Technical Report
Environmental
Monitoring Group

Guidelines for the Assessment of Groundwater Abstraction Effects on Stream Flow

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Environment Canterbury
Your regional council

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1.0 Introduction

1.1 The Stream Depletion Issue

The management of water resources has often focussed on groundwater and surface water as if they were separate resources. However, nearly all surface water features (streams, lakes, springs, wetlands, estuaries and the sea-coast) interact with groundwater. Consequently, effective water management requires a good understanding of the way in which interaction between groundwater and surface water takes place.

The interactions between groundwater and surface water take many forms, which affect both the quantity and the quality of water in both resources. These interactions are often difficult to observe and measure, which creates uncertainty regarding their magnitude, their effect and an appropriate form of management. One of the potentially most significant forms of interaction is the effect of groundwater pumping on surface waterways. This is commonly referred to as the stream depletion effect. Understanding this particular form of interaction is becoming increasingly important as greater abstraction demands are placed on both surface and ground waters.

1.2 Purpose of the Guidelines

This technical guideline has been prepared to help the recognition of situations where significant stream depletion effects may occur and to provide tools that quantify the effects of groundwater abstraction on surface waterways. These tools apply to the assessment of single pumping wells, or small groups of wells by using the superposition of single well effects. The cumulative effects of large numbers of wells is best dealt with through the preparation of regional groundwater flow models, which is beyond the scope of this guideline document.

The quantification of the stream depletion effect caused by groundwater abstractions is a necessary component of the Assessment of Environmental Effects which is required by the Resource Management Act when applying for a resource consent to take groundwater.

The purpose of the guideline is to assist both resource consent applicants and the Regional Councils and Unitary Authorities who assess these applications. It is also expected that the quantification of these effects will be a necessary first step to allow the development of effective water management policies to deal with stream depletion effects.

Natural groundwater systems are inherently complex and it is well recognised that any attempts to quantify their behaviour requires gross simplification of the natural variability. The effect of groundwater pumping on surface waterways is no exception.

This guideline presents a pragmatic approach to assessing the effect, while also recognising the approximate nature of the assessment methods.

1.3 Structure of the Guidelines

The guideline has been structured in the following way:

Section 2: General Concepts of Stream-Aquifer Interaction

A stand-alone section which outlines the typical settings where groundwater and surface water interact.

Section 3: Initial Screening of Sites

Definition of some general hydrogeologic criteria which define whether or not groundwater pumping is likely to create a depleting effect on surface waterways.

Section 4: Assessment Methods

A description of tools that are available to quantify the effect of groundwater pumping on surface waterways.

Section 5: Non-uniform Hydrogeological Settings

A description of some common hydrogeological settings that do not match the analytical assessment methods.

Section 6: Field Measurements to Assist Assessments

A description of field measurements that can be made to define the parameters needed for the assessment methods presented in Section 4.

Section 7: Management Implications

This section discusses an approach for assessing the stream depletion effect in resource consent applications and provides some preliminary comments on how quantification of these stream depletion effects can contribute to the development of effective water management policies.

Appendices: Typical Examples

Actual field measurements of groundwater pumping effects are presented and assessed in Appendices A – D. Appendix E presents a summary brochure of the guidelines.

This technical guideline provides a detailed assessment of the stream depletion issue. It is accompanied by a simpler introductory note which summarises the key information in a more user friendly format.

1.4 Summary of Symbols

A	a cross-sectional area perpendicular to the direction of water flow [L^2]
b	the thickness of the aquifer [L]
D	the depth of water in the stream [L]
H	the depth to the water table below the stream water surface [L]
$h_{aquifer}$	the elevation of water in the aquifer [L]
h_{stream}	the elevation of the stream water surface [L]
Δh	the difference in elevation between the water surface in the stream and the groundwater in the aquifer [L]
i	the hydraulic gradient between the stream water and the groundwater [dimensionless]
I	the seepage per unit area of streambed [L/T]
K	the hydraulic conductivity of the aquifer underneath and/or adjacent to the streambed [L/T]
K'	the vertical hydraulic conductivity of the streambed [L/T]
L	the length of a stream reach over which seepage is assessed [L]
ℓ	the perpendicular separation distance between a well and a stream which is approximated by a straight line [L]
lambda = λ	the streambed conductance (a measure of the hydraulic conductivity and dimensions of the streambed) [L/T]
M	the thickness of the streambed which has a hydraulic conductivity of K' [L]
Q	the abstraction rate from a well [L^3/T]
q	the flow of water between the stream and the aquifer, i.e. the stream depletion flow rate [L^3/T]
S	the storage coefficient of the aquifer (a measure of how much water is released from the pore space of the aquifer as water pressures fall) [dimensionless]
T	the transmissivity of an aquifer (a measure of how permeable the aquifer is) [L^2/T]
t	the length of time over which abstraction from a well takes place [T]
W	the width of a stream reach over which seepage is assessed [L]

1.5 Summary of Unit Symbols

[unit symbols are shown in square brackets]

L length (e.g. metres)

T time (e.g. days)

e.g. terms with units of [L^3/T] could be defined as m^3/day or litres/second (L/s).

NB consistent units must be used for all calculations.

2.0 General Concepts of Surface Water-Groundwater Interaction

2.1 Groundwater and Streams

The interaction between streams and groundwater takes place in two basic ways:

(a) Streams gain water from groundwater through the streambed when the elevation of the water table adjacent to the streambed is greater than the water level in the stream (shown schematically in Figure 1).

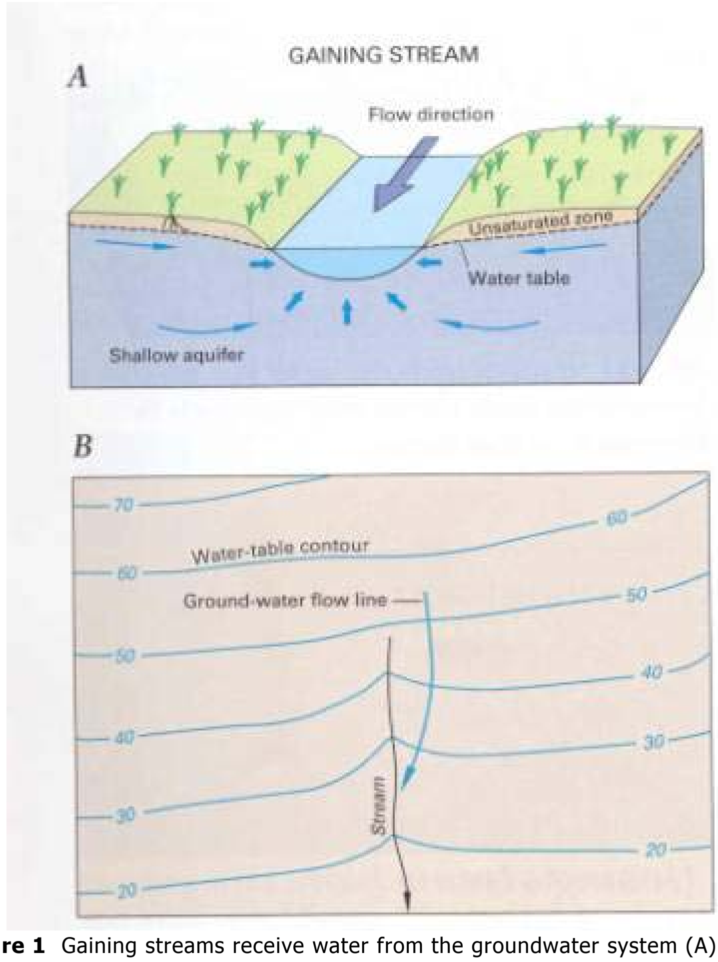


Figure 1 Gaining streams receive water from the groundwater system (A). This can be determined from water table contour maps because the contour lines point in the upstream direction where they cross the stream (B). (from USGS 1998)

(b) Streams lose water to groundwater by outflow through the streambed when the elevation of the water table is lower than the water level in the stream (shown schematically in Figure 2).

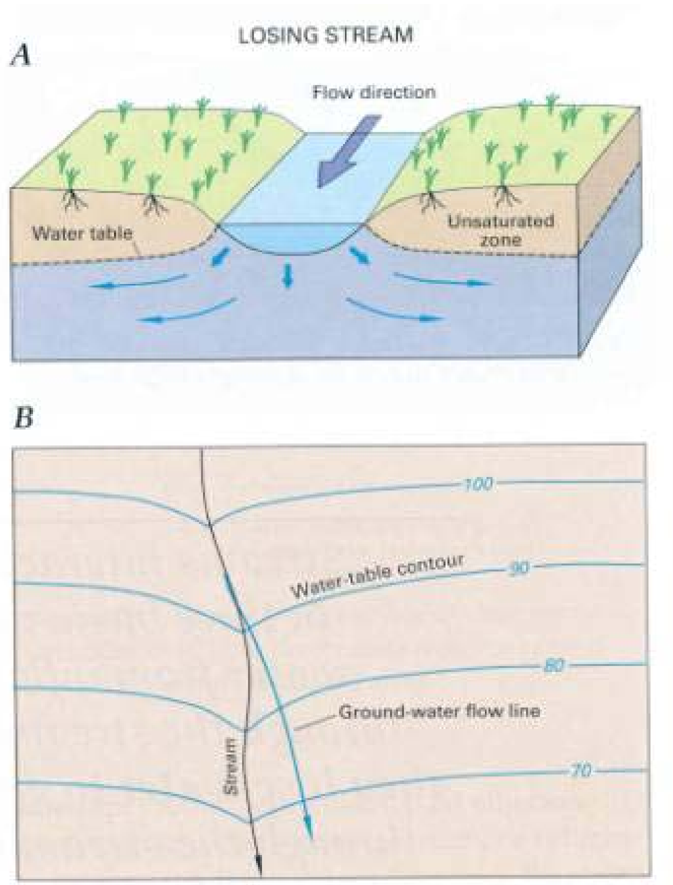


Figure 2 Losing streams lose water to the groundwater system (A). This can be determined from water table contour maps because the contour lines point in the downstream direction where they cross the stream (B). (from USGS 1998)

A third possibility is that streams may have no flow across their streambed if the stream water levels and groundwater levels are exactly equal. However, this is a relatively rare coincidence that is unlikely to occur over long reaches for prolonged periods.

Within any particular stream, it is not uncommon to have different reaches gaining or losing water, or the same reach losing or gaining water at different times of the year.

This variability in stream flow caused by groundwater seepage is often most pronounced around the headwaters of groundwater-fed steams or in stream reaches that periodically go dry because of seepage losses.

At times of low groundwater levels, surface flow may only occur due to surface flows in the upper catchment (e.g. from snow melt) or from stormwater runoff. This flow will contribute seepage through the streambed to the underlying water table. As the water table rises due to general recharge (e.g. rainfall infiltration) to the aquifer at large, the losing reach may become a gaining reach as the water table rises above the level of the streambed. Such a change means that the point where groundwater first contributes water to the stream shifts laterally in an upstream direction, as shown schematically in Figure 3.

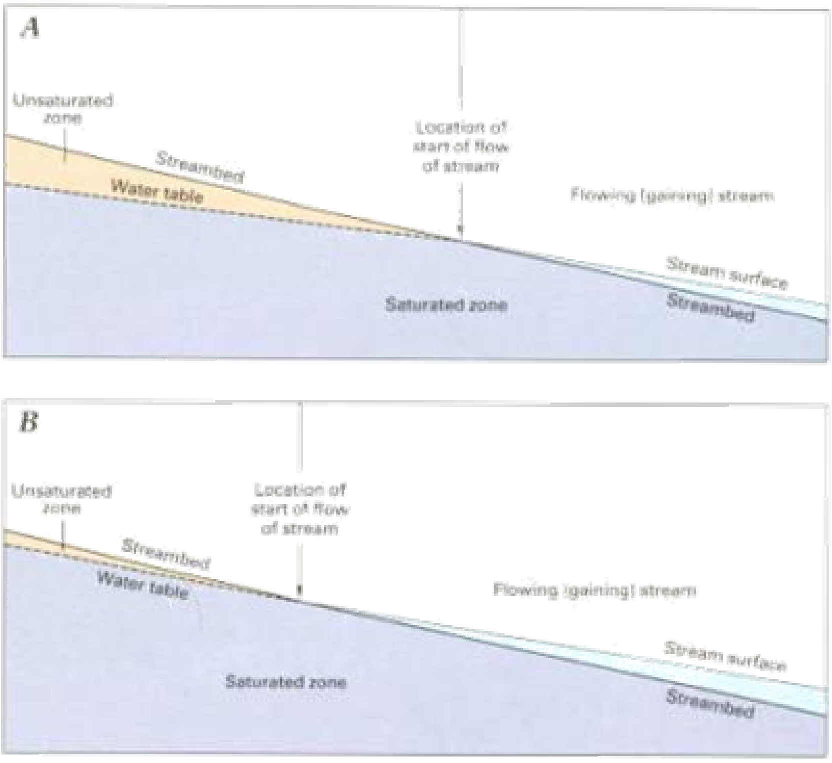


Figure 3 The location where stream flow begins can vary due to changes in the water table. Emergence of stream flow can change from a downstream location under low water table conditions (A) to an upstream location at high water table conditions (B). (from USGS 1998)

The hydraulic connection between the stream and the groundwater may be direct, as shown in Figures 1 and 2 or it may be disconnected by an intervening unsaturated zone, with streams losing water by seepage through a streambed down to a deep water table, as shown schematically in Figure 4. As with the gaining and losing reaches of a stream, the degree of connection to a stream can also change over different reaches within any one stream and from time to time over the same reach.

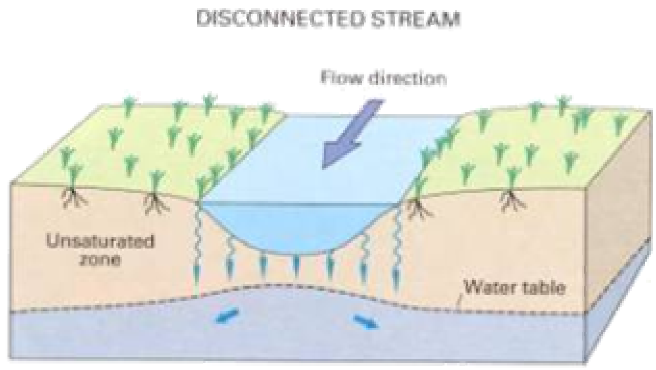


Figure 4 Disconnected streams are separated from the groundwater system by an unsaturated zone. (from USGS, 1998)

The rate of movement of water between a stream and the groundwater is most easily assessed by assuming that water moves vertically through the streambed. The flow of water can be defined by Darcy's equation as:

$$q = {}^{-}K'LWi$$

where q is the rate of flow between the stream and the aquifer [L^3/T]

K' is the vertical hydraulic conductivity of the steam bed [L/T]

L is the length of the stream reach over which seepage is assessed [L]

W is the average width of the stream over the reach which is assessed [L]

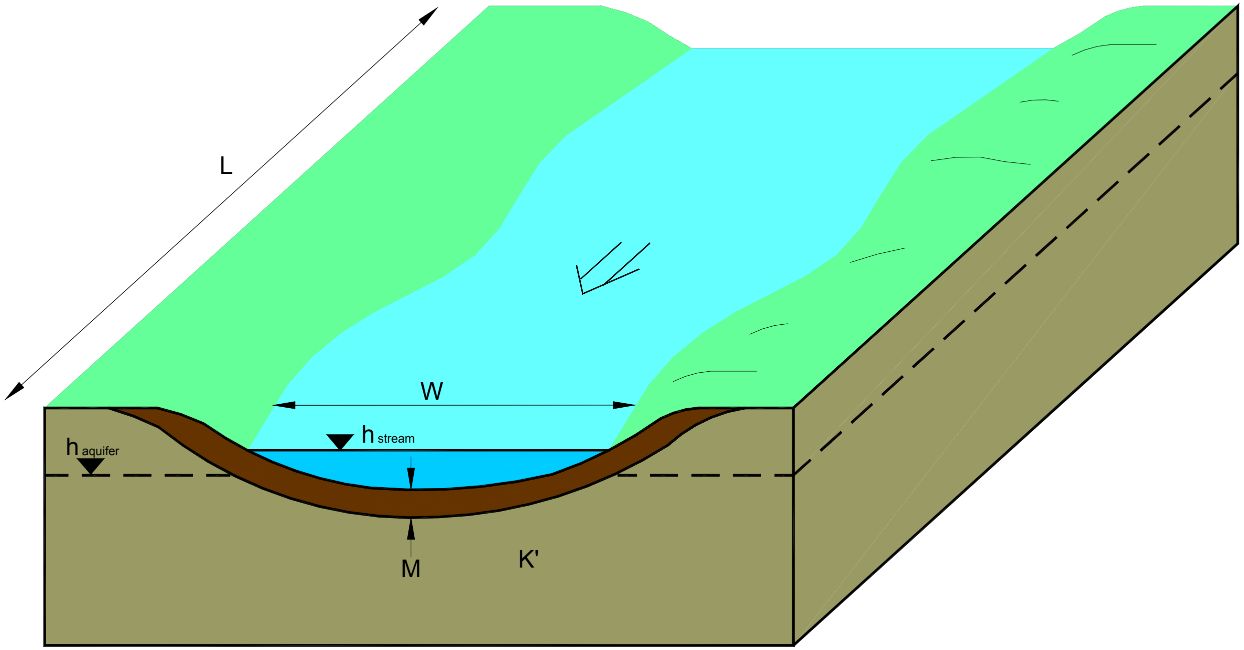
i is the hydraulic gradient between the stream and the groundwater which can be defined as:

$$i = \frac{h_{aquifer} - h_{stream}}{M}$$

where $h_{aquifer}$ is the elevation of the groundwater beneath the stream [L]
h_{stream} is the elevation of the stream water surface [L]
M is the thickness of the streambed [L]

Note: When the groundwater is disconnected from the surface water, as shown schematically in Figure 4, it is normally assumed that the stream loses water under a hydraulic gradient (i) of 1.

The parameters used in the Darcy equation calculation are shown schematically in Figure 5. In practice, the flow through a streambed will vary depending on small scale variability in the hydraulic conductivity of the streambed sediments and the groundwater flow pattern around the streambed. As with most groundwater analyses, it is not practical to quantify such small scale variability, and the seepage equation listed above forms a satisfactory basis to approximate the bulk movement of water through the streambed.



$$q = -K'LW\,i$$

$$i = \frac{h_{aquifer} - h_{stream}}{M}$$

Figure 5: Calculating the flow of water between a stream and an aquifer

2.2 Groundwater and Lakes

The type of interactions between groundwater and lakes are generally similar to interactions with streams. The main difference is that lakes have a much larger surface water and bed area. Furthermore, the slower throughflow rates in a lake often result in accumulations of low permeability sediments in the lake floor which can affect the distribution of seepage. As a result, the rate of seepage is often greatest around the lake margin where wave action may restrict the deposition of finer sediments.

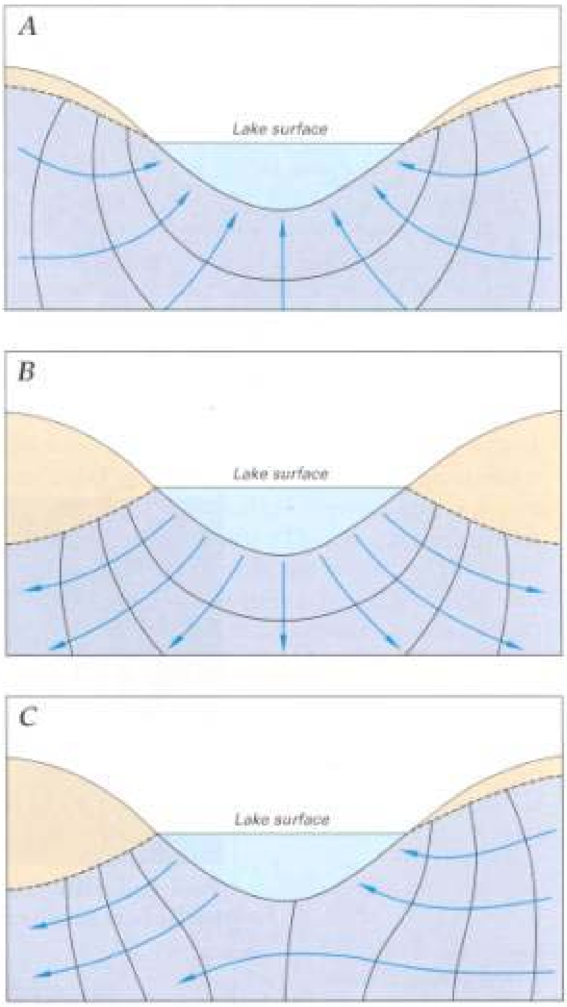


Figure 6 Lakes can receive groundwater inflow (A), lose water as seepage to groundwater (B) or both (C). (from USGS, 1998)

2.3 Interaction of Groundwater and Wetlands

Wetlands typically occur in areas where groundwater discharges to the land surface or in areas where ground conditions impede the drainage of water, as shown in Figure 7. For situations where impeded drainage occurs, stream depletion effects are unlikely to be significant because the layer of impeded drainage is also likely to inhibit the upward transmission of any pumping effects. However, in areas where groundwater springs discharge into wetlands, the pumping from underlying aquifers can affect the amount of groundwater discharge, as shown in the figures on the following page.

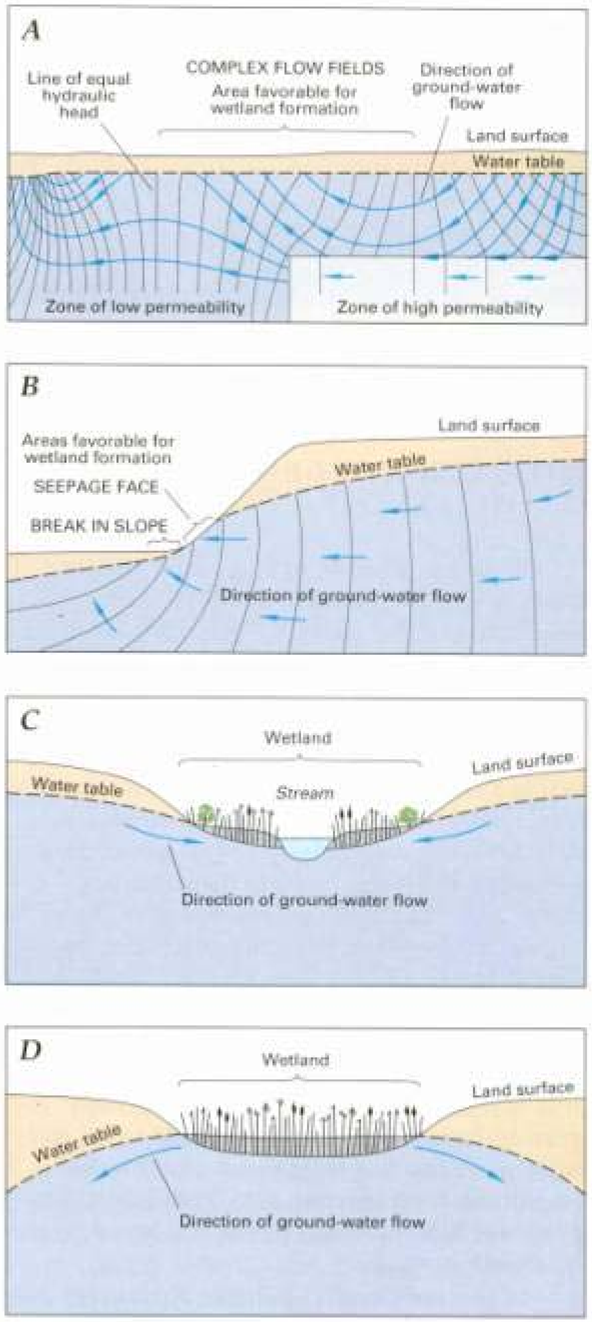


Figure 7 The source of water to wetlands can be from groundwater discharge where the land surface is underlain by complex groundwater flow fields (A), from groundwater discharge at seepage faces and at breaks in slope of the water table (B), from streams (C), and from precipitation in cases where wetlands have no stream inflow and groundwater gradients slope away from the wetland (D). (from USGS, 1998)

The stream depletion assessment methods that are described in Sections 3 – 6 of these guidelines focus on streams and wells. However, the same approach can be equally well applied to estimate effects on lakes, wetlands, or any surface water feature that can be approximated as a linear external source or sink to a groundwater flow system.

2.4 Interaction of Groundwater and Surface Water in Different Landscapes

The interaction of groundwater with surface water depends on the physiographic and climatic setting of the landscape. The United States Geological Survey (USGS, 1998) have described a conceptual landscape to define the most common types of groundwater-surface water interaction. The conceptual landscape is shown in Figure 8 and defines the following types of terrain: mountainous (M), river valleys (R, small and V, large), coastal (C), glacial and dune (G), and karst (K). Descriptions of typical interactions are presented in USGS (1998) and the remainder of the text in this section is drawn from that report.

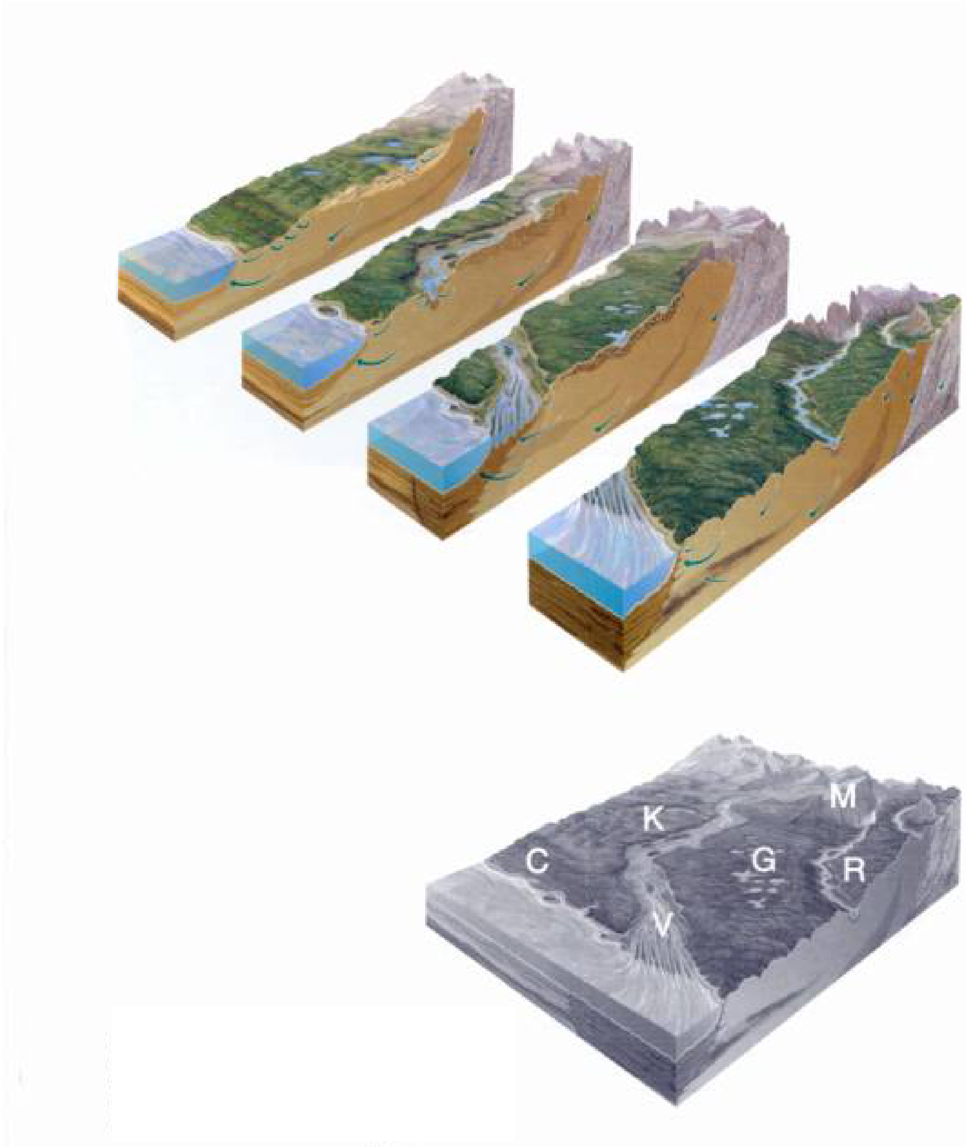


Figure 8 Groundwater and surface water interact throughout all landscapes from the mountains to the oceans, as depicted in this diagram of a conceptual landscape. M, mountainous; K, karst; G, glacial; R, small river valley; V, large river valley; C, coastal. (from USGS 1998)

2.4.1 Mountainous Terrain

The hydrology of mountainous terrain (area M of the conceptual landscape, Figure 8) is characterised by highly variable precipitation and water movement over and through steep land slopes. On mountain slopes, macropores created by burrowing organisms and by decay of plant roots have the capacity to transmit subsurface flow downslope quickly. In addition, some rock types underlying soils may be highly weathered or fractured and may transmit significant additional amounts of flow through the subsurface. In some settings this rapid flow of water results in hillside springs.

A general concept of water flow in mountainous terrain includes several pathways by which precipitation moves through the hillside to a stream (Figure 9).

Between storm and snowmelt periods, stream flow is sustained by discharge from the groundwater system (Figure 9A). During intense storms, most water reaches streams very rapidly by partially saturating and flowing through the highly conductive soils. On the lower parts of hillslopes, the water table sometimes rises to the land surface during storms, resulting in overland flow (Figure 9B). When this occurs, precipitation on the saturated area adds to the quantity of overland flow. When storms or snowmelt persist in mountainous areas, near-stream saturated areas can expand outward from streams to include areas higher on the hillslope. In some settings overland flow can be generated when the rate of rainfall exceeds the infiltration capacity of the soil (Figure 9C).

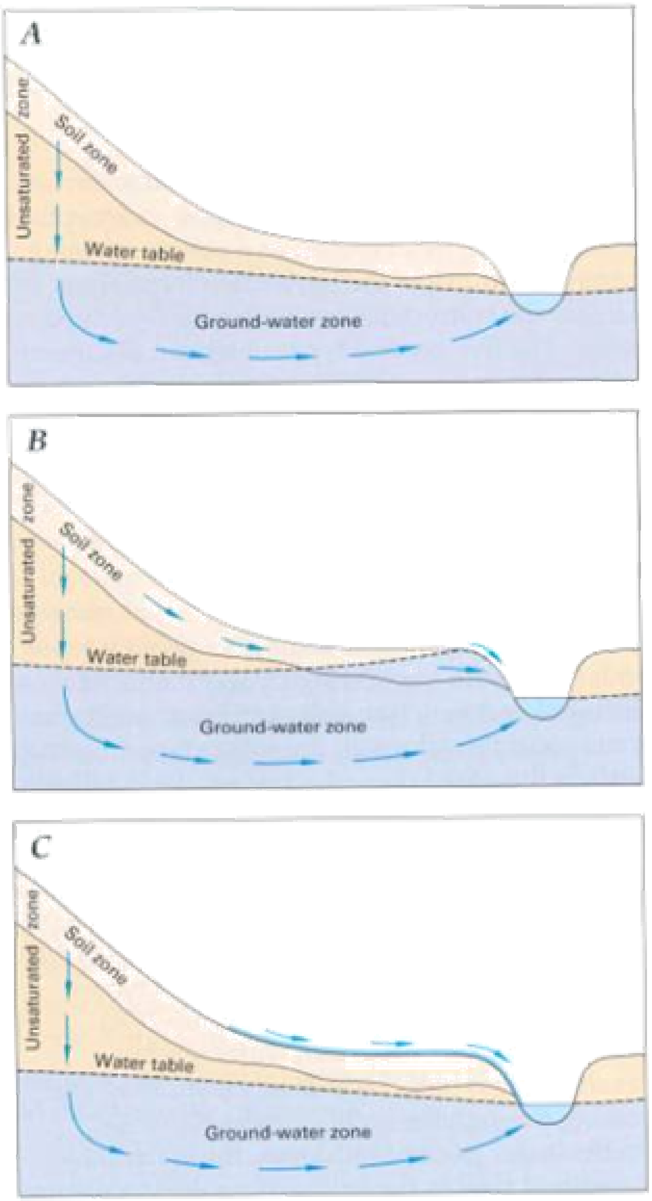


Figure 9 Water from precipitation moves to mountain streams along several pathways. Between storms and snowmelt periods, most inflow to streams commonly is from groundwater (A). During storms and snowmelt periods, much of the water inflow to streams is from shallow flow in saturated macropores in the soil zone. If infiltration to the water table is large enough, the water table will rise to the land surface and flow to the stream is from groundwater, soil water, and overland runoff (B). In arid areas where soils are very dry and plants are sparse, infiltration is impeded and runoff from precipitation can occur as overland flow (C). (Modified from Dunne, T, and Leopold, L B, 1978, Water in environmental planning: San Francisco, W H Freeman.) (Used with permission.) (from USGS 1998)

Near the base of some mountainsides, the water table intersects the steep valley wall some distance up from the base of the slope (Figure 10, left side of the valley). This results in perennial discharge of groundwater and, in many cases, the presence of wetlands. A more common hydrologic process that results in the presence of wetlands in some mountain valleys is the upward discharge of groundwater caused by the change in slope of the water table from being steep on the valley side to being relatively flat in the alluvial valley (Figure 10, right side of the valley). Where both of these water table conditions exist, wetlands fed by groundwater can be present.

Basement geology strata and structure are a key component which influences these characteristics and groundwater flow patterns in mountainous terrain.

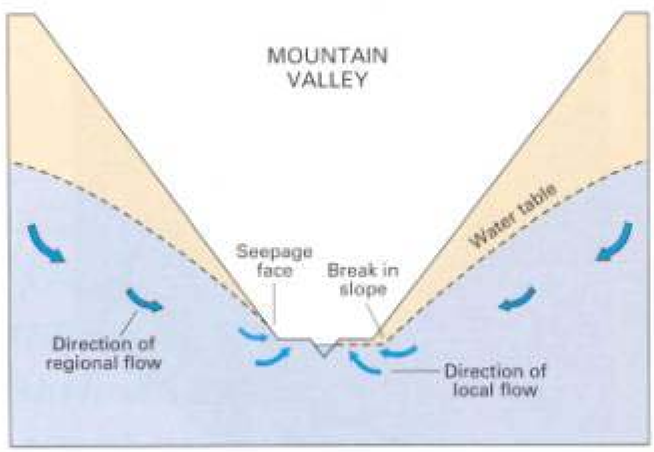


Figure 10 In mountainous terrain, groundwater can discharge at the base of steep sloped (left side of valley) at the edges of flood plains (right side of valley), and to the stream. (from USGS 1998)

Another dynamic aspect of the interaction of groundwater and surface water in mountain settings is caused by the marked longitudinal component of flow in mountain valleys. The high gradient of mountain streams, coupled with the coarse texture of streambed sediments, results in a strong down-valley component of flow accompanied by frequent exchange of stream water with groundwater (Figure11). The hyporheic zone shown in Figure 11 refers to the zone where water characteristics are intermediate between those of surface water and groundwater.

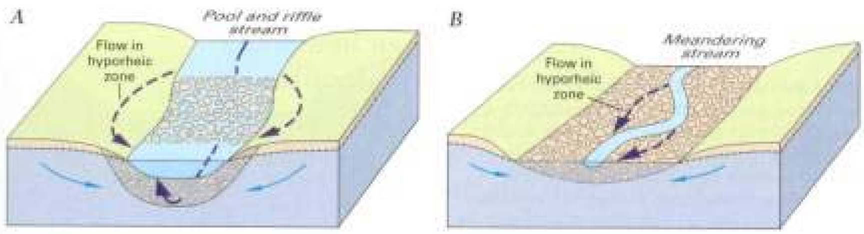


Figure 11 Surface water exchange with groundwater is associated with abrupt changes in streambed slope (A) and with streambed meanders (B). (from USGS, 1998)

Streams flowing from mountainous terrain commonly flow across alluvial fans at the edges of the valleys. Most streams in this type of setting lose water to groundwater as they traverse the highly permeable alluvial fans and seepage of water from the stream can be the principal source of aquifer recharge.

The most common natural lakes in mountainous terrain are those that are dammed by rock sills or glacial deposits high in the mountains. Whilst they receive much of their water from snowmelt, they also interact with groundwater much like the processes shown in Figure 10, and can be maintained by groundwater throughout the snow-free season.

2.4.2 River Valley Terrain

In some landscapes, stream valleys are small and they commonly do not have well-developed flood plains (area R of the conceptual landscape, Figure 8). However, major rivers (area V of Figure 8) have valleys that usually become increasingly wider downstream. Terraces and abandoned river meanders are common landscape features in major river valleys, and wetlands and lakes commonly are associated with these features.

The interaction of groundwater and surface water in river valleys is affected by the interchange of local and regional groundwater flow systems with the rivers and by flooding and evapotranspiration. Small streams receive groundwater flow primarily from local flow systems, which usually have limited extent and are highly variable seasonally. Therefore, it is not unusual for small streams to have gaining or losing reaches that change seasonally.

For larger rivers that flow in alluvial valleys, the interaction of groundwater and surface water usually is more spatially diverse than it is for smaller streams. Groundwater from regional flow systems discharge to the river as well as at various places across the flood plain (Figure 12). If terraces are present in the alluvial valley, local groundwater flow systems may be associated with each terrace, and lakes and wetlands may be formed because of this source of groundwater. Similarly, the occurrence of fault lines may create areas where significant exchanges between surface water and groundwater occur. At some locations, such as at the valley wall and at the river, local and regional groundwater flow systems may discharge in close proximity. Furthermore, in large alluvial valleys, significant down-valley components of flow in the streambed and in the shallow alluvium also may be present.

River valley alluvial deposits range in size from clay to boulders, but in many alluvial valleys, sand and gravel are the predominant deposits.

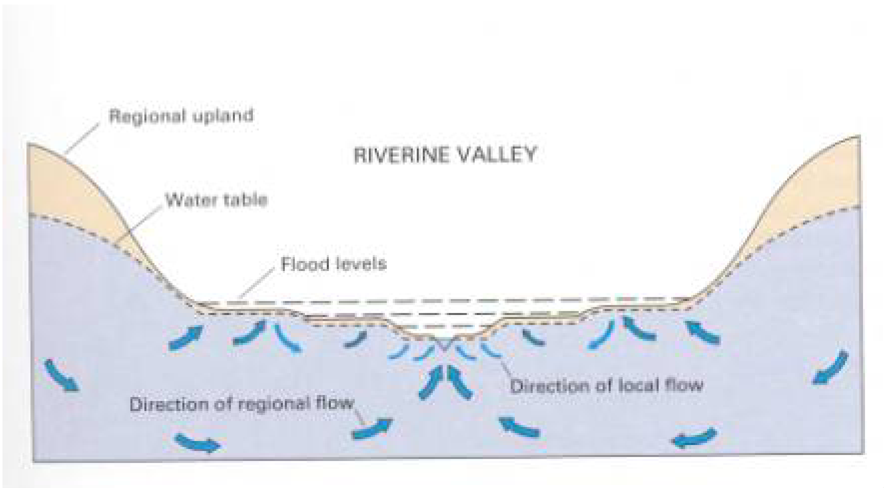


Figure 12 In broad river valleys, small local groundwater flow systems associated with terraces overlie more regional groundwater flow systems. Recharge from flood waters superimposed on these groundwater flow systems further complicates the hydrology of river valleys. (from USGS, 1998)

2.4.3 Coastal Terrain

Coastal terrain extends from inland scarps and terraces to the ocean (area C of the conceptual landscape, Figure 8). This terrain is characterised by (1) low scarps and terraces that were formed when the ocean was higher than at present; (2) streams, estuaries, and lagoons that are affected by tides; (3) ponds that are commonly associated with coastal sand dunes; and (4) barrier beaches and bars. Wetlands cover extensive areas in some coastal terrains.

The interaction of groundwater and surface water in coastal terrain is affected by discharge of groundwater from regional flow systems and from local flow systems associated with scarps and terraces (Figure 13), evapotranspiration, and tidal flooding. The local flow systems associated with scarps and terraces are caused by the configuration of the water table near these features. Where the water table has a downward break in slope near the top of scarps and terraces, downward components of groundwater flow are present; where the water table has an upward break in slope near the base of these features, upward components of groundwater flow are present.

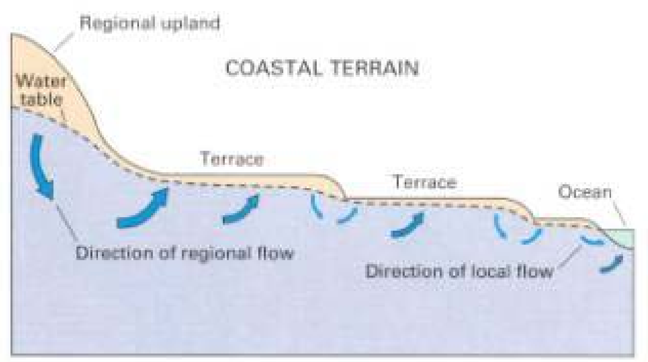


Figure 13 In coastal terrain, small local groundwater flow cells associated with terraces overlie more regional groundwater flow systems. (from USGS, 1998)

2.4.4 Glacial and Dune Terrain

Glacial and dune terrain (area G of the conceptual landscape, Figure 8) is characterised by a landscape of hills and depressions. Although stream networks drain parts of these landscapes, many areas of glacial and dune terrain do not contribute runoff to an integrated surface drainage network. Instead, surface runoff from precipitation falling on the landscape accumulates in the depressions, commonly resulting in the presence of lakes and wetlands. Because of the lack of stream outlets, the water balance of these "closed" types of lakes and wetlands is controlled largely by exchange of water with the atmosphere (precipitation and evapotranspiration) and with groundwater.

Lakes and wetlands in glacial and dune terrain can have inflow from groundwater, outflow to groundwater, or both (as shown in Figure 6). The interaction between lakes and wetlands and groundwater is determined to a large extent by their position with respect to local and regional groundwater flow systems. A common conception is that lakes and wetlands that are present in topographically high areas recharge groundwater, and that lakes and wetlands that are present in low areas receive discharge from groundwater. However, lakes and wetlands underlain by deposits having low permeability can receive discharge from local groundwater flows system even if they are located in a regional groundwater recharge area. Conversely, they can lose water to local groundwater flow systems even if they are located in a regional groundwater discharge area (Figure 14).

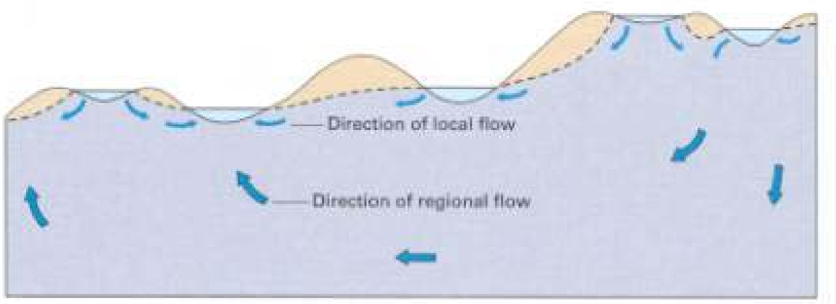


Figure 14 In glacial and dune terrain, local, intermediate, and regional groundwater flow systems interact with lakes and wetlands. It is not uncommon for wetlands that recharge local groundwater flow systems to be present in lowlands and for wetlands that receive discharge from local groundwater to be present in uplands.

Lakes and wetlands in glacial and dune terrain underlain by highly permeable deposits commonly have groundwater seepage into one side and seepage to groundwater on the other side. This relationship can be relatively stable because the water table gradient between surface water bodies in this type of setting is relatively constant. However, the boundary between inflow to lake or wetlands and outflow from it, termed the hinge line, can move up and down along the shoreline. Movement of the hinge line between inflow and outflow is a result of the changing slope of the water table in response to changes in groundwater recharge in the adjacent uplands.

Transpiration directly from groundwater can have a significant effect on the interaction of lakes and wetlands with groundwater in glacial and dune terrain. Transpiration from groundwater (Figure 15) has perhaps a greater effect on lakes and wetlands underlain by low permeability deposits than in any other landscape. The lateral movement of groundwater in low permeability deposits may not be fast enough to supply the quantity of water at the rate it is removed by transpiration, resulting in deep and steep-sided cones of depression. These cones of depression commonly are present around the perimeter of the lakes and wetlands (Figure 15).

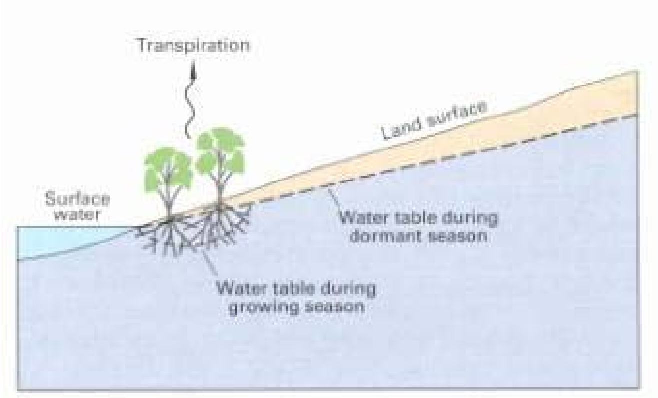


Figure 15 Where the depth to the water table is small adjacent to surface water bodies, transpiration directly from groundwater can cause cones of depression similar to those caused by pumping wells. This sometimes drains water directly from the surface water into the subsurface.

2.4.5 Karst Terrain

Karst may be broadly defined as all landforms that are produced primarily by the dissolution of rocks, mainly limestone and dolomite. Karst terrains (area K of the conceptual landscape, Figure 8) are characterised by (1) closed surface depressions of various sizes and shapes known as sinkholes; (2) an underground drainage network that consists of solution openings that range in size from enlarged cracks in the rock to large caves; and (3) highly disrupted surface drainage systems, which relate directly to the unique character of the underground drainage system.

Groundwater recharge is very efficient in karst terrain because precipitation readily infiltrates through the rock openings that intersect the land surface. Water can move at greatly different rates through karst aquifers; it moves slowly through fine fractures and pores and rapidly through solution-enlarged fractures and conduits. As a result, the water discharging from many springs in karst terrain may be a combination of relatively slow-moving water draining from pores and rapidly moving storm-derived water. The slow-moving component tends to reflect the chemistry of the aquifer materials, and the more rapidly moving water associated with recent rainfall tends to reflect the chemical characteristics of precipitation and surface runoff.

Water movement in karst terrain is difficult to predict because of the many paths groundwater takes through the maze of fractures and solution openings in the rock. Because of the large size of interconnected openings in well-developed karst systems, karst terrain can have true underground streams. These underground streams can have high rates of flow, in some places as great as rates of flow in surface streams. Furthermore, it is not unusual for medium-sized streams to disappear into the rock openings, thereby completely disrupting the surface drainage system, and to reappear at the surface at another place, which might even be in a different surface water catchment (Figure 16). Seeps and springs of all sizes are characteristic features of karst terrains. Springs having sufficiently large groundwater recharge areas commonly are the source of small- to medium-sized streams and constitute a large part of tributary flow to larger streams. In addition, the location where the steams emerge can change, depending on the spatial distribution of groundwater recharge in relation to individual precipitation events. Large spring inflows to streams in karst terrain contrast sharply with the generally more diffuse groundwater inflow characteristic of streams flowing across sand and gravel aquifers.

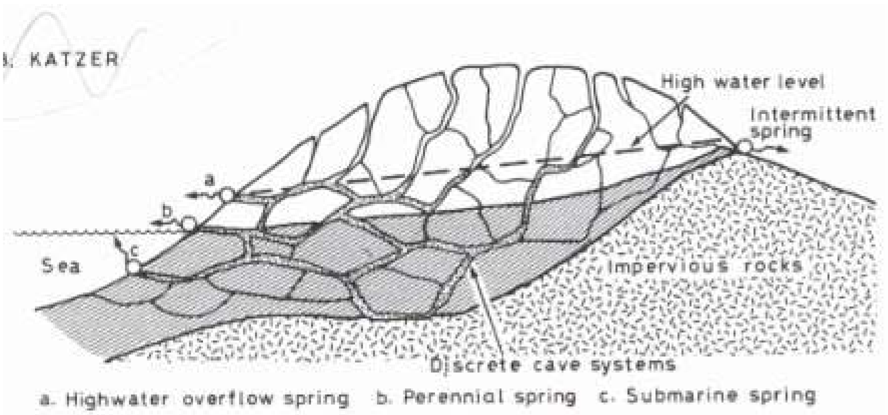


Figure 16 The karst groundwater system often operates as a network of independent subterranean rivers. (from Ford and Williams, 1989)

3.0 Initial Screening of Sites

3.1 When is Stream Depletion Likely to Occur?

Stream depletion effects that are induced by groundwater abstraction can occur in settings where a surface waterway occurs above and/or adjacent to a productive aquifer that is used for groundwater abstraction. They will occur provided that all three of the following criteria occur:

(i) Water can flow between the surface water and adjoining groundwater resource.

(ii) The rate of water movement between these two water bodies is dependent on the groundwater gradient adjacent to the stream.

(iii) The groundwater gradient adjacent to the stream is affected by groundwater abstractions.

The typical setting in which these circumstances occur is shown schematically in Figures 1 and 2. Example locations include:

» *Waipawa River, Hawkes Bay*

» *Papawai Stream, Wairarapa*

» *Little Sydney Stream, Tasman*

» *Omaka River, Marlborough*

» *Doyleston Drain, Canterbury*

For any potential setting where stream depletion effects may occur, an initial screening of water level data, and hydraulic conductivity data should be undertaken to determine whether or not stream depletion is likely to be a significant issue. The initial screening approach is described below. Whilst each of the parameters can be considered in isolation, it is important that the final conclusion must be consistent with the conceptual hydrogeological understanding for the general area.

3.1.1 Water Level Data

The depletion of stream flow by groundwater abstraction can occur if a stream is receiving groundwater flow (i.e. nearby groundwater levels are higher than stream water levels, Figure 1), if it is in equilibrium with groundwater (i.e. stream levels and groundwater levels are equal), or in some cases if a stream is losing flow to groundwater (i.e. groundwater levels are lower than stream levels, Figure 2).

In general, if long-term monitoring data indicates a correlation between groundwater levels and stream flow then it is likely that stream depletion effects have the potential to occur.

However, if groundwater levels are significantly deeper than the stream levels then the natural seepage loss from the stream will be caused by a hydraulic gradient of 1, acting vertically downwards, and will not be affected by any further lowering of the groundwater table that may be induced by groundwater abstractions (Figure 4).

Two guideline values are provided to assess the depth to water at which stream depletion effects will not occur. It is to be expected that both these water depth criteria shall be met before a conclusion can be reached that there is an absence of a stream depletion effect.

(i) Hunt (1997) describes a flow net analysis which shows that when a stream is perched above the water table a zone of uniform vertically downwards flow occurs. This vertical flow condition is expected to be reached when the depth to the water table below the stream surface (H) is five times the maximum depth of water in the stream (D), i.e. $H \geq 5D$, as shown in Figure 17. Under these circumstances, if H is increased due to drawdown from a pumping well it will not induce extra seepage from a stream.

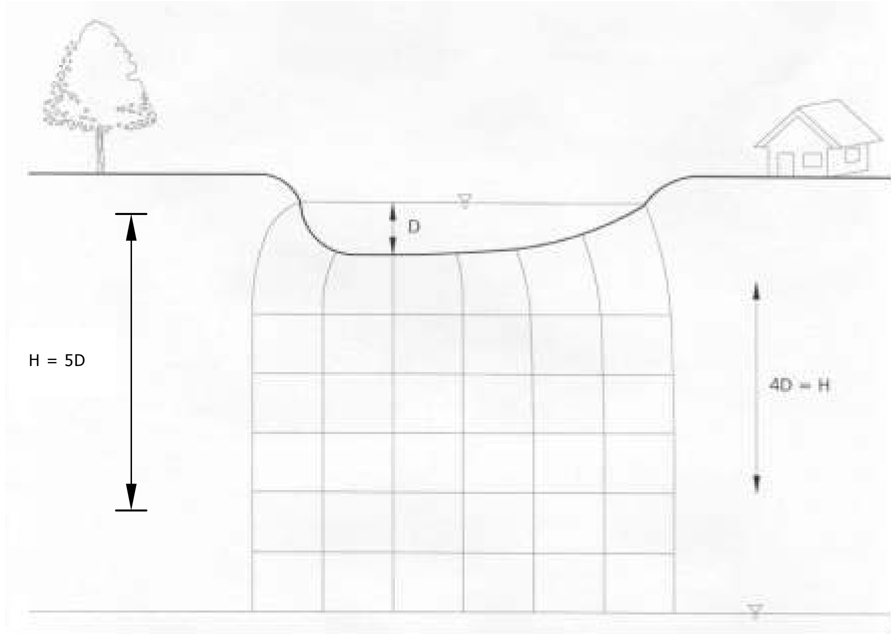
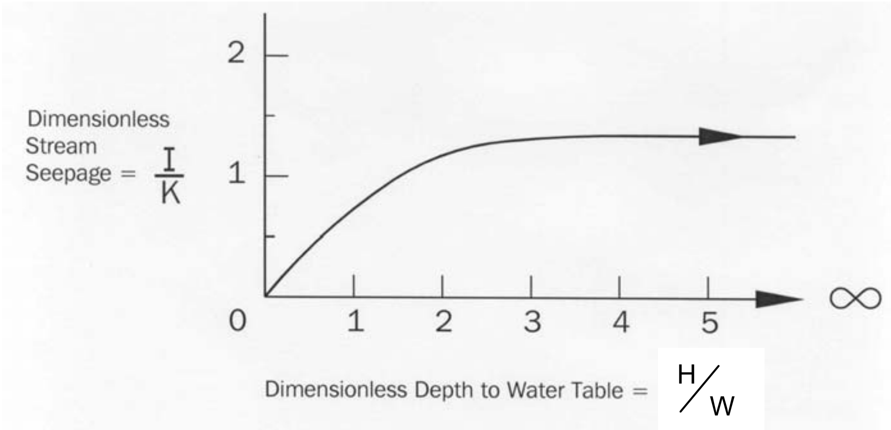


Figure 17 Flow net for seepage beneath a stream, showing that uniform vertical flow is reached by a depth of 5D below the stream surface (i.e. 4D below the streambed).

(ii) Bouwer (1997) describes stream seepage rates in relation to the dimensionless term H/W where H is the depth to groundwater and W is the width of the stream (Figure 18). If the depth to groundwater is more than twice the stream width (i.e. H ≥ 2W) then any further lowering of the groundwater table will not significantly increase stream seepage.



I = Seepage per unit area as stream seepage
K = Hydraulic conductivity of underlying strata
H = Vertical distance between the groundwater table and the water surface of the stream
W = Width of stream

Figure 18 Dimensionless plot of seepage (expressed as I/K) versus depth to groundwater (expressed as H/W) for clean stream channel (no clogging layer on bottom) - from Bouwer, 1997.

Bouwer's diagrams and notes regarding this work suggest that the measurement of the depth to the water table (H) can be made "at some distance" from the stream and therefore can be measured in wells which generally represent the water table elevation in the aquifer that is under consideration.

Groundwater level measurements can also be used to indicate layers of low hydraulic conductivity strata between the screened section of a well and the groundwater resource adjacent to a stream. A well which shows a water pressure that is significantly different than the water table, at the same location, provides an indication that a low permeability layer exists above the screened section of the well. Such a layer will reduce any drawdown effect that the pumping well creates on the shallow water table. These measurements would typically be supported by drillers well logs indicating the presence of low permeability fine grained layers, as discussed below.

3.1.2 Hydraulic Conductivity of Strata

The hydraulic conductivity of the strata between the stream and the screen location of an abstraction well has a significant bearing on the potential stream depletion effects. The hydraulic conductivity is primarily determined by the permeability of the strata – with coarse grained gravels having high values and fine grained silts and clays having very low values. In rock formations the hydraulic conductivity is largely determined by the number, size and continuity of fractures.

The hydraulic conductivity between the well and the stream may be reduced either by laterally extensive layers of low permeability strata extending between the screen and the stream, and/or the occurrence of a "clogging" layer within the streambed itself.

- Laterally Extensive Layers

  Laterally extensive layers would normally be indicated by several driller's logs showing the presence of fine grained strata that occur at similar depths between wells. This information can be used to demonstrate the continuity of the layer extending beneath the stream and above the screen of the pumping well. Similar conclusions on laterally extensive low permeability layers can be drawn from a consideration of the elevation of well screens. For example, well screens within a given area which occur in two discrete depth zones separated by a zone with no screens indicates low permeability strata of such an extent that abstraction from the deeper zone is less likely to have a significant effect on stream flow. This situation is shown schematically in Figure 19.

  If the low permeability layer between the well screen and the stream is likely to be leaky then pumping may affect the shallow water table and an associated stream to still create a stream depletion effect. It is the transmission of a pumping effect through to the shallow water table that will influence whether stream depletion effects occur. This must be judged by the results of other pumping tests in the area and the conceptual hydrogeological understanding for the groundwater system.

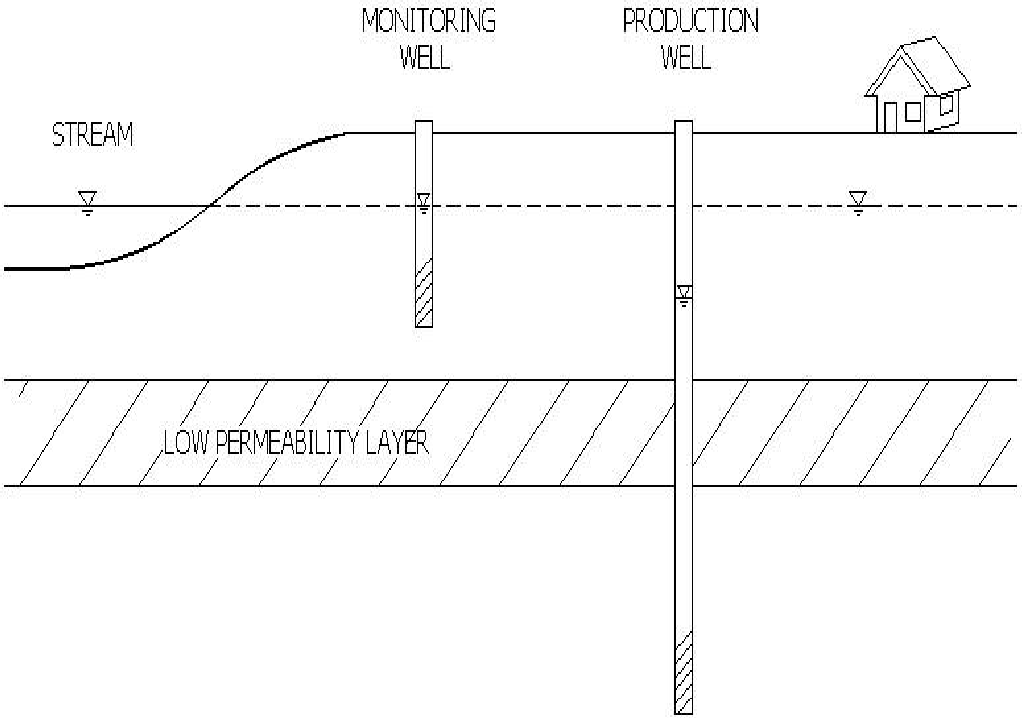


Figure 19: Indications of a deeper production aquifer that is hydraulically separated from the stream are provided by geological logs and water pressure me

- Streambed Hydraulic Conductivity

  Clogging layers within a streambed can form either from the natural strata through which the stream flows, or through the deposition of suspended sediments from the stream. Even in gravel bedded rivers, it is possible for finer grained sediment layers to form below the surface layer of well washed gravels. The presence of such layers is a significant factor that impedes the movement of water between the stream and the surrounding aquifer.

  Evidence of such low permeability layers may come from direct observation and probing of the streambed (to verify the thickness of the low permeability sediment) and excavation of test pits adjacent to the streambed. However, even if a clogging layer is present, there may still be discrete springs which allow water to move between the aquifer and the stream. Consequently, any conclusion about streambed clogging must be consistent with the conceptual hydrogeological understanding of the area.

3.1.3 Well Depth

Consideration has been given to defining a cut-off well depth below which stream depletion effects would not be significant – however, it is difficult to define such a depth. Whilst deeper wells are less likely to have a stream depletion effect the reason for this will be because of some intervening low hydraulic conductivity strata which should be demonstrated by the type of evidence described in sections 3.1.1 and 3.1.2. However, in the absence of any other information, if the only well in a particular area is a very deep well that may in itself be an indicator that the shallower strata has a low hydraulic conductivity that could not support a productive well. Such a conclusion would ideally be supported by a descriptive well drillers log.

The screening process described in section 3.1 has been summarised in a flow chart which is contained in the summary brochure of this guideline document which is presented in Appendix E. It sets out the consideration of geographical data, water level data and geological data which should be used to screen sites.

3.2 Parameters which Determine Stream Depletion Effects

The key parameters which indicate the magnitude of the stream depletion effect are:

» *Q – the abstraction rate from the well;*

» ℓ *– the separation distance between the well and the stream;*

» *t – the length of time over which the well is pumped;*

» *T – the transmissivity of the aquifer (a measure of how permeable the aquifer is)*

» *S – the storage coefficient of the aquifer (a measure of how much water is released from the pore space of the aquifer as water pressures fall)*

» *λ – the streambed conductance (a measure of the permeability and dimensions of the streambed and the hydraulic gradient across the streambed).*

The following panels have been prepared to describe these parameters and to indicate the range of values that are likely to occur if stream depletion effects are likely to be significant. This range of typical values can also be used as an initial screening procedure to determine whether stream depletion effects are likely to occur in any particular setting.

Parameter	Pumping Rate (Q)
Typical Units	m^3/day or L/s
Description	The average abstraction rate from a well over a fixed period of time. Note: The principle of superposition applies to stream depletion rates, therefore the effect of intermittent pumping can be simulated by the addition of effects resulting from a sequence of pumping and recovery. Jenkins (1977) concludes that "within quite large ranges of intermittency, the effects of intermittent pumping are approximately the same as those of steady, continuous pumping of the same volume." Therefore averaging of abstraction rates over a longer time period (e.g. an irrigation season) provides a useful estimate of stream depletion in many cases.
Typical Values	100 – 10,000 m^3/day Stream depletion effects increase with larger pumping rates.
Source of Data	• Direct measurement by the use of flow meters on abstraction wells. • Inferred rates from pump performance curves and readings of pump electricity meters. • Pumping rates are typically specified on resource consent applications.

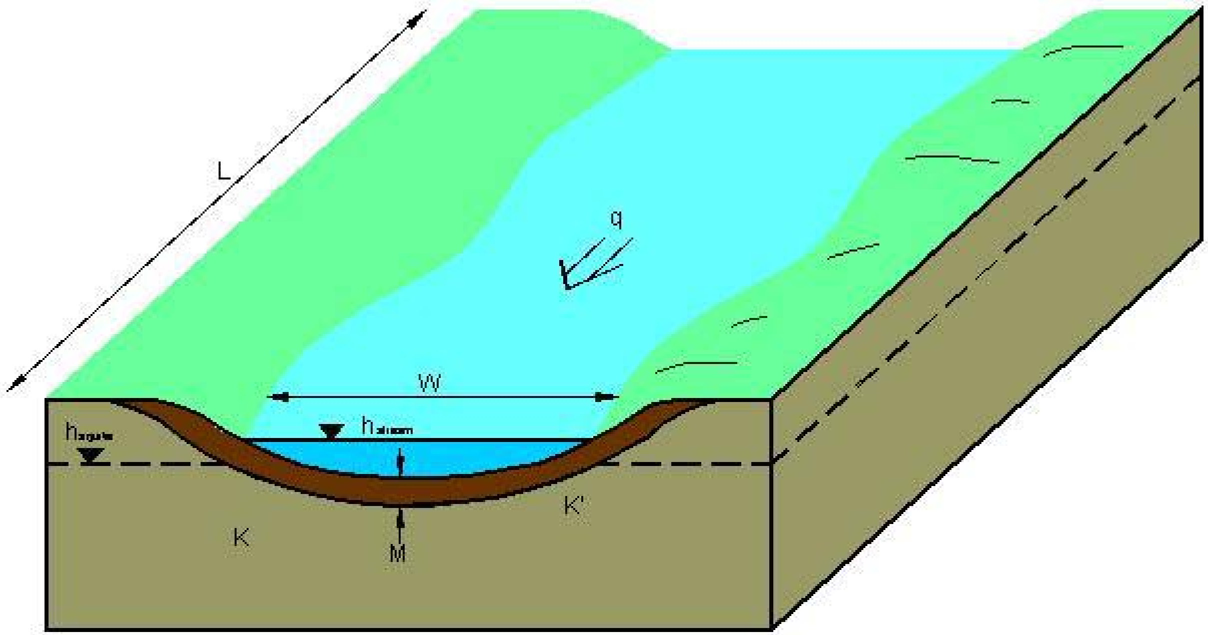
Parameter	Separation Distance (ℓ)
Typical Units	metres (m)
Description	The lateral separation distance from the abstraction well to the nearest edge of the stream water, measured perpendicular to the stream flow.
Typical Values	1 – 2,000 m Stream depletion effects increase with smaller separation distances. Where separation distances are greater than 2,000 m, stream depletion effects are unlikely to be significant.
Source of Data	• Topographic maps. • Aerial photos. • Direct measurement on the ground. Note: This measurement is clear cut for many streams, but is not so straightforward for braided rivers. In a braided river a judgement should be made as to the average location of the nearest river edge during the period of time over which the stream depletion effect is to be calculated. For example, if an assessment is being undertaken for a 35 year irrigation abstraction consent, it would be reasonable to assume that for at least one irrigation season the river may be at the closest edge of its braided bed width.

Parameter	Pumping Period (t)
Typical Units	days
Description	The duration of the pumping period of interest (see the note under "Pumping Rate" regarding intermittent pumping on page 32).
Typical Values	1 – 120 days (for irrigation wells). Continuous for pubic supply and industrial wells. Stream depletion effects increase with longer pumping periods.
Source of Data	• Direct measurement linked to monitoring of flow meters and/or electricity meters. • Inferred data from an irrigation design or commercial/industrial/reticulated supply requirements. • Pumping period details are typically specified on resource consent applications.

Parameter	Transmissivity (T)
Typical Units	m^2/day
Description	The transmissivity of the aquifer from which groundwater abstraction occurs (i.e. aquifer hydraulic conductivity x aquifer thickness).
Typical Values	10 – 10,000 m^2/day Stream depletion effects increase with higher values of transmissivity. Transmissivity values less than 10 m^2/day are unlikely to be sufficiently permeable to cause significant stream depletion effects.
Source of Data	• Pumping tests on abstraction wells – constant rate or step-drawdown. • Slug tests. • Estimates from specific capacity and/or geological logs from water wells. Note: The most reliable data comes from pumping tests on the well under investigation, provided that the test has used neighbouring observation wells and has been analysed in a way that takes the nearby stream into account (see section 6.1.2 and 6.2.3). Where multiple measurements of transmissivity are available from surrounding wells it is most appropriate to use the geometric mean of the values.

Parameter	Storage Coefficient (S)
Units	dimensionless
Description	The storage coefficient of the aquifer from which groundwater abstraction occurs (i.e. the volume of water released per unit volume of aquifer for each unit decline in the piezometric surface).
Typical Values	0.0005 – 0.3 Stream depletion effects increase with smaller values of storage coefficient. However, aquifers with values of S less than 0.0005 are likely to be confined. Under these circumstances there is unlikely to be sufficient hydraulic connection to a stream to cause significant stream depletion effects unless the stream channel penetrates the low permeability confining layer or there are discrete springs penetrating the confining layer.
Source of Data	• Pumping tests which utilise observation wells. • If no data is available, S = 0.1 is a typical value taken for settings where the hydrogeologic characteristics indicate the presence of an unconfined aquifer.

Parameter	Streambed conductance (λ)
Typical Units	m/day
Description	A measure of the vertical hydraulic conductance through the streambed to the underlying aquifer. Streambed conductance can be defined as: $$\lambda = \frac{K'W}{M}$$ where K' is the hydraulic conductivity of the strata in the streambed (m/day) W is the width of the streambed (m) M is the thickness of the streambed across which K' is measured (m) The relationship of these parameters is shown schematically in Figure 20.
Typical Values	0.01 – 5000 m/day Stream depletion effects increase with larger values of streambed hydraulic conductance. If the streambed conductance is less than 0.01 m/day then it is unlikely that stream depletion effects will occur.
Source of Data	• Gauging surveys (to determine gains or losses in flow along stream reaches) coupled with elevation surveys of stage height and groundwater levels. • Seepage meters. • Infiltration tests. • Excavation of test pits in dry streambeds for direct inspection of streambed strata.



$$\lambda = \underbrace{\frac{\Delta q}{L\Delta h}}_{\text{Gauging and Piezometric surveys}} = \underbrace{\frac{-K'W}{M}}_{\text{Infiltration/Seepage measurements}}$$

Figure 20: Streambed Conductance (λ)

3.3 Quantified Screening Criteria

If a stream and an adjacent aquifer have similar water levels (section 3.1.1) and are connected by permeable strata (section 3.1.2) then they fit the conceptual hydrogeologic settings in which stream depletion effects can occur.

To quantify how readily these effects will occur requires a consideration of the parameters described in section 3.1.3. For screening purposes, all the physical parameters of a site (excluding the abstraction schedule for the well, i.e. Q and t) can be combined into two screening parameters – the stream depletion factor (sdf) and the streambed conductance (λ).

3.3.1 Stream Depletion Factor (sdf)

The stream depletion factor was defined by Jenkins (1977) to describe the hydraulic connection between the stream and the pumping well. This term combines the aquifer parameters T and S with the separation distance to determine whether stream depletion effects will readily occur:

$$\text{sdf} = \frac{\ell^2 S}{T}$$

where ℓ is the separation distance between well and stream when the stream is approximated by a straight line [L]

T is the aquifer transmissivity [L^2/T]

S is the aquifer storage coefficient [dimensionless]

The effect of the sdf is shown graphically in Figure 21 for a well that pumps for 30 days and then is turned off for 30 days. (The data in Figure 21 assumes that the streambed offers no impedance to the flow of water between the stream and the aquifer (i.e. λ = 100 m/day)).

For settings with a small sdf the interaction between a pumping well and a stream is rapid and large (i.e. most of the well water is derived from the stream aquifer interaction within a short time from the commencement of pumping). However, as the sdf becomes larger, the stream depletion effect is more delayed and smaller. In some management situations an sdf value of 100 days has been used to differentiate between groundwater abstractions that can be managed to achieve some benefit in stream flow (sdf $\leq$ 100 days) and other abstractions that are less well connected to the stream (sdf > 100 days). Figure 21 shows that at a sdf value greater than 100 days the stream depletion rate is less than 20% of the well pumping rate after 30 days and the effect of shutting off the pump does not create an immediate benefit to the stream.

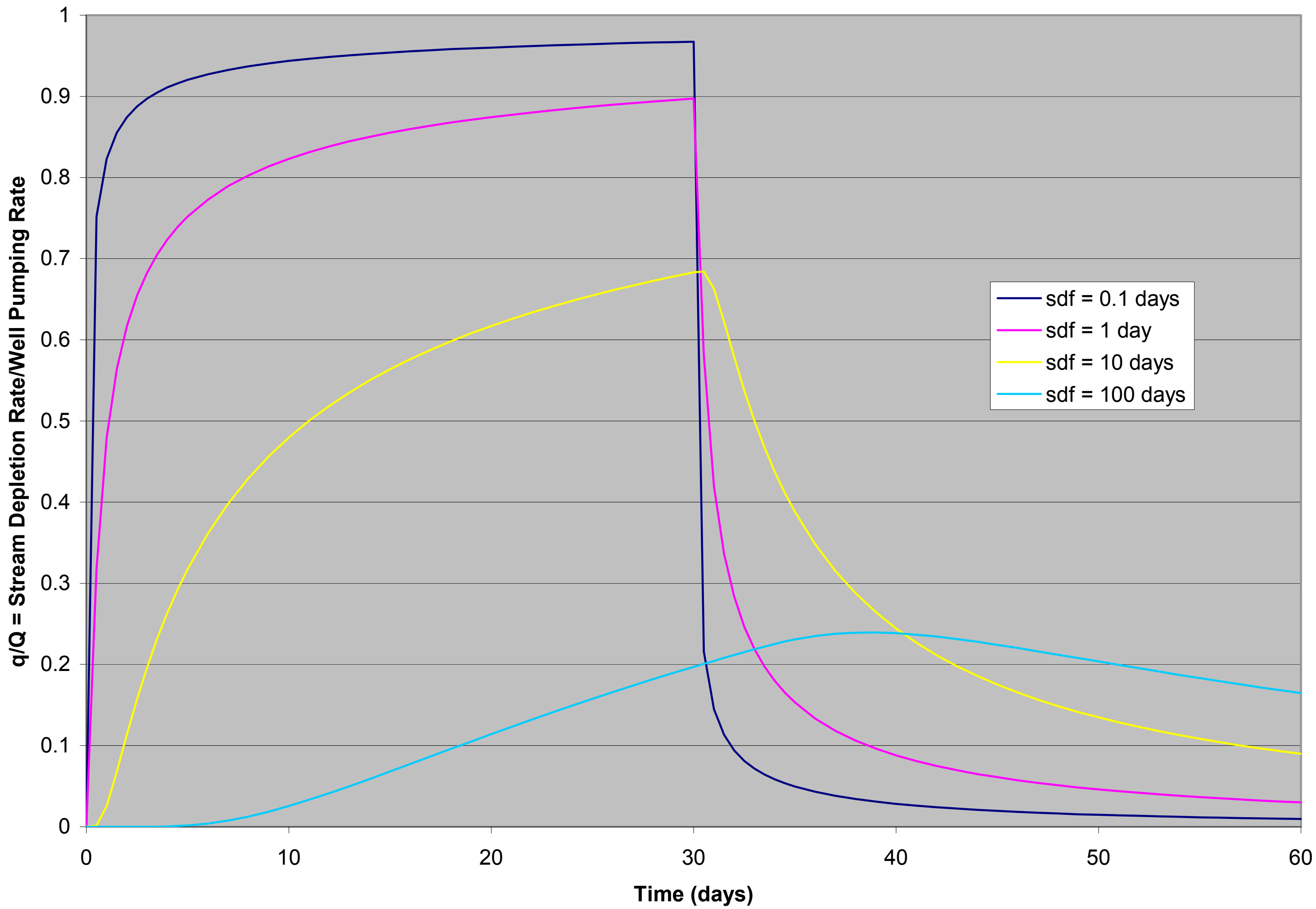


Figure 21: Relative stream depletion rate for a range of stream depletion factors (30 days pumping followed by complete shutdown, with lambda = 100 m/day)

3.3.2 Streambed Conductance (λ)

Streambed conductance is the term combining the factors that describe clogging in the streambed.

$$\lambda = \frac{K'W}{M}$$

where K’ is the hydraulic conductivity of the streambed [L/T];

W is the width of the streambed [L];

M is the thickness of the streambed [L].

The effect of λ is shown graphically in Figure 22 for a situation with a permeable aquifer (i.e. sdf = 0.1 days) and a well that pumps for 30 days and is then turned off for 30 days.

The curves show how the stream depletion rate reduces as λ decreases. This reduction becomes most noticeable when λ values are less than 10 m/day. When values of λ decline below 0.01 m/day it is unlikely that any significant stream depletion effects will occur.

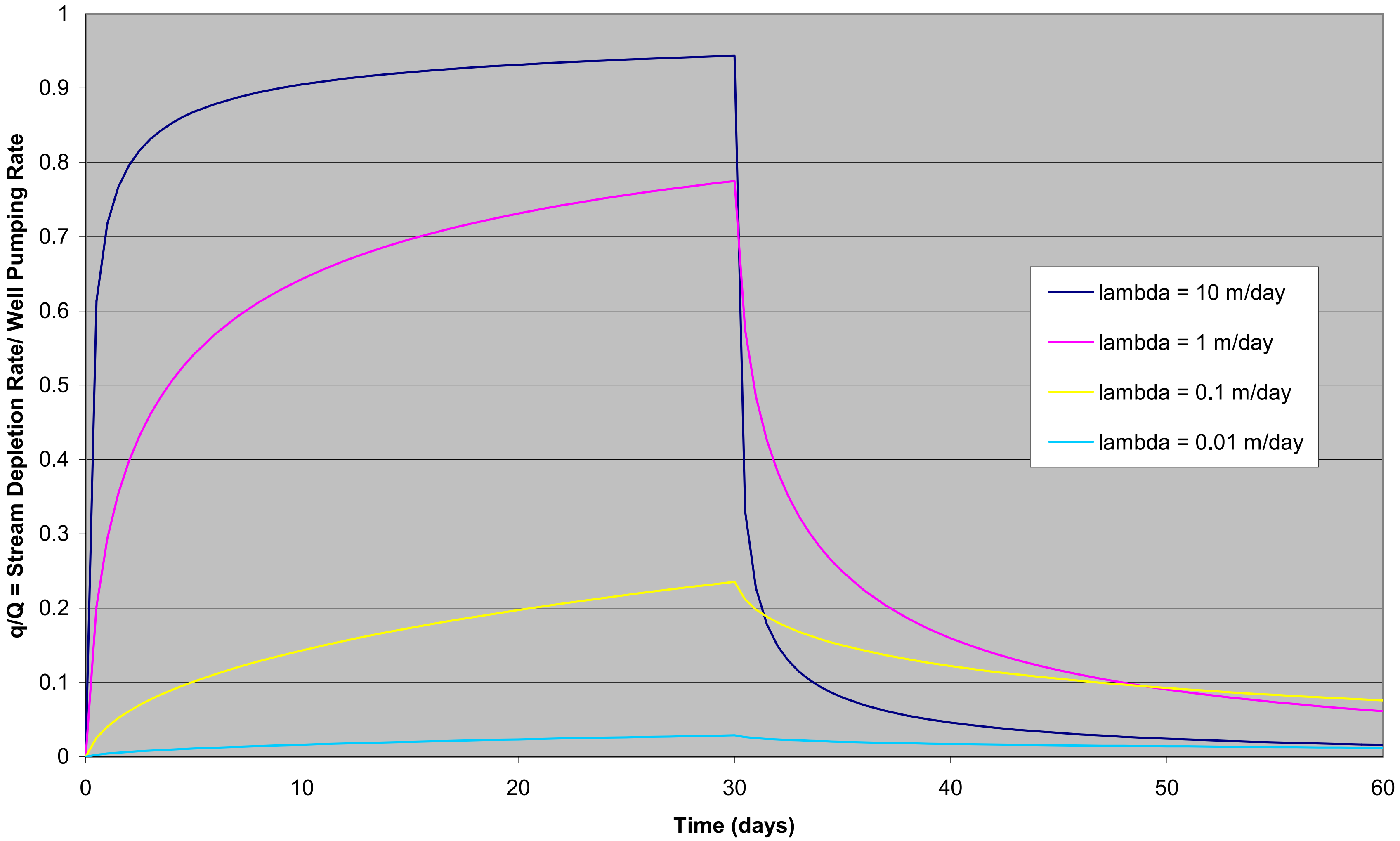


Figure 22: Relative stream depletion rate for a range of streambed factors (30 days pumping followed by complete shutdown, with sdf = 0.1 days)

4.0 Assessment Methods

This section presents quantitative tools which can be used to estimate the stream depletion effect of a pumping well on a nearby stream. As with all quantified groundwater assessments, they are a gross simplification of the complex variability that exists in naturally deposited groundwater systems. However, the methods described are considered to be the most appropriate means of estimating stream depletion effects.

Section 4.1 describes an analytical method developed by Dr Bruce Hunt of the University of Canterbury. It is appropriate for relatively uniform laterally extensive aquifer systems with a single continuous stream flow that can be approximated as a linear feature across the zone that is influenced by the pumping well. For other situations where the aquifer has well defined zones of variable parameters and/or where the stream connection to the aquifer is discontinuous, the effects can best be quantified by the use of a numerical model, as described in Section 4.2.

Before any quantified assessment is undertaken, it is essential that a realistic conceptual model of the stream and aquifer interaction is defined. The conceptual model must consider the following points:

» *the lateral extent of the aquifer and the location of any boundaries;*

» *the likely variability of the aquifer hydraulic conductivity and storativity;*

» *the location of stream flow that will occur for the duration of any calculation period;*

» *the nature of the streambed and its effect on water flow between the stream and the aquifer.*

This conceptual understanding will form the basis for selecting and applying the most appropriate method to quantify the stream depletion effect.

4.1 Analytical Equations

The original analytical equation to assess stream depletion effects was developed by Theis (1941) in the form of an integral which was evaluated with an infinite series. The equation was rewritten by Glover and Balmer (1954) using the complimentary error function, erfc, as follows:

$$\frac{q}{Q} = \mathrm{erfc}\left(\sqrt{\frac{S\ell^2}{4Tt}}\right)$$

where q is the stream depletion flow rate $[L^3/T]$;

Q is the constant flow rate abstracted at the well $[L^3/T]$;

S is the aquifer storage coefficient [dimensionless];

T is the aquifer transmissivity $[L^2/T]$;

t is the duration of the pumping period [T]; and

ℓ is the shortest distance between the well and the stream edge [L], based on the stream being approximated as a straight line feature.

The equation assumes that the stream fully penetrates the aquifer and forms a recharge boundary to the aquifer.

Glover and Balmer's form of the equation is the basis of a United States Geological Survey paper by Jenkins (1977). In outlining the purpose of his paper Jenkins observed that, "*The average user retreats in dismay when faced by the mysticism of 'line source integral', 'complimentary error function', or 'the second repeated integral of the error function'. The primary purpose of this report is to provide tools that will simplify the seemingly intricate computations and to give examples of their use.*"

Jenkins presented simple curves to estimate stream depletion effects and this approach, using the Glover and Balmer form of the equation has, until recently, been the most commonly used analytical tool for assessing stream depletion effects.

However, the use of the Glover and Balmer/Jenkins equation has been criticised based on comparative numerical modelling studies undertaken by Spalding and Khaleel (1991) and Sophocleous et al. (1995). They concluded that the equation tended to over-estimate stream depletion effects because, in reality, most streams only partially penetrate an aquifer and pumping wells may be able to create drawdown effects on the far side of the stream. Furthermore, many streambeds may have a "clogging nature" which has lower hydraulic conductance properties than the surrounding aquifer.

To address these concerns, Dr Bruce Hunt of the University of Canterbury has recently developed a new analytical equation which allows for the effects of both partial penetration and streambed clogging (Hunt, 1999). The Hunt equation is similar in form to the Glover and Balmer equation:

$$\frac{q}{Q} = \operatorname{erfc}\left(\sqrt{\frac{S\ell^2}{4Tt}}\right) - \exp\left(\frac{\lambda^2 t}{4ST} + \frac{\lambda\ell}{2T}\right)\operatorname{erfc}\left(\sqrt{\frac{\lambda^2 t}{4ST}} + \sqrt{\frac{S\ell^2}{4Tt}}\right)$$

where q is the stream depletion flow rate [L^3/T];

Q is the constant flow rate abstracted at the well [L^3/T];

S is the aquifer storage coefficient [dimensionless];

T is the aquifer transmissivity [L^2/T];

t is the duration of the pumping period [T];

ℓ is the shortest distance between the well and stream edge [L], based on the stream being approximated as a straight line feature; and

λ is the constant of proportionality between the seepage flow rate from the stream per unit length of streambed and the difference between stream and groundwater levels [L/T], as described in the panel on page 37 and in Figure 20.

Hunt's equation is based on the following assumptions:

» *The ratio of vertical to horizontal velocity components is small (the Dupuit approximation);*

» *The aquifer is of infinite extent and is homogeneous and isotropic in all horizontal directions;*

» *Drawdowns are small enough compared with saturated aquifer thicknesses to allow the governing equations to be linearised;*

» *The streambed cross section has horizontal and vertical dimensions that are small compared to the saturated aquifer thickness and the stream extends from y = - ∞ to y = ∞ along x = 0;*

» *The well flow rate, Q, is constant for 0 < t < ∞;*

» *Changes in water surface elevation in the river created by pumping are small compared with changes created in the water table elevation on the aquifer side of the semipervious layer;*

» *Seepage flow rates from the river into the aquifer are directly proportional to the change in piezometric head across the semipervious layer.*

These are all quite realistic assumptions for an analytical groundwater equation and are similar to those that are typically used for calculating drawdown around a pumping well. Consequently, the Hunt equation is recommended in this guideline as the most appropriate analytical tool to estimate stream depletion effects.

It is interesting to note that the equation has the following characteristics:

$$\frac{q}{Q} \to 0 \text{ as } t \to 0$$

i.e. when the well first starts pumping, there is relatively little water taken from the stream.

$$\frac{q}{Q} \to 1 \text{ as } t \to \infty$$

i.e. when the well has been pumping continuously for a very long time most of the well water is drawn from the stream.

$$\frac{q}{Q} = \operatorname{erfc}\left(\frac{\sqrt{S\ell^2}}{4Tt}\right) \text{when} \frac{\lambda\ell}{T} = \infty$$

i.e. the Hunt solution is equivalent to the Glover and Balmer/Jenkins solution if the streambed is assumed to be very conductive. Consequently, if no information is available on streambed clogging it will be conservative (from the stream point of view) to assume no clogging – in which case the Jenkins approach is a reasonable approximation.

To aid in the use of this equation, Figure 23 has been prepared to estimate stream depletion effects. Figure 23 shows the relation between the stream depletion factor (sdf) for pumping of duration t, and the rate of stream depletion q at time t, expressed as a ratio to the pumping rate from the well (Q). A family of curves is shown for different values of a dimensionless parameter referred to as the streambed factor (sbf) which reflects the effect of streambed clogging. It is defined as follows:

$$\text{sbf} = \frac{\lambda \ \ell}{T}$$

where λ is the streambed conductance [L/T]

ℓ is the separation distance between the stream and the well [L]

T is the aquifer transmissivity [L^2/T]

It is important to recognise that sbf is a dimensionless parameter to be used in conjunction with sdf in the graph in Figure 23. On its own, sbf is not a useful parameter for screening stream depletion effects due to the inconsistent influence of the parameters ℓ and T in the definition of sbf.

When sbf is 100 or greater, the curve in Figure 23 is the same curve that was presented in Jenkins (1977). However, when sbf reduces to values <10, noticeable reductions in stream depletion effects start to occur.

As noted in Jenkins (1977), the effects of intermittent pumping can quite reasonably be approximated by using the average pumping rate over the period of interest.

The following example has been prepared to demonstrate the application of Figure 23 to quantitatively assess stream depletion effects.

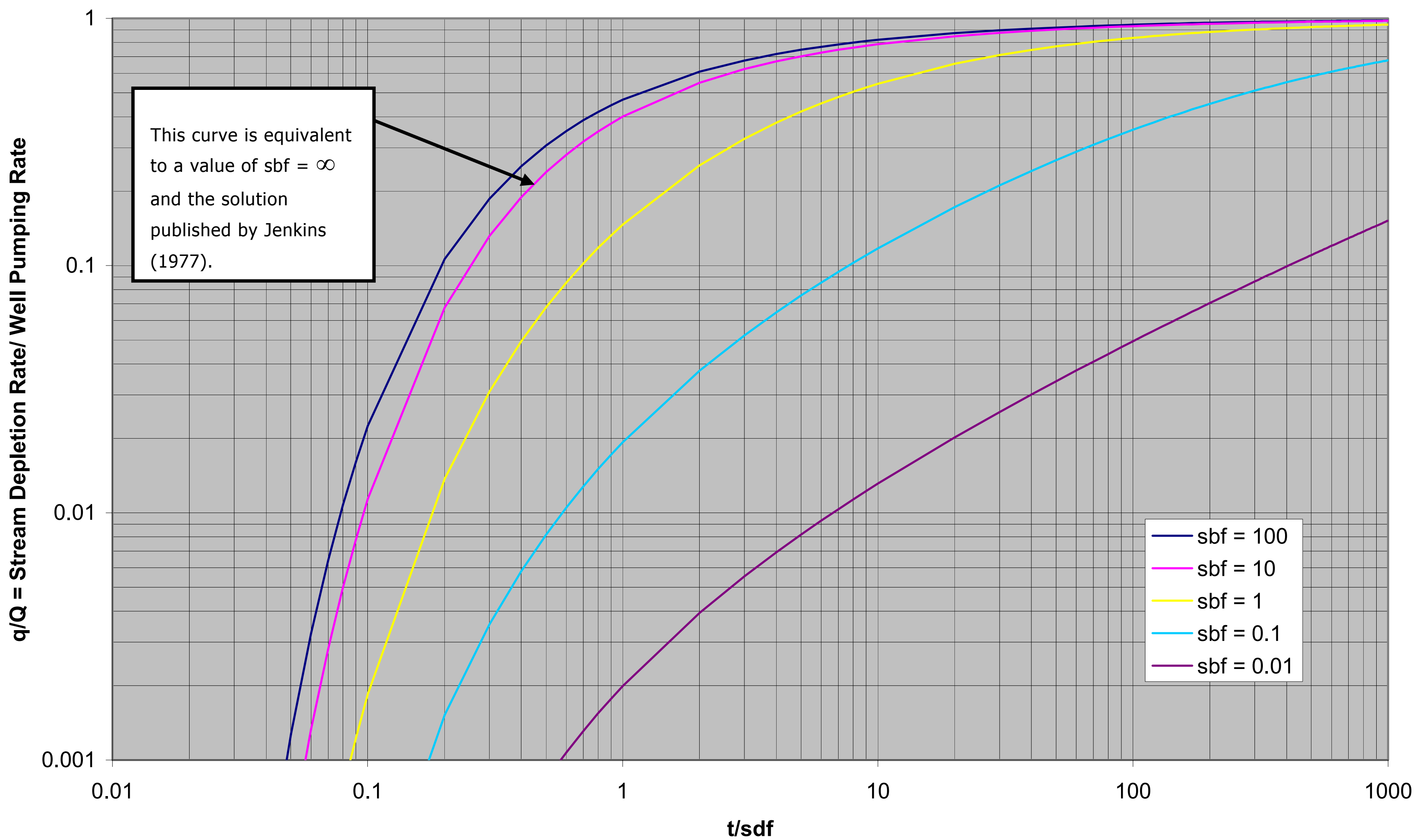


Figure 23: Curves to determine the rate of stream depletion

A shallow groundwater well in an unconfined aquifer pumps at 30 L/s for 20 hours per day, 4 days out of 5. The aquifer has a hydraulic conductivity of 80 m/day (K), a storage coefficient of 0.1 (S) and is 10 m thick (b). The well is located 200 m (ℓ) from a 5 m wide stream (W). The streambed is 1 m thick (M) and across this thickness the vertical hydraulic conductivity is one hundred times less than the hydraulic conductivity of the aquifer (i.e. K' = 0.8 m/day).

From this information we can determine the following:

Transmissivity = Kb = 80 m/day x 10 m = 800 m²/day

$$sdf = \frac{\ell^2 S}{T} = \frac{(200\,m)^2 \times 0.1}{800\,m^2\ /\ day} = 5\ \ days$$

$$\lambda = \frac{K' W}{M} = \frac{0.8\,m\ /\ day \times 5\,m}{1\,m} = 4\,m\ /\ day$$

$$sbf = \frac{\lambda\,\ell}{T} = \frac{4\,m\ /\ day \times 200\,m}{800\,m^2\ /\ day} = 1$$

With this information, the calculated stream depletion rate can be read off Figure 23 for the average pumping rate over different time periods, as tabulated below.

Pumping Period t (days)	*Average Abstraction Rate Q (L/s)*	$\frac{t}{sdf}$	$\frac{q}{Q}$	*Stream Depletion Rate q (L/s)*
0.83 (20 hours)	*30*	*0.167*	*0.009*	*0.3*
5	*20*	*1*	*0.15*	*3*
30	*20*	*6*	*0.45*	*9*
60	*20*	*12*	*0.57*	*11.4*
90	*20*	*18*	*0.64*	*12.8*

Using the principle of superposition, it can also be calculated that if after 60 days the abstraction ceased for 30 days the resultant effect on the stream depletion rate would be:

$q_{90} - q_{30}$ *=12. 8 – 9.0 = 3.8 L/s*

i.e. a reduction in the stream depletion rate of around 9 L/s, but a residual effect of around 4 L/s in the stream would still occur 30 days after pumping ceased.

4.2 Numerical Models

Inaccuracies in the estimates from the analytical equation can be caused by heterogeneity in both the aquifer and the stream. If the heterogeneity of the hydrogeologic system is such that it cannot be reasonably represented by the analytical equation described in section 4.1 then a numerical model could be used to assess the stream depletion effect.

There are a wide range of numerical models available for simulating groundwater flow. It is most important that a properly verified modelling code is utilised and that any modelling simulation is presented with appropriate checks to confirm its velocity. The discussion that follows focuses on the MODFLOW software (McDonald and Harbaugh, 1996) which is the most widely accepted groundwater flow model in New Zealand at the present time. However, the comments that are made below could be generically applied to any other modelling package that is utilised.

MODFLOW is a three-dimensional finite difference model for simulating the flow of water through a porous media. The area being modelled is divided vertically into layers and laterally, into a gird of rectangular cells with unique, uniform aquifer properties being defined for each cell. The movement of water between cells is calculated through the iterative solution to a sequence of finite difference equations.

It is beyond the scope of this guideline to give detailed instructions on the use of MODFLOW, however, for those persons with competence in the use of MODFLOW, the following notes outline the key features that should be present in a MODFLOW model that is developed to assess stream depletion effects. The MODFLOW programme is divided into a number of packages that define different characteristics for a particular aspect of the numerical simulation. Those packages required to formulate a stream depletion assessment are described below:

(i) The BASIC Package

The input data for the Basic package defines the following model characteristics:

» *it defines the model grid, i.e. the rows, columns and layers;*

» *it defines the time steps and stress periods for the simulation (a stress period is a period where the external model boundaries remain constant. Within each simulation, a stress period is divided up into a number of smaller time steps);*

» *it specifies the initial head distribution within the aquifer at the start of the simulation;*

» *it specifies the boundary conditions for the model grid.*

For the purposes of a stream depletion model it is important to remember that each model cell represents uniform aquifer properties. Consequently, a fine grid size (perhaps on the order of 10 m x 10 m) should be used in the vicinity of both the stream and the pumping well, where head change and groundwater fluxes may be greatest.

Similarly, a small time step must be used at the start of each stress period, although this can be easily accommodated by the use of MODFLOW's time step multiplier – a common approach is to divide each stress period into 10 time steps (NSTP) with a multiplier (TSMULT) of 1.5.

In contrast to the fine grid required around the pumping well and the stream, a coarser grid is required away from the area of interest so that the model can run efficiently, with the minimum number of rows, columns and layers. When expanding the grid size, the dimensions of each adjacent row or column should not increase by more than 50% between adjacent cells. Also, the horizontal dimensions of each grid cell should not be more than 10 times bigger in one direction than the other.

The external boundaries to the model should either coincide with real aquifer boundaries, or they should be placed at sufficient distances that they do not interfere with the accuracy of the simulation in the area of interest. The magnitude of any artificial boundary effects should be checked by running the model through two simulations, one without the well pumping and one with the well pumping. If the results show that a large proportion of the well water is coming from artificially located boundaries then the model grid must be redesigned to shift the artificial boundaries further away so that they do not interfere with the solution.

(ii) The BLOCK CENTRED FLOW Package

The input data for the Block Centred Flow Package defines the grid geometry, the hydraulic conductivity and storage coefficient parameters of each cell and the top and bottom elevations of the layers. These input parameters should be determined from the conceptual hydrogeological model, on the basis that the simplest model is the best. Ideally large groups of cells should have constant aquifer parameters and sharp contrasts in parameters should be avoided, unless there is a good hydrogeological basis for such a change.

Anisotropy in the aquifer parameters can be specified for those areas where this has been shown to occur.

(iii) RIVER Package and STREAM Package

Modflow has two packages that can be used to specify the stream: the RIVER package and the STREAM package. Both packages specify the cells in which the stream occurs, the stage height of water in the stream, the height of the bottom of the streambed and the conductance of the streambed. The flow between the stream and the aquifer, for each model cell, is shown graphically in Figure 24, and calculated as:

QRIV = CRIV (HRIV – h) for h > RBOT

i.e. the flow to or from the river varies depending on the head in the aquifer

QRIV = CRIV (HRIV – RBOT) for h ≤ RBOT

where: QRIV is flow to or from the stream

HRIV is the stage height of the stream

h is the head that the model calculates for the aquifer cell in which the stream occurs

RBOT is the height of the bottom of the streambed

CRIV is the hydraulic conductance of the stream-aquifer interconnection

$$CRIV = \frac{K' LW}{M}$$

where: K' is the hydraulic conductivity of the streambed [L/T];

L is the length of the streambed as it passes across the model cell [L]:

W is the width of the streambed as it passes across the model cell [L];

M is the thickness of the streambed [L]:

It is interesting to note that the value of CRIV per unit length of the stream $\left(\frac{CRIV}{L}\right)$ is the same value as the streambed conductance (λ) described in section 3.2.

The Modflow River and Stream packages assume that the flow from the river is constant when the head in the aquifer is lower than the base of the streambed. This is not entirely consistent with the guidelines of Hunt and Bouwer described in section 3.1.1. To allow for situations where this is thought to occur, it may be desirable to set RBOT at a lower elevation than the real streambed (i.e. down to the elevation below which constant leakage will occur).

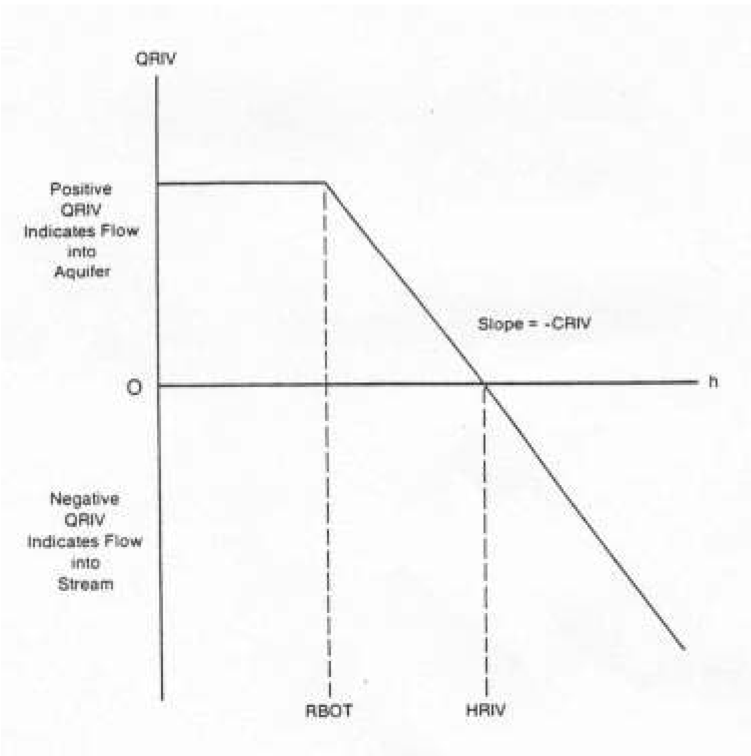


Figure 24 Plot of flow (QRIV) from a stream into or out of the modelled aquifer as a function of the head (h) in the modelled aquifer. RBOT is the elevation of the bottom of the streambed and HRIV is the elevation of the water in the stream (from McDonald and Harbaugh, 1983).

The RIVER package specifies a constant stage height for each stress period and operates under the assumption that the stage height remains constant, regardless of the movement of water between the stream and the aquifer. This package is most appropriate for simple simulations where the stream has a steady flow, significantly in excess of any possible stream depletion rates.

In contrast, the STREAM package provides the option of specifying an initial stage height and stream flow at the start of the model simulation. Throughout the model simulation, the stage height and stream flow are then determined by the interaction in flow between the aquifer and the steam. The STREAM package can allow the stream to go dry, if losses to the aquifer are sufficiently great.

(iv) WELL Package

The input to the well package specifies the location of the well and the pumping rate. A constant pumping rate is specified for each stress period.

(v) Solver Package

Modflow has a number of options for solver packages:

» *Strongly Implicit Procedure Package (SIP);*

» *Slice Successive Overrelaxation Package (SSOR);*

» *Preconditioned Conjugate-Gradient Package (PCG2);*

» *WHS Solver for Visual MODFLOW (WHS).*

Various solvers can be trialled, but all should give satisfactory solutions to stream depletion simulations. One of the key parameters for each solver is the specification of the head change criterion for convergence (HCLOSE). This determines that the iteration during each time step is concluded when the maximum absolute value of head change from all cells between the two most recent interactions is less than the value specified. In most cases a value of 0.001 m or less should be sufficient.

(vi) Running MODFLOW Simulations for Stream Depletion Assessments

In carrying out these modelling exercises it is important that the stream depletion effect caused by the pumping well must be carefully isolated from the other groundwater flow interactions that occur within a MODFLOW simulation. To achieve this it is recommended that stream depletion assessments should be run on the following basis:

» *Firstly, a steady-state simulation should be run with no well pumping. The water budget at the end of this simulation must show a low percentage error (ideally below 1%) to confirm that the model has run correctly. The resulting aquifer heads, stream flow and stage heights should then be used as the input data for the stream depletion simulation.*

» *The stream depletion simulation should be run with a minimum of three stress periods:*

  » *firstly with no well pumping;*

  » *secondly with the well pumping;*

  » *thirdly with no well pumping.*

The water budget output at the end of each stress period must be carefully checked. The percentage error must be low (ideally well below 1%) to confirm that the model has run correctly. Furthermore, the proportional distribution of water inflow and outflow to the model must also be carefully checked for each stress period. In particular, if a significant proportion of the pumped well water is sourced from artificially placed model boundaries *then an artificial result is obtained and the model must be run again, with more appropriate boundary definition.*

Once these checks have been satisfactorily complied with, the model output can be used to quantify the magnitude and timing of the stream depletion effect.

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. λ = 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

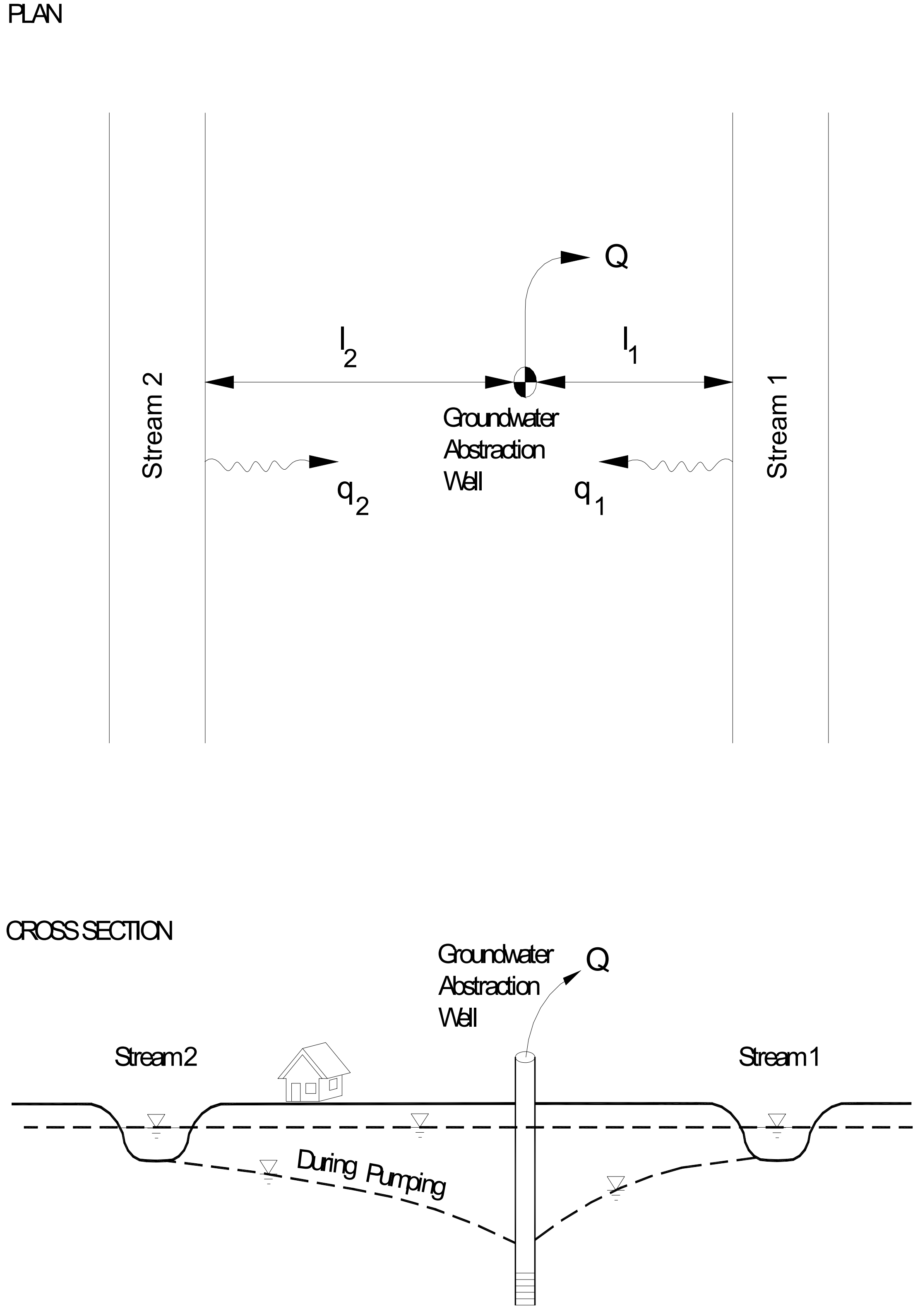


Figure 25: Schematic View of Two Streams Bordering a Groundwater Abstraction Well

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

$$q_{analytical}$$

or

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

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

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

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.

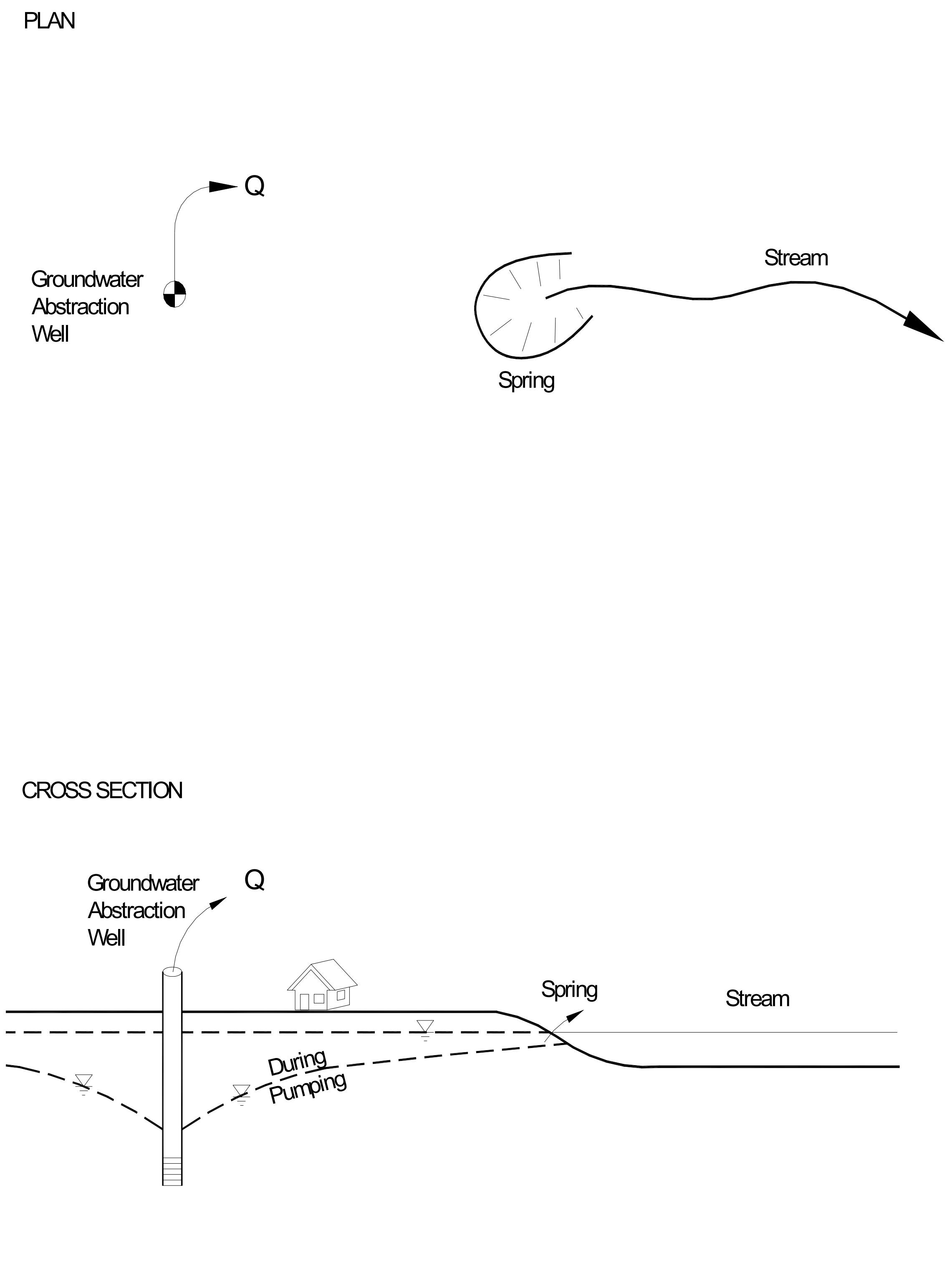


Figure 26: Schematic View of a Groundwater Abstraction Well Upstream of a Spring Fed Stream

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).

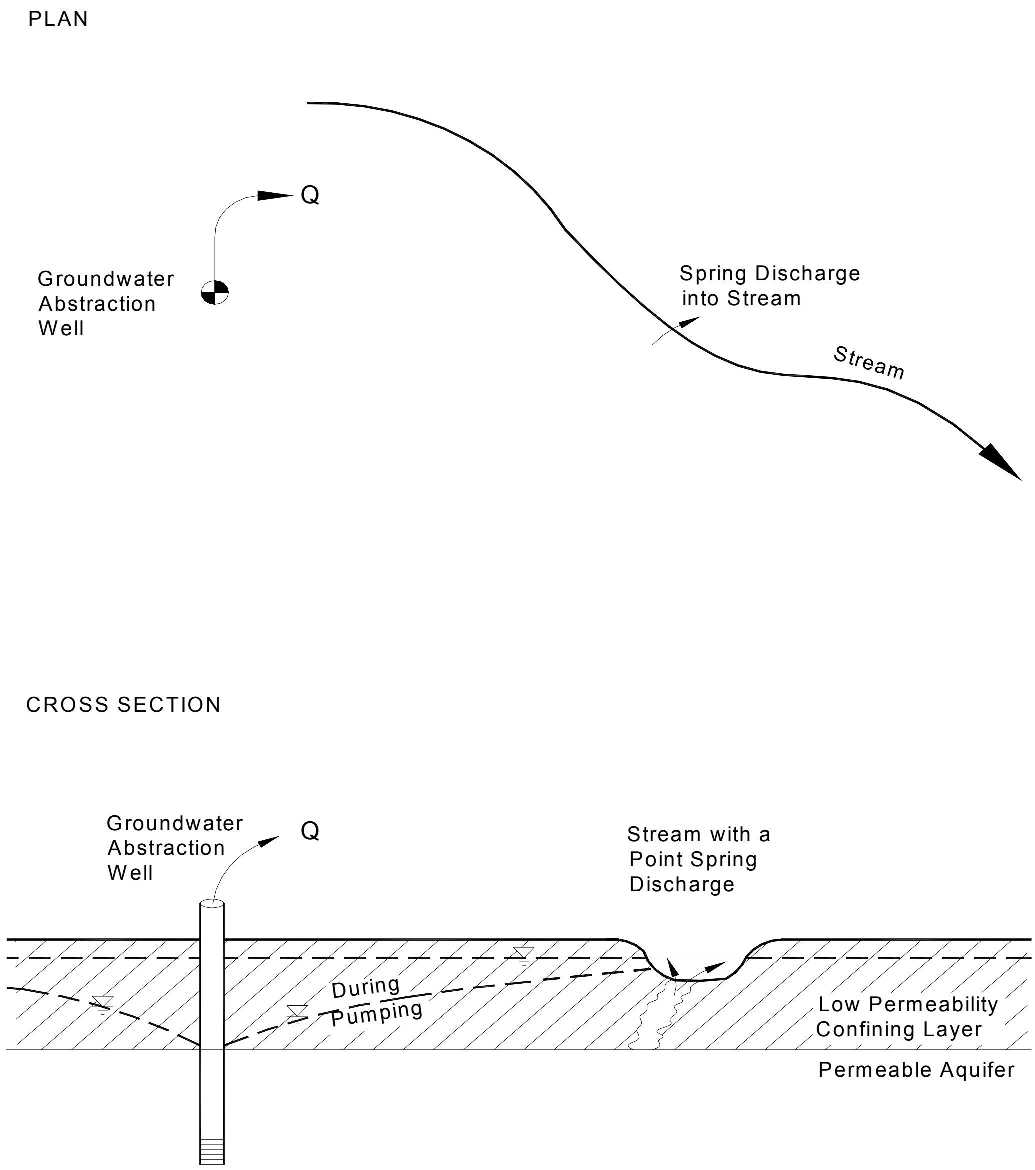


Figure 27: Schematic view of a Groundwater Abstraction Well Affecting an Artesian Spring

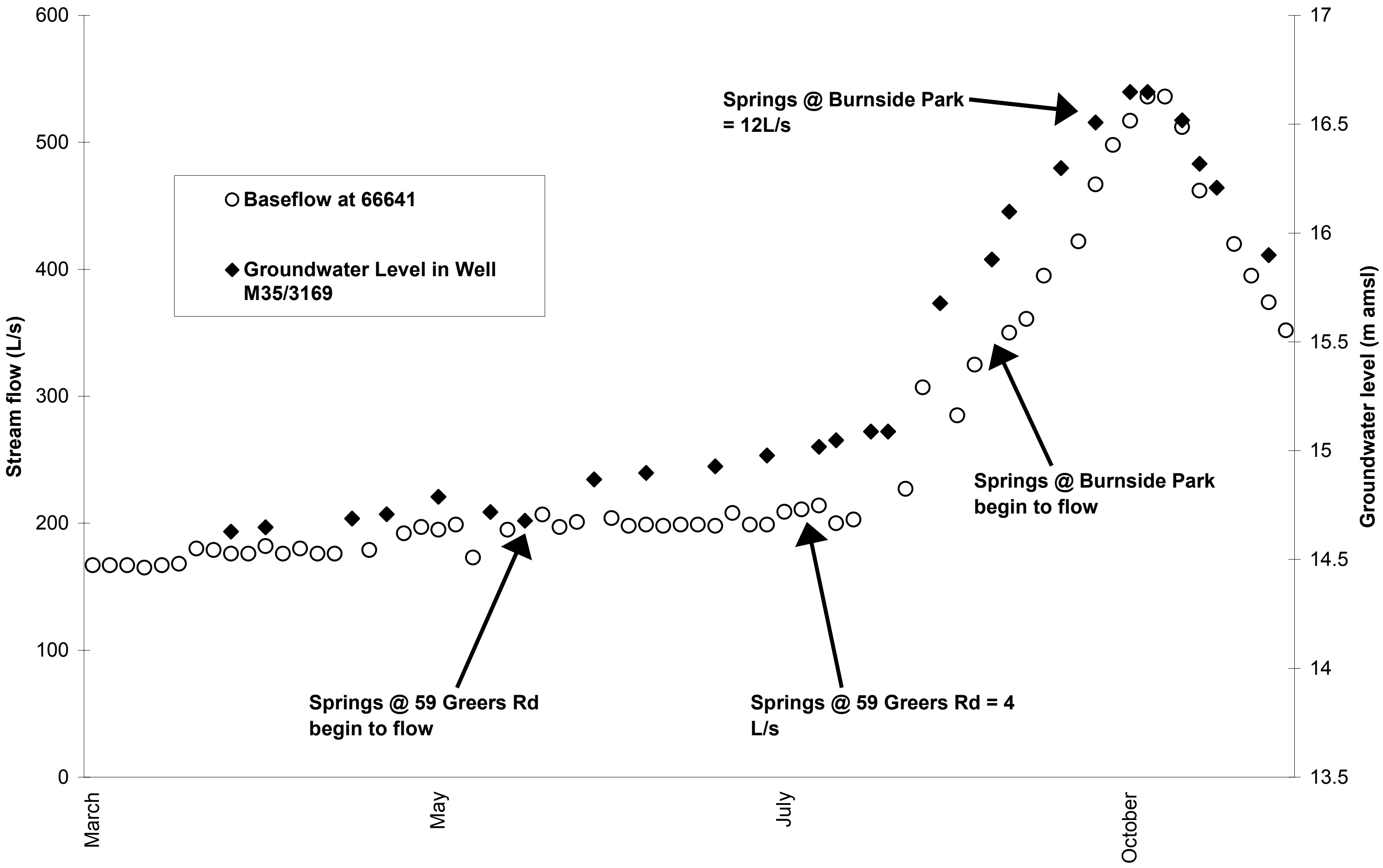


Figure 28: The discharge of two artesian spring sections in the Waimari Stream relative to baseflow rates and groundwater levels

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];

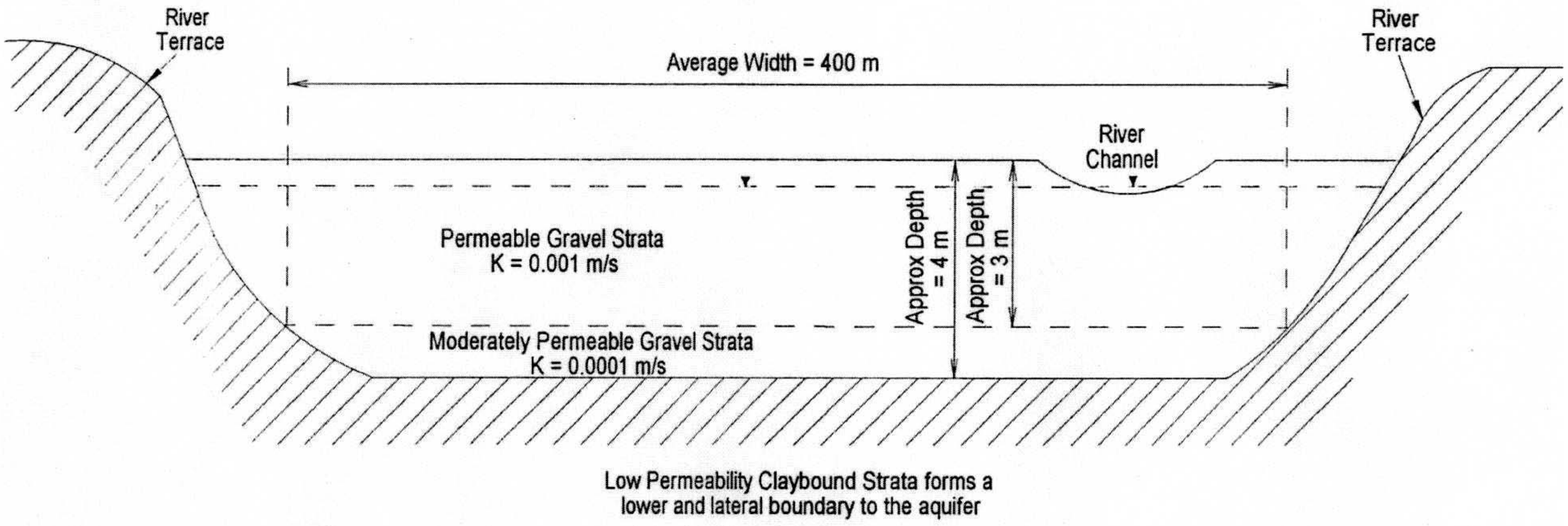
Width is the width of the aquifer (between the low permeability terraces) [L];

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 m/s x –0.009 x (3-Depth to Water) x 400 m
= 0.0036 m^3/s x (3-Depth to Water) = 311 m^3/day x (3-Depth to Water)

Moderately permeable gravels: Q = -KiA = -0.0001 m/s x –0.009 x 1 m x 400 m
= 0.00036 m^3/s = 31 m^3/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.

6.0 Collection of Field Measurements

6.1 Field Measurements to Assist Stream Depletion Assessments

The quantification of stream depletion effects can be greatly enhanced by the collection of relevant field data in the area under investigation. These field measurements are used to asses the following aspects of the stream depletion problem:

» *streambed conductance or the hydraulic conductivity of the streambed*

» *direct measurements of seepage across the streambed*

» *aquifer characteristics*

» *the flow regime between the stream and the aquifer*

There are four main types of field measurements which help the quantification of stream depletion effects. They are:

» *Gauging surveys combined with piezometric surveys*

» *Pumping tests*

» *Seepage measurements in streambeds*

» *Infiltration measurements in streambeds*

Each of the field measurement techniques is discussed in turn below.

6.1.1 Gauging and Piezometric Surveys

Combined stream gaugings and piezometric surveys involve measuring the flow of water in the stream at various locations along its length and measuring the level of water both within the stream and in wells adjacent to the stream. This gives information on the nature of the hydraulic interaction between the stream and the aquifer and allows a broad averaging of parameters because measurements are made across an extended length of streambed.

This type of survey is used to determine the streambed conductance (λ). If the width of the stream (W), the thickness of the clogging layer (M) and length of the stream reach (L) are known then the hydraulic conductivity can also be determined from the gauging survey. Streambed conductance is related to the streambed hydraulic conductivity as follows (Figure 20):

$$\text{Streambed conductance/unit length} = \frac{K'W}{M} = \frac{\Delta q}{L\Delta h}$$

where K' = hydraulic conductivity of the streambed [L/T];

W = width of streambed [L];

M = thickness of streambed [L];

Δq = change in flow rate in river between the upstream and downstream gauging sites [L^3/T];

L = length of streambed between the upstream and downstream sites [L];

Δh = difference in elevation between the water surface in the stream and the groundwater table adjacent to the stream (as determined from direct measurement or interpreted piezometric contours) [L].

A piezometric contour map can be constructed from a levelling survey to allow groundwater levels at the stream gauging locations to be approximated. Alternatively mini piezometers can be installed at the stream gauging location to allow direct measurement. The difference between the estimated groundwater level (as indicated from the piezometric map or measured directly) and the measured stream level at each gauging site gives Δh.

It is preferable for these surveys to be carried out during times when no pumping is taking place and relatively steady groundwater and surface water flow conditions exist.

It is also important that gauging surveys are conducted with great accuracy and that the results are interpreted with regard to the magnitude of potential errors relative to the magnitude of flow that are measured. This matter is discussed further in section 6.2.1.

6.1.2 Pumping Tests

A pump test comprises pumping water from a well whilst measuring the discharge of the well, the drawdown in the well and the drawdown in nearby observation piezometers at known distances from the pumped well. These measurements can be substituted into an appropriate well-flow equation allowing the hydraulic characteristics of the aquifer and in some cases of the stream to be calculated. Pumping tests are used in stream depletion assessments in three main ways.

Firstly, a pumping test carried out in a well that is not adjacent to a stream, allows the calculation of general hydraulic characteristics of an aquifer. In particular, the test allows a determination of:

» *the transmissivity and storage coefficient of the aquifer;*

» *the structure of the aquifer, for instance whether the aquifer is unconfined, confined or leaky, and/or has lateral boundaries;*

» *if the aquifer is leaky, a pump test can be used to determine the characteristics of the leaky low permeability layer. This helps to assess the transmission of drawdown effects through a semi-confined layer to determine the pumping influence on a shallow water table and stream.*

Secondly, where a pump test is conducted adjacent to a stream, the drawdown response in a well can be analysed using the Hunt match point method (Hunt 1999) which allows T, S and λ to be calculated.

Thirdly, in some particularly well controlled settings, pumping tests adjacent to streams can sometimes be used to directly measure stream depletion, where the stream flow is accurately gauged up and down gradient of the radius of influence of the wells pumping (as estimated by drawdown simulations). The difference between the gaugings before and after pumping can be inferred to have been caused by the well pumping. This method is best used in conjunction with the pump test analysis described above. Used alone the results can often be inconclusive because of the margin of error inherent in flow gauging surveys compared to the flow depletion rate, and also due to the changes in antecedant aquifer conditions and stream flows. These inaccuracies are discussed in section 6.2.

6.1.3 Infiltration Tests

Infiltration rings can be used to estimate the vertical hydraulic conductivity across the streambed. However, as they only measure values at a specific spot they are less representative than gauging surveys or pumping tests. As a result, the more measurements that can be made over as wide an area as possible will increase the reliability of these tests.

For this field method a double ring system is typically used to create vertical flow downwards through the streambed. This involves driving two steel rings into the streambed as shown in Figure 30.

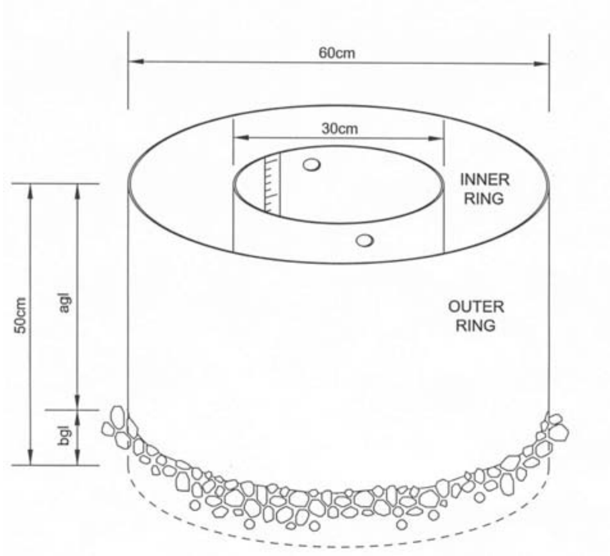


Figure 30 Double ring infiltrometer

Both rings are filled with water and the incremental volumes (V) of water required to keep the central ring at a constant level over the time of the test is recorded. The level of water in the outer ring is also kept constant but is not measured. The purpose of the outer ring is to ensure that flow from the central ring is predominantly vertical. Over the duration of the test the infiltration rate reduces and once it reaches a stable condition (i.e. a flattening of a graph of rate of water level decline vs time since start of test) the hydraulic conductivity of the streambed, K', is approximately proportional to the rate of infiltration, i.e.

$$K' = \frac{V}{t\ i\ A}$$

where K' = vertical hydraulic conductivity [L/T];

t = time [T];

V = the volume of water required to keep the water level constant over the time t [L^3];

A = the cross-sectional area of the ring [L^2];

i = hydraulic gradient [dimensionless].

The hydraulic gradient is determined by measuring the difference in head between the water in the ring and the hydraulic head in the ground along the seepage path, as monitored by a piezometer or tensionmeter. As an alternative approach, in dry streambeds this test can be carried out over areas where the depth to the water table is greater than 5 times the depth of water in the ring (as discussed in section 3.1.1) so that a vertical hydraulic gradient of 1 can be assumed.

The test can also be carried out in calm shallow water. In these circumstances only a single ring is required as the stream provides a relatively constant water level around the ring.

The analysis of the single ring test in a calm stream requires that a flow net be constructed, to determine the hydraulic gradient and hence the hydraulic conductivity from Darcy's equation. This approach was used in the Environment Canterbury and Pattle Delamore Partners Ltd study of the Ohoka Stream which is referred to in Appendix C. The construction and use of flow nets is discussed fully in many references (e.g. Craig 1992).

Because this test relies on the seepage of water from the ring into the ground it may not yield reliable data if there is an upward natural gradient acting across the test area.

6.1.4 Seepage Surveys

In areas of deeper water, the rate of seepage through the streambed can be measured directly using a seepage meter. While detailed designs differ, in essence, the seepage meter consists of a cylinder with one open end and one closed end (e.g. a 200 L drum cut in half). The open end is pressed into the streambed, so that the entire cylinder is submerged. The closed end is connected to a flexible bag (e.g. a rubber weather balloon) through which any seepage water will move (Figure 31). If the flow through the streambed is upwards then seepage water will accumulate in the flexible bag. If the flow is downwards then the bag is filled with water at the start of the test and the volume of water in the bag will reduce as the test proceeds.

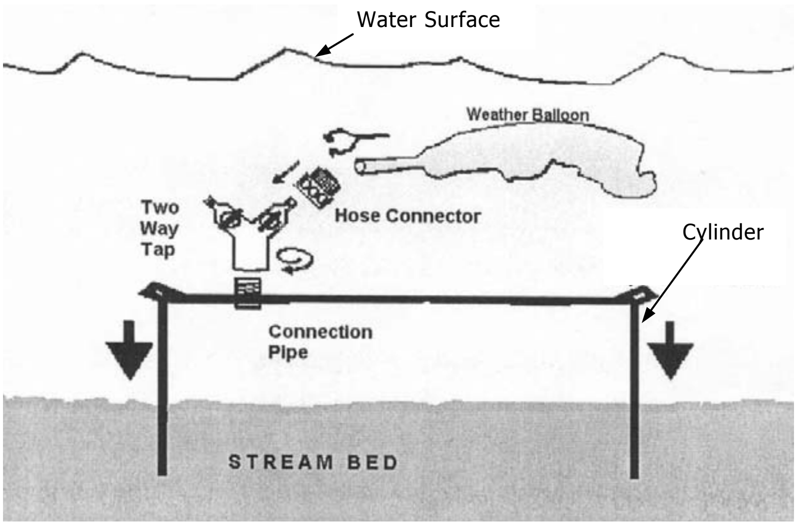


Figure 31 Seepage meter. (from ECan)

The hydraulic conductivity, K', of the streambed is determined using Darcy's equation:

$$K' = q / i A$$

where q = flow rate determined by seepage meter [L^3/T];

A = area of seepage meter [L^2];

i = $\Delta h/M$ = vertical hydraulic gradient, measured by the head difference between groundwater and the stream divided by the vertical distance (M) over which the head difference is measured [dimensionless];

K' = vertical hydraulic conductivity [L/T] .

The aquifer pressure beneath the stream may be measured by mini-piezometers within or adjacent to the streambed, or from nearby wells. However, the more distant the groundwater monitoring point (both laterally and vertically) the greater the level of uncertainty in the measurement of gradient.

As with the infiltration rings, seepage meters only make measurements at isolated points in the streambed, and a large number of measurements over a wide area of the streambed will increase the reliability of the results of these tests.

Appendices A – D have been prepared to present examples of field measurements and stream depletion assessments that have been made in different Regional Council areas of New Zealand.

6.2 Field Measurements that May Not Assist Estimates of Stream Depletion Effects

The field measurements described in section 6.1 are considered to be the most reliable means of gathering the key hydrogeologic data to allow an accurate assessment of stream depletion. This section briefly outlines some other field measurements that could be made, but may give misleading results due to incorrect assumptions.

6.2.1 Inaccurate Gauging Surveys

It is conceptually appealing to assess stream depletion effects by simply pumping a well for a period of time and monitoring the flow in a nearby stream to see the change that occurs.

Float methods and current meters can be used to measure the water flow rate and when this information is combined with the cross-sectional area of the flow channel, the flow rate can be estimated. However, this has not been found to be a reliable method of assessment due to measurement inaccuracies coupled with background fluctuations in stream flow compared to the relatively small effect from a pumping well, particularly over short pumping periods (i.e. less than 48 hours).

The Australian Water Resources Council note that under the most favourable conditions, margins of error in float measurement are about 10%. If non-uniform conditions occur within the stream (which is a more typical situation) then margins of error of 25% or more can be expected. A well designed current meter survey under favourable conditions may be able to achieve margins of error of about 5% to 10%.

An example of the background variability that occurs in streams is demonstrated from data reported in Weir (1999) on the Doyleston Drain which is presented as a case study in Appendix D. Figure 32 shows background monitoring results of stream flow and groundwater levels around the time of a carefully controlled pumping test. Weirs were installed in the stream to measure the stream flow as accurately as possible during a controlled pumping test. The results detect a measurable reduction in stream flow during the pumping period and this has been attributed to the stream depletion effect.

However, even under this most carefully controlled condition there is uncertainty due to the recovery of flow in the downstream weir before pumping stopped (perhaps due to external effects on the stream or to a timing offset in the flow recording device) and the larger reduction in stream flow after pumping stopped (perhaps due to changes in barometric pressure). This example is probably the best experimental field data that is available for the assessment of stream depletion effects, yet the uncertainties listed above highlight the difficulty of using gauging assessments over a pumping test as a means of producing convincing data to quantify stream depletion effects.

The fact of the matter is that stream flow is naturally quite variable and stream depletion effects from wells typically build up over a prolonged period of pumping (i.e. several days). Over the period of time that stream depletion effects occur it is likely that the background fluctuations in stream flow will be of sufficient magnitude to swamp the attempted stream depletion measurement, particularly when allowance has to be made for the inaccuracies in the gauging method. All these factors should be assessed before contemplating any gauging measurements to assess stream depletion effects.

6.2.2 Water Chemistry Analyses

Consideration may also be given to measuring the change in water quality characteristics that occurs between the surface water body and the groundwater. If such a parameter could be found then monitoring of the water quality in the pumped well could detect a change in the proportion of surface water being pumped from the well. For instance, a common example of the change in water chemistry between surface water and groundwater occurs in the pH value of the water:

Groundwater	Surface Water
low pH	high pH
$H^+ + HCO_3^-$	$H_2O + CO_2$ (gas)
$\equiv$	

One difficulty that exists with such chemical indicators of surface water is that many of the chemical parameters will typically undergo chemical transformation as they move through the streambed and into the subsurface environment. Consequently, by the time any water that originated from the surface water body reaches a well, its chemical composition will have been modified so that it can no longer be directly compared with the chemical composition of surface water.

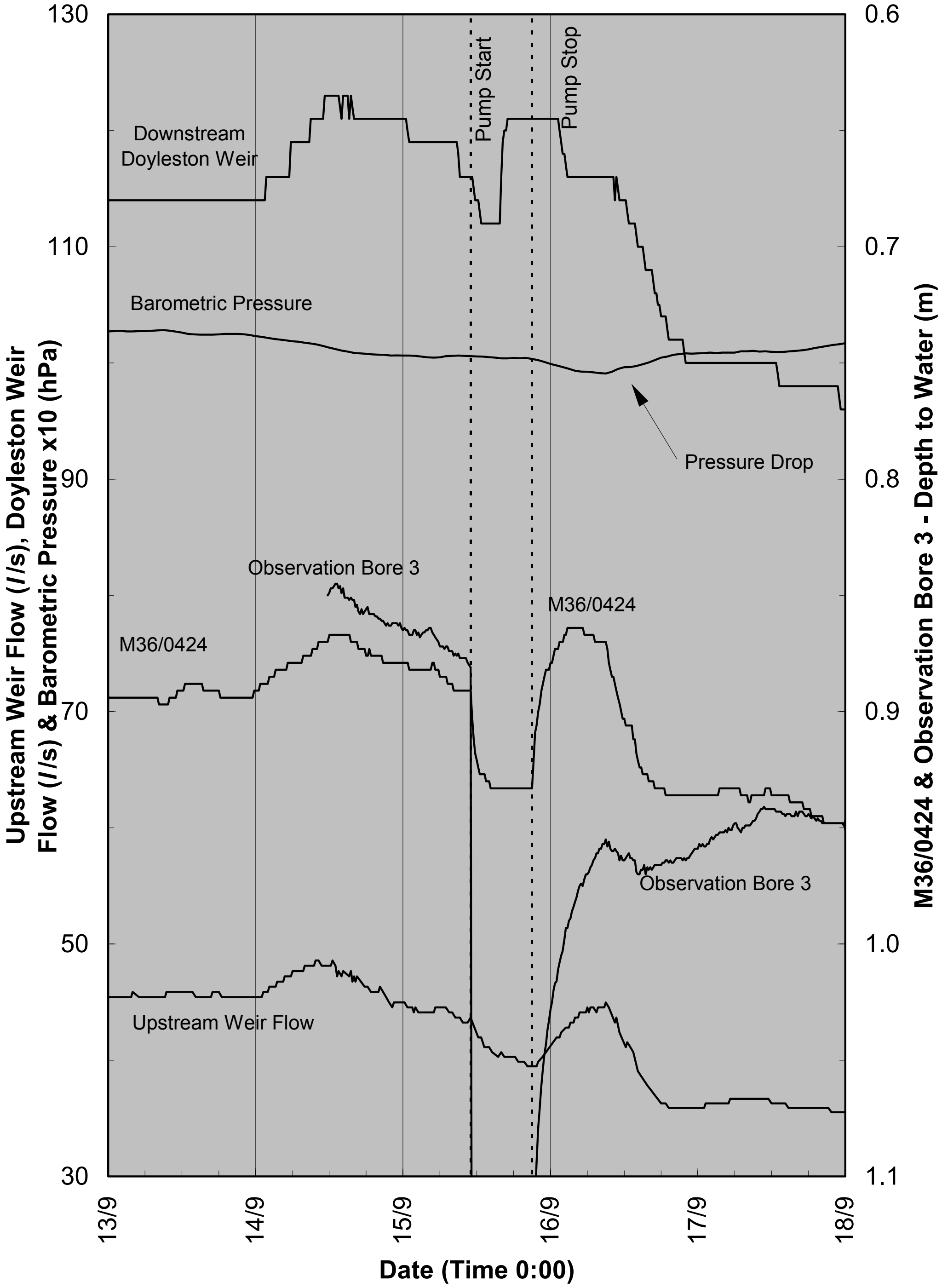


Figure 32: Variations in Stream flow (from Weir, 1999)

Of even greater significance though is recognition that stream depletion effects can occur without any stream water reaching the well. The stream depletion effect is caused by a pumping well creating a change in hydraulic gradient adjacent to the streambed which results in a loss in stream flow, or an equivalent reduction in groundwater seepage that would otherwise enter the stream.

This effect can occur without any surface water actually been drawn into the well. As a result, sampling the water quality of the pumping well for surface water indicators cannot be used as a reliable measure of stream depletion effects.

6.2.3 Pumping Test Results Near Streams

Pumping test are a common hydrogeological field technique to determine aquifer transmissivity and storage characteristics. They involve the pumping of an individual well at a constant rate for a period of several hours whilst the drawdown in water levels is measured both in the pumped well and in surrounding "observation wells", as described in section 6.1.2. These tests are typically analysed using methods that assume the aquifer is of infinite extent. However, if such tests have taken place in aquifers where stream depletion effects are expected to occur then it may be inappropriate to use aquifer parameters that have been analysed using the assumption of an infinite aquifer. In such circumstances it will be more appropriate to re-analyse the test data using the approach described in Hunt (1999).

It has also sometimes been misleadingly reported that because pumping tests have created very small drawdowns, there must be no stream depletion effect occurring. However, it is often the recharge from a stream that causes the small drawdown effects during pumping.

6.2.4 Water Divining

Water divining or dowsing is claimed to be a technique available to a few "gifted" people who have the ability to "divine" groundwater flow paths. This is most often carried out by the use of forked sticks, wire rods, hoops, pendulums or similar instruments. The instruments are typically held in a position where a small change in muscular tension (either deliberately or subconsciously) results in a deflection. The nature of the deflection is claimed to indicate the depth, direction and/or rate of water flow beneath the ground. These deflections occur as the diviner walks across the area of interest and/or surveys a map or aerial photograph of the area.

Whilst many wells have been successfully "divined" this has typically occurred in areas where groundwater can successfully be found without the help of a diviner. In fact, successful divining seems to rely on a general knowledge of existing geological and well performance characteristics in an area. Williamson notes that a book entitled, "The Modern Dowser: A Practical Guide to Divining", advises that, "It is useless to look for water where geology tells us there cannot be any. The dowser then must have a special knowledge of geology and especially that of the country where he is working." Williamson also states, "it has been found that the most successful diviners are those who are good observers and well experienced in the area in which they operate, their failures becoming less frequent as their experience increases. In fact, if the groundwater conditions are particularly favourable, they may not have had any failures at all".

Bowden et al. (1983) express the view that reliance on divining has "led to a great deal of wasted time and money". They state that, "to demonstrate their worth, diviners would have to show a success rate significantly better than that of a groundwater geologist, an experienced driller or indeed random chance. This has never been done and in fact the opposite is true. There is no acceptable scientific evidence that water divining works; whenever controlled tests have been done, the claims of the diviners have been disproved."

This view is clearly not shared by diviners and many others who have used their services to find successful groundwater sources. However, the lack of verifiable scientific evidence to support their claims means that divining does not provide reliable information on the movement of water between streams and pumping wells. On this basis it cannot be used to help in the assessment of the issues presented in this guideline.

7.0 Management Implications

It is expected that the quantification of stream depletion effects, described in these guidelines, will prove most useful in the assessment of resource consent applications and in the development of rules for the management of water resources.

7.1 Resource Consent Applications

As part of the Assessment of Environmental Effects that is required for a resource consent application, consideration should be given to whether stream depletion effects are likely to be an issue of concern. This should be done by assessing the screening criteria that are described in section 3.1 of these guidelines and summarised in the flow chart presented in Appendix E. These are relatively straightforward parameters, which should generally be judged on existing information. However, surveyed elevations of both stream levels and groundwater levels along with the installation of monitoring boreholes and/or test pits may help to improve the assessment.

If it is concluded that stream depletion effects are unlikely to be an issue of concern then no further consideration needs to be given to this matter.

On the other hand, if the indications are that adverse stream depletion effects may occur, then they must be considered further. A suggested starting point is to consider that an adverse effect on the stream could occur and propose appropriate consent conditions that avoid the potential adverse effect. If these conditions are acceptable to all parties, then no further investigation is needed.

However, if the conditions are considered undesirable, then there are a number of progressively more detailed steps that a consent applicant can undertake to improve the understanding of their potential stream depletion effect. These steps are described below. Depending on the level of information provided in the consent application, it is suggested that the points listed below can also be used by Regional Authority Officers to provide guidance in preparing a "Request for Further Information" as specified in Section 92 of the Resource Management Act (1991).

Step 1 The first step to quantifying the stream depletion effects is to assign a numerical value to the parameters described in section 3.2 of these guidelines. The definition of each parameter, its likely range and a data source are listed in the panels of section 3.2. It is unlikely that all parameters will be precisely defined. For those parameters that are not well defined then the range of values that are most likely to apply at a site should be considered. The definition of values must be based on credible sources of information.

Step 2 The known values (or range of likely values) from Step 1 should be used to calculate the value (or range of values) of the stream depletion factor (sdf) and streambed conductance (λ), as described in section 3.3. Using these values, consideration should be given to Figures 21 and 22 to determine whether the stream depletion effect is likely to be significant and to reconsider the proposed consent conditions.

Step 3 If further quantification is required then the values (or range of values) from Step 1 should be used to calculate the likely stream depletion effect using the techniques described in Sections 4 and 5 of this document. The appropriate calculation method must be consistent with the conceptual hydrogeologic model for the area. If assumptions and simplifications are to be made, they should be clearly stated along with their impact on the resulting calculation.

Step 4 If the accuracy of the result is still considered undesirable then more detailed site specific field investigation will be required. It is to be expected that by the time this point is reached the following parameters should have been well defined:

» *Pumping rate from the well (Q);*

» *Separation distance between the well and the stream (P);*

» *Pumping period (t)*

Assuming this is the case, then more detailed investigations will involve implementing the field methods described in section 6.1. It is recommended that a sensitivity analysis should be carried out on the calculations undertaken in Step 3 to see which of the poorly defined parameters have the biggest influence on the calculated stream depletion effect. Field investigations should be prioritised to focus on those parameters which have the greatest effect on the calculated result. The choice of field investigations is summarised below, based on the consideration of which issues are most sensitive.

Information Required	**Recommended Field Test**
• Improved definition of the interaction between the stream and the aquifer	• Gauging and piezometer survey (section 6.1.1) and/or • A detailed pumping test with several observation bores (section 6.1.2)
• Improved definition of the aquifer Transmissivity (T) and/or Storage Coefficient (S)	• A detailed pumping test with several observation bores (section 6.1.2)
• Improved definition of streambed conductance (λ)	• Gauging and piezometer survey (section 6.1.1) • Infiltration tests (section 6.1.3) in reaches which are dry or have calm, shallow water • Seepage meter surveys (section 6.1.4) in reaches with slow moving deep water • A detailed pumping test with several observation bores (section 6.1.2)

To carry out this field work in a reliable manner requires detailed site observations that may often be beyond the resources of an individual consent applicant. However, the results of the field tests may often have a wider applicability than the assessment of a single consent application. For example, gauging and piezometric surveys and direct infiltration/seepage measurements along a streambed reach could be used in the assessment of all groundwater abstractions within the catchment. Therefore, it may be more feasible for a group of water users, recreational interest groups, Territorial Authority and/or Regional Authority to arrange field investigations that allow an improved assessment of stream depletion effects along a particular stream reach in which all the parties have an interest.

7.2 Water Resource Management Plan Issues

It is beyond the scope of these guidelines to discuss the management options for groundwater abstractions which affect stream flow. However, it is worth pointing out that an effective management approach is dependent upon a knowledge of how the stream depletion effect occurs. Consequently, the assessment tools described in this guideline are an essential first step in determining the effectiveness of any management option.

In particular, the timing and severity of restrictions on groundwater pumping should be based on the degree of hydraulic connection between the pumping well and the stream. As shown in Figures 21 and 22, groundwater pumping restrictions may be quite beneficial to a stream for settings with a low sdf and a high λ. However, restrictions will be relatively ineffective if sdf is high (e.g. > 100 days) and/or if λ is low (< 0.01 m/day), particularly for low – moderate pumping rates.

Management policies must also take into consideration the characteristics of the surface waterway that could be depleted. Stream depletion effects only become a water management issue if the affected surface waterway has important values that are adversely affected by low flows which are contributed to by the stream depletion effect. This situation is typically identified by the stream having a minimum flow and/or a water allocation regime established for surface water users. The magnitude and timing of any stream depletion effects should be compared with the magnitude of the river flow to assess their significance.

This guideline sets out a variety of calculation tools and field investigation techniques which can be used to better understand and quantify the effects of a pumping well on nearby stream flow. These techniques range from simple screening observations to detailed quantitative evaluation. The level of detail that is required in each particular setting should be matched by the likely management implications for any particular groundwater abstraction.

The key parameters which allow the necessary understanding of this effect are the hydrogeologic characteristics of the aquifer and their interaction through the streambed. In the absence of any field data it is conservative (from the stream's perspective) to assume that the streambed has the same hydraulic conductivity as the adjacent aquifer. However, for streams where groundwater pumping effects have the potential to be an important water management issue then the implementation of field measurements (as described in Section 6.1) should be undertaken to aid in the improved management of the water resource. As noted in section 7.1, the resources of several interested groups could be pooled to allow field measurements along a full stream length thereby creating a very worthwhile benefit for improved water management within the catchment.

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Appendix A:

Case Study A – Papawai Stream, Wellington Regional Council

A1. Available Information to Set the Scene

In the Wairarapa, Wellington Regional Council (WRC) have assessed the effects of an irrigation abstraction well (reference number 5G/68/8/1) on the nearby Papawai Stream. WRC have provided the following information contained in two reports: Draft WRC report titled "Papawai Stream Water Allocation Current Information Knowledge", 1999; and "Papawai Stream and Groundwater Monitoring 1995-96 Irrigation Season" by Greg Butcher.

The Papawai Stream is a small springfed stream which originates approximately 100 m south of the intersection of Church Street and East Street in Greytown (Figure A1). At Fabian Road (around 2,000 m downstream of the headwaters) the mean flow is 307 L/s and the median flow is 268 L/s. Between November 1981 and February 1996 the mean annual daily low flow was 209 L/s.

The irrigation well is 8.2 m deep, located approximately 180 m from the south tributary of the Papawai Stream and is permitted to abstract water at a rate of around 40 L/s. It draws water from a shallow aquifer with an average transmissivity of 3,300 m^2/day. Groundwater quality data, environmental isotopes and groundwater hydrographs indicate that the aquifer is recharged by a combination of river flow and rainfall recharge.

Figure A2 shows the correlation between a groundwater monitoring well (located between the irrigation well and the Papawai Stream) and stream flow measured at Fabians Road. This figure indicates a strong relationship exists between measured groundwater levels and the flows in the Papawai Stream (particularly the base flows). The same monitoring well also shows the effect of pumping (at a rate of 40 L/s) from the nearby irrigation well, which appears to cause a drawdown effect of 0.1 – 0.2 m over a period of 80 days.

On the basis of this information the reports concluded that "any significant abstraction of shallow groundwater in the area could cause significant reduction in flow from the Papawai Stream" (Draft WRC report – Papawai Stream Water Allocation Current Information Knowledge, 1999). and "Results of monitoring during the 1995-96 irrigation season suggest that under current levels of groundwater pumping approximately an 18% reduction in the flow of the Papawai Stream may occur at Fabians Road' (Greg Butcher "Papawai Stream and Groundwater Monitoring 1995-96 Irrigation Season").

A2. Analysis

The conceptual hydrogeological setting of a shallow aquifer and a nearby stream, with flows that correlate with groundwater levels, indicate a setting where stream depletion effects have the potential to occur.

The analytical method for calculating stream depletion effects (section 4.1), can be applied to this situation using the following parameters:

Transmissivitiy of 3,300 m^2/day

a separation distance from the stream of 180 m

Storage coefficient of 0.1 (a typical value an unconfined aquifer)

These parameters define a stream depletion factor of 0.98 days which indicates that stream depletion effects are likely, provided that the streambed is conductive.

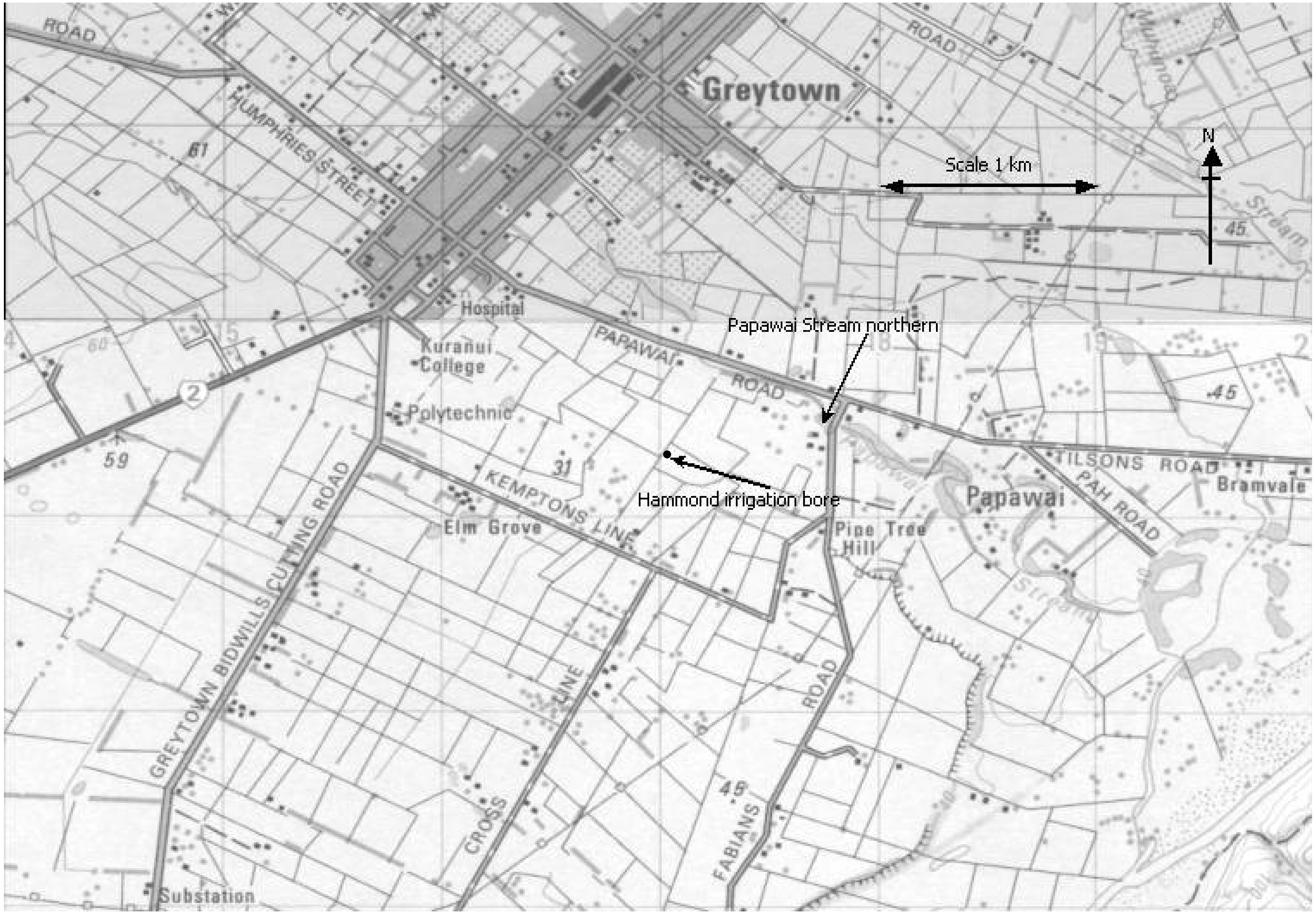
With regard to the streambed clogging layer it can be assumed for the setting described above that, in the absence of any other data, the aquifer transmissivity is indicative of the streambed conductance. Because the irrigation well is 8.2 m deep it is not unreasonable to assume a 10 m thick aquifer, giving a hydraulic conductivity of 330 m/day $\left(K = {}^{T}\!/_{b} = K'\right)$.

Assuming a stream bed thickness of 1 m, and a stream bed width of 10 m we are then able to calculate the stream bed conductance, λ as being 3,300 m/day.

Figure A3 shows the calculated effect of this abstraction on stream flow.

The parameters indicate that stream flow is estimated to reduce by 90% of the bore's average pumping rate after 30 days pumping – a result that is consistent with the Wellington Regional Council's assessment. Stream depletion rates would continue to be around 80% of the well abstraction rate after 30 days pumping even if the stream bed hydraulic conductivity was as low as 3.3 m/day (100 times less than the aquifer hydraulic conductivity, i.e. K' = 0.01 K).

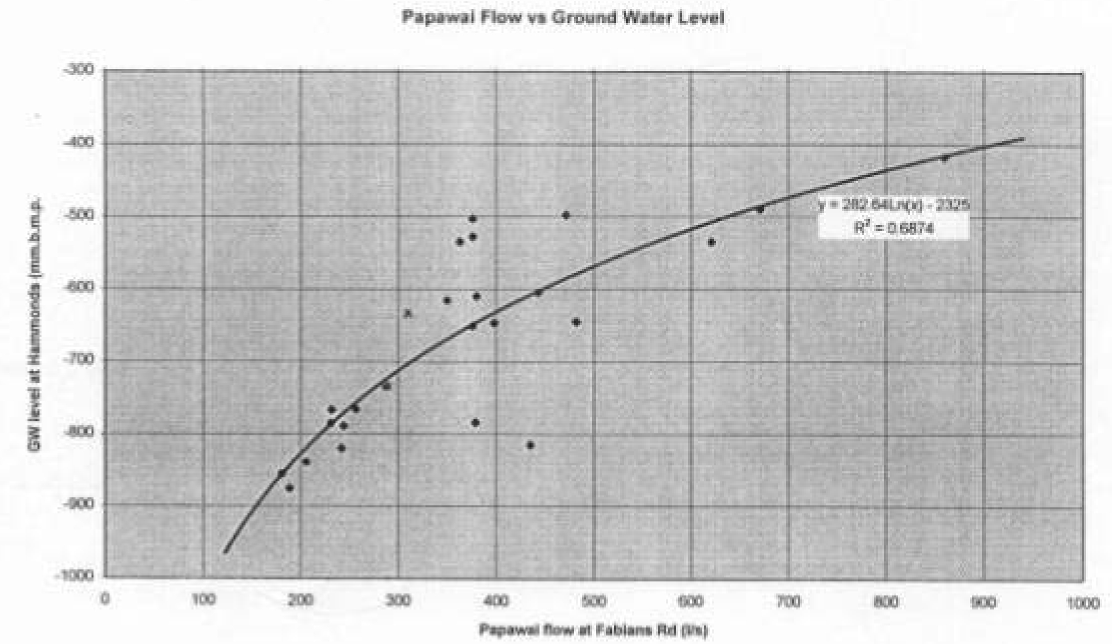
In this case, the quantitative assessment using the technique in this guideline supports the inferences made from the WRC's assessment of groundwater level and stream flow data, whilst giving a more realistic picture of the transient changes in stream depletion effects over a period of prolonged pumping.



NZMS 260 Map S26 & S27.

Sourced from Land Information New Zealand data.

Figure A1 Location Map for Case Study A



A plot using only the low flow gaugings is shown below.

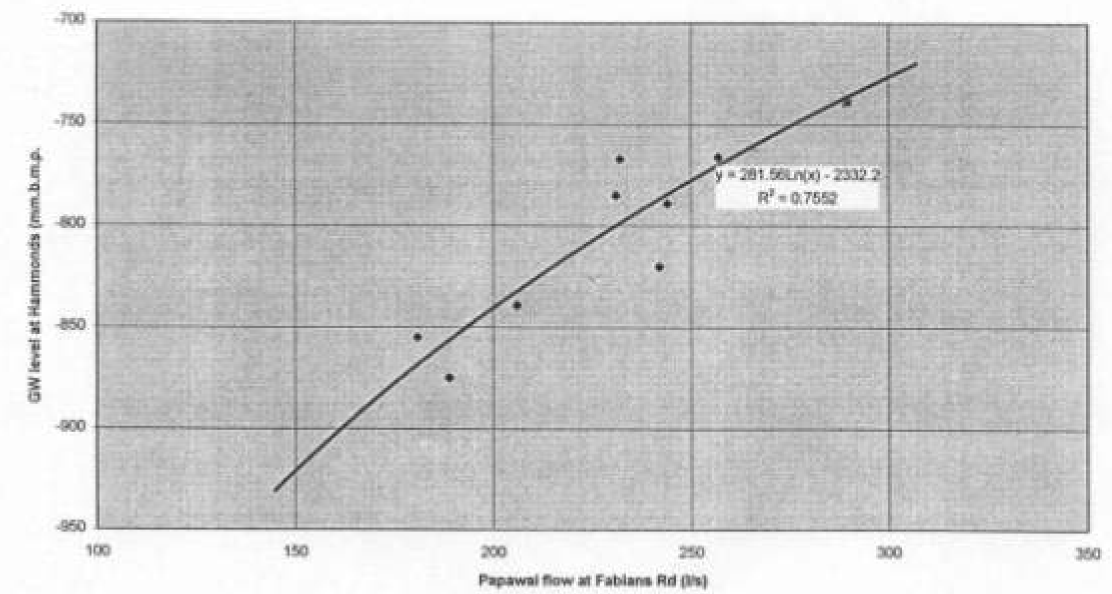


Figure A2 Correlations between Groundwater Level and Stream Flow

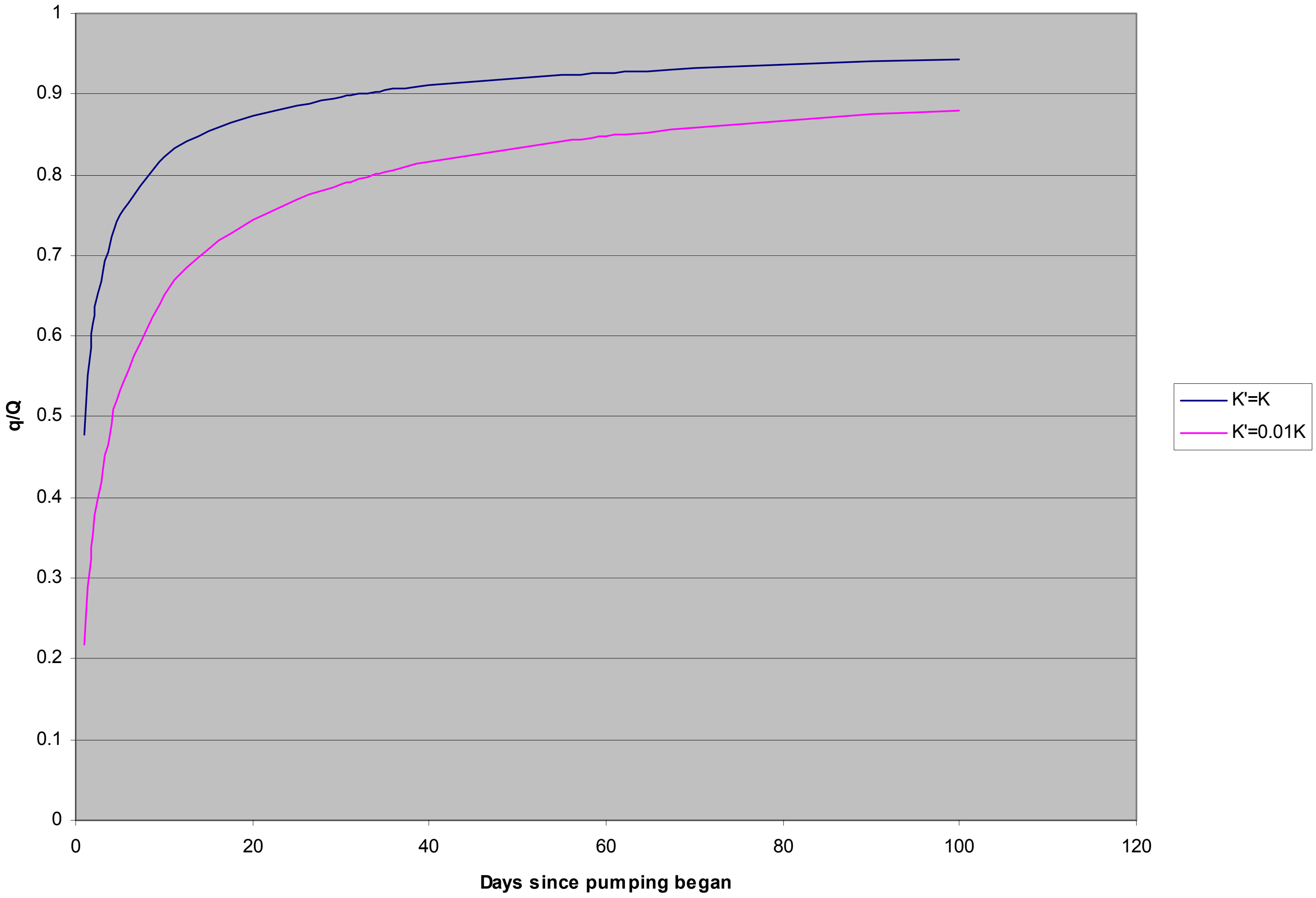


Figure A3: Calculated stream depletion effect

Appendix B:
Case Study B – Little Sydney Stream, Tasman District Council

B1. Available Information to Set the Scene

At Riwaka, near Motueka, Tasman District Council (TDC) have assessed the effects of seven irrigation abstraction wells (reference numbers WWD3286, WWD3292, WWD3346, WWD3290, WWD3291, WWD3402, WWD3138) located near the Little Sydney Stream. TDC have provided the results of the gaugings and details regarding the stream and the abstraction wells, plus a file note summarising their findings.

The Little Sydney Stream is a small stream in the Riwaka Valley (Figure B1). For several years there have been complaints that the stream has been going dry, especially along the Swamp Road and Factory Road areas in the drier summers. Irrigation abstraction has been considered a potential cause of this effect. While surface water abstractors would obviously have a direct effect on stream flows, the issue of whether adjacent groundwater users were also having an impact on flows in the stream was raised and investigated by a flow gauging survey, the results of which are summarised below.

Two series of gaugings were undertaken along the Little Sydney Stream (Figure B2) to investigate the effect groundwater users have on the stream as follows (refer to Figure B1 for gauging locations):

- Midway through the irrigation season (19/1/98), all irrigators ceased pumping for 24 hours. During this period of no abstraction the first gauging survey was carried out.
- On the following day (20/1/98), only groundwater abstractions took place and a second gauging survey was undertaken.

Stream flows during the gauging period varied from 9 to 21 L/s. The combined permitted peak rate of abstraction from the seven wells is around 100 L/s and the average combined weekly permitted rate is around 45 L/s. The permitted surface water peak abstraction rate from the stream is 108 L/s and the average weekly permitted rate is only around 2.6 L/s. Under the provisions of the Operative Motueka/Riwaka Plains Water Management Plan (1995), the allocation limit for surface water takes from the Little Sydney Stream is 29 L/s (calculated from the sum of weekly permit allocation). There is also a trigger for rationing for the Little Sydney Stream as well as a reduction in usage provision for the Little Sydney surface water users. The first trigger occurs where a flow of 60 L/s is measured at the State Highway 60 bridge north of Riwaka. The second trigger occurs at a flow of 15 L/s.

The wells adjacent to the stream are typically around 10 m deep and are screened in a sandy gravel aquifer which varies from 0 to 15 m thick. Some of the well logs indicate the presence of a clay layer between the aquifer strata and the ground surface, between 0.3 to 2.8 m thick. The closest wells are located as near as 20 m from the stream. The TDC staff expect that the aquifer transmissivity in the aquifer is around 500 – 800 m^2/day and storativity is expected to be around 0.001. Depth to groundwater has been measured on two occasions (in 1983 and 1985) in three wells in the vicinity of Little Sydney Stream (wells WWD3291, WWD3290 and WWD3135) varying from 0.25 to 2.53 m below ground level. Field observations indicate that the stream bed has a low conductivity (pers comm J Thomas, TDC).

The results of the gauging data comparing groundwater pumping effects during the period of no surface water abstraction are plotted in Figure B2. With sites 1 – 6 representing the flow changes from upstream (site 1) to downstream (site 6). On the day where no pumping occurred there was a slow decline in flows from upstream to downstream sites. On the day where only groundwater abstractions occurred , the river flow actually increased from site 1 – 3 and then declined to site 4, stabilised and then dropped again before site 6. TDC have suggested that the increase in flow from site 1 to site 3 may be attributed to runoff into the stream from the areas that were being irrigated.

On the basis of this information the TDC file note stated that *'The results indicate that the short term pumping of groundwater seems to not have a significant effect on the stream flow. It is pointed out that long term pumping of groundwater (i.e. continuously) may still have an effect. More details would be required to assess the long term impact.'*

B2. Analysis

In this setting there is uncertainty regarding the nature of the interaction between the groundwater and surface flow and the streambed conductance. This situation could be clarified by a combined gauging and piezometric survey as described in section 6.1.1.

The available stream gauging data is highly variable, and, with reference to Figure B2, the only change that could be attributed to stream depletion is the reduction in flow that occurs between sites 3 and 4. In this location the largest groundwater abstraction occurs from well 3346, which holds a consent to pump at a rate of 36 L/s for 3 days out of 7 (a long-term average rate of 15.4 L/s).

The analytical method for calculating stream depletion effects (section 4.1), can be applied to this situation using the following parameters:

Transmissivitiy of 800 m^2/day

a separation distance from the stream of 20 m

Storage coefficient of 0.001

These parameters define a stream depletion factor of 5 x 10^{-4} days which indicates that stream depletion effects will be likely if the streambed has a sufficiently high conductance (>0.01 m/day).

The description of stream bed clogging could be represented by assuming that the hydraulic conductivity of the stream bed is around 1,000 – 10,000 times less than the aquifer, i.e. K' = 0.08 – 0.008 m/day (based on K = 80 m/day for a 10 m thick aquifer). TDC plans indicate that along Swamp Road (between sites 3 and 4), the stream bed is around 4 m wide and is incised by around 2 m. This indicates that the low permeability layer (varying from 0.3 – 2.8 m thick), may be around 0.5 m thick beneath the stream bed. Using these values gives a stream bed conductance (λ) of 0.064 – 0.64 m/day.

To provide an indicative reproduction of the situation that was monitored by the TDC it was simulated that well 3346 pumped at its average long-term rate of 15.4 L/s for 60 days, stopped for 1 day and then re-commenced pumping for 1 day at its peak rate of 36 L/s. The curves for the two different stream bed conductance scenarios are plotted in Figure A3. They suggest that the 24 hour shut off and recommencement of pumping could have caused the variations in stream flow that are presented in Table B1, compared to the measured variation in flow.

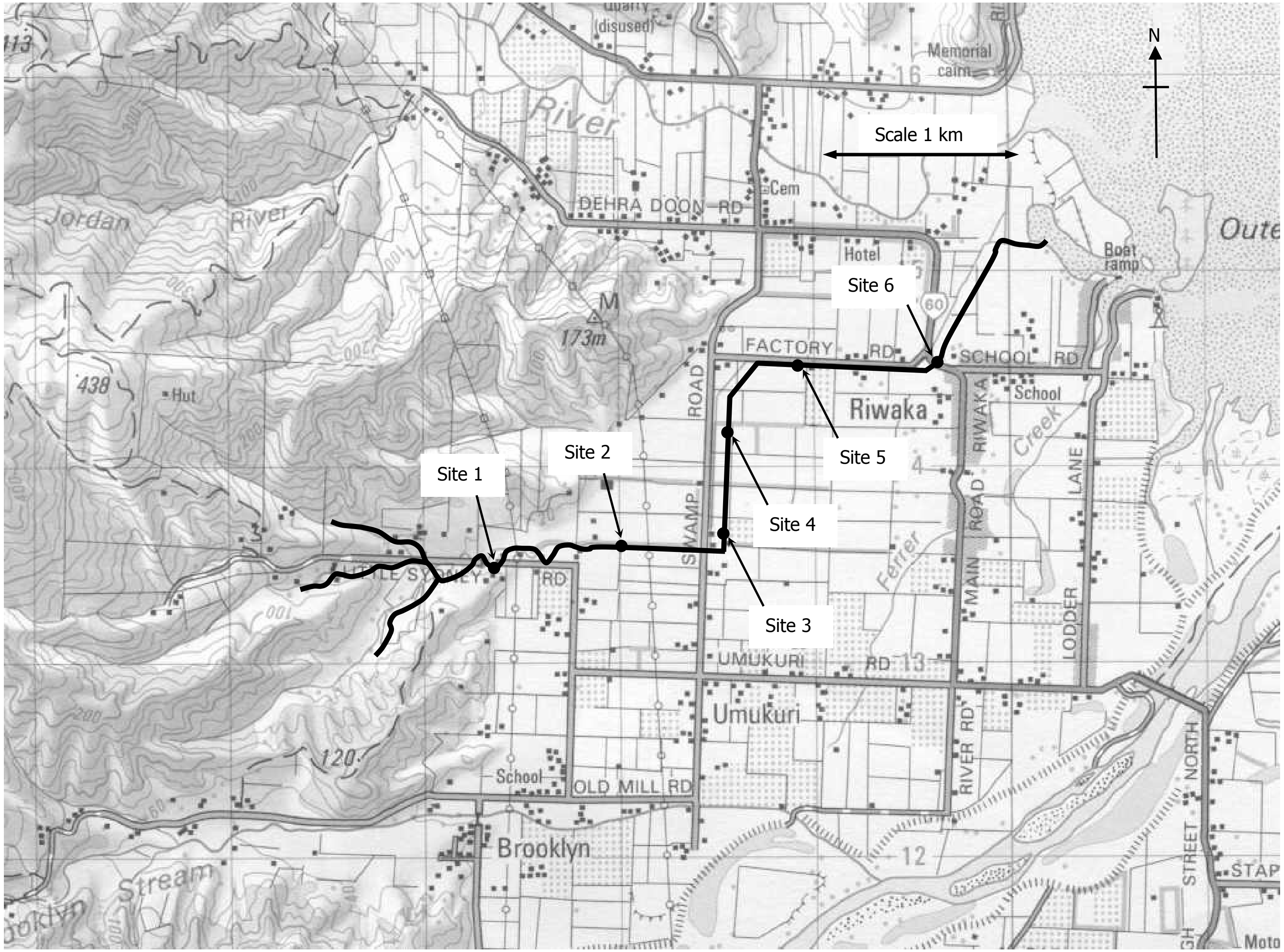
Table B1: Little Sydney Stream Assessment

Date	Status	Measured Changes in Stream Flow between Sites 3 and 4	Calculated Change in Stream Depletion Rate Caused by Well 3346	
			K' = 0.08 m/day	K' = 0.008 m/day
18/1/98	All pumping ceases for 1 day	- 3 L/s	- 5.1 L/s	- 0.5 L/s
19/1/98	Groundwater pumping recommences for 1 day	- 7 L/s	- 9.2 L/s	- 1.1 L/s

The simulation for K' = 0.08 m/day shows a change in stream flow of 4.1 L/s between 18 and 19 January which closely matches the observed change in flow of 4 L/s. The results suggest that groundwater abstraction could be a contributing factor to the variations in stream flow measured by the TDC. However, the overall variation in measurements is too large to draw any definite conclusion. This inconclusive result again highlights the uncertainty that typically occurs when stream depletion is assessed through gauging surveys, as discussed in section 6.2.1.

The calculations listed above show the significance of the stream bed conductance value for the combination of parameters that have been used in this example. Site specific field measurements of the stream bed conductance using the techniques described in section 6.1 of these guidelines would enhance the understanding of stream depletion effects at this site.

TDC have reported that recent numerical modelling of the Motueka Plain has suggested that groundwater abstractions have little effect on stream flow in this area. It is beyond the scope of these guidelines to review this numerical model, however, it is worth noting that the scale of a regional model (both in terms of grid size and time step) often fails to provide the detail necessary for the type of individual stream depletion assessments that are described in these guidelines.



NZMS 260 Map N26.

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Figure B1 Location Map for Case Study B

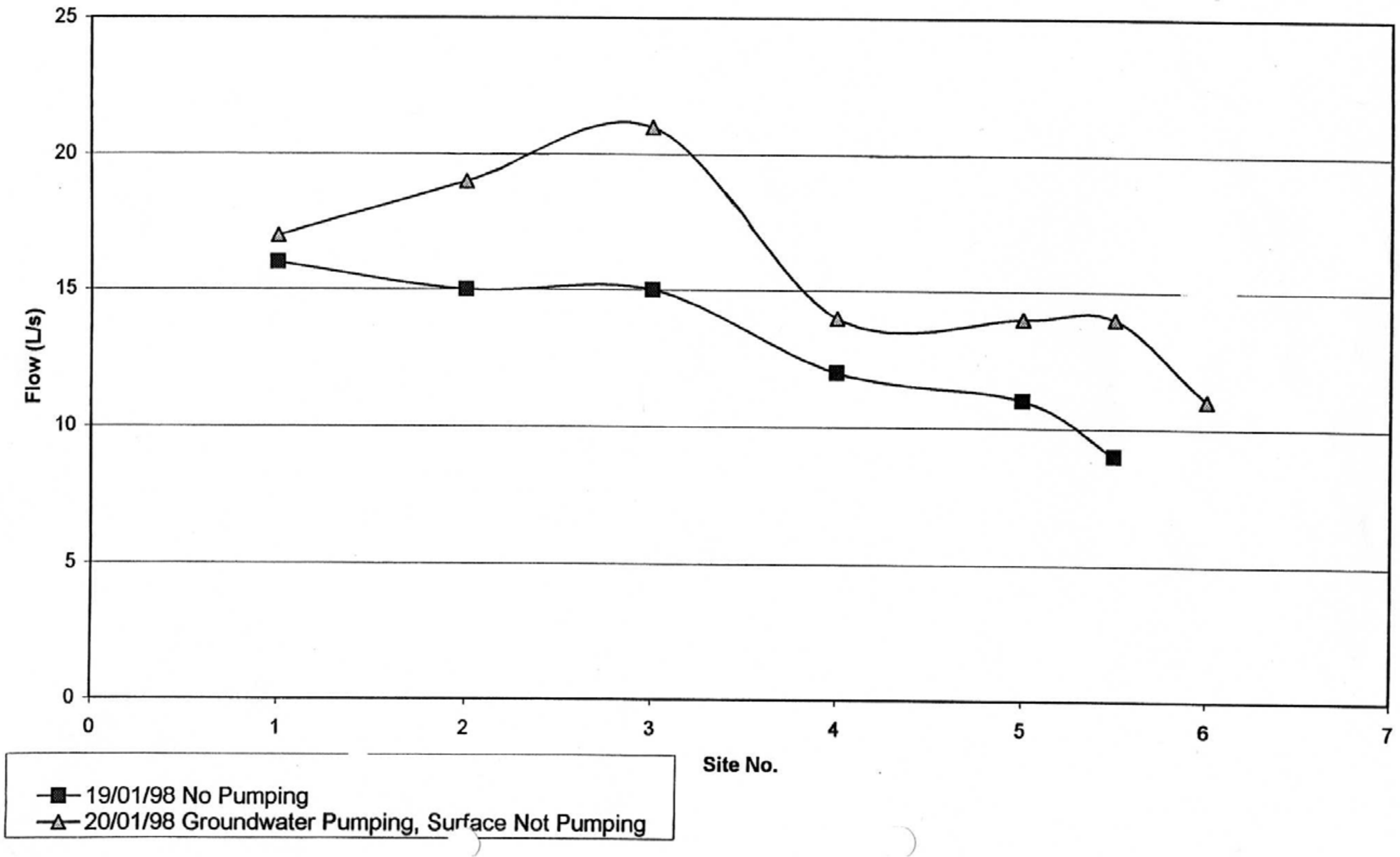

Figure B2. Little Sydney Gauging Results 1998

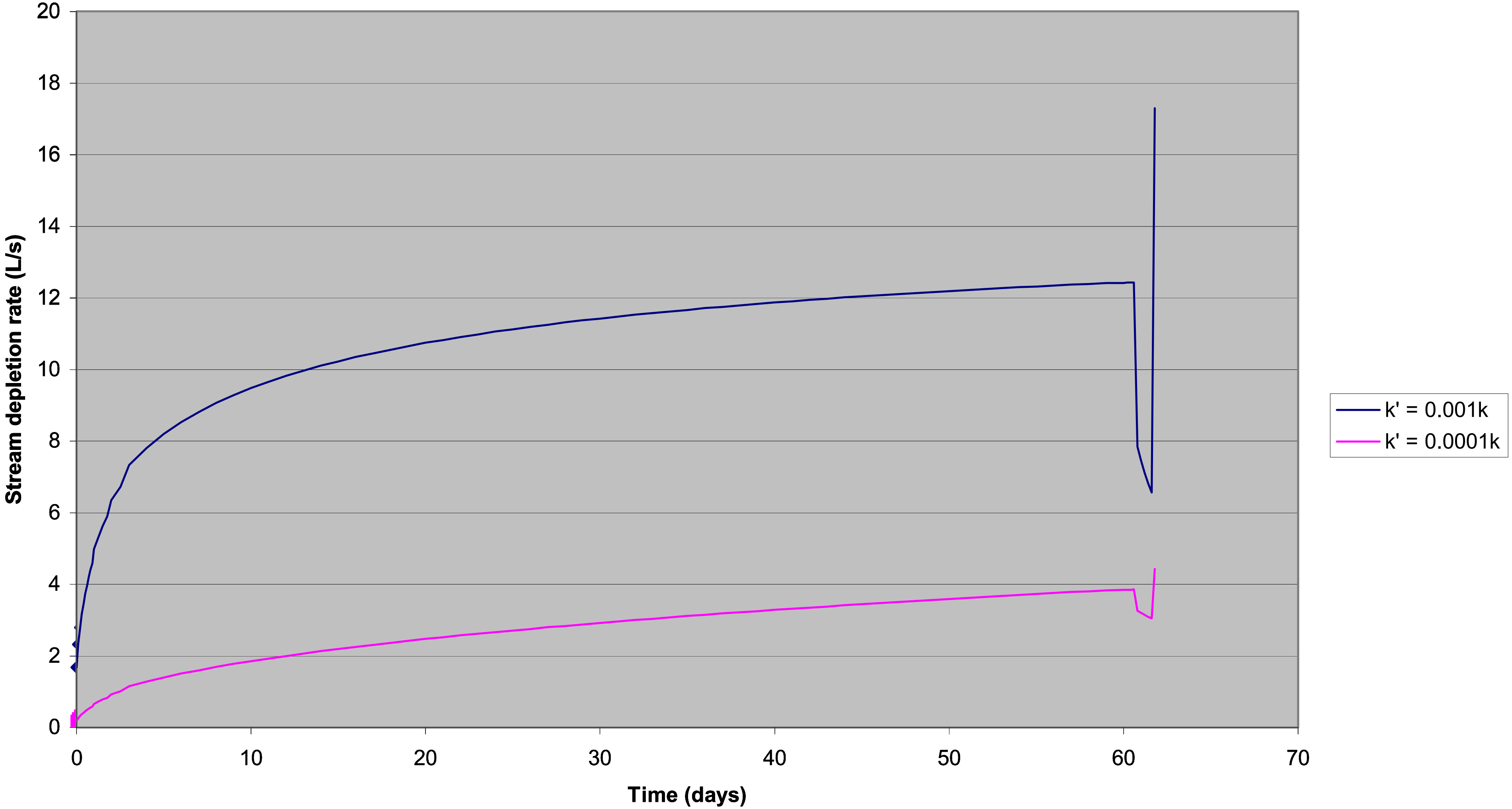

Figure B3: Possible stream depletion rate at Little Sydney Stream

Appendix C:
Case Study C – Ohoka Stream, Environment Canterbury

In Ohoka, Canterbury, Environment Canterbury and Pattle Delamore Partners Ltd have undertaken a number of simple field measurements in the Ohoka Stream, to assess the conductance of the stream bed, as described in PDP (1999).

These measurements were taken to allow an assessment of the effects of a number of irrigation abstraction wells on the Ohoka Stream.

The Ohoka Stream is one of many spring fed streams that emerge towards the coastal margin of the Canterbury Plains. Their emergence is associated with the western extent of low permeability estuarine and marine deposits which overlie the permeable gravel aquifers found over the Plains. The uppermost aquifer is unconfined to the west of these surficial fine grained sediments, and to the east the aquifer is confined by these overlying low permeability strata.

Concerns regarding the effect of low stream flows on the aquatic habitat of the stream motivated Environment Canterbury to place conditions which restrict abstractions at times of low flow, on both surface water abstractions from the stream and also on groundwater abstractions adjacent to the stream. Minimum flow restrictions for the Ohoka Stream occur at a flow of 300 L/s for "A" permits and at a flow of 800 L/s for "B" permits as measured at the confluence with the Kaiapoi River. During the testing period the flow in the Ohoka Stream was monitored at 260 L/s.

Stream bed conductance measurements were made to allow a better prediction of the magnitude of stream depletion effect caused by groundwater irrigators.

The surface aquifer connected to the stream occurs from around 3 to 24 m depth and water levels range from 0.04 to 7 m below ground level.

Three types of tests were carried out at different points along the streambed: (1) Ring infiltrometer tests; (2) a seepage meter test; and (3) a stream gauging and groundwater piezometric survey. The methods used for these tests are described in section 6.1. The results of the tests are presented in Table C1. They show that hydraulic conductivity of the Ohoka Streambed varied between 4×10^{-6} m/s and 3×10^{-5} m/s.

Table C1: Results of Streambed Conductance Tests

Test	Field Measurement	Calculated Coefficients (iA)	Hydraulic Conductivity = Q/iA (m/s)
Single Ring Tests	Flow to maintain constant head in ring Q (m^3/s)	Calculated from flow net analysis	
Mill Rd 1	7.2×10^{-7}	0.060	1×10^{-5}
Mill Rd 2	4.4×10^{-7}	0.033	1×10^{-5}
Mill Rd 3	3.8×10^{-7}	0.021	2×10^{-5}
Bradleys Rd 1	5.3×10^{-7}	0.028	2×10^{-5}
Bradleys Rd 2	2.3×10^{-7}	0.022	1×10^{-5}
Seepage Meter	Flow out of Seepage Meter (m3/s)	Calculated from piezometric survey	
Mill Rd	3.82×10^{-7}	0.0914	4×10^{-6}

Stream	Conductance $\Delta q/L\Delta h$ = $K^{l}W/M$ (m/s)	Width of Stream "W" (m)	Estimated Streambed thickness (m)	Hydraulic Conductivity "K" (m/s)
Gauging Survey				
Ohoka Stream	1.1×10^{-5}	2.7	1.0	4×10^{-6}
South Branch	2.6×10^{-5}	1.0	1.0	3×10^{-5}

The above values give an indication of the vertical hydraulic conductivity of the Ohoka Streambed. They can be compared with an aquifer hydraulic conductivity measurement in the area of approximately 1.4×10^{-3} (Moore 1995), a value that represents the horizontal hydraulic conductivity of the aquifer. Consequently the streambed hydraulic conductivity is around 100 to 1,000 times less than the aquifer.

The analytical method described in section 4.1 was used to estimate the amount of stream depletion caused by a well 15.4 m deep located 30 m from the Ohoka Stream, pumping at a rate of 26 litres per second.

For this example the following values were used:

Q = 26 L/s = 0.026 m^3/s (from ECan resource consent database)

S = 0.1 (assumed)

ℓ = 30 m (from ECan database)

T = 0.016 m^2/s (Moore 1995), with an aquifer thickness of 11.4 m

t = 30 days = 2,592,000 s

λ = 9×10^{-6} m/s (based on K' = 1×10^{-5} m/s, stream width of 0.9 m and stream bed thickness of 1 m) = 0.8 m/day

These values give a stream depletion factor of 0.065 days.

The stream depletion rate after pumping for 30 days calculated using the Hunt equation is equal to 4.5 L/s or 17% of the well pumping rate.

At this site, there is some uncertainty regarding the appropriate storage coefficient to be used. Consequently a range of storage coefficients down to 0.001 has been considered. The stream depletion rate calculated after 30 days pumping where the storage coefficient is 0.001 would be 73%. Because of the sensitivity of the solution at this site to the storage coefficient it is desirable that a pumping test (or tests) be undertaken so as to better characterise this parameter.

Appendix D:
Case Study D – Doyleston Drain, Environment Canterbury

In 1998 Environment Canterbury supported a ME research project (Weir, 1999) which involved a detailed pumping test adjacent to a springfed drain. The Doyleston Drain was excavated to drain farmland around Lake Ellesmere and is a linear feature with a 2.5 m wide gravel bed and near vertical grassy banks (Figure D1).

Driller's logs from the area indicate the presence of a surface low permeability silt/clay layer around 1.7 m thick overlying a sandy gravel layer to a depth of 2.8 m which, in turn, overlies a permeable gravel aquifer approximately 20 m thick. A pumping well (200 mm diameter, screened from 8.15 – 10.15 m deep) was installed 55 m from the drain, as shown in the cross-section in Figure D2.

To test for stream depletion effects a network of observation wells was established around a pumped well and a stream, as shown in Figure D3.

The well was pumped at a rate 17.5 L/s for 10 hours whilst water levels were measured in the pumped well and in nearby observation wells. In addition, stream flow was measured at weirs located 200 m upstream and 200 m downstream of the pumped well position.

The resulting drawdown from two observation wells located 25 m either side of the pumped well is plotted in Figure D4 and indicates the recharging effect of stream seepage (borehole 4 is located between the pumped well and the stream whereas borehole 2 is located on the opposite side of the pumped well). Measurements of stream flow are plotted in Figure D5. The stream flow loss between the two weirs measured during the pumping period was 1.2 L/s. In addition, the upstream weir showed a reduction in stream flow during the test of 4.8 L/s, resulting in a total estimated depletion rate for the test of 10.8 L/s ((4.8 x 2) + 1.2).

The test was analysed using these field measurements and the methods described in section 6.1.4 of this guideline. In addition, estimates of streambed conductance were made using gauging surveys (section 6.1.3) and infiltration tests (section 6.1.1). A summary of the test results are presented in Table D1. They show generally good agreement.

These detailed measurements on the Doyleston Drain reflect the magnitude of stream depletion effects that can occur and indicate that the methods of assessment described in sections 4.1 and 6.1 of this guideline are valid techniques to estimate the effect.

Table D1: Results

	T **m^2/day**	**S**	**sdf** **(days)**	λ **m/day**
Drawdown (match point) - Constant pumping - Constant recovery - Step	 1470 2680 2420	 0.00195 - 0.00169	 4×10^{-3} - 2×10^{-3}	 50.3 2.7 16.9
Stream Depletion Match Point	2070	0.00161	2×10^{-3}	7.1
Stream Gauging Survey	-	-	-	35.6
Infiltration Measurements	-	-	-	40.8
Geometric Mean	2070	0.0017	2×10^{-3}	16.9

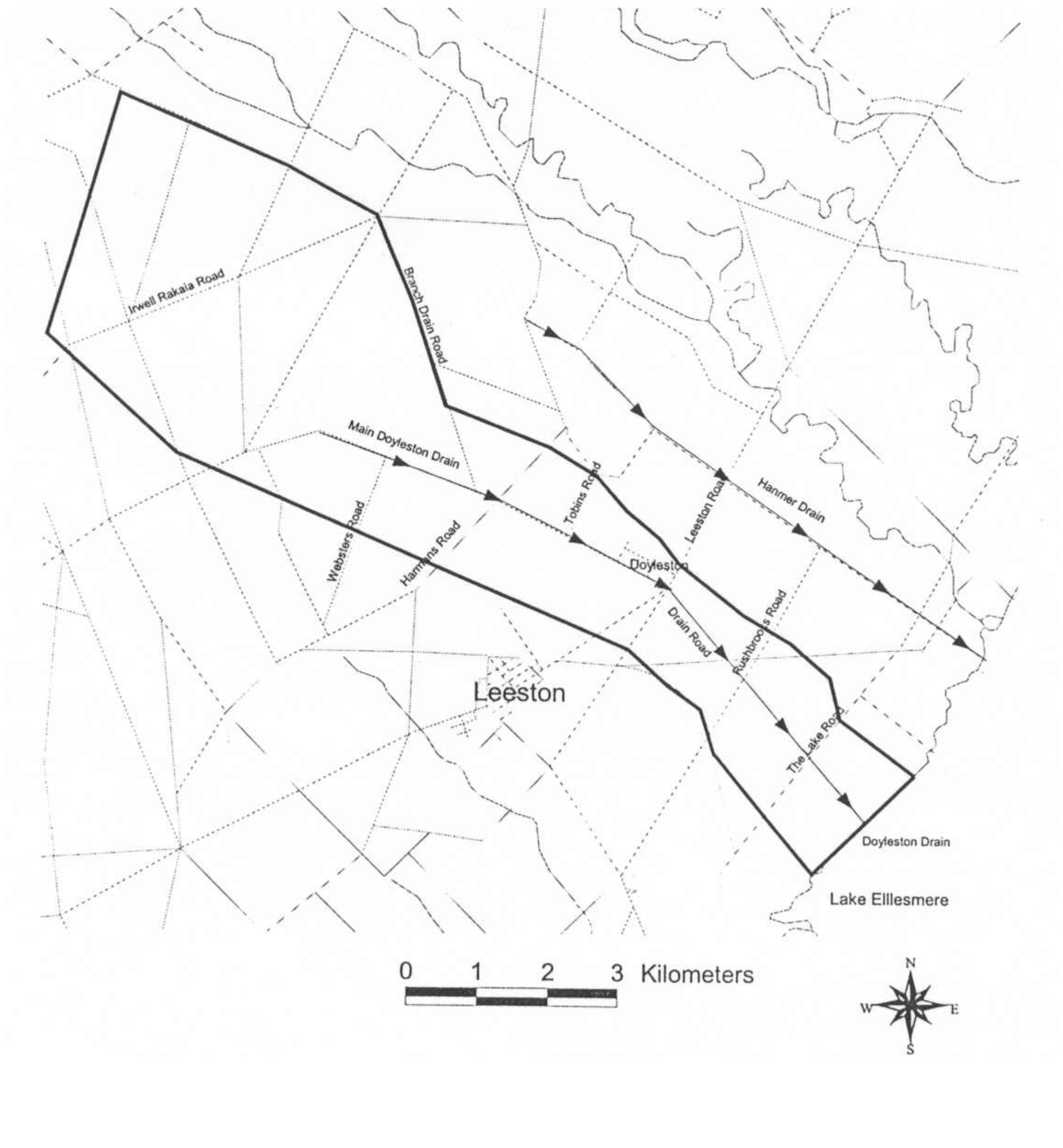

Figure D1: Location Map for Case Study D

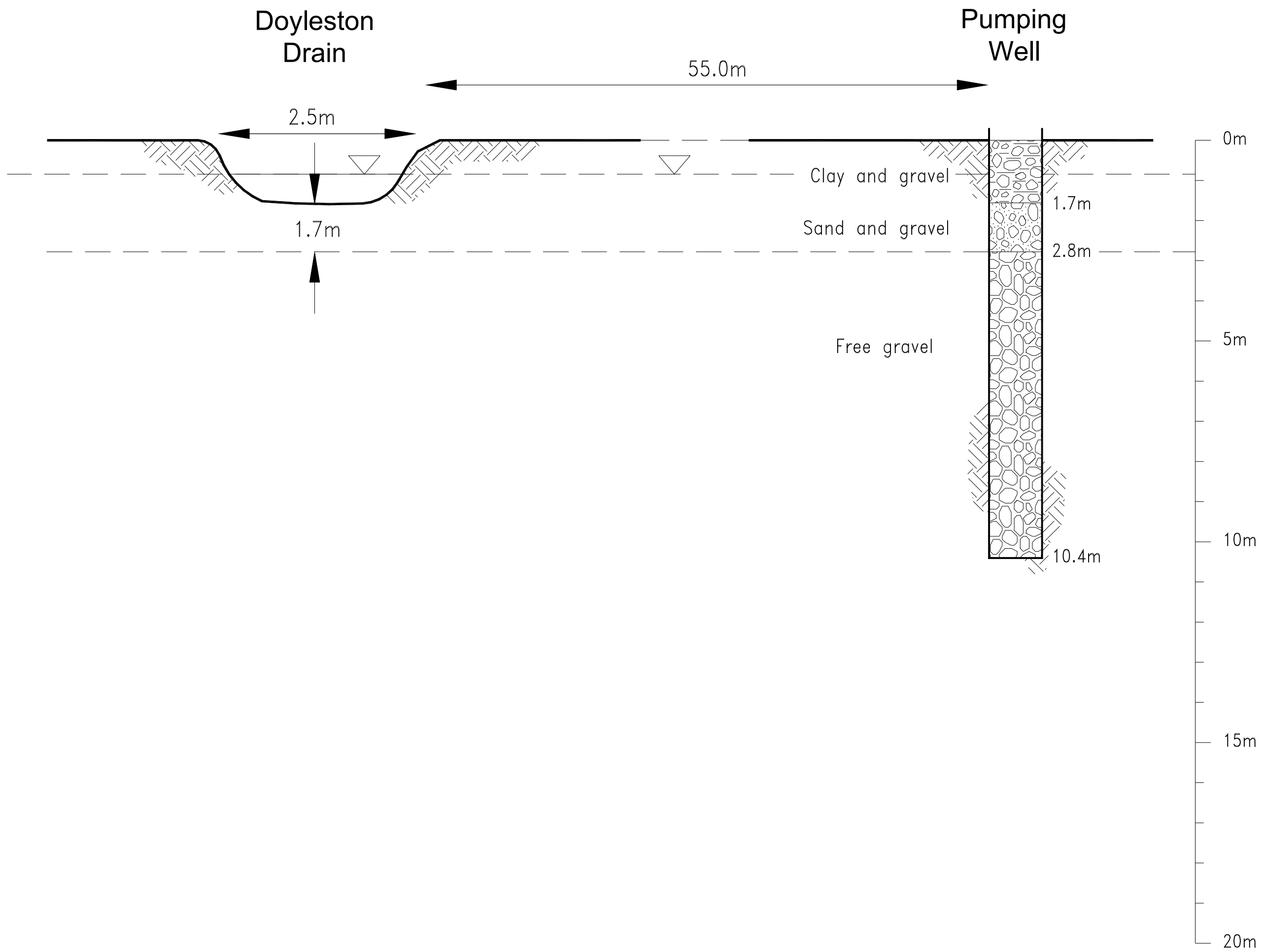

Figure D2: Conceptual Layout of Doyleston Drain Test Site

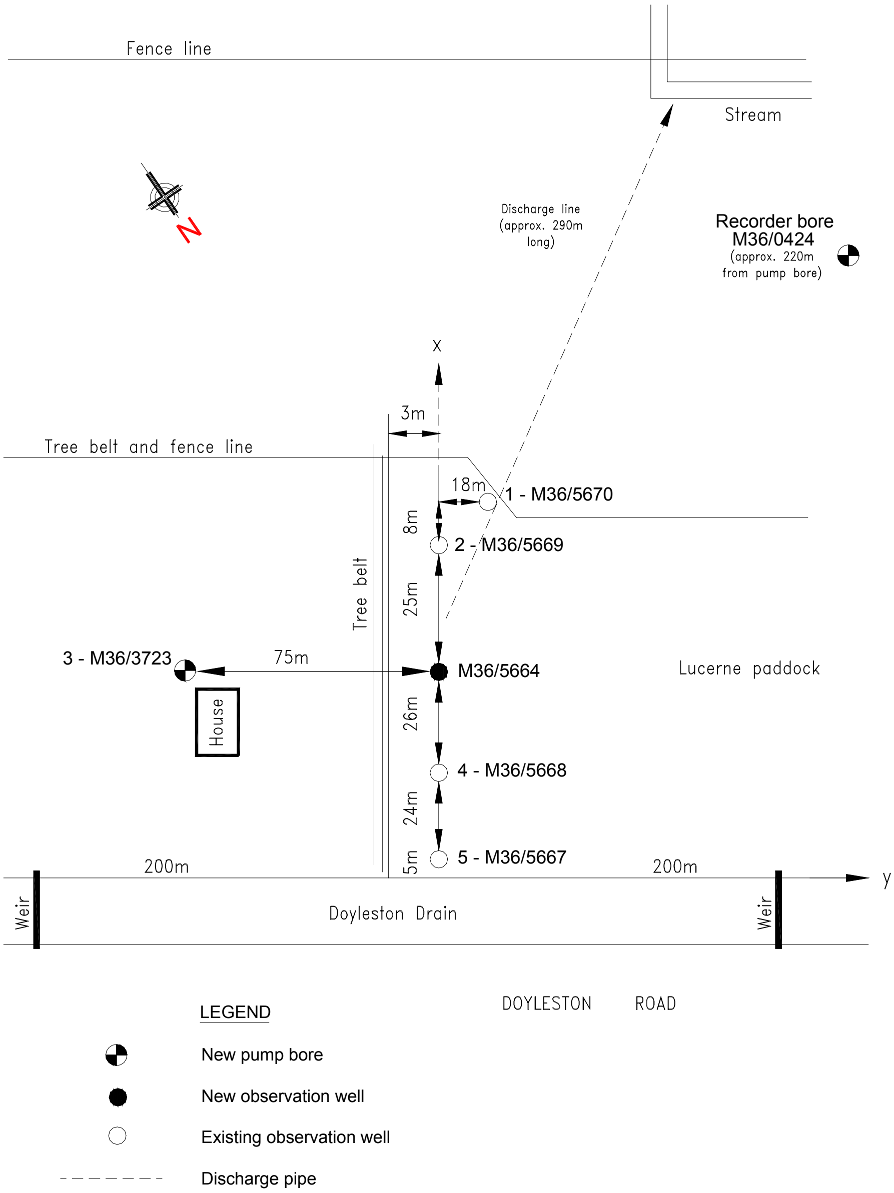

Figure D3: Schematic Layout of Test Site

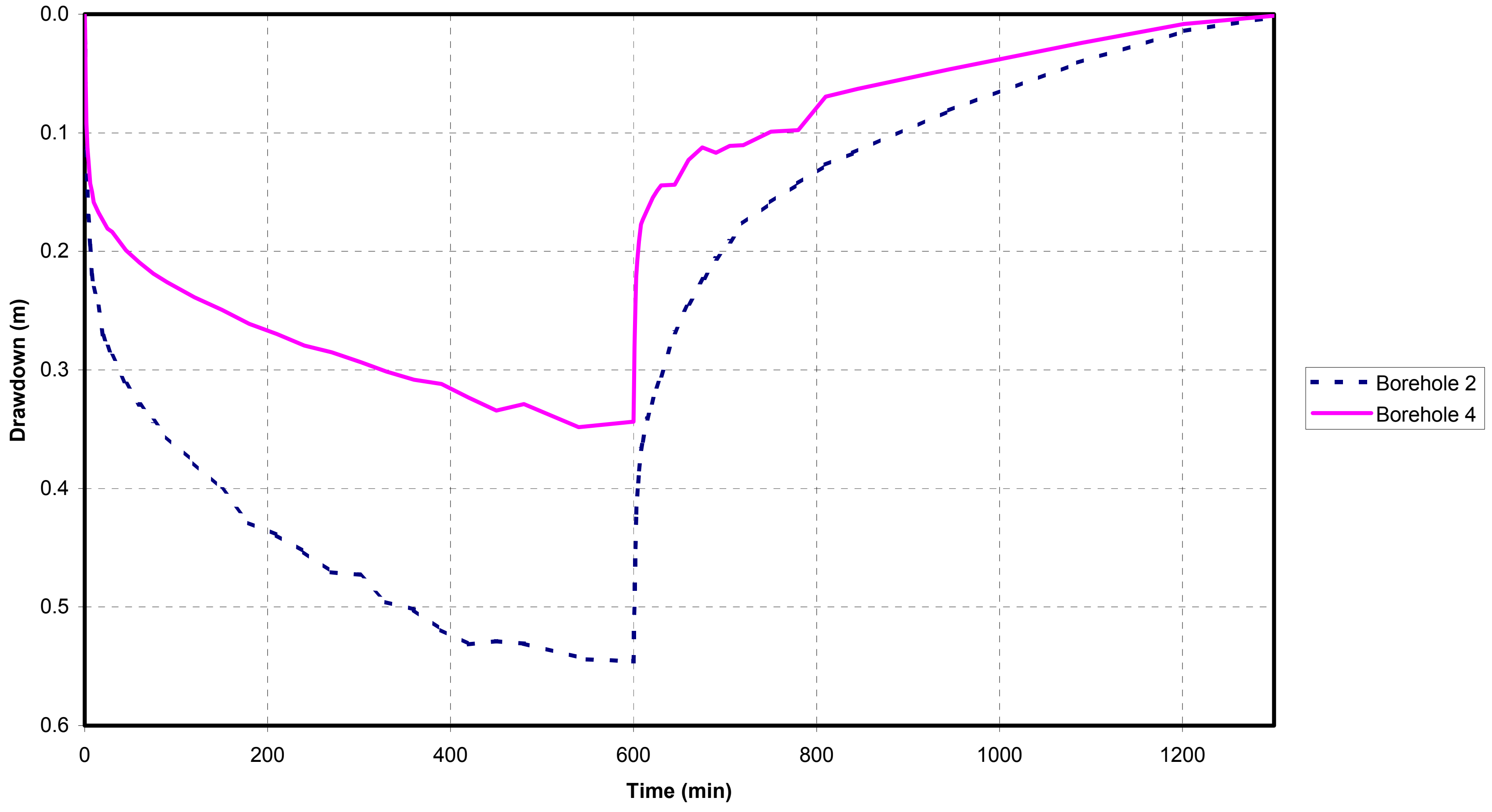

Figure D4: Drawdown in Observation Wells

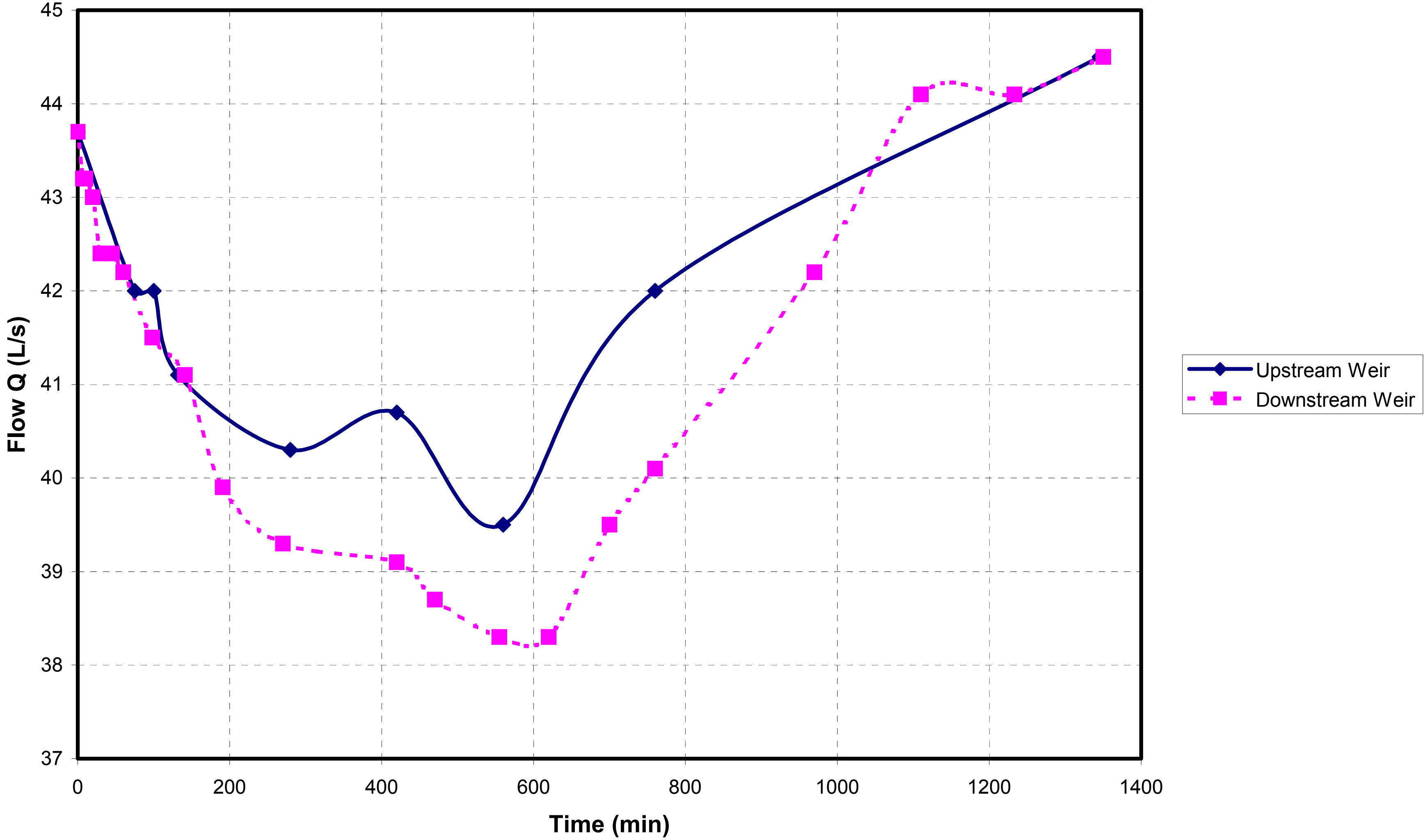

Figure D5: Inferred Change in Stream Flow

Quantifying stream depletion effects

In almost all surface water settings there is movement of water between the underlying groundwater and the surface water body. This occurs in streams, lakes, wetlands, estuaries and at the sea coast. Because of this interaction there are some situations in which wells pumping from permeable aquifers can reduce the flow or volume of water in the surface water way – this is called a "stream depletion" effect.

With the increasing demand on water resources, it is becoming important to understand and assess the stream depletion effect caused by pumping from groundwater.

The interactions of water movement between groundwater and surface water are difficult to observe and measure. This creates uncertainty regarding the magnitude of any surface water/groundwater interaction, the implications of the effect and an appropriate form of management. To reduce the uncertainty, this note has been prepared by Pattle Delamore Partners Ltd as part of an Environment Canterbury project to help in quantifying the stream depletion effect. This project was supported by the Ministry for the Environment's Sustainable Management Fund.

This note describes the following:

Part A a general screening process to identify settings where stream depletion may occur.

Part B an outline of the parameters needed to calculate stream depletion effects and a graph to allow a preliminary estimate to quantify the effect.

Part C an indication of field measurements that can be used to improve the accuracy of stream depletion assessments.

Part A comprises consideration of simple observations that can be evaluated by all water users. Parts B and C describe more complex issues that should only be considered with a technical adviser.

This note is based on a more comprehensive technical guideline which is available from Environment Canterbury, P O Box 345, Christchurch (ph 03-365-3828) or from Pattle Delamore Partners Ltd, P O Box 389, Christchurch (ph 03-379-3532).

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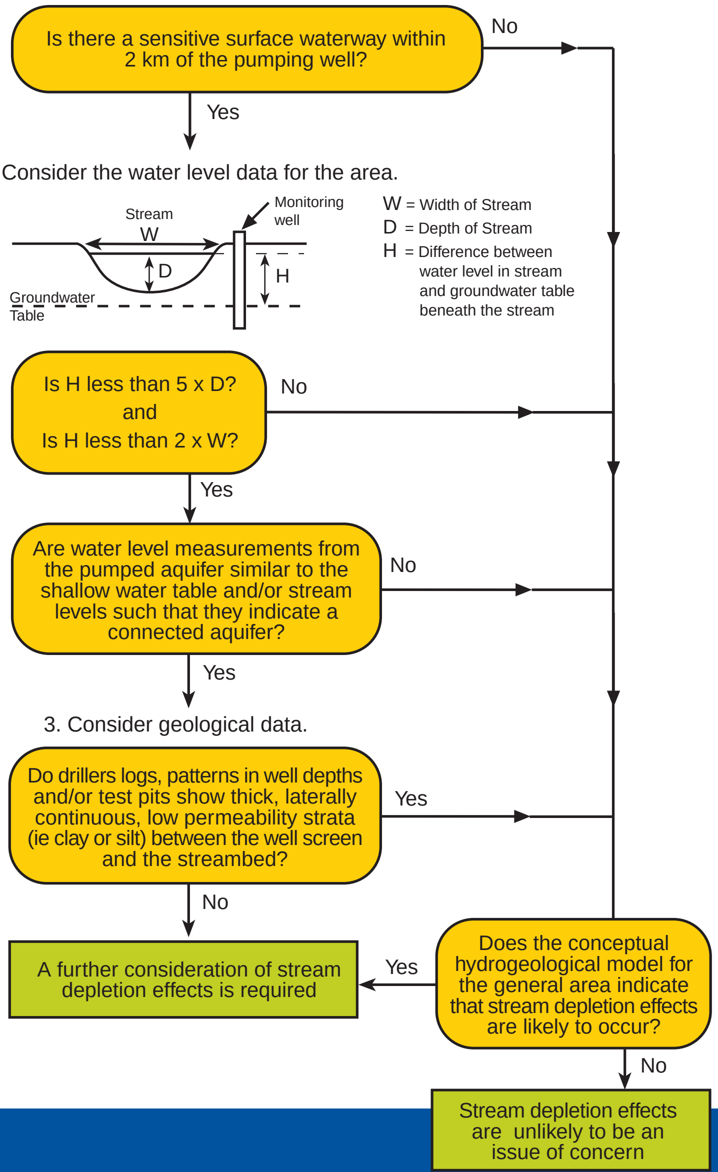
Part A – Is stream depletion a concern?

For any particular site, a review of information on well locations, water levels and borehole strata records can be used to answer the questions in the flow chart below. This helps determine whether stream depletion is likely to be an issue of concern.

Is pumping from a well likely to have a stream depletion effect?

1. Consider the location of the well in relation to nearby surface waterways



Part B – Calculating the stream depletion effect

The key parameters that are used to calculate the magnitude of stream depletion effects are listed below:

- Q the pumping rate from the well (m^3/day).
- ℓ the separation distance between the well and the stream (m).
- t the duration of time that the well is pumped (days).
- T the transmissivity of the aquifer – an indication of permeability. Values of T in productive aquifers typically range from 10 – 10,000 m^2/day.
- S the storage coefficient of the aquifer – an indication of how much water is stored in the strata. Where stream depletion is an issue, values of S typically range from 0.005 – 0.3.
- λ the hydraulic conductance of the streambed which is calculated by:

$$\lambda = \frac{\text{vertical hydralic conductance of the stream bed} \times \text{width of streambed}}{\text{thickness of streambed}}$$

where stream depletion is an issue, values of λ typically range from 0.1 – 5,000 m/day.

The values of these parameters can be used in equations to indicate the effect of groundwater pumping on nearby surface waterways.

Step 1. Consider the Stream Depletion Factor

An indicator of the aquifer conditions and separation distance to the well is provided by the "stream depletion factor"

$$sdf = \frac{\ell^2 \times S}{T}$$

If the value of sdf is greater than 100 days, then stream depletion effects are likely to develop very slowly, and not be large.

Step 2. Consider the Streambed Conductance

An indicator of the streambed conductance is the parameter, λ, described above. If λ is less than 0.01 m/day then the streambed is likely to have a sufficiently low hydraulic conductivity so that no large significant depletion effect could be caused by a pumping well.

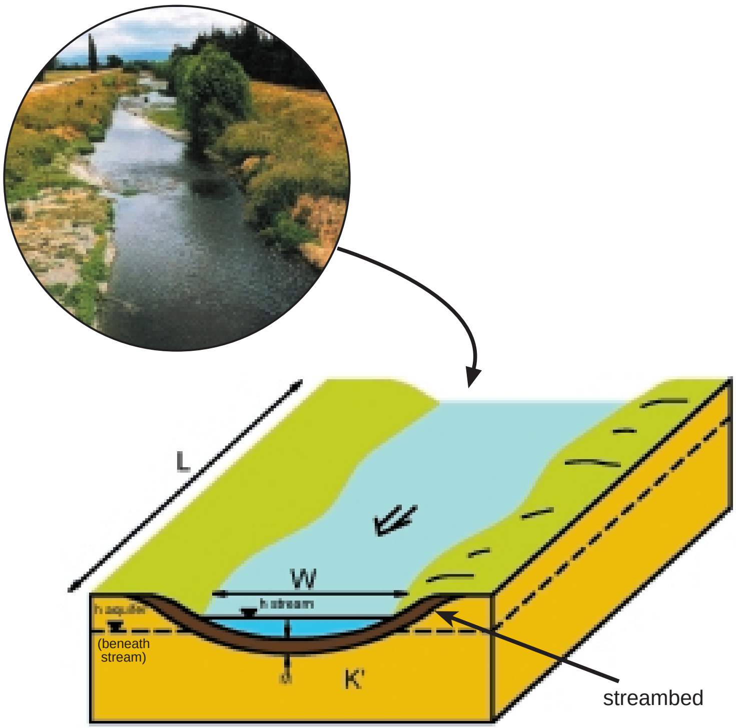
Step 3. Estimating the Stream Depletion Effect

For simple settings, where a surface waterway can be represented as a straight line passing through an adjacent aquifer, the following parameters can be calculated:

$$\frac{t}{sdf} = \frac{t^2 \times T}{\ell^2 \times S} \quad \text{and “stream bed factor” sbf} = \frac{\lambda \times \ell}{T}$$

The figures on the following pages describe this simple hydrogeological setting. The graph can be used to estimate the stream depletion effect by locating the value of t/sdf on the bottom horizontal axis and the correct sbf curve to read off a value of q/Q – the ratio of the rate of water drawn from the stream to the rate of water pumped from the well.

Some key parameters for calculating stream depletion effects



$$q = -K' \times L \times W \times i = -\lambda \times L \times (h_{aquifer} - h_{stream})$$

q	the flow of water between the stream and the aquifer (m^3/day)
K'	the vertical hydraulic conductivity of the streambed (m/day)
L	the length of the stream reach over which seepage is assessed (m)
W	the width of a stream reach over which seepage is assessed (m)
i	the hydraulic gradient between the stream and the aquifer

$$i = \frac{h_{aquifer} - h_{stream}}{M}$$

λ	the streambed conductance, a measure of the hydraulic conductivity and dimensions of the streambed (m/day)
$h_{aquifer}$	the level of the water in the aquifer (m)
h_{stream}	the level of the water in the stream (m)
M	the thickness of the streambed which has a hydraulic conductivity of K' (m)

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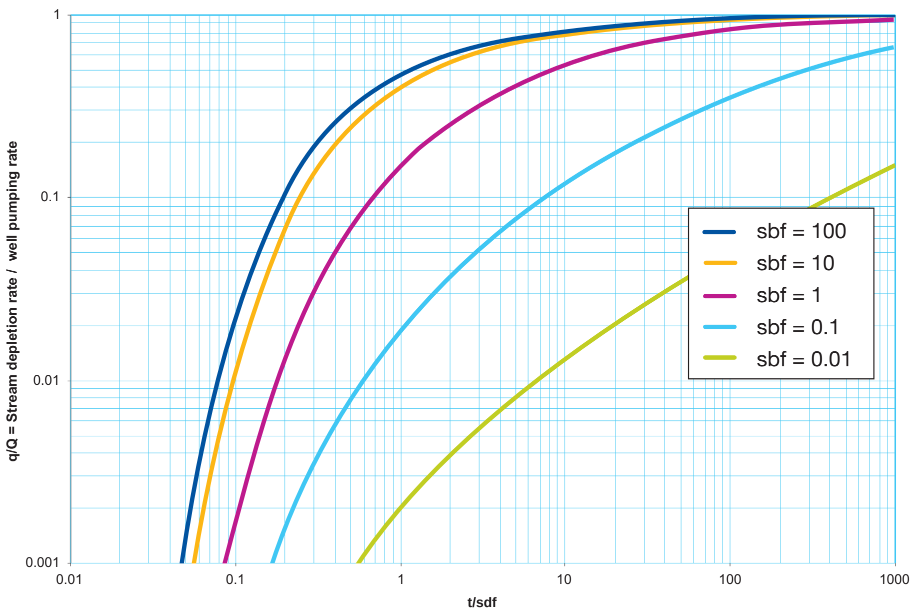
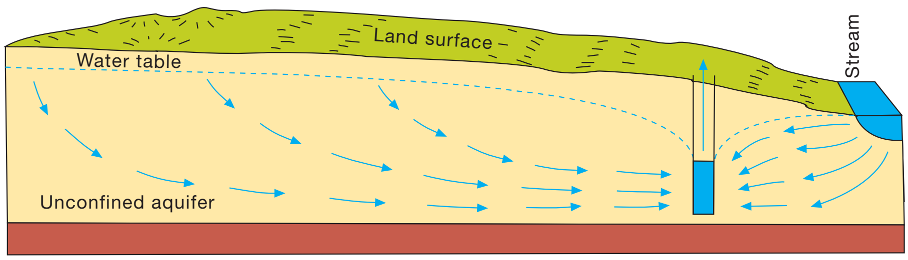
Determining the rate of stream depletion





Schematic setting for the curves shown above

For more complex hydrogeological settings, the reading from this graph may still be useful to indicate the range of stream depletion effects that may occur. However, if a more detailed assessment is required then other techniques such as numerical modelling should be used to better quantify the stream depletion effect. This may involve settings where the aquifer is not laterally extensive due to the presence of low permeability boundaries, multiple streams and/or streams that only flow in a portion of the area affected by a pumping well.

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Part C – Improving the accuracy of stream depletion assessment

Groundwater systems are inherently complex and it is well recognised that attempts to quantify their behaviour requires gross simplification of the natural variability. Assessing the effect of groundwater pumping on surface waterways is no exception. However, useful quantitative assessments of effects can be made by estimating the likely range of parameters that will occur. If these estimates indicate that an improved accuracy is desirable then the following field measurements can be made:

- Improved estimates of streambed conductance can be made by:
  - Gauging and piezometric surveys to assess the natural exchange of water between the surface and subsurface environments.
  - Infiltration measurements in streambed using seepage meters or infiltrometer rings.
- Improved estimates of aquifer parameters (Transmissivity and Storage Coefficient) and aquifer structure can be obtained from carefully controlled pumping tests. The analysis of such tests must allow for the effects of nearby surface waterways.
- Improved estimates of groundwater pumping can be achieved by the installation of flow meters on abstraction wells.

Some other field measurements that have been considered to improve stream depletion estimates but have typically not proven to be successful are:

- Gauging flows in a stream to directly measure the effect of a pumping well. The accuracy of the gauging method compared to the variability in surface flow and the time over which stream depletion effects develop often make this an unreliable assessment technique unless ideal circumstances exist.
- Water chemistry analyses are typically unreliable because stream depletion effects can occur without any surface water actually reaching the pumping well (the effect can occur simply by a change to the hydraulic gradient across the stream bed caused by groundwater pumping).
- Water divining, whilst relied on by many people to site water supply wells, has not been scientifically verified to provide a source of reliable information on stream depletion effects.

As with all groundwater assessments, the carefully controlled measurement of field parameters greatly enhances the conceptual understanding and quantitative assessment of effects.

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