HYDROGEOLOGICAL AND HYDROCLIMATIC CONTROLS ON SURFACE WATER–GROUNDWATER INTERACTIONS

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DECLARATION

I certify that this thesis does not incorporate without acknowledgment any material previously submitted for a degree or diploma in any other university; and that to the best of my knowledge and belief it does not contain any material previously published or written by another person except where due reference is made in the text.

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Edward Wallace Banks

CO-AUTHORSHIP

Eddie Banks is the primary author on this thesis and all the enclosed documents. Chapters 2 to 4 were written as independent manuscripts in which the co-authors provided intellectual supervision and editorial comment.

In addition, Dr. Andrew Love suggested investigating the importance of native vegetation in the interactions between surface water and groundwater discussed in Chapter 3. Prof. Craig Simmons suggested using a modelling approach to investigate the vegetation controls on variably saturated processes between surface water and groundwater in Chapter 4. Dr. Philip Brunner provided modelling support for the numerical model HydroGeoSphere which was used in Chapter 4.

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SUMMARY

Growing awareness about the benefits from more sustainable management and allocation of water resources has highlighted the need to manage surface water and groundwater systems as one integrated system. Whilst there has been a significant contribution to the knowledge and understanding of hydrogeological research of surface water–groundwater interactions in the past few decades, there are still specific knowledge gaps on how different types of systems (i.e. connected gaining and losing, and losing disconnected) function and interact at different spatial and temporal scales and in different hydrogeological environments.

This body of research addresses some of the complexities of surface water–groundwater interactions in fractured rock environments, at different spatial and temporal scales and investigates a number of the dominant controls (e.g. geology, topography and vegetation) that influence the exchange processes and dynamics between surface water and groundwater. Specifically, this work investigates: (1) the importance of groundwater from the fractured bedrock compared to groundwater from the saprolite zone in streamflow generation and why the bedrock interface cannot be considered a no-flow boundary. (2) surface water–groundwater interactions in a pristine catchment at a regional scale to determine the state of connection between surface water and groundwater along a river system from the catchment headwaters to the discharge point at the sea. (3) the vegetation controls on variably-saturated processes between surface water and groundwater and its impact on their state of connection.

The first part of this research was a field-based study examining surface water–groundwater interactions along a gaining river reach. The study investigated the importance of groundwater from the fractured bedrock compared to groundwater from the saprolite zone in streamflow generation and examined why the bedrock interface cannot be considered a no-flow boundary. The hypothesis was to determine whether the saprolite zone is hydraulically more active than the deeper bedrock zone. The findings of this study suggest that hydrologic conceptual models, which treat the saprolite-fractured bedrock interface as a no-flow boundary and do not consider the deeper fractured bedrock in hydrologic analyses, may be overly simplistic and inherently misleading in some surface water–groundwater interaction analyses. The results emphasise the need to understand the relative importance of subsurface flow activity in both of these shallow saprolite and

deeper bedrock compartments as a basis for developing reliable conceptual hydrologic models of these systems.

The second part of this research was also a field-based study in a pristine catchment which investigated the state of connection between surface water and groundwater along a river system from the catchment headwaters to the discharge point at the sea. The relative source and loss terms of the river and groundwater systems were assessed, as were their relative magnitude changes along the river, and how a fresh water river system in a pristine catchment covered by native vegetation exists in an otherwise saline regional groundwater system. Many surface water–groundwater interaction studies of different types of systems are either undertaken at the local or river reach scale, however, catchments encompass multiple types of systems at a regional scale. There has been very little research investigating how multiple river reaches function in the context of the entire regional river system from the headwaters to the sea or discharge point, and this study demonstrates the benefits of doing so.

The final part of this research used a fully coupled, physically based numerical model to demonstrate the vegetation controls on variably-saturated processes between a perennial river and an aquifer and its impact on their state of connection. By examining different conceptual models of catchments with different slopes and vegetation type (i.e. root depth) the research identified the conditions required for changes to vegetation to have the greatest effect on the flow regime and the presence of an unsaturated zone beneath a riverbed. The analysis also suggested that the flow regime and hydraulic response to the presence of vegetation and subsequent removal can be much greater in flatter catchments than those that are steep. Intuitively, this may appear plausible in a qualitative sense; however, it has not been demonstrated quantitatively. The results of the study therefore suggests that in addition to the well known influences of physical variables such as hydraulic conductivity or topography, the effects of vegetation need to be carefully considered when investigating surface water–groundwater interactions.

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CHAPTER 1

1. INTRODUCTION

1.1 THE RESEARCH PROBLEM

Over the past few decades there has been increasing concern about water scarcity and the impacts of water resource consumption on water dependent ecosystems (Sophocleous, 2002; Winter et al., 1998; Woessner, 2000). Growing awareness about the need for more sustainable management and allocation of water resources has highlighted the need, from both a scientific and policy perspective, to manage surface water and groundwater systems as one integrated system. Traditionally, surface water and groundwater have been managed separately which has resulted in double accounting (i.e., where one parcel of water is accounted for once as groundwater and then a second time as surface water baseflow) and over allocation of the water resources. This management regime has had serious impacts upon the environment and the long term viability of many water dependant ecosystems and industries. Successful management of surface water and groundwater as an integrated resource is greatly improved once the state of connection is determined. This requires an understanding of surface water–groundwater interactions, the state and type of connection (i.e. gaining, losing, and losing disconnection) and the role of groundwater in streamflow generation. It is inadequate to assume that pumping from a river that is considered disconnected at one location will not have an effect on the length of the river that is disconnected. The terminology that is used to describe disconnection has recently been described by Brunner et al. (2011), who identify the state of disconnection by an unsaturated zone under the stream or by showing that the infiltration rate from a stream to an underlying aquifer is independent of the watertable position.

The impacts from using surface water and groundwater from the different system types will vary depending on the connectivity state. In gaining river systems, such as in the catchment headwaters, increased groundwater extraction may impact on the state of connection between surface and groundwater systems, potentially causing a reduction in river flow and duration. Significant reductions in watertable elevation may ultimately cause groundwater levels to drop below the elevation of the surface water system such that it becomes a losing type system, eliminating baseflow and posing a threat to the sustenance of permanent pools. In a losing and disconnected surface water system, groundwater extraction is less likely to cause any further impact to streamflow conditions at the location where it is losing but will affect the state of connection along the length of the surface water system. It is therefore important that quantitative estimates of the groundwater contribution to the surface water system are understood to ensure that extraction volumes do not exceed the volume required to sustain river flow.

It is now more common to investigate and quantify surface water–groundwater interactions using a multi-disciplinary approach which includes physical hydrogeology (Hatch et al., 2006), hydrogeochemistry (Cook et al., 2003; Cook et al., 2006; Payn et al., 2009; Shand et al., 2007a; Shand et al., 2009), environmental tracers (Ellins et al., 1990; Ruehl et al., 2006; Stellato et al., 2008) and numerical modelling techniques (Bruen and Osman, 2004; Brunner et al., 2009a; Brunner et al., 2011; Brunner et al., 2010; Kalbus et al., 2009; Osman and Bruen, 2002). The combination of the different techniques helps constrain and identify contributing sources of solutes, preferential flow pathways and residence times at different spatial and temporal scales (Kalbus et al., 2006). The application of numerical modelling has been used more frequently in the last decade to examine and quantify the water flux exchange between surface water and groundwater in the different types of surface water–groundwater systems (i.e. connected gaining and losing, and losing disconnected).

While there has been an increasing focus upon integrated groundwater and surface water development, there are several areas requiring further research and investigation if sustainable management practices are to be achieved. This PhD focuses upon three specific knowledge gaps:

Firstly, there are significant complexities involved in understanding and managing fractured rock aquifer systems, which are inherently more difficult to research and conceptualise than sedimentary aquifer systems. Whilst the influence of topography on groundwater flow is important (McGuire et al., 2005; Wörman et al., 2006), the underlying geological structure can have a considerable effect on controlling the direction and contribution of groundwater to surface water systems (Fan et al., 2007). Often, to simplify the conceptualisation of fractured rock systems, the bedrock interface is considered to be a no flow boundary. However, this fails to consider the contribution from the fractured rock system to streamflow generation. In fractured rock aquifer systems, it is more likely that they will be complex systems whereby surface water features may receive water from one or more groundwater flow systems, whether they be local, intermediate, regional, shallow or deep. Understanding the functioning of fractured rock systems and their contributions to streamflow generation will help ensure that more reliable estimates of the location, volume and timing of fluxes between groundwater and surface water features can be made.

Secondly, most surface water–groundwater interaction studies are either undertaken at the local (metres to tens of metres) or river reach scale (hundreds of metres to a few kilometres) and specifically investigate one type of system or connection state. However, many catchments encompass multiple types of systems. There has been very little research investigating how individual river reaches function in the context of the entire regional river system (comprising multiple river reaches; tens of kilometres) from the headwaters to the sea or discharge point. Failing to consider the range of connection types along the entire length of a catchment can result in simplistic assessments and inappropriate management decisions resulting in poor outcomes for water quantity and quality. For example, a river reach at the top of the catchment may be found to be a gaining system; however the river system across the whole of the catchment may be a losing system overall, which would require different management approaches to that of a gaining system. Another important factor in assessing the different states of connection and the potential impacts on water quality is distinguishing between the contributing sources of groundwater to the river and the exchanges between local and regional groundwater systems.

Finally, vadose zone processes play an important role in surface water–groundwater interaction. For example, the presence of an unsaturated zone has a strong influence on biogeochemical processes of river systems (Bencala, 1993) and various ecological and hyporheic exchange processes (Brunke and Gonser, 1997; Findlay, 1995). Also, it is the presence of an unsaturated zone which controls the state of connection between surface water and groundwater. Despite recent advances in knowledge on the relationship between hydrogeological variables, the presence of an unsaturated zone and the state of connection, there is limited understanding of the role of vegetation (i.e. evapotranspiration) in forming an unsaturated zone between surface water and groundwater. The effects of evapotranspiration is significant because in regions of Australia, native vegetation clearance and land use modification (e.g. deep rooted vegetation replaced by shallow rooted crops) has had considerable impacts on the water balance and surface water and groundwater salinities as a result of increased recharge and the mobilisation of salt stored in the shallow regolith (Allison et al., 1990; Bell et al., 1990; Cartwright et al., 2004). Qualitatively, it may seem obvious that the presence of vegetation can alter the state of connection or that change to the rates of precipitation and evapotranspiration may, under certain conditions, have some effect on the state of connection. However, there is little understanding what the quantitative effects will be and the sensitivity of the state of connection (and associated exchange fluxes) to various controlling physical variables i.e. hydraulic conductivity of the aquifer and clogging layer, catchment slope and vegetation type (e.g. root extinction depth).

1.2 Research Aim

The broad aim of this PhD was to explore specific knowledge gaps in the hydrogeological research area of surface water–groundwater interactions. The PhD addresses some of the many complexities of surface water–groundwater interactions in fractured rock environments, at different spatial and temporal scales and investigates a number of the dominant controlling functions (e.g. geology, topography and vegetation) that influence the exchange processes and dynamics between surface water and groundwater using field and numerical modelling techniques. Specifically, this body of work investigates:

i. the importance of groundwater from the fractured bedrock compared to groundwater from the saprolite zone in streamflow generation and why the bedrock interface cannot be considered a no-flow boundary. It also distinguished between different contributing sources to streamflow and their relative hydraulic activity/responsiveness.

ii. surface water–groundwater interactions in a pristine catchment at a regional scale (tens of kilometres) to determine the state of connection between surface water and groundwater along a river system from the catchment headwaters to the discharge point at the sea. In addition, the research examined the relative source and loss terms of the river and groundwater system and how their relative magnitude changes along the river, and how a fresh water river system in a pristine catchment covered by native vegetation exists in an otherwise saline regional groundwater system.

iii. using a simple conceptual model based on realistic and representative parameter values, to what degree the presence of vegetation can cause an unsaturated zone to develop between a perennial river and an aquifer and therefore, its effects on the state of connection. More specifically, it investigated if evapotranspiration is a plausible mechanism to create an unsaturated zone underneath a riverbed and whether it can influence connected gaining and losing, and losing–disconnected type conditions. The physical controls of catchment slope, the hydraulic conductivity of the riverbed clogging layer and aquifer, and the vegetation root extinction depth (transpiration extinction depth) were also examined to determine how these variables influence the state of connection.

The specific research areas of surface water–groundwater interactions are addressed in three manuscripts and are contained in Chapters 2, 3 and 4 of this thesis. The three manuscripts have now been published in international hydrogeological journals. Presented and published conference proceedings as part of this PhD are shown in Appendix A.

[Chapter 2]

Banks, E. W., Simmons, C. T., Love, A. J., Cranswick R., Werner A. D., Bestland E. A., Wood, M. and Wilson, T. (2009). Fractured bedrock and saprolite hydrogeologic controls on groundwater/ surface-water interaction: a conceptual model (Australia). *Hydrogeology Journal* **17**: 1969-1989. doi: 10.1007/s10040-009-0490-7

[Chapter 3]

Banks, E. W., Simmons, C. T., Love, A. J. and Shand, P. (2011b). Assessing spatial and temporal states of connection between surface water and groundwater in a regional catchment: implications for regional scale water quality. *Journal of Hydrology* **404**(1-2): 30-49. doi: 10.1016/j.jhydrol.2011.04.017

[Chapter 4]

Banks, E. W., Brunner, P. and Simmons, C. T. (2011a). Vegetation controls on variably-saturated processes between surface Water and groundwater and its impact on their state of connection. *Water Resources Research* **47**(11): W11517. doi:10.1029/2011WR010544

1.3 Contribution of this PhD

This PhD explores the complexities of surface water–groundwater interactions in fractured rock environments, at different spatial and temporal scales using both field based and numerical modelling techniques. The research investigates several of the dominant controls (e.g. geology, topography and vegetation) that influence the exchange processes and dynamics between surface water and groundwater. It does so by examining: (i) the importance of groundwater from the fractured bedrock compared to groundwater from the saprolite zone in streamflow generation. Previous research has often over simplified this boundary condition in hydrological conceptual models, however, this research has shown that treating the bedrock interface as a no-flow boundary needs to be carefully evaluated in hydrologic analyses. The relative importance of subsurface flow activity in both of these shallow saprolite and deeper bedrock compartments is needed as a basis for developing reliable conceptual hydrologic models. (ii) regional scale surface water–groundwater systems. Most surface water–groundwater interaction studies have been undertaken at the river reach scale. Whilst many benefits have arisen from this type of investigation, a regional scale approach along a river, which was conducted as part of this research, provides greater insight into the changes to the state of connection from the catchment headwaters to the surface water discharge point (e.g. the sea) which can be of greater use to the management of the water resource. (iii) the vegetation controls on variably-saturated processes between surface water and groundwater and its impact on their state of connection. This may appear intuitively plausible in a qualitative sense; however, it has not been demonstrated quantitatively. The research therefore suggests that in addition to the well known influences of physical variables such as hydraulic conductivity or topography, the effects of vegetation (i.e. evapotranspiration) need to be carefully considered when investigating surface water–groundwater interactions. Specifically, changes in vegetation type and extent that may be associated with land use change or climate change are expected to have an impact on the state of connection and therefore on exchange fluxes, directions, water balances and water quality matters in such systems.

It is hoped that these research findings will broaden understanding of regional scale surface water–groundwater interactions. Effective management of surface water and groundwater resources requires this understanding.

CHAPTER 2

2. MANUSCRIPT I: FRACTURED BEDROCK AND SAPROLITE HYDROGEOLOGIC CONTROLS ON GROUNDWATER–SURFACE WATER INTERACTION: A CONCEPTUAL MODEL (AUSTRALIA)

PUBLISHED IN HYDROGEOLOGY JOURNAL, 2009

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Banks, E. W., Simmons, C. T., Love, A. J., Cranswick, R., Werner, A., Bestland, E., Wood, M., Wilson, T. (2009). Fractured bedrock and saprolite hydrogeologic controls on groundwater/ surface-water interaction: a conceptual model (Australia). *Hydrogeology Journal* **17**: 1969-1989. doi: 10.1007/s10040-009-0490-7.

ABSTRACT

Hydrologic conceptual models of groundwater/surface-water interaction in a saprolite-fractured bedrock geological setting often assume that the saprolite zone is hydraulically more active than the deeper bedrock system and ignore the contribution of deeper groundwater from the fractured bedrock aquifer. A hydraulic, hydrochemical, and tracer-based study was conducted at Scott Creek, Mount Lofty Ranges, South Australia, to explore the importance of both the deeper fractured bedrock aquifer system and the shallow saprolite layer on groundwater/surface-water interaction. The results of this study suggest that groundwater flow in the deeper fractured bedrock zone is highly dynamic and is an important groundwater flow pathway along the hillslope. Deep groundwater is therefore a contributing component in streamflow generation at Scott Creek. The findings of this study suggest that hydrologic conceptual models, which treat the saprolite-fractured bedrock interface as a no-flow boundary and do not consider the deeper fractured bedrock in hydrologic analyses, maybe overly simplistic and inherently misleading in some groundwater/ surface-water interaction analyses. The results emphasise the need to understand the relative importance of subsurface flow activity in both of these shallow saprolite and deeper bedrock compartments as a basis for developing reliable conceptual hydrologic models of these systems.

2.1 INTRODUCTION

Interactions between groundwater and surface water (gw–sw) form one component of the hydrological cycle and are largely controlled by the effects of physiography (topography and geology) and climate (Winter et al., 1998). Considerable research on this topic has been undertaken in sedimentary aquifer systems (e.g. Beyerle et al., 1999; Krause and Bronstert, 2007; Schilling et al., 2006) but very few studies are reported for fractured bedrock systems (e.g. Haria and Shand, 2006; Kahn et al., 2008; Manning and Caine, 2007; Sklash and Farvolden, 1979). The latter are substantially more complex owing to the geological heterogeneity of the fractured-rock aquifer. In addition, the deeper fractured bedrock aquifers are usually overlain by surficial weathered soil and saprolite material. Intuitively, these different geologic layers and the high level of heterogeneity in this system is expected to be critical in controlling both groundwater and surface water responses and hence the nature of the gw–sw interaction. However, the partitioning of the aquifer system between the soil (weathered material), saprolite (weathered material that retains structure of the parent rock) and fractured rock (unweathered bedrock) is often simplified. These geologic layers are usually not considered as explicit geologic features and hydrogeologic controls on gw–sw interaction. Some important questions arise: What is the groundwater flux through the saprolite zone compared to the deeper fractured bedrock system? How do these various geologic controls influence the nature of gw–sw interaction? Under what conditions do these various geologic controls need to be included in our hydrogeologic conceptual models? What level of simplification is permissible in a hydrologic conceptual model and what are the consequences of conceptual model choice?

The study of gw–sw connectivity under gaining-stream conditions in fractured-rock systems generally involves the application of one of at least three types of conceptual models, the most basic forms of which are as classified and described in the following (Figure 2.1). In the first type, the gw–sw interaction and streamflow response is interpreted using a single geological system in a bucket-type approach, which does not explicitly account for soil, saprolite or fractured bedrock (Figure 2.1a). Type 1 conceptual models implicitly or explicitly assume that the system is homogeneous, as is commonly the case in traditional hydrograph separation methods. In the second type, the surficial saprolite system dominates the hillslope hydrology (i.e. there is no groundwater flow in the deeper fractured bedrock aquifer and the saprolite-fractured bedrock interface is considered an impermeable boundary) and the gw–sw interaction and streamflow response of the gaining-stream is completely interpreted by assuming subsurface flow occurs in the surficial saprolite layer only (Figure 2.1b). Thirdly, both a saprolite layer and fractured bedrock aquifer system are explicitly considered as geologic features and hydrogeologic controls on gw–sw interaction (Figure 2.1c). The first conceptual model (Figure 2.1a) may be regarded as the simplest model, while the third (Figure 2.1c) may be regarded as the most complex. A critical challenge is that one rarely knows, apriori, which is the most appropriate conceptual model to choose for any given field site and what the consequences of conceptual model choice (simple or complex) are on the interpretation of gw–sw interaction data and associated analyses.

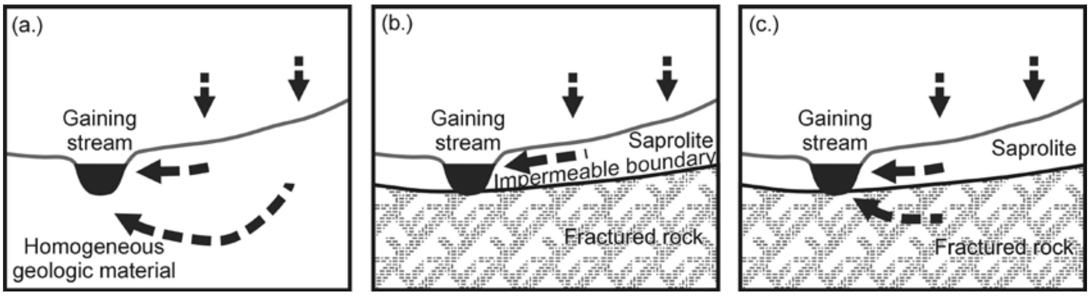


Figure 2.1 Conceptual models of groundwater/surface-water interaction under gaining stream conditions in a complex saprolite-fractured bedrock aquifer system: (a.) homogeneous system, (b.) subsurface flow in the surficial saprolite zone only, and (c.) subsurface flow occurs in both the surficial saprolite layer and deeper fractured bedrock aquifer.

Numerous studies of gw–sw connectivity have applied the first conceptual model (Figure 2.1a), particularly in the application of hydrograph separation methods (e.g. Chapman, 1999; Pinder and Jones, 1969), where the pathways of groundwater discharge are not explicitly determined and often considered as a 'blackbox' within a bucket type approach. The assumptions used in the second conceptual model (Figure 1b) can be seen in a wide range of catchment hillslope hydrological studies (e.g. Kirkby, 1988), which very often assume that subsurface flow occurs in the soil/saprolite zone only, rather than in the deeper fractured bedrock groundwater and hence that surficial flows in the saprolite zone are of far greater concern in the partitioning of streamflow. Gburek and Urban (1990) and Gburek et al. (1999) investigated a layered fractured zone along two hillslope cross-sections and noted that the soil zone rapidly transmitted rainfall to the groundwater and that lateral flow in the highly fractured zone above less permeable bedrock dominated the groundwater flow.

The geological controls on rapid lateral flow that may occur along the interface between the saprolite and fractured bedrock are largely unknown and there is still surprisingly little information on the importance of the fractured bedrock aquifer in streamflow generation. However, the very few recent studies which have considered the third conceptual model (Figure 1c; e.g. Haria and Shand, 2006; Shand et al., 2005) have shown that rising water tables in the fractured rock contribute to relatively rapid lateral flows in the saprolite zone and that deeper groundwater flows in the fractured bedrock play an active and critical role in streamflow generation. Distinguishing between shallow soil, saprolite and fractured bedrock groundwater contributions to streamflow generation is difficult and even more complicated with the existence of pre-event water. A study by Sklash and Farvolden (1979) of several watersheds characterised by surficial deposits of 1–20 m thickness overlying bedrock in Quebec, Canada, noted that groundwater can play an active and responsive role in streamflow generation. The study emphasised that it is not just surface-water runoff that contributes to peak flow events and that groundwater may also control surface water quality during these events.

Part of the difficulty in assessing these types of saprolite-fractured bedrock systems lies in the complexity associated with the geological heterogeneity inherent to all of fractured rock hydrology. It is well known that the rates of groundwater flow and connectivity in fractured rock aquifers are difficult to determine, and methods commonly used for porous media are often not applicable (Cook et al., 1996; Love et al., 2002). Multi-tracer approaches have been used in fractured rock systems (e.g. Genereux et al., 1993b; Shand et al., 2007a) and have proven invaluable in constraining contributing sources of solutes, preferential flow pathways and residence times in these types of systems. Groundwater noble gas, age and temperature data were used by Manning and Caine (2007) in an alpine bedrock catchment to evaluate the groundwater pathways and activity of deep fractured bedrock systems, and mean residence time of groundwater in the catchment.

The movement of water through soils and the saprolite through relict fractures and macropores has been shown to be important in several studies. McKay et al. (2005) found that the hydraulic conductivity and groundwater flow in a sedimentary rock saprolite was influenced significantly by the parent lithology, degree of infilling with clays and Fe/Mn oxides, and the physical nature of the macropores. McDonnell (1990) hypothesised that rapid downslope flow of pre-event water through continuously connected pipes extending from the hillslopes to the stream was responsible for observed runoff in a small headwater catchment in New Zealand. The importance of rapid transport pathways via fractures and macropores in saprolite and fractured rock was also shown in a study by Van der Hoven et al. (2005). Their study evaluated temporal variability in tracer data in the soils, saprolite and shallow fractured sedimentary rocks in Tennessee, USA.

The field-based study described in this paper investigates the role of the saprolite-fractured bedrock aquifer system on gw–sw interactions in the Scott Creek catchment in the Mount Lofty Ranges, South Australia. The monitoring period for this study was between July 2005 and December 2007. It builds upon some earlier hillslope studies in this area by Smettem et al. (1991) and Leaney et al. (1993) which investigated the influence of macropores on runoff processes and concluded that lateral subsurface flows are initiated above the soil–rock interface rather than the A–B soil horizon boundary once infiltration of the macropore system has been exceeded. It was implicitly assumed in those earlier studies that the main contributors to streamflow were lateral subsurface flow in the saprolite zone only and surface runoff. The role of deeper groundwater from the fractured bedrock zone was not considered. As a result, previous conceptual models of the saprolite-fractured bedrock interface at Scott Creek were considered to be no flow boundaries (i.e. to date, the second conceptual model (Figure 2.1b) as defined previously, has been used as the basis for previous hydrologic analyses at Scott Creek). Using a hydraulic, hydrochemical, and tracer-based approach, the field study reported here explores the importance of the deeper fractured bedrock aquifer system in the Scott Creek catchment to determine whether it exerts greater influence on gw–sw dynamics than has previously been reported. Our results provide conclusive evidence that the deeper fractured bedrock is hydraulically active and indeed quite possibly more so than the shallow saprolite system. This suggests that, unlike previous studies at this site, the deeper fractured bedrock system should be considered in hydrologic analyses pertaining to gw–sw connectivity in this system.

2.2 STUDY AREA AND BACKGROUND

The study site Scott Bottom is located near the bottom of the Scott Creek Catchment (SCC; Figure 2.2). The SCC is a relatively small (27 km^2) catchment in the Mount Lofty Ranges (MLR), South Australia and is located approximately 30 km southeast of Adelaide. The SCC is characterised by a temperate climate, experiencing warm dry summers and wet cool winters. Mean annual evaporation is 1,517 mm (Department of Water, Land and

Biodiversity Conservation, unpublished data, 2007). Mean annual rainfall ranges from 804 mm at the bottom of the catchment to 1,009 mm in the upper reaches of the catchment (based on 1968–2007 data; Department of Water, Land and Biodiversity Conservation, unpublished data, 2007).

The topography of SCC varies from steep slopes to gently undulating land. The main watercourse of Scott Creek runs in a north–south direction and minor tributaries that contribute to Scott Creek, which tend to be dry during the summer months, dissect the steep sloped valleys. The established gauging station on Scott Creek at the study site Scott Bottom, located near the bottom of the SCC, has a rectangular stepped V-notch weir with continuous streamflow records from 1964 until present. The mean annual flow is approximately 3,710 ML/year (based on 1979–2007 data; Department of Water, Land and Biodiversity Conservation, unpublished data, 2007). Due to increased rainfall from May to October each year, high flows usually occur during this period with maximum flow typically observed around August. Streamflow records were instantaneous but for this study data is presented on a daily interval. A pluviometer at the site with a 203 mm throat and 0.2 mm tipping bucket has recorded rainfall from 1991 through to the present day.

2.2.1 Geology and Hydrogeology

The SCC is characterised by moderate to steep topographic relief with soil (0–3 m thickness) and saprolite (1–20 m thickness) material underlain by fractured bedrock, which is exposed at the surface in some areas. The geology is structurally complex, with a diverse range of metamorphosed sedimentary formations including siltstone, sandstone and dolomite of the Adelaidean sequences of the Adelaide Geosyncline (Preiss, 1987; Figure 2.2). The Woolshed Flat Shale metasediment dominates the area around the study site at Scott Bottom. The eastern side of the catchment is predominantly Aldgate Sandstone (and to a lesser degree Skillogalee Dolomite) and the higher topographic areas on the western side of the catchment is dominated by Stonyfell Quartzite.

There are approximately 150 bores in the SCC. While some are completed in the shallow alluvial aquifers (<20 m depth), the majority of bores are located in the deeper fractured Woolshed Flat Shale, Aldgate Sandstone and Stonyfell Quartzite metasediments (bore depths of up to 190 m depth) due to higher yields and lower salinities in the deeper bedrock system. Yields range from 0.02 to 25 L/s, and of the bores that have salinity data records, 95% have an electrical conductivity (EC) less than 1,500 µS/cm, and 40% have an EC less than 500 µS/cm (James-Smith and Harrington, 2002).

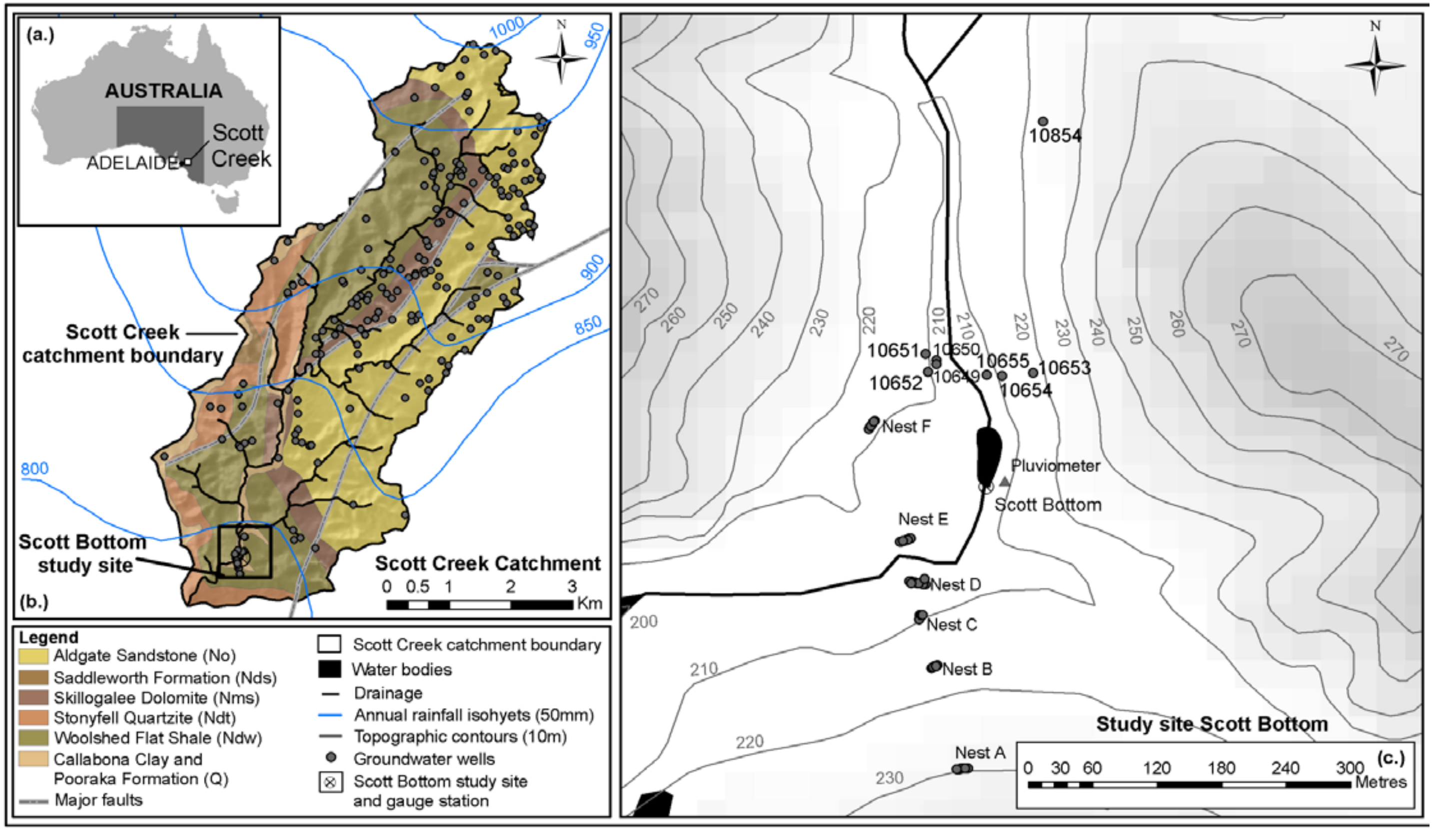


Figure 2.2 (a.) Map showing the location of Scott Creek Catchment in the Mount Lofty Ranges, South Australia. (b.) The geology, groundwater wells and location of Scott Bottom study site in the Scott Creek Catchment. (c.) Location of the nested piezometers, open groundwater wells, pluviometer (rainfall collector is adjacent to pluviometer) and Scott Bottom gauging station at the Scott Bottom study site.

2.3 Methods

An integrated approach was employed combining physical, hydrogeological, hydrochemical and environmental tracer measurements made in the field. The hydrogeological and hydraulic data was used to establish the groundwater flowpaths, hydraulic responses and relative flow activity of the shallow saprolite and deeper fractured bedrock geologic zones over the duration of the study period, between July 2005 and December 2007. The hydrochemistry and environmental tracers were used to identify the contributing end members in this flow system, mixing processes and for comparison with the hydraulic data in order to reinforce our hydraulic conceptual model of the system.

2.3.1 Piezometer Installation

Forty-two 50 mm PVC piezometers were installed at Scott Bottom in July 2005 to complement the existing monitoring well network at the site and to monitor the groundwater processes in the soil, saprolite and fractured bedrock zones. The existing network includes eight larger diameter open wells of up to 96 m depth completed in the fractured bedrock aquifer, which were constructed in 2002. The construction details and preliminary hydrogeologic investigation of these wells are described in Harrington (2004), Harrington et al. (2004) and James- Smith and Harrington (2002). The piezometers are arranged in six nests situated along a transect perpendicular to the creek valley, on an inferred groundwater flowpath covering a distance of about 330 m (Figure 2.3). There are four nests on the southern side (A, B, C and D) and two nests on the northern side of the creek (E and F), with depths varying from 1.5 to 28.5 m (Cranswick, 2005). Piezometer screen intervals were no greater than 3 m in length to target discrete groundwater flow zones in the soil, saprolite and fractured bedrock which allows representative samples from the aquifer to be obtained for geochemical analyses (Shapiro, 2002). The construction details of each piezometer are shown in Table 2.1. The ground elevation and piezometers were surveyed and the water table elevations were corrected to a reduced standing water level (RSWL) relative to the Australian Height Datum (mAHD; i.e. the mean sea level around the coast of the Australian continent). Manual water level measurements in the piezometers were conducted on a monthly basis from July 2005 until December 2007. Stream water level measurements were monitored in Scott Creek in between nests D and E (in line with the piezometer transect) on a monthly basis and compared with the continuous streamflow data from the gauge station located approximately 50 m upstream of the transect.

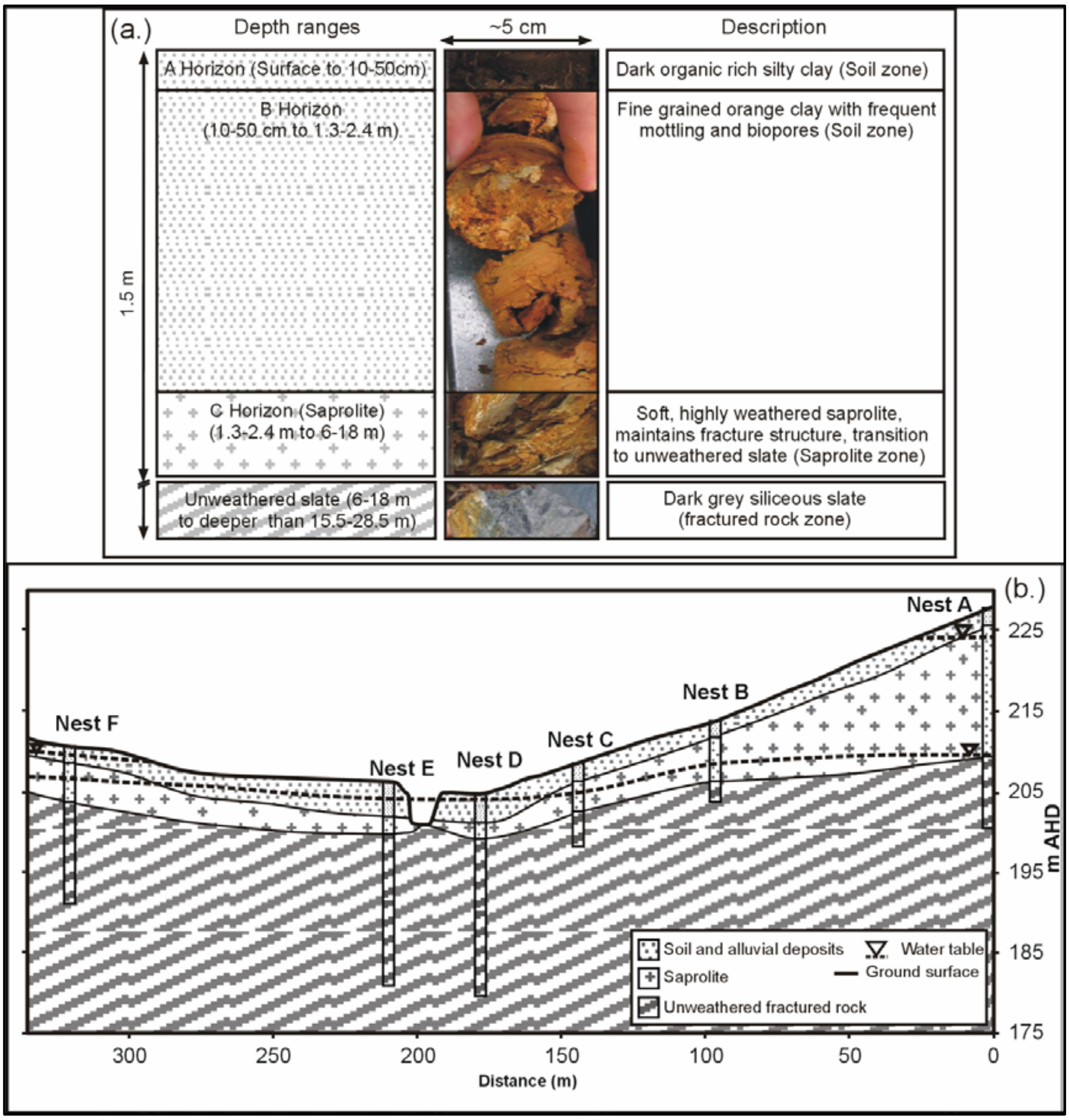


Figure 2.3 (a.) Summary of the drilling program, depth ranges and description of the A and B horizons (soil zone), saprolite and unweathered bedrock with photographs of each zone. (b.) Cross-section of the nested piezometer transect from nest F (northern extent) to nest A (southern extent) showing depths of piezometers and inferred lithology at the site. mAHD metres above Australian Height Datum; RSWL reduced standing water level.

2.3.2 Soil and Aquifer Hydraulic Conductivity

The hydraulic conductivity of the soil, saprolite and fractured bedrock was determined in order to estimate the relative activity of each geologic zone in controlling the hydrologic fluxes and gw–sw connections. Field-saturated hydraulic conductivity of the shallow soil zone was determined at the study site using the Guelph permeameter method which has been widely applied as a field technique for determining soil hydraulic properties (Bagarello and Giordano, 1999; Elrick and Reynolds, 1992). Tests were conducted next to the six piezometer nests at 5, 10 and 15 cm depth below ground. One test was also conducted at the base of an excavated soil pit on the sandstone plateau, 200 cm below ground level. When a constant depth of ponding (*H*) in mm is maintained in the borehole, the field-saturated hydraulic conductivity (K_{fs}) can be determined by:

$$K_{fs} = \frac{Q}{2\pi H^2}\left\{\ln\left[\left(\frac{H}{r}\right) + \left(\frac{H^2}{r^2} + 1\right)^{1/2}\right] - 1\right\} \qquad \text{(2.1) (Elrick and Reynolds, 1992)}$$

where *r* is the radius of the borehole, *Q* is the steady-state volumetric flow rate calculated from the cross sectional area (A) of the tube and measurement of the rate of fall in the reservoir over time (LT^{-1}). For this particular Guelph design and our experiment, a constant head value of 50 mm and radius of the borehole equal to 28 mm were used.

Single-well aquifer tests were conducted on six of the piezometers at nests D, E and F to characterise the hydraulic conductivity of the fractured bedrock zone so that groundwater fluxes could be estimated at the study site. Drawdown was measured in the piezometers located above and below the piezometer being pumped to monitor the vertical connection between the piezometers. Aquifer tests are typically more suited to sedimentary systems because the models for interpreting the test data represent the aquifer as a homogeneous and isotropic porous medium. However, their application to nested piezometers in fractured rock aquifers can provide valuable information on the vertical variation of hydraulic conductivity and can be used to derive other physical characteristics of the aquifer (Cook, 2003). Where aquifer tests could not be conducted, bail tests were done, using the method described by Hvorslev (1951).

The Cooper-Jacob straight-line method (Fetter, 2001) was used to determine the aquifer transmissivity, where:

$$T = \frac{2.3Q}{4\pi\Delta(h_0 - h)} \qquad (2.2)$$

T is the transmissivity [L^2T^{-1}], Q is the constant discharge from the pump [L^3T^{-1}] and $\Delta(h_0-h)$ is the drawdown per log cycle of time [L] (L is length and T is time). The transmissivity of the aquifer was used to determine the bulk hydraulic conductivity [LT^{-1}] K_b over the length of the screen interval, *b*.

2.3.3 Stream and Groundwater Sampling

A YSI multi-parameter meter was used to measure the pH, electrical conductivity (EC), dissolved oxygen (DO), redox potential and temperature in the creek and also during purging of the piezometers using a flow-through cell. Alkalinity was also measured in the field using a HACH titration kit. Prior to sampling the piezometers, the static water level was measured from top of casing (TOC) using an electric water level indicator. Samples were collected after purging the piezometers and once the physical parameters had stabilised, indicating that the sample was representative of the section of the aquifer sampled. Only the piezometers that were wet could be sampled and these are shown in Table 2.1.

Major ion analyses were conducted on the surface water and groundwater samples that were filtered through a 0.45-µm membrane filter. Cations were acidified with nitric acid (1% v/v HNO_3) to keep the ions in solution and analysed by a Spectro CIROS Radial Inductively Coupled Plasma Optical Emission Spectrometer at CSIRO Land and Water Analytical Services, Adelaide, South Australia. Anions were analysed using a Dionex ICS-2500 Ion Chromatograph.

Groundwater samples were collected for CFC-11 (trichlorofluoromethane, $CFCl_3$) and CFC-12 (dichlorodifluoromethane, CF_2Cl_2) analysis, using a N_2-pressurised gas bailer. Analysis of CFCs is by purge and trap gas chromatography. Corresponding analytical errors in apparent CFC ages are approximately ±2 years for ages less than 20 years, increasing to ±4 years for ages of 30 years and higher. The detection limit for both CFCs is approximately 5 pg kg^{-1}, which equates to an age of water dating to approximately the year 1961 (Busenberg and Plummer, 1992).

All isotopic concentrations were measured by isotope ratio mass spectrometry using a Europa Geo 20-20 at the CSIRO Land and Water Isotope Analysis Service in Adelaide, South Australia. δ^2H and $\delta^{18}O$ were analysed by H_2O reduction to H_2 (for δ^2H) by hot Uranium (Dighton et al., 1997) and CO_2 equilibrium for $\delta^{18}O$ (Socki et al., 1992). The results are reported as a deviation from Vienna Standard Mean Ocean Water (vs. VSMOW) in per mil (‰) difference using delta (δ) notation. The analytical precision for $\delta^{18}O$ and δ^2H is ±0.15

and ±1.5‰, respectively. The local meteoric water line (LMWL) for Adelaide, the closest rainfall station located approximately 30 km to the northwest of SCC with reasonable isotopic records is $\delta^2H=7.7\times\delta^{18}O+9.6$ (International Atomic Energy Agency (IAEA), 2001). Data from the Adelaide rainfall station, provided by the International Atomic Energy Agency (IAEA) Global Network of Isotopes in Precipitation (GNIP) service, were collected from 1962 and ceased in 1984 (International Atomic Energy Agency (IAEA), 2001). The GNIP-derived LMWL for Adelaide is in similar agreement with the LMWL calculated for Bedford Park ($\delta^2H=7.5\times\delta^{18}O+11.2$) at the western foothills of the MLR (Kayaalp, 2001) and the LMWL calculated for Scott Bottom ($\delta^2H=6.6\times\delta^{18}O+6.1$) based on several years of data between 2004 and 2007. Local rainfall at the Scott Bottom site was collected using a 200-mm-wide funnel inserted into a 1.25-L bottle held 1 m above ground in 100-mm PVC stormwater pipe next to the pluviometer at the site. Approximately 10 mm of paraffin oil was added to each new collection bottle prior to deployment to prevent isotopic fractionation from evaporation of the rainfall sample, which was collected on a monthly basis.

Table 2.1 Construction details of the nested piezometers along the hillslope transect and open wells at Scott Bottom.

Site ID	Lithology	Well depth	Mid-screen depth	Screen length	Date sampled	Temp	EC	DO	pH	Field Eh	HCO_3	Br	Cl	SO_4	Ca	K	Mg	Na	TDS	δ^2H	$\delta^{18}O$	CFC 11	CFC 11	CFC 12	CFC 12
		m	m	m		°C	µS/cm	mg/L		mV	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	‰ rel VSMOW	‰ rel VSMOW	pptv	age	pptv	age
A1	Woolshed Flat Shale	28.5	27.0	3.0	01/08/05	18.9	9650	2.8	7.2	98	331	11.6	3220	521	529	23	439	994	6069	-25.0	-4.69	43	1969	90	1968
A2	Woolshed Flat Shale	20.5	19.0	3.0	01/08/05				7.2		311	36.6	10100	780	1300	48	1550	2210	16336	N/S	N/S	N/S		N/S	
A8	Sandy Clay	1.5	1.1	0.8	01/08/05						40	0.4	232	30	19	2.6	13	136	474	N/S	N/S	N/S		N/S	
B4	Woolshed Flat Shale	10.5	9.5	2.0	28/07/05	18.5	2590	2.1	7.2	151	403	2.1	565	158	128	8	80	303	1647	-24.2	-4.78	68	1972	196	1974
B5	Woolshed Flat Shale	7.7	7.2	1.0	28/07/05	18.3	2590	2.0	7.2	166	401	2.1	565	156	127	8	79	301	1639	-24.2	-4.64	76	1972	266	1979
C4	Woolshed Flat Shale	10.5	10.0	1.0	28/07/05	17.9	2030	0.1	7.1	43	456	1.6	479	207	128	8	80	289	1649	-24.5	-4.85	26	1965	228	1976
C5	Woolshed Flat Shale	7.5	7.0	1.0	28/07/05				7.3		424	1.7	552	167	126	8	80	306	1664	-25.7	-4.88	71	1972	293	1980
D1	Woolshed Flat Shale	25.5	24.0	3.0	01/08/05	15.5	3310	0.6	7.3	-81	422	2.5	807	305	194	15	115	381	2242	-25.6	-4.76	<25	<1965	56	<1965
D2	Woolshed Flat Shale	20.5	19.0	3.0	28/07/05	15.5	3230	0.0	7.2	-74	356	2.4	748	299	176	10	117	327	2035	-24.6	-4.79	<25	1965	102	1969
D3	Woolshed Flat Shale	16.1	14.6	3.0	28/07/05	15.4	2770	3.0	7.4	-166	362	2.1	646	267	160	12	102	303	1855	-24.2	-4.68	37	1968	231	1976
D4	Woolshed Flat Shale	10.5	10.0	1.0	28/07/05	15.5	3150	0.0	7.2	-49	338	2.4	718	416	199	11	108	351	2142	-24.6	-4.87	<25	<1965	99	1969
D5	Woolshed Flat Shale	7.5	7.0	1.0	28/07/05	15.1	3290	1.6	7.3	18	319	2.4	715	435	198	11	107	355	2141	-26.3	-4.99	<25	<1965	108	1969
D6A	Saprolite	4.3	3.8	1.0	28/07/05	14.1	2350	2.6	7.2	37.5	473	1.3	413	247	135	9	85	252	1614	-26.1	-4.97	100	1974	340	1983
D6B	Saprolite	4.5	4.0	1.0	28/07/05	13.8	2770	2.7	7.2	34	394	1.7	523	366	162	8	101	286	1842	-26.8	-5.06	60	1971	247	1977
D7	Sandy Clay	2.5	2.0	1.0	01/08/05	14.3	4400	3.5	7.3	120	608	3.2	1062	476	268	16	175	447	3055	-26.6	-4.97	N/S		N/S	
D8	Sandy Clay	1.5	1.1	0.8	28/07/05		3490		7.9		198	2.8	1030	261	206	6	174	408	2287	-21.2	-4.37	N/S		N/S	

Table 2.1 continued

Site ID	Lithology	Well depth	Mid-screen depth	Screen length	Date sampled	Temp	EC	DO	pH	Field Eh	HCO_3	Br	Cl	SO_4	Ca	K	Mg	Na	TDS	δ^2H	δ^{18}0	CFC 11	CFC 11	CFC 12	CFC 12
		m	m	m		^oC	µS/cm	mg/L		mV	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	‰ rel VSMOW	‰ rel VSMOW	pptv	age	pptv	age
E1	Woolshed Flat Shale	25.5	24.0	3.0	01/08/05	15.6	3480	1.1	7.5	-18	339	2.4	750	514	184	17	101	433	2341	-28.5	-5.47	<25	<1965	263	1979
E2	Woolshed Flat Shale	20.5	19.0	3.0	29/07/05	15.5	4020	0.0	7.2	-43	441	3.1	985	322	216	12	151	412	2542	-23.45	-4.63	123	1976	85	1967
E3	Woolshed Flat Shale	15.5	14.0	3.0	29/07/05	15.3	3190	3.1	7.4	-60	402	2.6	827	288	186	18	120	361	2205	-23.3	-4.65	42	1968	101	1969
E4	Woolshed Flat Shale	10.5	10.0	1.0	29/07/05	15.6	3140	0.0	7.2	-43	362	2.3	755	248	168	9	117	316	1979	-23.7	-4.67	<25	<1965	103	1969
E5	Saprolite	7.5	7.0	1.0	01/08/05	15.1	2730	2.4	7.4	37	344	2.0	635	279	155	13	100	302	1831	-25.2	-4.93	<25	<1965	77	1967
E6	Saprolite	4.5	4.0	1.0	29/07/05	14.5	3650	5.4	7.2	129	328	2.6	877	343	176	11	136	377	2251	-23.1	-4.59	86	1973	259	1979
F2	Woolshed Flat Shale	20.0	18.5	3.0	29/07/05	16.5	3120	-0.1	7.1	-25	363	2.3	692	319	153	13	118	329	1988	-29.5	-5.35	<25	<1965	<50	<1965
F3	Woolshed Flat Shale	15.5	14.0	3.0	29/07/05	16.1	3090	0.1	7.1	-18	369	2.3	696	313	150	13	119	334	1997	-28.6	-5.53	<25	<1965	<50	<1965
F4	Woolshed Flat Shale	10.5	10.0	1.0	29/07/05	16.1	2790	0.8	7.2	-50	367	2.1	537	248	139	12	103	294	1702	-29.2	-5.3	<25	<1965	<50	<1965
F5	Saprolite	7.5	7.0	1.0	29/07/05	16.1	1006	4.0	7.6	78	255	0.5	155	54	67	3	37	88	660	-23.4	-4.64	106	1975	309	1982
F6	Saprolite	4.5	4.0	1.0	29/07/05	15.1	340	7.4	7.5	57	127	0.2	51	11	25	1	14	37	267	-19.8	-4.41	157	1980	457	1990
F7	Saprolite	2.5	2.0	1.0	29/07/05	15.5	842	7.5	7.9	13	340	0.3	81	35	23	1	31	130	641	-20.9	-4.4	N/S		N/S	
F8	Clay	1.5	1.1	0.8	14/11/05	15.7	2554				51	1.8	727	124	37	3	57	368	1369	-19.8	-4.3	N/S		N/S	
662710650	Woolshed Flat Shale	52.6	26.3	O/H	11/10/06		2561	0.02	6.8	-112	410	2.0	582	540	211	13	121	354	2233	-29.23	-5.46	N/S		N/S	
662710655	Woolshed Flat Shale	52.6	26.3	O/H	11/10/06		2767	0.01	12.	-33	372	1.9	565	369	192	10	124	303	1937	-28.49	-5.62	N/S		N/S	
662710653	Woolshed Flat Shale	58.2	29.1	O/H	13/02/07	19.8	1997	0.5	7.7	-24	420	1.8	559	267	157	9	131	244	1789	-30.33	-5.48	N/S		N/S	

Measured physical parameters, major ion chemistry, CFCs and the stable isotope results for the corresponding piezometers, open wells are also shown. pptv equivalent atmospheric concentration in parts per trillion volume; N/S no sample taken; O/H open hole.

2.4 Results and Discussion

This section discusses the hydraulic data along the hillslope transect and the surface water and groundwater chemistry results from the nested piezometers. It presents a conceptual model of groundwater flow dynamics and mechanisms of streamflow generation at the study site Scott Bottom. Importantly, the relative importance of subsurface flow activity in both of the shallow saprolite and deeper fractured bedrock geologic compartments is assessed.

2.4.1 Hydrogeological Characterisation

Drilling investigations for 42 shallow piezometers in July 2005 at Scott Bottom showed that there were four distinct soil/rock horizons (Figs. 2.2 and 2.3). From the ground surface to 10–50 cm depth exists an A horizon composed of loose, organic-rich silty clay soils. An abrupt textural contrast was observed between the A and B horizons, and is characteristic of a duplex soil (Chittleborough, 1992). The B horizon ranged in depth from 10–50 cm to 1.3–2.4 m before a gradual transition to fragmented saprolite. Varying degrees of mottling were observed in the B horizon with the presence of biopores decreasing distinctly with depth. The saprolite ranged in depth from 1.3–2.4 to 6–18 m below ground. Extensive weathering was observed in this zone with both visible leached profiles and iron oxide staining. Within the saprolite there was a transition with depth from unconsolidated and highly porous to progressively less weathered material, which retained the structure of the parent bedrock. Groundwater flow in the saprolite zone is expected to occur primarily through the permeable material, and from observation of drill logs and excavated pits, was extensively fractured with a fracture spacing less than 0.1 m. The unweathered bedrock (Woolshed Flat Shale) below the saprolite was a grey siliceous slate, containing frequent quartz veining and pyrite precipitates. Acoustic borehole televiewer (BHTV), geophysical and video camera logs of the existing 96 m deep open hole in the Woolshed Flat Shale at Scott Bottom identified a structurally complex deformed and inclined fold sequence.

Groundwater flow in the fractured bedrock is likely to be dominantly controlled by the fracture network density, geometry, connectivity and mineralization. According to the BHTV and calliper logs, there are at least 196 significant fractures and 5 dominant fracture sets. These fracture sets include bedding, bedding-parallel, two main joint sets and some random fractures. The dip and dip direction of these sets are: set A 39/220, set B 55/050, set C 53/292, set D 62/340 and set E 52/144, respectively. These results suggest that there is likely to be some strong bedding controls but no strong structural anisotropy, which is likely to be related to the complexity of the deformation and folding of the bedrock. Significant conductive fractures and fracture set spacings at intervals between 0.1 and 1 m over the entire depth of the hole were observed (Department of Water, Land and Biodiversity Conservation, unpublished data, 2007). These results matched the outcrop mapping of the Woolshed Flat Shale unit at a nearby road cutting and mine located less than 1 km from Scott Bottom. The average fracture spacing of the bedrock was 0.21 m. For this study the groundwater system has been partitioned, according to the change in geology, into three main groundwater flow zones: soil zone (0 to 1.3–2.4 m), saprolite zone (1.3–2.4 to 6–18 m) and fractured bedrock zone (deeper than 6–18 m). However, it is the relative activity of the saprolite zone compared to the fractured bedrock zone that is of greater interest in this study.

The single-well aquifer tests conducted in the piezometers constructed in the fractured bedrock (Woolshed Flat Shale) at different rates of constant discharge showed that the average bulk hydraulic conductivity was 5 m/day, and ranged from 1.5 m/day to 14 m/day. The monitored water levels in the piezometers constructed above and below the pumped piezometer only showed minimal drawdown during the 100-min aquifer tests. The lack of drawdown suggests that the vertical connection between piezometers via the fracture network is limited and/or the horizontal connection of the fracture network is extensive and that there is a high horizontal to vertical hydraulic conductivity anisotropy ratio. Changing the rate of constant discharge from ca. 13–50 m^3/day (maximum pumping rate) for the tests did not significantly affect the degree of connection between the pumped and monitored piezometers. For the tests with a higher discharge rate, the calculated hydraulic conductivity was less than one order of magnitude different when compared to the tests conducted with a lower discharge rate. This suggests that the rate of discharge has little effect on the estimated hydraulic conductivity of the fractured bedrock but that the test was limited by the maximum pumping rate (and hence sampling spatial scale), which could be employed here. Investigations in porous and heterogeneous carbonate rocks by Schulze-Makuch and Cherkauer (1998) found that increases in apparent hydraulic conductivity are dependent on the scale of the aquifer test and not the method of measurement. They also noted that the scale dependency is caused by heterogeneities and connectedness within the aquifer. As discussed earlier, the high fracture density and orientation of the major fracture sets within the Woolshed Flat Shale unit implies that there is laterally extensive, well-connected fractures along bedding and bedding parallel. In the vertical direction, fracture connectivity is likely to be limited to two major joint sets. The fracture density of the fractured bedrock zone (average=0.21 m) and the high measured hydraulic conductivity from the aquifer tests is typical of a highly conductive aquifer with the groundwater flow controlled by well connected and conductive fractures (Gburek et al., 1999). Direct connectivity between the fractured bedrock zone and Scott Creek at Scott Bottom is also evident. During the drilling of the piezometers into the fractured rock at nest D and during development of the well, a steady line of rising air bubbles in Scott Creek was observed indicating that there is a direct connection between the fractured rock aquifer and the creek.

Investigations by Mortimer (2009) and Skinner and Heinson (2004) in the Clare Valley (approximately 110 km to the north of Scott Bottom and in the same regional geology) found that near vertical structures bedding planes dominate the hydraulically conductive fractures and that near-horizontal structures are critical to fracture network connection. Mortimer (2009) also identified that groundwater flow in fractured rock aquifers can be affected by in-situ stress fields at depths less than 200 m. Whilst the palaeo-stress regimes play the primary role in creating the original properties of the fractured rock (orientation, fracture density, length, etc) and hence the dominant influence on groundwater flow, the more recent in-situ stress fields tend to alter fracture hydraulic aperture distributions and fracture network connectivity. The surface processes of weathering, erosion and unloading also contributes to spatial heterogeneity in the fractured rock and the depth dependent changes in fracture density have a significant influence on groundwater flow systems.

Single well aquifer tests were attempted on three of the piezometers that were located in the saprolite zone. However, the piezometers ran dry even at very low pumping rates. As an alternative, bail tests were conducted on these piezometers using the method described by Hvorslev (1951). The hydraulic conductivity in the saprolite zone ranged from 0.04–2.5 m/day and a mean of 1.5 m/day. Whilst it indicates considerable heterogeneity in the saprolite, it is generally less than the hydraulic conductivity values determined for the fractured bedrock zone.

Soil-saturated hydraulic conductivity was estimated to improve our understanding of the rainfall infiltration rates and the physical properties of the soil zone at the study site. The saturated hydraulic conductivity measurements using the Guelph permeameter in the soil zone between 5 and 200 cm depth at Scott Bottom were much lower than the single well aquifer tests of the fractured bedrock. Hydraulic conductivities ranged from 0.003 to 0.33 m/day, a mean of 0.045 m/day and the highest conductivity recorded at 200 cm depth. If these measurements reflect effective hydraulic conductivity of surface layer materials at the site, then only very slow groundwater recharge is expected through the unsaturated soil zone. This is qualitatively reinforced by the observations of overland flow at the site during significant rainfall events and further suggests that recharge (rates not known) must be occurring via preferential pathways. It is interesting to comment on the range of measured hydraulic conductivity values in the soil zone. A study by McKay et al. (2005) conducted in sedimentary rock saprolite in Tennessee, USA, showed that variations in hydraulic conductivity did not always correspond with changes in lithology, but a complex interaction of several factors, including parent bedrock lithology, the nature of macropores and degree of infilling with pedogenic clays and Fe/Mn oxides. Preferential weathering processes are thought to occur in highly fractured zones or in areas exposed to increased chemical or physical weathering. The potential origin and processes controlling the generation of the soils at Scott Bottom is discussed by Chittleborough (1992). The strong texture contrast between surface and subsurface horizons are important in controlling the resultant hydraulic properties of interest in soil flow analyses. Water movement in the soil zone appears to be controlled by infiltration rates through the A-horizon and the rate at which macropores are filled. A previous study by Leaney et al. (1993) concluded that flow through macropores was the major mechanism for infiltration and throughflow (horizontal movement of water through the unsaturated zone) at a site near Scott Bottom. Their results showed that the major component of throughflow at the site was precipitation associated with current storm/rainfall event rather than pre-existing water in the soil zone regardless of the magnitude of the rainfall event or season. Over the course of their study, it was observed that rainfall intensity often exceeded the infiltration rate of the soil zone during the winter months resulting in overland flow. It was not clear from drilling whether observed biopores or macropores of any other type, penetrate through the B-horizon to the saprolite. It would be expected, however, since the hillslope was only cleared of its native vegetation approximately 80 years ago, that remnant biopores from native vegetation might still be active (Harris, 1976).

2.4.2 Groundwater-Surface Water Interaction

The hydrogeological characterisation demonstrated the physical differences between the soil and saprolite zones with the fractured bedrock groundwater flow zone. The following section describes the direction and relative groundwater flow rates. Groundwater level time series data was collected over the study period to establish the hydraulic responses in each of the aquifer zones to rainfall events and hydraulic gradients at the site. Groundwater level data from the six piezometer nests along the hillslope transect and rainfall collected from 15 July 2005 until 31 December 2007 are shown as time series plots in Figure 2.4. Water levels measured at each of the nests are shown corrected to a RSWL relative to mAHD.

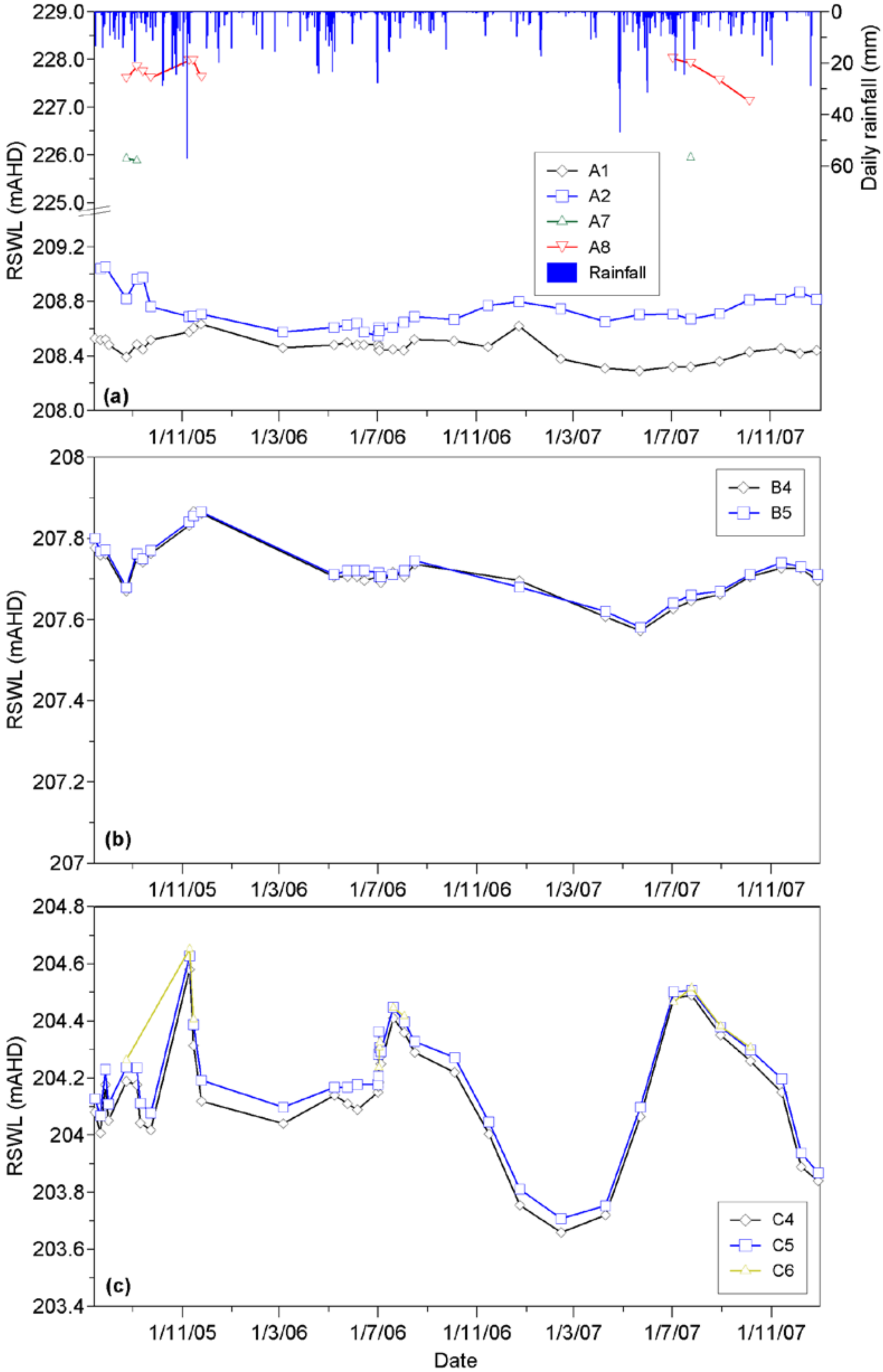
At nest A (the southern upslope end of the transect), only the two deepest piezometers A1 (28.5 m) and A2 (20.5 m) were wet for the duration of the study period (Figure 2.4a). The six shallower piezometers remained dry except for piezometer A7 (2.5 m) and A8 (1.5 m) when there was a significant late spring rainfall event in November 2005 and May through July of 2007. The vertical hydraulic gradient between the deeper piezometers A2 and A1 was in a downward direction indicating processes of groundwater recharge. The aquifer response to rainfall at nest A in the fractured bedrock zone was considerably dampened with a lag time between event and water-table/water-level rise in the order of weeks to months. The lag in response was in part due to the depth to the piezometric surface being greater than 20 m.

Similar to nest A, the shallower piezometers at nest B were dry throughout the study period except for the two deepest piezometers B4 (10.5 m) and B5 (7.65 m), located in the fractured bedrock. The vertical hydraulic gradient between these piezometers was in a downward direction for the duration of the study period (Figure 2.4b). At nest C, there was also a downward vertical hydraulic gradient between piezometers C4 (10.5 m) and C5 (7.5 m), located in the fractured bedrock. Piezometer C6 (4.5 m), located in the saprolite, was only wet during the higher rainfall events. The water level in C6 was very similar to C4 and C5 indicating that there is a strong hydraulic connection between the saprolite and fractured bedrock zone at this location (Figure 2.4c). An interesting feature of the hydraulic data at nest C was the larger piezometric surface response to rainfall and greater seasonal variation between head levels than at nests A and B. This is indicative of significant recharge processes most likely associated with rainfall events in combination with low aquifer storage.

Figure 2.4d shows a large aquifer response to rainfall events at all depths at nest D, located on the southern side of the creek. There was a large vertical hydraulic gradient upwards between the deepest piezometer D1 (25.5 m) and piezometer D6A at 4.5 m depth. Between the shallowest piezometer D8 (1.5 m) and D6A there was a vertical hydraulic gradient downwards. This indicates a convergence of shallow and deep groundwater within the saprolite zone. The piezometric surface elevations in nest D were all higher than the stage height in Scott Creek indicating that Scott Creek is gaining along this reach.

At nest E, on the northern side of Scott Creek, an upward vertical hydraulic gradient was observed between the deepest piezometer E1 (25.5 m) and E6 (4.5 m) (Figure 2.4e). The shallower piezometers E7 (2.5 m) and E8 (1.5 m) remained dry throughout the study period. Similar to nest D, the piezometric surface elevations in nest E were all higher than the stage height in Scott Creek indicating that Scott Creek is gaining along this reach. The aquifer response to rainfall at nest E was rapid and large at all depths. The observed trends in the water levels in each of the piezometers at nest E show three zones of hydraulic connectivity. These zones may relate to the physical characteristics and boundaries of the soil deposits, saprolite and the fractured bedrock at this location.

At nest F, the vertical hydraulic gradient observed between the deeper piezometers, F2 (20 m), F3 (15.5 m) and F4 (10.5 m), and the four shallower piezometers, F5 (7.5 m), F6 (4.5 m), F7 (2.5 m) and F8 (1.5 m) was upwards suggesting groundwater discharge from the fractured bedrock aquifer towards the overlying perched aquifer (Figure 2.4f). However, measured field parameters (EC, pH and temp) and observations of boggy soil conditions and permanent puddles of surface water for the duration of the study period indicated that groundwater discharge from the fractured rock zone is occurring at the ground surface, less than 10 m below nest F via a natural spring. The variations in the potentiometric surface of the fractured bedrock aquifer system at nest F is small with a dampened response to rainfall events and suggests that this aquifer may be semi-confined in this area of the site. The hydraulic head data in piezometers F2, F3 and F4 are almost identical indicating that they are hydrogeologically well connected. However, the data also suggests this group of F piezometers are hydraulically disconnected from the piezometers above. According to the drill hole logs of nest F, the transition zone between the saprolite and fractured bedrock occurs at about 10 m depth, which is in between the screen depth of piezometers F4 (10.5 m) and F5 (7.5 m). This transition zone is likely to form the base of the shallow perched aquifer above. On several occasions (October and November 2005), in response to heavy rainfall events at Scott Bottom, the water level of the perched aquifer was higher than the potentiometric surface of the fractured bedrock aquifer. This indicated possible recharge from the perched aquifer to the fractured bedrock aquifer. Frequent monitoring (days to weeks) of the water levels of the perched aquifer at nest F (F5, F6, F7 and F8) showed rapid responses to rainfall events with the water table rising to within 0.2 m of the ground surface and falling significantly over a period of weeks.



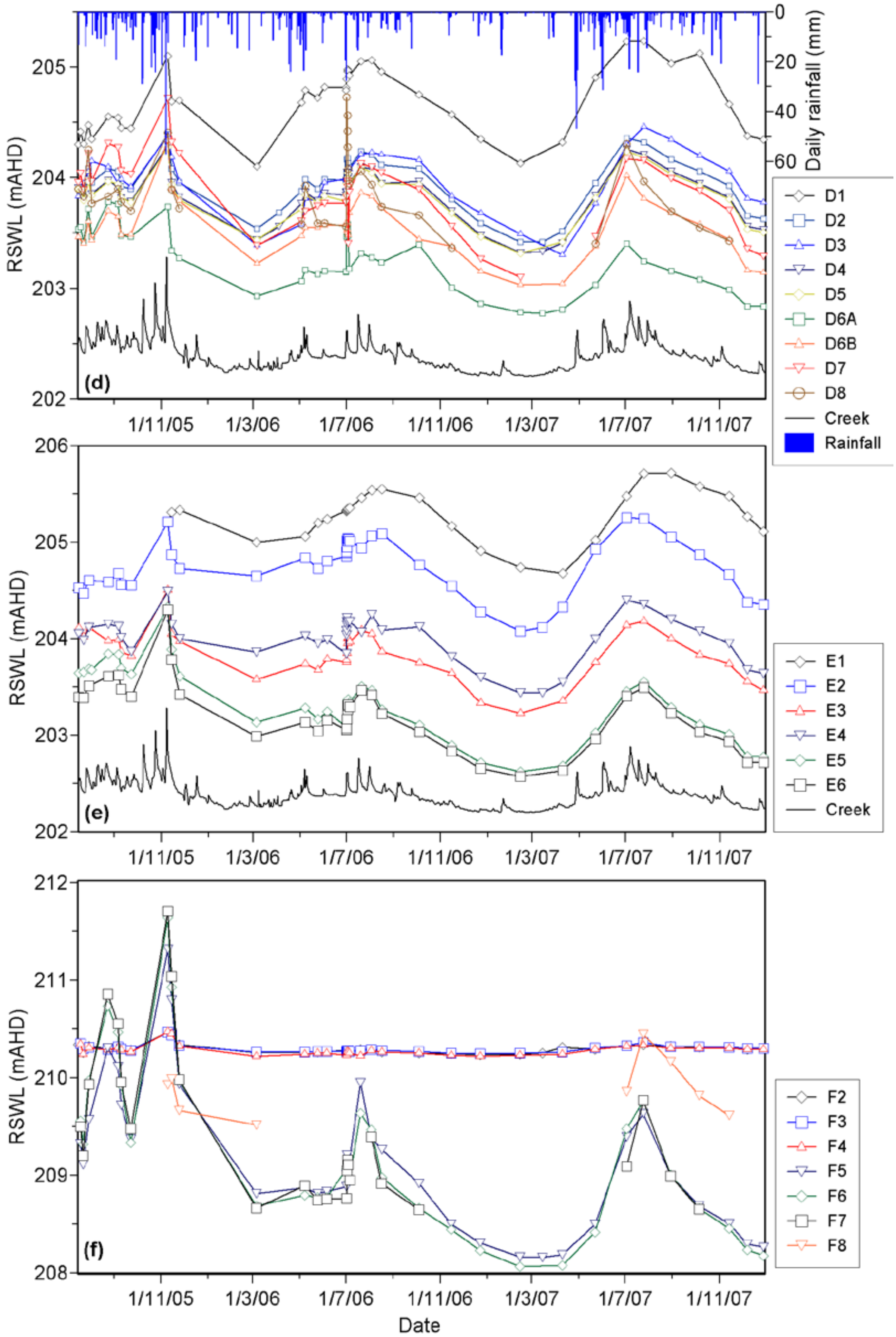


Figure 2.4 Manual water level data (corrected to RSWL mAHD) from the piezometer nests a (a), b (b), c (c), d (d), e (e) and f (f) at Scott Bottom from 15 July 2005 until 31 December 2007. Depths of piezometers are shown in the legend. Discontinuous hydrographs of some piezometers are a result of the water table falling below the base of screen. Automated rainfall data from the pluviometer and creek gauge height between nest D and E adjusted from the gauge station at the study site are also shown.

2.4.3 Hydraulic Conceptual Model

Observations of the hydraulic data from each of the piezometer nests along the hillslope transect show several interesting features which are described here as a basis for formulating the hydraulic conceptual model. At any point in time the hydraulic data shows the movement of groundwater from the elevated areas at the northern and southern extents of the transect towards the creek at the valley bottom, i.e. head gradients always point from the aquifer toward Scott Creek. This demonstrates that Scott Creek is a gaining stream. The groundwater flux to the creek is determined by the hydrogeological characteristics of the interface between the saprolite and bedrock with the creek bottom. Observations of fracture orientation and density and outcropping in the creek bed suggest that there is exchange between the fractured bedrock and the creek. Over the period of the study, the higher groundwater levels in July–September (winter) compared to the lower groundwater levels in March (baseflow only conditions) at nests D and E relative to the creek standing water level suggests that groundwater discharge to the creek would be greatest during July-September as a result of the larger hydraulic gradient. Baseflow conditions in Scott Creek typically occur from the beginning of November through to May and there are very few recorded occasions when flow has ceased completely.

In summary, the piezometer hydrographs of nests C, D, E and F show that there is a seasonal response of the potentiometric surface to groundwater recharge in the soil zone, saprolite zone and also in the fractured bedrock zone. This indicates that the fractured bedrock groundwater system is dynamic and hydraulically active. Hydraulically active refers to the responsiveness of the aquifer system to groundwater recharge and flow processes. The water level elevations in each piezometer are generally higher in July–September and have a lag time of several weeks after the rainfall event. More importantly, there are significant water level fluctuations in the fractured bedrock zone, similar to the soil and saprolite zones as observed in piezometric information. The limited hydraulic gradient data of the saprolite zone restricts any robust quantitative comparison between groundwater discharge volumes and flow velocities through the soil, saprolite and fractured bedrock zones. However, there is a discernible difference between the hydraulic conductivities of the soil, saprolite and fractured bedrock zones, which can be used to infer the relative activity of these groundwater zones. In particular, it is noted that hydraulic conductivity is a parameter, which can vary by many orders of magnitude (e.g. Love et al., 2002), but that field scale hydraulic gradients are typically less variable than the variability in K. Thus, hydraulic conductivity is a good proxy for groundwater velocity. For the surficial soil layer, the hydraulic conductivity ranges from 0.003–0.33 m/day, for the saprolite system hydraulic conductivity ranges from 0.04–2.5 m/day and in the deeper fractured bedrock it was found to range from 1.5–14 m/day (similar to a fine medium sand in a sedimentary system).

The measured horizontal hydraulic gradient in the fractured bedrock (10^{-2}) at the site is higher than what may be considered a typical field gradient (say approximately 10^{-3}) and comparable to the range in vertical hydraulic gradients between each of the groundwater flow zones at each of the nests (ca. 10^{-1}–10^{-3}). Combining the relatively high hydraulic conductivity data for the fractured bedrock aquifer with an average horizontal gradient of 10^{-2}, points to a highly active groundwater flow system in the bedrock aquifer. In contrast, the lower hydraulic conductivity of the saprolite and soil zone with a similar horizontal hydraulic gradient (based only on the gradient between nest F and E in the saprolite zone) suggests that the saprolite zone (and also the soil layer) is less transmissive than the fractured bedrock zone laterally. When these basic calculations are extrapolated to compute volumetric contributions by including layer thickness per unit width (i.e. average thickness of soil=1.85 m and saprolite=10 m), it is expected that the saprolite (and indeed soil) zones would be smaller contributors yet again. This is because they not only have smaller horizontal hydraulic conductivity and hydraulic gradients (and hence horizontal flow rates) but their effective contributing thicknesses are also significantly smaller than the deeper fractured bedrock aquifer (approximately 100 m). The calculated discharge of the soil, saprolite and fractured bedrock groundwater flow zones using the range of measured hydraulic conductivities, gradients and average thicknesses is 10^{-5}–10^{-3}, 10^{-3}–10^{-1} and 10–100 m^3/day, respectively. Whilst it is apparent that the hydraulic conductivity, groundwater flow speed, and groundwater volumetric discharge from the deeper fractured bedrock are expected to be larger than those which are likely to be encountered in the shallower saprolite and soil zones, further data would be required to verify these estimates.

2.4.4 Groundwater Chemistry Variation

The hydrogeological conceptual model has indicated an active fractured bedrock aquifer system at the study site. In the following sections, some of the hydrochemical and isotope results are compared to the hydraulic model. The groundwater samples collected between the 28 July and 2 August 2005 at Scott Bottom show chemical variability between the soil, saprolite and fractured bedrock zones and highlight the spatial differences along the hillslope transect (Figure 2.5 and Table 2.1). The plots in Figure 2.5 show the surface water and groundwater major ion (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, SO_4^{2-} and HCO_3^{-}) to chloride ratios versus chloride compared to the respective ion/chloride ratios of local rainfall collected at Scott Bottom. It is assumed that chloride behaves conservatively and therefore variations in ionic ratios are presumed to indicate addition or loss of other major ions via flow through the system. The groundwater samples from the six piezometer nests and groundwater samples from the open wells at the study site are presented in the plots relative to the lithology type at the sample depth. Surface water samples from Scott Creek taken at different times of the year are also shown for comparison (Table 2.2).

The plots of the major ion/chloride ratios versus chloride for the groundwater samples show broad linear trends of increasing individual ion concentrations relative to increased chloride concentration (Figure 2.5). The trends in Ca^{2+}, Mg^{2+}, Na^{+} and HCO_3^{-} lie slightly above the rainfall dilution line implying some degree of water–rock interaction and the weathering of primary silicate minerals. Other studies (e.g. Poulsen et al., 2006) have indicated that the majority (75%) of dissolved salts in the Mount Lofty Ranges are of marine origin (rainfall) and only a minor component (25%) comprised of terrestrial sources such as rock mineral weathering.

It would be expected that the degree of chemical weathering along the inferred groundwater flowpaths from the upslope area towards the creek (discharge zone) at the valley bottom would increase because of the longer water residence time in the aquifer. However, the hydrochemical data suggests that there are more complex Hydrogeochemical interactions taking place and that merging groundwater flowpaths and interaction between the groundwater flow zones is influencing the groundwater composition. It is worth noting that there is greater chemical variation (and in some cases higher concentrations) in the samples from the soil and saprolite zones compared to the fractured rock zone. This is shown more clearly in the depth profile plots of nest D, E and F in Figure 2.6. This may be a result of the spatial variability between atmospheric and weathering (clay soils and weathered fractured bedrock metasediments) derived sources, evaporative processes and variable rates of flushing of pore waters via discrete flow paths during recharge events. The more positive deuterium values close to the soil surface are characteristic of an evaporative process, however, in the δ^2H versus $\delta^{18}O$ plot (Figure 2.7) there is little indication of evaporative enrichment in all three groundwater zones and therefore it cannot be considered as a dominant factor to the higher solute concentrations. Whilst Scott Creek

Catchment has a large proportion of native vegetation, there are areas that have been cleared in the last 80 years including parts of Scott Bottom, which may have also affected the movement of salt in the landscape. The processes of salt mobilisation, increased recharge and stream salinisation are well documented in Australia (e.g. Peck and Hurle, 1973; Williamson et al., 1987). Vegetation clearance is likely to have resulted in increased recharge to the aquifer system and therefore the soil and saprolite zones may still be undergoing a transition to re-equilibrium.

The samples taken from the saprolite zone at nests D and E have a similar isotopic composition to the fractured bedrock zone, which may be a result of groundwater mixing before final groundwater discharge occurs to the creek from this zone, which was shown in the hydraulic data. In comparison, the groundwater samples F5, F6 and F7 from the saprolite zone at nest F have low dissolved solutes similar to rainfall and a more positive deuterium value thought to be a result of seasonal groundwater recharge events. The soil and saprolite groundwater samples are distinctly different to the samples taken from the piezometers at greater depth (F2, F3 and F4) located in the fractured bedrock, which have higher major ion concentrations, and more depleted isotopic compositions.

The dissolved oxygen and redox potential measurements from nests D, E and F show the groundwaters varying from oxidising to reducing conditions with depth, which coincide with the boundary between the saprolite and fractured bedrock zones at about 10 m (Figure 2.6). Similar changes of other physical parameters, including temperature and pH profiles, with increasing depth at nests D and E may be a result of the piezometer screens intersecting consistent fracture networks and hence similar preferential pathways of groundwater flow. The samples from the fractured bedrock at each of the nests plot in a cluster indicating that the groundwater in this zone is relatively well mixed (Figure 2.5). The samples with the higher chloride concentrations are from the deeper piezometers and the piezometers located at the valley bottom suggesting a greater groundwater residence time. The similarity of the major ion (Ca^{2+}, Mg^{2+}, Na^{+}, Cl^{-} and HCO_3^{-}) profiles at nests D, E and F illustrate that the groundwater is relatively well mixed within the fractured bedrock, and also reflects the upward hydraulic head gradients at these nests as a groundwater discharge zone (Figure 2.6). The deuterium composition of the fractured bedrock zone at nest F is considerably more depleted than in the fractured bedrock zone at nests D and E, which suggests a different source of recharge for these groundwaters. This source may be deeper groundwater flow from an intermediate flow system that has recharged the aquifer at a higher elevation. These different chemical signatures reinforce the hydraulic model described earlier, which indicated a perched separated shallow aquifer overlying the deeper fractured bedrock aquifer at this location. The creek is a mixed sample of groundwaters, soil waters and surface runoff and the chemistry at any given the groundwater pathways, residence time and mixing within the entire upstream catchment. The range of chloride concentrations and variability of the ion/chloride ratios of the samples from Scott Bottom clearly shows that the creek is comprised of a mix of different waters and that dilution and evaporative processes influence its temporal variability (Figure 2.5). The surface-water samples taken upstream and downstream of Scott Bottom show an increase in concentration downstream implying that the creek is gaining along this section. Whilst the source to the creek is not definitive in the hydrochemical data, there is evidence to suggest that there is a groundwater source from the fractured bedrock zone.

Table 2.2 Measured physical parameters, major ion chemistry and the stable isotope results for the surface water and local rainfall samples at Scott Bottom.

Site ID	Date sampled	Temp	EC	DO	pH	Field Eh	HCO_3	Br	Cl	SO_4	Ca	K	Mg	Na	TDS	δ^2H	$\delta^{18}0$
		^oC	μS/cm	mg/L		mV	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	‰ rel VSMOW	‰ rel VSMOW
Scott Creek	02/08/05	9.0	816	12.8	7.8	138	145	0.5	184	40	30	5	30	95	529	-23.2	-4.72
Scott Creek	14/11/05		442		7.4		93	0.1	75	16	16	4	15	53	271	-17.00	-3.79
Scott Creek	03/05/06						N/S	N/S	N/S	N/S	N/S	N/S	N/S	N/S	N/S	-23.6	-4.37
Scott Creek	15/11/06	15.0	1353	1.1	8.0	11	272	0.9	290	66	65	6	62	172	934	-20.6	-3.65
Scott Creek	18/12/06	18.2	1844	5.5	8.2	34	411	1.2	384	99	91	8	79	204	1277	-16.30	-3.29
Scott Creek	13/02/07	19.8	1998	0.5	7.7	-24	378	1.4	421	117	99	8	89	241	1354	-20.2	-3.38
Scott Creek	30/08/07	12.3	948	10.4	7.8	125	192	0.5	176	44	38	5	37	98	593	-23.0	-4.65
Scott Creek	15/03/07	16.6	2430	0.22	7.56	-6	368	1.7	495	165	123	9	117	271	1549	-18.5	-3.08
Scott Creek	20/06/07	8.1	855	14.7	8	92	148	0.5	173	75	37.5	4.5	40.7	110	589	-26.1	-4.67
Upstream	15/03/07	20.1	2425	0.78	7.9	11	358	1.6	478	162	122	9	113	267	1511	-20.0	-2.9
Downstream	15/03/07	16.1	2482	0.33	7.3	-124	408	1.8	547	153	119	10	109	305	1652	-16.5	-3.3
Upstream	20/06/07	8.0	850	9.83	7.9	96	144	0.5	173	78	37	5	41	108	586	-27.7	-4.6
Downstream	20/06/07	7.9	872	11.32	8.0	80	148	0.5	176	78	38	4	41	112	598	-25.7	-4.8
Monthly Rainfall	Jun-04															-1.70	-0.04
Monthly Rainfall	Oct-04															-21.20	-4.49
Monthly Rainfall	Aug-04															-26.00	-5.32
Monthly Rainfall	Feb-05															-9.00	-1.60
Monthly Rainfall	Mar-05															-19.00	-4.04
Monthly Rainfall	Apr-05															-13.90	-3.08
Monthly Rainfall	May-05															-3.90	-1.85
Monthly Rainfall	Jun-05															-24.50	-5.52
Monthly Rainfall	Jul-06															-28.8	-4.84
Monthly Rainfall	Aug-06															-8.3	-2.37
Monthly Rainfall	Sep-06															-15.6	-3.38
Monthly Rainfall	Oct-06															5.5	-0.21
Monthly Rainfall	Jan-07															-20.9	-3.68
Monthly Rainfall	Mar-07															-20.5	-4.13
Monthly Rainfall	Apr-07															-52.5	-7.43
Monthly Rainfall	May-07															-35.2	-5.16
Monthly Rainfall	Jun-07															-12.2	-2.43
Monthly Rainfall	Jul-07															-18.8	-4.01
Monthly Rainfall	Aug-07															-13.4	-3.10
Rainfall event	Jun-07															-13.0	-3.13
Rainfall event	Jul-07															-24.6	-4.98
Rainfall event	Jul-07															-18.3	-3.59

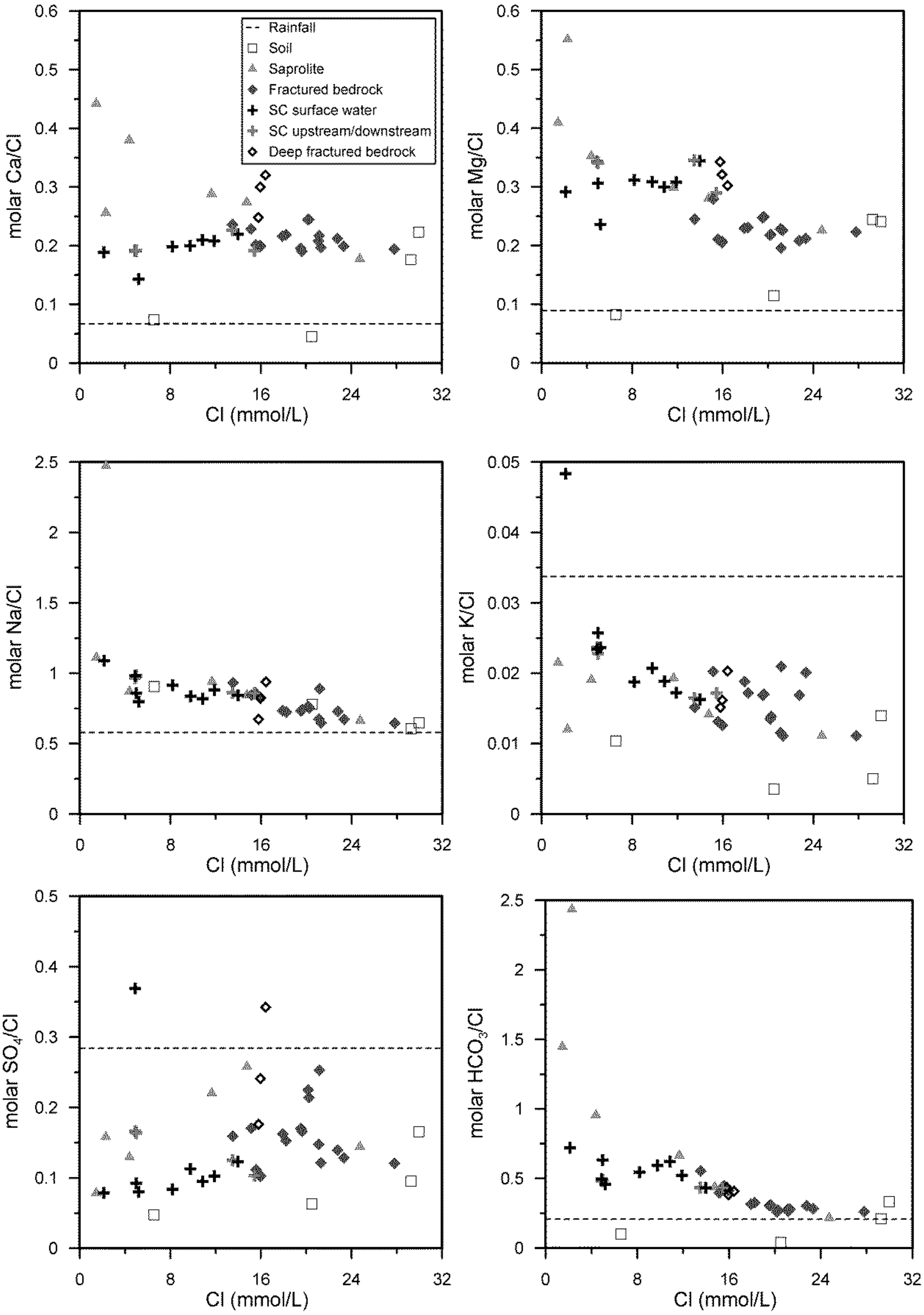


Figure 2.5 Composite diagrams of major ion (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, SO_4^{2-} and HCO_3^{-}) to chloride ratios versus chloride of surface water (August 2005 to June 2007) and groundwater collected between 28 July and 2 August 2005 at Scott Bottom.

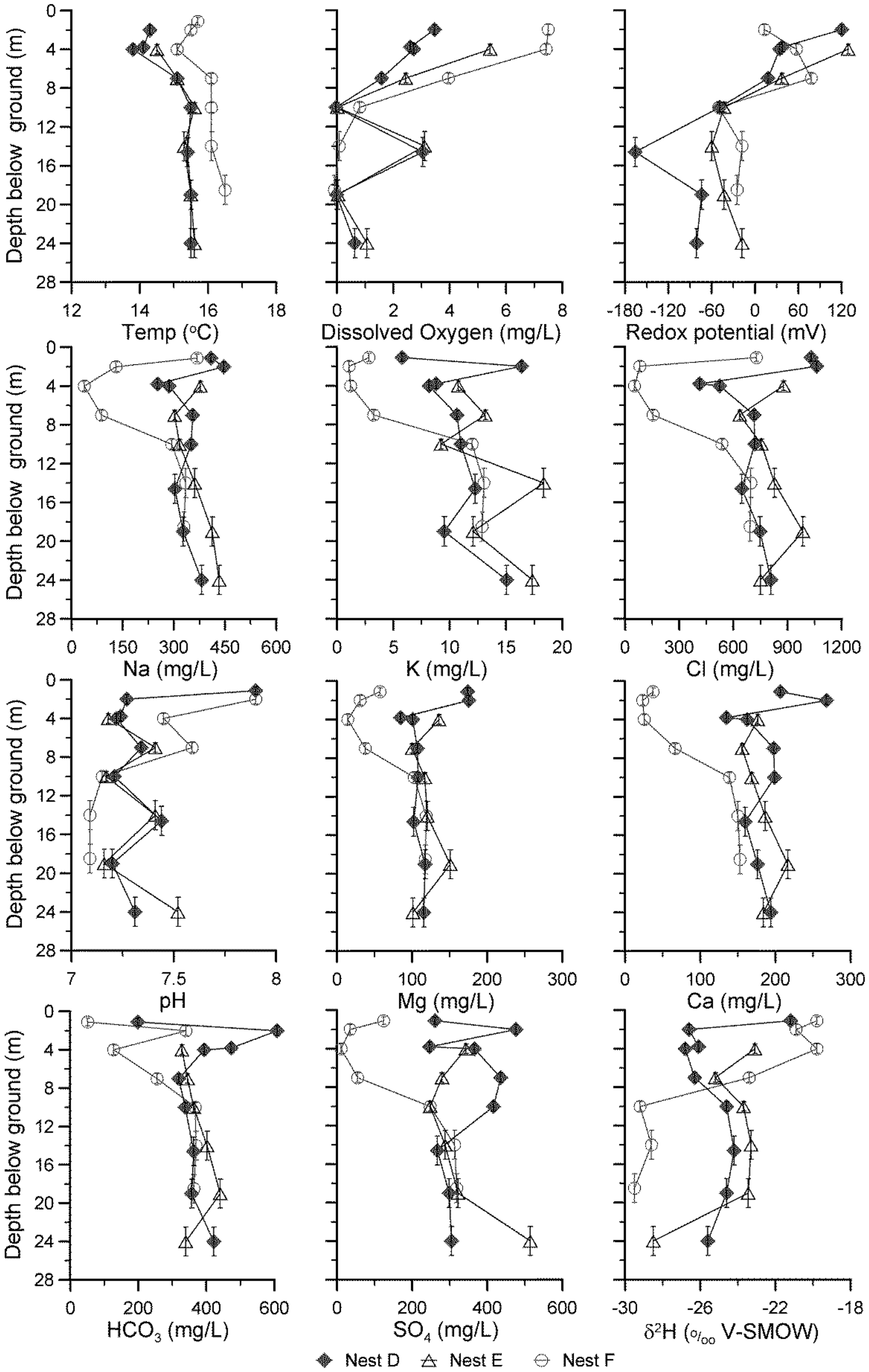


Figure 2.6 Temperature (Temp), dissolved oxygen, redox potential, pH, major ion and deuterium profiles of groundwater collected between 28 July and 2 August 2005 at nests D, E and F along the hillslope transect. The vertical error bars represent the length of the screened interval.

2.4.5 ORIGIN AND AGES OF GROUNDWATER

The isotopic ratios (δ^2H and δ^{18}O) of rainfall, surface water and groundwater samples from the Scott Bottom site are plotted in Figure 2.7, relative to the Adelaide LMWL and the calculated Scott Bottom LMWL, to investigate the potential source of the waters and in assisting the development of the hydrogeological conceptual model. The isotopic ratios of the groundwater samples from the nested piezometers range between −29.5 and −19.6‰ for δ^2H, and between −5.5 and −3.6‰ for δ^{18}O. The majority of samples plot closer to the Scott Bottom LMWL than the Adelaide LMWL and indicates that despite the limited isotopic rainfall data set at Scott Bottom it shows the likely origin of the groundwaters. The δ^2H and δ^{18}O of these samples are similar or more positive than the weighted average rainfall for Scott Bottom (δ^2H=−24‰ and δ^{18}O=−4.5‰). The more positive ratios of these samples are thought to represent a mixture of seasonal localised recharge events and/or small amounts of evaporation that may occur at high humidity. The three groundwater samples from the open wells at the study site (δ^2H range between −29.2 and 30.3‰) and samples from piezometers E1, F2, F3 and F4 plot close to the Scott Bottom LMWL and have a more depleted isotopic composition than the weighted average rainfall for Scott Bottom, which is indicative of diffuse recharge during cooler autumn and winter rainfall events, and/or altitude effects. Altitude effects on isotopic composition is a temperature-related affect and leads to a reduction in both δ^{18}O (~0.15–0.5‰ per 100 m of increase in altitude) and δ^2H (~1–4‰ per 100 m of increase in altitude) values (Clark and Fritz, 1997). The elevation of Scott Bottom is 250 m above sea level, about 200 m higher than the Adelaide GNIP station and located on the eastern side of the Mount Lofty Ranges. The surrounding hills adjacent to Scott Bottom are at an elevation of about 320 m. Therefore, the most depleted groundwater isotopic values mentioned in the preceding can most likely be attributed to altitude effects and represent a deeper intermediate groundwater flow system at the study site. Additionally, however, observations of higher rainfall and increased water table elevations during autumn and winter at Scott Bottom also provide evidence of seasonal recharge effects. Cooler rainfall would result in more depleted groundwater isotopic values relative to mean Adelaide rainfall, and therefore recharge to the deeper fractured bedrock zone may only occur during these periods.

The variability of the isotopic composition of the groundwater samples from the soil and saprolite zones is a result of seasonal recharge events of the current climate. This is comparatively different to the majority of the groundwater samples from the fractured bedrock zone, which plot in a relatively tight cluster and represent an isotopically well-mixed system where the seasonal signature has been lost to dispersive mixing. No significant deviation of the isotopic composition of the groundwater samples from the Scott Bottom LMWL was observed and suggests that there has been minimal isotopic fractionation by evaporative process prior to rainfall infiltration to the three groundwater flow zones. Similarly, the water samples collected during the study period from Scott Creek at Scott Bottom have isotopic compositions close to the LMWL. However, samples from Scott Creek collected during baseflow conditions (November–May) show signs of isotopic enrichment by evaporation, whilst samples collected during the winter period plot closer to the line and near the cluster of groundwater samples from the fractured bedrock zone and the weighted average rainfall at Scott Bottom (Figure 2.7; Table 2.2). The intersection of an evaporation trend line of the Scott Creek samples with the Scott Bottom LMWL confirms the likely sources to Scott Creek. The bulk monthly rainfall samples from the Scott Bottom site show seasonal isotopic variations as well as isotopically different rainfall events. The samples from the winter months generally fall close to the LMWL and below the average Adelaide rainfall, whilst the $\delta^{2}H$ and $\delta^{18}O$ from the summer months are generally more positive and plot near the LMWL and above the average rainfall as a result of warmer temperatures and greater evaporation during summer rainfall events (Coplen et al., 1999).

To summarise, the stable isotope data suggests that there are distinct differences between the soil, saprolite and fractured bedrock groundwater flow zones, and that there has been minimal evaporation of the water in each of these zones. Samples from some of the deepest piezometers (i.e. nests E and F) in the fractured bedrock zone appear to only be recharged during the cooler winter months after significant rainfall events and at higher elevation than Scott Bottom. These groundwater samples are likely to be sourced from a much deeper intermediate flow system because they have isotopic compositions similar to the deep open wells. The majority of the samples from the fractured bedrock zone have a similar composition to the mean-weighted Scott Bottom rainfall and represent an isotopically well-mixed system. The samples from the soil and saprolite zones show much greater variability and are influenced by seasonal recharge events. Surface-water samples at Scott Bottom, particularly those sampled during the summer months, have evolved from an isotopic composition similar to groundwater from the fractured bedrock zone as well as the soil and saprolite zones.

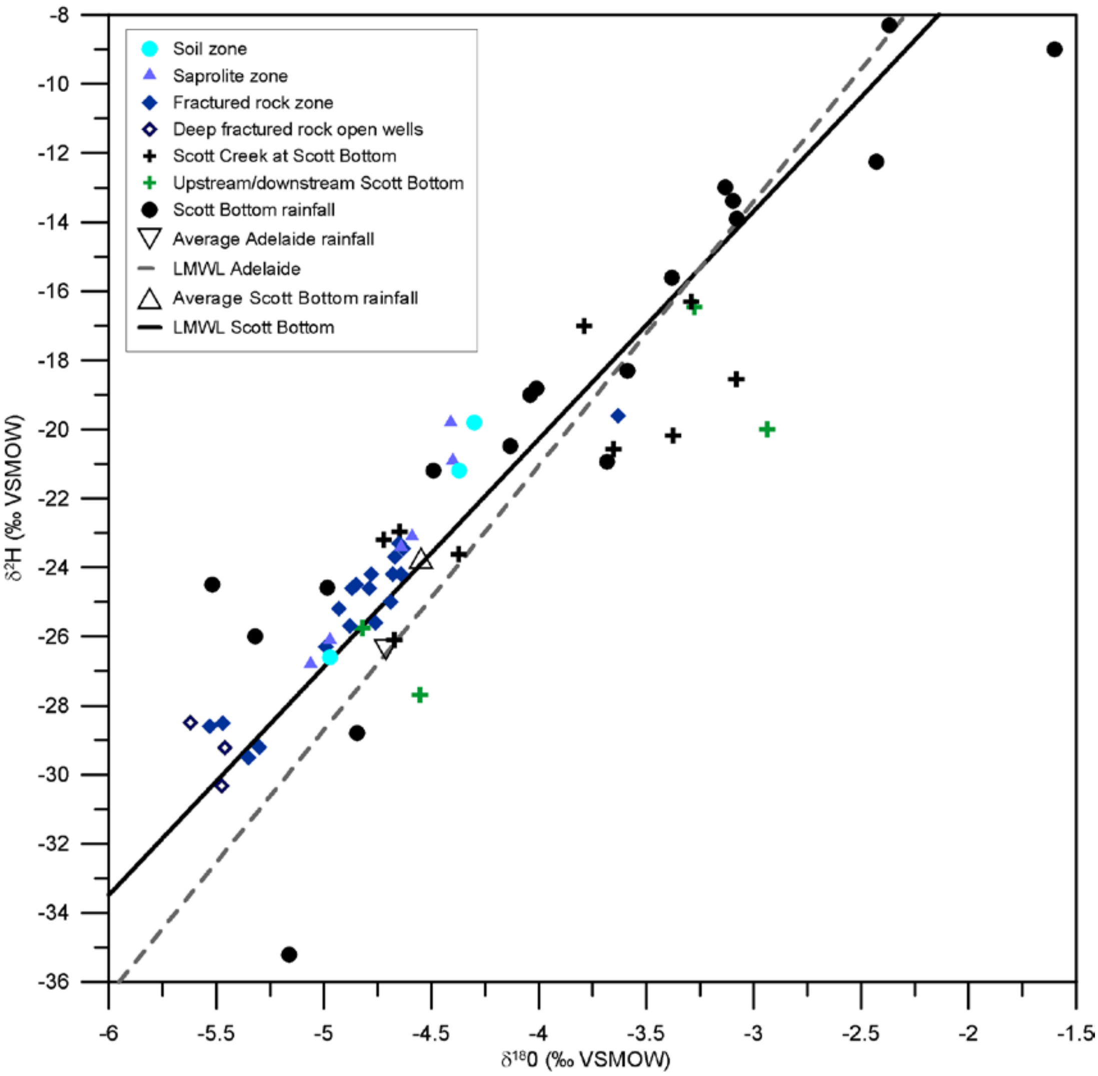


Figure 2.7 δ^2H versus δ^{18}O for local rainfall, Scott Creek and groundwater samples from the nested piezometers and open wells at Scott Bottom over the monitoring period. The LMWL for Adelaide is δ^2H=7.7 δ^{18}O+9.6 and the LMWL for Scott Bottom is δ^2H=6.6 δ^{18}O+ 6.1. The mean weighted rainfall for Adelaide is δ^2H=−26‰ VSMOW and δ^{18}O=−4.7‰ VSMOW and the mean weighted rainfall for Scott Bottom is δ^2H=−24‰VSMOW and δ^{18}O=−4.6‰ VSMOW.

CFCs were used to determine the apparent age of groundwater and to provide information on the groundwater flow processes, including depth of circulation and vertical connectivity. CFCs have been used as age indicators for groundwater studies since about 1979 (Szabo et al., 1996). Measurable concentrations of CFCs (apparent groundwater age less than 40 years old) were found in the fractured bedrock zone to depths of 20 m below the water table at the upper and lower elevations of the hillslope (Table 2.1). Detectable CFC concentrations indicate that there is active groundwater flow in the fractured bedrock zone. Although the screen intervals are short, there may also be mixing of older groundwaters with post-1950s groundwater from the shallower flow zones. The majority of the CFC-11 groundwater ages were older than the CFC-12 ages, which suggests that some retardation of CFC-11 has occurred in the unsaturated zone or degradation by microbial activity (Busenberg and Plummer, 1992). Observed fracture spacings of the fractured bedrock zone at Scott Bottom are typically small and certainly less than 1 m and as a result CFC concentrations in the fracture are expected to be in equilibrium with the pore water in the matrix (Cook et al., 2005; Cook et al., 1996). Therefore, the apparent groundwater ages can be used to indicate the minimum depth of circulation and may also be used to estimate groundwater recharge (Love et al., 2002).

CFC concentrations of samples taken from F2, F3 and F4 at nest F were below detection limit (5 pg kg^{-1}) indicating groundwater ages older than 40 years. This supports the results from the hydraulic and isotope data, which showed that the groundwater from the fractured bedrock zone at nest F is from a deeper intermediate flow system. Comparatively, CFC concentrations of sample F5 (7.5 m depth) and F6 (4.5 m depth) in the saprolite zone, represent groundwater that is relatively young (<20 years). The presence of young groundwater at this depth supports the hydraulic and hydrochemical evidence of the relatively fresh, perched aquifer at nest F.

Despite the upward hydraulic gradients in the fractured bedrock zone at nests D and E, the presence of CFCs in the fractured bedrock zone suggest that groundwater from the overlying soil and saprolite zones may be mixing with the groundwater below or that there is a source of young groundwater in the fractured bedrock up gradient of nests D and E that is moving laterally towards the stream. The hydraulics and hydrochemical data indicated that there is likely to be more dominant downward vertical groundwater flow in the soil and saprolite zones. Whereas in the fractured bedrock, the data indicated a relatively well mixed system where the dominant groundwater flow paths are horizontal and converge with groundwater flowpaths of the overlying flow zones prior to discharge to the creek or in the case of nest F at the ground surface.

Therefore, the study data has demonstrated that the fractured bedrock zone may play a more active role in streamflow generation at Scott Bottom than has previously been documented by other authors working at this site (e.g. Leaney et al., 1993) and internationally in analogous geologic systems (e.g. Manning and Caine, 2007; McDonnell, 1990; Shand et al., 2007a). These chemistry and tracer data reinforce the hydraulic data.

They are strongly suggestive of the fact that the implicit or explicit assumption of a no-flow boundary condition at the saprolite-fractured bedrock zone interface must be carefully evaluated in the conceptual model of groundwater flow and gw–sw interaction in these types of geologic systems.

2.4.6 GROUNDWATER RECHARGE

In the following, a chloride mass balance (CMB) technique is used to compare recharge rates at various locations throughout the site and in the different geologic compartments. This is useful as a further indicator of groundwater dynamics at the site. Evidence of active recharge in different geologic layers will provide another method for defining the hydraulic conceptual model (e.g., Is there any evidence of active recharge in the deeper fractured bedrock aquifer?). The CMB technique has been used successfully in sedimentary aquifer systems (e.g. Allison and Hughes, 1978) and has been suggested the most reliable technique for determining recharge rates to fractured rock aquifers systems (Cook, 2003). However, the recharge rate determined from CMB should be considered as a minimum rate because of the addition of other sources of chloride, which may occur, by rock weathering. Changes in environmental conditions (i.e. land clearing) will also impact on the equilibrium of chloride in the fractures with the rock matrix and may take a significant amount of time for the diffusion of salts from the matrix into the fractures to re-equilibrate. The subsequent steady state mass balance equation can be used to estimate recharge (R):

$$R = \frac{PC_p}{C_{gw}} \tag{2.3}$$

R is the estimated recharge [LT^{-1}], P is the average annual rainfall [LT^{-1}], C_p is the chloride concentration of rainfall [ML^{-1}], and C_{gw} is the chloride concentration of groundwater [ML^{-1}].

According to the long-term rainfall record, the average annual rainfall at Scott Bottom is about 800 mm. The average chloride concentration in rainfall from gauge stations in close proximity to SCC is about 9 mg/L. Using the measured chloride concentrations from the soil (4 samples), saprolite (6 samples) and fractured bedrock (19 samples) zones, the estimated recharge rate to the groundwater system in each of these zones is 7–31, 8–141, and 0.7–15 mm/year, respectively. The average rate of 18 mm/year (2% of rainfall) from all of these three zones is within the range of estimated recharge rates of 26 natural catchments in southern Australia documented by Scanlon et al. (2006) and similar to the estimated range of recharge rates (16–111 mm/year) in the western MLR (Green and Zulfic, 2008). There is significant variability in recharge over a large spatial scale and the range of rates from all three of these zones represents only 0.1 to 18% of the long-term average annual precipitation. As discussed earlier, there was greater variability in the hydrochemistry data in the soil and saprolite zones, and therefore the applicability of this method to these zones may not be appropriate as the system is unlikely to be in a state of equilibrium. More importantly, the recharge rates calculated from chloride data in the deeper fractured bedrock system indicate that there is effective and significant recharge to this zone, an observation that is entirely consistent with the conceptual model that recognises a dynamic and active fractured bedrock system.

2.5 Summary and Conclusions

Hydrological conceptual models of groundwater/surface water (gw–sw) interaction in a saprolite-fractured bedrock geological setting often do not consider the contribution of groundwater from below the bedrock interface (considering it a no flow or impermeable boundary). More recent studies have begun to address this issue (Haria and Shand, 2006; Manning and Caine, 2007; Shand et al., 2005). The purpose of this investigation was to determine the relative importance and contribution of both the soil-saprolite and fractured bedrock aquifer systems and their influence on gw–sw interaction. From the hydrogeology and hydrochemical results presented in this study, a conceptual model describing the significance of the fractured bedrock zone and its influence on the interactions between the groundwater system and Scott Creek at Scott Bottom was developed (Figure 2.8). This study provides an important demonstration of how geologic controls greatly influence groundwater flows.

The results presented here suggest that the relative activity of the groundwater in the fractured bedrock zone is indeed not inactive and plays just as important a role as that in the shallower soil and saprolite zones. Groundwater flow in the fractured bedrock zone is dynamic and may be an important flow pathway along the hillslope via a well-connected fracture network. The seasonal variation of the water levels in the nested piezometers and higher elevations during winter verify active groundwater recharge and discharge occurring along the hillslope and that groundwater movement is from the higher parts of the catchment to the valley bottom. Groundwater flow in the fractured bedrock aquifer appears to be significant. It is hypothesised that the hydraulic properties and high degree of connection between the soil, saprolite and fractured bedrock zones is a dominant control on the major contribution of groundwater to Scott Creek. The hydraulic and hydrochemical data suggest that the groundwater in the fractured bedrock zone is relatively well mixed and that there is some mixing of this water with the shallower groundwater in the soil and saprolite zones where the groundwater flow paths converge at the valley bottom prior to discharge to the creek. The data also shows that there is a hydraulic disconnection between the soil and saprolite zones with the deeper fractured bedrock in the upslope areas of the study site as a result of the increasing thickness of the soil and saprolite zones. Perched aquifers are established in the shallow soil zone and persist throughout the year. These shallow aquifer systems may play an important role in overland flow and throughflow to the creek as well as a 'solute' mixing zone of shallow and deep groundwater.

Previous studies (Leaney et al., 1993; Smettem et al., 1991) at the Scott Creek field site have assumed that subsurface flow occurs in the shallow saprolite zone only. In striking contrast, the results of this study are important because they clearly show that groundwater flow in the deeper fractured bedrock zone is highly dynamic and an important groundwater flow pathway along the hillslope and readily exchanges with the creek. As a result, deep groundwater from the fractured bedrock aquifer is therefore expected to be a significant component in streamflow generation at Scott Creek. These are important and new observations at the Scott Creek site, which profoundly alter the hydrogeologic conceptualisation of this system. Our results also reinforce the findings of a limited number of previous studies in similar geologic settings (Haria and Shand, 2006; Manning and Caine, 2007; Shand et al., 2005). This finding will influence the way in which gw–sw interaction is analysed at the Scott Creek site and will inform the development of quantitative numerical models of the system. More generally, however, this study suggests that hydrologic conceptual models that do not consider the deeper fractured bedrock in hydrologic analyses and hence treat the saprolite-fractured bedrock interface as a no flow boundary may be overly simplistic and inherently misleading in some groundwater/surface water interaction analyses such as the Scott Creek site presented here. The choice of conceptual model employed in the gw–sw interaction assessment (homogeneous, saprolite only, saprolite and deeper fractured bedrock) is critical. What is particularly problematic is that the choice of a conceptual model is often made implicitly without careful justification. Our results emphasise the need to understand the relative importance of subsurface flow activity in both of these shallow saprolite and deeper bedrock compartments as a basis for developing reliable conceptual hydrologic models of these systems. These choices are in turn crucial for understanding what level of simplification is permissible in conceptual model definition. Understanding fractured bedrock and saprolite hydrogeologic controls are clearly important for accurate quantitative flux analyses, streamflow interpretation and a range of other matters that arise in gw–sw interaction studies.

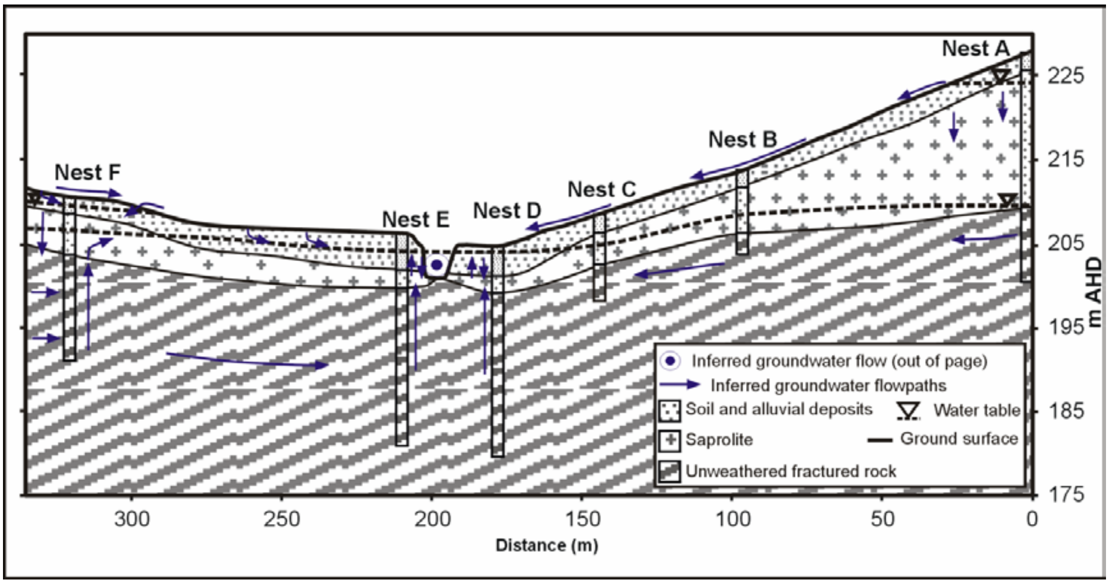


Figure 2.8 Conceptual model of the study site Scott Bottom. Arrows indicate direction of inferred groundwater flow.

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CHAPTER 3

3. MANUSCRIPT II: ASSESSING SPATIAL AND TEMPORAL STATES OF CONNECTION BETWEEN SURFACE WATER AND GROUNDWATER IN A REGIONAL CATCHMENT: IMPLICATIONS FOR REGIONAL SCALE WATER QUALITY

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ABSTRACT

It is common for surface water–groundwater interaction assessments to investigate river reaches at a local scale and as discrete individual systems, which are generally classified as connected (gaining and losing type systems) or disconnected (transitional or completely disconnected type systems). While these classifications are valid at any point in space and time, studies often fail to consider how individual river reaches function in the context of the entire regional river system (comprising multiple river reaches) and what implications this can have on water quantity and quality. In this study, spatial and temporal assessments were made in a regional catchment using hydraulic, hydrochemical, and tracer-based techniques to determine the source and loss terms of the river and groundwater system and how their relative magnitude changes along the river from the catchment headwaters towards the sea. Applying an entire regional river system assessment we demonstrate that the state of connection can change along river reaches, as well as take place concurrently at the same location. Water level data, together with salinity and stable isotope results showed that the relatively low salinity of the fresh water river system was maintained in an otherwise saline regional groundwater system by virtue of the lack of saline groundwater inputs in a river with a dominantly losing connectivity state. This losing state was strongly influenced by the high evapotranspiration of the native vegetation. By determining the state of connection between surface water and groundwater and understanding the variable and complex nature of contiguous river reaches of an entire regional river system more appropriate management practices can be employed.

3.1 INTRODUCTION

Assessments of surface water–groundwater interactions provide important information for water resource managers to evaluate water resource allocation needs and the impacts on groundwater dependent ecosystems (Sophocleous, 2002; Winter et al., 1998; Woessner, 2000). Surface water–groundwater interactions are commonly investigated at the river reach scale and generally classified as connected or disconnected type systems. Connected systems are either: (1) a gaining surface water system where groundwater discharges through the streambed to contribute to streamflow, or (2) a losing surface water system which loses water to (or recharges) the local groundwater system. Disconnected systems are defined by an unsaturated zone beneath the surface water system which loses water at a rate related to the hydrogeological properties of the streambed and the aquifer. If an unsaturated zone is maintained below a stream the infiltration rate would be independent of the hydraulic properties of an aquifer under any situation and changes in the watertable do not significantly affect the infiltration rates from the river. Disconnected systems can be: (1) completely disconnected, whereby changes in the watertable do not affect the infiltration rate from the river to the aquifer; or (2) in a 'transitional' state, part way between disconnected and connected whereby changes to the watertable may still affect the infiltration rate (Brunner et al., 2011).

The majority of surface water–groundwater interaction studies are of gaining type systems and have been well documented in the literature (Banks et al., 2009; McGlynn and McDonnell, 2003; Sklash and Farvolden, 1979). There have been relatively few studies investigating losing type systems compared to gaining systems (Ruehl et al., 2006). In the case of losing and disconnected type systems, the theoretical criteria for disconnection are easily defined, however observing and assessing them in the field is very difficult (Brunner et al., 2009a; Brunner et al., 2011; Desilets et al., 2008; Fox and Durnford, 2003). Disconnection is most often inferred on the basis of a theoretical assessment of head measurements rather than a direct field-based measurement (e.g., the measurement of an unsaturated zone beneath a river bed).

While the classifications of the different system types are valid, most studies fail to consider how individual river reaches function in the context of the entire regional river system (comprising multiple river reaches) from the headwaters to the sea or discharge point and what implications the system type can have on water quantity and quality. These classifications may only be valid at a particular point in space and time. Understanding the state of connection between surface water and groundwater along contiguous river reaches at a regional scale is critical for an accurate assessment and management of a catchments water resource. There are few, if any, studies which have examined these connectivity states, exchange fluxes, and implications for water quantity and quality at a regional scale and over time. Water quality considerations are important where fresh surface water resources lie above more saline regional groundwater. Another important factor in the classification of these system types is distinguishing between the contributing sources of groundwater to the river and the exchanges between local and regional groundwater systems. There are relatively few studies which have investigated the importance of shallow perched aquifers on sustaining river flow (Fleckenstein et al., 2006; Niswonger and Fogg, 2008), and far fewer studies in pristine catchments covered by native vegetation. This is significant because in regions of Australia, native vegetation clearance and landuse modification (e.g. deep rooted vegetation replaced by shallow rooted crops) has had considerable impacts on the water balance and surface water and groundwater salinities. This is a result of increased recharge and the mobilisation of salt stored in the shallow regolith as well as greater groundwater discharge to surface water systems (Allison et al., 1990; Bell et al., 1990; Cartwright et al., 2004). The impact on the water quality of the surface water and groundwater systems is largely determined by the state of connection between these systems.

There are various techniques that can be used to investigate surface water–groundwater connectivity and they typically depend on the type of connectivity condition being investigated and also the scale of the investigation and whether it is at the regional catchment scale or at the smaller scale such as the hyporheic zones of streams (Harvey et al., 1996; Harvey and Wagner, 2000). Numerous studies have applied hydrograph separation methods (Chapman, 1999; Pinder and Jones, 1969), where the pathways of groundwater discharge are not explicitly determined and often considered as a 'blackbox' within a single bucket type approach, whilst others have used numerical groundwater models to examine surface water- groundwater connectivity (Brunner et al., 2009b; Fox and Durnford, 2003; Osman and Bruen, 2002). Field based studies have used tracer approaches- electrical conductivity, temperature, stable isotopes and radiogenic tracers e.g., 4He and Rn-222 as an indicator of groundwater gaining conditions (Cook et al., 2003; Cook et al., 2006; Hatch et al., 2006; Hatch et al., 2010; Kalbus et al., 2006; Stellato et al., 2008).

In this paper, we investigate surface water–groundwater connectivity from the catchment headwaters to discharge point in the pristine Rocky River Catchment on Kangaroo Island, South Australia and test the hypothesis that the dominant groundwater source to the river is from the fractured rock aquifer system. Applying an entire regional river system assessment we demonstrate that the system types (or connectivity states) can change along stream reaches, vary depending on the season (variably gaining/losing systems) as well as take place concurrently at the same location. These concurrent connectivity types of systems where there is both gaining and losing type conditions at the same point in space and time have been rarely studied (Payn et al., 2009).

Spatial and temporal assessments were made along the entire length of the catchment using hydraulic, hydrochemical, and tracer-based techniques to determine the relative source and loss terms of the river and groundwater system and how their relative magnitude changes along the river from the catchment headwaters down to the sea. By determining the state of connection between surface water and groundwater and understanding the variable nature of contiguous river reaches of an entire regional river system, more appropriate management practices can be employed. The significance of this type of assessment in a pristine catchment is even greater because it can be used as a baseline to draw comparisons with catchments which have been cleared of native vegetation or undergone changes in landuse which ultimately affect the water and salt balances of a catchment. The primary objectives of this paper are: (1) to determine the state of connection between surface water and groundwater along a river system from the catchment headwaters to the discharge point at the sea, (2) to assess the relative source and loss terms of the river and groundwater system and how their relative magnitude changes along the river, and (3) to investigate the mechanisms by which a fresh water river system in a pristine catchment covered by native vegetation exists in an otherwise saline regional groundwater system.

3.1.1 Study Site Description and Geology

The Rocky River Catchment (RRC) is located within Flinders Chase National Park on Kangaroo Island, South Australia (Figure 3.1). The catchment covers an area of about 216 km^2 and is one of very few remaining catchments in South Australia, which is covered by native vegetation and has not been subject to land clearing (Henschke et al., 2003). The topography varies from relatively steep and narrow valleys in the top part of the catchment at an elevation of approximately 300 m above sea level to more subtle and wider valleys in the lower part of the catchment. Rocky River is approximately 40 km in length and its course travels in a southwest direction before discharging into Maupertuis Bay at Snake Lagoon. The river is approximately 2–4 m wide and 0.3–0.7 m deep, over the upper and middle reaches, although this does vary significantly between base flow conditions and the wetter winter months. At many locations along its length there are surface water pools, one of the biggest being the Platypus Pools. It is a semi-perennial river, flowing throughout the year in the headwaters to midway down the catchment whilst the lower part of the river occasionally ceases to flow during the drier summer months between January and April. The mean annual flow volume of Rocky River (based on data collected since 1970) at the gauge station located towards the bottom of the catchment (catchment area 189 km^2) is 1.37 x 107 m^3/y. The mean annual rainfall is 780 mm/y and the mean potential annual evaporation is 1400 mm/y. Rainfall is winter dominant with more than two thirds of the annual rainfall falling between the months of May and September (Department of Water Land and Biodiversity Conservation, 2009). It is worth noting that south eastern Australia has experienced significant drought in the last decade which has been evident by below average annual rainfall and decreases in river discharge and duration.

The geology of the RRC is characterised by a laterite plateau underlain by Cambrian and Neoproterozoic metasediments of the Kanmantoo Group (Figure 3.8). The Kanmantoo Group is composed of massively-bedded, medium to fine grained, grey feldspathic sandstone and metapelites with thin biotite laminations, which contain few joints and fractures (James and Clark, 2002). The lateritic plateau, which formed as a result of prolonged weathering and erosion, consists of a sandy A-horizon overlying a ferruginous zone up to 2 m thick. The ferruginous zone is composed of hydrated iron oxide nodules, cemented together with goethite sand and silt. This is underlain by a kaolinised zone, a few tens of metres thick and overlies the parent rock (Daily et al., 1974). Towards the coast and the bottom of the catchment, the Rocky River has incised through Quaternary calcarenite and limestone deposits, known as the Bridgewater Formation. The recent Quaternary sediments overlying the basement rocks in the RRC are characterised by a layer of unconsolidated sandy clay and clayey sand containing weathered sandstone fragments and minor gravel lenses.

3.2 METHODS

3.2.1 PIEZOMETER INSTALLATION

The interpretation of surface water–groundwater connectivity between Rocky River and the underlying aquifer systems was based on data obtained from 38 shallow, 50 mm PVC piezometers installed during the investigation (Table 3.1). The shallow piezometers (depth of piezometer ranging from 1.1 m to 45.5 m below the ground surface) were located at three study sites along the Rocky River and at one site in the headwaters of the catchment (Figure 3.1). Piezometer screen intervals were no greater than 3 m in length to target discrete groundwater flow zones in the regolith and fractured bedrock which allows representative depth samples from the aquifer to be obtained for geochemical analyses (Shapiro, 2002). The construction details of each piezometer are shown in Table 3.1. The ground elevation and piezometers were surveyed for this study and watertable elevations were corrected to a reduced standing water level (RSWL) relative to the Australian Height Datum (mAHD) (i.e., the mean sea level around the coast of the Australian continent). Manual water level measurements were taken in the river and shallow piezometers from August 2007 to April 2010 and referenced to AHD. Automatic data loggers (In-situ® aqua troll® 200) were installed in several of the piezometers at different depths to provide a continuous record of water level fluctuations over the monitoring period.

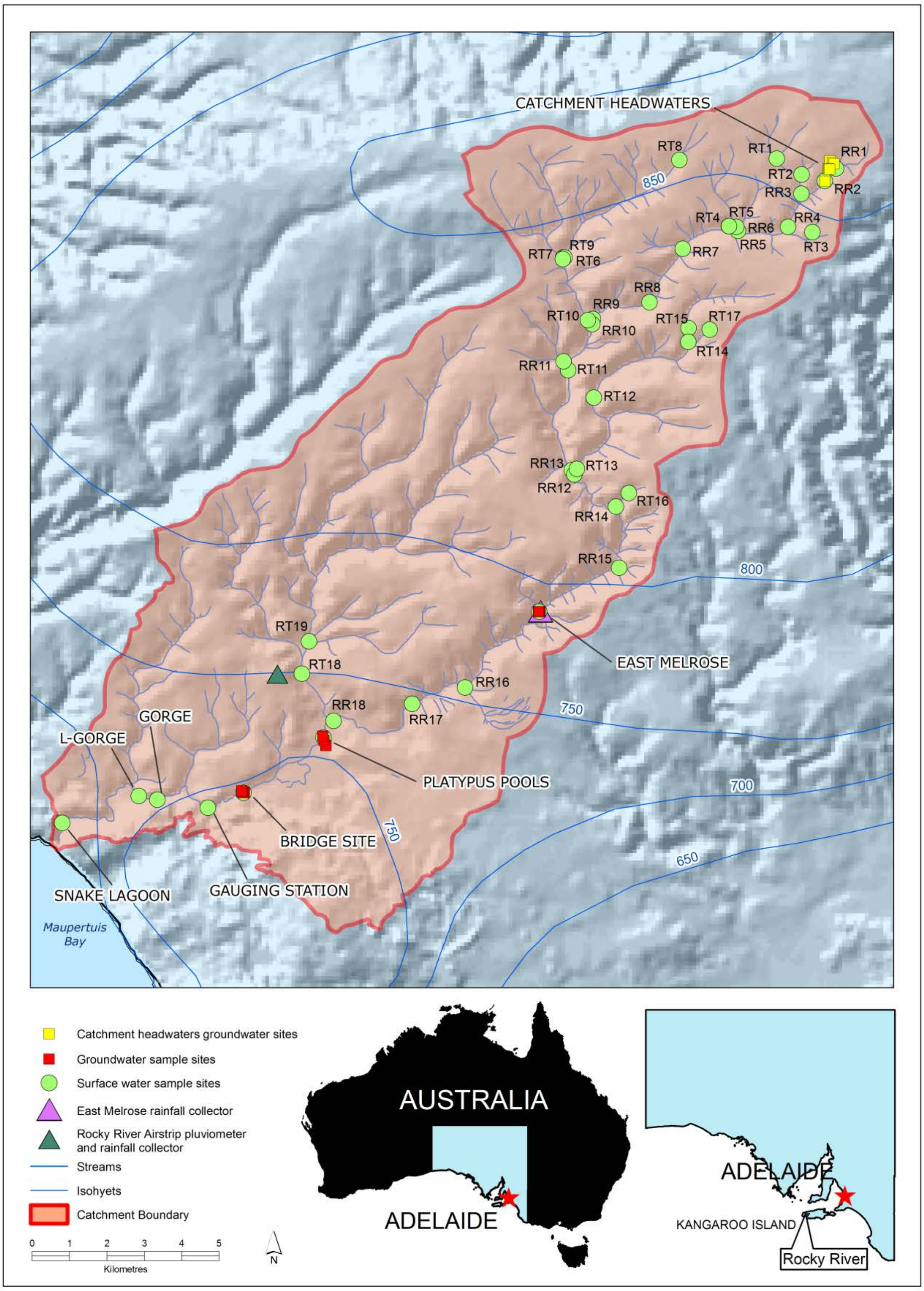


Figure 3.1 Location map of the surface water and groundwater sample locations in the Rocky River Catchment, Kangaroo Island, South Australia.

Table 3.1 Piezometer construction details at East Melrose, Platypus Pools, Bridge site and the catchment headwaters. n/a = not applicable.

Site	Piezometer site name	Eastings	Northings	Drill Date	Major lithology at screen	Ground elevation (mAHD)	Top of Casing (TOC) (mAHD)	Total depth (m)	production zone (m)
East Melrose									
Nest 1	M1a	662018	6024914	23/03/07	Sand	92.285	92.909	6	3 - 6
	M1b	662018	6024916	22/03/07	Sandstone	92.217	92.781	10	7 - 10
	M1c	662017	6024917	21/03/07	Sandstone	92.131	92.618	29.5	26.5 - 29.5
	M1d	662018	6024919	10/03/07	Sandstone	92.209	92.837	14.3	11.3 - 14.3
Nest 2	M2a	661984	6024916	23/03/07	Sandstone	86.673	86.558	7	5 - 7
	M2b	661980	6024916	23/03/07	Sandstone	86.581	86.482	5	3 - 5
Nest 3	M3a	661971	6024923	24/03/07	Sandstone	88.479	89.017	14	11 - 14
	M3b	661972	6024922	24/03/07	Sandstone	88.156	88.652	10	7 - 10
	M3c	661973	6024922	25/03/07	Sandstone	87.953	88.405	6.2	3.2 - 6.2
	M3d	661974	6024921	26/03/07	Sand	87.985	88.515	3	1 - 3
Nest 4	M4a	661998	6024916	26/03/07	Sand	89.248	89.942	4	2 - 4
	M4b	661997	6024916	26/03/07	Sand	88.918	89.509	3	1 - 3
River level monitoring gauge	MS	661985	6024919	n/a	n/a	87.568	89.622	n/a	n/a
Platypus Pools									
Nest 1	P1a	656297	6021350	01/04/07	Sand	58.456	59.023	9	6.5 - 9
	P1b	656295	6021349	01/04/07	Sand	58.525	59.217	15	12 - 15
	P1c	656293	6021347	31/03/07	Sand	58.500	59.141	28.5	25.5 - 28.5
	P1d	656298	6021347	07/03/07	Sandstone	58.564	59.247	45.5	42.5 - 45.5
Nest 2	P2a	656247	6021479	02/04/07	Sand	56.998	57.628	3	1 - 3
	P2b	656246	6021481	02/04/07	Clay	56.984	57.608	5	2 - 5
	P2c	656247	6021482	02/04/07	Clay	56.969	57.601	9	6 - 9
	P2d	656245	6021479	02/04/07	Clay	57.024	57.602	15	12 - 15
Nest 3	P3a	656214	6021595	04/04/07	Sand	56.238	56.735	5	2 - 5
	P3b	656214	6021597	04/04/07	Sand	56.313	56.923	8	5 - 8
	P3c	656213	6021599	04/04/07	Sand	56.307	56.909	14.5	11.5 - 14.5
	P3d	656213	6021594	02/04/07	Sand	56.204	56.637	27.5	24.5 - 27.5
River level monitoring gauge	PS	656227	6021555	n/a	n/a	56.570	58.438	n/a	n/a

Table 3.1 continued

Site	Piezometer site name	Eastings	Northings	Drill Date	Major lithology at screen	Ground elevation (mAHD)	Top of Casing (TOC)	Total depth (m)	production zone (m)
Bridge site									
Nest 1	B1a	654120	6020094	28/03/07	Sand	46.904	47.491	6.5	3.5 - 6.5
	B1b	654120	6020093	28/03/07	Sand	46.938	47.534	10	7 - 10
Nest 2	B2a	654088	6020080	20/03/07	Sand	46.536	48.490	3.2	1.2 - 3.2
	B2b	654089	6020079	19/03/07	Clay	46.530	48.525	6.5	4.5 - 6.5
Nest 3	B3a	654053	6020136	18/03/07	Sandstone	48.231	48.109	35	32 - 35
	B3b	654052	6020132	13/03/07	Sandstone	48.003	47.910	21.5	18.5 - 21.5
	B3c	654053	6020134	28/03/07	Sand	48.174	48.050	9.6	6.6 - 9.6
	B3d	654051	6020134	28/03/07	Sand	48.066	47.851	6.4	3.4 - 6.4
River level monitoring gauge	BS	654089	6020079	n/a	n/a	46.536	48.537	n/a	n/a
Catchment headwaters									
Transect 1	UTa	669755	6036955	02/12/09	Sand	283	283.77	1.23	0.63-1.23
	UTb	669822	6036910	02/12/09	Sand	280.5	281.26	1.86	1.06-1.86
	UTc	669872	6036882	02/12/09	Sand	277.6	278.46	1.13	0.53-1.13
Transect 2	UT1	669714	6036759	02/12/09	Sand	281	281.77	1.625	0.83-1.63
	UT2	669729	6036746	02/12/09	Sand	279.05	279.83	1.82	1.02-1.82
	UT3	669756	6036730	02/12/09	Sand	277.6	278.17	1.22	0.62-1.22

3.2.2 SAMPLING AND ANALYTICAL TECHNIQUES

Surface water and groundwater in the RRC was sampled between August 2007 and December 2009. Surface water sampling was conducted during five sampling rounds at five study sites (East Melrose, Platypus Pools, the Bridge site, gauge station and Snake Lagoon): round one (24–28 August 2007), round two (9–11 October 2007), round three (3–5 December 2007), round four (11–12 March 2008) and round five (3–8 December 2009). In addition, surface water at 45 locations across the catchment was sampled in round five during base flow conditions. Groundwater sampling was conducted at the study sites: East Melrose, Platypus Pools, the Bridge site and the catchment headwaters. This paper shows data of the groundwater sampled in round one and round five and surface water sampled in round five during base flow conditions. The data for other sample rounds is presented in Banks (2010).

A YSI® multi-parameter meter was used to measure pH, specific electrical conductance (SEC), dissolved oxygen (DO), redox potential (ORP) and temperature in the river and also during purging of the piezometers using a flow-through cell. The alkalinity (as HCO_3^-) was measured in the field using a HACH titration kit. Prior to sampling the piezometers, the static water level was measured from top of casing (TOC) using an electric water level indicator. Samples were collected after purging the piezometers and once the physical parameters had stabilised (usually within 20–40 minutes), indicating that the sample was representative of the section of the aquifer sampled. Only the piezometers that were wet could be sampled and these are shown in Table 3.1.

Major element analyses were conducted on the surface water and groundwater samples that were filtered through a 0.45 µm membrane filter in the field. Major cation and trace element samples were acidified with nitric acid (1% v/v HNO_3) and analysed by a Spectro CIROS Radial Inductively Coupled Plasma Optical Emission Spectrometer at CSIRO Land and Water Analytical Services, Adelaide, South Australia. Major anions were analysed using a Dionex ICS-2500 Ion Chromatograph. All ion balances were typically better than ±3 %.

All isotope compositions were measured by isotope ratio mass spectrometry using a Europa Geo 20-20 at the CSIRO Land and Water Isotope Analysis Service in Adelaide, South Australia. δ^2H and $\delta^{18}O$ were analysed by H_2O reduction to H_2 (for δ^2H) by hot Uranium (Dighton et al., 1997) and CO_2 equilibrium for $\delta^{18}O$ (Socki et al., 1992). The results are reported as a deviation from Vienna Standard Mean Ocean Water (vs. VSMOW) in per mil

(‰) difference using delta (δ) notation. The analytical precision for $\delta^{18}O$ and $\delta^{2}H$ is ±0.15 ‰ and ±1.5 ‰, respectively.

Groundwater and surface water samples were collected for analysis of ^{222}Rn activity. ^{222}Rn is a radioactive, inert gas that is generated from the decay of uranium and thorium series isotopes in the aquifer and has a half-life of approximately 3.82 days. Groundwater samples for ^{222}Rn analysis were collected directly from the pump outlet of the purged well using a syringe. A sample of 14 mL was transferred to a pre-weighed 22 mL Teflon-coated PTFE vial with 6 mL Packard NEN mineral oil scintillant, gently shaken for 30 seconds, sealed and the date and time recorded. Surface water samples for the measurement of ^{222}Rn activity were collected using a rapid field extraction method developed by Leaney and Herczeg (2006). Surface water samples were collected in 1.25 L polyethylene terephthalate bottles. Using a syringe, 50 mL of sample was removed from the bottle and then 20 mL of mineral oil scintillant was added from a pre-weighed scintillation vile. The bottle was shaken for four minutes so that the radon equilibrates between the water–air-scintillant phases. The bottle was left to stand for one minute, during which time the scintillant settles to the top of the water. The scintillant was returned to the vial using a glass nozzle, sealed and the date and time recorded. The samples were submitted to the Adelaide Isotope Laboratory within 3 days of sample collection and counted by liquid scintillation on a LKB Wallac Quantulus counter (Herczeg *et al.*, 1994).

Groundwater and surface water samples were collected for the analysis of the radiogenic isotopes of strontium. Strontium isotope ($^{87}Sr/^{86}Sr$) ratios were analysed at the University of Adelaide using a Finnegan Mat 262 thermal ionisation mass spectrometer (TIMS). Strontium was extracted from filtered water samples by evaporating water to leave a solid precipitate, which was then re-dissolved in hydrochloric acid and filtered through columns of Biorad cation exchange resin to isolate $SrCl_2$.

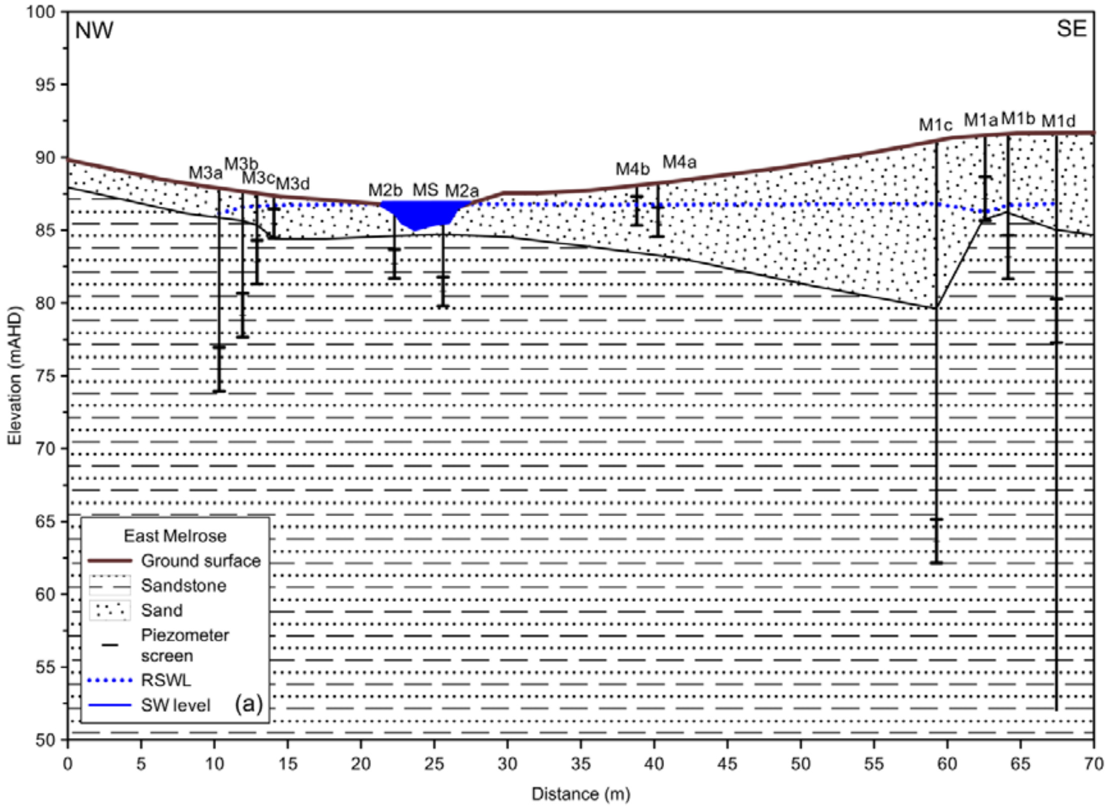
Local rainfall at the Rocky River airstrip and the East Melrose site were collected using a 200 mm wide funnel inserted into a 1.25L bottle held one metre above ground in 100 mm PVC stormwater pipe. Approximately 10 mm of paraffin oil was added to each new collection bottle prior to deployment to prevent isotopic fractionation from evaporation of the rainfall sample, which was collected on a monthly basis.

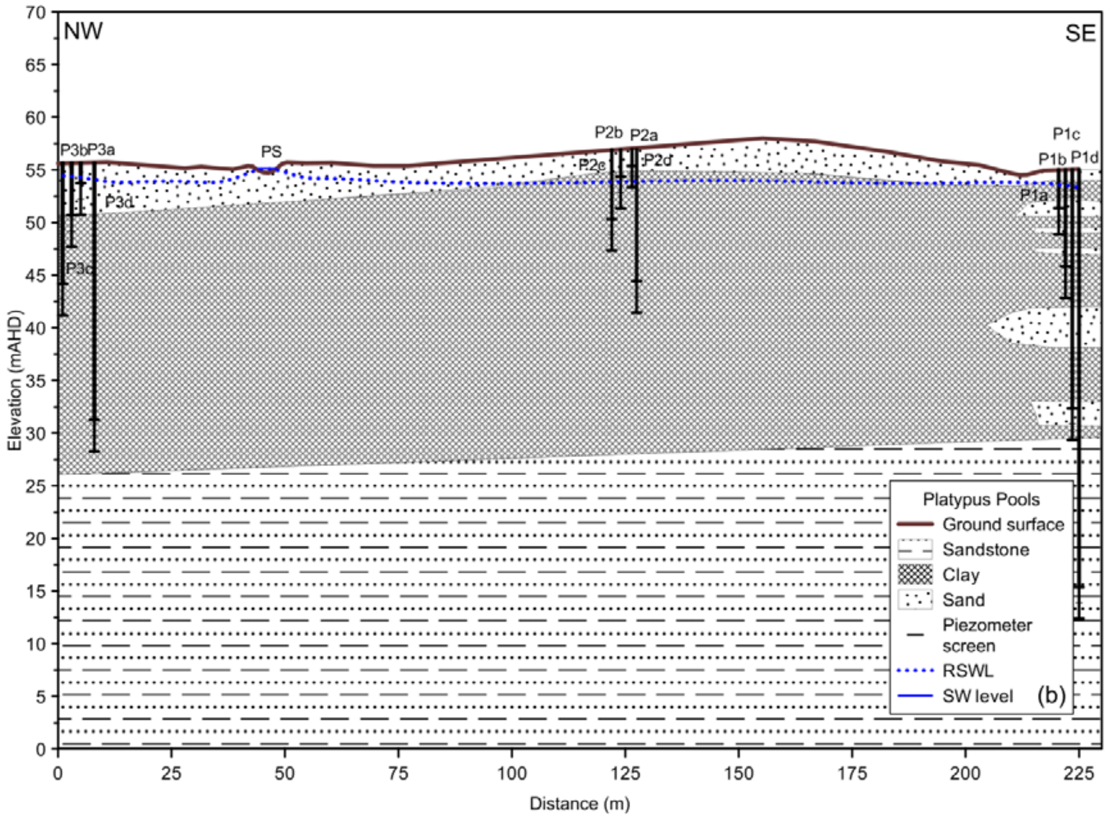
Manual river-flow gauging was conducted using a YSI® flow tracker to quantify river flow and thus determine the gaining and losing reaches of Rocky River at the five study sites and at the 45 locations across the catchment in December 2009.

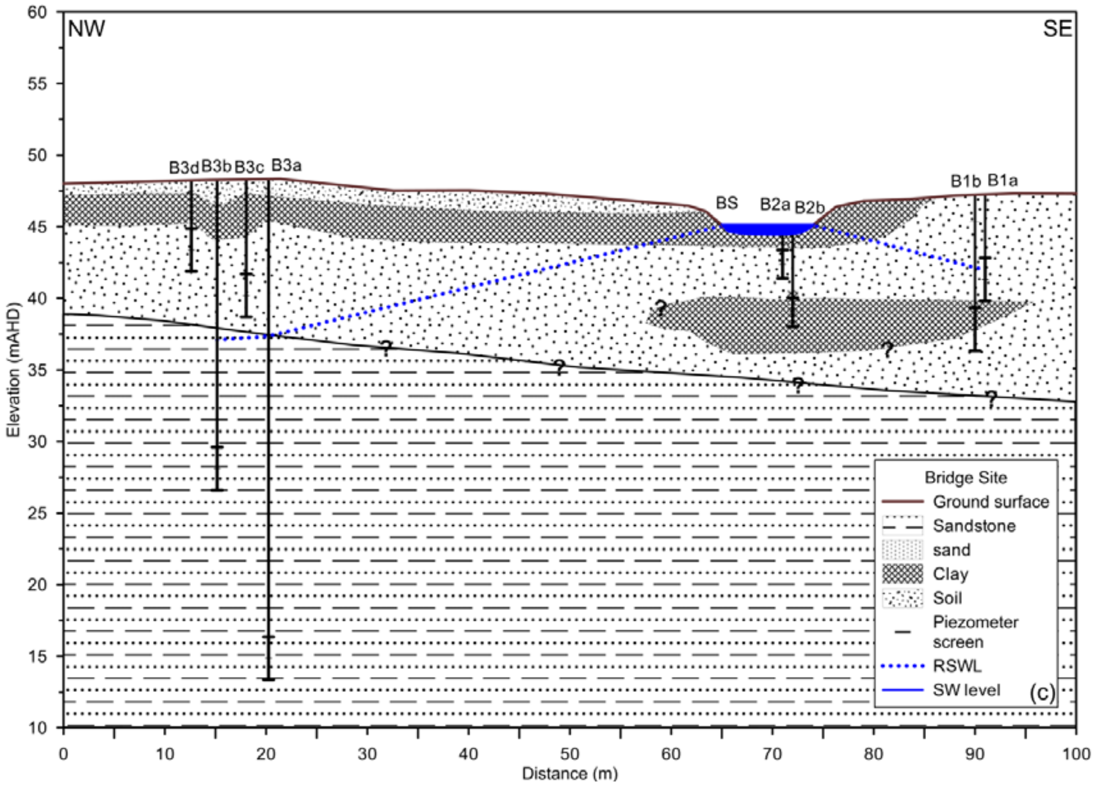
3.3 RESULTS AND DISCUSSION

3.3.1 HYDROGEOLOGICAL CHARACTERISATION

Drilling investigations for the piezometers at East Melrose, Platypus Pools and the Bridge study sites as well as hand augering in the catchment headwaters highlighted several distinct hydrogeological boundaries which could influence the state of connection between Rocky River and the aquifer systems beneath (Figure 3.2). The data suggested that there are three main groundwater flow zones; a localised shallow regolith aquifer, a deeper regional fractured rock aquifer and a shallow perched Quaternary aquifer located in catchment headwaters which will be referred to from this point forward. The regolith material at the study sites varied in thickness being absent at East Melrose and up to 25 m thick at the Platypus Pools. A number of thick (1–3 m) clay layers with high organic content were encountered at the Platypus Pools which were not observed elsewhere. At the Bridge site, unsaturated conditions in the regolith beneath the river bottom was observed, indicating that the river at this site is losing, forming a transitional or disconnected system (Brunner et al., 2011). In the catchment headwaters there were substantial Quaternary deposits of fine grained sand up to 3 m thick overlying an impermeable clay layer, which formed a shallow perched aquifer system. The existence of a low conductance clay layer would restrict vertical movement of groundwater to the regolith and fractured rock aquifer below and result in greater lateral movement of groundwater towards the river. What was also unique about the catchment headwaters were the riparian swamp systems (e.g. chains of ponds and hanging swamps) within the valley bottom of the river tributaries which appeared to have significant water storage capacity and reduced flow velocities in the river. Significant weathering and formation of clay-rich regolith at the top of the Kanmantoo Group bedrock is also likely to restrict infiltration to the deeper fractured rock aquifer and may also result in perched aquifer conditions in the overlaying regolith material. Bedrock exposure at the ground surface and in the river bed was observed mostly in the upper part of the catchment above East Melrose where the steepness in topography was much greater.


NW
SE
M3a
M3b
M3c
M3d
M2b
MS
M2a
M4b
M4a
M1c
M1a
M1b
M1d
Elevation (mAHD)
Distance (m)
East Melrose
Ground surface
Sandstone
Sand
Piezometer screen
RSWL
SW level
(a)


NW
SE
P3b
P3a
P3d
P3c
PS
P2b
P2a
P2e
P2d
P1c
P1b
P1d
P1a
Elevation (mAHD)
Distance (m)
Platypus Pools
Ground surface
Sandstone
Clay
Sand
Piezometer screen
RSWL
SW level
(b)



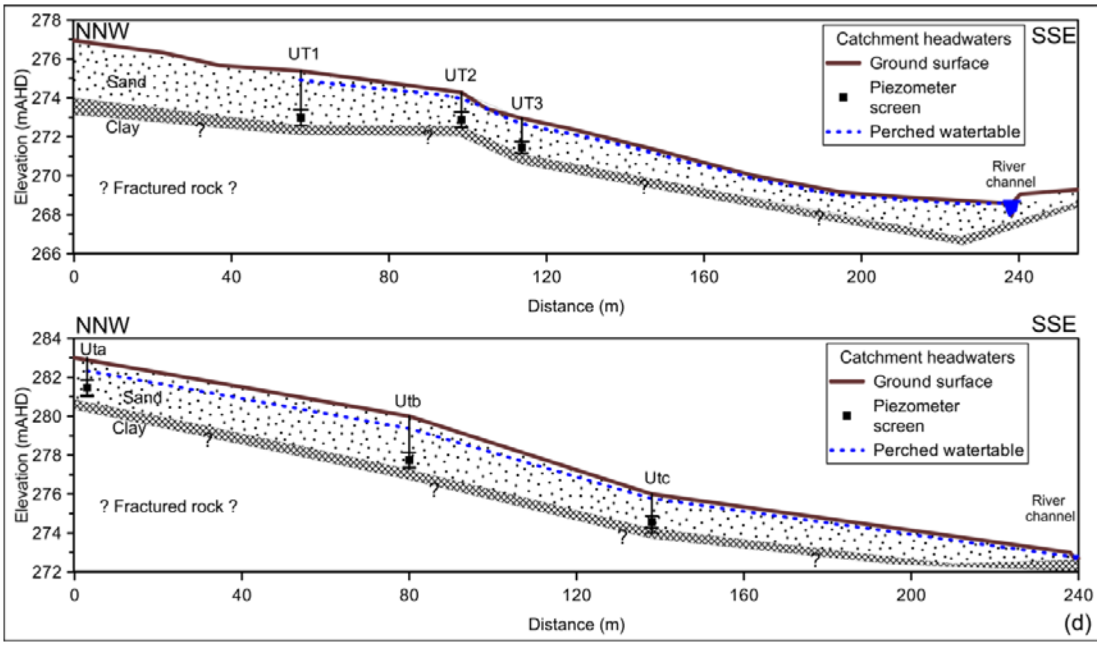


Figure 3.2 Hydrogeological cross sections at the study sites East Melrose (a), Platypus Pools (b), the Bridge site (c), and catchment headwaters (d). The view-point of the cross sections is looking up river.

3.3.2 VARIATIONS IN HYDRAULIC HEAD AND RIVER ELEVATIONS

Rainfall, groundwater and surface water level time series data collected from August 2007 until April 2010 at East Melrose, Platypus Pools and the Bridge study sites are shown in Figure 3.3, Figure 3.4 and Figure 3.5. Over the period of the study, there were higher groundwater levels in winter (July–September) compared to those in summer (December–March). The annual variation in hydraulic head in the aquifers at East Melrose ranged from up to 1.4 m, and less than 0.9 m of annual variation in hydraulic head was observed at M1b between September 2007 and September 2008 (Figure 3.3a). While the variation in hydraulic head at the Platypus Pools was up to 2.6 m and an annual variation of 2.6 m in hydraulic head was observed at P2c between July 2009 and April 2010 (Figure 3.4b). The variation in hydraulic head at the Bridge site was 0.3–1.7 m (Figure 3.5). The annual variation in the river ranged from 0.53 to 0.86 m at these three sites. The hydraulic time series data showed that the river standing water level followed a very similar trend to the groundwater levels in both the bedrock and regolith aquifers which suggests that there is a hydraulic connection between the river and the groundwater system below the river. Base flow conditions in Rocky River typically occurred from the beginning of November through to May and there were very few occasions when river flow ceased completely at East Melrose.

The transverse hydraulic flow assessments also indicated that Rocky River is a dominantly losing system at East Melrose (Figure 3.3b), Platypus Pools (Figure 3.4c) and the Bridge (Figure 3.5a) study sites, with the river level higher than the watertable beneath and directly adjacent to the river over the entire study period. As a result, surface water potentially recharges the unconfined aquifer systems beneath the river and the flux rate is controlled by the hydraulic conductivity of the river bed sediments and the hydraulic gradient between the river and the aquifer. There is no significant hydraulic connection between the regional fractured rock groundwater system and the river i.e. there is no significant groundwater contribution from the fractured rock aquifer to the river (M3a in Figure 3.3b, M2a and M2b in Figure 3.3c, and B3a and B3b in Figure 3.5a). There were two occasions when there was a hydraulic gradient reversal from losing to gaining at the Platypus Pools. These gradient reversals occurred during late winter through to spring in 2008 and 2009 when the river flooded and inundated the low lying riverbank and resulted in recharge to the unconfined regolith aquifer (P2a, P2b and P2c in Figure 3.4b). The higher hydraulic heads in some piezometers compared to the stage height of the river at East

Melrose is not a gradient reversal as these piezometers are located furthest and upslope from the river (e.g. M1b, M1c and M1d in Figure 3.3a).

Observations of the hydraulic data from each of the piezometers at the three study sites showed that there was lateral groundwater movement in the shallow regolith and deeper fractured rock aquifers from the more elevated areas towards the river at the valley bottom. The vertical hydraulic gradient within and between the piezometer nests was most often in a downward direction which indicated groundwater recharge and that river water was recharging the unconfined aquifers (i.e. regolith and fractured rock) below the river bed (Figure 3.3, Figure 3.4 and Figure 3.5). In some of the piezometers in the fractured rock (i.e. B3a and B3b) the fluctuation in the piezometric surface was relatively small suggesting that the fractured rock aquifer system at this location may be influenced by different recharge and discharge processes compared to the other sites where there was a greater variation in the hydraulic head within the fractured rock system. In order to sustain semi-perennial flow down the river, gaining conditions were confirmed upstream of the East Melrose site during the run of river assessment.

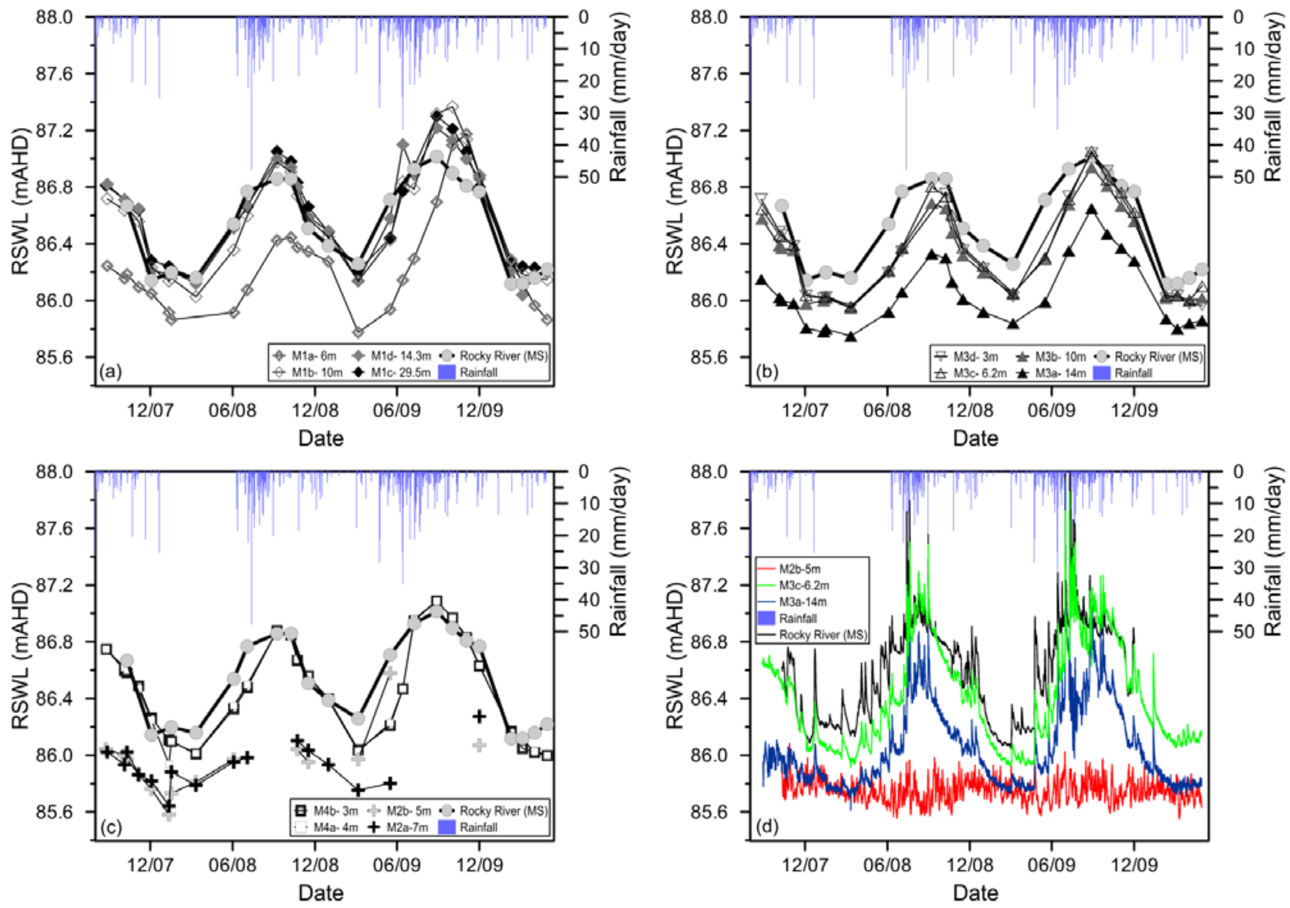


Figure 3.3 Rainfall and selected continuous and manual water level data from piezometers and Rocky River at East Melrose- nest M1 (a), nest M2 and M4 (b), nest M3 (c) and logger data (d) from August 2007 until April 2010.

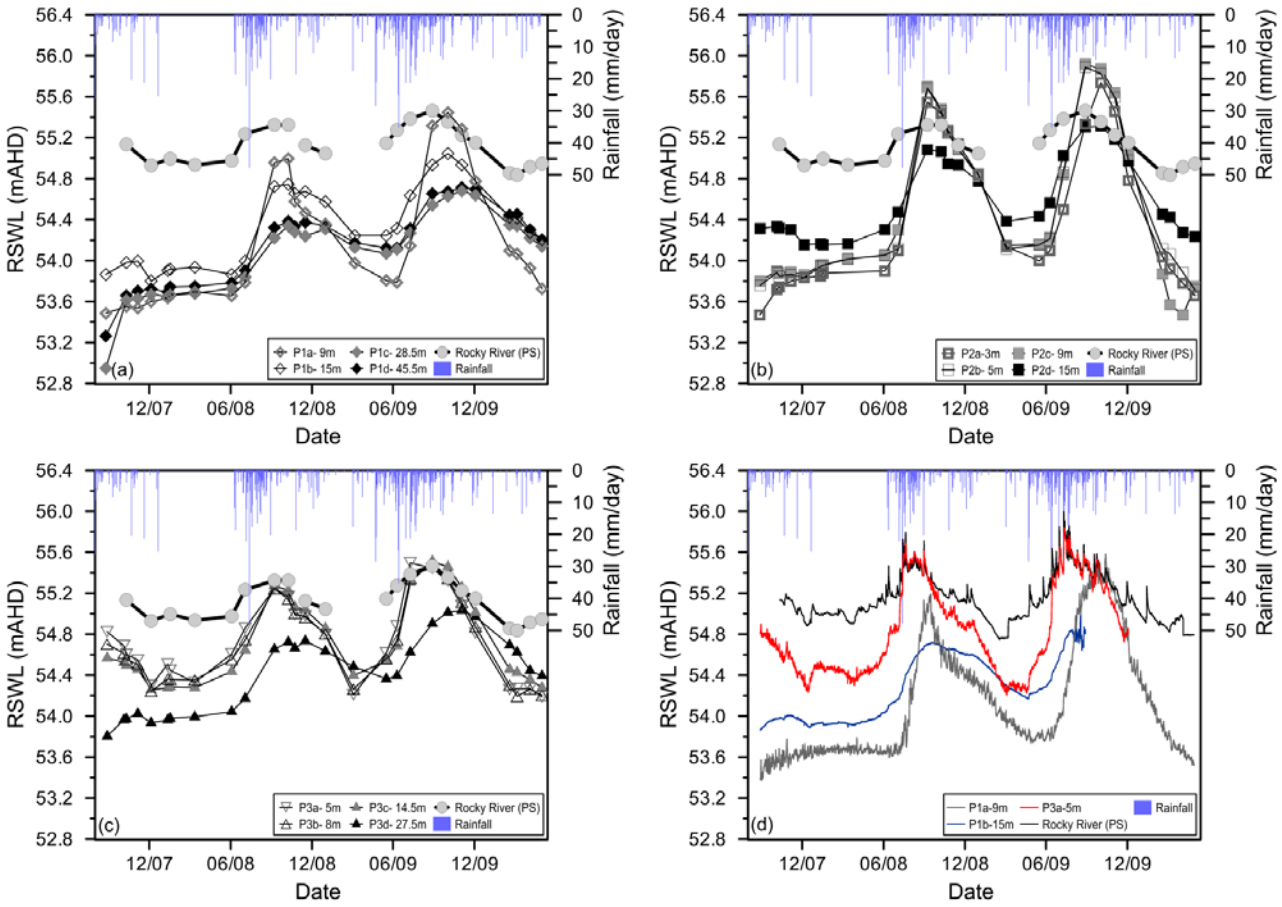


Figure 3.4 Rainfall and selected continuous and manual water level data from piezometers and Rocky River at Platypus Pools- nest P1 (a), nest P2 (b), nest P3 (c) and logger data (d) from August 2007 until April 2010.

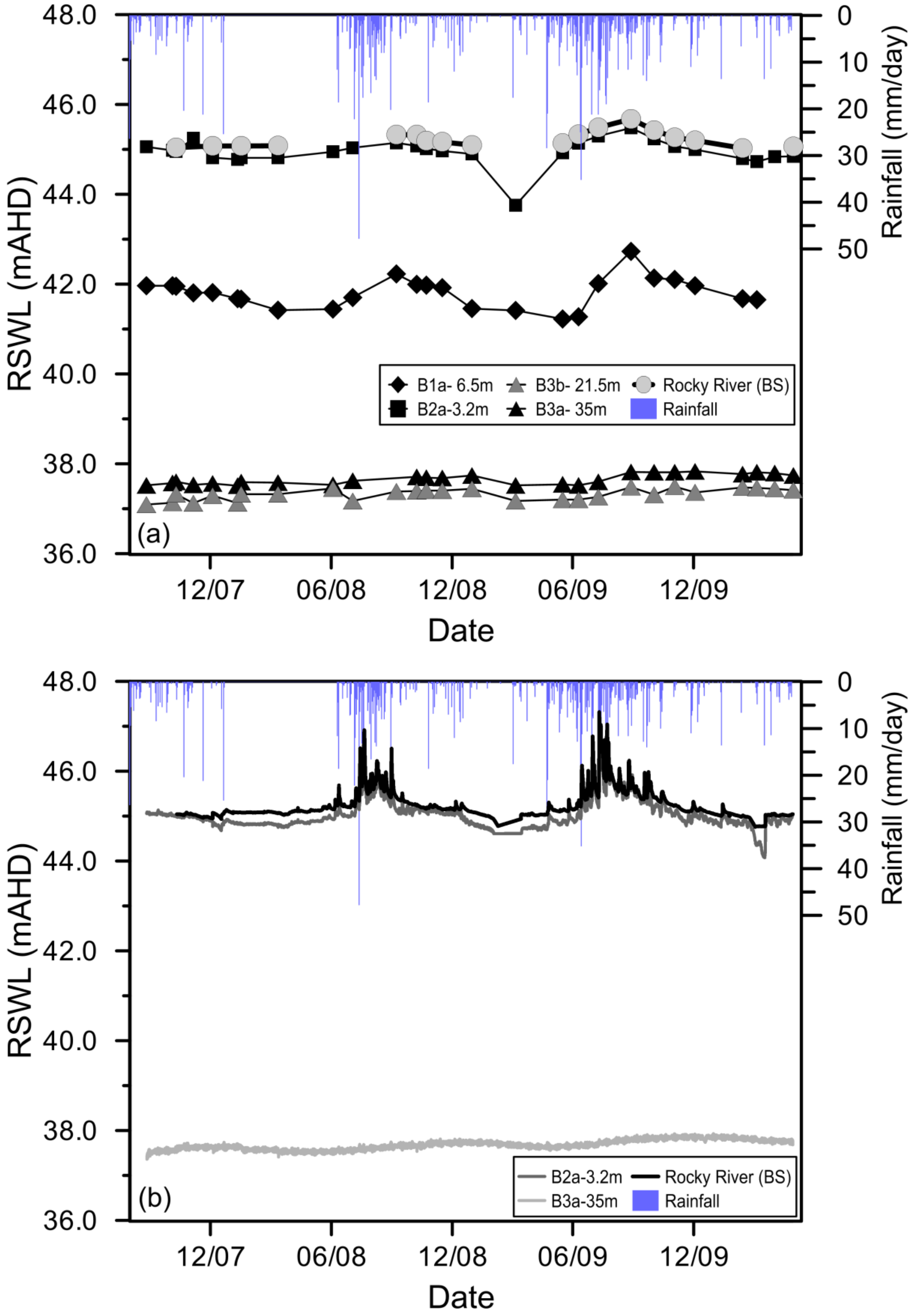


Figure 3.5 Rainfall and selected continuous and manual water level data from piezometers and Rocky River at the Bridge site- nest B1, B2 and B3 (a) and logger data (b) from August 2007 until April 2010.

3.3.3 Run of River Flow Assessment

The interaction between Rocky River and the three defined groundwater flow zones (a localised shallow regolith aquifer, a deeper regional fractured rock aquifer and a shallow perched Quaternary sand aquifer) can be attributed to changes to the river source and loss terms as a result of variation in groundwater discharge, variation in leakage through the riverbed to the aquifer below, contribution from surface runoff or surface water loss to evaporation. Flow gauging at locations along the length of Rocky River and its tributaries during round five sampling identified which reaches of the river were losing and gaining during base flow conditions (Figure 3.6a). The flow volume gradually increased from the catchment headwaters downstream as minor tributaries joined the main watercourse. Flow significantly increased at the 31 km mark as a result of the confluence of a major tributary in the north western corner of the catchment with the main watercourse. Flow peaked at the 26.5 km mark (RR12- 7560 m^3/day) before decreasing gradually to 11.6 km (RR18- 7271 m^3/day- just upstream of the Platypus Pools) where flow increased again with a minor tributary joining the main channel and also where there is a change in the geology from the Kanmantoo Group to the Quaternary Bridgewater formation. Along the last river reach flow decreased again. The temporal variability in river flow (according to flow gauging measured at different times of the year) and the state of connection during the other sample rounds also confirmed gaining conditions up stream of East Melrose and losing type conditions at East Melrose, Platypus Pools and the Bridge site and is discussed in further detail in Banks (2010).

In summary, longitudinal river flow measurements during the run of river assessment and continuous water level monitoring at the three study sites over the study period showed that Rocky River is a gaining type system in the catchment headwaters above East Melrose, a losing system in the middle of the catchment and a gaining system again downstream of the Bridge site where there is a change in the catchment's geology. The data showed that there is a seasonal response of the piezometric surface to groundwater recharge in the shallow regolith and deeper fractured rock aquifers and that groundwater discharge from the regional fractured rock aquifer to Rocky River is unlikely. Therefore, the decrease in river flow where it was observed to be losing can be attributed to a decrease in groundwater discharge from the shallow regolith aquifer system, increased leakage through the riverbed to the deeper regional aquifer below, or surface water loss to evaporation. Where the river losing rate is low (i.e. between 26.5 and 11.6 km) could be attributed to the unsaturated zone between the river and the underlying aquifer or the low hydraulic conductivity of the streambed. The water loss via infiltration and evaporation could also be balanced by inflow from the minor tributaries.

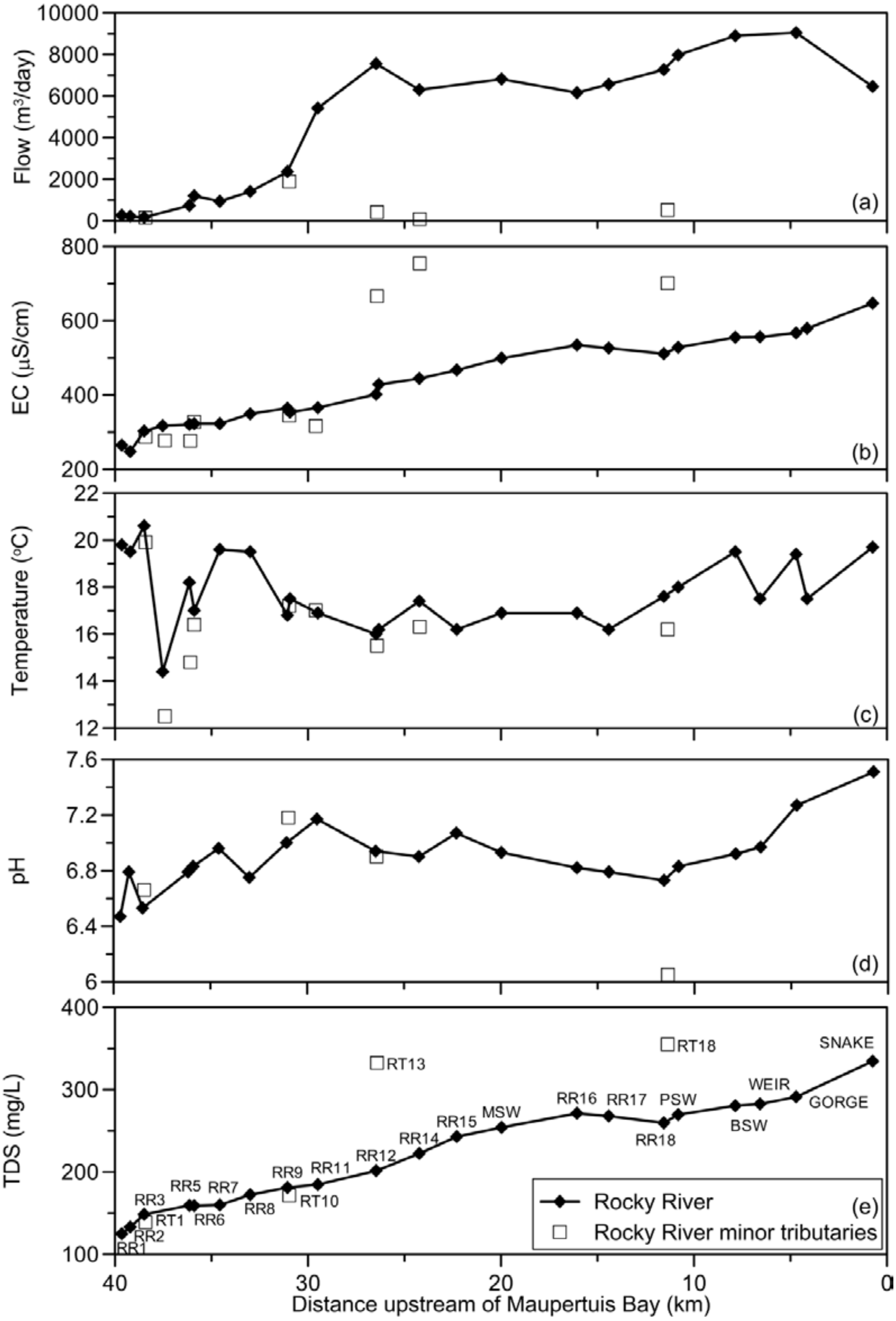


Figure 3.6 Run of river showing river flow (a), EC (b), temperature (c), pH (d) and TDS (e) in Rocky River, December 2009. Distances are measured upstream of the Rocky River outlet at Maupertuis Bay.

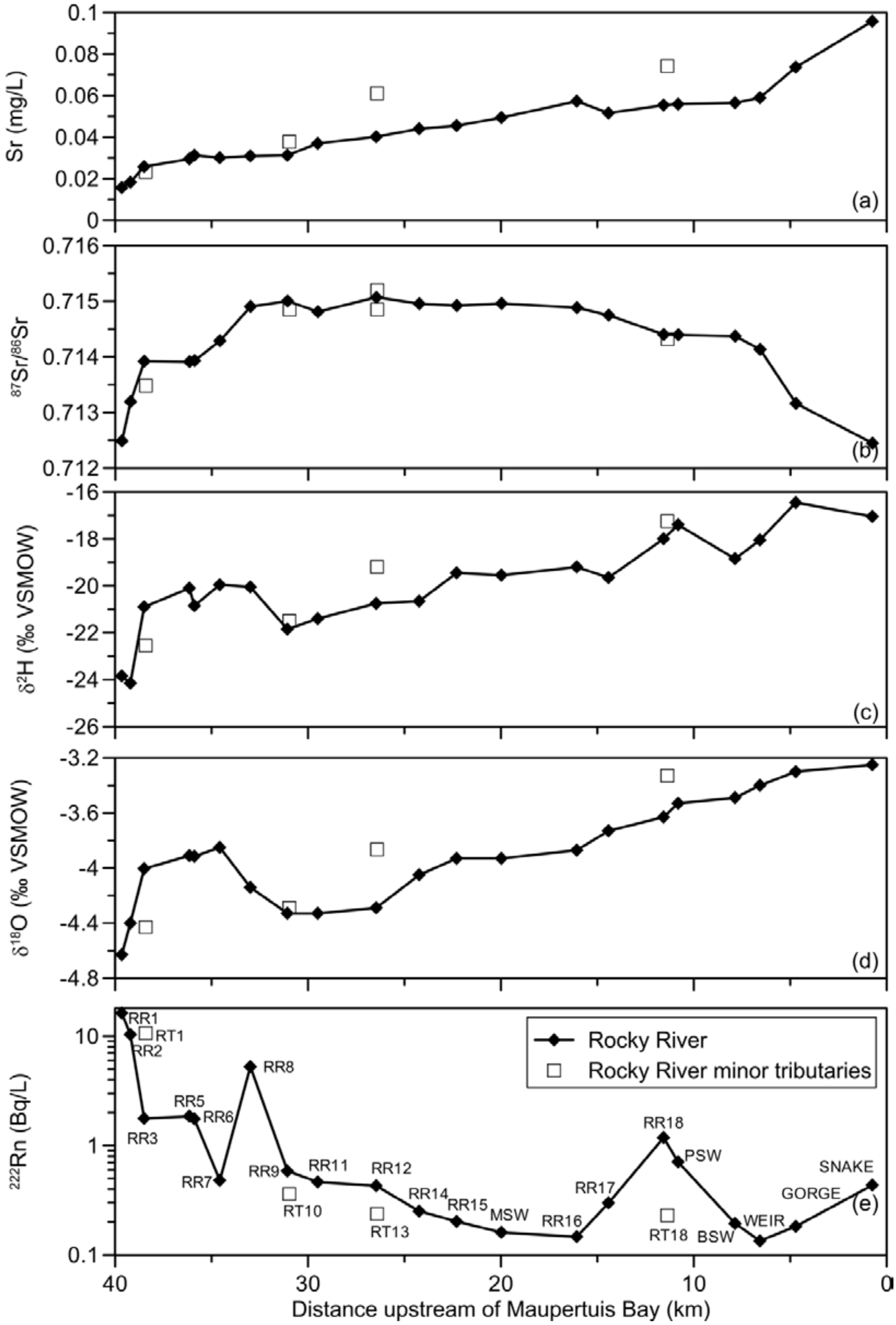


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Figure 3.7 Run of river showing strontium concentration (a), strontium isotope ratios (b), Deuterium (c), Oxygen-18 (d) and ^{222}Rn (e) in Rocky River, December 2009. Distances are measured upstream of the Rocky River outlet at Maupertuis Bay.

3.3.4 ELECTRICAL CONDUCTIVITY AND MAJOR IONS

The chemical composition of the shallow perched Quaternary, shallow regolith and deeper fractured rock groundwater from East Melrose, Platypus Pools, the Bridge site and the catchment headwaters is shown in Table 3.2. Only data from round one and round five are shown here; data from the other rounds and additional information about the hydrochemistry can be found in Banks (2010). The salinity of the groundwater samples from the fractured rock aquifer at the study sites ranged from 894 to 10401 μS/cm (according to bedrock aquifer sampling data at M2a and P1d) and the samples from the regolith zone ranged from 984 to 18910 μS/cm (according to regolith aquifer sampling at B1a and M1a) (Table 3.2 and Figure 3.8). In comparison, the salinity of the groundwater samples from the shallow perched Quaternary sand aquifer in the catchment headwaters ranged from 214 to 342 μS/cm (UT1-UT2). This groundwater was more similar to the salinity of Rocky River which ranged from 265 μS/cm in the catchment headwaters to 647 μS/cm at Snake Lagoon at the bottom of the catchment (Table 3.3 and Figure 3.6).

The run of river data shows clearly the steady increase in EC and the major ions from the top of the catchment down the river to Snake Lagoon (Figure 3.6 and Figure 3.8). The increase in salinity is a result of evapotranspiration and/or discharge of more saline groundwater to the river. As discussed earlier, continuous water level monitoring at the study sites (East Melrose, Platypus Pools and the Bridge) showed that the river was losing and therefore, the increase in salinity is most likely a result of evapotranspiration at these locations as there were no major tributaries along this section of the river to contribute to the observed salinity increase. Gaining conditions in the upper region of the catchment, above East Melrose, indicated a subsurface fresh water source to the river that was either from shallow sedimentary and/or fractured rock aquifer systems. The low salinity of the river water and water level data showed that the dominant subsurface source to the river in the catchment headwaters is from the shallow perched Quaternary sand aquifer and that there is minimal influence from the deeper saline fractured rock aquifer. Thick clay sediments that induce perched aquifer conditions in the catchment headwaters limit the possibility of the vertical movement of groundwater to the aquifers below. The proximity of the perched aquifer to the ground's surface and its shallow depth and low salinity would imply that there is a short groundwater residence time in this aquifer.

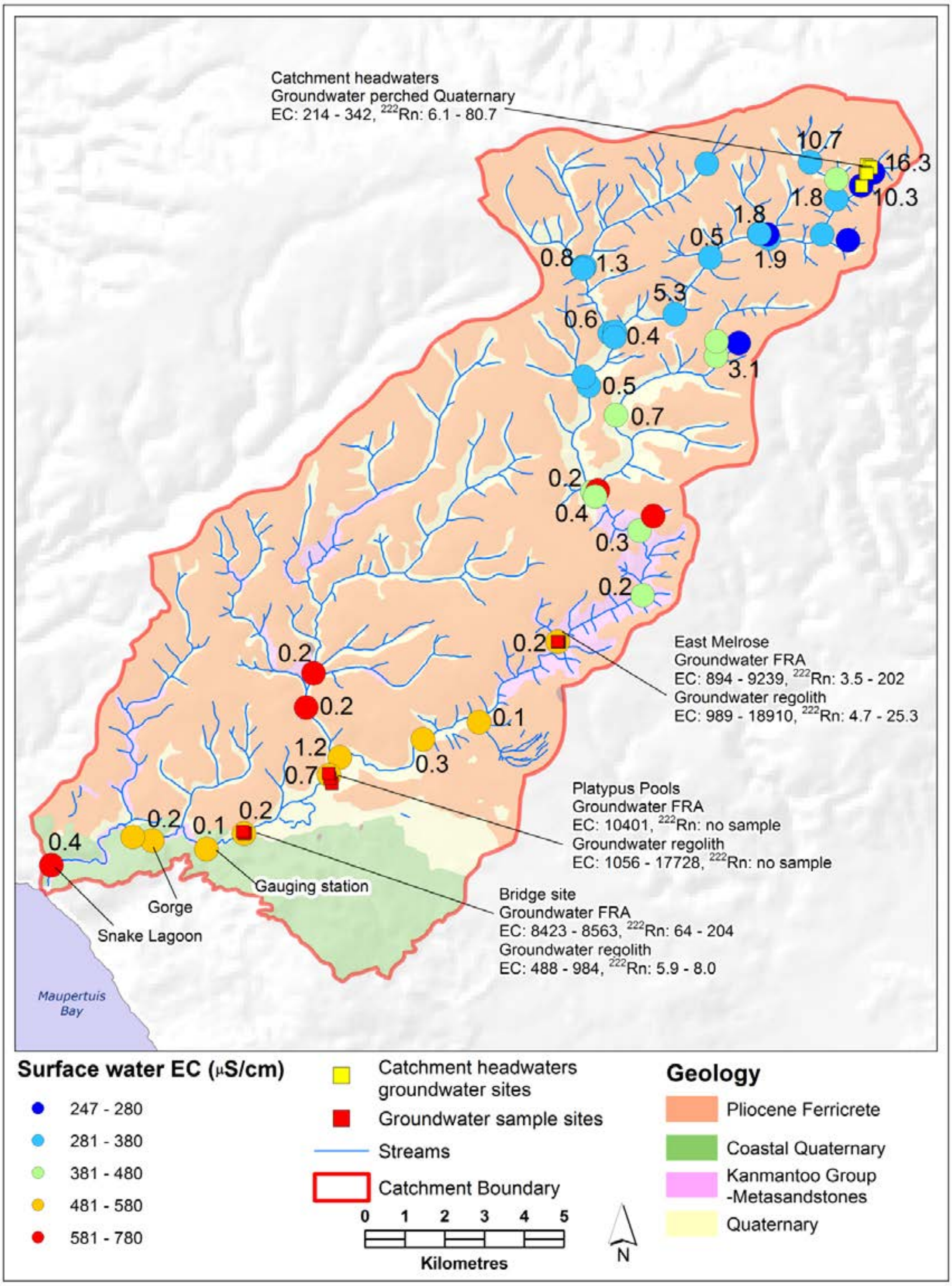

Figure 3.8 Surface water and groundwater electrical conductivity (EC- microS/cm) and ^{222}Rn (Bq/L) in the Rocky River Catchment. Also shown is the surface geology across the catchment.

3.3.5 STABLE ISOTOPE RATIOS

The stable isotope ratio data were used to infer the likely sources of subsurface water to Rocky River (Table 3.3). The δ^2H and $\delta^{18}O$ values of Rocky River and the groundwater samples from the shallow regolith and deeper fractured rock aquifers are plotted in Figure 3.9, relative to the local meteoric water line (MWL) for RRC from monthly rainfall samples collected during the study period Table 3.4) which is slightly different to the local MWL for Adelaide rainfall (IAEA, 2001). The deeper regional groundwater from the fractured rock aquifer falls slightly above the MWL and the majority of samples plot closely to the weighted mean precipitation for the Rocky River catchment (δ^2H = –23.5 ‰ and $\delta^{18}O$= –4.7 ‰). The groundwater sample from the fractured rock that had a more enriched isotopic composition is from the piezometer at the Platypus Pools site (45.5 m below ground) and is likely to represent water from different climatic conditions. Results of age dating showed that this sample had a carbon-14 composition of 21 % modern carbon (Banks, 2010). Groundwater from the regolith zone at East Melrose, Platypus Pools and the Bridge study sites had similar isotopic compositions to the groundwater samples from the fractured rock aquifer. The three more enriched samples from the regolith zone are from shallow piezometers directly adjacent to the large open surface water feature at the Platypus Pools and adjacent to the river at the Bridge site. These samples represent a mixture of groundwater and surface water because the hydraulic data showed that there was a gradient reversal between the river and aquifer when the area around these piezometers was flooded during the higher river flows in winter. The groundwater samples from the shallow perched aquifer in the catchment headwaters were similar to most of the samples from the regolith and fractured rock at the other study sites, however, some had isotope compositions that were more depleted and plotted below the weighted mean precipitation, indicative of winter rainfall events (Figure 3.9).

The surface water samples collected during the run of river assessment had δ^2H and $\delta^{18}O$ values that were higher than the weighted mean precipitation for RRC and plot along an evaporation line with a slope of 4.7 (Figure 3.9). This evaporative signature of the surface water is observed in Figure 3.7 which shows the δ^2H and $\delta^{18}O$ values increasing from the catchment headwaters down the river. A plot of δ^2H versus chloride concentration (Figure 3.10) shows the surface water samples plotting along an evaporation trend evolving from the samples of groundwater from the shallow Quaternary aquifer in the catchment headwaters. In comparison, most of the groundwater samples from the regolith and fractured rock are much more saline. The few groundwater samples from the regolith and fractured rock that plot close to the surface water samples represent a mixture of groundwater and surface water and are from piezometers that are located in close proximity to the river.

The variation in the stable isotopic compositions in the catchment headwaters (first 10 km) is most likely caused by the contribution from numerous tributaries which drain a range of surface water landscapes from hanging swamps to a chain of ponds. A simple steady–state mixing bucket model of Rocky River showed that an evaporation rate of 10–15 mm/day could result in the observed increase in salinity from the catchment headwaters to the catchment bottom. To satisfy this calculation, several assumptions had to be made which included an average river width of two metres and initial salinity of the river at the catchment headwaters of 265 μS/cm. The decrease in river flow along some reaches of the river cannot be attributed exclusively to surface evaporation and may be due to a loss of surface water to groundwater.

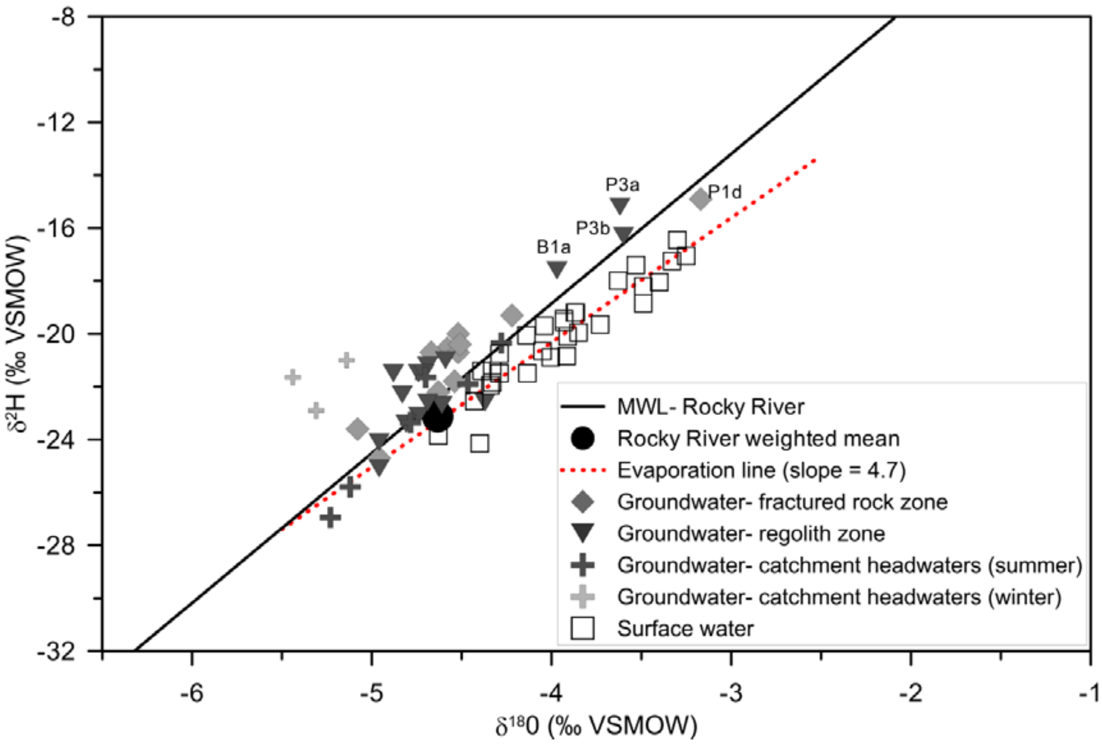

Figure 3.9 δ^2H versus δ^{18}O values of surface water from Rocky River and groundwater in the RRC. The MWL for Adelaide is $\delta^2H = 7.7\delta^{18}O + 9.6$ and the weighted mean precipitation, $\delta^2H = -26$ ‰ and $\delta^{18}O = -4.7$ ‰. The MWL for RRC is $\delta^2H = 5.7\delta^{18}O + 4.2$ and the weighted mean precipitation, $\delta^2H = -23$ ‰ and $\delta^{18}O = -4.7$ ‰.

Table 3.2 Physical chemistry, ion concentrations, stable isotopes of water, ^{222}Rn and $^{87}Sr/^{86}Sr$ ratios of groundwater from Rocky River Catchment. n/m indicates not measured.

Site ID	Sampling Date	Temp (^{o}C)	pH	Redox (ORP)	DO (mg/L)	SEC (μS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	$^{87}Sr/^{86}Sr$ ratio
M1a	27/08/07	16.0	7.0	-2	1.3	18910	343	5870	16.8	498	147	62	388	3220	10545	-5.0	-24.1	25.3	1.2	0.717685
M1b	27/08/07	16.7	6.7	35	0.9	9239	239	2700	7.6	266	74	42	139	1620	5087	-5.1	-23.6	27.5	0.6	0.718163
M1c	24/08/07	16.1	7.5	-131	4.4	7403	197	2280	6.4	119	189	46	127	1200	4164	-4.6	-22.2	3.5	1.2	0.713799
M1d	27/08/07	15.5	6.7	-19	0.4	7746	265	2210	6.3	204	126	42	144	1230	4227	-4.6	-20.6	94.5	0.7	n/m
M4a	24/08/07	12.7	6.6	155	6.7	989	25	239	0.5	52	6.7	7.2	19	151	500	-4.7	-21.5	4.7	0.06	0.716042
M4b	25/08/07	13.8	6.2	146	7.6	1978	27	510	1.2	75	16	8.7	32	304	973	-5.0	-25.1	7.5	0.1	0.715778
M2a	27/08/07	13.6	6.7	-140	0.1	894	61	195	0.5	32	5.9	8.2	8.6	151	462	-4.5	-20.7	202.0	0.05	0.716009
M2b	24/08/07	12.1	6.6	-165	1.0	1157	55	287	0.7	30	6.3	6.8	14	191	591	-4.5	-20.4	94.6	0.06	0.715688
M3a	25/08/07	14.7	6.7	-76	1.2	6189	265	1580	4.4	117	86	30	101	922	3105	-5.0	-24.7	182.0	0.5	0.715332
M3b	27/08/07	15.0	6.8	-119	0.2	6426	208	1780	5.0	120	34	31	84	1110	3372	-4.5	-21.8	87.3	0.3	0.716993
M3c	27/08/07	14.3	6.8	-151	0.3	6450	211	1630	4.4	110	47	32	80	1000	3115	-4.7	-20.7	6.1	0.4	0.714706
M3d	25/08/07	12.6	6.8	-120	2.0	8522	145	2470	6.8	282	40	37	156	1450	4586	-4.6	-21.0	9.4	0.4	0.714965
P1a	25/08/07	15.8	6.6	60	0.3	6591	130	1870	5.0	186	51	34	79	1150	3505	-4.4	-22.6	n/m	0.6	0.710834
P1b	25/08/07	16.4	6.6	4	0.7	17445	261	5630	15.9	494	356	50	427	2610	9845	-4.7	-21.5	n/m	3.2	0.712063
P1c	25/08/07	18.0	6.6	-20	0.2	17728	208	5920	17.1	406	719	56	588	2020	9935	-4.7	-23.1	n/m	3.2	0.715229
P1d	25/08/07	16.9	7.2	-56	0.3	10401	163	3320	10.1	130	308	44	267	1400	5643	-3.2	-14.9	n/m	1.8	0.716024
P2a	28/08/07	dry		dry																

Table 3.2 continued.

Site ID	Sampling Date	Temp (°C)	pH	Redox (ORP)	DO (mg/L)	SEC (µS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	$^{87}Sr/^{86}Sr$ ratio
P2b	28/08/07	15.2	6.4	124	3.9	6403	105	1830	4.9	162	40	32	84	1090	3348	-4.7	-22.6	n/m	0.5	0.710776
P2c	28/08/07	15.7	6.7	94	0.7	6943	141	1970	5.3	212	61	32	98	1180	3699	-4.8	-23.4	n/m	0.7	0.710930
P2d	28/08/07	15.5	6.5	49	0.3	10580	206	3220	8.9	285	152	39	209	1690	5810	-4.9	-21.5	n/m	1.4	0.712487
P3a	28/08/07	14.5	6.7	18	0.6	1056	198	187	0.8	4	29	7.4	26	119	574	-3.6	-15.2	n/m	0.2	0.713855
P3b	28/08/07	14.9	6.7	-15	0.4	1114	216	214	0.8	0.3	25	7.9	24	154	644	-3.6	-16.3	n/m	0.2	0.713920
P3c	28/08/07	14.9	6.7	34	0.4	7484	203	2180	6.2	188	133	34	161	1140	4047	-4.7	-21.2	n/m	1.1	0.713074
P3d	28/08/07	15.7	6.8	-41	0.2	6703	223	1950	5.6	111	221	36	206	778	3533	-4.8	-22.3	n/m	1.4	0.717242
B1a	26/08/07	16.0	5.4	27	0.9	984	19	163	0.5	173	30	4.2	25	106	520	-4.0	-17.6	5.9	0.3	0.713186
B1b	26/08/07	dry		dry																
B2a	26/08/07	12.5	6.4	-152	0.5	488	109	69	0.2	5	16	3.2	8.9	55	268	-4.6	-22.7	8.0	0.1	0.712686
B2b	26/08/07	dry		dry																
B3a	26/08/07	17.3	7.0	-48	0.3	8423	127	2610	7.8	141	364	32	263	915	4459	-4.5	-20.0	64.0	3.8	0.716210
B3b	26/08/07	16.9	7.2	-28	0.4	8563	146	2640	7.9	162	464	32	263	824	4538	-4.2	-19.3	204.0	3.2	0.716682
B3c	26/08/07	dry		dry																
B3d	26/08/07	dry		dry																

Table 3.2 continued.

Site ID	Sampling Date	Temp (^{o}C)	pH	Redox (ORP)	DO (mg/L)	SEC (μS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	$^{87}Sr/^{86}Sr$ ratio
UT1	13/08/08	10.3	6.5		8.4	280	5.4	70	<0.05	10.4	1.5	1.1	6.1	42	137	-5.3	-22.9	15.36	0.028	0.711166
UT2	13/08/08	10.3	6.7		8.9	287	6.3	69	0.05	15.9	1.6	1.2	5.7	46	146	-5.1	-21.0	n/m	0.028	0.712852
UT3	13/08/08	10.9	7.0		8.8	341	9.5	90	0.28	1.8	1.4	1.2	6.7	53	164	-5.4	-21.7	n/m	0.023	0.712304
UTa	03/12/09	18.5	6.1		2.1	301	12.3	76	0.306	12.5	4.3	2.0	4.9	39	150	-4.3	-20.3	74.5	0.028	0.715066
UTb	03/12/09	18.1	5.8		6.5	342	12.8	101	0.362	14.8	4.7	1.2	6.3	53	194	-4.5	-21.9	80.7	0.035	0.711009
UTc	03/12/09	18.0	5.5		6.0	331	8.7	92	0.582	8.3	3.1	0.9	4.1	49	167	-4.8	-23.3	6.1	0.024	0.710822
UT1	03/12/09	17.1	6.0		1.8	214	9.5	58	0.267	10.8	4.0	1.3	3.5	32	120	-5.1	-25.8	29.7	0.020	0.710384
UT2	03/12/09	17.7	5.6		1.5	342	7.8	100	0.313	11.5	2.3	0.7	8.5	47	178	-5.2	-26.9	18.2	0.031	0.710433
UT3	03/12/09	16.8	7.8		5.5	319	44.8	75	<0.5	10.4	10.2	0.9	6.0	44	191	-4.7	-21.6	n/m	0.050	n/m

Table 3.3 Physical chemistry, ion concentrations, stable isotopes of water, ^{222}Rn and ^{87}Sr/^{86}Sr ratios of surface water from Rocky River Catchment. n/m indicates not measured.

Sample sites	Date	Distance upstream of Maupertuis Bay (km)	Temp (°C)	pH	DO (mg/L)	SEC (μS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	δ^{18}O (‰VSMOW)	δ^{2}H (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	^{87}Sr/^{86}Sr
RR1	03/12/09	39.7	19.8	5.9	6.2	265	6.0	65.6	0.29	8.6	0.9	1.1	4.0	38.3	125	-4.6	-23.9	16.30	0.016	0.712489
RR2	03/12/09	39.2	19.5	5.9	7.9	247	7.9	70.1	0.37	7.0	1.7	1.3	4.3	40.0	133	-4.4	-24.2	10.35	0.018	0.713194
RR3	04/12/09	38.5	20.6	6.2	8.2	303	10.1	81.1	0.70	3.6	1.6	1.6	5.1	44.5	148	-4.0	-20.9	1.77	0.026	0.713922
RR4	04/12/09	37.5	14.4	6.4	6.6	317	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RR5	04/12/09	36.1	18.2	6.7	10.1	321	13.3	84.4	0.59	4.2	1.9	1.8	5.8	47.3	159	-3.9	-20.1	1.85	0.030	0.713910
RR6	04/12/09	35.9	17	6.6	8.9	322	12.2	85.6	0.61	4.1	1.9	1.8	5.8	46.9	159	-4.0	-20.9	1.75	0.031	0.713932
RR7	04/12/09	34.6	19.6	6.7	8.4	323	10.1	87.6	0.56	4.0	2.0	1.8	5.8	48.0	160	-3.9	-20.0	0.48	0.030	0.714290
RR8	04/12/09	33.0	19.5	6.4	8.6	349	12.3	93.3	0.53	5.6	2.0	1.7	6.5	50.3	172	-4.1	-20.1	5.27	0.031	0.714904
RR9	05/12/09	31.1	16.8	7.0	9.7	365	12.4	99.8	0.50	6.1	1.9	1.7	6.8	51.7	181	-4.3	-21.9	0.59	0.031	0.715002
RR10	05/12/09	30.9	17.5	7.5	10.0	354	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RR11	05/12/09	29.5	16.9	6.7	8.2	366	15.4	98	0.74	5.2	2.3	1.8	7.4	53.7	185	-4.3	-21.4	0.46	0.037	0.714812
RR12	06/12/09	26.5	16	6.8	9.1	402	16.5	107	0.72	6.1	2.6	1.8	7.8	58.8	201	-4.3	-20.8	0.43	0.040	0.715071
RR13	06/12/09	26.3	16.2	6.8	8.4	428	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RR14	06/12/09	24.2	17.4	6.8	8.3	444	15.9	121	0.77	7.8	2.6	2.0	8.5	63.6	222	-4.1	-20.7	0.25	0.044	0.714954
RR15	08/12/09	22.3	16.2	7.0	8.4	467	16.3	133	0.81	8.5	3.0	2.1	9.8	69.2	243	-3.9	-19.5	0.20	0.046	0.714922
MSW	07/12/09	20.0	16.9	6.8	7.6	499	15.9	139	0.81	10.1	3.1	2.2	10.1	72.6	254	-3.9	-19.6	0.16	0.049	0.714958
RR16	07/12/09	16.1	16.9	6.8	8.4	534	15.5	149	0.88	11.7	3.4	2.3	11.0	77.3	271	-3.9	-19.2	0.15	0.057	0.714886

Table 3.3 continued.

Sample sites	Date	Distance upstream of Maupertuis Bay (km)	Temp (^{o}C)	pH	DO (mg/L)	SEC (μS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	δ^{18}O (‰VSMOW)	δ^{2}H (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	^{87}Sr/^{86}Sr
RR17	07/12/09	14.4	16.2	6.7	8.0	526	15.0	149	0.89	11.0	3.2	2.3	10.7	75.6	268	-3.7	-19.7	0.30	0.052	0.714747
RR18	06/12/09	11.6	17.6	6.6	7.1	511	20.9	141	0.93	4.5	3.3	2.4	10.3	75.9	259	-3.6	-18.0	1.18	0.055	0.714402
PSW	05/12/09	10.8	18	6.5	6.6	528	19.1	152	0.98	4.7	3.4	2.6	10.5	76.0	270	-3.5	-17.4	0.71	0.056	0.714395
BSW	05/12/09	7.9	19.5	6.9	8.2	555	15.9	160	0.98	7.0	3.5	2.6	10.8	79.4	280	-3.5	-18.9	0.19	0.057	0.714371
WEIR	06/12/09	6.6	17.5	7.0	7.3	556	17.4	160	1.02	6.6	4.0	2.6	10.8	79.9	283	-3.4	-18.1	0.13	0.059	0.714134
GORGE	08/12/09	4.7	19.4	7.3	8.6	567	25.0	161	1.02	6.1	6.5	2.6	10.7	78.0	291	-3.3	-16.5	0.18	0.074	0.713162
L-GORGE	08/12/09	4.1	17.5	7.1	7.1	579	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
SNAKE	07/12/09	0.7	19.7	7.5	8.4	647	26.6	180	0.93	12.3	9.6	2.8	12.6	89.5	334	-3.3	-17.1	0.43	0.096	0.712449
RT1	04/12/09	38.4	19.9	6.1	5.5	287	9.3	73.3	0.62	6.2	1.4	1.5	4.6	41.5	139	-4.4	-22.6	10.66	0.023	0.713482
RT2	04/12/09	38.4	14.2	5.4	4.9	412	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT3	04/12/09	37.4	12.5	6.1	6.0	277	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT4	04/12/09	36.1	14.8	6.5	8.7	277	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT5	04/12/09	35.9	16.4	6.6	9.0	328	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT6	07/12/09	31.0	15.5	7.0	9.0	344	26.3	85.5	0.99	2.8	2.9	1.4	7.3	48.1	175	-4.4	-21.4	1.26	0.044	n/m
RT7	07/12/09	31.0	16	6.2	8.2	314	9.9	83.4	0.79	3.8	2.0	1.6	5.9	44.7	153	-4.1	-21.5	0.81	0.034	n/m
RT8	04/12/09	31.0	19.7	6.5	8.4	350	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT9	07/12/09	31.0	15.7	6.7	8.7	331	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m

Table 3.3 continued.

Sample sites	Date	Distance upstream of Maupertuis Bay (km)	Temp (^{o}C)	pH	DO (mg/L)	SEC (μS/cm)	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	^{222}Rn (Bq/L)	Sr (mg/L)	$^{87}Sr/^{86}Sr$
RT10	05/12/09	31.0	17.2	7.4	9.1	344	17.1	89.7	0.80	4.1	2.4	1.6	6.8	49.0	172	-4.3	-21.5	0.36	0.038	0.714851
RT11	05/12/09	29.6	17	6.2	8.2	316	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT12	05/12/09	26.4	16.9	6.9	9.2	460	20.8	123	0.59	5.4	2.6	2.0	8.4	67.7	231	-4.0	-19.7	0.68	0.046	0.715201
RT13	06/12/09	26.4	15.5	6.7	7.6	666	16.9	183	0.81	16.7	3.6	2.7	12.1	96.5	332	-3.9	-19.2	0.24	0.061	0.714850
RT14	05/12/09	26.4	17.5	5.9	6.9	278	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT15	05/12/09	26.4	16.3	6.1	3.2	414	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT16	06/12/09	24.2	16.3	6.1	0.5	754	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m	n/m
RT17	04/12/09	26.4	16.8	6.1	5.0	409	20.9	110	0.51	3.7	2.7	1.6	7.9	59.4	207	-4.3	-22.0	3.13	0.045	n/m
RT18	07/12/09	11.4	16.2	6.1	8.2	701	5.8	203	0.92	24.5	4.4	3.0	13.3	99.7	355	-3.3	-17.3	0.23	0.074	0.714320
RT19	06/12/09	11.4	19.6	6.0	7.5	722	5.5	202	0.93	25.1	4.7	3.0	14.4	101.0	357	-3.5	-18.2	0.21	0.074	n/m

Table 3.4 Physical chemistry, ion concentrations, stable isotopes of water and $^{87}Sr/^{86}Sr$ ratios of monthly rainfall from Rocky River Catchment. n/m indicates not measured.

Sample sites	Date	pH	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	Sr (mg/L)	$^{87}Sr/^{86}Sr$
Rain RR airstrip	05/11/07	5.4	4.1	12.6	0.05	2.5	0.8	0.2	0.9	7.4	29	n/m	n/m	0.009	n/m
Rain RR airstrip	03/12/07	6.1	5.1	8.1	0.05	1.8	0.9	1.8	0.6	4.7	23	n/m	n/m	0.006	n/m
Rain RR airstrip	Feb-Apr 2008	6.3	4.3	15.2	0.03	2.2	1.6	0.6	1.0	9.3	34	-20.5	-4.7	0.014	n/m
Rain RR airstrip	14/05/08	6.4	3.8	9.4	0.05	1.6	0.7	0.2	0.7	6.2	23	-23.9	-4.9	0.006	n/m
Rain RR airstrip	03/06/08	6.1	3.2	11.2	0.05	1.8	0.5	0.2	0.9	7.0	25	-26.2	-5.3	0.008	n/m
Rain RR airstrip	30/06/08	6.5	3.7	12.2	0.05	1.9	0.8	0.3	0.9	7.9	28	-31.5	-6.2	0.008	0.70985
Rain RR airstrip	14/08/08	6.2	4.8	16.8	0.05	2.7	0.8	0.4	1.3	11.8	39	-18.0	-4.3	0.013	n/m
Rain RR airstrip	07/09/08	6.4	3.8	19.6	0.05	3.7	1.2	0.7	1.4	12.6	43	-13.6	-3.5	0.014	n/m
Rain RR airstrip	08/10/08	6.5	55.7	10.0	0.05	4.6	1.3	3.8	1.0	3.7	80	-8.2	-2.3	0.012	n/m
Rain RR airstrip	16/11/08	7.7	79.6	57.0	0.09	5.0	13.5	2.6	6.3	36.3	200	10.1	3.5	0.062	0.70994
Rain RR airstrip	04/03/09	5.9	3.0	14.0	0.05	2.0	0.8	0.3	0.8	4.8	26	-24.2	-4.7	0.007	0.71003
Rain RR airstrip	17/05/09	6.7	5.6	7.1	0.05	1.2	1.1	0.2	0.4	1.6	17	-29.6	-5.6	0.005	n/m
Rain RR airstrip	10/06/09	6.4	4.2	10.0	0.05	1.6	0.8	0.1	0.6	3.1	21	-39.1	-6.9	0.005	n/m
Rain RR airstrip	11/07/09	6.3	4.0	15.0	0.05	2.1	0.7	0.2	0.9	4.8	28	-25.3	-4.8	0.006	0.70974
Rain RR airstrip	03/12/09	6.8	6.8	32.7	1.03	5.6	2.3	0.8	2.1	16.7	69	-12.0	-2.7	0.060	n/m
Rain East Melrose	05/11/07	5.9	4.8	12.0	0.05	2.0	0.7	0.8	0.9	7.1	28	n/m	n/m	0.005	n/m
Rain East Melrose	03/12/07	5.5	4.1	6.0	0.05	1.3	0.6	0.6	0.4	3.4	17	n/m	n/m	0.007	n/m

Table 3.4 continued.

Sample sites	Date	pH	HCO_3 (mg/L)	Cl (mg/L)	Br (mg/L)	SO_4 (mg/L)	Ca (mg/L)	K (mg/L)	Mg (mg/L)	Na (mg/L)	TDS (mg/L)	$\delta^{18}O$ (‰VSMOW)	$\delta^{2}H$ (‰VSMOW)	Sr (mg/L)	$^{87}Sr/^{86}Sr$
Rain East Melrose	17/01/07	7.1	85.0	11.9	0.05	0.1	7.5	9.4	3.4	9.0	126	-10.1	-2.9	0.075	n/m
Rain East Melrose	14/02/08	6.6	6.5	13.9	0.05	0.4	1.2	0.8	1.0	9.0	33	-19.4	-4.5	0.013	0.71168
Rain East Melrose	03/06/08	6.9	11.7	8.9	0.05	2.4	1.0	1.2	0.8	6.6	33	-25.4	-5.4	0.012	0.71255
Rain East Melrose	30/06/08	7.6	43.4	11.6	0.05	1.0	2.3	1.7	1.2	7.7	69	-30.7	-6.2	0.026	0.71307
Rain East Melrose	14/08/08	7.9	60.6	16.7	0.05	6.2	3.0	4.5	1.6	11.1	104	-17.0	-4.3	0.035	n/m
Rain East Melrose	07/09/08	6.9	11.1	20.4	0.05	5.1	2.4	1.7	1.5	13.7	56	-13.5	-3.7	0.031	n/m
Rain East Melrose	08/10/08	3.8	0.0	14.0	0.05	5.3	2.0	5.4	1.8	4.8	33	-8.4	-1.8	0.016	n/m
Rain East Melrose	16/11/08	6.2	7.2	7.0	0.05	1.6	0.8	0.8	0.5	2.6	21	-13.9	-2.6	0.010	0.71147
Rain East Melrose	31/12/08	6.3	20.7	18.0	0.05	3.4	2.4	3.1	1.4	6.9	56	-28.5	-4.4	0.015	
Rain East Melrose	07/03/09	6.3	4.4	11.0	0.05	1.7	0.6	0.6	0.7	3.2	22	-24.6	-4.8	0.006	0.71161
Rain East Melrose	17/05/09	6.4	13.0	6.6	0.05	0.3	0.5	0.7	0.4	1.5	23	-40.3	-7.0	0.006	0.71109
Rain East Melrose	14/06/09	6.5	20.8	11.0	0.05	0.2	1.2	1.4	0.8	3.4	39	-32.1	-5.9	0.013	
Rain East Melrose	11/07/09	6.7	35.0	13.0	0.05	0.6	1.8	2.2	1.0	4.4	58	-24.4	-4.8	0.019	0.71250
Rain East Melrose	01/12/09	7.2	27.3	20.6	1.06	4.7	1.6	1.4	1.5	10.7	76	-18.0	-3.4	0.143	

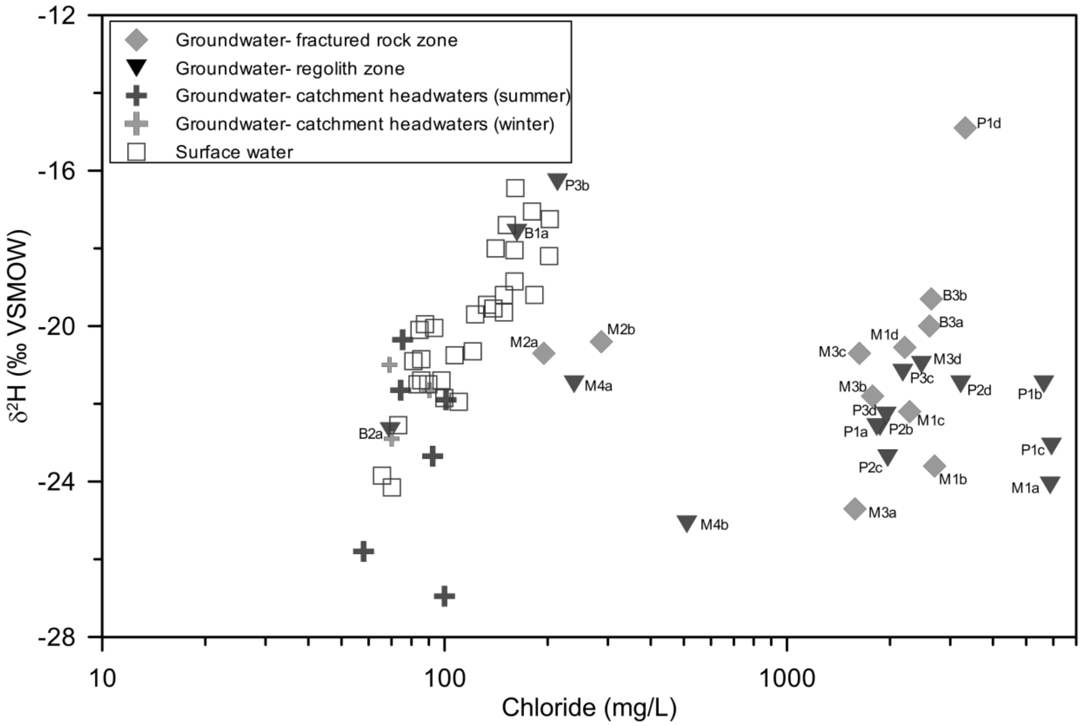

Figure 3.10 δ^2H versus chloride of surface water from Rocky River and groundwater in the RRC.

3.3.6 ^{222}RADON

The ^{222}Rn activities in the surface water and groundwater are shown in Table 3.2 and Table 3.3. The groundwater from the fractured rock aquifers had ^{222}Rn activities that ranged from 9 to 200 Bq/L which was higher than the ^{222}Rn activities of the groundwater from the regolith zone which ranged from 4 to 10 Bq/L. The groundwater from the shallow perched Quaternary sand aquifer in the catchment headwaters had ^{222}Rn activities that ranged from 6 to 80 Bq/L. The measured ^{222}Rn activities in the river ranged from zero up to 16 Bq/L. The hydraulic and river discharge measurements showed that Rocky River was gaining from the catchment headwaters to the 26.5 km mark (several km above East Melrose) and this was supported by the high ^{222}Rn activities measured in the upper part of the catchment which gradually decreased from 16 Bq/L to less than 0.2 Bq/L by the time it reached East Melrose (Figure 3.7e and Figure 3.8). The steady decrease in ^{222}Rn activities between the 26.5 km and 16.1 km mark suggests there was minimal to no groundwater contribution to streamflow which corresponded to the observed losing type conditions along this reach of the river. A spike in ^{222}Rn activity near the Platypus Pools may be a result of in-situ ^{222}Rn production in the hyporheic zone of the river and the large permanent pools at this location. The locations of high ^{222}Rn activity indicate localised groundwater discharge to the river (in this investigation we assume that the groundwater has a high ^{222}Rn activity and reflects the lithology of the major aquifers sampled). The surface water in the tributaries and main river channel in the catchment headwaters had high ^{222}Rn activities for relatively small flows compared to further down the catchment. The ^{222}Rn activity downstream of the groundwater discharge location declines due to its short half-life (3.82 days) and loss to the atmosphere by gas exchange. The rate of gas loss is controlled by the gradient of stream (controlling degassing rate), as well as volume of discharge, stream profile and streambed roughness (Ellins *et al.*, 1990). The steady ^{222}Rn activities along some reaches of Rocky River suggest a balance between steady groundwater discharge and the de-gassing and radioactive decay of ^{222}Rn in the river.

3.3.7 ^{87}Sr/^{86}Sr Ratios

Strontium isotope ratios have been used successfully in surface water–groundwater interaction studies to distinguish different input sources and help identify likely flow pathways (Land et al., 2000; Shand et al., 2007a; Shand et al., 2009). The ^{87}Sr/^{86}Sr ratios of the bulk monthly rainfall, surface water and groundwater in the catchment are shown in Table 3.2, Table 3.3 and Table 3.4. The analysed monthly rainfall collected at East Melrose and the Rocky River airstrip had ^{87}Sr/^{86}Sr ratios from between 0.7097 and 0.7130 over the study period, with some samples displaying evaporative enrichment and contribution from windblown dust with a higher radiogenic strontium component (Figure 3.11). Rainfall in coastal areas has a similar ^{87}Sr/^{86}Sr to modern seawater (0.709) and strontium concentrations that are typically several orders of magnitude lower (Aberg *et al.*, 1989). The ^{87}Sr/^{86}Sr ratios in the groundwater from the regolith zone ranged from 0.7108 to 0.7177, displaying a slightly larger range than groundwater from the fractured rock aquifer which ranged from 0.7138 to 0.7182. The groundwater from the shallow perched aquifer in the catchment headwaters had ^{87}Sr/^{86}Sr ratios that ranged from 0.7103 to 0.7150. The ^{87}Sr/^{86}Sr ratios in the surface water had a range of 0.7124 to 0.7152 that was between rainfall and groundwater. The ^{87}Sr/^{86}Sr ratios in the surface water showed an increasing trend from the catchment headwaters to above East Melrose where it stabilised before decreasing towards Snake Lagoon (Figure 3.7b). The considerable decrease in the ^{87}Sr/^{86}Sr ratio below the weir to Snake Lagoon corresponded to an increase in the strontium and calcium concentration and pH. This is likely to be related to carbonate dissolution from the carbonate rocks (calcarenite) that outcrop in this part of the catchment, and is in agreement with Shand et al (2008). Whilst rainfall has a similar ^{87}Sr/^{86}Sr ratio to calcarenite, dilution by rainfall is unlikely as it would be expected that both the strontium concentration and isotope ratio would decrease and this was not observed.

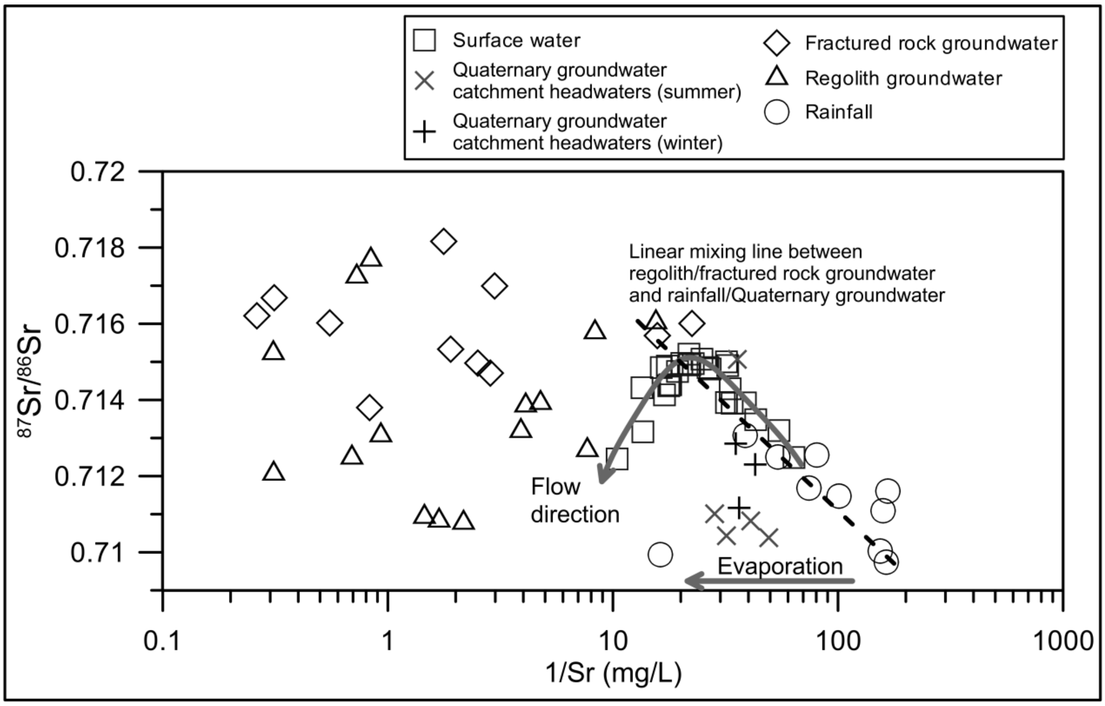

Figure 3.11 $^{87}Sr/^{86}Sr$ ratio versus 1/Sr (mg/L) for Rocky River, groundwater from the study sites, shallow groundwater in the catchment headwaters and local rainfall.

The plot of $^{87}Sr/^{86}Sr$ ratio versus 1/Sr shows the groundwater from the shallow perched aquifer from the catchment headwaters and surface water samples from Rocky River plot in between the rainfall samples and the groundwater samples from the regolith and fractured rock aquifers (Figure 3.11). The groundwater samples collected in winter from the shallow perched aquifer had $^{87}Sr/^{86}Sr$ ratios similar to rainfall and the surface water in the catchment headwaters and were more radiogenic than the groundwater samples collected in summer which also had an evaporated rainfall signature. This suggests that there is a lag time between precipitation moving through the perched aquifer and into the river on the order of 6 months, as it would be expected that the winter strontium isotopic composition would be less radiogenic than during the summer.

The two groundwater samples from the fractured rock beneath the river at East Melrose (M2a and M2b) plot closely to the surface water samples (Figure 3.11). This implies that there is some mixing between the lower strontium concentration surface water with the groundwater and supports the losing type stream conditions at this location. The majority of the surface water samples fall along a mixing line between the rainfall and shallow Quaternary perched groundwater with the average strontium composition of the groundwater (0.7146) from the regolith and fractured rock aquifers (Figure 3.11). The surface water samples that plot on the downward trending flow direction line represent the lower part of the catchment where the strontium isotopic and concentration signature of the surface water is influenced by the change in geology from the sandstone to limestone lithology, as discussed previously.

To examine the relative proportions of the contributing sources to Rocky River a strontium isotope end-member mixing analysis was used which is described below.

$$^{87/86}\mathrm{Sr_M} = {}^{87/86}\mathrm{Sr_A} \,.\, f\frac{\mathrm{Sr_A}}{\mathrm{Sr_M}} + {}^{87/86}\mathrm{Sr_B}.\ \frac{\mathrm{Sr_B}(1-f)}{\mathrm{Sr_M}} \qquad (3.1)$$

(Faure, 1986)

In Equation 3.1, $^{87/86}Sr_A$, $^{87/86}Sr_B$ and $^{87/86}Sr_M$ are the strontium isotope ratios of end-members A and B and the mixture M. The terms Sr_A, Sr_B and Sr_M are strontium concentrations in A, B and M, and *f* is the fraction of the end-member A in the mixture. A range of values for rainfall and groundwater from the study were used for the end members of the strontium isotope ratios and strontium concentrations. The mixing equation determined that the surface water was comprised dominantly of the rainfall/shallow perched Quaternary groundwater end member (greater than 95 %) compared to only a small portion from the regolith/fractured rock groundwater (less than 5%). In Equation 3.1 the strontium concentration is more sensitive than the strontium isotope ratio, however, the results suggest that the rainfall and shallow perched groundwater end member is the greatest contributor to the strontium signature in the river.

3.3.8 Conceptual Model

The new conceptual model of the near stream environment in the gaining river reaches of the catchment headwaters is a perched river system which is hydraulically connected to a fresh shallow perched Quaternary sand aquifer which overlies a clay layer of low conductance on top of the regional aquifer in the fractured bedrock (Figure 3.12-a). The water source to the river is from winter rainfall, falling between May and October with occasional summer rain and the slow discharge from the shallow perched Quaternary aquifer system in the catchment headwaters. This shallow perched aquifer system, buffered by a multitude of in-stream swamp systems in the catchment headwaters has a large water storage capacity to provide year-round base flow to the surface water system without the need for connection to the regional groundwater system. These swamp systems significantly reduce stream flow velocities and result in an increase in residence time of the water in the surface water system.

At the transition point along the river from gaining to losing type conditions (e.g. East Melrose) there was no Quaternary perched aquifer and little to zero groundwater contribution from the shallow regolith zone. Outcrops of bedrock were frequent along the riverbed but were not a significant contributing groundwater source (Figure 3.12-b). The Platypus Pools was uniquely different to other parts of the catchment because it showed both gaining and losing type conditions. The river at this location was comprised of a series of permanent ponds established within the low lying landscape and thick regolith zone underlain by fractured bedrock (Figure 3.12-c). Along the losing river reaches of the catchment downstream of the Platypus Pools is a perched river system which is hydraulically connected to a relatively fresh shallow aquifer in the fluvial sediments which overlies a more saline regional aquifer in the fractured bedrock (e.g. the Bridge site). In some locations there was an unsaturated zone between the river and the fractured rock aquifer system and was characterised as either a transitional or disconnected system (Figure 3.12-d). Towards the bottom of the catchment, below the Bridge site (7.9 km), there was a considerable change in the river flow conditions and water chemistry which corresponded to a change in the geology and potentially another groundwater source to the river (Figure 3.12-e).

An annual catchment water balance indicated that 90 % of the rainfall is lost to groundwater recharge and/or evapotranspiration with only 10 % exiting the catchment via Rocky River. High evapotranspiration rates from the native vegetation in a dominantly losing type system are also likely to contribute to maintaining a fresh river system by keeping the watertable at a lower elevation than what would be observed if the catchment was cleared. In this state, deeper regional saline groundwater from the fractured rock aquifer cannot discharge into the river as base flow. This hypothesis is currently being examined in a separate numerical modelling study and model results suggest this is indeed a plausible process.

The Rocky River catchment provides a baseline assessment of a pristine, undisturbed catchment. Adjacent catchments (e.g. Cygnet River and Timber Creek) and others within the Mount Lofty Ranges South Australia, which have been cleared of vegetation, have become salinised river systems with surface water EC values of up to 20,000 µS/cm (Shand et al., 2008). There is limited understanding of the impacts from the removal of native vegetation on surface water–groundwater connectivity in these types of systems. This is important because in regions of Australia, native vegetation clearance has had considerable impacts on surface water and groundwater salinities, through the mobilisation of salt stored in the shallow regolith as well as increased groundwater discharge (Allison *et al.*, 1990). A previous investigation in the RRC had identified that there was a low salinity subsurface source that sustained semi-perennial flow down the river most years and through the summer (Shand et al., 2007b). At the time, there were no groundwater monitoring wells in the RRC but it was hypothesised that conduits and macropore flow existed between the underlying fractured rock groundwater system and the surface water system as an important mechanism for bypassing the salt stored beneath native vegetation. Our study has found that the dominant low salinity subsurface source is most likely to be from the shallow Quaternary perched aquifer system and that the catchment headwaters are very important in streamflow generation down the catchment.

A semi-impermeable boundary comprised of clay material, as a result of bedrock weathering processes, provides ideal conditions for the establishment of perched aquifer systems. The perched aquifer is able to provide base flow to the river which is not in contact with the regional groundwater system. Similar to a study by Niswonger and Fogg (2008), the perched aquifer provided gaining conditions and maintained base flow for longer periods. These shallow systems are maintained only by seasonal rainfall and therefore they are very susceptible to exploitation and sensitive to climate change. This study particularly highlights the importance of headwater systems in controlling and maintaining stream flows. Therefore, it is critical that headwater areas be given much more importance in managing river systems at the catchment scale.

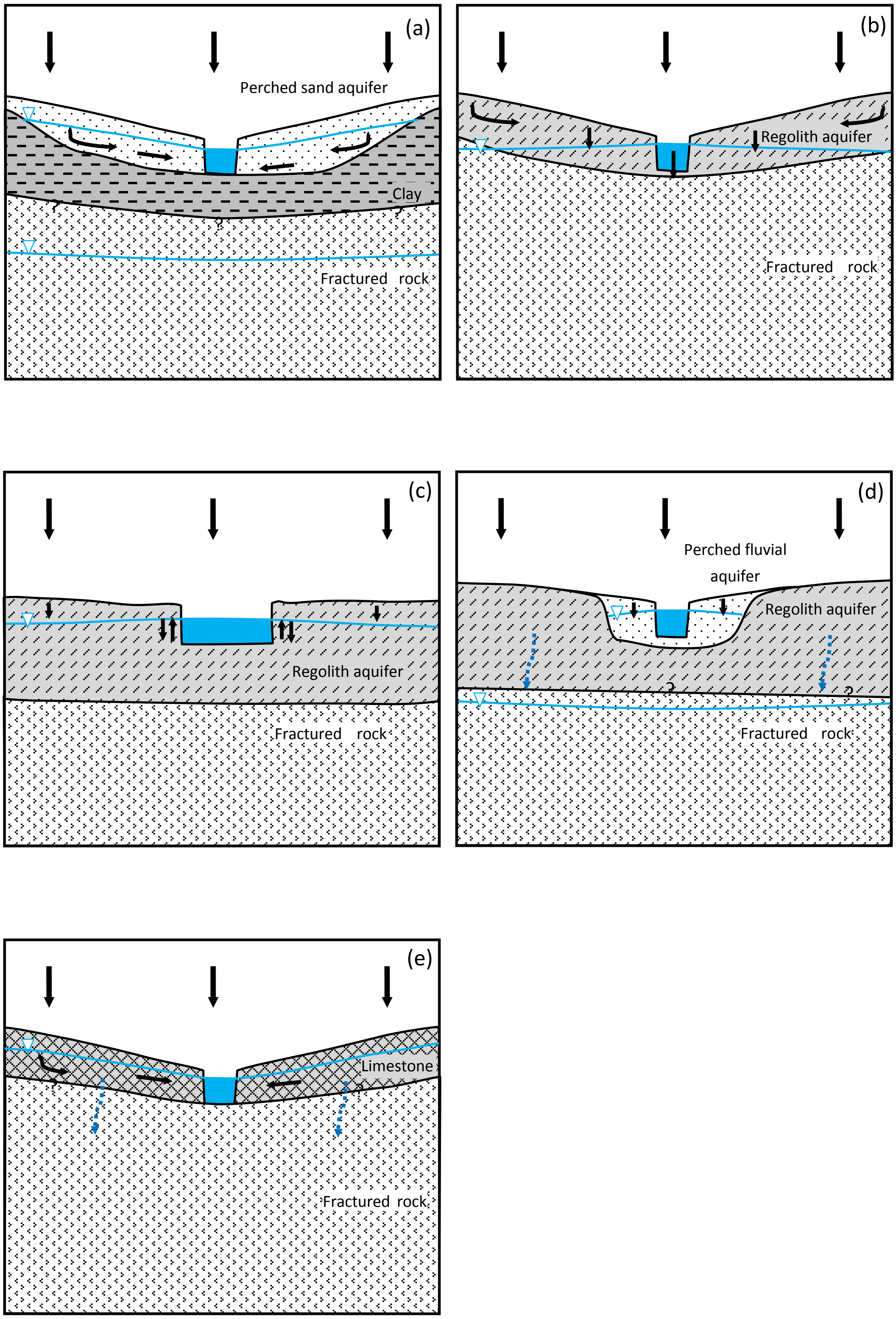

Figure 3.12 Conceptual model along different stream reaches of the RRC. (a) catchment headwaters, (b) East Melrose, (c) Platypus Pools, (d) the Bridge Site, and (e) Snake Lagoon.

3.4 CONCLUSIONS

There are very few entire river system studies which have shown how surface water–groundwater interactions can change along river reaches, as well as the effects that the state of connection can have on the catchment water and salt balances. This study demonstrated how hydrogeological and hydroclimatic controls influence the state of connection in a pristine catchment and also the complexities of fractured rock environments. Flow rate and water level data, together with salinity and stable isotope results, showed that the river system was not connected to the regional saline groundwater aquifer system and can generally be classified as either a transitional or disconnected type system. The relatively low salinity of the fresh water river system can be maintained in an otherwise saline regional fractured rock groundwater system by virtue of the dominantly losing connectivity state (i.e. there is little to no groundwater discharge from the deeper regional system to the river as base flow). The hydrochemistry showed that the dominant groundwater source to the river system was from the shallow sedimentary perched aquifer system in the catchment headwaters and not from the regional fractured rock aquifer system as previously thought (Shand et al., 2007b). The perched aquifer system provided base flow down the length of Rocky River and maintained flow for longer periods. Groundwater discharge from the fractured rock aquifer system to the river is unlikely because the groundwater from this system was much more saline than the low salinity of Rocky River.

We hypothesise that there are important vegetation (evapotranspiration) controls on the state of connection between surface water and groundwater of this system. Further research to quantify the effects of vegetation clearance and landuse change on the groundwater levels in the perched and fractured rock aquifers, state of disconnection, and saline accessions from groundwater to the river is essential to ensure more appropriate management practices can be employed. This is especially important for understanding the impacts of land use change and climate change on the quantity and quality of surface water and groundwater in large regional scale catchments.

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CHAPTER 4

4. MANUSCRIPT III: VEGETATION CONTROLS ON VARIABLY-SATURATED PROCESSES BETWEEN SURFACE WATER AND GROUNDWATER AND THEIR IMPACT ON THE STATE OF CONNECTION

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ABSTRACT

The vadose zone plays an important role in surface water–groundwater interaction and exerts strong influences on biogeochemical, ecological, and hyporheic processes. It is also the presence of an unsaturated zone that controls the state of connection between surface water and groundwater. Despite recent advances on how hydrogeological variables affect surface water–groundwater interactions, there is limited understanding of the hydroclimatic effects of precipitation and evapotranspiration. More specifically, there is a need for a physically based understanding on the changes that may occur in response to changes in vegetation. While it may seem qualitatively obvious that the presence of vegetation can cause an unsaturated zone to develop underneath a riverbed and alter the state of connection, it has so far not been demonstrated quantitatively. Also, the influence of variables such as root extinction depth, topography, and the influence of land clearance has so far not been explored. In this study, fully coupled, physically based 2-D transient homogeneous models were used to simulate the impact of land clearance and revegetation on the state of connection of a perennial river system. The simulations showed that the presence of vegetation can create an unsaturated zone between a river and an aquifer and affect the state of connection and that the removal of deep-rooted vegetation from a catchment may have a significant impact on the state of connection as well as the condition of the water resource.

4.1 INTRODUCTION

Vadose zone processes play an important role in surface water–groundwater interaction. For example, the presence of an unsaturated zone has a strong influence on biogeochemical processes of river systems (Bencala, 1993), the fate of nutrients (Boulton et al., 1998), and various ecological and hyporheic exchange processes (Brunke and Gonser, 1997; Findlay, 1995). Also, it is the presence of an unsaturated zone which controls the state of connection between surface water and groundwater. The state of connection has received greater attention in the last decade (Brunner et al., 2011; Brunner et al., 2009b; Fox and Durnford, 2003; Harvey and Wagner, 2000; Vazquez-Sune et al., 2007), in response to concerns about water scarcity and the sustainable management and allocation of water resources (Sophocleous, 2002; Winter et al., 1998).

In natural environments, surface water–groundwater systems are influenced by physical (hydrogeological), topographical (terrain and landform), and hydroclimatic (precipitation and evapotranspiration) variables and are generally classified as connected or disconnected systems. Connected systems are either (1) gaining, where groundwater discharges through the riverbed to contribute to river flow, or (2) losing, where water infiltrates from the river to the groundwater system. Losing systems can sometimes be disconnected. Disconnected systems show flow losses through an unsaturated zone, and as a result changes in the water table do not significantly affect the infiltration rates from the river. The flow regime between connected and disconnected is called transitional and is the state between the initial development of an unsaturated zone and the point where the infiltration rate no longer changes in response to a further decline in the water table. In transitional and disconnected systems, an unsaturated zone under the riverbed is present (Brunner et al., 2009a; Brunner et al., 2011).

In a recent study, Brunner et al. (2009a) used a theoretical and modelling approach to examine the most important hydrogeological parameters that influence the state of connection between surface water and groundwater and developed a set of criteria to determine whether a system can become disconnected or not in the presence of a low conductivity streambed. Reisenhauer (1963) showed that a disconnection is possible even in the absence of a clogging layer due to capillary effects. However, we are unaware of any field documentation to support such a case, and Peterson and Wilson (1988) concluded this is unlikely to occur. Typically, an unsaturated zone (and therefore a disconnection) can develop in the presence of a streambed with a lower hydraulic conductivity than the underlying aquifer. Such low conductivity layers (subsequently called a clogging layer) are widespread and have been observed in rivers throughout all climatic zones. Low conductivity layers can develop due to biological clogging (Treese et al., 2009) or sedimentary processes (Schälchli, 1992). Hatch et al. (2010) showed using field data that heterogeneity in the streambed can result in the presence of both saturated and unsaturated zones adjacent to one another. Frei et al. (2009) reached the same conclusion through simulating the effect of heterogeneity on surface water–groundwater interaction. Other modelling studies (Niswonger and Fogg, 2008) have investigated the hydrogeological effects on perched stream aquifer type systems, while Desilets et al. (2008) focused on the pathways and rates of infiltration as stream–aquifer systems transition from connected to disconnected and how these effect local groundwater flow patterns. It has also been shown that groundwater pumping adjacent to surface water systems can have a considerable effect on the interaction between surface water and groundwater (Moore and Jenkins, 1966; Spalding and Khaleel, 1991; Su et al., 2007). Fox and Durnford (2003) showed that groundwater pumping adjacent to a surface water body can induce an unsaturated zone and a disconnection between surface water and groundwater.

Despite these recent advances in knowledge on the relation between hydrogeological variables, the presence of an unsaturated zone and the state of connection, there is limited quantitative understanding of the role of vegetation (i.e., evapotranspiration) in forming an unsaturated zone between surface water and groundwater. It is also unclear, in a precise quantitative cause and effect manner, how land clearance or revegetation affects the state of connection. In Australia, evapotranspiration is an important hydroclimatic process because native vegetation clearance has had considerable impacts on surface water and groundwater salinities, through the mobilization of salt stored in the shallow regolith as well as increased groundwater discharge (Allison et al., 1990; Allison and Forth, 1982; Cook et al., 1994; George et al., 1997). Williamson et al. (1987) compared water and salt balances from cleared and uncleared catchments in south-western Australia and showed that in the cleared catchments there was a significant increase in streamflow and exported salt, and that the water and salt balances had not yet reached a new equilibrium.

There is an analogy that trees behave like groundwater pumps and that they are able to remove or intercept a large portion of precipitation input in a catchment water balance. While trees and pumps might have similar effects on the water balance, the way they extract water from the aquifer is fundamentally different. It is not intuitively obvious how a change in the water table relates to a tree's ability to consume groundwater (Butler et al., 2007; Loheide et al., 2005; Shafroth et al., 2005), let alone effect its ability to create an unsaturated zone between a river and an aquifer or alter the state of connection. Several studies have calculated the amount of water certain vegetation types transpire on an annual basis (Farrington et al., 1994; Salama et al., 1994). Results from these studies found that 11.4–18.0 m^3 of water was transpired per year per eucalyptus tree in Western Australia. Banks et al. (2011b) hypothesized that there may be important vegetation (evapotranspiration) controls on the state of connection between surface water and groundwater in a pristine catchment on Kangaroo Island, South Australia. Their study suggested that the presence of vegetation was a fundamental difference in controlling the surface water–groundwater interactions compared to adjacent catchments that had been cleared of vegetation and which had significantly more saline surface water (Henschke et al., 2003; Shand et al., 2007c). The results of the study by Banks et al. (2011b) also showed that the relatively low salinity of the fresh water river system can be maintained in an otherwise saline regional groundwater system by virtue of the dominantly losing state of connection. It was hypothesized that this losing state was created and maintained by vegetation cover in the pristine catchment. Qualitatively, it may seem obvious that the presence of vegetation can alter the state of connection or that a change to the rates of precipitation and evapotranspiration may, under certain conditions, have some effect on the state of connection. However, there is little understanding what the quantitative effects will be and the sensitivity of the state of connection (and associated exchange fluxes) to various controlling physical variables. Such variables include the hydraulic conductivity of the aquifer and clogging layer, catchment slope, and root extinction depth.

The aim of this paper is to explore the hypothesis: can the presence of vegetation create an unsaturated zone or even a disconnection between a river and an aquifer using reasonable and representative vegetation and hydroclimatic variables? We further explore how land clearance can affect the state of connection. In this context we study the following question: is evapotranspiration a plausible mechanism to create an unsaturated zone underneath a riverbed and how does it influence connected gaining and losing, and losing–disconnected type conditions? The physical controls of catchment slope, the hydraulic conductivity of the riverbed clogging layer and aquifer, and the vegetation root extinction depth (transpiration extinction depth) are also examined to determine how these variables influence the state of connection. Understanding how vegetation type and cover (and hence evapotranspiration) relates to an unsaturated zone and affects the state of connection provides valuable information on what the impacts of vegetation clearance, revegetation or changes in land use are likely to be on surface water–groundwater connection and the consequential effects on water quality. Similarly, climate change impacts can be qualitatively inferred from these general cause and effect type relationships.

4.2 NUMERICAL MODELLING

4.2.1 CONCEPTUALISATION OF DISCONNECTED SYSTEMS

Whether a surface water–groundwater system is connected, disconnected, or in transition between the two states has profound implications on how changes in the water table affect the exchange fluxes in the system. In the presence of a clogging layer within the streambed, lowering the water table can result in an unsaturated zone under the clogging layer; the system first enters a transition mode. Further lowering the water table increases the infiltration flux, and a maximum value for the current hydraulic system is approximated. This upper limit of infiltration (corresponding to an upper limit of suction under the clogging layer) can be calculated. However, in reality this upper limit is only approximated and therefore a cut-off value that separates transition from disconnection has to be defined. In this study, the cut-off between disconnected and transitional type systems was defined at 1% of the upper limit of pressure head (suction; γ^*_p) at full disconnection. The pressure head at full disconnection was determined using the following equation, and it can be solved graphically (Osman and Bruen, 2002) or numerically (Brunner et al., 2009a).

$$K_c \, \frac{(h_c + d - \gamma^*_p)}{h_c} = K_a k_r(\gamma^*_p) \qquad (4.1)$$

where γ^*_p is the pressure head that develops at the interface between the clogging layer and the aquifer at full disconnection, K_c is the hydraulic conductivity of the clogging layer, K_a is the hydraulic conductivity of the aquifer, h_c is the thickness of the clogging layer, d is the depth of the river, k_r is the relative hydraulic conductivity of the aquifer, which is derived from a relationship between pressure and hydraulic conductivity according to pressure-saturation curves from the work of van Genuchten (1980). The left-hand side of Equation 4.1 is the infiltration flux through the clogging layer calculated following Darcy's law. For disconnected systems the infiltration under the clogging layer is driven by gravity drainage, therefore, the infiltration flux equals the hydraulic conductivity of the aquifer (right-hand side of the equation $K_a\, k_r$). Equation 4.1 cannot be solved analytically due to the highly non-linear relations between pressure and relative hydraulic conductivity.NUMERICAL MODEL: HYDROGEOSPHERE

Surface water–groundwater interactions were simulated using the groundwater flow model HydroGeoSphere (HGS) (Therrien et al., 2010). HydroGeoSphere is a physically based, numerical model describing fully integrated surface and unsaturated and saturated flow in the subsurface. The capability of HGS to model flow in the unsaturated zone using the Richards equation as well as disconnection between surface water and groundwater has significant benefits over modelling codes that do not explicitly consider the unsaturated zone as outlined in the manuscript by Brunner et al. (2010). We limit the description of HGS to the conceptualization of evapotranspiration. For further details on the code and a recent software review the reader is referred to Therrien et al. (2010) and Brunner and Simmons (2011).

Evapotranspiration (ET) is modelled as a combination of plant transpiration and evaporation, and affects both the surface and subsurface flow domains. Transpiration from vegetation occurs within the root zone of the subsurface and is a function of the leaf area index (*LAI*) [dimensionless], nodal water (moisture) content (θ) [dimensionless] and a root distribution function (*RDF*) over a prescribed extinction depth. Water content is simulated as saturation because it is more stable and always varies between zero and one, while in reality moisture content varies from zero to a value equal to porosity. As we will discuss later in the base case setup, the effects of vegetation clearance and revegetation are simulated by modifying the extinction depth, e.g., a small depth value is used for a change from native vegetation to shallow rooted pasture crops and a zero depth value is used for no vegetation (land clearance). The rate of transpiration (T_p) is estimated using the following relationships (Kristensen and Jensen, 1975):

$$T_p = f_1(LAI)\, f_2(\theta)\, RDF\, \left[E_p - E_{can}\right] \qquad (4.2)$$

where E_p is the reference potential evapotranspiration which may be derived from pan measurements or computed from vegetation and climatic factors [L T^{-1}] and E_{can} is the tree canopy evaporation [L T^{-1}]. E_p can also be described as the amount of water that would be removed through evapotranspiration if the watertable was at the ground surface. The value and description of E_p has followed the notation and conceptualisation of Therrien et al. (2010) and Kristensen and Jensen (1975). The vegetation function (f_1) correlates the transpiration (T_p) with the leaf area index (LAI) in a linear fashion and is expressed as:

$$f_1(LAI) = \max\{0, \min[1, (C_2 + C_1 LAI)]\} \tag{4.3}$$

The root zone distribution function (RDF) is defined by the relationship:

$$RDF = \frac{\int_{C_1}^{C_2} r_F(z)dz}{\int_0^{Lr} r_F(z)dz} \tag{4.4}$$

The moisture content (θ) function (f_2) correlates T_p with the moisture state at the roots and is expressed as:

$$f_2(\theta) = \begin{cases} 0 & \text{for } 0 \le \theta \le \theta_{wp} \\ f_3 & \text{for } \theta_{wp} \le \theta \le \theta_{fc} \\ 1 & \text{for } \theta_{fc} \le \theta \le \theta_o \\ f_4 & \text{for } \theta_o \le \theta \le \theta_{an} \\ 0 & \text{for } \theta_{an} \le \theta \end{cases} \tag{4.5}$$

where:

$$f_3 = 1 - \left[\frac{\theta_{fc}-\theta}{\theta_{fc}-\theta_{wp}}\right]^{\frac{C_3}{E_p}} \tag{4.6}$$

$$f_4 = 1 - \left[\frac{\theta_{an}-\theta}{\theta_{an}-\theta_o}\right]^{\frac{C_3}{E_p}} \tag{4.7}$$

and where C_1, C_2, and C_3/E_p are dimensionless fitting parameters, L_r is the effective root length [L], z is the depth coordinate from the soil surface [L], ϑ_{fc} is the moisture content at field capacity, ϑ_{wp} is the moisture content at the wilting point, ϑ_o is the moisture content at the oxic limit, ϑ_{an} is moisture content at the anoxic limit and r_F (z) is the root extraction function [L^3 T^{-1}] which typically varies logarithmically with depth. Below the wilting-point moisture content, transpiration is zero; transpiration then increases to a maximum at the field-capacity moisture content. This maximum is maintained up to the oxic moisture content, beyond which the transpiration decreases to zero at the anoxic moisture content. When available moisture is larger than the anoxic moisture content, the roots become inactive due to lack of aeration (Therrien et al., 2010).

In HGS, evaporation from the soil surface and subsurface soil layers is a function of nodal water content and an evaporation distribution function (*EDF*) over a prescribed extinction depth. The model assumes that evaporation (E_s) occurs along with transpiration, resulting from energy that penetrates the vegetation cover and is expressed as (Therrien et al., 2010):

$$E_s = \alpha^*(E_p - E_{can})[1 - f_1(LAI)]\, EDF \qquad (4.8)$$

The wetness factor (α*) is given by:

$$\alpha^* = \begin{cases} \frac{\theta - \theta_{e2}}{\theta_{e1} - \theta_{e2}} & \text{for } \theta_{e2} \leq \theta \leq \theta_{e1} \\ 1 & \text{for } \theta > \theta_{e1} \\ 0 & \text{for } \theta < \theta_{e2} \end{cases} \qquad (4.9)$$

where ϑ_{e1} is the moisture content at the end of the energy–limiting stage (above which full evaporation can occur) and ϑ_{e2} is the limiting moisture content below which evaporation is zero.

4.2.3 Conceptual Model

The conceptual model employed here was based on the research outcomes of a field-based study which investigated the state of connection between a fresh water river, a perched sedimentary aquifer and a saline fractured rock aquifer system in the pristine Rocky River catchment on Kangaroo Island, South Australia (Banks et al., 2011b). The long-term mean annual precipitation for this catchment is 780 mm y^{-1} and the mean reference potential annual evapotranspiration is 1400 mm y^{-1}. Evapotranspiration is greater than precipitation during the summer months whilst in winter there is potential for groundwater recharge when precipitation exceeds evapotranspiration (Figure 4.1).

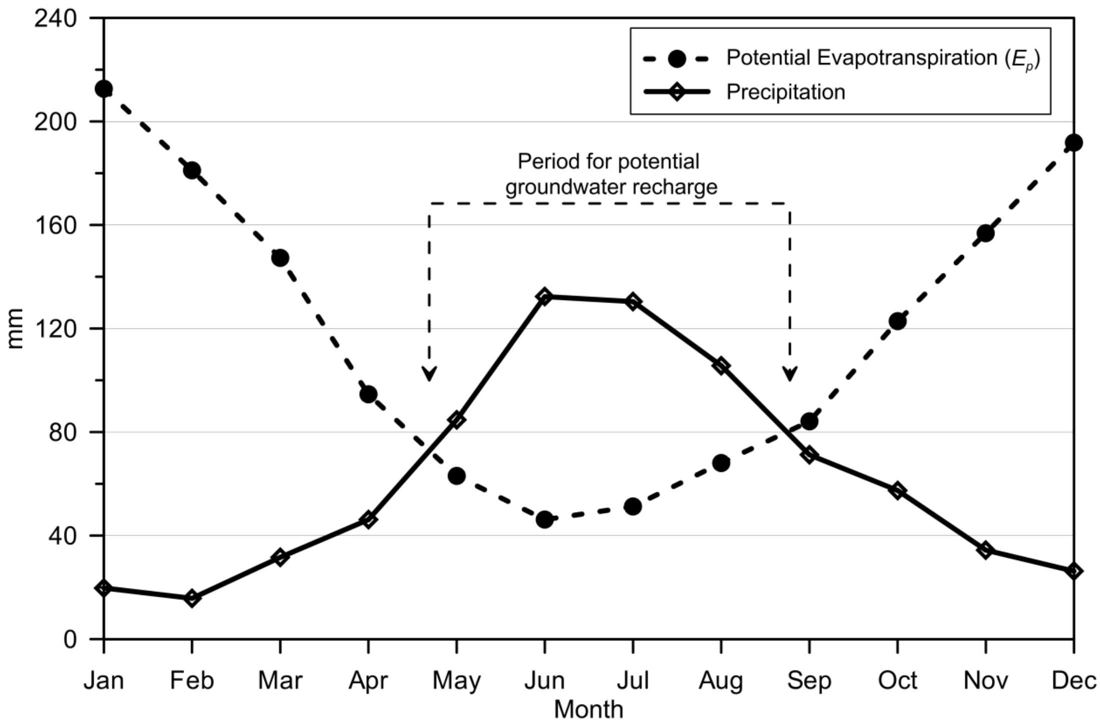

Figure 4.1 Monthly precipitation and reference potential evapotranspiration (E_p) for Rocky River catchment.

We defined a conceptual model that was used for all simulations (Figure 4.2). A base case was defined by assigning properties such as hydraulic conductivities, the slope of the catchment and the evaporation and vegetation parameters to the conceptual model. We then varied the hydraulic conductivity of the aquifer and the clogging layer as well as the slope and the evaporation and vegetation extinction depths to understand how these parameters relate to the influence of vegetation. While a number of other parameters could be varied additionally (e.g., porosity, the retention functions of the aquifer or surface properties such as rill storage height), the analysis was limited to these key parameters in order to keep the interpretation focused.

To analyse how vegetation affects the interaction between surface water and groundwater, forcing functions have to be applied to the system. A common way to conceptualise forcing functions in surface water–groundwater systems is to apply head boundaries to the watertable (Bruen and Osman, 2004; Brunner et al., 2009a; Osman and Bruen, 2002). As pointed out by several authors (Freeze, 1974; Lewandowski et al., 2009; Panday and Huyakorn, 2004), natural systems are often flux controlled, and therefore hydraulic heads are not themselves the forcing function. Instead, hydraulic heads are the response to changes in fluxes such as recharge, evaporation, transpiration or pumping. In order to study the effect of vegetation on surface water–groundwater interaction, we therefore applied precipitation and evapotranspiration as forcing functions. This approach is conceptually different to the aforementioned modelling approaches by Bruen and Osman (2004) and Osman and Bruen (2002).

We addressed the questions raised in the introduction with a homogeneous 2-D model (Figure 4.2) and do not consider any transience in the surface water domain. A simple conceptual model was chosen for a number of reasons: (1) In a more complicated setting (e.g. heterogeneous streambed and aquifer), it would be very difficult to examine and deconvolute the responses in the flow regime to changes in evaporation and transpiration processes; and (2) perennial river conditions (defined by a constant head) maintained 100 % saturation in the clogging layer. Simulating the temporal dynamics of ephemeral or intermittent streams and the wetting and drying cycles of the clogging layer undermine a unique and unambiguous interpretation of the systems response to the effects of vegetation and hence evapotranspiration.

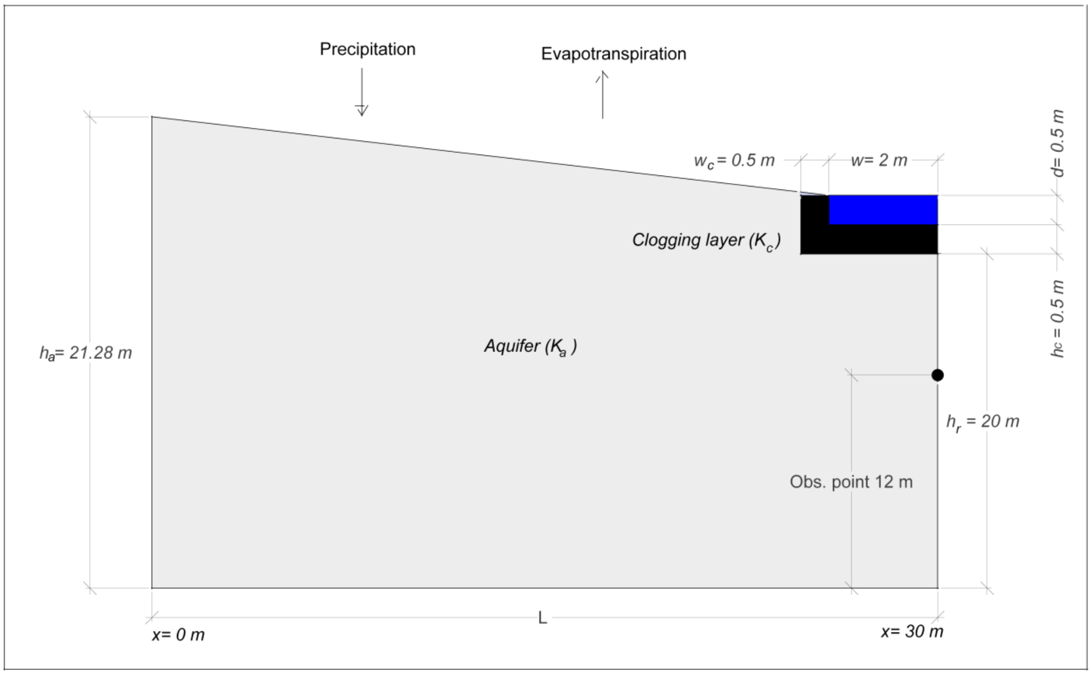

Figure 4.2 Conceptual model of the 2-D surface water–groundwater system. The river and aquifer are separated by a clogging layer (h_c) that is 0.5 m thick and 0.5 m wide at the river edge (w_c). The clogging layer has a hydraulic conductivity (K_c) that is less than the hydraulic conductivity of the aquifer (K_a). The river is defined by a constant head boundary with a water depth (*d*) of 0.5 m and has a width (*w*) of 2 m. *L* is the length (30 m) of the model in the x-direction. The height of the left-hand side of the model (h_a) at x = 0 is 21.28 m and the right-hand side (h_r) at x = 30 is 20.0 m. The left- and right-hand side and the base of the model are all no flow boundaries. The observation point directly beneath the centre of the river at 12 m elevation is also shown.

The model boundary conditions were designed according to the conceptual model in Figure 4.2. Because of the symmetry of the conceptual model only the left-hand side of the catchment was represented in the model domain to reduce computational time. The left- and right-hand side and the base of the model were all no flow boundaries. The river was perennial and represented as a constant head boundary. The horizontal extent of the model domain was designed in a way so that there were no significant impacts of the boundary conditions on the near river environment. The model discretisation was fine enough to ensure grid-independent results and to provide an appropriate level of vertical detail of the unsaturated zone. To ensure that this was the case, 60 subsurface layers were used. For example, below the river the vertical discretisation was set to 0.1 m from the top of the model domain down to 15.5 m elevation. Below 15.5 m elevation to the base of the model domain the layer thickness was increased and ranged from 0.2 up to 4 m in the base layer. The horizontal discretisation increased from 0.02 m near the river edge to 1 m at the left boundary of the model domain. The datum elevation (0 m) is located at the base of the aquifer.

The river was represented as a channel 0.5 m deep (vertical direction = 20.5 – 21 m) and 2 m wide (x – direction = 28 – 30 m). A slope between the river and the top left boundary was defined and the values provided in section 4.2.4. The slope was varied in the other scenarios tested. The clogging layer extended 0.5 m beneath the river channel and 0.5 m upslope from the edge of the river bank. The choice of this thickness is not critical for the following reasons: The ratio between the hydraulic conductivity of the clogging layer and its thickness is the first-order control of infiltration flux from the river to the aquifer. Given that the hydraulic conductivity of the clogging layer can vary by many orders of magnitude there is a large degree of freedom in choosing the thickness of the clogging layer.

The identification of the presence of an unsaturated zone is straight forward: if the water table drops below the clogging layer, an unsaturated zone develops. To determine whether the system is disconnected or not the hydraulic head beneath the centre of the river needs to be evaluated (Brunner et al., 2009a). Observation nodes were used to obtain point specific model outputs such as the hydraulic head or the position of the water table. For the purposes of this study, an observation borehole was located 12 m above the reference datum beneath the centre of the river. We used the hydraulic head of this borehole to approximate the location of the water table (defined through pressure = zero). This approximation was implemented to accelerate post-processing the large amount of model output data. A systematic comparison of a representative number of models was carried out and revealed that the largest deviation between the hydraulic head at an elevation of 12 m and the true location of the water table beneath the centre of the river was at greatest around a centimetre. We therefore considered the hydraulic head at this borehole location as a sufficiently accurate approximation of the location of the watertable.

4.2.4 Base Case Setup

The base case scenario was a transient model setup for a period of 7304 days (~ 20 years) using an initial time step of 0.1 days, a maximum time step of 1 day and a maximum time step multiplier of 1.25. The initial conditions of the model were determined numerically from a dynamic steady state under vegetated type conditions (evapotranspiration taking place). Based on these initial conditions the model was run for a period of ten years (3652 days) at which point the vegetation was removed or modified (through a change of the extinction depth) and the model was run for a further ten years. The slope of the catchment was 1cm m^{-1}. As a result, the ground elevation at the edge of the river (x – direction = 28 m) was 21 m and at the left boundary (x – direction = 0) was 21.28 m.

The physical properties of the clogging layer and the more hydraulic conductive homogeneous aquifer were based on representative literature values (Carsel and Parrish, 1988; Freeze and Cherry, 1979). The soil moisture retention curve for the aquifer, defined by the van Genuchten parameters α and β, was kept constant (as opposed to the hydraulic conductivity which is varied around the base case). No retention curve needed to be defined for the clogging layer because it remained saturated for the entire simulation due to perennial river conditions (i.e., saturation of the clogging layer occurs when the hydraulic conductivity of the clogging layer is less than the hydraulic conductivity of the aquifer).

Equation 4.1 was used to determine the critical watertable beneath the centre of the river that defines the border between transition and disconnection of surface water and groundwater. In the base case, this critical watertable was at 18.68 m above the base of the model (reference datum = 0 m), or 1.32 m below base of the clogging layer (i.e., watertable depth below a disconnected infiltration zone) for the physical parameters used.

Evapotranspiration was dynamically simulated as a combination of evaporation (Equation 4.8) and transpiration (Equation 4.2) processes by removing water from all model cells of the surface and subsurface flow domains within the defined zone of the evaporation and root extinction depths. To simulate evaporation only, the transpiration process was shut down by changing the root extinction depth and LAI to zero (i.e., from the last day of the first ten year period [day 3652] to the next day of the following ten year period [day 3653] there was no transpiration in the base case). The daily reference potential evapotranspiration (E_p) rate (Equation 4.2 and 4.8) was based upon the historical average daily reference potential evapotranspiration data for the Rocky River catchment (Figure 4.1) and transpiration parameters, typical of native vegetation in southern Australia (Table 4.1). The evapotranspiration processes were simulated for the entire model duration beginning on day 1. For example the E_p for the first day in the simulation is 0.0065 m and represents the average daily potential reference evapotranspiration for 1 January from the weather station from over 30 years of historical data. The E_p value for each day of the year is then repeated again for the next year and so on for the entire model duration.

Precipitation was simulated for the entire model duration beginning on day 1. The daily precipitation values used in the model were based on the historical daily average precipitation for the Rocky River catchment. Precipitation was not simulated over the river channel because the river was set with a constant hydraulic head boundary condition and therefore the additional water on top of the constant head boundary would cause erroneous results in the surface water domain. The stage height elevation of the river was set to 21 m, corresponding to a surface water depth of 0.5 m to provide an unlimited source of water to the aquifer system (i.e., perennial river), to ensure the clogging layer remained saturated and to isolate the effects of evapotranspiration on the state of connection. Surface runoff can only occur if the watertable rises above the ground surface and exceeds the assigned rill storage height. The rill storage was sufficiently high (0.01 m) to prevent overland flow.

The aquifer was composed of an isotropic homogeneous sand with a saturated hydraulic conductivity of 1.0 m d^{-1} and van Genuchten (van Genuchten, 1980) parameters (α = 4.0, β = 1.4; (Carsel and Parrish, 1988)) (Table 4.1). The clogging layer beneath the river had a saturated hydraulic conductivity of 0.005 m d^{-1}. The extinction depths for evaporation and transpiration were 1 m and 5 m, respectively, with both processes modelled over these depths using a quadratic decay function. Transpiration extinction depth (root depth) varies largely between different types of vegetation, ranging from over 60 metres in the case of *Boscia albitrunca* and *Acacia erioloba* found in the central Kalahari, Botswana to less than half a metre for shallow rooted cereal crops (Canadell et al., 1996; Schenk and Jackson, 2002; Shah et al., 2007). The extinction depth chosen for the base case scenario (5 m) was selected to represent an average extinction depth of native vegetation that is commonly found in Australia (e.g., eucalyptus and acacia species) (Robinson et al., 2006; Stone and Kalisz, 1991). The leaf area index (LAI) used to describe the transpiration function was set to 1.5 and is typical of native vegetation in southern Australia (Ellis and Hatton, 2008). The limiting saturation constants for evaporation and transpiration and other model parameters are shown in Table 4.1.

Table 4.1 Notation, units and selected model parameters. [a]The values are modified for different simulations.

Symbol	Description of conceptual model[a]	Units
d	depth of river with constant head	m
h_a	thickness of saturated/ unsaturated aquifer at model boundary	m
h_r	thickness of saturated/ unsaturated aquifer at x = 30	m
h_c	thickness of clogging layer	m
K	hydraulic conductivity	m d^{-1}
K_a	hydraulic conductivity of aquifer	m d^{-1}
K_c	hydraulic conductivity of clogging layer	m d^{-1}
Q	vertical flow rate through the clogging layer	m d^{-1}
L	distance to lateral model boundary from centre of river	m
w	width of river from centre of river to river edge	m
w_c	width of clogging layer at river edge	m

Model parameters (fixed for all simulations)	Value	Units
Maximum time step	1	days
Minimum time step	0.1	days
Sand porosity	0.25	
Clay porosity	0.38	
Specific storage sand	0.001	
Specific storage clay	0.01	
Residual water content sand (θ)	0.04	
Residual water content clay (θ)	0.04	
Van Genuchten alpha (α) for sand and are defined by van Genuchten (1980)	4.0	m^{-1}
Van Genuchten beta (β) for sand and are defined by van Genuchten (1980)	1.4	
Reference potential evapotranspiration (E_p)- daily data taken from Rocky River weather station		m
Tree canopy evaporation (E_{can})	0	m
Evaporation extinction depth defined by a quadratic decay evaporation distribution function (*EDF*)	1.0	m
Evaporation limiting saturation sand (min)	0.05	
Evaporation limiting saturation clay (min)	0.25	
Evaporation limiting saturation sand (max)	0.9	
Evaporation limiting saturation clay (max)	0.9	
Transpiration extinction depth defined by a quadratic decay root distribution function (*RDF*)	5.0	m
Leaf area index (LAI)	1.5	$m^2\ m^{-2}$
Transpiration fitting parameter (c1)	0.6	
Transpiration fitting parameter (c2)	0.0	
Transpiration fitting parameter (c3)	1.0	
Transpiration limiting saturation (wilting point)	0.05	
Transpiration limiting saturation (field capacity)	0.1	
Transpiration limiting saturation (oxic limit)	0.8	
Transpiration limiting saturation (anoxic limit)	0.95	
Rill storage height	0.01	m

4.3 Results and Discussion

4.3.1 Base Case

Here we describe the base case model and the different scenarios using a subset of representative and realistic hydrogeological, hydroclimatic, and vegetation variables (Table 4.2).

Table 4.2 Hydrogeological, evaporation and vegetation variables used in the different model scenarios.

Simulation	Hydraulic conductivity aquifer (K_a)(m d^{-1})	Hydraulic conductivity aquifer (K_c)(m d^{-1})	Catchment slope (-)	Evaporation extinction depth (m)	Transpiration extinction depth (m)
Figure 4.3	1	0.005	0.01	1	5
Figure 4.4-a	1	0.005	0.02	1	5
Figure 4.4-b	1	0.005	0.08	1	5
Figure 4.4-c	1	0.005	0.32	1	5
Figure 4.5-a	4	0.005	0.01	1	5
Figure 4.5-b	10	0.005	0.01	1	5
Figure 4.5-c	1	0.0005	0.01	1	5
Figure 4.6	1	0.005	0.01	1	1, 2, 7

Figure 4.3 shows the hydraulic head versus time at an observation point at 12 metres elevation located directly beneath the centre of the river for the transient principal model (K_a = 1, K_c = 0.005 and slope = 0.01 [1 cm m^{-1}]). The dashed horizontal line shown in Figure 4.3 (and subsequent figures) describes the position of the water table (i.e., zero pressure head) directly beneath the centre of the river at disconnection between the river and the aquifer, and is also the point of maximum flux calculated using equation 4.1. A total head at an observation point beneath the centre of the river and above this horizontal line indicates a system in transition between a connected and disconnected regime. A total head at an observation point beneath the centre of the river below this horizontal line is a system that is disconnected and can also be described as a system with a deep water table.

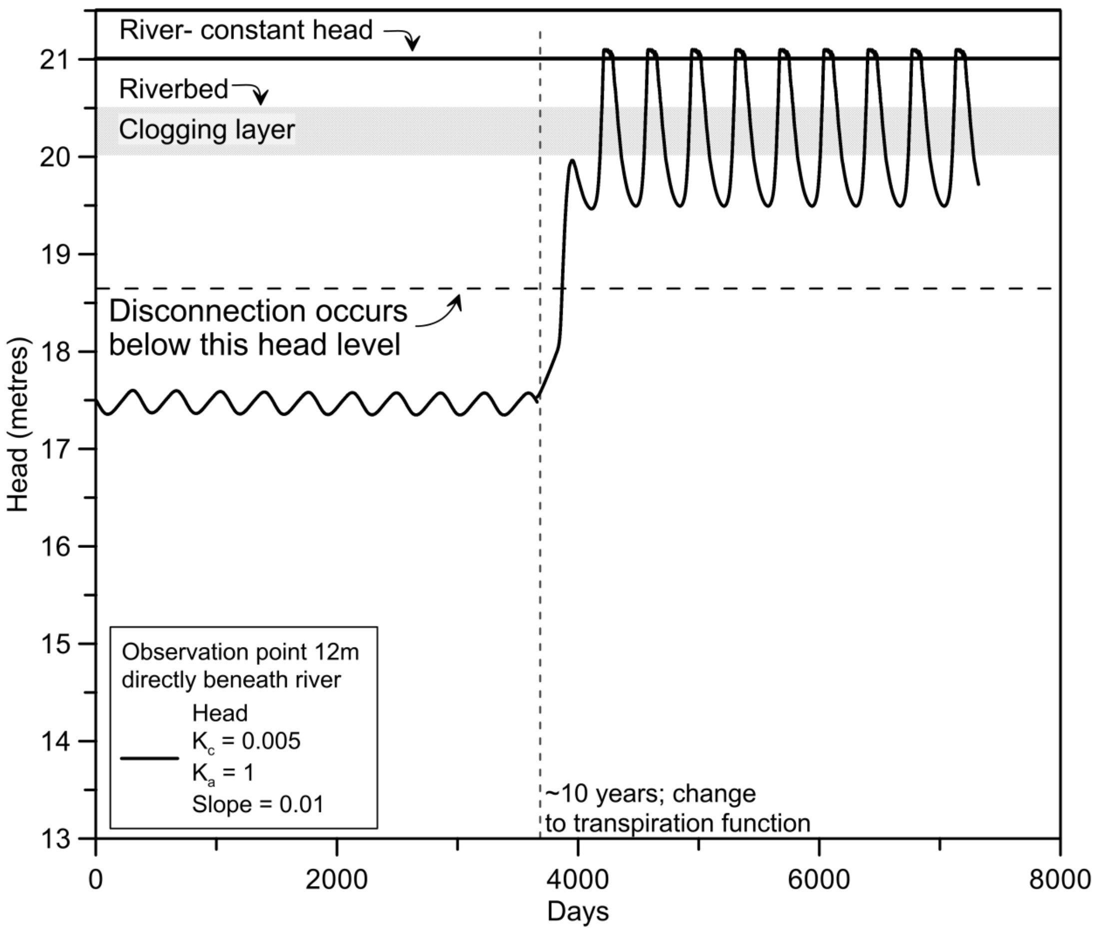

Figure 4.3 Hydraulic head at an observation point at 12 m elevation directly beneath the centre of the river (x = 30 m) for the transient model. The catchment slope of the 2-D model is 0.01. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1}.

As mentioned in section 4.2.3, the calculated hydraulic head at this observation point is a very close approximation of the watertable of the aquifer. The results showed that the model reached a quasi-steady state between precipitation and evapotranspiration over the first 10 year period and the river and aquifer were in a losing disconnected type regime. The results also showed that the presence of vegetation, through evapotranspiration, was able to cause and maintain an unsaturated zone beneath the river. It is worth noting that the aquifer was still responsive to the seasonal fluctuations in precipitation and evapotranspiration (Figure 4.3) with higher head levels in winter and lower head levels in summer. At the end of the 10 year period, transpiration was set to zero (i.e., the trees were removed) and the state of connection of the model changed from a losing disconnected type system to one that was connected and seasonally gaining and losing. Recall that evaporation was still simulated after transpiration was turned off. The hydraulic head fluctuated by approximately 1.5 metres with a maximum at winter which is associated with gaining type conditions and a minimum at summer which is associated with losing type conditions. The change from disconnected to connected status was quite rapid with the water table rising several meters in a matter of days after the removal of vegetation.

4.3.2 SCENARIOS

The model's sensitivity to the parameters of catchment slope, the hydraulic conductivity of the aquifer and the clogging layer, and the transpiration function was examined using several different model scenarios (Table 4.2) which are shown in Figure 4.4, Figure 4.5 and Figure 4.6.

To explore the impact of catchment slope, the slope gradient of the base case scenario was increased from 0.01 to 0.02 to 0.08 and up to 0.32 (Figure 4.4). Increasing the catchment slope increased the thickness of the vadose zone away from the river and limited the availability of soil moisture. Hence, the amount of water removed by evaporation and transpiration was limited to the functions' prescribed extinction depths of 1 and 5 metres, respectively. In the model with a catchment slope of 0.08 (Figure 4.4), the river and aquifer were connected under a vegetated catchment regime and there were gaining and losing connected type conditions in response to the seasonal variation in precipitation and evapotranspiration. Once the transpiration was shut down after ten years (by setting the root extinction depth and LAI to zero), the watertable was much closer to the ground surface but there were still seasonal gaining and losing connected type conditions. Figure 4.4, with a catchment slope of 0.32, shows that the river and the aquifer were connected. However, the river was losing under a vegetated catchment regime and transitioned to gaining and losing type conditions when transpiration was set to zero. It is also worth noting that the minimum and maximum values in the hydraulic head levels between the seasons were not as great as those observed in the scenario with a catchment slope of 0.08. The increased slope resulted in a wider vertical extent of the unsaturated zone and therefore changed the response in evaporation and transpiration which led to a dampened seasonal response in the hydraulic head (Figure 4.4).

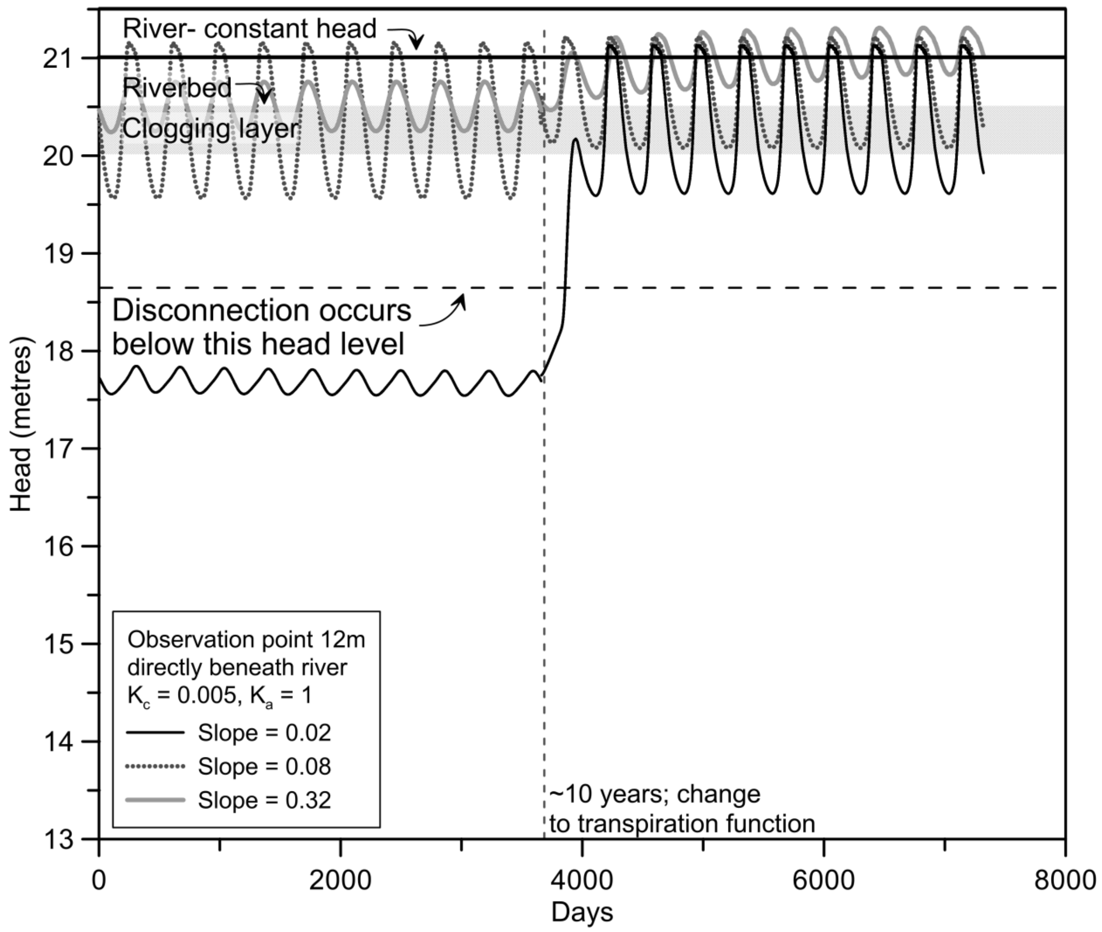

Figure 4.4 Hydraulic head at an observation point at 12 m elevation directly beneath the centre of the river (x = 30 m) showing the sensitivity of the catchment slope on the transient 2-D model. Three model scenarios are shown with different values for the catchment slope. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1} for all three models.

To examine the system's response to hydraulic conductivity, the conductivity of the aquifer and the clogging layer were varied by several orders of magnitude and were shown to have a significant effect on the state of connection between the river and the aquifer (Figure 4.5). When the conductivity of the clogging layer was kept constant at 0.005 m d^{-1} and the conductivity of the aquifer was modified from 1 to 4 and up to 10 m d^{-1}, the extent that the evaporation and transpiration functions lowered the watertable of the aquifer was increased. Under vegetated conditions (first ten year period) the increase in the hydraulic conductivity of the aquifer still resulted in disconnected conditions. Once transpiration was shut down (set to zero) after the ten year period the river and aquifer became connected again, however, in the scenario where the hydraulic conductivity of the aquifer was 10 m d^{-}

[1] there were periods during the summer when evaporation was high and the river and the aquifer became disconnected (Figure 4.5).

When the hydraulic conductivity of the aquifer was kept constant at 1 m d^{-1} and the conductivity of the clogging layer was modified from 0.005 m d^{-1} (base case scenario) to 0.0005 m d^{-1} there was a significant effect on the state of connection between the river and the aquifer. In the scenario where the clogging layer conductivity was 0.0005 m d^{-1} the initial condition of the watertable was considerably low and once the transpiration was set to zero after the ten year period it took a considerable amount of time for the river and aquifer to become connected (Figure 4.5). The model simulation time was extended for this scenario to show that the model reached a quasi-steady state again. In comparison, when the hydraulic conductivity of the clogging layer was increased above the value used in the base case scenario (0.005 m d^{-1}) the river and aquifer remained connected.

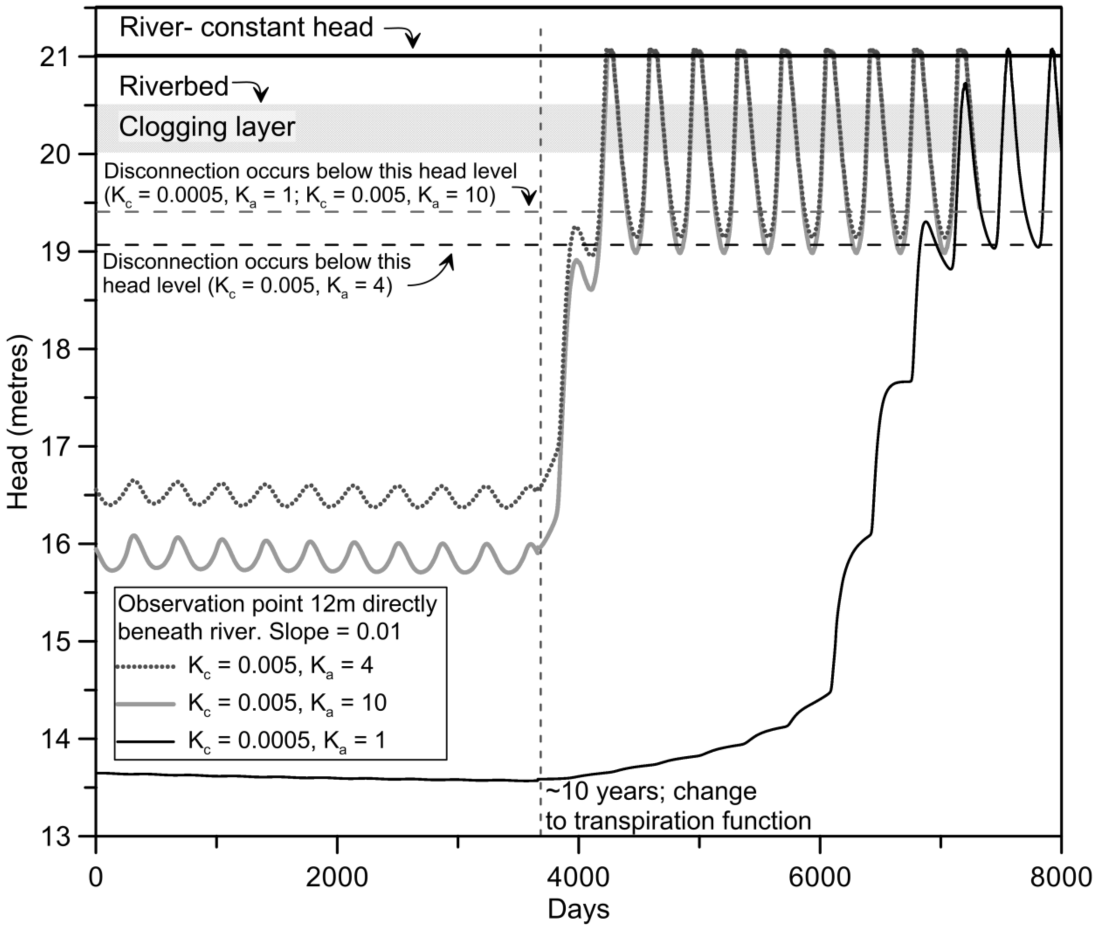

Figure 4.5 Hydraulic head at an observation point at 12 m elevation directly beneath the centre of the river (x = 30 m) showing the sensitivity of the hydraulic conductivity (m d^{-1}) of the aquifer (K_a) and clogging layer (K_c) on the transient 2-D model. Three model scenarios are shown with different values for K_a and K_c. All three models have a catchment slope gradient of 0.01.

To explore the effect of transpiration on causing an unsaturated zone to develop underneath the river bed and alter the state of connection in the base case scenario, the transpiration extinction depth was modified from zero (i.e., no transpiration) to 1 metre, 2 metres and 7 metres depth without changing the transpiration function values and LAI (Figure 4.6). The first ten year period of the simulation was with evapotranspiration as simulated in the base case scenario (solid dashed line shows transpiration extinction depth 5 m). After the ten year period different transpiration extinction depths were simulated for another ten years (shown by four different line weights in Figure 4.6). The results showed that the greater the extinction depth (i.e., greater depth of the plant root zone) the more water was removed from the aquifer and the more likely that the river and the aquifer would transition to a disconnected type system. The change from a disconnected to connected system (defined by the hydraulic head level at the observation point crossing the determined line of disconnection) was rapid when there was no transpiration (less than 200 days), while changing to a shallower extinction depth there was a time lag (of at least 600 days) before the system reached a new quasi-steady state.

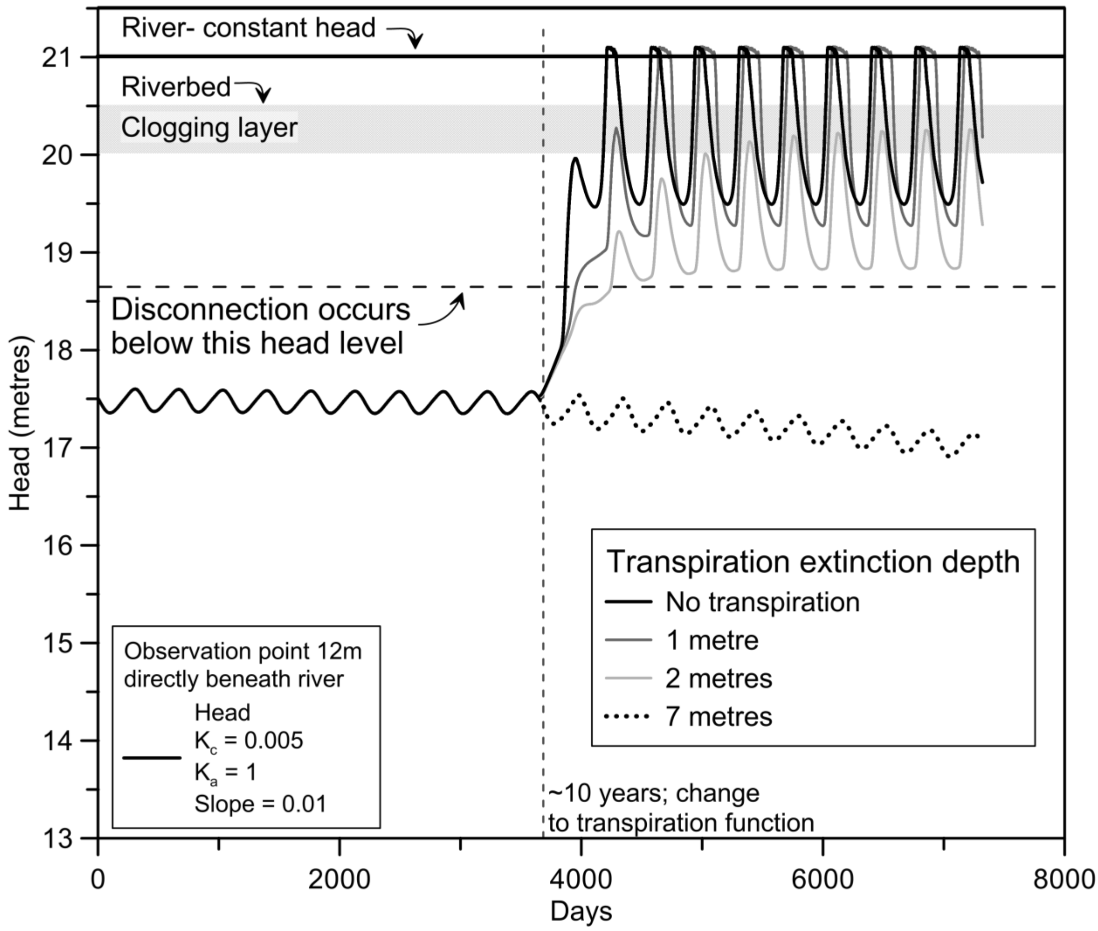

Figure 4.6 Hydraulic head at an observation point at 12 m elevation directly beneath the centre of the river (x = 30 m) showing the sensitivity of the transpiration extinction depth function on the state of connection between surface water and groundwater. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1}, and the catchment slope gradient is 0.01 for all models.

4.3.3 EFFECTS OF EVAPOTRANSPIRATION ON THE PRESENCE OF AN UNSATURATED ZONE AND THE STATE OF CONNECTION

The modelling presented in this study has shown that evapotranspiration can cause and maintain an unsaturated zone between a perennial river and aquifer system and in some cases a state of disconnection. Removing native deep rooted vegetation and replacing it with shallow rooted vegetation (i.e., modification of the transpiration function) can have a substantial effect on the state of connection and is more likely to change from a disconnected to a connected type system. While the present work attempted to evaluate the effects of evapotranspiration on the development of an unsaturated zone and the state of connection, we only addressed a few of the possible scenarios that may be observed in nature. For example, the temporal and spatial dynamics of ephemeral and intermittent rivers and their contrasting wetting and drying cycles compared to perennial rivers are also likely to have a significant impact on the infiltration flux from the river to the aquifer beneath and the state of connection (Hatch et al., 2010; Niswonger et al., 2008). In our study, a constant head in the river was used in the conceptual model to maintain 100 % saturation in the clogging/stream bed layer to reduce system complexity so that we could accurately test our hypothesis.

Ultimately, the processes of evaporation and transpiration were restricted to the prescribed extinction depths in the model domain which was investigated in the scenario models shown in Figure 4.6. Therefore, the steepness of the catchment slope had a considerable effect on the amount of water that could be removed by evaporation and transpiration processes due to the thickness of the vadose zone increasing with greater distance away from the river. In the base case scenario, the slope of the catchment was small (1 cm m^{-1}) and therefore there was minimal influence of the slope on the evaporation and transpiration functions. Increasing the slope increased the thickness of the vadose zone and decreased the depth of available soil moisture to transpiration and evaporation processes. It is worth noting that in most real systems the transpiration capacity of vegetation communities would also change along the slope in response to the available soil moisture and this has not been addressed in our study. However, the analysis of the catchment slope sensitivity does provide some insight as to where the greatest changes in the state of connection may occur in different types of catchment settings in response to a change in vegetation. For example, there would be a greater impact in catchments that are flat compared to ones that are steep.

In the different model scenarios described here, the initial watertable elevation of the aquifer was important in influencing the state of connection. In model scenarios where the initial watertable elevation (located at the observation point directly beneath the centre of the river) was well below the bottom of the clogging layer and the river, there was complete capacity of the evaporation and transpiration functions to remove water. In comparison, when the initial watertable elevation was relatively shallow and close to the ground surface, the transpiration function was severely limited by complete saturation of the vegetation root zone.

The effect of evapotranspiration on the development of an unsaturated zone beneath a riverbed and the state of connection depended largely on the hydraulic conductivity of the clogging layer beneath the river being less than the hydraulic conductivity of the aquifer. When the hydraulic conductivity of the clogging layer was large, the river continuously replenished the aquifer and no rate of evapotranspiration could induce an unsaturated zone below the clogging layer. In comparison, when a smaller conductance value was used, the processes of evapotranspiration were able to create an unsaturated zone below the clogging layer and in some cases resulted in disconnected type conditions. This illustrates the complex interplay between the various controlling variables and processes. The analysis of the described conceptual model was for a homogeneous system with homogeneous hydraulic conductivities. Simplifying the system complexity was necessary in order to remove any of the confounding effects (i.e., heterogeneity within the clogging layer and aquifer) to ensure clarity of the specific hypothesis that was being examined. It is worth noting that heterogeneity within the clogging layer and aquifer can be an important control on river seepage temporally (in response to streambed scouring) as well as its spatial distribution along the channel (Fleckenstein et al., 2006; Hatch et al., 2010; Niswonger and Fogg, 2008). Frei et al. (2009) also noted that spatial and temporal heterogeneity within alluvial sediments can cause distinct patterns and dynamics of river seepage in rivers overlying a deep watertable (i.e., disconnected systems), and that most seepage occurs along preferential flow zones.

All of the model scenarios described in this study have used the historical average precipitation and reference potential evapotranspiration data for the Rocky River Catchment. According to Figure 4.1 there is a period from mid-April until the end of August where precipitation is greater than evapotranspiration which would be a time when groundwater recharge could occur. The seasonal trends in precipitation and evapotranspiration in Figure 4.1 are representative of hydroclimatic conditions in many parts of southern Australia. The simulations showed that there was a strong seasonal response in the aquifer to changes in precipitation and evapotranspiration, and it was only when the watertable was at a considerable depth (~6.5 m) below the riverbed that the seasonal response was not observed (Figure 4.5; K_c = 0.0005). The seasonal variations between summer (hydraulic minimum) and winter (hydraulic maximum) were more pronounced after the removal of vegetation when the watertable was closer to the ground surface as a result of increased recharge to the aquifer.

4.3.4 Trees as Groundwater Pumps

The analogy of trees as groundwater pumps and their potential to influence the exchange fluxes between surface water and groundwater has been well established (Butler et al., 2007; Loheide et al., 2005). However, so far it has not been demonstrated in a quantitative and systematic way if evapotranspiration can cause an unsaturated zone to develop underneath a riverbed and in some instances cause a disconnection. The results of this study support the earlier hypothesis proposed by Banks et al. (2011b), which suggested that the river system in the Rocky River catchment maybe fresher (less saline) than rivers in adjacent catchments in apparently similar geologic and climatic settings. The low salinity in the river is maintained by virtue of the fact that the Rocky River catchment is pristine and covered by native vegetation which creates losing conditions, while the others are cleared and are likely to be gaining systems. The vegetation controls are a plausible explanation for different state of connection and poorer water quality of the adjacent cleared catchments which were studied previously by Henschke et al. (2003) and Shand et al. (2007c).

According to the water balance of the Rocky River Catchment, the annual precipitation input volume is 147.4 x 10^6 m^3 yr^{-1} (based on precipitation of 780mm yr^{-1} and gauged catchment area of 189 km^2) and the annual streamflow discharge is 1.4 x 10^6 m^3 yr^{-1}. Therefore, 146 x 10^6 m^3 yr^{-1} of the catchments' precipitation is lost to evapotranspiration and/or groundwater recharge (assuming that streamflow represents surface runoff only). It can be assumed that there is zero to very little groundwater recharge beneath native vegetation (Allison and Hughes, 1983; Leaney and Allison, 1986), and therefore 0.77 x 10^6 m^3 per square kilometre would be removed via transpiration from the catchment water balance, which would equate to approximately 53,986 trees per km^2 (using the average water use of a eucalyptus tree equal to 14.7 m^3 per tree per year (Farrington et al., 1994). The high density (and therefore high evapotranspiration) has the potential to maintain the lower elevation of the watertable beneath the river and hence a disconnected type system. In comparison, in the adjacent cleared catchments to the pristine Rocky River Catchment, a decrease in evapotranspiration through the removal of native vegetation and replacement with shallow rooted vegetation has evidently resulted in a change to a connected type system and salinisation of the water resource as a result of increased recharge and a rising watertable. Our results also showed that with the removal of vegetation and a rising watertable the seasonal variations between summer (hydraulic minimum) and winter (hydraulic maximum) were more pronounced.

4.4 CONCLUSION

By using a simple conceptual model based on realistic and representative parameter values, we have demonstrated that the presence of vegetation is a plausible mechanism for causing an unsaturated zone to develop between a perennial river and an aquifer. Vegetation can therefore also affect the state of connection between surface water and groundwater and in some instances create a disconnection. This may appear intuitively plausible in a qualitative sense; however, it has not been demonstrated quantitatively. Our study therefore suggests that in addition to the well known influences of physical variables such as hydraulic conductivity or topography, the effects of vegetation need to be carefully considered when investigating surface water–groundwater interactions. By examining different conceptual models of catchments with different slopes and vegetation type (i.e., root depth) we provided insights into the conditions where changes to vegetation can affect the flow regime and the presence of an unsaturated zone. Our analysis showed that the flow regime and hydraulic response to the presence of vegetation and subsequent removal can be much greater in flatter catchments than those that are steep.

Given the importance of vegetation on surface water–groundwater interactions, changes in vegetation can have considerable consequences to shifting the state of connection. Such changes can be associated with land clearance, revegetation or climate change. In catchments in southern Australia where the aquifer systems are often saline or there is a significant salt stored in the unsaturated zone, changing from a losing disconnected to gaining connected type system results in serious water quality issues because saline groundwater discharges to the surface water system. In the longer-term, the change in vegetation in pristine catchments can lead to the salinisation of the surface water resource. This link between pristine and cleared catchments, and the resulting state of connection or disconnection, may be important in explaining differences in observed river water quality between catchments in similar geographic, geologic and climatic locations. The results of this current study also appear to support the hypothesis raised in the earlier Banks et al. (2011b) study that land clearance may be the key factor to maintain low levels of salinity in the Rocky River, while the other rivers in adjacent catchments are more saline.

The purpose of this study was to demonstrate that vegetation can (under reasonable and representative conditions) create an unsaturated zone and therefore affect the state of connection. Even though we did not develop a generalised theoretical framework on the effects of evapotranspiration on the state of connection between surface water and groundwater, the findings of this study are likely to be applicable to other catchments where land clearance or revegetation occurred or is occurring.

Further work carried out in 3-D would be useful to explore the effects of evapotranspiration on the state of connection along the length of the river (compared to the cross section of the river as discussed in this study) and how the interface between connected and disconnected regimes migrates up and down the river channel in both space and time as a function of vegetation clearance or land use change. 3-D analysis could also provide insight into the influence of more natural meandering river geometries and how this may affect the convergence and divergence of groundwater flow paths near the river. Additional simulations might explore the effect of different vegetation types throughout catchments (e.g., vegetation with increasing extinction depth up slope) but the basic outcomes of this study are not expected to be significantly altered in those more complex cases.

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APPENDIX A

6. PUBLISHED CONFERENCE PROCEEDINGS

6.1 HYDROGEOCHEMICAL INVESTIGATIONS OF INTERACTIONS BETWEEN GROUNDWATER AND SURFACE WATER IN A FRACTURED ROCK ENVIRONMENT, MOUNT LOFTY RANGES, SOUTH AUSTRALIA

PRESENTED AT WATER DOWN UNDER CONFERENCE, ADELAIDE, SOUTH AUSTRALIA, APRIL 2008

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ABSTRACT: The management and allocation of natural water resources requires an understanding of the connections between groundwater and surface water systems to ensure provisions are made for environmental requirements and to prevent the double accounting of connected water resources. Evaluating fluxes between surface water and groundwater in fractured rock aquifers presents more difficulty than in sedimentary aquifers. Groundwater flow paths in fractured rock are difficult to determine and methods commonly used for porous media are not applicable. However, the analysis of hydrogeochemical indicators in a number of groundwater and surface water samples can be used to determine the relative proportions of groundwater discharge and surface runoff present in surface water systems and the likely source aquifer(s) of the groundwater. The catchments of the Eastern and Western Mount Lofty Ranges fall under separate water management jurisdictions and were treated as two independent study areas. Investigations were undertaken using two different approaches to address the large spatial scale of the two areas. For the Western Mount Lofty Ranges three representative sub-catchments with a combined area of 95 km^2 were selected, allowing a relatively high density of sample points.

For the Eastern Mount Lofty Ranges, sample points were distributed across 6 catchments covering the whole 1600 km^2 study area. The former enabled detailed 'run-of-river' analyses, which revealed how water quality and groundwater inflow rates vary along the length of a creek during different hydrological events. Variations in hydrochemical characteristics showed strong correlations to the creek's underlying geology and presence of major fault zones. The larger scale approach employed in the Eastern Mount Lofty Ranges enabled differing hydrochemical characteristics to be attributed to different geology types and provided estimates of groundwater contributions to surface water at individual points spread across the entire study area. However, there was insufficient data density to describe in detail the interactions between surface water and groundwater within each catchment.

INTRODUCTION

Traditionally, surface water and groundwater have been, and generally continue to be, managed as separate resources in Australia. It is well understood that in some cases there are strong hydraulic connections between these two resources, which is spatially and temporally variable (Brodie et al., 2007; Kalbus et al., 2006; Winter et al., 1998). An understanding of these interactions is of particular importance in water allocation planning if double accounting of resource volumes and subsequent over-allocation of water are to be avoided.

In areas of steep topographic relief and dry temperate climate such as the Mount Lofty Ranges (MLR), groundwater discharge into streams is a significant component to the catchment water balance, particularly during the summer months and drought conditions. Many of the surface water features are perennial, dependent on groundwater, and support a diverse range of flora and fauna.

The hydrogeology of the MLR is dominated by fractured rock aquifer (FRA) systems with minor alluvial aquifers in the valley bottoms. Interactions between groundwater and surface water in fractured rock environments are complex and difficult to determine at a catchment scale. Hydraulic connections are related to intersections between the underlying fracture network and streambed. Methods commonly used for porous media are not applicable (Cook et al., 1996; Love et al., 2002; Oxtobee and Novakowski, 2003). However, multi-tracer hydrochemical approaches have recently been used in fractured rock environments (Shand et al., 2007a; Van der Hoven et al., 2005) and have proven valuable in constraining contributing sources of solutes, preferential flow pathways and residence times.

This study describes the results of investigations of the interactions between groundwater and surface water in the Western Mount Lofty Ranges (WMLR) and Eastern Mount Lofty Ranges (EMLR) catchment areas using two different approaches to address the large spatial scale of the study areas. Investigations were undertaken in the Cox, Lenswood and Kersbrook Creek Catchments in the WMLR and the Angas, Bremer and Finniss River catchments in the EMLR. This paper only presents results from the Cox Creek Catchment and the Bremer River Catchment.

STUDY CATCHMENTS

The Cox Creek Catchment (CCC) is situated approximately 20 km east of Adelaide in the Western Mount Lofty Ranges (Figure) and has a catchment area of 28.8 km^2. Surface drainage is from the higher northern boundary of the catchment (approx. 630 mAHD), through steep topography to the southeast where it discharges into the Onkaparinga River at approximately 320 mAHD. The headwater tributaries of Cox Creek converge less than a kilometre upstream of a streamflow gauging station, which measures flow only from the upper 4.3 km^2 of the catchment. Average annual rainfall is 1189 mm/y, the majority of which falls between May and October.

The Bremer River Catchment (BRC) is approximately 23 km to the southeast of the CCC in the Eastern Mount Lofty Ranges (Figure 2). The catchment covers a total area of approximately 590 km^2. Topographically, the BRC is highest along the northwestern margin, with a maximum elevation of approximately 480 mAHD. Surface drainage is dominated by the Bremer River, which runs from north to south along the eastern side of the catchment and discharges into Lake Alexandrina at approximately mean sea level. Average annual rainfall ranges from 766 mm/y at Mount Barker in the western part of the catchment, to 383 mm/y at Hartley on the lowland plain in the southeast.

The geology of the CCC and the BRC includes several of the stratigraphic sequence associated with the Adelaide Geosyncline. In the CCC, the Neoproterozoic Emeroo Subgroup dominates the north and south of the catchment and is separated by the Archean Barossa Complex, which lies across the middle of the catchment (Figure 1). Major fault lines are present along the margins of the different geological units, traversing in a northeast-southwest direction. The Emeroo Subgroup consists of quartzite, sandstone and dolomite, whilst the Barossa Complex is characterised by metamorphic rocks with retrograde metamorphism. The majority of wells are completed in the Emeroo and Mundallio Subgroup, which typically have high yields and good quality water compared to aquifers in the Barossa Complex.

In the BRC the Neoproterozoic Burra Group, Umberatana Group and Wilpena Group occur in the west, while the eastern half of the catchment is predominantly underlain by the metasediments of the Cambrian Kanmantoo Group. The lower, eastern and southeastern parts of the catchment are characterised by unconsolidated Quaternary clays, sands and gravel deposits typical of the lower Murray Basin overlying the basement of Kanmantoo Group metasediments. Groundwater in the majority of the BRC exists primarily in fractured rock aquifers of the Kanmantoo Group. The groundwater potentiometric surface elevation tends to correlate with land surface elevation and is highest in the northwest area of the catchment, and lowest in the southeast, such that groundwater flow throughout the majority of the BRC is in a south-easterly direction.

METHODOLOGY

Surface water sampling in the CCC was conducted during four sampling rounds: round 1 (7-9 December 2005), round 2 (6-7 March 2006), round 3 (26-27 July 2006) and round 4 (31 October- 1 November 2006). Groundwater sampling was conducted during round 1 and round 3. In the BRC three rounds of surface water sampling were conducted, spaced at approximately four month intervals: round 1, in September 2006, round 2, in January 2007, and round 3, in May 2007. Two rounds of groundwater sampling were conducted, in November 2006 and May 2007.

A YSI® multi-parameter meter was used to measure the pH, specific electrical conductivity (SEC), dissolved oxygen (DO), redox potential and temperature in the creek and also during purging of the groundwater bores. The total alkalinity (as HCO_3^-) was also measured in the field using a titration kit. Major ion samples were filtered in the field using a 0.45µ membrane filter. Samples for major cations and trace element samples were preserved with nitric acid (1% v/v HNO_3) and analysed by Inductively Coupled Plasma Emission Spectrometry (ICP-ES). Major anions were analysed by Ion Chromatography (IC).

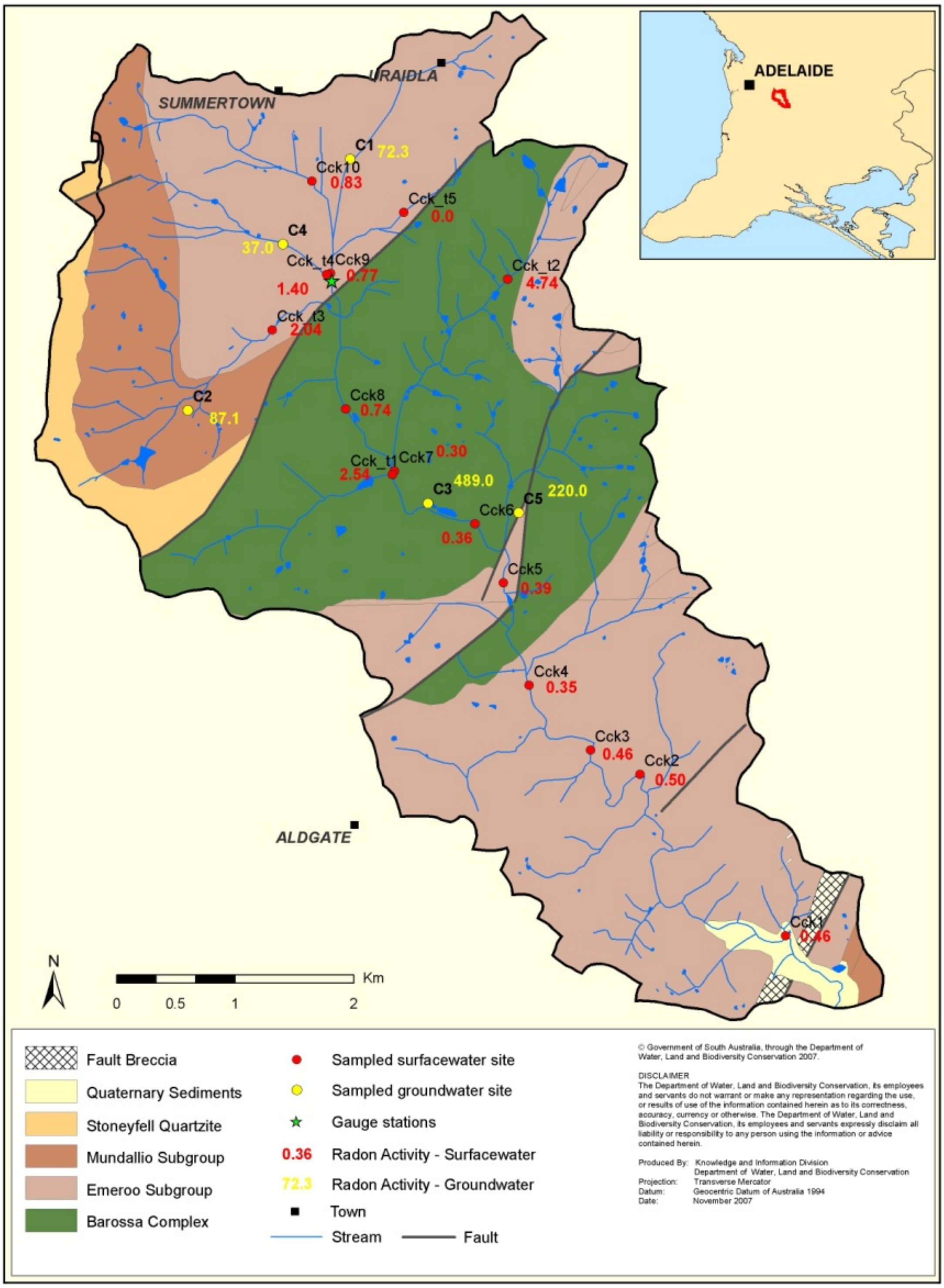

Figure 1. Cox Creek Catchment surface geology, spatial distribution of ^{222}Rn activities (Bq/L) of surface water at the sample locations, and groundwater wells in round 1 (December 2005).

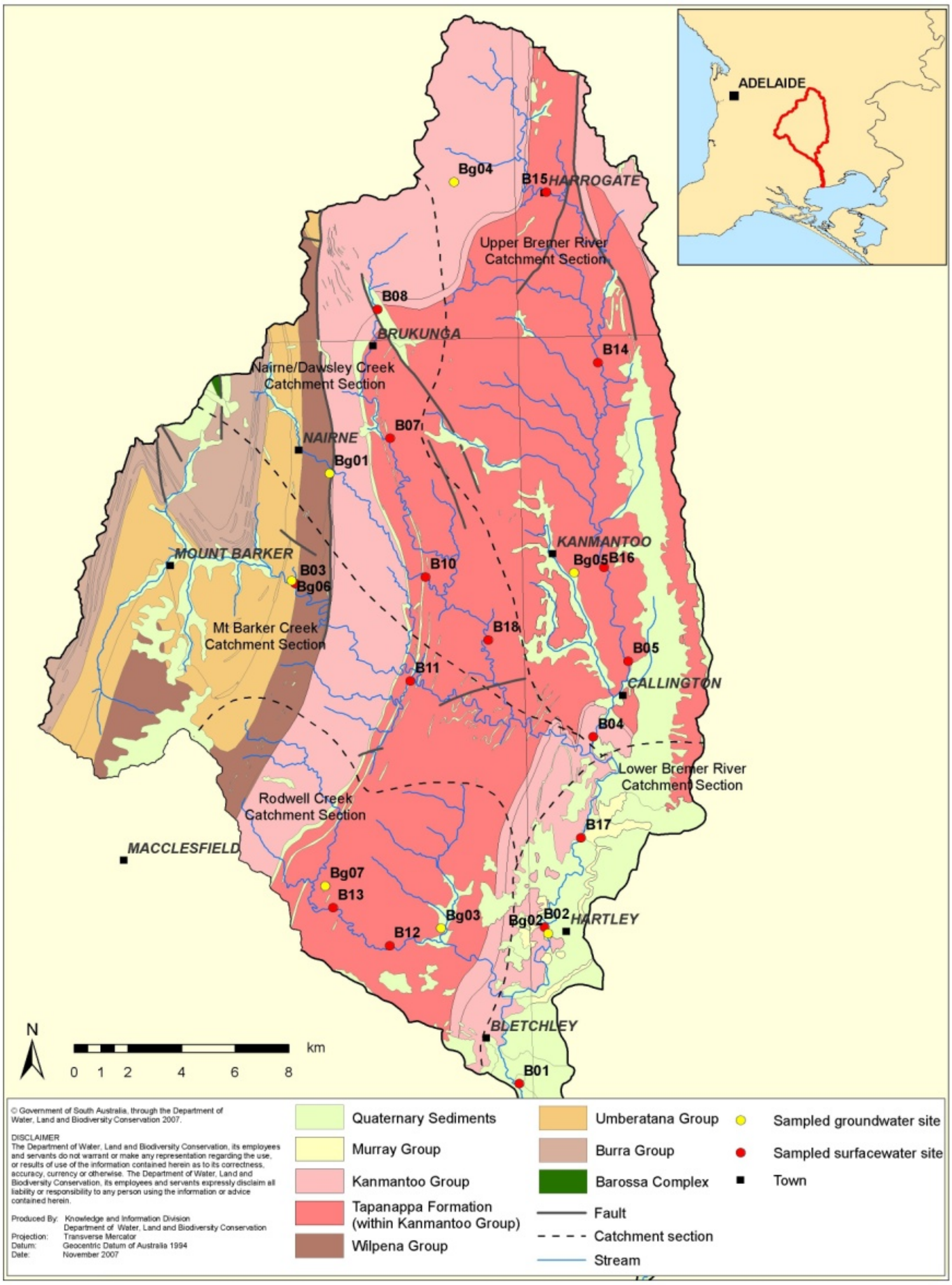

Figure 2. Bremer River Catchment surface geology and location of surface and ground water sample locations. Dashed lines show division of catchment into sections as described in the text.

Samples were analysed for the stable isotopes of the water molecule at the CSIRO Land and Water Isotope Analysis Service in Adelaide. These are expressed in delta notation as δ^2H and δ^{18}O, normalized to Vienna Standard Mean Ocean Water (vs. VSMOW). Adelaide is the closest rainfall station to the MLR with rainfall isotopic data provided by the International Atomic Energy Agency (IAEA) Global Network of Isotopes in Precipitation (GNIP) service.

Samples for analysis of radon-222 (^{222}Rn) activity and the radiogenic isotopes of strontium were also collected. Groundwater samples for ^{222}Rn analysis were collected directly from the pump outlet of the purged well using a syringe. A sample of 14 mL was transferred to a pre-weighed 22 mL Teflon-coated PTFE vial with 6 mL Packard NEN mineral oil scintillant, gently shaken for 30 seconds and sealed. Surface water samples for the measurement of ^{222}Rn activity were collected using a rapid field extraction method developed by Leaney and Herczeg (2006). The samples were submitted to the Adelaide Isotope Laboratory within 3 days of sample collection and counted by liquid scintillation on a LKB Wallac Quantulus counter. Strontium isotope ratios ($^{87/86}$Sr) were analysed at the University of Adelaide using a Finnegan Mat 262 thermal ionisation mass spectrometer (TIMS). Strontium was extracted from filtered water samples by evaporating water to leave a solid precipitate, which was then re-dissolved in hydrochloric acid and filtered through columns of Biorad cation exchange resin to isolate $SrCl_2$.

Manual stream gaugings were carried out at each of the sampling locations at the different sampling rounds. A pigmy flow meter (OTT) was used to conduct the instantaneous gaugings. In some locations in the BRC it was clear from visual observations where water emerges from the ground or disappears into it and provided important information regarding the location and direction of groundwater exchange.

Results and interpretation

Cox Creek Catchment

Gauged streamflow data of Cox Creek shows that the mean annual flow is approximately 1180 ML/year (DWLBC, 2009). High flows usually occur from May to October with maximum flow observed around July and August. Manual measurements of stream flow conducted during the four rounds at the sampling locations indicated that changes in flow between locations were similar during the four rounds, with higher flows occurring during rounds 1 and 3 in response to a series of rainfall events prior to measurement. The flow at sites Cck7 and Cck3 were consistently higher than at other sites, suggesting that the creek may be gaining groundwater at these locations.

Hydrochemical and isotopic variations in Cox Creek

The variation in the water chemistry along Cox Creek is a result of different waters mixing from different flow pathways, geochemical reactions and residence times within the catchment. Significant changes in major and trace element concentrations along the length of the creek may indicate changes in water sources resulting from changes in the underlying geology, or the influence of fault zones that the creek traverses. The chloride concentrations ([Cl^-]) in Cox Creek (Figure(a)) were higher during round 2 (72-98 mg/L) and represent baseflow conditions when there was minimal rainfall. In comparison, [Cl^-] in round 3 (49-69 mg/L) are much lower as a result of dilution by low–chloride winter rainfall and surface runoff. The [Cl^-] shows a decreasing trend in the downstream direction from sample site Cck9 to Cck7, then gradually increasing from Cck7 to Cck1 in rounds 1, 3 and 4. The trend is similar in round 2 except that the [Cl^-] are higher and there is a noticeable increase in [Cl^-] between Cck8 and Cck4. A small tributary (Cck_t3) provides a low [Cl^-] inflow, which diluted the concentration in Cox Creek between Cck9 and Cck8 during all four rounds. There is also a low [Cl^-] contribution from the tributaries Cck_t1 and Cck_t2 in round 3, between locations Cck8 and Cck7, which significantly dilute the [Cl^-] in Cox Creek downstream to Cck7, after which they increase again.

^{222}Rn activities measured at the individual sites are higher in round 2 and round 4 compared to round 1 and 3, but all show similar trends along the length of Cox Creek (Figure 3 (b)). The high ^{222}Rn activity indicates localised groundwater influx to the stream. The ^{222}Rn activity downstream of the influx declines rapidly due to its short half-life (3.82 days) and loss to the atmosphere by gas exchange. The rate of gas loss is controlled by the gradient of stream, as well as volume of discharge, stream profile and streambed roughness (Ellins et al., 1990). ^{222}Rn activities remain fairly constant between sample sites Cck10 and Cck8 followed by a sharp decrease in activity between Cck8 to Cck7. Between Cck7 and Cck6 the activity is steady then gradually increases to Cck2 where it begins to decrease again to Cck1. During baseflow conditions (round 2) ^{222}Rn activities vary from 0.6 Bq/L at Cck7 to 3.4 Bq/L at Cck2 compared to winter (round 3) where the activity varies from 0.39 Bq/L and 0.43 Bq/L, respectively at the same locations. The high ^{222}Rn activities in the two tributaries sampled at Cck_t1 and Cck_t2 (2000 m upstream of confluence with Cox Creek) have minimal influence on the ^{222}Rn activity in Cox Creek, reflecting the small contribution of flow from these tributaries (Figure 1).

The ^{222}Rn activities in the groundwater are order of magnitude higher than in the surface water samples and reflect the mineralogy of the aquifer (Love et al., 2002). The variation in the groundwater ^{222}Rn activities in the CCC appears to be related to the two major rock types. Groundwater samples C1, C2 and C4 located in the sandstone, quartzite and dolomite of the Emeroo and Mundallio Subgroup have lower activities (37.9-87.1 Bq/L) compared to samples C3 and C5 located in the metamorphosed gneisses and schists of the Barossa Complex which have much higher activities (220-489 Bq/L). The similar trends in the ^{222}Rn activity data during different sample rounds implies that the connection between the groundwater system and Cox Creek exists throughout the year and that the differences in activity in Cox Creek between seasons are a result of dilution by rainfall and surface water runoff, which have low to zero ^{222}Rn activity. The constant ^{222}Rn activities along some creek reaches indicates a balance between constant groundwater inflow and the de-gassing and radioactive decay of ^{222}Rn in creek water.

The deuterium ratio (δ^2H) (Figure 3 (c)) is more negative between Cck10 to Cck9 as a result of surface water contribution above Cck9 from one main tributary (Cck_t5) and/or groundwater inflow between these points. Downstream of Cck9 the δ^2H becomes progressively more positive. Between Cck4 and Cck1, the trends in the δ^2H during different rounds highlight the varying degrees of surface evaporation from the creek at different times of the year. This is very clear in round 2 (March 06), which shows the δ^2H significantly more positive, reflecting baseflow conditions, higher evaporation, and minimal rainfall along the length of Cox Creek at that time of year.

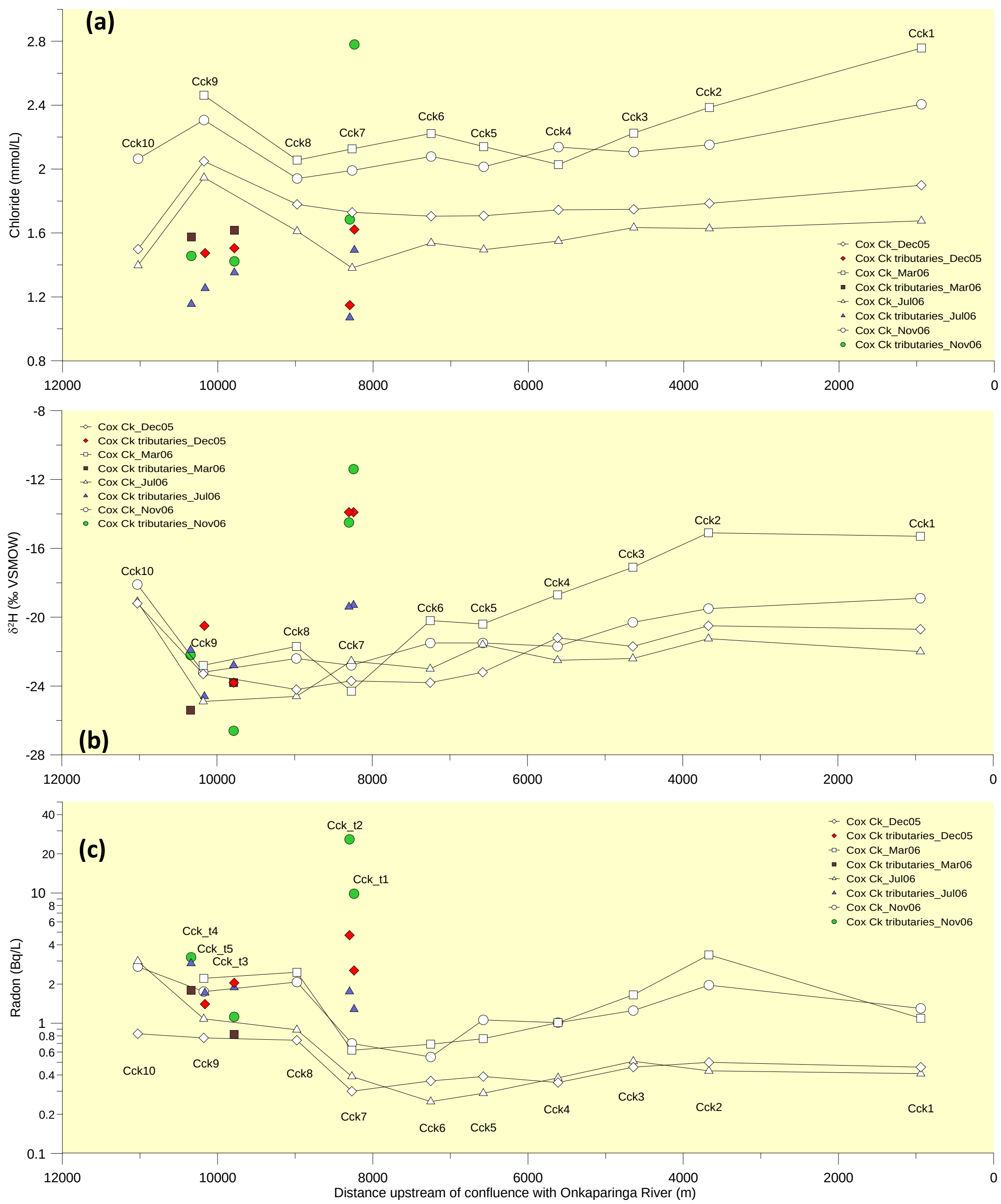

Figure 3. Spatial variation of chloride (a), deuterium (b) and ^{222}Rn (c), in Cox Creek in Dec. 2005, Mar. 2006, Jul. 2006 and Nov. 2006.

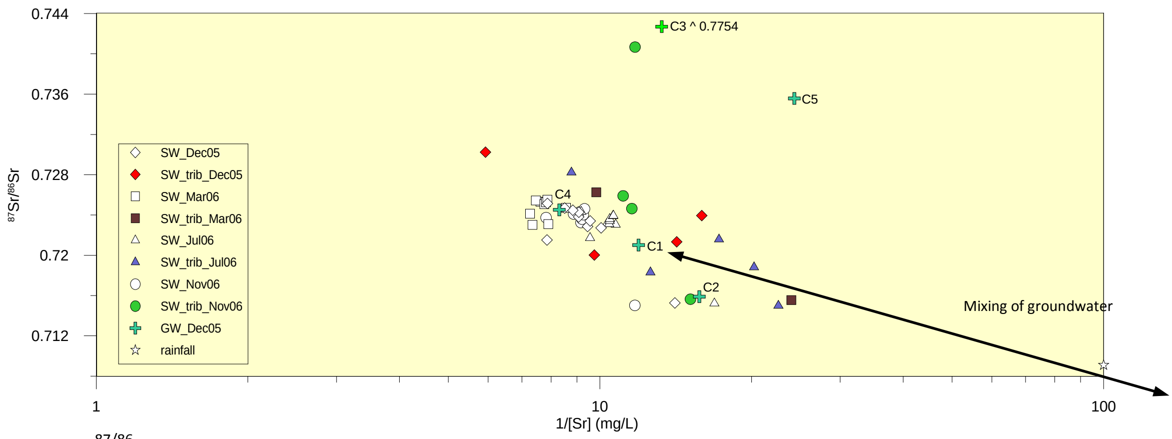

Figure 4. $^{87/86}Sr$ ratio versus the 1/[Sr] in Cox Creek and the sampled groundwater bores.

The plot of the $^{87}Sr/^{86}Sr$ ratio versus the reciprocal of the strontium concentration (1/[Sr]) shows that the surface water samples tend to lie on a mixing line between the expected end members of local rainwater and groundwater from the Emeroo Subgroup and more towards the groundwater end member (Figure 4). The differences in the $^{87}Sr/^{86}Sr$ ratios of the groundwater samples are controlled by variations in the initial atmospheric inputs, mineralogy of the rock along flowpaths, source of easily weathered strontium and residence time. The groundwater samples C4 (0.7245) and C1 (0.7210) were sampled from bores located upstream of Cck9 in the Emeroo Subgroup and their $^{87}Sr/^{86}Sr$ ratios are similar to the average ratio (0.7240) at all 4 sampling stages along the length of Cox Creek between sites Cck9 to Cck1. This suggests that there is a well-mixed and relatively homogeneous source of strontium and that the source of groundwater is dominantly from aquifers located in the Emeroo Subgroup, which dominates the hydrochemical signature in Cox Creek.

Figure 5 shows δ^2H versus the oxygen isotopes ratio ($\delta^{18}O$) of Cox Creek surface and groundwater samples relative to the Adelaide local meteoric water line (LMWL). The majority of the surface water samples lie close to and above the LMWL for Adelaide ($\delta^2H = 7.66.\delta^{18}O + 9.57$), with a composition similar to, or more positive than, the weighted average rainfall for Adelaide (δ^2H= -26.4 ‰ and $\delta^{18}O$= -4.7 ‰) (IAEA/WMO, 2001). These relatively positive values are thought to result from a mixture of seasonal localised recharge events and/or small amounts of evaporation occurring in conditions of high humidity (Coplen et al., 1999). Samples from sites Cck4, Cck3 and Cck1 in round 2 fall below the LMWL as a result of isotopic enrichment caused by evaporation. Samples Cck_t1 and Cck_t2, collected from small tributaries to Cox Creek, have a more isotopically enriched composition and reflect evapo-concentration of the water due to longer residence times in a series of pools and in-stream dams before the confluence with the main creek.

The proximity of the groundwater samples to the LMWL suggests that groundwater recharge has occurred fairly rapidly with minimal isotopic fractionation by evaporative process prior to rainfall infiltration. Groundwater samples (C1, C2, C4 and C5) have a more depleted isotopic composition than the weighted average rainfall, which is indicative of diffuse recharge occurring during cooler autumn and winter rainfall events and altitude effects, which result in depletion of both $\delta^{18}O$ (~0.15 to 0.5 ‰ per 100 m) and $\delta^{2}H$ (~1 to 4 ‰ per 100 m) values (Clark and Fritz, 1997). All of the groundwater samples (excluding C3) have a $\delta^{2}H$ that is more negative than the surface water samples from Cox Creek. The more positive $\delta^{2}H$ of sample C3 may be due to the sampled bore being located down-gradient of a leaking dam. The $\delta^{2}H$ of samples C1 and C4, sampled from bores located above Cck9, are $\delta^{2}H$ = -25.2 and -26.5 ‰, respectively indicating that groundwater here is a possible source of the more negative $\delta^{2}H$ end member to Cox Creek in the upper reaches of the catchment. However, the higher [Cl^-] of sample C1 indicates that the contribution of groundwater inflow from this area would have to be small given that the surface water samples have a much lower [Cl^-].

Groundwater flowpaths in CCC tend to be dominated by flow towards the creek, and may vary due to heterogeneities in the soil and bedrock. Comparing ^{222}Rn activities, isotope and hydrochemical stream data and geology across the catchment shows that the major fault systems are important zones where the contribution of groundwater input to Cox Creek may be greater than at other reaches along the creek. Downstream of Cck9, Cox Creek traverses a major fault between the Emeroo Subgroup and the Barossa Complex and corresponds with an increase in ^{222}Rn activity (Figure 1). Cox Creek traverses another fault system near Cck5, possibly another groundwater input zone, before crossing the Barossa Complex/Emeroo Subgroup boundary upstream of Cck4.

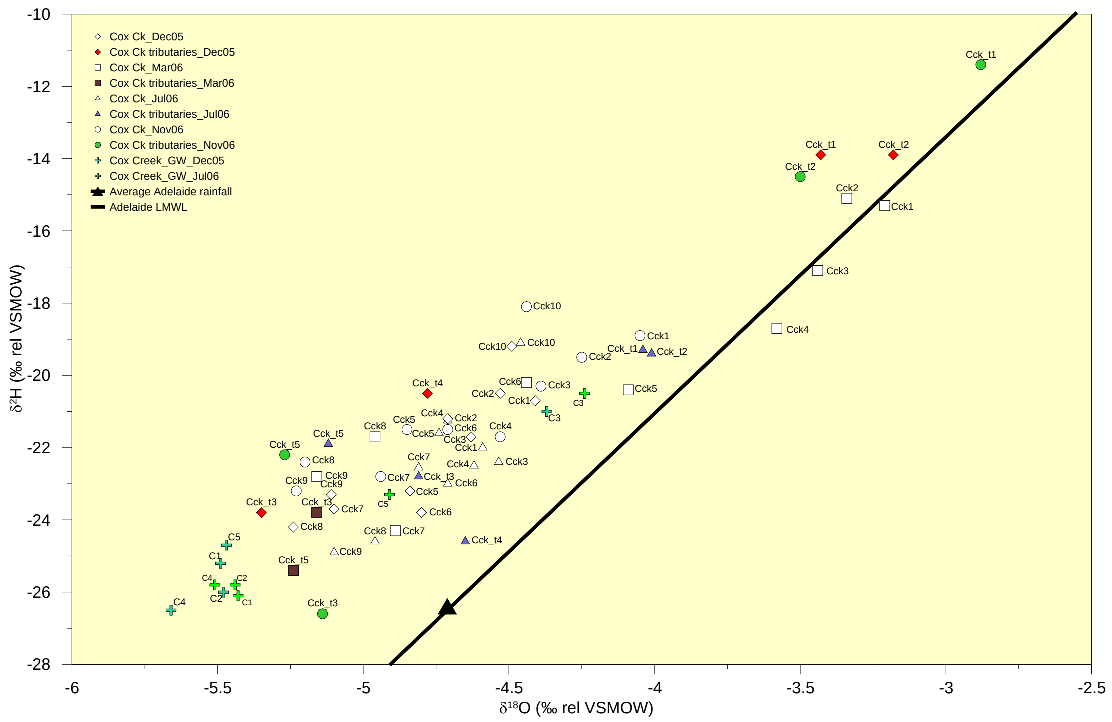

Figure 5. δ^2H versus δ^{18}O for Cox Creek Catchment. The LMWL for Adelaide is $\delta^2H = 7.7\ \delta^{18}O + 9.6$. The weighted average rainfall for Adelaide is δ^2H = -26.4 $_{‰VSMOW}$ and δ^{18}O = -4.7 $_{‰VSMOW}$.

Bremer River Catchment

Through the period of the BRC sampling program, many of the creeks in the catchment were not flowing. The Bremer River itself was only flowing below the confluence with Mount Barker Creek, during the first round of sampling and therefore a run-of-river analysis was not appropriate for this catchment. Instead, the relative contributions of groundwater inflows and surface runoff at individual locations were interpreted from the position of surface water sample data on scatter graphs of a number of hydrochemical variables. Sample numbers and locations are shown on Figure 2.

The ^{222}Rn activity (Figure 6) was used as an indicator of groundwater inflow occurring either recently or contemporary with the time of sampling and occurring fairly close to the sampling point. ^{222}Rn activity values of greater than 0.5 Bq/L in surface water samples were taken to indicate some degree of nearby and recent groundwater inflow. Activities below 0.5 Bq/L were not taken to imply that groundwater is not contributing to the water found at any location, but that it has had sufficient time to degas from the water or has undergone radioactive decay to the extent that the remaining ^{222}Rn activity in the water is very low and cannot be distinguished from the activity that may result from emanation of ^{222}Rn from the hyporheic zone.

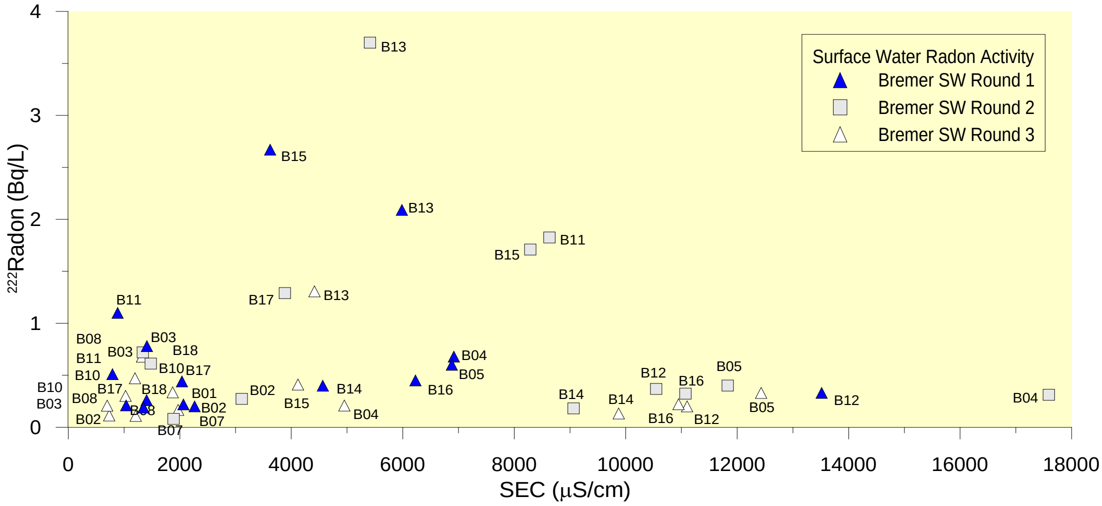

Figure 6. ^{222}Rn activities (Bq/L) of surface water samples in the Bremer River Catchment.

The $^{87/86}$Sr ratios were plotted against 1/[Sr] (Figure 7). The location of surface water samples on this plot in relation to the expected end members of local groundwater and local rainwater were taken to imply evapo-concentration and mixing proportions of water from these two sources. An assumption made here is that the $^{87/86}$Sr ratio of the sampled water reflects the $^{87/86}$Sr ratio of the dissolved Sr in the source water or mix of sources. Groundwater may develop an $^{87/86}$Sr isotope signature that is distinct from rainwater and reflective of the geology of the aquifer when it has a long enough residence time in an aquifer to dissolve sufficient strontium from the rock to dominate the $^{87/86}$Sr ratio in the water. Where a water sample appeared to be a mix of groundwater and surface water sources, Equation 1 (Faure, 1986) was applied to estimate relative concentrations from each source:

$$^{87/86}Sr_{mix} = (^{87/86}Sr_{gw}) \cdot (Srgw.f / Sr_{mix}) + (^{87/86}Sr_{rw}) \cdot (Sr_{rw}.(1-f) / Sr_{mix}) \qquad \text{(Equation 1)}$$

In Equation 1, Sr_{gw} and Sr_{rw} and Sr_{mix} are the concentrations of Sr in, respectively, the groundwater and rainwater end members and the mixture of the two, while $^{87/86}Sr_{gw}$, $^{87/86}Sr_{gw}$, and $^{87/86}Sr_{gw}$ are the $^{87/86}$Sr ratios of these three components, and f is the fraction of groundwater in the mix.

Deuterium plotted against the chloride concentration (Figure 8Figure) was used to identify the degree of evaporation of surface water samples and to indicate whether this was likely to be a result of the evaporation of groundwater or surface runoff. For the purposes of this assessment, the BRC was considered as five distinct sections: the Upper Bremer River, Lower Bremer River, Mount Barker Creek, Nairne/Dawesley Creek, and Rodwell Creek. These areas are outlined in Figure 2.

The groundwater samples from the BRC all lie in a cluster on the plot of δ^2H versus [Cl^-], while the surface samples lie on an evaporation line drawn from the groundwater samples in the direction of more positive δ^2H and greater [Cl^-]. In the *Upper Bremer River* the permanence of the pools through the exceptionally dry period of the field study suggests that there must be an inflow of groundwater that is approximately equal to the evaporation from the pools. Some samples from the summer and autumn samplings show very positive δ^2H and high [Cl^-], as would be expected after a long period of evaporation without flushing by a flowing river. The Sr isotope data support this interpretation. When plotted on the graph of $^{87/86}Sr$ against the reciprocal of the Sr concentration, all of the surface water samples have a similar $^{87/86}Sr$ ratio but lie to the left of the groundwater samples, indicating high Sr concentrations resulting from evaporation of the water with the same $^{87/86}Sr$ ratio as the local groundwater. Groundwater from wells in the Tapanappa Formation FRA in this section was found to have relatively high ^{222}Rn activities of between 200 and 400 Bq/L. The majority of the surface water samples from this area had low ^{222}Rn activities of less than 0.5 Bq/L, commensurate with a low rate of groundwater inflow.

In the *Nairne / Dawesley Creek* section the four surface water sample locations had low ^{222}Rn activities in all three sampling rounds, even when there was significant flow through the creeks such that water residence times were low. These results suggest that there is minimal inflow of groundwater to these creeks. This is supported by the plot of δ^2H versus [Cl^-], in which the surface water samples from this section all lie in a zone above and to the right of rainwater samples collected at the nearby Mount Pleasant pluviometer station, suggesting that the majority of this water is derived from surface runoff. On the plot of $^{87/86}Sr$ versus 1/[Sr], these samples all lie on a line of dilution between the samples of groundwater from this section and the expected Sr isotopic characteristics of evapo-concentrated local rainwater. Because of the mass difference between the very low concentrations of Sr in rainwater (typically < 0.005 mg/L) and the relatively high concentrations of Sr in the Bremer catchment groundwater (0.4 – 1.5 mg/L), any movement toward the rainwater end-member in the $^{87/86}Sr$ versus 1/[Sr] relationship requires a high degree of dilution of groundwater by rainwater. A mass balance calculation using both Sr concentrations and $^{87/86}Sr$ isotope ratios according to Equation 1 indicates the surface water samples from the Nairne/Dawesley Creek subcatchment to be a mix of 21% or less of evapo-concentrated groundwater and 79% or more of evapo-concentrated rainwater. The creeks in this section of the catchment are surmised to be mostly dependent on rainwater runoff rather than on groundwater inflows. It is likely that much of this flow arises from the areas of higher rainfall in the northwestern part of the BRC.

In the *Mount Barker Creek Subcatchment,* the groundwater samples had ^{222}Rn activities of 60 - 118 Bq/L, whereas surface water samples had a maximum ^{222}Rn activity of only 0.78. This suggests that either there is a low rate of groundwater inflow here or that any groundwater inflow occurs some distance upstream. However, the creek at this location flows throughout the year, implying some groundwater inflow to sustain surface flows. The $^{87/86}$Sr ratios of the two surface water samples from this subcatchment lie on a dilution line between the $^{87/86}$Sr isotopic characteristics of the groundwater and the expected Sr isotopic characteristics of evapo-concentrated local rainwater. Application of the mass balance model (Equation 1) indicates groundwater fractions of 26% and 19% in these samples. The plot of δ^2H versus [Cl$^-$] supports this finding, indicating all but one of the surface water samples result from either evaporation of local rainwater, or a mix of local rainwater with a low percentage of local groundwater.

In the *Rodwell Creek* section, water was present at both of the surface water sample locations throughout the year, suggesting a groundwater inflow to maintain the permanent pools that exist at these locations. Collectively, the hydrochemical and isotopic data for the Rodwell Creek samples strongly suggest that surface water at these locations is derived largely from groundwater discharge. On the plot of δ^2H versus [Cl$^-$], the surface water samples from locations B12 and B13 in all three sampling rounds lie on a line of evapo-concentrated groundwater samples from wells Bg03 and Bg07, indicating evaporation of groundwater inflow only. The ^{222}Rn activities of the spring, summer and autumn samples from B13 have ^{222}Rn activities of 2.1 Bq/L, 3.7 Bq/L and 1.3 Bq/L respectively, indicating some groundwater inflow throughout the year. The activities at B12 are lower than at B13, being all less than 0.5 Bq/L, indicating that if the surface water at B12 is derived from groundwater inflows, then the residence time of the water here is higher than at B13. On the plot of $^{87/86}$Sr versus 1/[Sr], all surface water samples from B12 and B13 lie directly to the left of the groundwater samples from wells Bg03 and Bg07, suggesting that they result solely from evapo-concentration of groundwater discharged from the FRA in the Tapanappa Formation.

The *Lower Bremer River* Section of the BRC has significantly lower rainfall than the areas of higher elevation in the west and north of the catchment. At times when flow occurs in this section of the river, it is mostly as a result of flows from upstream subcatchments, primarily Mount Barker Creek. All but one of the surface water samples from this section had ^{222}Rn activities of less than 0.5 Bq/L. The exception was an activity of 1.3 Bq/L at location B17 during the summer sampling round, indicating that water flowing in that section of the river during summer is at least partly derived from groundwater inflowing immediately upstream of B17. The plot of δ^2H versus [Cl$^-$] suggests that water at B17 has the characteristics of evapo-concentrated rainwater in the spring and summer, while samples from B01 and B02 had these characteristics at all times that water was present. It appears that water occurring in the lower reaches of the Bremer River, south of B17 is due mostly to surface runoff. Flow may only occur here after heavy rain, or during the winter when significant amounts of water are flowing from Mount Barker Creek into this section of the Bremer River. During the autumn sampling round, the river had a significant flow of approximately 20 L/s at location B02, but this flow was observed to be all lost to groundwater within approximately 1 km south of that point. Further south, at location B01, water was only flowing during the spring sampling round. It is inferred from these observations that the section between B02 and B01 is a losing reach of the river and that flow only occurs at B01 during winter and after rain, when the rate of flow through B02 is greater than the rate at which water can drain to the groundwater system between B02 and B01. Upstream of B02 the river receives some local groundwater inflow in the summer, which is supplemented by outflows from Mount Barker Creek during all other seasons.

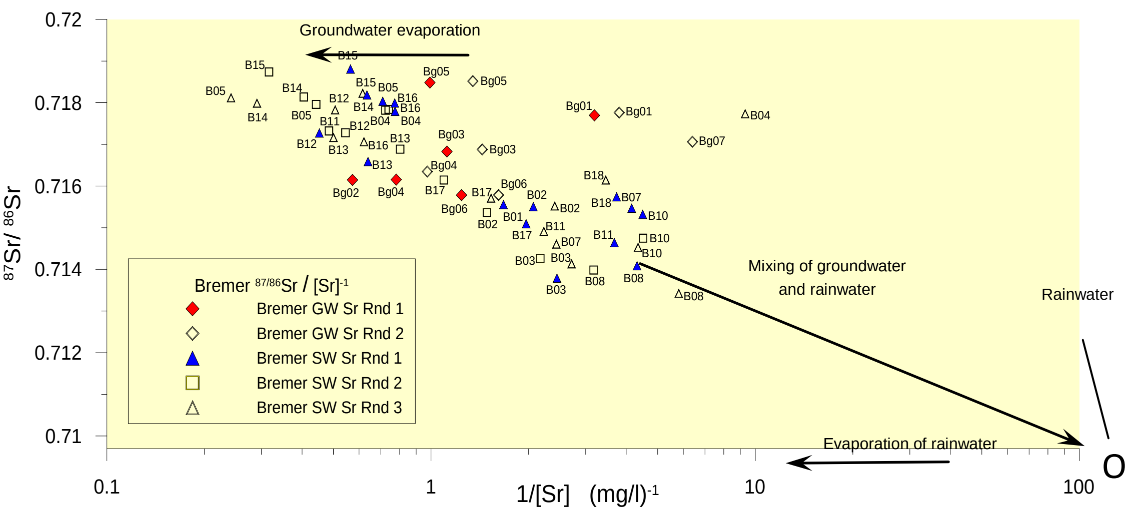

Figure 7. $^{87/86}$Sr ratio versus the 1/[Sr] in surface and groundwater samples from the BRC.

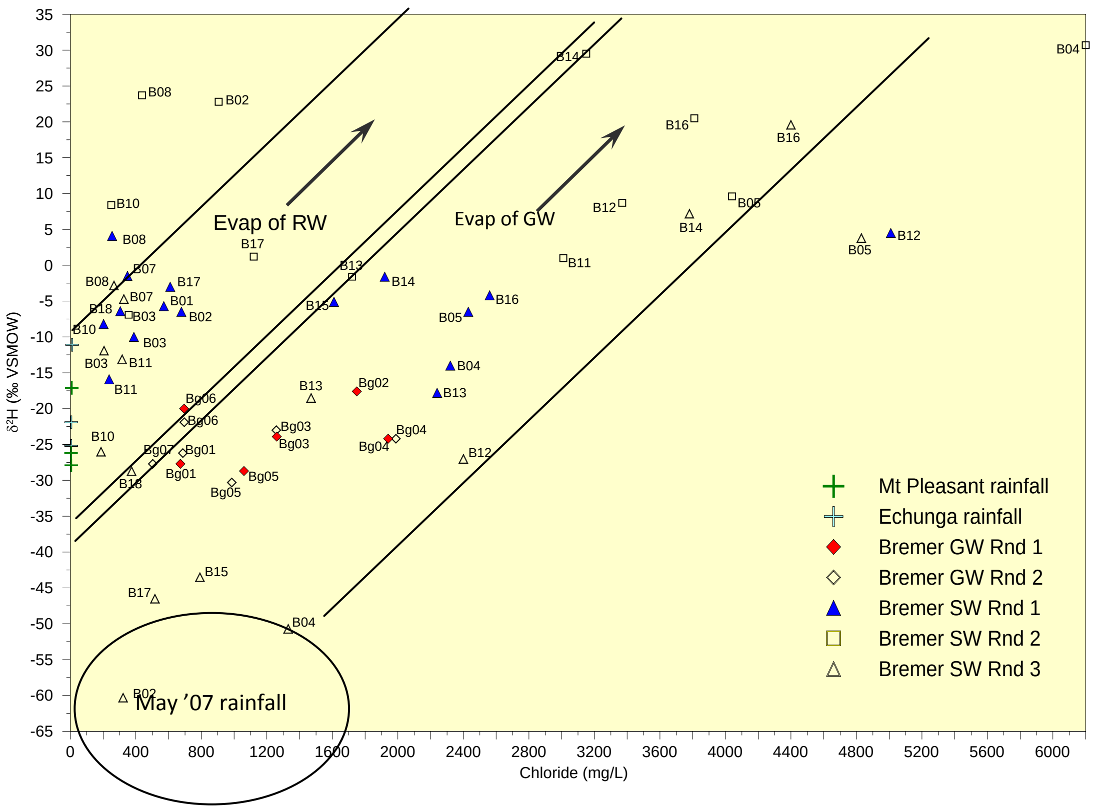

Figure 8. δ^2H versus [Cl$^-$] of surface water samples in the Bremer River catchment.

Discussion

The hydrochemical, radon and isotope 'run of river' data has shown consistency in the trends between the four sampling rounds and suggests that there is a steady baseflow contribution of groundwater to Cox Creek along its length regardless of the season. The temporal variations in the radon activities, isotopes and ion concentrations are a result of rainfall dilution and evaporative processes and contribution from small ephemeral streams to Cox Creek. The spatial variations in the data are a result of groundwater inflow contributions to the creek, which are related to the geology, zones of major faulting in the vicinity of the creek, topography, surface water residence times and land use. The findings indicate that there are groundwater 'hotspots' over a scale of one to two kilometres along Cox Creek which could not be easily identified on a potentiometric map constructed from the limited head data from the catchment. The steep topography of the catchment was also a limiting factor in the generation of a potentiometric map. Conducting four sampling rounds at different flow conditions has shown consistency in the results as to where there are significant groundwater inflows to the creek and not a result of pre or post rainfall events if sampling had been done only once. It has also shown the importance of flow pathways and residence times of the small ephemeral tributaries to Cox Creek, which were in most instances sustained by groundwater.

The results from the five sections of the BRC indicate there are significant surface water streams with contributions for at least part of the year from groundwater inflows. The year-round presence of water in pools along the course of the upper Bremer River appears to be particularly dependent on groundwater inflows. In the Rodwell Creek section, the agricultural dams in the higher-rainfall areas of the upper catchment prevent flow to the lower catchment. The year-round presence of water in the lower Rodwell Creek is therefore also largely dependent on groundwater inflows. In the higher-elevation and higher-rainfall subcatchments of the Nairne, Dawesley and Mount Barker Creeks there is a mix of contributions from groundwater inflows and rain runoff. As may be expected, in locations where water was present during the summer sampling round there was a higher proportion of groundwater. However, in the Nairne Creek, which has fewer agricultural dams in its upper catchment, flow persisted even through the summer and much of this flow appeared to be derived from surface runoff of recent rainfall, suggesting a delayed interflow of water from prior rain events.

Key to the management of water resources in these catchments is the understanding of groundwater <u>contributions</u> to surface streams and the groundwater <u>dependence</u> of the streams. This distinction between contributions and dependence is important when managing water resources in the catchment with a view to protecting surface water ecological assets. Groundwater dependence cannot be assumed to be only a feature of streams in low-rainfall catchments. In the higher-rainfall catchment of Cox Creek it is apparent that at all times of year groundwater provides some proportion of the water flowing in the creek. This proportion is greater in the summer, when the groundwater contribution may provide the majority of water. Hence, Cox Creek is dependent on groundwater inflows for its perennial flow condition. In contrast, there is rarely flow in the Upper Bremer River, however the permanent pools here are dependent on groundwater inflows. In the Nairne Creek there was flow for the whole year of the sampling program, however, this flow did not appear to be dependent on groundwater inflows at any time of year.

The analysis of the BRC has shown that there can be differing degrees of surface water–groundwater interaction among the subsections of just one relatively large catchment.

There is considerable similarity between the lithology and topography of the three subcatchments of the western BRC, and in the absence of other data, a resource manager may be led to believe that one of these subcatchments could be used as an example that indicates the likely nature of interactions in the other two. However, even among these topographically similar and adjacent catchments, the variations observed in the BRC study show that a common rule-of-thumb cannot be applied to predict the likely extent of interactions in any one of the catchments. There is clearly a minimum density of sample locations, below which the findings of a study may not be useful even for regional groundwater resource management. The investigation of the BRC was undoubtedly testing that lower limit of sample point density. With only approximately one sample point for each 8 km of stream, in some areas the likely interactions between surface water and groundwater had to be inferred from other knowledge of the catchment, such as the presence of springs or dams, rather than determined definitively from the hydrochemical results. If definitive results are required to enable the pinpointing of individual reaches of net gain or loss from the stream, then the higher sample point density and subsequent run-of-river analysis of data, as employed in the CCC, is likely to be necessary.

Conclusion

The two different approaches used in the WMLR and EMLR have improved the understanding of the interactions between groundwater and surface water in fractured rock environments. Whether the two studies provide sufficient spatial resolution of the surface water – groundwater interactions at the catchment scale to be useful in planning the water resources in the MLR must be considered. In managing the groundwater resources there is a requirement to identify 1) where surface water bodies with significant ecological assets exist, 2) where those ecological assets may be affected by increases in extraction of groundwater for irrigation / industrial purposes and 3) where there may be areas of over allocation of the resources to current and future users. This type of study aims to be able to provide answers to the second and third of these requirements, and we surmise that the more intensive study has achieved this aim. The higher density of sample points used in the Cox Creek study allowed a 'run of river' approach was useful to identify 'hotspots' of groundwater inflow to the creek. The more generalised interpretations that have been possible with the low spatial density of sample locations employed in the BRC study have not been able to identify such hotspots of groundwater inflow, but have been able to identify the degree to which subcatchments of the BRC have surface streams that are dependent on groundwater inflows.

Ultimately, in this type of study, the more sample points that are used, the more informative are the results. In practice the spatial density of sample locations will be limited and a trade-off must be made between resources available and the quality of the results achieved. The comparison of these two studies shows that the spatial density of samples used in the CCC, with approximately 1–2 km separation between points provides a significantly more satisfactory result. However, useful information for natural resource management decisions can also come from studies that cover larger areas with lower sample locations densities.

6.2 Assessing Connectivity between a Fresh Water River and Saline Aquifer in a Pristine Catchment, Kangaroo Island, South Australia

Presented at American Geophysical Union Western Pacific Meeting Conference, Cairns, Queensland, August 2008

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The importance of surface water-groundwater interactions has received greater attention in the last decade, in response to water resource allocation needs and the impacts on groundwater dependent ecosystems. Assessments of connectivity (e.g., losing, gaining, flow-through or disconnected conditions) are becoming increasingly commonplace in water resource management. Many connectivity assessments are either model based and theoretical in nature, or based primarily on hydraulic gradient information alone. Some have used EC or Radon-222 activity as an indicator of groundwater gaining conditions. Losing stream conditions are possibly more challenging. In the case of completely disconnected conditions, the theoretical criteria for disconnection are poorly defined. Disconnection is inferred on the basis of a theoretical assessment of head measurements rather than a direct field-based measurement (e.g., the measurement of an unsaturated zone beneath a river bed). This study, attempts to assess the state of connectivity using hydraulic data, chemistry and tracer approaches to provide multiple lines of evidence in support of a working hypothesis that a river system is disconnected from groundwater.

The Rocky River Catchment on Kangaroo Island, South Australia is unique, as it is one of the few environmentally pristine catchments covered by native vegetation remaining in South Australia. Rocky River flows perennially and despite what one would expect for a catchment

covered by native vegetation, the river is extremely fresh (EC <500 μS/cm) compared to the groundwater of the regional fractured rock aquifer system (EC >6000 μS /cm). Three nested piezometer transects were installed in the Rocky River Catchment. Preliminary findings indicate that at each of the nested piezometer transect sites the river is disconnected from the aquifers beneath and behaves as a losing river system. The hydraulic gradient between the shallow aquifer directly beneath the river and adjacent the river is relatively steep (approximately 0.2) and suggests a hydraulic disconnection. This hypothesis is supported by soil augering in the river channel at one of the transects, which clearly showed that there is an unsaturated zone in the shallow aquifer beneath the river and above the regional aquifer. While this provides primary evidence for the disconnection hypothesis, more data is required before making broader generalisations about the larger system.

The emerging conceptual model of this system is that the river is fed by winter rainfall, falling between May and October. Given that the river continues to flow all year around, this suggests a subsurface fresh water inflow either from shallow lateral flow or a source upstream of the study sites. Soil chloride profiles taken across the catchment indicated that there are large stores of salt present at shallow depth. Further work is required to assess the water storages and volumes required to sustain perennial flows. A key component of our working hypothesis is that flow may occur via macropore pathways and lateral flow in deeper regolith or bedrock, effectively by-passing the salt store.

An assessment of pristine versus cleared catchments would be vital in assessing what affect clearing this catchment would have on the groundwater levels, state of disconnection, and hence saline accessions from groundwater to the river.

6.3 Assessing Surface water–groundwater Connectivity Using Hydraulic and Hydrochemical Approaches in Fractured Rock Catchments, South Australia

Presented at Water– Rock Interaction 13 Conference, Guanajuato, Mexico, August 2010

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ABSTRACT: In Australia, native vegetation clearance has had considerable impacts on surface water and groundwater salinities. The impact on surface water-groundwater connectivity is less understood. A hydraulic, hydrochemical, and tracer-based study was conducted at two contrasting fractured rock catchments in South Australia. Results indicate that connectivity was variable across each of the catchments. The influence of the fractured rock aquifer was minimal in the pristine, uncleared Rocky River catchment, whereas in the cleared, mixed land-use Cox Creek catchment, the fractured rock aquifer played a more significant role. The results emphasise the need to understand the importance that the impacts of land-use change (particularly vegetation clearance) can have on surface water-groundwater connectivity.

Introduction

The importance of surface water-groundwater interactions has received greater attention in the last decade, in response to water resource allocation needs and the impacts on groundwater dependent ecosystems. It is becoming increasingly evident that there are often strong hydraulic connections between these two resources (Banks et al., 2009; Winter et al., 1998). Considerable research has been undertaken in sedimentary aquifer systems but very few studies are reported for fractured bedrock systems (Haria and Shand, 2006; Kahn et al., 2008). The latter are substantially more complex owing to the geological heterogeneity of the fractured rock aquifer (FRA).

Rates of groundwater flow and connectivity in FRA are difficult to determine, and methods commonly used for porous media are often not applicable (Cook et al., 1996). Multi-tracer approaches have been used in fractured rock systems (e.g. Genereux et al., 1993a; Shand et al., 2007a) and have proven invaluable in constraining contributing sources of solutes, preferential flow pathways and residence times in these types of systems.

This study examines the results of two contrasting catchments in the greater Mount Lofty Ranges (MLR) of South Australia using a multi-tracer approach. The aim was to improve understanding of surface water-groundwater connectivity in fractured rock catchments and evaluate potential differences between cleared and uncleared catchments.

STUDY SITES

COX CREEK CATCHMENT

The Cox Creek catchment (CCC) is situated approximately 20 km east of Adelaide. The catchment covers an area of 28.8 km^2 and land use is a mix of agriculture, small farms and forest. Average annual rainfall is 1189 mm/y, the majority of which falls between May and October. The geology of CCC includes several stratigraphic sequence associated with the Adelaide Geosyncline. The Neoproterozoic Emeroo Subgroup (quartzite, sandstone and dolomite) dominates the north and south of the catchment and is separated by the Archean Barossa Complex (metamorphic rocks with retrograde metamorphism), which lies across the middle of the catchment (Figure 1).

ROCKY RIVER

The Rocky River Catchment (RRC) is located within Flinders Chase National Park on Kangaroo Island, South Australia. The catchment covers an area of about 216 km^2 and is one of very few remaining catchments in South Australia still covered by native vegetation (Figure 2). The Rocky River is semi-perennial, flowing all year round in the mid to upstream part of the catchment whilst the lower part usually ceases to flow during the summer months. The average annual rainfall is 780 mm/y. The geology is dominated by a laterite plateau underlain by Cambrian and Neoproterozoic metasediments. Towards the bottom of the catchment, the river has incised through Quaternary calc-arenite and limestone deposits.

METHODS

SAMPLING AND ANALYTICAL TECHNIQUES

Surface water and groundwater sampling was conducted along the length of the creeks during a 12 month period. A YSI® multi-parameter meter was used to measure the pH, specific electrical conductance (SEC), dissolved oxygen (DO), redox potential (Eh) and temperature in the creek and in groundwater bores. The alkalinity (as HCO_3^-) was measured in the field using a HACH titration kit. Surface water and groundwater samples were analysed for major elements, stable isotopes, ^{222}Rn, and $^{87}Sr/^{86}Sr$. All analyses were done using standard analytical techniques. Manual flow gauging was conducted using a pigmy flow meter (OTT) and YSI® flow tracker at sampling locations in both catchments.

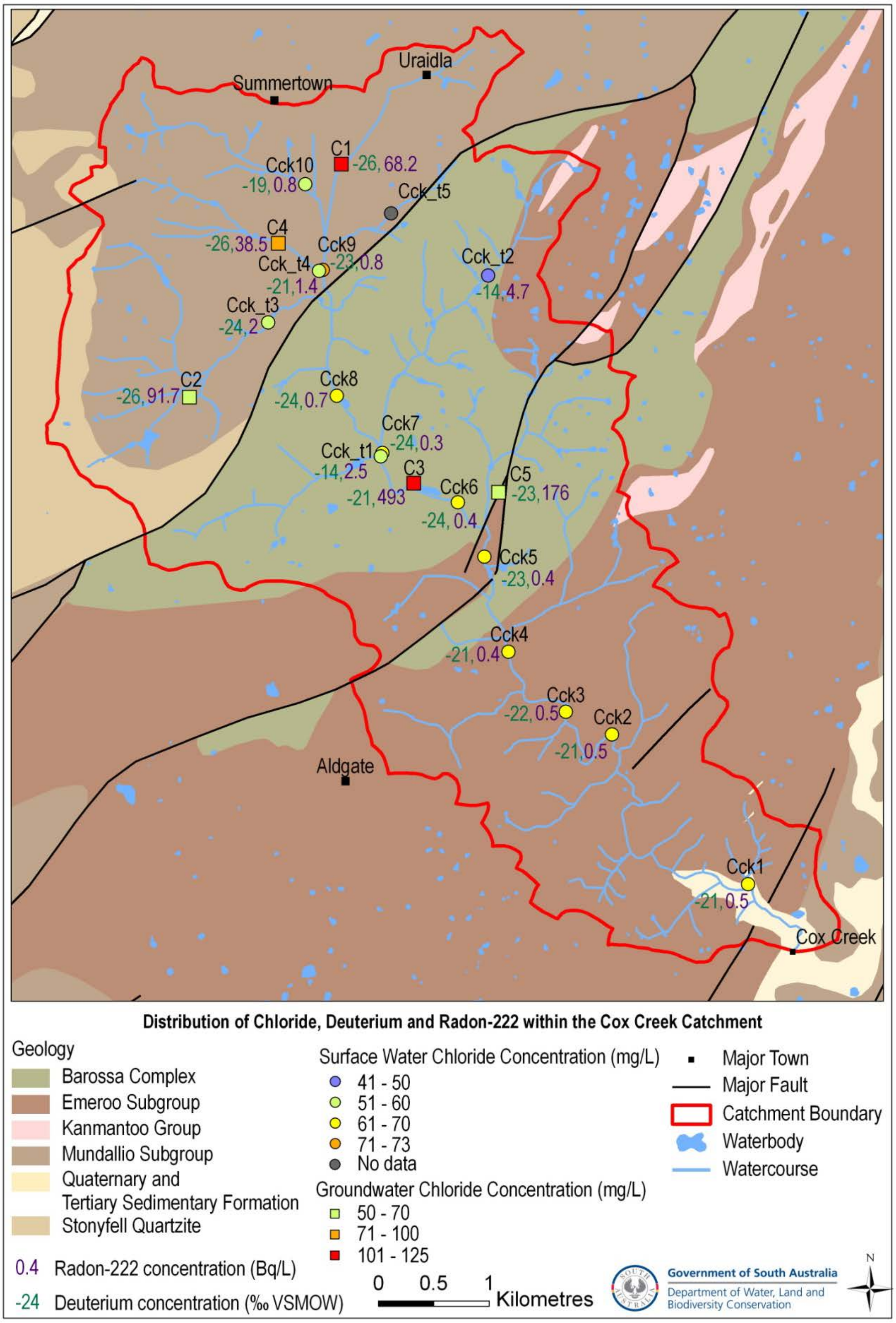

Figure 1. Spatial distribution of ^{222}Rn (Bq/L), deuterium (‰ relative to Vienna Standard Mean Ocean Water- VSMOW) and chloride (mg/L) concentrations of surface water and groundwater in the Cox Creek Catchment, December 2005.

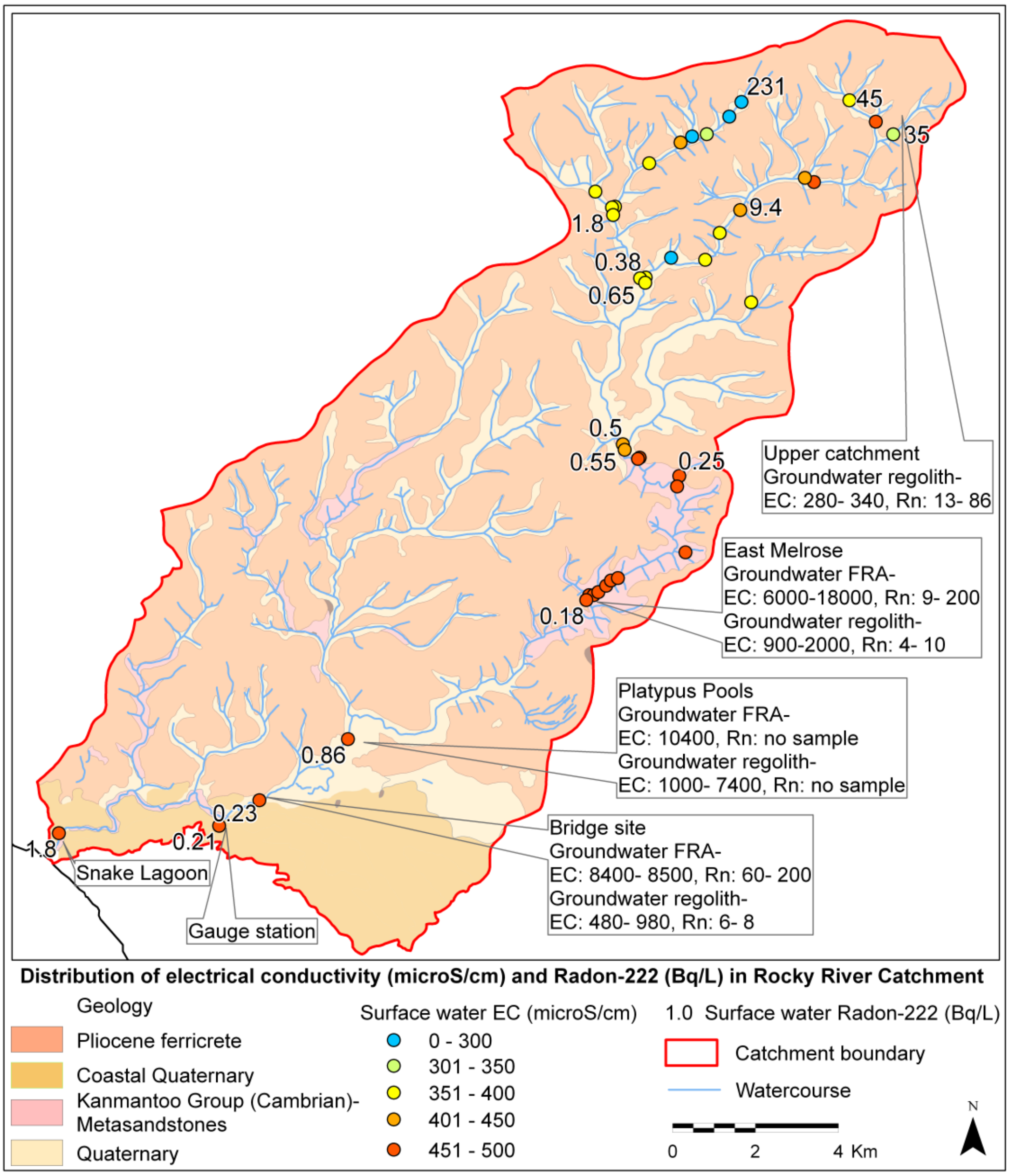

Figure 2. Spatial distribution of electrical conductivity (EC μS/cm) and ^{222}Rn (Bq/L) of surface water and groundwater in the Rocky River Catchment.

RESULTS AND DISCUSSION

COX CREEK

Significant changes in solute concentrations and environmental tracers along the length of Cox creek may indicate locations of groundwater discharge to the creek and the influence of the underlying geology and/or fault zones that traverse the creek (Figure 1). The trends in the ^{222}Rn activity during the sampling period were similar, and suggest that there is connectivity between the groundwater system and Cox Creek (Figure 3). The locations of high ^{222}Rn activity indicate localised groundwater influx to the stream (in this investigation we assume that the groundwaters have a constant high ^{222}Rn activity and reflect the lithology of the major aquifers sampled). The ^{222}Rn activity downstream of influx declines rapidly due to its short half-life (3.82 days) and loss to the atmosphere by gas exchange. The constant ^{222}Rn activities along some creek reaches suggests a balance between a consistent groundwater inflow and the de-gassing and radioactive decay of ^{222}Rn in creek water. During baseflow conditions (Dec-Mar), ^{222}Rn activities in the stream varied from 0.6 Bq/L at Cck7 to 3.4 Bq/L at Cck2, compared to winter (Jul) where the activity only varied from 0.39 Bq/L at Cck7 and 0.43 Bq/L at Cck2 (Figure 3). The lower values in winter are likely to be a result of dilution by rainfall and surface runoff.

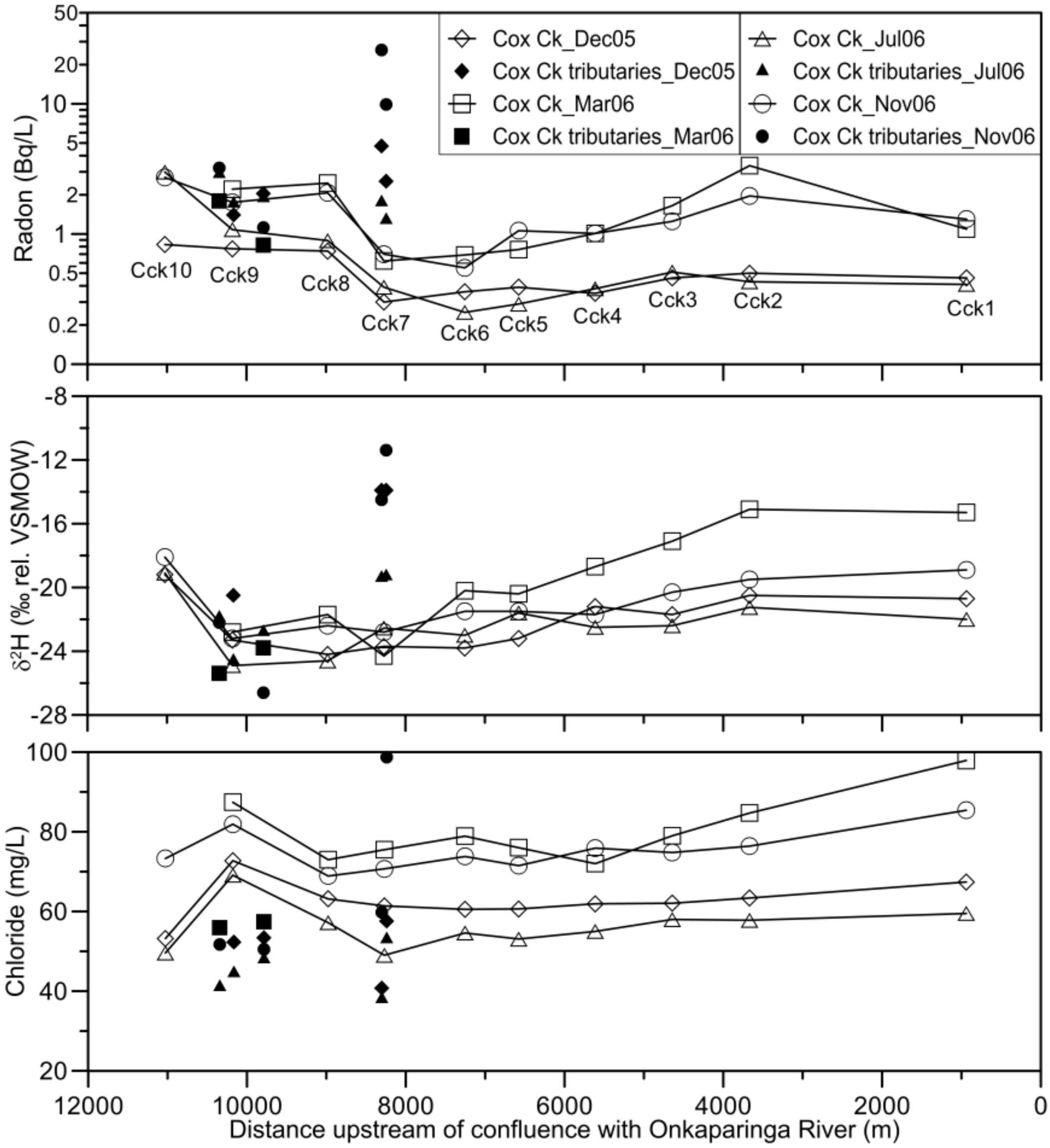

Figure 3. ^{222}Rn (a), deuterium (b) and chloride (c) in Cox Creek in December 2005, March 2006, July 2006 and November 2006. Distances are measured upstream of confluence with the Onkaparinga River (m).

The ^{222}Rn activities in the groundwater were an order of magnitude higher than in the surface water samples and reflect the mineralogy of the aquifer (Love et al., 2002). Groundwater from the FRA in the sandstone, quartzite and dolomite units had lower ^{222}Rn activities (37.9–87.1 Bq/L) compared to groundwater from the FRA in the metamorphosed gneisses and schists (220–489 Bq/L).

The chloride concentrations (Cl) in Cox Creek showed a decreasing trend in the downstream direction from sample site Cck9 to Cck8/Cck7, and then gradually increased to Cck1. This suggests discharge of a low salinity groundwater end member in the top part of the catchment. This was supported by the stable isotopes of the water molecule which showed the groundwater had a more depleted isotopic composition than Cox Creek (Figure 1). Downstream of sample site Cck8 the δ^2H became progressively more positive indicating surface evaporation. The higher Cl of groundwater sample C1 compared to the other samples in the top part of the catchment indicates that the contribution of groundwater from this area of the top part of the catchment would have to be small given that the surface water samples have a much lower Cl.

The plot of the ^{87}Sr/^{86}Sr ratio versus the reciprocal of the strontium concentration (1/Sr) shows that the surface water samples trend between local rainwater and groundwater (Figure 4). A more detailed analysis suggests that groundwater from aquifers located in the top part of the catchment dominate the hydrochemical signature in Cox Creek and are likely to be the main source of groundwater to the creek. Evaluation of the ^{87}Sr/^{86}Sr ratio versus δ^{18}O provided further evidence, which indicated that the majority of the samples from Cox Creek had an evaporated groundwater signature.

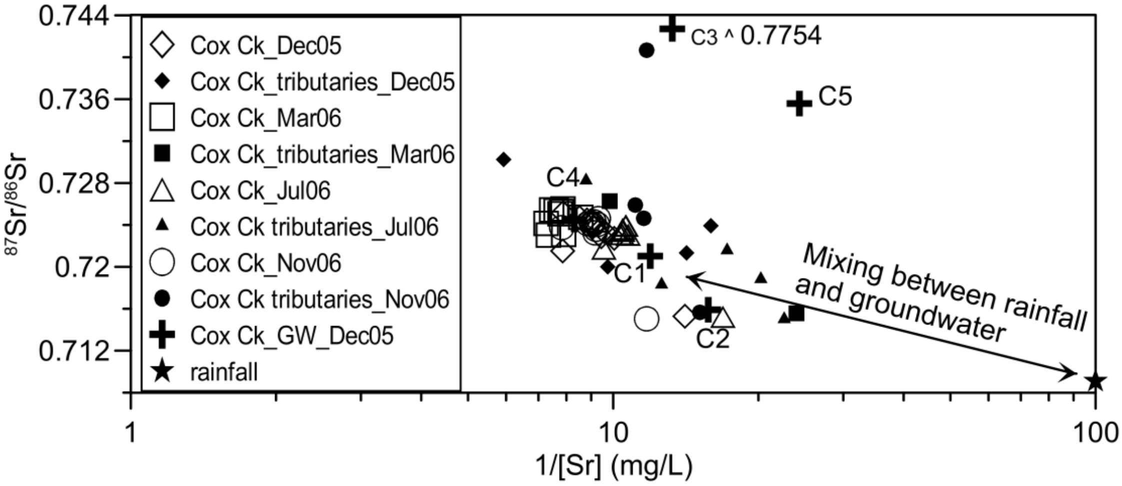

Figure 4. ^{87}Sr/^{86}Sr ratio versus 1/Sr in Cox Creek and the sampled groundwater bores.

ROCKY RIVER

Hydraulic and hydrochemical results showed both gaining and losing sections of the Rocky River and in some locations the river is disconnected from the aquifers beneath. The increase in EC from the top of the catchment down to Snake Lagoon is a result of evapotranspiration and/or groundwater discharge (Figure 2). Continuous water level monitoring showed that at the study sites (East Melrose, Platypus Pools and the Bridge) the river was losing at these locations and therefore, the increase in EC can only be a result of evapotranspiration. In the upper region of the catchment, the river must be gaining to sustain perennial flow at and above East Melrose. Below East Melrose during summer periods flow has sometimes ceased (Figure 5).

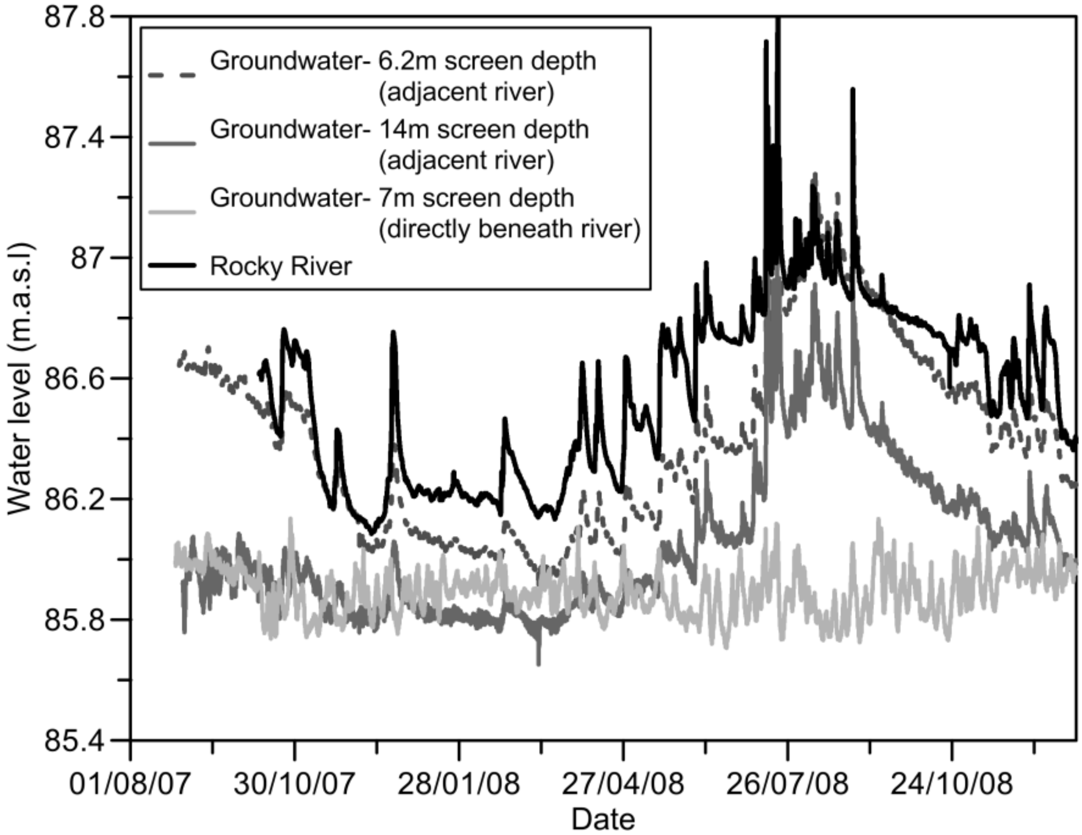

Figure 5. Continuous water level data (m above sea level) from monitored piezometers and Rocky River at East Melrose from August 2007 until December 2008.

Gaining conditions in the upper region of the catchment indicates a subsurface fresh water source to the river that is either from shallow sedimentary and/or fractured rock aquifer systems. There is no substantial winter surface runoff to sustain perennial river flow. Groundwater samples from the fractured rock aquifer had much higher salinities (>6000 μS/cm) than the river (<500 μS/cm). Soil chloride profiles taken across the catchment also indicated that there are large stores of salt present at shallow depth. However, groundwater sampling in shallow perched sand aquifers in the upper reaches of the catchment had similar salinities and ^{222}Rn activities to the river, suggesting that these shallow aquifers are the subsurface source to Rocky River (Figure 2). The ^{222}Rn activity in the river reduced to less than 0.2 Bq/L by the time it reached East Melrose suggesting minimal to no groundwater contribution below this point.

The plot of $^{87}Sr/^{86}Sr$ ratio versus 1/Sr shows the shallow groundwater and surface water from the upper region of the catchment plot close to the rainfall samples, whilst the surface water samples at and below East Melrose have a higher ratio and Sr concentration. The groundwater samples from the fractured rock have a large variation in the $^{87}Sr/^{86}Sr$ ratio however, the groundwater samples from the fractured rock sampled beneath the river plot closely to the surface water samples supporting the losing stream conceptual model (Figure 6).

The emerging conceptual model of this system is that winter rainfall replenishes the shallow perched sand aquifers located in the upper reaches of the catchment which gradually drain and discharge into the tributaries of Rocky River. The contribution from the FRA is minimal. A high evapotranspiration rate from the native vegetation in a dominantly losing system is likely to maintain a fresh river system because in this state, deeper regional saline groundwater from the FRA cannot discharge into it as baseflow.

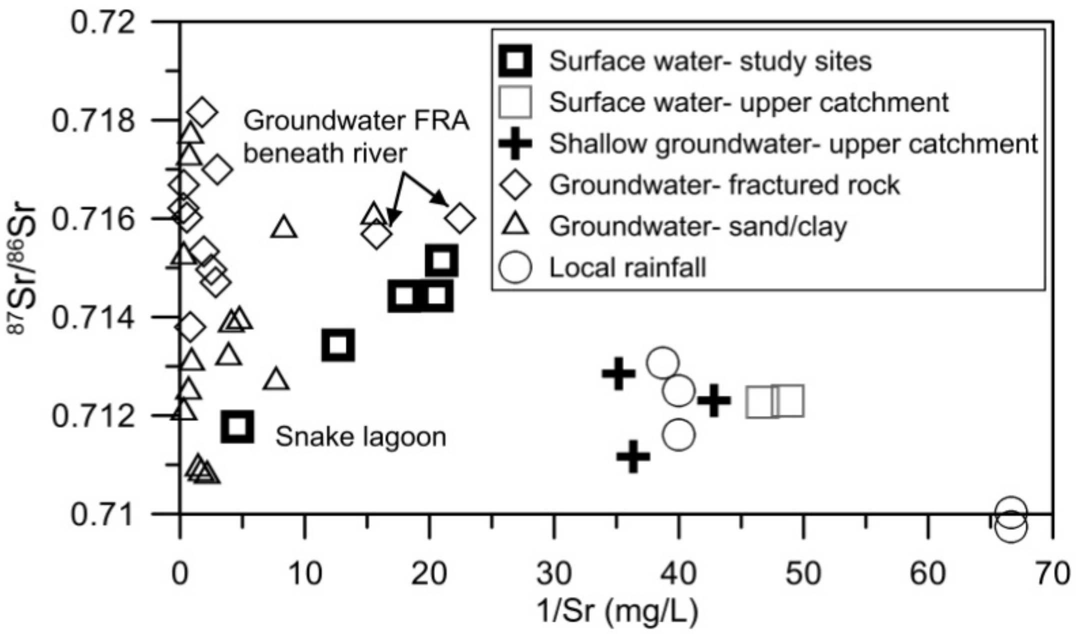

Figure 6. $^{87}Sr/^{86}Sr$ ratio versus 1/Sr for surface water, groundwater and local rainfall in Rocky River Catchment.

CONCLUSION

This study investigated surface water-groundwater connectivity in fractured rock catchments at two sites in South Australia. Results indicate that connectivity was variable across each of the catchments. The influence of the fractured rock aquifer was minimal in the pristine, uncleared RRC whereas in the cleared, mixed land-use CCC, the fractured rock aquifer played a more significant role. This result calls for further research investigating the impacts of land-use change (particularly vegetation clearance) on surface-water groundwater connectivity.

6.4 ASSESSING SURFACE WATER GROUNDWATER CONNECTIVITY – FROM THE HEADWATERS TO THE SEA, KANGAROO ISLAND, SOUTH AUSTRALIA

PRESENTED AT GROUNDWATER 2010- THE CHALLENGE OF SUSTAINABLE MANAGEMENT CONFERENCE, CANBERRA, AUSTRALIAN CAPITAL TERRITORY, NOVEMBER 2010

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Assessments of surface water-groundwater connectivity provide important information for water resource managers to evaluate water resource allocation needs and the impacts on groundwater dependent ecosystems. It is common for connectivity assessments to investigate stream reaches as discreet systems which are generally classified into 3 types: (1) A gaining surface water system where groundwater discharges through the streambed to contribute to streamflow, (2) A losing surface water system which loses water to (or recharges) the groundwater system, and (3) A disconnected system which is defined by an unsaturated zone beneath the surface water system and which loses water at a rate related to the hydrogeologic properties of the streambed and the aquifer. While these classifications are valid, studies often fail to consider how individual stream reaches function in the context of the entire river system (comprising multiple stream reaches) from the headwaters to the sea or discharge point. Applying an entire river system assessment we demonstrate that the system types (or connectivity conditions) can change along

stream reaches, vary depending on the season (variably gaining/losing systems) as well as take place concurrently at the same location and time. These concurrent connectivity types of systems have been rarely studied. By understanding the variable nature of connectivity of an entire river system, more appropriate management practices can be employed. This study investigated connectivity between a fresh water river, a perched sedimentary aquifer and a saline fractured rock aquifer system in the pristine Rocky River Catchment on Kangaroo Island, South Australia. We present preliminary results and insights here. Longitudinal and transverse assessments were made across the entire catchment using hydraulic, hydrochemical, and tracer-based techniques to determine the relative source and loss terms of the river and groundwater system and how their relative magnitude would change along the river from the catchment headwaters towards the sea. Despite what one would expect for a catchment covered by native vegetation in southern Australia, the river is relatively fresh (EC <500 μS/cm) compared to the groundwater of the regional fractured rock aquifer system (EC >6000 μS/cm). Semi-perennial river flow and gaining conditions in the headwaters of the catchment indicated that there is a subsurface fresh water source from either the shallow sedimentary and/or fractured rock aquifer system. Water level data together with salinity and stable isotope results confirmed that the dominant source to the river in the headwaters is from the shallow sedimentary aquifer and that there is minimal influence from the deeper fractured rock aquifer. i.e. the river is gaining from the shallow sedimentary system and has the potential to lose water to the disconnected regional aquifer system below. From the headwaters down towards the bottom of the catchment there was a steady increase in river salinity. River volumetric flows steadily increased from the headwaters downstream then decreased and increased again. This reflects localised changes in the river source and loss terms along the river. The increase in river salinity is likely to be a result of evapotranspiration and/or discharge of higher salinity groundwater. Longitudinal streamflow measurements and continuous water level monitoring at three study sites showed that groundwater discharge from the saline regional aquifer to the river was unlikely. Therefore, the decrease in river flow can be attributed to a decrease in groundwater discharge from the shallow sedimentary system, increased leakage through the riverbed to the regional aquifer below, or surface water loss to evaporation. The stable isotope results indicated that surface evaporation is responsible for the increase in river salinity but is not the major contributor to reduced river flow. Our results show that the relatively low salinity of the fresh water river system can be maintained in an otherwise saline regional groundwater system by virtue of the dominantly losing connectivity state. We hypothesise that there are important vegetation (evapotranspiration) controls on the surface water-groundwater connectivity state of this system.

6.5 EFFECTS OF LAND CLEARANCE AND REVEGETATION ON THE STATE OF CONNECTION BETWEEN SURFACE WATER AND GROUNDWATER

PRESENTED AT 11TH AUSTRALASIAN ENVIRONMENTAL ISOTOPE CONFERENCE AND 4TH AUSTRALASIAN HYDROGEOLOGY RESEARCH CONFERENCE, 12-14TH JULY, 2011, CAIRNS, QLD

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The importance of surface water–groundwater interactions and their state of connection has received greater attention in the last decade, in response to concerns about water scarcity and the sustainable management and allocation of water resources. The state of connection between surface water and groundwater in natural environments is influenced by physical, topographical and hydroclimatic variables. There is limited understanding of the hydroclimatic effects of precipitation and evapotranspiration on surface water–groundwater interaction. More specifically, there is a need for a physically based understanding on the changes that may occur in response to changes of vegetation. Whilst it may seem qualitatively obvious that changes in vegetation could affect the state of connection between surface water and groundwater, it has so far not been demonstrated quantitatively. Also, the influence of parameters such as root extinction depth, topography or the influence of land clearance on the state of connection have so far not been explored. 2-D transient and steady-state models were used to simulate the impact of land clearance and revegetation on the state of connection of a perennial (constant head) river with a clogging layer and homogeneous aquifer. The simulations showed that changes in vegetation can affect the state of connection and that the removal of deep-rooted vegetation from a catchment may have a significant impact on the state of connection between surface water and groundwater as well as the condition of the water resource.