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Abstract: Dewatering associated with mining below water table to achieve dry mining conditions may exert significant pressure on water balance in terms of lowering the water table and change in the dynamics of interactions between surface water and groundwater. The discharge of surplus mine water into ephemeral streams may also affect the water balance, by elevating groundwater levels and altering the exchange rate between streams and underlying aquifers. However, it is unclear whether volumes and recharge processes are within the range of natural variability. Here, we present a case study of an ephemeral creek in the semi-arid Hamersley Basin of northwest Australia that has received continuous mine discharge for more than six years. We used a numerical model coupled with repeated measurements of water levels, chloride concentrations and the hydrogen and oxygen stable isotope composition ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) to estimate longitudinal evapotranspiration and recharge rates along a 27 km length of Weeli Wolli Creek. We found that chloride increased from 73 to 120 mg/L across this length, while $\delta^{18}\text{O}$ increased from -8.2‰ to -7.00‰. Groundwater is directly connected to the creek for the first 13 km and recharge rates are negligible. Below this point, the creek flows over a highly permeable aquifer and water loss by recharge increases to a maximum rate of 4.4 mm/d, which accounts for ~65% of the total water discharged to the creek. Evapotranspiration losses account for the remaining ~35%. The calculated recharge from continuous flow due to surplus water discharge is similar to that measured for rainfall-driven flood events along the creek. Groundwater under the disconnected section of the creek is characterised by a much lower Cl concentration and more depleted $\delta^{18}\text{O}$ value than mining discharge water but is similar to flood water generated by large episodic rainfall events. Our results suggest that the impact of recharge from continuous flow on the water balance of the creek has not extended beyond 27 km from the discharge point. Our approach using a combination of hydrochemical and isotope methods coupled with classical surface flow hydraulic modelling allowed evaluation of components of the water budget otherwise not possible in a highly dynamic system that is mainly driven by infrequent but large episodic floods.

**Evaluating recharge to an ephemeral dryland stream using a hydraulic model
and water, chloride and isotope mass balance**

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Abstract

Dewatering associated with mining below water table to achieve dry mining conditions may exert significant pressure on water balance in terms of lowering the water table and change in the dynamics of interactions between surface water and groundwater. The discharge of surplus mine water into ephemeral streams may also affect the water balance, by elevating groundwater levels and altering the exchange rate between streams and underlying aquifers. However, it is unclear whether volumes and recharge processes are within the range of natural variability. Here, we present a case study of an ephemeral creek in the semi-arid Hamersley Basin of northwest Australia that has received continuous mine discharge for more than six years. We used a numerical model coupled with repeated measurements of water levels, chloride concentrations and the hydrogen and oxygen stable isotope composition ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) to estimate longitudinal evapotranspiration and recharge rates along a 27 km length of Weeli Wolli Creek. We found that chloride increased from 73 to 120 mg/L across this length, while $\delta^{18}\text{O}$ increased from -8.2‰ to -7.00‰. Groundwater is directly connected to the creek for the first 13 km and recharge rates are negligible. Below this point, the creek flows over a highly permeable aquifer and water loss by recharge increases to a maximum rate of 4.4 mm/d, which accounts for ~65% of the total water discharged to the creek. Evapotranspiration losses account for the remaining ~35%. The calculated recharge from continuous flow due to surplus water discharge is similar to that measured for rainfall-driven flood events along the creek. Groundwater under the disconnected section of the creek is characterised by a much lower Cl concentration and more depleted $\delta^{18}\text{O}$ value than mining discharge water but is similar to flood water generated by large episodic rainfall events. Our results suggest that the impact of recharge from continuous flow on the water balance of the creek has not extended beyond 27 km from the discharge point. Our approach using a combination of hydrochemical and isotope methods coupled with classical surface flow hydraulic modelling allowed evaluation of components of the water budget otherwise not possible in a highly dynamic system that is mainly driven by infrequent but large episodic floods.

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1. Introduction

Accurate quantification of recharge rates and dynamics, as well as understanding of the key processes that determine surface-groundwater interactions, underpins optimal management and protection of water resources, particularly in more arid (dryland) regions. The recharge into ephemeral and intermittent streams in dryland regions are often complex owing to a high degree of natural variability in hydrological conditions, significant evaporative loss and often diverse geological landscapes (Morrice et al., 1997; Wroblicky et al., 1998; Anderson and Acworth, 2009; Skrzypek et al., 2013). For example, rivers and creeks in the subtropical, semi-arid Hamersley Basin of northwest Australia are characterised by episodic flows that occur in response to intense rainfall events associated with cyclones or tropical lows, coupled with dissected terrain and a largely impermeable geology, particularly at the stream headwaters (Dogramaci et al., 2012). Peak surface flow rates generated from some of these ephemeral rivers and creeks can reach thousands of cubic metres per second after such events, which then decline over weeks to months depending on patterns of supplementary rainfall and the nature of the catchment in terms of size, recharge through the regolith and topography (Ruprecht and Ivanescu, 2000). Mean annual rainfall deficits based on pan evaporation across the basin also often exceed 2500 mm/yr (Charles et al., 2013). These factors contribute to high spatial and temporal heterogeneity of recharge-discharge mechanisms across any one catchment, which in turn presents considerable challenges for estimating rates and volumes of recharge.

In general, the magnitude of recharge from ephemeral streams to groundwater aquifers is dependent on the amount of water infiltrating into the streambed as a flood wave from an episodic rainfall event progresses in the downstream direction (Vasquez-Sune et al., 2007; Noorduijn et al., 2014). The rate of recharge depends on a number of factors including hydraulic properties of stream sediments and the underlying aquifer as well as topography, hydraulic gradient, temporal variation in precipitation, local groundwater flow patterns and climate (Cey et al., 1998, Oxtobee and Novakowski, 2002, Kumar et al., 2009; Jiménez-Martínez et al., 2013). The hydraulic connection

between surface water and underlying aquifers also has a direct effect on the rate of recharge (Brunner et al., 2011). Given these complexities and the highly unpredictable nature of stream flow in arid and semi-arid environments, the infiltration-recharge process for ephemeral streams has generally been calculated using a combination of approaches, including a range of analytical and numerical models (e.g., Abdulrazzak and Sorman, 1983; Morel-Seytoux et al., 1990; Lamontagne et al., 2014). However, numerical models on their own are of limited usefulness because multiple scenarios can be considered from the same input data (Moore and Doherty, 2006). Alternatively, a combination of hydrochemical tracers and isotope water mass balance approaches may provide greater insight into recharge processes and have been applied extensively to quantify the rates of groundwater inflow to rivers (Cook et al., 2003); distinguish sources of groundwater recharge (Blasch and Bryson, 2007); and calibrate numerical models (Taniguchi et al., 1999).

Major dissolved ions, particularly chloride, and stable isotopes of water, can be used to constrain numerical models, thus helping to develop a more realistic understanding of the study system. Chloride is a conservative tracer as it is primarily derived from rainfall, is very soluble and does not precipitate out of solution unless in very high concentrations (>300 g/L; Hem, 1985), which means it is an ideal indicator of evaporation and evapotranspiration in surface and groundwater (Skrzypek et al., 2013). The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ signatures of water also become progressively more enriched with increasing evaporation (Clark and Fritz, 1997). Transpiration losses have generally been assumed to have relatively little influence on the stable isotope composition of the water sources taken up by plants as uptake and transport within most plants occurs without isotope fractionation (Flanagan et al., 1991; Farquhar and Lloyd 1993). While recent direct measurements of transpired $\delta^{18}\text{O}$ have demonstrated that transpired water is often not in an isotopic steady-state, especially under highly variable environmental conditions (Song et al., 2013), this is mainly of concern to partitioning ecosystem water fluxes over short time intervals. Over longer timeframes, the relative contribution of evaporation versus transpirational losses may theoretically be resolved from observing the variation in chloride and isotope composition of the

source waters (Herzceg et al., 2001; Cartwright et al., 2012). The usefulness of conservative tracers in hydrology also relies on the assumption that they move at the same speed as the water body and their concentration fluctuations are dependent only on evaporation and mixing (Drever, 1988; Kirchner et al., 2010).

Groundwater sustains agriculture, industry and domestic needs in many parts of the world, particularly in dryland regions where surface water is limited. Consequently, most studies of recharge to groundwater in arid environments have focused on reducing the risks of declining groundwater levels associated with abstraction. However, there have been fewer studies of how ephemeral systems may respond to artificial recharge and potential rising groundwater levels. Here, we investigate the influence of continuous discharge of surplus mining water on the overall water budget of Weeli Wolli Creek, a naturally ephemeral, losing stream in the Hamersley Basin of northwest Australia. Continuous discharge into Weeli Wolli Creek since 2007 has created permanent pools in areas that previously held water only for brief periods of time, with likely increased recharge to surrounding groundwater systems. We sought to estimate evaporation and transpiration losses from Weeli Wolli Creek and ultimately recharge to the underlying aquifer as a basis for understanding stream-aquifer interactions and assessing potential impacts on the water balance of the underlying groundwater.

2. Catchment characteristics

2.1 Climate

The climate of the region is hot and semi-arid, with a loosely bimodal rainfall distribution. Most rainfall occurs in the austral summer between December and March, with occasional winter rainfall. The average long-term rainfall at the site (1937-2013) is 250 mm, although rainfall varies considerably from the average from year to year. The largest recorded single rainfall event associated with a cyclone resulted in >500 mm of rainfall in 24 hours at Newman weather station

(December 2002, www.bom.gov.au). In the summer months (December to March) the average maximum temperature is often over 40°C, while during the winter months it falls to about 25°C.

2.2 Physiography

The Weeli Wolli Creek catchment covers an area of approximately 4,000 km². The upper catchment is characterised by relatively wide, flat plains, surrounded by rugged hills of outcropping Banded Iron Formation (BIF). Weeli Wolli Creek and its tributaries drain in a northeast direction, converging at a narrow valley system approximately 30 km down gradient from the headwaters (Fig. 1). The lower catchment downstream of this valley joins with Marillana Creek to the west and drains on to a wide alluvial floodplain (20 to 60 km wide), with a meandering channel that can alter its course following large flood events (Fig. 1).

Upper Weeli Wolli Creek (section A; Fig. 2) is underlain by the relatively impermeable Weeli Wolli Formation for 12 km extending from the upper outlets of the mine water discharge (discharge area 1-DP1) to the lower discharge point (discharge area 2-DP2). The impermeable nature of the formation has been established from prior drilling, which also demonstrated a K value two orders of magnitude less than the CID aquifer (Kirkpatrick and Dogramaci, 2010). This upper section A contrasts to the lower part of the creek (section B, Fig. 2), which is underlain by a Palaeochannel Iron Deposit (CID) aquifer that is characterised by a relatively high hydraulic conductivity ($K \sim 10$ m/d; Kirkpatrick and Dogramaci, 2010). The narrow ~500 m wide and 100 m thick) CID aquifer is hosted within the impermeable Weeli Wolli Formation for most its length. The Brockman Iron Formation ($K \approx 5$ m/d) underlies the lower part of section B, before the creek leaves the Hamersley Ranges and drains into the Fortescue Valley. In general, the creek alluvium is comprised of inter-bedded layers of clays to sands and gravels to cobbles. The coarse alluvium of the creek bed is highly conductive ($K \approx 1000$ m/d), resulting in rapid recharge to the underlying alluvium during creek flows (Peck and Associates, 1995). However, the the volume of the creek bank storage is relatively restricted and negligible as the thickness of alluvium sediments ranges

from ~ 5 m along section A to 40 m along section B. As the creek flows from section A to section B the depth of the unsaturated zone increases (measured water levels along the creek) and the hydrologic regime transitions from a connected to a disconnected creek (Brunner et al., 2009). The transition coincides with where the creek enters the area underlain by the permeable CID aquifer (Fig. 1, 2). The water table depth increases progressively downstream from 0 m below ground level at the transition point to 29 m at the mouth of the Fortescue Valley (Fig. 3).

2.3 Hydrology and flood frequency

Total rainfall records were obtained from Hope Downs (10 km up gradient of DP1) and runoff records from Waterloo Bore gauging station (4 km downstream from the last sampling point in section B-Australian Bureau of Meteorology, www.bom.gov.au). Prior to mining in the region, except for a relatively short section downstream from Weeli Wolli spring in section A, the natural creek condition was ephemeral and only flowed during and immediately after high volume and intense rainfall events, usually in the summer. Runoff in the Weeli Wolli catchment occurs only for precipitations over 36 mm per day (referred to as the mean rainfall threshold for runoff commencement; Charles et al., 2013).

Flood frequency analysis indicates large flooding events ($\geq 100 \text{ m}^3/\text{s}$) have occurred once every two to three years at Waterloo Bore gauging station (Australian Bureau of Meteorology, www.bom.gov.au), with major floods occurring across the basin at least every four years (Rouillard et al., 2014). A TUFLOW 2D hydraulic model suggests at section B of the Weeli Wolli Creek a peak stream flow of $110 \text{ m}^3/\text{s}$ for a 50% AEP (Annual Exceedance Probability) rainfall event (50 mm/9 hour). A 10 % AEP rainfall event (110 mm/12 hour) can produce $1160 \text{ m}^3/\text{s}$. During these events Weeli Wolli Creek has continuous flow over its entire 60 km length.

The ephemeral nature of Weeli Wolli Creek has been altered over recent years due to the continuous discharge of $0.6 \text{ m}^3/\text{s}$ surplus water from mine dewatering. Water is discharged to the creek at two locations; the primary discharge (Discharge point 1 – DP1) is located 12 km upstream

from the secondary discharge (Discharge area point 2 – DP2) (Fig. 1). The primary discharge (DP1) into Weeli Wolli Creek commenced in December 2007 and the secondary discharge (DP2) commenced in October 2007. A total volume of approximately $22 \times 10^7 \text{ m}^3$ had been discharged from October 2007 to December 2013 ($17 \times 10^7 \text{ m}^3$ from DP1 and $50 \times 10^7 \text{ m}^3$ from DP2). Continuous flow is now visible as surface water for 24 km downstream of DP1, but fluctuates from 24 km to 27 km owing to variability in discharge rates (Fig. 4).

3. Methodology

3.1 Hydrochemical analyses

In June 2010, December 2010 and July 2011, we collected surface water samples from twelve points (WW1-WW20) located at discrete intervals of 1-2.5 km downstream from the primary discharge (DP1) along the creek, encompassing a total length of 24 km (Fig. 2). In addition, five groundwater samples were collected from monitoring bores along Weeli Wolli Creek in sections B (Fig. 2), after first pumping out three casing volumes or after field measurements of EC, pH, Eh and temperature stabilised (INSITU Multi parameter Water Quality TROLL 9500 - data not shown). Details of the bores are provided in Table 3. The longitudinal extent of surface saturation along Weeli Wolli Creek has been recorded by ground-based field observations from February 2007 to May 2012 for a total of 83 observations, whilst discharge rates to the creek have been recorded monthly at discharge points DP1 and DP2 (Fig. 2).

All water samples were analysed for Cl concentration and stable hydrogen and oxygen isotope compositions ($\delta^{18}\text{O}$ and $\delta^2\text{H}$). The Cl concentrations were measured using a Discrete Analyser in Water by Aquakem Cl anions (method Ref. APHA 4500, with a reporting limit of 1 mg/L). The analytical uncertainty is 0.5 mg/L and is based on deionised water spiked with a concentration of chloride range of 25 to 150 mg/L.

Stable hydrogen and oxygen isotope compositions ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) of water samples were analysed at the West Australian Biogeochemistry Centre at The University of Western Australia

using an Isotopic Liquid Water Analyser Picarro L1115-i (Picarro, Santa Clara, CA, USA) as per the method of Skrzypek and Ford (2014). The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values of samples were normalized to the VSMOW international scale (Vienna Standard Mean Ocean Water), following a three-point normalization (Skrzypek, 2013) based on three laboratory standards calibrated against international reference materials using the VSMOW-SLAP scale (Coplen, 1996). All results are reported in permil (‰) using standard δ -notation. The analytical uncertainty (one standard deviation) was determined to be <0.10‰ for $\delta^{18}\text{O}$ and <1.0‰ for $\delta^2\text{H}$.

3.2. Calculation of total water loss (q)

The total loss from surface water along the saturated section of the creek was simulated using a continuity equation as follows:

$$R = (Q_1 + Q_2 + I) - (E + T + S) \quad \text{Eq.1}$$

where R is the recharge, Q_1 and Q_2 are measured flow rates at DP1 and DP2, E is evaporation calculated from the isotope mass balance, T is transpiration, calculated from the difference between chloride and isotope mass balance, S is storage (i.e., bank storage) and is assumed to be negligible, and I is runoff from rainfall and flooding events and assumed to be zero for dry periods (no run off).

The total loss parameter q (recharge and evapotranspiration) was calculated using a kinematic wave routing model for channels (Lighthill and Whitman, 1955). The kinematic wave approximation is a simplification of the Saint-Venant equations (Saint-Venant, 1877), which assumes that the momentum equation is reduced to $S_f = S_0$, where S_f is the friction slope and S_0 is the stream bed slope. In a losing (disconnected) stream the kinematic approximation can be described by the continuity equation (Chow et al., 1988):

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} + q = 0 \quad \text{Eq.2}$$

where Q is the total flow along the channel, A is the cross section of the channel and q is the total water loss per unit length. The rating curve of the channel can be described as:

$$A = \alpha Q^\beta \quad \text{Eq.3}$$

where α and β are parameters determined from the Manning equation (Manning, 1891). In particular, for a rectangular channel $\alpha = \left(\frac{nP_3^2}{\sqrt{S_0}}\right)^{\frac{3}{5}}$ and $\beta = 0.6$ where P is the wetted perimeter of the cross section. Equation 1 can be modified to include the α and β parameters as follows:

$$\frac{\partial Q}{\partial x} + \alpha\beta Q^{\beta-1} \frac{\partial Q}{\partial t} = -q \quad \text{Eq.4}$$

Equation 3 has been solved numerically with a finite difference linear scheme (Mahmood and Yevjevich, 1975; Overton and Meadows, 1976) as follows:

$$Q_i^j = \frac{\left[\frac{\Delta t}{\Delta x} Q_i^{j-1} + \alpha\beta Q_{i-1}^j \left(\frac{Q_i^{j-1} + Q_{i-1}^j}{2} \right)^{\beta-1} - q_i \Delta t \right]}{\frac{\Delta t}{\Delta x} + \alpha\beta \left(\frac{Q_i^{j-1} + Q_{i-1}^j}{2} \right)^{\beta-1}} + S_i^j \quad \text{Eq.5}$$

where i and j are the spatial ($\Delta x = 100\text{m}$) and temporal ($\Delta t = 1 \text{ day}$) discretisation indices, S_i^j is the additional discharges into the creek at the distance $x = i \Delta x$ and time $t = j \Delta t$, and q_i is the loss per unit length assumed constant for each discretised i -th segment of the channel. The parameters α and β were calculated with the assumption of a rectangular section of 25 m width (B), constant streambed slope of $S_0 = 0.01$, constant manning coefficient of $0.088 \text{ m}^{1/3}/\text{s}$ (n) and a wet perimeter approximately equal to B. Equation 4 can further be simplified by aggregating discretised sections with similar hydraulic characteristics that influence the evapotranspiration and recharge to underlying aquifer, i.e. using only two discrete intervals (q_1 and q_2) for each of the two different sections of the creek.

3.3 Calculation of evaporation and transpiration using $\delta^{18}\text{O}$ and chloride mass balance

Water loss due to evaporation for each of the sampling intervals was calculated using a modified Craig-Gordon model (Gibson and Reid, 2010) as the ratio of evaporation to inflow (E/I) (Eq. 6). The E/I ratio describes what percentage of inflowing water is evaporating as it passes between consecutive sampling points along the creek. The total evaporative loss was estimated as

the cumulative percent of losses for each sampling interval (Fig 2), based on the difference in the $\delta^{18}\text{O}$ between the start and end of each section.

$$E/I = \left[\frac{(\delta_L - \delta_P)}{(\delta^* - \delta_L) \times m} \right] \quad \text{Eq. 6}$$

where δ_P represents inflowing water and δ_L represents outflowing water. The δ^* parameter is the limiting isotope enrichment factor (Eq. 7) calculated using: humidity (h), isotope composition of moisture in ambient air (δ_A) and total enrichment factor (ϵ) :

$$\delta^* = \frac{h \times \delta_A + \frac{\epsilon}{1000}}{h - \frac{\epsilon}{1000}} \quad \text{Eq. 7}$$

The m enrichment slope factor is defined by Eq. 8:

$$m = \frac{h - \frac{\epsilon}{1000}}{1 - h + \frac{\epsilon_k}{1000}} \quad \text{Eq. 8}$$

where h is air humidity given as a fraction, ϵ_k is the kinetic fractionation and ϵ is the total fractionation factor (Gibson and Reid, 2010)). The total water volume loss was estimated using Cl mass balance, which was calculated for each of the intervals of the creek using the following equation:

$$V_L = (1 - Cl_P / Cl_L) \times 100 [\%] \quad \text{Eq. 9}$$

where V_L is the water loss calculated as a percentage, Cl_P and Cl_L are Cl concentration in inflowing and outflowing water for each sampling interval respectively. The isotope model estimates evaporative losses only, while chloride model estimates total water losses from both evaporation and transpiration. The difference between volume losses from isotope model and chloride model reflects water losses due to transpiration. The predicted Cl increase along creek line just due to evaporation, when transpiration would be zero, was calculated as Cl_e follow equation:

$$Cl_e = Cl_P / (1 - E/I) \quad \text{Eq.10}$$

3.4. Estimation of groundwater recharge

We have estimated recharge rates to the underlying aquifers for the two main sections of the creek (A and B) based on the calculated evapotranspiration rate and the total water loss along the

saturated section. First, we calculated the Cl concentration for each consecutive sampling interval assuming that the Cl concentration increases proportionally to the inverse of the percentage of the remaining water:

$$Cl(x) = \begin{cases} \left(\frac{xq'_1}{Q_1} + 1 \right) Cl_1, x \leq L \\ \left(\frac{(x-L)q'_2}{Q_1-Lq_1+Q_2} + 1 \right) Cl_{L+}, x > L^+ \end{cases} \quad \text{Eq. 11}$$

where Cl_1 and Q_1 are the initial chloride concentration and the discharge volume into the creek, whilst Cl_{L+} is the chloride concentration at discharge at the beginning of section B ($x = L^+$) calculated as the weighted average of the modelled chloride concentration prior to DP2 ($x = L$) and the measured chloride concentration at DP2 (Cl_2):

$$Cl_{L+} = \frac{(Q_1-Lq_1)Cl(L)+Q_2Cl_2}{Q_1-Lq_1+Q_2} \quad \text{Eq.12}$$

Discharge Q_1 was $1 \text{ m}^3/\text{s}$ and Q_2 was $0.28 \text{ m}^3/\text{s}$. Cl_1 was 73 mg/L and Cl_2 was 100 mg/L .

q'_1 and q'_2 are the evapotranspiration rates for the two sections of the creek that were calibrated using the measured Cl concentrations in each section.

Second, we subtracted the calculated evapotranspiration rates (q'_1 and q'_2) from the total water losses calculated by the hydraulic model (q_1 and q_2) model to estimate the infiltration rates.

3.5. The water table fluctuation (WTF) method to calculate recharge

The water table fluctuation (WTF) method was used to calculate groundwater recharge (Healy and Cook, 2002) after flood events to compare with recharge rate calculated by Cl and isotope mass balance approaches. The WTF method in unconfined aquifers is based on calculating the rate of rise of water level in response to a given rainfall due to recharge water arriving at the water table. The method is only valid for unconfined aquifers and the underlying assumptions are that monitoring bores are screened (see Table 3) at or close to water table, and the water arriving at the water table goes immediately into storage (Healy and Cook, 2002). Recharge is calculated as:

$$R = S_y \Delta h / \Delta t \quad \text{Eq. 13}$$

where S_y is specific yield (0.25), Δh is water table level increase at Δt time. In order to minimise uncertainty in Cl and $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values on a flood time scale and consequently of recharge water, extensive sampling of flood water and runoff was made during the main cyclone event in 2012 for several creeks (Coondiner Creek, Fellman et al., 2012; Fortescue River, Dogramaci and Skrzypek, 2014) and Marillana creek (unpublished report). Cl concentration varies during flooding, with relatively high Cl concentrations with the onset of runoff due to mobilisation of dry salt, followed by a gradual decrease to ~ 20 mg/L by the end of flood. However, the calculated average Cl concentration falls within a relatively narrow range (20 mg/L to 50 mg/L). Therefore, we expect that the maximum Cl concentration in floodwater will not exceed the higher end of the spectrum of 45 mg/L.

4. Results

4.1. Modelling and hydrochemistry

We initially assumed that the water loss parameter (q) was constant along the entire creek channel. The calibration was carried out with the code PEST (Doherty, 2005), by changing the loss parameter q with a range of variability between 0.01 and 20 $\text{m}^3/\text{d}/\text{m}$. The model was calibrated by minimising the sum of the square residuals between the calculated distance at which $Q = 0$ and the observed surface water expression, hence the RMS is expressed in km. The observations of the water front (the point where the surface water is no longer visible are expressed in terms of distances at which the water disappears in the channel. Therefore, the flow and recharge at this location are zero. The calibrated q is 3.53 $\text{m}^3/\text{d}/\text{m}$ (Root Mean Square RMS of 4.52 km and a Normalised RMS of 38.8%).

Given the difference in hydrogeological settings between sections A and B (Fig. 2), we then re-ran the water mass balance model using two water loss parameters, where q_1 was assigned to Section A and q_2 to Section B. The calibration of the model as two sections (Doherty, 2005) results

in a q_1 of 0.017 m³/d/m and q_2 of 9.41 m³/d/m (RMS of 2.07 km and a NRMS of 17.8%). The higher rate in section B coincides with the location where the saturated creek crosses and flows over the Channel Iron Deposit (CID) aquifer (Fig. 2).

The calculated saturated lengths of the two-parameter model more closely matched the observed data than the one-parameter model and overall had a stronger statistical fit (Fig. 4). The Cl concentration in discharge water at DP1 and DP2 has been fairly constant since discharge commenced and are 74 ± 2 mg/L and 89 ± 2 mg/L, respectively (Table 1). The Cl concentration consistently increased between the most upstream (WW20) and lower (WWE) sampling points (Fig. 5A, B); by 30 mg/L in June 2010, 34 mg/L in December 2010 and 46 mg/L in July 2011 (Table 1). The differences among sample times mainly reflect varying evaporative losses under different seasonal conditions before mixing with the secondary discharge at the beginning of section B (Cl =100 mg/L). Below the mixing at DP2, Cl increased from 89 mg/L to 120 mg/L (Table 1). The Cl concentration for groundwater from bores 99YJWB01 and 99YJWB04 from 1991 and pre-dating groundwater discharge, ranges from 28 mg/L to 110 mg/L. The Cl concentration trend in bore 99YJWB01 has remained nearly constant over the last 24 years at approximately 40 mg/L, whereas 99YJWB04 has increased gradually, from 40 mg/L in 1991 to 110 mg/L (Fig. 6).

The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of water collected at DP1 were -8.57‰ and -57.5‰ and at DP2 7.92‰ and -54.9 ‰ (Fig. 7A, B). The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of surface water at the end of section B were -7.00‰ and -52.0‰. The isotopic composition of the surface water also showed a gradual enrichment in heavy isotopes (^2H and ^{18}O) following the Local Evaporation Line (LEL) with slopes of 4.60 in June 2010; 4.32 in December 2010 and 5.18 in July 2011. These slopes are consistent with a long-term dataset for the whole Hamersley Basin (5.16 with long-term mean temperature 24°C and RH 31%; Dogramaci et al., 2012) as well as for short-term pan evaporation experiments in the region (4.07 at 29.3°C, RH 23%, and 4.67 at 24.8°C, RH 29%; Skrzypek et al., 2014). The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of groundwater are similar to surface water except at the monitoring bore 99YJWB01, where isotope composition is slightly more depleted (Table 1).

The Cl concentration measurements were used to calibrate the loss parameters due to evapotranspiration (q'_1 for section A and q'_2 for section B; equation 10). The calculated q'_1 value is 0.62 m³/d/m, which is higher than the total loss calibrated with the hydraulic model ($q_1 = 0.017$ m³/d/m). Both results suggest that the aquifer feeds the creek in section A with a rate of 0.60 m³/d/m. In contrast, the calculated q'_2 for section B was 2.66 m³/d/m compared to the total loss of $q_2 = 9.41$ m³/d/m. Therefore, we deduced that the difference of 6.75 m³/d/m is due to infiltration. Transpiration rates were calculated by subtracting water loss calculated using the stable isotope mass balance from that calculated Cl mass balance (Table 2).

Based on the water loss rates calculated from isotope and Cl mass balance, we calculated the percentage of water loss by transpiration, evaporation and recharge (Table 1). The maximum cumulative evapotranspiration loss in section A is 6.8 % of the inflow volume compared to 35 % in section B. In contrast, the total cumulative water loss by evaporation in section A is 6.1 % of the inflow volume compared to 4.0 % in section B. The total water loss due to recharge accounts for 65% in section B, whilst there is negligible recharge in section A. The percentages of evapotranspiration and infiltration calculated with the water loss parameters calibrated using equation 10 were slightly different (68% and 32%, respectively). This difference is because the best fitted representation of equation 10 underestimates the calculated Cl concentration at WWE (24 km down gradient from DP1; Figure 8).

The saturation of the streambed increases longitudinally with time due to continuous flow, which in turn depends on the rate of discharge and hydraulic conductivity of the creek bed (Brunner et al., 2011). At Weeli Wolli Creek, we would expect that recharge increases longitudinally due to the increased thickness of unsaturated zone of the underlying aquifer along the creek. Our study confirms that the change in recharge does not increase gradually but rather rapidly between section A and B. This occurs due to the relatively abrupt transition from a connected stream where the surface water is connected to water table in the 5 m thick sediments (section A), to a disconnected creek where the water table in the underlying thick CID aquifer is separated from the creek by an

unsaturated zone (section B), resulting in rapid increase in the recharge rates (Table 2) and substantial reduction in the flow. The reduction in flow rate also coincides with a dense forest of *Melaleuca* and riparian eucalypts lining the section B of the creek, which in turn results in greater transpirational losses compared to section A, where riparian vegetation is relatively sparse.

4.2. Recharge from artificial discharge compared to flood events

Long-term water table data from the CID aquifer from seven monitoring bores (Table 3) located along section B of the creek (Fig. 3; JSE37, 99YJWB04, D08YJ126, YM119, D08YJ216, D08YJ165 and 99YJWB01) were used to compare recharge derived from the discharge with recharge after rainfall events (Fig. 8). The depth of water table decreased longitudinally down the creek from 4 m below ground surface at bore JSE37 to 29 m down gradient at bore 99YJWB01. Groundwater levels are shown for three dates (25/07/2010, 27/07/2012 and 19/04/2013), a period that captured a major cyclone, illustrating the response of groundwater in the creek to rainfall and flooding events (Fig. 3). We calculated recharge using the WTF method after Cyclone Heidi (8- 13 January 2012). The total rainfall recorded for this period was 225 mm. Maximum discharge of 430 m³/s was recorded on 13 January, 2012 at Waterloo Bore gauging station, 4 km downstream from the last sampling point in section B of Weeli Wolli creek (Australian Bureau of Meteorology, www.bom.gov.au). Flows resulted in significant recharge along the creek, with water table rise ranging from 0.16 m in monitoring bore JSE37 (the mid-part of the creek section B) to 10 m in the monitoring bore D08YJ165 (the bottom part of the creek section B) (Fig. 3 & 9). The rate of recharge in response to Cyclone Heidi increased downstream at bore JSE37 toward bore 99YJWB01 (Fig 9), reflecting the thickness of unsaturated zone under the creek. The first four bores in this sequence (JSE37, 99YJWB04, D08YJ126 and YM119) have a similarly shallow water table ranging from 4.2 m to 6.3 m, whilst the next three bores are characterised by a deeper water table, ranging from 17.6 m to 29 m. Accordingly the recharge rate increases down gradient from < 1 mm/d at bore JSE37 to ~ 20 mm/d at bore 99YJWB01. The positive correlation between the rate of

recharge and thickness of unsaturated zone is due to the potentially large storage capacity associated with relatively deep water table. The recharge rates obtained with the WTF method have been multiplied by a representative recharge width in order to compare them with the recharge volumes per unit length calculated by Cl mass balance (Eq. 11) and water mass balance (Eq. 5). The recharge volumes per unit length calculated by WTF method range between $<0.1 \text{ m}^3/\text{d}/\text{m}$ and $8.4 \text{ m}^3/\text{d}/\text{m}$ (Fig. 9). Assuming a linear variation of the recharge rate along the creek from the location in which the creek is disconnected from the aquifer, we calculated a maximum rate of $11.8 \text{ m}^3/\text{d}/\text{m}$, which is consistent with the maximum rate estimated by the WTF method (Fig. 10).

5. Discussion

Overall, the water, Cl and $\delta^{18}\text{O}$ mass balance models suggest that evapotranspiration over the length of Weeli Wolli Creek accounts for 35% of the total discharge; therefore the remaining 65% is lost via recharge to the underlying CID aquifer. Temporal Cl concentration data from local groundwater, which pre-dates discharge of mine surplus water, illustrates how the recharge from adjacent Weeli Wolli Creek has gradually increased the Cl concentration up to 14 km from the primary discharge location. Combined modelling of water loss with hydrochemical trends of surface water, suggests that the recharge occurs in the lower section of the creek (section B) where it crosses and flows adjacent to the permeable CID aquifer underlain by the permeable Brockman Iron Formation. Although the discharge of surplus water resulted in a significant change to the local natural water balance, no impact on groundwater chemistry has been observed further than 24 km downstream of the discharge outlet DP1 (Fig. 6).

Our results also suggest that the flow and recharge from large and short-lived flood events may be masking the recharge from surplus water in the bottom part of section B where the creek is not connected to the underlying aquifer. The total recorded discharge since 2007 to May 2013 was $\sim 22 \times 10^7 \text{ m}^3$. A comparison of predicted maximum discharge rates from a series of natural rainfall events calculated using rainfall and runoff data and TUFLOW 2D hydraulic models suggests that

the volume of discharge water released into the creek is much less than the volume of water flowing through the system during a flash flood. For example, the large scale flooding after Cyclone Heidi (January 2012) resulted in runoff rate of $> 400 \text{ m}^3/\text{s}$ at the Waterloo gauging station and the accumulation of $> 30 \times 10^7 \text{ m}^3$ of water in the Fortescue Marsh (Fig. 1) as the result of runoff from the upper part of the Fortescue River, including Weeli Wolli Creek (Skrzypek et al., 2013). This single event delivered to the marsh more water than the total discharge from the mine its operation time (2007-2013, $20 \times 10^7 \text{ m}^3$).

The prevailing effect of recharge from flooding events on the water balance of Weeli Wolli Creek over mine discharge is evident when observing long term Cl concentration in groundwater. The data for groundwater from bores 99YJWB01 and 99YJWB04 shows how the infiltration of nearby creek water has affected the groundwater composition in the connected section (section A and upper half of section B). While both 99YJWB01 and 99YJWB04 are located in section B, the thickness of unsaturated zone increases down gradient from $\sim 4 \text{ m}$ at 99YJWB04 to $\sim 29 \text{ m}$ at 99YJWB01 bore. Both bores have an initial Cl concentration of 40 mg/L ; however, the Cl concentration trend in bore 99YJWB01 remains nearly constant with time, whereas 99YJWB04 increases gradually, trending toward the similar Cl concentration of the surplus mine water discharged to the creek 100 mg/L at DP2 (Fig. 11). The much lower Cl concentration in monitoring bore 99YJWB01 and significantly more depleted ^{18}O value ($\delta^{18}\text{O} -8.8\text{‰}$) compared to surface water (Fig. 7A, B) suggests recharge to the aquifer from high intensity rainfall events (Dogramaci et al., 2012; Skrzypek et al., 2013). Previously, we measured Cl concentrations in surface water pools fed by flooding at different parts of the Hamersley Basin; our results suggest a typical range for floodwater of 20 mg/L to 50 mg/L (unpublished data). Consistently lower Cl concentrations in groundwater at 99YJWB01 ($<50 \text{ mg/L}$) compared to discharge mine water (100 mg/L at DP2) suggest the impact of mine discharge has negligible influence on water geochemistry in this part of the creek. Interestingly, mining discharge coincidentally started at the end of the wettest period (1995-2006) over the whole last century (Charles et al., 2013; Rouillard et al., 2014). Therefore, it is

likely the initial recharge of surplus mining water to groundwater occurred when the water table was the closest to the surface it has been for the last 100 years. The discharge of $\sim 22 \times 10^7 \text{ m}^3$ of surplus water since 2007 into Weeli Wolli Creek has resulted in changes to groundwater levels in section B where the CID aquifer underlies the creek and consequently the increase in the rate of groundwater through flow. The groundwater storage calculated for the CID aquifer for section B is approximately $60 \times 10^6 \text{ m}^3$ ($500/2 \text{ m wide} \times 80 \text{ m deep saturated thickness based on } > 10,000 \text{ drill hole records along the CID aquifer; length of CID from intersection of Weeli Wolli creek up to the fluvial plain } 12 \text{ km; and the average specific yield of } 25 \%$). If 65 % of the total discharge volume to date ($22 \times 10^7 \text{ m}^3$) infiltrates into the underlying CID aquifer, this equates to $\sim 13 \times 10^7 \text{ m}^3$ of additional recharge in to Weeli Wolli Creek since 2007. Therefore the aquifer should have been flushed twice since 2007. Given that the rate of lateral flow through the CID aquifer (calculated from Darcy's equation $\sim 10 \text{ mm/d}$) is higher than the average recharge rate (4.4 mm/d), it is not anticipated that the water table should rise (in the monitoring bores along the entire lower 7 km section of the creek). In fact, long-term observations of the monitoring bores within the CID confirm that the response of groundwater is constrained to the saturated stretch of the Weeli Wolli Creek and does not extend further than 8 km downstream from the start of section B (Fig 3).

The results of this study suggest that the storage capacity of the aquifer is much larger at the mouth of the Weeli Wolli Creek draining into Fortescue Valley. Water level rises of as much as 10 m have been observed in the bores located within section B of the creek after intense rainfall and flooding events (Fig 9). The relatively constant (40 mg/L) concentration of Cl since 1998 coupled with the rapid response to floods suggests the dominance of infiltration from flood events on Cl concentration (Fig 6). If the entire section B of the CID aquifer were the natural conduit for through flow, the Cl concentration in groundwater would have been the same as surface water (100 mg/L). This suggests that the CID aquifer at this location is characterized by a much higher transmissivity than the modeled estimate for groundwater storage. Therefore, we hypothesize that the aquifer storage capacity at the mouth of Fortescue Valley is many times larger than $60 \times 10^7 \text{ m}^3$. We thus

propose a new model for this catchment that accounts for the change in hydraulics along its length and which better explains why discharge into Weeli Wolli Creek has little impact on the overall water balance of the aquifer. This model is based on the assumption that the storage of the aquifer will increase dramatically (> ten-fold) at the mouth of Fortescue Marsh fluvial plain comparing to the upper part. In this section the formation underlying the CID is fractured and mineralised Brockman Iron Formation (cross-section C-C', Fig. 11) that is characterised by a much higher transmissivity compared to Weeli Wolli Formation and un-mineralised Brockman Iron Formation underlying the upper part of the catchment (cross-section B-B', Fig. 11). Because the volume of water recharging the aquifer is insignificant compared with the volume of groundwater storage, a change in Cl concentration is not observed.

6. Conclusions

Overall, our study demonstrates that a combination of hydrochemical and isotope methods with classical surface water flow mathematical modelling is useful for confirming hydrogeological models otherwise not possible in regions with extreme climate variability, especially where hydrology is driven by infrequent flash floods. The controlled nature of this study in terms of volume of discharge and the extent of surface water expression and extensive transient sampling of the surface water and groundwater for Cl concentration and $\delta^{18}\text{O}$ and $\delta^2\text{H}$ composition allowed partitioning of components of water balance that otherwise would not have been possible by water mass balance method alone. We conclude that although more than $22 \times 10^7 \text{ m}^3$ of mine water discharge has changed the nature of the connectedness of Weeli Wolli creek to the underlying aquifer in the upper section associated with the impermeable Weeli Wolli Formation, there has been negligible impact on the overall water balance of the broader catchment. Most of the recharge to the groundwater aquifer along Weeli Wolli Creek occurs where the creek flows adjacent or across the CID aquifer. The calculated rates of recharge are also commensurate with large flooding events that occur in the region every few years. Changes in surface water connectivity and sustained periods of

saturation may nevertheless have localised and probably relatively short-term impacts (until the next flood) on the ecology of streams. However, the importance of no or low (surface) flow periods in ephemeral streams to overall stream functioning (from an eco-hydrological perspective) is largely unknown as most studies of the ecohydrological impacts of altered flows have focused on reduced rather than increased flows associated with drawdown. By increasing understanding of the overall water balance, our study shows that management of artificial discharge to ephemeral creeks in dryland environments, at least from a water balance perspective, can be optimised if there is high storage capacity of the underlying aquifer..

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660

- an ephemeral creek in the semi-arid Australia received mine discharge for >6 years
- numerical modelling, Cl and isotope mass balance were used to assess impact
- Cl increased from 73 to 120 mg/L and $\delta^{18}\text{O}$ from -8.2‰ to -7.0‰ along the creek
- water loss by recharge accounts for ~ 65% and evapotranspiration by ~35%
- combined multi-method approach allowed successful calculation of water budget

- 1 Figure 1. Geological map showing, the Weeli Wolli Creek system relative to the Fortescue
- 2 Valley.
- 3 Figure 2. The Weeli Wolli Creek study area, showing sample points, nearby groundwater
- 4 bores and discharge locations (DP1 and DP2). The cross-section shows the variation in
- 5 geology below the creek.
- 6 Figure 3. Cross-section of lower part of the section B (7 km) showing water levels in
- 7 monitoring bores adjacent to Weeli Wolli Creek. Note the lower water table where the creek
- 8 transitions from connected to disconnected state.
- 9 Figure 4. Observed variation in the extent of the surface water expression in Weeli Wolli
- 10 Creek (during the dry seasons) and the calculated water extend front. The single water loss
- 11 parameter scenario $q = 3.53 \text{ m}^3/\text{d}/\text{m}$ for the whole creek. The two-parameter water loss model
- 12 parameter scenario $q_1 = 0.017 \text{ m}^3/\text{d}/\text{m}$ for section A and $q_2 = 9.41 \text{ m}^3/\text{d}/\text{m}$ for section B (Fig.
- 13 2).
- 14 Figure 5. A - Chloride concentration in surface water and groundwater along Weeli Wolli
- 15 Creek. B - Cumulative % chloride increase along the creek, illustrating the change in the rate
- 16 that chloride concentration increases in section B.
- 17 Figure 6. Cl concentration in groundwater at bores 99YJWB01 and 99YJWB04 since July-
- 18 1998. The gradual increase in Cl concentration from 2008 in bore 99YJWB04 is due to
- 19 mixing of recharge (100 mg/L) with groundwater.
- 20 Figure 7. A - Relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in surface water and groundwater from
- 21 Weeli Wolli Creek. B - Local Meteoric Water Line and Local Evaporation Line (yellow line)
- 22 for the Hamersley Basin as defined by Dogramaci et al. 2012 and results from Weeli Wolli
- 23 Creek sampling programs.
- 24 Figure 8. The Cl mass balance model assuming water loss is due to evaporation only. The Cl
- 25 concentration for section A closely follows the observed values. The large difference between
- 26 modelled and observed values in section B is due to water loss by recharge into the
- 27 underlying CID aquifer.
- 28 Figure 9. Hydrographs from the monitoring bores along Weeli Wolli Creek, illustrating the
- 29 change in water level in response to rainfall. Note the water level rise after Cyclone Heidi

30 (Jan 2012). Rainfall data collected from Hope Downs weather station adjacent to the creek
31 (Easting: 714213 and Northing: 7457937).

32 Figure 10. Calculated recharge rates downstream for section B using the Water Table
33 Fluctuation method (WTF) and water, Cl and isotope mass balance for data collected from
34 monitoring bores during a rainfall event in 2012.

35 Figure 11. 3D geology of the mouth of Weeli Wolli creek where it meets the Fortescue
36 Valley, illustrating the variation in hydrogeology and the resultant increase in potential
37 groundwater storage. The two transects highlight the difference between the previous
38 conceptual models that was characterized by relatively much smaller groundwater storage
39 compared to the new hypothesized model proposed by this study.

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Table 1. Measured Cl, $\delta^{18}\text{O}$, $\delta^2\text{H}$ and calculated water losses from isotope and Cl mass balances along Weeli Wolli Creek (2 July 2013 sampling). Cumulative values were calculated separately for Sections A and B. Evaporative loss was calculated using an isotope mass balance model as a ratio of evaporation to inflow (E/A) for each section using 22°C, 44% of relative humidity (RH) and $\delta^{18}\text{O} = -17.08\text{‰}$ ambient air (calculated from recent rainfall in June 2011 $\delta^{18}\text{O} = -7.45\text{‰}$). The Cl_e was calculated by subtraction of the transpirational water loss from that of evaporative water loss for each interval of the creek.

Surface Water	Name	Distance from DP1	Cl [mg L ⁻¹]	$\delta^{18}\text{O}$ [‰, VSMOW]	$\delta^2\text{H}$ [‰, VSMOW]	Cumulative volume loss (%) due to evaporation (Isotope mass balance)	Cumulative volume loss (%) due to evapotranspiration (Cl mass balance)	Calculated Cl_e due to evaporation [mg L ⁻¹]
DP1	DP1	0	73	-8.57	-57.5	-	-	-
Section A	WW020a	0.4	74	-8.24	-57.1	0	-	-
	WW018	3.1	75	-8.20	-56.9	0.2	1.4	74.1
	WW016	5.2	77	-8.06	-56.5	0.8	4.1	74.6
	WW014	7.0	79	-8.03	-57.2	0.9	6.8	74.7
	WW012	9.8	79	-7.98	-55.3	1.1	6.8	74.8
	WW010	11.3	79	-7.98	-54.7	1.1	-	74.8
DP2	DP2	12.4	100	-7.92	-54.9	-	-	-
Section B	WW008a	13.2	89	-7.93	-55.0	0	-	-
	WW006	15.3	90	-7.77	-53.9	0.6	1.1	89.6
	WW001	18.0	96	-7.69	-54.6	1.0	7.9	89.9
	WW003	23.2	110	-7.28	-50.8	2.8	23.6	91.5
	WW003a	23.6	110	-7.31	-51.5	2.7	23.6	91.4
	WWE	24.1	120	-7.00	-52.0	4.0	34.8	92.7
Ground water	WWA	17.9	95	-7.38	-53.3	-	-	-
	99YJWB04	18.0	102	-7.87	-55.3	-	-	-
	YM119	20.5	106	-7.77	-54.9	-	-	-
	D08YJ188	22.8	130	-7.13	-50.8	-	-	-
	99YJWB01	26.0	48	-8.83	-60.6	-	-	-

Table 2. Average calculated water balance components in m³/day/m length of the saturated stretch of Weeli Wolli Creek.

Water Budget Component	Section A	Section B
Transpiration	0.27	2.43
Evaporation	0.35	0.23
Recharge	-0.60	6.75
Total	0.02	9.41

Table 3. Bore construction details. The ALL, CID and WW denotes to Alluvium, Channel Iron Deposit and Wellie Wolli Formations.

Bore ID	Distance from DP1 (km)	Depth (m)	Screened Lithology	Screen Interval (m)
JSE37	17.9	78	ALL/CID/WW	6 to 78
WWA	17.9	15	ALL	2 to 15
99YJWB04	18.0	40	ALL/CID	6 to 40
D08YJ126	19.5	70	ALL/CID	6 to 70
D08WJ188	20.4	80	ALL/CID	6 to 80
YM119	20.7	60	ALL/CID	6 to 60
D08YJ216	21.1	96	ALL/CID/WW	6 to 96
D08YJ165	23.9	90	ALL/CID	6 to 90
99YJWB01	25.1	68	ALL/CID	6 to 68