8.

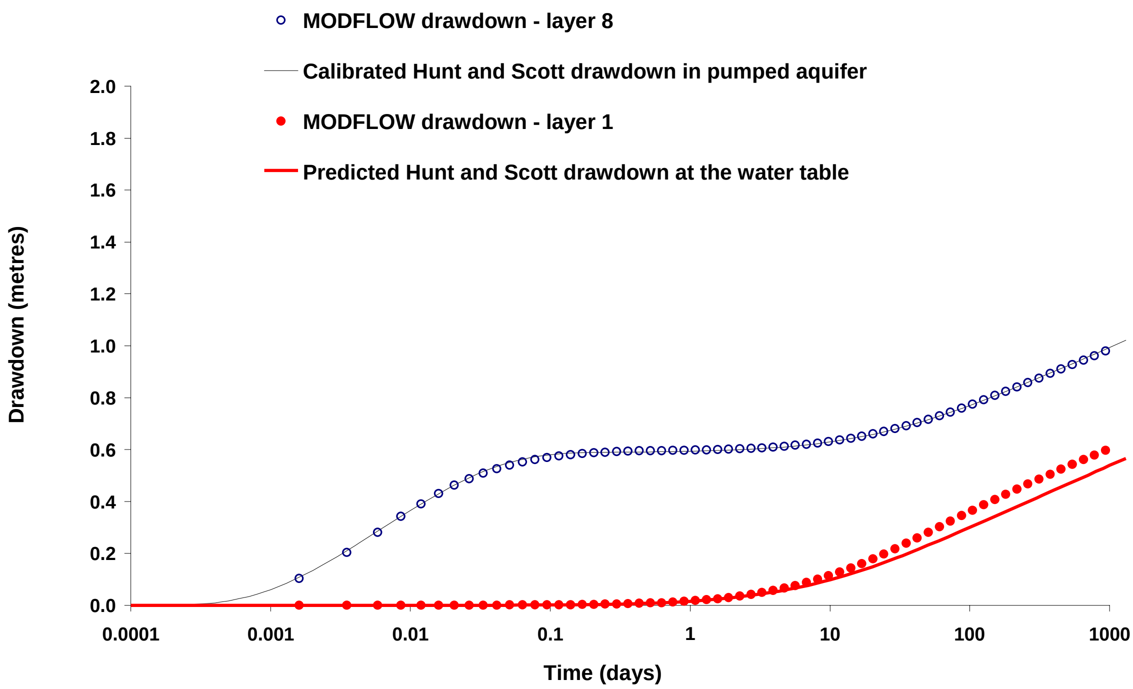


Figure 9-8: Results of curve-fitting for simulation 3

The parameters required to achieve this fit between the drawdown predicted with the Hunt and Scott (2007) solution and the drawdown generated in the MODFLOW model in layer 8 are shown in Table 9.3.

As with simulation 2, the Hunt and Scott (2007) pumped aquifer transmissivity and elastic storage coefficient are similar to the MODFLOW model parameters, and the effective aquitard conductance value for the overlying aquitard are identical (refer to Table 9.3). The Hunt and Scott (2007) interpreted specific yield value for this simulation is identical to the MODFLOW value. The total Hunt and Scott (2007) transmissivity is about 10% less than the MODFLOW value, as for simulation 2. The simulation 3 parameters are generally similar to those for simulation 2, differences being due to the contribution of flow from the underlying layers as well as overlying layers.

The predicted drawdowns at the water table using these simulation 3 parameters do not correlate as well as they do for the simulation 2 analysis, with the Hunt and Scott (2007) analytical solution underestimating the MODFLOW modelled drawdowns beyond ten days of pumping by about 10%. This difference is well within the typical precision of aquifer parameter estimates derived from aquifer tests. The difference also illustrates the usefulness of having a well screened at the water table to calibrate the analysis of drawdown data for a deeper observation well. While it is often not practical to run a test for long enough to observe drawdown at the water table, a record which shows the absence of a drawdown effect is useful for model calibration purposes.

Based on this comparison the Hunt and Scott (2007) solution, when used to analyse drawdown recorded in a pumped aquifer with both underlying and overlying aquifers, can provide realistic short and long-term predictions of drawdown within the pumped aquifer where no significant changes in aquifer recharge or discharge occur.

The Hunt and Scott (2007) solution may underestimate drawdown at the water table and over-estimate the lateral extent of the shallow drawdown effect. The differences are likely to be within aquifer test experimental uncertainty, depending on how the analysis and predictions have been carried out.

As with simulation 2, the predicted flows are essentially identical beyond one day of pumping (Figure 9-9), which is relevant in terms of water budgeting for resource management purposes.

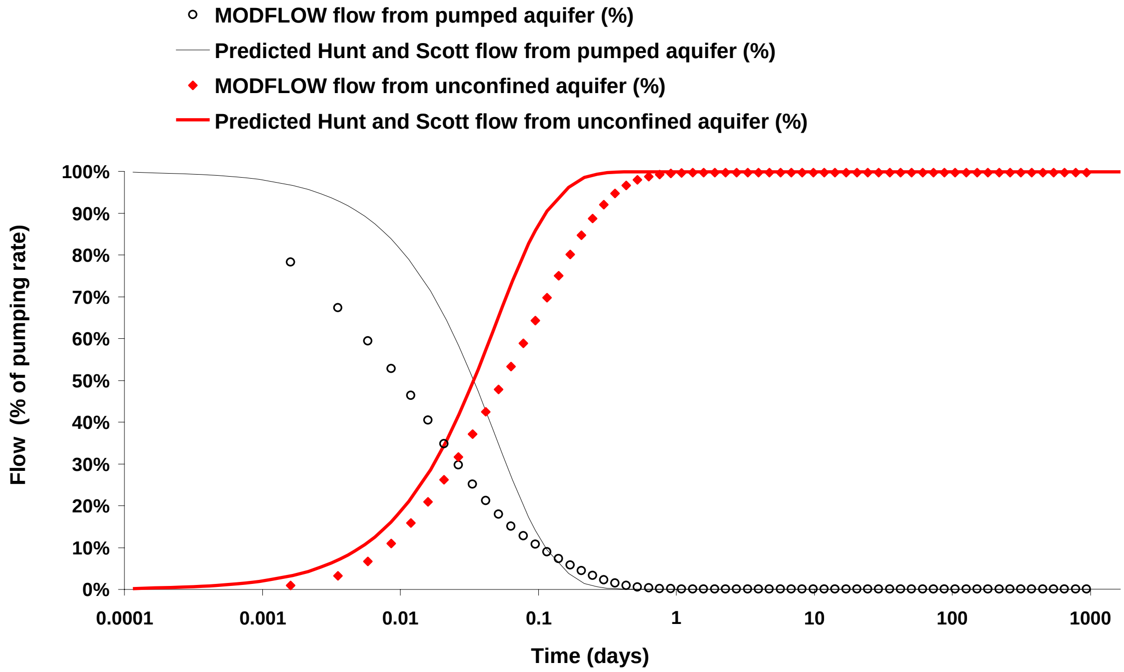


Figure 9-9: Comparison between MODFLOW flows and those predicted using the parameters from the drawdown analysis obtained with the Hunt and Scott (2007) solution for simulation 3

Based on the modelling result, the Hunt and Scott (2007) solution can reliably predict longer term volumetric changes within the aquifer containing the water table.

9.2.4 Analysis of simulation 5 (pumping from a well screened at the base of an anisotropic unconfined aquifer) using the Hunt and Scott (2007) solution

The results of a similar investigation for simulation 5 (pumping a well screened at the base of an anisotropic unconfined aquifer) are illustrated in Figure 9-10. This shows the fit obtained between the drawdown calculated with the Hunt and Scott (2007) solution and the MODFLOW generated drawdown in the pumped aquifer 200 m from the pumped well.

The parameters required to achieve this fit between the drawdown predicted with the Hunt and Scott (2007) solution and the drawdown generated in the MODFLOW model in layer 14 are shown in Table 9.3. The interpreted parameters are similar to those for simulation 2. The effective aquitard conductance is exactly the same as the effective value derived from the MODFLOW model. The sum of the Hunt and Scott (2007) transmissivities for the pumped and water table aquifers is equivalent to the transmissivity of that of the unconfined anisotropic aquifer in MODFLOW.

Using these parameters, the short-term predicted drawdowns at the water table do not correlate as well as they do for simulation 2, but better than they do for simulation 3. The modelled drawdowns at the water table are slightly underestimated by the analytical solution beyond fifty days.

As with simulation 2, the predicted flows are essentially the same beyond the first day of pumping (Figure 9-11). These flow effects after one day of pumping have particular relevance to water budgets, and from that perspective are of significance for resource management.

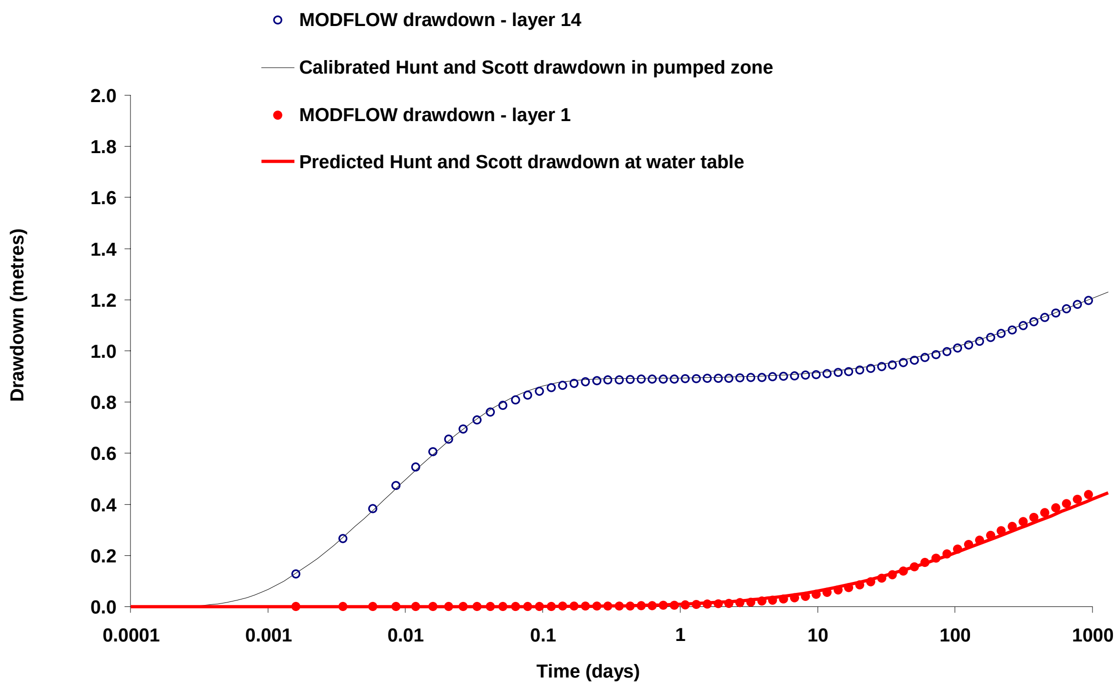


Figure 9-10: Results of curve-fitting for simulation 5

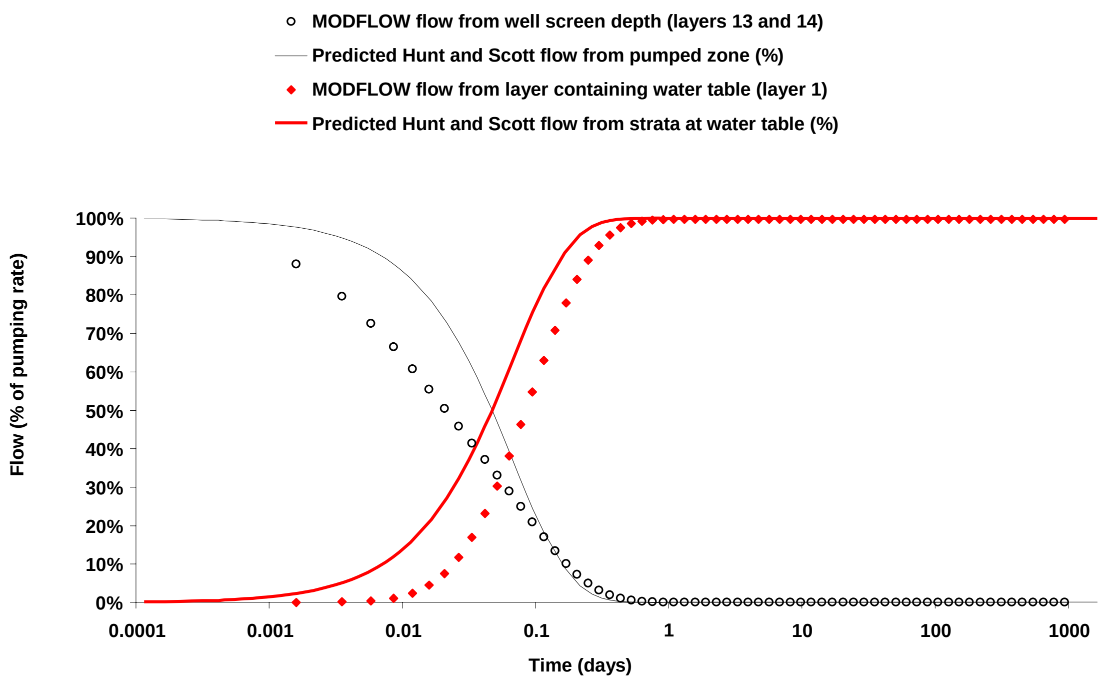


Figure 9-11: Comparison between MODFLOW flows and those predicted using the parameters from the drawdown analysis obtained with the Hunt and Scott (2007) solution for simulation 5

Based on the modelling results, the Hunt and Scott (2007) solution can also be used for drawdown analysis in an anisotropic unconfined aquifer. It can provide realistic long-term predictions of drawdown within the pumped strata, provided no significant changes in aquifer recharge or discharge occur, but may underestimate drawdown at the water table and over-estimate the lateral extent of the shallow drawdown effect. The Hunt and Scott (2007) solution can reliably predict volume changes within the aquifer containing the water table, beyond small pumping durations provided no significant changes in aquifer recharge or discharge occur.

Despite these results, it is more appropriate to use an analytical solution for the drawdown analysis that specifically models flows in an anisotropic aquifer, such as the solution described in Zhan and Zlotnik (2002).

The Zhan and Zlotnik (2002) solution was therefore used to analyse the drawdown generated in the MODFLOW model for simulation 5 at 200 m from the pumped well in layer 14. The fit obtained is shown in Figure 9-12.

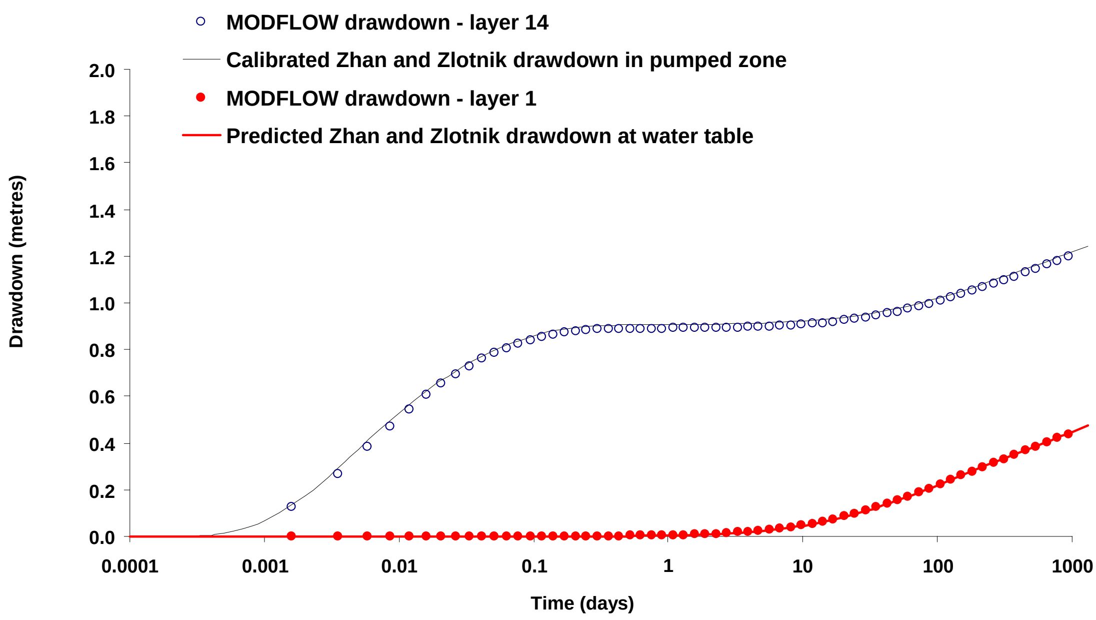


Figure 9-12: Results of curve-fitting for Simulation 5 with the Zhan and Zlotnik (2002) solution

The parameters in the Zhan and Zlotnik (2002) solution required to achieve the fit shown in Figure 9-12 are provided in Table 9.4. Comparison of these values with the Hunt and Scott (2007) values used in the MODFLOW model for simulation 5 (Table 9.2) shows that, with the exception of the specific storage which is only 10% larger, the values are identical. This indicates that the Zhan and Zlotnik (2002) solution can provide reliable estimates of hydrogeological parameters through the analysis of aquifer test data from an unconfined anisotropic aquifer. Volumetric changes can be estimated numerically using the results of the drawdown analysis.

Table 9.4 Derived parameters from curve matching of Zhan and Zlotnik (2002) solution with simulation 5 drawdowns

Horizontal hydraulic conductivity (m/day)	Specific storage (m^{-1})	Anisotropy ratio (Kv/Kh) (-)	Specific yield (S_y) (-)	Saturated thickness (m)
103	0.000011	0.0005	0.1	35

9.2.5 Summary of the appropriateness of the Hunt and Scott (2007) solution

The comparisons presented above indicate that aquifer parameters derived using the Hunt and Scott (2007) analytical solution for drawdown analysis, allow for reliable predictions of longer-term effects for the type of settings modelled, given the typical uncertainties in the parameters used in such simulations.

The disparities identified between the analytical solution and the MODLFOW model approaches are small even after the total simulation period shown (1000 days). On a time-scale of a pumping test, the differences are likely to be insignificant in relation to the uncertainties associated with the measurement of drawdown in the field. Errors in typical aquifer test data are likely to be of much more significance than those introduced by using the Hunt and Scott (2007) solution, rather than a more complex model, for many test analyses.

Where pumping tests are of insufficient duration to observe the complete characteristic leaky aquifer drawdown curve (Figure 5-3), there will be uncertainty introduced into longer term drawdown predictions by assuming values for parameters not determinable during the analysis. The uncertainties introduced by these assumptions are likely to be more significant than errors introduced by use of the Hunt and Scott (2007) solution. The modelling comparison has shown that longer-term volumetric simulations are equivalent.

As with any model, the Hunt and Scott (2007) analytical solution does have limitations. An alternate model that better describes the hydrogeological setting in which the assessment is taking place should be used where such a model is available. The Zhan and Zlotnik (2002) solution can be more appropriate for an anisotropic system, and a solution such as the Hunt (2003) stream depletion solution will be more appropriate where both pumping-induced leakage and interaction with a surface waterway are occurring over the period of assessment.

Where a hydrogeologic system cannot be reasonably represented by a simple analytical model, due to the level of complexity, a numerical model may be required.

10 Conclusions

This report shows that pumping from a well at any depth can induce changes in the water levels and storage volumes in strata throughout the system. These changes ultimately modify the recharge to and/or discharge from the system.

The geology of the Canterbury Plains allows groundwater flow both parallel with the major sedimentary bedding structure and across this structure (vertically) as there are no laterally extensive impermeable layers within the main groundwater system.

While most data on leakage effects in Canterbury have been collected from the Plains aquifers, as outlined in Section 2**,** there are other geological environments. Pumping-induced leakage from strata other than the pumped aquifer is likely to occur within all of the recognised groundwater systems within Canterbury that are dominated by alluvial gravels.

Pumping from any depth within alluvial gravel systems has the potential to create a fall in the level of the water table over an affected area that is equivalent to the volume of water pumped from the well. The reduction in storage at the water table will not always equal the volume of water pumped from the well if there are other sources of water present, such as a hydraulically connected surface water body. In this instance, the depletion of water as a result of the pumping will be apportioned between a reduction in storage at the water table and a depletion of flow in the hydraulically-connected surface water body.

While pumping from any depth can cause the same volumetric change at the water table, the effects on piezometric heads will be different in magnitude and timing, depending on the depth of the strata from which groundwater abstraction is occurring. For example, stream depletion effects may develop slowly from a deep well and the effects may be distributed between a number of waterways due to the small but widespread effect pumping from deeper layers has on the level of the water table. If all abstractions were to be concentrated at a particular depth (deep or shallow) this could create adverse cumulative effects, such as drawdown interference or cumulative depletion of a particular stream. Therefore, as with location, the depth of takes must be considered in the management of the resource.

A key relationship between leakage and the local hydrogeology, is that the lower the effective vertical hydraulic conductivity, and the deeper the well, the smaller but more widespread the drawdowns will be in the shallowest aquifer. Where the effective vertical hydraulic conductivity is higher and the well is shallower, greater drawdowns will occur in the shallow aquifer but will tend to be more localised.

Reported aquitard conductances (Figure 8-3) show a decreasing trend with depth. In effect this illustrates that as the depth of take increases, the additional thickness of saturated sediment through which groundwater has to travel from the unconfined water table to the pumped well screen, increases the impedance to flow.

The magnitude of the effective vertical hydraulic conductivity values inferred from aquifer tests is much smaller than typical horizontal hydraulic conductivity values in Canterbury. This supports the concept that the Canterbury Plains groundwater system, over its entire saturated thickness, behaves anisotropically i.e. with greater resistance to vertical flow than horizontal flow.

Measurement of long-term effects of abstraction is difficult, therefore modelling of these effects is often required. The numerical modelling described in this report was carried out to investigate interaction between geologic units at different depths in a stratified system and does not represent a particular part of the Canterbury Plains. General principles evident in the modelling results have important implications for the management of Canterbury's groundwater systems.

Based on the comparison with the MODFLOW model, the Hunt and Scott (2007) solution can be used for drawdown analysis in a multi-layered system and in an anisotropic unconfined aquifer, to provide realistic long-term predictions of drawdown within the pumped aquifer. However, such predictions assume that there is no induced change in aquifer recharge or discharge over the period of assessment.

The Hunt and Scott (2007) solution may be expected to reliably predict longer term volumetric changes within the aquifer containing the water table for situations where there is no induced change in aquifer recharge or discharge over the period of assessment.

The theory and examples presented in this report will not apply to strata that are variably saturated with depth as pumping from deeper layers will not increase the natural drainage from such hydraulically-disconnected strata. Beneath perched aquifers such variably saturated strata may occur.

The analytical models reviewed in this report may be overly simplistic for some applications. When modelling level and flow effects for systems where boundary effects and groundwater-surface water interaction are relevant over the period of assessment, an appropriate technique should be used to incorporate these. There are a variety of analytical and numerical models that can be used for these more complex settings.

11 Recommendations

Recommendations based on the information in this report are as follows:

- In the management of any groundwater system, consideration needs to be given to the hydrogeological properties of the system and the magnitude and timing of drawdowns in shallower and deeper levels in response to pumping from a particular depth range. Of equal importance is consideration of the way the whole system (including flow between different levels and changes in recharge and discharge) will be affected by groundwater abstractions.
- Shallow and deep strata should not be managed as disconnected entities as the flow of water between them can be substantial.
- Effective management of a groundwater resource requires recognition that the effects on piezometric heads will be different in magnitude and timing depending on the depth of the strata from which groundwater abstraction is occurring. For example, stream depletion effects may develop more slowly from a deeper well, and the effects may be distributed between a number of waterways, due to the small but widespread effect pumping from deeper layers has on the level of the water table.
- Individual abstractors pumping from deep layers may cause very small changes in the level of the water table. However, those deep abstractions can eventually affect the water table level and change recharge or discharge components of the system, such as stream and spring flows. The magnitude and location of these changes (i.e. which streams could be affected and to what degree) is dependent on the location of the groundwater takes, their depth and the hydrogeological properties of the system. These characteristics need to be taken account of in groundwater resource management.
- For each of the recognised groundwater systems within Canterbury, leakage between strata is only one of a number of issues that needs to be incorporated into management decisions. The modelling of leakage presented here uses a simplistic setting where the aquifers are assumed to be of infinite extent and there are no recharge sources. The complexities of each system need to be considered when management systems are being developed and an integrated approach is necessary if sustainable outcomes are to be achieved.
- Whether the sedimentary sequence is characterised as being layered or as simply displaying a degree of anisotropy is not usually of great relevance in the assessment of leakage effects.
- The hydrogeological terms used to label sediments (e.g. aquitards, aquifers) should not influence decisions made on the management of groundwater or assessments of groundwater flow. Those decisions should be made based on actual or potential effects that arise from the abstraction of groundwater from the resource
- The use of the Hantush and Jacob (1955) solution to analyse aquifer tests, whilst valid as an indicator of pumping-induced leakage, cannot estimate realistic long-term drawdowns in a groundwater system and a more complete and relevant method such as the Hunt and Scott (2007) or Zhan and Zlotnik (2002) solutions should be used. Where necessary, a method that allows for changes in the recharge and discharge components of the system should be used.
- Careful analysis of appropriate aquifer tests provides useful information for the assessment of drawdown effects and of long-term effects on the hydrogeological system. Such tests, although not precise, are the best and most practical tool for assessing the magnitude and timing of pumping-induced leakage effects related to an individual abstraction.

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13 References

Aitchison-Earl, P., Scott, D., Sanders, R., 2004. Groundwater allocation limits: Guidelines for the Canterbury Region. Environment Canterbury Report No. U04/02, 60 p.

Aitchison-Earl, P. and Smith M. 2008. Aquifer test guidelines (2nd edition) Environment Canterbury Technical Report R08/25, 57 p.

Amin, E. 2005. Determination of the rate and volume of leakage using the slopes of time-drawdown data. Environmental Geology. 47. p 558-564.

Anderson, B. 1994. Groundwater between the Selwyn and Rakaia rivers, Canterbury, New Zealand, M.Sc. thesis, University of Otago, 285 p.

Ashworth, P.J., Best, J.L., Peakall, J., Lorsong, J.A. 1999. The influence of aggradation rate on braided alluvial architecture: field study and physical scale- modelling of the Ashburton River gravels, Canterbury Plains, New Zealand. Special Publication of the Association of Sedimentology. 28, 333 – 346.

Bear, J. 1979. Hydraulics of groundwater. Dover Publications Inc. Mineola, New York. 569 p.

Bidwell, V., 2003. Groundwater management tools: analytical procedure and case studies. MAF Technical Paper No: 2003/06.

Boulton, N. 1955. Unsteady radial flow to a pumped well allowing for delayed yield from storage. Proceedings General Assembly Rome: International Association Sci. Hydrologists Publication 37, p 472-477.

Boulton, N.S. 1963. Analysis of data from non-equilibrium pumping tests allowing for delayed yield from storage. Proc. Inst. Civil Eng., 26, Paper No. 6693: 469-482.

Boulton, N. S. 1973. The influence of delayed drainage on data from pumping tests in unconfined aquifers. Journal of Hydrology, 19, p 157-169.

Brown, L J. 2001. Canterbury; In Groundwaters of New Zealand, M R Rosen and P A White (eds) New Zealand Hydrological Society Inc, Wellington. p 441-459.

Brown, L J. and J.H. Weeber 1992. Geology of the Christchurch urban area, 1:25 000 geological map, IGNS Map 1 and booklet (104 p), Institute of Geological and Nuclear Sciences.

Cooley, R. L., and Case, C. M. 1973. Effect of a water table aquitard on drawdown in an underlying pumped aquifer. Water Resources Research, 9(2), p 434–447.

Dann, R., Close, M., Pang, L., Flintoff, M., Hector, R. 2008. Complementary use of tracer and pumping tests to characterise a heterogeneous channelized aquifer system in New Zealand. Hydrogeology Journal. DOI 10.1007/s10040-008-0291-4.

Darcy, H., 1856. Les Fontaines Publiques de la Ville de Dijon, Victor Dalmont, Paris. 647 p.

Davey, G. 2006. Definition of the Canterbury Plains aquifers. Environment Canterbury Technical Report U06/10, 81 p.

Environment Canterbury. 2008a. Final decision of Canterbury Regional Council independent commissioners, 69 Rakaia-Selwyn Groundwater Allocation Zone applications, Environment Canterbury, 174 p.

Environment Canterbury. 2008b. Final decision of Canterbury Regional Council independent commissioners, 41 Selwyn-Waimakariri Groundwater Allocation Zone applications, Environment Canterbury, 275 p.

Freeze, R. A. and Cherry, J. A., 1979. Groundwater, Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 604 p.

Hantush, M.S. and C.E. Jacob. 1955. Non-steady radial flow in an infinite leaky aquifer. Transactions American Geophysical Union, 36, 95-100.

Hunt, B. 2003. Unsteady stream depletion when pumping from semi-confined aquifer. ASCE Journal of Hydrologic Engineering, Vol. 8, p 12-19.

Hunt, B. 2004. A spring-depletion solution. ASCE Journal of Hydrologic Engineering, Vol.9, 144-149.

Hunt, B., and Scott, D. 2005. Extension of the Hantush and Boulton solutions. Journal of Hydrologic Engineering, Vol.10, 223–236.

Hunt, B., Scott, D. 2007. Flow to a well in a two-aquifer system. Journal of Hydraulic Engineering. Vol. 12, No. 2, 146–155.

Hunt, B., Smith, M. 2008. Improved spring depletion solution and analysis. Journal of Hydraulic Engineering. Vol. 13, No. 7, 577–584.

Hunt, B., 2008a. Groundwater analysis using *Function.xls*. Manual to accompany Hunt's Excel spreadsheet software Function.xls. Version updated 15th August 2008. Downloadable from the website http://www.civil.canterbury.ac.nz/staff/bhunt.asp

Hunt, B. 2008b. Stream depletion for streams and aquifers with finite widths. Journal of Hydraulic Engineering. Vol. 13, No. 2, 80–89.

Jenkins, C. T. 1968. Techniques for computing rate and volume of stream depletion by wells. Ground Water, 6(2), 37-46.

Johns, R., Semprini, L., Roberts, P. 1992. Estimating aquifer properties by nonlinear least-square analysis of pump test response. Ground Water. Vol. 30, 68-77.

Johnson, G., Cosgrove, D., Frederick, D. 2001. A numerical model and spreadsheet interface for pumping test analysis. Ground Water. Vol. 39, 582-592.

Kruseman, G.P. and de Ridder, N.A. 1991. Analysis and evaluation of pumping test data, International Institute for Land Reclamation and Improvement. Wageningen, The Netherlands, 377 p.

Lough, H. 2004. The evaluation of hydrogeological parameters for stream depletion assessments. Thesis submitted in partial fulfilment of the requirements for the Degree of Master of Engineering at the University of Canterbury, Christchurch, New Zealand. 301 p.

Lough. H.K. and Hunt, B. 2006. Pumping test evaluation of stream depletion parameters. Ground Water, 44, 540.

McDonald, M. G., and Harbaugh, A. W. 1988. A modular three-dimensional finite-difference ground-water flow model. USGS techniques of water-resources investigations, Book 6, Chap. A1, United States Geological Survey, Reston, Va.

Neuman, S. P., and Witherspoon, P. A. 1968. Theory of flow in aquicludes adjacent to slightly leaky aquifers. Water Resources Research, 4, 103-112.

NCCB (North Canterbury Catchment Board). 1983. Interim report on the groundwater resource of the Central Plains. North Canterbury Catchment Board and Regional Water Board.

Pattle Delamore Partners Ltd. 2002. Groundwater Model Audit Guidelines. Prepared for Ministry for the Environment, New Zealand. 72 p + appendices.

Pattle Delamore Partners Ltd and Environment Canterbury. 2000. Guidelines for the Assessment of Groundwater Abstraction Effects on Stream Flow. Environment Canterbury Technical Report. U00/11, 114 p.

Scott, D.M. and B. Hunt, 2007. A water budget perspective of leaky aquifer response to pumping. Rotorua New Zealand Hydrological Society annual conference abstract.

Shulmeister, J. 2007. Evidence presented at Selwyn-Waimakariri Groundwater Allocation Zone consent hearing before Commissioners.

Sinclair Knight Merz. 2008. Inventory of aquifer test data for the Central Plains. Prepared for Environment Canterbury, Technical Report R08/80, 43 p.

Sinclair Knight Merz. 2009. Inventory of aquifer test data for the Ashburton and Valetta Groundwater Allocation Zones. Prepared for Environment Canterbury, Technical Report R09/12, 30 p.

Theis, C. 1935. The relation between the lowering of the piezometric surface and the rate of duration of discharge of a well using ground-water storage. Trans. American Geophysical Union 2, 519-524

Theis, C. 1957. The source of water derived from wells essential factors controlling the response of an aquifer to development. Groundwater Water Notes Hydraulics No.34. United States Department of Interior Geological Survey Water Rescues Division. Ground Water Branch.

Trinchero, P., Sanchez-Vila, X., Copty, N., Findikakis, A., 2008 A new method for the interpretation of pumping tests in leaky aquifers. Ground Water, 46, p 133-143.

Williams, H., Scott, D.M. and Bidwell, V., 2008. Adaptive management of groundwater in the Rakaia-Selwyn Groundwater Allocation Zone; Environment Canterbury Technical Report R08/64, 93 p.

White, P., 2009. Geology and vertical groundwater flow, Central Plains, Canterbury. Environment Canterbury Technical Report R09/5, 73 p.

Zhan H., and Zlotnik., V. A. 2002. Groundwater flow to a horizontal or slanted well in an unconfined aquifer, Water Resources Research, 38 (7), 1108.

Appendix A: Details of wells used to assess effective vertical hydraulic conductivity (Figure 8-4)

PUMPED WELL NUMBER	PUMPED WELL DEPTH (mbgl)	TOTAL SCREENED INTERVAL (mbgl)	REPORTED AQUITARD CONDUCTANCE K'/B' (per day)	ESTIMATED SATURATED THICKNESS B' (m)	CALCULATED EFFECTIVE VERTICAL HYDRAULIC CONDUCTIVITY K' FROM K'/B' DATA (m/day)
L36/0913	174.5	121 to 174.5	1.2E-04	64	7.7E-03
L37/1209	43.1	37.1 to 43.1	1.4E-03	32	4.4E-02
K37/1505	59.11	53.11 to 59.11	2.4E-04	47	1.1E-02
K37/1752	95.23	89 to 95	2.3E-04	87	2.0E-02
L37/0220	70.4	64.3 to 70.4	2.6E-04	39	1.0E-02
M35/10908	145.8	142.8 to 145.8	3.7E-04	138	5.1E-02
M35/12026	102	87.7 to 101.7	9.2E-05	78	7.2E-03
M35/3487	21	no info	3.2E-03	17	5.5E-02
K37/1765	54	51 to 54	4.2E-03	44	1.8E-01
K37/1768	104	98 to 104	7.5E-04	97	7.3E-02
M35/11522	30	no info	4.3E-05	26	1.1E-03
M35/11591	25.45	21.45 to 24.45	2.8E-04	16	4.5E-03
K37/1907	48.5	45.5 to 48.5	3.1E-04	45	1.4E-02
L37/0144	40	no info	3.6E-04	32	1.1E-02
L37/1187	71.5	63.5 to 71.5	3.5E-04	61	2.1E-02
L37/1282	75.6	67.6 to 75.6	1.0E-03	51	5.1E-02
M37/0242	19.6	16.3 to 19.6	2.1E-02	16	3.3E-01
K36/0649	60	54 to 60	3.9E-03	49	1.9E-01
K36/0665	216.2	206.2 to 216.2	5.6E-05	175	9.8E-03
K37/1703	119.4	113.4 to 119.4	2.0E-05	111	2.3E-03
K37/2079	50	40.755 to 46.755	8.3E-04	36	3.0E-02
K38/1700	64.5	58.5 to 64.5	3.0E-05	58	1.7E-03
K38/1701	66.9	60.9 to 66.9	1.1E-04	60	6.6E-03
L36/0326	99	96 to 99	1.3E-04	60	7.6E-03
L36/1107	53.17	50.17 to 53.17	1.5E-04	48	7.2E-03
L36/1533	77.08	71.08 to 77.08	5.6E-04	49	2.8E-02
L36/1576	89.97	83.97 to 89.97	8.7E-04	58	5.1E-02
L36/1801	191.27	170.89 to 191.27	1.5E-04	143	2.1E-02
L37/0243	25.9	22.85 to 25.9	4.2E-02	14	5.8E-01
L37/0629	95	no info	2.5E-03	54	1.4E-01
L37/0907	48	42 to 48	6.6E-05	39	2.6E-03
L37/0970	86.4	81.1 to 86.4	2.3E-03	45	1.0E-01
L37/1042	95.5	89.5 to 95.5	4.5E-03	50	2.3E-01
L37/1055	111.5	105.5 to 111.5	2.0E-04	70	1.4E-02
L37/1257	77.19	71.19 to 77.19	1.6E-03	57	9.1E-02
M33/0252	46.5	30.6 to 42.9	1.6E-04	30	4.7E-03
M34/5623	122.5	116.5 to 122.5	1.0E-06	113	1.1E-04
M34/5633	75	67 to 73	2.7E-04	61	1.7E-02
M35/0469	24.4	17.7 to 24.4	1.5E-04	16	2.4E-03
M35/9628	120.25	114.25 to 120.25	2.1E-04	64	1.3E-02
M36/1968	43	no info	1.2E-04	27	3.3E-03
M36/3548	13.1	no info	7.6E-02	8	6.1E-01
M36/6986	62.9	56.9 to 62.9	7.3E-04	27	2.0E-02

PUMPED WELL NUMBER	PUMPED WELL DEPTH (mbgl)	TOTAL SCREENED INTERVAL (mbgl)	REPORTED AQUITARD CONDUCTANCE K'/B' (per day)	ESTIMATED SATURATED THICKNESS B' (m)	CALCULATED EFFECTIVE VERTICAL HYDRAULIC CONDUCTIVITY K' FROM K'/B' DATA (m/day)
M36/7817	14.3	13.57 to 14.3	3.0E-02	12	3.6E-01
M36/8220	50	47 to 50	2.8E-04	27	7.6E-03
O32/0019	12.8	9.2 to 12.8	1.0E-02	8	8.3E-02
M36/7711	150	118 to 148	3.0E-05	122	3.7E-03
M36/7711	150	118 to 148	1.5E-04	122	1.8E-02
L36/1622	36	37 to 52	7.5E-04	25	1.9E-02
L36/2159	81	74 to 81	2.2E-04	71	1.6E-02
L36/2159	81	74 to 81	1.5E-04	71	1.1E-02
L36/1709	63.5	57 to 63	3.0E-05	54	1.6E-03
L36/1709	63.5	57 to 63	4.0E-05	54	2.2E-03
L36/1069	48	43 to 48	2.5E-03	38	9.5E-02
L36/1531	239	no info	5.8E-04	83	4.8E-02
L37/1474	65	59 to 65	7.0E-04	56	3.9E-02
L36/1028	94.3	85.3 to 94.3	1.9E-04	83	1.6E-02
L37/1171	30	25 to 30	4.4E-03	25	1.1E-01
K36/0436	111.2	90 to 111.2	2.0E-04	85	1.7E-02
K37/1305	81.16	72.71 to 81.16	3.5E-04	75	2.6E-02
K37/1659	88.76	68.06 to 88.76	1.4E-05	72	1.0E-03
K37/2386	54.6	46.4 to 54.6	1.9E-02	44	8.4E-01
K37/1312	78	65.89 to 78	4.4E-04	67	3.0E-02
K37/3054	101.35	86.35 to 101.35	1.6E-03	56	9.0E-02
L36/1469	146.65	125 to 146.65	1.1E-03	95	1.1E-01

Note:
Results reported for constant rate tests classed by ECan as being reliable with accurate analysis (i.e. Reliability ≤2, Parameter Accuracy ≤3, Single Screen). Some wells appear more than once because there was either more than one aquifer test or more than one observation well.

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