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Examining the spatial and temporal variation of groundwater inflows to a valley-to-floodplain river 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 groundwater-surface water 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 gradient toward the river and hence causes high baseflow to the river during wet (high flow) periods. In the lower catchment, lower rainfall and finer-gained sediments reduce the magnitude and variability of hydraulic gradient between the aquifer 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 catchment. 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 2000 % 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 flows.

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

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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 groundwater-surface water (GW–SW) 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; (3) and 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 GW–SW 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.

GW–SW 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 GW–SW interactions from both qualitative and quantitative perspectives (e.g., 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 reaches secular equilibrium with ^{226}Ra in groundwater 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 inflows into rivers, especially where the difference in major ion concentrations between groundwater and surface water is small, such as in many upper catchment streams.

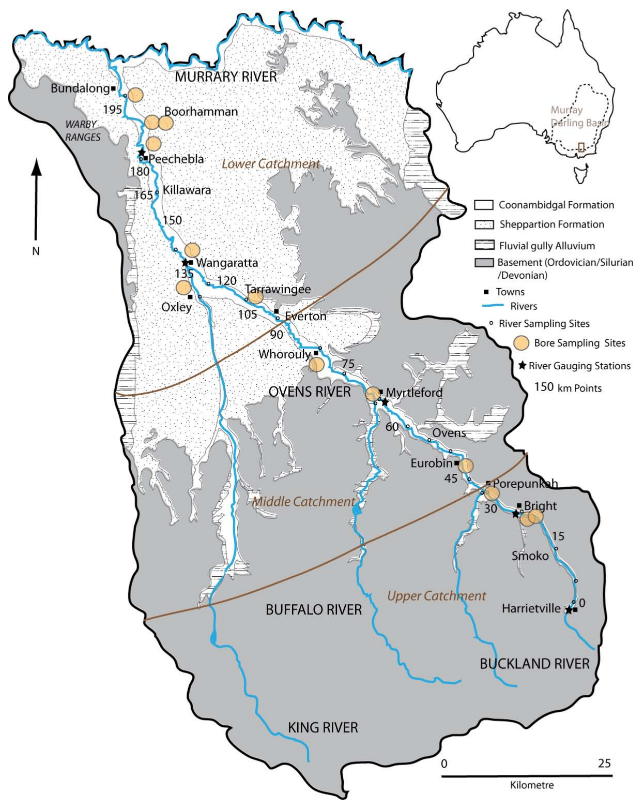

Fig. 1. Map of the Ovens River Catchment showing surface geology, sampling sites and gauging stations. Data from van den Berg and Morand (1997); Victorian Water Resource Data Warehouse (2011).

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

$$Q\frac{dC_r}{dx} = I(C_i - C_r) + wEC_r + F_{\mathrm{h}} - kdwC_r - \lambda dwC_r \quad (1)$$

(Cook et al., 2006; Mullinger et al., 2007). In Eq. (1), Q is the stream discharge (m^3 day^{-1}), C_r is the ^{222}Rn activity within the stream (Bq m^{-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 (Bq m^{-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. (1). Equation (1) can also be used for other tracers. For major ions, such as Na or Cl that do not degas to the atmosphere or decay, the last two terms on the right-hand side are redundant.

This study uses ^{222}Rn activities and major ion geochemistry in conjunction with physical hydrological data to determine the GW–SW relationships and the contribution of baseflow along the Ovens River (Fig. 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 (2001–2009) (van Dijk et al., 2013) and the 2010 Victorian floods; these are typical of floods that recur on average every 10 to 20 yr (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 GW–SW exchange and the baseflow fluxes to the river remains unknown. The results will provide an important background for future GW–SW studies in this and other catchments in the Murray–Darling Basin, and elsewhere.

2 Study area

2.1 The Ovens River

Located in the south-east margin of the 1 061 469 km^2 Murray–Darling Basin, the Ovens Catchment (Fig. 1) occupies just 7813 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; 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 2 m 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 (Fig. 1), the valley broadens and transitions into open alluvial flood plains. The river in the lower catchment has a low gradient of less than 1 m 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 (Fig. 1). 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.

2.2 Groundwater along the Ovens River

The stratigraphy of the Ovens Catchment comprise 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 170 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 (Fig. 1). 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 undistinguishable from those of the underlying Quaternary fluvio-lacustrine 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 where it is incised into the basement, for example, in Smoko, Bright and Myrtleford. The surface aquifers receive recharge through direct infiltration on the valley floors and via exposed and weathered bedrock at the margins of valley. Vertical head gradients throughout the Ovens catchment are downward except for a few locations within a few tens of metres of the river in the upper and middle catchment where they are upwards (Victorian Water Resource Data Warehouse, 2011). 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).

2.3 Climate and land use

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 10 % of the total water resource available in the catchment (Victorian Government Department of Sustainability and Environment, 2010c).

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. 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 floodplains. During March 2011 and June 2011, detailed 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; 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 (Bq m^{-3}). ^{222}Rn was degassed from a 500 mL Buchner flask via a closed circuit

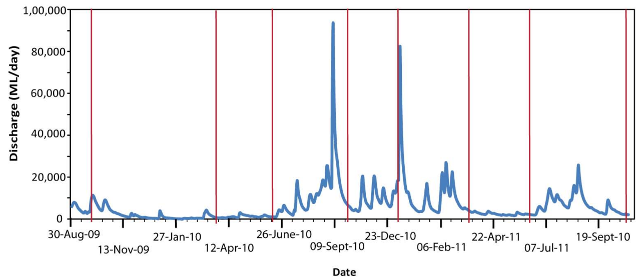

Fig. 2. Variation in discharge of the Ovens River at Peechelba (Fig. 1) during the study period (September 2009 to October 2011; Victorian Water Resource Data Warehouse, 2011). Low flow condition prior to June 2010 was due to drought, followed by several flood events in September to December 2010. Times of sampling are indicated by the red lines.

of a known volume for 5 minutes. Counting times were 3 or 4 cycles over 2 h for river water and 20 min for groundwater samples. Based on replicate analyses, precision of ^{222}Rn activities is within 3 % at 10 000 Bq m^{-3}, increasing to around 8 % at 200 Bq m^{-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 ICP-MS 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 %. 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).

4 Results

4.1 River discharge

Between September 2009 and June 2010, the discharge of the Ovens River at Peechelba was between 160 and 4360 ML day^{-1} with several moderately high flow events up to 11 420 ML day^{-1} during the 2009 winter (Fig. 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 2010–2011 La Niña event. The river flow returned to 1910 to 3800 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 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 (6635 ML day^{-1}) and March 2011

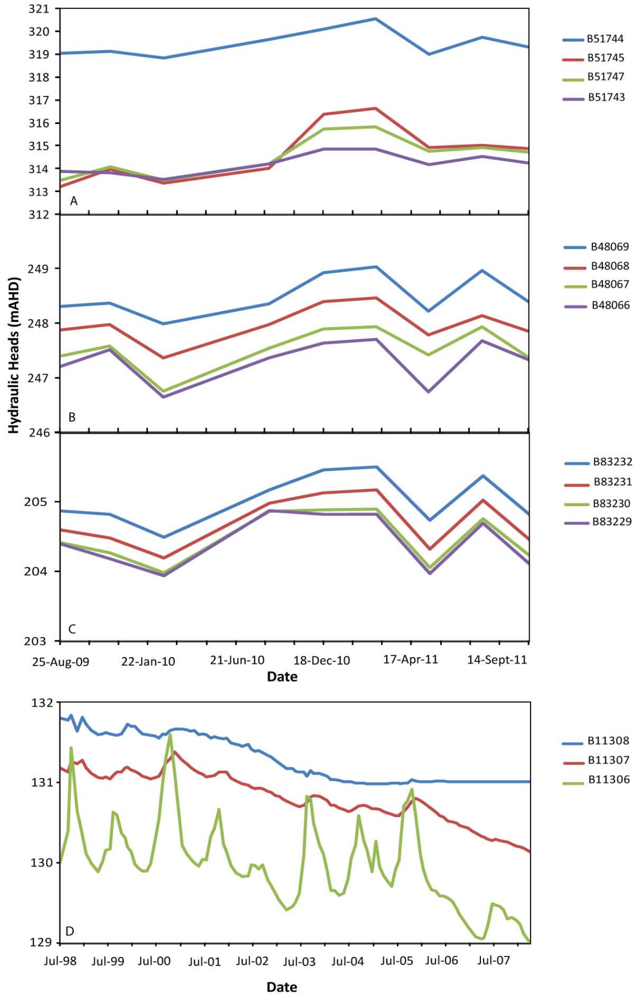

Fig. 3. Seasonal variation in bore hydrographs at **(A)** Bright in the upper catchment, **(B)** Eurobin and **(C)** Myrtleford in the middle catchment, and **(D)** Peechelba in the lower catchment (Victorian Water Resource Data Warehouse, 2011). Bores are listed in order of decreasing distance from 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.

(4894 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}, 1114 ML day^{-1}, 2292 ML day^{-1} and 2606 ML day^{-1}, respectively, and these sampling rounds represent low flow periods.

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. 3a) (Victorian Water Resource Data Warehouse, 2011). The annual hydraulic head variation at Bright in the upper

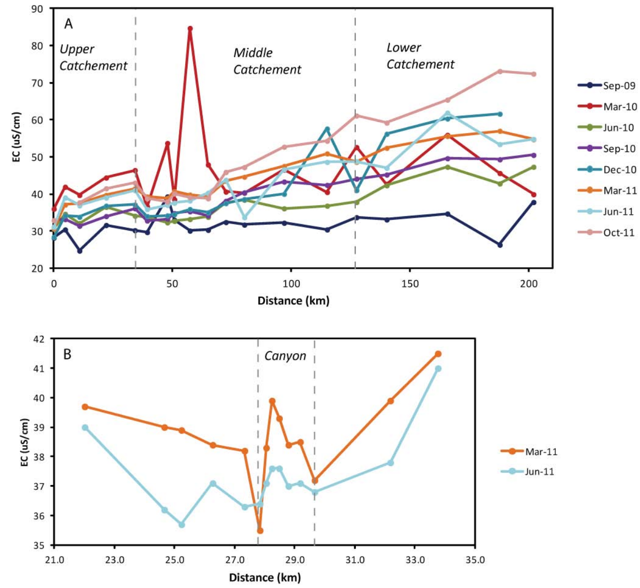

Fig. 4. (A) EC values along the course of the river (Table 1). EC values gradually increase downstream. **(B)** EC values along the Bright–Porepunkah reach in March and June 2011. Distinct EC peaks at 28.5 km, followed by a gradual increase in EC values in both sampling rounds.

catchment between 2009 and 2011 was 0.5 to 3.0 m. There was a strong lateral head gradient of $\sim 7 \times 10^{-3}$ 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. 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, historical data at Peechelba indicates the hydraulic heads in the flood plains (B11308 & B11307) varies by only a few millimetres per year (Fig. 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 is $\sim 10^{-4}$.

4.3 Electrical conductivity

The EC values of the Ovens River increase from $\sim 30\,\mu\mathrm{S\,cm}^{-1}$ in the upper catchment to 37 to 55 $\mu\mathrm{S\,cm}^{-1}$ at Peechelba in the lower catchment (Fig. 4). There is always an increase in the EC values in the first 5 km river reach from Harrietville. However, most of the increase in EC values occurs from the middle catchment downstream.

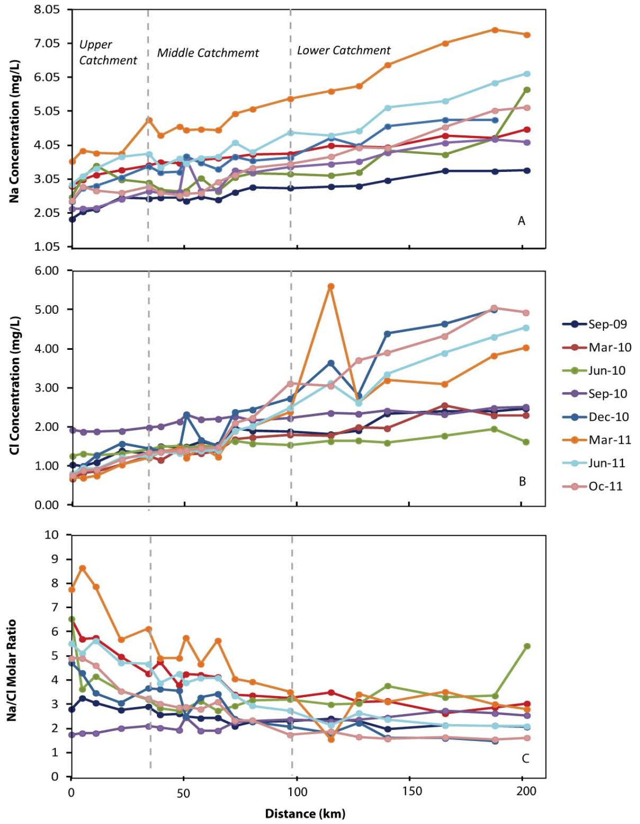

Fig. 5. Variation in Na **(A)** and Cl **(B)** concentrations, and Na/Cl ratios **(C)** along the river. Na and Cl concentrations increase progressively downstream, whereas the Na/Cl ratios decrease downstream.

Higher EC values (35 to 55 µS cm^{-1}) were recorded in March 2010, March 2011 and June 2011 at the end of summer or during low flow. There is a small increase in EC in the Bright–Porepunkah river section (2.8 µS cm^{-1} in March 2011, 1.2 µS cm^{-1} in June 2011) in the canyon (at 28 km) followed by a progressive increase in EC values downstream towards Porepunkah.

The EC values of shallow groundwater (< 50 m) increases 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 1200 µS cm^{-1} in the lower catchment (Table 2). EC values of groundwater before and after the 2010 Victorian floods were similar.

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 1). The relative mass abundance of the anions are HCO_3 (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 EC, the concentrations of major ions progressively increase downstream. For example, sodium concentrations increase 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. 5a). Likewise, chloride concentrations rise from 0.70 to 1.90 mg L^{-1} in the upper catchment to 1.60 to 4.90 mg L^{-1} at Peechelba (Table 3) (Fig. 5b). Molar Na/Cl ratios decrease downstream from 2.80 to 8.60 in the upper catchment to 1.60 to 3.50 at ∼ 135 km and then remain at a similar level in the lower catchment (Fig. 5c). The only exception to this trend is the September 2010 sampling round where the Na/Cl ratio was between 1.70 and 2.70 along the entire river. Other cation/Cl ratios have similar trends to Na/Cl.

Groundwater in the upper and middle catchment is 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 2). Groundwater in the lower catchment is 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 % 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 are mainly between 1.0 and 3.9 but locally as high as 11, whereas the more saline groundwater from the lower catchment has a Na/Cl ratio of 0.8 to 1.5 which is close to those of local rainfall (Blackburn and McLeod, 1983).

4.5 Radon activities

The Ovens River at uppermost site (0 km) in Harrietville has consistently low ^{222}Rn activities (112 to 245 Bq m^{-3}), whereas other river reaches in the upper catchment have ^{222}Rn activities between 373 and 2903 Bq m^{-3} (Fig. 6a). The highest ^{222}Rn activity is 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 Bq m^{-3} (March 2010) and 817 Bq m^{-3} (June 2010) at 28 km in the canyon (Fig. 6b), 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 1243 Bq m^{-3}. A small stream and spring on the floodplain at Porepunkah had ^{222}Rn activities of 2663 Bq m^{-3} (March 2010) and 8083 Bq m^{-3} (June 2010), and 10 488 Bq m^{-3} (March 2010) and 50 450 Bq m^{-3} (June 2010), respectively. In the middle catchment, ^{222}Rn activities generally decrease downstream from 601 to 2174 Bq m^{-3} to 231 to 440 Bq m^{-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.

Table 1. EC values of the Ovens River. Notes: nm – not measured; a – distance from the uppermost sampling site, Harrietville.

					EC (µS cm^{-1})							
Site No.	Site Name	Easting	Northing	Distance (km)[a]	September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
Ovens River												
SW19	Bundalong	042 773	6 007 835	202	37.8	39.9	47.3	50.5	nm	54.7	54.8	72.4
SW18	Peechelba	043 119	5 997 555	187	26.3	45.6	42.8	49.3	61.6	57.0	53.4	73.1
SW17	Killawara	043 417	5 984 065	165	34.7	55.9	47.3	49.6	60.3	55.6	61.9	65.4
SW16	Wangaratta	043 861	5 956 675	140	33.1	42.7	42.3	45.2	56.2	52.4	47.0	59.2
SW15	South Wang	044 237	5 974 395	127	33.7	52.6	37.8	44.1	41	48.6	48.8	61.1
SW14	Tarrawingee	045 113	5 970 115	115	30.4	40.5	36.8	42.4	57.6	50.9	48.6	54.3
SW13	Everton	045 732	5 966 695	97	32.2	46.5	36.1	43.3	40	47.5	46.6	52.7
SW12	Whorley	046 451	5 959 045	80	31.7	40.3	38.4	40.4	38.6	44.6	33.7	47.2
SW11	Whorley East	047 013	5 966 685	72	32.4	40.6	37.7	38.1	37.5	43.5	43.8	45.9
SW10	Myrtleford	047 453	5 952 805	65	30.3	47.9	33.8	34.2	35	39.2	40.3	38.8
SW9	Salziers Ln	047 837	5 949 855	57	30.1	84.7	33.2	35.3	35.9	39.8	38.2	39.1
SW8	Ovens	048 239	5 947 285	50	33	38.5	32.7	34.8	34.5	40.7	37.6	40.1
SW7	Eurobin	048 684	5 945 025	47	39.3	53.7	32.3	33.3	34.2	38.9	37.0	37.9
SW6	Porepunkah N	048 833	5 941 035	39	29.7	36.6	33.7	32.8	34	39.3	35.9	38.8
SW5	Porepunkah	049 182	5 938 925	34	30.2	46.4	34.1	36.1	37.1	41.5	41.0	43.0
SW4	Bright	049 936	5 935 505	22	31.6	44.4	36.5	34.0	36.8	39.7	39.0	41.5
SW3	Smoko	050 568	5 927 365	11	24.7	39.7	32.1	31.4	33.8	37.4	36.9	37.7
SW2	Trout Farm	050 560	5 918 975	5	30.4	41.9	34.6	33.2	34.2	37.1	39.0	
SW1	Harrietville	050 566	5 917 175	0	28.2	36.0	30.6	32.9	28.4	30.7	31.1	32.9
Tributaries												
SW22	King @ Oxley	044 442	5 966 735		33.0	37.7	42.6	43.7	43.1	50.8	85.2	51.9
SW21	Buffalo @ Myrtleford	047 182	5 954 215		34.8	37.3	37.6	36.6	34.6	39.2	39.4	41.0
SW20	Buckland @ Porepunkah	049 049	5 938 625		33.2	34.1	30.7	30.0	32.8	33.2	31.6	34.9
Bright–Porepunkah												
SW5		049 182	593 892	33.8						41.5	41.0	
BC8		0 493 028	5 937 880	32.2						39.9	37.8	
BC7		0 494 855	5 936 181	29.7						37.2	36.8	
BC1		0 495 297	5 936 079	29.2						38.5	37.1	
BC2		0 495 387	5 935 740	28.8						38.4	37.0	
BC3		0 495 519	5 935 501	28.5						39.3	37.6	
BC4		0 495 640	5 935 350	28.2						39.9	37.6	
BC5		0 495 703	5 935 212	28.0						38.3	37.1	
BC6		0 496 330	5 935 344	27.8						35.5	36.4	
BU4		0 496 796	5 925 419	27.3						38.2	36.3	
BU1		0 497 316	5 935 587	26.3						38.4	37.1	
BU2		0 498 227	5 935 171	25.2						38.9	35.7	
BU3		0 498 714	5 935 382	24.7						39.0	36.2	
SW4		049 936	5 935 505	22.0						39.7	39.0	
Bright–Porepunkah (At Location BC8)												
SR	Spring-fed Stream									48.6	49.0	
SS	Spring									64.1	52.2	

River reaches in the lower catchment have the lowest ^{222}Rn activities, ranging between 80 and 754 Bq m^{-3}. Elevated ^{222}Rn activities of between 699 and 745 Bq m^{-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 is minimal, with a maximum difference of ~200 Bq m^{-3} between sampling rounds.

The ^{222}Rn activities of groundwater are 30 000 to 110 000 Bq m^{-3} in the upper catchment, 20 000 to 42 000 Bq m^{-3} in the middle catchment, and 10 000 to 20 000 Bq m^{-3} in the lower catchment (Table 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 plains. There were no statistically-significant differences in groundwater ^{222}Rn activities between the sampling rounds even after the 2010 floods.

Table 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.

Bore No.	Location	Catchment	Distance (km)[a]	Easting	Northing	Bore Screen Depth (m)[b]	Distance to River (km)	September 2009 EC ($\mu S\,cm^{-1}$)	September 2009 Cl ($mg\,L^{-1}$)	March 2011 EC ($\mu S\,cm^{-1}$)	March 2011 Cl ($mg\,L^{-1}$)
B51743	Bright	Upper	22	499 291	5 935 508	5–11	0.0206	82.1	2.80		
B51747	Bright	Upper	22	499 190	5 935 414	2–20	0.160	58.1	3.10		
B1	Bright	Upper	22	499 270	5 935 517	2–4	0.0095			94.7	2.95
B2	Bright	Upper	22	499 260	5 935 513	2–4	0.0156			82.0	2.87
B51745	Bright	Upper	22	499 139	5 935 375	5–11	0.225	63.1	2.16		
B51744	Bright	Upper	22	498 933	5 934 911	6–12	0.725	56.3	2.73		
B51737	Bright	Upper	22	498 445	5 935 658	36–42	0.261	110.6	2.41		
B51738	Bright	Upper	22	498 397	5 935 420	58–63	0.0262	199.5	2.39		
B51735	Bright	Upper	22	498 391	5 935 314	30–42	0.0708	74.4	3.58		
B51736	Bright	Upper	22	498 382	5 935 299	20–26	0.0824	53.1	3.29		
B109461	Bright	Upper	22	497 818	5 935 267	20–26	0.339	82.8	3.42		
B109462	Bright	Upper	22	497 818	5 935 267	45–51	0.339	100.4	3.48		
B88271	Porepunkah	Upper	34	493 294	5 938 062	8–14	0.219	99.5	3.98	106.5	3.75
B88274	Porepunkah	Upper	34	493 256	5 938 067	35–53	0.210	64.3	1.85		
B48069	Eurobin	Middle	47	487 803	5 944 698	5–8	0.506	128.5	3.59	122.0	4.07
B48068	Eurobin	Middle	47	487 657	5 944 643	7–13	0.357	74.3	3.52	68.9	3.42
B48067	Eurobin	Middle	47	487 519	5 944 594	12.0	0.203	92.1	4.14	77.4	3.24
B48066	Eurobin	Middle	47	487 411	5 944 553	9–15	0.0914	77.9	3.90	70.3	3.97
B83232	Myrtleford	Middle	65	474 884	5 953 288	6–12	0.447	106.5	9.33		
B83231	Myrtleford	Middle	65	474 704	5 953 010	8–14	0.126	49.1	2.08		
M1	Myrtleford	Middle	65	474 605	5 952 919	4–6	0.0049			90.0	2.37
M2	Myrtleford	Middle	65	474 605	5 952 936	4–6	0.0215			67.6	2.43
B83229	Myrtleford	Middle	65	474 607	5 952 916	8–14	0.0128	86.7	2.01		
B83230	Myrtleford	Middle	65	474 604	5 952 937	8–14	0.0359	45.1	1.87		
B102783	Whorouly	Middle	80	464 087	5 959 833	5.1–11.27	0.0048	106.9	10.32		
T1	Tarrawingee	Lower	115	451 112	5 970 209	5–7	0.0056			367.0	46.89
T2	Tarrawingee	Lower	115	451 121	5 970 212	5–7	0.0142			364.0	47.78
T3	Tarrawingee	Lower	115	451 136	5 970 245	6–8	0.0509			315.0	45.96
B110738	Oxley	Lower	125	444 240	5 966 742	18.5–44	0.0048	106.9	10.32		
B11326	Wangaratta	Lower	140	439 879	5 982 755	23.7	2.79	106.3	15.42		
B11493	Wangaratta	Lower	140	439 422	5 982 189	16.5	2.13	1341	192.03		
B302296	Boorhamman East	Lower	165	437 925	5 992 950	70.5–77	3.52	920	134.21		
B11323	Boorhamman East	Lower	165	437 924	5 992 953	17.4	3.52	567	95.15		
B50788	Boorhamman	Lower	170	442 072	5 999 081	60–72	9.93	536	89.17		
B50789	Boorhamman	Lower	170	442 072	5 999 081	18–30	9.93	3800	654.13		
B11306	Peechelba	Lower	187	432 684	5 994 603	15.8	0.469	12020	2331.00		
B11311	Bundalong	Lower	202	427 007	6 005 559	16.2	0.3780	1194	163.46		
B11310	Bundalong	Lower	202	427 237	6 005 560	13.9	0.191	2270	389.07		

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. There are no occurrences of halite in the Ovens Valley, and Cl 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 (Fig. 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 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 ratios.

Overall, the high ^{222}Rn activities in the upper and middle catchment of the Ovens River (Fig. 6) together with the progressive downstream increase in EC values (Fig. 4) and major ion concentrations (Fig. 5a and b) 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 Cl, 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,

Table 3. Cl concentrations of the Ovens River.

Site No.	Site Name	Easting	Northing	Distance (km)[a]	Cl ($mg\,L^{-1}$) September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
					Ovens River							
SW19	Bundalong	042 773	6 007 835	202	2.47	2.30	1.62	2.51	nm	4.03	4.54	4.93
SW18	Peechelba	043 119	5 997 555	187	2.39	2.29	1.95	2.48	5.00	3.82	4.30	5.05
SW17	Killawara	043 417	5 984 065	165	2.39	2.55	1.77	2.32	4.64	3.10	3.89	4.32
SW16	Wangaratta	043 861	5 956 675	140	2.34	1.96	1.59	2.41	4.39	3.19	3.34	3.90
SW15	South Wang	044 237	5 974 395	127	1.91	1.99	1.65	2.33	2.80	2.62	2.61	3.71
SW14	Tarrawingee	045 113	5 970 115	115	1.81	1.78	1.63	2.35	3.64	5.61	3.12	3.05
SW13	Everton	045 732	5 966 695	97	1.88	1.80	1.54	2.22	2.72	2.38	2.48	3.11
SW12	Whorley	046 451	5 959 045	80	1.9	1.73	1.57	2.17	2.44	2.01	2.02	2.23
SW11	Whorley East	047 013	5 966 685	72	1.96	1.68	1.63	2.26	2.38	1.89	1.90	2.10
SW10	Myrtleford	047 453	5 952 805	65	1.54	1.37	1.51	2.2	1.50	1.23	1.40	1.47
SW9	Salziers Ln	047 837	5 949 855	57	1.61	1.32	1.52	2.19	1.66	1.49	1.39	1.45
SW8	Ovens	048 239	5 947 285	50	1.49	1.30	1.47	2.32	2.30	1.21	1.39	1.40
SW7	Eurobin	048 684	5 945 025	47	1.47	1.43	1.51	2.13	1.41	1.44	1.32	1.38
SW6	Porepunkah N	048 833	5 941 035	39	1.5	1.15	1.48	2.01	1.38	1.36	1.35	1.35
SW5	Porepunkah	049 182	5 938 925	34	1.31	1.25	1.42	1.98	1.44	1.21	1.25	1.34
SW4	Bright	049 936	5 935 505	22	1.39	1.03	1.32	1.89	1.56	1.03	1.21	1.15
SW3	Smoko	050 568	5 927 365	11	1.09	0.85	1.28	1.88	1.27	0.75	0.92	0.90
SW2	Trout Farm	050 560	5 918 975	5	0.99	0.82	1.31	1.87	1.00	0.69	0.95	0.89
SW1	Harrietville	050 566	5 917 175	0	1.03	0.67	1.25	1.92	0.78	0.71	0.81	0.76
					Tributaries							
SW22	King @ Oxley	044 442	5 966 735		2.77	2.10	2.21	2.31	3.02	3.20	3.14	3.46
SW21	Buffalo @ Myrtleford	047 182	5 954 215		2.12	1.64	1.61	2.31	2.21	3.34	1.80	1.87
SW20	Buckland @ Porepunkah	049 049	5 938 625		1.52	1.05	1.57	1.76	1.39	1.18	1.17	1.20
					Bright–Porepunkah (at location BC8)							
SR	Spring-feed stream									1.57	2.15	
SS	Spring									1.85	2.21	

2011; CA (Central Asian) Water-Info, 2013; NSW Government WaterInfo, 2013; USGS Water Data for USA, 2013).

5.1 Baseflow fluxes calculation using ^{222}Rn activities

Groundwater influxes to the river for the sampling rounds were calculated by rearranging Eq. (1) using ^{222}Rn activities from Table 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 vary from 1.2 to 8.0 m in winter and 0.3 to 6.7 m in summer, and river widths range from 15 to 100 m in winter and 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 5, ^{222}Rn activities of 76 000 Bq m^{-3}, 32 000 Bq m^{-3} and 19 000 Bq m^{-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 $\sim$300 Bq m^{-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):

$$k = 4.87\times10^{-4}\left(\frac{v}{d}\right)^{0.85} \tag{2}$$

$$k = 9.301\times10^{-3}\left(\frac{v^{0.5}}{d^{1.5}}\right), \tag{3}$$

where v is the average stream velocity (m day^{-1}) derived from the stream discharge, river depth and river width data. Equation (2) generally produces higher k values (and hence yields greater groundwater fluxes) than Eq. (3). The k values for the winter months are 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

Table 4. ^{222}Rn activities of the Ovens River.

					Radon ($Bq\,m^{-3}$)							
Site No.	Site Name	Easting	Northing	Distance (km)[a]	September 2009	March 2010	June 2010	September 2010	December 2010	March 2011	June 2011	October 2011
Ovens River												
SW19	Bundalong	042 773	6 007 835	202	199	242	109	119	nm	30	159	161
SW18	Peechelba	043 119	5 997 555	187	699	325	127	205	123	179	118	148
SW17	Killawara	043 417	5 984 065	165	230	754	130	211	105	83	185	135
SW16	Wangaratta	043 861	5 956 675	140	138	693	169	227	136	161	267	225
SW15	South Wang	044 237	5 974 395	127	227	193	179	240	127	186	318	224
SW14	Tarrawingee	045 113	5 970 115	115	222	206	227	385	169	257	363	387
SW13	Everton	045 732	5 966 695	97	297	263	239	439	231	275	440	325
SW12	Whorley	046 451	5 959 045	80	245	293	450	514	296	185	374	548
SW11	Whorley East	047 013	5 966 685	72	1296	623	562	536	413	654	555	414
SW10	Myrtleford	047 453	5 952 805	65	2318	639	628	674	230	500	643	413
SW9	Salziers Ln	047 837	5 949 855	57	640	293	538	744	225	567	625	662
SW8	Ovens	048 239	5 947 285	50	715	545	569	746	408	301	1337	530
SW7	Eurobin	048 684	5 945 025	47	1669	533	530	970	297	928	945	867
SW6	Porepunkah N	048 833	5 941 035	39	1544	573	552	770	399	930	803	942
SW5	Porepunkah	049 182	5 938 925	34	2174	1040	1005	1126	601	983	1243	1155
SW4	Bright	049 936	5 935 505	22	2903	630	707	754	654	580	1032	1007
SW3	Smoko	050 568	5 927 365	11	2781	1063	844	641	793	631	707	1062
SW2	Trout Farm	050 560	5 918 975	5	2659	1265	1168	679	783	373	1403	1103
SW1	Harrietville	050 566	5 917 175	0	222	112	238	213	169	224	245	226
Tributaries												
SW22	King @ Oxley	044 442	5 966 735		419	155	142	211	107	102	146	209
SW21	Buffalo @ Myrtleford	047 182	5 954 215		605	620	637	503	358	378	678	682
SW20	Buckland @ Porepunkah	049 049	5 938 625		339	450	740	824	644	267	928	1165
Bright–Porepunkah												
SW5		049 182	593 892	33.8						983	1243	
BC8		0 493 028	5 937 880	32.2						881	1060	
BC7		0 494 855	5 936 181	29.7						317	835	
BC1		0 495 297	5 936 079	29.2						469	627	
BC2		0 495 387	5 935 740	28.8						688	640	
BC3		0 495 519	5 935 501	28.5						905	817	
BC4		0 495 640	5 935 350	28.2						740	712	
BC5		0 495 703	5 935 212	28.0						374	617	
BC6		0 496 330	5 935 344	27.8						455	682	
BU4		0 496 796	5 925 419	27.3						371	792	
BU1		0 497 316	5 935 587	26.3						680	657	
BU2		0 498 227	5 935 171	25.2						564	738	
BU3		0 498 714	5 935 382	24.7						555	562	
SW4		049 936	5 935 505	22.0						580	1032	
At location BC8												
SR	Spring-fed stream									2652	8083	
SS	Spring									10 488	50 450	

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 4) and the discharge near the exit of the tributary.

The calculations indicate that most reaches are gaining ($I > 0\,m^3\,m^{-1}\,day^{-1}$), 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), 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 $m^3\,m^{-1}$

Table 5. ^{222}Rn activities of groundwater in the Ovens Catchment.

							Radon (Bq m^{-3})					
Bore No.	Location	Catchment	Easting	Northing	Bore Screen Depth (m)[b]	Distance to River (km)	September 2009	March 2010	June 2010	March 2011	June 2011	October 2011
B51737	Bright	Upper	498 445	5 935 658	36–42	0.261	116 750		100 230			
B51743	Bright	Upper	499 291	5 935 508	5–11	0.0206	75 880	59 000	67 210	64 500	72 375	82 500
B51744	Bright	Upper	498 933	5 934 911	6–12	0.725				85 875	26 125	23 200
B51747	Bright	Upper	499 190	5 935 414	2–20	0.160	50 650					
B1	Bright	Upper	499 270	5 935 517	2–4	0.0095				28 225	90 125	71 750
B2	Bright	Upper	499 260	5 935 513	2–4	0.0156				39 563	73 250	76 750
B88271	Porepunkah	Upper	493 294	5 938 062	8–14	0.219	58 880			48 125		
B48066	Eurobin	Mid	487 411	5 944 553	9–15	0.0914	42 910	34 740	45 620	28 350		
B48067	Eurobin	Mid	487 519	5 944 594	12	0.203	28 150	25 010	24 360	30 925		
B48068	Eurobin	Mid	487 657	5 944 643	7–13	0.357				31 163		
B48069	Eurobin	Mid	487 803	5 944 698	5–8	0.506				30 088		
B83229	Myrtleford	Mid	474 607	5 952 916	8–14	0.0128	26 180		25 980			
B83230	Myrtleford	Mid	474 604	5 952 937	8–14	0.0359	31 260		35 870			
M1	Myrtleford	Mid	474 605	5 952 919	4–6	0.0049				10 325	19 400	18 788
M2	Myrtleford	Mid	474 605	5 952 936	4–6	0.0215				19 500	30 238	25 063
B102873	Whorouly	Mid	464 087	5 959 833	5.1–11.27	0.513	28 660					
B110738	Oxley	Lower	444 240	5 966 742	18.5–44	0.0048	31 090					
T1	Tarrawingee	Lower	451 112	5 970 209	5–7	0.0056				5988	15 150	23 300
T2	Tarrawingee	Lower	451 121	5 970 212	5–7	0.0142				14 938	15 738	18 325
T3	Tarrawingee	Lower	451 136	5 970 245	6–8	0.0509				16 325	22 738	18 867
B11326	Wangaratta	Lower	439 879	5 982 755	23.7	2.79	35 100					
B11493	Wangaratta	Lower	439 422	5 982 189	16.5	2.13	21 900					
B11323	Boorhamman East	Lower	437 924	5 992 953	17.4	3.52	14 360		15 210			
B50789	Boorhamman	Lower	442 072	5 999 081	18–30	9.93	9260					
B11306	Peechelba	Lower	432 684	5 994 603	15.8	0.469	12 980	10 460	14 890			
B11310	Bundalong	Lower	427 237	6 005 560	16.2	0.191	21 360					
B11311	Bundalong	Lower	427 007	6 005 559	13.9	0.3780	18 290	18 130				

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. 7a); 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 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. 7b). 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. 8a), 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. 8b). 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 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. (3) lowers the estimates of groundwater influxes in individual reaches by 11 to 70 %, with an average of 43 %. The largest percentage changes occur in the gaining reaches where groundwater inflows are lower (c.f., Cook et al., 2003, 2006). 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} with a mean of 190 000 m^3 day^{-1}.

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 Bq m^{-3}, 7500 Bq m^{-3} and 8400 Bq m^{-3}, respectively. An increase in the groundwater ^{222}Rn activity reduces groundwater influxes in individual reaches by 19 to 31 %,

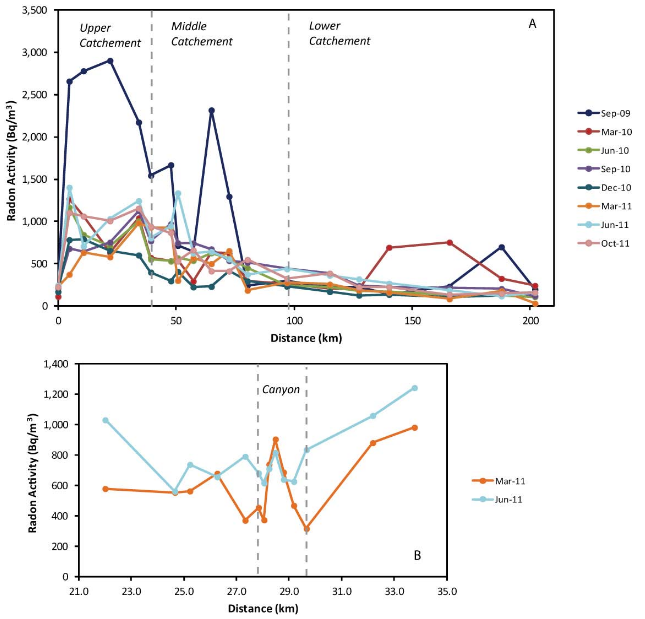

Fig. 6. (A) Variation of ^{222}Rn activities in the river (Table 4). High ^{222}Rn activities are recorded in the upper catchment and decrease down the valley. Temporal variation in the ^{222}Rn activities is minimal in the lower catchment. **(B)** ^{222}Rn activities along the Bright–Porepunkah reach in March and June 2011. Distinct ^{222}Rn peaks at 28.5 km (at which higher EC values are also found (Fig. 4b), followed by a gradual increase in ^{222}Rn activity in both sampling rounds.

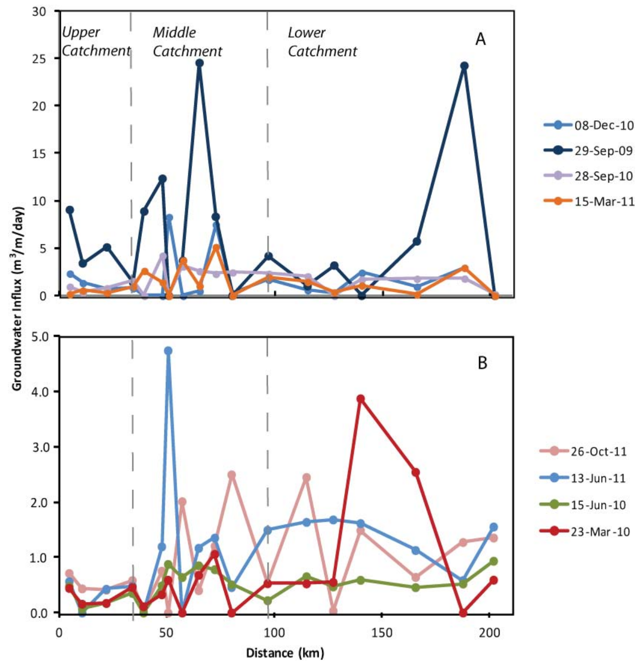

Fig. 7. Groundwater influxes calculated from ^{222}Rn activities, based on high k values, in flow conditions of 4894 to 18 520 ML day^{-1} **(A)** and of 995 to 2606 ML day^{-1} **(B)**. High baseflow occurs in the upper catchment and often increases during high flow conditions. High baseflow also occurs 65 to 72 km and 166 to 188 km.

whereas a decrease in the groundwater ^{222}Rn activity increases groundwater influxes by 31 to 81 %. The calculated cumulative groundwater inflows are 74 000 to 1 000 000 m^3 day^{-1} if the higher groundwater ^{222}Rn activities are used and 170 000 to 2 200 000 m^3 day^{-1} if the lower groundwater ^{222}Rn activities are used.

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 are then subtracted from the measured river ^{222}Rn activities and the groundwater influxes recalculated. For September 2009, the background river ^{222}Rn activities are 220 Bq m^{-3}, 175 Bq m^{-3} and 130 Bq m^{-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 %.

5.2 Baseflow fluxes calculation using Cl concentrations

Groundwater inputs to the river were also calculated by Cl concentrations (Table 3) via

$$I = \left(Q\frac{d\mathrm{Cl}_r}{dx} - wE\mathrm{Cl}_r \right) \Big/ (\mathrm{Cl}_i - \mathrm{Cl}_r) \tag{4}$$

(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 (Table 2), the chloride 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. 9a). 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 Cl concentrations are higher than those based on ^{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 1414 m^3 m^{-1} day^{-1}. The best match between ^{222}Rn- and Cl-derived groundwater influxes are the

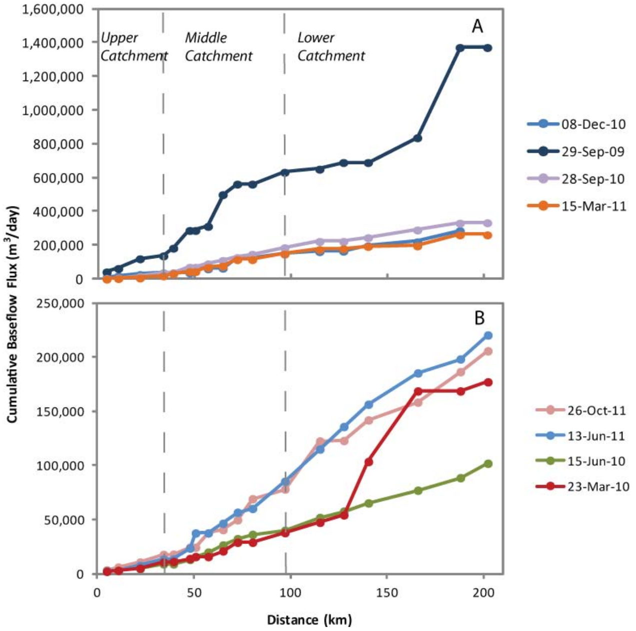

Fig. 8. Cumulative baseflow estimated from ^{222}Rn activities, based on high k values, in flow conditions of 4894 to 18 520 ML day **(A)** and of 995 to 2606 ML day **(B)**. High cumulative baseflow usually occurs in high flow conditions.

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 1400 $m^3\,m^{-1}\,day^{-1}$ (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), 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 ($C_i - C_r$) or ($Cl_i - Cl_r$) become smaller, 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.

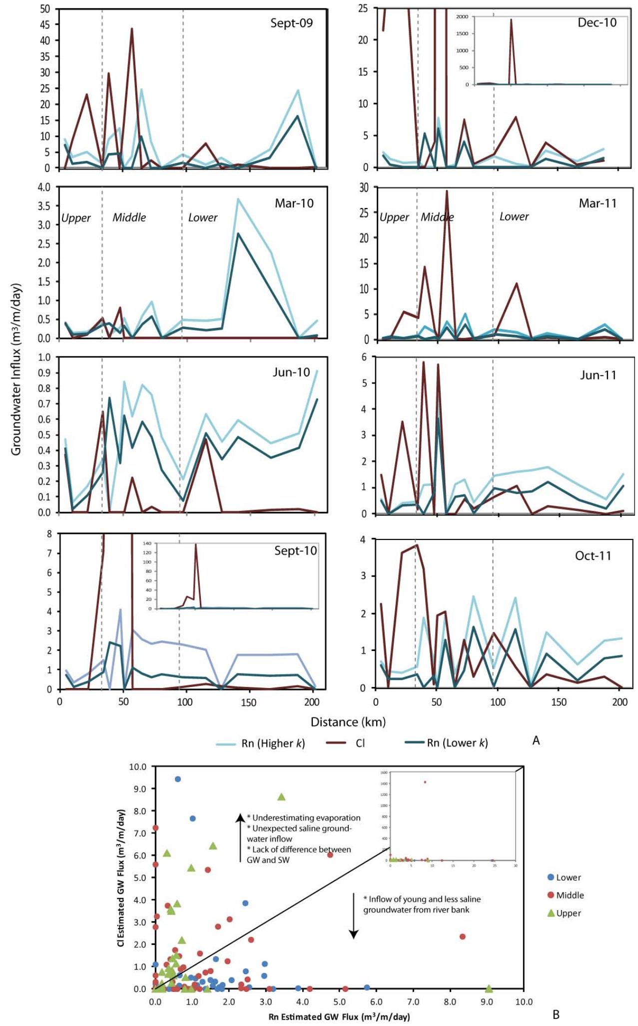

Fig. 9. (A) Comparison of baseflow fluxes in each sampling round, estimated from ^{222}Rn activities and Cl concentrations using Eqs. (1) and (4), respectively. For Rn, baseflow fluxes based on higher and lower k values are shown. **(B)** A scatter plot of all ^{222}Rn- and Cl-derived groundwater influxes (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.

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 and interflow (Nathan and McMahan, 1990; Eckhardt, 2005). The Nathan and McMahon filter (1990) is

$$b_k = y_k - \left[\alpha f_{k-1} + \frac{(1+\alpha)}{2}(y_k - y_{k-1})\right], \quad (5)$$

where 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 are 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):

$$Q_{t+1} = \alpha Q_t. \quad (6)$$

The calculated α 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

$$b_k \frac{(1-\mathrm{BFI}_{\max})ab_{k-1} + (1-\alpha)\mathrm{BFI}_{\max} y_k}{1-\alpha \mathrm{BFI}_{\max}}, \quad (7)$$

where b is the filtered baseflow ($b_k \leq y_k$) and $\mathrm{BFI}_{\max}$ is the maximum value of the baseflow index (BFI) that can be modelled by the algorithm. $\mathrm{BFI}_{\max}$ cannot be measured but is assigned based on the catchment lithology and river flow regime. Eckhardt (2005) proposed $\mathrm{BFI}_{\max}$ 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 $\mathrm{BFI}_{\max}$ 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.

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

$$Q_{\mathrm{gw}} = Q_{\mathrm{dn}} - Q_{\mathrm{up}} + \sum Q_{\mathrm{out}} - \sum Q_{\mathrm{in}} \quad (8)$$

(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 (evaporation and extraction) and Q_{in} is inputs to the reach (rainfall, tributaries and irrigation drainage). The groundwater during the low flow periods was calculated using Eq. (8) from parameters listed in Table 6. The calculations also subtract flow input from three main tributaries. Like the 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. These estimates are higher than the corresponding fluxes derived from ^{222}Rn activities; 4600 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.

5.5 Variations in baseflow

5.5.1 Spatial variations in baseflow

The baseflow fluxes derived from ^{222}Rn activates indicate moderate groundwater inflows in the upper catchment. In the upper catchment, the narrow valley creates a high hydraulic gradient of $\sim 7 \times 10^{-3}$ between the alluvial aquifers and the river producing the observed groundwater inflows (Fig. 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. 4b and 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. 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.

Table 6. Parameters used for calculating the net groundwater flux (Q_{gw}) during low flow conditions by differential flow gauging using Eq. (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).

Location	Parameters	March 2010	June 2010	June 2011	October 2011
Peechelba	Q_{dn} (m^3 day^{-1})	995 191	1 113 730	2 291 906	2 605 936
	Q_{up} (m^3 day^{-1}))	281 073	384 569	626 881	812 098
	Q_{out} (Evaporation) (m^3 day^{-1})	12 200	2440	2440	9760
	Q_{in} (Tributaries) (m^3 day^{-1})	384 757	503 800	1 168 102	1 159 343
	Q_{in} (Rainfall) (m^3 day^{-1})	0	0	0	0
	Q_{gw} (m^3 day^{-1})	341 561	227 801	499 363	644 255
Myrtleford	Q_{dn} (m^3 day^{-1})	281 073	384 569	626 881	812 098
	Q_{up} (m^3 day^{-1}))	108 119	142 785	274 903	340 353
	Q_{out} (Evaporation) (m^3 day^{-1})	2303	461	461	1842
	Q_{in} (Tributaries) (m^3 day^{-1})	119 485	152 338	214 891	273 636
	Q_{in} (Rainfall) (m^3 day^{-1})	0	0	0	0
	Q_{gw} (m^3 day^{-1})	55 772	89 907	137 548	19 9951
Bright	Q_{dn} (m^3 day^{-1})	108 119	142 785	274 903	340 353
	Q_{up} (m^3 day^{-1}))	54 021	90 172	59 184	152 869
	Q_{out} (Evaporation) (m^3 day^{-1})	550	110	110	440
	Q_{in} (Tributaries) (m^3 day^{-1})	15 819	18 623	43 630	63 689
	Q_{in} (Rainfall) (m^3 day^{-1})	0	0	0	0
	Q_{gw} (m^3 day^{-1})	38 829	34 100	172 199	124 235

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. 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. 7). 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. 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.

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. 7a and 8a). 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 (120 to 180 mm yr^{-1} (Cartwright and Morgenstem, 2012)) with an annual fluctuation of up to 3 m in the water table (Fig. 3a). The rising groundwater elevations, which is referred to as hydraulic loading, increases the hydraulic head gradient toward the river and thus causes 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. 8a). 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 reduces 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 (Fig. 2). The groundwater inflows in the upper catchment can be low during 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. 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.

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

In the upper and to some extent, the middle catchment, the groundwater fluxes estimated from Cl concentrations are often greater than those based on ^{222}Rn activities by 30 to 2000 % (Fig. 9). The possible reasons for the discrepancy would be underestimating evaporation, ignoring local saline groundwater inputs or/and lack of difference in Cl concentrations between the groundwater and surface water. 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 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 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. 9). If the assigned groundwater Cl concentrations in 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. 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 returns, interflow and draining of pools on the floodplains into the slow flow 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 components contribute to the flow of the Ovens River. That other components, such as bank returns, 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. This difference 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 net groundwater discharge (*outflow – inflow*) and thus can consist of groundwater, delayed bank returns and interflows. 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 high groundwater inflows from differential flow gauging may suggest that interflow and water from unsaturated zone provide a significant amount of discharge to the river. The large soil zone in the upper and middle catchments is likely supplies water to the river, maintaining the river flow during dry and baseflow conditions 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 GW–SW 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 groundwater (like regional groundwater, river bank, water from the unsaturated zone) the study focuses on.

6 Conclusions

The GW–SW 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, the total of 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 rapid 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 utilise 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.

Supplementary material related to this article is available online at http://www.hydrol-earth-syst-sci.net/17/4907/2013/hess-17-4907-2013-supplement.zip.

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