Chapter 9 - Aquifer Hydraulic Properties

Introduction

Aquifers are natural systems and as a result they have a high degree of variability. This variability makes accurately describing the hydraulic properties of an aquifer difficult as unlike a river channel we can't see what is going on underground. Indirect measurements of aquifer water levels are used to determine the direction of groundwater flow, to measure aquifer hydraulic properties and understand its chemistry. It is important to recognise that the parameters and classifications of aquifers are a gross simplification of the natural groundwater system. One of the challenges in describing an aquifer is how to characterise the large degree of natural variability that occurs into meaningful and manageable definitions. For the purposes of analysis, groundwater systems are generally described in simplified terms and for most purposes this is adequate, particularly at a regional or catchment scale. The simplification of observed strata is often depicted in a conceptualised form (Fig. 9.1).

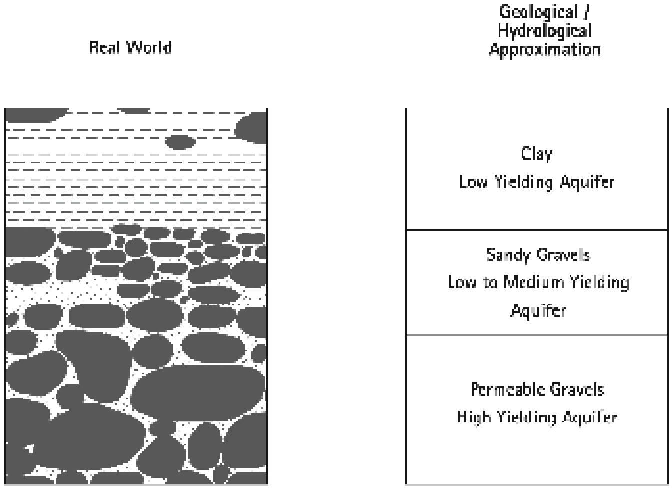

Figure 9.1: Simplifying nature

Grain size and sorting

The basis of information that is used to describe an aquifer's structure is gained from the carefully recorded observations of the sediments encountered during well drilling. For unconsolidated materials these are generally described in terms of the grain size of individual particles such as sand or gravel. A standard classification of grain size has been developed over time (Table 9.1). While these standardised classification systems exist, there is still a degree of subjectivity when it comes to describing material in the field and terms will vary from driller to driller. It is sometimes difficult to know whether terms such as gravel or shingle on well logs mean the same thing, especially if the wells were drilled by different drillers or in different eras.

The degree of uniformity of rocks and gravels forming a geological formation is also important to the characteristics of the aquifer it hosts. A poorly sorted formation consisting of a mixture of grain sizes from clays to boulders will tend to form less permeable aquifers compared to an equivalent geological formation comprising of well sorted gravels. This is because the strata won't be as efficient for storing or transmitting groundwater, which are the fundamental functions of an aquifer.

Description of Material	Grain Size (millimetres)
Boulders	Greater than 200
Cobbles	60 – 200
Coarse gravel	20 – 60
Medium Gravel	6 – 20
Fine Gravel	2 – 6
Coarse sand	0.6 – 2
Medium sand	0.2 - 0.6
Fine sand	0.06 - 0.2
Silt	0.004 - 0.06
Clay	0.004 or less

Table 9.1: Grain size classification

Aquifers and aquitards

Aquifers - unconfined versus confined

A water bearing layer of sediment that is sufficiently permeable to allow water to be abstracted in sufficient volume to be a useful water supply, is called an aquifer.

Aquifers come in many different configurations but are broadly classed depending on their geological structure as either unconfined or confined (Fig. 9.2). Confined aquifers are overlain by a low permeability stratum such as clay that isolates them from surface processes such as rainfall or landuse effects. A common characteristic of water wells tapping confined aquifers in Marlborough are water pressures greater than atmospheric pressure. These are called artesian wells and commonly free-flow when the well is uncapped.

Unconfined aquifers do not have any overlying barriers isolating them from surface processes and water pressure is in equilibrium with the atmosphere forming the surface known as the water table. They are effectively open to the land-surface and natural climatic processes. As is often the case with nature, most aquifers have a structure somewhere in-between which is neither impermeable nor totally porous to flow either. The degree of confinement largely controls how aquifers are recharged, the quality of the groundwater and the yield of wells tapping them.

Both confined and unconfined aquifers are present in Marlborough. The Wairau Aquifer near Renwick is unconfined which allows it to receive recharge from infiltrating rainfall and from Wairau River flow losses via gravity drainage. By contrast the coastal Wairau Aquifer

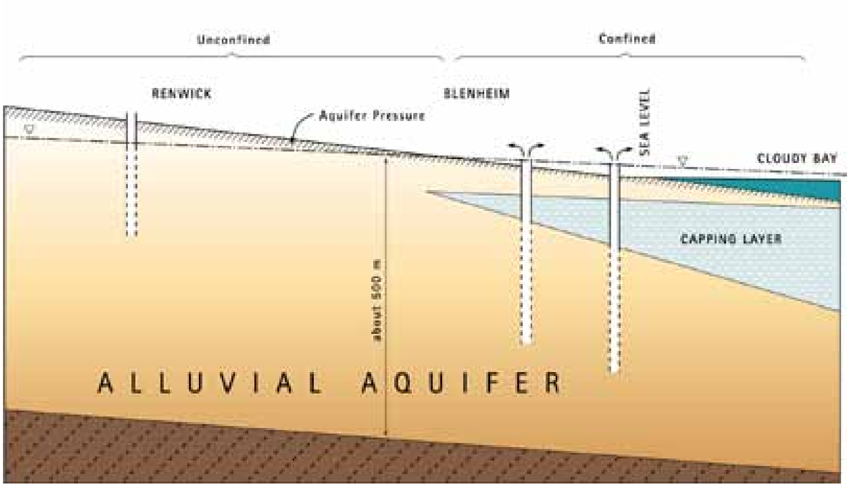

Figure 9.2: Wairau Plain aquifer structure

east of SH1 is confined by a continuous bed of clay sediments which isolates it from the atmosphere and the land surface.

Wairau Aquifer groundwater flows from west to east, although this can take several decades depending on the route. This means the confined aquifer receives recharge from a considerable distance away.

There are advantages and disadvantages associated with each aquifer type. Unconfined aquifers are recharged more rapidly when it rains for example, but are more susceptible to surface pollution from overlying land uses, which can include microbes, nitrates and sulphates.

A common disadvantage of confined aquifers is the slower rate of groundwater flow and depleted oxygen levels. This characteristically leads to more mineralised groundwater due to the increased interaction with local rocks. Also, some nuisance minerals which have generally been present in an insoluble form can become soluble due to changes in the chemical conditions in confined aquifers. These commonly include iron, manganese and sometimes arsenic.

Confined aquifers are less affected by surface microbial contamination due to the natural barrier formed by the low permeability capping strata and flow rates that are slower. Both these factors mean that microbial contaminants don't persist for as long underground. Potential problem chemicals such as nitrate and sulphate can also be transformed into less harmful forms on encountering the reducing conditions commonly associated with local confined aquifers.

Aquitards and confining layers

Aquitard is the name given to the low permeability strata that confine the aquifers.

Most of New Zealand's aquitards that are found in coastal areas were created when the sea invaded the land during warmer inter-glacial periods. These distinctive geological formations make useful marker beds or clocks, and under a microscope this fine grained clay or mud material consists of a distinctive array of broken shells or marine fossils that indicate the climatic conditions that existed at that time.

The coastal aquifer systems around New Zealand consist of alternating bands of river gravels, sandwiched between confining layers of marine sediments (Fig. 9.3).

Aquifer storage and flow

The two most important characteristics of aquifers that determine the way in which they respond to changes in their use are the way in which groundwater flows through the strata and the way in which water is released from storage within the strata.

Groundwater flow

The mathematical expression of groundwater flow was first developed in 1856 by a French hydraulics engineer named Henri Darcy. The empirical relationship that now bears his name is a universally accepted method in assessing groundwater flow. Darcy's equation is used to calculate the velocity and volumetric rate of flow through a porous medium such as gravel or sand.

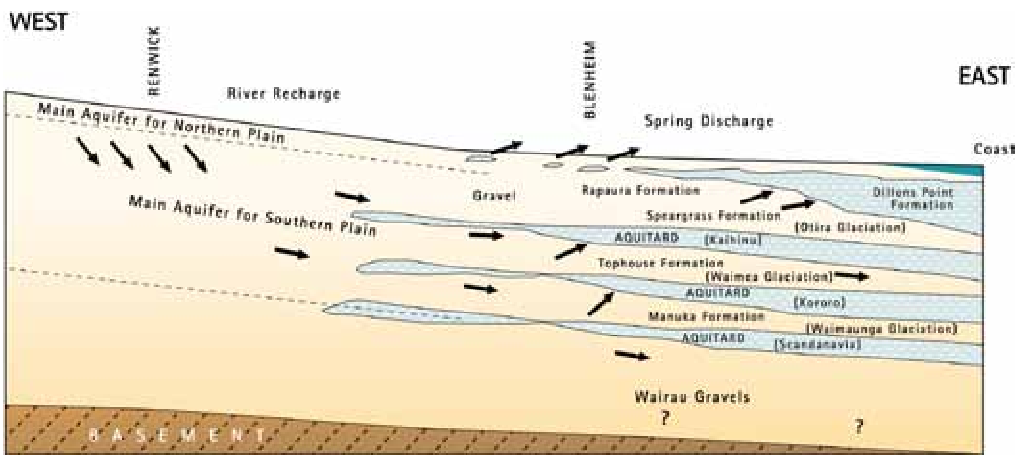

Figure 9.3: Alternating sequence of marine and land derived sediments beneath the Wairau Plain

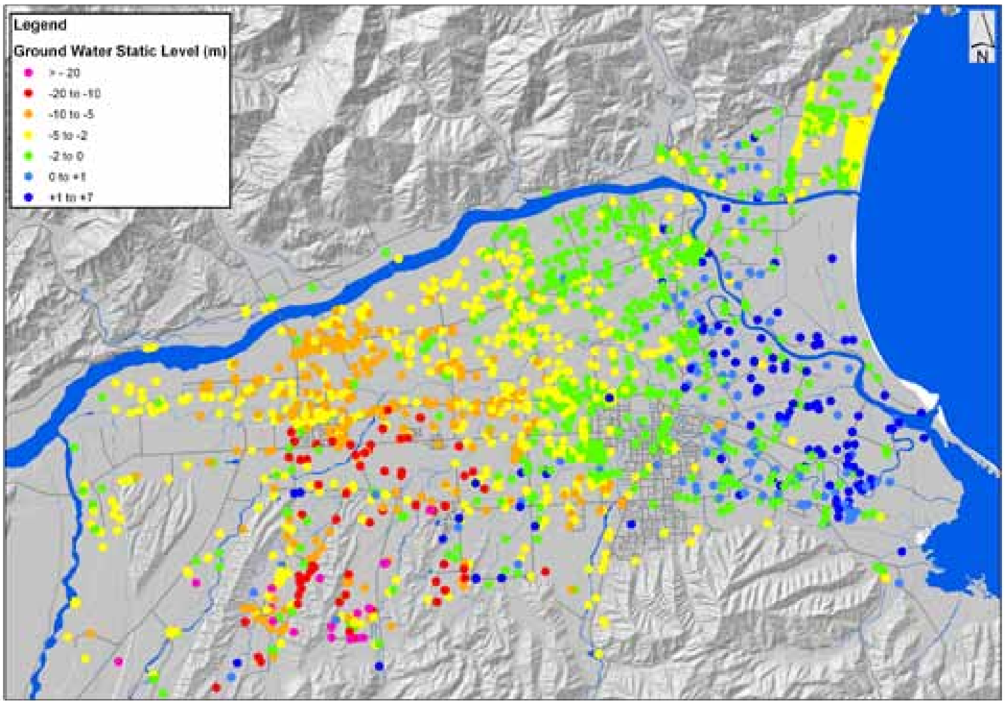

Figure 9.4: Wairau Plain static well level

Darcy's equation states that the rate of flow through a porous medium depends on a physical property of the medium, called its hydraulic conductivity, and is also proportional to the slope of the water surface or hydraulic gradient. The hydraulic gradient is a measure of the energy available to overcome the resistance to flow, and the greater the gradient the more energy is available. Darcy's equation only applies to laminar not turbulent flow. However because of the low velocities normally associated with natural groundwater flow, it is normally valid under all conditions. The higher the hydraulic conductivity and the larger the hydraulic gradient, the greater the rate of groundwater flow through an aquifer.

Hydraulic conductivity has the units of metres/second (m/s) or metres/day (m/day), and is related to permeability. While hydraulic conductivity reflects the characteristics of both the water and the geology, the fluid characteristics such as density, temperature or salinity are less important. The hydraulic conductivity may also be considered as a measure of the resistance to flow, where the greater the hydraulic conductivity, the lower the resistance to flow.

Static well level and hydraulic gradient

The depth to the water table in a well which is not being pumped is known as its static level. It is a very useful measurement and the MDC has archived the static water level observations observed by drillers for several thousand Wairau Plain well records (Fig. 9.4).

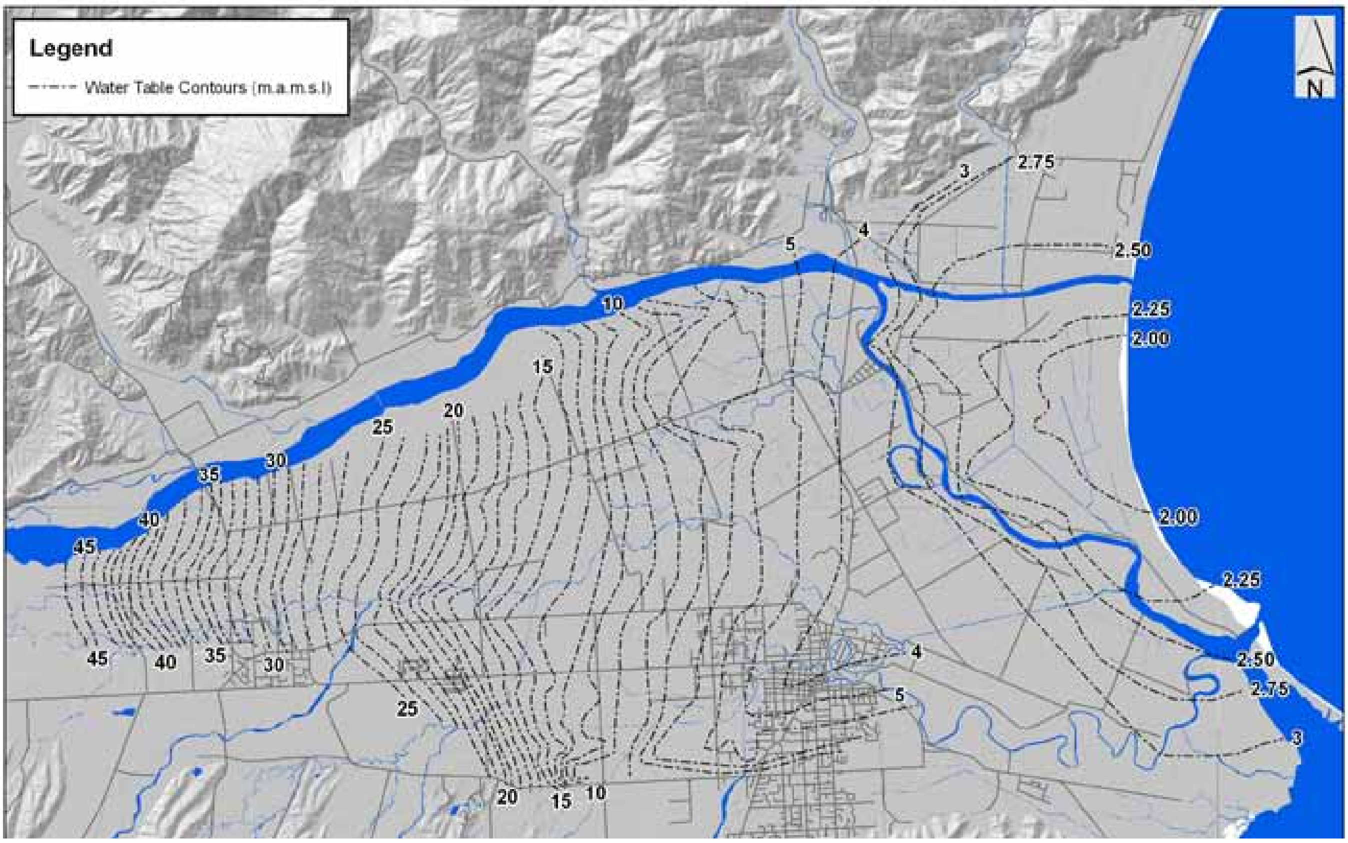

Figure 9.5: Flow net showing contours of groundwater elevation

The static levels of wells tapping aquifers beneath the western or southern Wairau Plain are generally below ground, whereas those in the east or nearer the Cloudy Bay coast tend to be near or higher than the ground surface. This reflects natural changes in aquifer structure from unconfined to confined aquifer conditions and the relative elevation of the recharge source. Deeper static water level are associated with deep wells tapping the Southern Valleys Aquifers, with the maximum being close to 30 metres below the surface. A clear pattern exists across the Wairau Plain whereby static levels increase towards the coast. Static levels greater than zero indicate artesian water. In general terms a water table below ground level indicates an aquifer recharge zone, whereas artesian pressures indicate a discharge area where there is a net loss of groundwater.

Measuring the depth to the water table is therefore one of the most important measurements that can be made of a groundwater system. These water level measurements can be mapped to form a groundwater flow net which shows water level elevations. This enables the recognition of aquifer flow patterns (Fig. 9.5). Just as a topographical map describes the contours and shape of the land surface, a water table contour map describes the water table or piezometric surface. It provides a visual picture of the nature of the resource that exists out of sight beneath the ground. They show how and where the energy within a groundwater system is distributed, and the location of aquifer boundaries. Because water flows from areas of high to low elevation, we can map the regional direction of groundwater flow. Assuming the gravels forming the aquifer are homogeneous, the direction of groundwater flow is at right angles to the contours and towards the east or north-east.

Flow nets are created by linking points of equal groundwater elevation to form contours called equipotentials. A water table or piezometric surface contour is a line along which either the water table is at the same elevation or the piezometric surface has equal pressure. No flow occurs along a contour as no hydraulic gradient exists to drive the flow.

Material	Hydraulic Conductivity (m/day)
Gravel	150 – 86,000
Sand	2.5 – 45
Peat	5.7
Loess	0.08
Silt/Clay	0.08 - 0.0002
Schist	0.2
Slate	0.00008
Kaituna Gravel	32,575*
Unconfined Wairau Aquifer	100 – 5,000

Table 9.2: Values of hydraulic conductivity

Flow occurs at right angles to these contours and is proportional to the gradient between them. The greater the distance between contours, the slower the flow. The path a particle of water follows is represented by a flow or stream line with the zone between called a stream tube. Groundwater flow within a stream-tube is assumed to be constant unless water is added from a river or lost due to pumping or via spring flow. Because these maps work in accordance with Darcy's equation, the higher the permeability of the alluvium forming the aquifer in a particular area, the narrower the stream tube required to pass a certain volume of water. This is analogous to fluid flow in a pipe. Streamtubes are wider for lower permeability material and converge for higher permeability gravels.

Where there are a number of wells pumping from the aquifer, a water table contour map will indicate the effect these are having on aquifer levels and how this effect is distributed. Over the years there have been many water table contour maps constructed for various Marlborough aquifers. They are all unique to the seasonal conditions prevailing at the time, reflecting the cumulative effects of pumping, recharge and drainage. For example maps prepared from surveys during the 1997/98 drought exhibit relatively low groundwater levels with depressions caused by pumping. Conversely in winter when there is no pumping for crop irrigation, the aquifer surface is relatively flat.

Volumetric groundwater flow can be calculated for the streamtube formed between adjoining pairs of flow lines if the hydraulic conductivity of the aquifer is known, because the difference in groundwater elevation from the flow net defines the hydraulic gradient.

Hydraulic conductivity

Typical values of hydraulic conductivity for different materials vary from several hundreds of metres per day for gravels, down to imperceptible rates for consolidated hard rocks such as schist.

While the dimensions of hydraulic conductivity have the units of velocity (distance divided by time), it is in fact a flux representing a discharge of water per unit area under a hydraulic gradient of 1, with the full units of $m^3/day/m^2$. In other words, the volume of water passing through a square cross-sectional area of the aquifer in a given time. It can't be used to describe the rate of groundwater flow on its own without knowledge of the hydraulic grade, which is the driving force causing water to move and the pore space through which the water movement occurs.

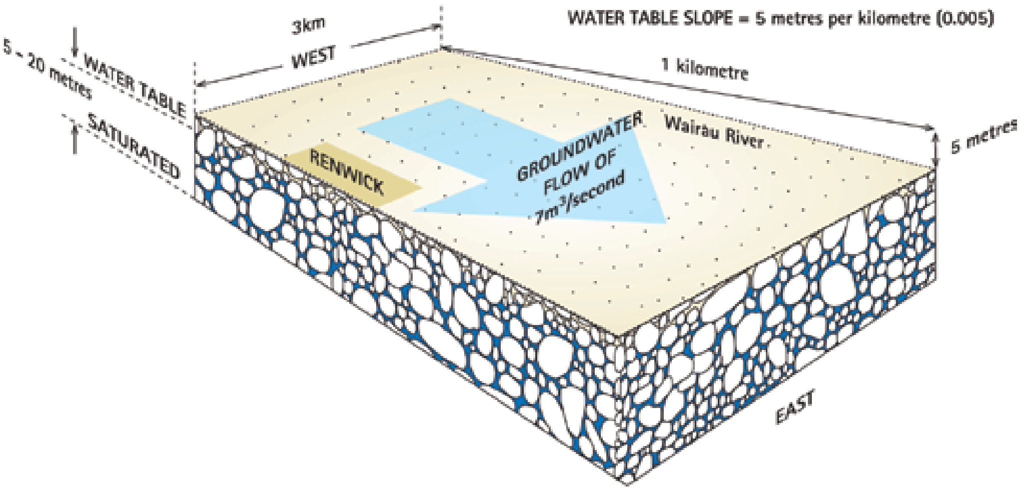

Figure 9.6: **Wairau Aquifer recharge throughflow**

Thorpe (1992) described New Zealand values ranging from as low as 10^{-8} m/day for clay up to 86,400 m/day for coarse gravels. Measured values for Canterbury alluvial aquifers of around 1,700 m/day are likely to generally reflect the gravel material encountered in Marlborough, although there will always be local variability (NCCB – 1986). In 1986 a tracer experiment measured hydraulic conductivities of 32,575 m/day for extremely transmissive gravels on the north bank of the Wairau River (White – 1986) (Table 9.3).

Hydraulic conductivity is usually higher in the horizontal plane for local alluvial sediments. This reflects the way they were deposited in nature by rivers. In areas where vertical groundwater flow is significant such as the coastal Wairau Aquifer, the vertical hydraulic conductivity is the most important for determining the direction of groundwater flow.

The most accurate means of estimating hydraulic conductivity is through aquifer testing because the results reflect the average value across the geological formation and take into account natural variability. There is uncertainty as to whether laboratory tests reflect real world conditions because the material used has been disturbed.

Volumetric flow

The principles discovered by Darcy may seem theoretical, but they actually work well for real world aquifer flow situations. Darcy's equation is used on a day to day basis by MDC staff to calculate the volumetric flow and the velocity of groundwater for various purposes. For example, in a local case study the throughflow of groundwater in the 3,000 metre width of permeable gravels between the Renwick Terrace and the Wairau River has been calculated (Fig. 9.6). The slope of the groundwater table has been measured as 5 metres per kilometre, or 0.005, but the hydraulic conductivity and aquifer depth are uncertain. Measurements of aquifer transmissivity show the average value is about 2,000 m^2/day, however this is generally recognised as under-representing nature due to wells that only partially penetrate the main water bearing layers of the Wairau Aquifer. The three kilometre wide cross-section shown in Figure 9.6 corresponds to a throughflow of 30,000 m^3/day. However we know from loss gaugings that around 7 m^3/s (604,800 m^3/day) is lost from the Wairau River and passes through this area of the aquifer (Table 9.4).

To pass this volume of groundwater requires aquifer transmissivity values of 20,000 to 30,000 m^2/day, which are high by New Zealand standards and much higher than historically measured values in the unconfined Wairau Aquifer. Because transmissivity equals hydraulic conductivity multiplied by aquifer thickness, this implies hydraulic conductivities of between 700 and 3,000 m/day assuming an aquifer thickness of 10 to 30 metres.

Groundwater travel times

Knowledge of the rate at which groundwater moves is useful for assessing the time it has been underground, its origin, and for identifying sources of landuse contamination or the risk of groundwater contamination from land uses. For example, local values of hydraulic conductivity are particularly relevant in relation to the movement of contaminants through aquifers in the event of a petrol spill.

Location	Hydraulic Gradient	Hydraulic Conductivity (m/day)	Seepage Velocity**
Unconfined Wairau Aquifer	0.005	820	16
Recent Wairau River gravels at Kaituna (north-bank)	0.0033	32,575	443

*** Based on porosity of 0.25*

Table 9.3: **Marlborough groundwater velocities**

Wairau River Recharge To Wairau Aquifer
Wairau River channel flow loss to groundwater of 7 m^3/s
Hydraulic gradient of $5x10^{-3}$
Cross-sectional area at right angles to flow of $9x10^4$ m^2
Hydraulic conductivity of 1344 m/day
Porosity of 0.25
Average seepage velocity of 27 m/day

Table 9.4: **Unconfined Wairau Plain Aquifer properties**

When considering travel times, it is important to recognise that flow does not occur through the entire cross-sectional area that is used in Darcy's equation. In reality, we know it only passes through the pore spaces and not the solids. The correct velocity is therefore calculated by dividing the Darcy velocity by the porosity, which in turn is called the seepage velocity.

Even this is a simplification of reality as we cannot see the microscopic pathway that water has to flow through.

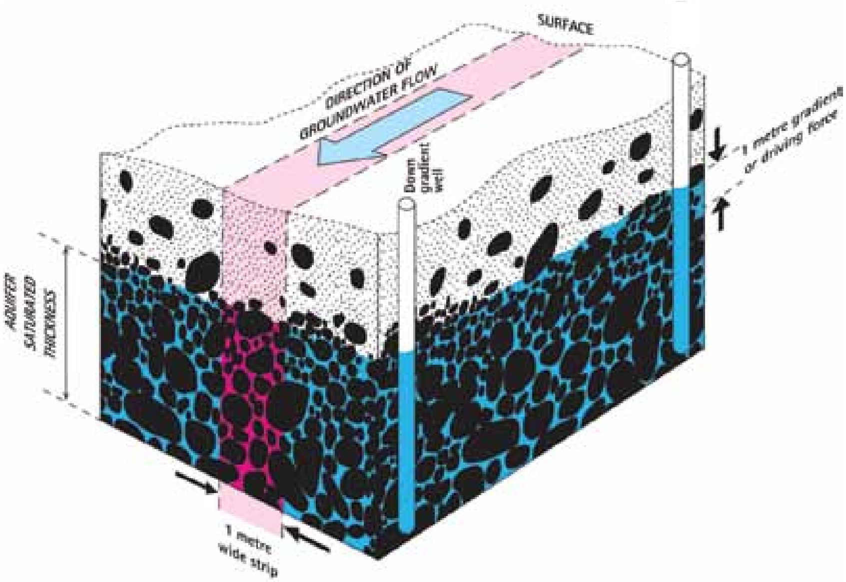

Figure 9.7: Definition of aquifer transmissivity

Porosity

Porosity is the percentage of the geological formation hosting an aquifer not occupied by solids. In other words, the proportion of solids to voids or gaps in a sedimentary formation. The higher the porosity the more groundwater an aquifer can store.

While clay has a high porosity of 50% or half of its volume available in the form of interstices to store water, the gaps are microscopically small meaning water can only move slowly between them. Furthermore due to the large surface area, water is tightly retained by capillary forces. Clays do not make good aquifers whereas gravels with a lower porosity of 10% do. This is because groundwater is not as tightly held in the pores of a gravel formation and pores are more interconnected, allowing water to drain freely or move towards a well when a pump is operating.

A porosity of 25 percent is commonly assumed to represent the unconfined aquifer. An indirect method of deriving groundwater velocities is to divide average groundwater residence times by the distance travelled. Shallow groundwater in the unconfined Wairau Aquifer has a mean residence time of less than five years, but is generally less than one year old.

During this time it covers a distance of up to ten kilometres after having left the Wairau River channel as leakage and flowing east before re-emerging as spring flow. This means groundwater is flowing at a rate of about 30 metres per day which is in agreement with the velocity calculated in Table 9.4.

One of the complicating factors in assessing groundwater flow rates in local alluvial gravels is the variability of the material and the presence of preferred pathways where velocities can be orders of magnitude higher and which act like pipes. This increases the productivity of wells, but also provides an avenue for contaminants to move more quickly towards a well.

Aquifer transmissivity

Transmissivity is a measure of the capability of the aquifer to transmit groundwater through a one metre wide band over its full depth, under a one metre or unit gradient (Fig. 9.7). The units of transmissivity are m^3/day/metre, which simplifies to m^2/day. Another way to think of transmissivity is the ease with which water can be extracted from the aquifer. If there is considerable resistance to groundwater flow through the host sediments the transmissivity will be low. Transmissivity is equivalent to hydraulic conductivity multiplied by the aquifer thickness.

Aquifer transmissivity is most easily determined from aquifer pumping tests. Pumping tests are an intensive and costly activity and they are usually only carried out to resolve a specific issue. As a result, less than 2% of the wells drilled on the Wairau Plain have been subject to aquifer pumping tests (Table 9.5).

Aquifer Area	Number Of Pump Tests	Mean (m^2/day)	Median (m^2/day)	Standard Deviation (m^2/day)
Riverlands Aquifer	16	297	290	129
Wairau Aquifer, Confined	20	3,705	2,970	2,964
Wairau Aquifer, Springs Area	14	5,014	3,289	4,704
Wairau Aquifer, Unconfined	14	5,138	4,852	4,186
Woodbourne	3	3,456	3,917	3,051
North Bank	7	5,441	4,400	5,495
Rarangi Shallow Aquifer	7	590	518	335
Southern Valleys	7	296	55	378

Table 9.5: Marlborough transmissivity values

Aquifer tests are relatively rare in Marlborough with well productivity tests being far more common. The latter are usually performed after screening and developing a well. This productivity test involves measuring the drawdown after pumping the well at a constant rate for anywhere between 30 and 120 minutes. The test usually concludes when the driller feels that the well level has stabilised.

Knowledge of the variation in transmissivity throughout an aquifer is useful for identifying boundaries where values will typically be lower than elsewhere. It will also generally reflect well productivity and indicates the likely well yield that can be expected in an area, can be used to calculate volumetric groundwater flow or velocity, and for predicting the likely effect of pumping on springs or other wells (Fig. 9.8).

The transmissivity of unconfined aquifers varies seasonally depending on the volume of groundwater. For areas like Woodbourne where there are large seasonal variations in the saturated thickness of the aquifer, transmissivity can vary from very high in spring, to very low for tests carried out during summer. Tests at the Base Woodbourne tower wells during September 1982 indicated a transmissivity value of 7,095 m^2/day, while testing later on that season during January 1983 when well levels were four metres lower, gave a transmissivity value of just 625 m^2/day (Fig. 9.9).

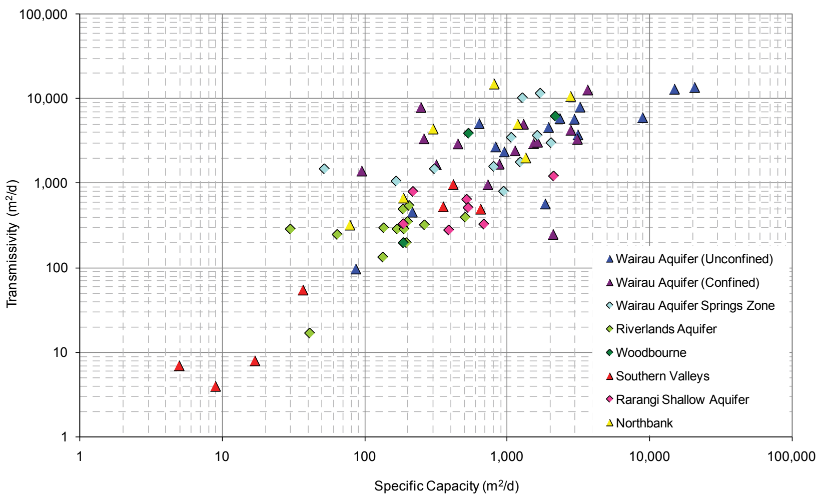

Figure 9.9: Transmissivity versus specific capacity for Wairau Plain aquifers

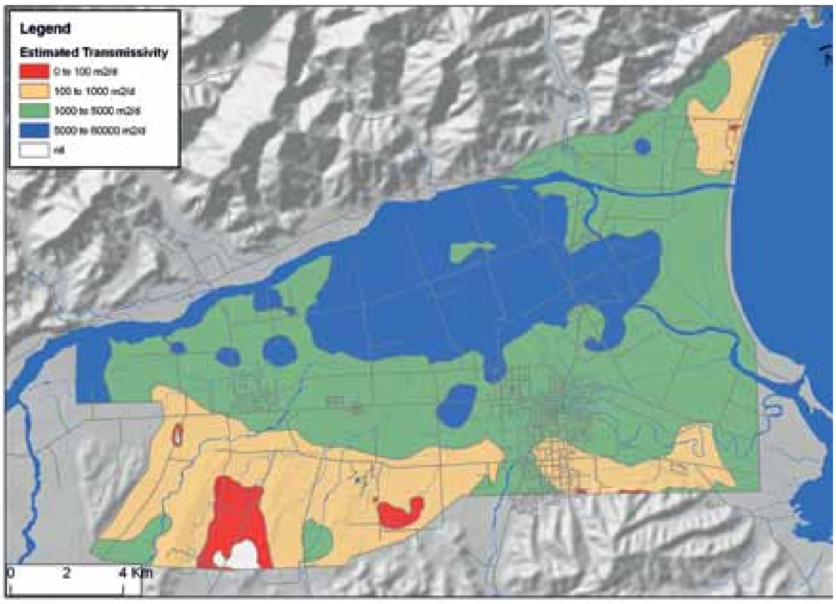

Figure 9.10: Aquifer transmissivity estimated from well specific capacity

In the central areas of the Wairau Plain from Renwick to the Cloudy Bay coast, transmissivity values are generally in the range from 2,000 to 10,000 m^2/day. Along the southern fringe of the Wairau Plain through to south-eastern Blenheim, values can be several orders of magnitude less than during summer, but similar in wetter seasons (Fig. 9.10).

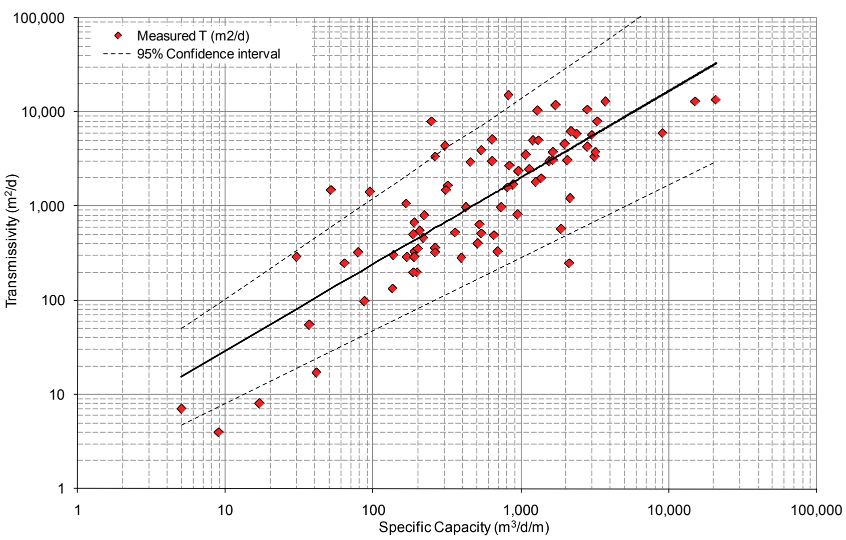

Figure 9.8: Empirical relationship between transmissivity and specific capacity for Wairau Plain

Values for the Riverlands Aquifer are typically an order of magnitude lower at around 300 m^2/day. Further south in the Southern Valleys Catchments, aquifer transmissivity is typically less than 100 m^2/day, and commonly much lower.

Well specific capacity

The pumping rate divided by the change in well water level is known as the specific capacity of the well, with a higher figure equating to a higher producing well, and therefore by implication, aquifer. It normally has the units of m^3/hour per metre of drawdown.

There is generally a good correlation between well performance and aquifer yield unless the well is inefficient and a greater drawdown occurs than otherwise reflects the aquifer properties. It is useful to develop a relationship between specific capacity and transmissivity. In areas where no pumping test results exist, transmissivity can be derived from the more abundant specific capacity information. Over 1,300 (about 30%) of wells on the Wairau Plain have been tested for productivity. Of these, only 80 have also had aquifer pumping tests carried out on them.

The theoretical relationship between specific capacity and transmissivity is linear on a log scale (Walton - 1970). Several authors have also developed empirical or observed relationships between specific capacity and transmissivity (Logan - 1964, Driscoll - 1986, Razack and Huntley - 1991, Bal - 1996). An empirical relationship for the Wairau Plain has been determined by linear regression of the log-transformed information and is expressed as:

Transmissivity= 3.526 x Specific Capacity$^{0.92}$

The correlation isn't perfect with a third of the observed variance in specific capacity unexplained by transmissivity alone. This variance reflects differences in the duration of pumping, well development or construction, storage in the well casing and extra drawdown caused by well inefficiencies, all of which affect the value of specific capacity.

Vertical flow - aquitards and aquicludes

An aquifer has been defined as a water bearing layer which produces economically useful amounts of water. Conversely aquicludes and aquitards form natural barriers to flow. Aquicludes are incapable of transmitting groundwater because they are formed of impermeable material. There are probably no complete aquicludes, as water can always move through any strata, albeit at an infinitesimally slow rate for some sediments. However, bedrock such as the strata of the Pukaka Ranges forming the northern boundary of the Rarangi Shallow Aquifer is a local example that is close to being an aquiclude (Fig. 9.11).

Aquitards on the other hand are geological formations which allow some groundwater flow, but it is small relative to that of an adjoining aquifer. The eastern Wairau Aquifer for example, is confined by a bed of fine grained marine clays or silts that restrict but don't altogether stop upwards flow through the aquitard, even though it can be tens of metres thick. The most extensive example of an aquitard in Marlborough is the Dillons Point Formation, which provides the confining layer over the top of the coastal Wairau Aquifer.

Aquitard properties

A number of properties are used by hydrologists to define how leaky an aquitard is but the one most commonly used in Marlborough is the aquitard leakage term (ALT). Aquitard leakage is defined as the hydraulic conductivity (K') of the confining bed divided by its thickness (B') (Hazel - 1975).

Values of ALT are calculated from pumping test measurements but are often problematic to measure accurately in coastal aquifers because of tidal fluctuations. On the Wairau Plain the values tend to become smaller to the east with the thickening of the aquitard, although higher values may be found in areas where there are discontinuities in the layer (Table 9.6).

Very low ALT values mean the aquifer is highly confined either because the aquitard is thick, its hydraulic conductivity is low, or a combination of both. Low

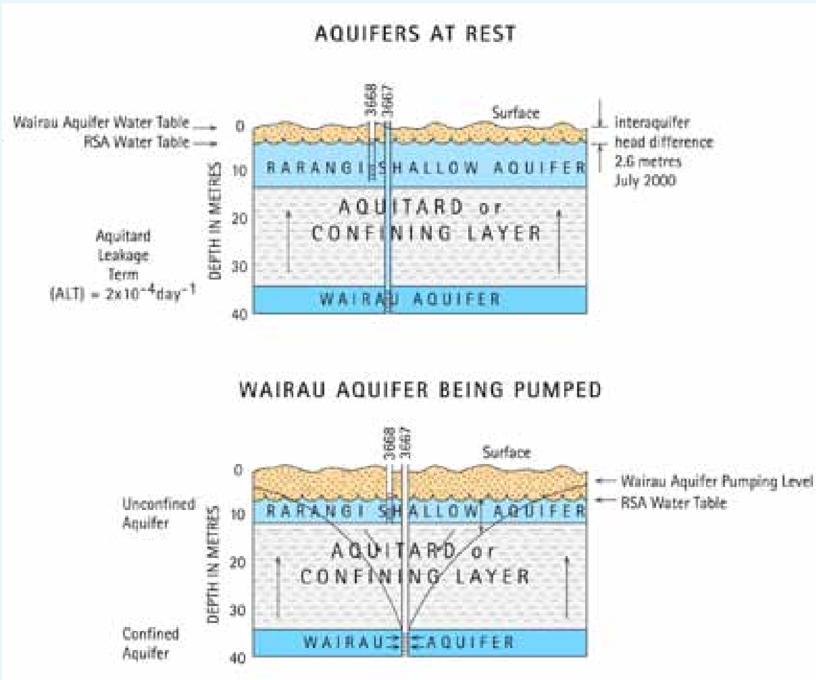

Figure 9.11: Aquitard leakage.
The paired MDC monitoring wells 3668 and 3667 are shown intercepting the Rarangi Shallow Aquifer (RSA) and Wairau Aquifers respectively, which are separated by the aquitard or confining layer. The shaded blue area represents the static level of groundwater in each aquifer. Aquifer levels measured in these wells in July 2000 showed levels were 2.6 metres higher in the confined Wairau Aquifer relative to the RSA. This vertical discontinuity in aquifer level at the same location indicates that upward flow predominates over horizontal groundwater flow in the coastal area of Cloudy Bay.

Source	Pumped Well	Test Date	Observation Wells	Aquifer	Transmissivity (m^2/day)	Storativity	Specific yield Of Unconfined Aquifer	Aquitard Leakage Term K′/B′ ($days^{-1}$)	Vertical Conductivity K′ (m/day)	Streambed Conductance λ (m/day)
Assessment of Wairau Aquifer Groundwater Abstraction PDP report C01552402 prepared for Rarangi Holdings Trust	4735	2006	4743	Wairau Aquifer	3,312	8×10^{-5}		1.08×10^{-3}	0.028	
			3961		4,320	3×10^{-5}		2.16×10^{-4}	0.006	
			3667		3,312	4×10^{-5}		6.48×10^{-4}	0.016	
			4639		6,912	8×10^{-5}		6.48×10^{-4}	0.018	
Assessment of Proposed Groundwater Abstraction by PF Olsen Group Ltd, PDP report C02052500	4785	March 2008	1733	Wairau Aquifer	8,000*	4×10^{-4}		2×10^{-4}	0.002	
			1789							
Mid 2010 MDC Grove Road municipal supply wellfield redevelopment test			10081	Wairau Aquifer	2,337	1.1×10^{-3}	-	0.17		
			10082		1,899	5×10^{-4}		0.105		
			2993		672	2.8×10^{-3}		-		
			10083		5,040	2×10^{-4}		0.01		
			3145		5,859	2×10^{-4}		0.031		
			0544		4,608	1.5×10^{-4}		1×10^{-5}		
MDC Mills & Ford Road East test: PDP re-analysis October 2004	4404	2004	0306	Wairau Aquifer	10,368	1×10^{-4}	0.2	0.0216		432
			4405		14,112	1.1×10^{-3}	0.2	0.0288		432
			2004		11,520	8×10^{-4}	0.2	0.216		432
MDC Drain N test: PDP re-analysis October 2004	3958	2004	4441	Wairau Aquifer	3,888	9×10^{-4}	0.08	0.1152	1.34	-
			274		3,528	1.5×10^{-3}	0.11	0.1584	1.01	
Kapiti Views Trust PDP report: Second Pump Test on Kapiti Views Well	4706	2008	1426	Southern Springs	3,800	1.3×10^{-4}	0.002	0.02		
			1849		5,000	5×10^{-4}	0.008	0.2		
			3848		4,100	8×10^{-4}	0.01	0.2	1.88	
			2511		6,000	2.8×10^{-4}	0.0025	0.065		
			1384		4,600	5×10^{-4}	0.0025	0.05		
			949		4,400	2.4×10^{-4}	0.0025	0.0055		
PDP Report CJ845 prepared for JPS Trust	1428	2003	1731	Southern Springs	1,600	0.15			10	5
			1404		1,500	4.5×10^{-3}			0.1	0.6
			2347		1,800	0.12			8	4
Aquifer Testing at Okaramio PDP report C01559400	4383	2004	Ob1 and Ob2	Pelorus Gravels	3,456	3.0×10^{-3}	0.02	0.13	4.73	7.2
Pump Test on Kaituna-Tuamarina Road - PDP report	4650	?????	4707	Waikakaho River Gravels Aquifer	2,000	2×10^{-4}	0.1	0.12		1
MDC Middle Renwick Road wellfield: PDP re-analysis October 2004	3120	2007	3119	Wairau Aquifer	3,050	1.3×10^{-4}	0.22	0.04	0.25	1080
			612		4,392	9.5×10^{-5}	0.3	0.04	0.72	1080
Anderson Test – PDP report CJ741	O28w/0124	1997	?	Wairau Valley Aquifer	940	-	0.016	0.03		1440

Table 9.6: Aquifer hydraulic properties

values imply a low rate of vertical flow through the confining layer for a given hydraulic gradient across it.

We can also use knowledge of the (ALT) to calculate the vertical flow velocity of groundwater across the aquitard. The daily rate of upwards groundwater flow is about 1mm/day, which represents a very slow rate when compared with local gravels aquifers. Knowledge of the ALT can also be used to calculate the upwards volumetric groundwater flow through the confining layer across cross-sectional areas of the Wairau Aquifer east of SH1 of 50 square kilometres. It predicts a daily upflow of 300 l/s. This is likely to be an order of magnitude estimate only, but appears sensible based on the 500 l/s rate shown in the Wairau Aquifer water balance.

The natural rate of upwards groundwater flow will respond to localised pumping, and variability in natural factors such as atmospheric pressure and river flow, all of which will affect the pressure gradient across the aquitard.

Three dimensional groundwater flow

Until now we have restricted discussion to one or two dimensional groundwater flows, and assumed relatively homogeneous geology. This approximation works well for the case of the unconfined Wairau Aquifer where groundwater flow is essentially horizontal. However, the same approach can't be taken in areas where there are multiple aquifer layers or significant vertical flow such as in parts of the Southern Valleys Aquifers, or the confined Coastal Wairau Aquifer.

In these areas structural differences caused by geology, such as the appearance of confining layers, creates pressure differentials within aquifers, or between water bearing layers of varying depth. Symptomatic of this are varying groundwater pressures in the same well depending on the depth at which measurements are made.

It is likely that variations in aquifer pressure with depth partly explain a 0.5 metre difference in well level between the centrally located MDC Bar well 1733, and its northern counterpart MDC well 3667 at Rarangi.

MDC Well Number	Lower Water Bearing Layer Depth (metres below surface)	Static Water (metres below surface)	Upper Water Bearing Layer Depth (metres below surface)	Static Water (metres below surface)	Hydraulic Grade (metres)
4135 near Wairau Bar	Water bearing from 48.3 to 51.6	0.3	Water bearing from 44.9 to 47.2	1.3	1 upwards
4785 near Wairau Bar	Water bearing from 49.1 to 51.5	0.17	Poor water bearing from 45.1 to 47.2	0.9	0.73 upwards
3667 north of Diversion	Water bearing from 37.5 to 40.16	0.8	Poor water bearing from 30.1 to 33	3.37	2.57 upwards

Table 9.7: Vertical groundwater pressure gradient

Examples of increasing groundwater pressure with depth below the surface can be seen at three separate coastal Wairau Plain wells (Table 9.7). The first two wells are centrally located and intercepted two distinct confined water bearing layers over their 52 metre depth. Both exhibit an upwards gradient of around one metre. The third well is slightly shallower at 40 metres depth and has a 2.57 metre difference in groundwater level across neighbouring water bearing layers.

Another factor to consider when interpreting observations of well water level is the homogeneity of the material forming the aquifer, or the degree of uniformity. Because groundwater flow speeds up or slows down depending on the permeability of the geology, differences in the shape of a flow net in a heterogeneous aquifer will not necessarily only reflect aquifer flow or gradient. The widening of a stream tube for example, could be explained by decreasing permeability for the same grade.

Groundwater storage

Storage is an important characteristic of an aquifer because it allows water to be stored throughout the year. This is analogous to a bank account which allows withdrawals between deposits as long as there are savings.

The volume of groundwater stored in an aquifer depends on the porosity, but not all of the water stored in the pore spaces can be extracted, or will drain under the force of gravity. For this reason when we define aquifer storage what is really meant is yield.

Specific yield refers to the fraction of water that will naturally drain out of a porous rock, while specific retention is the residue that is bound by capillary forces. The sum of the two equals the porosity of a particular rock type. The specific yield of clays is low at only 3% compared to values of around 20% for gravels or sands. In other words gravels drain much more freely.

The larger the specific yield, the more water an aquifer produces for each metre a water table is lowered by pumping. However this concept of specific yield only applies to unconfined aquifers where changes in storage represent an actual dewatering or drainage of the pore spaces.

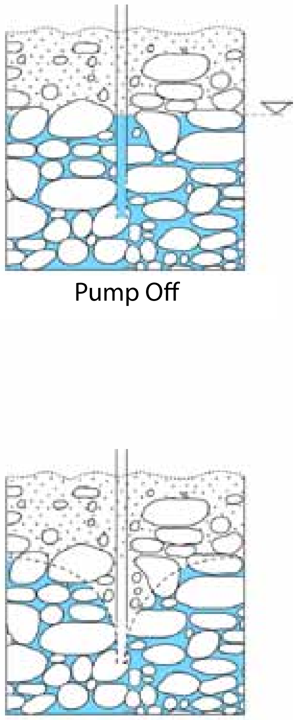

Figure 9.12: Unconfined aquifer response to well pumping

Unconfined aquifers are not capped and can store and release water by changes in the elevation of the surface of the water table (Fig. 9.12). Unconfined aquifers are more productive for the equivalent fall in aquifer level, and generate effects over a much more localised area than confined aquifers (Fig. 9.13).

To reflect this different mechanism operating in confined aquifers, its storage is referred to as storativity or storage coefficient. This distinguishes it as a secondary effect of aquifer compaction or expansion.

A confined aquifer is normally always full of water, even when being pumped, and has very little capacity to store more. This is because groundwater is pressurised by the overlying clay capping layer which acts as a seal,

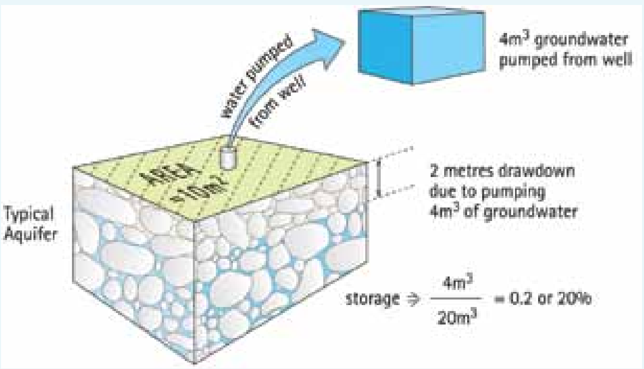

Figure 9.13: Conceptual diagram of aquifer storage for an unconfined aquifer. If a typical 10 m^2 area of the Wairau Aquifer near Renwick yields 4 m^3 of groundwater resulting from a two metre fall in well level, the yield (specific yield) would equal 20%. In contrast, under typical confined aquifer conditions, the same two metre drawdown is likely to generate only about 0.01 m^3 of water from storage. This equates to a yield (storativity) of 0.05%.

in conjunction with recharge water entering at a higher elevation. Extra water can only enter if the water pressure increases and this has the effect of inflating the aquifer slightly. This is called the elastic storage component and is reversible, meaning a confined aquifer stretches to accommodate extra recharge and relaxes when water is pumped from it.

We know confined aquifers are under pressure because the water levels in wells tapping them sit near or above groundlevel and sometimes free flow (artesian). This is analogous to inserting a series of straws into a water bed and watching the water level rise above the surface of the bed and squirt out.

When a confined aquifer is pumped groundwater levels fall in the pumped and neighbouring wells, but the water level usually remains above the confining layer meaning groundwater continues to be pressurised. Because the water in the confined aquifer is pressurised the pores spaces are perpetually saturated. As water is pumped, the groundwater supplying the well originates from a slight expansion of the water itself, and from the fine pore spaces of clay layers forming the confining layer or interbedded layers within the aquifer. (Fig. 9.14)

Pumping reduces the aquifer pressure which in turn causes the compressible material forming the aquifer such as the clay confining layer, to contract. This squeezing forces water towards the well. Normally the reduction in the thickness of the confining layer or clay beds isn't large enough to be propagated to the surface, which explains why landowners don't notice the changes.

This means the aquifer itself only acts as a conduit to deliver water to the well from distant parts of the aquifer and only produces a fraction of the groundwater volume compared to the drainage of pore spaces in an unconfined aquifer for the same pumping rate. Furthermore, groundwater is scavenged over a large aquifer area which explains why the cone of influence generated by pumping a confined aquifer is so large, with a radius that may extend up to several kilometres from the well. If sufficient groundwater is extracted from a confined aquifer and the groundwater level falls below the geological beds confining it, it behaves as an unconfined aquifer and produces far more groundwater.

The degree of confinement also determines the way an aquifer behaves when it is pumped. Typically the storativity of confined aquifers is between 10^{-4} and 10^{-5}, whereas the specific yield of an unconfined aquifer falls in the range from 0.1 to 0.3. The four orders of

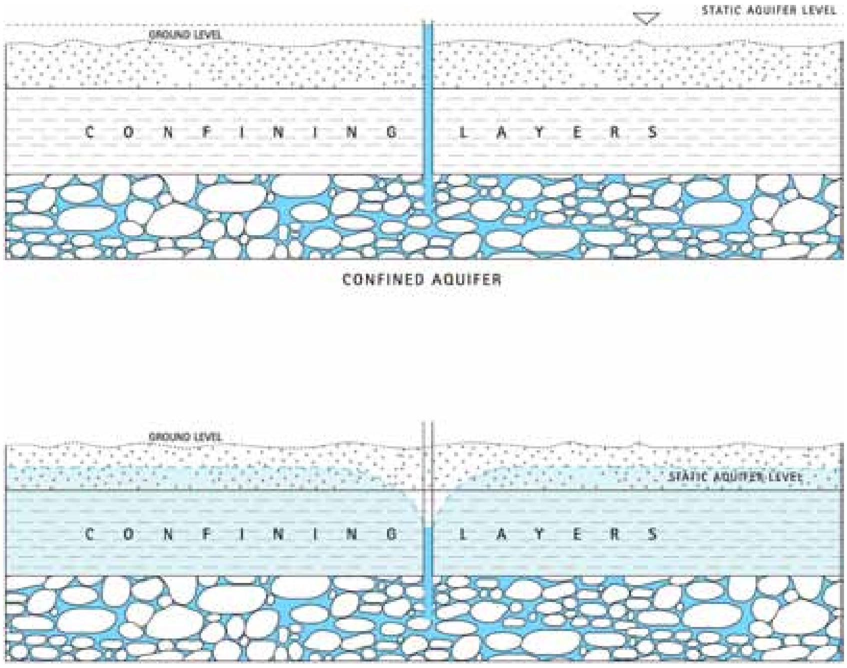

Figure 9.14: Confined aquifer response to well pumping

magnitude difference reflects the greater storage associated with draining pore spaces compared to the elastic changes in confined aquifers which release very little water. This makes storage a good indicator of the type of aquifer structure one is dealing with. As a consequence of their low storage, confined aquifers respond quickly to recharge events or pumping, whereas in unconfined aquifers, longer pumping periods are required to achieve the same drawdown (Table 9.8).

Aquifer subsidence

A potential consequence of over-pumping a confined aquifer can be a permanent change to the structure of the geological formation, which can lead to subsidence at the surface. There is no evidence of subsidence occurring in Marlborough to date.

Land surface subsidence was potentially a consequence of the 15 metre fall in Benmorven Aquifer levels experienced during the 2000/01 summer drought. The MDC sought specialist advice and was advised the risk was low due to the high gravel content of the local aquifers.

Area	#	Mean	Median	Standard Deviation
Riverlands Aquifer	14	8.8×10^{-5}	8.0×10^{-5}	8.5×10^{-5}
Wairau Aquifer, Confined	19	3.2×10^{-4}	1.0×10^{-4}	3.7×10^{-4}
Wairau Aquifer, Springs Area	11	0.026	0.001	0.054
Wairau Aquifer, Unconfined	4	0.097	0.119	0.068
Woodbourne	3	0.102	0.005	0.172
North Bank	2	0.150	0.150	0.212
Rarangi Shallow Aquifer	6	0.103	0.075	0.098
Southern Valleys	1	0.010	0.010	-

Table 9.8: Summary storage values for Wairau Plain aquifers

The Benmorven Aquifer was of particular concern due to the presence of thick clay beds which were considered to be compressible. According to expert opinion the highest risk is associated with strata formed of highly compressible materials such as peat and clay. Peat is common in the Bells Road area underlying what was once known as the Great Fairhall Swamp.

Changes in aquifer storage

Storage coefficient can be used to calculate the volume of water contained within an aquifer or that will be released when it is pumped. For example, the volume of groundwater added to storage as a result of a one metre rise in well levels across the 7,000 ha middle section of the Wairau Aquifer, is equal to a volume of 7 million m^3 of water. This is equal to the aquifer area multiplied by the rise in aquifer level of one metre by the average aquifer storage coefficient of 0.1. However, not all of the groundwater stored in an aquifer can be accessed. Most is needed to provide the gradient for groundwater to flow or to maintain spring flow, and keep the seawater interface in a safe position. Some water will also be so tightly held by natural capillary, friction or suction forces, that no pump will be able to extract all of it. This is analogous to plant roots trying to extract soil moisture which is tightly bound to soil particles in dry conditions.

Other aquifer properties

There are also a number of assumptions made when simplifying groundwater systems so we can practically analyse them. The first concept relates to uniformity in space or direction within an aquifer. It is often assumed aquifers are homogeneous and isotropic, meaning their properties are the same everywhere and independent of the direction of flow. By contrast a heterogeneous and anisotropic aquifer is made up of material that varies depending on location and the direction of water flow. Anisotropy is a common property of the alluvial type aquifers found in Marlborough and is typical of aquifer gravels deposited by the Wairau River (Fig. 9.15). Because they aren't round they tend to lie with their flat side down, making it easier for groundwater to flow horizontally through the gaps, but flow is slower vertically.

Bouwer (1978) found that the vertical hydraulic conductivity may only be 10% to 20% of the horizontal conductivity, and that this phenomenon is the rule

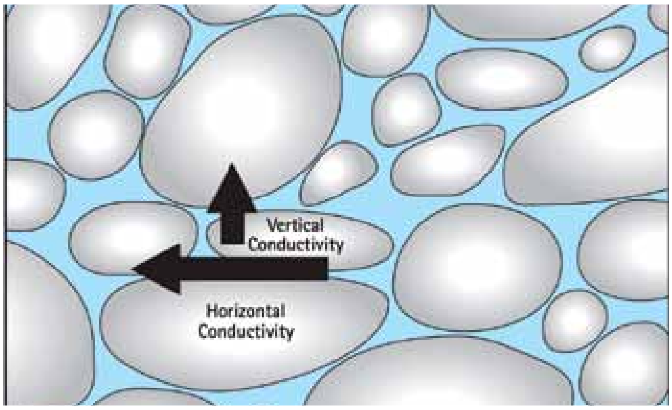

Figure 9.15: Anisotropy

rather than the exception for undisturbed alluvial deposits. It is likely that the degree of anisotropy is even higher for alluvial aquifers on the north-bank of the Wairau River formed by schist derived alluvium because of the platy nature of this material. In reality most aquifers are heterogeneous and anisotropic.

Another principle in common use involves the uniformity of flow conditions with time. Steady state means the rate of flow is constant with time. Conversely, non-steady state or transient conditions mean that inputs or outputs to the system vary over time. At a local scale a well may reach steady state when its pumping rate equals the recharge rate of the surrounding aquifer. At a regional scale steady state conditions occur when the volume of water stored in an aquifer isn't changing because the natural drainage rate is balanced by recharge. This situation is most likely to occur in winter. Steady state means equilibrium conditions exist.

Finally, it is important to consider that all aquifers have boundaries. Real world aquifers aren't of infinite extent. It is important to know how groundwater is affected by boundaries or barriers such as confining layers which may restrict its movement. Boundaries are formed by less permeable sediments such as greywacke and schist basement material.

Solid boundaries formed by mountains, such as the Richmond or Pukaka Ranges, influence the behaviour of wells close by. These types of boundaries reduce well yield compared to a well operating in the centre of the aquifer. Conversely aquifers bounding the Wairau River benefit directly from natural recharge, or when a pumping well induces channel flow into the aquifer from a nearby stream such as Doctors Creek.

Soil properties

Any discussion of aquifer recharge or patterns of crop water use must include the role of the soil layer. This is because soils form the unsaturated or so called vadose zone separating recharge water arriving as rain at the land surface from the underlying groundwater table.

The water content of soil also potentially affects plant growth and influences a wide range of soil properties or functions such as its aeration status, temperature, cohesion, strength, microbial activity, nutrient availability and its leaching potential of surface contaminants such as pesticides or fertilisers.

The thickness of the unsaturated layer and the size and arrangement of its particles are the physical parameters which largely determine how water traverses any particular soil. The upper one metre or so of the soil profile influences land surface recharge the most, and is also the layer that is the most clearly understood.

To fully understand the complexities of water availability in soils it is also important to understand the concept of soil water potential. The soil water potential is the sum of all the forces acting on soil water including gravitation, matric, pressure and solute potential. The relationship between soil water content and soil water potential provides information about the water holding capacity and drainage characteristics of a soil.

The amount of water held by any particular soil is predominantly influenced by soil texture and structure. The soil texture and structure governs the size of the pore spaces which in turn controls how tightly the water is bound. For any given rain event, heavy soils made up of fine grained, clay or silt sized grains will retain more water compared to a thin, freer draining soil.

This is commonly referred to as the soil moisture retention curve and is unique for each category of soil type (Fig. 9.16). Clay loam soil has higher water contents for any given soil water potential compared to either the silt loam or sandy loam soil.

This is because the clay soil is finer textured with greater amounts of small pores which can hold onto more water at a higher suction (Fig. 9.17). At the other extreme the sandy loam soil has greater amounts of larger sized pores and the water in this soil can be removed at relatively lower suctions (Fig. 9.18).

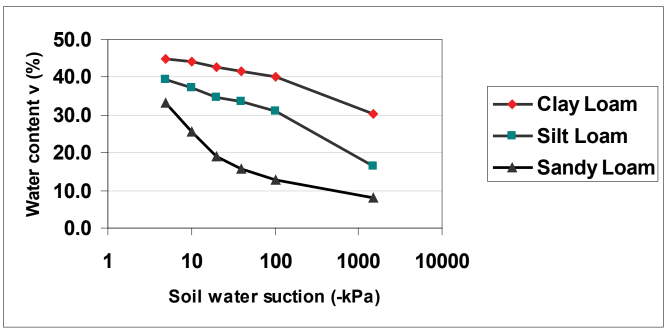

Figure 9.16: Soil moisture release characteristic curves showing the effect of soil texture

Figure 9.17: Clay loam soil.

The difference in the distribution of pores between clay, silt and sandy loam soils is reflected in the shape of the moisture characteristic curves. The clay soil has a gradual curve which indicates a fairly uniform distribution of pore sizes. In contrast the sandy soil loses a lot of moisture over a more restricted pore size range while the silt loam is somewhere between the two soils.

Figure 9.18: Sandy loam soil.

Soil water classification

While the soil water potential concept allows a precise definition of the soil water status, for practical application in the field a series of more general terms have been defined to classify the soil water content.

Saturated water content

The maximum amount of water a soil can hold is called its saturated water content. It occurs when all the pores in a soil are full of water and no air remains in the soil. It is governed by the total number of pores in a soil although in reality it is rarely reached as most soils have some air trapped as bubbles within soil aggregates.

Field capacity (FC)

When a saturated soil begins to drain it is the largest pores which empty first. After several days when the drainage of macropores is complete and the soil water content has become relatively stable, the soil is said to be in a state called field capacity. Field capacity is often related to the soil moisture content at a potential of -10 kPa.

Permanent wilting point (PWP)

Over time if the soil receives no further moisture, drainage to groundwater, transpiration by plants and evaporation from the soil will cause it to dry to below field capacity. As it does so water becomes more difficult for a plant to abstract through its root system because it is held at greater suction by the soil.

The stage where the amount of water in the soil is such that plants are permanently wilted is called the permanent wilting point. It is typically related to the soil moisture content at a potential of -1500 kPa. The remaining water is held so tightly within micropores and sorbed onto soil particles, it is unavailable to the plant.

Soil type	Horizon	Field Capacity[1]	Permanent Wilting Point[2]	Total Water Available
Kaiapoi sandy loam	A	32.1	11.5	20.6
	B	26.1	9.1	17.0
Paynter clay loam	A	38.6	26.7	12.0
	B	44.2	30.3	13.9
Wairau silt loam	A	37.8	16.4	21.5
	B	35.0	19.5	15.5
Renwick silt loam	A	36.4	17.6	18.7

[1] = moisture content -10 kPa (0.1 bar)
[2] = moisture content at -1500 kPa (15 bar)

Table 9.9: Approximate moisture contents (% v/v) of selected soil on the Wairau Plain.

Available water (AW)

The amount of water a soil can store for plant growth is numerically equal to the amount of water held at field capacity minus the amount held at permanent wilting point (PWP) (Table 9.9).

AW% = FC (% v/v) – PWP (% v/v)

References

Bal, A.A., 1996. Valley fills and coastal cliffs buried beneath an alluvial plain: evidence from variation of permeabilities in gravel aquifers, Canterbury Plains, New Zealand. Journal of Hydrology (NZ), 35(1):1-27.

Begg, J. G. and Johnstone, M.R. (compilers), 2000. Geology of the Wellington area. Institute of Geological and Nuclear Sciences 1:250 000 Geological Map 10. Institute of Geological and Nuclear Sciences, Lower Hutt, 1 sheet + 64p.

Begg, J.G. and Mazengarb, C., 1996. Geology of the Wellington Area: scale 1:50,000. Institute of Geological and Nuclear Sciences Geological Map 22, Lower Hutt.

Bouwer, H. 1978. Groundwater Hydrology

Broadbent, M. and Callander P.F. 1991. A resistivity survey near Waimakariri River, Canterbury Plains, to improve understanding of local groundwater flow and of the capabilities of the survey method. New Zealand Journal of Geology and Geophysics, 1991, Vol. 34: 441-453

Brown, L. J., 1981a. Water Well Data Northern Marlborough, Part Sheet P28 Wairau Plain (Geology and water wells), 1:25000. New Zealand Geological Report 93.

Brown, L.J., 1981b. Late Quaternary geology of the Wairau Plain, Marlborough, New Zealand. New Zealand Journal of Geology and Geophysics 24: 477-490.

Cunliffe, J.J., 1988. Water and Soil Resources of the Wairau Volume Two: Groundwater. Marlborough Catchment and Regional water Board, 107p.

Davidson, P.W., and Scott, D., 1994. Wairau Aquifer Resource and Issues. Technical Report Prepared for Marlborough District Council.

Environment Canterbury. 2000. Guidelines for the Assessment of Groundwater Abstraction Effects on Stream Flow, Technical Report ROO/11 First Edition Environment Monitoring Group

Freeze, A. F., & Cherry, J.A. 1979. Groundwater

Hazel, C.P. 1975. Groundwater Hydraulics. Lectures presented by Irrigation and Water Supply Commission, Queensland to the Australian Water resources Council's groundwater school Adelaide 1975

Krusemann, G.P. and de Ridder, N.A., 1970. Analysis and evaluation of pumping test data. International Institute for Land Reclamation and Improvement, 200p.

Lensen, G.J., 1962. Sheet 16: Kaikoura, Geological Map of New Zealand. DSIR, Wellington.

Logan, J. 1964. Estimating transmissivities from routine tests of waterwells. Groundwater 2: 35-37.

North Canterbury Catchment Board and Regional Water Board, 1986. The Christchurch Artesian Aquifers

Ota, Y., Brown, L.J., Berryman, K.R., Fujimori, T., Miyauchi, T., 1995. Vertical tectonic movement in northeastern Marlborough: stratigraphic, radiocarbon, and paleoecological data from Holocene estuaries. New Zealand Journal of Geology and Geophysics 38: 269-282.

Rae, S.N, 1988. Water and Soil Resources of the Wairau Volume One. (Groundwater Resources), Marlborough Catchment and Regional Water Board

Rattenbury, M.S., Townsend, D.B., and Johnstone, M.R. (compilers), 2006. Geology of the Kaikoura area. Institute of Geological and Nuclear Sciences 1:250 000 Geological Map 13. Institute of Geological and Nuclear Sciences, Lower Hutt, 1 sheet + 70p.

Razack, M.D. and Huntley, D., 1991. Assessing transmissivity from specific capacity data in a large and heterogeneous alluvial aquifer. Groundwater 29 (6): 856-861.

Stewart, M.K. 2004b. Age dating of groundwater in the Rarangi area. Report written for Marlborough District Council. Institute of Geological Sciences, Lower Hutt.

Suggate, R.P., 1965. Late Pleistocene geology of the northern part of the South Island. New Zealand Geological Survey Bulletin 77, 91p.

Taylor, C.B., Brown, C.J., Cunliffe, J.J. and Davidson, P.W., 1992. Environmental tritium and 18O applied in a hydrological study of the Wairau Plain and its contributing mountain catchments, Marlborough, New Zealand. Journal of Hydrology (Amsterdam) 138: 269-319.

Thorpe, H.R, 1992. The hidden resource. In Groundwater's of New Zealand. M.R. Rosen and P.A. White (eds). New Zealand Hydrological Society Inc., Wellington. Pp. 161-183.

Todd, D.K. 1980. Groundwater Hydrology

TSE Group Consultants, 1979. Waitaki NZ Refrigerating Ltd Report on Investigations for the supply of water to the proposed new freezing works at Blenheim

Ward, R.C. 1975. Principles of Hydrology

Walton, W.C., 1970. Groundwater Resource Evaluation. McGraw-Hill Book Company, 664p.

Wellman, H.W., 1955. New Zealand Quaternary tectonics. Geologische Randschau. 43, 248-257.

White, P.1986. Shallow Groundwater Flow Direction and Velocity at the Renwick Refuse Tip, Marlborough. Report No. WS 1109, Water and Soil Science Centre Christchurch, Ministry of Works and Development for the National Water and Soil Conservation Organisation

Wilkinson, S.J., 2005. A hydrological study of the Rarangi area with reference to arsenic contamination. Report submitted in partial fulfilment for the degree of BSc (Hons) in Engineering Geology, University of Canterbury.

Wilson, S.R., 2008. Riverlands Groundwater Model and Aquifer Sustainability Assessment. Technical report prepared for Marlborough District Council by Water Matters Limited, Wellington, 42p