5.0 Non-uniform Hydrogeological Settings

The analytical equation described in section 4.1 and plotted in Figure 23 is an easily applied solution for a stream which can be represented as a straight line near a well in a laterally extensive aquifer. For more complex hydrogeologic settings a numerical modelling approach is recommended, as described in section 4.2.

Because the use of numerical modelling is a specialised and often expensive process, this section of the guideline has been prepared to provide some simple indicative approaches for common hydrogeological settings which do not readily fit the requirements of the analytical equation. The situations considered are:

» *A well bounded on either side by two streams (section 5.1);*

» *A well located upstream of the headwaters of a springfed stream (section 5.2);*

» *A well located near an artesian spring that penetrates through a low permeability surface confining layer (section 5.3);*

» *A well located on river flats formed by recent alluvial gravels which are bounded on either side, and underneath by lower permeability strata (section 5.4).*

It is important to recognise that the measures described in sections 5.1 – 5.4 do not represent a detailed assessment of each situation, but rather provide a preliminary indication that can be used to assess the likely significance of stream depletion in the setting that is described.

5.1 A Well Bounded by Two Streams

This situation is shown schematically in Figure 25. The assumption is made that both streams have a similar bed conductance.

A set of numerical modelling simulations have been undertaken (PDP, 1995) for the situation where each stream has very good hydraulic connection to the aquifer (i.e. $\lambda = 5{,}000$ m/day). This is expected to be a conservative assessment whereby the streams have the maximum impact on the pumping well response.

The modelled aquifer was 20 m thick with a hydraulic conductivity of 50 m/day and a storage coefficient of 0.1. Simulations assumed a well pumping at 20 L/s for 30 days followed by a further period of no pumping for 30 days. Two different settings were considered and the position of the well was varied between the two streams which were set at 1,000 m and then 1,800 m apart for each of the different settings.

The results of these simulations show that the effect of a second stream bounding the pumping well is to reduce the stream depletion ratio from the single stream assessment and to increase the overall stream depletion ratio from the two streams combined, compared to a situation where only one stream is present. For the simulations carried out in this study the reduction in 30 day stream depletion ratios as

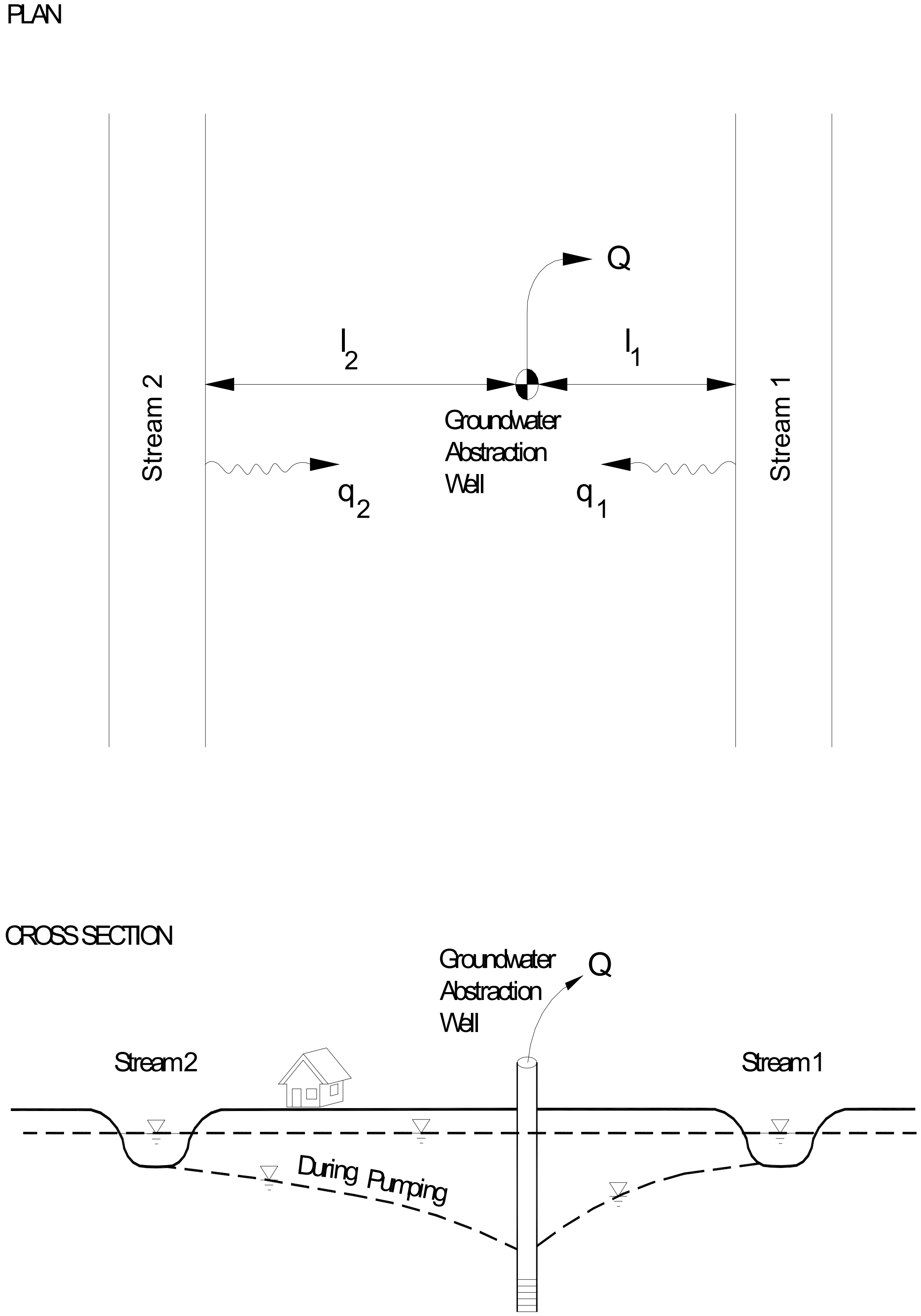


Figure 25: Schematic View of Two Streams Bordering a Groundwater Abstraction Well

a result of introducing a second stream is generally small for the closest stream (less than 12% reduction) compared to the values calculated using the analytical equation described in section 4.1.

For the more distant of the two streams there was very good agreement between the model and analytical equation when the streams were 1,800 m apart. However, when the streams were 1,000 m apart, the analytical equation significantly over-estimated the stream depletion ratio for the more distant stream of the pair.

The simulations have also shown that when pumping rates reduce, there is a faster reduction in stream depletion ratios for a two stream situation compared to a one stream situation. Consequently, the analytical equation will over estimate the stream depletion ratios that occur after pumps have been switched off.

For the purposes of preliminary assessment, the result of these simulations indicate that the estimates using the analytical equation become less accurate when the combined stream depletion effect from the two streams is greater than 90% of the well pumping rate. Under these circumstances the numerical model indicates that the stream depletion rate from each of the two bordering streams is proportional to their separation distance from the pumping well.

In practice, this means that after 30 days pumping, the cone of depression from the pumping well has extended out to the streams which provide a source of recharge to the aquifer on either side of the well. The cone of depression is in a steady state so that no more water is drawn from aquifer storage, and all the well water is drawn from stream seepage.

Consequently, the following rules can be used to correct the analytical calculations:

- When two streams border the well and $q_{1,\text{analytical}} + q_{2,\text{analytical}} < 0.9\,Q$ then the analytical calculation can be used for each stream.
- When two streams border the well and $q_{1,\text{analytical}} + q_{2,\text{analytical}} \geq 0.9\,Q$ and $< Q$ the smaller of the following two options are used:

$$q_{\text{analytical}}$$

or

$$q_1 = \left(\frac{\ell_2}{\ell_1 + \ell_2}\right)\left(q_{1,\text{analytical}} + q_{2,\text{analytical}}\right) \text{ and } q_2 = \left(\frac{\ell_1}{\ell_1 + \ell_2}\right)\left(q_{1,\text{analytical}} + q_{2,\text{analytical}}\right)$$

- When two streams border the well and $q_{1,\text{analytical}} + q_{2,\text{analytical}} \geq Q$ then the following approximation should be used:

$$q_1 = \left(\frac{\ell_2}{\ell_1 + \ell_2}\right) Q \text{ and } q_2 = \left(\frac{\ell_1}{\ell_1 + \ell_2}\right) Q$$

where: ℓ_1 = distance between pumping well and stream 1 [L];

$q_{1,analytical}$ = stream depletion rate from stream 1, calculated by the analytical equation described in section 4.1 [L^3/T];

ℓ_2 = distance between pumping well and stream 2 [L];

$q_{2,analytical}$ = stream depletion rate from stream 2, calculated by the analytical equation described in section 4.1 [L^3/T];

Q = pumping rate from the well [L^3/T].

The lesser of these three options has been found to give the best match to the numerical simulations reported in PDP (1995).

5.2 An Upstream Well

This situation is shown schematically in Figure 26. It will typically occur in areas around the head waters of streams, or for sections of streams which periodically go dry and re-emerge further downstream. In these circumstances, the source of stream flow originates from the intersection of the water table with the streambed. As the water table fluctuates due to climatic and pumping effects, the flow in the stream varies and the position of the headwaters moves laterally, as shown in Figure 3. In these situations a decision is often required regarding the location of the stream section to be considered in the analysis. Typically the analysis would consider the section of stream which was flowing around the time leading up to potentially adverse low flow situations.

To make a comparison of the stream depletion assessment for this situation, indicative numerical simulations have been carried out using information from wells adjacent to the Selwyn River in the Central Canterbury Plains (PDP, 1999). The Selwyn River occasionally has flow along its full length, however, it also has long periods where it is dry across most of the Plains, with surface flow only emerging from groundwater discharge at the downstream end.

The model assumed a homogeneous, isotropic aquifer with a storage coefficient of 0.1 and a transmissivity which was varied between simulations from 500 – 2,000 m^2/day. The streambed was assumed to have good hydraulic connection to the aquifer.

Simulations covering all wells pumping from a variety of locations across the stream length show that up to 15% of the pumped water came from the stream when there was continuous flow. However, this reduced to 0.6% when the stream emerged from its lower reaches. Simulations looking at different groupings of wells were undertaken. It was found that only wells within about 2 km of the stream emergence had any effect on stream flow.

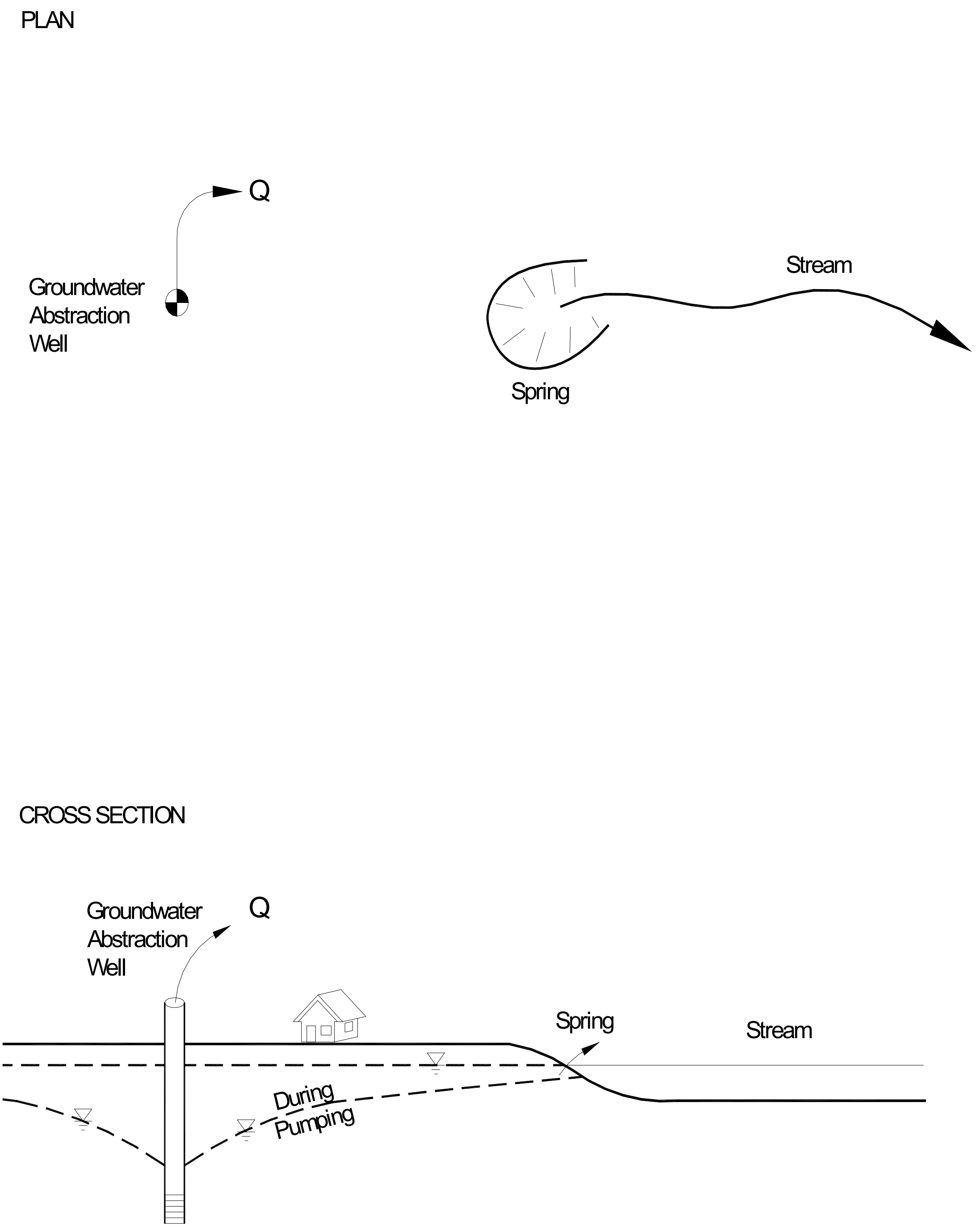


Figure 26: Schematic View of a Groundwater Abstraction Well Upstream of a Spring Fed Stream

In general terms, the distance at which upgradient pumping effects cease to be significant is difficult to define. As an indicative screening criteria it can be assumed that the stream depletion effect of an upgradient well will be less than half the effect calculated by the analytical equation in section 4.1. This is because the analytical equation assumes a continuous stream flow adjacent to the well.

If this screening assessment suggests that the stream depletion effect is significant and further characterisation is necessary then it would be appropriate to consider using a numerical model on a case-by-case basis.

5.3 Wells Located Near Artesian Springs

This situation is shown schematically in Figure 27. It occurs in settings such as the Avon River it its reaches through the University of Canterbury. Cameron (1993) describes observable flowing springs within the stream channel in the University grounds underlain by a 1 – 10 m thick fine grained confining layer which is breached by a permeable "pipe" structure. This "pipe" permits the upward transmission of water from the underlying artesian gravel aquifer.

An indicative estimate of stream depletion effects can be made by calculating the drawdown effect of the pumping well at the point where the spring emerges, by using the Theis equation or Hantush leaky aquifer equation, where appropriate (Domenico and Schwartz, 1990). The assumption is made that the discharge from the spring has established a stable piezometric pattern onto which the drawdown effect from a pumping well can be superimposed.

The discharge rate from the spring will vary in accordance with fluctuations in groundwater pressures. An example of this is seen in Cameron's monitoring of springs in western Christchurch (Figure 28). Consequently, the magnitude of the calculated drawdown effect can be compared with the expected range of water level fluctuations to provide an indication of the significance that the pumping effect might have.

For example, if a point spring discharge occurs in an area where groundwater levels annually fluctuate over a 2 m range and an abstracting well creates a calculated drawdown interference effect of 0.2 m then, as a first approximation, it can be estimated that the spring flow will reduce by around 10% of its range of normal seasonal flow variability as a result of the well abstraction.

5.4 A Well Bounded by Lower Permeability Terraces

This situation is shown schematically in Figure 29 and represents a common setting where rivers have down cut down through older strata and formed a permeable alluvial aquifer bounded below and at the sides by lower permeability deposits. An example of such a situation is the Kowai River in North Canterbury which is described in PDP (1996).

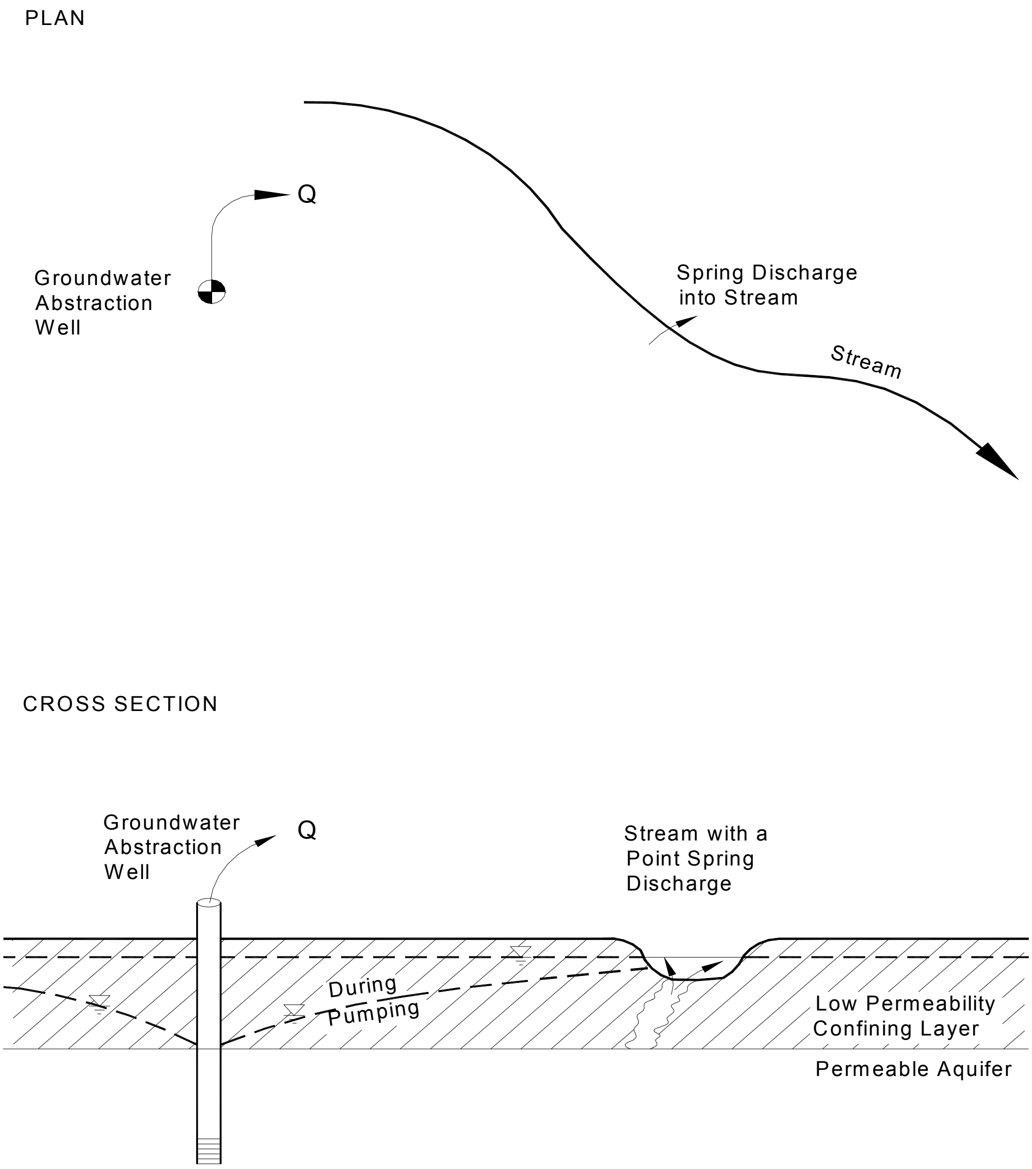


Figure 27: Schematic view of a Groundwater Abstraction Well Affecting an Artesian Spring

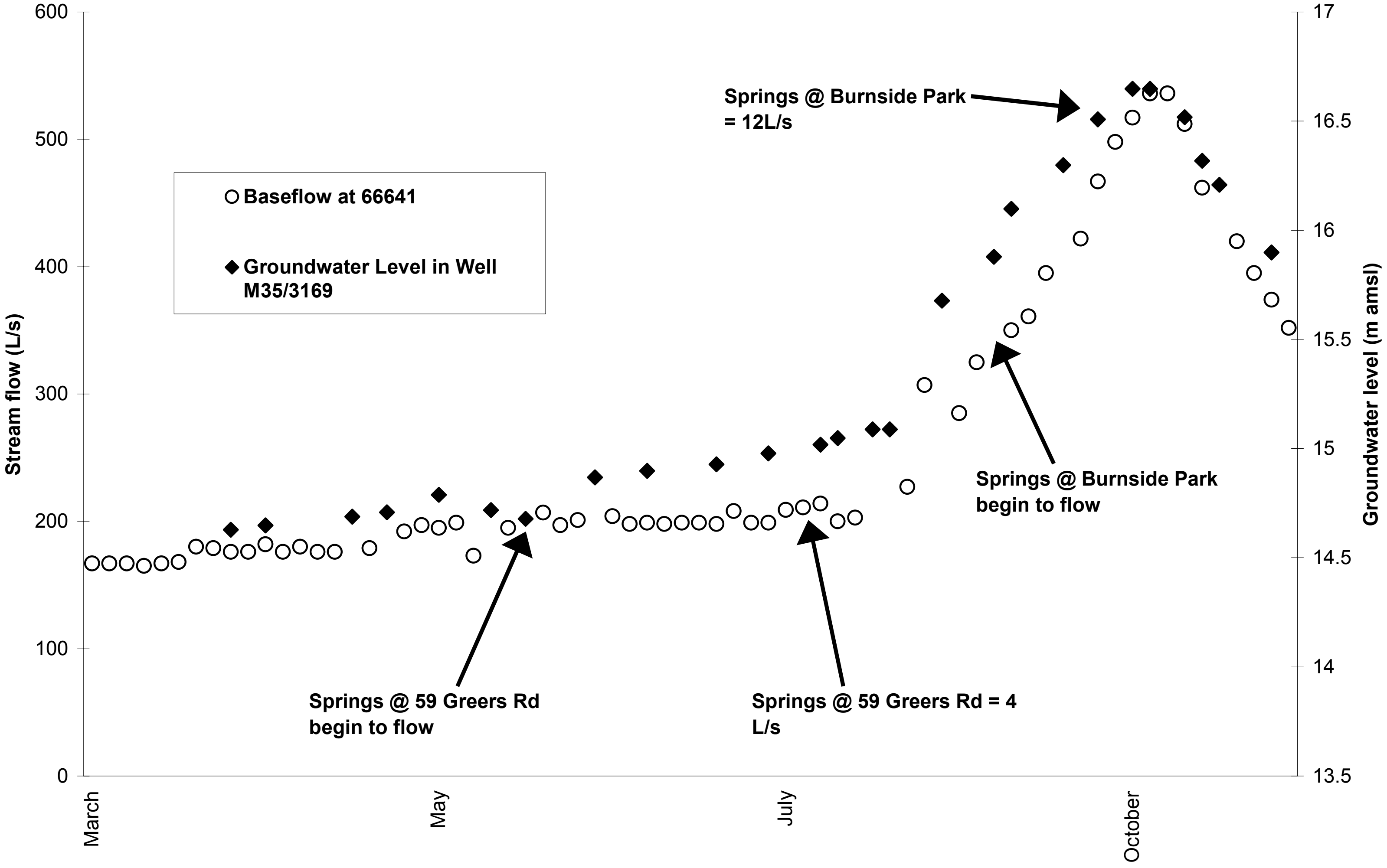


Figure 28: The discharge of two artesian spring sections in the Waimari Stream relative to baseflow rates and groundwater levels

Due to the bounded nature of the aquifer, it is possible to estimate the total groundwater resource that interacts with stream flows. If the contrast in hydraulic conductivity between the alluvial valley and surrounding terraces is great (e.g. 2 orders of magnitude or more) an estimate can be made by assuming that the surrounding strata is impermeable and the total water flow can be quantified by a measure of the stream flow and the groundwater flow, as estimated by using Darcy's equation:

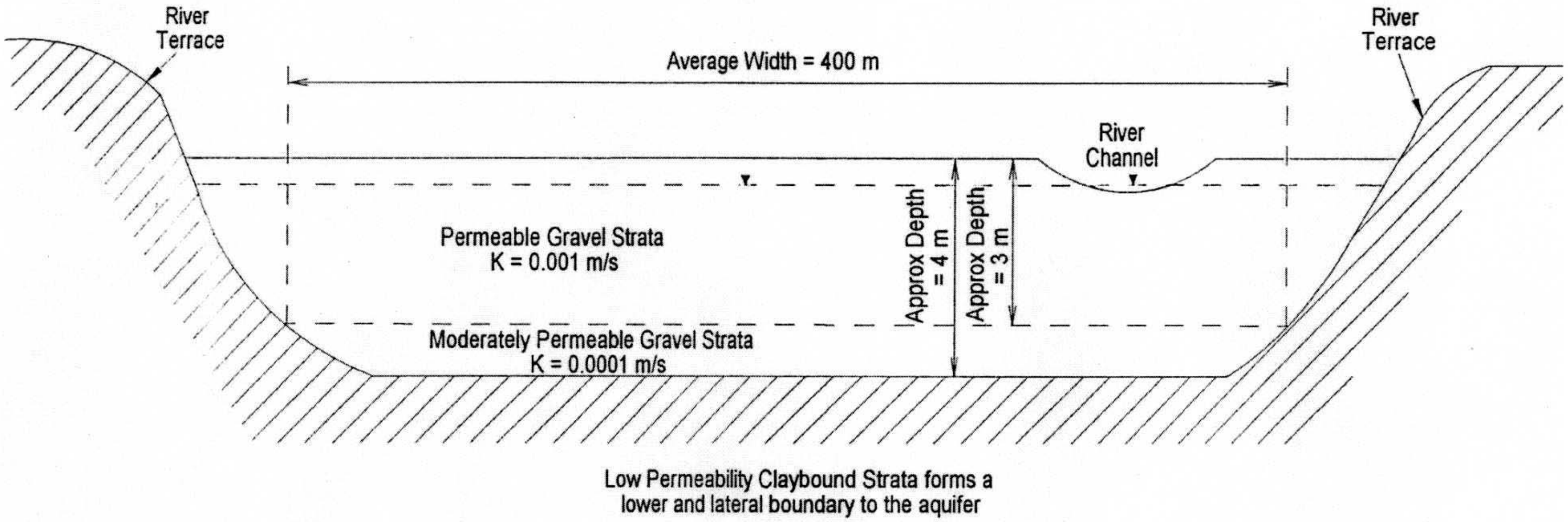
$$Q_{total} = Q_{stream} + K\ i\ A$$

where	Q_{total}	is the total flow through the water resource $[L^3/T]$;
	Q_{stream}	is the surface flow $[L^3/T]$;
	K	is the hydraulic conductivity of the strata adjacent to the stream [L/T];
	i	is the gradient of the water table [dimensionless];
	A	is the cross-sectional area of the shallow aquifer, perpendicular to the direction of groundwater flow $[L^2]$.

In addition, if the range of groundwater level fluctuations that coincide with variations in surface flow is known then the rate of water released from storage by a fall in water level over a given period of time can be estimated, due to the bounded nature of the aquifer.

$$Q_{storage} = \frac{Width \times L \times \Delta WL \times S}{t}$$

where	$Q_{storage}$	is the flow of water released from storage caused by a fall in water levels (Δh) over a period of time (t) $[L^3/T]$;
	Width	is the width of the aquifer (between the low permeability terraces) [L];
	L	is the length of the river valley over which the drawdown effects of a pumping well may extend (this could be estimated from the Theis drawdown equation) [L];
	t	is the time period over which the fall in water level occurs [T];
	ΔWL	is the fall in groundwater level that coincides with times of surface flow [L];
	S	is the storage coefficient for the aquifer [dimensionless].



Note: Assuming an average water table gradient (i) of –0.009 the following aquifer through flow is estimated:

Permeable gravels: $Q = -KiA = -0.001\ \text{m/s} \times -0.009 \times (3\text{-Depth to Water}) \times 400\ \text{m}$
$= 0.0036\ \text{m}^3/\text{s} \times (3\text{-Depth to Water}) = 311\ \text{m}^3/\text{day} \times (3\text{-Depth to Water})$

Moderately permeable gravels: $Q = -KiA = -0.0001\ \text{m/s} \times -0.009 \times 1\ \text{m} \times 400\ \text{m}$
$= 0.00036\ \text{m}^3/\text{s} = 31\ \text{m}^3/\text{day}$

Figure 29: Calculation of available water from the North Branch of the Kowai aquifer (from Pattle Delamore Partners Ltd, 1996)

Due to the characteristics of a well abstraction and its cone shaped drawdown pattern, it is not possible to abstract all the aquifer throughflow and storage. The combination of Q_{total} and $Q_{storage}$ represents a ballpark estimate of the total water resource in which abstractions will affect surface flow. If the hydraulic conductivity of the streambed strata is similar to the surrounding aquifer then groundwater abstractions cannot exceed the combination of Q_{total} and $Q_{storage}$ without contributing to a significant depletion of streamflow. The apportionment of pumped water between streamflow losses and aquifer loses can be estimated by use of a numerical model.