To what extent do long-duration high-volume dam releases influence river–aquifer interactions? A case study in New South Wales, Australia

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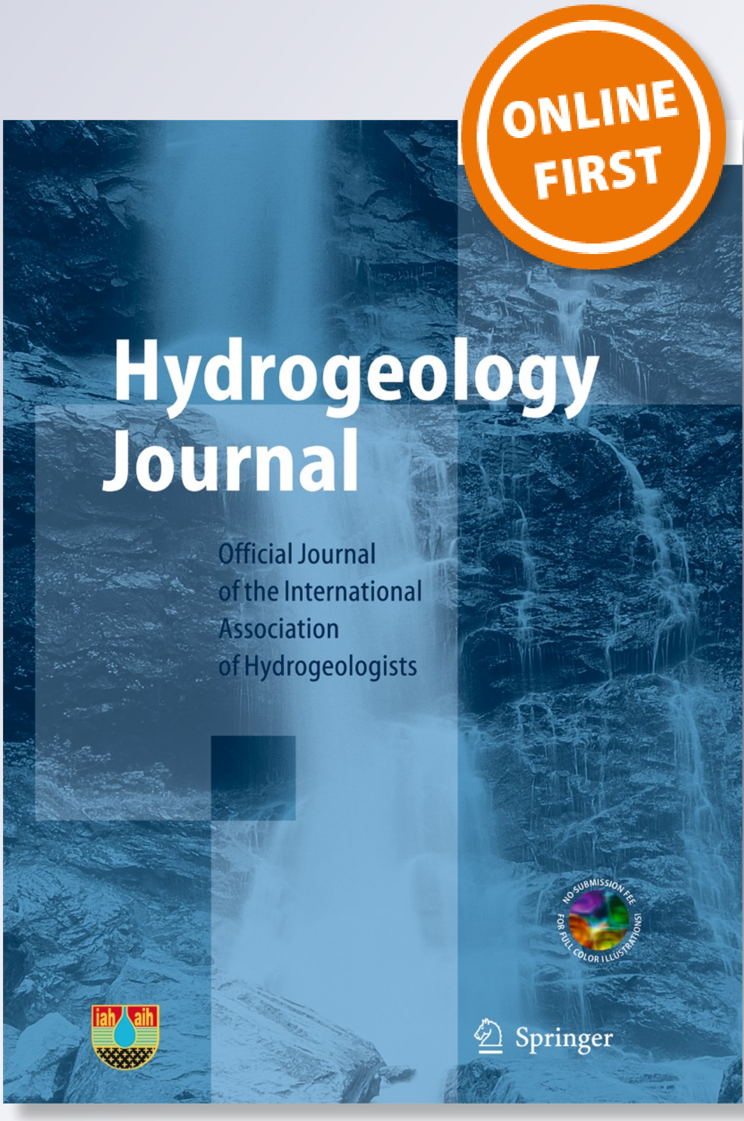

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To what extent do long-duration high-volume dam releases influence river–aquifer interactions? A case study in New South Wales, Australia

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Abstract Long-duration high-volume dam releases are unique anthropogenic events with no naturally occurring equivalents. The impact from such dam releases on a downstream Quaternary alluvial aquifer in New South Wales, Australia, is assessed. It is observed that long-duration (>26 days), high-volume dam releases (>8,000 ML/day average) result in significant variations in river–aquifer interactions. These variations include a flux from the river to the aquifer up to 6.3 m^3/day per metre of bank (at distances of up to 330 m from the river bank), increased extent and volume of recharge/bank storage, and a long-term (>100 days) reversal of river–aquifer fluxes. In contrast, during lower-volume events (<2,000 ML/day average) the flux was directed from the aquifer to the river at rates of up to 1.6 m^3/day per metre of bank. A groundwater-head prediction model was constructed and river–aquifer fluxes were calculated; however, predicted fluxes from this method showed poor correlation to fluxes calculated using actual groundwater heads. Long-duration high-volume dam releases have the potential to skew estimates of long-term aquifer resources and detrimentally alter the chemical and physical properties of phreatic aquifers flanking the river. The findings have ramifications for improved integrated management of dam systems and downstream aquifers.

Keywords Surface water hydrology · Dam releases · Groundwater/surface-water relations · Water management · Australia

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Introduction

Climate change and increasing world population will, throughout the 21st century, place increased stress on the worlds' water resources (Loáiciga 2003; Oki and Kanae 2006). It is estimated that 1.7 billion people live in areas where groundwater resources or groundwater-dependant ecosystems are under threat (Gleeson et al. 2012). The measurement and quantification of recharge to aquifers is an essential component for integrated water-resource management. In Australia and also internationally, groundwater resources play a critical role in supporting both rural and urban populations. The overexploitation and increasing development of these important resources is also a global risk, especially in light of competing interests from the domestic, agricultural and industrial sectors. Clearly, an improved understanding of the extent, capacity and myriad interactions that groundwater resources have with other natural and anthropogenic systems is key in developing an appropriate and sustainable management strategy. In the absence of intervention, it is widely believed that groundwater will be mismanaged and misallocated, resulting in either exhaustion of supplies or reaching a point where the cost of pumping additional water becomes uneconomical (Koundouri 2004).

Throughout the 20th century, anthropogenic features such as dams have resulted in alterations of the natural interaction between groundwater resources and river systems (Larned et al. 2008). Successful future management of water resources requires an understanding of how these anthropogenic features impact on groundwater–river systems. Furthermore, an ability to predict the likely response of an interconnected groundwater–river system to variations in how these systems are managed may allow more efficient utilisation of this limited resource.

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Although the inflow and mixing of water from a river system into an adjacent aquifer during periods of short-duration (less than 7 days) and high-volume dam release have previously been characterised (Ramírez-Hernández et al. 2013), assessments of the impact of long-duration high-volume dam releases (more than 5,000 ML/day for more than 7 days) on river–aquifer interaction are limited. For short-duration high-volume releases, research has suggested that net reduced flow would result in reduced groundwater recharge and falling groundwater levels in the riparian zones (Nilsson and Berggren 2000). Other studies have quantified the volume of daily hyporheic exchange (up to 1 m^3 per metre of bank per day) and the extent of water-table fluctuations into the riparian zone (up to 30 m; Sawyer AH et al. 2009). It has been suggested by Chen and Chen (2003) that although detailed studies have been completed on bank storage, analyses have rarely focussed on determining the zone of an aquifer where stream water infiltrated during the flood had displaced the pre-existing groundwater.

A long-term study of the impact of environmental release events (from 2,000 ML for 7 days to 13,000 ML for 4 days) on aquifer recharge was undertaken within the Colorado River (USA) by Ramirez Hernandez et al. (2013), with a correlation identified between transient elevation of groundwater heads and flood events. Research focusing on the Namoi River system in Australia (McCallum et al. 2013; McCallum et al. 2014) indicated that the relationship between stream stage, event duration and river loss are not necessarily straightforward. It was found that the loss ratio (flow loss/total river flow) was greater for smaller flows than larger flows with similar duration (McCallum et al. 2014). This indicates that antecedent aquifer conditions may play a significant role in the amount of recharge–bank storage for a particular flow event. Studies of the El Khairat aquifer in the Tunisian Sahel suggest that the construction of a dam has resulted in long-term recharge of the surrounding aquifer (which was depleted through over extraction) and a corresponding shift in the location of regional saltwater intrusion (Ketata et al. 2014).

Assessments of the impact of long-duration high-volume dam releases on river–aquifer interaction are therefore needed. Such assessments also need to be considered in the context of the possible effects of climate change, which might include increased variability and intensity of rainfall (Senior et al. 2002). The subsequent impact of variable hydrological forcing may result in more frequent high-volume releases from dams to prevent catastrophic overflow during extreme events as well as longer duration dam releases to sustain agricultural irrigation in dry conditions. As such, developing a better understanding of flood recharge rates, subsurface residence times and long-term impacts of flood recharge on river–aquifer exchanges warrants further research focus (Simpson and Meixner 2012).

Apart from fundamental hydrological implications, understanding river–aquifer interactions during flood events is also important from a water quality perspective. Solute exchange between a stream and aquifer during periods of high stream flow (floods) can be particularly important for the quality of riparian groundwater and subsequently the quality of baseflow fed stream flow (Baillie et al. 2007). In highly saline groundwater environments, the role of fluctuating water tables and discharge of shallow groundwater into rivers is central to the potential mobilisation and export of salts (Baskaran et al. 2009). A study of a portion of the Murray River in Australia, described the system as one which transitions from being dominantly losing to being variably gaining due to diminishing surface-water flows. During drought conditions, groundwater inflows to the river removed water from low salinity groundwater lenses, degrading these lenses (Cartwright et al. 2011).

The application of basin wide multiple criteria analysis (MCA) or multi-scale modelling for water resource planning and management has become commonplace over the past 30 years. However the application of these methods is limited by the explicitness of trade-offs and the quantity of useful information available (Hajkowicz and Collins 2007; Schmoldt and Peterson 2000; Victoria et al. 2005). A key criterion for water resource planning in a catchment is integration and assessment of all potential inputs and outputs to the hydrologic cycle. Changes to the water budget caused by dam releases are able to be planned and accounted for. However, the influence of long-duration high-volume dam releases is an input for which the effect has not to date been suitably assessed or characterised.

Further development of the processes resulting from long-duration high-volume dam releases may allow prediction of river–groundwater responses during such events by applying existing analytical models. The general relationship between river stage and aquifer head response is well recognised, with early attempts to predict aquifer response to river stage fluctuations employing analytical methods. These included approximations of changes in the groundwater head based on river stages using damped sinusoids and knowledge of the approximate aquifer diffusivity (Cooper and Rorabaugh 1963; Marino 1973). Methods for prediction of aquifer diffusivity based on the aquifer response to varied river stages have also been developed, with the diffusivity value then being used in recharge and flux calculations. Available methods include development of type curves to predict the diffusivity of the aquifer based on monitoring well response to changes in river stage (Pinder et al. 1969) or development of explicit expressions using closed form solutions (Singh 2003, 2004; Srivastava 2006). Criss and Criss (2012) present a generic head estimation methodology that does not require prior knowledge of the diffusivity for modelling the aquifer response to change in river stage. The head estimation equation uses three variable parameters (calibrated using an existing data set) to allow prediction of groundwater head change based on changes to river stage and effective daily precipitation. The methodology presented by (Criss and Criss 2012) has not been tested for long-duration high-volume dam release

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scenarios. As only river stage, precipitation and an initial set of head data is required for the model, it is an attractive option for long-term head estimation and presents a useful approach for catchment managers to predict groundwater response to rainfall and river stage variations with minimal data input required.

The aims of this study are (1) to estimate the influence of long-duration high-volume dam releases on water-table fluctuations for the first time, and (2) to assess the applicability of existing methods on quantifying stream–aquifer exchange. To achieve these aims, analysis of groundwater, river and dam release levels over a 2-year period was undertaken on an alluvial aquifer down-gradient of the Burrendong Dam, situated on the Macquarie River in central west New South Wales (NSW), Australia. Further, the Criss and Criss (2012) head estimation equation was implemented together with a volumetric flow equation to determine whether this model could predict river–aquifer interactions using only initial conditions and predicted or scheduled precipitation and dam release values.

The benefits of this study are the identification (and characterisation) of a previously neglected, but significant, common anthropogenic event with potential long-term impacts on groundwater sustainability. The modelling component was undertaken to determine if an existing model, like that utilised in this study, could provide catchment managers with a tool for the prediction of the effects of long-duration high-volume dam releases on the flux of water between down-gradient aquifer–river systems using only system starting conditions and dam release volumes as inputs; these are data-sets which may typically be available in most regulated catchments.

The Wellington field site

The research was undertaken at the University of New South Wales-Wellington Research Station (UNSW-WRS) located near the town of Wellington in the central west of NSW, Australia (Fig. 1). The climate zone of the region is characterised as a hot, dry zone with cool winter and has an average January temperature of more than 30 °C and an average July temperature of less than 14 °C. The annual average rainfall in the area is approximately 650 mm (BOM 2012).

The field site is located on the southern side of the Macquarie River, which is a major tributary within the Murray-Darling Basin (Fig. 1). Burrendong Dam is located approximately 15 km to the south-east of the UNSW-WRS, and the Macquarie River flows almost 25 km between Burrendong Dam and the study site. The dam has a storage capacity of 1,188,000 ML and releases to the Macquarie River via a gated concrete chute, with releases controlled by environmental, agricultural and drinking water supply requirements (State Water Corporation 2012). Throughout spring and summer, dam releases for agricultural irrigation typically govern the river flow. Over winter, the releases are minimal and are aimed at maintaining environmental flows. Based on the dam release data collected for 2011 and 2012, the average seasonal release is as follows: spring (September to November) 1,990 ML/day; summer (December to February) 2,530 ML/day; autumn (March to May) 218 ML/day; and winter (June to August) 390 ML/day. It should be noted that these release figures are variable and dependant on climate conditions.

The Macquarie River at the UNSW-WRS is located in a shallow valley overlaying an assumed fault zone within the Lachlan fold belt. Ordovician metamorphic units outcrop steeply next to the river on the north-eastern side, whereas Devonian metamorphic units outcrop gently some 500–600 m to the south-west of the river (Scott et al. 1999). The area between the river and the outcropping Devonian units is in-filled with Quaternary alluvial units that vary in size from silts and clays to gravel and cobble units. A site map with location of the river and bores is shown in Fig. 2a and a schematic cross section of the geology is shown in Fig. 2b.

A Silurian limestone unit is possibly present below the alluvium within the fault zone between the Ordovician and Devonian units (Fig. 2b). The alluvial units between the bedrock outcrops form an aquifer of between 10 and 25 m thickness. The alluvial sequence consists of an unconfined aquifer and a semi-confined aquifer. The semi-confined aquifer seems to be present within 100 m of the river and is the result of discontinuous lenses of silty and clayey gravel at depths of approximately 10 m below ground level (mbgl). These lenses have not been identified in bores further than 100 m from the river bank. However, from approximately 200 to 500 m from the river bank, a thick clay layer with low permeability is found at shallow depth (approximately 0.5–6 mbgl; see Fig. 2b). The clay layer is presumed to inhibit surface infiltration into this part of the aquifer.

Methodology

A conceptual hydro-stratigraphic model was based on a review of drilling logs, site walkover, aquifer and slug test results and long-term monitoring of rainfall, dam releases, river levels and groundwater levels. Following is a description of these data and the approaches used to develop the river–aquifer system conceptual model.

Aquifer characterisation

The alluvial aquifer was characterised through a drilling program, including installation of eight groundwater monitoring wells using a combination of auger, Tubex and air hammer drilling techniques. The location of wells was determined by differential global positioning system (DGPS), and distance to the river is shown in Table 1 along with well characteristics such as screened interval. The location of the wells in relation to the river is shown in Fig. 2a,b. A review of drill cuttings allowed for accurate borehole logs to be completed, with this

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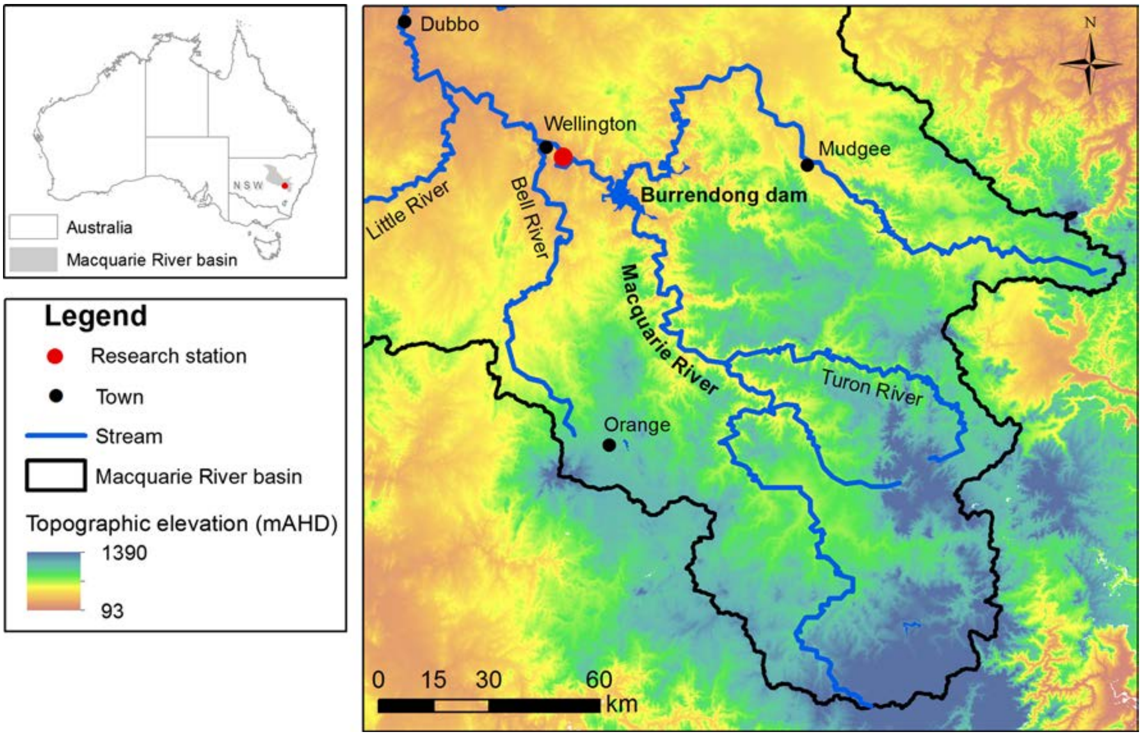

Fig. 1 Site location in New South Wales, Australia. AHD is the Australia Height Datum

information used to construct the cross section in Fig. 2b showing the thickness and geometry of aquifer horizons.

To determine the aquifer characteristics, a 96-h aquifer test and individual well slug tests were undertaken. Aquifer transmissivity (*T*) was calculated using the recovery curve from the extraction well and boreholes BH02 and BH04. The coefficient of storage (*S*) was calculated for individual wells using time drawdown graphs and the methodology outlined in Driscoll (1986).

Boreholes BH05, BH06, BH07 and BH08 were outside the cone of drawdown and therefore only slug test data were available for these locations. Borehole BH03 is believed to be in direct connection with the river and, therefore, the aquifer test was not able to cause a significant drawdown at this location and the slug test did not provide reasonable results. Slug test data was analysed using the Hvorslev method (Hvorslev 1951).

Water level and electrical conductivity data

Solinst level logger pressure transducers were installed in the wells to monitor water level and temperature at 30-min intervals over a period of 18 months. A barometric pressure logger was also installed to allow data sets to be corrected for barometric pressure. A permanent river level and temperature logging station was also installed. Daily dam release values during the investigation period were obtained from the New South Wales Office of Water. Daily rainfall data were collected from a weather station located 400 m to the south-west of the study site and river.

Electrical conductivity (EC) was also monitored in four groundwater bores over the period 30 April 2012 to 28 February 2013. Data were collected at 30-min intervals from bores BH01, BH02, BH05 and BH08. Boreholes BH05 and BH08 are located in the unconfined gravel aquifer 130 and 330 m from the Macquarie River, respectively. Borehole BH02 and BH01 were located within a semi-confined sandy clay unit 20 and 80 m from the Macquarie River, respectively. To determine the extent of river–aquifer interaction, the EC data were compared to river stage fluctuations, dam release levels and rainfall data.

Groundwater flux estimation

The flux of water exchange between the river and the aquifer was estimated using a volumetric flow equation based on Darcy's law (Fetter 1999):

$$Q = \left(T \times \frac{dh}{dx}\right) \quad (1)$$

where Q is the volumetric flow rate in m^3/day for a unit metre of river bank, T is the transmissivity and dh/dx is the groundwater gradient per time step. A positive value represents water moving into the aquifer from the river, while a negative value represents water flowing into the river from the aquifer. When applying this formula, it is assumed that the aquifer is isotropic and homogenous with a consistent thickness. The hydraulic conductivity used in

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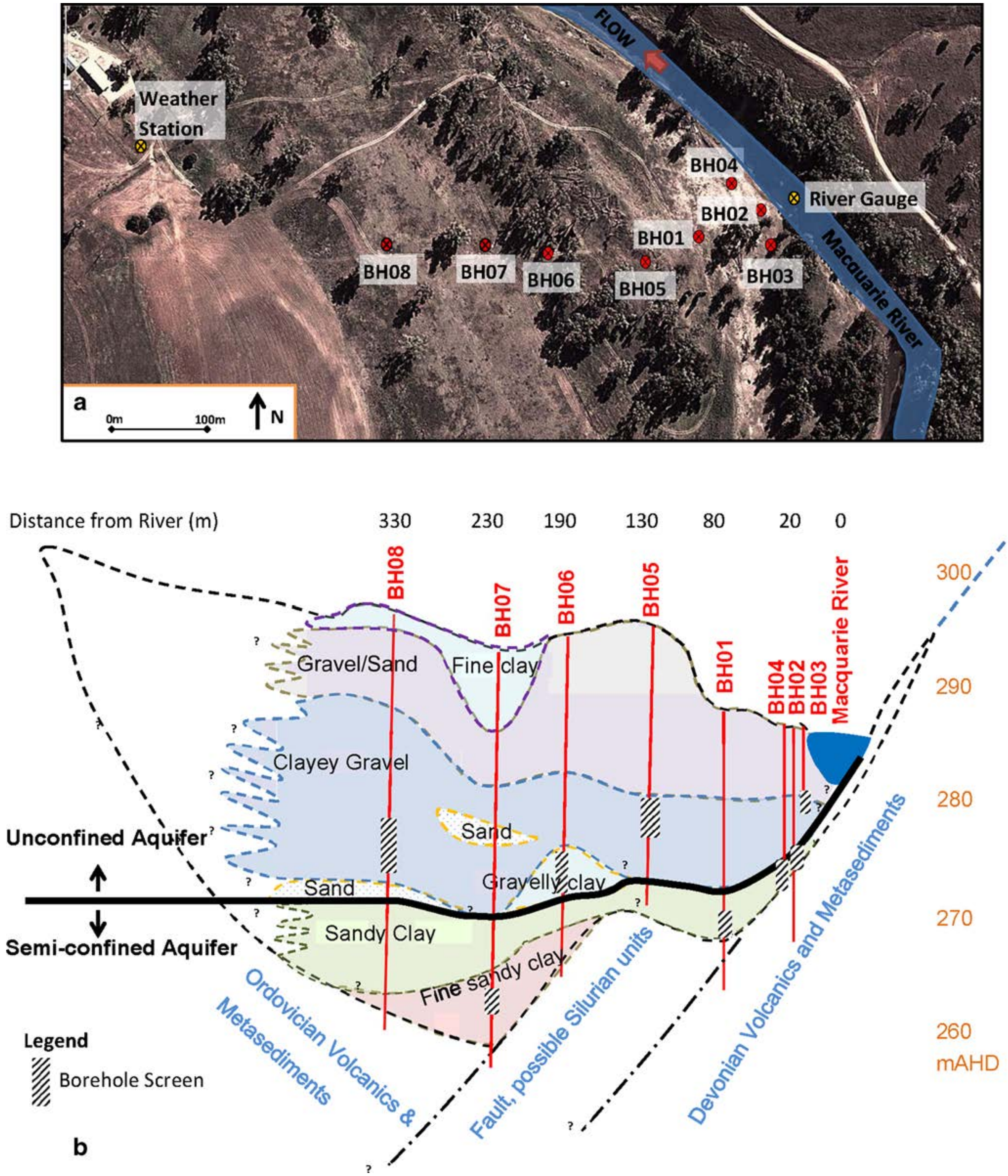

Fig. 2 **a** Borehole locations relative to *Macquarie River*. **b** Interpolated stratigraphic cross section showing screened depth in each borehole and the relative location of the unconfined and semi-confined aquifers

Eq. (1) was based on the average measured hydraulic conductivity between boreholes BH05 and BH02. The transmissivity was based on an average aquifer thickness of 10 m determined from the borehole logs.

Head estimation

Head estimation was undertaken on borehole BH05. This borehole was chosen as it was screened across the unconfined aquifer, had the most detailed data set, and

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Table 1 Measured hydrogeological characteristics at the Wellington Research Station

Borehole	Distance from river (m)	Storage coefficient, S	Hydraulic conductivity, K (m/day)	Transmissivity, T (m^2/day)	Screened interval (mAHD)	Aquifer thickness, from borehole logs (m)
BH01	80	1.7×10^{-3}	8.0	40	282–287	5
BH02	20	2.7×10^{-4}	16.0	39	274–277	3–5
BH03	10	–	–	–	278–281	5–6
BH04	30	6.0×10^{-5}	34.0	37	272–276	5–6
BH05	130	–	34.4	413	275–281	12–13
BH06	190	–	42.2	506	272–276	12–13
BH07	230	–	1.2	6	261–264	5–6
BH08	330	–	27.0	405	273–278	15

the borehole location was far enough away from the river (130 m) to ensure measured fluctuations in head were representative of long-term changes to bank storage/aquifer recharge. Borehole BH03 was initially used, however sensitivity analysis suggested that head variations in this location were almost completely driven by the river stage, with minimal input from other model parameters. As prediction of heads at borehole BH05 utilises all components of the model, it was considered a more representative location for the analysis.

Initially two methods were considered for head estimation: (1) the analytical solution for head response in a semi-infinite aquifer presented by Singh (2004) and (2) the simplified volumetric flux equation presented by Criss and Criss (2012). Initial investigations of the Singh method found that the head prediction was accurate for daily fluctuations identified in the boreholes close to the river such as borehole BH02, but the long-term trends observed in borehole BH05 were not satisfactorily modelled. The Criss and Criss method was found to provide more reproducible head estimation and was therefore implemented.

Criss and Criss (2012) present a simplified volumetric flux equation with three variable parameters to allow prediction of groundwater head change based on changes to river stage and effective daily precipitation (Criss and Criss 2012).

$$dh/dt = a(s-h) + b\left(P_{eff}\right) + c \tag{2}$$

where dh/dt is the predicted daily change in head (m/day), s is the daily value for actual river stage (m), h is the simulated groundwater head (m), P_{eff} is the effective precipitation (m/day) and a (day^{-1}), b (dimensionless) and c (m/day) are free parameters.Preliminary estimates of the free parameters are calculated using least squares fits to relate dh/dt to $(S - h)$ and P_{eff} respectively. Parameter a is initially estimated by reducing the data set to include values with no effective precipitation and using the following equation:

$$dh/dt \approx a(S-h) \tag{3}$$

Parameter b is then calculated at a point where there is known effective precipitation and actual change in groundwater head by substituting the estimated value of a in the following equation:

$$dh/dt \approx a(S-h) + bP_{eff} \tag{4}$$

Once the initial parameters for a and b have been derived using least squares fitting, they are used in the following equation to determine an initial estimate of c

$$c = -a(S-h) - b\,P_{\mathrm{eff}} + (\mathrm{h_f} - \mathrm{h_i})/T \tag{5}$$

where h_{f} and h_i represent the final and initial groundwater levels (m) and T(day^{-1}) is the total length of the record. The method described here allows initial parameter estimation with sufficient accuracy to allow integration of Eq. (2). This equation is integrated using a modified version of the Adams-Bashforth-Moulton (ABM) fourth order method (Butcher 2008), which uses the predictor and corrector equations identified below:

Predictor

$$h^*_{n+1} = h_n + \frac{\Delta t}{24}\left(55h'_{n+1} - 55h'_{n-1} + 37h'_{n-2} - 9h'_{n-3}\right) \tag{6}$$

Corrector

$$h_{n+1} = h_n + \frac{\Delta t}{24}\left(9h'^{*}_{n+1} + 19h'_n - 5h'_{n-1} - h'_{n-2}\right) \tag{7}$$

where h_n denotes the height of the water table at step n, * indicates an initial predicted value, Δt represents the step size, and h' the derivative of the water-table height given by Eq. (2). The parameter estimates are refined by undertaking additional least squares fitting. The parameters are then further refined by comparison of initial predicted heads to actual well values (over a specified initial period).

Results

The results of the aquifer characterisation testing are described in section 'Aquifer characterisation' and are

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shown in Table 1. The accuracy of the parameters governs the performance of this method and as would be expected, the longer the initial data set used to refine the parameters, the better the accuracy. The estimated parameters were found to be more accurate if the groundwater data used for the calibration included representative climatic variations such as floods and low river stage events. In this instance, a 40-day data set was used to estimate the initial parameters. Once the parameters were derived, numerical integration of Eq. (2) utilising Eqs. (6) and (7) can be completed producing a predicted groundwater head function. This head function uses complete river stage data and precipitation records along with actual head data for an initial 4 days, with subsequent head values predicted from Eqs. (2), (6) and (7). Once the parameters have been refined, Eqs. (2), (6) and (7) therefore allow prediction of well head response to different dam release/precipitation scenarios. When combined with Eq. (1), the flux of water moving between the river and aquifer can also be predicted for these scenarios. Criss and Criss (2012) have noted that the method is not able to accurately predict the head response during flood events causing overland flow, therefore, the very large natural flood events where significant overland flow occurs were not included. The aim of using modelled head levels to predict river–aquifer flux was to determine if the effects of future dam release events on river–aquifer flux could be accurately modelled and predicted using only starting conditions and the anticipated dam release values.

Two long-duration high-volume flood events occurred over the 18-month study period, with one in December 2010 and the other in April/May 2012 (Fig. 3a, events 2 and 3). Prior to the 2010 flood event the area had been in a long-term drought, with dam levels at less than 40 %, minimal river flow and local aquifers significantly depleted. Figure 3a (event 1) shows that the aquifer prior to the 2010 flood event was more responsive to rainfall events (when water was not released from the dam). For the period December 2010 to April 2012, minimal response to rainfall events was identified in the river (R^2=<0.40). However, the response of the river system to dam releases was found to correlate strongly (R^2=0.94).

Figure 3b shows the groundwater levels and dam releases during a 5-month period that includes the May/April 2012 flood event, along with several smaller releases. Based on visual inspection, the type of response in the wells can be split into three groups: (1) borehole BH03, which is in direct connection with the river and responds simultaneously with any river level change; (2) boreholes BH01, BH02, BH04 and BH07, which show rapid response to major changes in river levels; and (3) boreholes BH05, BH06 and BH08, which show a delayed response to river levels, and only respond to the long-duration high-volume events.

Boreholes BH05, BH06 and BH08 show minimal response to dam releases until such a point as the river level is higher than the groundwater level in the locations of these boreholes. This occurs twice in Fig. 3b, once during January 2012 and a second time during March 2012 (events 1 and 2). In both instances, as the groundwater gradient is reversed, the groundwater levels in these wells slowly increases, then subsides as the river level returns to base flow. However, the level in boreholes BH05, BH06 and BH08 remains at a higher level than prior to the groundwater gradient reversal, as shown in Fig. 3b (event 3). As there was a definite change in groundwater levels which was sustained for up to 12 months (Fig. 3a, following event 2), following the return of the river system to equilibrium, the rise in groundwater level is considered to represent a measureable amount of recharge to the aquifer.

A pressure response to river stage at boreholes BH01, BH02, BH03 and BH04 is shown in Fig. 3b. Over the period from 22 August to 1 December 2011, the dam releases average approximately 2,000 ML/day and boreholes BH01, BH02, BH03, BH04 and the river stage all show a level response to the dam releases (Fig. 3b, event 4). Using borehole BH04 as an example, the initial water level was 282.79 m AHD and coincided with a dam release of 100 ML/day. The level of borehole BH04 peaked at a level of 283.79 m AHD and, at the same time, the dam release was 3,916 ML/day. On 1 December 2011, the dam release returned to 194 ML/day and, at the same time, the level of borehole BH04 returned to 282.71 m AHD. This pattern was also evident in boreholes BH01, BH02 and BH03.

Groundwater away from the river typically has EC values of 0.7–0.8 mS/cm, while river water typically is within the 0.2–0.3 mS/cm range. Variation in EC at boreholes BH02, BH05 and BH08 were observed to respond to river stage (Fig. 4), while response to rainfall events was not apparent. River stage has been noted previously to be dependent on dam releases, with only minor variation in stage due to average rainfall events. The EC variations include a reduction in EC during high-volume release events and an increase during low volume periods. This would suggest flow in the unconfined aquifer consists of gradient controlled groundwater flow during low river stage, and inflow of river water during high stage events, resulting in a combination of bank storage and recharge.

A variation in EC was measured in boreholes from the river bank and up to 330 m away (Fig. 4). The variation in EC has been calculated as a percentage of the maximum EC value in a line of boreholes running from the river. These include: BH02 (20 m from river bank) where, following a long-duration high-volume flood event, EC decreased by up to 25 % of the total value; borehole BH05 (130 m from river bank) where EC varied up to 15 % of the total value; and borehole BH08 (330 m from river bank) where EC varied by up to 7 % of the total. Fluctuations observed in borehole BH01 were <2 %, which would be expected as this borehole samples water from a semi-confined aquifer. The EC variations in borehole BH08 could indicate that the river water–groundwater mixing zone extends at least as far as 330 m from the bank.

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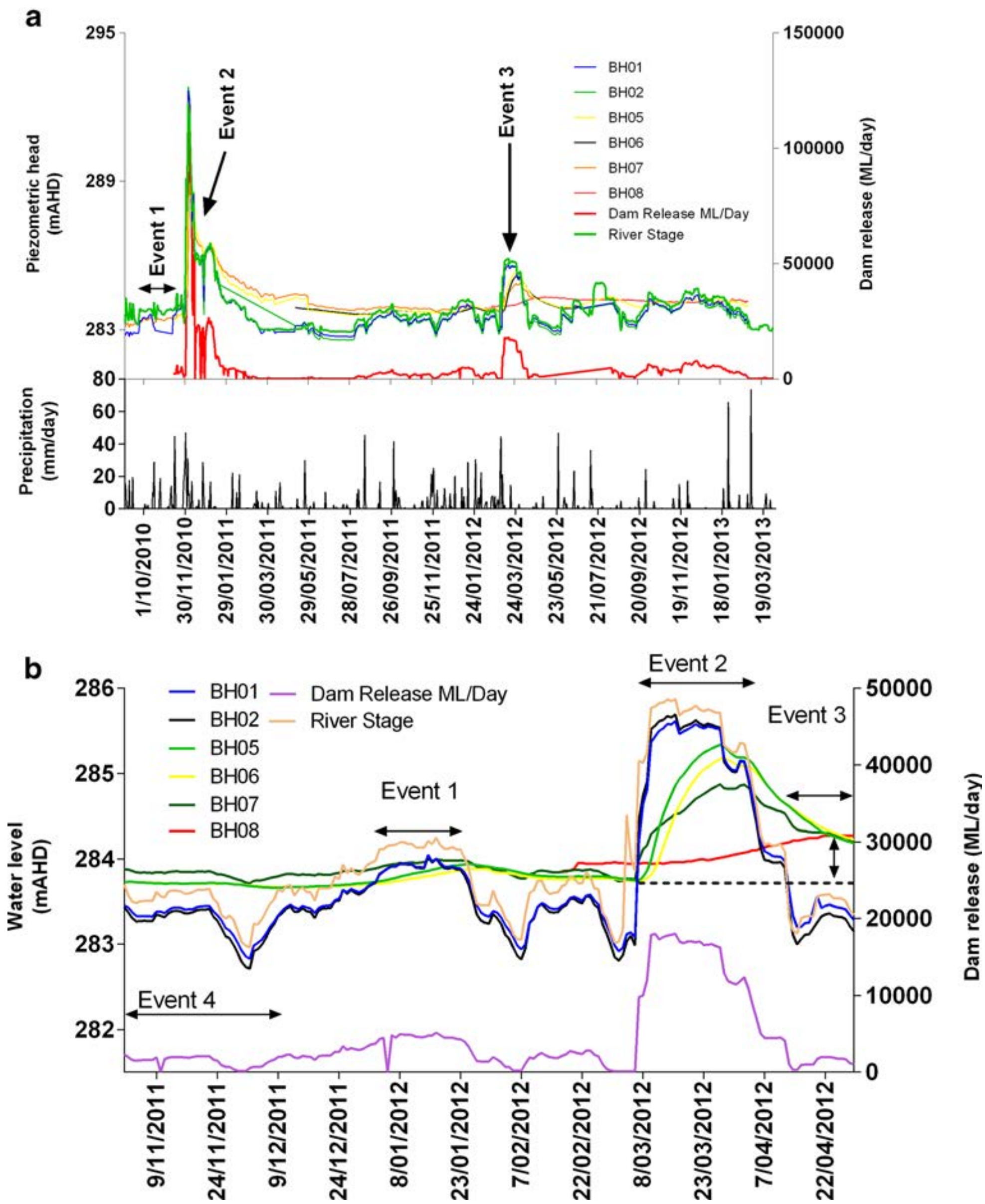

Fig. 3 **a** Comparison of groundwater-level response in boreholes *BH01*, *BH02*, *BH05*, *BH08*, *BH07* and *BH08* to dam release and rainfall events through the period 13 November 2010 to 6 April 2013 **b** Comparison of river-stage and groundwater-level response in boreholes *BH1*, *BH2*, *BH5*, *BH6*, *BH7* and *BH8* to dam release through the period 3 November 2011 to 1 May 2012

Groundwater flux estimation

Groundwater exchanges between the unconfined aquifer and the river were examined over the period 13 November 2010 to 28 February 2013. Fluxes were calculated between the river bank and borehole BH05 (130 m from the bank). Due to limited data availability, only short-term fluxes were calculated for the transects between boreholes BH06 and BH05 and boreholes BH08 and BH06. Fluxes calculated for these locations did show a response to the dam release undertaken between 7 March 2012 and 13 April 2012. Flux values between the river and borehole BH08 for targeted periods are shown in Table 2, with the flux response to dam release shown in Fig. 5a,b.

In general, there is a negative flux (movement of water from the river into the unconfined aquifer) recorded in response to the large long-duration dam release events. The maximum negative flux (movement of water into the aquifer) was measured during the 12,000 ML 37-day event, with peak flux of −4.42 m^3/day per metre of bank (−87 m^3/m cumulative flux over 37 days) inflow between the river and borehole BH05, peak flux of −0.95 m^3/day per metre of bank (−14 m^3/m cumulative flux over 37 days) between borehole BH05 and BH06, and peak flux of −0.64 m^3/day per metre of bank (−10 m^3/m cumulative flux over 37 days) days) measured between borehole BH06 and BH08. Interestingly, the flux was positive (movement from the aquifer to the river) between the river and borehole BH05 following a 26-day 39,000-ML event, likely due to the fact that overland flow was occurring during this

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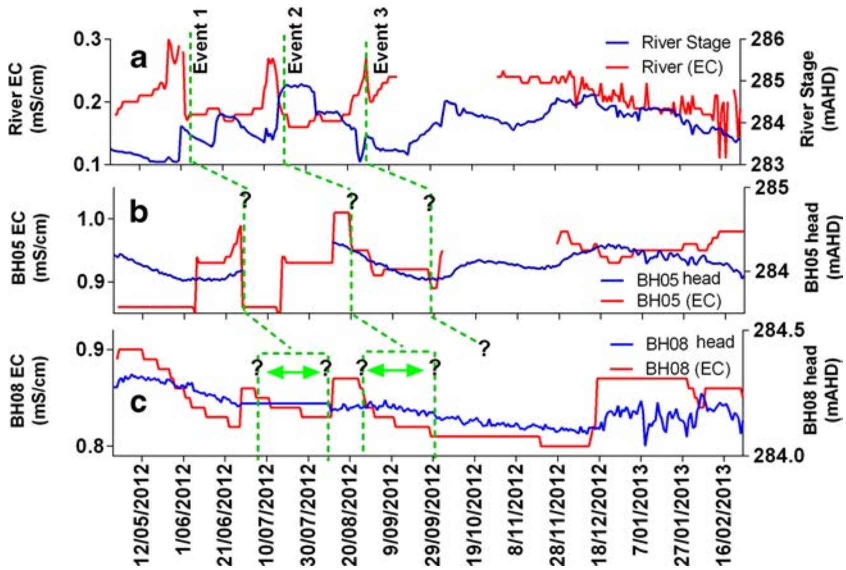

Fig. 4 Electrical conductivity (*EC*) response to river stage (30 April 2012 to 22 February 2013) in **a** the river, **b** borehole BH05 and **c** borehole BH08. The three events (*1*, *2*, *3*) are discussed in detail in the text

period. The flux measured during a 61-day 8,000-ML event produced a cumulative positive flux of 0.98 m^3/m over 61 days between borehole BH05 and the river, whilst the cumulative flux further away from the river, between borehole BH05 and BH06 and between BH06 and BH08, was −12.91 m^3/m over 61 days and −5.32 m^3/m over 61 days respectively. The flux measured between borehole BH05 and the river is less consistent, with flux both towards and away from the river recorded during periods of dam release greater than 8,000 ML/day. This is considered to be due to the highly variable river stage compared to the relatively stable groundwater level at borehole BH05, resulting in highly variable gradients. These gradients then influence the volumetric flow calculations. As the gradients between boreholes BH08, BH06 and BH05 are all taken from groundwater head data, the flux calculations in these locations are considered to be more representative as the groundwater system was more stable in these locations.

Table 2 Flux estimation and characteristics of representative dam release periods

Start date (dd/mm/yy)	Finish date (dd/mm/yy)	Total duration (days)	Average dam release (ML/day)	Total dam release (ML)	Average river stage (m AHD)	Max flux (m^3/day/m)	Min flux (m^3/day/m)	Cumulative volume (m^3)	Rain during period (mm)
Flux between borehole BH05 and river bank (m^3/day per metre of bank; negative values represent water movement away from the river)									
1/12/10	27/12/10	26	39,864	1,076,352	287.52	1.99	−6.33	2.59	179
7/03/12	12/04/12	37	12,731	471,074	285.27	0.86	−4.42	−86.99	20
1/03/12	30/04/12	61	8,225	493,516	284.55	1.63	−2.06	0.98	133
25/11/12	18/01/13	55	6,006	330,938	284.39	0.25	−0.56	−11.32	23
7/09/11	1/12/11	86	1,949	167,505	283.60	0.81	−0.35	12.17	227
12/04/12	30/04/12	18	1,434	25,820	283.44	1.63	0.65	17.98	7
23/06/11	20/08/11	60	320	18,960	283.05	0.88	0.17	40.06	101
Flux between borehole BH06 and BH05									
1/12/10	27/12/10	26	39,864	1,076,352	287.52	NA	NA	NA	179
7/03/12	12/04/12	37	12,731	471,074	285.27	0.0089	−0.95	−14.18	20
1/03/12	30/04/12	61	8,225	493,516	284.55	0.12	−0.95	−12.91	133
25/11/12	18/01/13	55	6,006	330,938	284.39	NA	NA	NA	23
7/09/11	1/12/11	86	1,949	167,505	283.60	0.02	−0.06	−1.16	227
12/04/12	30/04/12	18	1,434	25,820	283.44	0.12	0.006	1.42	7
23/06/11	20/08/11	60	320	18,960	283.05	0.11	0.05	5.78	101
Flux between borehole BH08 and BH06									
1/12/10	27/12/10	26	39,864	1,076.352	287.52	NA	NA	NA	179
7/03/12	12/04/12	37	12,731	471,074	285.27	0.52	−0.64	−10.57	20
1/03/12	30/04/12	61	8,225	493,516	284.55	0.59	−0.64	−5.32	133
25/11/12	18/01/13	55	6,006	330,938	284.39	NA	NA	NA	23
7/09/11	1/12/11	86	1,949	167,505	283.60	NA	NA	NA	227
12/04/12	30/04/12	18	1,434	25,820	283.44	0.42	−0.37	1.52	7
23/06/11	20/08/11	60	320	18,960	283.05	NA	NA	NA	101

NA data not available for this period

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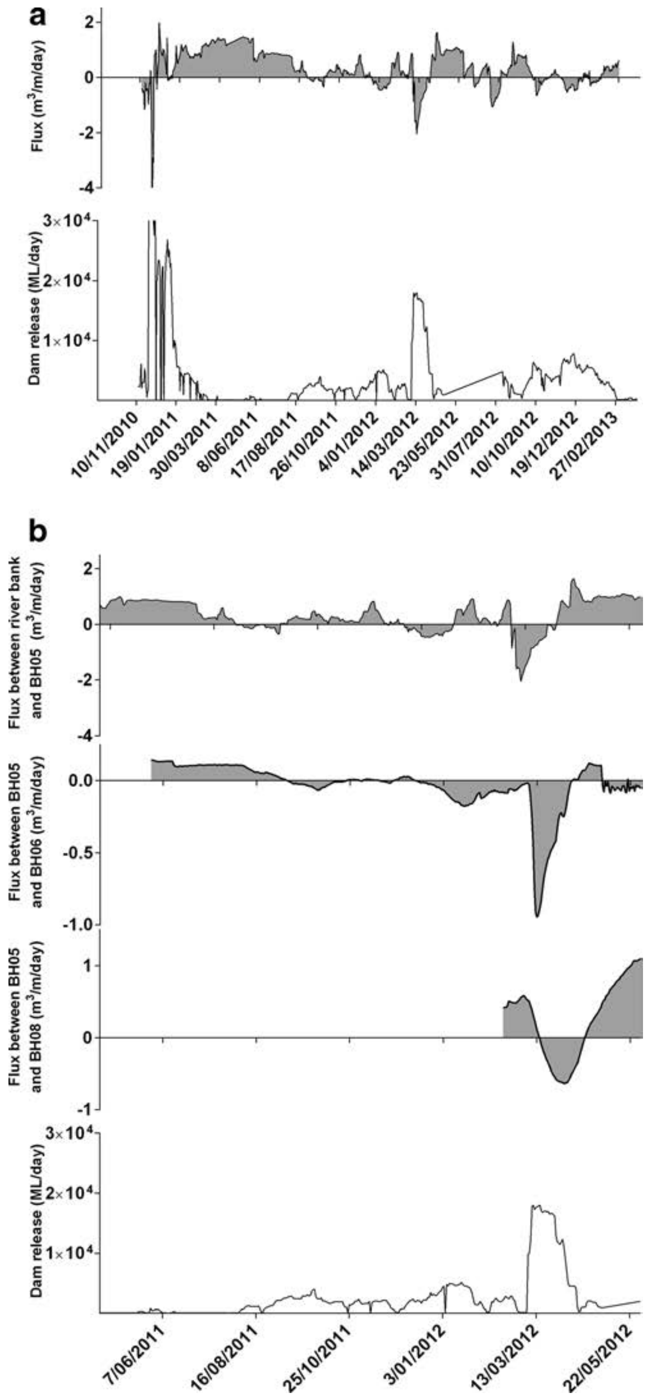

Fig. 5 **a** Calculated flux of groundwater between the river and borehole BH05 shown against dam-release volumes from 13 November 2010 to 28 February 2013. **b** Calculated flux of groundwater at given locations, shown against dam-release volumes from 2 June 2011 to 1 June 2012; *negative values* represent water movement away from the river

Head estimation and water-table fluctuation prediction

The Criss and Criss (2012) head prediction equation was applied to boreholes BH05, BH06 and BH08, with the aim to calibrate the model to each location and allow water-table fluctuations to be made. The overall goal of this was to enable modelled water-table fluctuations to be used for prediction of river–aquifer flux during future dam release events. Unfortunately, robust calibration at borehole BH08 could not be achieved, as comparison with actual head data achieved an R^2 of only 0.30.

Borehole BH05 was modelled over the period 13 November 2010 to 28 February 2013, while borehole BH06 was modelled over the period 10 May 2011 to 28 June 2012. In each instance, the variable parameters were estimated using the initial 40 days of actual head data. Following integration, additional calibration of the parameters was undertaken. Initial head estimations were compared with the existing groundwater data set (875 days) and further parameter refinement was undertaken through least squares fitting. Calibration of borehole BH05 and borehole BH06 was more successful than for BH08, with predicted versus observed head R^2 values of 0.76 and 0.77 for borehole BH05 and BH06, respectively. The parameters used in the calibrated model runs are presented in Table 3. Graphs of the predicted heads against observed heads are presented in Fig. 6a,b.

Using the predicted heads, groundwater–river flux at borehole BH05 and BH06 was estimated using the volumetric flow equation. The flux values calculated using the predicted heads were then compared to flux values calculated using field measured heads to provide a cumulative recharge percentage error for each time period. Results are shown in Table 4 and in Fig. 7a,b, respectively. The aim of this was to assess whether the predicted heads would provide suitable data for future flux estimations based on different dam release scenarios.

Discussion

River–aquifer interaction: observations and characterisation

A review of the dam release and river stage data revealed several events that involved long-duration high-volume releases, averaging in excess of 8,000 ML/day. The three largest of these events spanned 26, 37 and 61 days and released a total of 1,076 GL, 471 GL and 493 GL respectively during the periods. These events represent significant anthropogenic induced flow events, with a combination of duration and volume which would not be likely under natural conditions. Indeed, the volume of water released in these events is several magnitudes higher than would be expected as a result of normal rainfall induced floods. Records of annual river flow at a river gauge 65 km downstream from Burrendong Dam show the average annual flow prior to the dam construction was 730,141 ML and 880,368 ML following construction of the dam (Ren et al. 2010).

Although the long-term total volume of water moving along the river has not changed as a result of the presence of the dam (and in fact may be less due to evaporation in the dam), the flow conditions have changed with high-level long-duration dam releases resulting in a larger volume of water being driven further into the phreatic

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Table 3 Calibrated parameters used in model runs

Parameter	Borehole BH05	Borehole BH06
a (day^{-1})	0.256	0.036
b	0.009	0.0003
c (m/day)	0.0026	0.0009
Predicted vs actual head R^2	0.76	0.77
Predicted vs actual head slope	0.94	1.09

aquifer at those times. This type of dam release is likely to become more commonplace as catchment managers attempt to control variations to dam water influx caused by climate change. The resultant water-table fluctuations could lead to changes in water quality within an aquifer due to infiltration floodwater with a high content of organic matter into the aquifer system. Other impacts could include incorrect rainfall recharge estimations or erroneous estimates of available long-term aquifer supply, leading to mismanagement of aquifer resources.

Interpretation of the response in the semi-confined lower aquifer to rainfall and dam releases combined with the cross section presented in Fig. 2 and the conceptual

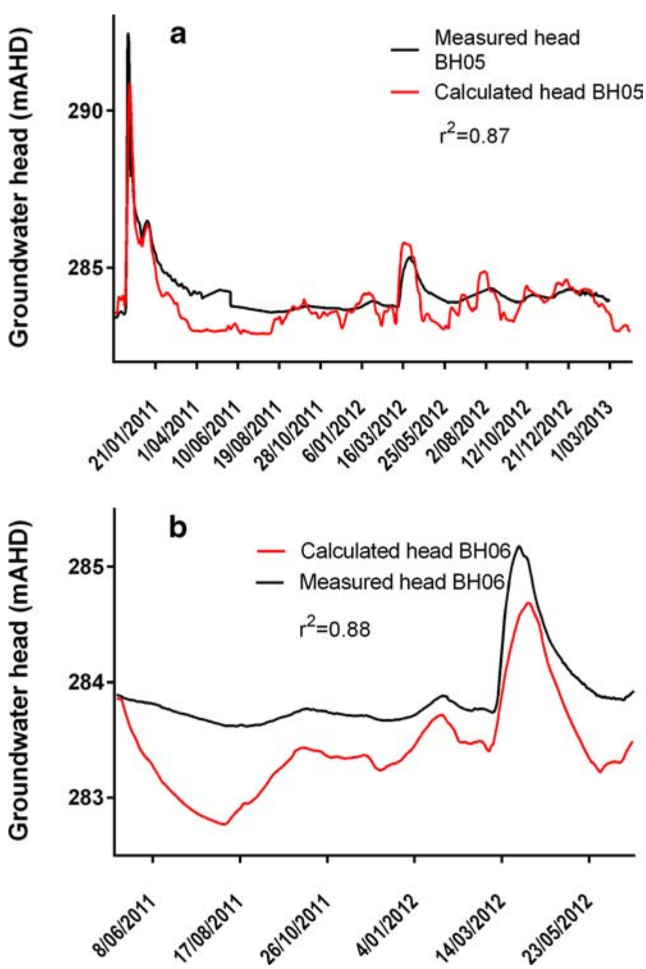

Fig. 6 Groundwater head in two boreholes, showing measured against modelled values: **a** period 13 November 2010 to 28 February 2013 for borehole BH5; **b** period 10 May 2011 to 29 June 2012 for BH6

model in Fig. 8 suggests boreholes BH01, BH02, BH04 and BH07 all show a pressure response to the elevated river level following a long-duration high-volume dam release. The pressure response observed in the semi-confined aquifer is representative of a direct increase in pressure from the river to the clayey aquifer (Chen 2007). When considered in conjunction with storage coefficients of 10^{-3}–10^{-5} for this aquifer, together with the lack of elevated aquifer levels after the river stage has receded, it suggests that the river–aquifer flux caused by river level changes is minimal, as the observed head changes are the result of a pressure response rather than inflow of water into the aquifer.

In contrast, boreholes BH05, BH06 and BH08, which are all in the upper (phreatic) aquifer, show minimal response to dam releases until river level is higher than the groundwater level at the location of these bores. This occurs twice: once during January 2012 and a second time during March 2012 (Fig. 3b, events 1 and 2). In both instances, as the groundwater gradient is reversed, the groundwater level in these wells slowly increases, and this level again subsides slowly after the river level returns to base flow and remains at a higher level than prior to the groundwater gradient reversal (Fig. 3b, event 3). As there was a definite and sustained change in groundwater levels when the river stage returned to pre-release levels, the rise in groundwater level is considered to represent a measureable flux of water from the river to the aquifer.

The observed EC values (Fig. 4) show fluctuations in river EC in response to dam releases and resultant fluctuations in EC in the monitored groundwater wells. Figure 4a shows an inverse relationship between EC and river stage, suggesting that at low river stage, groundwater with a higher EC is discharging into the river increasing surface-water EC. During high river stages (dam release events) the river EC drops and where there is a flux of river water moves into the aquifer, groundwater EC decreases.

From 12 May 2012 through to September 2012, there are two periods of high-volume long-duration dam releases (see events 1 and 2 in Fig. 4). During these periods, the river EC drops by >30 % and a subsequent decrease in EC occurs in boreholes BH05 and BH08.

Event 1 (Fig. 4) shows an increase in river flow following a period of low river level and approximately 25 days later there is a sharp decrease in EC of 20 % at borehole BH05. Between an additional 9–27 days, there is a less clear and smaller decrease in EC of 5 % at borehole BH08. This pattern is replicated at event 2 (Fig. 4) where again there is a significant dam release and a decrease of 40 % in EC in the river. As there had been a period of low flow in the river, EC in boreholes BH05 and BH08 had risen from 0.84 to 1.01 mS/cm and from 0.83 to 0.87 mS/cm respectively, due to groundwater flow of higher EC towards the river. Approximately 25 days after this dam release the groundwater EC at borehole BH05 decreased by 10 %, and for a longer period than event 1, reflecting the volume and duration of the dam release; event 2). Again, the decrease in EC at borehole BH08 is less abrupt

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Table 4 Predicted Flux using Modelled Data

Start date	Finish date	Max flux (m^3/day/m)		Min flux (m^3/day/m)		Cumulative flux (m^3/m)		
		Measured	Predicted	Measured	Predicted	Measured	Predicted	% error
Flux between borehole BH05 and river bank (m^3 per metre of bank, negative values represent water movement away from the river								
01/12/10	27/12/10	1.99	2.73	−6.33	−5.55	2.59	−6.59	139
07/03/12	12/04/12	0.86	0.80	−4.42	−1.49	−86.99	−1.62	5270
01/03/12	30/04/12	1.63	0.97	−2.06	−1.49	0.98	−0.48	304
25/11/12	18/01/13	0.25	0.28	−0.56	−0.29	−11.32	−1.73	554
07/09/11	01/12/11	0.81	0.32	−0.35	−0.22	12.17	1.11	996
12/04/12	30/04/12	1.63	0.97	0.65	−0.25	17.98	2.60	592
23/06/11	20/08/11	0.88	0.17	0.17	−0.34	40.06	−2.19	1929
Flux between borehole BH06 and BH05								
07/03/12	12/04/12	0.0089	0.85	−0.95	−4.42	−14.18	−86.99	84
01/03/12	30/04/12	0.12	2.34	−0.95	−4.42	−12.91	−62.55	79
07/09/11	01/12/11	0.02	0.25	−0.06	−1.49	−1.16	−58.97	98
12/04/12	30/04/12	0.12	2.34	0.006	0.85	1.42	24.98	94
23/06/11	20/08/11	0.11	0.18	0.05	−1.28	5.78	−20.58	128

and can be seen to commence approximately 10 days after the decrease is noted in borehole BH05, while the largest decrease at borehole BH08 is seen approximately 27 days after the initial decrease in borehole BH05. Event 3 (Fig. 4) is a short dam release after a short period of low river flow, which lowers the river EC by 25 %. While this short fluctuation can be seen in borehole BH05 (increasing the EC by approximately 6 % approximately 26 days after the river fluctuation occurred), the event is not of sufficient duration or volume to be observed at borehole BH08.

Based on the observed EC variations, there is a lag of approximately 26 days for an EC response to reach borehole BH05. Due to increasing advective transport distances, there is a linked but smaller response at borehole BH08 commencing approximately 35 days after the event and continuing until approximately 52 days after the river-stage-fluctuation event. The viability of this lag period has been calculated using field data. Calculations have assumed that the EC response shown in Fig. 4 is due to the advective transport of dissolved ions driven by the change in gradients between aquifer and river. The calculations also assume no pre-existing EC gradients within the alluvium and that the path of flow is a direct line from the river towards Boreholes BH05 and BH08. The groundwater linear flow rate required to result in the observed response 130 m from the river (borehole BH05) within 26 days can be estimated to be 5 m/day based on Eq. (8):

$$q = \frac{X}{t} \tag{8}$$

where X is the distance (m) and t is time (days). Ancillary measurements of hydraulic conductivity indicated a value of approximately 34 m/day (BH05). During event 2, the groundwater gradient between the river and borehole BH05 is approximately 0.0008 m/m. Using an advective transport flow equation (Eq. 9; Fetter 1999):

$$V_x = \frac{K}{n_e} \times \frac{dh}{dl} \tag{9}$$

where V_x is the average linear velocity (m/day), K is hydraulic conductivity (m/day), n_e is the effective porosity and dh/dl is hydraulic gradient (m/m). Applying this equation to a hydraulic conductivity of 34 m/day, a gradient of 0.0008 m/m and an effective porosity of 0.1 (estimated value which could range from 0.2 to 0.01), the linear velocity would be approximately 0.3 m/day. This is a magnitude below the required 5 m/day and could be explained by variations in the aquifer characteristics such as effective porosity, aquifer thickness or advective flow paths. It is also possible that part of the reason is pre-existing low EC water and, hence, EC gradients in the alluvium from previous events which are leading to an

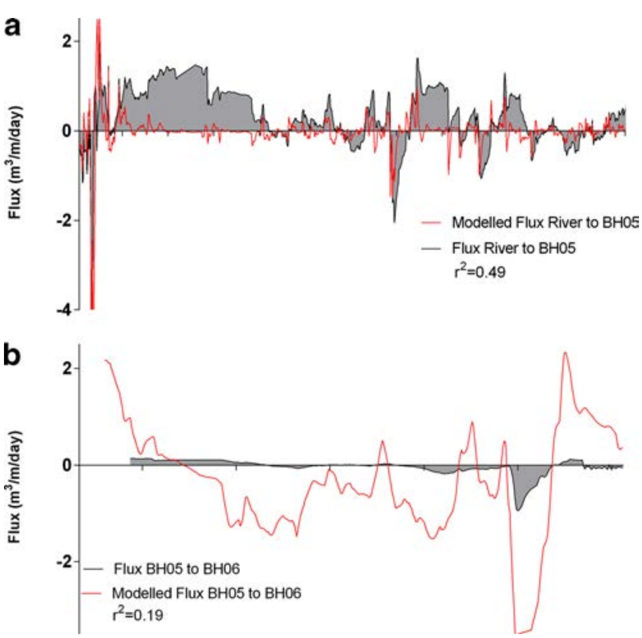

Fig. 7 Modelled and calculated flux of groundwater movement **a** between the river and borehole BH5 for the period 13 November 2011 to 28 February 2013, and **b** between boreholes BH5 and BH6 for the period 10 May 2011 to 29 June 2012. *Negative values* represent movement away from the river

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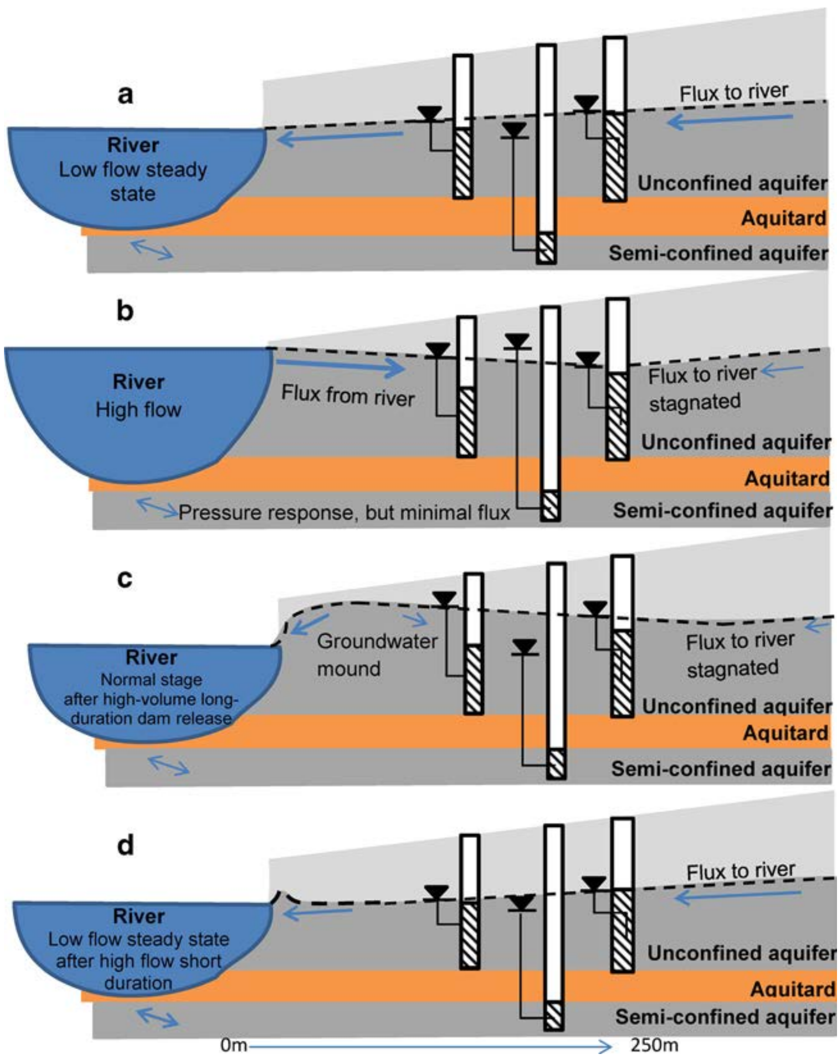

Fig. 8 A conceptual model representing the status of stream–aquifer interactions as a function of volume and duration of dam releases. **a** Prior to the dam release, the river has low flows and it is in a steady-state condition with the groundwater system. During this period, the direction of the flux is from the aquifer to the river (gaining condition). **b** During high-volume dam releases, the river stage increases and the direction of the flux is towards the aquifer system (losing condition). During these large events, the lateral extent of surface-water/groundwater interaction increases and it can extend up to 330 m, as has been observed at this site. **c** If high-volume dam releases continue for long durations, a groundwater mound will develop in the vicinity of the river. **d** Moderate-volume short-duration dam releases may result in an increase in river flux to the groundwater system for a short period of time but overall the main direction of the flux is from the aquifer to the river

overestimation of the groundwater flow rates (by Eq. 8); however insufficient data has been collected by this study to provide a detailed explanation.

The groundwater EC data suggest that advective transport of salts between groundwater and river water is influencing the alluvium water quality as far as 330 m from the river bank and this is primarily due to the flux of water from the river into the aquifer. This flux is a result of long-duration high-volume dam releases with the flux and distance of impact directly related to the duration and volume of the dam release. In instances where river water may contain contaminants, the potential for groundwater quality to become impacted during this process is a management risk. As the phenomenon is characterised by long-duration dam releases, the recharging river waters are held within the aquifer for an extended period of time (which is unlikely to occur in an unregulated system). The presence of recharging river water in the aquifer for long periods of time increases the chemical reaction times. The resulting changes in water quality may have a beneficial or a negative impact on the aquifer and could include changes to redox state, dissolved oxygen concentration, degradation of organic contaminants, nitrate reduction, dissolution of minerals or sorption of organic compounds onto organic material within the aquifer. The extent and level of such an impact has not been addressed in detail during this study, but presents an important topic for future consideration.

Based on the field data, the river–aquifer system is one for which river stage and flow is controlled predominantly by dam releases. Average rainfall recharge is thought to have minimal influence on the river stage (although heavy

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rainfall events will, through the release practice, influence the timing of dam releases). At least two alluvial aquifers that include the upper unconfined aquifer and the lower confined aquifer have been identified. Both aquifers are influenced by river stage changes, which in turn are generally controlled by dam release events. The confined aquifer shows a pressure response to river stage changes, but this response shows minimal lag and corresponds almost directly to river stage changes. River flux into this confined aquifer due to river stage changes are therefore thought to be minimal. While it is also possible that the observed response is representative of a high flux scenario, this is considered unlikely due to the rapid recession of groundwater correlating with recession of the river stage; this concept has not been investigated in detail in this study. The upper aquifer receives a much larger river–aquifer flux and resulting increased river stage, especially during periods of long-duration high-volume dam releases. The flux is driven by a reversal of the groundwater gradient in the unconfined aquifer, with the accumulated flux being a function of the height of the river stage and the duration of the river stage change.

Estimated river–aquifer flux

It can be seen in Fig. 5 that, during low-flow events, the exchange flux remains positive and with up to 40 m^3 per metre of bank of water moving from borehole BH05 towards the river. This would suggest that the groundwater system drains towards the river, with fluctuations in the river stage as a result of dam releases capable of reversing this. Using the 61-day 8,000-L release event as an example, it can be seen from Fig. 5 and Table 2 that, although the river level varied during the release period, resulting in variations between positive and negative flux in the unconfined aquifer immediately adjacent to the river, there remained a persistent negative flux within the unconfined aquifer up to 330 m from the river over a 61-day period.

Head estimation and predicted recharge and flux

The aim of using modelled head levels to predict river–aquifer flux was to determine if the effects of future dam release events on river–aquifer flux could be accurately modelled and predicted using only starting conditions and the anticipated dam release values. Predictive data such as this could be utilised by catchment managers and down-gradient operations utilising groundwater from influenced aquifers, e.g. irrigators could plan increased groundwater extraction based on expected increased aquifer storage during a high-volume dam release.

The head estimation undertaken using the methodology presented in Criss and Criss (2012) resulted in head predictions for boreholes BH05 and BH06 with average errors of 0.29 m (borehole BH05 over a 769-day period) and 0.40 m (borehole BH06 over a period of 413 days). Figure 7a,b shows that the head predictions generally fluctuate around the observed head. Head predictions are more closely related to the river stage variability, due to the model bias to this term, whereas observation data shows that such fluctuations have been significantly diminished in boreholes BH05 and BH06.

The flux and recharge calculated using the head prediction model were found to be significantly different to the actual calculated recharge and flux, with errors generally in excess of 100 %. The average errors of 0.29 and 0.40 m for head estimation in boreholes BH06 and BH05 (Fig. 6a,b) could be considered acceptable for long-term head prediction. However, the consistent bias of the predicted heads compounds errors when applied to gradient calculations used in the volumetric flow equation. Based on the calibrated weights of the variable parameters, the model is almost completely driven by river stage fluctuations with minimal influence from precipitation driven recharge events. The predicted heads for boreholes BH05 and BH06 show a high variability similar to the river stage, whereas the observed head changes in these areas are much more subtle. The inability of the model to account for time lag and dissipation in head response, which becomes increasingly important with distance from the river, is the likely cause of error in the head prediction model. Errors in head prediction were also expected as a result of inputs from overland flow, although this was not significant compared to the identified time lag and dissipation issues.

Generalised conceptual model

A generalised conceptual model of groundwater river interactions in response to high-volume long-duration dam releases is presented in Fig. 8. The model is considered generally applicable to unconfined aquifers in direct connection to river systems, and is dependent on the presence of an up-gradient dam system which controls the river stage. During normal flow (average flow in regulated river maintaining environmental flow), the river is a gaining system with groundwater moving into the river. During short-duration high river stages (Fig. 8b,d) resulting from flood or short duration dam releases, there is some flux of river water into the aquifer. This is a function of the event duration, and where events last less than 5 days, the lateral extent of river water into the aquifer is expected to be less than 30–40 m. The high-volume long-duration dam release event (Fig. 8b,c) is characterised by a high river stage, maintained for a long period of time (in this example, events of longer than 26 days with a consistent increase in river head of greater than 2.5 m were measured). The groundwater gradient in the immediate vicinity of the river is reversed, and assuming a conservative hydraulic conductivity of 10 m/day, the river recharge would be expected to reach at least 250 m from the river. The scenario is dependent on the hydraulic conductivity and connectivity of the aquifer. The relationship between the river level during the event and the pre-existing groundwater gradients also plays an important role in the process. However, this case study is considered relevant to the majority of alluvial aquifers flanking river

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systems and, in many instances, the hydraulic conductivity could be expected to be up to 30 m/day, potentially resulting in zones of influence extending up to 800 m from the river.

Conclusions

A conceptual model of river–aquifer interaction with river stage influenced by dam releases was developed. The conceptual model was used to identify the periods with high fluxes from the river to the aquifer and to determine the required conditions for aquifer recharge/bank-storage to occur. The recharge/bank-storage events were quantified using a volumetric flow equation.

Evidence for recharge/bank storage to the aquifer was found in groundwater wells as far as 330 m from the river and was dependant on reversal of the groundwater gradient. These conditions were only met during high-volume (>8,000 ML/day) long-duration (>26 days) dam release events. These events are a unique anthropogenic occurrence and it is unlikely that the natural system could reproduce the resulting river aquifer interactions.

The volume of flux identified is potentially a significant source of flow loss and aquifer recharge during managed release scenarios that should be considered by water managers attempting to monitor conjunctive water uses or implement management-modelling systems such as multiple criteria analysis or multi-scale modelling. In river systems where controlled dam releases have contributed to stagnant or saline aquifers, it is possible that the effects of these dam releases could be utilised as a flushing mechanism for the aquifers.

Testing of an analytical model developed by Criss and Criss (2012) showed head estimations comparable to observed results and the model was considered capable of providing long-term head prediction. However, coupling of the head estimation model with a volumetric flow equation failed to adequately predict flux.

Overall, the study has identified the significant influence that long-duration high-volume dam releases have on river–aquifer interactions. The long-term impact of these events on aquifer health and sustainability remains to be considered in detail. As the effects of climate change become more apparent and population pressures increase, intelligent, effective and sustainable water-resource management will become increasingly important. Recognition of the effects of long-duration high-volume dam releases on river–aquifer interaction is a critical element of understanding the larger interconnected system. While this study has identified and undertaken preliminary estimations of the influence on recharge/bank storage and river–aquifer flux, the actual chemical and physical influences of these releases on aquifers is poorly characterised. Future investigations should include assessment of dam releases in different river systems and an assessment of the impact of the increased recharge/bank storage on the natural aquifer flow, i.e. whether stagnation occurs and whether this influences the chemical and biological properties of the groundwater.

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