Surface water and groundwater interactions in the Ovens River

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A thesis submitted for the degree of Doctor of Philosophy at
Monash University in 2016
School of School of Earth, Atmosphere and Environment

Declaration

This thesis contains no material which has been accepted for the award of any other degree or diploma at any university or equivalent institution and that, to the best of my knowledge and belief, this thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

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Date: 15th June 2016

Thesis including published works declaration

In accordance with Monash University Doctorate Regulation 17.2 Doctor of Philosophy and Research Master's regulations the following declarations are made:

I hereby declare that this thesis contains no material which has been accepted for the award of any other degree or diploma at any university or equivalent institution and that, to the best of my knowledge and belief, this thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

This thesis includes pages 35 - 90 original papers published in peer reviewed journals. The core theme of the thesis is river-groundwater interactions The ideas, development and writing up of all the papers in the thesis were the principal responsibility of myself, the candidate, working within the School of School of Earth, Atmosphere and Environment under the supervision of Professor Ian Cartwright.

The inclusion of co-authors reflects the fact that the work came from active collaboration between researchers and acknowledges input into team-based research.

In the case of chapter 2 my contribution to the work involved the following:

Thesis chapter	Publication title	Publication status	Nature and extent of candidate's contribution	Co-author name(s) Nature and % of Co-author's contribution	Co-author(s), Monash student Y/N
Chapter 2	Examining the spatial and temporal variation of groundwater inflows to a valley-to-floodplain river using ^{222}Rn, geochemistry and river discharge: Ovens River	Published	88% data collection, analysis, reporting	Ian Cartwright (8%): Supervisory role, data collection assistance, review of manuscript Joseph Braden (2%): Data collection assistance Selene de Bree (2%): Data collection	No Yes Yes

I have renumbered sections of submitted or published papers in order to generate a consistent presentation within the thesis.

Student signature: [redacted] **Date:** 15th June 2016

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the student's and co-authors' contributions to this work. In instances where I am not the responsible author I have consulted with the responsible author to agree on the respective contributions of the authors.

Main Supervisor signature: [redacted] **Date:** 15th June 2016

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Contents

List of Figures

List of Tables

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Abstract

This thesis investigated surface water-groundwater (SW-GW) interactions in the Ovens River, southeast Australia. The Ovens River is hosted within a valley with coarse-grained Quaternary alluvial deposits in the upper catchment and flows across an alluvial floodplain with fine-grained and mature Quaternary to Tertiary sediments in the lower catchment. Electrical conductivity (EC), major ions and radon (^{222}Rn) indicate that the Ovens River is dominantly gaining in the upper catchment and fluctuates between gaining and losing in the lower catchment. The groundwater inflow in the Ovens River, estimated through ^{222}Rn mass balance, is 2 *to* 17% of the total discharge. The groundwater inflows in the upper catchment are higher during high flow periods. The spatial variation of SW-GW interaction is due to the differences in distribution of rainfall, topography and aquifer lithology across the catchment. The temporal variation of baseflow over 10 years in the Ovens River was studied by flow duration curve (FDC), graphical and filter-based hydrograph separation, and EC-derived chloride-based chemical mass balance (CMB). Baseflow fluxes derived from the hydrograph separation methods are significantly greater than those determined by the chloride-based CMB and FDC by 35% *to* 200%. The differences are greatest during and following high flow events. These differences suggest that discharge from transient water stores, such as river banks, the unsaturated zone, and pools or disconnected channels on the floodplain, contribute a significant proportion of baseflow in the Ovens Catchment. The differences in baseflow estimates are caused by the fact that baseflow estimated by hydrograph separation consists of both groundwater inflow and discharge from transient water stores, while CMB and FDC yield groundwater inflows only. Bank storage occurs along most of the river banks at the Ovens River in the middle and lower catchments, except for the steep mid-to-lower valley section of the upper catchment. The indications of bank storage in these areas include the reversed hydraulic head in river banks, the reduction of EC in river banks, the shift in the Na/Cl ratios, major ion concentrations and stable isotope values of the near-river groundwater toward those of the river water following high flow events, and the similarity in tritium concentrations between the water in river banks and the river water at high and moderate flows. Bank storage in the Ovens catchment occurs mainly in areas that have a relatively lower regional hydraulic gradient toward the river and high-to-moderate hydraulic conductivity river bank sediments. The knowledge gained from this thesis will improve our understanding of river-groundwater interactions and has management implications for the Ovens River and other similar rivers.

Chapter 1

Introduction

Examining the interactions between surface water and groundwater is a challenging but important step in understanding hydrological process and for managing surface water and groundwater resources. Surface water-groundwater (SW-GW) interaction is an integral part of the water cycle and can alter the quality and quantity of the water resources in the two respective systems (Winter *et al*., 1998; Brodie *et al.*, 2008). However, these interactions have only been studied in detail in recent years (Cey *et al.*, 1998; Cook et al., 2003; Lamontagne *et al.*, 2005; Andersen and Acworth, 2009; Cartwright *et al.*, 2011; 2014; McCallum *et al.*, 2012; 2013; Unland *et al.*, 2013). This neglect has led to a long-standing water resource management practice in which surface water and groundwater resources are managed separately (National Groundwater Committee, 2004). The consequence of such practice had been detrimental. For example, groundwater had been often allocated without considering the connected rivers, lakes and wetlands in the catchment, resulting in the depletion of water in the surface water bodies, which eventually affects surface water users and ecosystems (Weber and Perry, 2006; Evans and Merz, 2007; Glennon, 2012; McCallum *et al.*, 2013). The lack of understanding of SW-GW interaction had also led to the failure of appreciating the implications of some environmental problems. Land clearing in Australia, for example, has resulted in rising water tables which causes detrimental effects such as dryland salinity (Cartwright *et al.*, 2004). However, dryland salinity also salinizes the previously fresh rivers due to the increase in the saline groundwater inflows which damages their ecological health (Callow and Smettem, 2007).

In this introductory chapter, an overview of SW-GW interaction in a riverine setting and on the commonly used methods for investigating river-groundwater interaction will be presented. Major issues in river-groundwater interaction will also be identified. This chapter will then describe how these issues will be addressed by stating overall aims of this thesis and brief descriptions of three studies on river-groundwater interaction.

1.1 Overview of topics

In essence, all surface water and groundwater systems on the Earth are connected as water moves from one system to another in the water cycle (Winter *et al.*, 1998; Sophocleous, 2002). However, the degree of connectivity between the two systems varies and is controlled largely by geomorphology, geology and climate (Sophocleous, 2002). Surface water and groundwater systems are considered to be hydraulically connected when the streambed is directly in contract with the underlying aquifer via a zone of water saturated sediments (Winter *et al.*, 1998). In connected systems, the SW-GW interaction can be grouped into two categories: a gaining system where the regional water table is above the water level of the surface water system, and the surface water system receives groundwater as baseflow; and a losing system where the water level in the surface water system is above the regional water table, and water from the surface system infiltrates to the underlying aquifer as recharge (Fig. 1.1a & b). The definition of a disconnected system is,

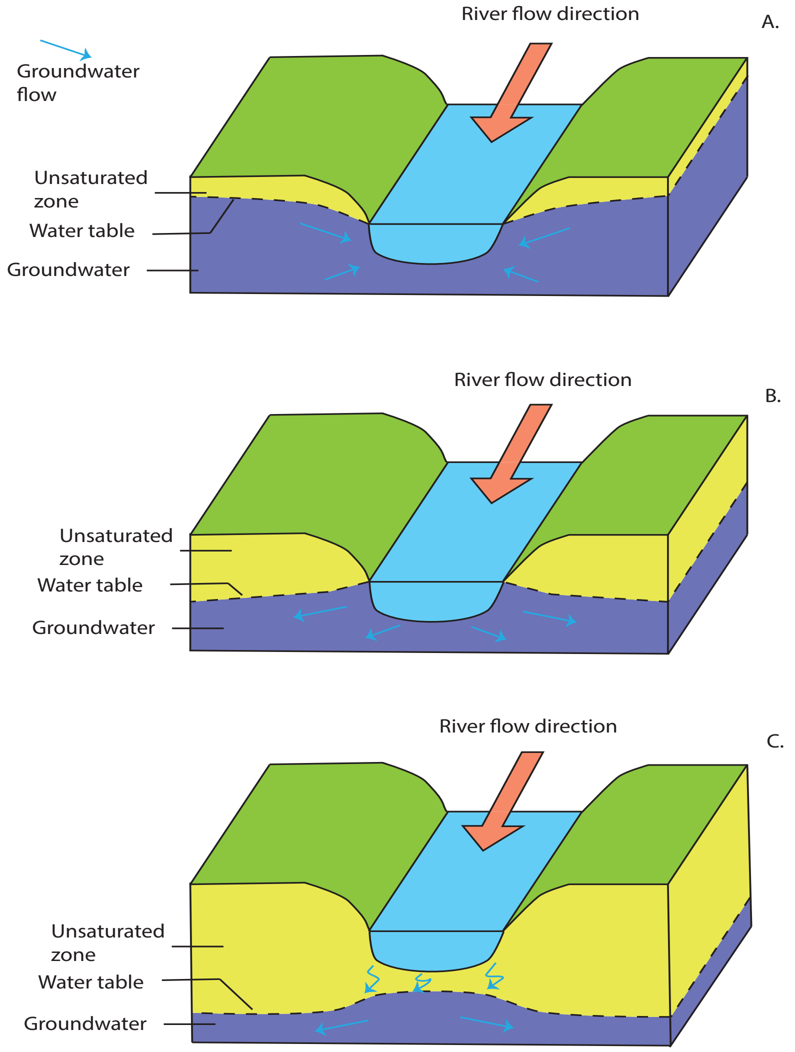


Figure 1.1 Types of surface water and groundwater interactions in rivers. Connected systems: gaining stream (A) and losing streams (B). Disconnected system (C). Adapted from Winter *et al.* (1998).

however, less clear. A common consensus is that surface water and groundwater systems are considered as a disconnected system when an unsaturated zone exists beneath the surface water system, and the infiltration rate from the surface water system across its width does not change in response to the lowering of water table (Fig. 1.1c) (Brunner *et al.*, 2009; 2011).

The understanding of SW-GW interaction in riverine settings at small scales has greatly improved due to the increasing number of studies focusing on individual river reaches (Genereux *et al.*, 1990; Cey *et al.*, 1998; Lambs, 2004; Cox *et al.*, 2006; Andersen and Acworth, 2009). These studies have been helpful in conceptualising river-groundwater interactions and defining some of the major factors that control river-groundwater interactions. The three main controls on river-groundwater interactions are the basin morphology and the position of the river channel within the landscape, the hydraulic conductivities of the river channel and adjacent alluvial aquifer, and the river stage relative to the level of the water table in the adjacent aquifer (Sophocleous, 2002; Pritchard, 2005). Several studies have showed that incised river reaches with steep banks or in regions of steep topography receive a greater volume of groundwater inflow because of a high hydraulic gradient existing between the river and the water table in these areas (Unland *et al.*, 2013; Atkinson *et al.*, 2015; Cartwright and Gilfedder, 2015). The rate and volume of SW-GW exchange in rivers increases with increasing hydraulic conductivity (Morrice *et al.*, 1997; Chen and Chen, 2003). Any factors affecting the water table or the stream height will also alter the connectivity. These factors include groundwater extraction, stream regulation and alteration, climate change and land use. For instance, low rainfall in recent years in south-east Australia has resulted in a significant drop in the water level of the Murray River, and this has led to reduced recharge of the adjacent low-salinity groundwater lenses in the basin (Cartwright *et al.*, 2011).

Many operational decisions on environmental flow and water allocation are typically made at the reach-scale, resulting many hydrological studies on river-groundwater interaction being conducted at smaller scales, such as several river reachs or subcatchments (Lovell, 2009). However, only a few studies have considered the connectivity between a river and its underlying aquifers in context of the entire catchment, from the headwaters to the discharge point (Braaten and Gate, 2003; Covino and McGlynn, 2007; Bank *et al.*, 2011). It is often easy to overlook the fact that individual river reach functions in the context of the entire river system (comprising multiple river reaches) from the headwaters to the sea or discharge point (Bank *et al.*, 2011). Understanding river-groundwater interactions at a catchment scale provides a more accurate assessment of the water and salt balances within a catchment. This translates to a better water allocation for both human and environmental consumption. One important implication of acknowledging river-groundwater interaction at a catchment scale is that the connectivity between a river and its underlying aquifers can vary along the river because of the variation in basin geomorphology and local climate along the river (Braaten and Gate, 2003; Payn *et al.*, 2009). Water can be gained in some reaches while being lost to the underlying aquifers in other reaches. The variation of river-groundwater interactions in several rivers in the Murray-Darling Basin in New South Wales was mapped using hydrometric data (Braaten and Gate, 2003). The data indicated that the narrow alluvial valley and high rainfall in the mid-sections of the Murray-Darling catchments produce shallow water table, creating a strong hydraulic connection between the rivers and its aquifer. In contrast, when the constricted mid-sections of the Murray-Darling catchments enter the wider semi-arid plains of the lower valleys, the water table falls as the result of the wider topography. The surface water and groundwater systems in the lower valleys then become disconnected.

The categorization of SW-GW interactions described above is useful in understanding and conceptualising river-groundwater interaction. However, it may also provide an incorrect assertion that the connectivity between a river and its underlying aquifers is static. The connectivity can alter with time, often in response to seasonal and decadal changes in rainfall. Many streams recharge the adjacent aquifers during high flow periods when the stream stage is higher than the regional water table. During low flow periods, the stream level drops below the regional water table, resulting in groundwater influx to the river from the adjacent aquifers. There have been studies examining how the river-groundwater connectivity varies with seasonal changes (Malcolm *et al.*, 2005; Song et al., 2006; Baskaran *et al.*, 2009; Cartwright *et al.*, 2011). However, some of these studies have limited sampling rounds (mostly one during the high flow period and one during the low flow period), while others may use one set of geochemistry data to infer the possible changes in the connectivity status between the two systems in different flow periods. The temporal variation

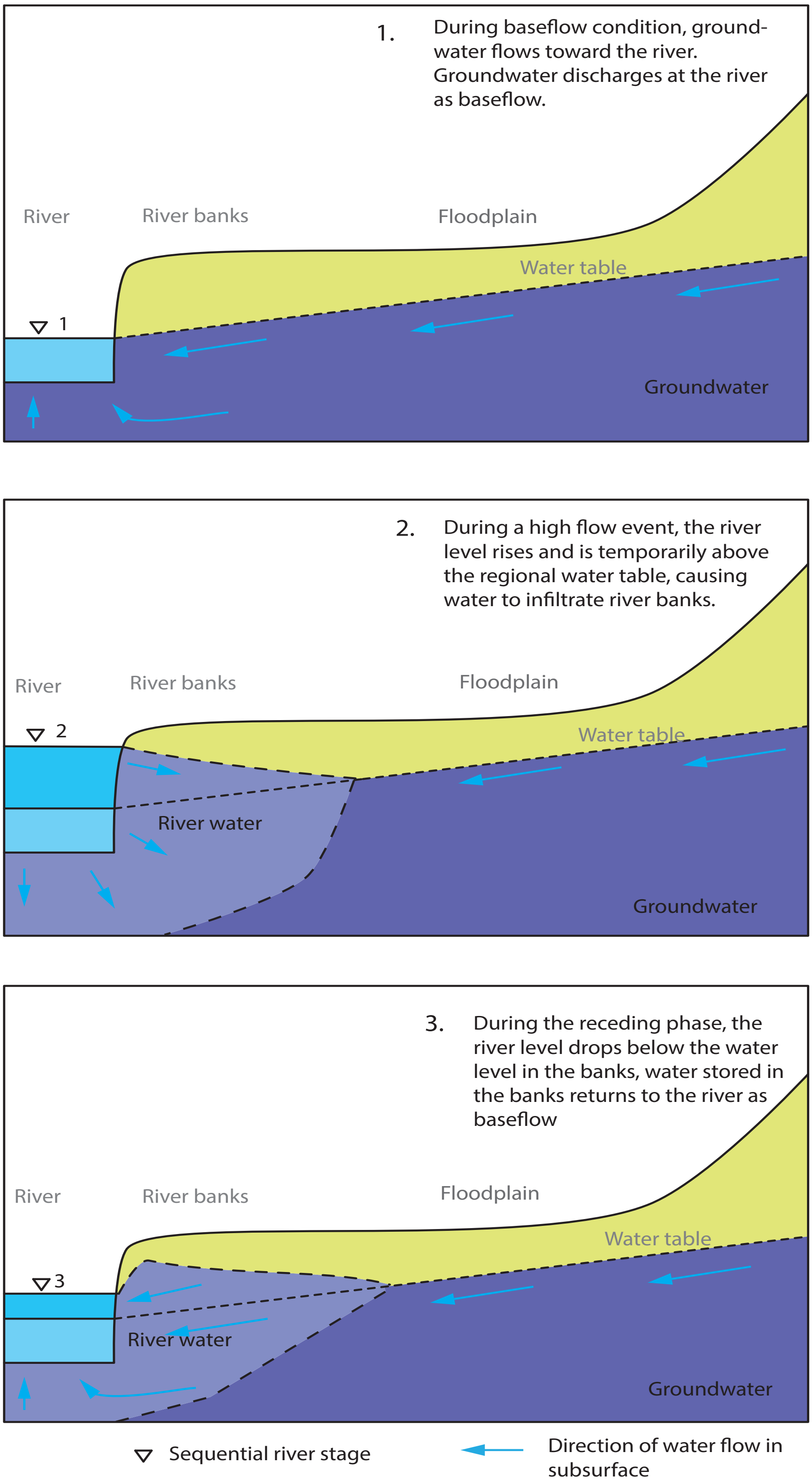
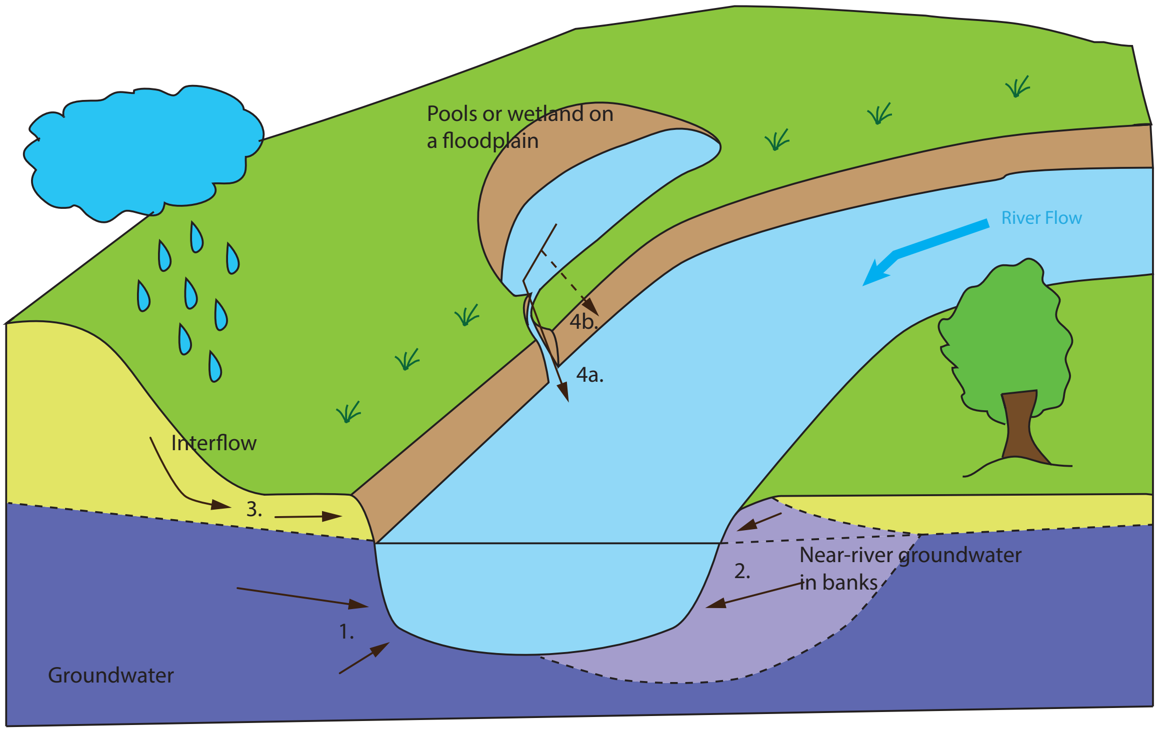


Figure 1.2 Process of bank storage. Adapted from Winter et al. (1998).

in river-groundwater interactions can also occur over a short to medium timeframe (Winter *et al.*, 1998). One example is a storm event. During a storm, the increase in the stream stage temporarily reverses the hydraulic gradient away from the stream in a gaining stream, causing water to infiltrate from the stream into the stream banks. As the stream stage declines during the recession period, the pre-storm hydraulic gradient is re-established, and the stored water in the banks is discharged back into the river (Fig. 1.2). Bank storage can attenuate the flood wave during the onset of an event flow as well as contributing flow to the stream between flow events. The discharge from river banks can occur over a period of days to months. Bank storage has implications for protecting riparian and riverine ecosystems, and estimating catchment water balance (Lambs, 2004; Lamontagne *et al.*, 2005). Bank storage and its transient impacts on rivers have been mainly studied using analytical and numerical techniques, but field studies, particularly those based on geochemistry, in these areas is still lacking (Squillace, 1996; Whiting and Pomeranets, 1997; Chen and Chen, 2003). Recent studies have further explored the effects of river bank slope and unsaturated flow on bank storage (Doble *et al.*, 2012) and the solute movement within river banks during the rise and fall of river stage (McCallum *et al.*, 2010) using a combination of numerical modelling and field observations. The ability of using geochemistry to define bank storage is important because only geochemistry can accurately describe the movement of river water within river banks through water mixing between river water and regional groundwater (Welch *et al.*, 2013; 2014).

River-groundwater interaction is sometimes viewed as comprising "two buckets" (that is groundwater and surface water) with water being transferred from one bucket to another (Song *et al.*, 2006; Akiyama *et al.*, 2007). The "two buckets" framework may be valid in some hydrological systems, but the majority of hydrological systems contain other water stores aside from groundwater and surface water (Fig. 1.3). Some studies have examined the contribution of various sources of groundwater inflow to a stream, primarily from regional and local aquifers



Various baseflow components:

1. Groundwater inflow from regional aquifers
2. Bank return flow
3. Water from the unsaturated zone (inter-flow)

4a. drainage from pools or wetlands on a floodplain via channel
4b. drainage from pools or wetlands on a floodplain via the sediments

Figure 1.3 Various sources of water stores that contribute to the river flow during baseflow condition in a catchment. Groundwater is one of the common baseflows. Other includes bank return flow, water from unsaturated zone and water from pools on a floodplain.

(Cook *et al.*, 2003; Brodie *et al.*, 2007; Smerdon *et al.*, 2013). As suggested earlier, water contained in stream banks is another important component in a catchment. Recently recharged water may mix with regional groundwater in the river banks during high flow condition and is returned to the river as parts of baseflow during the recession period. There are also other near-stream water reservoirs, including the unsaturated zone, wetlands and pools on the floodplains (Griffiths and Clausen, 1997; Hofmann, 2011; Cartwright *et al.*, 2014). Water from these reservoirs can interact with the river in addition to the interaction between the stream and its underlying aquifers. If these components are not carefully accounted for, it can lead to a wrong conclusion on the actual proportion of groundwater inflow in a river. For example, an increase in streamflow between two gauging stations is often interpreted as groundwater influx along the river reach. However, in reality, the water may be derived from groundwater inflows, bank flow return, discharge from the saturated zone, drainage from pools on floodplains or a combination of all these sources. Without a better understanding of these water components, it will not be possible to appreciate the true nature of river-groundwater interaction in a catchment. This, in turn, can lead to incorrect catchment water balance, resulting in an inappropriate water management plan.

One important aspect of river-groundwater interaction is the quantification of water exchange between the river and its aquifer. There are several methods of estimating baseflow flux, and these methods operate over a range of temporal and spatial scales (Kalbus *et al.*, 2006; Brodie *et al.*, 2007). One method which is often seen as a benchmark method is to use the Darcy Law. However, the quantity of water exchanged between rivers and aquifers may not be just controlled by the hydraulic conductivity of aquifers but also that of river bed sediments. These sediments are commonly finer grained and of lower hydraulic conductivity than the aquifers. Estimating hydraulic conductivity can be difficult since the magnitude and direction of hydraulic conductivity are dependent on the highly anisotropic and heterogeneous nature of aquifer and river channel materials (Fetter and Fetter, 2001). Furthermore, the Darcian approach may sometimes be too insensitive to estimate SW-GW exchange if the aquifer transmissivity (and therefore groundwater flow) is large in relative to surface water flow (Andersen and Acworth, 2009). Finally, the baseflow flux is not linearly related to changes in the hydraulic head gradient because of large variations in river flow, channel geometry, wetted perimeter and clogging layers in river channels (Sophocleous, 2002).

One alternative and commonly used method in estimating baseflow is hydrograph separation. A stream hydrograph is a time-series record of river discharge, showing the two major flow components; quickflow and baseflow. The technique relies on surface runoff varying over short time periods, whereas baseflow is assumed to vary over longer timescales. Hydrograph separation separates baseflow from quickflow by either graphical approaches or digital filters. The graphical approach systematically draws connecting lines between the lowest points of the streamflow hydrograph at which surface runoff stops and the river is mainly fed by baseflow (Sloto and Crouse, 1996; Aksoy *et al.*, 2009). The sequence of these connecting lines defines the baseflow flux. The filter-based approach adapts the filtering technique in signal analysis and processing; it derives the low-frequency baseflow by low-pass filtering the hydrograph to remove the high frequency responses of the hydrograph which are usually associated with the variable quickflow (Nathan and McMahan, 1990; Eckhardt, 2005). Hydrograph separation is a popular method in the water resource industry because it is easy to use, and the discharge datasets are abundant and available without a significant cost. However, hydrograph separation has no physical basis (Hewlett and Hibbert, 1967; Freeeze, 1972). Furthermore, baseflow determined by this method not only includes groundwater inflows but also discharge from all transient water stores (for example, bank return flows, the unsaturated zone, and draining of billabongs and pools on floodplains) because the latter also contributes to the river on timescales longer than surface water (Nathan and McMahan, 1990; Evans and Neal, 2005; Cartwright *et al.*, 2014). Finally, hydrograph analysis is not suitable for highly regulated rivers. Hydrograph separation has been used in a number of Australian studies, such as those in the Murray-Darling Basin, and the quality of results varies, depending on the completeness of flow data and site selection (Neal *et al.*, 2004; Ivkovic *et al.*, 2005; Cook *et al.*, 2010; CSIRO and SKM, 2010).

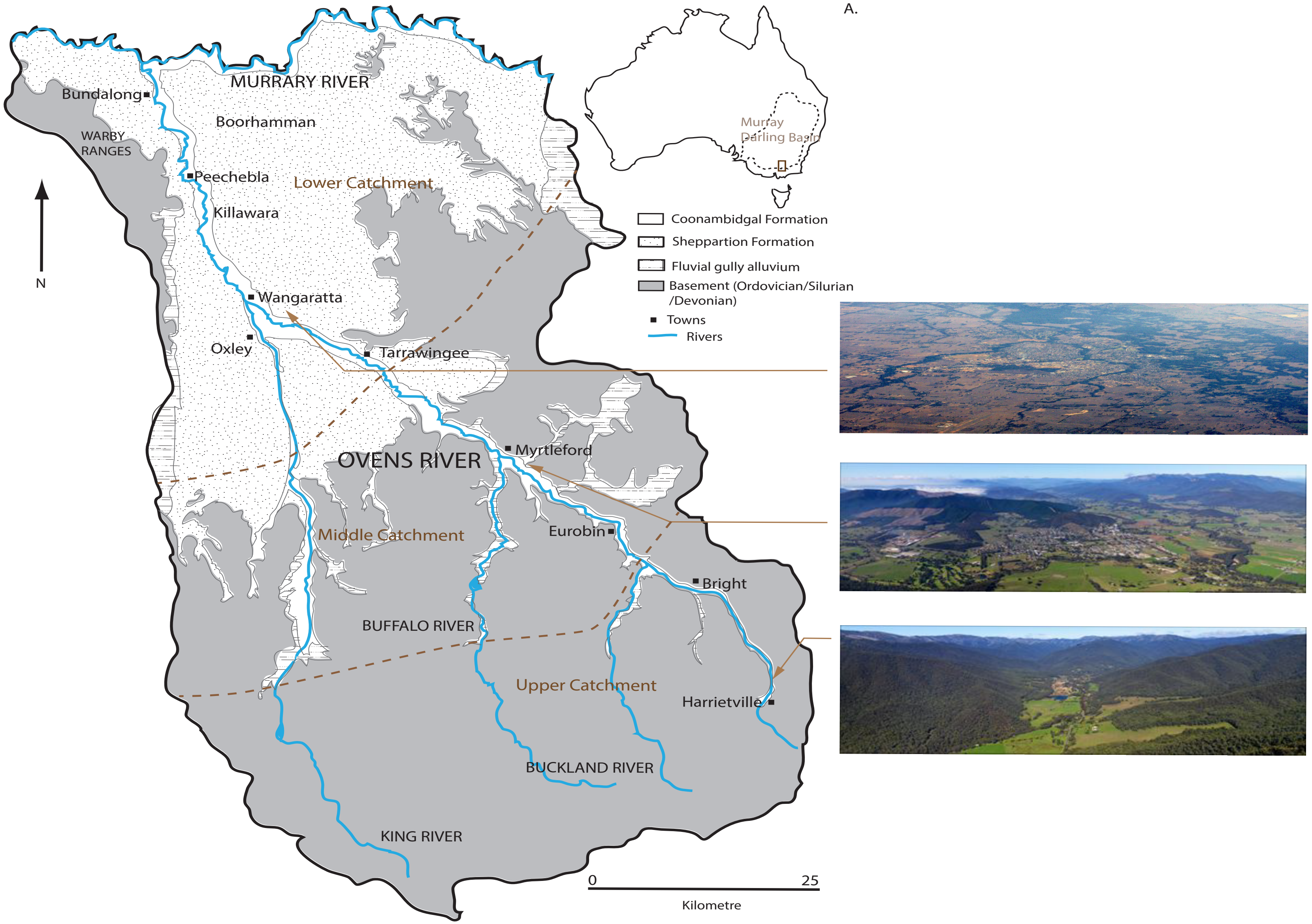
Baseflow can also be estimated by river geochemistry data through chemical mass balance (CMB). The CMB can be based on a variety of data, such as electrical conductivity (EC), major ions, stable and radioactive isotopes, gases, nutrients, and contaminants. One advantage of using river geochemistry is that these chemical components evolve along the water flow path, providing more accurate information on water evolution, residence times or mixing ratios that would otherwise be difficult to determine using physical hydrological data (Cook, 2013). Geochemistry data is also capable to provide information on both the temporal and spatial variations in baseflow (Cook *et al.*, 2003; Cartwright *et al.*, 2011; Gilfedder *et al.*, 2012; Cook, 2013). The former is achieved by measuring the stream chemistry at one location over time, while the latter is done by longitudinal stream chemical sampling.

Another important aspect of using geochemistry in estimating baseflow is that it usually yields estimates of groundwater inflow rather than the total baseflow flux because the geochemistry of water in transient water stores is similar to that of surface water from which they are derived. The accuracy of baseflow estimation using chemistry data-based methods depends on several factors, including the difference in the concentrations of the chemical tracer between groundwater and surface water, the ability of accurately characterising the source and sinks of the tracer, and the ability of predicting any in-stream processes that alter the concentration of the tracer. Chloride (Cl) is often used in preference to other geochemical tracers because it behaves conservatively, and it is the most abundant element in water. EC is very often used in chemistry surveys because it can be easily measured over a period of time. Another useful tracer is radon (^{222}Rn). ^{222}Rn is the decay product of ^{226}Ra in the uranium decay series with a relatively short half-life of 3.825 days. Groundwater often has a relatively high ^{222}Rn activity because aquifer matrices have a high level of radium-bearing minerals, and the activity of ^{222}Rn in groundwater reaches secular equilibrium with ^{226}Ra in the aquifer matrix over a few weeks (Cecil and Green, 1999). Conversely, the ^{222}Rn activity in surface water is usually low because of low dissolved ^{226}Ra activities, the relatively short half-life of ^{222}Rn and the rapid degassing of ^{222}Rn to the atmosphere. These unique characteristics enable ^{222}Rn to be used for identifying groundwater inflow in surface water systems. Unlike Cl and EC, ^{222}Rn does not accumulate in water and is therefore a better technique in identifying zones of groundwater inflow along the river. The use of ^{222}Rn as a groundwater tracer has increased over the last few decades as methods for measuring Rn in the field, including the ability to measure continuously, have improved (Ellins *et al.*, 1990; Mullinger *et al.*, 2007; Cartwright *et al.*, 2011; Gilfedder *et al.*, 2012; Cartwright *et al.*, 2015).

In general, SW-GW interaction studies employ one method to quantify groundwater influx. Only few studies have attempted to quantify groundwater influx by using two or more methods (Unland *et al.*, 2013; Cartwright *et al.*, 2014). Some studies use multiple techniques to calibrate one method with another method(s) for estimating baseflow (Stewart *et al.*, 2007; Gonzales *et al.*, 2009). These studies indicate that various methods often provide very different groundwater influxes with physical techniques, such as hydrograph separation, producing much higher estimates than other methods. Reconciling these differences may provide a further insight into river-groundwater interaction and derives more accurate flux estimations for the purpose of water allocation (Hofmann *et al.*, 2011).

1.2 Aims

This thesis aims to explore the issues described above through three related studies of river-groundwater interactions in the Ovens River. The Ovens Catchment is located in the southeast fringe of the Murray-Darling Basin in Victoria, Australia (Fig. 1.4a). It is characterised by the transition from alpine valleys with coarse-grained Quaternary alluvial deposits in the southeast to an alluvial floodplain with fine-grained and mature Quaternary to Tertiary sediments in the northwest (Fig. 1.4a & b). Rainfall declines from the southeast alpine region to the northwest, which is semi-arid. Although the Ovens catchment only occupies 0.7% of the Murray-Darling Basin, it supplies 11% of the total flow to the socially, economically and ecologically important Murray River (CSIRO, 2008). Understanding the river-groundwater interactions in the catchment



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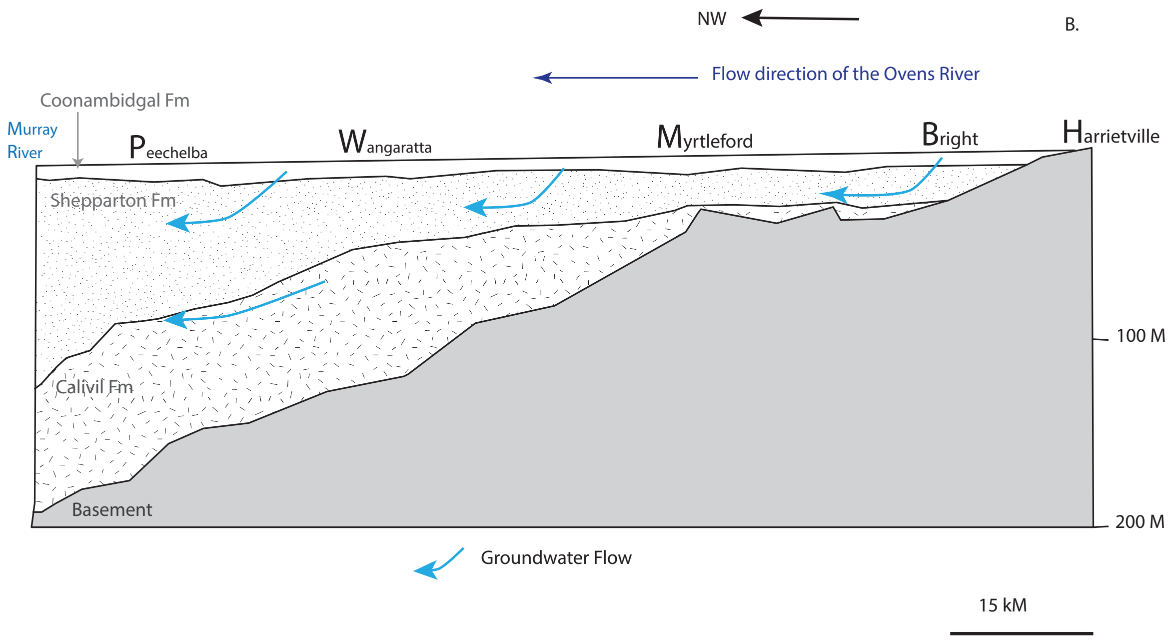

Figure 1.4 (cont.) (A) The township, the surface geology and the catchment physiograph of the Ovens catchment. The photos show the changes in catchment physiology, from narrow valleys in the upper catchment to a mature and wide floodplain in the lower catchment. (B) The simplified geology cross-section of the Ovens Catchment along the Ovens River, showing the geometry of the major stratigraphic units. Data from Lawrence (1988), van den Berg and Morand (1997) and Water Measurement Information System (2013).

will help us to understand the hydrological system in the catchment and in turn assists in better managing the water resources in the catchment and at the Murray River. The knowledge gained in the three studies may be also applicable to other catchments and enhances our general understanding of river-groundwater interaction.

The specific aims of this thesis to

- examine the spatial and temporal variations of groundwater influxes along the Ovens River using environmental tracers and discharge data;
- calculate and compare the magnitude of the long-term baseflow in the Ovens Catchment between numerical techniques and chemical mass balance; and
- characterise the bank storage in the Ovens Catchment in terms of location, size and timing of recharge and discharge.

1.3 Description of studies and structure of the thesis

These aims are achieved by presenting the results from three separate but interrelated studies. The study in chapter 2 explored the spatio-temporal variation of river-groundwater interactions in the Ovens River at catchment scale using ^{222}Rn through longitudinal stream chemical sampling at various flow conditions over 26 months. Results from ^{222}Rn were interpreted along with groundwater hydraulic heads, EC, and major ion geochemistry in the catchment. The controls of river-aquifers interactions were explained. Baseflow fluxes were estimated using ^{222}Rn-based CMB and briefly compared with other baseflow estimation methods, including Cl-based CMB, differential flow gauging and hydrograph separation.

The study in chapter 3 utilised the EC-derived Cl concentration and discharge datasets collected near the discharge point of the Ovens River to estimate long-term baseflow flux in the river over a 10-year period. In addition, the relationship between Cl concentrations and discharge was explored to further constrain the characteristics of baseflow in the river. Unlike the study in chapter 2 which focuses primarily on the spatial aspect of river-groundwater interactions, this study examined the temporal variation of baseflow in the river over an extended period of time. Baseflow estimates were calculated using the flow duration curve, graphical and filter-based hydrograph separation and Cl-based CMB. These baseflow estimates were compared, and possible reasons for the differences in baseflow estimates and their implications were given.

While baseflow commonly include groundwater, it also contains discharge from transient water stores, such as river banks. The study in chapter 4 examined and defined the process of bank storage at the Ovens River using hydraulic heads, geochemistry (including EC, major ion chemistry and stable isotopes) and tritium. The controls of the bank storage process in the catchment were explained.

Chapter 5 is the conclusion chapter of this thesis, providing a summary of findings from the three studies and the implications of these findings on river-groundwater interaction and water catchment management.

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Chapter 2

Examining the spatial and temporal variation of groundwater inflows using ^{222}Rn, geochemistry and river discharge: the Ovens River, southeast Australia

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Abstract

Radon (^{222}Rn) and major ion geochemistry were used to define and quantify the catchment-scale surface water-groundwater interactions along the Ovens River in the southeast Murray-Darling Basin, Victoria, Australia, between September 2009 and October 2011. The Ovens River is characterized by the transition from a single channel within a mountain valley in the upper catchment to a multi-channel meandering river on flat alluvial plains in the lower catchment. Overall, the Ovens River is dominated by gaining reaches, receiving groundwater from both alluvial and basement aquifers. The distribution of gaining and losing reaches is governed by catchment morphology and lithology. In the upper catchment, rapid groundwater recharge through the permeable aquifers increases the water table. The rising water table, referred to as hydraulic loading, increases the hydraulic head gradients toward the river and hence causes high baseflow to the river during wet (high flow) periods. In the lower catchment, lower rainfall and finer-grained sediments reduce the magnitude and variability of hydraulic gradients between the aquifers and the river, producing lower but more constant groundwater inflows. The water table in the lower reaches has a shallow gradient, and small changes in river height or groundwater level can result in fluctuating gaining and losing behaviour. The middle catchment represents a transition in river-aquifer interactions from the upper to the lower catchments. High baseflow in some parts of the middle and lower catchments is caused by groundwater flowing over basement highs. Mass balance calculations based on ^{222}Rn activities indicate that groundwater inflows are 2 *to* 17% of total flow with higher inflows occurring during high flow periods. In comparison to ^{222}Rn activities, estimates of groundwater inflows from Cl concentrations are higher by up to 2,000% in the upper and middle catchment but lower by 50 *to* 100% in the lower catchment. The high baseflow estimates using Cl concentrations may be due to the lack of sufficient difference between groundwater and surface water Cl concentrations. Both hydrograph separation and differential flow gauging yield far higher baseflow fluxes than ^{222}Rn activities and Cl concentrations, probably indicating the input of other sources to the river in additional to regional groundwater, such as bank return flow.

KEYWORDS: River-groundwater interactions, Groundwater inflow, Radon, Chemical mass balance, Ovens River

2.1 Introduction

Defining the relationship between rivers and adjacent groundwater systems is a crucial step in developing programs and policies for protecting riverine ecosystems and managing water resources. Rivers interact with various water stores, such as groundwater in local and regional aquifers, water in river banks, water in the unsaturated zone, and soil water (Turner *et al.*, 1987; Genereux *et al.*, 1993; Winter *et al.*, 1998; Oxtobee and Novakowski, 2002; Sophocleous, 2002; Lamontagne *et al.*, 2005). Losing streams recharge groundwater, while gaining streams receive groundwater as baseflow. The status of a river can vary along its course with topography, for example rivers may be gaining in narrow valleys in the hills but losing when they flow across the broad plains (Winter *et al.*, 1998; Braaten and Gate, 2003; Bank *et al.*, 2011; Guggenmos *et al.*, 2011). Furthermore, the direction and magnitude of water fluxes can change over time; a gaining stream, for instance, can become a losing one if the river rises above the water table during a storm event (Todd, 1980; Winter *et al.*, 1998; Cartwright *et al.*, 2011; Rosenberry *et al.*, 2013). The three main controls on catchment-scale surface water-groundwater (SW-GW) interactions are: (1) the basin morphology and the position of the river channel within landscape; (2) the hydraulic conductivities of the river channel and adjacent alluvial aquifer; and (3) the relation of the river stage to the level of the water table in the adjacent aquifer, which is closely related to precipitation patterns (Woessner, 2000; Sophocleous, 2002; Ransley *et al.*, 2007). Without a sound understanding of SW-GW interactions in a catchment, it is not possible to identify potential pathways for water contamination and to calculate hydrologic budgets for water allocation. The latter has become an important issue in Australia because of the growing demands from both humans and the environment in a drought-affected continent.

2.1.1 Quantifying SW-GW interactions

SW-GW interactions can be investigated by several techniques. Hydrograph separation is a straightforward method for assessing baseflow at a catchment scale. However, it cannot be used for losing or highly-regulated systems, and the slowflow component isolated by the method may aggregate several water storages (such as, bank return flow or interflow) rather than representing only regional groundwater inflow (Griffiths and Clausen, 1997; Halford and Mayer, 2000; Evans and Neal, 2005). Geochemistry, such as major ion concentrations, stable isotopes and radiogenic isotopes, may also be used to quantify groundwater inflows in gaining streams (Brodie *et al.*, 2007; Cook, 2012). The requirements for using geochemical tracers to quantify groundwater inflows are that the concentration of the tracer in groundwater is significantly different to that in river water and that concentrations in groundwater are relatively homogeneous (or that any heterogeneities are known). Radon (^{222}Rn) is a powerful tracer for examining SW-GW interactions from both qualitative and quantitative perspectives (Ellins *et al.*, 1990; Cook *et al.*, 2006; Baskaran *et al.*, 2009; Cartwright *et al.*, 2011). ^{222}Rn is a radiogenic isotope produced from the decay of ^{226}Ra in the uranium decay series. The ^{222}Rn activity in surface water is usually low because of low dissolved ^{226}Ra activities, the relatively short half-life of ^{222}Rn (3.825 days) and the rapid degassing of ^{222}Rn to the atmosphere. Groundwater has ^{222}Rn activities that are commonly two to three orders of magnitude higher than those of surface water due to the near-ubiquitous presence of U-bearing minerals in the aquifer matrix. Due to the short half-life, the activity of ^{222}Rn in groundwater reaches secular equilibrium with ^{226}Ra over two to three weeks (Cecil and Green, 1999). The high contrast between groundwater and surface water activities makes ^{222}Rn a useful tracer of groundwater inflow into rivers, especially where the difference in major ion concentrations between groundwater and surface water is small, such as in many upper catchment streams.

The change in ^{222}Rn activities in a gaining stream (dC_r/d_x) is governed by groundwater inflow, in-stream evaporation, hyporheic exchange, degassing, and radioactive decay:

$$Q\frac{dC_r}{dx} = I(C_i - C_r) + wEC_r + F_h - kdwC_r - \lambda dwC_r \qquad \textit{Eq. 2.1}$$

(Cook *et al.*, 2006; Mullinger *et al.*, 2007). In Eq. 2.1, Q is the stream discharge (m^3 day^{-1}), C_r is the ^{222}Rn activity within the stream (Bqm^{-3}), x is distance in the direction of flow (m), I is the groundwater inflow rate per unit of stream length (m^3 m^{-1} day^{-1}), C_i is the ^{222}Rn activity in the inflowing groundwater, F_h is the flux of ^{222}Rn from hyporheic zone (Bqm^{-1} day^{-1}), w is the width of the river surface (m), d is the mean stream depth (m), E is the evaporation rate (m day^{-1}), k is the gas transfer coefficient (day^{-1}), and λ is the radioactive decay constant (0.181 day^{-1}). Groundwater inflow can be calculated by rearranging Eq. 2.1. Equation 2.1 can also be used for other tracers. For major ions, such as sodium or chloride, that do not degas to the atmosphere or decay, the last two terms on the right-hand side are redundant.

2.1.2 Aims

This study uses ^{222}Rn activities and major ion geochemistry in conjunction with physical hydrological data to determine the SW-GW relationships and the contribution of baseflow along the Ovens River (Fig. 2.1) from its upper catchment to its discharge point at the Murray River. The study covers a period of 26 months that include the end of the Millennium drought (2001 *to* 2009) (van Dijk et al., 2013) and the 2010 Victorian floods; these are typical of floods that recur on average every 10 *to* 20 years (Comrie, 2011). From hydraulic heads and river heights, CSIRO (2008) indicated that the Ovens River is gaining in the upper catchment, alternately gaining and losing in the middle catchment and mainly losing in the lower catchment. However, the precise distribution of gaining and losing reaches, the temporal variation of SW-GW exchange and the baseflow fluxes to the river remain unknown. The results will provide an important background for future SW-GW studies in this and other catchments in the Murray-Darling Basin, and elsewhere.

2.2 Study area

Located in the south-east margin of the 1,061,469 km^2 Murray-Darling Basin, the Ovens Catchment (Fig. 2.1) occupies just 7,813 km^2 but contributes 6 *to* 14% of the total flow of the Murray River (CSIRO, 2008). The Ovens River is the main river in the Ovens Catchment;

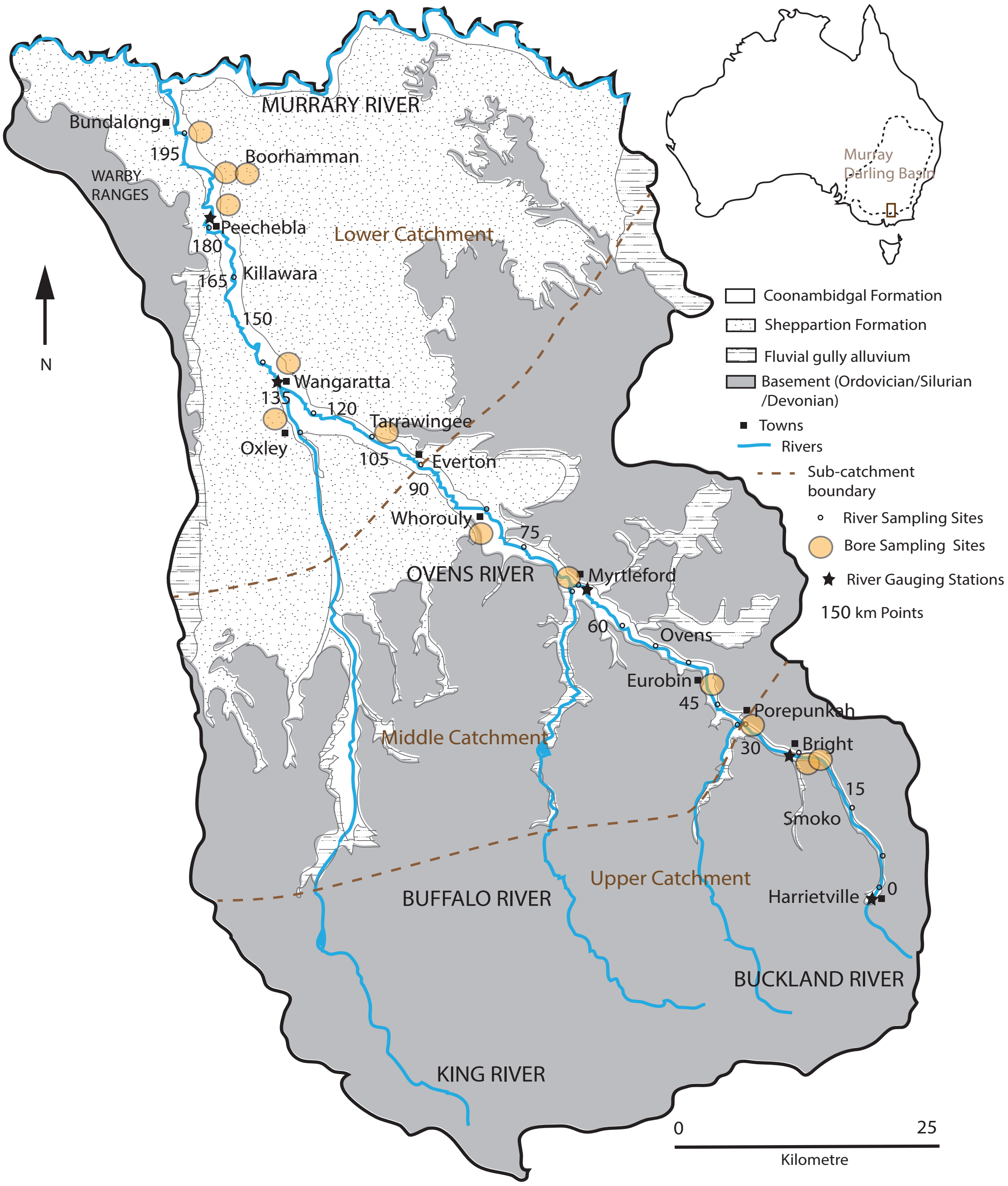

Figure 2.1 Map of the Ovens River Catchment showing surface geology, sampling points and gauging stations. Data from van den Berg and Morand (1997); Water Resources Data Warehouse (2011)

it is approximately 202 km long and originates on the northern flanks of the Victorian Alps and flows north-westwards. The catchment is characterised by multiple narrow V-shaped mountain valleys in the upper catchment and broad flat alluvial flood plains in the lower catchment. In the upper catchment, the river is 5 *to* 10 m wide and 1 *to* 2m deep. It has small rapids with a steep channel gradient of around 6.5 m km^{-1} (Victorian Government Department of Sustainability and Environment, 2010a). Downstream of Porepunkah, the valley broadens and transitions into open alluvial flood plains. The river in the lower catchment has a low gradient of less than 1m km^{-1} and develops a network of meandering and anastomosing channels downstream of Everton (Victorian Government Department of Sustainability and Environment, 2010a). In its lower reaches, it is 40 *to* 50 m wide and up to 8 m deep. It flows past the Warby Ranges before discharging to the Murray River at Bundalong. The Ovens River is perennial and receives water from three main tributaries: the Buckland, Buffalo and King Rivers. The monthly discharge at the Peechelba gauging station located toward the discharge point varies between 200 and 30,200 ML day^{-1} with high flow occurring in Australian winter months (June to September) (Victorian Water Resource Data Warehouse, 2011). The river in the upper and middle regions is unregulated, but the flow downstream is partially regulated due to the storages on the Buffalo and King tributaries.

The stratigraphy of the Ovens Catchment comprises Palaeozoic basement overlain by Tertiary to recent fluviatile sediments (Lawrence, 1988; van den Berg and Morand, 1997). The depth to basement in the upper and middle catchments is generally 10 *to* 50 m, while the depth to basement is up to 200 m in the lower catchment. Several basement highs and local outcrops exist at Myrtleford in the middle catchment and between Killawarra and Peechelba in the lower catchment. The basement predominantly consists of metamorphosed Ordovician turbidites intruded by Silurian and Devonian granites that form a fractured-rock aquifer with a hydraulic conductivity of 0.3 *to* 10 m day^{-1} and a transmissivity of $< 10\ m^2\ day^{-1}$ (Slater and Shugg, 1987).

The overlying sediments consist of, from the base to top, the Calivil Formation, the Shepparton Formation and the Coonambidgal Formation. However, these formations grade into each other, and their boundaries are often not well defined. The sedimentary cover has the maximum thickness in the lower catchment, and thins and pinches out over basement highs and in the valleys toward the highlands. The terrestrial Tertiary Calivil Formation has a thickness of up to 45 m. It does not crop out and occurs between 20 and 100 m below ground surface. It comprises consolidated gravel, sand silt, clay and cobbles with a hydraulic conductivity of 5 *to* 50 m day^{-1} (Shugg, 1987; Cheng and Reid, 2006). The alluvial deposits of the Holocene Coonambidgal Formation in the river valleys are contiguous with and indistinguishable from those of the underlying fluvio-lacustrine Quaternary Shepparton Formation. The Shepparton Formation and Coonambidgal Formation together are up to 170 m thick and form a complex heterogeneous unconfined to confined aquifer of clay and silt, and "shoestring lenses" of sand and gravel (Tickell, 1978). The alluvial sediments vary from unsorted cobbles and coarse gravels with fragments of basement rocks and minerals upstream to mature fine sands and silt downstream that are dominated by quartz and feldspar. The hydraulic conductivity of the Shepparton and Coonambidgal Formations is 0.1 *to* 10 m day^{-1} with an average of 0.2 *to* 5 m day^{-1} (Tickell, 1978). The Ovens River is hosted within the Coonambidgal Formation, except for several upstream locations, such as Smoko, Bright and Myrtleford, where it is incised into the basement. The surface aquifers receive recharge through direct infiltration on the valley floors and via exposed and weathered bedrock at the margins of valley. The vertical head gradients throughout the Ovens Catchment are generally downward, while the vertical head gradients within a few tens of metres of the river in the upper and middle catchments are upwards (Victorian Water Resource Data Warehouse, 2011). The regional groundwater flow is northwest parallel to the major valleys. The groundwater has a total dissolved solids (TDS) content of 100 *to* 500 mg L^{-1} which is higher than that of the Ovens River (TDS of 25 *to* 48 mg L^{-1}) (Victorian Water Resource Data Warehouse, 2011).

The climate of Ovens catchment is mainly controlled by the topography. The average rainfall decreases from 1127 mm in the alpine region at Bright to 636 mm on the alluvial plains in Wangaratta with most rainfall occurring in winter months (Bureau of Meteorology, 2011). During the Millennium drought (particularly 2006 *to* 2009), rainfall in the Ovens Catchment was between 40 and 80% of the long-term average (Victorian Government Department of Sustainability and Environment, 2010b, c). Potential evaporation increases northwards and ranges from 0 *to* 40 mm $month^{-1}$ to 125 *to* 200 mm $month^{-1}$ in winter and summer, respectively (Bureau of Meteorology, 2013). The riverine plains and alluvial flats are primarily cleared for agricultural use, while the hills and mountains are covered by native eucalyptus and plantation forests. Water extraction from both surface and groundwater resources is relatively low, being 5% of the total water resource available in the catchment (Victorian Government Department of Environment and Primary Industries, 2013).

2.3 Sampling and analytical techniques

Eight "run-of-river" sampling rounds took place over a period of 26 months (September 2009, March 2010, June 2010, September 2010, December 2010, March 2011, June 2011, and October 2011). Sample sites are designated by distance downstream of the uppermost sampling site at Harrietville (Fig. 2.1). Areas between Harrietville and Porepunkah (0 *to* 34 km), between Porepunkah and Everton (34 *to* 97 km), and between Everton and Bundalong (97 *to* 202 km) are defined as upper, middle and lower sub-catchments, respectively. These sub-catchments broadly represent the mountain valley, the transition from valley to alluvial plains and the broad flat alluvial flood plains. During March 2011 and June 2011, detailed electrical conductivity (EC) and ^{222}Rn surveys were also made between Bright and Porepunkah (22 *to* 34 km). This section includes a 2 km long and 2 *to* 4 m deep canyon in the basement, followed by a transition to the alluvial floodplain. River samples were collected from approximately 1 m above the riverbed using a collection beaker attached to a pole. In September 2009, groundwater was sampled from the Coonambidgal and Shepparton Formations close to the Ovens River from observation bores that are 2 *to* 50 m deep with 1 *to* 2 m screens; the geological composition in shallow depth consists of gravels with minor sand and clay, while that in lower depth is weathered siltstone. Some bores were re-sampled in the following rounds. The groundwater was sampled using an impeller pump set at the screened interval and at least 3 bore volumes of water were purged prior to sampling. EC was measured in the field using a TPS meter and electrode that was calibrated onsite. ^{222}Rn activities were measured using a portable in-air monitor (RAD-7, Durridge Co.) following methods described by Burnett and Dulaiova (2006) and expressed as Becquerels of radioactivity per cubic metre of water (Bqm^{-3}). ^{222}Rn was degassed from a 500 mL Buchner flask via a closed circuit of a known volume for 5 minutes. Counting times were 3 or 4 cycles over 2 hours for river water and 20 minutes for groundwater samples. Based on replicate analyses, precision of ^{222}Rn activities is within 3% at 10,000 Bqm^{-3}, increasing to around 8% at 200 Bqm^{-3}. Anion concentrations were measured on filtered and unacidified samples using a Metrohn ion chromatograph at Monash University. Cations were analysed using a Varian Vista ICP-AES at the Australian National University or a ThermoFinnigan OptiMass 9500 ICPMS at Monash University on samples that were filtered and acidified to pH <2. The precision of major ion concentrations based on replicate analyses is ±2 %. The charge balance errors were -16 *to* 17% for the surface water samples and -5 *to* 4% for the groundwater samples. Stable isotopes were measured at Monash University using a Finnigan MAT 252 and ThermoFinnigan Detla Plus Advatange mass spectrometers. δ^{18}O values were determined via equilibration with He-CO at 32 ^{o}C for 24 *to* 48 hours in a ThermoFinnigan Gas Bench. δ^{2}H was measured by reaction with Cr at 850 ^{o}C using an automated Finnigan MAT H/Device. δ^{18}O and δ^{2}H values were measured relative to internal standards calibrated using IAEA SMOW, GISP and SLAP. Data were normalized following Coplen (1988) and are expressed relative to V-SMOW, where δ^{18}O and δ^{2}H of SLAP are -55.5‰ and -428‰, respectively. Precision

(1σ) based on replicate analysis is: $\delta^{18}O = \pm0.1‰$ and $\delta^{2}H = \pm1‰$. River discharge for Harrietville (0 km), Bright (22 km), Myrtleford (65 km), Wangaratta (140 km) and Peechelba (187 km) was obtained from the Victoria Water Resources Data Warehouse (2011). Groundwater elevations in the catchment were also obtained from the Victoria Water Resources Data Warehouse (2011).

2.4 Results

2.4.1 River discharge

Between September 2009 and June 2010, the discharge of the Ovens River at Peechelba was between 160 and 4,360 ML day^{-1} with several moderately high flow events up to 11,420 ML day^{-1} during the 2009 winter (Fig. 2.2) (Victorian Water Resource Data Warehouse, 2011). Multiple extremely high flow events of up to 93,570 ML day^{-1} occurred between August 2010 and March 2011 that resulted from the 2010 *to* 2011 La Niña event. The river flow returned to 1,910 *to* 3,800 ML day^{-1} in the period of between March and June 2011, followed by multiple moderately high flow events of up to 25,850 ML day^{-1} between July and September 2011. The September

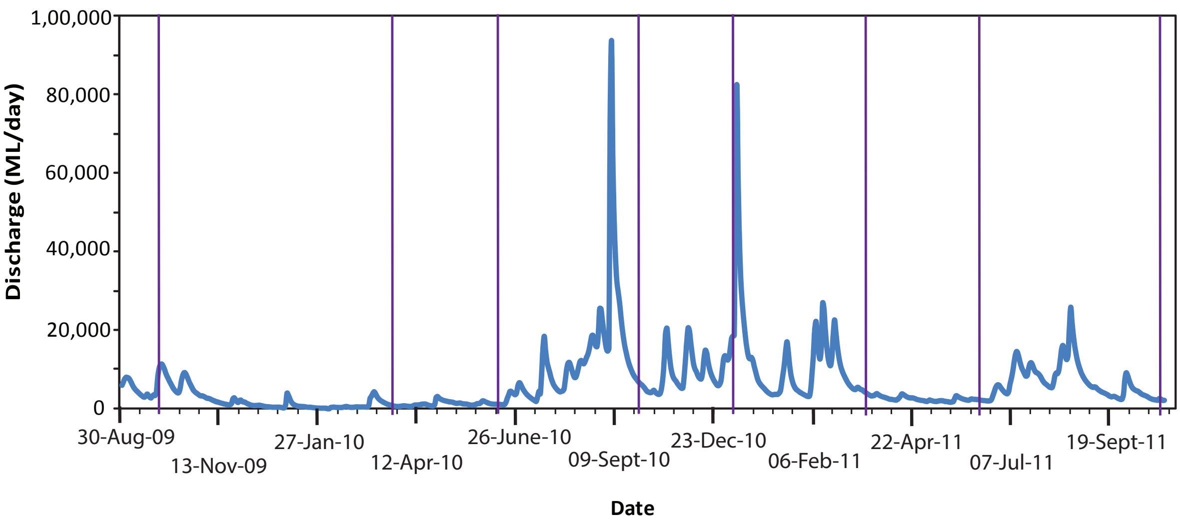

Figure 2.2 Variation in discharge of the Ovens River at Peechelba during the study period (September 2009 to October 2011) (Victorian Water Resource Data Warehouse 2011). Low flow condition prior to June 2010 was due to drought, followed by several flood events in September to December 2010. Times of sampling are indicated by the purple lines.

2009, September 2010, December 2010, and March 2011 sampling rounds all took place during high flow conditions with the September 2009 (10,178 ML day^{-1}) and December 2010 (18,520 ML day^{-1}) rounds occurring on the rising limb of a flow event, and the September 2010 (6,635 ML day^{-1}) and March 2011 (4,894 ML day^{-1}) rounds occurring on the receding limb of a flow event. The discharge in the March 2010, June 2010, June 2011 and October 2011 sampling rounds were 995 ML day^{-1}, 1,114 ML day^{-1}, 2,292 ML day^{-1} and 2,606 ML day^{-1}, respectively, and these sampling rounds represent low flow periods.

2.4.2 Groundwater levels

The hydrographs of shallow bores (< 20 m deep) at Bright indicate that recharge occurred on the valley alluvial plain in June 2010 *to* February 2011 and June *to* September 2011 (Fig. 2.3a) (Victorian Water Resource Data Warehouse, 2011). The annual hydraulic head variation at Bright in the upper catchment between 2009 and 2011 was 0.5 *to* 3.0 m. There was a strong lateral head gradient of ~7×10^{-3} between the edge of valley (B57144) and the river bank (B51747 & B51743) towards the river. There were several head reversals between the bores in the bank prior to June 2010. As with the upper catchment, there was recharge at Eurobin and Myrtleford in the middle catchment in the same period (Fig. 2.3b and c). However, the annual hydraulic head variation was only 0.5 *to* 1.0 m. Furthermore, the lateral head gradient toward the river in the middle catchment was lower (2×10^{-3} *to* 4×10^{-3}). The head gradients reversed in the river bank at Myrtleford during recharge periods (May 2009 and August 2010). No data is available for the groundwater level near the river in the lower catchment during the study period. However, the historical data at Peechelba indicates that the hydraulic heads in the flood plains (B11308 & B11307) varied by only a few millimetres per year (Fig. 2.3d). In contrast, the hydraulic head in the river bank (B11306) shows a greater variation of up to 1.5 m. The lateral head gradient toward the river in the lower catchment was ~10^{-4}.

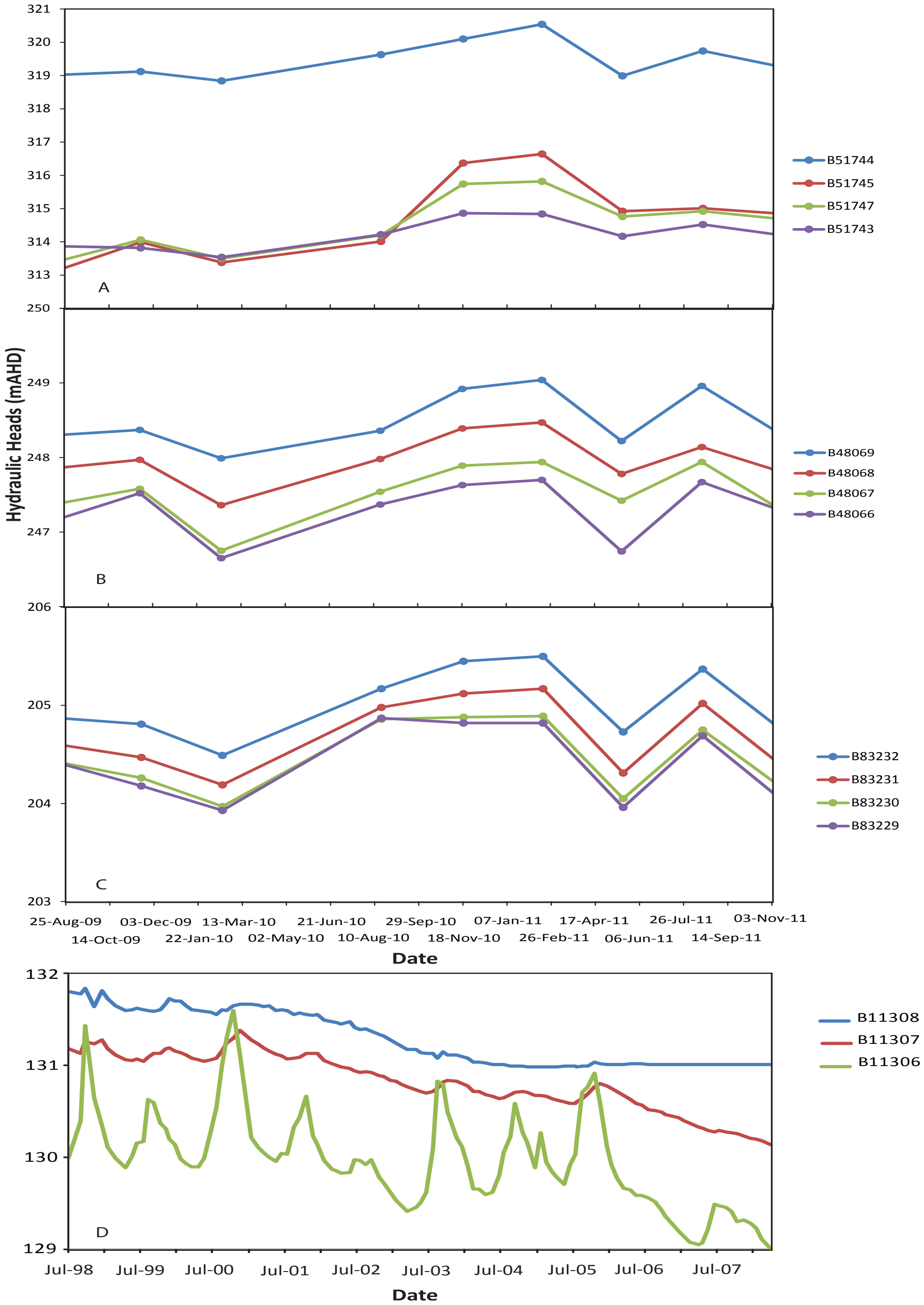

Figure 2.3 Seasonal variation in bore hydrographs at (A) Bright in the upper catchment, (B) Eurobin and (C) Myrtleford in the middle catchment, and (D) Peechelba in the lower catchment (Victorian Water Resource Data Warehouse 2011). Bores listed in order of decreasing distance from the river. Annual head variations and hydraulic gradients toward the river in the catchment decrease downstream. mAHD (m Australia Height Datum) refers to metres above mean sea level.

2.4.3 Electrical conductivity

The EC values of the Ovens River increased from ~30 μS cm^{-1} in the upper catchment to 37 *to* 55 µS cm^{-1} at Peechelba in the lower catchment (Fig. 2.4). There was always an increase in the EC values in the first 5 km river reach from Harrietville. However, most of the increase in EC values occurred from the middle catchment downstream. Higher EC values (35 *to* 73 µS cm^{-1}) were recorded in March 2010 and March 2011 at the end of summer, and June 2011 and Oct 2011 which had a discharge lower than the long-term average discharge. There was a very high EC value at 55 m in March 2010. Despite the high EC value, there is no a corresponding increase in the Cl concentration and ^{222}Rn activity at the same location. There was a small increase in the EC values in the Bright-Porepunkah river section (2.8 µS cm^{-1} in March 2011, 1.2 µS cm^{-1} in June

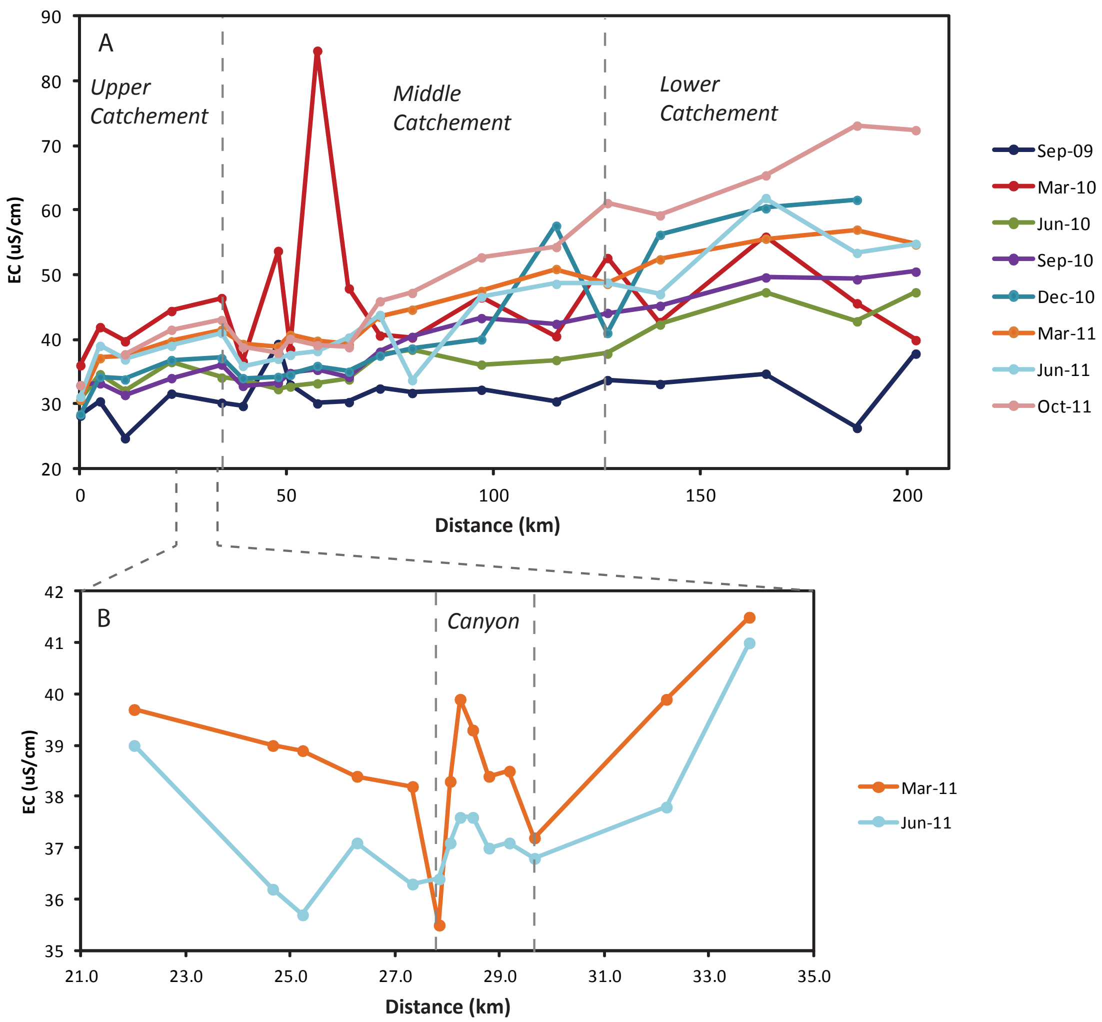

Figure 2.4 (A) EC values along the course of the river (Table 2.1). EC values gradually increase downstream. (B) EC values along the Bright-Porepunkah reach in March and June 2011. Distinct EC peaks at 28.5 km, followed by a gradual increase in EC values in both sampling rounds.

Site No.	Location	Easting	Northing	Distance (km)a	EC (µS/cm)							
					September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
Ovens River												
SW19	Bundalong	042773	6007835	202	38	40	47	51		55	55	72
SW18	Peechelba	043119	5997555	187	26	46	43	49	62	57	53	73
SW17	Killwara	043417	5984065	165	35	56	47	50	60	56	62	65
SW16	Wangaratta	043861	5956675	140	33	43	42	45	56	52	47	59
SW15	Wangaratta S.	044237	5974395	127	34	53	38	44	41	49	49	61
SW14	Tarrawingee	045113	5970115	115	30	41	37	42	58	51	49	54
SW13	Everton	045732	5966695	97	32	47	36	43	40	48	47	53
SW12	Whorley	046451	5959045	80	32	40	38	40	39	45	34	47
SW11	Whorley E.	047013	5966685	72	32	41	38	38	38	44	44	46
SW10	Myrtleford	047453	5952805	65	30	48	34	34	35	39	40	39
SW9	Salziers Ln	047837	5949855	57	30	85	33	35	36	40	38	39
SW8	Ovens	048239	5947285	50	33	39	33	35	35	41	38	40
SW7	Eurobin	048684	5945025	47	39	54	32	33	34	39	37	38
SW6	Porpunhah N.	048833	5941035	39	30	37	34	33	34	39	36	39
SW5	Porpunkah	049182	5938925	34	30	46	34	36	37	42	41	43
SW4	Bright	049936	5935505	22	32	44	37	34	37	40	39	42
SW3	Smoko	050568	5927365	11	25	40	32	31	34	37	37	38
SW2	Trout Farm	050560	5918975	5	30	42	35	33	34	37	39	38
SW1	Harrietville	050566	5917175	0	28	36	31	33	28	31	31	33
Tributaries												
SW22	King @ Oxley	044442	5966735		33	38	43	44	43	51	85	52
SW21	Buffalo @ Myrtleford	047182	5954215		35	37	38	37	35	39	39	41
SW20	Buckland @ Porpunkah	049049	5938625		33	34	31	30	33	33	32	35
Bright- Porepunkah												
SW5		049182	5938925							42	41	
BC8		493028	5937880							40	38	
BC7		494855	5936181							37	37	
BC1		495297	5936079							40	37	
BC2		495387	5935740							38	37	
BC3		495519	5935501							39	38	
BC4		495640	5935350							40	38	
BC5		495703	5935212							38	37	
BC6		496330	5935344							35	36	
BU4		496796	5925419							38	36	
BU1		497316	5935587							38	37	
BU2		498227	5935171							39	36	
BU3		498714	5935382							39	36	
SW4		049936	5935505							40	39	
Bright- Porepunkah (At Location BC8)												
SR	Spring-fed stream									49	49	
SS	Spring									64	52	

Table 2.1 EC values of the Ovens River. Notes: empty cell – not measured; a - Distance from the uppermost sampling site, Harrietville.

Bore No.	Location	Catchment	Distance (km)a	Easting	Northing	Bore Screen Depth (m)b	Distance to River (km)	September 2009		March 2011	
								EC (μS/cm)	Cl (mg/L)	EC (μS/cm)	Cl (mg/L)
B51743	Bright	Upper	22	499291	5935508	5-11	0.0206	82	2.80	76	3.32
B51747	Bright	Upper	22	499190	5935414	2-20	0.16	58	3.10	60	4.51
B1	Bright	Upper	22	499270	5935517	2-4	0.0095			95	2.95
B2	Bright	Upper	22	499260	5935513	2-4	0.0156			82	2.87
B51745	Bright	Upper	22	499139	5935375	5-11	0.225	63	2.16		
B51744	Bright	Upper	22	498933	5934911	6-12	0.725	56	2.73		
B51737	Bright	Upper	22	498445	5935658	36-42	0.261	111	2.41		
B51738	Bright	Upper	22	498397	5935420	58-63	0.0262	200	2.39		
B51735	Bright	Upper	22	498391	5935314	30-42	0.0708	74	3.58		
B51736	Bright	Upper	22	498382	5935299	20-26	0.0824	53	3.29		
B109461	Bright	Upper	22	497818	5935267	20-26	0.339	83	3.42		
B109462	Bright	Upper	22	497818	5935267	45-51	0.339	100	3.48		
B88271	Porepunkah	Upper	34	493294	5938062	8-14	0.219	100	3.98	107	3.75
B88274	Porepunkah	Upper	34	493256	5938067	35-53	0.21	64	1.85		
B48069	Eurobin	Middle	47	487803	5944698	5-8	0.506	129	3.59	122	4.07
B48068	Eurobin	Middle	47	487657	5944643	7-13	0.357	74	3.52	69	3.42
B48067	Eurobin	Middle	47	487519	5944594	12.0	0.203	92	4.14	77	3.24
B48066	Eurobin	Middle	47	487411	5944553	9-15	0.0914	78	3.90	70	3.97
B83232	Myrteford	Middle	65	474884	5953288	6-12	0.447	107	9.33		
B83231	Myrteford	Middle	65	474704	5953010	8-14	0.126	49	2.08		
M1	Myrteford	Middle	65	474605	5952919	4-6	0.0049			90	2.37
M2	Myrteford	Middle	65	474605	5952936	4-6	0.0215			68	2.43
B83229	Myrteford	Middle	65	474607	5952916	8-14	0.0128	87	2.01		
B83230	Myrteford	Middle	65	474604	5952937	8-14	0.0359	45	1.87		
B102783	Whorouly	Middle	80	464087	5959833	5-11	0.0048	107	2.06		
T1	Tarrawingee	Lower	115	451112	5970209	5-7	0.0056			367	47
T2	Tarrawingee	Lower	115	451121	5970212	5-7	0.0142			364	48
T3	Tarrawingee	Lower	115	451136	5970245	6-8	0.0509			315	46
B110738	Oxley	Lower	125	444240	5966742	19-44	0.0048	106	27		
B11326	Wangaratta	Lower	140	439879	5982755	23.7	2.79	1341	298		
B11493	Wangaratta	Lower	140	439422	5982189	16.5	2.13	920	134		
B302296	Boorhaman E.	Lower	165	437925	5992950	71-77	3.52	567	115		
B11323	Boorhaman E.	Lower	165	437924	5992953	17.4	3.52	536	96		
B50788	Boorhaman	Lower	170	442072	5999081	60-72	9.93	3800	923		
B50789	Boorhaman	Lower	170	442072	5999081	18-30	9.93	12020	3830		
B11306	Peechelba	Lower	187	432684	5994603	16	0.469	1194	299		
B11311	Bundalong S.	Lower	202	427007	6005559	16	0.378	2270	628		
B11310	Bundalong S.	Lower	202	427237	6005560	14	0.191	2250	570		

Table 2.2 EC values and Cl concentrations of groundwater in the Ovens Catchment. Notes – b: if single value is reported, the value represents the depth of the middle of the bore screen below ground surface.

2011) in the canyon (at 28 km) followed by a progressive increase in the EC values downstream towards Porepunkah. The EC values of shallow groundwater (< 50 m) increased down catchment from 50 *to* 100 μS cm^{-1} in the upper catchment to 100 *to* 400 μS cm^{-1} in the middle catchment and to 520 *to* 1,200 μS cm^{-1} in the lower catchment (Table 2.2). EC values of groundwater before and after the 2010 Victorian floods were similar.

2.4.4 Major ion chemistry

The cations in the Ovens River are in the following order of mass abundance: Na (36 *to* 58%), Mg (15 *to* 30%), Ca (18 *to* 29%) and K (4 *to* 22%) (Supplement; Table S2.1). The relative mass abundance of the anions were HCO (48 *to* 90%), Cl (3 *to* 44%), SO_4 (1 *to* 16%) and NO_3 (0.5 *to* 7%) (although HCO_3 were not measured for all sample rounds). As with the

Site No.	Location	Easting	Northing	Distance (km)a	Cl (mg/L)							
					September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
Ovens River												
SW19	Bundalong	042773	6007835	202	2.47	2.30	1.62	2.51	nm	4.03	4.54	4.93
SW18	Peechelba	043119	5997555	187	2.39	2.29	1.95	2.48	5.00	3.82	4.30	5.05
SW17	Killwara	043417	5984065	165	2.39	2.55	1.77	2.32	4.64	3.10	3.89	4.32
SW16	Wangaratta	043861	5956675	140	2.34	1.96	1.59	2.41	4.39	3.19	3.34	3.90
SW15	Wangaratta S.	044237	5974395	127	1.91	1.99	1.65	2.33	2.80	2.62	2.61	3.71
SW14	Tarrawingee	045113	5970115	115	1.81	1.78	1.63	2.35	3.64	5.61	3.12	3.05
SW13	Everton	045732	5966695	97	1.88	1.80	1.54	2.22	2.72	2.38	2.48	3.11
SW12	Whorley	046451	5959045	80	1.9	1.73	1.57	2.17	2.44	2.01	2.02	2.23
SW11	Whorley E.	047013	5966685	72	1.96	1.68	1.63	2.26	2.38	1.89	1.90	2.10
SW10	Myrtleford	047453	5952805	65	1.54	1.37	1.51	2.2	1.50	1.23	1.40	1.47
SW9	Salziers Ln	047837	5949855	57	1.61	1.32	1.52	2.19	1.66	1.49	1.39	1.45
SW8	Ovens	048239	5947285	50	1.49	1.30	1.47	2.32	2.30	1.21	1.39	1.40
SW7	Eurobin	048684	5945025	47	1.47	1.43	1.51	2.13	1.41	1.44	1.32	1.38
SW6	Porpunhah N.	048833	5941035	39	1.5	1.15	1.48	2.01	1.38	1.36	1.35	1.35
SW5	Porpunkah	049182	5938925	34	1.31	1.25	1.42	1.98	1.44	1.21	1.25	1.34
SW4	Bright	049936	5935505	22	1.39	1.03	1.32	1.89	1.56	1.03	1.21	1.15
SW3	Smoko	050568	5927365	11	1.09	0.85	1.28	1.88	1.27	0.75	0.92	0.90
SW2	Trout Farm	050560	5918975	5	0.99	0.82	1.31	1.87	1.00	0.69	0.95	0.89
SW1	Harrietville	050566	5917175	0	1.03	0.67	1.25	1.92	0.78	0.71	0.81	0.76
Tributaries												
SW22	King @ Oxley	044442	5966735		2.77	2.10	2.21	2.31	3.02	3.20	3.14	3.46
SW21	Buffalo @ Myrtleford	047182	5954215		2.12	1.64	1.61	2.31	2.21	3.34	1.80	1.87
SW20	Buckland @ Porpunkah	049049	5938625		1.52	1.05	1.57	1.76	1.39	1.18	1.17	1.20
Bright- Porepunkah (At Location BC8)												
SR	Spring-fed stream									1.57	2.15	
SS	Spring									1.85	2.21	

Table 2.3 Cl concentrations of the Ovens River

EC, the concentrations of major ions progressively increased downstream. For example, sodium concentrations increased from 1.80 *to* 3.50 mg L^{-1} in the upper catchment to 3.10 *to* 7.30 mg L^{-1} in the lower catchment (Fig 2.5a). Likewise, chloride concentrations rose from 0.70 *to* 1.90 mg L^{-1} in the upper catchment to 1.60 *to* 4.90 mg L^{-1} at Peechelba (Fig. 2.5b). Molar Na/Cl ratios decreased downstream from 2.80 *to* 8.60 in the upper catchment to 1.60 *to* 3.50 at ~135 km downstream and then remained at a similar level in the lower catchment (Fig. 2.5c). The only exception to this trend was the September 2010 sampling round where the Na/Cl ratio was between 1.70 and 2.70

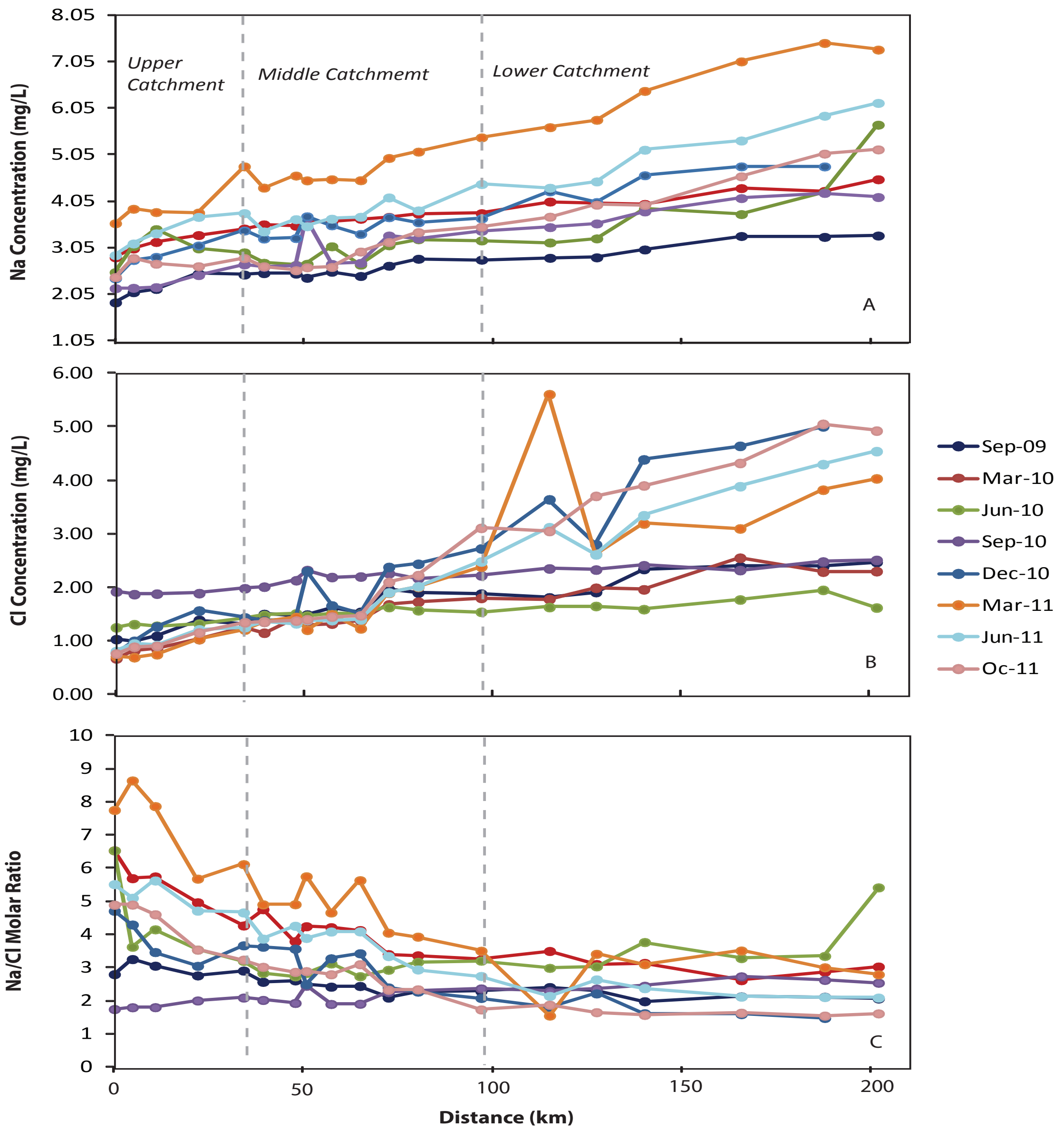

Figure 2.5 Variation in Na (A) and Cl (B) concentrations, and Na/Cl ratios (C) along the river. Na and Cl concentrations increase progressively downstream whereas the Na/Cl ratios decrease downstream.

along the entire river. Other cation/Cl ratios have similar trends to Na/Cl.

Groundwater in the upper and middle catchment was dominantly of a mixed magnesium and sodium or potassium bicarbonate type. Na comprises 22 *to* 58% of the total cations (by mass) with 20 *to* 43% Mg, 16 *to* 30% Ca and 3 *to* 21% K; HCO_3 accounts for 64 *to* 95% of anions with 5 *to* 18% Cl, 1 *to* 20% SO_4 and < 1 *to* 25% NO_3 (Supplement; Table S2.2). Groundwater in the lower catchment was a sodium or potassium chloride type with relative cation concentrations of 38 *to* 83% Na, 4 *to* 54% Ca, 2 *to* 27% Mg, and < 1 *to* 3% K, and relative anion concentrations of 29 *to* 64% Cl, 20 *to* 68% HCO_3, < 1 *to* 16% SO_4 and < 1 *to* 4% NO_3. Molar Na/Cl ratios of the low salinity (TDS < 100 mg L^{-1}) groundwater from the upper and middle catchments were mainly between 1.0 and 3.9 but locally as high as 11, whereas the more saline groundwater from the lower catchment had a Na/Cl ratio of 0.8 to 1.5 which is close to those of local rainfall (Blackburn and McLeod, 1983).

2.4.5 Stable isotopes

The Ovens River had $\delta^{18}O$ values of -5.8 *to* -7.5‰ and δ^2H values of -37 *to* -44‰. The $\delta^{18}O$ and δ^2H values define arrays lying to the left of the Melbourne meteoric water line (MMWL) with slopes of 3 to 6 (Fig. 2.6a). The deviation to the left of the MMWL is also apparent in the groundwater from the southeast Murray Basin and is probably due to local climatic differences between Melbourne (which is on the coast) and the inland Murray Basin (Cartwright and Morgenstem, 2012; Cartwright *et al.*, 2008; Leaney and Herczeg, 1999). Both $\delta^{18}O$ and δ^2H values increased downstream by 1.0 *to* 1.5‰ and by 3 *to* 6‰, respectively with most increases occurring between 0 and 75 km (Fig. 2.6b). The stream and the spring at 34 km in Porepunkah both had $\delta^{18}O$ values of -6.0 *to* -6.3‰ and δ^2H values of -34 *to* -36‰.

As with the Ovens River, the $\delta^{18}O$ and δ^2H values of groundwater in the Ovens Catchment

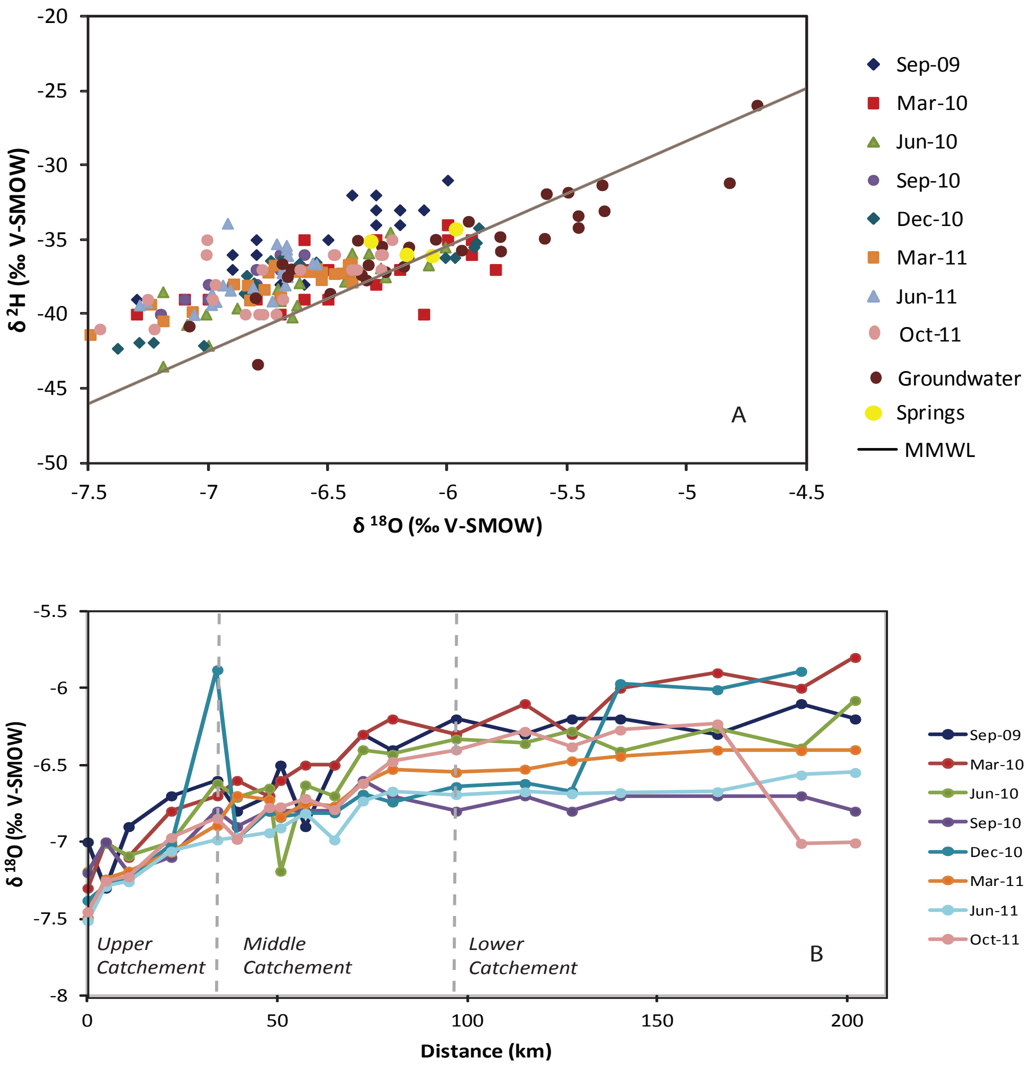

Figure 2.6 (A) $\delta^{18}O$ vs $\delta^{2}H$ values of the Ovens River and of groundwater in the Ovens catchment (Supplement; Tables S2.1 - S2.3). MMWL = Melbourne Meteoric Water Line. (B) Variation in $\delta^{18}O$ with distance along the river.

also plot to the left of the MMWL. The $\delta^{18}O$ values range from -4.5 to -7.5 ‰, while the $\delta^{2}H$ values are -30 *to* -40 ‰ (Fig. 2.6a). These are similar to those modern rainfall in southeast Australia ($\delta^{18}O$ = -6 ‰ and $\delta^{2}H$ = -35 ‰) (International Atomic Energy Agency, 2011) and therefore suggest recent recharge. Groundwater $\delta^{18}O$ and $\delta^{2}H$ values were similar in all sampling rounds.

2.4.6 Radon activities

The Ovens River at uppermost site (0 km) in Harrietville had consistently low ^{222}Rn activities (112 to 245 Bqm^{-3}), whereas other river reaches in the upper catchment had ^{222}Rn activities between 373 and 2,903 Bqm^{-3} (Fig. 2.7a). The highest ^{222}Rn activity was commonly recorded at 4.8 km, with the exception of the September 2010 and March 2011 sampling rounds. ^{222}Rn activities in the upper catchment were highest in September 2009, June 2011 and October 2011 and lower in March 2011, September 2011 and December 2011. In the Bright-Porepunkah river section, there was a significant ^{222}Rn peak of 905 Bqm^{-3} (March 2010) and 817 Bqm^{-3} (June 2010) at 28 km in the canyon (Fig. 2.7b), which is the site where the small increase in EC values was observed. The ^{222}Rn activities in the last 1.5 km of this river section were 881 to 1,243 Bqm^{-3}. A small stream

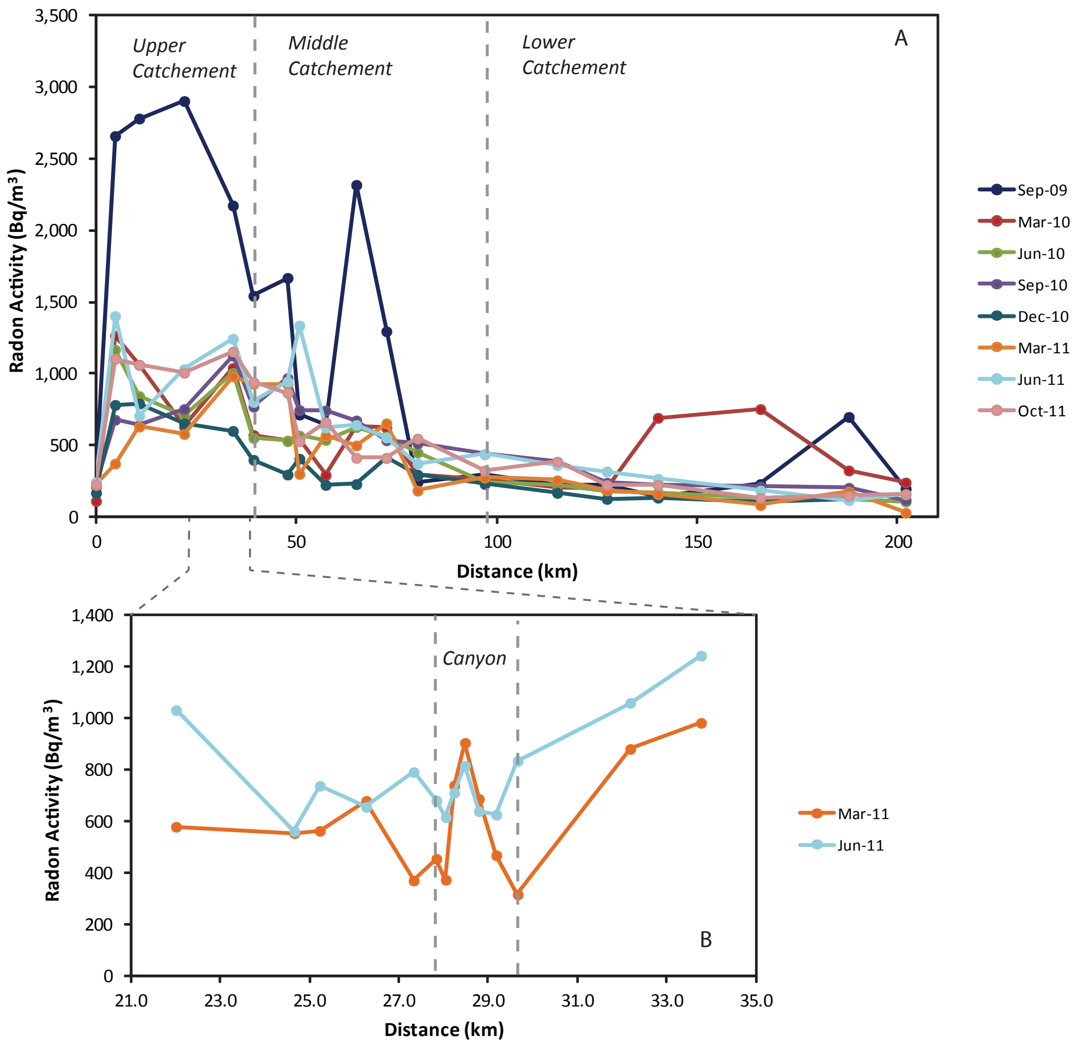

Figure 2.7 Variation of ^{222}Rn activities in the River (Table 2.4). High ^{222}Rn activities are recorded in the upper catchment and decrease down the valley. Temporal variation in the ^{222}Rn activities is minimal in the lower catchment. (B) ^{222}Rn activities along the Bright-Porepunkah reach in March and June 2011. Distinct ^{222}Rn peaks at 28.5 km (at which higher EC values are also found (Fig. 2.4c)), followed by a gradual increase in ^{222}Rn activity in both sampling rounds.

Site No.	Location	Easting	Northing	Distance (km)a	Radon (Bq/m3)							
					September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
Ovens River												
SW19	Bundalong	042773	6007835	202	199	242	109	119	nm	30	159	161
SW18	Peechelba	043119	5997555	187	699	325	127	205	123	179	118	148
SW17	Killwara	043417	5984065	165	230	754	130	211	105	83	185	135
SW16	Wangaratta	043861	5956675	140	138	693	169	227	136	161	267	225
SW15	Wangaratta S.	044237	5974395	127	227	193	179	240	127	186	318	224
SW14	Tarrawingee	045113	5970115	115	222	206	227	385	169	257	363	387
SW13	Everton	045732	5966695	97	297	263	239	439	231	275	440	325
SW12	Whorley	046451	5959045	80	245	293	450	514	296	185	374	548
SW11	Whorley E.	047013	5966685	72	1296	623	562	536	413	654	555	414
SW10	Myrtleford	047453	5952805	65	2318	639	628	674	230	500	643	413
SW9	Salziers Ln	047837	5949855	57	640	293	538	744	225	567	625	662
SW8	Ovens	048239	5947285	50	715	545	569	746	408	301	1337	530
SW7	Eurobin	048684	5945025	47	1669	533	530	970	297	928	945	867
SW6	Porpunhah N.	048833	5941035	39	1544	573	552	770	399	930	803	942
SW5	Porpunkah	049182	5938925	34	2174	1040	1005	1126	601	983	1243	1155
SW4	Bright	049936	5935505	22	2903	630	707	754	654	580	1032	1007
SW3	Smoko	050568	5927365	11	2781	1063	844	641	793	631	707	1062
SW2	Trout Farm	050560	5918975	5	2659	1265	1168	679	783	373	1403	1103
SW1	Harrietville	050566	5917175	0	222	112	238	213	169	224	245	226
Tributaries												
SW22	King @ Oxley	044442	5966735		419	155	142	211	107	102	146	209
SW21	Buffalo @ Myrtleford	047182	5954215		605	620	637	503	358	378	678	682
SW20	Buckland @ Porpunkah	049049	5938625		339	450	740	824	644	267	928	1165
Bright- Porepunkah												
SW5		049182	5938925							983	1243	
BC8		493028	5937880							881	1060	
BC7		494855	5936181							317	835	
BC1		495297	5936079							469	627	
BC2		495387	5935740							688	640	
BC3		495519	5935501							905	817	
BC4		495640	5935350							740	712	
BC5		495703	5935212							374	617	
BC6		496330	5935344							455	682	
BU4		496796	5925419							371	792	
BU1		497316	5935587							680	657	
BU2		498227	5935171							564	738	
BU3		498714	5935382							555	562	
SW4		049936	5935505							580	1032	
Bright- Porepunkah (At Location BC8)												
SR	Spring-fed stream									2653	8083	
SS	Spring									10488	50450	

Table 2.4 ^{222}Rn activities of the Ovens River.

Bore No.	Location	Catchment	Easting	Northing	Screen Depth (m)b	Distance to River (km)	Radon (Bq/m3)					
							September 2009	March 2010	June 2010	March 2010	June 2010	October 2010
B51737	Bright	Upper	498445	5935658	36-42	0.261	116750		100230			
B51743	Bright	Upper	499291	5935508	5-11	0.0206	75880	59000	67210	64500	72375	82500
B51744	Bright	Upper	498933	5934911	6-13	0.725				85875	26125	23200
B51747	Bright	Upper	499190	5935414	2-20	0.16	50650					
B1	Bright	Upper	499270	5935517	2-4	0.0095				28225	90125	71750
B2	Bright	Upper	499260	5935513	2-4	0.0156				39563	73250	76750
B88271	Porepunkah	Upper	493294	5938062	8-14	0.219	58880			48125		
B48066	Eurobin	Mid	487411	5944553	9-16	0.0914	42910	34740	45620	28350		
B48067	Eurobin	Mid	487519	5944594	8-15	0.203	28150	25010	24360	30925		
B48068	Eurobin	Mid	487657	5944643	3-7	0.357				31163		
B48069	Eurobin	Mid	487803	5944698	5-8	0.506				30086		
B83229	Myrtleford	Mid	474607	5952916	8-14	0.0128	26180		25980			
B83230	Myrtleford	Mid	474604	5952937	8-14	0.0359	31260		35870			
M1	Myrtleford	Mid	474605	5952919	4-6	0.0049				10325	19400	18788
M2	Myrtleford	Mid	474605	5952936	4-6	0.0215				19500	30236	25063
B102873	Whorouly	Mid	464087	5959833	5-11	0.513	28660					
B110738	Oxley	Mid	444240	5966742	44-48	0.0048	31090					
T1	Tarrawingee	Mid	451112	5970209	5-7	0.0056				5988	15150	23300
T2	Tarrawingee	Mid	451121	5970212	5-7	0.0142				14938	15738	18325
T3	Tarrawingee	Mid	451136	5970245	6-8	0.0509				16325	22738	18867
B11326	Wangaratta	Lower	439879	5982755	24	0.0509	35100					
B11493	Wangaratta	Lower	439422	5982189	17	2.13	21900					
B11323	Killawarra	Lower	437924	5992953	17	3.52	14360		15210			
B50789	Peechelba	Lower	442072	5999081	18-30	9.93	9260					
B11306	Peechelba	Lower	432684	5994603	16	0.469	12980	10460	14890			
B11310	Bundalong	Lower	427237	6005560	16	0.191	21360					
B11311	Bundalong	Lower	427007	6005559	14	0.378	18290	18130				

Table 2.5 ^{222}Rn activities of groundwater in the Ovens Catchment.

and spring on the alluvial plain at Porepunkah had ^{222}Rn activities of 2,663 Bqm^{-3} (March 2010) and 8,083 Bqm^{-3} (June 2010), and 10,488 Bqm^{-3} (March 2010) and 50,450 Bqm^{-3} (June 2010), respectively. In the middle catchment, ^{222}Rn activities generally decreased downstream from 601 *to* 2,174 Bqm^{-3} to 231 *to* 440 Bqm^{-3}, with several ^{222}Rn peaks occurring between 47 and 63 km. High ^{222}Rn activities were recorded in September 2009 and June 2011, whereas ^{222}Rn activities were lowest in December 2010. River reaches in the lower catchment had the lowest ^{222}Rn activities, ranging between 80 and 754 Bqm^{-3}. Elevated ^{222}Rn activities of between 699 and 745 Bqm^{-3} were recorded 140 and 187 km in September 2009 and March 2010. The temporal variation in the ^{222}Rn activities in the lower catchment was minimal, with a maximum difference of ~200 Bqm^{-3} between sampling rounds. The ^{222}Rn activities of groundwater were 30,000 *to* 110,000 Bqm^{-3}

in the upper catchment, 20,000 *to* 42,000 Bqm^{-3} in the middle catchment, and 10,000 *to* 20,000 Bqm^{-3} in the lower catchment (Table 2.5). The decreasing trend in the groundwater ^{222}Rn activities across the catchment reflects a change in lithology from immature sediments in the alluvial valleys, containing abundant U-bearing fragments of granitic and metamorphic material to more mature, weathered sediments that are dominated by quartz and feldspar on the flood plains. There were no statistically significant differences in groundwater ^{222}Rn activities between the sampling rounds even after the 2010 floods.

2.5 Discussion

The geochemistry of the Ovens River allows the major geochemical process to be defined and the distribution and magnitude of groundwater inflows to be calculated. The downstream increase in the $\delta^{18}O$ and $\delta^{2}H$ values (Fig. 2.6b) and the observation that the $\delta^{18}O$ and $\delta^{2}H$ values of the Ovens River define arrays with gradients of 3 *to* 6 indicate that evaporation has occurred (Fig. 2.6a). However, the changes in $\delta^{18}O$ and $\delta^{2}H$ values may also be accounted for by the different rainfall compositions across the catchment where the high altitude rainfall in the upper catchment is depleted, while the low altitude rainfall in the lower catchment is enriched. If evaporation does occur, the magnitude of increases in $\delta^{18}O$ and $\delta^{2}H$ values can be achieved by just < 10% total evaporation (Gonfiantini, 1986). Thus, in-stream evaporation is a minor role in concentrating river ions process and is insufficient to explain the downstream increases in TDS or EC (Fig. 2.4). There are no occurrences of halite in the Ovens Valley, and chloride in groundwater and surface water is derived from rainfall (Cartwright et al., 2006). Since the molar Na/Cl ratio of the rainfall in the region is 1.0 *to* 1.3 (Blackburn and McLeod, 1983), the high river Na/Cl ratios (Fig. 2.5c) and other cation/Cl ratios in the upper reaches of the Ovens River is probably caused by surface runoff and throughflow containing ions derived from physical and chemical weathering of silicate minerals on land surface or in the unsaturated zone within the upper catchment. The decrease in river Na/Cl ratios down the catchment is probably due to the influxes of both groundwater in the valley aquifers and surface runoff from the middle and lower catchments, both of which have relatively low Na/Cl ratio.

The agricultural and human activities along the Ovens River has moderately affected the water quality of the river in the middle and lower catchments as reflected by the elevated NO_3 concentrations (up to 0.78 mg L^{-1}) (Supplement; Table S2.1). However, the majority of river NO_3 concentrations are below 0.50 mg L^{-1}, and the highest EC values of the river is only 84 µS cm^{-1} (Table 2.1). As a results, the river chemistry in the Ovens River is affected by the human activities to a minor degree but should not have a significant impact on using chemistry to derive the volume of baseflow. More importantly, the ^{222}Rn activities in the river are not affected by agricultural and human activities. The ^{222}Rn derived baseflow thus does reflect the actual estimates of baseflow in the river.

Overall, the high ^{222}Rn activities in the upper and middle catchment of the Ovens River (Fig. 2.7) together with the progressive downstream increase in the EC values (Fig. 2.4) and major ion concentrations (Fig. 2.5) suggest that the Ovens River receives groundwater inflows. ^{222}Rn is used to identify gaining reaches and to calculate baseflow in this study because the difference of ^{222}Rn activities between groundwater and river water in the Ovens Catchment is 2 *to* 3 orders of magnitude, whereas the relative difference in the EC values and the concentrations of major ions between the groundwater and river water is much smaller. For comparison, baseflow fluxes are also calculated using chloride, and estimated from the hydrographs (Nathan and McMahan, 1990; Eckhardt, 2005) and differential flow gauging. Assessing other methods of estimating baseflow is valuable because river discharge and major ion data are far more extensive than ^{222}Rn data (e.g., Victorian Water Resource Data Warehouse, 2011; Central Asian Water-Info, 2013; NSW Government WaterInfo, 2013; USGS Water Data for USA, 2013).

2.5.1 Baseflow fluxes calculation using ^{222}Rn activities

Groundwater influxes to the river for the sampling rounds were calculated by rearranging Eq. 2.1 using ^{222}Rn activities from Table 2.4. Stream discharges of individual reaches were estimated by linear interpolation of the discharge at the five gauging stations. River depths and widths were estimated in the field; river depths varied from 1.2 to 8.0 m in winter and from 0.3 to 6.7 m in summer, and river widths ranged from 15 to 100 m in winter and from 7 to 90 m in summer. Evaporation rates are 1.3×10^{-3} m day^{-1} and 6.0×10^{-3} m day^{-1} for winter and summer months, respectively (Bureau of Meteorology, 2013). Based on the data in Table 2.5, ^{222}Rn activities of 76,000 Bqm^{-3}, 32,000 Bqm^{-3} and 19,000 Bqm^{-3} were assigned to groundwater from the upper, middle and lower catchments, respectively. Hyporheic exchange can also cause an elevation in ^{222}Rn activity in rivers where ^{222}Rn activities are less than ~300 Bqm^{-3} or where groundwater has a low ^{222}Rn activity (Lamontagne and Cook, 2007; Cartwright *et al.*, 2011; Cook, 2012). Failing to account for hyporheic exchange may result in the overestimation of groundwater inputs. The Ovens River and groundwater have generally high ^{222}Rn activities, and errors associated with not accounting for hyporheic exchange are likely to be small; thus initially the F_h term was omitted.

Gas exchange coefficients (k) were estimated using the modified gas transfer models of O'Connor and Dobbins (1958) and Negulescu and Rojanski (1969) as described by Mullinger *et al.* (2007)

$$k = 4.87 \times 10^{-4} \left(\frac{v}{d}\right)^{0.85} \qquad \textit{Eq. 2.2}$$

$$k = 9.301 \times 10^{-3} \left(\frac{v^{0.5}}{d^{1.5}}\right) , \qquad \textit{Eq. 2.3}$$

where v is the average stream velocity (m day^{-1}) derived from the stream discharge, river depth and river width data. Equation 2.2 generally produces higher k values (and hence yields greater groundwater fluxes) than Eq. 2.3. The k values for the winter months were generally 3.0 *to* 8.0 day^{-1} in the upper catchment, decreasing to 0.2 *to* 1.0 day^{-1} in the lower reaches. Lower values were obtained for the summer months, from 3.0 *to* 4.0 day^{-1} in the upper catchment to 0.2 *to* 0.3 day^{-1} in the lower catchment. High values of k in the upper catchment reflects the high velocities due to the shallow river depth and steep channel gradient, while low values of k in the lower catchment is the result of lower velocities due to the greater river depth and low channel gradient. Low-gradient rivers elsewhere generally have k values of 0.5 *to* 2.5 day^{-1} (Raymond and Cole, 2001; Cook *et al.*, 2003; Cartwright *et al.*, 2011), while shallow and turbulent rivers have k values of up to 34 day^{-1} (Mullinger *et al.*, 2007). Thus, the calculated k values for the Ovens River are within the range recorded in other studies. In this study, each sub-catchment was assigned an average value of k based on the k values from all individual reaches within the sub-catchment. The impact of tributary mixing on ^{222}Rn activities was calculated by combining the ^{222}Rn activity and the discharge at the sampling site downstream of the confluence with the ^{222}Rn activity (Table 2.4) and the discharge near the exit of the tributary.

The calculations indicate that most reaches are gaining (i.e. $I > 0$ m^3 m^{-1} day^{-1}), except for one reach at 11 km in the upper catchment in June 2011 and a few reaches in the middle and lower catchments. Losing reaches generally occur in the reaches between 48 and 57 km, between 115 and 117 km, and between 118 and 202 km. Based on the higher k values from Eq. 2.2, the baseflow fluxes are 0.4 *to* 9.0 m^3 m^{-1} day^{-1} with a mean of 1.0 m^3 m^{-1} day^{-1} for the upper catchment, 0.3 *to* 24.4 m^3 m^{-1} day^{-1} with a mean of 2.3 m^3 m^{-1} day^{-1} for the middle catchment and 0.2 *to* 24.1 m^3 m^{-1} day^{-1} with a mean of 1.1 m^3 m^{-1} day^{-1} for the lower catchment during high flow periods (September 2009, September 2010, December 2010, March 2011) (Fig. 2.8a); and 0.1 *to* 0.7 m^3 m^{-1} day^{-1} with a mean of 0.4 m^3 m^{-1} day^{-1} for the upper catchment, 0.1 *to* 2.5 m^3 m^{-1} day^{-1} with a mean of 0.6 m^3 m^{-1} day^{-1} for the middle catchment and 0.1 *to* 3.8 m^3 m^{-1} day^{-1} with a mean

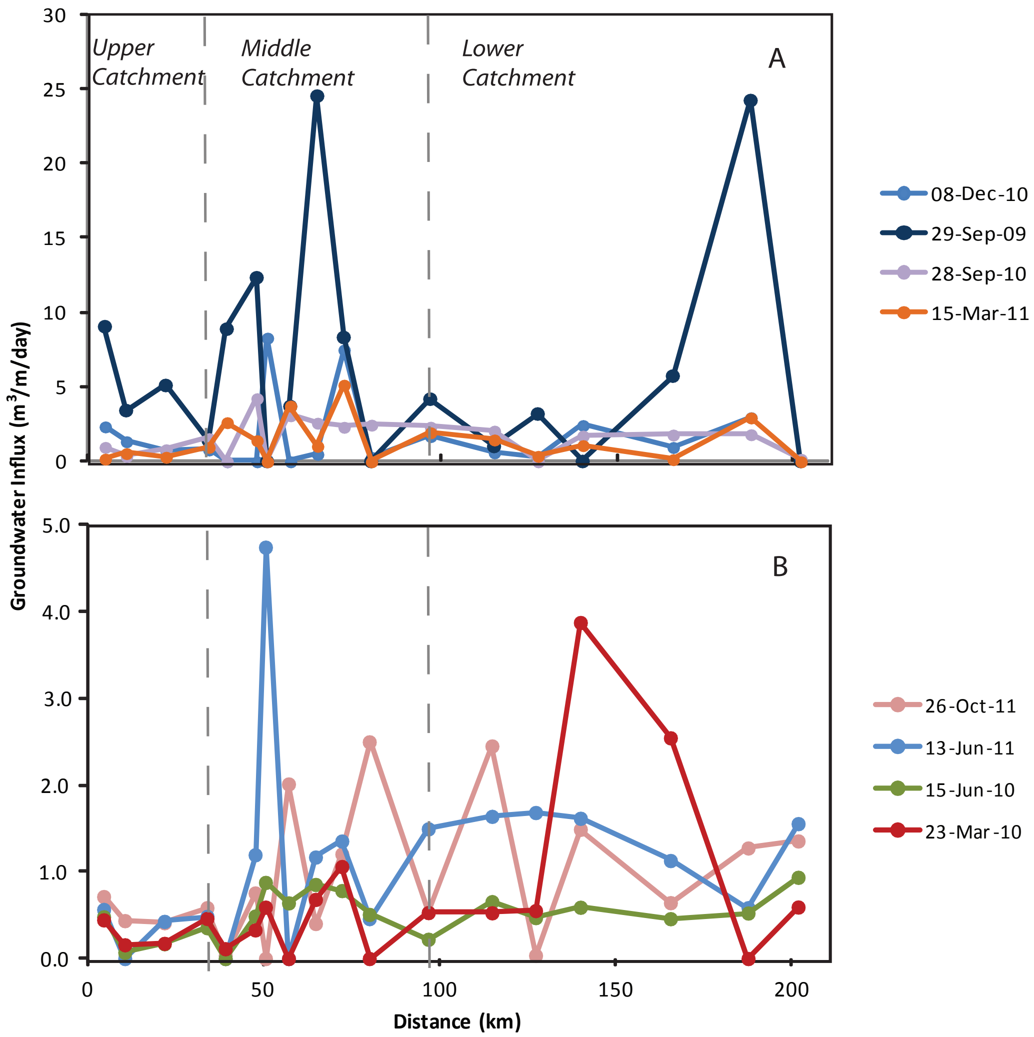

Figure 2.8 Groundwater influxes calculated from ^{222}Rn activities, based on the high k values, in flow conditions of 4894 to 18520 ML day^{-1} (A) and of 995 to 2606 ML day^{-1} (B). High baseflows occur in the upper catchment and often increase during high flow conditions. High baseflows also occur 65 to 72 km and 166 to 188 km.

of 0.8 m^3 m^{-1} day^{-1} for the lower catchment during low flow periods (March 2010, June 2010, June 2011 and October 2011) (Fig. 2.8b). The highest groundwater influxes generally occur in the middle catchment. There are very high groundwater inputs (up to 24 m^3 m^{-1} day^{-1}) at several locations (65 to 72 km and 166 to 188 km) in the middle and lower catchments. Furthermore, groundwater inputs, particularly in the upper and middle catchments, often increase during the high flow periods. This increase in groundwater influxes during the high flow periods is also reflected by the higher cumulative groundwater influxes in September 2009 (1,400,000 m^3 day^{-1}), December 2010 (290,000 m^3 day^{-1}) and September 2010 (330,000 m^3 day^{-1}) (Fig. 2.9a), compared to the low flow periods in March 2010 (180,000 m^3 day^{-1}), June 2010 (110,000 m^3 day^{-1}) and June 2011 (220,000 m^3 day^{-1}) (Fig. 2.9b). Although the cumulative groundwater fluxes increase during high flow periods, the proportional contribution of groundwater to the river is generally greater during low flow periods. September 2009 is the exception, where both the total groundwater and the proportion of groundwater (13%) are high. The cumulative groundwater inflow for the catchment during the study period was 110,000 *to* 1,400,000 m^3 day^{-1} with a mean of 370,000 m^3 day^{-1}, or 2 *to* 17% of total flow. Repeating the calculations with lower *k* values from Eq. 2.3 lowers the estimates of groundwater influxes in individual reaches by 11 *to* 70%, with an average of 43%.

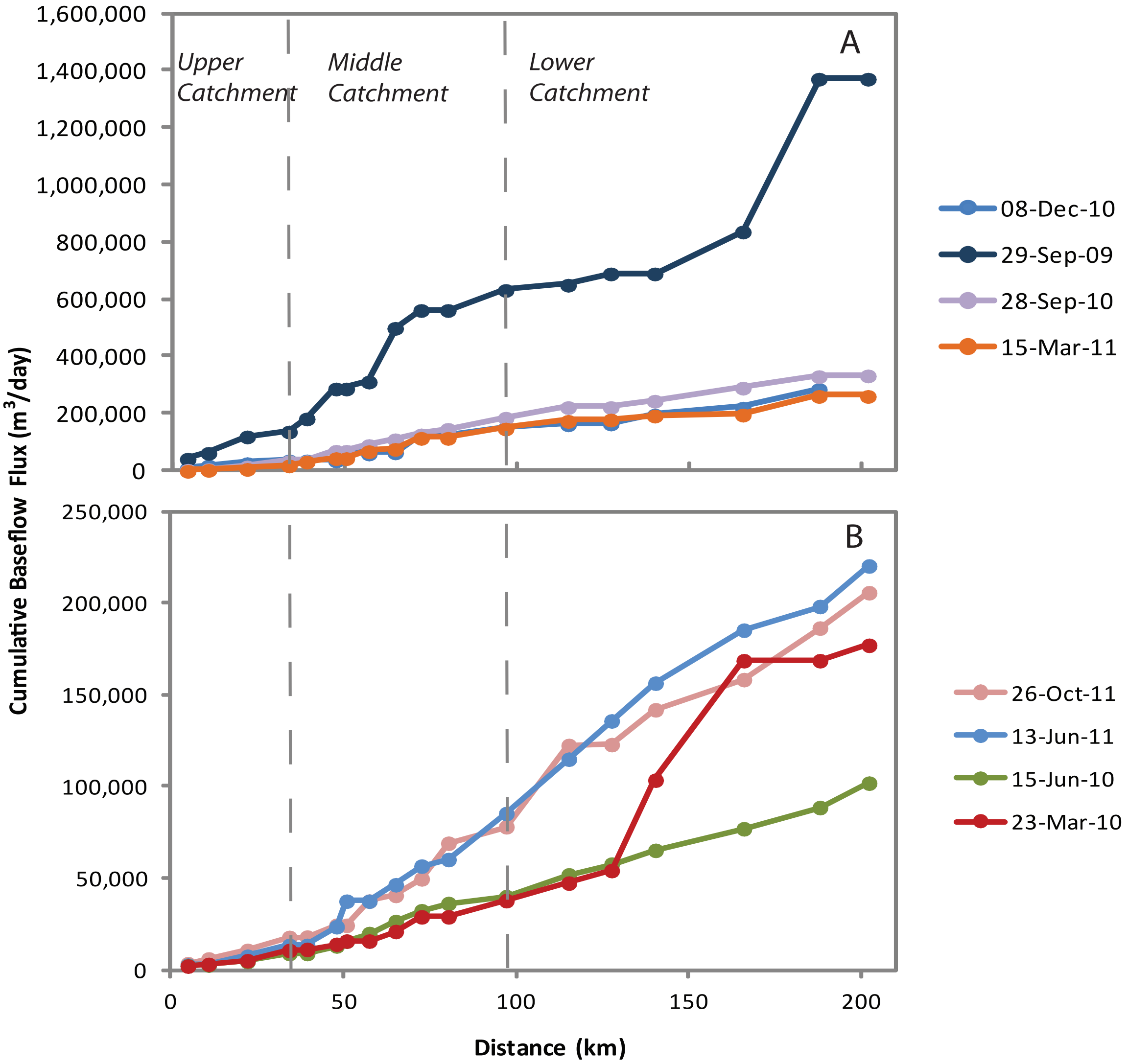

Figure 2.9 Cumulative baseflow estimated from ^{222}Rn activities, based on high k values, in flow conditions of 4894 to 18520 ML day^{-1} (A) and of 995 to 2606 ML day^{-1} (B). High cumulative baseflow usually occur in high flow conditions.

The largest percentage changes associated with the lower *k* values occur in the gaining reaches where groundwater inflows are lower (c.f. Cook *et al.*, 2003, 2006). Thus, the ability to estimate the amount of groundwater inflow in rivers with low groundwater inflows can be affected if the *k* value is not accurately constrained. The lower *k* estimates also result in an additional two reaches in the upper catchment and several reaches in the middle and lower catchment being interpreted as losing.

Overall, the lower *k* values reduce the calculated cumulative groundwater influxes to 77,000 *to* 680,000 m^3 day^{-1} for various flow conditions (compared to 102,000 *to* 1,400,000 m^3 day^{-1} when the higher *k* values are used). The calculations were also repeated by assigning different groundwater ^{222}Rn activities (±1 standard deviation of the sub-catchment ^{222}Rn activity) to understand the impact of the spatial variation in ^{222}Rn groundwater activities on groundwater inflows. The standard deviations of groundwater ^{222}Rn activity in the upper, middle and lower catchments are 29,400 Bqm^{-3}, 7,500 Bqm^{-3} and 8,400 Bqm^{-3}, respectively. An increase in the groundwater ^{222}Rn activity reduces groundwater influxes in individual reaches by 19 *to* 31%, whereas a decrease in the groundwater ^{222}Rn activity increases groundwater influxes by 31 *to* 81%. The calculated cumulative groundwater inflows for various flow conditions are 74,000 *to* 1,000,000 m^3 day^{-1} if the higher groundwater ^{222}Rn activities for various flow conditions are used and 170,000 *to* 2,200,000 m^3 day^{-1} if the lower groundwater ^{222}Rn activities are used. Therefore, the variation of groundwater ^{222}Rn end-member concentrations can lead to a high degree of uncertainty in determining the amount of groundwater inflow in rivers. The impact of ignoring hyporheic flow was assessed by assuming that the background ^{222}Rn activities in losing reaches were maintained by hyporheic exchange (c.f. Cartwright *et al.*, 2011). These background river ^{222}Rn activities were then subtracted from the measured river ^{222}Rn activities, and the groundwater influxes recalculated. For September 2009, the background river ^{222}Rn activities were 220 Bqm^{-3}, 175 Bqm^{-3} and 130 Bqm^{-3} for upper, middle and lower catchments, respectively. The revised groundwater influxes in individual reaches are 3 *to* 58% lower. The larger discrepancies occur in some reaches of the middle and lower catchments that have low calculated groundwater influxes. However, these reaches only contribute a small proportion of baseflow to the catchment, and thus these large discrepancies will only have a small effect on the catchment-scale groundwater inflow. The overestimation on the cumulative groundwater inflow in September 2009 due to ignoring hyporheic flow is ~17 %.

2.5.2 Baseflow fluxes calculation using Cl concentrations

Groundwater inputs to the river were also calculated by Cl concentrations (Table 2.3) via

$$I = \left(Q\frac{dCl_r}{dx} - wECl_r\right) \Big/ (Cl_i - Cl_r) \qquad \textit{Eq. 2.4}$$

(Cartwright *et al.*, 2011), where Cl_r and Cl_i are Cl concentrations in the river and groundwater, respectively. The spatial variation of groundwater chloride concentrations are different from those of ^{222}Rn, the Cl concentrations of groundwater used in the calculations are 3.25 mg L^{-1} for 0 *to* 65 km, 45 mg L^{-1} for 65 *to* 127 km and 275 mg L^{-1} for 127 *to* 202 km. Compared to the ^{222}Rn mass balance calculations, the Cl mass balance calculations indicate fewer gaining reaches in the upper, middle, and lower catchments (Fig. 2.10a). Additionally, the locations of high groundwater inflow are not always the same as those predicted from the ^{222}Rn activities. The groundwater influxes for the upper and middle catchments based on the Cl concentrations are higher than those based on the ^{222}Rn activities. Conversely, the Cl mass balance often yields lower groundwater influxes in the lower catchment than those calculated using ^{222}Rn activities. Several reaches in the middle catchment have extremely high calculated baseflow of up to 1,414 m^3 m^{-1} day^{-1}. The best match between ^{222}Rn- and Cl-derived groundwater influxes are the ones in the upper catchment in March

2010 and June 2010, and the ones in the lower catchment in December 2012. During the high flow periods, groundwater influxes of 0.5 *to* 34.8 $m^3 m^{-1} day^{-1}$ (a mean of 3.3), 0.1 *to* 1 400 $m^3 m^{-1} day^{-1}$ (a mean of 1.1) and 0.1 *to* 13.2 $m^3 m^{-1} day^{-1}$ (a mean of 0.3) are predicted for the upper, middle and lower catchment, respectively. For the low flow periods, groundwater influxes are lesser: 0.3 *to* 3.4 with a mean of 0.8 $m^3 m^{-1} day^{-1}$ for the upper catchment, 0.1 *to* 6.0 with a mean of 0.5 $m^3 m^{-1} day^{-1}$ for the middle catchment and 0.1 *to* 0.8 with a mean of 0.2 $m^3 m^{-1} day^{-1}$ for the lower catchment. Overall, the cumulative groundwater inflows based on the Cl mass balance calculations are 4 *to* 28% of total flow. If the assigned groundwater Cl concentrations are increased by 1 standard deviation (0.9, 17 and 120 mg L^{-1} for the upper, middle and lower catchments, respectively), the calculated groundwater inflows decrease by 25 *to* 37%. Decreasing the assumed groundwater Cl concentrations by a similar amount results in a 50 *to* 58% increase in groundwater inflows. For both the ^{222}Rn and Cl mass balance calculations, decreasing the groundwater end-member concentration (C_i & Cl_i) makes a greater change in groundwater fluxes compared with increasing C_i or Cl_i. As the difference between groundwater and river concentrations (C_i – Cr) or (Cl_i – Cl_r) become smaller, the uncertainties in C_i or Cl_i, produce larger relative errors in the calculated groundwater inflows. This is apparent in the Cl mass balance calculations because the difference between Cl_i and Cl_r is already low.

2.5.3 Baseflow fluxes calculation via hydrograph separation

Recursive digital filters separate the slowflow component of the hydrograph (assumed to be mainly baseflow) which has a low frequency from the high frequency signals associated with surface runoff (Nathan and McMahan, 1990; Eckhardt, 2005). The Nathan and McMahon filter (1990) is

$$b_k = y_k - \left[\alpha f_{k-1} + \frac{(1+\alpha)}{2}(y_k - y_{k-1})\right] , \qquad \textit{Eq. 2.5}$$

where b is the baseflow, y is the total stream discharge, f is the filtered quick flow, k is the time step, and α is the recession constant. Daily discharge data used in the calculations are between October 2000 and October 2011 from the three gauging stations: Bright (22 km), Myrtleford (65 km) and Peechelba (187 km). α is the gradient of the falling limb of a hydrograph and was determined via linear regression following Eckhardt (2008):

$$Q_{t+1} = \alpha Q_t \quad . \qquad \textit{Eq. 2.6}$$

The calculated values are 0.976 for Bright, 0.970 for Myrtleford and 0.967 for Peechelba. The filter was applied in three passes (forward, backward, forward) across the hydrograph as suggested by Nathan and McMahon (1990). The calculated percentages of baseflow in the May to October (wet) and November to April (dry) periods are 47 and 83% at 22 km, 51 and 78% at 65 km, 49 and 79% at 187 km, respectively. The Eckhardt (2005) filter is

$$b_k = \frac{(1-\mathrm{BFI}_{max})\alpha b_{k-1} + (1-\alpha)\mathrm{BFI}_{max} y_k}{1-\alpha\,\mathrm{BFI}_{max}} \quad , \qquad \textit{Eq. 2.7}$$

where b is the filtered baseflow ($b \leq y$), and BFI_{max} is the maximum value of the baseflow index (BFI) that can be modelled by the algorithm. BFI_{max} cannot be measured but is assigned based on the catchment lithology and river flow regime. Eckhardt (2005) proposed BFI_{max} values of 0.8 for perennial streams with porous aquifers, 0.5 for ephemeral streams with porous aquifer and 0.25 for perennial steams with hard rock aquifers. Considering the change from the large area of bedrock in the upper catchment to the sedimentary aquifers in the lower catchment, area-weighted BFI_{max} values of 0.31 for the upper catchment, 0.36 for the middle catchment and 0.47 for the lower catchment were assigned. The filter was applied in a single pass across the hydrograph as suggested by Eckhardt (2005). In comparison to the Nathan and McMahon filter, the Eckhardt filter produces lower percentages of baseflow: 36 and 52% at 22 km, 43 and 58% at 65 km, and 54 and 66% at 187 km, in the May to October and November to April periods, respectively. However, these values are still substantially higher than those estimated by ^{222}Rn activities: 3 and 2% at 22 km, 10 and 9% at 65 km, and 16 and 12% at 187 km.

2.5.4 Baseflow fluxes calculation via differential flow gauging

Groundwater inflow can also be estimated using differential flow gauging. When surface runoff is negligible, the groundwater flux to a river can be calculated from

$$Q_{gw} = Q_{dn} - Q_{up} + \sum Q_{out} - \sum Q_{in} \qquad \textit{Eq. 2.8}$$

(Brodie et al., 2007), where Q_{gw} is the groundwater flux, Q_{dn} is the river discharge at the downstream site, Q_{up} is the river discharge at a upstream site, Q_{out} is outputs from reach (such as, evaporation and extraction) and Q_{in} is inputs to the reach (such as, rainfall, tributaries and irrigation drainage). The groundwater during the low flow periods was calculated using Eq. 2.8 from parameters listed in Table 2.6. The calculations also subtract flow input from three main tributaries. Like the

Location	Parameters	23/03/2010	15/06/2010	13/06/2011	27/10/2011
Peechelba	Qdn (m^3/day)	995,191	1,113,730	2,291,906	2,605,936
	Qup (m^3/day)	281,073	384,569	626,881	812,098
	Qout (Eavopration) (m^3/day)	12,200	2,440	2,440	9,760
	Qin (Tributaries) (m^3/day)	384,757	503,800	1,168,102	1,159,343
	Qin (Rainfall) (m^3/day)	0	0	0	0
	Qgw (m^3/day)	341,561	227,801	499,363	644,255
Myrtleford	Qdn (m^3/day)	281,073	384,569	626,881	812,098
	Qup (m^3/day)	108,119	142,785	274,903	340,353
	Qout (Eavopration) (m^3/day)	2,303	461	461	1,842
	Qin (Tributaries) (m^3/day)	119,485	152,338	214,891	273,636
	Qin (Rainfall) (m^3/day)	0	0	0	0
	Qgw (m^3/day)	55,772	89,907	137,548	199,951
Bright	Qdn (m^3/day)	108,119	142,785	274,903	340,353
	Qup (m^3/day)	54,021	90,172	59,184	152,869
	Qout (Eavopration) (m^3/day)	550	110	110	440
	Qin (Tributaries) (m^3/day)	15,819	18,623	43,630	63,689
	Qin (Rainfall) (m^3/day)	0	0	0	0
	Qgw (m^3/day)	38,829	34,100	172,199	124,235

Table 2.6 Parameters used for calculating the net groundwater flux (*Qgw*) during low flow conditions by differential flow gauging using Eq. 2.8. Discharge data were obtained from Victorian Water Resource Data Warehouse (2011), and evaporation was estimated based on the surface area of river and data from Bureau of Meteorology (2013).

hydrograph separation, the discharge data come from the three gauging stations. The calculated net groundwater influxes at Bright, Myrtleford and Peechelba were 34,100 *to* 172,199 m^3 day^{-1}, 55,772 *to* 199,951 m^3 day^{-1} and 227,801 *to* 644,255 m^3 day^{-1}, respectively (Table 2.6). The net groundwater influxes for all the locations were much higher on 13/06/2011 and 27/10/2011 when the annual rainfall along with the river discharge was much higher in 2011. These estimates are higher than the corresponding fluxes derived from ^{222}Rn activities; 4,600 *to* 11,000 m^3 day^{-1} for Bright, 16,000 *to* 39,000 m^3 day^{-1} for Myrtleford and 62,000 *to* 150,000 m^3 day^{-1} for Peechelba.

2.5.5 Variations in baseflow

2.5.5.1 Spatial variations in baseflow

The baseflow fluxes derived from ^{222}Rn activates indicate moderate to high groundwater inflows in the upper catchment. In the upper catchment, the narrow valley creates a high hydraulic gradient of ~7×10^{-3} between the alluvial aquifers and the river, producing the observed groundwater inflows (Fig. 2.3a). The majority of the groundwater inflows occur in the first few river reaches (0 *to* 11 km) and between 31 and 34 km at Porepunkah. Between 0 and 11 km, the river is located at the edge of the valley, and it is likely that groundwater discharges to the river at these break of slopes as a result of the topography. The river reach at 28 *to* 30 m is in a moderately steep canyon. As the flow leaves the canyon, it cuts a deep channel through shallow sediments on the alluvial valley plains at Porepunkah, and the Porepunkah site is in an area with springs and a spring-fed stream. Groundwater inflows in the upper and middle catchments are also derived directly from the basement aquifer as evidenced by the presence of ^{222}Rn and EC peaks in the canyon (28.4 *to* 28.7 km) (Figs. 2.4b and 2.6b). The magnitude of groundwater influx from the basement aquifer may be large (up to 16 m^3 m^{-1} day^{-1} in March 2011). Since fractured bedrock aquifers often have very limited storativity, they deplete very quickly; groundwater inflows in this zone were lower toward the end of autumn in June 2011.

Groundwater fluxes in the middle catchment are locally lower or higher than those in the upper catchment. In general, the lateral head gradient toward the river in this region is lower due to the widening of the valley (Fig. 2.3b & c). The aquifer sediments also have lower hydraulic conductivities, and both these factors can cause a reduction of groundwater influxes to the river. However, some sections of the river in the middle catchment are moderately incised with steep banks. These reaches are likely to have higher groundwater inflows. Groundwater inflows are reduced in the lower catchment. This is the result of the shallow hydraulic gradient between the river and the groundwater in the open and flat alluvial flood plains in a semi-arid environment (Fig. 2.3d). Furthermore, groundwater inflows are likely to be restricted by the less conductive alluvial sediments.

Despite the widening of alluvial plains, several locations in the middle and lower catchments (between 65 and 72 km and between 166 and 188 km) receive significant baseflow (up to 24 $m^3\ m^{-1}\ day^{-1}$) (Fig. 2.8). This gaining behaviour is probably caused by basement highs that deflect groundwater flow and induce upward head gradients. Between 65 and 72 km several large outcrops of bedrocks occur near the river, while the river meanders close to the Warby Ranges between 166 and 188 km (Fig. 2.1). The losing reaches are generally located in the middle and lower catchments. At these locations, the difference between the river and the water table is usually small due to the increasing flatness of the topography. Thus, any small changes in river height or groundwater level can result in the observed fluctuating gaining and losing behaviour along these sections of the river.

2.5.5.2 Temporal variations in baseflow

Groundwater inflows in the upper catchment and some parts of the middle catchment increase during high flow periods (Figs. 2.8a and 2.9a). The increased rainfall over autumn and winter produces high surface runoff and also recharges the groundwater. The recharge rate in the coarse sediments of the upper Ovens is high with a recharge rate of 120 *to* 180 mm yr^{-1} (Cartwright and Morgenstem, 2012) and an annual fluctuation of up to 3 m in the water table (Fig. 2.3a). The rising groundwater elevations, which is referred to as hydraulic loading, increase the hydraulic head gradients toward the river and thus cause greater groundwater inflows. However, the magnitude of groundwater inflows do not always increase proportionally with river flows. For instance, the discharge in December 2010 was greater than that in September 2009, and yet the December 2010 round had a lower cumulative groundwater influx than the September 2009 round (Fig. 2.9a). The lower baseflow fluxes may be caused by the high river stage as a result of multiple floods in the previous winter/spring months that reduce the hydraulic gradient between the river and the adjacent groundwater. In contrast, the river was relatively dry in September 2009 after a period of drought, allowing a greater hydraulic gradient to be developed during the recharge period and thus producing a greater amount of baseflow which results in the large Rn peaks observed in figure 2.7. The groundwater inflows in the upper catchment can be low during extended low flow periods. The coarse aquifer sediments enable relatively quick drainage of groundwater into the river during winter and spring months. As a result, the water table near the river can drop significantly during dry periods (Victorian Water Resource Data Warehouse, 2011), resulting in less groundwater influxes to the river or losing reaches.

The baseflow fluxes in the lower catchment are similar at both high and lower flow conditions. The constant baseflow fluxes are probably caused by the limited fluctuation in the water table. The water table in this region varies by 0.5 *to* 1.5 m near the river and less than a few millimetres away from the river (Fig. 2.3d). The lower variation in the water table elevation is due to the lower recharge rate of 30 *to* 40 mm yr^{-1} on the floodplains (Cartwright and Morgenstem,

2012) which is the result of reduced rainfall, flat topography and the low hydraulic conductivity of the alluvial sediments. Since the water table near the river does fluctuate, it is possible for the river to recharge the adjacent aquifers and river banks during high flow conditions.

2.5.6 Comparing Rn with Cl concentrations, hydrograph separation and differential flow gauging

In the upper and to some extent the middle catchments, the groundwater fluxes estimated from the Cl concentrations are often greater than those based on the ^{222}Rn activities by 30 *to* 2,000% (Fig. 2.10a). The possible reasons for the discrepancy would be underestimating evaporation, ignoring local saline groundwater inputs or lack of difference in Cl concentrations between the groundwater and surface water (Fig. 2.10b). However, underestimating evaporation is unlikely since evaporation rates in this catchment are low. Saline groundwater input is probably not the sole reason for the overestimation. If the assigned groundwater Cl concentration was increased to 5 mg L^{-1} which is the highest Cl concentration in the upper catchment, the groundwater influxes would only decrease by 20 *to* 50%. The likely reason for the discrepancy is due to the similarity between the groundwater and river Cl concentrations in the upper and middle catchment. If the difference in the two end-member concentrations is small, it requires a significant input of groundwater in order to detect a rise in river Cl concentrations. It also magnifies any calculation and measurement errors in mass balance calculations, particularly in the groundwater end-member concentration (Cook, 2012). The large increases in the river Cl concentrations in some reaches (and thus the large estimated groundwater influxes) may be due to accumulation of Cl over several reaches or may come from other sources, such as water in the unsaturated zone or in pools on the riverine plain.

In the lower catchments, the Cl-derived baseflow fluxes are usually lower than those estimated by ^{222}Rn by 50 *to* 100% (Fig. 2.10a). If the assigned groundwater Cl concentrations in

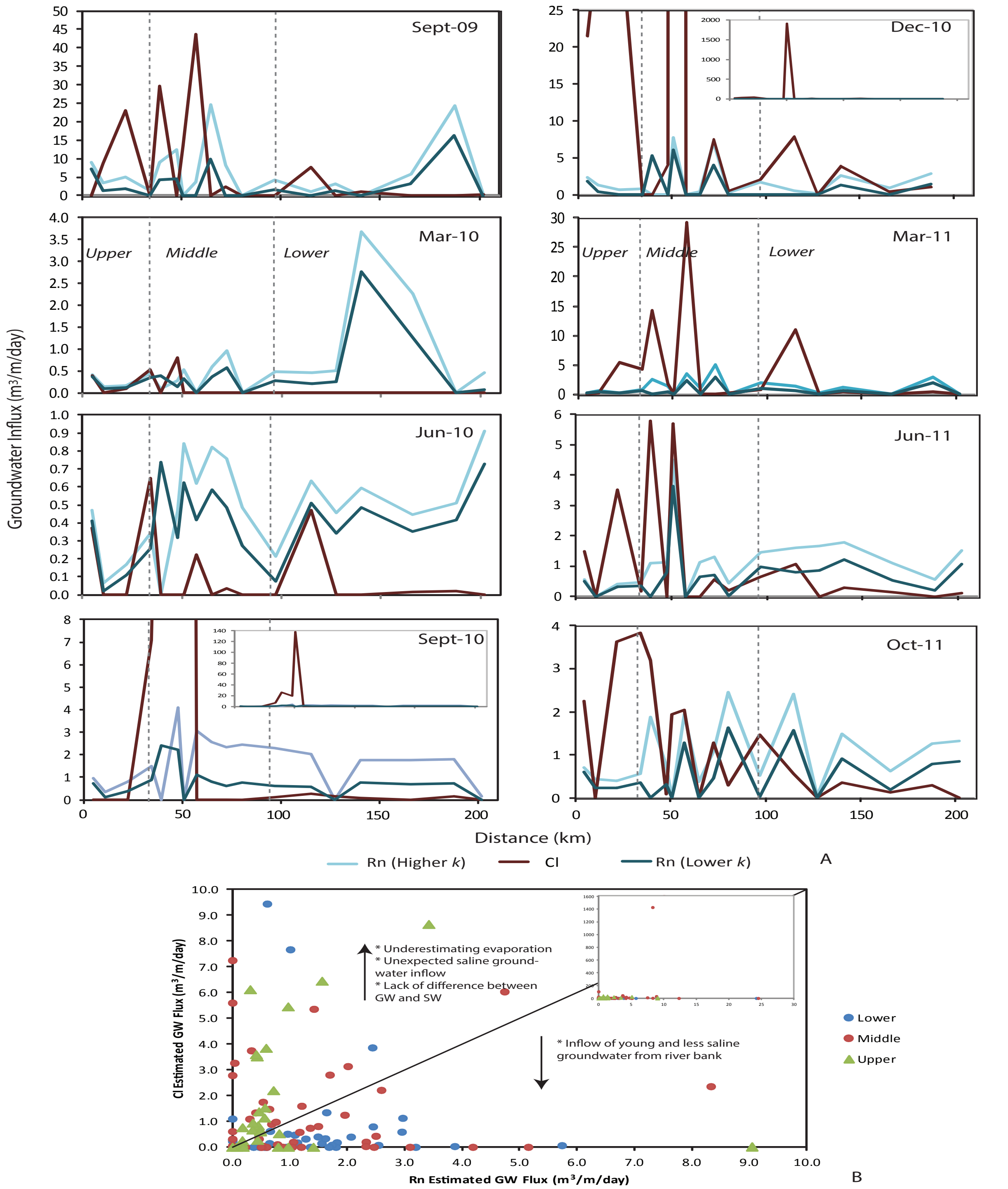

Figure 2.10 (A) Comparison of baseflow fluxes in each sampling round, estimated from ^{222}Rn activities and Cl concentrations using Eqs. 2.1 and 2.4 respectively. For Rn, baseflow fluxes based on higher and lower *k* values are shown. (B) A scatter plot of all ^{222}Rn- and Cl-derived groundwater influxes (for Rn only those based on higher *k* values are shown in the graph). If the ^{222}Rn and Cl derived groundwater influxes agree to each other, they should be plotted in a straight line. Data points above the line indicate that Cl-derived groundwater inflows exceed ^{222}Rn-derived groundwater inflows, and vice versa. The possible reasons for the discrepancies are given in the graph. Both (A) and (B) shows Cl concentrations generally yield higher baseflow fluxes in the upper catchment but lower baseflow fluxes in the lower catchment. The inserts shows the original scale of the enlarged graph. Lower = lower catchment, Middle = middle catchment, Upper = upper catchment.

the calculations are reduced, the baseflow estimates would progressively increase, matching ones derived from the ^{222}Rn activities. This may indicate that the majority of baseflow in this area is derived from less saline water in the mid-channel bars and river banks rather than the more saline regional groundwater (Fig. 2.10b). Using regional groundwater compositions in the mass balance calculations would underestimate the total groundwater discharge to the river but correctly identify the amount of regional groundwater discharge if the groundwater discharge comprises both bank storage and regional groundwater (McCallum *et al.*, 2010; Cartwright *et al.*, 2011). In comparison to hydrograph separation, ^{222}Rn mass balance produces consistently lower total baseflow fluxes across the catchment. The ^{222}Rn mass balance calculations require that the groundwater is in secular equilibrium with the aquifer sediments (Cook, 2012). Thus, recently recharged groundwater may not be adequately accounted for by ^{222}Rn mass balance. Hydrograph separation, however, aggregates groundwater, bank return flow, interflow and draining of pools on the floodplains into the slowflow component (Griffiths and Clausen, 1997; Halford and Mayer, 2000; Evans and Neal, 2005). The discrepancy between ^{222}Rn mass balance and hydrograph separation probably indicates that in addition to regional groundwater, other delayed flow components contribute to the flow of the Ovens River. That other components, such as bank return flow, contribute to the river is also suggested by the Cl data, as discussed above. The uncertainty in assigning BFI_{max} in the Eckhardt filter and the assumptions behind the hydrograph separations may also contribute to the discrepancy.

The groundwater inflows calculated based on ^{222}Rn activities are lower than those derived from differential flow gauging. It is unlikely that the difference is caused by not taking the water abstraction along the Ovens River in the calculations because the use of surface water in the catchment is less than 2% of the total surface water storage (Victorian Government Department of Environment and Primary Industries, 2013). The difference, however, may be due to unaccounted surface runoff during rainfall in the catchment leading to the sampling. Furthermore, baseflow derived from differential flow gauging is the total baseflow flux (outflow – inflow) and thus can consist of groundwater, delayed bank returns and interflows. This is probably why the baseflow fluxes were much greater in June 2011 and October 2011 since the high rainfall in 2011 resulted in greater groundwater inflows via hydraulic loading and bank return flow. On the other hand, groundwater inflow estimated by ^{222}Rn may not include some short-medium term water stores, such as delayed bank returns and interflows, due to insufficient time to reach secular equilibrium. The higher groundwater inflows from differential flow gauging may suggest that interflow and water from the unsaturated zone provide a significant amount of discharge to the river. The large soil zone in the upper and middle catchments is likely to supply water to the river, maintaining the river flow during low flow periods since the calculated baseflow in the catchment is only 2 to 17% of total flow. As these comparisons indicate, numerical methods based on flow data are likely to provide larger groundwater inflow estimates compared to chemical mass balance. Although these physical methods can isolate the delayed flows from surface runoff, they cannot separate various components of delayed flow. On the other hand, the delayed components such as groundwater, interflow, or bank return flows may have a different geochemistry. Therefore, chemical tracer based methods may be able to track the different components of delayed flow. When considering methods for studying SW-GW interaction, the availability of data is an important factor. Flow data is often a first choice since it is readily available. But it is equally important to consider the aims of the study and what particular components of baseflow (like regional groundwater, river bank, water from the unsaturated zone) the study focuses on.

2.6 Conclusions

The SW-GW interactions at the Ovens River were investigated using chemical tracers and flow data. Groundwater inflow is controlled by topography and aquifer lithology. Although the groundwater often constitutes the highest proportion of the river flow during baseflow conditions, total groundwater inflow increases with river flow. The increase in total groundwater inflow is caused by hydraulic loading where recharge during high rainfall conditions produces a rapidly raising water table and increases the hydraulic gradient between the groundwater and the river. The effect of hydraulic loading is likely to be common in areas with steep topography and permeable aquifers, and is greatest during the receding phase of river flow. The understanding of hydraulic loading in this study also shows that while it is important for any river-aquifer interaction studies to examine changes in river height, the fluctuation in water table needs to also be carefully considered, especially if the aquifer is responsive to rainfall. This study shows that in a catchment where the difference in the major ion geochemistry between groundwater and surface water is minimal, ^{222}Rn is a good tracer of groundwater inflows. However, the inclusion of chloride concentrations and discharge data in this study allows other possible sources of water inflowing into the river to be identified. Since each method utilises different physical and chemical properties of groundwater to trace groundwater, it is better to adapt a multiple-technique approach in order to provide a more completed view on the relationship of a river to its adjacent groundwater system.

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Table S2.1 Geochemistry of the Ovens River in the study period of September 2009 to October 2011.

Site No.	Location	Easting	Northing	Distance (km)	EC (µS/cm)	δ18O SMOW	δ2H SMOW	Cl (mg/L)	NO3 (mg/L)	SO4 (mg/L)	HCO3 (mg/L)	Na (mg/L)	Mg (mg/L)	K (mg/L)	Ca (mg/L)
Septermber 2009															
Ovens River															
SW19	Bundalong	042773	6007835	202	38	-6.2	-33	2.47	0.61	1.24		3.31	1.23	0.95	1.47
SW18	Peechelba	043119	5997555	187	26	-6.1	-33	2.39	0.47	1.33	4	3.28	1.28	0.83	1.57
SW17	Killwara	043417	5984065	165	35	-6.3	-32	2.39	0.53	1.25		3.3	1.24	0.79	1.54
SW16	Wangaratta	043861	5956675	140	33	-6.2	-33	2.34	0.57	1.1		3.01	1.26	0.78	1.55
SW15	Wangaratta S.	044237	5974395	127	34	-6.2	-33	1.91	0.62	1.01	5	2.84	1.46	0.62	1.48
SW14	Tarrawingee	045113	5970115	115	30	-6.3	-33	1.81	0.47	0.91	12	2.83	1.47	0.6	1.45
SW13	Everton	045732	5966695	97	32	-6.2	-34	1.88	0.59	0.99		2.79	1.42	0.58	1.46
SW12	Whorley	046451	5959045	80	32	-6.4	-32	1.9	0.6	0.99		2.81	1.39	0.62	1.45
SW11	Whorley E.	047013	5966685	72	32	-6.3	-34	1.96	0.78	1.11		2.66	1.43	0.58	1.45
SW10	Myrtleford	047453	5952805	65	30	-6.5	-35	1.54	0.36	0.73	21	2.44	1.38	0.49	1.25
SW9	Salziers Ln	047837	5949855	57	30	-6.9	-36	1.61	0.43	0.78		2.53	1.48	0.48	1.37
SW8	Ovens	048239	5947285	50	33	-6.5	-37	1.49	0.44	0.79	10	2.4	1.49	0.46	1.47
SW7	Eurobin	048684	5945025	47	39	-6.7	-37	1.47	0.27	0.8		2.49	1.49	0.47	1.33
SW6	Porpunhah N.	048833	5941035	39	30	-6.8	-35	1.5	0.37	0.85		2.5	1.47	0.47	1.37
SW5	Porpunkah	049182	5938925	34	30	-6.6	-38	1.31	0.36	0.84	23	2.47	1.64	0.44	1.49
SW4	Bright	049936	5935505	22	32	-6.7	-38	1.39	0.38	0.83	74	2.49	1.7	0.43	1.6
SW3	Smoko	050568	5927365	11	25	-6.9	-37	1.09	0.01	0.73	18	2.16	1.55	0.36	1.49
SW2	Trout Farm	050560	5918975	5	30	-7.3	-39	0.99	0.2	0.78	11	2.09	1.68	0.36	1.49
SW1	Harrietville	050566	5917175	0	28	-7.0	-38	1.03	0.24	0.49		1.87	1.35	0.31	1.43
Tributaries															
SW22	King @ Oxley	044442	5966735		33	-6.2	-33	2.77	0.56	1.01		3.09	0.84	0.88	1.7
SW21	Buffalo @ Myrtleford	047182	5954215		35	-6.0	-31	2.12	0.7	1.08		2.65	1.33	0.64	1.41
SW20	Buckland @ Porpunkah	049049	5938625		33	-6.8	-36	1.52	0.26	0.73		2.49	1.41	0.49	1.25
March 2010															
Ovens River															
SW19	Bundalong	042773	6007835	202	40	-5.8	-37	2.30	0.04	1.16		4.52	1.77	0.85	1.58
SW18	Peechelba	043119	5997555	187	46	-6.0	-34	2.29	0.02	1.05		4.27	1.73	0.82	1.61

SW17	Killwara	043417	5984065	165	56	-5.9	-35	2.55	0.18	0.87	4.33	1.73	0.73	1.59
SW16	Wangaratta	043861	5956675	140	43	-6.0	-34	1.96	0.09	0.78	3.99	1.7	0.71	1.57
SW15	Wangaratta S.	044237	5974395	127	53	-6.3	-35	1.99	0.09	0.97	4.01	1.72	0.67	1.55
SW14	Tarrawingee	045113	5970115	115	41	-6.1	-40	1.78	0.09	0.90	4.04	1.7	0.69	1.54
SW13	Everton	045732	5966695	97	47	-6.3	-38	1.80	0.20	0.95	3.79	1.61	0.6	1.58
SW12	Whorley	046451	5959045	80	40	-6.2	-37	1.73	0.17	0.97	3.78	1.59	0.62	1.53
SW11	Whorley E.	047013	5966685	72	41	-6.3	-35	1.68	0.24	0.95	3.71	1.56	0.65	1.51
SW10	Myrtleford	047453	5952805	65	48	-6.5	-37	1.37	0.13	0.77	3.67	1.49	0.57	1.58
SW9	Salziers Ln	047837	5949855	57	85	-6.5	-39	1.32	0.24	0.77	3.61	1.56	0.54	1.56
SW8	Ovens	048239	5947285	50	39	-6.6	-35	1.30	0.18	0.75	3.58	1.56	0.5	1.52
SW7	Eurobin	048684	5945025	47	54	-6.7	-39	1.43	0.16	0.68	3.52	1.52	0.51	1.54
SW6	Porpunhah N.	048833	5941035	39	37	-6.6	-37	1.15	0.11	0.62	3.55	1.51	0.53	1.5
SW5	Porpunkah	049182	5938925	34	46	-6.7	-40	1.25	0.08	0.71	3.46	1.52	0.47	1.51
SW4	Bright	049936	5935505	22	44	-6.8	-39	1.03	0.01	0.70	3.32	1.35	0.43	1.68
SW3	Smoko	050568	5927365	11	40	-7.1	-39	0.85	0.02	0.79	3.17	1.49	0.4	1.57
SW2	Trout Farm	050560	5918975	5	42	-7.0	-39	0.82	0.01	0.86	3.03	1.47	0.39	1.57
SW1	Harrietville	050566	5917175	0	36	-7.3	-40	0.67	0.02	0.58	2.84	1.42	0.36	1.55
Tributaries														
SW22	King @ Oxley	044442	5966735		38	-5.9	-36	2.10	0.11	0.54	3.65	1.41	0.54	1.41
SW21	Buffalo @ Myrtleford	047182	5954215		37	-6.0	-35	1.64	0.35	0.85	3.21	1.58	0.52	1.39
SW20	Buckland @ Porpunkah	049049	5938625		34	-6.6	-39	1.05	0.11	0.61	3.43	1.38	0.35	1.31
June 2010														
Ovens River														
SW19	Bundalong	042773	6007835	202	47	-6.1	-37	1.62	0.12	0.84	5.69	1.56	0.90	1.68
SW18	Peechelba	043119	5997555	187	43	-6.4	-37	1.95	0.15	0.71	4.24	1.52	0.77	1.63
SW17	Killwara	043417	5984065	165	47	-6.3	-38	1.77	0.18	0.65	3.77	1.46	0.66	1.59
SW16	Wangaratta	043861	5956675	140	42	-6.4	-37	1.59	0.13	1.01	3.89	1.52	0.64	1.73
SW15	Wangaratta S.	044237	5974395	127	38	-6.3	-37	1.65	0.11	0.75	3.25	1.60	0.54	1.48
SW14	Tarrawingee	045113	5970115	115	37	-6.4	-37	1.63	0.09	0.88	3.16	1.56	0.52	1.44
SW13	Everton	045732	5966695	97	36	-6.3	-36	1.54	0.01	0.70	3.20	1.56	0.52	1.45
SW12	Whorley	046451	5959045	80	38	-6.4	-38	1.57	0.09	0.78	3.23	1.54	0.51	1.45

SW11	Whorley E.	047013	5966685	72	38	-6.4	-36	1.63	0.01	0.86	3.10	1.40	0.52	1.30
SW10	Myrtleford	047453	5952805	65	34	-6.7	-39	1.51	0.09	0.90	2.68	1.53	0.37	1.28
SW9	Salziers Ln	047837	5949855	57	33	-6.6	-39	1.52	0.17	0.97	3.08	1.67	0.37	1.37
SW8	Ovens	048239	5947285	50	33	-7.2	-39	1.47	0.04	0.88	2.70	1.64	0.35	1.39
SW7	Eurobin	048684	5945025	47	32	-6.7	-40	1.51	0.02	0.79	2.68	1.60	0.34	1.32
SW6	Porpunhah N.	048833	5941035	39	34	-6.7	-38	1.48	0.18	0.75	2.72	1.61	0.34	1.33
SW5	Porpunkah	049182	5938925	34	34	-6.9	-40	1.42	0.35	0.85	2.95	1.96	0.34	1.67
SW4	Bright	049936	5935505	22	37	-7.0	-42	1.32	0.11	0.62	3.03	1.88	0.32	1.71
SW3	Smoko	050568	5927365	11	32	-7.1	-41	1.28	0.02	0.47	3.45	1.84	0.33	1.36
SW2	Trout Farm	050560	5918975	5	35	-7.0	-40	1.31	0.11	0.49	3.08	1.99	0.36	1.71
SW1	Harrietville	050566	5917175	0	31	-7.2	-44	1.25	0.09	0.41	2.52	1.58	0.21	1.50
Tributaries														
SW22	King @ Oxley	044442	5966735		43	-6.0	-36	2.21	0.31	0.32	4.15	1.20	0.87	2.08
SW21	Buffalo @ Myrtleford	047182	5954215		38	-6.2	-35	1.61	0.24	0.95	2.94	1.35	0.54	1.31
SW20	Buckland @ Porpunkah	049049	5938625		31	-6.6	-38	1.57	0.11	0.54	3.01	1.53	0.37	1.21
Septermber 2010														
Ovens River														
SW19	Bundalong	042773	6007835	202	51	-6.8	-37	2.51	0.32	0.37	4.14	1.86	1.0	1.92
SW18	Peechelba	043119	5997555	187	49	-6.7	-37	2.48	0.66	0.77	4.23	1.85	0.96	1.91
SW17	Killwara	043417	5984065	165	50	-6.7	-37	2.32	0.71	0.82	4.12	1.80	0.94	1.89
SW16	Wangaratta	043861	5956675	140	45	-6.7	-36	2.41	0.53	0.61	3.83	1.77	0.90	1.88
SW15	Wangaratta S.	044237	5974395	127	44	-6.8	-37	2.33	0.62	0.72	3.57	1.85	0.70	1.65
SW14	Tarrawingee	045113	5970115	115	42	-6.7	-37	2.35	0.29	0.34	3.49	1.83	0.71	1.64
SW13	Everton	045732	5966695	97	43	-6.8	-37	2.22	0.66	0.77	3.41	1.82	0.68	1.63
SW12	Whorley	046451	5959045	80	40	-6.7	-36	2.17	0.43	0.50	3.25	1.79	0.65	1.62
SW11	Whorley E.	047013	5966685	72	38	-6.6	-36	2.26	0.83	0.96	3.31	1.79	0.63	1.42
SW10	Myrtleford	047453	5952805	65	34	-6.8	-38	2.2	0.6	0.70	2.72	1.63	0.53	1.35
SW9	Salziers Ln	047837	5949855	57	35	-6.8	-38	2.19	0.38	0.44	2.70	1.68	0.53	1.40
SW8	Ovens	048239	5947285	50	35	-6.8	-38	2.32	0.47	0.55	3.69	1.71	0.56	1.46
SW7	Eurobin	048684	5945025	47	33	-6.8	-38	2.13	0.2	0.23	2.67	1.62	0.50	1.31
SW6	Porpunhah N.	048833	5941035	39	33	-6.9	-38	2.01	0.36	0.42	2.64	1.61	0.49	1.33

SW5	Porpunkah	049182	5938925	34	36	-6.8	-37	1.98	0.43	0.50	2.69	1.80	0.49	1.50
SW4	Bright	049936	5935505	22	34	-7.1	-39	1.89	0.37	0.43	2.46	1.67	0.45	1.49
SW3	Smoko	050568	5927365	11	31	-7.2	-40	1.88	0.24	0.28	2.20	1.63	0.37	1.47
SW2	Trout Farm	050560	5918975	5	33	-7.0	-39	1.87	0.47	0.55	2.18	1.72	0.38	1.44
SW1	Harrietville	050566	5917175	0	33	-7.2	-40	1.92	0.01	0.01	2.18	1.38	0.32	1.29
Tributaries														
SW22	King @ Oxley	044442	5966735		44	-6.7	-37	2.31	0.57	0.66	3.87	1.34	1.1	1.99
SW21	Buffalo @ Myrtleford	047182	5954215		37	-6.7	-37	2.31	1.32	1.53	2.86	1.58	0.62	1.40
SW20	Buckland @ Porpunkah	049049	5938625		30	-7.0	-38	1.76	1.36	1.58	2.51	1.40	0.48	1.15
December 2010														
Ovens River														
SW19	Bundalong	042773	6007835	202										
SW18	Peechelba	043119	5997555	187	62	-5.9	-36	5.00	0.45	1.17	4.80	2.27	3.00	3.03
SW17	Killwara	043417	5984065	165	60	-6.0	-36	4.64	0.46	1.23	4.80	2.05	2.64	3.16
SW16	Wangaratta	043861	5956675	140	56	-6.0	-36	4.39	0.46	1.04	4.61	1.97	2.62	3.02
SW15	Wangaratta S.	044237	5974395	127	41	-6.7	-37	2.80	0.33	0.97	4.03	1.84	0.80	2.09
SW14	Tarrawingee	045113	5970115	115	58	-6.6	-37	3.64	0.57	1.01	4.26	1.59	2.65	2.93
SW13	Everton	045732	5966695	97	40	-6.6	-37	2.72	0.31	1.08	3.68	1.74	0.98	1.96
SW12	Whorley	046451	5959045	80	39	-6.7	-36	2.44	0.29	0.92	3.59	1.79	0.70	1.72
SW11	Whorley E.	047013	5966685	72	38	-6.7	-36	2.38	0.41	0.85	3.70	1.83	0.73	1.79
SW10	Myrtleford	047453	5952805	65	35	-6.8	-38	1.50	0.33	0.69	3.34	1.81	0.60	1.64
SW9	Salziers Ln	047837	5949855	57	36	-6.8	-38	1.66	0.29	0.73	3.53	1.93	0.65	1.74
SW8	Ovens	048239	5947285	50	35	-6.8	-37	2.30	0.23	0.70	3.72	1.88	0.64	1.65
SW7	Eurobin	048684	5945025	47	34	-6.8	-38	1.41	0.24	0.66	3.25	1.84	0.52	1.55
SW6	Porpunhah N.	048833	5941035	39	34	-7.0	-39	1.38	0.09	0.65	3.24	1.85	0.51	1.58
SW5	Porpunkah	049182	5938925	34	37	-5.9	-35	1.44	0.24	0.77	3.42	2.06	0.52	1.79
SW4	Bright	049936	5935505	22	37	-7.0	-42	1.56	0.13	0.69	3.10	2.03	0.51	1.88
SW3	Smoko	050568	5927365	11	34	-7.2	-42	1.27	0.07	0.76	2.85	2.00	0.42	1.74
SW2	Trout Farm	050560	5918975	5	34	-7.3	-42	1.00	0.07	0.89	2.78	2.10	0.42	1.77
SW1	Harrietville	050566	5917175	0	28	-7.4	-42	0.78	0.03	0.45	2.39	1.65	0.30	1.59
Tributaries														

SW22	King @ Oxley	044442	5966735		43	-5.9	-34	3.02	0.40	0.97	4.14	1.83	0.85	2.00
SW21	Buffalo @ Myrtleford	047182	5954215		35	-6.6	-37	2.21	0.26	0.92	3.28	1.50	0.61	1.72
SW20	Buckland @ Porpunkah	049049	5938625		33	-6.9	-39	1.39	0.03	0.62	3.11	1.66	0.46	1.60
March 2011														
Ovens River														
SW19	Bundalong	042773	6007835	202	55	-6.4	-38	4.03	0.66	1.26	7.31	2.35	1.64	3.84
SW18	Peechelba	043119	5997555	187	57	-6.4	-37	3.82	0.05	1.12	7.45	2.32	1.61	3.79
SW17	Killwara	043417	5984065	165	56	-6.4	-37	3.10	0.55	1.42	7.06	2.37	1.57	3.76
SW16	Wangaratta	043861	5956675	140	52	-6.4	-37	3.19	0.40	0.97	6.42	2.00	1.44	3.72
SW15	Wangaratta S.	044237	5974395	127	49	-6.5	-37	2.62	0.39	1.09	5.80	2.13	1.21	3.50
SW14	Tarrawingee	045113	5970115	115	51	-6.5	-37	5.61	6.25	1.31	5.64	1.64	1.66	3.37
SW13	Everton	045732	5966695	97	48	-6.5	-37	2.38	0.32	1.02	5.42	2.04	1.13	3.30
SW12	Whorley	046451	5959045	80	45	-6.5	-38	2.01	0.34	0.94	5.12	2.02	1.03	3.13
SW11	Whorley E.	047013	5966685	72	44	-6.6	-37	1.89	0.40	0.85	4.98	1.97	1.04	3.10
SW10	Myrtleford	047453	5952805	65	39	-6.8	-38	1.23	0.34	0.84	4.50	2.01	0.90	2.86
SW9	Salziers Ln	047837	5949855	57	40	-6.8	-37	1.49	0.47	0.68	4.52	2.23	0.87	2.87
SW8	Ovens	048239	5947285	50	41	-6.8	-38	1.21	0.50	0.79	4.50	2.20	0.87	2.83
SW7	Eurobin	048684	5945025	47	39	-6.7	-37	1.44	0.43	0.60	4.60	2.36	0.88	2.87
SW6	Porpunhah N.	048833	5941035	39	39	-6.7	-39	1.36	0.31	0.56	4.34	2.26	0.78	2.73
SW5	Porpunkah	049182	5938925	34	42	-6.9	-38	1.21	0.28	0.79	4.80	2.24	0.89	3.48
SW4	Bright	049936	5935505	22	40	-7.1	-40	1.03	0.12	0.71	3.80	2.23	0.68	2.68
SW3	Smoko	050568	5927365	11	37	-7.2	-40	0.75	0.09	0.84	3.81	2.33	0.56	2.85
SW2	Trout Farm	050560	5918975	5	37	-7.2	-39	0.69	0.02	0.00	3.89	2.37	0.66	3.03
SW1	Harrietville	050566	5917175	0	31	-7.5	-41	0.71	0.05	0.44	3.57	2.07	0.42	2.59
Tributaries														
SW22	King @ Oxley	044442	5966735		51	-6.4	-37	3.20	0.45	1.15	6.18	2.20	1.21	3.42
SW21	Buffalo @ Myrtleford	047182	5954215		39	-6.6	-37	3.34	1.24	1.55	4.45	2.17	0.91	2.95
SW20	Buckland @ Porpunkah	049049	5938625		33	-6.8	-39	1.18	0.17	0.45	4.31	1.84	0.72	2.43
June 2011														
Ovens River														

SW19	Bundalong	042773	6007835	202	55	-6.5	-37	4.54	0.56	1.67	6.16	2.36	1.07	2.496
SW18	Peechelba	043119	5997555	187	53	-6.6	-37	4.30	0.63	1.65	5.89	1.99	0.99	2.697
SW17	Killwara	043417	5984065	165	62	-6.7	-36	3.89	0.63	1.37	5.35	1.98	0.89	2.345
SW16	Wangaratta	043861	5956675	140	47	-6.7	-38	3.34	0.60	1.31	5.16	2.07	0.94	2.802
SW15	Wangaratta S.	044237	5974395	127	49	-6.7	-37	2.61	0.50	1.29	4.47	2.13	0.84	2.691
SW14	Tarrawingee	045113	5970115	115	49	-6.7	-35	3.12	0.37	0.84	4.34	1.31	0.99	2.329
SW13	Everton	045732	5966695	97	47	-6.7	-37	2.48	0.65	1.27	4.42	2.15	0.72	2.186
SW12	Whorley	046451	5959045	80	34	-6.7	-37	2.02	0.70	1.17	3.85	2.05	0.70	2.11
SW11	Whorley E.	047013	5966685	72	44	-6.7	-39	1.90	0.53	1.02	4.13	2.08	0.73	2.151
SW10	Myrtleford	047453	5952805	65	40	-7.0	-39	1.40	0.59	0.87	3.71	2.00	0.53	1.732
SW9	Salziers Ln	047837	5949855	57	38	-6.8	-38	1.39	0.50	0.88	3.67	2.08	0.55	1.94
SW8	Ovens	048239	5947285	50	38	-6.9	-38	1.39	0.54	0.85	3.51	2.09	0.51	1.825
SW7	Eurobin	048684	5945025	47	37	-6.9	-38	1.32	0.40	0.73	3.66	2.04	0.48	1.723
SW6	Porpunhah N.	048833	5941035	39	36	-7.0	-39	1.35	0.33	0.80	3.40	1.83	0.44	1.807
SW5	Porpunkah	049182	5938925	34	41	-7.0	-39	1.25	0.41	0.75	3.79	2.06	0.45	2.175
SW4	Bright	049936	5935505	22	39	-7.1	-40	1.21	0.14	0.87	3.71	2.22	0.47	2.043
SW3	Smoko	050568	5927365	11	37	-7.3	-39	0.92	0.27	0.72	3.36	2.21	0.39	2.088
SW2	Trout Farm	050560	5918975	5	39	-7.3	-39	0.95	0.33	0.97	3.14	2.33	0.35	1.911
SW1	Harrietville	050566	5917175	0	31	-7.5	-40	0.81	0.25	0.62	2.89	1.86	0.38	2.026
Tributaries														
SW22	King @ Oxley	044442	5966735		85	-6.7	-35	3.14	0.63	1.36	4.82	2.36	0.78	2.367
SW21	Buffalo @ Myrtleford	047182	5954215		39	-6.7	-36	1.80	0.50	0.99	3.98	1.95	0.66	2.106
SW20	Buckland @ Porpunkah	049049	5938625		32	-6.9	-34	1.17	0.46	0.72	3.41	1.75	0.44	1.484
October 2011														
Ovens River														
SW19	Bundalong	042773	6007835	202	72	-7.0	-35	4.93	0.00	1.98	5.16	1.74	1.12	2.76
SW18	Peechelba	043119	5997555	187	73	-7.0	-36	5.05	0.05	2.55	5.07	1.71	1.09	2.64
SW17	Killwara	043417	5984065	165	65	-6.2	-35	4.32	0.01	2.11	4.58	1.55	1.00	2.53
SW16	Wangaratta	043861	5956675	140	59	-6.3	-36	3.90	0.01	1.34	3.97	1.54	1.01	2.61
SW15	Wangaratta S.	044237	5974395	127	61	-6.4	-37	3.71	0.02	1.40	3.97	1.72	0.97	2.45
SW14	Tarrawingee	045113	5970115	115	54	-6.3	-36	3.05	0.03	1.35	3.72	1.65	0.94	2.42
SW13	Everton	045732	5966695	97	53	-6.4	-37	3.11	0.13	1.23	3.50	1.51	0.93	2.28

SW12	Whorley	046451	5959045	80	47	-6.5	-36	2.23	0.08	1.18	3.38	1.50	1.02	2.44
SW11	Whorley E.	047013	5966685	72	46	-6.6	-37	2.10	0.02	1.04	3.17	1.47	0.89	2.23
SW10	Myrtleford	047453	5952805	65	39	-6.8	-40	1.47	0.18	0.84	2.97	1.46	1.06	2.17
SW9	Salziers Ln	047837	5949855	57	39	-6.7	-40	1.45	0.02	0.84	2.63	1.47	0.75	1.95
SW8	Ovens	048239	5947285	50	40	-6.8	-40	1.40	0.06	0.79	2.63	1.46	0.74	1.93
SW7	Eurobin	048684	5945025	47	38	-6.8	-37	1.38	0.11	0.71	2.56	1.40	0.72	1.91
SW6	Porpunhah N.	048833	5941035	39	39	-7.0	-39	1.35	0.01	0.71	2.65	1.44	0.72	1.92
SW5	Porpunkah	049182	5938925	34	43	-6.8	-40	1.34	0.01	0.78	2.83	1.69	0.73	2.28
SW4	Bright	049936	5935505	22	42	-7.0	-38	1.15	0.32	0.86	2.65	1.64	0.69	2.26
SW3	Smoko	050568	5927365	11	38	-7.2	-41	0.90	0.01	0.87	2.70	1.66	0.70	2.39
SW2	Trout Farm	050560	5918975	5	38	-7.3	-39	0.89	0.09	0.96	2.82	1.67	0.73	2.44
SW1	Harrietville	050566	5917175	0	33	-7.5	-41	0.76	0.02	0.60	2.42	1.43	0.61	2.20
Tributaries														
SW22	King @ Oxley	044442	5966735		52	-6.5	-36	3.46	0.02	0.88	4.04	1.05	1.27	2.75
SW21	Buffalo @ Myrtleford	047182	5954215		41	-6.3	-37	1.87	0.11	1.21	2.74	1.26	0.81	2.08
SW20	Buckland @ Porpunkah	049049	5938625		35	-6.7	-39	1.20	0.01	0.63	2.43	1.18	0.68	1.74

Table S2.2 Geochemistry of groundwater in the Ovens Catchment in September 2009 and March 2010

Bore No.	Location	Catchment	Easting	Northing	Bore Screen Depth (m)	EC (µS/cm)	δ18O SMOW	δ2H SMOW	Cl (mg/L)	NO3 (mg/L)	SO4 (mg/L)	HCO3 (mg/L)	Na (mg/L)	Mg (mg/L)	K (mg/L)	Ca (mg/L)
September 2009																
B51743	Bright	Upper	499291	5935508	5-11	82	-6.2	-36	2.80	2.89	0.55	40	5.35	4.30	0.75	3.37
B51747	Bright	Upper	499190	5935414	2-20	58	-6.1	-35	3.10	5.36	0.21	13	3.03	2.01	0.61	1.66
B51745	Bright	Upper	499139	5935375	5-11	63	-6.2	-37	2.16	2.41	0.75	19	2.25	2.51	0.85	1.83
B51744	Bright	Upper	498933	5934911	6-13	56	-4.7	-26	2.73	7.33	1.00	30	5.21	3.11	0.99	3.20
B51737	Bright	Upper	498445	5935658	36-42	111	-6.4	-35	2.41	0.14	0.62	60	7.57	6.02	0.81	4.64
B51738	Bright	Upper	498397	5935420	58-63	200	-6.3	-37	2.39	0.01	0.27	88	17	6.21	0.95	4.70
B51735	Bright	Upper	498391	5935314	30-42	74	-6.3	-35	3.58	0.79	0.60	28	4.88	2.69	0.84	2.46
B51736	Bright	Upper	498382	5935299	20-26	53	-6.2	-36	3.29	1.30	0.26	23	4.53	2.48	0.72	1.69
B109461	Bright	Upper	497818	5935267	20-26	83	-6.3	-37	3.42	4.52	0.54	43	5.88	4.04	1.30	4.19
B109462	Bright	Upper	497818	5935267	45-51	100	-6.2	-35	3.48	0.40	0.30	44	5.88	4.41	0.91	4.58
B88271	Porepunkah	Upper	493294	5938062	8-14	100	-6.4	-37	3.98	0.42	2.57	49	3.53	6.93	0.75	4.90
B88274	Porepunkah	Upper	493256	5938067	35-53	64	-6.8	-39	1.85	0.22	0.78	30	4.24	3.59	0.70	2.14
B48069	Eurobin	Middle	487803	5944698	5-8	129	-5.9	-34	3.59	0.07	9.60	35	8.39	3.35	1.14	2.36
B48068	Eurobin	Middle	487657	5944643	7-13	74	-5.4	-31	3.52	0.09	2.63	20	3.39	2.36	1.99	1.73
B48067	Eurobin	Middle	487519	5944594	12.0	92	-5.5	-32	4.14	14	0.71	15	4.16	3.57	1.08	2.91
B48066	Eurobin	Middle	487411	5944553	9-15	78	-5.6	-32	3.90	17	0.69	9	3.28	2.92	1.35	2.45
B83232	Myrteford	Middle	474884	5953288	6-12	107	-5.3	-33	9.33	2.59	6.64	33	11	3.97	1.67	3.17
B83231	Myrteford	Middle	474704	5953010	8-14	49	-6.7	-37	2.08	1.46	3.19	24	3.53	2.48	0.72	2.09
B83229	Myrteford	Middle	474607	5952916	8-14	87	-6.5	-39	2.01	0.71	0.43	38	3.13	4.06	0.80	4.12
B83230	Myrteford	Middle	474604	5952937	8-14	45	-6.7	-37	1.87	0.57	1.56	22	3.21	2.13	0.59	1.86
B102783	Whorouly	Middle	464087	5959833	5-11	107	-6.7	-37	2.06	0.28	0.07	63	10	5.67	0.66	4.52
B110738	Oxley	Middle	444240	5966742	19-44	106	-7.1	-41	27	0.05	0.16	35	15	0.78	2.71	22
B11326	Wangaratta	Lower	439879	5982755	23.7	1341	-5.5	-33	298	30	25	389	192	60	3.07	33
B11493	Wangaratta	Lower	439422	5982189	16.5	920	-4.8	-31	134	0.00	15	322	134	27	0.64	12
B302296	Boorhaman E.	Lower	437925	5992950	71-77	567	-5.5	-34	115	0.03	0.91	123	95	4.52	2.53	7.40
B11323	Boorhaman E.	Lower	437924	5992953	17.4	536	-5.6	-35	96	0.62	4.80	155	89	3.93	1.40	27
B50788	Boorhaman	Lower	442072	5999081	60-72	3800	-5.9	-36	923	1.67	214	333	654	61	7.02	59
B50789	Boorhaman	Lower	442072	5999081	18-30	12020	-6.8	-43	3830	3.44	879	1250	2331	356	13	357

B11306	Peechelba	Lower	432684	5994603	16	1194	-6.3	-38	299	2.36	38	151	163	51	1.70	11
B11311	Bundalong S.	Lower	427007	6005559	16	2270	-5.8	-36	628	36	172	218	389	57	2.38	20
B11310	Bundalong S.	Lower	427237	6005560	14	2250	-5.8	-35	570	31	159	220	378	61	3.14	25
March 2011																
B1	Bright	Upper	499270	5935517	2-4	95	-5.9	-34	2.95	0.48	0.35	32	7.34	3.82	0.85	4.52
B2	Bright	Upper	499260	5935513	2-4	82	-6.1	-31	2.87	0.78	0.30	24	7.42	3.56	0.92	4.28
B51743	Bright	Upper	499291	5935508	5-11	76	-6.2	-35	3.32	3.00	0.46	33	9.25	4.54	1.05	5.20
B51744	Bright	Upper	498933	5934911	6-12	60	-6.0	-34	4.51	6.23	1.46	24	6.97	3.29	1.54	3.31
M1	Myrtleford	Middle	474605	5952919	4-6	90	-6.5	-35	2.37	0.29	4.96	17	7.46	3.18	0.96	3.44
M2	Myrtleford	Middle	474605	5952936	4-6	68	-6.5	-34	2.43	3.54	4.87	23	8.21	4.67	1.39	5.25
T1	Tarrawingee	Lower	451112	5970209	5-7	367	-5.5	-32	47	0.50	0.24	52	52	8.13	2.25	9.51
T2	Tarrawingee	Lower	451121	5970212	5-7	364	-5.5	-33	48	1.05	0.36	45	55	7.39	3.76	11
T3	Tarrawingee	Lower	451136	5970245	6-8	315	-5.3	-31	46	1.53	0.28	45	54	8.73	3.15	10
B88271	Porpunhah	Upper	493294	5938062	8-14	107	-6.3	-33	3.75	0.38	2.47		6.36	8.52	1.09	8.21
B48066	Eurobin	Middle	487411	5944553	9-15	70	-5.4	-32	3.97	18.99	0.64		6.19	4.13	2.38	4.01
B48067	Eurobin	Middle	487519	5944594	12.0	77	-5.3	-31	3.24	21.64	0.74		6.29	4.40	1.50	4.78
B48068	Eurobin	Middle	487657	5944643	7-13	69	-5.5	-30	3.42	18.51	0.96		7.50	4.47	1.16	4.00
B48069	Eurobin	Middle	487803	5944698	5-8	122	-5.9	-33	4.07	0.10	7.50		13	4.22	1.37	3.78

Table S2.3 Geochemistry of spring and spring-fed stream at Location BC8

Site No.	Type of Sample	EC (uS/cm)	δ18O SMOW	δ2H SMOW	Cl (mg/L)	NO3 (mg/L)	SO4 (mg/L)	HCO3 (mg/L)	Na (mg/L)	Mg (mg/L)	K (mg/L)	Ca (mg/L)
March 2011												
SR	Spring-fed stream	49	-6	-34	1.57	2.15	1.75		5.78	2.83	1.67	3.34
SS	Spring	64	-6.1	-36	1.85	0.02	1.02		4.91	4.15	2.01	5.72
June 2011												
SR	Spring-fed stream	49	-6.2	-36	2.15	1.62	1.56		3.94	2.96	1.19	2.4
SS	Spring	52	-6.3	-35	2.21	1.26	0.65		4.13	3.26	0.91	2.9

Chapter 3

Using continuous discharge and river chloride concentrations to estimate long-term baseflow contribution to the Ovens Catchment, southeast Australia

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Abstract

The flow duration curve (FDC), local minimum method, recursive digital filters and chloride (Cl)-based chmeical mass balance (CMB) were used to estimate the contribution of baseflow in the Ovens Catchment in the southeast Murray-Darling Basin, Victoria, Australia. The total baseflow flux between 2004 and 2014 calculated using two recursive digital filters was ~46% of the total discharge

with an annual baseflow of 48 *to* 50%, 46 *to* 47% and 37 *to* 46% of the annual discharge for 2008, 2012 and 2013, respectively. Slightly higher estimates were obtained from the local minimum method: 52% of the total discharge for 2004 *to* 2014 with an annual baseflow flux of 57%, 48% and 52% of the annual discharge for 2008, 2012 and 2013, respectively. However, the baseflow fluxes from the FDC and CMB are significantly lower. The FDC calculated that total baseflow was 13% of the total discharge between 2004 and 2014 with an annual baseflow of 33%, 30% and 15% of the annual discharge for 2008, 2012, and 2013, respectively. The CMB calculated that total baseflow was 9.5% of the total discharge between 2004 and 2014 with an annual baseflow of 8.4%, 10% and 8.0% of the annual discharge for 2008, 2012, and 2013, respectively. The differences in the baseflow estimates between the hydrograph separation methods and CMB are often greater during high flow periods. It is likely that in addition to groundwater, water from transient water stores, such as river banks, the unsaturated zone, and pools and/or disconnected channels on the floodplain, contributes a significant portion of baseflow in the Ovens Catchment. Since discharge from transient water stores exhibits the low-frequency characteristic of groundwater in a hydrograph, it is aggregated with groundwater inflow in the hydrograph separation methods. This leads to high baseflow estimates in hydrograph separation methods. Moreover, discharge from transient storage is often highest during and following high flow events when the transient water storages are replenished during the onset of high flow events. This phenomenon results in greater differences in the baseflow estimates during and after high flow events between the techniques. The variation in the Cl concentrations of the Ovens River in respect to river discharge also suggests discharge from transient water storage in the Ovens Catchment. The increase in the river Cl concentrations prior to some flood peaks may be the result of flushing saline water from the unsaturated zone and evaporated pools or disconnected channels on the floodplains. The persistent low Cl concentrations in the river following flood peaks suggest the influx of low salinity water from transient storage into the river. Without discharge from transient storage, the Cl concentrations in the Ovens River would increase rapidly after flood peaks. Overall, discharge from transient water store is an important process in the Ovens Catchment and needs to be considered when managing the riverine environment in the catchment.

KEYWORDS: River-groundwater interactions, Baseflow, Groundwater inflow, Transient water stores, Hydrograph separation, Flow duration curve, Chemical mass balance, Ovens River

3.1 Introduction

Comprehensive knowledge of the various sources of water in rivers and the proportion of water contributed from each source is vital to understand catchment hydrology and for protecting our rivers and their riparian ecosystems. The water in a river may be divided into two broad components: quickflow and baseflow (Fig. 3.1) (Gordon *et al.*, 2004; Brodie *et al.*, 2007). Quickflow includes direct precipitation and overland flow that contribute to rivers during and immediately following a rainfall event; this is generally termed "event water". As well as event water, quickflow can also include older water from the unsaturated zone or pools on the floodplain displaced by the event water, or groundwater mobilised by hydraulic loading associated with recharge on the floodplains or in the valleys (Sklash and Farvolden, 1979; Kirchner, 2003; Fiori, 2012). Baseflow (also known as delayed flow) reflects water sources that have been stored in a catchment for a period of time after a rainfall event and are then discharged back to sustain the river between rainfall events (Hall, 1968; Price, 2011; Cartwright *et al.*, 2014). Baseflow commonly includes groundwater, but there may also be contributions from transient water stores that drain slowly to the river, for example, river banks, the unsaturated zone and pools on a floodplain (Mulholland, 1993; Griffiths and Clausen, 1997; Sophocleous, 2002; Chen and Chen, 2003; Price, 2011; Cartwright *et al.*, 2014).

The relative importance of the potential water sources that contribute to rivers is influenced by many factors, including fluvial geomorphology, topography and geology of the

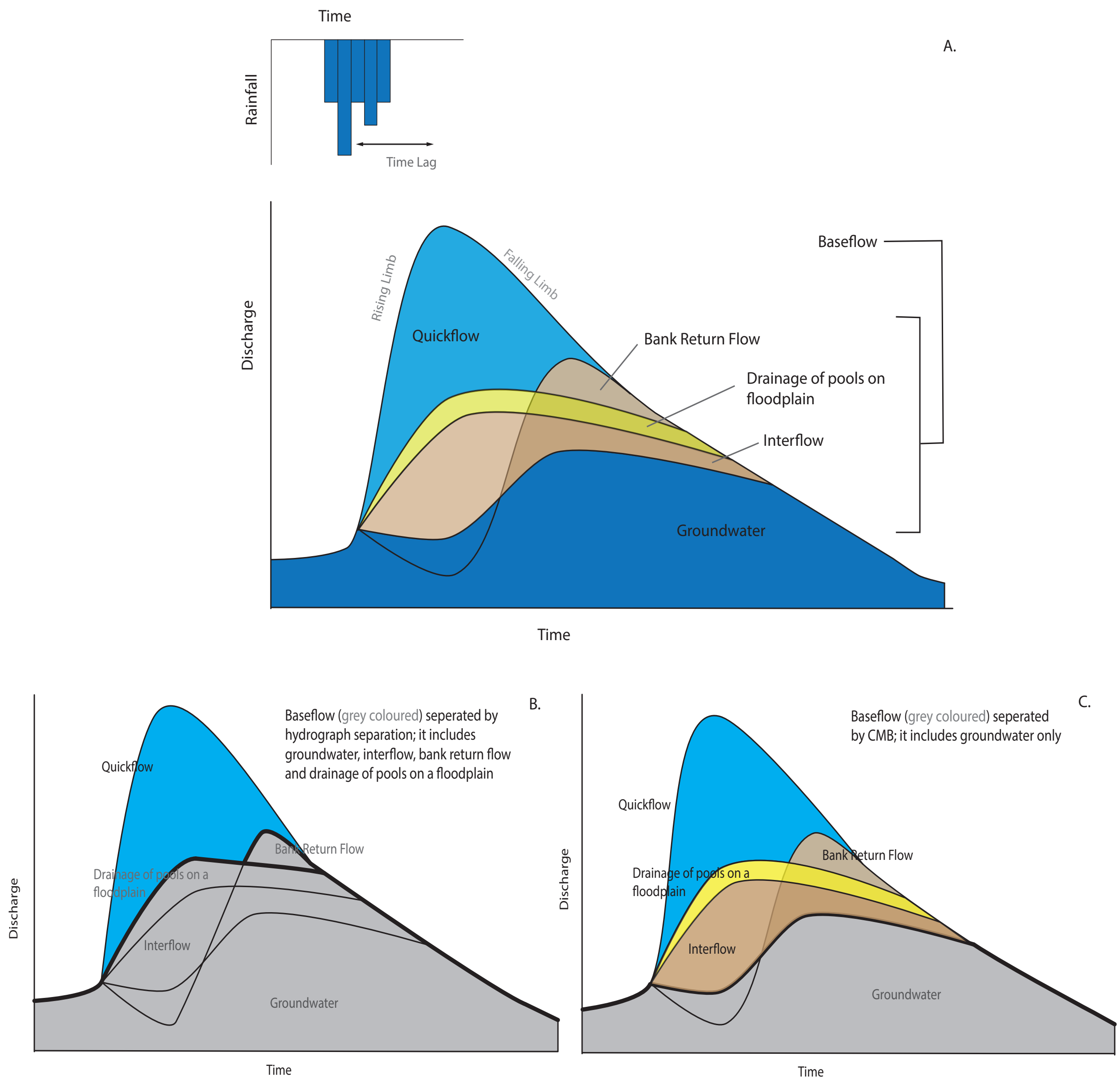

Figure 3.1 A) A typical river hydrograph, showing various flow components in a river. Modified from Freeze and Cherry (1979) and Brodie et al. (2007). B) Baseflow components (grey coloured) separated by hydrograph separation methods. It includes all delayed flows: groundwater inflow, bank return flow, flow from the saturated zone and flows from pools on a flooplains. C) Baseflow component (grey coloured) derived from CMB. It includes groundwater inflow only.

catchment, the regional groundwater flow pattern, and the distribution and amount of rainfall. Rivers with steep banks and coarse-grained bank sediments are likely to have a greater potential of bank storage, and bank return flows may contribute substantial volumes of water to the river as part of the baseflow component when river levels decline after rainfall events (Whiting and Pomeranets, 1997; Chen and Chen, 2003). Catchments with a steep topography and thin soils generally contribute large volumes of water to the river via overland flow following a rainfall event but less volumes of water via interflow through the unsaturated zone (Akbarimehr and Naghdi, 2012; El Kateb *et al*., 2013). Catchments with a mature floodplain generally have many disconnected channels. These disconnected channels can become potential water stores during rainfalls and later provide flow to the river after rain events (Newson, 1994; Makaske, 2001; Cartwright *et al.*, 2014). Rivers are likely to receive a greater volume of groundwater inflow when the regional groundwater flow is perpendicular to the river rather than parallel to the river (Atkinson *et al.*, 2015). Bank recharge and bank return flow are likely to be significant when there is a large increase in the river stage as a result of a heavy rainfall event (Chen and Chen, 2003). As indicated earlier, the contribution of various water sources in rivers varies throughout a rainfall event. It can also change over a longer timescales, e.g. decades. A decrease in rainfall associated with drought may cause a reduction in the contributions of surface runoff and of bank return flow in the river, thereby increasing the proportion of groundwater in the baseflow component (Mahe, 1997; Brodie *et al.*, 2007). Deforestation can increase the contribution of surface runoff in rivers but reduces infiltration and groundwater recharge, leading to groundwater making a lower contribution to baseflow (Mumeka, 1986; Winter *et al.*, 1998; Gholami, 2013). Additionally, deforestation in the undulated part of the catchment may increase the rate at which the surface runoff supplies the river, causing a rapid increase in the river level during rainfall events which in turn increases bank storage (Mumeka, 1986; Olang and Fürst, 2011). Conversely, land clearing on floodplains may cause an increase in groundwater recharge and in turn raises the water table, which causes greater amounts of groundwater inflow during low flow periods (Dahlhaus *et al.*, 2000; Scanlon *et al.*, 2006).

When a river relies on baseflow to maintain its flow throughout during dry periods, it is important to limit the use of groundwater nearby. If groundwater is saline or polluted, it can have negative impacts on the quality of river water. It is important, therefore, to understand the inputs to the river, especially the proportion of baseflow, and the relative amount of water from the various stores that contribute to the baseflow in order to protect and manage river systems.

3.1.1 Determining sources of water in the river

The sources of water in river and their relative contribution to the river flow can be determined by various techniques. Techniques based on the river hydrograph, such as frequency analysis or hydrograph separation, are commonly used in studying the sources of water in rivers (Evans and Neal, 2005; Brodie *et al.*, 2007). Frequency analysis examines the relationship between magnitude and frequency of streamflow discharge over a period of time, and one type of frequency analysis is flow duration curve (FDC). FDC describes the percentage of time that a given flow rate is equalled or exceeded (Searcy, 1959; Vogel and Fennessey, 1995). The low-flow portion of a FDC (i.e. below median flow) represents low flow conditions and can provide qualitative or semi-quantitative estimates of baseflow fluxes (Smakhtin, 2001; Brodie *et al.*, 2007). Hydrograph separation can may be done graphically or by using digital filters. The graphical approach, such as local minimum method, estimates baseflow by extrapolating the minimum discharges within a specific time window which is determined by the catchment area (Pettyjohn and Henning, 1979; Sloto and Crouse, 1996; Aksoy *et al.*, 2009). Filter-based hydrograph separation relies on the principle that the high frequency variability of the streamflow in a hydrograph is mainly caused by quickflow and can be filtered out by passing a low-pass filter to retain the low-frequency baseflow (Nathan and McMahan, 1990; Eckhardt, 2005; Brodie *et al.*, 2007). These hydrographic analysis methods are simple to use and can be readily automated to produce reproducible results. Furthermore, a large volume of discharge datasets covering an extended period of time is commonly available for many river systems. Discharge data integrates the effects of climate, topography and geology on the flow upstream of the gauge, and thus is able to provide an averaged temporal variation of the composition of river flows for the entire catchment upstream of the gauge (Searcy, 1959).

These techniques, however, do not have any hydrological basis and do not generally take the mechanisms of quickflow and baseflow, and the factors that controlling these two flow compnonets, such as catchment slope, size and geomorphology into consideration (Hewlett and Hibbert, 1967; Freeeze, 1972; Hornberger *et al*., 2014). Additional parameters have been added to some hydrograph separation algorithms in order to improve the accuracy of estimating baseflow fluxes. One example of these parameters is the maximum value of the baseflow index (BFI_{max}) which is controlled by the catchment geomorphology, but the BFI_{max} parameter is often determined subjectively (Eckhardt, 2005). Hydrograph separation methods tend to aggregate all the delayed flows into the baseflow component (Fig. 3.1) (Nathan and McMahan, 1990; Halford and Mayer, 2000; Evans and Neal, 2005; Cartwright *et al.*, 2014); thus assuming that the baseflow is dominated by groundwater will overestimate the groundwater inflows.

Hydrochemistry and environmental tracers can also be used to determine the sources of water in rivers since the chemical composition of various water sources is often different. Chemical mass balance (CMB) has been used to estimate mixing ratios of river water and groundwater (Cey *et al.*, 1998; Yu and Schwartz, 1999; Cartwright *et al.*, 2011; Cook, 2013). As indicated in chapter 2, CMB yields better estimates when the chemical constituent used has significant different concentrations in river water and in groundwater, the chemical constituent is conservative (i.e. its concentration does not change due to in-river processes) or the rate of production or loss for the chemical constituent is well understood. Since the transient water stores that contribute to baseflow are derived from the river or surface runoff, they share similar chemical characteristics with quickflow. CMB does not account for these flows and thus gives estimates of groundwater inflow and not the total baseflow flux (Fig. 3.1) (McCallum *et al.*, 2010). Longitudinal stream chemical sampling is effective in constraining the spatial variation in the relative proportions of surface runoff and groundwater in rivers, but often not the temporal variation (Cartwright *et al.*, 2011; Cook, 2013; Unland *et al.*, 2013; Atkinson *et al.*, 2015). Some chemical constituents can be measured continuously in order to derive the temporal variability of groundwater inflow, but often require an advanced set-up. One possible alternative is to utilise electrical conductivity (EC) in rivers. EC provides a general indicator of water chemistry and generally has a close correlation with the chloride (Cl) concentration in water which is a conservative ion (Norton and Friedman, 1985; Abyaneh *et al.*, 2005; Sanford *et al.*, 2011). It can be simply measured continuously for an extended period of time with EC loggers, and many long-term EC datasets commonly from sites where discharge is measured also exist, which allow comparison of CMB and techniques based on the hydrograph.

The concentration of the dissolved ions in rivers varies in response to an increase in discharge. Previously it was thought that the decrease in river salinity was merely a dilution of river chemistry by quickflow. However, the variation of the dissolved ion concentrations in rivers is governed by the mixing of dissolved ions from various components of quickflow and baseflow, and the sequence in which these various flow components contribute to the river during a flow event (Evans and Davies, 1998, Carroll *et al.*, 2007). Prior to a flow event, the river is dominated by groundwater inflow. During the early phase of a flow event, surface runoff dominates, followed by the inflowing water from transient water stores, such as river banks and the unsaturated zone. With the diminishing of the event flow during the receding phase, the groundwater inflow is re-establish gradually. Such a sequence of water contribution in a river results in a clockwise circular pattern of river Cl concentrations in respect to discharge, referred to as clockwise hysteresis, where the concentrations at a given discharge during the rising phase of a flow event are higher than those at the same discharge during the receding phase (Evans and Davies, 1998). Other sequences of water contributions or different proportions of contribution from various flow components can alter the shape and the rotational direction of the hysteresis loop. The hysteresis in the discharge vs. concentrations relationship has been used to determine the mixing of various sources of water in rivers and the relative timing of this mixing during flow events (House and Warwick, 1998; Rose, 2003; Cartwright *et al.*, 2014).

3.1.2 Aims

The aim of this study is to use continuous datasets of river discharge and EC to identify and estimate various sources of baseflow contributing to the flow of the Ovens River, southeast Australia, over a 10-year period. Baseflow fluxes were constrained and compared using four methods; frequency analysis (mainly FDC), graphical and filter-based hydrography separations, and Cl-based CMB. Moreover, the relationship between EC-derived Cl concentrations and discharge was explored in order to further constrain the characteristics of baseflow in the Ovens River. The spatial variations of groundwater inflows in the Ovens River were examined using major ion chemistry and ^{222}Rn in chapter 2. The temporal aspects of groundwater inflows and of other sources of baseflow from this study will further enhance our knowledge of the hydrological processes, assisting in managing the river in this and other catchments in the Murray-Darling Basin. Showing the ability of using the combination of discharge and EC datasets to study the sources of water to rivers is beneficial to environmental management since these datasets are readily available without a significant cost.

3.2 Hydrological setting

The Ovens River is the major river in the Ovens Catchment along with three major tributaries; Buckland, Buffalo and King Rivers (Fig. 3.2a). The Ovens Catchment is 7813 km^2 in size and occupies 0.7% of the Murray-Darling Basin, southeast Australia (CSIRO, 2008). Despite

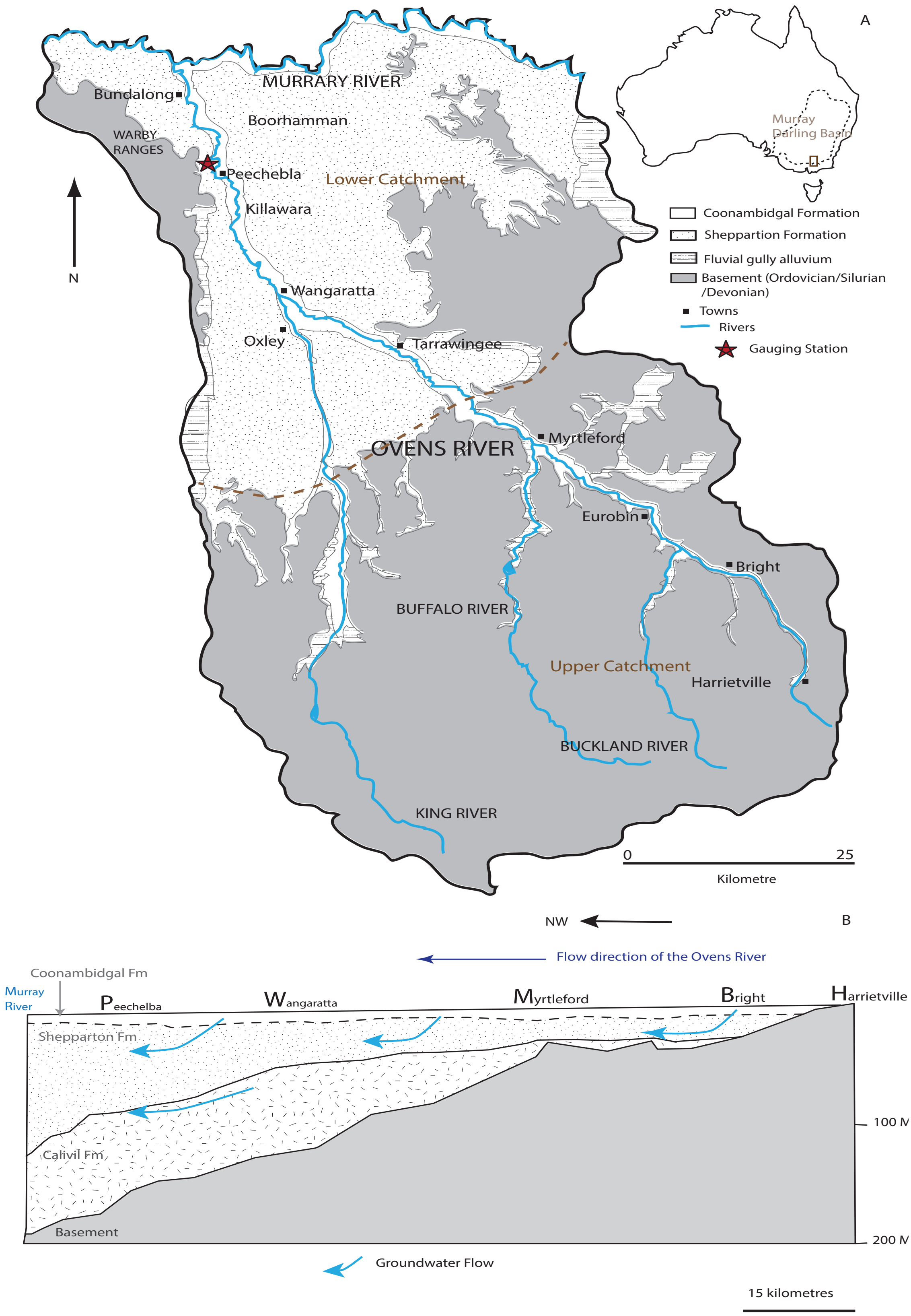

Figure 3.2 A) The map of the Ovens River with surface geology, township and the location of the Peechelba gauging station. The river flows north-westwards and discharges to the Murray River. B) Simplified geological cross-section of the Ovens Catchment along the Ovens River. Data from Lawrence (1988), van den Berg and Morand (1997) and Water Measurement Information System (2013).

its small size, the Ovens River supplies 11% of the total flow to the Murray River (CSIRO, 2008). The catchment is characterised by narrow and steep-sided valleys in the upper catchment and a well-developed floodplain and extensive regions of low topography in the lower catchment. The catchment comprises Palaeozoic basement rocks overlaid by Tertiary to recent fluviatile sediments that are up to 210 m thick (Fig. 3.2b) (Lawrence, 1988; van den Berg and Morand, 1997). The basement consists of metamorphosed Ordovician turbidites together with Silurian and Devonian granite intrusions, forming a fractured-rock aquifer with a hydraulic conductivity of 0.01 *to* 1 m day^{-1} (Slater and Shugg, 1987). The deepest sediments belong to the terrestrial Tertiary Calivil Formation and are up to 50 m thick on the floodplain. The Calivil Formation contains consolidated gravel, sand, silt and cobbles with a hydraulic conductivity of 5 *to* 50 m day^{-1} (Shugg, 1987; Cheng and Reid, 2006). The overlying sediments are the fluvio-lacustrine Quaternary Shepparton Formation and the Holocene Coonambidgal Formation which are contiguous and indistinguishable in the catchment. The Shepparton Formation contains intercalated lenses of fine to coarse sands and gravel. These two formations together are up to 170 m thick and contain mostly sand, gravel, silt and clay with fragments of basement rocks and minerals that form heterogeneous, unconfined to confined aquifers. The hydraulic conductivity of the Shepparton and Coonambidgal Formations vary from 0.1 to 60 m day^{-1} with an average of 0.2 *to* 5 m day^{-1} (Tickell, 1978). The sediments in of the Shepparton and Coonambidgal Formations are coarser grained in the valleys and become finer grained and more mature on the floodplain. Groundwater from these two formations interacts with the Ovens River. The surface aquifers receive recharge through direct infiltration on the valley floor, and via exposed and weathered bedrock at the margins of the valleys. The vertical head gradients throughout the Ovens Catchment are generally downward, while the vertical head gradients within a few tens of metres of the river are upwards (Water Measurement Information System, 2013). The hydraulic gradient of groundwater toward the Ovens River decreases down the catchment. The regional groundwater flow in the catchment is dominantly northwest, parallel to the valleys (Cartwright and Morgenstern, 2012; Water Measurement Information System, 2013). The Cl concentration of groundwater in the surface aquifer ranges from 2.8 to 2331 mg L^{-1} and generally increases down catchment (Water Measurement Information System, 2013).

The average rainfall decreases from approximately 1130 mm in the upper catchment to 640 mm on the alluvial plain with most rainfall occurring in the Australian winter months (June *to* September) (Bureau of Meteorology, 2013). Potential evaporation increases northwards and ranges from 0 *to* 40 mm $month^{-1}$ to 125 *to* 200 mm $month^{-1}$ in winter and summer, respectively (Bureau of Meteorology, 2013). The Ovens River drains the northern flank of the Victorian Alps and flows north-westwards (Fig. 3.2a). It is a single channel confined within a steep-sided valley south of Myrtleford and then develops into a network of meandering and anastomosing channels north of Whorouly before discharging to the Murray River. The Buckland and Buffalo Tributaries are entirely hosted within upper catchment valleys and joins with the Ovens River at Porepunkah and Myrtleford, respectively. The King River is the second longest river in the catchment. While the upper section of the King River is hosted within a valley, the lower section flows across a floodplain and joins with the Ovens River at Wangaratta. The Ovens River is unregulated itself, but two of the major tributaries, Buffalo and King Rivers, have a water storage in the upper reaches (Lake Buffalo of 23.5 GL and Lake William Hovell of 13.5 GL, respectively) (Goulburn-Murray Water, 2013). The mean monthly discharge at Peechelba is between 4,090 and 414,793 ML day^{-1} with high flows occurring in winter (Water Measurement Information System, 2015). The Cl concentrations of the Ovens River vary from 0.7 to 2.2 mg L^{-1} in the upper catchment and from 1.6 to 5.0 m L^{-1} on the floodplain (Chapter 2). The change in the river Cl concentrations reflects both the groundwater discharge from the surface aquifers and the longitudinal increase in groundwater Cl concentrations from the upper to lower catchments.

The Riverine Plain and alluvial flats in the Ovens Catchment are primarily cleared for agricultural activities, including grazing, horticulture, vineyards, and orchards, while the hills and mountains are covered by native eucalyptus and plantation forests. Water extraction from both surface and groundwater resources is relatively low, being 5% of the total water resource available in the catchment (Victorian Government Department of Environment and Primary Industries, 2013).

3.3 Data sources and methods

River discharge, river geochemistry and groundwater geochemistry data are from the Victorian Water Measurement Information System (2015) and unpublished Victoria Department of Primary Industries data. River discharge and EC are monitored continuously (at intervals between 15 and 60 minutes) at the Peechelba gauging station, which is the last gauging station on the Ovens River before the river drains to the Murray River (Fig. 3.2a). As the Peechelba gauging station is the lowermost gauge in the catchment, information on the sources of water to the river over time from the whole catchment is captured. In addition to the continuous EC measurements, spot measurements of geochemistry, including Cl concentrations, has also been collected (between 1983 and 2005) at a wide range of river flows. Measurements of the river EC and Cl concentrations are well correlated with a R^2 value of 0.96 (Fig. 3.3). Thus, the Cl concentrations of the Ovens River can be estimated with a high degree of certainty from the EC values. The stream hydrograph and the calculated Cl concentrations are used to estimate baseflow inputs using frequency analysis, hydrograph separation by graphical and filtering methods, and CMB. Baseflow fluxes from these calculation are also compared with those from the Rn mass balance (chapter 2).

3.3.1 Frequency analysis

The flow duration curve (FDC) displays the percentage of time during which the specified stream discharge is equalled or exceeded over the monitoring period. The low-flow

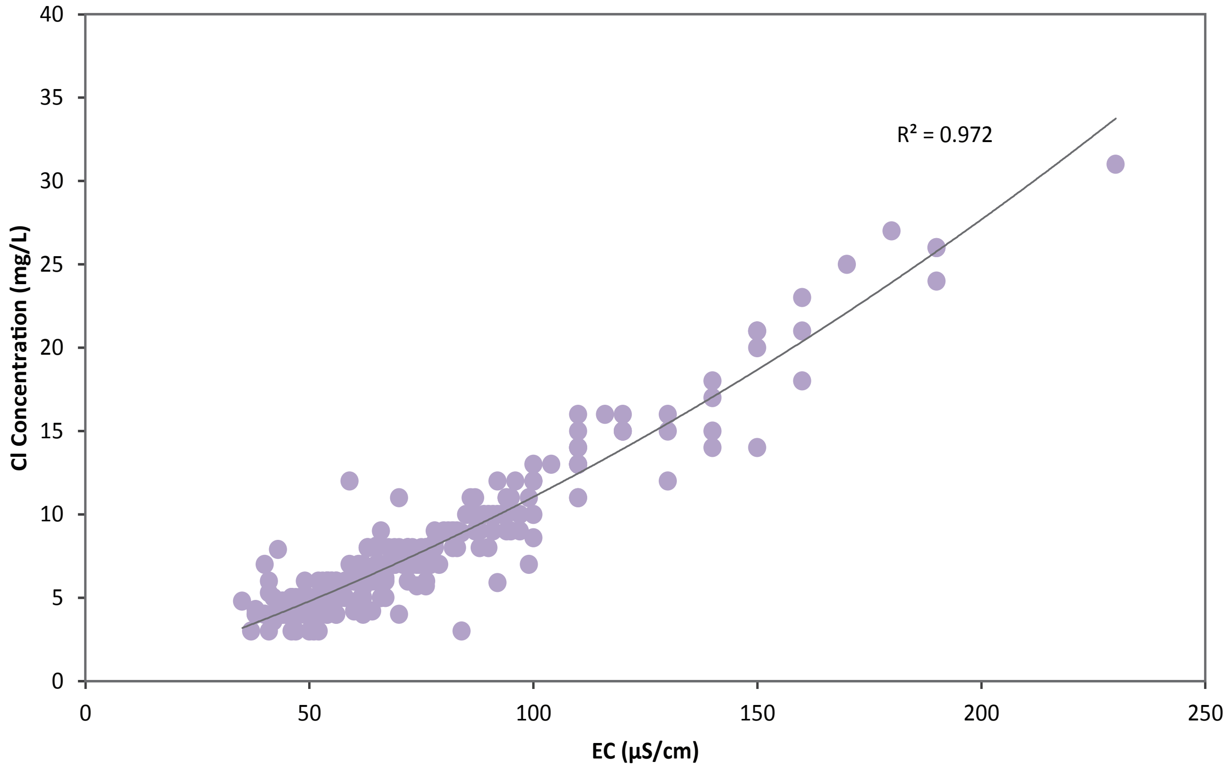

Figure 3.3 Correlation between the EC values and Cl concentrations in the Ovens River. Data from Water Measurement Information System (2013) and unpublished data from the Victorian Department of Primary Industry.

portion of a FDC is often used to characterise baseflow since streams at these flows are mainly dominated by baseflow. The part of a FDC with a flow below the median flow (the flows equalled or exceeded 50% of the time, Q50) arbitrarily represents low flow conditions (Smakhtin, 2001). If this part of the curve has a gentle slope, baseflow contribution is interpreted to be significant, while a steep slope represents an only minor or variable baseflow input (Searcy, 1959; Brodie et al., 2007). Various indexes have been derived from the low-flow part of a FDC, such as Q70, Q90 or Q95. One commonly used index is Q90/Q50, which is the ratio of the discharge which is equalled or exceeded 90% of the time to median flow. Q90/Q50 indicates the proportion of river flow originating from groundwater (Cross, 1949; Searcy, 1959; Smakhtin, 2001). The Q90/Q50 index eliminates the absolute flow unit, which can be affected by the size of the catchment, by dividing the Q90 by the median flow, allowing the comparison of baseflow across catchments with difference sizes.

3.3.2 Hydrograph separation

3.3.2.1 Graphical method

The local minima method in graphical hydrograph separation derives baseflow fluxes by interpolating the data points linearly between minimum discharges that occur within a specific number of days. The discharge windows are defined as 0.5 (2 N^* – 1) days (Sloto and Crouse, 1996; Aksoy *et al.*, 2009), where N is the number of days over which surface runoff occurs which relates to the catchment size (A in square miles) by the empirical relationship, $N = A^{0.2}$ (Sloto and Crouse, 1996).

3.3.2.2 Filter-based method

Recursive digital filters estimates the low-frequency baseflow fluxes from a stream hydrography by removing the high frequency signals of surface runoff. The two filters used in this study are

$$b_k = y_k - \left[\alpha f_{k-1} + \frac{(1+\alpha)}{2}(y_k - y_{k-1})\right] \qquad \textit{Eq. 3.1}$$

(Lyne and Hollick, 1979) and

$$b_k = \frac{(1-BFI_{max})\alpha b_{k-1} + (1-\alpha)BFI_{max} y_k}{1-\alpha BFI_{max}} \qquad \textit{Eq. 3.2}$$

Eckhardt, 2005; Eckhardt, 2008). In both equations, the y is total stream flow on day k, b is the filtered baseflow ($b < y$) on day k, and α is the recession constant. α is estimated from the falling limbs of the hydrograph by calculating $y_k + 1 = \alpha\, y_k$ for every stream discharge value that is part of a recession period of at least five days (Nathan and McMahan, 1990; Eckhardt, 2008). f_{k-1} in the Lyne and Hollick filter is the filtered quickflow on k – 1 day. In the Eckardt filter, BFI_{max} is the maximum value of the baseflow index (BFI) that can be modelled by the algorithm. BFI_{max} cannot be measured but is subjectively assigned based on the catchment lithology and the river flow regime (Eckhardt, 2005). The Lyne and Hollick filter is widely used in Australia (Evans and Neal, 2005). The Eckardt filter is similar to the Lyne and Hollick filter, but can further constrain the calculated baseflow flux by considering the geomorphological characteristics of the catchment via setting the BFI_{max} parameter.

3.3.3 Chemical mass balance

The relative contribution of groundwater in a river can be estimated by mass balance using stream geochemistry:

$$b_k = y_k \frac{C_r - C_{sw}}{C_{gw} - C_{sw}} \qquad \text{Eq. 3.3}$$

(Cey *et al.*, 1998; Yu and Schwartz, 1999), where C_r, C_{sw}, and C_{gw} are the concentrations of the chemical component in the river, surface runoff, and groundwater, respectively. One assumption in this approach is that the chemical characteristic of each water component remains unchanged over time (Stewart *et al.*, 2007). This method yields more accurate baseflow fluxes when the dissolved constituent is conservative, and its concentration is significantly higher in groundwater than in surface water (Stewart *et al.*, 2007; Kish *et al.*, 2010; Cook, 2013). One commonly used dissolved constituent in hydrological studies is Cl which is both conservative and a common major ion in both groundwater and surface water.

3.4 Results

This study examined the data collected between 2004 and 2014 with a focus on several high flow events in 2008, 2012 and 2013 (Fig. 3.4). These three years (2008, 2012 and 2013) were selected to examine how baseflow in the Ovens Catchment varies with various magnitudes of flow events. The magnitude of flow events in 2008 (in the late stage a decade-long drought) was significantly lower than that of flow events in 2012 and 2013 when southeast Australia experienced La Niña events. Furthermore, the data sets, particularly EC values, for these years is either completed or nearly completed.

3.4.1 Rainfall

The mean annual rainfall at Eurobin in the upper catchment and at Wangaratta in the lower catchment ranged from 449 to 1439 mm year^{-1} and from 283 to 728 mm year^{-1}, respectively (Bureau of Meteorology, 2015) (Table 3.1; Fig. 3.4a). While 2006 *to* 2009 and 2013 had rainfall below the long-term average (1146 mm year^{-1} for Eurobin (1910 *to* 2013) and 609 mm year^{-1} for Wangaratta), 2004, 2005, 2011, 2012 and 2014 were close to or above the long-term average (Bureau of Meteorology, 2015). In the years that had a below long-term average rainfall, southeast Australia was experiencing an extended period of drought, often referred to the Millennium drought (van Dijk et al., 2013; Bureau of Metrology, 2015).

3.4.2 Discharge

The annual discharge for the Ovens River at Peechelba between 2004 and 2014 varied from 474,234 to 2,031,126 ML year^{-1} (Table 3.1; Fig. 3.4b). The maximum discharge and median (Q50) discharge between 2004 and 2014 were 93,573 ML day^{-1} and 1,424 ML day^{-1}, respectively (Fig. 3.5). High flow usually occurs between June and October, while low flow occurs between December and March (Fig. 3.4). This yearly pattern in discharge variation mostly repeated throughout the ten-year period, even during the low discharge years. There was a prolonged period with a relatively low discharge (< 1,000 ML day^{-1}) between March 2006 and May 2010. The discharge dropped below 100 ML day^{-1} between October 2006 and April 2007, and was close

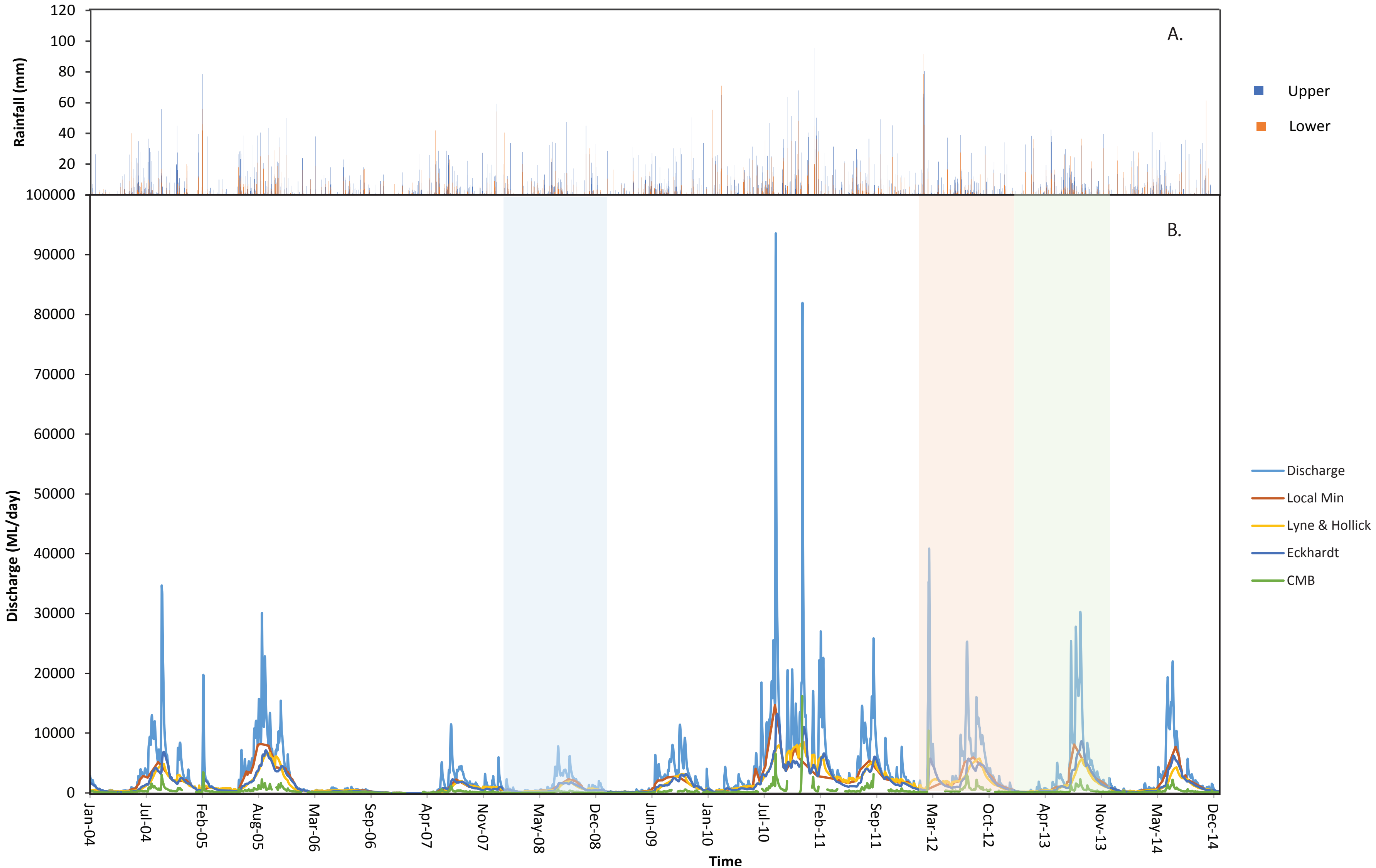

Figure 3.4 See the next page for the figure caption.

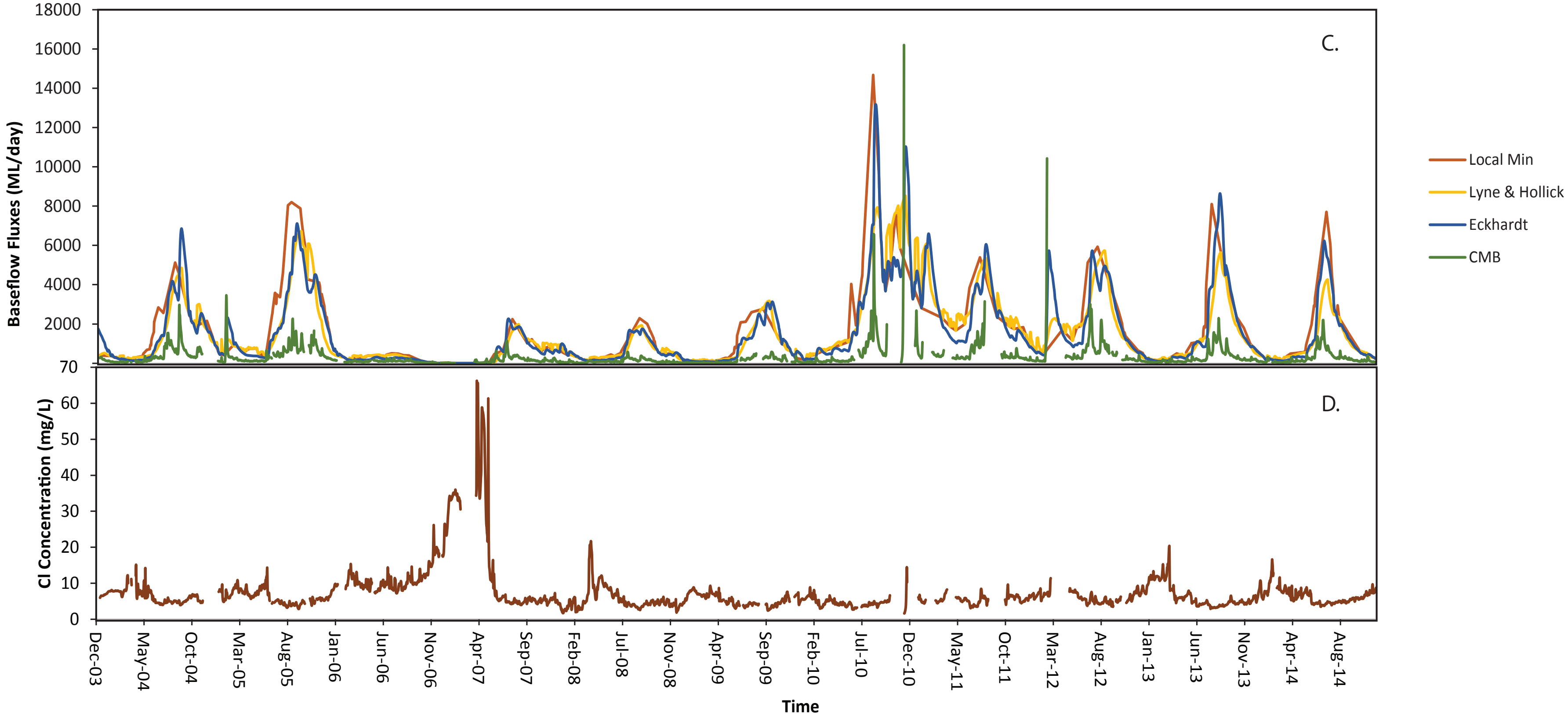

Figure 3.4 (cont.) A) The amount of rainfall at Eurobin in upper catchment (Upper) and at Peechelba in the lower catchment (Lower) in the period of between 2004 and 2014. Data from Bureau of Meteorology (2014). B) The discharge of the Ovens River at Peechelba between 2004 and 2014. High flow events in the shadowed Years (2008, 2012 & 2013) are selected for studying the Cl concentration-discharge relationship C) The baseflux fluxes between 2004 and 2014, calculated by local minimum (Local Min) (Sloto and Crouse, 1996), Lyne and Hollick (1979) and Ekchardt (2005) digital filters, and chemical mass balance (CMB). D) The EC-derived Cl in the river in the same period. Data from Water Measurement Information System (2014).

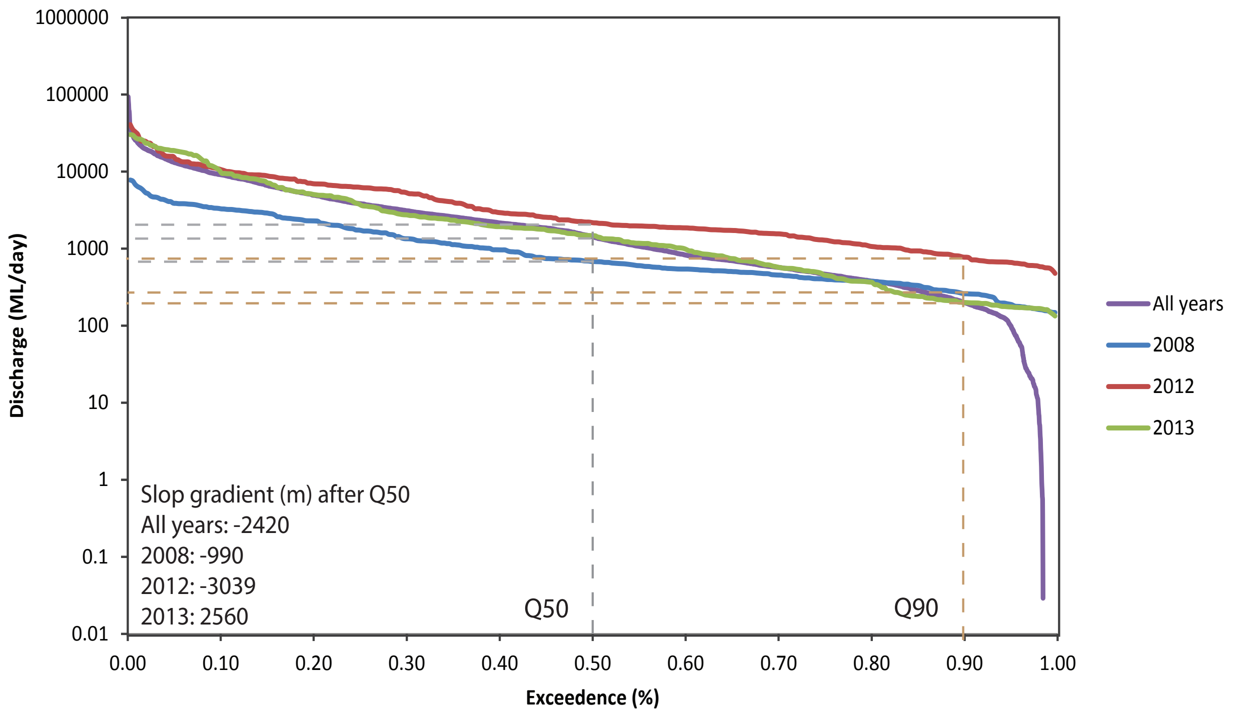

Figure 3.5 The FDCs for the Ovens River in the years between 2004 and 2014, and in 2008, 2012 and 2013 with Q50 and Q90. The section below Q50 for 2008 is relatively flatter than that for all years and 2012 and 2013.

to zero between January 2007 and March 2007. Several large event flows (25,848 *to* 93,573 ML year^{-1}) occurred in the second half of 2010, and 2011, and first half of 2012. In 2008, 2012 and 2013, the annual discharge in 2008 was 474,234 ML, which is considerably lower than that in 2012 and 2013. The annual discharge for 2012 and 2013 were 1,677,648 ML and 1,341,387 ML, respectively. The maximum discharge and Q50 discharge are: 7,783 and 148 ML day^{-1} for 2008; 40,862 and 2,255 ML day^{-1} for 2012; and 30,321 and 1,479 ML day^{-1} for 2013, respectively (Fig. 3.5). The yearly variation in discharge for these three years is similar to the one observed throughout the past 10 years (Figs. 3.7 *to* 3.9).

3.4.3 Chloride concentrations

The Cl concentrations in the Ovens River at Peechelba, which are derived from the river EC values, in the ten-year study period were between 1.6 and 65.7 mg L^{-1}. The mean and the medium river Cl concentrations were 7.2 and 5.7 mg L^{-1}, respectively. Low Cl concentrations

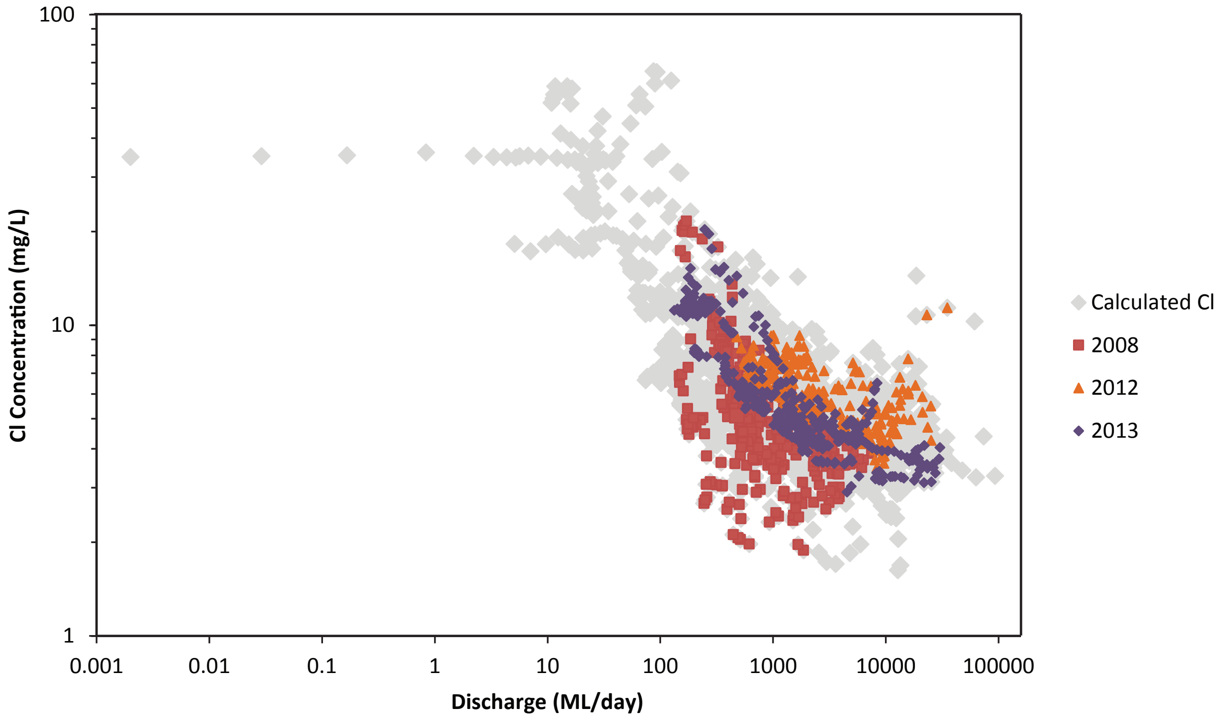

Figure 3.6 The calculated Cl concentrations of the Ovens River at Peechelba in respect to discharge. There is a inverse relationship between the river Cl concentration and river discharge. Data from Water Measurement Information System (2014).

are generally associated with high discharge, while high Cl concentrations are associated with low discharge (Fig. 3.6). The Cl concentrations in the Ovens River throughout the 10 years were generally below 15 mg L^{-1} (Fig. 3.4c). A concentration above 15 mg L^{-1} only occurred three times, beginning in November 2006, March 2008 and March 2013. In November 2006, the Cl concentrations remained above 15 mg L^{-1} for six months and at one stage (in April 2007) increased up to 65.7 mg L^{-1}. These high Cl concentrations followed 10 months of low discharge at a time when the discharge was close to zero. In the three years (2008, 2012 and 2013), the river Cl concentrations throughout the high flow period between May and October were mostly around or below 5 mg L^{-1}, even in 2008 which was one of the dry years (Figs. 3.7c *to* 3.9c). The river Cl concentrations increased from ~6 to 21 mg L^{-1} during the low flow period between November and May in each calendar year. There were occasions where Cl concentrations rapidly increased during or following a high flow event (such as June 2008, March 2012 and June 2013).

Overall, there is an inverse relationship between discharge and river Cl concentrations, i.e. high discharge is associated with low river Cl concentrations or vice versa (Fig 3.6). However, the Cl concentrations were generally higher at any given discharge during the rising phase of an event flow than during the falling phase, particularly during those discrete flow events. Such a variation in river Cl concentrations in respect to discharge over a flow event produces a clockwise hysteresis loop in the discharge-Cl concentrations diagrams (Figs. 3.10 *to* 3.12). These hysteresis loops, however, were not always clockwise in the subsequent high flow events of multiple flow events.

3.5 Calculating baseflow fluxes and expected river Cl concentrations

River discharge and Cl concentrations were used to estimate baseflow fluxes using the FDC, local minimum method and two recursive digital filters (Lyne and Hollick, and Eckardt) from hydrograph separation methods, and CMB. The baselfow estimates from all four methods were then compared with particular emphasis on the discrepancy in baseflow estimates and how the discrepancy varies throughout a year. In addition, the expected Cl concentrations in the river were calculated based on the baseflow fluxes estimated from the hydrograph separation methods. The observed and expected river Cl concentrations were then compared to gain an insight into the possible multiple baseflow components in the Ovens River.

As some lands in the Ovens Catchment have been cleared for agricultural activities, some impacts on the Ovens River are observed in term of the low-to-moderately elevated river NO_3 concentrations (Chapter 2). However, the EC in the river remains low, and thus, the changes in the river EC values (and hence the river Cl concentrations) are likely to reflect the natural changes in the catchment rather than being the results of human activities. There are possibly inaccuracies in deriving baseflow fluxes from hydrograph separation because the Ovens River in the lower catchment is partly regulated by the water storages in the upper reaches of the tributaries. Regulated flows interferes with the baseflow signal in gauged streamflow data. Finally, as the surface water abstraction is less than 2% of the total surface water storage, it thus has only a minimal impacts on baseflow calculations.

3.5.1 Frequency analysis method

The FDC for the Ovens River for the period of 2004 to 2014 has a moderately steep curve below Q50 (Fig. 3.5). The FDC for 2013 closely resembles to that of the 2004 *to* 2014 period. The FDCs for 2008 have a slightly shallow slope below Q50. The Q90/Q50 index for the entire period between 2004 and 2014 was 0.13, while the Q90/Q50 indexes for 2008, 2012, and 2013 were 0.33, 0.30 and 0.13, respectively. The rainfall in the Ovens Catchment in 2008 was significantly below the long-term average, and the catchment in 2008 had a significant lower annual discharge than in 2012 and 2013.

3.5.2 Hydrograph separation methods

The local minimum method was applied to the discharge data of Peechelba between 2004 and 2014. The catchment size at Peechelba is 6230 km^2, giving 2 N* = 11. The estimated baseflow fluxes for the entire period (2004 *to* 2014), 2008, 2012 and 2013 were 7,107,381 ML year^{-1} (52% of the total discharge), 272,704 ML year^{-1} (57% of the annual total discharge), 813,162 ML year^{-1} (48% of the annual total discharge) and 704,036 ML year^{-1} (52% of the annual total discharge), respectively (Table 3.1; Figs. 3.4b & 3.7b *to* 3.9b). Baseflow fluxes generally increase in the period of high flow during the Australian winter between June and September. The high baseflow during the winter period probably reflects groundwater recharge in the Ovens valley that results in hydraulic loading and increase in hydraulic gradients, causing a greater amount of groundwater discharge to the river (Chapter 2). In addition, years with a higher volume of

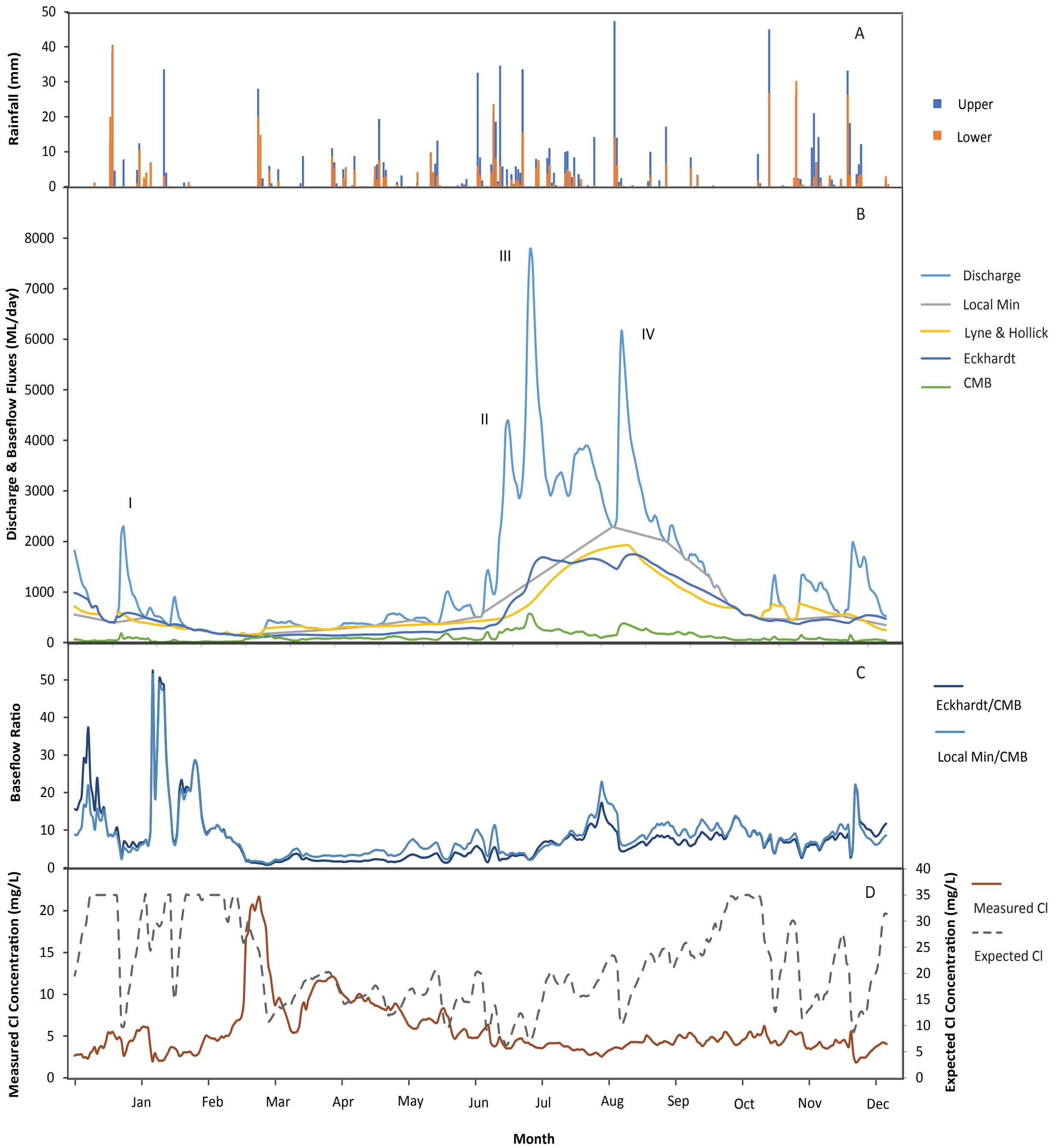

Figure 3.7 A) The amount of rainfall at Eurobin in upper catchment (Upper) and at Peechelba in the lower catchment (Lower) in 2008. Data from Bureau of Meteorology (2014). B) Variation in discharge of the Ovens River at Peechelba in 2008 with baseflow estimates calculated by local minimum (Local Min) (Sloto and Crouse, 1996), Lyne and Hollick (1979) and Eckhardt (2005) digital filters, and chemical mass balance (CMB). C) Ratios of baseflow fluxes derived by local minimum and the Eckhardt digital filter to those estimated by CMB. The ratio (or the difference in baseflow fluxes between the two methods) is larger during high flow periods. D) Variation in river Cl concentrations (Calculated based on EC and Expected) at Peechelba in 2008. The expected Cl concentrations often increase rapidly following high flow events, while the calculated Cl concentration remains low for a period of time. The Cl concentrations vs discharge relationship for the flow events indicated by the Roman numerals are examined later in the chapter.

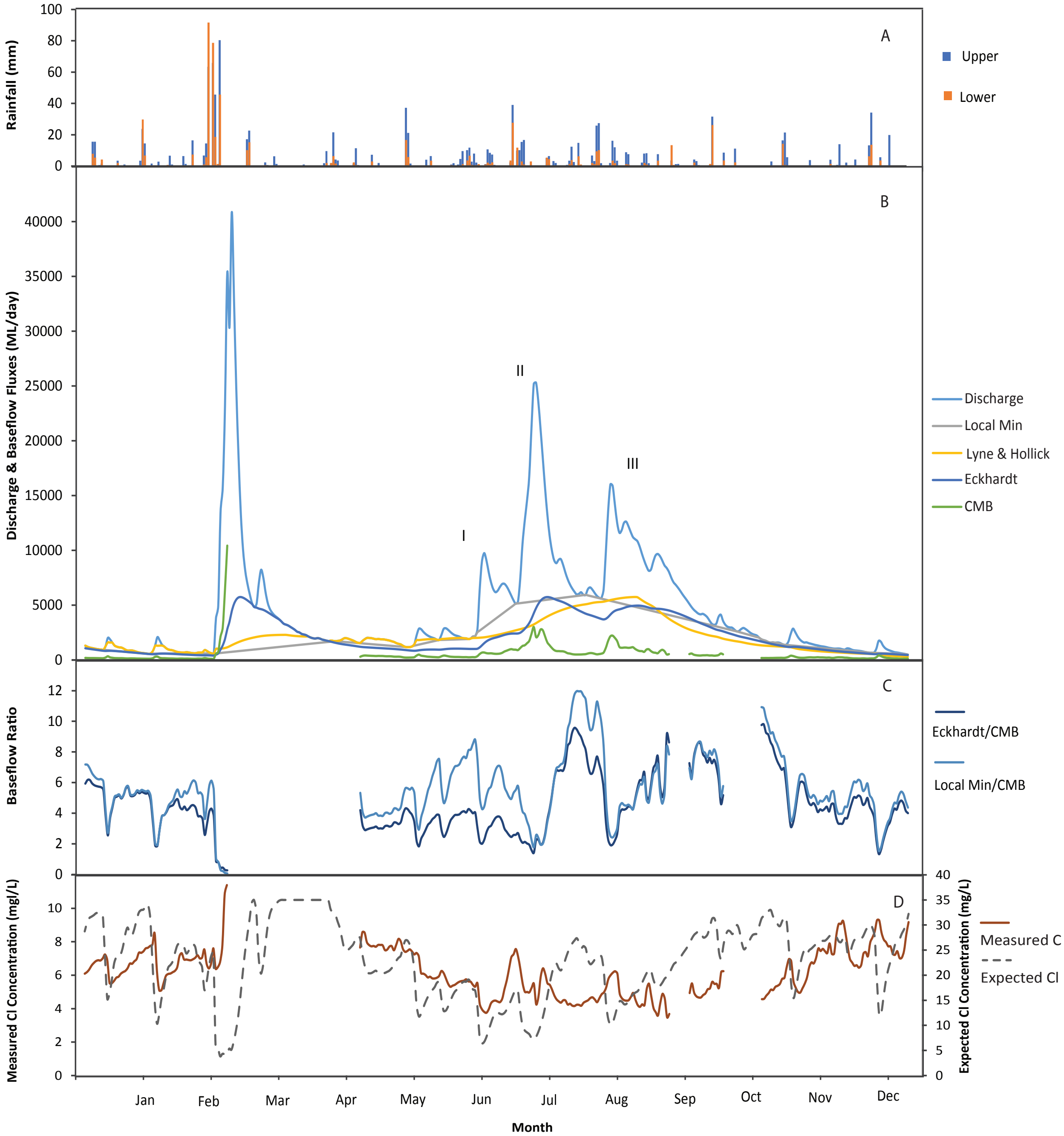

Figure 3.8 A) The amount of rainfall at Eurobin in upper catchment (Upper) and at Peechelba in the lower catchment (Lower) in 2012. Data from Bureau of Meteorology (2014). B) Variation in discharge of the Ovens River at Peechelba in 2012 with baseflow estimates calculated by local minimum (Local Min) (Sloto and Crouse, 1996), Lyne and Hollick (1979) and Ekchardt (2005) digital filters, and chemical mass balance (CMB). C) Ratios of baseflow fluxes derived by local minimum and the Ekchardt digital filter to those estimated by CMB. D) Variation in river Cl concentrations (Calculated based on EC and Expected) at Peechelba in 2012.

discharge and a higher yearly rainfall generally has a greater net volume of baseflow. High rainfall results in more recharge and in turn a greater increase in hydraulic gradients with a higher volume of groundwater inflow. A lower net baseflow occurs during the low-flow periods (such as the

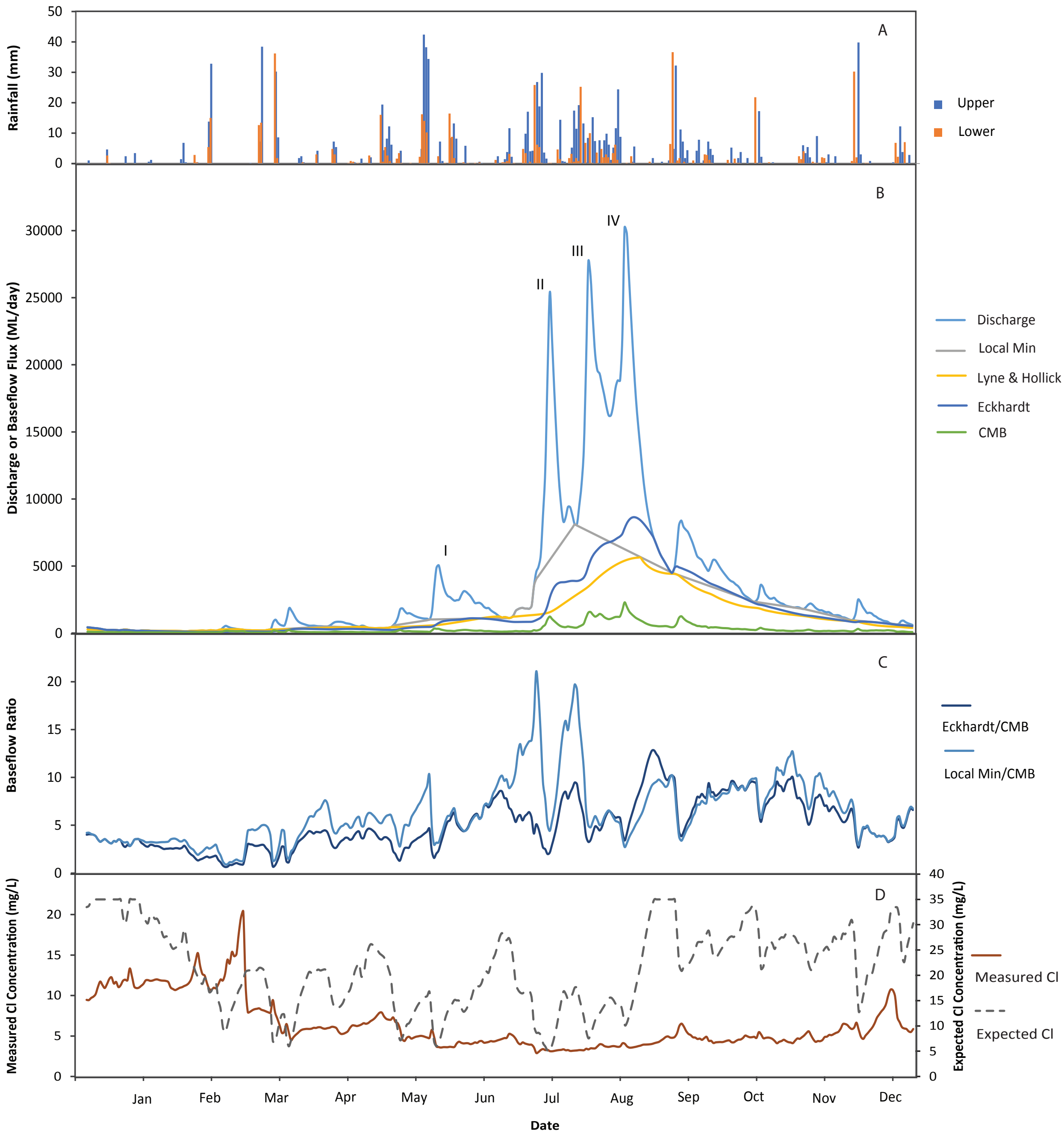

Figure 3.9 A) The amount of rainfall at Eurobin in upper catchment (Upper) and at Peechelba in the lower catchment (Lower) in 2013. Data from Bureau of Meteorology (2014). B) Variation in discharge of the Ovens River at Peechelba in 2013 with baseflow estimates calculated by local minimum (Local Min) (Sloto and Crouse, 1996), Lyne and Hollick (1979) and Ekchardt (2005) digital filters, and chemical mass balance (CMB). C) Ratios of baseflow fluxes derived by local minimum and the Ekchardt digital filter to those estimated by CMB. D) Variation in river Cl concentrations (Calculated based on EC and Expected) at Peechelba in 2013.

drought between 2006 and 2009), and in the Australian summer and autumn, but the percentage of baseflow in the river during these periods is often larger because majority of flow in the river is derived from baseflow when surface runoff is at minimum.

Year	Rainfall (Upper Catchment) (mm)	Rainfall (Lower Catchment) (mm)	Discharge (ML/year)	Baseflow Flux (ML/year and % of the annual flow, respectively)				
				FDC	Local Minimum	Lyne and Hollick	Ekchardt	CMB
2004	1,147	492	1,243,949		614,782 49	518,289 42	583,441 47	144,002* 11
2005	1,130	728	1,722,190		1,053,375 61	794,702 47	801,198 47	172,989* 10
2006	449	283	136,765		112,082 81	99,587 72	81,303 59	33,274* 24
2007	954	442	502,676		228,268 45	221,182 44	220,768 44	52,377 10
2008	908	489	474,234	180,614 33	272,704 57	235,381 49	229,026 48	39,986 8.4
2009	993	421	703,793		373,438 53	319,826 45	326,529 46	47,591* 6.8
2010	1,439	N/A	2,705,631		1,338,311 49	1,161,865 43	1,126,730 42	196,806* 7.3
2011	1,317	699	2,031,126		997,550 49	1,154,814 56	1,026,387 51	191,428* 9.4
2012	1,166	605	1,167,648	593,695 30	813,162 48	762,643 45	795,802 47	197,921 12
2013	960	492	134,387	181,865 15	704,036 52	501,208 37	623,131 46	107,281* 8
2014	1,005	598	991,120		599,672 60	386,673 39	467,761 47	100,751* 10
Total								
	Total Rainfall (Upper Catchment) (mm)	**Total Rainall (Lower Catchment) mm**	**Total Discharge**	**Total Baseflow Flux (ML/year, and % of the total flow, respectively)**				
				FDC	**Local Minimum**	**Lyne and Hollick**	**Ekchardt**	**CMB**
2004-2014	11,480	N/A	13,530,519	2,051,745 13	7,107,381 52	6,159,168 46	6,282,076 46	1,284,410 9.5

Table 3.1 Summaries of rainfall, discharge, baseflow fluxes for the Ovens River at Peechelba between 2004 and 2014. Rainfall data are collected at Eurobin in the upper catchment and Wangaratta in the lower catchment (Bureau of Meteorology, 2014). Discharge data from Water Measurement Information System (2014). Baseflow fluxes are calculated by FDC, local minimum (Sloto and Crouse, 1996), Lyne and Hollick (1979) filter, Ekchardt (2005) filter and CMB. * represents years with an incomplete EC records, and baseflow fluxes were derived from the average baseflow flux over the dates when EC data is available, providing the missing data is less than 20% in that year. N/A = Not Available.

Both the Lyne and Hollick, and Eckhardt filters were applied to the discharge data between 2004 and 2014. The Lyne and Hollick filter (Eq. 3.1) was applied in a multi-pass approach (forward, backward and forward with a month of reversed discharge data prior and after the study period) as suggested by Nathan and McMahon (1990) and Ladson *et al*. (2013). In contrast, the Echardt filter (Eq. 3.2) was applied in a single pass across the hydrograph as suggested by Eckhardt (2005). For both filters, α value was assigned as 0.97 using linear regression of the hydrograph data as described by Eckhardt (2008). In regard to the BFI_{max} term in the Eckhardt filter, a value of 0.2 *to* 0.25 is suggested for perennial streams on crystalline basement and 0.8 for perennial streams with porous aquifers (Eckhardt, 2005; Eckhardt, 2008). Given that the geology of the Ovens Catchment includes bedrock aquifers in the upper catchment and the sedimentary aquifers in the lower catchment, and that the area of the upper catchment is slightly bigger than that of lower catchment (Fig. 3.2a), an area-weighted BFI_{max} value of 0.47 (0.25 × 60% + 0.8 × 40%) was adopted for the Eckhardt filter (the use of one BFI_{max} value is perfered because the boundary between the two subcatchments defined by the geology is not well-defined, and thus the individual discharge for the two subcatchments are not easily estimated). The Lyne and Hollick filter estimated that baseflow contributed 6,159,168 ML (45% of the total flow) to the flow of the Ovens River for the period between 2004 and 2014. The estimated baselfow fluxes in 2008, 2012 and 2013 were 235,381 ML year^{-1} (50% of the total annual flow), 762,643 ML year^{-1} (46% of the total annual flow) and 501,208 ML year^{-1} (37% of the total annual flow), respectively (Table 3.1; Figs. 3.4b & 3.7b *to* 3.9b). The adjustable term in the Lyne and Hollick is α, and if α was increased or decreased by 5%, the calculated baseflow for the entire study period increases or decreases by approximately 1.4%. The estimated total baseflow fluxes based on the Eckhardt filter was 6,282,076 ML (46% of the total discharge) for the entire study period. The estimated baselfow fluxes in 2008, 2012 and 2013 were 229,026 ML year^{-1} (48% of the total annual discharge), 795,802 ML year^{-1} (47% the total annual discharge) and 623,131 ML year^{-1} (46% the total annual discharge), respectively (Table 3.1; Figs. 3.4b & 3.7b *to* 3.9b). If the BFI_{max} value was increased or decreased by 10%, the estimated baseflow for the entire study period would increase or decrease by approximately 9.5%. Thus, BFI_{max} produces a greater uncertainty in calculating baseflow fluxes than α.

3.5.3 Chemical mass balance

The derived Cl concentrations were used to calculate baseflow fluxes using the CMB (Eq. 3.3). CMB yields an estimate of groundwater component in baseflow and not total baseflow flux because the geochemistry of groundwater inflow, Cl concentration for example, differs to that of other components of baseflow which is similar to the geochemistry of river water. Cl in a river derives from rainfall via overland flow, halite dissolution and/or groundwater inflows, while in-stream evaporation may also increase Cl concentrations. There are no occurrences of halite, and the low Cl/Br ratios in the groundwater and surface water (< 1000) in the region prelude halite dissolution as a major source of Cl in the Ovens River (Cartwright *et al.*, 2006). The $\delta^{18}O$ and $\delta^{2}H$ values of the Ovens River, which were measured at various flow conditions, including at the end of the Millennium drought (2001 *to* 2009) (van Dijk et al., 2013) and the 2010 Victorian floods, define evaporation trends and generally increase downstream (Chapter 2). These observations indicate that in-stream evaporation occurs in the catchment. However, the shift in $\delta^{18}O$ is less than 2‰ during the sampling rounds, implying < 10% in-stream evaporation in the catchment (c.f. Gonfiantini, 1986). Thus, evaporation is likely to account for a small increase in Cl concentrations in the river. Overall, the variation of Cl concentrations in the river reflects the chemical balance between groundwater and surface water.

The Cl concentrations in groundwater in the upper and lower catchments vary from 1.8 to 9.3 mg L^{-1} and from 96 to 923 mg L^{-1}, respectively. As groundwater that feeds the river is often different to the regional groundwater (McCallum *et al.*, 2010; Cartwright *et al.*, 2011; Cook, 2013), and there are limited near-river groundwater bores in the lower catchment, it is difficult to derive an average Cl concentration of near-river groundwater for the Ovens Catchment. Adopting the common approach that has been used in many studies (Yu and Schwartz, 1999; Gonzales *et al.*, 2009), the highest Cl concentration in the Ovens River over the low-flow period during which the river is mostly likely to be fed by mainly groundwater was used to define the groundwater end-member. The highest groundwater concentration between 2004 and 2014 was 35 mg L^{-1} when the river discharge was close to zero (Fig. 3.6). The Cl concentration of surface runoff was initially assumed to be 1.5 mg L^{-1}, which is the Cl concentration in rainfall in the region (Blackburn and McLeod, 1983). Several years have incomplete records of EC data for calculating the annual total baseflow flux. For these years, the annual total baseflow flux was estimated by calculating the daily average baseflow flux over the number of days when EC data is available, and then multiplying this daily average baseflow flux by the number of days in that year. The estimated baseflow fluxes for 2008, 2012 and 2013 were 39,986 ML $year^{-1}$ (8.4 % of the total annual discharge), 197,921 ML $year^{-1}$ (11% of the total annual discharge) and 107,281 $year^{-1}$ (8% of the total annual discharge), respectively (Table 3.1, Figs. 3.4 & 3.7 *to* 3.9). The total baseflow for the 2004 *to* 2014 period was 1,284,409 ML (9.5 % of the total discharge). The variation in baseflow fluxes derived from the CMB is similar to those based on the hydrograph separation methods; the total volume of baseflow is higher during high flow periods than during low-flow periods (Figs. 3.4b & 3.7b *to* 3.9b). Increasing the groundwater Cl component in the CMB to 90 mg L^{-1} (which is the low end of measured groundwater Cl concentration in the lower catchment) would reduce the baseflow flux in 2008 to 15,136 ML $year^{-1}$ (or by 45%). However, this would imply that river at low discharge would always have a considerable component of surface water. Decreasing the groundwater Cl end-member by 10% would increase the baseflow flux in 2008 to 44,651 ML $year^{-1}$ (or by 12%). However, a lower average groundwater term would result in negative baseflow fluxes when the river has a high Cl concentration. Cl concentrations in groundwater, especially near-river groundwater, may vary on yearly basis depending on the amount of recharge in each year. The temporal variation in Cl concentrations for the near-river groundwater in the Ovens Catchment ranges from 5 to 35% (Chapter 2; Water Measurement Information System 2015). Taking the variability of the Cl groundwater end-member into consideration and using the maximum Cl concentration in the river in that year (21 mg L^{-1} in 2008 for example) as the Cl groundwater end-member, the groundwater influx in 2008 would increase to 68,694 ML $year^{-1}$ (or by 55%). Surface runoff may have higher salinity and thus a higher Cl concentration than rainfall. If the surface runoff term was increased to 2.5 mg L^{-1} (the minimum Cl concentration in the river), the calculated baseflow flux in 2008 is reduced to 26,625 ML $year^{-1}$ (or by 33%). Finally, the calculated baseflow fluxes would be reduced by 6 *to* 8 % if a maximum in-stream evaporation of 10% had occurred.

The CMB equation (Eq. 3.3) can be used in reverse to estimate the Cl concentration in a river if the proportion of baseflow to total discharge is known, and the groundwater is assumed to be the only component of baseflow. In this study, the proportions of baseflow to total discharge are from the Eckhardt filter, and the Cl concentrations in surface runoff and in groundwater are again assumed to be 1.5 mg L^{-1} and 35 mg L^{-1}, respectively. The overall variation in the predicted river Cl concentration shares some similarities with the observed pattern; the river has high Cl concentrations during low flow conditions but low Cl concentrations during high flow conditions (Figs. 3.7c *to* 3.9c). The expected river Cl concentrations generally decrease with increasing discharge. There were several occasions where the expected river Cl concentrations increaseto 35 mg L^{-1} (i.e. the assumed groundwater end-member Cl concentration) and remain for a period of time (shown as flatted peaks in figures. 3.7 *to* 3.9). This indicates the river composes of 100% of groundwater. Although the observed Cl concentrations remain low following the high flow period in July to August, the predicted Cl concentrations steadily increase after high flow events. The discharge vs Cl concentrations relationships based on the predicted Cl concentrations define hysteresis loops (Figs. 3.10 *to* 3.12). However, these loops are anticlockwise with lower concentrations during the rising limb of a hydrograph than during the falling limb. In addition, these loops are steeper than the observed ones. Increasing the Cl concentration of surface runoff term would reduce the values of the expected river Cl concentration but would not alter the shape

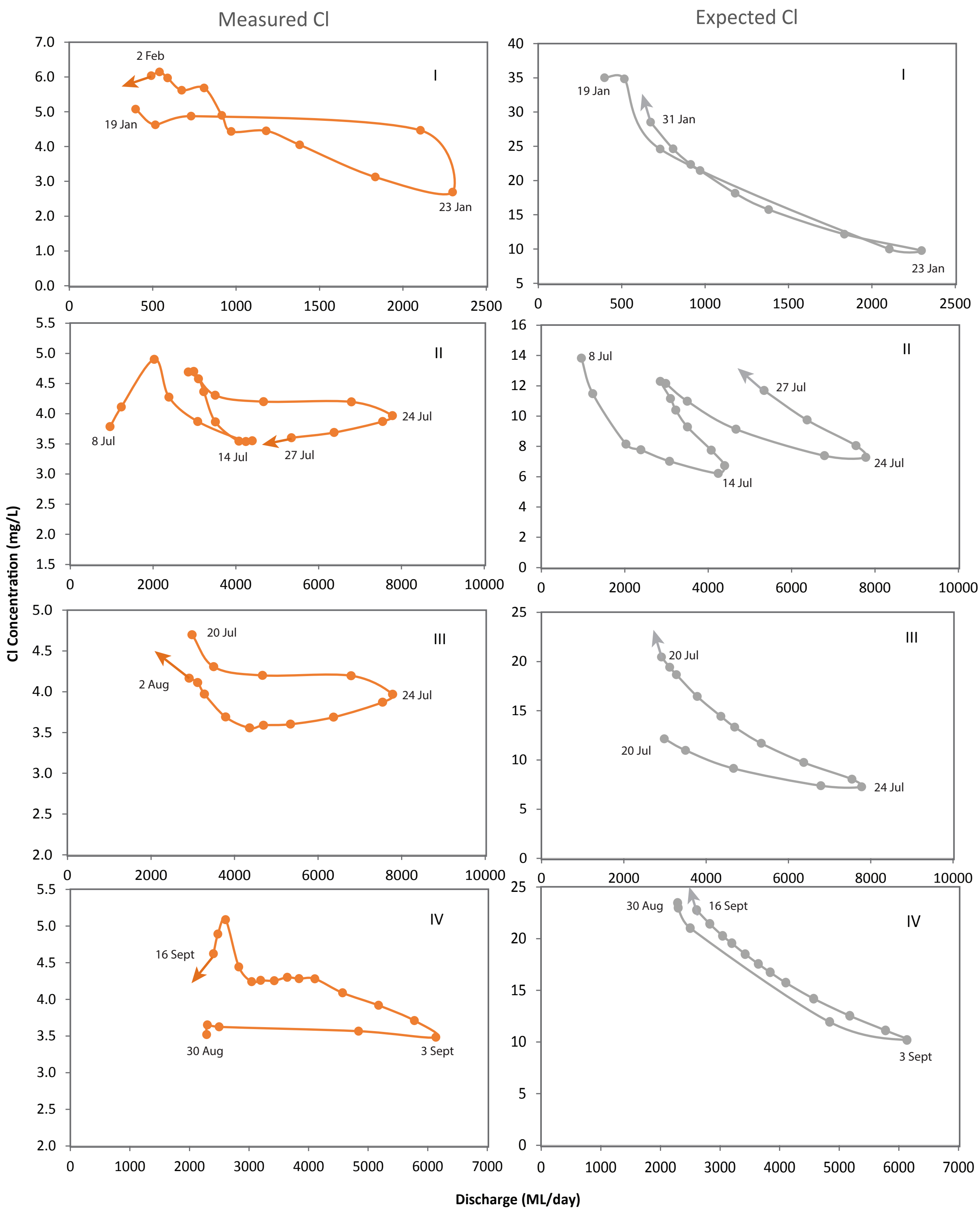

Figure 3.10 River Cl concentrations at Peechelba in respect to discharge during several high flow events in 2008. The Roman numerals refer to the particular flow events indicated in Fig. 3.7b; the arrow heads indicate the change in Cl concentrations and discharge with time. The predicted Cl concentration is based on the baseflow fluxes by the Ekchardt (2005) digital filter.

and the rotational direction of the discharge vs. Cl concentration hysteresis loops. In other words, it does not affect the interpretation on the porportion of various flow component that contributes

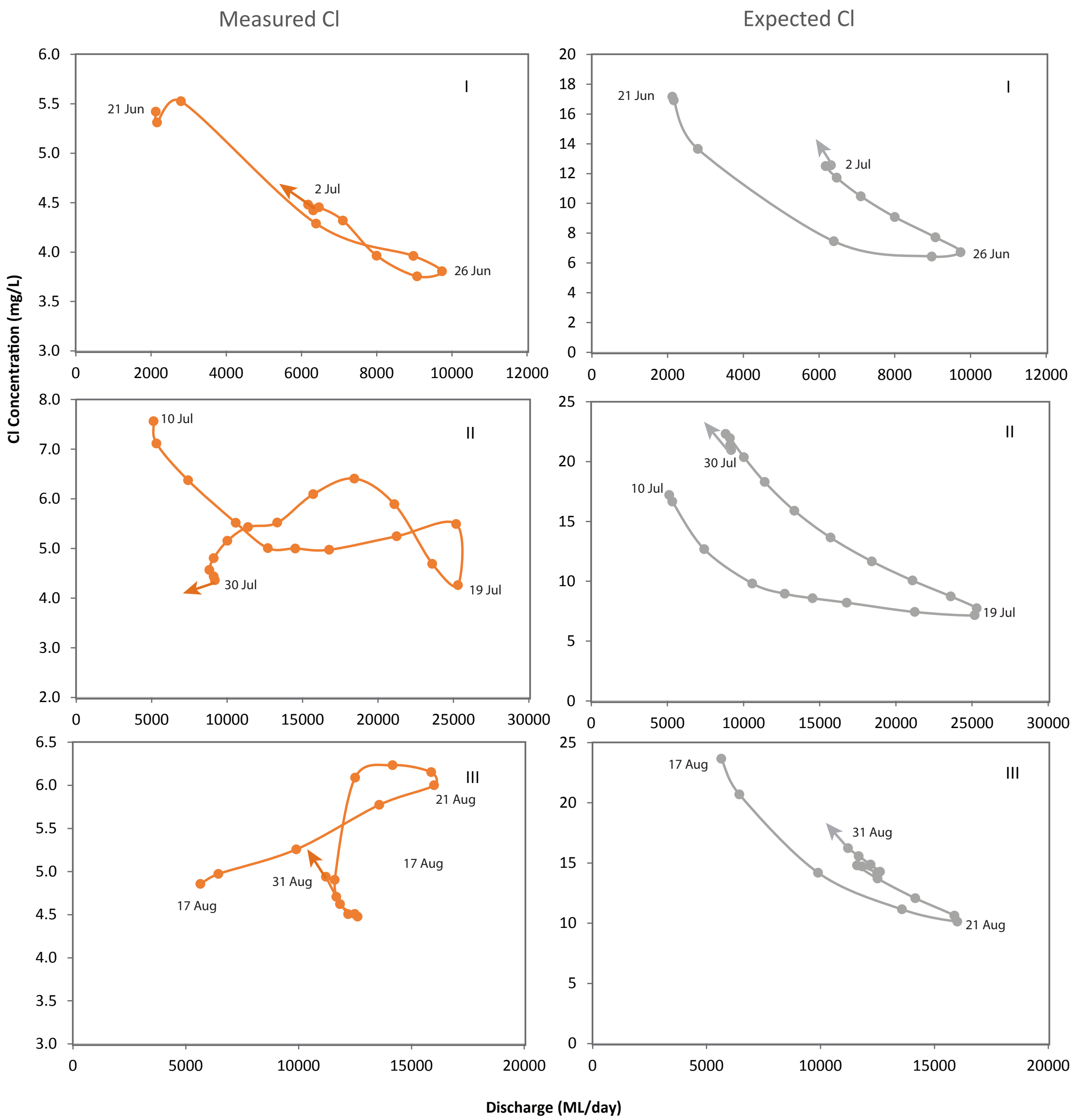

Figure 3.11 River Cl concentrations at Peechelba in respect to discharge during several high flow events in 2012. The Roman numerals refer to the particular flow events indicated in Fig. 3.8b; the arrow heads indicate the change in Cl concentrations and discharge with time. The predicted Cl concentration is based on the baseflow fluxes by the Ekchardt (2005) digital filter.

to the river or the sequence of flow contribution by various flow components during a high flow event.

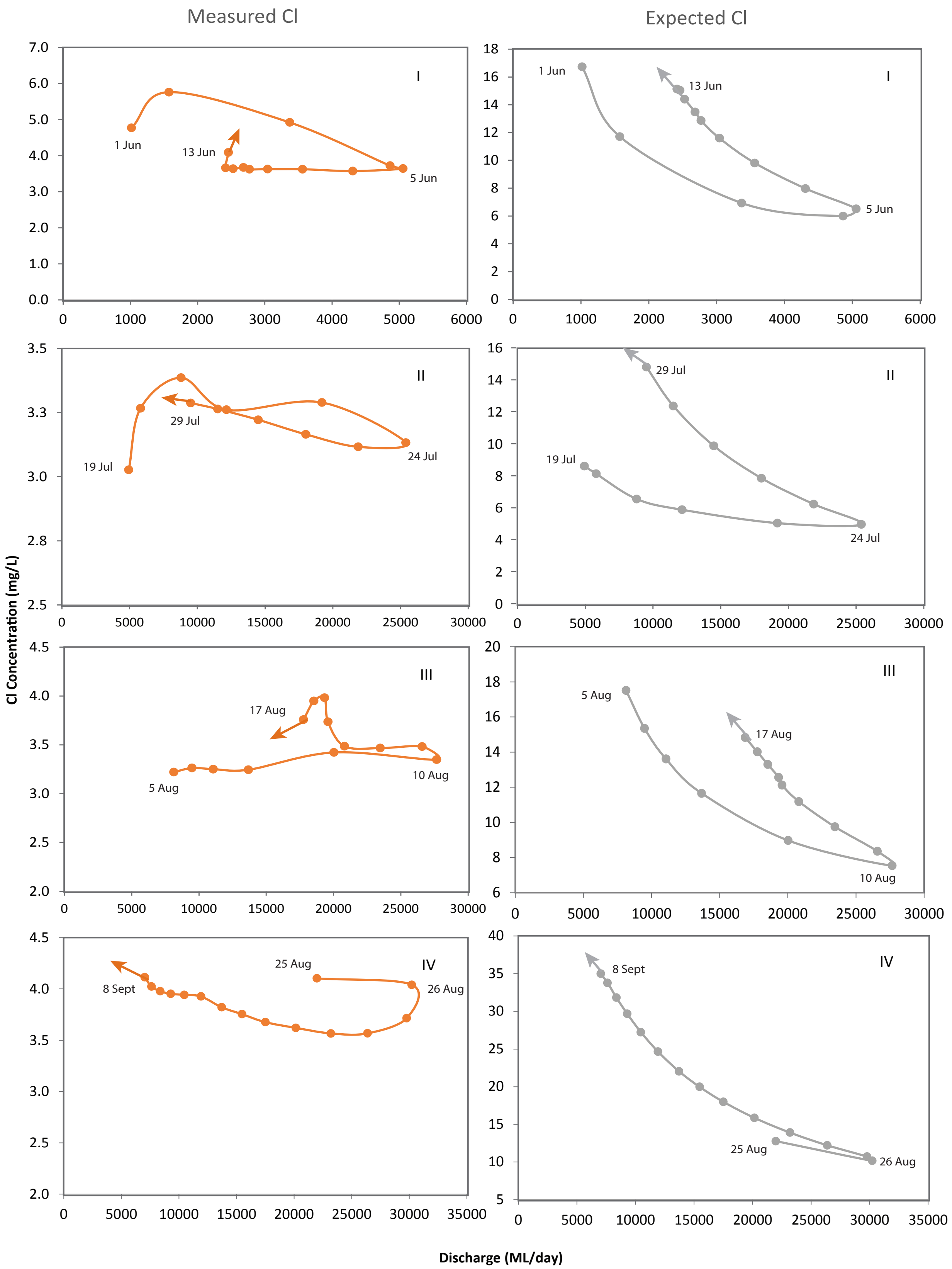

Figure 3.12 River Cl concentrations at Peechelba in respect to discharge during several high flow events in 2013. The Roman numerals refer to the particular flow events indicated in Fig. 3.9b; the arrow heads indicate the change in Cl concentrations and discharge with time. The predicted Cl concentration is based on the baseflow fluxes by the Ekchardt (2005) digital filter.

3.6 Discussion

3.6.1 Comparison of baseflow fluxes

The FDC indicates that baseflow constitutes a minor source for the Ovens River with a Q90/Q50 index of 0.15 for the 2004 *to* 2014 period. The Q90/Q50 index increased to 0.33 during the dry year in 2008. The relative baseflow fluxes based on Q90/Q50 are similar to those estimated from the river Cl concentrations which is 9.5% for the entire 10-year period, respectively.

All three algorithms in hydrograph separation methods produces similar baseflow fluxes for the Ovens Catchment with a higher baseflow fluxe by the local minimum method (Table 3.1). The total volume of baseflow for the entire period (2004 *to* 2014) by the local minimum, the Lyne and Hollick filter, and the Eckhardt filter were 7,107,381 ML (51% of total discharge), 6,159,168 ML (46% of total discharge) and 6,282,076 ML (47% of total discharge), respectively. The estimates from the hydrograph separation methods are, however, significantly higher than those based on the FDCs (especially for the entire study period and the wet years) and the CMB (Table 3.1). For example, the percentage differences in the 2008 (dry year) and the 2012 (moderately high flow year) baseflow fluxes between the FDC and the Eckhardt filter are 35% and 140%, respectively. Likewise, the percentage differences in the 2008, 2012 and 2013 baseflow fluxes between the two estimates from the CMB and the Eckhardt filter are approximately between 120% and 140%. Additionally, during the period of high discharge (from June to September) and the times at which a high flow event occur after a dry period (such February or May), the difference in the baseflow estimates is larger with a hydrograph separations:CMB ratio of up to 52:1 (Figs. 3.7c *to* 3.9c). In contrast, both the hydrograph separation methods and the CMB produce similar baseflow fluxes during low flow periods in December to March. The results from uncertainty analysis suggest that the uncertainty in assigning some of the surface water or groundwater variables cannot explain the large differences. The discrepancy is likely contributed by the assumptions that inherent to each technique (i.e. whether the technique defines baseflow using the chemical composition of baseflow or the rate at which baseflow is discharged) rather than the technique itself.

3.6.1.1 Reasons for difference in baseflow estimates between the techniques

There is a large difference in the baseflow estimates between the FDC and hydrograph separation methods, especially for the wet years. The difference is probably caused by aggregating discharge from transient storage (such as river banks and the unsaturated zone) with groundwater inflow in the hydrograph separation methods. The discharge from transient water stores occurs for period of weeks or months after the surface runoff has ceased, and thus it is difficult to separate them from groundwater inflow, which also occurs in the period of weeks to months, in the hydrograph separation methods (Nathan and McMahan, 1990; Evans and Neal, 2005; Brodie *et al.*, 2007). In contrast, the baseflow fluxes from FDC are derived from the 90th percentile streamflow (Q90) which is the value of daily streamflow that is exceeded 90% of the time. At Q90, those transient water stores, which are replenished at high flow conditions (mostly at the 10th percentile streamflow or below, < Q10), are usually depleted. Consequently, the baseflow at this flow condition (Q90) almost exclusively consists of groundwater. The baseflow fluxes from FDC therefore may reflect the amount of groundwater inflow and not the total volume of baseflow (i.e. combination of flows from transient storage and groundwater).

Large discrepancies in the baseflow estimates also exits between the CMB and hydrograph separation methods. As discussed above, the hydrograph separations cannot differentiate discharge from transient water stores from groundwater inflow because they both represent delayed water inputs. However, water from the transient water stores is derived from surface runoff or river water and has a low salinity that is similar to surface runoff or river water. Thus, the CMB groups discharge from these transient stores with the surface water (or quickflow) component, making the baseflow fluxes derived from the CMB to only consist of groundwater inflow. The salinity (or Cl concentration) of the transient water stores in the Ovens Catchment could be increased by a few processes, such as mineral dissolution and evapotranspiration. As discussed earlier, the lack of halite in the Ovens Catchment precludes the dissolution of halite as a source of Cl in water from the transient water stores. The dissolution of silicate minerals, particularly plagioclase which is common in the aquifers in the Ovens Catchment, could increase the salinity of transient water stores. However, it requires approximately 100 years to increase the Na^+/Cl^- ratio from 1 to between 2 and 11 (Cartwright and Morgenstern, 2012). The period over which the water is stored in the transient water store (usually from days to months) is thus too short for dissolution of silicate minerals to increase the salinity significantly (Squillace, 1996; McCallum *et al.*, 2010; Cartwright *et al.*, 2014). Finally, the salinization effect of evapotranspiration on these water stores is likely to be minimum because water in some transient water stores, particularly river banks, are much deeper than water in the soil zone. The observation that the discrepancy between CMB and the hydrograph separation methods is greater during the high flow periods further suggests a significant contribution of transient storage in the catchment. This relates to the fact that transient storage is recharged at the onset of high flow conditions and then discharges the stored water into the river during and after high flow periods. The discrepancy in baseflow estimates for the Ovens Catchment exists not only between hydrograph separation and Cl-based CMB but also between hydrograph separation and ^{222}Rn-based CMB (Chapter 2). The ^{222}Rn concentrations in the Ovens River indicate that the baseflow flux for the Ovens catchment ranges from 2 to 17% of the total flow (Chapter 2) which is similar to the baseflow estimates obtained from the FDCs and Cl-based CMB. These differences support the idea that groundwater is only a small component of baseflow with the majority of baseflow being derived from transient water stores in the Ovens Catchment.

3.6.2 Contribution of transient water stores

That the variation of Cl concentrations in the Ovens River in respect to discharge show a clockwise hysteresis loop (Figs. 3.10 *to* 3.12) indicates a contribution from transient water stores to the river flow in the Ovens Catchment. The clockwise hysteresis loops indicate the sequence of flow contribution is overland flow and precipitation, water from transient water stores, and then groundwater. The low salinity water from overland flow during the rising phase of the flow event decreases the Cl concentrations in the Oven River. However, there were occasions, especially in autumns following a dry period, where the Cl concentrations in the river increased during the rising limb of a flow event. The reason for such increase in the river Cl concentration is likely to be the flushing of saline water from the unsaturated zone and/or from the stagnant pools on floodplains into the river channel (Squillace, 1996; Evans and Davies, 1998; McCallum *et al.*, 2010; Zabaleta and Antigüedad, 2012; Aubert *et al.*, 2013; Cartwright *et al.*, 2014). After the flood peaks, water from the low salinity transient stores is discharged to the river, leading to the salinity in the Ovens River remaining low for a period of time. The low salinity in the Ovens River following the flood peak is shown by the low river Cl concentrations after the highest river discharge in the clockwise discharge vs. Cl concentration hysteresis loops. The discharge vs. Cl concentration hysteresis loop of some multiple and consecutive flow events is, however, sometimes anticlockwise (Figs. 3.10c 3.11c & 3.12c). As the catchment may not have returned to the pre-event flow equilibrium prior to the next high flow event during a consecutive flow event, the normal sequence of discharge from various water stores or the amount of discharge from various water stores may be altered during the second or third flow events. This altered sequence of discharge from various water stores affects the rotational direction of some hysteresis loops in these consecutive flow events (Evans and Neal, 2005). The difference between the observed and expected variations in the river Cl concentrations further indicates the contribution of transient storage to the river (Figs. 3.10 *to*

3.12). The expected discharge vs. Cl hysteresis loops are much steeper than those observed, and that suggests that when groundwater was the only component of baseflow, the rate at which the Cl concentrations in the river return to the pre-event flow would be faster. Since the Cl concentrations in the Ovens River do not increase rapidly following the high flow events, water from low salinity transient stores likely contributes to the river for maintaining a low level of Cl concentrations when the overland flow is diminishing after a rain event.

There are several transient water stores possibly contributing to the flow of the Ovens River. One is bank storage. Bank storage is the water that is recharged into the bank at high river-stage and then returned to the river as a part of baseflow during the receding phase of flow events. Bank return flow can sustain over a period of weeks to months (Squillace, 1996; McCallum *et al.*, 2010). Thus, bank return flow may partly explain why the Cl concentrations in the Ovens River remain low for a period of time after the pass of high flow events. Disconnected anastomosing channels and the unsaturated zone are also likely to provide a source of water between flow peaks. There are numerous disconnected anastomosing channels in the lower Ovens Catchment. These channels are likely be filled with water directly from rainfall or overbank recharge and later discharge the stored water back to the river directly as outflow over several days to weeks or through the unsaturated zone over a periods of months. The contribution of groundwater inflow in the Ovens River is less than 20%, but the river does not become dry during periodical or extended dry periods. River banks and reconnected channels are mostly drained during these dry periods. This observation suggests that the unsaturated zone in the catchment may provide a significant amount of water to the river during dry periods.

3.6.3 Relationship between the difference in baseflow estimates and discharge

If the difference in baseflow estimates between the hydrograph separation methods,

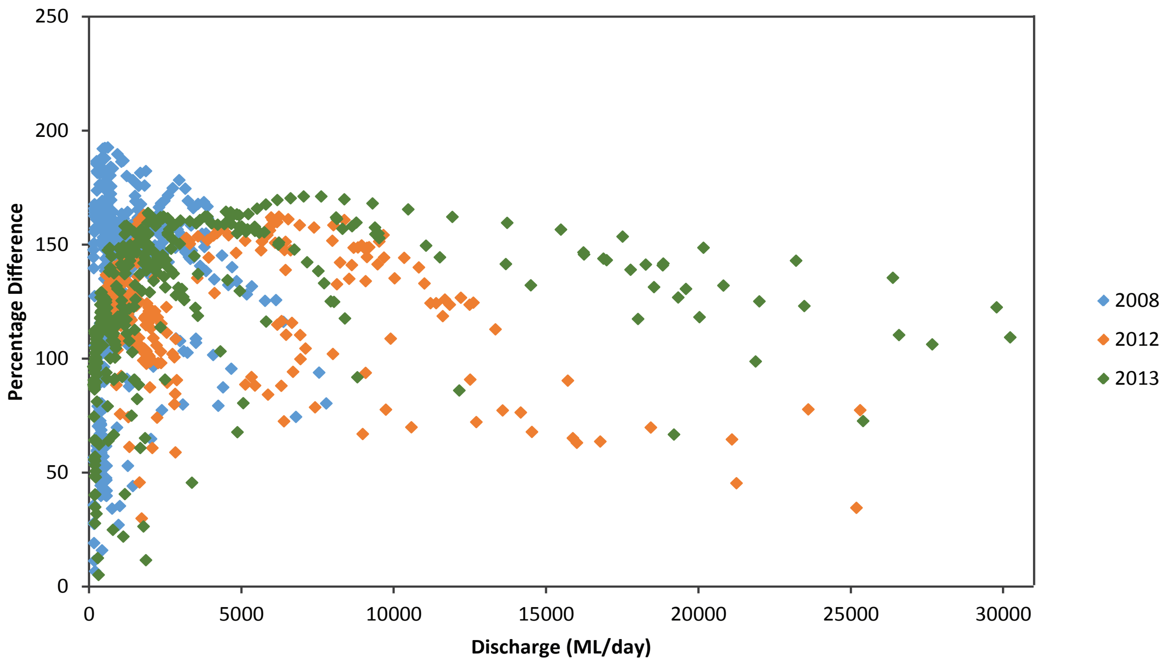

Figure 3.13 Percentage difference in the baseflow estimates between the Ekchardt (2005) digital filter and CMB for the years of 2008, 2012 and 2013. No clear relationship can be defined between the two variables.

and CMB generally relates to discharge, such a relationship may be further explored to predict the volume of transient storage in a catchment for a given discharge. However, there are no clear relationships between the daily discharge and the percentage difference in the baseflow estimates between the two techniques (Fig. 3.13). The lack of relationship suggests that the volume of transient storage generated by a surface runoff event depends on more than just the volume of river flow. Other factors that should be considered are the volume of transient storage prior to a high surface runoff event, the timing between the high surface runoff events and the frequency of high surface runoff events, and whether or not these transient water stores interact with other water stores in the catchment, for example, the interaction between bank storage and regional groundwater (Cartwright *et al.*, 2014). Moreover, Cartwright *et al.* (2014) proposed that the difference in baseflow estimates between the hydrograph separation methods and CMB should be considered as the minimum estimates of potential transient storage in the catchment since the flushing of saline water during the onset of high low events are not considered in the CMB calculation.

3.7 Conclusions

The hydrograph separation methods indicated that between 37% and 70% of total discharge in the Ovens Catchment is derived from baseflow, whereas FDC and CMB indicated that the baseflow only constitutes only 13% to 33% and 8% to 20% of the total river flow, respectively. These differences in baseflow estimation are larger during high flow periods. Hydrograph separation reports high baseflow fluxes probably because it aggregates several delayed flows together, including groundwater inflow, interflow, bank return flow and drainage from pools on the floodplain, and yields the total baseflow flux in rivers. In contrast, CMB separates groundwater inflow from other delayed flow because the geochemistry of groundwater inflow differs to that of other delayed flows. Thus, it yields the groundwater inflow in rivers. FDC yields the groundwater inflow probably because it estimates baseflow flux by using the 90th percentile streamflow during which the baseflow is almost exclusively groundwater with minimum influxes from the depleted transient water stores. The relatively low groundwater inflow in the Ovens River was also reported elsewhere. Groundwater inflow in the mid-catchment (at Myrtleford) was estimated to be 6% of the median annual streamflow based on the hydraulic heads and river heights (CSIRO, 2008). Other low-lying catchments in southeast Australia, such as Barwon Catchment, also report a large difference in baseflow estimates between hydrograph separation and CMB with percentage difference of between 4% and 28% (Cartwright *et al.*, 2014). This suggests that discharge from transient water stores occurs in many catchments regardless of the catchment physiography, but the relative contribution of various transient water stores in rivers is likely to be different for each catchment.

This study shows that continuous datasets of discharge and EC are useful in estimating baseflow flux and identifying various sources of water contributing to a river. The advantage of using discharge and EC data is their ease of measuring at different flow conditions with automatic equipment. Consequently, discharge and EC data can provide temporal information on the sources of water in a river at different conditions over an extend period of time. This study also demonstrates that hydrograph separation methods can overestimate groundwater inflow if discharge from transient storage makes up a significant portion of baseflow. Thus, hydrograph separation methods should not be used alone for investigating the interaction between rivers and sub-surface water resources in a catchment. It is also important to differentiate groundwater inflow from discharge from transient water stores in baseflow estimation because each of them represents different water stores and provides information on different hydrological processes in a catchment. Studying the groundwater inflow component in baseflow provides information on the impact of groundwater extraction on nearby rivers. Knowledge of transient water stores, particularly river banks, is vital in understanding flooding in catchments. In conclusion, in addition to groundwater, water from transient water storage is a significant source of water for the Ovens Catchment and should be taken into consideration when managing the riverine environment in the catchment. For example, transient water storage is normally depleted during low flow periods and will not able to contribute the flow of the Ovens River during these periods. Therefore, a significant reduction in water abstraction in the Ovens Catchment will be required in order to maintain sufficient flow in the Ovens River during low flow periods.

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Chapter 4

Defining the nature of river-river bank interaction in a river valley using hydraulic heads, geochemistry and tritium: Ovens River, southeast Australia

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Abstract

Hydraulic heads, major ion geochemistry and environmental isotopes (C, O and H) were used to define the river-river bank interactions in the Ovens River in southeast Australia. In the uppermost section of the upper catchment, the

decrease in hydraulic heads away from the river, the increase in salinity away from the river, and the similarity of the Na/Cl ratios and ^3H concentrations between the near-river groundwater and the river water imply constant recharge by the river through the river bank. At locations where the river flows close to the steep slopes at the edge of the valley, bank recharge only occurs during high flow conditions when the high river stage reverses the upward head gradient associated with the break of slope. The periodic bank recharge at these locations is evidenced by the reversed hydraulic gradients, the reduction in EC in the river banks during high flow conditions, and the high ^3H concentrations of the near-river groundwater. In the middle to lower sections of the upper catchment (the steepest part of the valley), the lack of variation of the near-river groundwater EC values, the similar Na/Cl ratios between the near-river groundwater and the regional groundwater, and the low ^3H concentrations imply that there is little bank recharge probably due to the persistent high hydraulic gradients toward the river in the valley. In the middle and lower catchments, sporadic bank recharge is suggested by the reversal of the head gradients, the reduction in the EC values of the near-river groundwater, the shift in the Na/Cl ratios, major ion composition and stable isotopic signature of the near-river groundwater toward those of the river water, and the high ^3H concentrations following high flow conditions. Overall, bank recharge occurs in the uppermost section of upper catchment, and in the middle and lower catchments because these areas have a relatively lower regional hydraulic gradient toward the river and a high to moderate hydraulic conductivity in river bank sediments, prompting bank infiltration following high conditions. The combination of geochemical parameters and head data may be used to provide a comprehensive understanding of the locations of bank storage that helps to understand the hydrogeology of the catchment.

KEYWORDS: River-groundwater interaction, Bank storage, Hydraulic heads, Geochemistry, Tritium, Ovens River

4.1 Introduction

Understanding the relationship between rivers and groundwater at a local scale within the river bank is important in advancing our overall understanding of surface water-groundwater interaction, and further enhances our ability to protect water resources and riparian ecology. In losing rivers, the river banks are continuously recharged by the river. The recharged water in the banks then flows into the adjacent alluvial aquifer. By contrast, river banks in gaining rivers are only recharged during high flow events when the river elevation is above the regional water table. When the river level subsides, the water stored in the bank returns to the river as part of baseflow (i.e. water that has been stored in the catchment is discharged to a river between rainfall events). This process is referred to as bank storage (Singh, 1968; Winter *et al.*, 1998). Bank storage has implications for constraining baseflow, managing water resources and protecting riparian ecology. If a river bank is frequently recharged by the river, the chemistry of the groundwater in the bank is likely to reflect a mixture of river water and regional groundwater. The near-river groundwater chemistry is likely to vary over time, reflecting the variable proportions of water derived from the river and regional groundwater. Such chemical variability can make it difficult to characterise the chemistry of the near-river groundwater which subsequently makes defining the groundwater end-member in a chemical mass balance model for estimating baseflow more difficult (McCallum *et al.*, 2010; Cook, 2013; Unland *et al.*, 2013). Specifically, estimates of groundwater inflows to the river based on regional groundwater chemistry will underestimate the total baseflow. In terms of management, if a river is mainly fed by the return flow from river banks, protection of the river banks from contamination is essential for maintaining the water quality of the river (Squillace *et al.*, 1993; Lynch *et al.*, 2014). Furthermore, baseflow derived from bank storage is transient and may not be able to sustain the stream flow over a long baseflow period (Kondolf *et al.*, 1987; Squillace, 1996; Whiting and Pomeranets, 1997). A different type of flow management is thus required for this type of river to ensure a sufficient environmental flow during low flow periods. Bank storage can mitigate the magnitude of flood waves, reducing the severity of flooding downstream (Winter *et al.*, 1998). Bank storage may also alter the amount of water and nutrients available to riparian vegetation, which is vital to the overall riverine ecosystem (Bourg and Bertin, 1993; Burt *et al.*, 2002; Lamontagne *et al.*, 2005). The fluctuation in the local water table associated with bank storage, for example, may affect the growth of groundwater dependant plants in the riparian zone. Bank infiltration may create an anoxic zone within the banks through the bacterial degradation of organic matter (Hiscock & Grischek, 2002; Gunkel & Hoffmann, 2009). The anoxic environment in turn can promote denitrification and/or phosphorus mobilisation.

The physical process of bank storage is controlled by a number of factors, including the frequency and duration of floods, the hydraulic conductivity of the sediments in the river bank, the shape of river channel, the presence of clogging layers on the river bank or bed, the slope of river bank and the gradient of regional water table. A flood wave of short duration can lead to a faster rate of water exchange between the river and the river bank but a lower volume of bank storage (Chen and Chen, 2003; Ha et al., 2008). River banks with a high hydraulic conductivity have a greater potential volume of bank storage, but the bank return flow from these permeable banks is also faster (Whiting and Pomeranets, 1997; Chen and Chen, 2003). In alluvial plains, vertical hydraulic conductivity is often less than horizontal hydraulic conductivity, and where the anisotropy is large the vertical expansion of the bank storage zone will be limited (Whiting and Pomeranets, 1997; Chen and Chen, 2003). However, the addition of water into bank storage through the bottom of a river bed can be significant when the river is wide and shallow (Squillace, 1996). Deep narrow rivers with a wide floodplain have a greater capacity for bank storage (Whiting and Pomeranets, 1997) than shallow broad rivers or rivers with narrow floodplains. The presence of a clogging layer delays the bank infiltration and bank flow return, and reduces the volume of bank storage (Ha *et al.*, 2008). Shallow bank slopes may increase both the bank infiltration rate and the volume of bank storage, but delays bank return flow (Doble et al., 2012). A gradient toward the river reduces the lateral distance of river water penetration and the volume of bank storage as well as increasing the rate of bank return flow.

4.1.1 Assessing bank storage

Comparing hydraulic heads in the river bank with river levels can be used to define the direction of flow between the river bank and the adjacent river, and to calculate the exchange potential between the river and groundwater in the river bank (Squillace, 1996; Brodie *et al.*, 2007; Lewandowski *et al.*, 2009; Banzhaf *et al.*, 2011). However, hydraulic heads alone do not always reveal the extent of water movement into the river bank because an increase in hydraulic head during a high flow event may represent the displacement of existing near-river groundwater which causes the water table to rise at some distance away (Welch *et al.*, 2013; 2014). A more reliable indication of the actual water movement into the bank is given by changes to groundwater chemistry (Bourg and Bertin, 1993; Duval and Hill, 2006; Allen *et al.*, 2010; Banzhaf *et al.*, 2011; Majumder *et al.*, 2013). As groundwater generally has higher solute concentrations than surface water, infiltration from the river may result in a decline in major ion concentrations and electrical conductivity (EC) in the river bank. The most direct way in tracing water mixing is to use ^{18}O, ^{2}H and ^{3}H as these stable or radioactive isotopes are parts of the water molecule. Providing that there is a difference in δ^{18}O and δ^{2}H values between regional groundwater and river water, the δ^{18}O and δ^{2}H signatures of the groundwater in a river bank may be used to constrain mixing between regional groundwater and infiltrating river water (McKenna *et al.*, 1992; Négrel *et al.*, 2003; Lambs, 2004; Lamontagne *et al*., 2005; Baskaran *et al*., 2009).

4.1.2 Quantifying residence time in river banks

^{3}H, with a half-life of 12.32 years, can be used to track groundwater recharged over the last 100 years (Cook and Böhlke, 2000; Kazemi *et al.,* 2006; Cartwright and Morgenstern, 2012). Mixing of young infiltrating water from a river with older regional groundwater will increase the ^{3}H concentration of groundwater in the river bank. As the atmospheric ^{3}H concentrations following the nuclear tests (the so called ‘‘bomb pulse’’) were several orders of magnitude lower in the southern hemisphere than in northern hemisphere, the ^{3}H activities of the bomb pulse water have now decayed below the modern precipitation (Clark and Fritz, 1997; Morgenstern *et al.*, 2010; Morgenstern and Daughney, 2012). The low atmospheric ^{3}H concentration allows a unique residence time to be calculated for water from a single ^{3}H measurement in southern hemisphere waters. Even where residence times are not reliably calculated, older water in the southern hemisphere has lower ^{3}H concentrations than younger water, allowing relative residence times to be established. Understanding the residence time of near-river groundwater may further assist in constraining river-river bank interactions.

The calculation of residence time of groundwater in river banks or aquifers using ^{3}H concentrations is based on the lumped parameter models (LMPs). Groundwater in an aquifer flows in different pathways of various lengths, and that results in water samples collected in bores having a range of residence times rather than a discrete age. The LMPs yield a mean residence time by taking the aquifer geometry and flow configurations in the aquifer into account, and treating the groundwater sample as comprising many individual parcels of water that have followed a different flow path and hence taken a different amount of time to reach the sampling point (Maloszewski and Zuber, 1982; 1992; Cook and Bohlke, 2000; Zuber *et al.*, 2005). For a steady-state groundwater system, the ^{3}H measured concentration at time t, $C_{meas}(t)$ is given by

$$C_{meas}(t) = \int_{-\infty}^{t} C_{prec}(t')\, e^{-\lambda(t-t')} g(t-t')dt' \quad , \qquad \textit{Eq. 4.1}$$

where C_{prec} *(t')* is ^{3}H concentration of precipitation at time *t'*; *t'* is the time of recharge; λ is the decay constant (0.0567 year^{-1} for ^{3}H); $t - t'$ is the residence time of the water sample; and $g(t - t')$ is residence time distribution function, which relates to the distribution of groundwater flow paths in the aquifer (Małoszewski and Zuber, 1982; Jurgens *et al.*, 2012). The exponential-piston flow model (EPM) describes the mean residence time of groundwater in unconfined to semi-confined aquifers. In these aquifers, the groundwater flow consists of two segments of flow: a segment of exponential flow and a segment of piston flow (Małoszewski and Zuber, 1982; Jurgens *et al.*, 2012). Exponential flow assumes that uniform recharge occurs in an unconfined aquifer of constant thickness, and groundwater travels downwards away from the recharge zone with a vertical stratification of groundwater residence times. In piston flow, groundwater flows linearly with no mixing leading to a situation where all water travelling at any particular time at a particular point in the aquifer has the same residence time. The residence time distribution function of the EPM is as follows:

$$EPM_{(t-t')} = \frac{n}{\tau_s} e^{\left[\frac{n(t-t')}{\tau_s}\right]}, \ \text{for}\ t \ \geq \ \tau_s \left(1 - \frac{1}{n}\right); 0 \quad , \qquad \textit{Eq. 4.2}$$

where n is EPM ratio+1 and τ_s is the mean age of water in the system (Małoszewski and Zuber, 1982; Jurgens *et al.*, 2012). The EPM ratio is the ratio of piston flow to exponential flow models, and varies from 0 for exponential flow to 1 for piston flow.

4.1.3 Aims

The results from chapters 2 and 3 indicate that water from transient water stores contributes a significant amount of baseflow in the Ovens Catchment. One of these transient water stores can be river banks in the catchment. This study uses major ions, stable (^{18}O & ^{2}H) and radiogenic (^{3}H & ^{14}C) isotopes in conjunction with hydraulic heads to characterise the river-river bank interactions at several locations in the Ovens Catchment, southeast Australia. In particular, the focus will be on determining whether the river bank is recharged by the river, the timing of bank infiltration and the lateral extent of bank infiltration. The paper also examines the geomorphological controls of bank recharge. As the Ovens Catchment is typical of valley-to-floodplain catchments, knowledge of river-river bank interaction from this study may be transferred to all river valleys in southeast Australia and globally.

4.2 Study area

4.2.1 Hydrological setting

The Ovens Catchment is located in the southeast of Murray Basin, extending northward from the Victorian Alps to the Riverine Plain, a large low-lying alluvial floodplain adjacent to the Murray River (Fig. 4.1) (Lawrence, 1988). The catchment can be subdivided into upper, middle and lower catchments which are broadly characterised by narrow and steep-sided valleys, broader valleys with a strip of floodplains and an extensive region of very low-gradient floodplains, respectively. The Ovens Catchment comprises Palaeozoic basement rocks overlain by Tertiary to recent fluviatile sediments that are up to 210 m thick. Several basement highs and outcrops exist in the catchment. The basement consists of metamorphosed Ordovician turbidites together with Silurian and Devonian granite intrusions which form a fractured-rock aquifer with a hydraulic conductivity of 0.01 *to* 1 m day^{-1} (Slater and Shugg, 1987; Heislers, 1993; van den Berg and Morand, 1997). The deepest sediments belong to the terrestrial Tertiary Calivil Formation; these have a maximum thickness of ~45 m and thin out toward the upper catchment. The Calivil Formation contains consolidated gravel, sand, silt and cobbles with a hydraulic conductivity of 5 *to* 50 m day^{-1} (Shugg, 1987; Cheng and Reid, 2006). The overlying sediments are the fluvio-

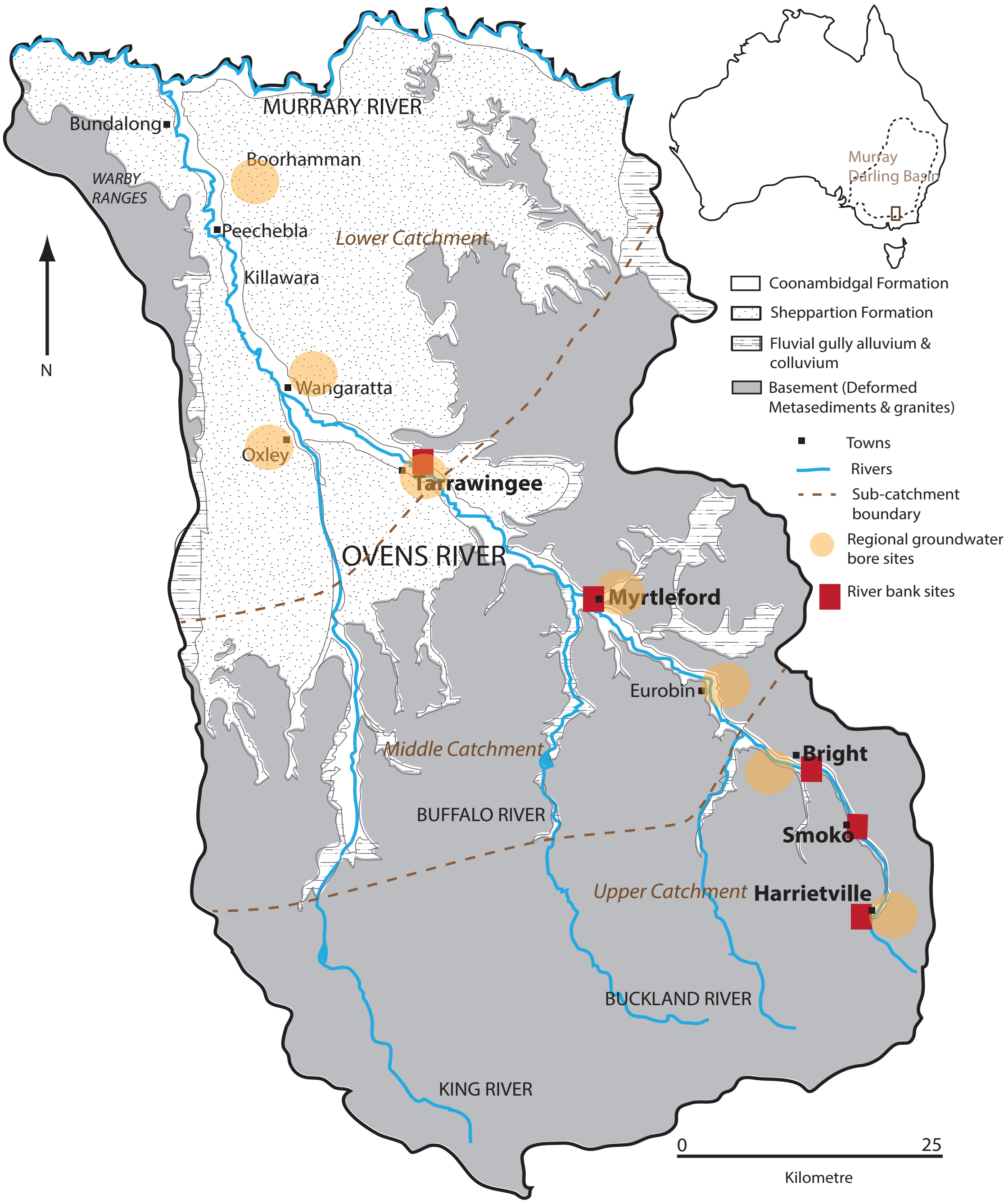

Figure 4.1 Map of the Ovens Catchment showing the surface geology, the regional groundwater sampling sites, and the river bank study sites. Data from van den Berg and Morand (1997); Victorian Water Measurement Information System (2014).

lacustrine Quaternary Shepparton Formation and the Holocene Coonambidgal Formation which are contiguous and indistinguishable in the Ovens catchment. These two formations together are up to ~170 m thick and contain mostly sand, gravel, silt and clay with fragments of basement rocks and minerals that form heterogeneous unconfined to confined aquifers (Tickell, 1978; Slater and Shugg, 1987; Lawrence, 1988); groundwater from these two formations interacts with the Oven

River. The sediments of the Shepparton and Coonambidgal Formations are coarser grained and immature in the upper valleys and become finer grained and more mature in the lower catchment. The hydraulic conductivity of the Shepparton and Coonambidgal Formations varies from 0.1 to 60 m day^{-1} with an average of 0.2 *to* 5 m day^{-1} (Tickell, 1978; Shugg, 1987). The surface aquifers receive recharge through direct infiltration on the valley floors, and via exposed and weathered bedrock at the margins of the valleys. The vertical head gradients throughout the Ovens Catchment are generally downward, while the vertical head gradients within a few tens of metres of the river in the upper and middle catchments are upwards (Victorian Water Resource Data Warehouse, 2011). The regional groundwater flow is northwest, parallel to the valley.

The average rainfall decreases from 1,127 mm in the alpine region at Bright to 636 mm on the alluvial plains in Wangaratta with most rainfall occurring in the Australian winter months (June *to* September) (Bureau of Meteorology, 2013). Potential evaporation increases northwards and ranges from 0 *to* 40 mm month^{-1} in winter to 125 *to* 200 mm month^{-1} in summer (Bureau of Meteorology, 2013). The Ovens River is the major river in the catchment and is perennial with a length of approximately 202 km (Fig. 4.1). It has a single channel confined within a steep-sided valley south of Myrtleford and then develops into a network of meandering and anastomosing channels north of Whorouly before discharging to the Murray River. The mean monthly discharge at Peechelba is between 4,090 and 414,793 ML day^{-1} with high flows occurring in winter (Water Measurement Information System, 2014). Groundwater inflows to the Ovens River during winters are higher in the upper and middle catchments due to hydraulic loading caused by recharge of the valley aquifers, while there is a lower and constant groundwater discharge into the river in the lower catchment. Groundwater input to the river based on radon mass balance is between 2 and 17% of the total discharge (Chapter 2).

The Riverine Plain and alluvial flats in the Ovens Catchment are primarily cleared for agricultural activities, including grazing, horticulture, vineyards, and orchards, while the hills and mountains are covered by native eucalyptus and plantation forests. Water extraction from both surface and groundwater resources is 10% of the total water resource available in the catchment (Victorian Government Department of Sustainability and Environment, 2010).

4.2.2 Description of study sites

This study focuses on river reaches at Harrietville, Smoko, Bright, Myrtleford, and Tarrawingee (Fig. 4.1). Harrietville, Smoko and Bright are located in the upper Ovens Catchment, while Myrtleford and Tarrawingee are located in the middle and lower catchments, respectively. The observation bores in these locations are within ~50 m of the river bank and are arranged in transects perpendicular to the river with the screens defined as shallow (3 *to* 8 m), medium (9 *to* 11 m) or deep (21 *to* 59 m) (Table 4.1). Details of river morphology and river bank lithology are summarised in table 4.1 and illustrated in figure 4.2. Briefly, the Ovens River at Harrietville is located close to the western margin of the valley with gravel river banks. At Smoko, the Ovens River is located adjacent to a cliff on the eastern margin of the valley with a low-lying river bank. The river bank is similar to those at Harrietville. At Bright, the Ovens River runs through the centre of the valley with steep banks. The river banks at Bright consist of mainly gravel with a small proportion of clay. At Myrtleford, the Ovens River runs through the eastern part of a broader valley with gravel and sandy river banks. The Ovens River at Tarrawingee is situated in the middle of a broader valley with a much wider floodplain. The river bank lithology is a combination of sand and clay with a minor amount of gravel.

4.3 Sampling and analytical methods

In this study, regional, intermediate groundwater and near-river groundwater are defined as groundwater in the area at least 50 m away from the river channel, groundwater in the area

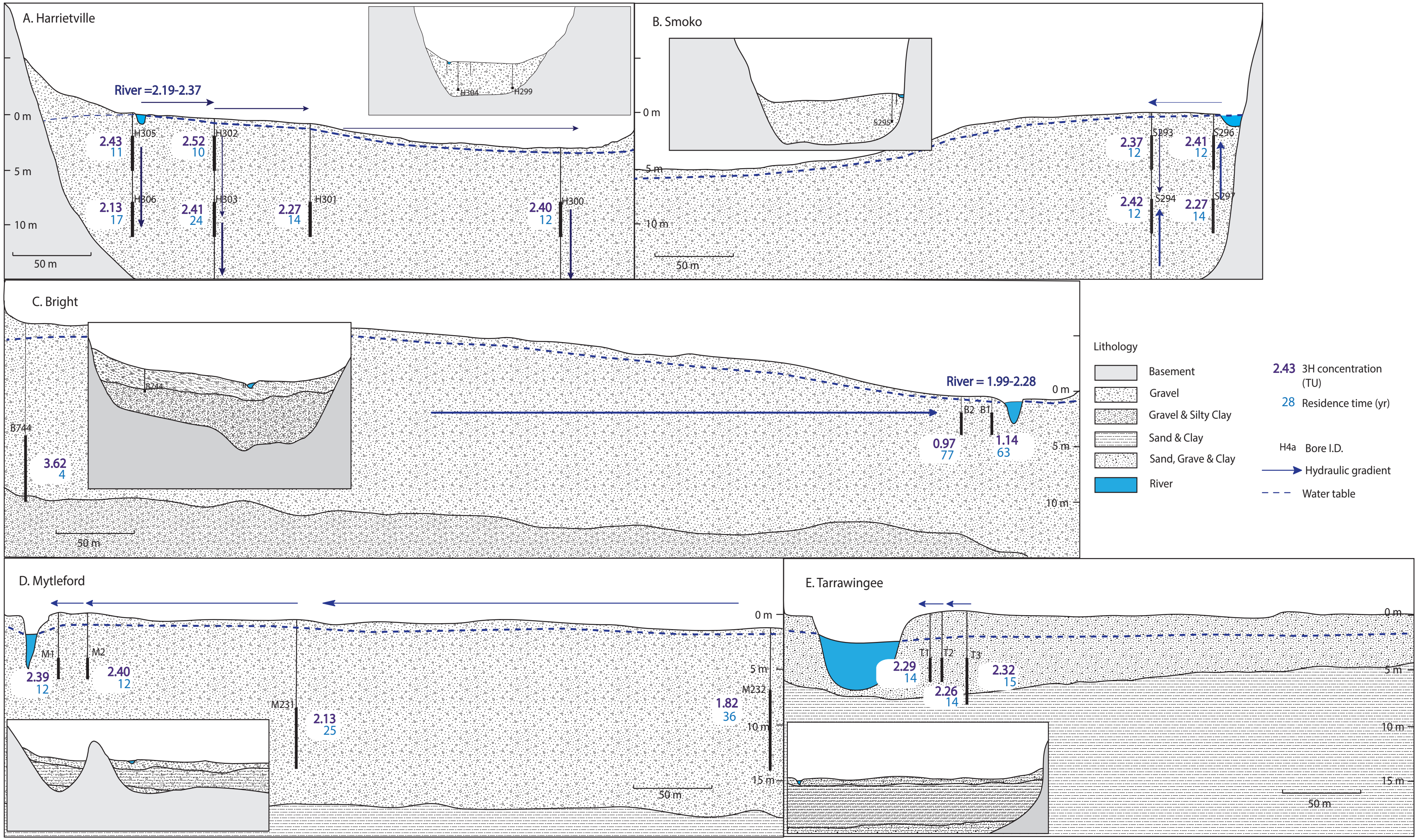

Figure 4.2 Cross sections of the five river bank study sites: (A) Harrietville, (B) Smoko, (C) Bright, (D) Myrtleford and (E) Tarrawingee. Each cross section shows the location of the river in relation to the valley (inserts), lithology of the river bank, groundwater bore localities, hydraulic gradients (during low flow conditions), ^{3}H concentrations and groundwater residence times. Hydraulic gradient (dark blue arrows) are based on the spot measurements of the hydraulic head data in Tables 4.2 *to* 4.4. ^{3}H concentrations from Tables 4.5 *to* 4.8.

Catchment	Location	Northing	Easting	Bore ID.	Screen Depth (m)	Distance to River (m)	Bank lithology	River and bank morphology
Upper	Harrietville	5918093	505568	H306	3-5	0.5	0 - 1 m – Soil	Located > 80 m from the western side of narrow valley
Upper	Harrietville	5918094	505567	H305	9-11	0.5	1 - 25 m – Gravel	A single river channel
Upper	Harrietville	5918153	505613	H302	3-5	34	> 25 m – Weathered siltstone basement	Consists of river gravel and cobbles
Upper	Harrietville	5918153	505611	H303	9-11	34		River depth, less than 1 m deep
Upper	Harrietville	5918153	505610	H304	23-25	34		Riverbank slope 60 – 45^{o}
								Relatively flat topography for the valley alluvial plain with a maximum difference of < 5 m
Upper	Smoko	5927062	505039	S296	3-5	0.5	0 - 0.5 m – Soil	Located against valley cliff at the eastern margin of the valley
Upper	Smoko	5927060	505039	S297	9-11	0.5	0.5 - 23 m – Gravel	Similar river geomorphology to Harrietville
Upper	Smoko	5927033	505021	S293	3-5	41	> 23 m – Weathered siltstone basement	Increased topography for valley alluvial plain (a difference of up 10 m over < 1 km)
Upper	Smoko	5927034	505020	S294	9-11	41		
Upper	Smoko	5927035	505018	S295	21-23	41		
Upper	Bright	5935517	499270	B1	2-4	10	0 - 1 m – Soil	Located in the middle of the valley
Upper	Bright	5935513	499260	B2	2-4	16	1 - 11 m – Gravel	Consists of river gravel, cobbles and sand
							11 - 70 m – Gravel & silty clay	River depth, less than 1 m deep
							> 70 m – Weathered siltstone basement	Riverbank slope 45 – 90^{o}
								Further increase in the topography of the valley alluvial plain (8 m difference in elevation over 700 m between the river bank and the valley margin)
Middle	Myrtleford	5952920	474605	M1	4-6	8	0 - 1.5 m – Soil	Located in the western side of the broader valley with an alluvial plains of up 6 km wide
Middle	Myrtleford	5952936	474606	M2	4 6	22	1.5 - 16 m – Gravel	More meandering with exposed gravel bed
							16 - 36 m – Sand and clay	Consists of river gravel and sand
							36 - 46 m – Weathered siltstone	River depth ,up to 4 m deep
							> 46 m – Fractured siltstone	Riverbank slope 70 – 90^{o}
								Decreased topography - an increase in elevation of 30 m over a distance of 6 km
Lower	Tarrawingee	5970210	451112	T1	5-7	6	0 - 2 m – Soil	Located in the middle of floodplains with an alluvial plains of up 30 km wide
Lower	Tarrawingee	5970212	451121	T2	5-7	14	2 - 12 m – Gravel, sand and clay	Meandering river with a anastomosing channel
Lower	Tarrawingee	5970245	451136	T3	6-8	52	12 - 92 m – Sand and clay sand and grave	Consists of sand and mud
							92 - 102 m – Clay, coarse sand and grave	River depth of 3 to 6 m
							> 102 m – Weathered bedrock	Riverbank slope 80 – 90^{o}

Table 4.1 Details of bores, riverbank lithology, and river geomorphology for each study site.

between 10 and 50 m from the river channel and groundwater in the area less than 10 m away from the river channel, respectively. Groundwater elevations and EC of the intermediate groundwater and near-river groundwater were recorded at 30 minute-intervals using Aqua TROLL® 200 loggers. Groundwater elevation was corrected for barometric pressure using a Rugged Baro TROLL®.

Rising slug head tests were performed in some of the bores located along the river banks. These tests were performed by pumping the bore for approximately 15 min at a rate of 4 L min^{-1} and then allowing the hydraulic head to recover. The recovery was recorded using a Rugged TROLL® 200 logger logging at 1 second interval. The hydraulic head measurements during the recovery were used to calculate the hydraulic conductivity using the simplified Hvorslev method outlined by Fetter (1994).

Regional groundwater samples were collected in September 2009 or May 2012. Near-river and intermediate groundwater was sampled periodically; March 2011, June 2011, October 2011 and May 2012 which represent nearly baseflow conditions, although high flow

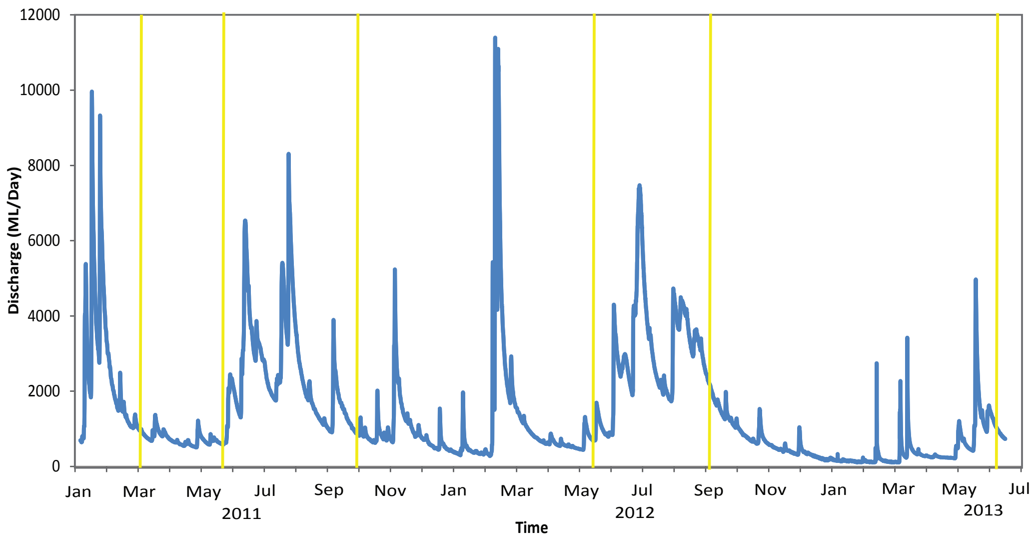

Figure 4.3 Discharge of the Ovens River (measured at Myrtleford in the middle catchment) during each groundwater sampling ground. Data from Water Measurement Information System (2013). The lines represent the sampling rounds. High flow events usually occur in Australian winter (June *to* October). Several high flow events occurred in the summer periods (e.g. February 2011 and March 2012) due to the La Niña event.

events occurred 1 *to* 2 months prior to the March 2011, October 2011 and May 2012 sampling campaigns, and September 2012 and June 2013 which represent high flow conditions (Fig. 4.3). ^{3}H concentrations and ^{14}C activities were measured in May 2012 and September 2012, respectively. Prior to groundwater sampling, the depth to water was determined at each bore using an electric measuring tape with a precision of ±1 cm. The groundwater was sampled using an impeller pump or a Bennett piston pump (for deeper bores) set at the screened interval and at least three bore volumes of water were purged prior to sampling. River samples were collected at Harrietville, Bright, Smoko and Porepunkah in the upper catchment in September 2012, December 2013 and February 2014 for ^{3}H analysis. Additional measurements of ^{3}H concentrations and ^{14}C activities of regional groundwater in the catchment are from Cartwright and Morgenstern (2012). EC and pH were measured in the field using a calibrated TPS WP-81 conductivity/pH meter and probes with accuracy of ±0.01 for pH and of ±0.2% for EC. Alkalinity was determined using a Hach digital titrator and reagents with a precision of ±5%. Cations were analysed using a ThermoFinnigan OptiMass 9500 ICP-MS at Monash University on samples that were filtered through 0.45 μm cellulose nitrate filters and acidified to pH < 2 with double distilled 16M nitric acid. Drift during ICP-MS analysis was corrected using internal Sc, Y, In and Bi standards, and replicate analyses indicate a precision of ±5%. Anion concentrations were measured on filtered and unacidified samples using a Metrohm ion chromatograph at Monash University. The precision of anion concentrations estimated by replicate analysis is ±2%. The charge balance errors for the groundwater samples were -8 *to* 20%. Stable isotopes were measured at Monash University using Finnigan MAT 252 and ThermoFinnigan DeltaPlus Advantage mass spectrometers. $\delta^{18}O$ values were determined via equilibration with He-CO at 32 °C for 24 *to* 48 hours in a ThermoFinnigan Gas Bench. $\delta^{2}H$ was measured by reaction with Cr at 850 °C using an automated Finnigan MAT H/Device. $\delta^{18}O$ and $\delta^{2}H$ values were measured relative to internal standards calibrated using IAEA SMOW, GISP and SLAP. Data were normalized following Coplen (1988) and are expressed relative to V-SMOW, where $\delta^{18}O$ and $\delta^{2}H$ of SLAP are -55.5‰ and -428‰, respectively. Precision (1σ) based on replicate analysis is: $\delta^{18}O = \pm 0.1‰$ and $\delta^{2}H = \pm 1‰$. ^{3}H was measured by liquid scintillation spectrometry at National Isotope Centre, New Zealand on samples that were distilled and electrolytically enriched. ^{3}H concentrations are expressed in tritium units (TU), and the precision at an average tritium concentration of 4 TU is ±0.06 TU, and the detection limit is ±0.03 TU (Morgenstern and Taylor, 2009). The ^{14}C activities were measured on graphitised samples on a National Electrostatics single stage accelerator mass spectrometer at the Australian National University. ^{14}C activities are expressed as percent modern carbon (pMC) and the precision of $^{14}C/^{12}C$ ratios is ±0.5%.

Periodical sampling in this study resulted in limited data on the geochemistry of groundwater and thus affects the ability of interpreting the variability of the groundwater geochemistry over a shorter time scale as well as the ability of integrating geochemistry data with discharge or EC data which collected over a shorter interval. For example, the Cl concentrations of groundwater in response to river discharge may have been more variable than reported in this study. One way to address this problem is to compare the Cl concentrations collected from the sampling rounds with the EC values of groundwater (see the results section in section 4.4.4) which were collected at a much shorter interval before the Cl concentrations are used in the analysis.

4.4 Results

4.4.1 Hydraulic conductivities

The hydraulic conductivity yielded from the rising head slug tests for the river banks of the Ovens River at Harrietville, Bright, Myrtleford and Tarrawingee were 29, 22, 22 and 11 m day^{-1}, respectively. The hydraulic conductivity of river banks along the Ovens River decreases down along the course of river.

4.4.2 Groundwater elevations and hydraulic gradients

At Harrietville, the lateral hydraulic gradients were away from the river, ranging from 0.008 *to* 0.023 in the river bank to 0.001 *to* 0.017 in the valley (Fig. 4.4aiv, Table 4.2). The vertical hydraulic gradient immediately adjacent to the river was 0.017 *to* 0.082 downward until it was reversed to a gradient of ~0.017 upward after February 2013 (Fig. 4.4aiii). Following the reversed gradient in February 2013, the hydraulic heads in the region dropped by up to 0.99 m (Table 4.2). At Smoko, the lateral gradient was 0.003 *to* 0.006 away from the river (Fig. 4.4biv). The vertical gradient was much greater, ranging from 0.003 to 0.024 upward. The vertical gradient was reversed to a maximum of 0.008 downward for 3 months during the winter in 2012 (Fig. 4.4biii). At Bright, the lateral gradient in the valley was ~0.008 toward the river (Table 4.2). The horizontal gradient in the bank was more variable, from 0.003 to 0.010 toward the river (Fig. 4.4civ) but did not show any reversals. At Myrtleford, the lateral gradients in the valley and in the bank were 0.002 *to* 0.009 toward the river (Fig. 4.5aiv). At Tarrawingee, the lateral gradients in the river bank were 0.001 *to* 0.005 toward the river (Fig. 4.5biv). However, the lateral gradient at Myrtleford and Tarrawingee were sometimes reversed during high flow conditions. These reversed hydraulic gradients lasted from several minutes to several hours. The longest periods of reversed hydraulic gradients at Myrtleford and Tarrawingee were 57 hours and 100 hours, respectively, and these gradient reversals occurred during the high flow events (up to 11,395 ML day^{-1} measured at Myrtleford) in March 2013. Overall, during low flow conditions, the hydraulic gradients are away from the river at Harrietville, while they are toward the river at Smoko, Bright, Myrtleford and Tarrawingee. During high flow conditions, the gradient at Smoko, Myrtleford and Tarrawingee can be reversed for a period of several hours to several days (Myrtleford and Tarrawingee) to several weeks (Smoko).

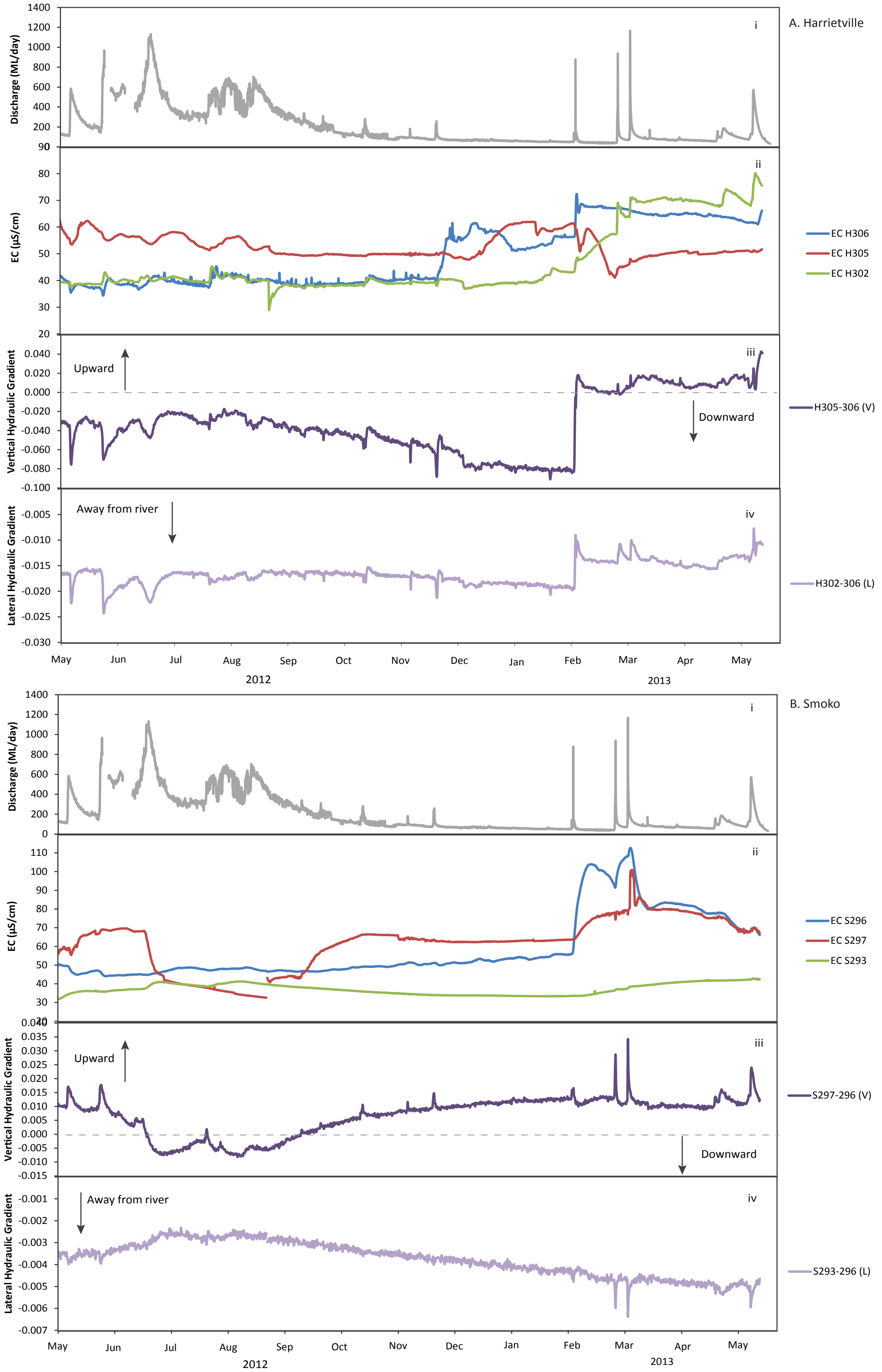

Figure 4.4 See the next page for the figure caption.

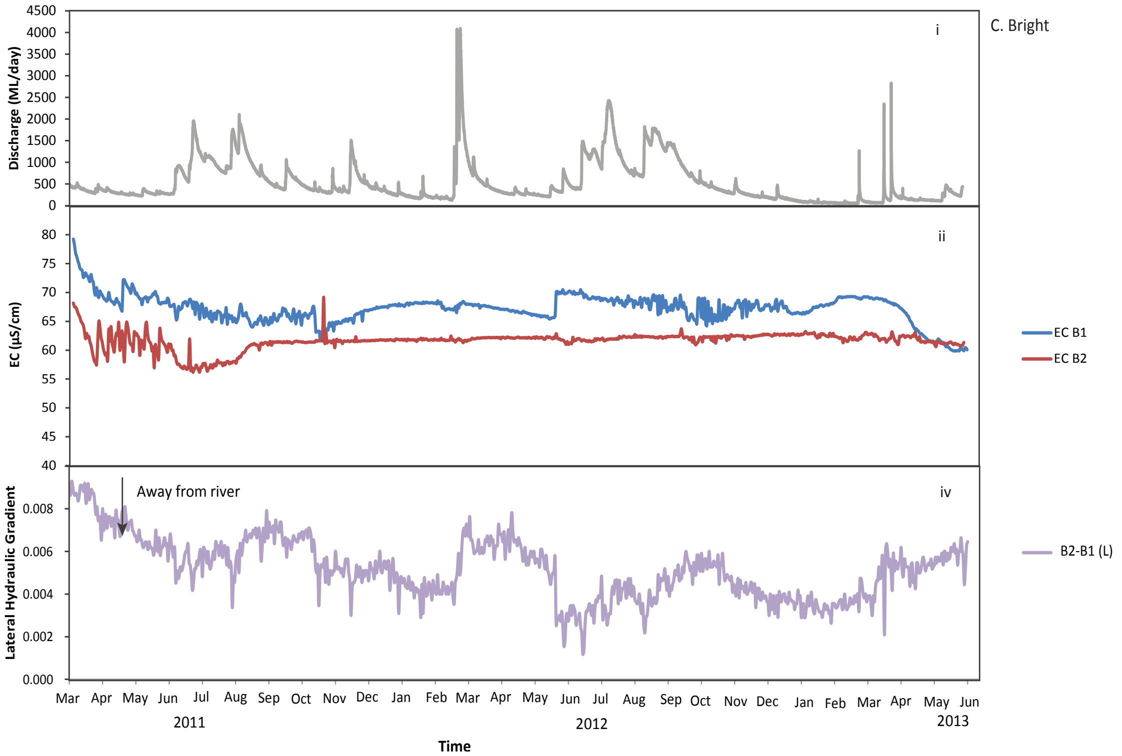

Figure 4.4 (cont.) Vertical (iii) and lateral (iv) hydraulic gradients based on the head data from loggers and continuous EC values of near-river and intermediate groundwater (ii) in respect to discharge (i) for Harrietville (A), Smoko (B) and Bright (C) in the upper catchment. The vertical gradient was reversed for several weeks at Smoko during the winter months. The EC values of near-river groundwater at Harrietville were lower during high flow events at Harrietville, while that of near-river groundwater at Smoko were lower following the head reversals during the winter months. In contrast, no head reversals at Bright, and the EC at the Bright river bank was relatively constant. Discharge data from Water Measurement Information System (2013).

4.4.3 Electrical conductivity

The EC values of the regional groundwater increased down catchment from 30 *to* 200 µS cm^{-1} in the upper and middle catchments to 50 *to* 12,020 µS cm^{-1} in the lower catchment (Tables 4.2 *to* 4.4). At Harrietville and Smoko, the continuous EC values of the shallow near-river and intermediate groundwater were 38 *to* 110 and 33 *to* 101 µS cm^{-1}, respectively (Figs. 4.4aii & bii). The lower EC values at both locations were usually recorded following high river flows in the winter months. At Harrietville, the EC values of the near-river groundwater increased with depth except for a short period in late December 2012 and the period after February 2013. The change in the spatial variation of EC after February 2013 was coincided with the vertical head reversal in

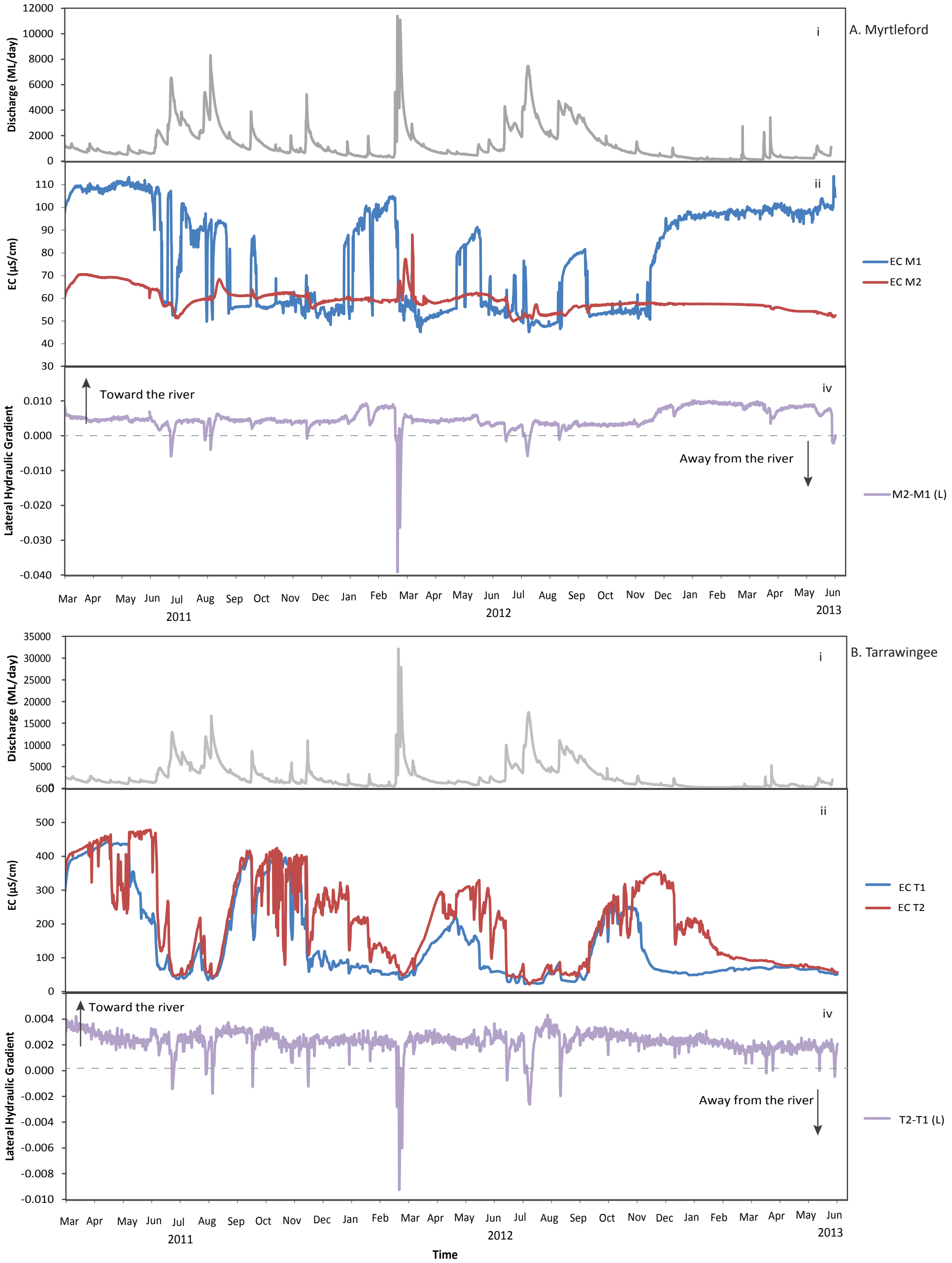

Figure 4.5 Lateral (iv) hydraulic gradients based on the head data from loggers and continuous EC values of near-river and intermediate groundwater (ii) in respect to discharge (i) for Myrtleford (A) Tarrawingee (B) in the middle and low catchments, respectively. The lateral gradient was reversed for several hours at Myrtleford and Tarrawingee following some high flow events. The EC values of near-river groundwater decreased following the high flow events and head reversals. Discharge data from Water Measurement Information System (2013).

February 2013. At Smoko, the vertical EC trend of the near-river groundwater changed from increasing with depth to decreasing with depth over the study period. Additionally, the EC values of the near-river groundwater at the medium depth decreased during winter months in 2012 when the vertical gradient was reversed. At Bright, the continuous EC values of the near-river and intermediate groundwater were generally between 60 and 79 $\mu S\ cm^{-1}$ and between 60 and 68 $\mu S\ cm^{-1}$, respectively. These EC values generally did not vary in response to the hydraulic gradient in the river bank (Fig. 4.4cii). At Myrtleford, the continuous EC values of the near-river and intermediate groundwater were 45 *to* 114 and 52 *to* 88 $\mu S\ cm^{-1}$, respectively (Fig. 4.5aii). At Tarrawingee, the continuous EC values of the near-river groundwater were 25 *to* 434 $\mu S\ cm^{-1}$, whereas those of the intermediate groundwater were 37 *to* 473 $\mu S\ cm^{-1}$ (Fig. 4.5bii). Spot measurements also showed a reduction in the regional groundwater EC at 50 m away from the river in May 2013 (Table 4.4). The low EC values at Myrtleford and Tarrawingee were recorded following the high flow conditions. Many episodes of EC reduction occurred during and after the hydraulic head reversals in the river banks, including July 2011, August 2011, February 2012, and July 2012.

In summary, the continuous EC values of the near-river and intermediate groundwater at all sites are in the lower range of the regional groundwater EC values. The EC values of the near-river groundwater at Harrietville, Smoko, Myrtleford and Tarrawingee often decreased during and after high flow conditions with a significant reduction of 40 *to* 80% at Myrtleford and Tarrawingee. By contrast, the EC values of the near-river and intermediate groundwater at Bright were relatively stable throughout the two-year period.

4.4.4 Major ion chemistry

The dominant cation in the regional groundwater in the upper and middle catchments was Na^+ (41% of total cations on a mass basis) followed by Mg^{2+} (27%), Ca^{2+} (25%) and K^+

Date	Site	Bore ID.	EC (µS/cm)	pH	DO (mg/L)	Na (mg/L)	K (mg/L)	Mg (mg/L)	Ca (mg/L)	Cl (mg/L)	HCO3 (mg/L)	SO4 (mg/L)	NO3 (mg/L)	Head (mADH)
Regional Groundwater														
Sept 2009	Bright	B747	58	5.9	1	3.03	0.61	2.01	1.66	3.10	13	0.21	5.36	313.93
Sept 2009	Bright	B745	63	5.8	1	2.25	0.85	2.51	1.83	2.16	19	0.75	2.41	313.65
Sept 2009	Bright	B744	56	5.9	6	5.21	0.99	3.11	3.20	2.73	30	1.00	7.33	319.65
Sept 2009	Bright	B737	111	5.9	1	7.57	0.81	6.02	4.64	2.41	60	0.62	0.14	311.53
Sept 2009	Bright	B735	74	6.9	1	4.90	0.80	2.70	2.50	3.60	28	0.60	0.80	310.61
Sept 2009	Bright	B736	53	5.6	2	4.53	0.72	2.48	1.69	3.29	23	0.26	1.30	310.64
Sept 2009	Bright	B461	83	5.9	3	5.88	1.30	4.04	4.19	3.42	43	0.54	4.52	310.61
Sept 2009	Bright	B462	100	5.9	1	5.88	0.91	4.41	4.58	3.48	44	0.30	0.40	306.42
May 2012	Harrietville	H301	57	6.3	3	5.28	0.52	2.16	2.78	1.52	25	0.96	1.06	500.24
May 2012	Harrietville	H300	71	6.3	3	6.19	0.69	2.94	3.35	1.90	28	1.06	3.71	499.32
May 2012	Harrietville	H299	133	6.3	2	10.93	2.05	6.28	6.17	1.89	79	0.58	0.15	498.85
May 2012	Porepunkah	P289	57	5.5	tr	4.31	0.87	1.73	2.55	2.22	21	0.62	0.15	293.23
Jun 2013	Harrietville	H301	60											499.25
Jun 2013	Harrietville	H300	80											499.00
Jun 2013	Harrietville	H299	142											498.10
Near-river and Intermediate Groundwater														
May 2012	Harrietville	B306	46	5.9	2	3.16	0.40	2.09	2.48	1.17	18	0.81	1.81	501.24
May 2012	Harrietville	B305	59	5.7	3	4.11	0.47	2.27	3.39	1.39	25	0.79	1.45	501.03
May 2012	Harrietville	B302	43	5.9	2	3.39	0.34	2.01	2.39	1.09	19	0.84	1.38	500.70
May 2012	Harrietville	B303	63	6	2	5.44	0.60	2.24	3.53	1.36	29	0.99	1.23	500.66
May 2012	Harrietville	B304	94	6.7	4	5.54	1.97	3.16	5.47	2.39	45	1.52	0.72	498.01
Sept 2012	Harrietville	B306	43			2.51	0.46	1.79	1.90	1.07	20	0.74	0.53	501.32
Sept 2012	Harrietville	B305	53			3.97	0.44	2.12	2.37	1.30	25	0.69	2.52	501.15
Jun 2013	Harrietville	B306	70			4.28	0.53	3.24	3.05	1.54	26	0.73	6.37	500.70
Jun 2013	Harrietville	B305	56			4.50	0.47	2.29	2.08	1.51	19	0.76	2.36	500.94
Jun 2013	Harrietville	B302	78			5.01	0.62	3.75	3.63	2.03	25	1.10	5.03	500.18
May 2012	Smoko	B296	50	6.2	3	2.73	0.35	1.53	2.24	0.96	18	2.29	0.05	405.63
May 2012	Smoko	B297	58	6.3	1	2.86	0.39	1.88	2.57	0.95	34	2.33	0.14	405.84
May 2012	Smoko	B293	53	5.9	1	2.95	0.92	1.77	2.22	1.45	23	0.77	0.36	405.58
May 2012	Smoko	B294	78	6.2	2	3.98	0.58	2.67	6.27	2.11	24	0.63	0.17	405.55
May 2012	Smoko	B295	47	5.9	tr	3.47	0.50	2.00	2.64	1.07	24	0.91	0.42	405.59
Sept 2012	Smoko	B296	36			2.04	0.34	1.31	1.90	0.87	28	1.55	0.00	406.96
Sept 2012	Smoko	B297	35			2.05	0.33	1.24	1.91	0.86	27	2.28	0.07	406.35
Jun 2013	Smoko	B296	74			2.46	0.68	3.19	3.08	1.42	24	0.98	0.13	405.53
Jun 2013	Smoko	B297	71			2.61	0.64	3.09	2.88	1.31	38	1.07	0.12	405.76
Jun 2013	Smoko	B293	47			3.18	0.37	2.03	1.79	1.25	23	0.95	0.24	405.46
Mar 2011	Bright	B1	95			7.34	0.85	3.82	4.52	2.95	32	0.35	0.48	314.02
Mar 2011	Bright	B2	82			7.42	0.92	3.56	4.28	2.87	24	0.30	0.78	314.14
Jun 2011	Bright	B1	74			5.63	0.88	3.72	3.86	2.47	32	0.46	3.12	313.81
Jun 2011	Bright	B2	69			5.05	0.80	3.42	3.58	2.27	35	0.53	2.91	313.77
Oct 2011	Bright	B1	72			3.63	0.79	2.51	3.27	2.42	35	0.47	1.75	313.93
Oct 2011	Bright	B2	67			3.54	0.86	2.16	3.14	2.16	42	0.43	1.64	313.99

Table 4.2 See next page for the table caption

May 2012	Bright	B1	65	5.7	1	3.93	0.56	2.75	3.18	2.19	27	0.43	1.63	313.82
May 2012	Bright	B2	73	5.6	1	4.02	0.57	2.86	3.67	2.07	34	0.30	1.46	313.88
Sept 2012	Bright	B1	65			4.11	0.56	2.73	2.79	3.88	33	0.93	1.62	314.11
Sept 2012	Bright	B2	61			3.86	0.54	2.53	2.90	3.60	40	0.63	1.50	314.17
Jun 2013	Bright	B1	73			4.48	0.71	3.21	2.84	2.42	33	0.40	3.21	313.86
Jun 2013	Bright	B2	91			3.81	0.59	2.67	2.73	1.92	30	0.25	0.60	313.93

Table 4.2 (cont.) Geochemistry of the regional, near-river and intermediate groundwater, and hydraulic heads (spot measurements) in the upper catchment (Harrietville, Smoko and Bright). Data for regional groundwater in Bright from Cartwright & Morgenstern (2012).

Date	Site	Bore ID.	EC (µS/cm)	pH	DO (mg/L)	Na (mg/L)	K (mg/L)	Mg (mg/L)	Ca (mg/L)	Cl (mg/L)	HCO_3 (mg/L)	SO_4 (mg/L)	NO_3 (mg/L)	Head (mADH)
Regional Groundwater														
Sept 2009	Eurobin	E069	129	6.5	3	8.39	1.14	3.35	2.36	3.59	35	9.6	0.07	248.77
	Eurobin	E068	74	6.2	1	3.39	1.99	2.36	1.73	3.52	20	2.63	0.09	248.37
	Eurobin	E067	92	6	2	4.16	1.08	3.57	2.91	4.14	15	0.71	14.22	247.74
	Eurobin	E066	78	5.6	3	3.28	1.35	2.92	2.45	3.9	9	0.69	17.11	247.66
	Myrtleford	B232	107	6	1	10.9	1.67	3.97	3.17	9.33	33	6.64	2.59	204.9
	Myrtleford	B231	49	6.1	4	3.53	0.72	2.48	2.09	2.08	24	3.19	1.46	205.04
Near-river and Intermediate Groundwater														
Mar 2011	Myrtleford	M1	90			7.46	0.96	3.18	3.44	2.37	17	4.96	0.29	204.27
	Myrtleford	M2	68			8.21	1.39	4.67	5.25	2.43	23	4.87	3.54	204.36
Jun 2011	Myrtleford	M1	108			3.09	0.77	2.06	2.52	1.52	29	4.94	0.71	203.95
	Myrtleford	M2	71			3.12	0.87	2.46	3.2	1.56	25	4.33	2.55	204.05
Oct 2011	Myrtleford	M1	77			3.63	0.79	2.51	3.27	2.42	33	0.47	1.75	204.11
	Myrtleford	M2	68			3.54	0.86	2.16	3.14	2.16	22	0.43	1.64	204.19
May 2012	Myrtleford	M1	77	5.8	3	4.12	0.68	2.55	2.83	2.35	29	4.22	1.2	204
	Myrtleford	M2	65	5.8	2	3.39	0.77	2.79	3.29	1.66	22	3.62	4.05	204.07
Sept 2012	Myrtleford	M1	66			2.92	0.49	2.12	2.03	1.33	32	4.42	1.26	204.45
	Myrtleford	M2	59			2.88	0.59	2.32	2.43	1.29	20	3.98	3.04	204.53
Jun 2013	Myrtleford	M1	114			4.55	0.73	2.94	2.33	1.36	16	8.2	8.25	204.18
	Myrtleford	M2	57			3.41	0.82	2.41	2.25	1.25	23	3.1	3.11	204.21

Table 4.3 Geochemistry of the regional, near-river and intermediate groundwater, and hydraulic heads (spot measurements) at Myrtleford in the middle catchment. Data for regional groundwater from Cartwright & Morgenstern (2012).

(8%), while the dominant anion was HCO_3^- (82%) followed by Cl^- (8%) and SO^- (4%) (Fig. 4.6a). The composition of the regional groundwater in the upper and middle catchments was similar to that of river water (Chapter 2). In the lower catchment, the relative proportion of Na^+ in the groundwater increased at the expense of Ca^{2+} and Mg^{2+} (81% vs. 13% and 5%, respectively). The dominant anion was Cl^- (59%) and then HCO3$^-$ (22%) (Fig. 4.6a). Similar trends in major ion geochemistry have been observed elsewhere in the Murray Basin and correspond to changes

Date	Site	Bore ID.	EC (µS/cm)	pH	DO (mg/L)	Na (mg/L)	K (mg/L)	Mg (mg/L)	Ca (mg/L)	Cl (mg/L)	HCO_3 (mg/L)	SO_4 (mg/L)	NO_3 (mg/L)	Head (mADH)
Regional Groundwater														
Sept 2009	Oxley	O738	106	9.8	1	15	2.71	0.78	22	27	35	0.16	0.05	149.16
Sept 2009	Wangaratta	W326	1341	6.5	1	192	3.07	59.79	33	298	389	25.34	30	136.24
Sept 2009	Wangaratta	W493	920	6.7	1	134	0.64	27.08	12	134	322	14.8	0	136.11
Sept 2009	Boorhaman E.	BE2296	567	9.4	1	95	2.53	4.52	7.40	115	123	0.91	0.03	132.01
Sept 2009	Boorhaman E.	BE323	536	6.3	1	89	1.40	3.93	27	96	155	4.8	0.62	132.06
Sept 2009	Boorhaman	BH788	3800	7.0	2	654	7.02	61.35	59	923	333	214	1.67	128.58
Sept 2009	Boorhaman	BH789	12020	6.5	1	2331	13	355.7	357	3830	1250	879	3.44	129.08
Sept 2009	Peechelba	PE306	1194	6.1	2	163	1.70	51	11	299	151	37.86	2.36	109.26
Sept 2009	Bundalong S.	BS310	2270	6.6	1	389	2.38	56.61	20	628	218	172	36	125.41
Mar 2011	Tarrawingee	T3	315			53	3.15	8.73	10	46	45	0.28	1.53	154.35
Jun 2011	Tarrawingee	T3	475			57	3.59	8.85	11	81	156	0.15	0.11	153.88
Oct 2011	Tarrawingee	T3	485			41	5.28	7.38	7.95	71	94	0.09	0.11	153.95
May 2012	Tarrawingee	T3	464	6.4	1	41	1.95	6.39	6.65	77	127	0.14	0.04	153.87
Sept 2012	Tarrawingee	T3	353			33	2.3	4.41	5.12	57	85	0.17	0.02	154.72
Jun 2013	Tarrawingee	T3	96			8.25	0.74	1.36	1.17	2.92	31	0.2	0.04	154.01
Near-river and Intermediate Groundwater														
Mar 2011	Tarrawingee	T1	367			52	2.25	8.13	9.51	47	52	0.24	0.50	154.09
Mar 2011	Tarrawingee	T2	364			55	3.76	7.39	11	48	45	0.36	1.05	154.17
Jun 2011	Tarrawingee	T1	235			25	1.38	4.72	5.77	30	42	0.70	2.07	153.80
Jun 2011	Tarrawingee	T2	145			49	2.02	6.68	8.93	72	42	0.13	1.09	153.86
Oct 2011	Tarrawingee	T1	364			30	5.19	6.71	8.55	50	43	0.11	0.02	153.82
Oct 2011	Tarrawingee	T2	448			38	5.12	7.48	8.42	64	71	0.08	0.02	153.89
May 2012	Tarrawingee	T1	74	6.2	2	5.33	0.69	1.35	1.98	4.79	31	1.08	0.06	153.80
May 2012	Tarrawingee	T2	253	6.2	2	68	13	13.05	21	34	81	0.27	0.07	153.87
Sept 2012	Tarrawingee	T1	82			7.24	0.75	1.36	1.67	12.	29	0.62	0.01	154.37
Sept 2012	Tarrawingee	T2	147			11	0.75	2.22	2.34	21	59	0.22	0.01	154.61
Jun 2013	Tarrawingee	T1	53			3.56	0.8	1.67	1.45	1.93	19	0.90	0.1	154.03
Jun 2013	Tarrawingee	T2	63			4.52	1.06	1.82	1.69	1.92	25	0.86	0.35	154.09

Table 4.4 Geochemistry of the regional, near-river and intermediate groundwater and hydraulic heads (spot measurements) at Tarrawingee in the lower catchment. Data for regional groundwater, expect for Tarrawingee, from Cartwright & Morgenstern (2012).

caused by ion exchange, mineral dissolution, and calcite precipitation that occur as groundwater salinity increases by evapotranspiration (Herczeg *et al.*, 2001). There were no major differences between the chemical compositions of the near-river or intermediate groundwater and the regional groundwater at Harrietville, Smoko, Bright and Myrtleford (Fig. 4.6a). At Tarrawingee, the near-river and intermediate groundwater after high flow periods (such as May 12, September 12 & June 13) had lower relative proportions of Na^+, K^+ and Cl^-, and higher relative proportions of Ca^{2+} and HCO_3^-, which closely resembles the composition of the river water (Fig. 4.6b).

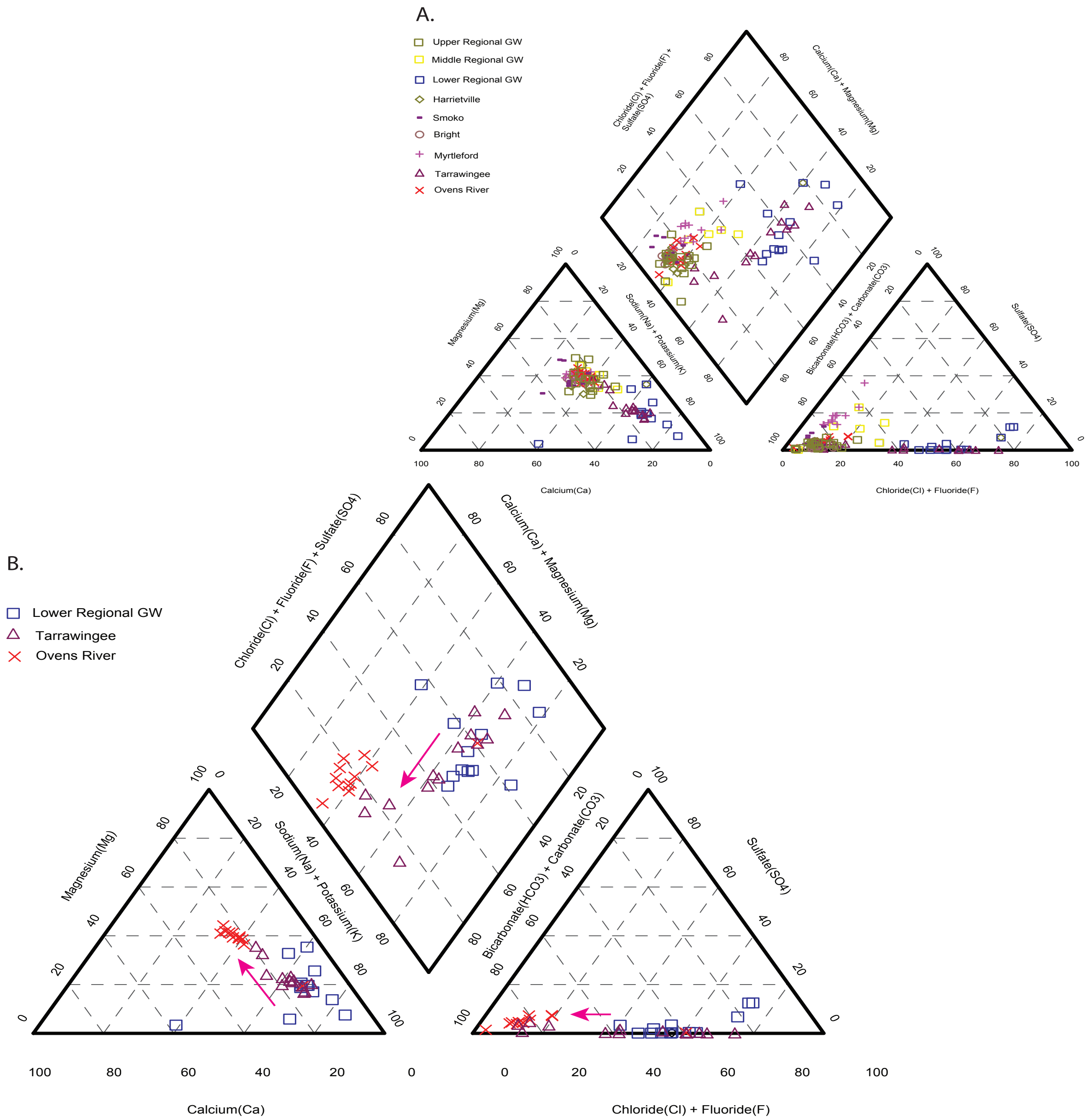

Figure 4.6 (A) Piper diagram of the near-river, intermediate and regional groundwater, and river water in the Ovens catchment. (B) Only the near-river, intermediate and regional groundwater at Tarrawingee and the river water are shown. The chemical composition of near-river and intermediate groundwater at Tarrawingee shifted toward that of river during some sampling rounds (such as, May 12, Sept 12 & June 13), indicating bank infiltration (red arrows). Even though bank infiltration occurs at Harrietville, Smoko and Myrtleford, such shift is not apparent for these areas in the Piper diagram (A) because of the overlaps of the chemical compositions between the river water and the regional groundwater chemical compositions in the upper and middle catchments. Data from Tables 4.2 – 4.4 with additional data from Cartwright & Morgenstern (2012).

The Cl concentrations of the regional groundwater were 1.1 *to* 9.3 mg L^{-1} in the upper and the middle catchments and 2.1 *to* 3,830 mg L^{-1} in the lower catchment (Tables 4.2 *to* 4.4). At Harrietville and Smoko, the Cl concentrations of the near-river and intermediate groundwater

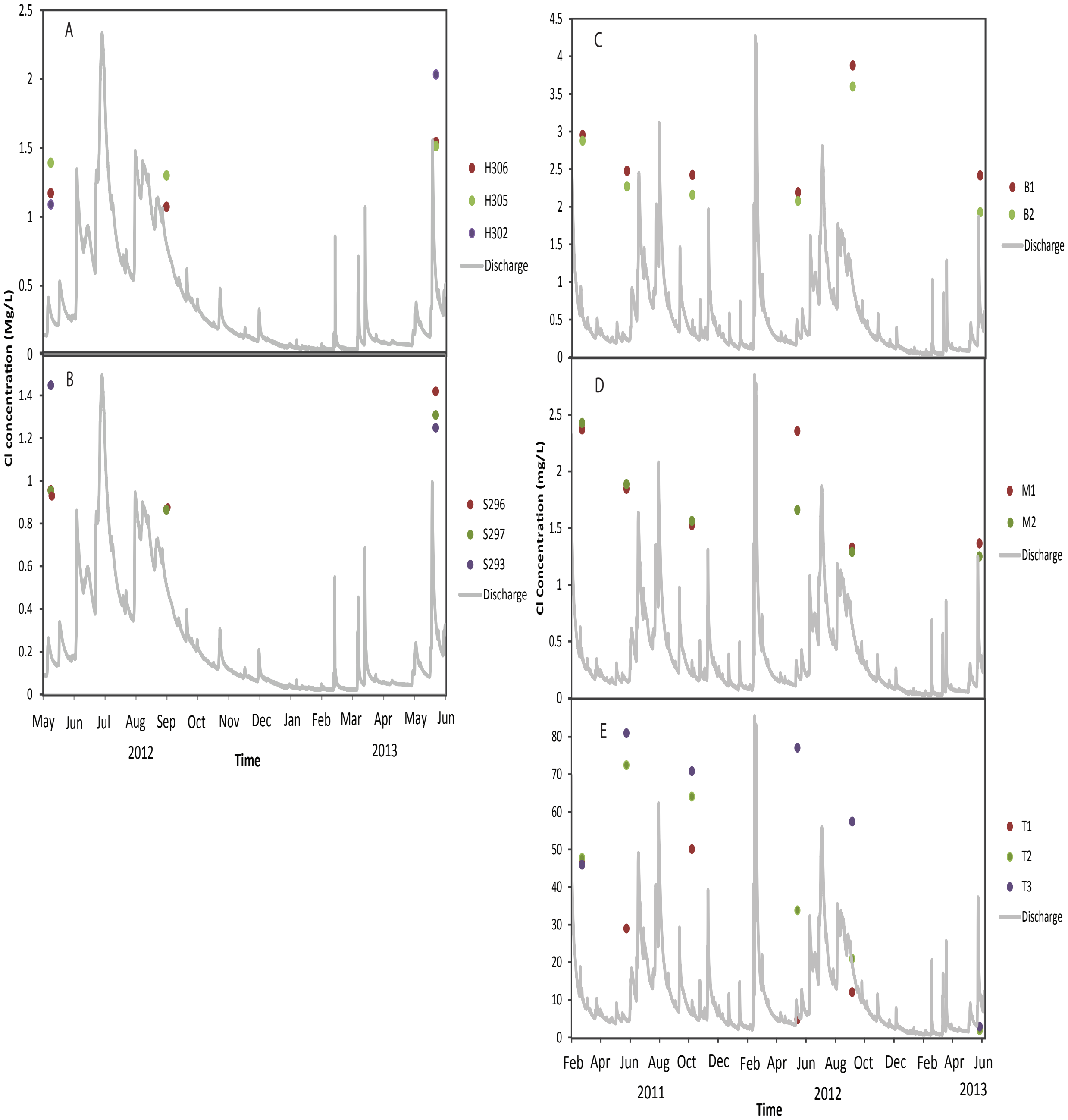

Figure 4.7 Temporal variations of the Cl concentrations of the near-river and intermediate groundwater for (A) Harrietville, (B) Smoko, (C) Bright, (D) Myrtleford and (E) Tarrawingee in respect to the discharge of the Ovens River. Cl concentrations of near-river and intermediate groundwater at river banks such as Harrietville (A), Smoko (B), Myrtleford (D) and Tarrawingee (E) were lower after high flow events. Data from Tables 4.2 – 4.4. Discharge data from Water Measurement Information System (2013).

generally increased with distance from the river (from 1.1 *to* 1.5 to 1.2 *to* 2.0 mg L^{-1} at Harrietville and from 0.9 *to* 1.4 to 1.3 *to* 1.5 mg L^{-1} at Smoko). At Harrietville, the Cl concentrations of the near-river groundwater increased with depth, except for June 2013 in which the vertical trend of the Cl concentrations was reversed (Fig. 4.7a). At Smoko, the Cl concentrations of the near-river groundwater at shallow and medium depths were similar at the beginning the study period. They then increased with depth in June 2013 (Fig 4.7b). The Cl concentrations of the near-river and intermediate groundwater at Bright ranged from 1.9 to 3.8 mg L^{-1} with the high Cl concentrations occurring in September 2012 (Fig. 4.7c). At Myrtleford, the Cl concentrations of the near-river and intermediate groundwater fluctuated between 1.3 and 2.4 mg L^{-1}. The low Cl concentrations in the river banks of Myrtleford were recorded after high flow events (Fig. 4.7d). At Tarrawingee, the Cl concentrations of the near-river groundwater were generally above 47 mg L^{-1} but fell to below 12 mg L^{-1} in May 2012, September 2012 and June 2013. Likewise, the Cl concentrations of the intermediate groundwater were also generally above 48 mg L^{-1} but were reduced to 21 *to* 34 mg L^{-1} in May 2012 and September 2012, and 2 mg L^{-1} in June 2013 (Fig 4.7e). In summary, the spatial and temporal variations in the Cl concentrations of the near-river and intermediate groundwater in the Ovens Catchment are similar to those of the EC values which were collected over a shorter time interval. The Cl concentrations of the near-river and intermediate groundwater at Harrietville, Smoko, Myrtleford and Tarrawingee decreased following high flow conditions. At Bright, the Cl concentrations of near-river and intermediate groundwater do not decrease in response to high flow conditions.

The majority of the regional groundwater in the upper, middle and lower catchment had molar Na/Cl ratios of 1.4 *to* 4.5, 1.5 *to* 4.0 and 0.8 *to* 1.5, respectively (Fig. 4.8). One groundwater sample which was collected in a deep bore in the upper catchment contained a very high Na/Cl (8.5). The molar Na/Cl ratios of the shallow, less-saline near-river and intermediate groundwater at Harrietville and Smoko were 3.6 *to* 4.7 (Figs. 4.8a & 4.8b). The near-river and intermediate groundwater at Bright had Na/Cl ratios between 1.7 and 4.0 (Fig. 4.8c). At Myrtleford, the molar Na/Cl ratios of the near-river and intermediate groundwater were relatively higher, ranging between 2.4 and 5.2 (Fig. 4.8d). At Tarrawingee, the molar Na/Cl ratios of the near-river and intermediate

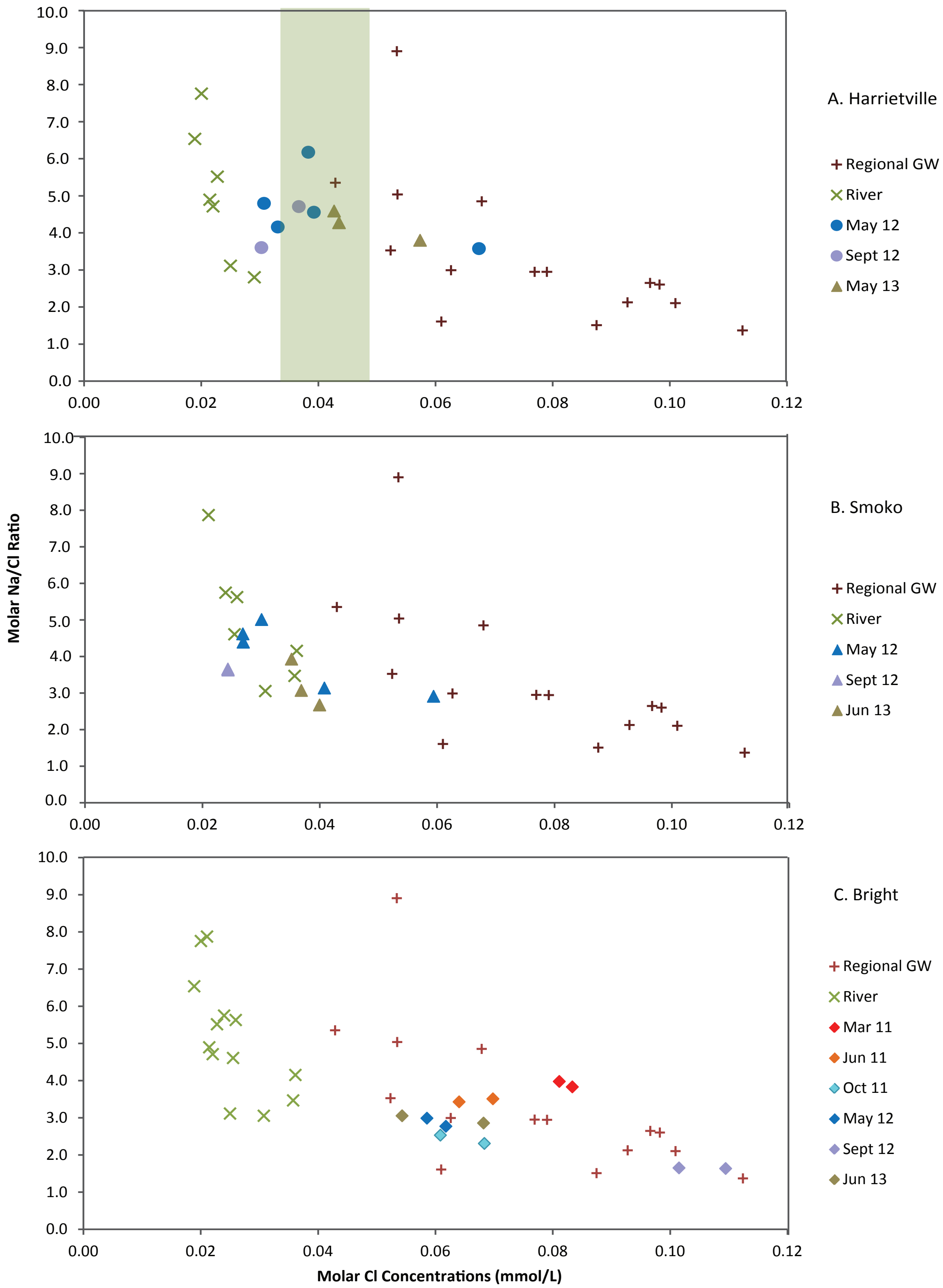

Figure 4.8 See the next page for the figure caption.

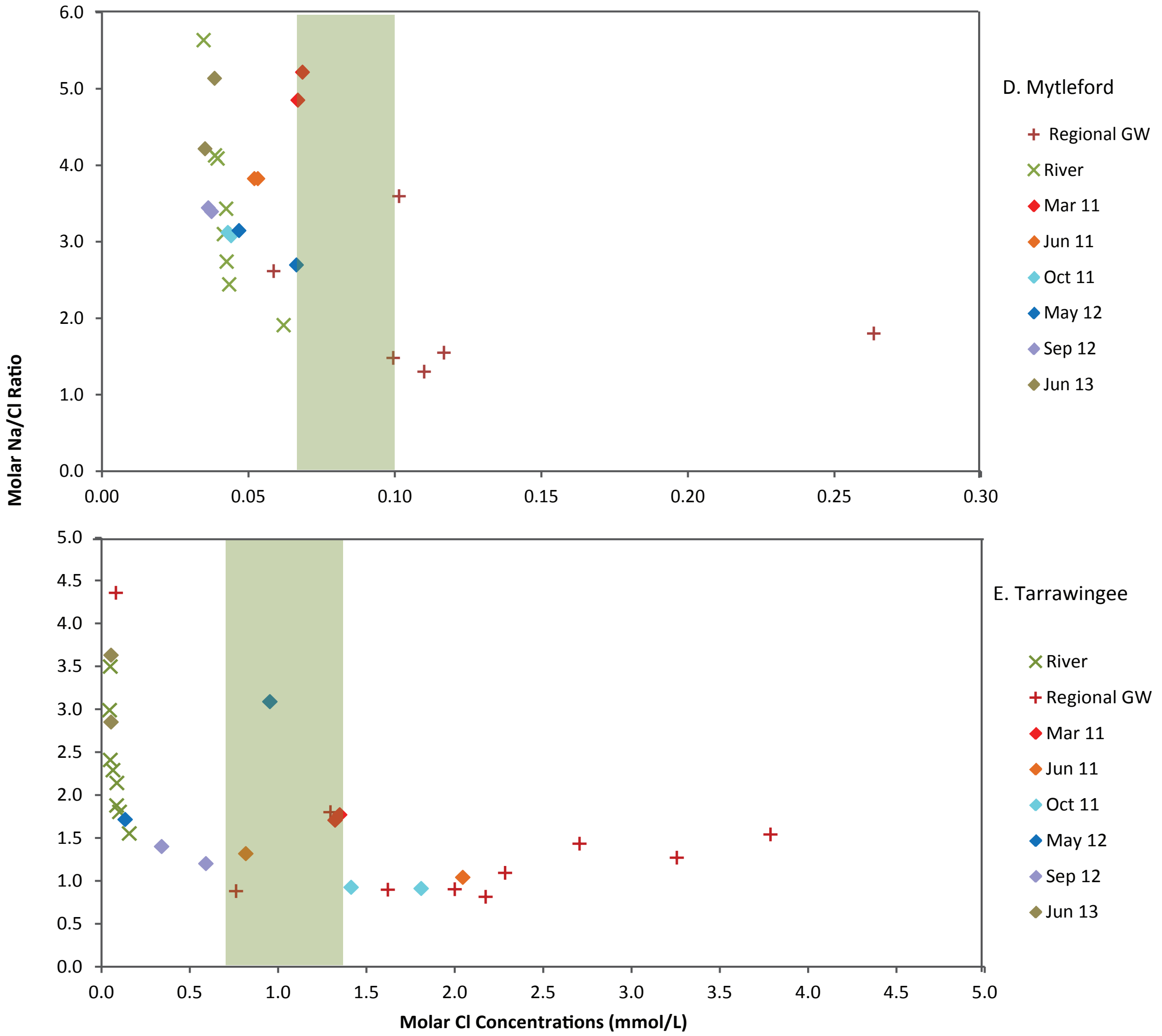

Figure 4.8 (cont.) Molar Na/Cl ratios of the groundwater at Harrietville (A), Smoko (B), Bright (C), Myrtleford (D) and at Tarrawingee (E). Ratios based on the Groundwater data from (Table 4.2 *to* 4.4), and the river data is from Chapter 2. River bank recharge is indicated when the Na/Cl ratios of the near-river and intermediate groundwater is close to that of the river, such as those at Harrietville (A), Smoko (B), Myrtleford (D) and at Tarrawingee (E). Some of the near-river or intermediate groundwater data lying in the intermediate range between the river water and regional groundwater indicate possible mixing of infiltrated river water with regional groundwater in the river banks (shown in light green).

groundwater in several sampling rounds (such as, June and Oct 2011) were usually below 1.3 but increased increase up to 3.6 after high flow events (such as May 2012 and June 2013) (Fig. 4.8e). Overall, the molar Na/Cl ratios of the near-river and intermediate groundwater at Harrietville, Smoko, Myrtleford and Tarrawingee are generally higher those of the regional groundwater and within, or close to, the ranges of the river water (3.0 *to* 8.0 for the upper and middle catchments and 1.5 *to* 3.5 for the lower catchment, Chapter 2). In comparison to Myrtleford, the Na/Cl ratios of the near-river and intermediate groundwater at Tarrawingee are more frequently close to those of the regional groundwater or in the intermediate range between the river water and the regional groundwater. The near-river and intermediate groundwater at Bright consistently has a molar Na/Cl ratio in the range of regional groundwater.

4.4.5 Stable isotopes

Although there are topographic and climatic differences between the upper and lower catchments, the $\delta^{18}O$ and $\delta^{2}H$ values of the groundwater in the Ovens Catchment define a single group that intersects the global meteoric water line near the average isotopic composition of precipitation in Melbourne (Fig. 4.9a). The $\delta^{18}O$ and $\delta^{2}H$ values of the groundwater in the Ovens Catchment lie to the left of the global meteoric water line and also the local meteoric water line for Melbourne; similar deviations are apparent in the stable isotope ratios of both groundwater and surface water from elsewhere in the Murray Basin and are probably caused by the local climatic differences between Melbourne (which is on the coast) and the inland Murray Basin (Cartwright *et al.*, 2010). The $\delta^{18}O$ and $\delta^{2}H$ values of the regional groundwater in the Ovens Catchment were -4.7 *to* -7.0‰ and -25 *to* -45‰, respectively, and these partially overlap the ranges of $\delta^{18}O$ and $\delta^{2}H$ for the water in the Ovens River ($\delta^{18}O$ = 6.0 *to* -7.5‰; $\delta^{2}H$ = -35 *to* -45‰, Chapter 2). The $\delta^{18}O$ and $\delta^{2}H$ values of most near-river and intermediate groundwater span across the ranges of stable isotopes of the river water and regional groundwater (Fig 4.9a). The $\delta^{18}O$ and $\delta^{2}H$ values of the near-river and intermediate groundwater at Tarrawingee after high flow events (such as May 12, Sept 12 & June 13) were less than -6.3‰ and -35.9‰, respectively. These $\delta^{18}O$ and $\delta^{2}H$ values were very similar to those of the river water at Tarrawingee (Fig. 4.9b).

4.4.6 Tritium

The ^{3}H concentrations of the regional groundwater at Harrietville were 0.18 *to* 2.27 TU,

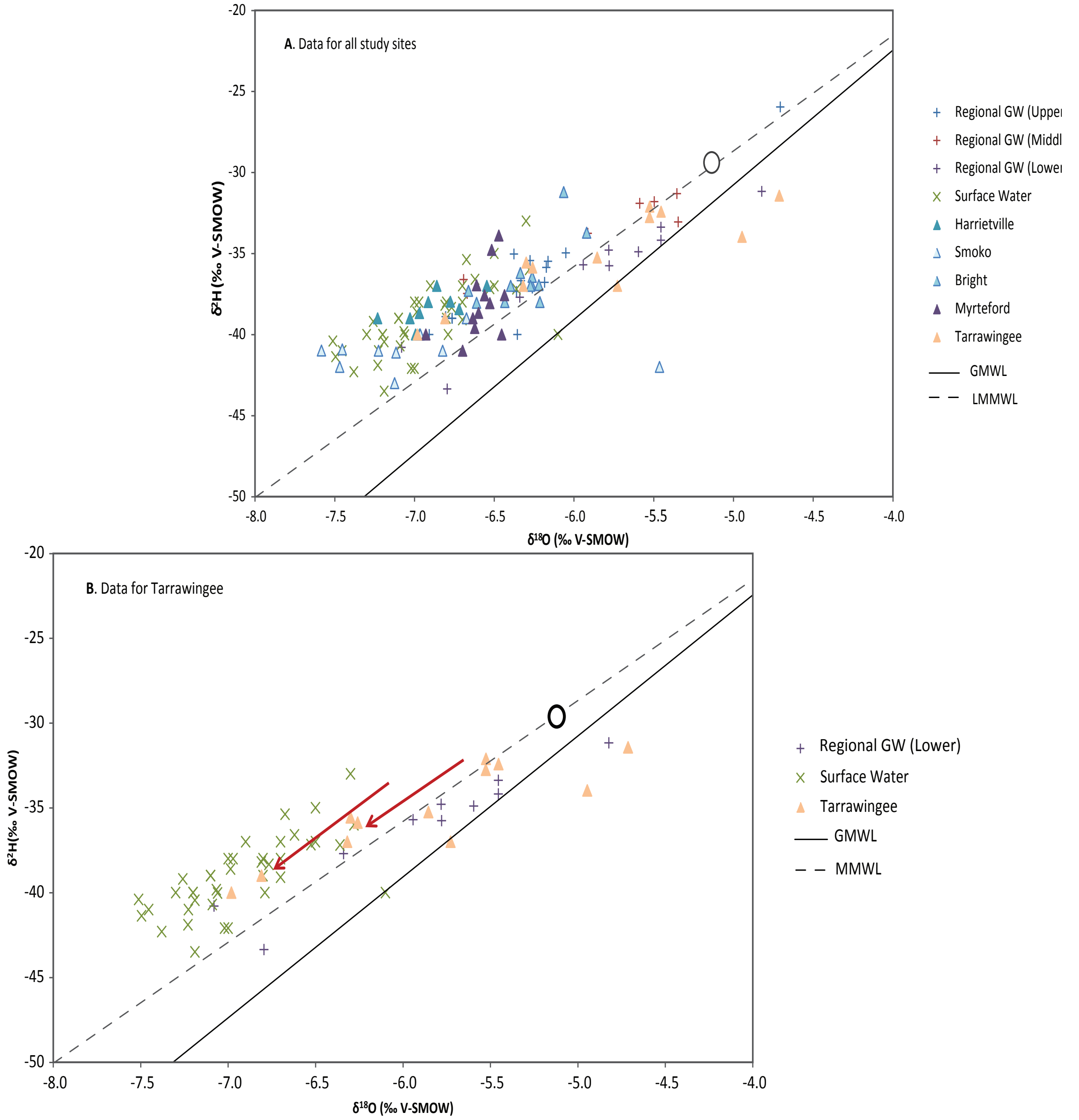

Figure 4.9 (A) $\delta^{18}O$ v. $\delta^{2}H$ values of the groundwater and river water in the Ovens Catchment. (B) Only the near-river, intermediate and regional groundwater at Tarrawingee and the river water are shown. The circle is the mean weighted average of rainfall in Melbourne. The $\delta^{18}O$ and $\delta^{2}H$ values of near-river and intermediate groundwater at Tarrawingee during some sampling rounds such as May 12, Sept 12 & June 13 decreased and were close to those of river (red arrows), implying bank recharge. It is difficult to use $\delta^{18}O$ and $\delta^{2}H$ distinguish bank infiltration at other river bank sites due to the lack of difference in $\delta^{18}O$ and $\delta^{2}H$ values between the river water and the regional groundwater (A). Data from Tables 4.5 *to* 4.7 with additional data from Cartwright & Morgenstern (2012) and Chapter 2.

while the regional groundwater in other locations in the upper catchment had ^{3}H concentrations of 0.071 *to* 3.62 TU (Cartwright and Morgenstern, 2012) (Table 4.5). In the middle and lower catchments, the ^{3}H concentrations of the regional groundwater were 1.70 *to* 2.13 TU and 0.6 *to*

0.2 TU, respectively (Tables 4.6 & 4.7) (Cartwright and Morgenstern, 2012). At Harrietville, the ^{3}H concentrations of the near-river and intermediate groundwater decreased with depth from 2.43 to 2.13 TU but increased with distance from 2.43 to 2.52 TU (Fig. 4.2a). At Smoko, the ^{3}H concentrations of the near-river and intermediate groundwater generally decreased vertically and laterally, from 2.41 to 2.27 TU and from 2.41 to 2.37 TU, respectively (Fig. 4.2b). At Bright, the ^{3}H concentrations of the near-river groundwater and immediate groundwater were 0.97 *to* 1.13 TU (Fig. 4.2c). At Myrtleford, the ^{3}H concentration of the near-river groundwater and immediate groundwater was 2.39 and 2.40 TU, respectively (Fig. 4.2d). At Tarrawingee, the ^{3}H concentrations of the near-river groundwater and immediate groundwater were between 2.26 and 2.32 TU (Fig. 4.2e).

The ^{3}H concentrations of the Ovens River in the upper catchment following several high flow events in September 2013 varied from 2.24 to 2.37 TU. The river ^{3}H concentrations were 2.27 *to* 2.28 TU during moderate discharge in December 2013, and 1.99 *to* 2.19 TU during very low flow conditions in February 2014 (Table 4.8). The ^{3}H concentrations of the river water generally decreased downstream, and this ^{3}H reduction trend probably indicates the input of older groundwater along the course river.

In summary, the groundwater ^{3}H concentrations generally increase with distance from the river at Harrietville, while they decrease with distance from the river at Smoko. For other locations, they are similar to each other within the river bank. The ^{3}H concentrations of the near-river groundwater and immediate groundwater at Harrietville, Smoko, Myrtleford and Tarrawingee are similar to those of the river water at high flows. Additionally, the ^{3}H concentrations of the near-river groundwater and immediate groundwater at Myrtleford and Tarrawingee are higher than those of the regional groundwater in their respective locations. The ^{3}H concentrations of the near-river groundwater and immediate groundwater at Bright, on the other hand, are lower than those

Date	Location	Bore No.	δ18O SMOW	δ2H SMOW	δ13C (PDB)	3H (TU)	3H Analytical uncertainty (±TU)	14C (pMC)	MRT (Year)
Regional Groundwater									
Sept 2009	Bright	B747	-6.1	-35	-20.5				
Sept 2009	Bright	B745	-6.2	-37	-24.4				
Sept 2009	Bright	B744	-4.7	-26	-20.9	3.62		99.9	4.5
Sept 2009	Bright	B737	-6.4	-35	-17.6	1.21		88.6	69
Sept 2009	Bright	B735	-6.3	-35	-19.5	1.39		99	56
Sept 2009	Bright	B736	-6.2	-36	-23.4	1.81		105.5	37
Sept 2009	Bright	B461	-6.3	-37	-12.8			89.9	
Sept 2009	Bright	B462	-6.2	-35	-15.4	1.42		94.1	55
May 2012	Harrietville	H301	-6.8	-39	-24.3	2.27	0.05	93.5	14
May 2012	Harrietville	H300	-6.8	-39	-26.5	2.41	0.05	93.8	12
May 2012	Harrietville	H299	-6.9	-40	-28.2	0.18	0.02	71.7	189
May 2012	Porepunkah	P289	-6.4	-40	-21.8	0.02		83.7	456
Near-river and Intermediate Groundwater									
Mar 2011	Bright	B1	-5.9	-34					
Mar 2011	Bright	B2	-6.1	-31					
Jun 2011	Bright	B1	-6.2	-37					
Jun 2011	Bright	B2	-6.3	-37					
Oct 2011	Bright	B1	-6.3	-36					
Oct 2011	Bright	B2	-6.3	-36					
May 2012	Harrietville	H306	-7.0	-39	-27.5	2.43	0.05	102.2	11
May 2012	Harrietville	H305	-6.5	-37	-26.7	2.13	0.05	99.2	17
May 2012	Harrietville	H302	-7.0	-40	-21.9	2.52	0.05	100.7	10
May 2012	Harrietville	H303	-6.9	-38	-26.8	1.93	0.04	92.8	24
May 2012	Harrietville	H304	-7.0	-40	-25.2	2.32	0.05	74.7	13
May 2012	Smoko	S296	-7.1	-43	-19.4	2.41	0.05	103.6	12
May 2012	Smoko	S297	-5.5	-42	-25.1	2.27	0.05	104.3	14
May 2012	Smoko	S293	-7.0	-40	-31.4	2.37	0.05	105.6	12
May 2012	Smoko	S294	-6.7	-39	-21.9	2.42	0.05	104.6	12
May 2012	Smoko	S295	-6.8	-41	-24.7	2.32	0.05	104.8	13
May 2012	Bright	B1	-6.2	-38	-25.8	1.14	0.03	95.9	63
May 2012	Bright	B2	-6.3	-37	-26.0	0.97	0.03	94.8	77
Sept 2012	Harrietville	H306	-7.2	-39					
Sept 2012	Harrietville	H305	-6.9	-37					
Sept 2012	Smoko	S296	-7.2	-41					
Sept 2012	Smoko	S297	-7.5	-42					
Sept 2012	Bright	B1	-6.4	-37					
Sept 2012	Bright	B2	-6.4	-38					
Jun 2013	Harrietville	H306	-6.8	-38					
Jun 2013	Harrietville	H305	-6.7	-38					
Jun 2013	Harrietville	H302	-7.0	-39					
Jun 2013	Smoko	S296	-7.6	-41					
Jun 2013	Smoko	S297	-7.1	-41					
Jun 2013	Smoko	S293	-7.5	-41					
Jun 2013	Bright	B1	-6.7	-37					
Jun 2013	Bright	B2	-6.6	-38					

Table 4.5 Stable and radioactive isotopes of the regional, near-river and bank groundwater in the upper catchment (Harrietville, Smoko and Bright), and mean resident times (MRT) of groundwater. Data for regional groundwater in Bright from Cartwright & Morgenstern (2012).

Date	Location	Bore No.	δ18O SMOW	δ2H SMOW	δ13C (PDB)	3H (TU)	3H Analytical uncertainty (±TU)	14C (pMC)	MRT (Year)
Regional Groundwater									
Sept 2009	Eurobin	E069	-5.9	-34	-15.9	1.7		76.1	41
Sept 2009	Eurobin	E068	-5.4	-31	-22.4	2.34		106.6	20
Sept 2009	Eurobin	E067	-5.5	-32	-21.9	2.69		104.2	13
Sept 2009	Eurobin	E066	-5.6	-32	-22.8	2.29		100.8	21
Sept 2009	Myrtleford	B232	-5.3	-33	-20.7	1.82		101.8	37
Sept 2009	Myrtleford	B231	-6.7	-37	-21.3	2.13		96.5	26
Near-river and Intermediate Groundwater									
Mar 2011	Myrtleford	M1	-6.5	-35					
Mar 2011	Myrtleford	M2	-6.5	-34					
Jun 2011	Myrtleford	M1	-6.6	-38					
Jun 2011	Myrtleford	M2	-6.5	-38					
Oct 2011	Myrtleford	M1	-6.6	-37					
Oct 2011	Myrtleford	M2	-6.6	-40					
May 2012	Myrtleford	M1	-6.6	-39	-27.1	2.39	0.05	100.4	12
May 2012	Myrtleford	M2	-6.5	-40	-29.3	2.40	0.05	100.7	12
Sept 2012	Myrtleford	M1	-6.7	-41					
Sept 2012	Myrtleford	M2	-6.9	-40					
Jun 2013	Myrtleford	M1	-6.4	-38					
Jun 2013	Myrtleford	M2	-6.6	-39					

Table 4.6 Stable and radioactive isotopes of the regional, near-river and bank groundwater in the upper catchment (Harrietville, Smoko and Bright), and mean resident times (MRT) of groundwater. Data for regional groundwater in Bright from Cartwright & Morgenstern (2012).

Location	3H (TU)	3H Analytical Uncertainty (±TU)	MRT (Year)	River Discharge (ML/Day)
Harrietville	2.37	0.02	13	2461
Bright	2.28	0.02	14	2461
Porepunkah	2.24	0.02	15	2461
Harrietville	2.27	0.04	14	806
Bright	2.28	0.04	14	806
Harrietville	2.19	0.05	15	91
Smoko	2.09	0.04	17	91
Bright	1.99	0.04	20	91

Table 4.8 Tritium concentrations of the Ovens River in the upper catchment, the discharge rate (measured at Bright) at the time of sampling, and MRTs of the river water. Discharge data from Water Measurement Information System (2014).

Date	Location	Bore No.	δ18O SMOW	δ2H SMOW	δ13C (PDB)	3H (TU)	3H Analytical uncertainty (±TU)	14C (pMC)	MRT (Year)
Regional Groundwater									
Sept 2009	Oxley	O738	-7.1	-41					
Sept 2009	Wangaratta	W326	-5.5	-34	-13.1	0.3		93.9	115
Sept 2009	Wangaratta	W493	-5.6	-35	16.7	0.02		75.3	172
Sept 2009	Boorhaman E.	BE2296	-5.9	-36	-10.2				
Sept 2009	Boorhaman E.	BE323	-6.8	-43	-15.4	0.94		91.2	110
Sept 2009	Boorhaman	BH788	-5.5	-33	-12.2	0.16		30.3	160
Sept 2009	Boorhaman	BH789	-4.8	-31	-14.6	0.037		88.8	88
Sept 2009	Peechelba	PE306	-6.3	-38	-15.1	0.26		103.3	111
Sept 2009	Bundalong S.	BS310	-5.8	-35	-14.2	0.49		95.9	104
Mar 2011	Tarrawingee	T3	-5.3	-31					
Jun 2011	Tarrawingee	T3	-4.6	-30					
Oct 2011	Tarrawingee	T3	-5.0	-32					
May 2012	Tarrawingee	T3	-4.6	-31	-17.4	2.32	0.05	102.3	15
Sept 2012	Tarrawingee	T3	-5.6	-35					
Jun 2013	Tarrawingee	T3	-4.1	-34					
Near-river and Intermediate Groundwater									
Mar 2011	Tarrawingee	T1	-5.5	-32					
Mar 2011	Tarrawingee	T2	-5.5	-33					
Jun 2011	Tarrawingee	T1	-5.9	-35					
Jun 2011	Tarrawingee	T2	-4.7	-31					
Oct 2011	Tarrawingee	T1	-5.5	-32					
Oct 2011	Tarrawingee	T2	-4.9	-34					
May 2012	Tarrawingee	T1	-6.3	-37	-20.8	2.29	0.05	105	14
May 2012	Tarrawingee	T2	-5.7	-37	-19.2	2.26	0.05	103.7	14
Sept 2012	Tarrawingee	T1	-7.0	-40					
Sept 2012	Tarrawingee	T2	-6.8	-39					
Jun 2013	Tarrawingee	T1	-6.3	-36					
Jun 2013	Tarrawingee	T2	-6.3	-36					

Table 4.7 Stable and radioactive isotopes of regional, near-river and bank groundwater at Tarrawingee in the lower catchment, and MRTs of groundwater. Data for regional groundwater, except for Tarrawingee, from Cartwright & Morgenstern (2012).

of the river water and regional groundwater at the location.

4.4.7 Carbon-14

The activities of ^{14}C of the regional groundwater at Harrietville and Porepunkah were 71.7 *to* 93.5 pMC, and 83.7 pMC, respectively. The activities of ^{14}C of the near-river and intermediate groundwater in the river banks along the Ovens River were between 92.8 and 105.65

pMC. The river bank that had low ^{14}C activities was Bright, ranging from 94.8 to 95.9 pMC. Some near-river or intermediate groundwater samples at Harrietville also had a ^{14}C of < 100 pMC, but they occurred at a depth of > 6 m. Much of the groundwater had ^{14}C activities greater than 100 pMC in the Ovens Catchment, meaning that it contains a component of water recharged during or after the bomb pulse in the 1950s and 1960s.

4.5 Discussion

This section defines the nature of river-river bank interactions at each location based on geochemistry and ^{3}H concentrations in conjunction with hydraulic heads, followed by discussing the groundwater residence times in the river banks. The discussion will be concluded by relating these river-river bank interactions to the catchment geomorphology.

4.5.1 River-river bank interactions

At Harrietville, the observations that hydraulic gradients are generally downward and away from the river imply that the river is losing with river water infiltrating into the river banks at all times. The mixing between low salinity river water and the regional groundwater has led to the trend of increasing EC and Cl concentrations with depth (Figs. 4.4aii & 4.7a). Furthermore, bank infiltration during high flow conditions results in lower EC in the river banks when low salinity surface runoff in the river is most prevalent. Bank recharge also explains the observation that the molar Na/Cl ratios of near-river groundwater are generally higher than those of the regional groundwater but similar to those of the river water (Fig. 4.8a). It also explains the similarity between the near-river groundwater ^{3}H concentrations and the river ^{3}H concentrations (Fig. 4.2a). That the river is losing at Harrietville is consistent with the low radon (^{222}Rn) activities in the Ovens River in this area (Chapter 2). The intermediate groundwater at Harrietville has higher Cl concentrations and higher ^{3}H concentrations but lower molar Na/Cl ratios (Figs. 4.7, 4.2 &

4.8), and this observation implies that the near-river groundwater flows further into the aquifer and mixes with the recently recharged regional groundwater that has higher Cl concentrations, lower molar Na/Cl ratios and slightly higher ^{3}H concentrations. Since February 2013, the vertical head gradient in the river bank had been reversed with a change in the vertical trends of the EC values and Cl concentrations (from increasing with depth to decreasing with depth) (Figs. 4.4a & 4.7). The change in flow regime occurred immediately after the short high flow event in late February 2013. A bushfire took place in the headwaters at the Ovens River in January 2013, and the subsequent flow events in late February 2013 washed down ashes and sediments downstream. It is likely that these ashes and sediments increased the thickness of the clogging layer, resulting the observed change in the flow regime at the river banks. Consistent with this hypothesis, the heads in the bores in this region dropped during this event which would be expected since there was diminished recharge of the groundwater by the river.

At Smoko, the dominant upward head gradient implies that the overall groundwater flow is upward, and groundwater is discharged at the river banks during low flow periods. The pattern of increase groundwater Cl concentrations toward the river channel in the river bank during low flow condition further supports this flow regime as groundwater mixes with the saline regional groundwater nearby while flowing toward the river (Fig. 4.7b). The fact that elevated ^{222}Rn activities were observed in the Oven River at this locality also supports the conclusion that the river is gaining (Chapter 2). Although the reach is gaining, the observed vertical head gradient reversal between July and September 2012 indicates that river water can infiltrate into the bank (Fig. 4.4biii). Bank infiltration is supported by the observations that the EC values (particularly the EC of the groundwater at medium depth) and Cl concentrations in the near-river groundwater were significantly lowered following high flow events in those periods (Figs. 4.4bii & 4.7b). The similarity in the molar Na/Cl ratios between the near-river and intermediate groundwater, and the river water further suggests bank recharge (Fig. 4.8b). Bank recharge has led to the high ^{3}H concentrations in the river bank (Fig. 4.2b).

At Bright, the hydraulic gradient is always toward the river and is seldom reversed (Fig. 4.4civ). This observation indicates that the river banks at this location are not recharged by the river. Without bank recharge, the observed EC values and the Cl concentrations of the near-river and intermediate groundwater are relatively constant and do not decrease in response to the rising river stage (Figs 4.4cii & 4.7c). The lack of bank recharge has also led to the low groundwater ^{3}H concentrations (in comparison to the river ^{3}H concentrations) (Fig. 4.2c) and a ^{14}C activity of less 100 pMC within the river bank. Finally, the observation that the Na/Cl ratios of the near-river and intermediate groundwater are close to those of the regional groundwater implies that the groundwater in the banks originates from the regional groundwater rather than the river via bank recharge (Fig. 4.8c).

At Myrtleford, the observation that the hydraulic gradient is toward the river with increasing EC values toward the river indicates that groundwater flows toward the river and is discharged at the river banks (Figs. 4.5aiv & 4.5aii). Gaining conditions are also implied by the high ^{222}Rn activities in the river (Chapter 2). Despite the generally gaining conditions, the river banks are periodically recharged by the river as demonstrated by the hydraulic gradient reversals in the bank during high flow conditions (Fig. 4.5iv). Bank recharge has also led to the ^{3}H concentrations of the near-river and intermediate groundwater being higher than those of regional groundwater at this location (Fig. 4.2d). The reduction in salinity of the near-river and immediate groundwater (EC values and Cl concentrations) and the higher molar Na/Cl ratios of the near-river and immediate groundwater further indicate bank recharge (Figs. 4.5aii, 4.7d & 4.8d). The observation that the variation in the EC values of the immediate groundwater is smaller than that of near-river groundwater indicates the lateral limitation of bank infiltration with a distance of only

10 *to* 50 m from the river channel. The Na/Cl ratios of the near-river and intermediate groundwater in March 2011 and May 2012 decrease with increasing Cl concentration (Fig. 4.8d). This change in the Na/Cl ratios may represent mixing between the recently recharged near-river groundwater water with a high Na/Cl ratio and a low Cl concentration, and the regional groundwater with a low Na/Cl ratio and a high Cl concentration. Only these two sampling rounds show such a mixing trend probably because these two rounds took placed three to two months after a major flood event with the infiltrated water starting to mix the regional groundwater. Moreover, the rapid raising river stage associated with the high flow event was likely to create a high seepage velocity in the bank, resulting in strong dispersion which leads to greater mixing (McCallum *et al.*, 2010).

At Tarrawingee, the hydraulic gradient indicates that during low flow conditions the groundwater flows toward the river (Fig. 4.5biv). Bank recharge is also evident because of the reversed hydraulic gradient, and the decrease in the EC values and Cl concentrations following some of the high flow events (Figs. 4.5biv, 4.5bii & 4.7e). Bank infiltration can occur as far as 50 m away from the river channel as demonstrated by the reduced salinity at Bore T3 in May 2013. Besides the changes in the EC values and Cl concentrations, bank infiltration led to a shift in the major ions composition, the molar Na/Cl ratios and the $\delta^{18}O$ and $\delta^{2}H$ values of the near-river, intermediate and regional (at 50m away from the river channel) groundwater from those of the regional groundwater toward those of the river water (Figs. 4.6b, 4.8e & 4.9b). In comparison to the Myrtleford site, the Na/Cl ratios of near-river and intermediate groundwater at Tarrawingee are more frequently close to that of the regional groundwater or in the intermediate range between that of the river water and that of the regional groundwater (Fig. 4.8e). This observation suggests that bank recharge may be less frequent Tarrawingee than at Myrtleford probably due to the lower conductivity of bank sediments at Tarrawingee (22 m day^{-1} at Myrtleford vs. 11 m day^{-1} at Tarrawingee). However, the low conductivity of bank sediments decreases the rate of bank return flow following flow events (Whiting and Pomeranets, 1997; Chen and Chen, 2003). This may partly explain the slow return of EC in the river bank after some flow events (Fig 4.5bii).

4.5.2 Groundwater residence times in river banks

The residence time of groundwater in the river banks was calculated from the ^{3}H concentrations (Tables 4.5 *to* 4.7) using TracerLPM (Jurgens et al., 2012), which is a programmed spreadsheet for evaluating groundwater age from environmental tracer data based on the LMPs (Eq. 4.1). An exponential piston-flow model (Eq. 4.2) was used in the calculations as it is appropriate for unconfined to semi-confined aquifers where the bores are screened below the water table (Maloszewski and Zuber, 1982; Morgenstern *et al.*, 2010; Cartwright and Morgenstern, 2012). The calculation of residence times using the LMPs requires two parameters: the ^{3}H concentration in precipitation and the EPM ratio (Eq. 4.1). The pre-1955 ^{3}H concentration is assumed to be that of the modern precipitation in the Ovens catchment. The ^{3}H concentrations of precipitation immediately following the bomb pulse and in the subsequent years are based on the mean weighted average of ^{3}H concentrations of precipitation in Melbourne (International Atomic Energy Agency, 2013). Finally, the ^{3}H concentration of the modern precipitation in the Ovens catchment is estimated to be 4.0 TU, which is higher than the observed values in Melbourne (2.8 *to* 3.2 TU) (International Atomic Energy Agency, 2013; Tadros *et al.*, 2014). This estimate is based on the evidence that the ^{3}H concentrations of groundwater in the catchment are as high as 3.7 TU (Cartwright and Morgenstern, 2012), and that there is a difference in the distribution of rainfall ^{3}H concentrations between inland and coastal catchments (Tadros *et al.*, 2014). The EPM ratio is set at 0.3 which is based on aquifer lithology, bore depths and widths of bore screens (Cartwright and Morgenstern, 2012).

The calculated mean residence times of near-river and intermediate groundwater in the Ovens Catchment were between 10 and 77 years (Fig. 2 & Tables 4.5 *to* 4.7). Analytical uncertainty of ^3H produces an uncertainty of ±1 *to* 3 years with a greater impact on the older groundwater with a residence time of great than 67 year. Reassigning the ^3H concentration for the modern precipitation to 3.2 TU (closer to the observed values in Melbourne) reduces the residence times by 6 *to* 8 years but has minimal impacts on the older groundwater. Varying the EPM ratio between 0.1 and 0.3 results in a residence time difference of ± 1 *to* 2 years for all the groundwater samples regardless of their residence time. Groundwater with the oldest mean residence time (63 and 77 years) occurred at Bright. As indicated earlier, the river bank at Bright is rarely recharged by the river and therefore contains older groundwater. The recently recharged near-river and intermediate groundwater at shallow and medium depths had a residence time of less than 17 years old, and that is above or similar to the residence time of the river water measured at high and moderate flows (13 *to* 15 years for the river water at high and moderate flows, measured in the upper catchment). This observation further suggests that these river banks are recharged over the high flow events during which the river water comprises of a higher amount of surface runoff with a lower residence time and a higher ^3H concentration. As the residence times of the recently recharged near-river groundwater in river banks are linked to the residence time of the river water, they are affected by the release of older water from the catchment storages in the headwaters and upper catchment. Therefore, the calculated ages do not just reflect the residence time in the river bank but the sum of residence times in the bank and in other catchment storages upstream prior to bank infiltration. Finally, the model choice of the LMPs, and the assignment of values for the ^3H concentrations in precipitation and the EPM ratio can contribute uncertainties in the calculated mean residence times. For example, the ages of river water at high and moderate flows could be much lower than the calculated ages (13 *to* 15 years) if the values for the ^3H concentrations in precipitation are much lower than the assumed value. However, these uncertainties have little effects on establishing relative residence times for the groundwater at the river bank sites since older water always have lower ^3H concentrations than younger water. It is the relative differences in the residence times (or ^3H concentrations) of near-river and intermediate groundwater that allow the identification of bank recharge, rather than being based on the absolute residence times (or ^3H concentrations) of the near-river and intermediate groundwater.

4.5.3 Controls of river bank recharge

The river banks at Harrietville, Smoko, Myrtleford and Tarrawingee may be recharged by the river at high flows, while the river bank at Bright is not recharged by the river even following very high river stages. At Harrietville, the river is located near the valley margin with a relatively higher elevation in relation to the rest of valley alluvial plain (Fig. 4.2a). Additionally, the river banks at Harrietville are not incised steeply and are permeable. These combined factors are unlikely to produce or maintain a high water table in relation to the river elevation, resulting in recharge to the aquifer through the river bank by the river. At Smoko, the break of slope at the adjacent valley cliff produces an upward hydraulic gradient, causing groundwater discharge at bank and at the base of the river (Fig. 4.2b). During high flow conditions, this upward gradient can be reversed, resulting in recharge to the bank and the underlying aquifer. At Bright, the river runs through the centre of valley with a steep lateral head gradient across the valley (Fig. 4.2c). In this part of the steep valley, a strong regional hydraulic gradient develops towards the river, preventing the reversal of head gradients in the river banks. Consequently, there is little bank recharge even during high flow conditions. At Myrtleford and Tarrawingee, bank recharge occurs because the reduced regional hydraulic gradient in the broader valley and in the floodplain can be reversed at a high river stage (Fig. 4.2d & e). Although the regional hydraulic gradient decreases in the broader valley of the middle catchment, it is still relatively high in relation to the river. Consequently, the lateral bank infiltration at Myrtleford in the middle catchment is very limited (10 *to* 50 m from the river), whereas the lateral bank infiltration can occur as far as 50 m away from the river channel at Tarrawingee in the lower catchment. The bank sediments at Myrtleford are coarser than at Tarrawingee. The infiltration and exfiltration rates, the volume of bank storage and the period over which the river water is stored in the bank are likely to differ between the two sites. The coarse sediments at Myrtleford in the middle catchment increase the frequent infiltration into the riverbank but also promote the return of the stored bank water into the river following the high flow events. The finer sediments at Tarrawingee in the lower catchment may reduce the frequency of bank infiltration, but the finer sediments combined with the lower regional hydraulic gradient together slow down the process of bank return.

4.6 Conclusions

A combination of geochemistry, ^{3}H and hydraulic heads has revealed that most river banks in the Ovens catchment are either continuously or periodically recharged by the river. The only exception is Bright (the mid to lower section of the upper catchment) where the river bank contains old regional groundwater and is rarely recharged by the river. The spatial variation in river bank recharge is related to the catchment topography and river bank lithology. Infiltration from the river can extend at least 10 m *to* 50 m in the middle catchment and possibly 50 m away from the river in the lower catchment. Therefore, the bank storage is limited in the narrow and broader valley in the upper and middle catchments. One management implication of bank storage is to attenuate flooding during high flow events. Since the narrow and broader valleys are shown to have limited bank storage, it is important to develop strategies or build infrastructures to reduce damages from possible flooding in these areas and the areas downstream. Flooding has caused severe damages to townships like Myrtleford in the middle catchment in the past decade (Goulburn Broken Catchment Management Authority, 2016).

High rainfall over winter months in the alpine regions combined with the large rainfall events in summers (such as those caused by La Niña in 2010/11 and 2011/12) promotes bank infiltration in the Ovens Catchment. Bank infiltration often provides a high quality of groundwater source along the river. However, for catchments in an alpine region (in the case of the upper and middle Ovens Catchment), it is relatively less important since direct groundwater recharge for these areas is high. Furthermore, the steep hydraulic gradient in alpine valleys prevents an extensive lateral infiltration of river water, restricting the fresh infiltrated water to small areas along the river channel. Bank recharge is, however, more important for arid/semi-arid catchments (for example, the lower Ovens Catchment) where the recharge rate on the floodplain is much lower, and the hydraulic gradient toward the river in the floodplain is low. Another implication of river-river bank interactions is that low saline baseflow derived from these banks in the Ovens Catchment may play important role in reducing the salt load in the salt affected Murray River downstream. The results and findings of this geochemistry-based field study are complimentary to the existing numerical and analytical studies of river-river bank interaction, assisting in understanding bank storage in the context of overall surface water-groundwater interaction.

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Supplement

Table S4.1 Details of bores located in the regional groundwater

Site	Catchment	Bore ID.	Easting	Northing	Screen Depth (m)b	Distance to River (m)
Bright	Upper	B747	499190	5935414	2-20	160
Bright	Upper	B745	499139	5935375	5-11	225
Bright	Upper	B744	498933	5934911	6-12	725
Bright	Upper	B737	498445	5935658	36-42	261
Bright	Upper	B735	498391	5935314	30-42	70.8
Bright	Upper	B736	498382	5935299	20-26	82.4
Bright	Upper	B461	497818	5935267	20-26	339
Bright	Upper	B462	497818	5935267	45-51	339
Harrietville	Upper	H301	505634	5918160	9-11	60
Harrietville	Upper	H300	505760	5918148	9-11	260
Harrietville	Upper	H299	505758	5918149	23-25	260
Porepunkah	Upper	P289	493292	5938063	53-59	
Eurobin	Middle	E069	487803	5944698	5-8	506
Eurobin	Middle	E068	487657	5944643	7-13	357
Eurobin	Middle	E067	487519	5944594	12.0	203
Eurobin	Middle	E066	487411	5944553	9-15	91.4
Myrtleford	Middle	B232	474884	5953288	6-12	447
Myrtleford	Middle	B231	474704	5953010	8-14	126
Oxley	Lower	O738	444240	5966742	19-44	
Wangaratta	Lower	W326	439879	5982755	23.7	2790
Wangaratta	Lower	W493	439422	5982189	16.5	2130
Boorhaman E.	Lower	BE2296	437925	5992950	71-77	3520
Boorhaman E.	Lower	BE323	437924	5992953	17.4	3520
Boorhaman	Lower	BH788	442072	5999081	60-72	9930
Boorhaman	Lower	BH789	442072	5999081	18-30	9930
Peechelba	Lower	PE306	432684	5994603	16	469
Bundalong S.	Lower	BS310	427237	6005560	14	191

Chapter 5

Conclusions

Rivers interact with subsurface water stores at various degrees in various flow conditions in the water cycle. In order to manage these connected water resources effectively and efficiently, the interactions between these water components need to be understood thoroughly. This thesis presented three studies that investigated different aspects of the interactions between surface water and groundwater at the Ovens River in southeast Australia using a combination of physical and chemical hydrological data.

5.1 Summary of the studies

In chapter 2, the spatio-temporal variation of groundwater inflows to the Ovens River was defined and quantified by using river ^{222}Rn activities. The study showed that the Ovens River is dominantly gaining in the upper catchment and fluctuates between gaining and losing in the lower catchment. The middle catchment represents a transition in river-aquifer interactions from upper to lower catchment. The distribution of gaining and losing reaches in the Ovens Catchment is governed by distribution of rainfall, topography and aquifer lithology. In the upper catchment, frequent groundwater recharge occurs through the permeable aquifers in the narrow valley, raising the water table. This process, referred to as hydraulic loading, increases the hydraulic head gradients toward the river. As a result, higher and variable groundwater inflows occur in the upper catchment, especially during high flow events in winters when most precipitation occurs. In the lower catchment, the lower rainfall and the fine-gained sediments reduce both the magnitude and variability of the hydraulic gradient between the aquifer and the river, producing lower and relatively constant groundwater inflows during low periods. Since the water table in the lower catchment has a shallow gradient toward the river, small changes in river height can result in losing behaviour in the lower catchment during high flow conditions. The study, based on the ^{222}Rn mass balance, estimated groundwater inflows in the Ovens River to be 2 *to* 17% of the annual discharge, which are lower than those from Cl-based chemical mass balance, hydrograph separation methods or differential flow gauging.

In chapter 3, flow duration Curves (FDC), graphical and filter-based hydrograph separation, and Cl-based chemical mass balance (CMB) were used to estimate and constrain the long-term baseflow flux near in the lower reaches of the Ovens Catchment. Between 2004 and 2014, the net baseflow contribution to the Ovens Catchment calculated using the FDC, local minima method from the graphical hydrograph separation, recursive digital filters from the filter-based hydrograph separation and chloride mass balance was 13%, 59%, 46% and 9.5% of the total discharge, respectively. Thus, baseflow fluxes estimated by the hydrograph separation techniques are significantly greater than those by the Cl-based CMB and FDC with a difference percentage of between 35% and 200%. Furthermore, the difference in the baseflow estimates between these techniques is larger during and following high flow events. These differences are interpreted as indicating that discharge from transient water stores, such as river banks, the unsaturated zone, and pools or disconnected channels on floodplains, is a significant proportion of the baseflow in the Ovens catchment. Discharge from these transient water stores displays the long wavelength variability of groundwater inflows and thus is coupled with the groundwater inflow in the hydrograph separation techniques. In contrast, discharge from these water stores has a low salinity and is thus not grouped with the groundwater inflows in the CMB. Likewise, the FDC yields the total groundwater inflow probably because it estimates baseflow flux by using the 90th percentile streamflow during which the baseflow is almost exclusively groundwater with minimum influx from the depleted transient storage. As discharge from these transient water stores occur mainly during and following high flow events, this causes the observed larger difference in baseflow estimates during high flow conditions. The contribution of transient water stores in the river is further supported by the persistent low Cl concentrations in the river following flood peaks.

In chapter 4, the importance of bank storage and bank return flows in the Ovens catchment was defined by using hydraulic heads, geochemistry (including EC, major ion chemistry and stable isotopes) and tritium. The study indicated that the Ovens River at Harrietville in the upper catchment is continuously recharged by the river (i.e. is a losing reach), and that most of the river banks in the middle and lower catchments are periodically recharged by the river. The indications of bank recharge in these locations are the reversed hydraulic heads in banks, the reduction in EC in banks, the shift in the Na/Cl ratios, major ion composition and stable isotopic signature of the near-river groundwater toward those of the river water during following high flow events, and the similarity in ^{3}H concentrations between the water in the river banks and the river water at moderate and high flow conditions. Significant bank storage occurs in areas that have relatively lower regional hydraulic gradients toward the river and that contain river bank sediments with higher hydraulic conductivities. These conditions prompt bank infiltration during high flow conditions. The area that has no bank recharge is the steep mid-to-lower valley section of the upper catchment. The high water table gradient in this area prevents the reversal of hydraulic gradient in river banks in this area that would normally occur during high flow conditions, resulting in no recharge in these river banks.

In summary, chapter 2 indicated that groundwater inflows are low with 2 *to* 17% of the annual discharge. However, chapter 3 suggested that other components of baseflow, such bank return flow drainage of pools or disconnected channels on floodplains and influx from the unsaturated zone, contribute to the flow of the Ovens River significantly during low flows conditions (probably up to 49% of the total discharge based on the difference in baseflow estimates between the numerical techniques and the chemical mass balance). As indicated in chapter 4, bank storage is locally important in the middle and lower catchment of the Ovens River.

5.2 Management implications of river-groundwater interaction for the Ovens Catchment

The location of all the studies was in the Ovens Catchment which is part of the socially, economically and ecologically important Murray Darling Basin. The results from these studies will help to understand the interactions between the river and its surrounding sub-surface water stores in the catchment and to appreciate the possible implications of such interactions on the water resources in the Ovens Catchment and on the Murray River downstream.

The studies indicated that although the Ovens River is hydraulically connected to its underlying surface aquifers, groundwater influx from these aquifers is relatively minor, varying from 2 to 17% of the total flow. One possible reason for the low groundwater inflow from regional aquifers could be due to the fact that the regional groundwater in the Ovens Catchment flows parallel to the river along the valley rather than toward the river (Victorian Government Department of Sustainability and Environment, 2010; Water Measurement Information System

2013). Groundwater probably flows down the Ovens valleys and is discharged to the Murry River along the northern boundary of the Ovens Catchment. The low groundwater inflows imply that in additional to groundwater, the Ovens River equally and probably more relies on discharge from transient water stores to maintain flow between rainfall events. As suggested in the study, river banks in the middle and lower catchments are one of the transient water stores. Therefore, it is vital to protect the near-river groundwater in the Ovens Catchment from pollution, salinization and over-allocation. As the scale of impacts of near-river groundwater and regional groundwater on the Ovens River are different, it is highly advisable to assess and manage these groundwater zones independently based on the degree of interactions with the river. The division of groundwater resources into zones based on the degree of their connectivity with surface water for management is actively advocated in Australia (Evans and Merz, 2007). The studies also showed that bank storage does not occur in the narrow valleys and is limited in the broader valleys in the Ovens Catchment. Since bank storage attenuates flooding during high flow events, it is important to develop strategies to reduce the negative impacts of possible flooding in these areas.

5.3 Wider implications of river-groundwater interaction

The knowledge gained from the three studies in this thesis is not only applicable to the Ovens Catchment but also to other catchments, having implications for understanding the process of river-groundwater interactions and managing water resources in catchments locally and globally.

5.3.1 Understanding and investigating river-groundwater interaction

Some river-groundwater interaction studies in the past have mainly focused on examining the changes in river height while treating hydraulic heads in adjacent aquifers as relatively constant. Assuming hydraulic heads as constant may be reasonable for catchments with fine-grained aquifers or for studying interactions between surface water and deep groundwater but may not be so for catchments with highly transmissive aquifers as illustrated in chapter 2. It is thus important for any studies to examine fluctuations in both river height and water table for determining the degree of river-groundwater interactions.

Using hydrograph separation methods is often the first choice of method in river-groundwater interaction studies because of the easy access to discharge data and the low cost associated with the methods (Brodie *et al.*, 2007; Cook *et al.*, 2010). As shown in chapter 3, the results from hydrograph separation methods provide valuable information on how baseflow varies in the catchment at various flow conditions over a long period of time. At the same time, the chapter also concluded that hydrograph separation methods produce much higher baseflow fluxes when compared to other techniques. Therefore, hydrograph separation methods should not be used alone for catchments with a large volume of transient water stores in order to avoid overestimating groundwater inflow. In addition, the choice of methods for constraining baseflow should depend on the focus of the study. For example, if the aim of a study is to examine the impact of groundwater extraction on nearby rivers, it is more appropriate to derive the total groundwater inflow using chemical mass balance rather than the total baseflow using hydrography separation. Chapters 2 and 3 illustrate that using multi-geochemistry tracers and/or multi-techniques is the best approach to constraint and discern various components of baseflow in rivers.

Chapter 2 demonstrated the usefulness of longitudinal stream chemical sampling technique in depicting the variation of baseflow flux along a river. Longitudinal stream chemical sampling is particularly useful in determining groundwater inflow hotspots, and the chemistry and residence time of groundwater discharge (Cook, 2013). Understand the variation of baseflow flux along rivers is vital if the practice of making water allocation at river reach scale is to continue. Many geochemical tracers, including major ions, stable isotopes and radiogenic isotopes can be used in many chemical sampling techniques. The choice of geochemical tracers should be based on the relative concentrations of the tracers in surface water and groundwater, the behaviour of the tracer through the water cycle, and the ability to quantifying the tracer's concentration in the groundwater end-member. As seen in chapter 2, it was difficult to use Cl in qualifying groundwater inflow in the upper catchment of Ovens Catchment where groundwater and river water have similar Cl concentrations. ^{222}Rn, on the other hand, is shown to be a good tracer of groundwater inflow, particularly in alpine areas where recently recharged groundwater and river water have similar major ion chemistry and stable isotopic signature. The accuracy of ^{222}Rn-derived baseflow fluxes can be hampered by issues such as heterogeneity of ^{222}Rn in groundwater, rate of ^{222}Rn degassing and hyporheic exchange. Though hyporheic exchange has a very limited impact on quantifying the cumulative groundwater inflow in a catchment because, as shown in chapter 2, reaches that are greatly affected by hyporheic exchange usually have low groundwater inflow, contributing only a small proportion of the total baseflow in the whole catchment. Furthermore, as more studies on refining ^{222}Rn as a groundwater tracer are carried out, quantitating groundwater discharge with ^{222}Rn will be more reliable. Regardless, ^{222}Rn is a good natural environment tracer for defining groundwater discharge areas in surface water systems.

5.3.2 Water resources and riverine environment management

The three studies in this thesis re-affirmed that the interaction of surface water and subsurface water stores vary spatially and occurs at different scale, from reach to catchment scales (Winter *et al.*, 1998; Braaten and Gate, 2003; Guggenmos *et al.*, 2011). As a result, the quantity and the source of baseflow can be different from one reach to the other within a catchment. Therefore, it is important to continue the current practice that environmental flow and water allocation are assessed, determined and reviewed at the reach-scale. The different proportions of various baseflow components in the Ovens River indicated by chapters 2 and 3 implies that different subsurface water resources can interact with rivers at various degrees. The near-river groundwater such as bank storage and alluvial aquifers is likely to interact with the Ovens River more regularly than the regional groundwater. Over-allocation of these groundwater resources will thus have a negative effect on the river at various degrees and at different rates. These sub-surface water resources should be assessed and managed independently while being seen as parts of the connected water resource in a catchment.

While studying river-groundwater interactions and managing water resources at river reach scale are important, it is equally important to appreciate these interactions at a catchment scale. The knowledge gained from chapters 2 and 3 provides a valuable overview of river-groundwater interactions in the whole Ovens catchment. This knowledge helps us in better understanding the water balance and salt load within catchment for developing a more comprehensive and balanced water policy. Water balance in catchments cannot be accurately estimated unless the knowledge of how multiple and contiguous river reaches of an entire river system interact with subsurface water stores is known (CSIRO, 2008; Bank *et al.*, 2011). Groundwater in semi-arid regions, such as Australia, often has high concentrations of dissolved solutes due to evaporation in a semi-arid climate, high transpiration rates of the native vegetation and recharge from poorly drained saline lakes (Herczeg *et al.*, 2001; Cartwright *et al.*, 2004). Input of such saline groundwater can have negative impacts on the quality of rivers nearby. The knowledge of the river-groundwater interaction at a regional scale assists in understanding how solutes may be mobilised between the two systems within a catchment for developing preventative strategies (Barton *et al.*, 2006). Reviewing water management policies that are made at river reach scale in the context of river-aquifers interaction at a catchment scale will reduce the negative impacts of fragmentation of water resources along the river (Braaten and Gate, 2003; CSIRO 2008).

The surface water-groundwater interaction has been classified into categories (gaining versus losing systems) for the purpose of conceptualisation (Winter 1998). Such categorization may create a notation that surface water-groundwater interaction is static and encourages the allocation of water to be based on whether a particular river is classified as gaining or losing. Yet, river- groundwater interaction, as indicated in chapter 2, can vary temporally, from during a high flow event to throughout a seasonal cycle. It is crucial to incorporate the issue of temporal variability in river-groundwater interaction in determining water allocation along the river. Water allocation for consumption purpose and environment needs to be flexible and adjustable in response to seasonal changes (Victorian Government Department of Sustainability and Environment, 2004; Lovell, 2009).

Despite of the common assertion that groundwater is the majority component of baseflow in rivers (American Ground Water Trust, 2003; Gordon *et al.*, 2004; Leap, 2007), chapter 3 showed that groundwater inflow only makes up a minor proportion of baseflow with the majority of baseflow being derived from transient water stores. Other catchments in southeast Australia were also reported to have a groundwater inflow of only between 10 and 30% of the total discharge (Unland *et al.*, 2013; Cartwright *et al.*, 2014; Atkinson *et al.*, 2015). The low groundwater inflows imply that the sustainability of river flow depends on not only groundwater but also transient water stores. As highlighted in chapters 3 and 4, near-river groundwater in the banks can be an important water store. Over allocation of near-river groundwater in these river banks can therefore reduce the ability of a river to maintain flow between high flow events. Transient water stores may not always have considered as a practical water resource because of their relatively short storage times, but the importance to environment flows should be not ignored.

It is important to differentiate groundwater inflow from inflows from transient water stores in baseflow estimation because each of them provides information on different hydrological processes in a catchment and have different implications for environmental management.

Understanding groundwater inflow in rivers assists in appreciating the connectivity between rivers and its adjacent aquifers and assessing the impact of groundwater extraction on nearby rivers. On the other hand, understanding discharge from transient water stores, particularly bank storage, is necessary in order to understand flooding in catchments. Such understanding is vital in protecting rivers and surrounding floodplains (that is for maintaining sufficient overbank flow to floodplain wetlands while preventing major flooding).

The chapter 2 highlighted that the highest groundwater inflow occurs days to weeks following a heavy rainfall because hydraulic loading increases the hydraulic gradient between the rising regional water table and the receding river level. On the other hand, groundwater constitutes the highest proportion of the river flow during baseflow conditions. This observation have implications for water pollution and salt load management in rivers. High groundwater influx during high flow implies that if the groundwater is for example saline, the salt load in the river will increase during high flow events due to the increase in saline groundwater inflow through hydraulic loading. This increase is in addition to the increase in salt load caused by the flushing associated with the high flow events. Therefore, the management of salt load or pollutants in rivers from groundwater inflow needs to be addressed at the time of high flow events as well as at baseflow conditions.

5.4 Final Remarks

In conclusion, the studies covered in this thesis have raised and re-emphasized several important points in understanding river-groundwater interactions. These issues have implications for both understanding river-groundwater interactions in general and protecting riverine ecosystems and managing water resources in catchments. The understanding of river-groundwater interactions can be further improved by expanding the scope of the studies in this thesis. Some examples include defining the volume and rate of water discharge from unsaturated zone in either floodplains or headwaters using geochemistry, coupling geochemistry and numerical modelling in assessing the volume of bank return flow, and reducing the uncertainties in using ^{222}Rn to estimate baseflow by addressing heterogeneity of ^{222}Rn in groundwater, degassing rate of ^{222}Rn in river and the water flux from hyporheic and parafluvial zones. This thesis will not provide the answers to all questions on river-groundwater interactions, but it will help to understand some aspects of river-groundwater interactions for the purpose of better catchment management.

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