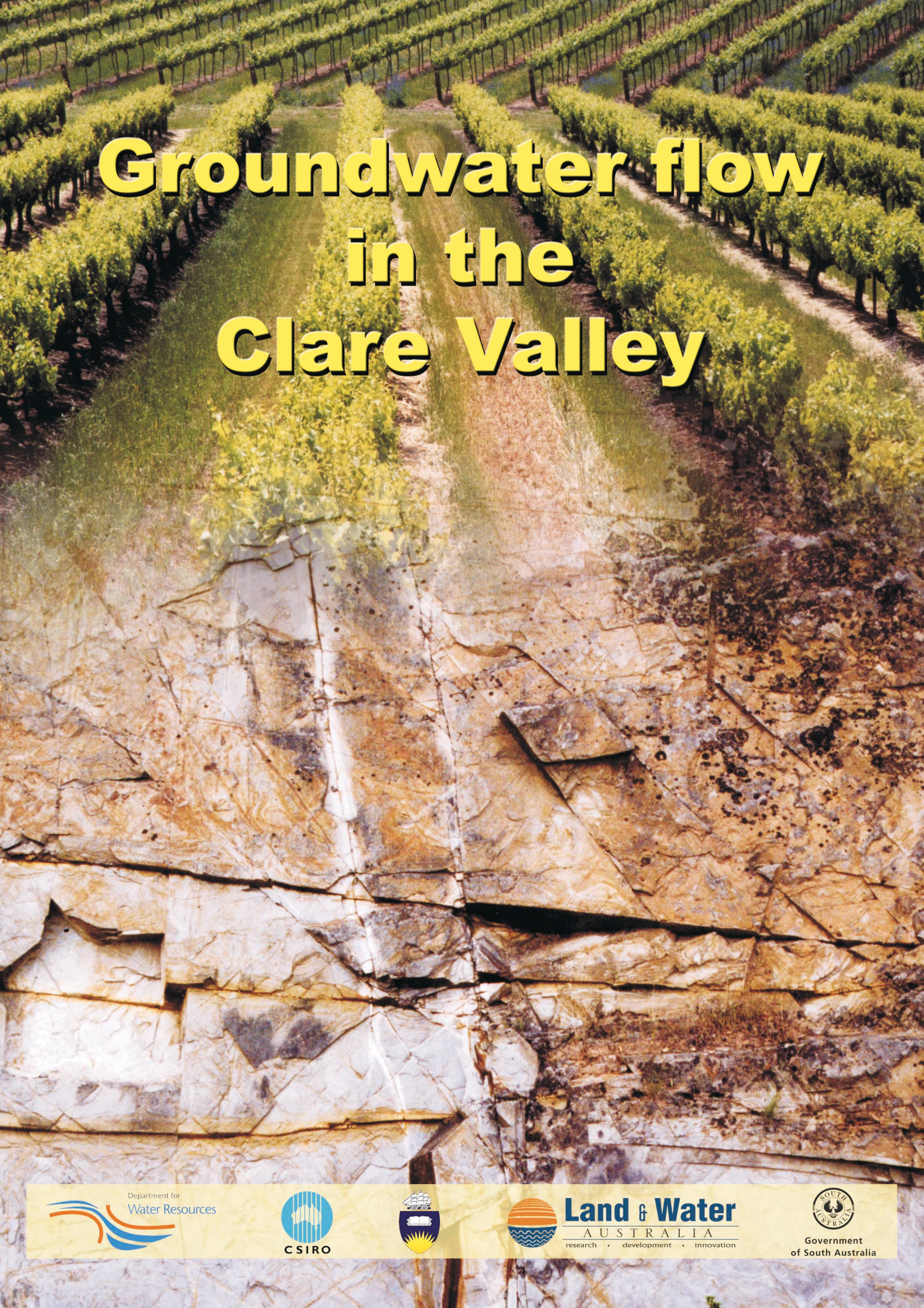
Groundwater flow in the Clare Valley



Department for Water Resources
CSIRO
Land & Water Australia
research • development • innovation
Government of South Australia

DWR02.03.0002

Cover photos

Top: 048371 Vineyard in the Hutt River Catchment.

Bottom: 048375 Saddleworth Formation outcrop near Clare.

Groundwater flow in the Clare Valley

Andrew J. Love
Peter G. Cook
Glenn A. Harrington
Craig T. Simmons

Department for Water Resources
CSIRO
Land & Water Australia
research · development · innovation
Government of South Australia

FOREWORD

South Australia's water resources are fundamental to the economic and social wellbeing of the State. Water resources are an integral part of our natural resources. In pristine or undeveloped situations, the condition of water resources reflects the equilibrium between rainfall, vegetation and other physical parameters. Development of surface and groundwater resources changes the natural balance and causes degradation. If degradation is small, and the resource retains its utility, the community may assess these changes as being acceptable. However, significant stress will impact on the ability of a resource to continue to meet the needs of users and the environment. Degradation may also be very gradual and take some years to become apparent, imparting a false sense of security.

Management of water resources requires a sound understanding of key factors such as physical extent (quantity), quality, availability, and constraints to development. The role of Resource Assessment Division of the Department for Water Resources is to maintain an effective knowledge base on the State's water resources, including environmental and other factors likely to influence sustainable use and development, and to provide timely and relevant management advice.

Bryan Harris
Director, Resource Assessment Division
Department for Water Resources

EXECUTIVE SUMMARY

The fractured rock aquifers in the Clare Valley of South Australia have been extensively researched over the past four years to provide estimates of groundwater fluxes which will ensure sustainable development of the expanding wine industry and protection of the environment. The project has developed new techniques for estimating aquifer recharge rates and horizontal and vertical flow velocities.

Hydraulic testing has indicated that the aquifers can be divided into two broad zones — a relatively permeable zone in the upper 20–40 m, and a deeper low-permeability regional zone. Fracture mapping of rock exposed at the surface and drill core indicate that fractures are closely spaced (generally less than 0.5 m) in the upper zone. The movement of water depends upon not only the fracture spacing, but also on their size and connection. The size of the fractures and their spacing tend to decrease with depth. Well testing has indicated that hydraulic conductivity varies by many orders of magnitude, but generally decreases with depth. The deeper flow system becomes progressively isolated from the upper system due to a decrease in fracture density.

Prior to the clearance of native vegetation, groundwater recharge rates in many parts of Australia were much lower than they are today. We have estimated the pre-clearing recharge rate in the Clare Valley to be <2 mm/yr from chloride mass balance methods. Current recharge rates cannot be estimated using this approach, as there is still some residual pre-clearing salt stored in the aquifer matrix.

We have obtained estimates of recharge rates at two different sites using a combination of aquifer properties and groundwater dating with a suite of environmental tracers (chlorine-36, carbon-14, chlorofluorocarbons (CFCs) and tritium). Recharge rates were determined to be in the range 50–75 mm/yr. In other areas of the Clare Valley, where mean annual rainfall is less, we would expect lower rates of annual recharge. Groundwater dating has also revealed rapid changes in groundwater age over short vertical distances in the aquifer. This suggests that the upper and lower flow systems are relatively isolated from each other.

Many wells in the region exhibit step-like increases in Electrical Conductivity (EC) with increasing depth below the water table. We attribute these steps to the locations where some major fractures intersect the well. In order to determine horizontal flow velocities through the well, we have developed a new method (well dilution) that involves disturbing the EC profile, and monitoring the time it takes for the disturbed profile to return to it's original position. The time that it takes for the mixed profile to return to the original profile is related to the flow rate. Flow rates estimated from this method range over four orders of magnitude from 1 mm/day to 10 m/day.

We have also developed another technique for estimating horizontal flow that involves comparing radon concentrations in unpurged and purged wells. If the flow rate through the well is very low then any radon delivered to the well from the aquifer will decay to zero in the well column. Significant radon concentrations will be present in the well if the rate of supply of radon to the well by horizontal flow exceeds the rate of radon decay. The range of groundwater flow rates obtained using this approach was within the range of those obtained using the well dilution method.

A surface water-groundwater chloride mass balance indicates that the mean groundwater discharge to the Eyre Creek is <10 mm/year. A similar result was obtained using a method which utilises seasonal fluctuations in aquifer water level data to separate stream flow data. Thus, if mean annual recharge is 50–75 mm/yr, groundwater pumping for irrigation and discharge to creeks are both less than 10 mm/yr, then the remaining 30–55 mm/yr of recharge must leave the catchments by deep groundwater outflow.

CONTENTS

1. INTRODUCTION

Characteristics of fractured rock aquifers

Fractured rock aquifers underlie approximately 40 % of the Australian continent (Fig. 1.1) and are becoming increasingly important water supplies as sedimentary aquifers are becoming fully utilised. Water supplies from these systems are used for irrigation, livestock supplies, small town water supplies, mining and remote communities throughout Australia. Despite their obvious importance, very little is known about how water is stored and transported in these systems.

Groundwater behaves very differently in fractured rock aquifers compared with porous media aquifers.

In a porous media aquifer, the underground water is stored in, and moves through, gaps between individual mineral grains or mineral assemblages (Fig. 1.2). Unconsolidated sand and gravel aquifers form the best examples of porous media aquifers.

Fractured rock aquifers consist of two components — the fractures and the matrix blocks. Fractures include cracks, joints and faults and can vary in length from centimetres to kilometres. The majority of groundwater flow occurs only through the fractures whereas the matrix blocks act as storage reservoirs. Fractured rock aquifers include basalt, granite, metamorphics, shale, slate, and some dolomites and limestones.

Obtaining reliable groundwater supplies in fractured rock aquifers is usually complicated by irregular, unpredictable distributions of fractures, causing large spatial variations in bore yield. Experimental and theoretical studies have shown that the flow rate increases dramatically as the width (aperture) of the fracture increases. For example, if the size of the aperture is doubled, the flow rate increases by a factor of 8 (Snow, 1968).

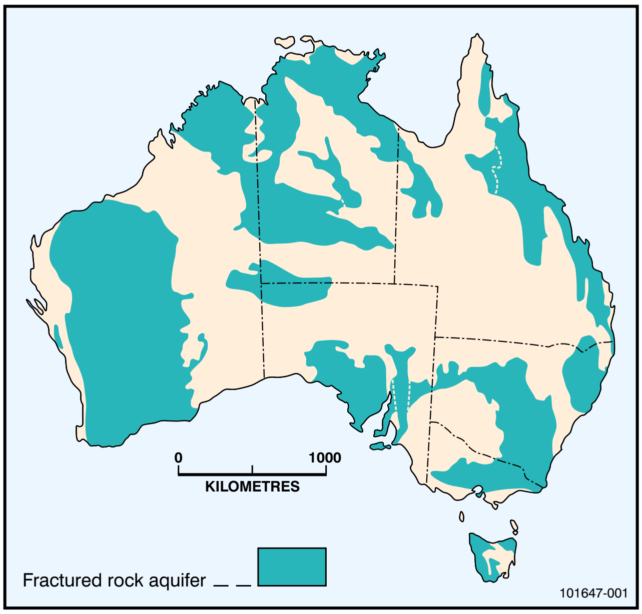


Fig. 1.1 Major fractured rock provinces in Australia (after Jacobsen and Lau, 1987).

Many of the traditional methods for estimating groundwater recharge and flow rates in porous, sedimentary aquifers cannot be directly transferred to fractured rock systems. Fractured rock aquifers are characterised by high spatial variability in hydraulic conductivity, making traditional hydraulic methods for estimating groundwater flow difficult to apply. Specific yield may also be extremely variable and difficult to measure, making groundwater recharge estimation from bore hydrographs unreliable. Sufficiently accurate data on aquifer thickness or porosity is not available for reliable determination of aquifer storage.

A lack of understanding and predictive capability of groundwater movement in and recharge to fractured rock aquifers has resulted in groundwater management being largely problematic. The major problem is our inability to measure the parameters that control the flow system (e.g. location and size of fractures). Management has often been reactionary, reliant on monitoring results from salinity and groundwater levels to assess the long term sustainability of a fractured rock system. Because of low porosity, and large spatial and temporal variability of water levels and salinity, monitoring in fractured rock aquifers needs to occur over longer time scales than for sedimentary aquifers. By the time monitoring has detected changes in the hydraulic balance of a system, it may be politically difficult to reduce allocations to the required level. Prediction of contaminant migration is similarly difficult, but may be particularly important due to potential high transport velocities.

The Clare Valley

The Clare Valley is renowned for the production of premium quality wines, traditionally from dry grown grapes, harvested from the numerous vineyards situated throughout the region. Over the past few decades the demand for increased grape yield and volume from the district has led to increased irrigation from underground and surface water resources. As well as expansion of the viticultural industry, there is no reticulated water supply between Auburn and Clare, which places addition demand on the groundwater resources in this area. Current extraction from groundwater resources is 1950 ML/a and surface water is 1700 ML/a. If we distributed the groundwater extraction over the entire area of the Clare Valley this would equate to an areal extraction rate of 4 mm/a. However, in areas of intense irrigation activity, the areal extraction rate may be up to three times this value for individual sub-catchments.

The increasing demand for additional groundwater and surface water resources to be used for irrigation resulted in

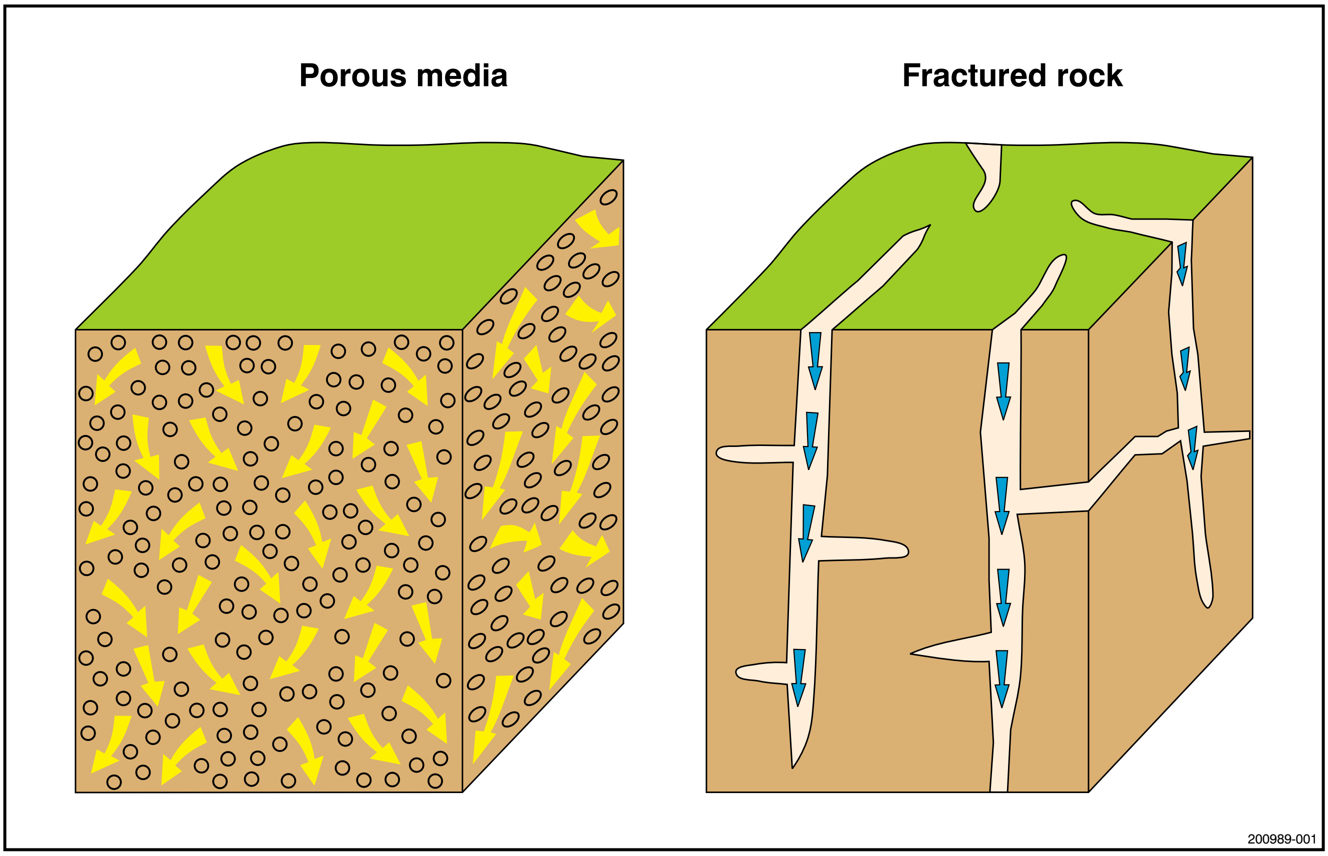


Fig. 1.2 The difference between a sand aquifer (porous medium) and a fractured rock aquifer is that, in the former, water moves between an interconnected network of gaps between the sand grains. In the latter, water must find its way through an array of irregularly spaced, and often unconnected, cracks in the rock.

regulating water use in the Clare Valley. A local management committee was established and the Clare Valley was proclaimed in July 1995 as a Prescribed Water Resources Area under the Water Resources Act of South Australia (Fig. 1.3). This resulted in a water allocation plan for the region, which has subsequently been updated as of December 2000. The aim of the plan is to sensibly control the increase demand for water resources by ensuring the resource is sustainable in the long term. Since then each well requires a licence and is metered so that extraction can be accurately determined. Each well or group of wells is given a zone of influence around the well, and new licence wells cannot be established within an existing zone. The zone of influence is calculated to be a circle around the licence well based on the capacity of the well (volume irrigated) divided by an assumed recharge rate of 30 mm/a (Fig. 1.4). To ensure that excessive well interference does not occur production wells can not be located closer than 200 m apart. As well, a zone of influence protects environmental sensitive permanent pools in creeks.

Groundwater in the Clare Valley is stored in fractured rock aquifers in the Adelaide Geosyncline. However to date, traditional methods for characterising groundwater systems in porous media aquifers have not been successfully transferred to fractured rock systems. The challenge for the future is to develop new techniques, which will allow us to improve our technical understanding of these systems. An improved technical understanding will underpin successful management of these resources into the next century. Without estimates of safe yield, further expansion of irrigation in the district may be limited due to uncertainty of suitable long-term supplies.

As a result of this, a collaborative research project to study the sustainability of groundwater in the Clare Valley was established between Department for Water Resources, CSIRO Land and Water and Flinders University. Land and Water Australia (formerly Land and Water Resources Research and Development Corporation) and the Clare Valley viticulture industry provided additional funding for this project.

The specific objectives of this project were:

- To develop more reliable methods for estimating recharge rates and horizontal flow rates in fractured rock aquifers.
- Apply these methods to the Clare Valley to determine whether future development can take place and the levels at which extraction can be comfortably sustained.
- Disseminate any improved techniques for parameter estimation in fractured rock aquifers to other water managers and researchers.

The following three chapters of this report develop a model of the groundwater flow systems in the Clare

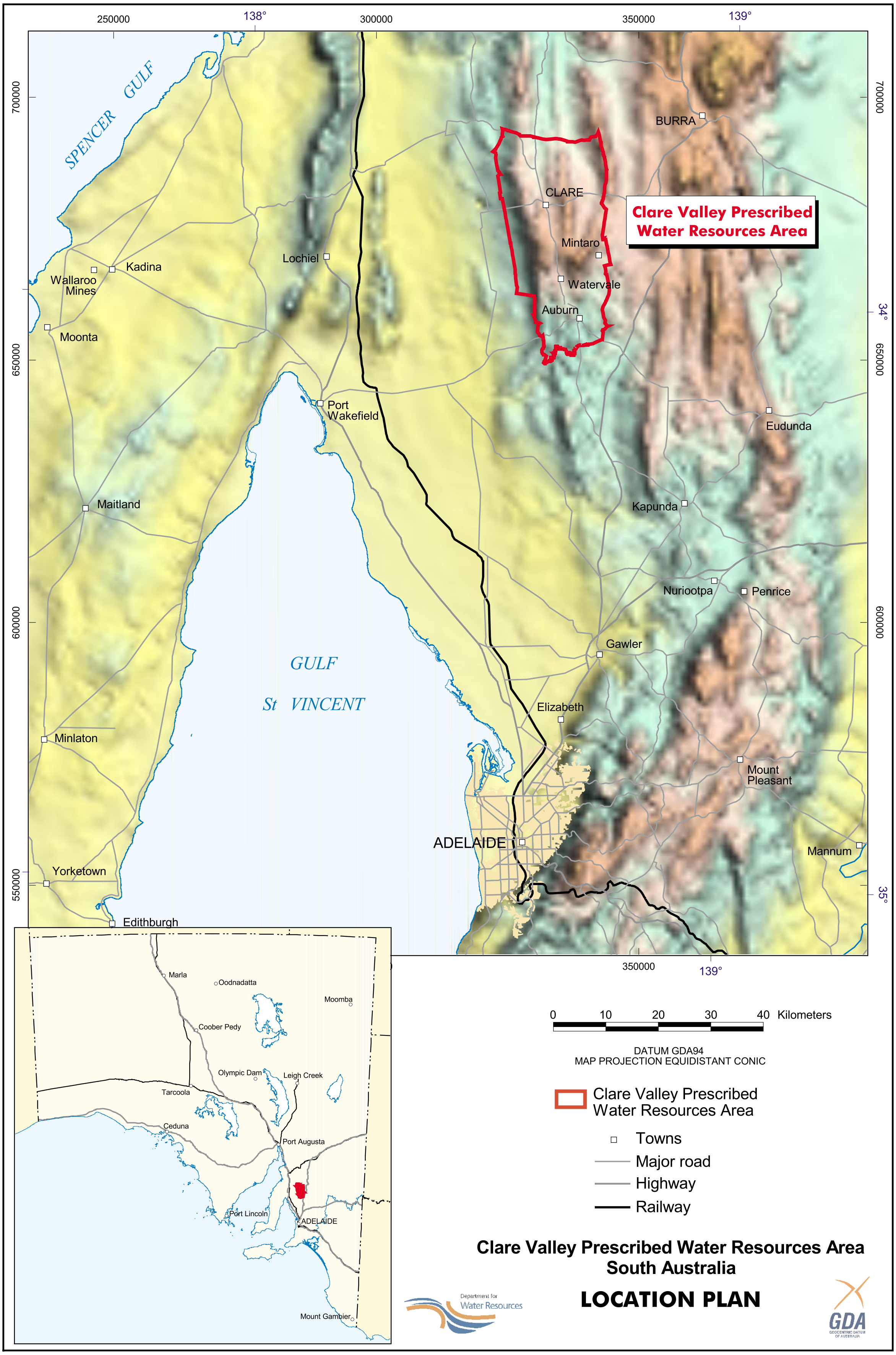


Figure 1.3

Publishing Services AV : 200527_001

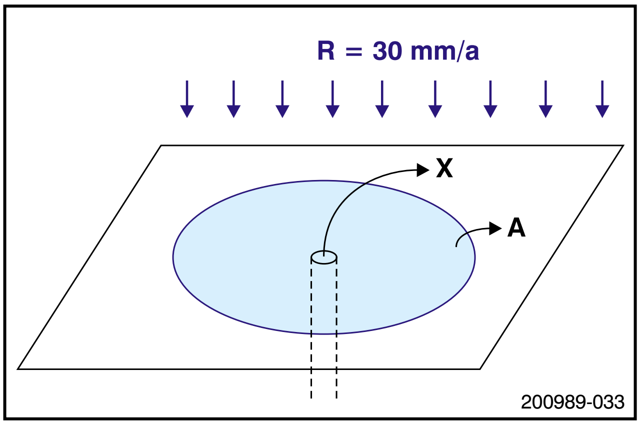


Fig. 1.4 *The groundwater zone of influence is calculated from A=X/R; where A is the area, X is the extraction rate of the well and R is the assumed recharge rate.*

Valley. Chapter 2 describes the hydrogeology, groundwater monitoring results, drilling program and presents a preliminary salt and water balance. Chapter 3 characterises the various properties of both the matrix and fractures using a range of laboratory and field techniques. Chapter 4 develops a number of novel approaches to estimating recharge rates, horizontal groundwater flow rates and groundwater discharge to creeks.

2. HYDROGEOLOGY OF THE CLARE REGION

Study area

The Clare Valley is situated approximately 100 km north of Adelaide within the northern Mount Lofty Ranges (Fig. 1.3). The study area covers approximately 700 square kilometres and incorporates a number of catchments and sub-catchments within the Clare Valley Prescribed Water Resources Area (Fig. 2.1). The two main surface water catchments within the Valley are the Wakefield and Broughton River Systems, which are separated by an east-west topographic high approximately 2 km north of Penwortham. Surface water flows primarily towards the north and south from this topographic high.

Physiography

The Clare Valley is not a single valley; rather it comprises a number of small catchments consisting of valleys that are sub-divided by a series of ridges. The Clare Valley is bounded to the west by a north-northwest trending fault ridge, with an elevation of approximately 400 m (a remnant of tectonic activity within the region); west of this ridge the terrain becomes relatively flat. To the east of the Valley, the pronounced ridges display elevations in excess of 550 m above sea level. Further eastward, the terrain becomes softly undulating. The numerous valleys and ridges can be attributed to a difference in weathering capability of the underlying rock types. The harder rock types, such as quartzite and sandstone, are less vulnerable to weathering, and thus form the pronounced ridges seen within the region. Softer rock types (e.g. dolomites and shales) are more easily weathered and form the lower lying areas.

There are a number of drainage paths within the region; the Hill River, Eyre Creek, Hutt River and River Wakefield being the most pronounced. The catchment areas for these surface drainage systems are illustrated in Figure 2.1. There are no perennial streams and flow within the watercourses occurs for less than half of the year, sometimes only intermittently depending on the amount and duration of rainfall events.

Climate

The climate is one of hot, dry summers and cool to cold, moist winters. Daily maximum temperatures average around 30°C in summer and 14°C in winter, with June through to August being the wetter months. Mean annual rainfall is 653 mm at Watervale, 629 mm at Clare and 587 mm at Auburn, while the mean annual evaporation at Clare is ~ 1975 mm (Bureau of Meteorology data base). The higher rainfall areas generally correspond to areas of higher elevation.

Land use

Native vegetation is confined mainly to the tops of ridges where soils are thin and hence productive agriculture is unsustainable. The most common native tree species are blue gum *(Eucalyptus leucoxylon)*, peppermint gum *(E. odecrata)* and sheoak *(Casuarina stricta)* (Forbes, 1965). Land use today is dominated by perennial pasture, vineyards and orchards.

Geological setting

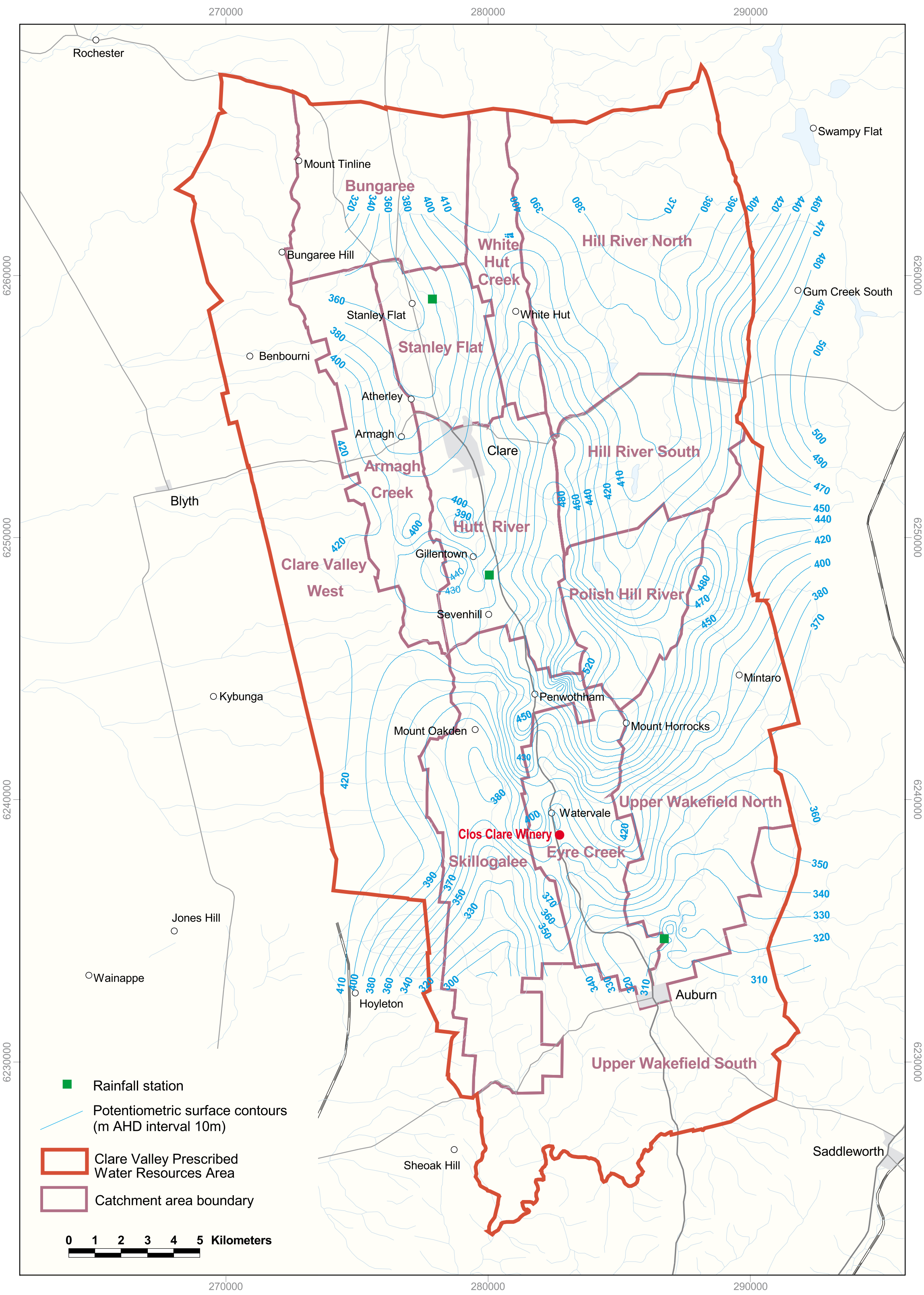
The Clare Valley lies within the Northern Mount Lofty Ranges and forms part of the larger Adelaide Geosyncline. The Adelaide Geosyncline is a failed continental rift valley that is exposed today as ranges extending from Kangaroo Island in the south to Freeling Heights in the north of South Australia. The Clare Valley consists pre-dominantly of Neo-proterozoic rocks of the Burra and overlying Umberatana Groups (650–780 million years BP). Rock types within the area consist primarily of shale, siltstone, sandstone, dolomite and quartzite (Fig. 2.2). Approximately 400-500 million years ago in the Paleozoic–Ordovician period, the original rock types (i.e. mudstone, siltstone, sandstone and carbonate) were subjected to a period of intense heat, pressure and tectonic activity (collectively known as the Delamerian Orogeny, Preiss, 1987). During this time the original rocks underwent slight chemical and physical alteration (low grade metamorphism), where the original porosity within the rocks was reduced and new mineral growths were established, resulting in the rock types observed within the area today. In addition, intense earth movements caused folding and faulting and subsequent brittle fracturing. A second period of tectonic activity occurred during the Tertiary period which resulted in uplift and secondary rejuvenation of earlier tectonic features.

Hydrogeology

There are two main aquifer types within the Clare Valley region — minor alluvial/colluvial aquifers which occur mainly to the north of Clare in the vicinity of Stanley Flat, and substantial fractured rock aquifers throughout the remainder of the region. Within fractured rock aquifers, the fractures act as conduits for groundwater flow whereas the surrounding matrix blocks act as storage reservoirs. Fractures may be continuous or discontinuous depending on their nature and may vary in scale from centimetres to kilometres. The yield of groundwater from a particular well is therefore dependent on the amount, aperture and orientation of fractures intercepted.

Hydrostratigraphic units

The principle hydrogeologic units within the Clare Valley are the Gilbert Range Quartzite, Mintaro Shale, Rhynie Sandstone, Saddleworth Formation, Skillogalee Dolomite and Undalya Quartzite (Fig. 2.3). Fracturing in the region is considered to be ubiquitous and groundwater can flow



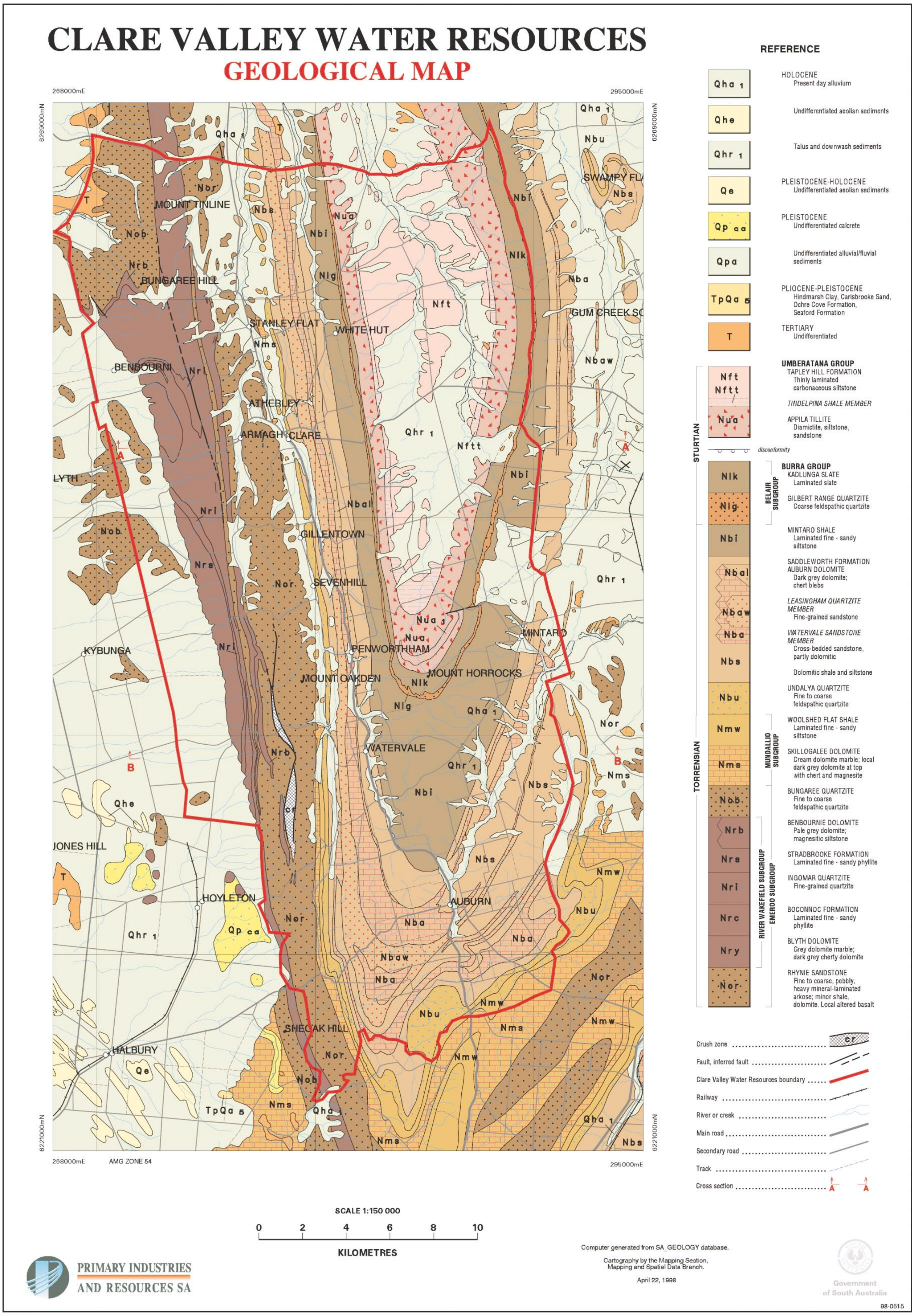
Clare Valley Prescribed Water Resources Area
South Australia
POTENTIOMETRIC SURFACE CONTOURS

Department for Water Resources

GDA

Figure 2.1

Publishing Services AV : 200527_002



J101087

Fig. 2.2

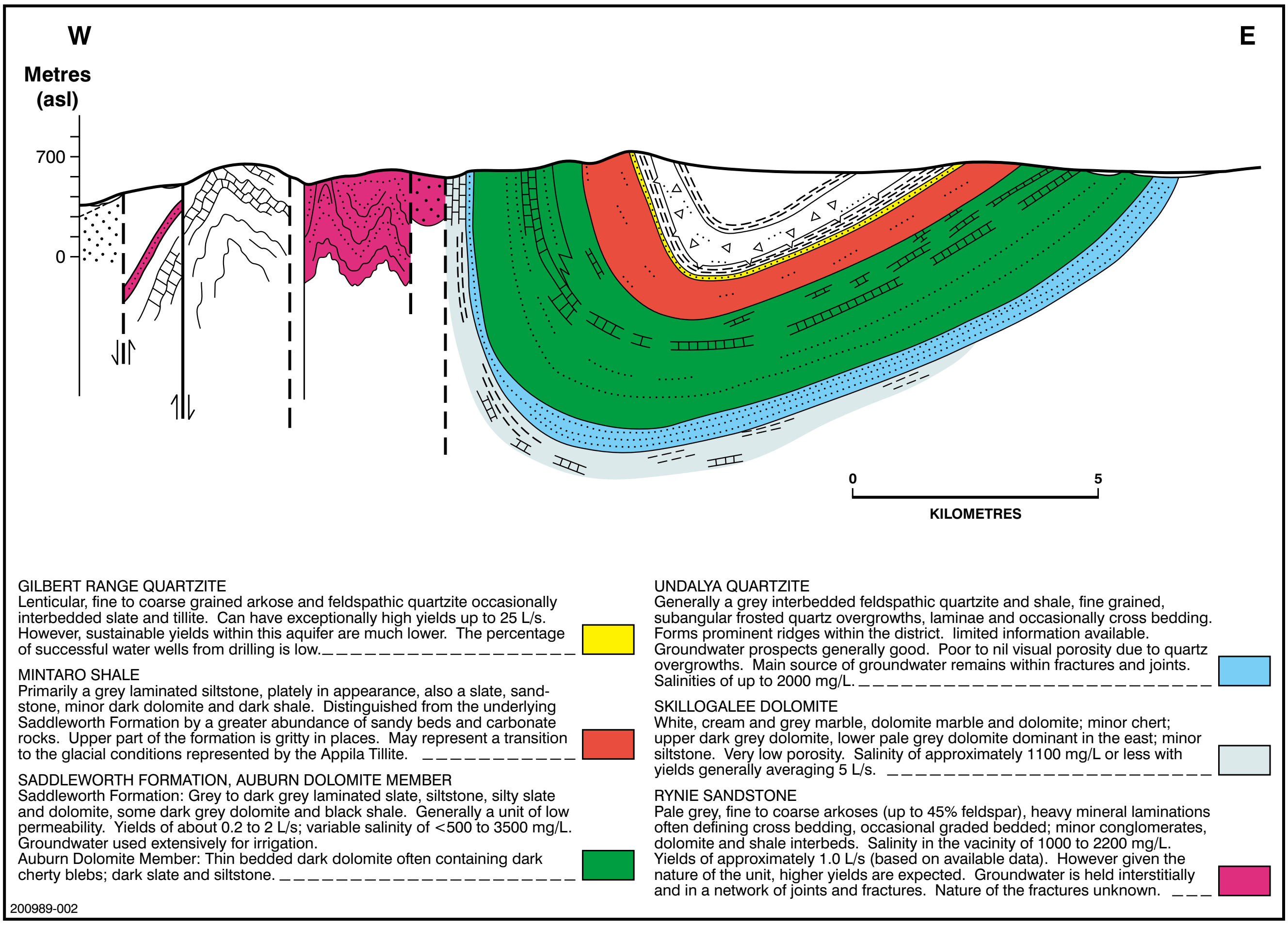


Fig. 2.3 *Schematic geological cross-section along A–A across the Hill River Syncline (Fig. 2.2).*

across hydrogeologic units. Variation in yield from individual wells is likely to be the result of fracturing or other structural constraints rather than rock type.

Water table contours

A potentiometric surface map was produced utilising water table depth data collected by staff of the Department of Environment, Heritage and Aboriginal Affairs (DEHAA) in September–October 1995 (Fig. 2.1). Only a small percentage of wells had corresponding surveyed elevation data — the remaining elevation data was estimated from topographic maps. Horizontal hydraulic gradients vary from approximately 10 m/100 m on steep ridge slopes to much flatter gradients of 0.5 m/100 m in the valley floors.

The potentiometric map indicates that the water table follows a subdued form of the topography and that groundwater flows approximately in the same direction as surface water. An east-west groundwater divide exists approximately 2 km north of Penwortham (coinciding with the topographic and surface water divide). Both groundwater and surface water flows primarily towards the north and south from this divide. However, detailed groundwater flow orientations cannot be constructed from the water table contours due to the highly anisotropic nature of the geological medium.

Groundwater salinity and well yields

Figure 2.4 depicts the spatial variation in groundwater salinity across the Clare region. The data used to construct this map was taken from SA_GEODATA and is the first data recorded in the data base, corresponding to the time that the well was drilled. Groundwater salinity varies from less than 500 mg/L to in excess of 7000 mg/L (Fig. 2.4). In areas of dense bore development groundwater salinity can vary considerably between adjacent bores (e.g. near Watervale and Mintaro). The best quality groundwater is associated with higher rainfall and higher topography between Watervale and Sevenhill. Outside the Clare Valley Water Resource boundary, the water quality generally exceeds 3000 mg/L with many bores greater than 7000 mg/L, and therefore is not suitable for many irrigation practices.

Well yields vary from less than 0.1 L/s up to 25 L/s depending on the number and characteristics of fractures and joints encountered. There are no trends between well yield and location nor rock type. The sustainability of groundwater supply from the higher yielding wells is unknown; for example, in some of the quartzite ridges, a number of high yielding bores have failed to maintain their supplies.

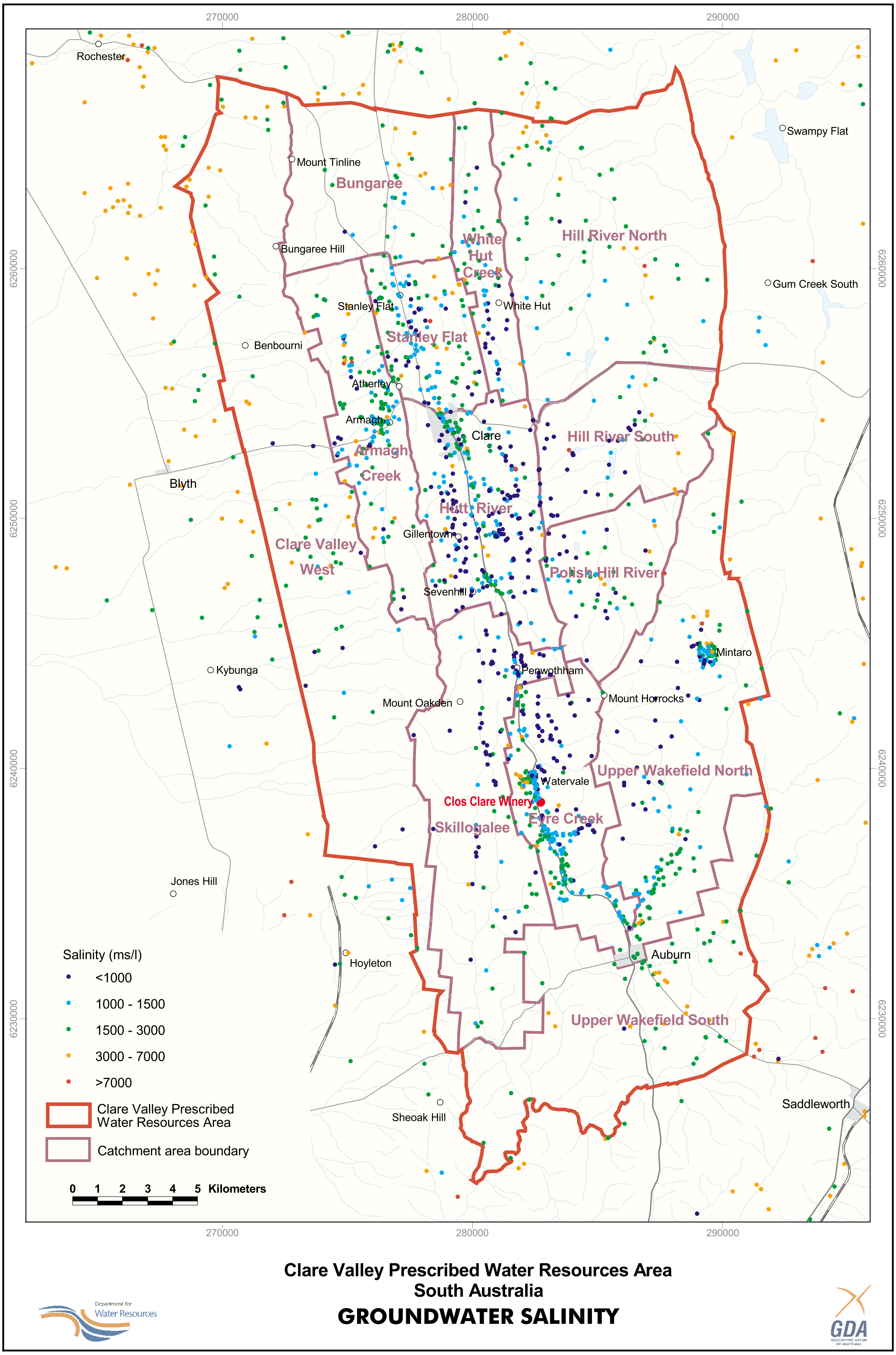


Figure 2.4

Publishing Services AV : 200527_010

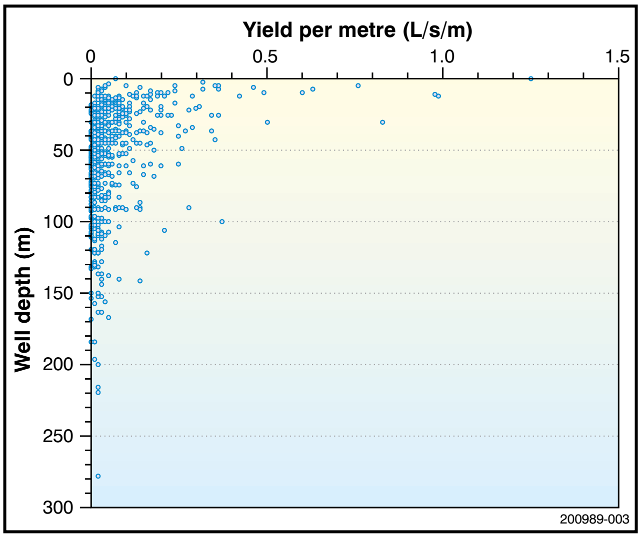


Fig. 2.5 *Bore yields per metre depth. In construction of this graph the yield of the well was divided by the total well depth (L/s/m) and plotted against the total depth of the well. Data from SA_GEODATA.*

There is no obvious correlation between salinity and well yield, however there is a noticeable relationship between well yield and total depth of the well. The yield per metre of well clearly decreases with depth (Fig. 2.5) indicating that the highest yielding part of the aquifer is the upper 20–30 m, which is most likely related to enhanced weathering of this zone.

Catchment scale water and salt budgets

Water and salt budgets were developed for two areas within the Clare Valley Prescribed Water Resources Area (CVPWRA), namely the Eyre Creek sub-catchment and the combined area of the Hutt River, Armagh Creek and Stanley Flat sub-catchments (Fig. 2.1). The Eyre Creek catchment has relatively high topographic relief and therefore high mean annual rainfall compared with the rest of the Valley.

Water balances were calculated for each area. The only input is rainfall and outputs were evapotranspiration, surface run-off and natural groundwater flow out of the catchment. Salt balances were calculated in the same way, with the input coming from rainfall and outputs via surface run-off that is focused into streams and lateral groundwater flow out of the catchments. Further details regarding the design of the water and salt budgets, as well all input data are presented in Harrington and Love (2000).

The Eyre Creek catchment is relatively well-instrumented containing over 100 years of reliable rainfall record at Watervale, a stream flow and salinity gauging station on the Eyre Creek near Auburn (Fig. 2.6), numerous ground-water level and salinity monitoring wells, and some aquifer transmissivity estimates from pumping tests. Nevertheless, estimating components of the water balance such as evapo-transpiration and lateral groundwater discharge from fractured rock aquifers can yield large errors in the final budgets. Thus a detailed error analysis was also performed.

The area encompassed by the Hutt River, Armagh Creek and Stanley Flat sub-catchments has generally lower topographic relief than the Eyre Creek catchment and is located in the northern part of the study area; hence it is referred to from this point onwards as the Northern catchments. Higher errors were expected for the water and salt budgets for this area compared with the Eyre Creek catchment because of the paucity of field data.

Results of each water and salt balance calculation are presented with estimates of associated errors in Table 2.1. The range of water In/Out ratios for the Eyre Creek and Northern catchments is 0.8 to 2.3 and 0.5 to 2.3 respectively. If groundwater recharge via rainfall was exactly matching discharge via natural outflow and pumping, then the water In/Out ratio would be 1. Therefore, the range of ratios obtained for each catchment are inconclusive. Similarly, the range of salt In/Out ratios obtained for each catchment are inconclusive.

Table 2.1 *Results of the water and salt budget calculations for two sub-catchments within the Clare Valley (Harrington and Love, 2000). Errors associated with each parameter are presented as percentages, while the overall results are presented as a range which incorporates all of these errors.*

	Eyre Creek	Hutt River Armagh Creek Stanley Flat
Water In (ML/yr)		
Rainfall	23419±5%	65090±5%
Total	22250–24600	61850–68350
Water Out (ML/yr)		
Surface discharge	346±10%	2773±20%
Evapotranspiration	15755±40%	42489±40%
Dam storage	967±25%	1757±25%
Groundwater pumping	187±20%	811±20%
Groundwater discharge	373±1000%	5676±1000%
Total	10700–27950	30250–128400
Water In/Out	**0.80–2.30**	**0.48–2.26**
Salt In (tonnes/yr)		
Rainfall + dryfall	104±15%	306±15%
Total	89–120	260–352
Salt Out (tonnes/yr)		
Surface discharge	22±20%	208±50%
Groundwater discharge	218±1000%	6465±1000%
Total	40–2423	750–71427
Salt In/Out	**0.04–3.0**	**0.004–0.47**

Root mean square (RMS) errors for each water and salt budget were determined after estimating the likely error associated with each individual parameter used in the budget calculations (Harrington and Love, 2000). However, the RMS errors (20 to 25 % for each budget) are negligible compared with the ranges of Water In/Out and Salt In/Out ratios presented in Table 2.1.

Groundwater monitoring

The Department for Water Resources is responsible for routine monitoring of groundwater levels (~190 wells) and salinity (~120 wells) throughout the Clare Valley. Approximately 17% of the monitoring wells are used for irrigation, while the remainder are either commercial, research, stock or domestic wells. Measurements are taken at least four times per year to enable the assessment of both the long- and short-term health of the groundwater resource.

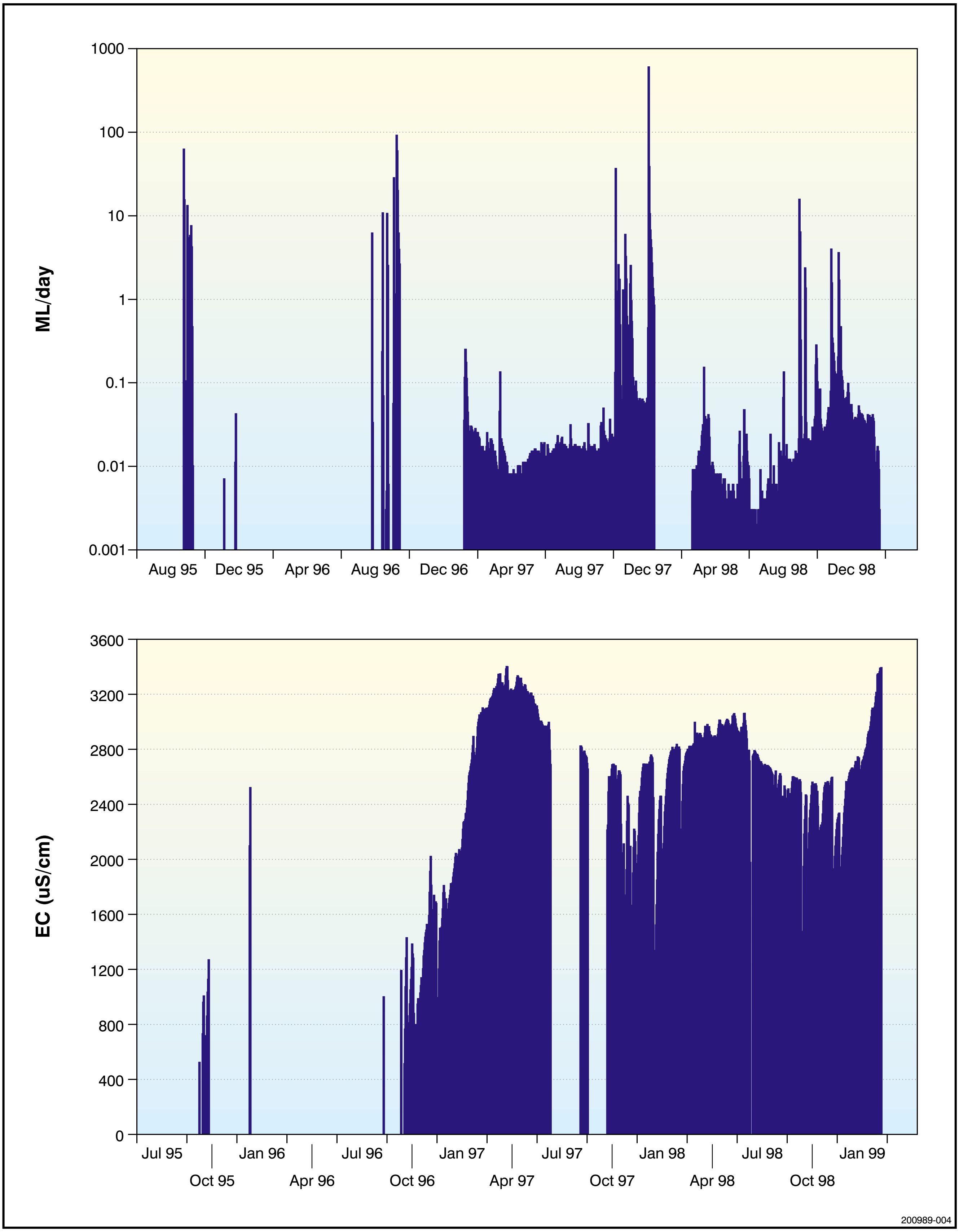


Fig. 2.6 *Daily stream flow and salinity data for the Eyre Creek recorded at the Auburn gauging station. The gaps in the graphs represent periods of no flow or times when the flow and/or EC measuring instruments were not operating.*

Water levels

Groundwater levels in the Clare Valley vary considerably throughout the year due to climatic conditions and characteristic properties of the fractured rock aquifers. The magnitude of seasonal fluctuations are perhaps more pronounced than many other parts of the State due to the low storage capacity of the aquifers in the region. That is, relatively small additions or extractions of groundwater from this resource can result in significant variations in the water table elevation. Such variations are evident in the two hydrographs shown in Figure 2.7. Both hydrographs reveal the annual variation in water tables, as well as the impact of significant rainfall (and hence recharge) events. For example, well 6630-0440 shows a peak in water table elevation at January 1993 that corresponds to the end of a period of above average monthly rainfalls. This plot also shows how the water table at February 2000 was the lowest it had been for 10 years, which was due to below average annual rainfall over the previous three years. Subsequent to this date, above average rainfall during winter and spring 2000 resulted in the water table recovering to the highest level since late 1993. Regardless of the dynamic nature of water levels in these aquifers, very few bore hydrographs exhibit any long-term (i.e. >10 year) declines in water levels.

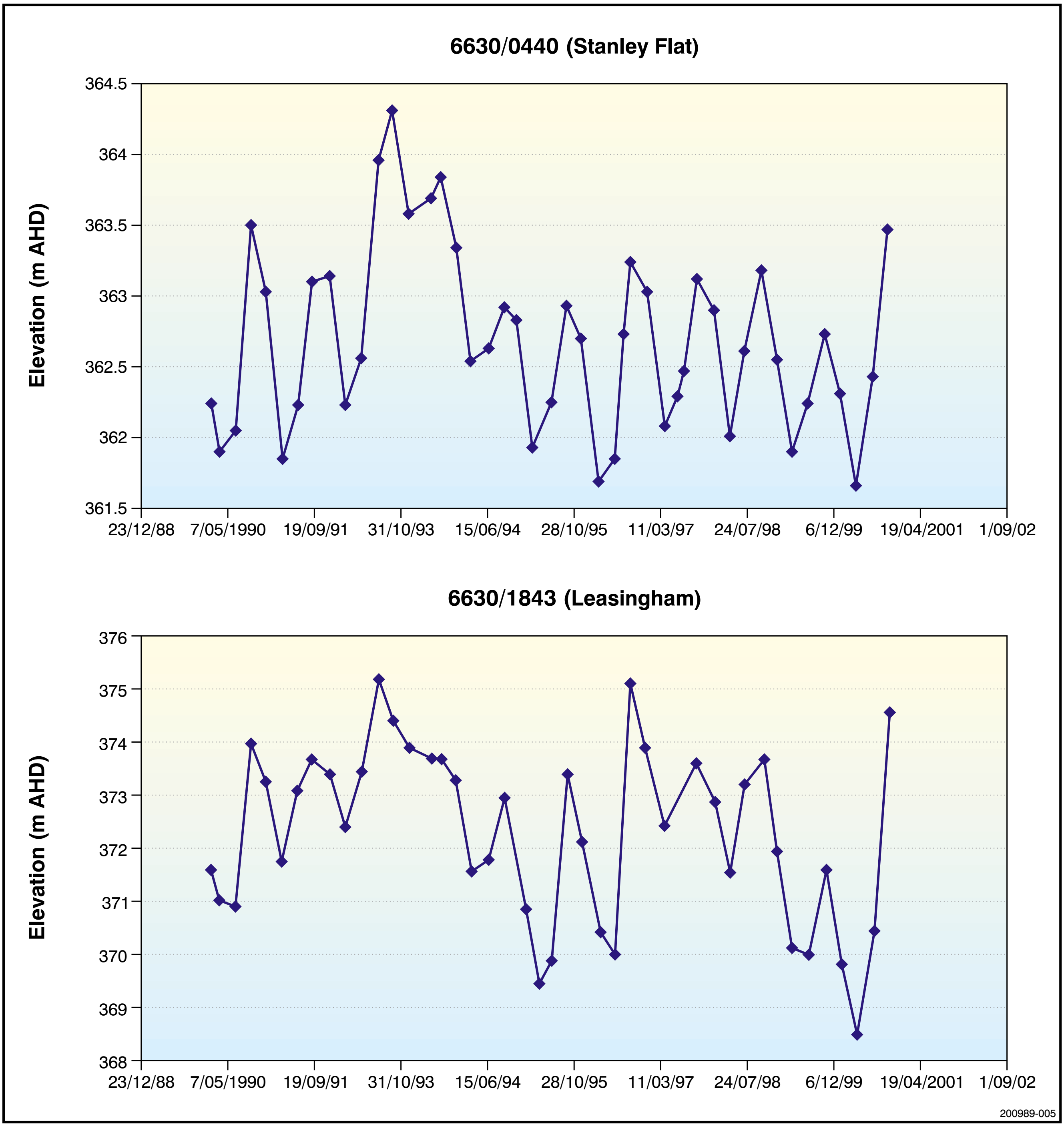


Fig. 2.7 *Hydrographs of representative monitoring wells near Stanley Flat and Leasingham.*

Salinity

Many groundwater wells within the Clare region exhibit an increase in salinity with increasing depth below the water table (e.g. Fig. 2.8). Since the time that land was cleared of native vegetation (50–80 years ago), the rates of groundwater recharge have increased dramatically, which has most likely resulted in increased flushing of accumulated salts from the aquifer. Thus, the salinity-depth trend observed in many wells is probably due to greater leaching of salt from the aquifer in the upper 10–30 m where we assume that fracture density is relatively high. Therefore, during periods of low rainfall, when groundwater pumping for irrigation is at a maximum, water tables are often drawn down to the extent that a large proportion of the pumped water is sourced from deeper in the aquifer, where the salinity is generally higher. Conversely, at the end of Spring when water tables have recovered due to recharge from rainfall, the average salinity of pumped water will be at a minimum as the majority of water is sourced from the uppermost part of the aquifer. This seasonal trend in groundwater salinity associated with transient water tables is shown diagrammatically for well 6630-1833 in Fig. 2.9.

Drilling program

Four phases of drilling were undertaken in the Clare Valley at different stages throughout the project. During these phases a total of 29 wells were drilled — 8 were completed as nests of piezometers and the remainder as open holes with surface casing only. A summary of the well location and construction details for all wells is presented in Table 2.2, and locations of research sites are shown in Figure 2.10.

Detailed geological logs were prepared from either chip samples or core material for each well and are presented elsewhere (Morton and Love, 1998; Morton et al., 1998; Clarke et al., 2001). The latter reference also contains fracture spacing and orientation data measured from a core at the Wendouree site.

Most of the wells drilled during this project were logged for salinity, temperature and pH using a YSI® Sonde at different times of the year, including immediately after drilling was completed. Temperature anomalies and rapid spatial changes in electrical conductivity (EC) have previously been used to infer the location of active hydraulic fractures (Drury and Jessop, 1982; Ge 1998). As drilling tends to mix the well column, each well was left undisturbed for a period of six months after it was drilled, so that it had time to stabilise. After this time, many wells were logged again for electrical conductivity, pH and temperature.

EC, pH and temperature profiles from the open well 6630-2877 in June 2000 are shown in Fig. 2.8. Two rapid vertical changes in EC occur at depths of 38 and 70 m. The EC step at 38 m changes from 1.9 $mS.cm^{-1}$ to 3.75 $mS.cm^{-1}$ over a vertical distance of only 4 m, while the EC step that starts at 70 m with an EC of 3.75 $mS.cm^{-1}$ changes to 5.75 $mS.cm^{-1}$ at 76 m. We believe these rapid vertical changes in EC in the well represent the location of groundwater flow entering the well via fractures. Major anomalies of the temperature gradient also correspond to step changes in the EC profile starting at 38 m and 70 m. Further confirmation that major fractures intersect the well at 38 m and 70 m is evident by the rapid changes in pH at these locations indicating the in-flux of new water into the well.

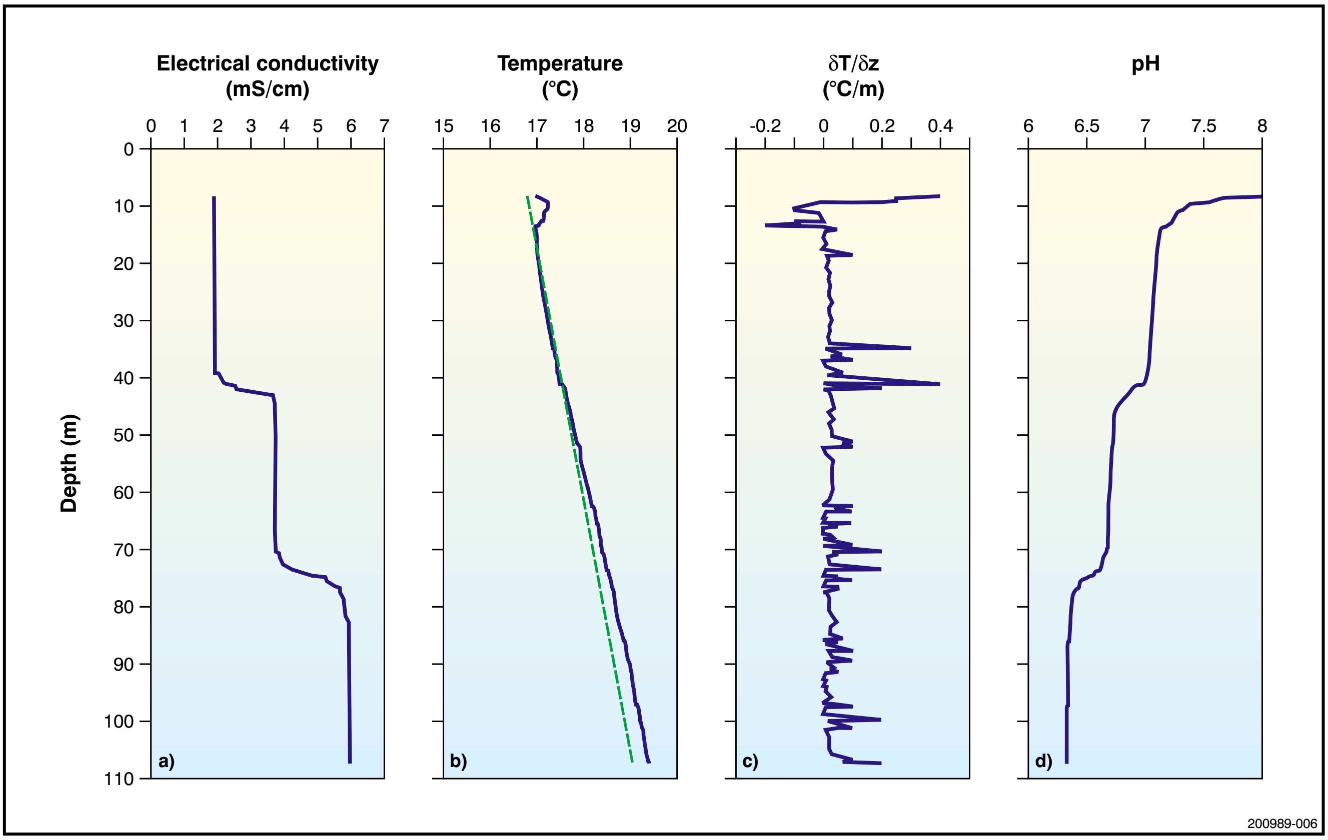


Fig. 2.8 Salinity, temperature, temperature gradient and pH versus depth, well 6630/2877.

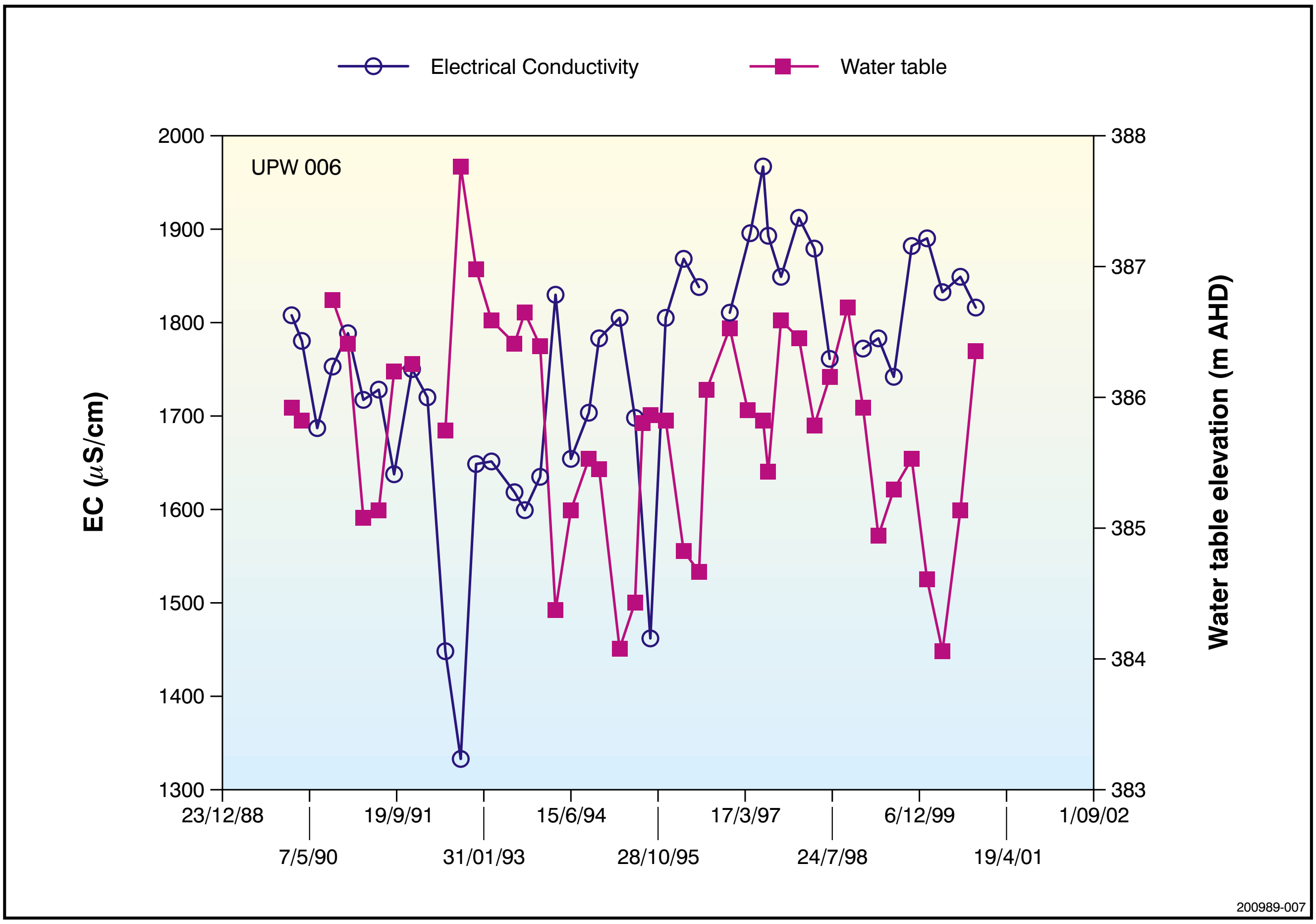
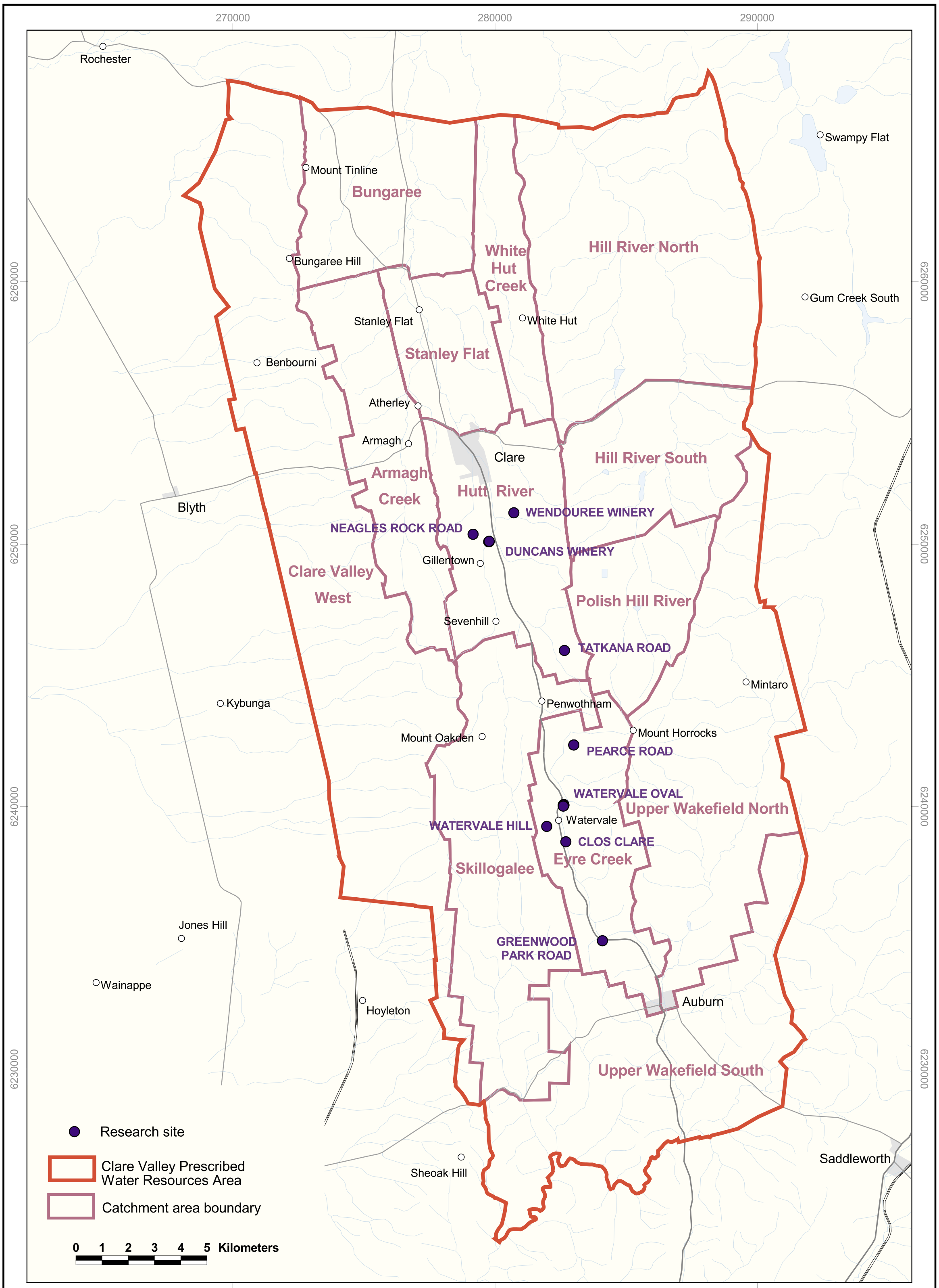


Fig. 2.9 Salinity and water table variations in monitoring well 6630/1833 near Leasingham.

Table 2.2 *Location and construction details and wells completed during four drilling phases in the Clare Valley, 1995–2000.*

Drill Phase	Permit No.	Unit No.	Obs. No.	Location	Geological Unit	Total Depth (m)	Cased to (m)	Slotted Interval (m)	SWL (m)
1	36386	6630-2625	CLR 94	Wyndham Park	Auburn Dolomite?	99	5.8		15.3
1	36387	6630-2621	CLR 96	Neagles Rock Road (10")	Skillogalee Dolomite	58.6	5.8	N/A	23.75
	1	6630-3130	CLR 137					54 - 56	
	2	6630-2734	CLR 138					44 - 45	
	3	6630-2735	CLR 139					38.75 - 39.7	
	4	6630-2736	CLR 140					34 - 34.5	
	5	6630-2737	CLR 141					29 - 29.5	
	6	6630-2738	CLR 142					26 - 26.5	
1	36388	6630-2622	CLR 95	Neagles Rock Road (8")	Skillogalee Dolomite	131.5	5.8	N/A	23.52
	1	6630-3129	CLR 133						
	2	6630-2739	CLR 134					95 - 100	
	3	6630-2740	CLR 135					81.5 - 84.5	
	4	6630-2741	CLR 136					63.2 - 66.2	
1	36389	6630-2620	CLR 93	Duncans Winery (1)	Undalya Quartzite	58.5	4	N/A	5.17
	1	6630-2704	CLR 128					50 - 53	
	2	6630-2703	CLR 129					40 - 43	
	3	6630-2702	CLR 130					30 - 33	
	4	6630-2701	CLR 131					18 - 21	
	5	6630-2700	CLR 132					9 - 12	
1	36390	6630-2619	CLR 92	Duncans Winery (2)	Undalya Quartzite	63	3.8		5.19
1	36391	6630-2590	CLR 91	Duncans Winery (3)	Undalya Quartzite	63	3	N/A	5.37
	1	6630-2709	CLR 123					50 - 53	
	2	6630-2708	CLR 124					40 - 43	
	3	6630-2707	CLR 125					30 - 33	
	4	6630-2706	CLR 126					18 - 21	
	5	6630-2705	CLR 127					9 - 12	
1	36392	6630-2584	UPW 53	Pearce Road (8")	Mintaro Shale	99.5	5	N/A	6.3
	1		UPW 63					94.25 - 99.25	
	2	6630-2726	UPW 64					75 - 80	
	3	6630-2727	UPW 65					54 - 57	
	4	6630-2728	UPW 66					35 - 38	
1	36393	6630-2624	UPW 54	Pearce Road (10")	Mintaro Shale	26.4	5	N/A	6.1
	1		UPW 67					25 - 26	
	2	6630-2729	UPW 68					19.6 - 20.6	
	3	6630-2730	UPW 69					15.2 - 16.3	
	4	6630-2731	UPW 70					12.8 - 13.8	
	5	6630-2732	UPW 71					10.85 - 11.36	
	6	6630-2733	UPW 72					8 - 8.5	
1	36394	6630-2627	UPW 57	Clos Clare Winery	Saddleworth Formation	90.6	5.8		3
1	36584	6629-1594	UPW 55	Greenwood Road	Woolshed Flat Shale	90.2	3.7		80
1	36585	6630-2626	UPW 56	Watervale	Saddleworth Formation	140.7	5.8		8
2	41496	6630-2792	CLR 104	Neagles Rock Road (8")	Skillogalee Dolomite	128	11		23
2	41497	6630-2793	CLR 103	Wendouree (10")	Auburn Dolomite	39.9	2.5		4
2	41498	6630-2794	CLR 105	Wendouree (8")	Auburn Dolomite	116.8	5.5		4
2	41499	6630-2795	UPW 59	Pearce Road (8")	Mintaro Shale	100	7		5
2	41500	6630-2796	UPW 60	Watervale Oval (8")	Mintaro Shale	99.2	9.4		3.1
2	41501	6630-2797	UPW 61	Watervale Oval (8")	Mintaro Shale	99.5	7.5		5.2
1	36385	6630-2623	CLR 97	Wendouree	Saddleworth Formation	117.5	4	N/A	4.5
3	1	6630-2839	CLR 143					95 - 98	
3	2	6630-2840	CLR 144					80 - 86	
3	3	6630-2841	CLR 145					71 - 74	
3	4	6630-2842	CLR 146					63 - 66	
3	44454	6630-2793	CLR 103	Wendouree	Saddleworth Formation	60.3	2.5 (Phase 2)	N/A	4.0
3	1		CLR 147					51.5 - 53.5	
3	2	6630-3045	CLR 148					44 - 47	
3	3	6630-3046	CLR 149					35 - 38	
3	4	6630-3047	CLR 150					27 - 30	
3	5	6630-3048	CLR 151					18 - 21	
3	6	6630-3049	CLR 152					9 - 12	
3	44455	6630-2874		Wendouree	Saddleworth Formation	90.1	5.5		4.0
3	44456	6630-2875		Wendouree	Saddleworth Formation	90.0	7.5		5.5
3	44457	6630-2876		Wendouree	Saddleworth Formation	90.0	5.5		5.5
3	44458	6630-2877		Wendouree	Saddleworth Formation	222.0	6.0		5.0
4	53287	6630-3062		Watervale Oval	Auburn Dolomite	99.2	6.5		6.5
4	53293	6630-3063		Watervale Oval	Auburn Dolomite	99.2	4.0		7.2
4	53294	6630-3065		Watervale Oval	Auburn Dolomite	99.2	5.2		5.7
4	53289	6630-3066		Wendouree	Saddleworth Formation	100.0	6.5		7.0
4	53296	6630-3067		Wendouree	Saddleworth Formation	100.0	6.0		7.0
4	53297	6630-3068		Wendouree	Saddleworth Formation	95.5	13.0		7.5



Clare Valley Prescribed Water Resources Area
South Australia

LOCATION OF THE CLARE VALLEY RESEARCH SITES

Department for Water Resources

GDA

Figure 2.10

Publishing Services AV : 200527_009

3. FRACTURE AND MATRIX CHARACTERISTICS

Introduction

The aim of this chapter is to characterise matrix and fracture properties of the aquifers at several sites in the Clare Valley. This basic data will be used in subsequent chapters to develop a conceptual model of the groundwater flow systems in the Clare Valley. In porous media aquifers a basic understanding of groundwater flow can be gained by knowledge of the distribution of hydraulic conductivity, porosity and the storage properties of the aquifer. In fractured rock aquifers (FRA) flow is controlled by the geometry of the system, the fracture apertures (orthogonal distance between fracture walls), fracture density, orientation, connectivity of fractures and porosity of the matrix and fractures. Many of these parameters are difficult to characterise.

In order to obtain an understanding of the range of different properties of the Clare Valley fractured rock aquifers, we have used a several surface and sub-surface techniques. These include fracture mapping of outcrop and core, porosity and permeability testing of cores, down hole geochemical logging and conventional aquifer testing.

Fracture mapping

Core

Rock cores were obtained from a carbonaceous dolomite aquifer at Wendouree Winery (the Auburn Dolomite member of the Saddleworth Formation, Well Unit Number 6630-3068). The coring was performed to map fracture spacings and orientations (Clarke et al., 2001). Cores of up to 3 m (Fig 3.1) were cut using a diamond-drill bit and a split tube core barrel. We assumed that the core represents a vertical scan line and the apparent fracture spacing was measured using a standard tape. Average bedding orientation of 345° 80° W was determined by measurements at exposed formation adjacent to the site. After identifying the bedding plane on the core, the core was rotated so that the bedding plane of the core had the same orientation as the nearby exposed rock. The core was then fixed in position and from this reference point, fracture plane orientations were mapped using a Brunton compass.

The rock cores consisted of a dark grey to black, fine grained dolomite with silty layering. The core is dominated by ragged grains of micron size dolomite that have been modified by low grade metamorphic recrystallisation. Clastic particles occur in minor amounts including quartz, felspar, pyrite and muscovite. Some of the fracture walls have mineral skins that consist of soft grey clay or yellow -brown iron staining. The average fracture spacing for the entire core was 0.33 m, however it is not possible to distinguish whether the fractures are open or closed due to disturbance while drilling and coring. A map of the

Fig. 3.1 *Selected section of core from well number 6630-3068.* *(Photo 47750)*

fracture plane orientation is shown in Figure 3.2. Numerous steeply dipping (>60°) and shallow dipping (20°–40°) fracture planes are present. A distinctive feature of the core is the decrease in frequency of vertical to sub-vertical dipping fractures below 60 m (in particular no bedding plane fractures were observed below this depth).

Outcrop

The best exposure of aquifer for fracture mapping purposes occurs in an outcrop of Saddleworth Formation behind the Foodland® Supermarket at Clare (Fig. 3.3). The outcrop has the approximate dimensions of 50 m in length with a height of approximately 6 m. This outcrop is situated on the west limb of the Hill River Syncline (Fig. 2.3), at this location the aquifers are on end with the majority of fractures and bedding planes orientated vertically.

Outcrop data indicates that five sets of fracture planes are present, four steeply dipping sets and one gently dipping set. The face of the outcrop is a bedding plane fracture that is vertically dipping and strikes at 331°. A second set of vertical fractures strikes at 239°. A set of conjugate shear fractures has the same trend as the second set of vertical fractures with dips approximately 60° E. The only shallow dipping set has a strike of 345° and a dip of 27° W.

The fracture density in outcrop was measured along vertical and horizontal scan lines. This was achieved by simply measuring the interval between all fractures (both open and closed) along the scan line. From this we determined an apparent fracture spacing which was converted to a true fracture spacing by accounting for the angular distortion of the fracture plane to the tape to give the true fracture orientation. The results from the Supermarket outcrop are shown in Figure 3.4, which

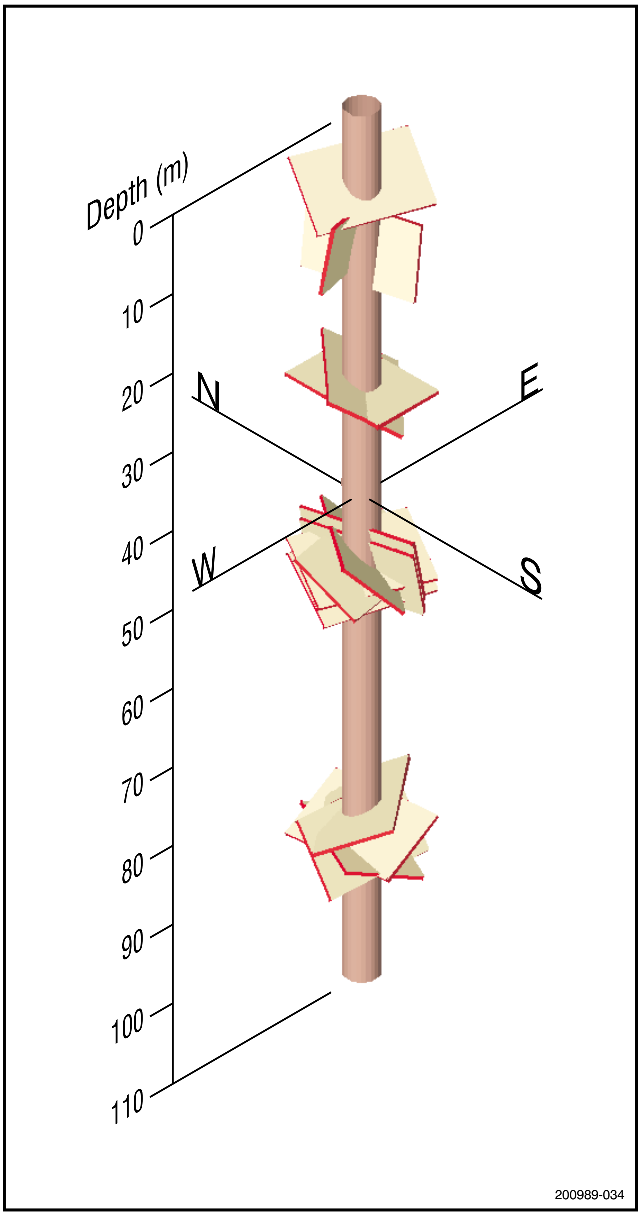


Fig. 3.2 Core mapping at Wendouree showing the orientation of fracture planes intersecting the well.

indicates that the mean spacing of the open fractures is between 0.13 and 0.16 m.

Fracture aperture is the orthogonal distance between adjacent fracture walls. Fracture apertures were measured along horizontal and vertical scan lines using an automotive feeler gauge and lines of known width. Caution needs to be used when interpreting fracture apertures measured from exposed rock as they have been affected by the release of overburden pressure. This generally results in apertures measured in outcrop being larger than those measured in the sub-surface. The distribution of fracture apertures measured at the Supermarket outcrop (Saddleworth Formation) and the Clare Quarry (Gilbert Range Quartzite) is shown in Figure 3.5. The mean fracture aperture distribution ranges from 100–200 μm in the Saddleworth Formation. The largest apertures measured range up to 6 mm and are associated with one of the local shear zones. The mean aperture distribution for the quartzite ranges from 800 μm to 1.2 mm. A conceptual model of the fracture network is presented in Figure 3.6 showing the major orientations measured in outcrop. The decrease in fracture density with depth is assumed.

Electrical conductivity, pH, temperature and radon profiles in open wells

Downhole measurements of electrical conductivity (EC), temperature, radon and pH have provided valuable information for developing a model of the groundwater flow system in the Clare Valley. Interpretation of these logs have provided information on the location of water flowing into the well, the spacing between major fractures as well as partitioning of major flow zones.

A distinctive feature of the many of the research wells in the Clare Valley is that the groundwater is stratified with respect to salinity (Fig. 2.8). In a number of wells, salinity increases by a factor of four over a vertical distance of only 100 m. The unique feature of this stratification is that salinity changes abruptly, with rapid changes (up to 3000 μScm^{-1}) occurring in a step-like fashion over vertical distances of only 1–5 m. At the same depth as the EC discontinuity we often observe an anomaly in the temperature and pH profiles, indicating water of a different composition entering the well (see Fig. 2.8).

We present two field examples from Wendouree (P/N 36385, at a later date this well was converted to a series of nested piezometers, refer to Table 2.2) and Watervale

Fig. 3.3 Outcrop of Saddleworth Formation at Clare Foodland. (Photo 48370)

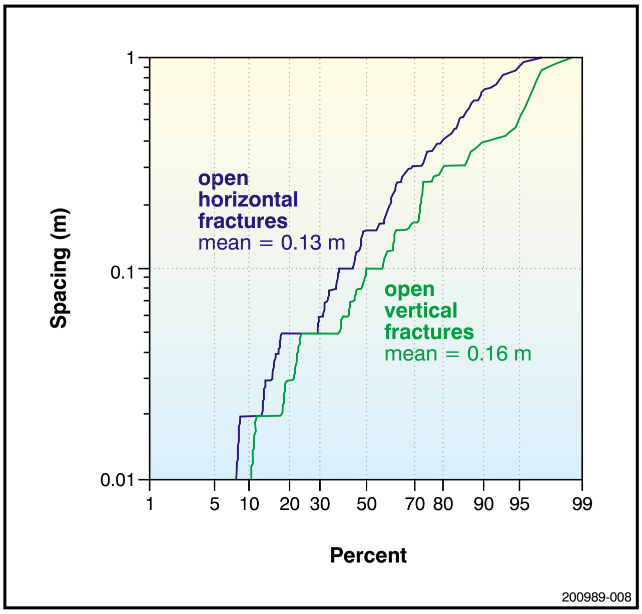


Fig. 3.4 Horizontal versus vertical fracture spacings from the Saddleworth Formation at the supermarket outcrop.

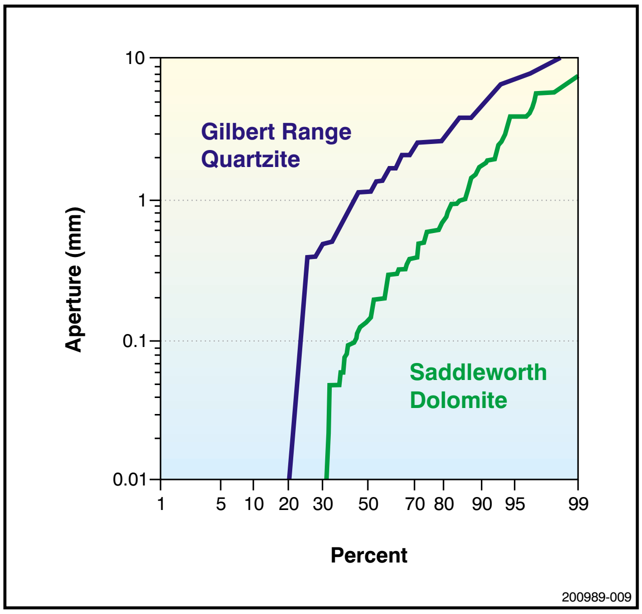


Fig. 3.5 Distribution of fracture apertures measured with a feeler gauge on exposures of the Gilbert Range Quartzite and Saddleworth Formation at the supermarket outcrop, Clare Valley, South Australia. The distributions are truncated at the lower end because of the inability to measure very small fractures.

(6630-2626). At Wendouree Winery (Fig. 3.7) step-like changes in EC of between 300–1500 μS/cm occur over vertical distances of only 1–3 metres. At the same depths a small temperature spike of 0.08–0.12°C is observed. These changes in EC and temperature occur at depths of 36, 52, 77 and 82 m below the water table. We believe these represent locations of major groundwater inflow to the bore via fractures.

Figure 3.8 shows the vertical electrical conductivity and temperature gradient in an uncased well at Watervale (6630-2626). A series of step like increases in EC discontinuities are observed each of which coincide with a temperature anomaly. As discussed previously these are believed to represent the location of major fractures intersecting the well. The highest radon concentration occurs at 75 m with a secondary peak at 45 m depth. If we assume that the concentration of radon within the aquifer is uniform then the radon peak at 75 m must correspond to the highest flow. Very low radon concentration below 80 m, indicates a very low flow rate through the well.

Matrix properties

Porosity and hydraulic conductivity

Laboratory porosity and permeability measurements were conducted on rock cores from several different lithologies and locations throughout the Clare Valley. Porosity was determined by helium porosimetry, mercury bulk volume measurements and gravimetric water content. For helium porosimetry, helium is injected into the rock pores under constant pressure and volume, where the drop in pressure and volume from the reference is measured in an evacuated cylinder. The basic principal for mercury porosimetry is to inject mercury into the pore spaces under incremental pressure changes to produce a distribution of the pore sizes. For gravimetric porosity the rock cores were re-saturated for 20 days and then oven dried at 105°C.

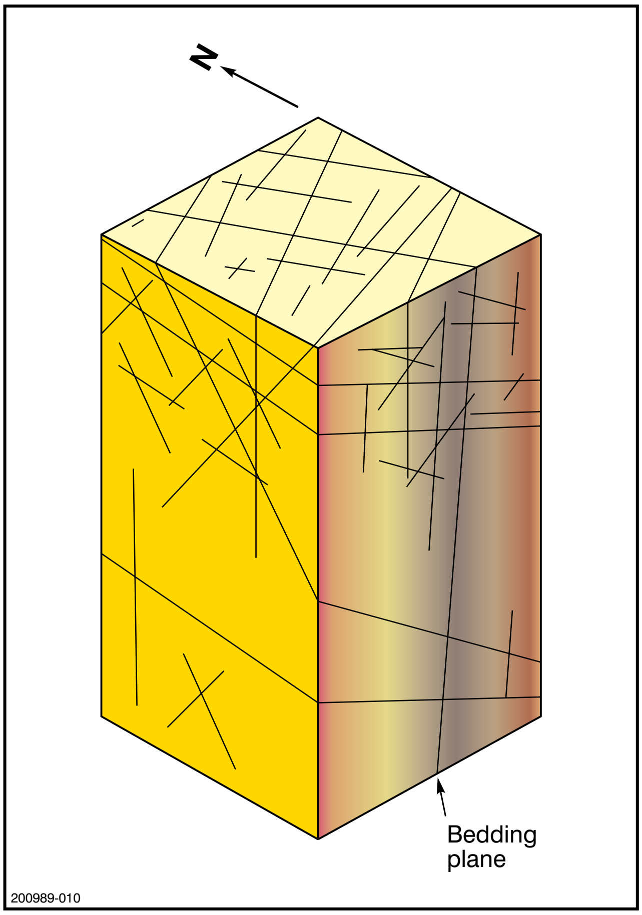


Fig. 3.6 Conceptual model of fractured slate aquifer. Fracture orientations are based on outcrop measurements. The decrease in fracture with depth is assumed.

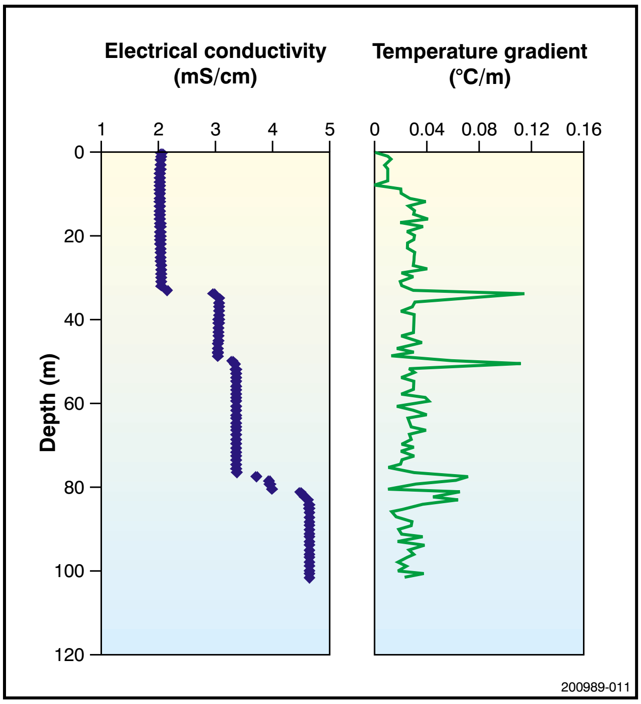


Fig. 3.7 Open well profiles of EC and temperature gradient measured at Wendouree (P/N 36385 prior to piezometers).

Porosity measurements from the helium method are considerably higher than those from the mercury and gravimetric methods (Table 3.1), although no rock core was analysed at the same depth for both methods. The discrepancy between helium and mercury measurements can in part be attributed to helium being more accessible to smaller size pores, which would result in larger values of porosity. The gravimetric porosity most likely represents minimum values due to inherent difficulties in saturating and drying of low permeability materials. We would expect that the effective porosity accessible to water to be less than the helium porosity as nearly all of the dissolved constituents in groundwater are present in ionic form and therefore would not be able to access the same volume of the rock mass as helium.

Hydraulic conductivity determined by triaxial cell tests on the rock cores were extremely low ranging from 2.4 x 10^{-8} to 2.8 x 10^{-13} m/s. This indicates that the matrix porosity is not well connected and therefore minimal advective flow would occur in the matrix. The matrix acts as storage for water and salt and any transport from the matrix to the fractures only occurs via diffusion.

Aquifer pumping tests

Pump-packer tests were performed at Wendouree Winery (6630-2794), and Watervale Hill (6630-2626). The aquifer at each of these sites was isolated by dual packers (Fig. 3.9) into 5 m zones with the packed interval purged with an Grundfos MP1 submersible pump. The pump-packer string was then lowered from the water table in consecutive 5 m intervals down to the bottom of the well. Approximately 20 constant discharge tests of around 120 minutes duration were conducted at each well. The tests were analysed using the Jacobs straight-line method. The draw-down data versus log time indicates a straight line for late time which is consistent with Jacob's approximation of the Theis solution (Cooper and Jacob 1946, Meier et al. 1998). Hydraulic conductivity varies by over 5 and 3 orders of magnitude at Wendouree and Watervale Hill respectively (Fig. 3.10). The profiles indicate that there is large spatial variation of hydraulic conductivity in the aquifer, with values varying by an order of magnitude from consecutive intervals tested. The highest values of hydraulic conductivity generally occur in the top 25 m and decrease with depth, with the exception of the high conductivity values measured between 70–75 m depth at both sites. The high value of hydraulic conductivity between 70–75 m at Watervale Hill corresponds to a high ^{222}Rn measured from the open borehole (refer to Figure 3.8).

Aquifer pumping tests were also conducted at the nested piezometers at Wendouree (P/N 36385 and P/N 44454), Pearce Road (P/N 36392 and 36393), Neagles Rock (P/N 36387 and 36388) and Duncan's Winery (P/N 36391). Results are shown in Figure 3.11. As with the pump-packer tests, the data from the piezometers were analysed using Jacob's straight-line method.

At the Pearce Road nested piezometer site (Fig. 3.11), the hydraulic conductivity varies from 10^{-3} to >10^{2} m/day. The highest hydraulic conductivity value occurs in the piezometer slotted between 33–38 m depth. For this piezometer it was not possible to induce any draw-down

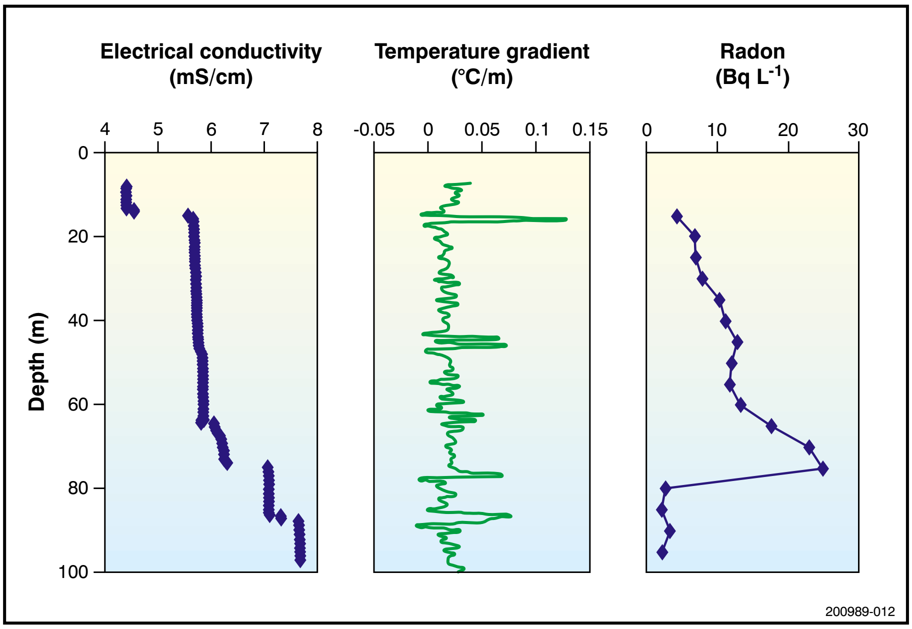


Fig. 3.8 Open well profiles of EC and temperature gradient measured at Watervale (well 6630/2626).

Fig. 3.9 *Pump/packer set-up. (Photo 47940)*

even when pumping at a maximum rate. The piezometers completed in the top 40 m have hydraulic conductivities that are orders of magnitude greater than those completed below 70 m. Hydraulic conductivity at the Neagles Rock piezometers varies by only two orders of magnitude (10^{-2}–1 m/day) with no apparent spatial relationship with depth. Data from Duncan's Winery shows a slight increase in hydraulic conductivity with depth from 1 to 5 m/day. The shallowest piezometer at this site (9–12 m) could not be tested due to an extremely low well yield.

The frequency distribution for all hydraulic conductivity data collected from the pump-packer, piezometer and rock core permeability tests, reveals a range in hydraulic conductivity over 10 orders of magnitude (Fig. 3.12). The relationship between hydraulic conductivity and depth from all of the pumping test results obviously shows a large variation in magnitude, but a general decrease in hydraulic conductivity with depth (Fig. 3.13). The average hydraulic conductivity for the pumping tests is between 1 and 10 m/day.

We can equate our measured values of bulk hydraulic conductivity from the pumping tests to discrete fracture properties from the following:

$$K_b = \frac{\rho g(2b)^3}{12\mu(2B)} \qquad (3.1)$$

where K_b is the bulk hydraulic conductivity over the test interval, ρ is the fluid density, g is acceleration due to gravity, $2b$ is the fracture aperture, μ is the kinematic viscosity and $2B$ is the fracture spacing (which is equal to the length of the test interval if we assume only one fracture). Alternatively, by rearranging equation 3.1 we can estimate an equivalent fracture aperture ($2b_{eq}$) from the pumping test:

$$2b_{eq} = \left(\frac{12\mu(2B)K_b}{\rho g}\right)^{1/3} \qquad (3.2)$$

If we assume that all groundwater flow occurs in fractures and there is no flow in the matrix, then the effective porosity for groundwater flow is the ratio of fracture aperture to fracture spacing (i.e. $2b/2B$). Hence, the groundwater velocity through the fractures (v) can be determined by combining equation (3.1) with Darcy's Law to give

$$v = \frac{\rho g}{12\mu}\left(2b_{eq}\right)^2 \frac{dh}{dx} \qquad (3.3)$$

The average value of bulk hydraulic conductivity obtained from consecutive pump-packer tests over 5 m intervals at Wendouree is approximately 15 m/day in the top 50 m. If we assume there is only one fracture over each test interval (i.e. $2B$ = 5 m), then the equivalent fracture aperture ($2b_{eq}$) is ~ 1500 μm and the groundwater velocity through the fracture is 560 m/day (dh/dx = 0.005). However if we assume a more realistic fracture spacing of 0.33 m and assume all the fractures are identical, then this would convert to an equivalent aperture of 460 μm and a flow velocity through the fracture of 50 m/day.

For depths greater than 50 m, we use an average hydraulic conductivity of 0.001 m/day and obtain $2b_{eq}$ = 46 μm and v = 0.5 m/day ($2B$ = 5 m). However if we assume a fracture spacing of 0.33 m, then we calculate $2b_{eq}$ = 18 μm and v = 0.08 m/day.

Table 3.1 *Porosity and permeability measurements on rock cores from the Clare Valley. All helium and mercury analysis was performed at CSIRO Petroleum Resources, except the sample denoted with * which was analysed at the University of Texas at Austin. Gravimetric porosity was analysed at CSIRO Land and Water.*

Site	**Well ID**	**Depth** (m)	**Helium porosity** (%)	**Mercury porosity** (%)	**Gravimetric porosity** (%)	**Water permeability** (m/s)	**Gas permeability** (m/s)
Watervale Oval	41501	26.3	1.06				2.8×10^{-13}
		55.1	2.36				
		76.7		0.19			
Wendouree	41498	8.2	4.76				
		24.51			0.49		
		24.55		0.09	0.45		
		70.5			0.69		
		70.8		0.26			
Pearce Road	41499	8.7	2.24				6.3×10^{-13}
		9.25			0.4		
		9.29			0.38		
		26.0		0.4*			
Neagles Rock	41496	40.4	13.1			6×10^{-10}	2.4×10^{-8}
		40.4	12.7				

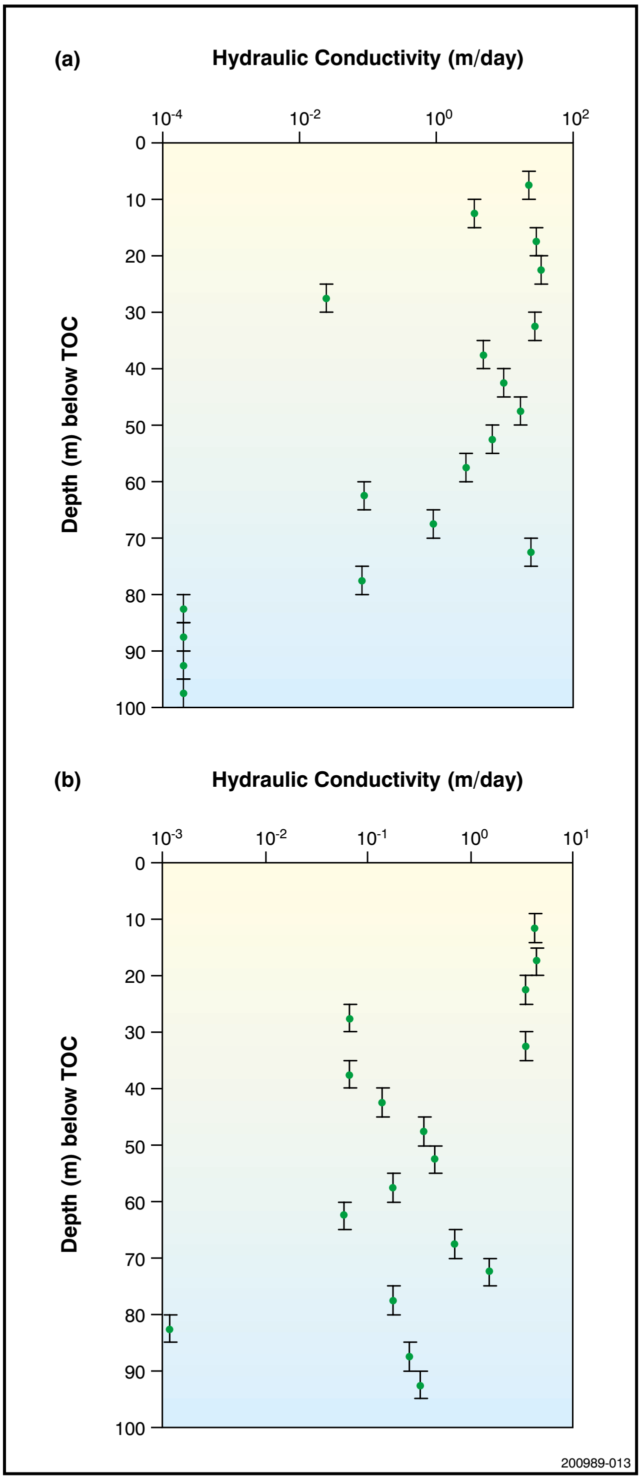


__Fig. 3.10__ Results from the pump/packer testing of hydraulic conductivity at (a) Wendouree 6630/2794 and (b) Watervale Hill (6630/2626).

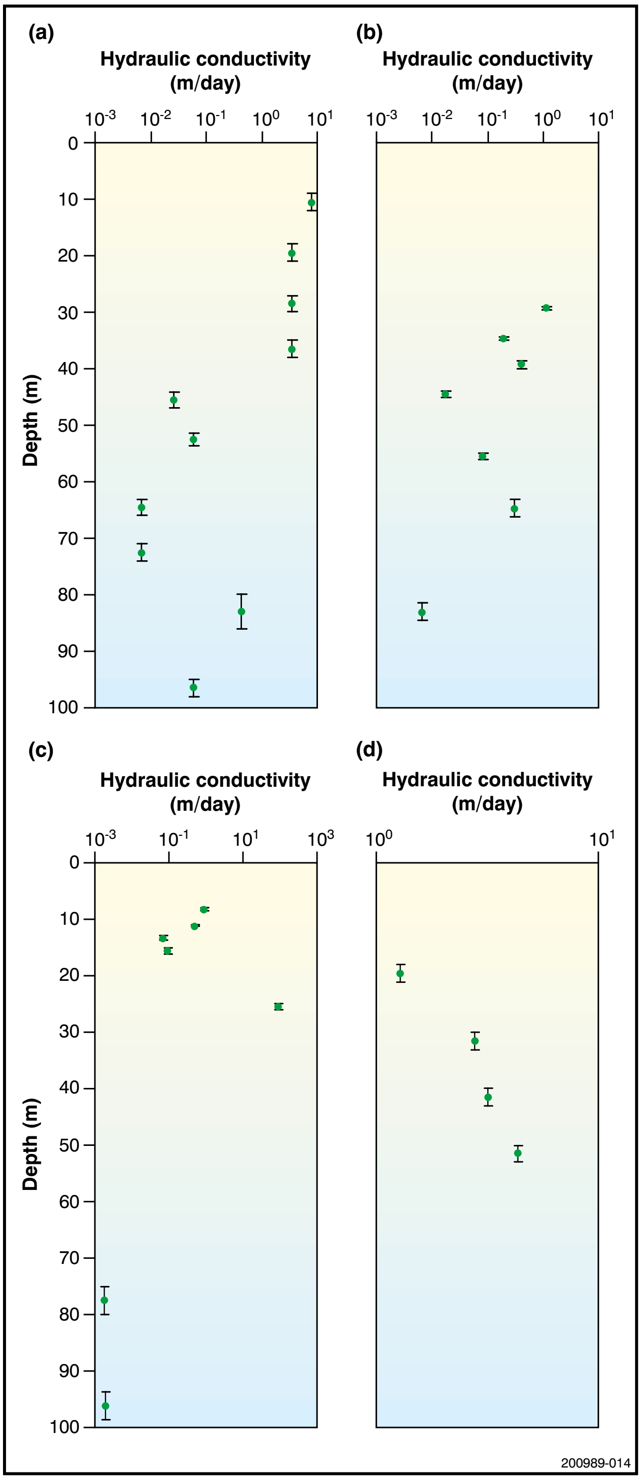


__Fig. 3.11__ Hydraulic conductivity values measured on the nested piezometer sites (a) Wendouree Winery, (b) Neagles Rock, (c) Pearce Road and (d) Duncans Winery.

Connection test

A pump-packer test was conducted at the Wendouree field site (6630-2794) at a depth of between 60–65 m and changes in water levels were measured in the piezometer nest located 14 m to the north-east (Fig. 3.14). The packed interval was pumped at a rate of 10 L/min for 460 minutes and water levels were measured in the piezometer nest before and at the end of the pumping test.

The pumping test indicates a strong vertical connection between the top four piezometers (<40 m) and the packed interval between 60–65 m. The piezometers located between 45 and 55 m display a lower degree of connection with the pumped well, while increased connection is recorded at the piezometers located at 65 and 75 m. The bottom two piezometers record no connection with the pumped interval. This suggests a high angle fracture connects 60 m depth in the pumped well with shallow

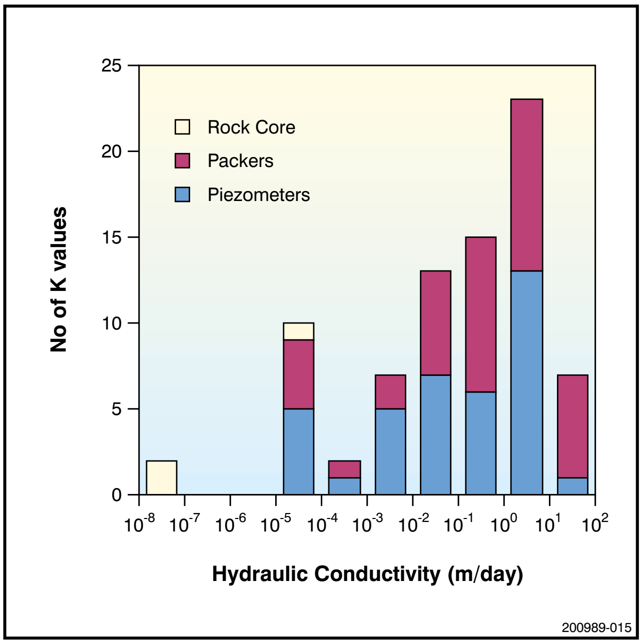


Fig. 3.12 Histogram showing the distribution of hydraulic conductivity in the Clare Valley.

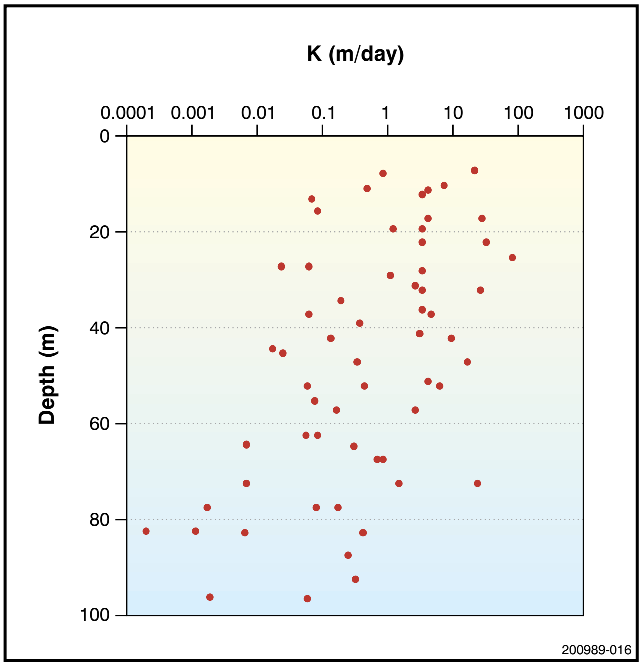


Fig. 3.13 Hydraulic conductivity versus depth for all pumping test measurements in the Clare Valley.

(<40 m) depth at the neighbouring piezometers. The lowest two piezometers are hydraulically isolated.

Anisotropy

In most cases fractures have a preferred orientation which may cause hydraulic conductivity to vary with direction. This will mean that the groundwater flow will often not be in the same direction as the hydraulic gradient (Schreiber et al., 1999). In principle, the flow direction can differ from the direction of maximum hydraulic gradient by up to 90°. Where there is a strong preferred orientation of fractures, the draw-down cone can be distorted from the concentric shape normally observed in porous media aquifers. However in many situations it is not possible to map the draw-down contours due to a lack of observation wells.

In order to assess anisotropy an 8 hour aquifer test was performed at both Watervale Oval and Wendouree. The production bore for the Watervale Oval test (Well 6630-2796) was pumped at 2 L/s for the first 100 minutes and then at 1.5 L/s for the remaining 380 minutes. Total draw down was measured in the production well and 4 observation wells at the site (Fig. 3.15a). A plot of total draw-down versus radial distance from the pumped well (Fig. 3.15b) reveals a relatively smooth piezometric level during pumping. This curve suggests that at least in the direction of the observation wells that the cone of depression is similar to what we might expect to observe in a porous media aquifer with low transmissivity. From the results of this test we can not infer any conclusions about anisotropy at this site. A larger array of observation wells may have detected some degree of anisotropy but the cost of drilling extra wells was prohibitive.

The aquifer test at Wendouree indicates a different response from the test at Watervale Oval. The production well was pumped at 3 L/s for 100 minutes and then at 7 L/s for the next 320 minutes and drawdown was measured in the production well and 7 observation wells (Fig. 3.16a). A plot of radial distance from the pumping well versus total draw down (Fig. 3.16b) indicates that at the same distance from the production well we observe different amounts of total draw down. A line of best fit through the observation wells reveals a large scatter of data. Observation wells that are orientated southwest and northeast from the production well, have the largest drawdown with respect to their radial

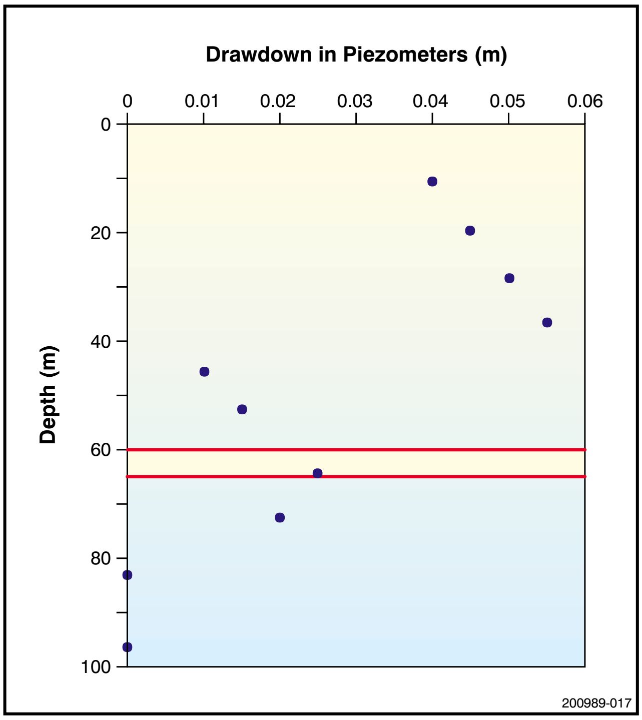


Fig. 3.14 Connection test at Wendouree Winery (lines delineate packed zone where pump was located).

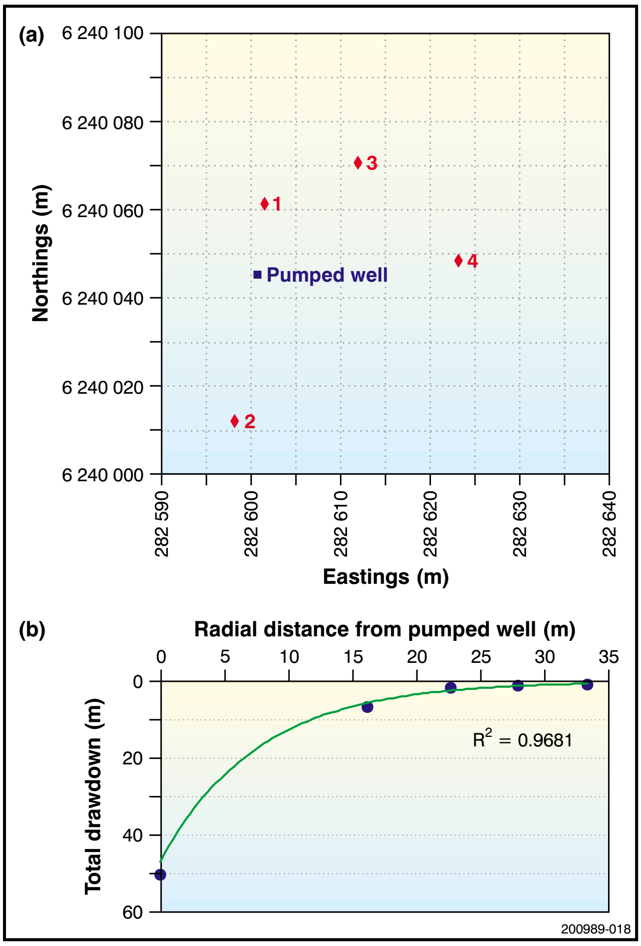


Fig. 3.15 (a) Location of wells at Watervale Oval (datum GDA 94) (b) Drawdown versus radial distance at Watervale Oval.

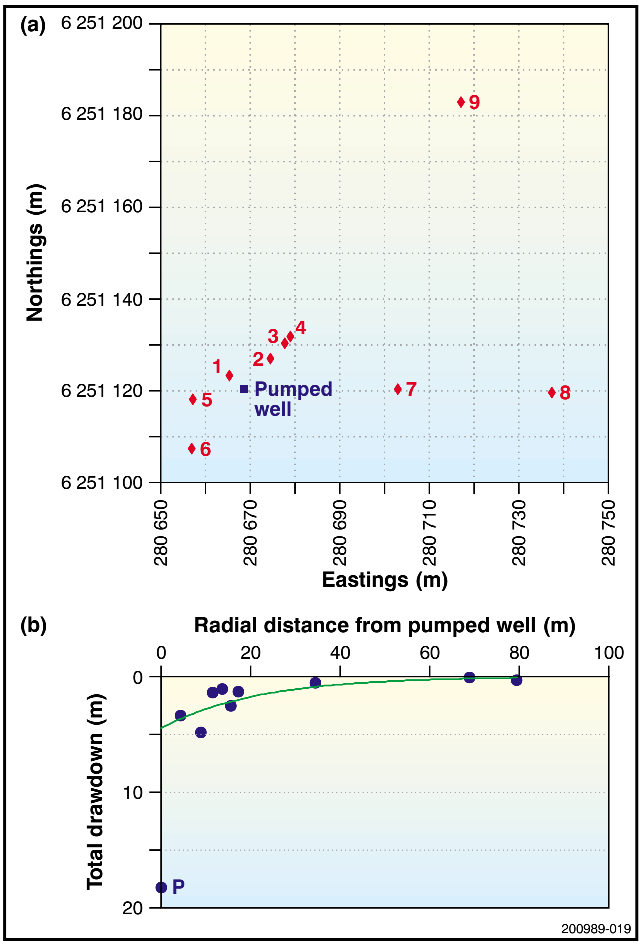


Fig. 3.16 (a) Location of wells at Wendouree Winery (datum GDA 94) (b) Drawdown versus radial distance at Wendouree Winery.

distance from the production well. Observation wells east of the production well have relatively smaller drawdown than the SW-NE orientated wells. The results from this test indicate that there is a preferred anisotropy in the NE-SW direction relative to E-W. However it is possible that the direction of major draw down was not recorded due to a lack of observation wells. For example there were not any observation wells located in the direction of the major bedding plane fracture measured from rock cores and the Supermarket outcrop at this site (i.e. NW–SE).

Summary of fracture and matrix properties

The fractured rock aquifers of the Clare Valley are extremely variable in terms of their physical characteristics. Groundwater flow is confined to the fracture network. Hydraulic conductivity varies over many orders of magnitude with a general decrease in values with depth. Major fractures that intersect open wells have been identified by rapid changes in electrical conductivity, temperature, pH and radon peaks.

The fractured rock aquifers can be broadly divided into two major flow zones. An upper highly transmissive system characterised by high rates of groundwater flow and larger apertures and a lower less transmissive flow system, characterised by reduced fracture apertures and relatively slower horizontal groundwater flow rates.

Field mapping and hydraulic tests indicate that the groundwater flow system is anisotropic. However at this stage it is not possible to predict the major directions of anisotropy and therefore results of this part of the study cannot be transferred into water allocation policy.

4. GROUNDWATER FLOW

Introduction

In porous media aquifers, groundwater flow rates can be determined by using Darcy's Law with estimates of hydraulic conductivity and hydraulic gradient. Alternatively, if groundwater ages can be determined using environmental tracers (^3H, ^{14}C, ^{36}Cl and CFCs), then estimates of vertical and horizontal flow rates can be determined. However in fractured rock aquifers neither method is straightforward. Hydraulic conductivity is highly variable and so estimates of flow velocity derived from Darcy's Law are unreliable. Similarly, matrix diffusion complicates interpretation of environmental tracers. This chapter develops new methods for estimating groundwater flow and recharge rates, that combine both hydraulic and tracer approaches.

Groundwater ages and depth of circulation

In unconfined aquifers, vertical profiles of groundwater age have been widely used to estimate rates of vertical groundwater flow, and hence aquifer recharge (Cook and Böhlke, 1999). However, in fractured rocks, groundwater ages obtained with environmental tracers usually do not represent the hydraulic age of the water within the fractures, because of the effects of matrix diffusion on tracer concentrations (Grisak and Pickens, 1980). Matrix diffusion is the process where solutes move from the fractures into the rock matrix or out of the matrix into the fractures depending on the direction of concentration gradient. The amount of matrix diffusion that occurs will depend on the fracture apertures and spacing, the matrix porosity and the diffusion coefficient of the tracer in the rock matrix. In the case of vertical, planar, parallel fractures, environmental tracer concentrations can be related to aquifer recharge rates provided that information is available on these fracture and matrix properties.

Where fracture orientations are more complex, groundwater ages may no longer provide information on aquifer recharge, but may provide valuable information on the depth of circulation of the groundwaters (Fig. 4.1). Groundwater ages beneath horizontal or sub-horizontal fractures should be related to the leakage rate beneath the fracture, rather than the aquifer recharge rate. This situation is analogous to groundwater ages obtained from confining beds (or confined aquifers), which may reflect the rate of leakage through the confining layer, rather than the recharge rate to the overlying aquifer.

Figure 4.2 depicts vertical profiles of carbon-14 obtained from the piezometer nest at Wendouree (PN 36385 and 44454). Step-like decreases in ^{14}C activity versus depth coincide with similar discontinuities in the electrical conductivity profile (Fig. 3.7), and are believed to reflect the locations of at least some of the major fractures that intercept the well. The ^{14}C activities represent relative groundwater ages, and hence the profile depicts abrupt increases in age of the groundwater below major fractures. It shows that progressively less water is moving through the aquifer below each of these fractures. At Wendouree, the ^{14}C data indicates rapid circulation of groundwater in the upper 40 m of the aquifer, with relatively little flow between 40 m and 60 m depth, and less still below 60 m (Love, et al. 1999). At Pearce Road, ^{14}C profiles show that rapid circulation of groundwater occurs in the upper 30 m of the aquifer (Fig. 4.3).

Vertical flow rates and aquifer recharge

There are relatively few established and reliable methods for estimating recharge rates in fractured rock aquifers. For this project, a number of new methods were developed, and the results obtained with these are described below.

Groundwater age gradients

Cook and Simmons (2000) used ^{14}C and CFC-12 concentrations to estimate vertical flow rates at the Pearce Road site, by assuming that flow occurred through vertical, planar, parallel fractures, and that matrix properties were spatially uniform. In the case of a constant source, conservative tracer, subject to radioactive decay, concentrations within the fracture (c) are related to the vertical flow rate within the fractures, V_w, by (Neretnieks, 1981)

$$\ln\left(\frac{c}{c_0}\right) = \frac{-\lambda z}{V_w}\left[1+\frac{\theta_m D^{1/2}}{b\lambda^{1/2}}\tanh(BD^{-1/2}\lambda^{1/2})\right] \qquad (4.1)$$

where c_0 is the initial concentration, b is the fracture half-aperture, B is the fracture half-spacing, θ_m is the matrix porosity, D is the diffusion coefficient within the matrix, λ is the decay constant, and z is depth. Equation 4.1 can be used to relate ^{14}C concentrations within fractures to vertical groundwater flow rates, where $\lambda = 1.21 \times 10^{-4}$ yr^{-1} is the decay constant for ^{14}C. Equation 4.1 also approximates the relationship between V_w and CFC-12 concentration, where the decay constant, λ, is replaced by the exponential growth rate for CFC-12, $k \approx 0.06$ yr^{-1}. Expressed in terms of apparent age gradients, equation 4.1 becomes

$$V_w = \left[1+\frac{\theta_m D^{1/2}}{b\lambda^{1/2}}\tanh(BD^{-1/2}\lambda^{1/2})\right] \Big/ \left[\frac{\delta t_a}{\delta z}\right] \qquad (4.2)$$

where t_a is the apparent ^{14}C or CFC-12 age. The mean volumetric flow rate through the fracture, Q_v is then given by

$$Q_v = V_w \frac{b}{B} \qquad (4.3)$$

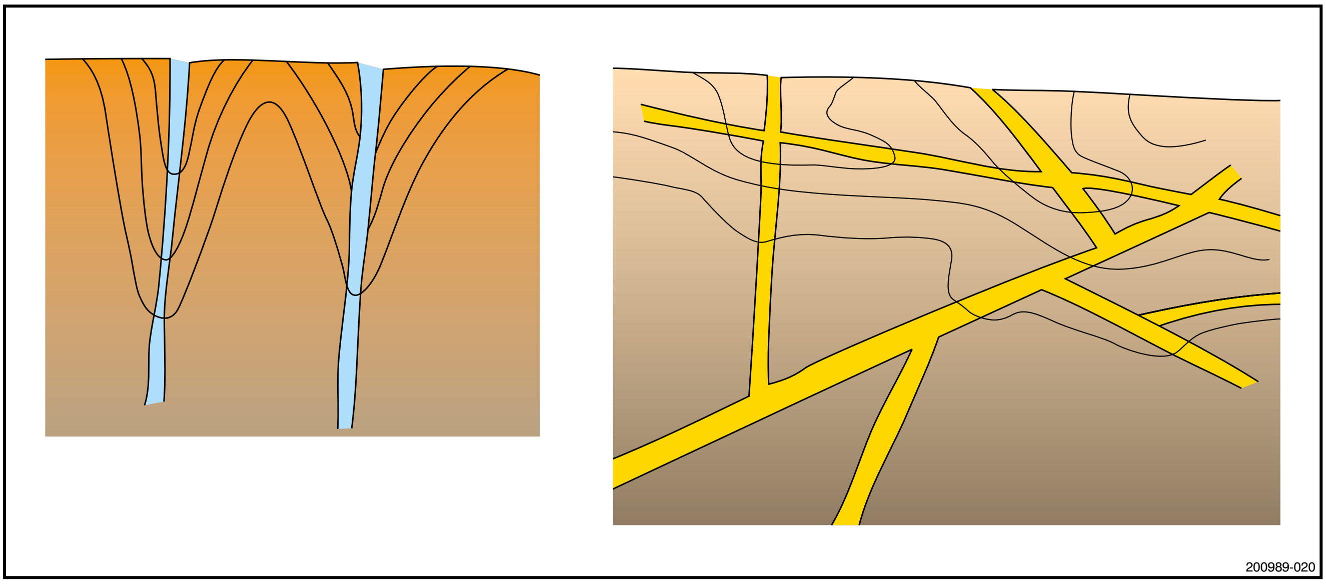


Fig. 4.1 *Schematic representation of environmental tracer concentrations in fractured rocks. The contours are lines of equal tracer concentrations. Groundwater moves through fractures, and carries dissolved solutes with it. However, solutes also move between the fractures and the matrix by diffusion processes (termed 'matrix diffusion'). For simple planar, parallel fractures, apparent groundwater ages can be converted to water ages if information on fracture spacing, aperture and matrix porosity and diffusion coefficent is known. For more complex fracture geometries, corrections for matrix diffusion cannot be accurately made, although environmental tracer concentrations still contain valuable information on the depth of circulation of the groundwaters.*

At or close to the water table, we assume that Q_v is the aquifer recharge rate (R). Figure 4.3 shows vertical ^{14}C and CFC-12 profiles obtained from a piezometer nest at Pearce Road. The vertical CFC-12 age gradient apparent in the upper six piezometers (i.e., between 5 and 25 m depth) is approximately $\delta t_a/\delta z = 0.1$ yr m^{-1} (2 years over 20 m). A mean fracture spacing of $2B = 0.16$ m and hydraulic conductivity of $K = 0.1$ m d^{-1} was assumed, based on outcrop measurements and single-well pumping tests, respectively. These values together imply a mean fracture aperture of $2b = 64\ \mu$m (Equation 3.1). Values for matrix diffusion coefficient and matrix porosity were assumed to be $D = 10^{-4}$ m^2 yr^{-1} and $\theta_m = 0.02$, respectively. Thus $V_w = 256$ m yr^{-1}, and hence Q_v (or R) = 102 mm yr^{-1}. A sensitivity analysis on this calculation showed that estimated recharge rate is particularly insensitive to the estimated hydraulic conductivity. In particular, an order of magnitude error in the hydraulic conductivity would result in a factor-of-two error in the estimated fracture aperture and minimum fracture velocity, but less than a 5% error in the estimated recharge rate. The estimated recharge rate is sensitive to the fracture spacing, and to the matrix porosity and matrix diffusion coefficient (Cook and Simmons, 2000).

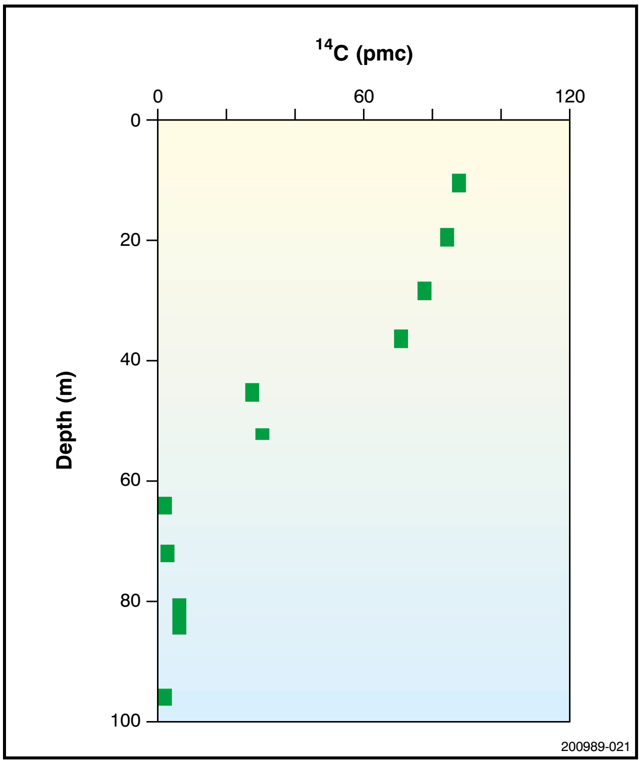


Fig. 4.2 *Carbon-14 profile obtained from piezometer nest at Wendouree (PN 36385 and 44454).*

Below 30 m, the hydraulic conductivity is approximately 10^{-3} m d^{-1}. Using the same fracture spacing of $2B = 0.16$ m, gives $2b = 14\ \mu$m. The decrease in ^{14}C activity of 75 pmc occurs over a depth of approximately 50 m, giving $\delta t_a/\delta z = 230$ yr m^{-1}. Assuming $D = 10^{-4}$ m^2 yr^{-1} and $\theta_m = 0.02$ gives $V_w = 0.5$ m yr^{-1}, and hence Q_v = 0.05 mm yr^{-1}. Of course, at these depths it may no longer be reasonable to assume that the fracture spacing is similar to that measured in near-surface exposures. If, instead, we assume $2B = 1.0$ m, this gives $2b = 25$ m, $V_w = 3.1$ m yr^{-1}, and $Q_v = 0.08$ mm yr^{-1}. In either case, the calculation confirms that the rate of leakage below approximately 30 m depth is very low.

^{3}H and ^{36}Cl Balance

Cook and Robinson (2002) used an alternative approach, to estimate vertical flow rates in the shallow flow system at Pearce Road and Wendouree, based on ^{3}H and ^{36}Cl

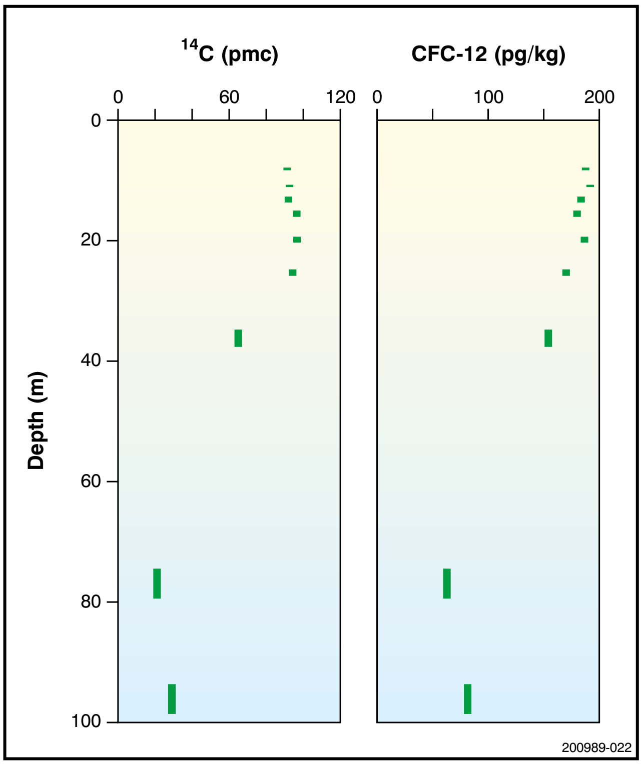


Fig. 4.3 Vertical profiles of ^{14}C and CFC-12 obtained at Pearce Road (PN 36392 and 36393).

concentrations. High ^{3}H and ^{36}Cl concentrations reflect rainfall from the 1950s and 1960s, when thermonuclear bomb testing was being carried out. Because ^{3}H is lost during evapotranspiration with negligible fractionation, while ^{36}Cl is retained within the soil, comparison of ^{3}H and ^{36}Cl concentrations allows estimation of the evaporation rate, and hence the aquifer recharge rate. Assuming a system of identical, vertical, planar, parallel fractures, both ^{3}H and ^{36}Cl profiles in fractures should exhibit an asymmetrical bulge-shaped profile, with highest concentrations close to the land surface, but persistent, above-background concentrations extending to some depth. The depth of the maximum concentration of ^{3}H and ^{36}Cl need not coincide, because highest atmospheric concentrations of these tracers were off-set in time, and also because their diffusion coefficients differ slightly. The depths of the maximum concentrations of both tracers will increase as the matrix porosity and tortuosity decreases, and as the fracture spacing increases. The value of the maximum concentration will increase as the tortuosity increases and as the fracture spacing decreases. The value of the maximum concentration is independent of the matrix porosity, and both the depth and value of the maximum concentration are independent of the fracture aperture. However, variations in matrix porosity, tortuosity and fracture spacing affect ^{3}H and ^{36}Cl profiles in exactly the same way, so that the ratio of ^{3}H to ^{36}Cl peak depths is unaffected by fracture and matrix parameters. In contrast, increases in recharge rate result in increases in depth of the maximum concentrations of both ^{3}H and ^{36}Cl, but result in a decrease in ^{36}Cl concentrations, without affecting the value of the maximum ^{3}H concentration. Thus comparison of ^{3}H and ^{36}Cl concentration may allow the recharge rate to be estimated, without knowledge of fracture and matrix parameters.

Figure 4.4 compares measured ^{3}H and above-background ^{36}Cl concentrations with those produced from numerical simulations for flow through vertical, planar, parallel fractures. Apertures of different fractures were assumed to follow a log-normal distribution. Parameters for the simulations were matrix porosity, matrix diffusion coefficient, fracture spacing, dispersion coefficient within the fracture, vertical head gradient, and mean and variance of the fracture aperture distribution. For each simulation, parameters were chosen to optimise the fit to the ^{36}Cl data, and then these same parameters were used to simulate ^{3}H transport. For Pearce Road, acceptable fits to the ^{36}Cl data could be obtained with recharge rates that varied between 30 and 75 mm yr^{-1}, but only those simulations with recharge rates between 60 and 75 mm yr^{-1} also provided an acceptable fit to the ^{3}H data. For Wendouree (not shown here), acceptable fits to both the ^{36}Cl and ^{3}H data were also possible only for simulations with recharge rates between 60 and 75 mm yr^{-1}. Thus this approach constrained recharge rates to within these relatively narrow limits (Cook and Robinson, 2002).

Chloride mass balance

Groundwater recharge rates within fractured rock systems can also be estimated using a chloride mass balance approach. If the chloride that occurs in groundwater is deposited by dust particles and precipitation, with negligible amounts contributed from rock weathering or anthropogenic sources (i.e. fertilisers), and negligible runoff, then under steady state conditions

$$C_P P = C_G R \qquad (4.4)$$

where C_P is the chloride concentration of precipitation (including dust particles), C_G is the chloride concentration in groundwater, P is the precipitation rate and R is the recharge rate.

In the Clare Valley, however, it is unlikely that steady state conditions exist (Love et al. 1999). Much of the area has been cleared of native Eucalyptus vegetation within the past 100 years, and so recharge rates will have increased (Allison et al., 1990). Within fractured systems, the time required for leaching of the stored salt and establishment of the new equilibrium will largely be determined by the rate of diffusion of salts from the matrix into the fractures. If fractures are widely-spaced, then this time can be extremely long.

Cook et al. (1996) noted that the time for equilibrium of solutes between the matrix and the fractures is given by

$$t \geq \frac{0.4911\,(2B)^2}{D} \qquad (4.5)$$

where t is time (years); $2B$ is the fracture spacing and D is the effective diffusion coefficient for the matrix (assumed to be 10^{-4} m^2/yr). Therefore, for $2B = 1$ m, $t \geq 5000$ years and for $2B = 0.1$ m, $t \geq 50$ years.

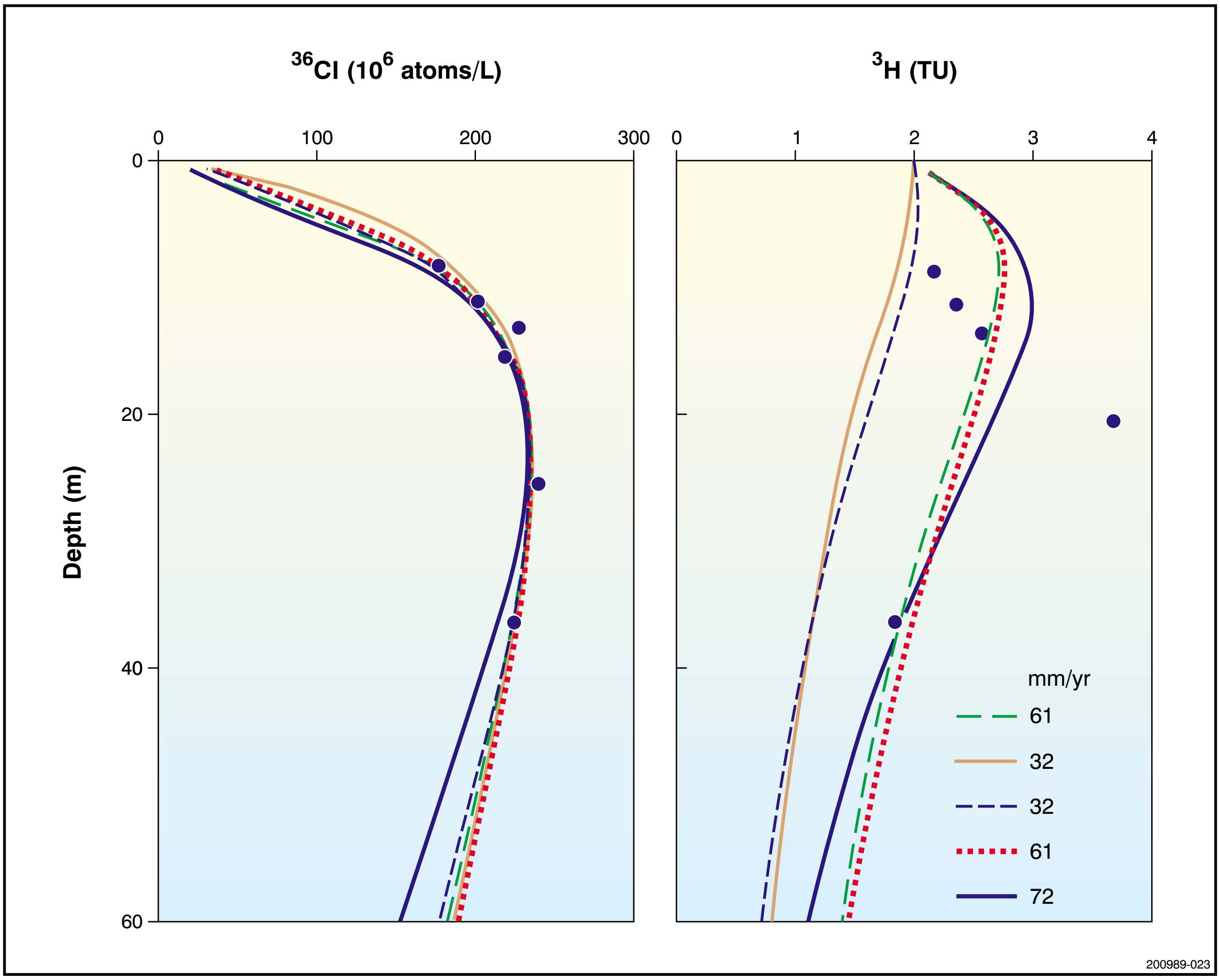


Fig. 4.4 *Comparison of field data of ^{3}H and above-background ^{36}Cl concentrations at Pearce Road with results of numerical simulations.*

It is likely that much of the chloride present within the groundwaters in the Clare Valley represents residual chloride which pre-dates clearing of the native vegetation. The spatial variability of salinity that is observed throughout the valley may represent variable rates of leaching of this stored salt. The increases in electrical conductivity with depth observed at sites throughout the valley may thus reflect a decrease in leaching with depth, perhaps due to lower flow rates, and larger fracture spacings.

If this hypothesis concerning residual salt is correct, then the chloride mass balance should provide a lower bound on the recharge rate at each site. At Pearce Road, the chloride concentration in groundwater above 12 m depth is approximately 65 mg L^{-1}, giving a recharge rate of $R > 50$ mm yr^{-1} ($P = 650$ mm yr^{-1}, $C_P = 5$ mg L^{-1}). At Wendouree, the chloride concentration in the shallow piezometers is approximately 350 mg L^{-1}, giving a recharge rate of $R > 9$ mm yr^{-1}. The higher measured chloride concentration at Wendouree may indicate a slower rate of flushing of salt, which might result from a larger fracture spacing or smaller matrix diffusion coefficient.

If we assume that the concentration of chloride in the matrix and the fractures are in equilibrium at depth then an estimate of pre-clearing recharge can be determined. Based on groundwater dating many deeper sections of the Clare Valley aquifers have not received modern day recharge. Chloride concentrations in these groundwaters are in the range of 2000–5000 ppm, which corresponds to a pre-clearing recharge rate <2 mm/year if rainfall was the same as present.

Horizontal flow velocities

Introduction

The traditional approach for estimating horizontal flow velocities, involves calculating flow rates as the product of the hydraulic head gradient and the aquifer hydraulic conductivity (Darcy's Law). However, the high spatial variability of hydraulic conductivity in fractured rock aquifers (see Chapter 3) makes this method unreliable. Also, hydraulic conductivity usually indicates the degree of fracturing in the vicinity of the well, whereas regional flow rates are largely determined by the degree of connection of fractures over large distances.

For example, at the Pearce Road site, the regional potentiometric surface slopes east to west, with a gradient of approximately 4 x 10^{-2} (40 m in 1 km). The hydraulic conductivity of the piezometer screened between 35 and 38 m is more than 10^2 m d^{-1} (transmissivity > 3 x 10^2 m^2 d^{-1}), giving a flow rate in excess of 12 m^2 d^{-1} for this zone.

Horizontal groundwater flow rates and aquifer recharge rates are related by

$$Q_H = Rx \qquad (4.6)$$

where Q_H is the horizontal flow rate; x is the distance to the flow divide, measured along the flow line; and R is the mean recharge rate along that flow line. Even ignoring flow through other intervals, based on a distance to the flow divide of only 1.5 km, the flow rate of 12 m^2 d^{-1} between 35 and 38 m depth would suggest that recharge over this distance must be 2900 mm/yr, which is five times higher than the mean annual rainfall. Thus Darcy's Law cannot be used for estimating regional flow rates at this site.

An alternative approach, which has been used successfully both in porous media and fractured rocks, is point dilution (Drost et al., 1968). This method involves isolating a section of a well using packers, adding a tracer to the groundwater within the packed-off section, and monitoring the decrease in tracer concentration with time as it is diluted by the ambient groundwater flow. Usually a recirculation pump or impeller system is used to ensure that the groundwater within the packed-off zone is well-mixed. Provided that the zone is well-mixed, then the concentration should decrease exponentially with time, according to

$$c(t) = c_\infty + (c_0 - c_\infty) \exp\left(\frac{-2qt}{\pi r}\right) \qquad (4.7)$$

where c_∞ is the concentration in ambient groundwater, c_0 is the concentration in the well immediately after injection of the tracer, q is the groundwater flow rate, r is the well radius, and t is time. While this method has been used with some success, it does require the use of inflatable packers. Also, ensuring that the packed-off zone is well-mixed is not always easy. Instead, we have designed two slight modifications of the point dilution method, which do not require the use of inflatable packers, and so are more suitable for regional investigations.

Radon method

^{222}Rn is a radioactive inert gas, with a high solubility in water, and a half-life of approximately 3.8 days. It is generated within the aquifer by radioactive decay of uranium and thorium series isotopes in the aquifer matrix, and its concentration in groundwater will depend on the aquifer mineralogy and pore space geometry. If the horizontal flow rate is sufficiently high, then the concentration of radon measured in the screened interval of an unpurged piezometer and in a purged piezometer should be the same, and equal to that within the aquifer. If the horizontal flow rate is very low, the water in the well will be stagnant, and so most of its radon will have decayed to background. Thus the ratio of radon concentration within the unpurged well to that in the aquifer (measured after purging the well) will depend on the amount of radioactive decay which occurs as the water moves through the well under natural flow conditions. It can be shown that, for a perfectly mixed well, the groundwater flow rate, q, (m yr^{-1}), is given by

$$q = \frac{c}{c_0 - c}\,\frac{\lambda \pi r}{2} \qquad (4.8)$$

where c and c_0 are the radon concentrations in the well and in the aquifer respectively, r is the well radius, and λ is the decay constant for radon [0.18 day^{-1}]. Figure 4.5 shows the solution to this equation for a well radius of 0.1 m. It suggests that radon concentrations should enable horizontal flow velocities to be estimated, provided that these are less than approximately 40 m yr^{-1}.

Figure 4.6 shows radon concentrations measured at Pearce Road both before ($^{222}Rn_{UP}$) and after purging ($^{222}Rn_P$) on 17 November 1997. Before purging, the highest radon concentration (6.3 Bq L^{-1}) occurred in the piezometer screen centred at 11 m depth, with only background concentrations (<0.5 Bq L^{-1}) below 40 m. Radon concentrations for purged wells increase from 10–13 Bq L^{-1} above 12 m depth, to over 80 Bq L^{-1} below 30 m. The increase in radon concentration with depth presumably reflects a change in mineralogy.

The ratio of the radon concentration measured before purging to that measured after purging ranges between zero and 0.63. Horizontal flow velocities have been calculated from these ratios. Unpurged and purged radon concentrations have been taken to represent c and c_0 in Equation 4.5. Slight corrections to this equation have been made to account for the volume occupied by the gravel pack in the piezometer nests, and for the volume occupied by PVC pipe. The ratio c/c_0 is greater than 0.1 above 20 m depth, with a maximum value of 0.63 at 11.1 m. The corresponding flow rate is estimated to be greater than 0.5 m yr^{-1} above 20m depth, and as much as 9.7 m yr^{-1} at

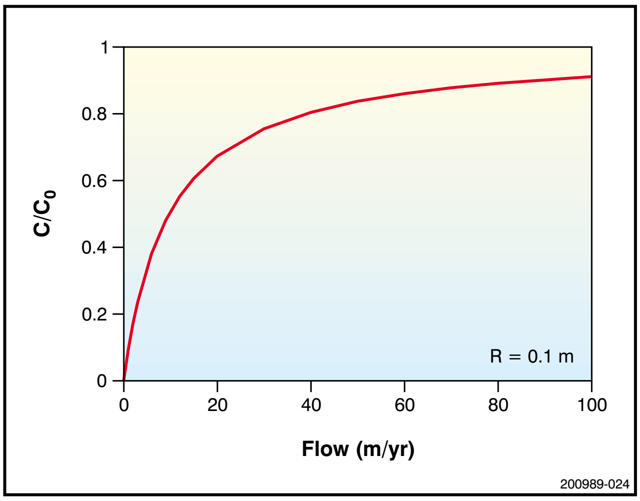


Fig. 4.5 Theoretical relationship between radon concentration in an unpurged well, relative to that in the aquifer, and groundwater flow rate for a bore radius of 0.1 m (from Cook et al., 1999).

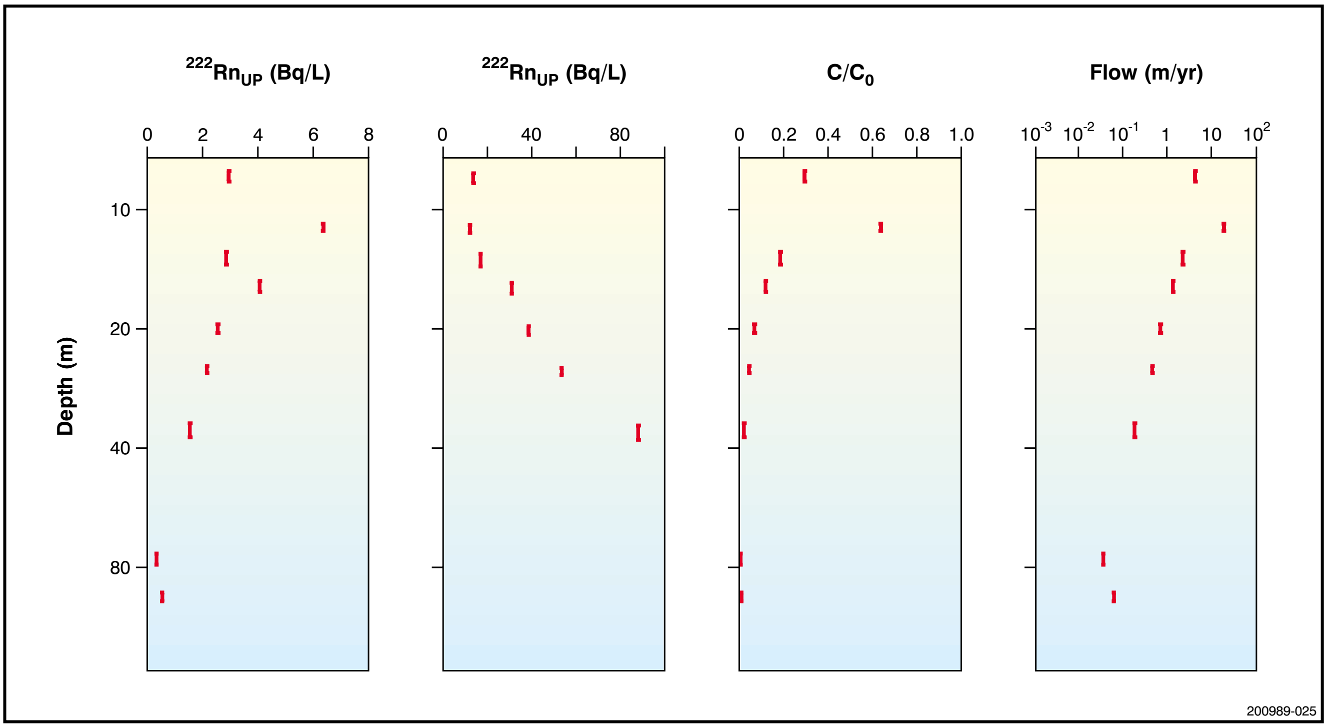


Fig. 4.6 *^{222}Rn concentrations measured in piezometers at Pearce Road (PN 36392 and 36393, refer to Table 2.2) before and after purging, together with inferred horizontal flow rates. Vertical bars indicate the length of the screened intervals. (Depth is plotted on a logarithmic scale.)*

11 m depth. A linear interpolation between the piezometer screens, allows calculation of a total volumetric flow rate through the well (to 100 m depth) of 0.13 $m^2\ d^{-1}$ at the time of sampling (November 1997). At Wendouree, the total volumetric flow rate through the well in January 2001 was estimated using this method to be 0.28 $m^2\ d^{-1}$.

Well dilution

In the well dilution method, we use a recirculation pump to add tracer to the well or to alter the tracer profile already present in the well, and then monitor changes in concentration over time (Love et al., 1999, 2002). However, unlike standard point dilution tests, this method seeks to examine changes in flow rate with depth, rather than to mix the zone and measure a mean flow rate for the zone. Under conditions of horizontal flow, the changes in concentration with time at each depth should be given by Equation 4.7.

Figure 4.7 depicts results of a well dilution experiment on well 6630-2877 at Wendouree. Under ambient conditions, a step-like electrical conductivity profile occurs within the well. Salinity increases with depth from approximately 2 mS cm^{-1} near the water table, to almost 6 mS cm^{-1} below 70 m depth. In October 1998, the water within the well was mixed, by pumping water from the bottom of the well, and re-injecting it at the water table. After mixing for 240 minutes, the pump was withdrawn from the well, and salinity profiles were then measured over time, for a period extending over one month. The time for the electrical conductivity to return to its original profile allows estimation of horizontal flow rates as a function of depth, and also of vertical flow rates within the well.

In May 2000, high flow rates are apparent through fractures at 17, 22, 28 and 38 m depth, although only flow through fractures at 28 and 38 m is apparent in October 1998. Lower flow rates are apparent below 40 m depth, as evidenced by the longer times required for the electrical conductivity profile to return to its original form.

Mean annual flow rates

Figure 4.8 shows the temporal variations in unpurged radon concentration, and calculated groundwater flow rate, for three piezometers at Pearce Road over a nine-month period during 1997. A very high variation in flowrate is inferred from the radon data, particularly in the shallower piezometer, which shows a variation of approximately two orders of magnitude. While the variations in flow rate roughly follow observed variations in rainfall and water table depth, it is apparent that increases in groundwater flow rate occur before the water table rises. The seasonal high water tables also lag behind the rainfall. Both of these observations are attributed the storage effect of the piezometers, which means that the piezometers respond only slowly to changes in the water table. Simmons et al. (Simmons et al., 1999) noted that short-term variations in aquifer water level are significantly attentuated within the well, particularly where the well radius is large. This arises where the permeability of the aquifer is low, and the storativity of the aquifer is very low relative to the storativity of the piezometer or well. Longer-term variations in aquifer water level (such as annual cycles) will be less attenuated. While not directly considered by Simmons et al., the same process should cause

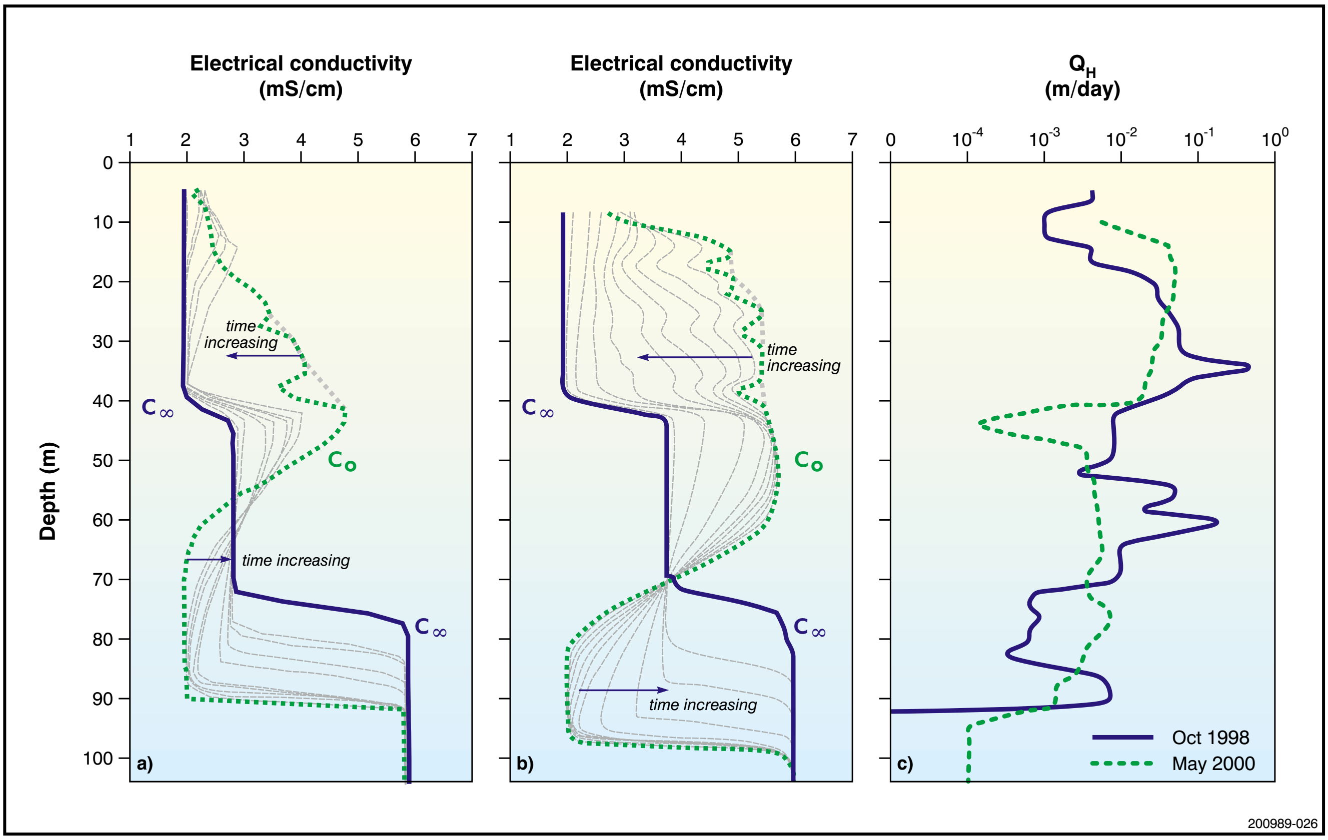


Fig. 4.7 Results on well dilution experiments carried out at Wendouree. (a) and (b) show the time series for the October 1998 and May 2000 experiments. The dark, solid line is the measured electrical conductivity profile immediately before disturbing the well. The solid broken line is the profile measured immediately after disturbance. The thin dashed lines represent the different monitoring times. (c) shows the horizontal flow rates through the well estimated from this data. After Love et al. (2002).

groundwater hydrographs measured in piezometers to lag behind true water table variations. Variations in groundwater flow will respond to actual water table variations, and hence may precede variations in water level as seen in piezometers. By combining this data on seasonal variability of flow rate, with the depth information in Figure 4.6, we estimate a total flow through the aquifer at Pearce Road over the two-year period 1997–98, to be 0.21 m^2 d^{-1}. Based on a distance to the flow divide of 1.5 km, this is consistent with an aquifer recharge rate of 50 mm/yr (Eq. 4.6).

The variation in flow rate throughout the year apparent from the radon data is also very large, being more than two orders of magnitude for piezometer 36393-6, and more than three orders of magnitude for piezometer 36393-3. Seasonal variations in water level over this time period are only ±2 m, while the regional gradient is 40 m in 1 km. The seasonal water level variation thus induces only very minor change in the regional groundwater gradient. This may be explained by a highly permeable zone close to the land surface. While at any particular site, high hydraulic conductivity zones may occur at greater depth, on a regional scale, these zones must only be connected through the shallow zone. Thus, when the water table drops, although these fractures remain saturated, they are no longer connected regionally, and so flow through them decreases markedly (Fig. 4.9).

Groundwater discharge

Groundwater discharge may occur as baseflow to streams, through springs, groundwater pumping, and lateral flow out of the region. Averaged over the region, groundwater pumping is estimated to be approximately 4 mm/yr. We have attempted to estimate groundwater discharge to streams within the Eyre Creek catchment using hydrographic and chloride mass balance approaches.

Hydrographic method

Aquifer heads are generally highest in the Clare Valley during October or November (Fig. 4.10) which can be up to several months after the time of peak stream flow (Fig. 2.6). Therefore hydraulic gradients and groundwater discharge rates to streams will also peak during October–November, and the discharge rate to the stream will be approximately equal to the total stream flow at this time of the year.

Aquifer heads fall from the maximum in November to a minimum value in May due to low rainfall (and hence recharge) plus groundwater pumping for irrigation over this period (Fig. 4.10). Thus, the groundwater discharge rate will be lowest at approximately May each year. We assume that all streamflow is groundwater inflow in November. It is very difficult to estimate the proportion of total stream flow that is

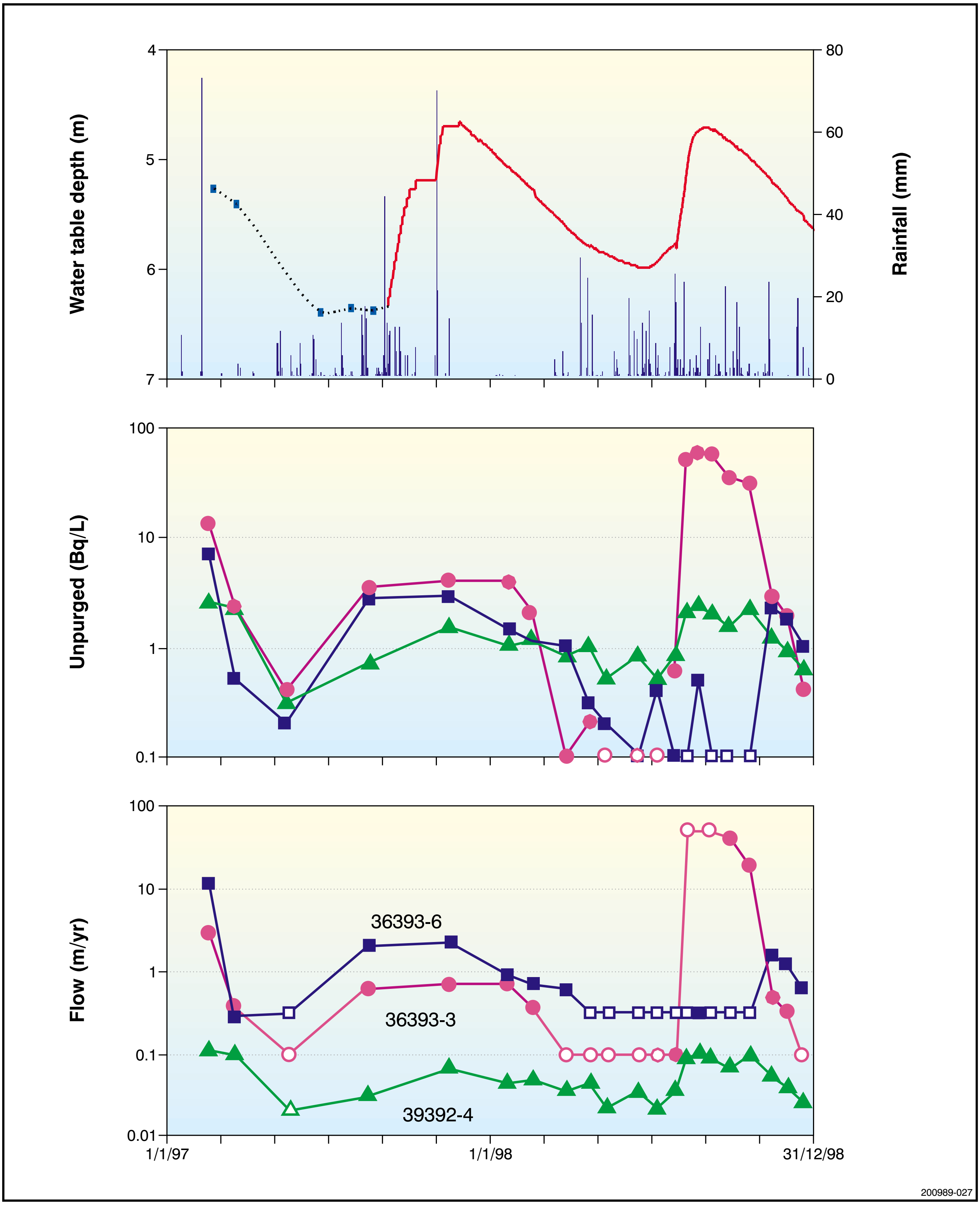


Fig. 4.8 *Temporal variation in rainfall, water table depth, unpurged radon concentration and inferred horizontal flow rate at Pearce Road between January 1997 and December 1998. Greatest variation in unpurged radon concentration, and hence groundwater flow rate is observed for the shallowest piezometer. The maximum flow rate that can be reliably estimated with this technique is approximately 50 m yr^{-1}. The minimum flow rate is determined by the detection limit for radon measurement, and on the radon concentration in the purged piezometer. Based on a detection limit of 0.5 Bq L^{-1}, the minimum flow rate that can be reliably measured is 0.3 m yr^{-1} for P/N 36393-6, 0.1 m yr^{-1} for P/N 36393-3, and 0.02 m yr^{-1} for P/N 36392-4. Open circles denote these minimum and maximum values.*

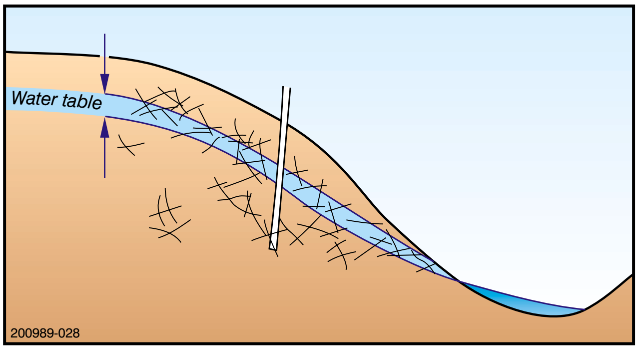


Fig. 4.9 Schematic representation of flow within the Clare Valley. While large fractures can occur at any depth, the intensity of fracturing decreases with depth. On a regional-scale, connections occur only through the shallow, highly fractured zone. While the well in the mid-slope position intercepts a number of fractures, when the water table falls their connection to the regional flow system is broken, and so flow through these fractures decreases markedly, even though the regional hydraulic gradient is still very high.

derived from groundwater discharge when the aquifer heads are at their lowest. By choosing a value between 0% and 100% for this parameter, a linear trend was used to estimate the proportion of total stream flow that is base flow at any other time of the year (Fig. 4.11). These proportions were then converted into volumetric discharge rates using the peak groundwater discharge. That is, the percentage of stream flow that is groundwater at any time of the year is given by

$$\%\ groundwater = min\ \% + \frac{(max\% - min\%).t}{t^*} \qquad (4.9)$$

and hence, the volumetric groundwater discharge rate at that time is

$$Discharge = (\%\ groundwater) \times (max.\ groundwater\ discharge) \qquad (4.10)$$

where *min%* and *max%* are the minimum and maximum proportions of stream flow that are derived from groundwater, respectively; t is time since the minimum or maximum discharge rate occurred (whichever is lower); and t^* is the total time between minimum and maximum discharge.

With this approach, we were able to estimate the annual volume of groundwater discharging to the Eyre Creek as a function of the minimum proportion of groundwater in stream flow (Fig. 4.12 (a)) using both mean and median monthly flow data from Auburn. Because the minimum % groundwater in stream flow occurs in May (before the region's wettest months) it is likely that some groundwater is present in the stream at this time. Supposing this fraction is 20–40%, we obtain an annual groundwater discharge volume of between 15 and 25 ML/yr or just 4.3–7.2% of the mean annual stream flow. These values correspond to a mean areal groundwater discharge to streams of 0.4–0.7 mm/yr.

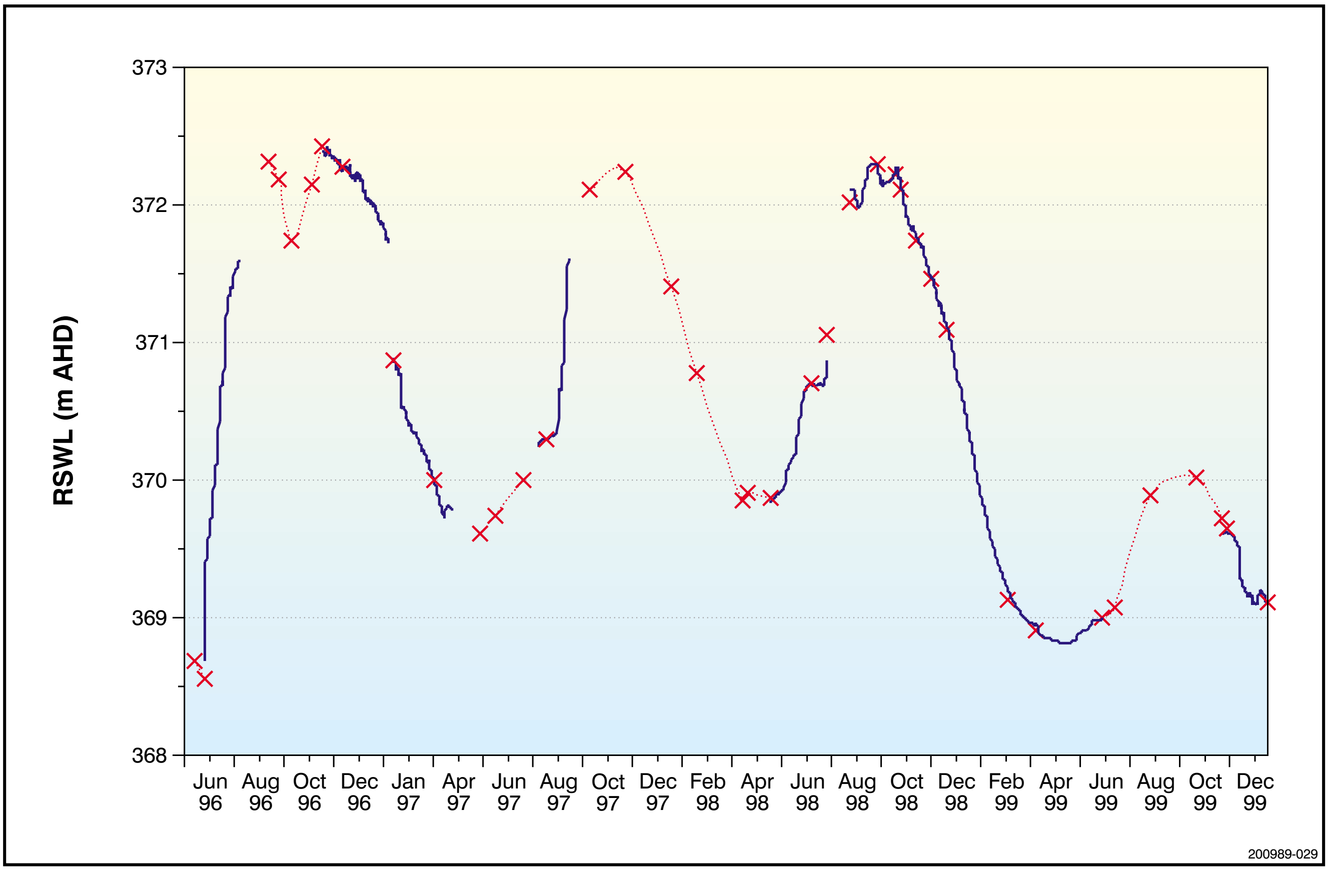


Fig. 4.10 Hydrograph showing seasonal fluctuations of water table elevation in well 6630 2627 at Clos Clare Winery, located approximately 100 m east of Eyre Creek.

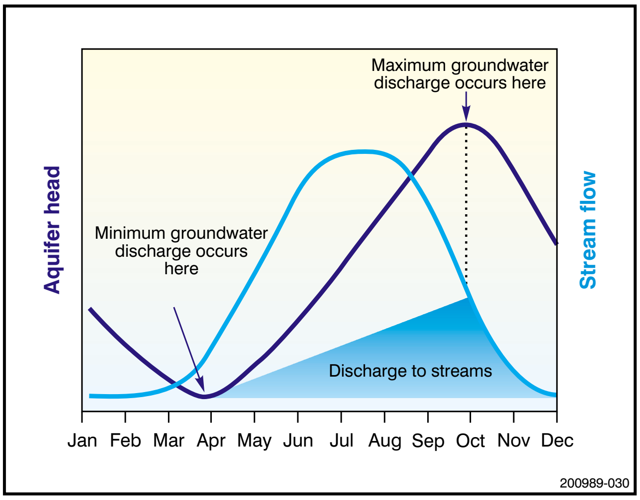


Fig. 4.11 *Schematic representation of seasonality in stream flow and aquifer head elevations observed throughout the Clare Valley. Maximum groundwater discharge to streams occurs in October–November each year, and is approximately equal to the total stream flow rate at this time. Minimum groundwater discharge to streams occurs in May when aquifer heads are lowest; the discharge rate may be any proportion of the total stream flow at this time, and accordingly we have left this parameter as a variable in the discharge calculations.*

Chloride mass balance method

The total mass of Cl in stream water is equal to the mass of Cl introduced from surface run-off plus the mass of Cl added via groundwater discharge.

$$Cl_S . Q_S = Cl_R . Q_R + Cl_{GW} . Q_{GW} \quad (4.11)$$

where Cl is the chloride concentration; Q is the volumetric flow rate; and subscripts S, R and GW represent stream flow, run-off and groundwater respectively. Thus, the total annual groundwater discharge to streams can be determined from daily stream flow and salinity data, and a re-arrangement of Equation 4.11:

$$Q_{GW} = \frac{Q_S(C_S - C_R)}{(C_{GW} - C_R)} \quad (4.12)$$

Estimates of groundwater discharge determined using the chloride mass balance method were performed using both mean and median annual salt loads through the Eyre Creek. It was assumed that the concentration of chloride in surface run-off (C_R in Equation 4.12) is negligible compared with the chloride concentration in both stream flow (C_S) and groundwater (C_{GW}). This assumption is likely to be valid most of the year. The only time C_R will approach C_S is after very intense rainfall events when the majority of stream flow is derived from surface run-off. Most of the groundwater in the vicinity of the Eyre Creek has a salinity in the range 1000–1500 mg/L (Morton et al., 1998) which equates to chloride concentrations of the order 500 mg/L. Figure 4.12 (b) indicates that a groundwater chloride concentration of 500 mg/L corresponds to an annual groundwater discharge of 96 ML/yr using the mean annual salt load or 44 ML/yr using the median data. These estimates are a factor of three to four higher than those obtained using the hydrographic method, and hence correspond to between 12.7% and 27.8% of the mean annual stream flow. Furthermore, they relate to mean areal groundwater losses of between 1.2 mm/yr and 2.7 mm/yr over the Eyre Creek catchment.

One limitation of the hydrographic and chloride mass-balance approaches is that they both neglect the volume (and chloride concentration) of water removed from the Eyre Creek for irrigation. However, because most of the creek water that is used for irrigation is pumped out during times of high flow and low Cl (to supplement existing supplies such as dams) when the groundwater component is likely to be minimal, the actual volume of groundwater lost to streams is not likely to be significantly different from that calculated above. Another

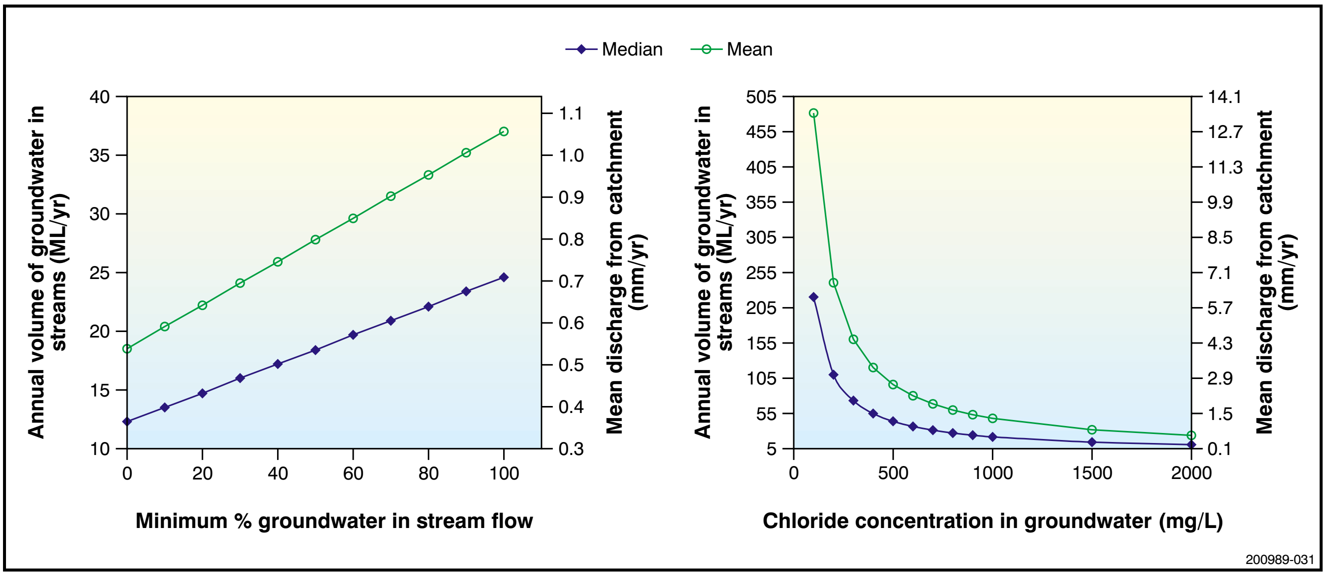


Fig. 4.12 *Calculated values of total annual groundwater discharge to streams obtained using (a) bore hydrographs, and (b) chloride mass balance.*

limitation that is common to both approaches is that the Eyre Creek flow and salinity data set only covers the last four years, which means any long-term average discharge calculations are not possible.

Summary of recharge and discharge

Groundwater age gradients immediately below the water table were used to estimated a recharge rate of 110 mm yr^{-1} at Pearce Road, although this value is very sensitive to the assumed values for matrix porosity, diffusion coefficient and fracture spacing. These latter values were not accurately determined. ^{3}H and ^{36}Cl profiles enabled estimation of a recharge rate of approximately 60–75 mm/yr at both Pearce Road and Wendouree. These results are also consistent with the chloride mass balance that estimated recharge rates to be greater than 50 mm/yr at Pearce Road and greater than 9 mm/yr at Wendouree. Over the two year period 1997–98, the mean horizontal groundwater flow rates at Pearce Road was estimated to be 0.21 m^2/day, which is consistent with an aquifer recharge rate of 50 mm/yr over the distance to the flow divide of 1.5 km.

At Pearce Road and Wendouree, aquifer recharge is thus estimated to be between 50 and 75 mm/yr. There is little information on how recharge may vary throughout the region, due to variations in rainfall, soil type, and landuse. Groundwater discharge to streams is at most 10% of this, and discharge via springs is likely to be even smaller. Averaged over the region, groundwater extraction is estimated to be approximately 4 mm/yr. Thus, even allowing for spatial variability, more than half of the aquifer recharge is unaccounted for. It appears most likely that there is a significant groundwater flow out of the region, both to the south and to the north.

5. CATCHMENT WATER BALANCE

In this project we have obtained estimates of groundwater flow velocity, recharge rates and groundwater discharge to streams in the Clare Valley fractured rock aquifers by a number of established and new methods. The mean annual recharge rate has been estimated to be 50–75 mm/yr at two sites, however these values maybe lower in areas of lower mean annual rainfall and different soil and land-use types. Groundwater discharge to creeks has been estimated to be less than 10 mm/yr in the Eyre Creek catchment. Current groundwater extraction for irrigation is less than 10 mm/yr over the entire Clare Valley. The remaining 30–55 mm/yr of recharge leaves the Clare Valley as lateral groundwater flow. Figure 5.1 is a schematic representation of major findings of this study in terms of the components of the groundwater budget.

The Clare Valley Prescribed Water Resources Area Water Allocation Plan was released in December 2000 to ensure long-term sustainable development of the surface water and groundwater resources in the area. In this plan, allocation of groundwater is based upon an assumed groundwater recharge rate of 30 mm/yr. Whilst our study has shown that recharge exceeds this rate at two sites, the value of 30 mm/yr was chosen as a more conservative estimate of the mean recharge over the entire area considering different soil types, mean annal rainfall and land uses.

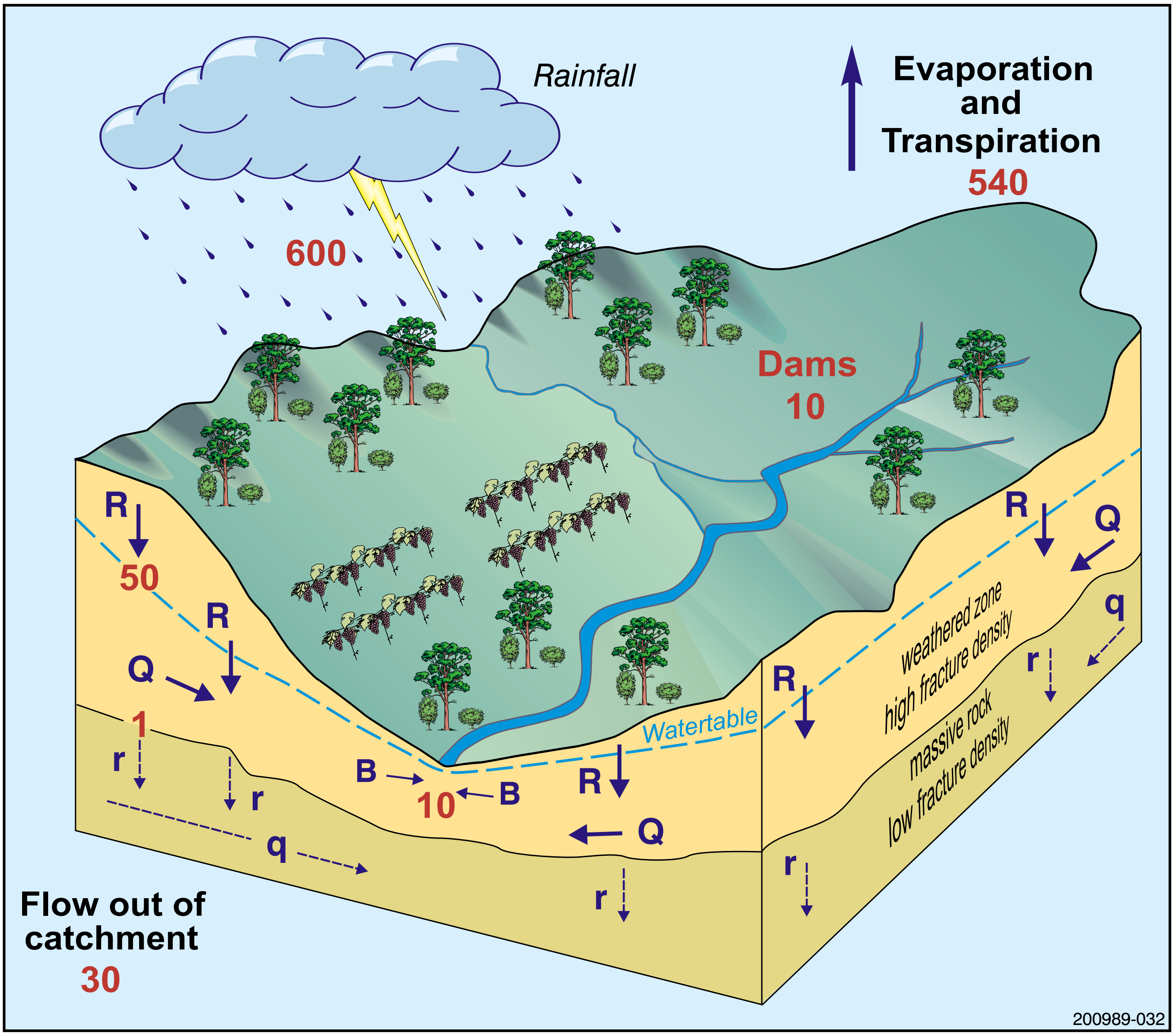


*Fig. 5.1 Conceptual block diagram of groundwater flow systems in the Clare Valley. **R** denotes groundwater recharge, **Q** is horizontal flow for the upper system, **B** is groundwater baseflow to creeks **q** is horizontal flow for the lower flow system and **r** denotes minimal connection between the upper and lower flow systems. Fluxes are in mm/yr.*

REFERENCES

Allison, G.B., Cook, P.G., Barnett, S.R., Walker, G.R., Jolly, I.D, and Hughes, M.W., 1990. Land clearance and river salinisation in the western Murray Basin, Australia. *Journal of Hydrology*, 119:1-20.

Cooper, H. H. Jr., and Jacob, C.E., 1946. A generalised graphical method for evaluating formation constants and summarising well field histories*. Eos, Trans. AGU,* 27(4), 526-534.

Cook, P.G., and Bohlke, J-K., 1999. Determining timescales for groundwater flow and solute transport. In Environmental Tracers in subsurface hydrology eds, Peter G Cook and Andrew Herczeg Kluwer Academic Publishers.

Cook, P.G., Love, A.J. and Dighton, J.C., 1999. Inferring groundwater flow in fractured rock from dissolved radon. *Ground Water*, 37(4):606-610.

Cook P.G. and Simmons C.T., 2000. Using environmental tracers to constrain flow parameters in fractured rock aquifers: Clare Valley, South Australia. In B. Faybishenko (ed.) Dynamics of Fluids in Fractured Rocks: Concepts and Recent Advances. AGU Monograph, 122:337-347.

Cook, P.G and Robinson, N.I., 2002. Estimating groundwater recharge in fractured rock from environmental ^{3}H and ^{36}Cl: Clare Valley, South Australia (submitted to Water Resource Research).

Clarke, D., Harrington, G.A and Love, A.J., 2001. Clare Valley Groundwater Resources. Progress Report 4. Drilling Phase III and IV, Department for Water resources, DWR Report Book 2001/001.

Drost, W., Klotz, D., Koch, A., Moser, H., Neumaier, F. and Rauert, W., 1968. Point dilution methods of estimating ground water flow by means of radioisotopes. *Water Resources Research*, 4:145-146.

Drury, M.J., and A.M. Jessop, 1982. The effect of a fluid-filled fracture on the tempertaure profile in a borehole. *Geothermics* 11, no. 3:145-152.

Ellins, K.K., Roman-Mas, A. and Lee, R., 1990. Using ^{222}Rn to examine groundwater/surface discharge interactions in the Rio Grande de Manati, Puerto Rico. *Journal of Hydrology*. 115:319-341.

Ge, Shemin., 1998. Estimation of groundwater velocity in localized fracture zones from well temperature profiles. *Journal of Volcanology and Geothermal Research*. 84: 93-101.

Grisak, G.E. and Pickens, J.F., 1980. Solute transport through fractured media. 1. The effect of matrix diffusion. *Water Resources Research*, 16(4):719-730.

Harrington, G.A. and Love, A.J., 2000. Water and Salt dynamics in the Clare Valley. DWR Report Book 2000/046.

Jacobsen, G. and Lau, J.E. (compliers), 1987. Hydrogeology of Australia (1: 5 000 000 scale map) Bureau of Mineral Resources, Canberra.

Love, A.J., and Herczeg, A.L., 2001. Exchange of solutes between primary and secondary porosity in a fractured rock aquifer induced by a change in land-use. *Water-Rock Interactions 2001, Cidu, (ed) Swets & Zeitlinger, Liasse* (1): 557-560.

Love et al., Cook. G.P., Simmons, C.T., 2002. Estimating groundwater flow rates in heterogeneous aquifers using the Well Dilution method (submitted to *Water Resources Research*).

Love A.J., Cook, P.G., Halihan, T and Simmons, C.T., 1999. Estimating groundwater flow rates in a fractured rock aquifer, Clare Valley, South Australia. In: The Institution of Engineers Australia, 25th Hydrology and Water Resources Symposium. Brisbane, Australia, July 6-8, 1999.

Love A.J., Cook P.G., Herczeg A.L., and Simmons C.T., 1999. Environmental tracers as indicators of groundwater flow and evolution in a fractured rock aquifer, Clare Valley, South Australia. In: International Symposium on Isotope Techniques in Water Resources Development and Management, International Atomic Energy Agency Vienna, Austria, 10-14 May 1999.

Meier, P.M., Carrera, J. and Sanchez-Vila, X., 1998. An evaluation of Jacob's method for the interpretation of pumping tests in heterogeneous formations. *Water Resources Research,* 34(5): 1011-1025.

Morton, D., Love, A.J., Clarke, D., Martin, R., Cook, P.G. and McEwan, K., 1998. Clare Valley Groundwater Resources. Progress Report 1. Hydrogeology, Drilling and Groundwater Monitoring. Primary Industries and Resources SA, Report Book 98/00015.

Morton, D. and Love, A.J., 1998. Clare Valley Groundwater Resources. Progress Report 2. Drilling Phase 2 Primary Industries and Resources SA, Report Book 98/00014.

Neretnieks, I., 1981. Age dating of groundwater in fissured rock: influence of water volumes in micropores, *Water Resources Research*: 17(2):421-422.

Preiss, W.V. (Compiler), 1987. The Adelaide Geosyncline — late Proterozoic Stratigraphy, sedimentation palaeontology and tectonics. *Bull. Geol. Surv. S. Aust.*, 53.

Schreiber, M.E., Moline, G.R. and Bahr, J.M., 1999. Using hydrochemical facies to delineate groundwatewr flow paths in fractured shale. *Groundwater Monitoring Review*, 19(1):95-109.

Simmons, C.T., Hee Hong Wye, D., Cook, P.G. and Love, A.J., 1999. Signal Propagation and periodic response in aquifers: the effect of fracture signal and measuring methods. *In* the Institution of Engineers, 25th Hydrology and Water Resource Symposium. Brisbane, Australia, July 6-8 Vol. 2:727-732.

Snow, D.T., 1968. Rock fracture spacings, openings and porosities. *J. Soil Mech*. Found. Div. Proc. Amer. Soc. Civil. Engrs., 94 p. 73-91

ACKNOWLEDGMENTS

This project was funded by major contributions from the Resource Assessment Division of the Department for Water Resources (formerly PIRSA, Groundwater Services), CSIRO Land and Water and Land and Water Australia (formerly Land and Water Resources Research and Development Corporation).

We gratefully acknowledge the contribution from the following:

Overseas Scientists

Todd Halihan (University of Texas)
Roland Purtschert (University of Berne)
Bill Sanford (Colorado State University)
Kip Solomon (University of Utah)
Rene Therrien (Universite Leval)

Project Management

Graham Allison (LWA)
Nick Schofield (LWA)

Wine and Industry Community

Clare and Gilbert Valleys Council
Duncans Wines
R. and S. Eaton
Flatview Vineyards
Gullyview Vineyards Pty Ltd
Jim Barry's Wines
Leasingham Wines
LK Vineyard
Petaluma Ltd
Rosehill Vineyards
Sevenhill Cellars
Southcorp Wines
Taylors Wines
Mt Surmon Vineyard

Field and Technical Support

Jeff Graham (DWR)
Jim Safta (DWR)
Rick Fennell (DWR)
David Clarke (DWR)
Russ Martin (DWR)
Robyn Gill (DWR)
Keith Brown (DWR)
Deborah Clarke (DWR)
Dawn Morton (DWR)
John Dighton (CSIRO, Land and Water)
Kerryn McEwan (CSIRO, Land and Water)
Andrew Herczeg (CSIRO, Land and Water)
Neville Robinson (CSIRO, Land and Water)
Lincoln Patterson (CSIRO, Petroleum)
Tony Brady (Wendouree Cellars)

Australian Universities

Graham Heinson (FUSA, Adelaide)
Tom Wilson (FUSA, Adelaide)
Chris Smitt (FUSA, Adelaide)
Brooke Warren (FUSA, Adelaide)
Davis Hee Hong Wye (FUSA, Adelaide)
Fritjof Fagerland (Adelaide)
Damien Skinner (Adelaide)

Resource Assessment Division

Department for Water Resources
Level 11, 25 Grenfell Street, Adelaide
GPO Box 1047, Adelaide SA 5001

Phone	National	(08) 8463 6800
	International	+61 8 8463 6800
Fax	National	(08) 8463 6999
	International	+61 8 8463 6999
Website	www.dwr.sa.gov.au	