Estimating seepage flux from ephemeral stream channels using surface water and groundwater level data

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Abstract Seepage flux from ephemeral streams can be an important component of the water balance in arid and semi-arid regions. An emerging technique for quantifying this flux involves the measurement and simulation of a flood wave as it moves along an initially dry channel. This study investigates the usefulness of including surface water and groundwater data to improve model calibration when using this technique. We trialed this approach using a controlled flow event along a 1387 m reach of artificial stream channel. Observations were then simulated using a numerical model that combines the diffusion-wave approximation of the Saint-Vénant equations for streamflow routing, with Philip's infiltration equation and the groundwater flow equation. Model estimates of seepage flux for the upstream segments of the study reach, where streambed hydraulic conductivities were approximately \(10^2\) m d\(^{-1}\), were on the order of \(10^{-4}\) m\(^2\) d\(^{-1}\) m\(^{-2}\). In the downstream segments, streambed hydraulic conductivities were generally much lower but highly variable \((-10^{-3} to 10^{-7}\) m d\(^{-1}\)). A Latin Hypercube Monte Carlo sensitivity analysis showed that the flood front timing, surface water stage, groundwater heads, and the predicted streamflow seepage were most influenced by specific yield. Furthermore, inclusion of groundwater data resulted in a higher estimate of total seepage estimates than if the flood front timing were used alone.

1. Introduction

Groundwater recharge from ephemeral and intermittent systems plays an important role in sustaining water resources in arid and semi-arid environments [Sophocleous, 2002]. Quantifying the contribution from ephemeral streams to the water balance is important in such environments, where surface water resources are limited and in great demand [Winter et al., 1998]. In particular, this requires knowledge of the spatial variability of recharge along the ephemeral stream channels that can be used to inform management strategies for the capture and storage of the water resource [Callegary et al., 2007]. However, the limited predictability of streamflow events in arid and semi-arid environments hinders the application of traditional field-based techniques to estimate seepage and recharge.

Direct and indirect methods have been developed that assess recharge in arid environments at the point/local and regional scales [Goodrich et al., 2004]; these methods include the use of geochemical tracers [Harrington et al., 2002; Vanderzalm et al., 2011], analytical and numerical modeling of groundwater mounding [Abdulrazzak, 1983; Walters, 1990; Mudd, 2006; Chenini and Ben Mammou, 2010], reach-scale water balances [Keppel and Renard, 1962; Knighton and Nanson, 1994; Ruehl et al., 2006], surface geophysics [Pool and Eychaner, 1995; Callegary et al., 2007], and water, solute, and heat transport modeling in the unsaturated zone [Conant, 2004; Su et al., 2004; Dahan et al., 2007, 2008; Kalbus et al., 2008; Dahan et al., 2009; Shanafiel et al., 2010]. The accuracy of upsampling point and downscaling regional seepage estimates has also proven challenging [Cushman, 1986; Wood, 2009].

Advancements in remote monitoring technology have led to the development of techniques involving the use of ephemeral river flood event data to quantify surface water infiltration, and hence groundwater recharge [Niswonger et al., 2008]. Previously, this concept had been applied in agricultural fields at the furrow scale to determine infiltration rates [Scaloppi et al., 1995; Camacho et al., 1997]; only recently has it been applied to larger natural and artificial channels to aid with catchment water management [Niswonger et al., 2008].
Mudd [2006] demonstrated theoretically that infiltration has an appreciable impact on the flood front movement, i.e., slowing flood front movement where infiltration is high. This approach has been further developed to incorporate variably saturated flow. Furthermore, it has been modified for both the kinematic and diffusion-wave approximation of the Saint-Vénant equations for the flood wave routing [Niswonger and Prudic, 2004, 2005; Shanafield et al., 2012].

Niswonger et al. [2005, 2008] developed a method that combined the kinematic wave approximation of the Saint-Vénant equations and Philip’s infiltration equation and was linked to the groundwater flow equation. This provided a framework enabling the use of streamflow-front velocity to determine streambed hydraulic conductivity for an initially dry channel. This package was developed from the streamflow routing package (SFR2) for MODFLOW-2000 for the simulation of gravity-induced unsaturated flow beneath streams [Niswonger and Prudic, 2004]. The kinematic wave approximation is the simplest approximation of the Saint-Vénant equations and assumes that the water surface is equal to the bed slope, i.e., the pressure force, local and convection acceleration force terms are negligible. This method proved successful in steeper downhill sloped ephemeral streams with fast flow [Niswonger et al., 2005].

The kinematic wave approximation can be an inappropriate simplification where the channel slope is shallow and where bed slope can be locally negative [Maidment, 1992]. In these instances including the pressure force term, as a minimum (i.e., diffusion-wave approximation or full Saint-Vénant equations), is necessary to account for the shallow or negative slopes and the resultant backwater effects on upstream flows and/or surface water levels. The more complicated full Saint-Vénant equations can prove unsuitable due to numerical instabilities and have a high computational demand; therefore, the diffusion-wave approximation proves to be the most suitable model [Chow et al., 1988].

For this study, the diffusion-wave approximation of the Saint-Vénant equations provided the most appropriate solution due to its ability to cope with shallow bed slope and negative gradients [Maidment, 1992]. Similarly, the predicted average kinematic wave number for the study reach was below the accepted value for the use of the kinematic wave model [Woolhiser, 1974]. The maximum Froude number for the reach was 0.07, also indicating that the most appropriate Saint-Vénant approximation is the diffusion wave [Vieira, 1983].

Several models for diffusion-wave routing have already been developed. DAFLOW [Jobson and Harbaugh, 1999] and MODFLOW-SWR1 [Hughes et al., 2012] are diffusion-wave models; however, they do not simulate unsaturated water movement. As a result, they are best suited to connected systems where surface water-groundwater interactions are governed by Darcy’s law. Similar to the Niswonger et al. [2008] kinematic model, Shanafield et al. [2012] linked a diffusion-wave routing of surface flow with the Philip’s equation for infiltration and MODFLOW for groundwater response. This model was therefore chosen for this study because it allowed for the advancement of a flood front along an initially dry, mild sloped channel and the resultant aquifer response to be used to calibrate for changes in hydraulic conductivity along the study reach.

The aim of this study was to investigate how the inclusion of surface water and groundwater head data in the calibration of a coupled surface water flow-infiltration-groundwater flow model aids the estimation of seepage flux. This was achieved using a controlled flow event in an artificial channel, and a recently developed model that couples the diffusion-wave approximation to Philip’s infiltration and MODFLOW [Shanafield et al., 2012]. Previous applications of the model have not considered stream stage or groundwater head data in the calibration processes, which we show is valuable. We assess the sensitivity of the flood front flow, surface water stage, and groundwater head objective functions as well as the estimated total seepage to various input parameters using a Latin Hypercube Monte Carlo sensitivity analysis [Saltelli et al., 2008].

2. Background

2.1. Theory

The diffusion-wave approximation combines the continuity and momentum equations, assuming that acceleration forces can be neglected [Moussa and Bocquillon, 1996]. Following Maidment [1992], the continuity equation can be expressed as

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = 0 \quad (1)$$
and the momentum equation can be expressed as

$$g\left(\frac{\partial h_s(t)}{\partial x(t)}\right) - g(S_o - S_f) = 0$$

(2)

where $Q$ is flow ($L^3 T^{-1}$), $A$ is the cross-sectional area of the channel ($L^2$), $h_s$ is the water surface elevation in the channel ($L$), $x$ is the length in the direction of flow ($L$), $S_o$ is the channel slope ($L T^{-1}$), $S_f$ is the friction slope ($L T^{-1}$), and $t$ is time ($T$).

The combination of these equation yields the one-dimensional diffusion-wave approximation given as [Panday and Huyakorn, 2004]

$$\frac{\partial}{\partial x} \left( k_s \frac{\partial h_s}{\partial x} \right) - \frac{\partial}{\partial t} (Bh_s) - Aq_g = 0$$

(3)

where $B$ is the cross-sectional top width ($L$), $k_s$ is the channel conductance term ($L^3 T^{-1}$), and $q_g$ is the flux per unit volume seepage to the subsurface ($T^{-1}$). This equation has the same form as the groundwater flow equation [Panday and Huyakorn, 2004], making it possible to solve using sparse-matrix solvers identical to those in MODFLOW [Harbaugh, 2005].

The proportion of infiltration occurring during the initial wetting up of the streamed profile has been shown to be significant when compared to the total seepage loss [Battle-Aguilar and Cook, 2012; Blasch et al., 2004]. This is not considered when a constant infiltration rate is applied uniformly along the channel in question and can result in errors in total seepage loss estimates [Walters, 1990; Hamed et al., 1996; Morin et al., 2009]. Philip’s infiltration equation [Philip, 1957, 1958] was used to incorporate the transient infiltration rate during the onset of flow, driven by the large capillary gradient at the flood front. The assumptions of this approach include one-dimensional, vertical, homogeneous flow and can be written as

$$I = \frac{1}{2} S t_w^{0.5} + K_s$$

(4)

where $I$ is infiltration ($L T^{-1}$), $t_w$ is the time since initial channel wetting by flow ($T$), $K_s$ is the saturated hydraulic conductivity of channel sediments ($L T^{-1}$), and $S$ is sorptivity ($L T^{-1/2}$), which is a function of the stream depth, sediment water retention curve, relative hydraulic conductivity, and the initial and saturated water content of the streambed, $\theta_i$ and $\theta_s$. Following Philip [1958], the effects of stream depth on sorptivity are also incorporated, and sorptivity is expressed as

$$S^2 = 2K_s(\theta_i - \theta_s) \left[ \int_0^{\psi} K_c(\psi) d\psi + h_i \right]$$

(5)

where $\psi$ is the capillary pressure potential ($L$) and $K_c(\psi)$ is the relative hydraulic conductivity as a function of capillary pressure potential (dimensionless). Philip’s infiltration equation was coupled to the Brooks and Corey [1964] soil-water characterization curve to estimate the changes in saturation and capillary gradient as the flow moved down gradient. This analytical solution assumes that both the hydraulic conductivity and soil water diffusivity are a function of sediment water content (rather than depth). Therefore, these assumptions are only valid where there is no significant vertical heterogeneity in the streamed sediments.

When the groundwater table is lower than the streamed, infiltration to the groundwater is incorporated into the seepage term as

$$Aq_g = PI$$

(6)

where $P$ is the channel wetted perimeter ($L$) and $I$ is the infiltration rate ($L T^{-1}$). When the water table rises to the base of the streambed and above, the seepage to the groundwater is calculated as

$$Aq_g = K_s P (h_g - h_i) / b$$

(7)

where $K_s$ ($L T^{-1}$) is the hydraulic conductivity of channel sediments, $h_g$ ($L$) is the groundwater head in the underlying cell, $I$ is the length of the reach ($L$), and $b$ is the average streamed thickness ($L$). Combining the
surface and subsurface flow equations enables estimation of the groundwater response to a surface flood event [Keppel and Renard, 1962]. To simulate the connection between surface water and groundwater flow, the diffusion-wave approximation and Philip's infiltration equations are combined and coupled with the groundwater flow equation [Panday and Huyakorn, 2004]. The surface water level and groundwater heads are combined into one matrix and determined simultaneously using a Newton-Krylov solver in the NWT package [Niswonger et al., 2011].

The model developed by Shanafield et al. [2012] couples the one-dimensional diffusion-wave approximation with Philip's infiltration equation [Philip, 1957, 1958] and MODFLOW-NWT [Niswonger et al., 2011] providing a useful tool for the simulation of within channel flow events and associated groundwater response (Figure 1).

2.2. Field Site

This study was performed on the Western Reflows Floodway (WRF), a newly graded, unlined artificial water conveyance channel located in the South East of South Australia (lat. 37°13′16 long. 140°29′21) (Figure 2). The channel was designed to convey water northward to the RAMSAR listed Coorong wetlands to help maintain ecological water requirements [Harding, 2007]. Construction of the trapezoidal channel was completed in 2010 and remained unused until a controlled flow event in May 2011.

The study focused on a 1387 m reach of the WRF north of the Callendale Flow Regulator (CFR) on Drain M (lat. 37°13′41.26 long. 140°29′27.57). The reach is located within a 7-year-old blue gum (Eucalyptus globulus) plantation; however, the channel itself is bare in the upstream section and sparsely vegetated with grass and reeds in the downstream section. At the field site, the Semaphore Sand member, an unconsolidated white bioclastic quartz-carbonate sand of modern beaches and transgressive dune fields extends from the surface to approximately 5 m depth [Fairclough, 2010]. During construction, a clay layer was observed beneath the channel; the depth to this layer decreases downstream from 0.5 to 2.0 m below the base of the channel. The WRF was designed to convey flows of less than 3 m$^3$ s$^{-1}$ and has a uniform base width of 20 m with ~45° banks. The average slope of the study reach is 0.0002 m m$^{-1}$. The direction of regional groundwater flow is toward the northeast and depth to groundwater in the area is approximately 4 m below ground level (bgl). However, beneath the channel the depth to groundwater is less (0.5–2 m bgl) suggestive of water table mounding due to increased recharge along the bare channel relative to the adjacent forest plantation.

3. Methodology

3.1. Field Methods

A controlled flow event was performed at the site from 19 May to 2 June 2011 by closing the CFR on Drain M and allowing water to flow into the WRF. The study reach was divided into 15 segments; flood front ($ff$) timing and surface water level (i.e., stage) were monitored at each segment boundary. Prior to the event, two additional monitoring sites, located approximately 500 m apart, were instrumented to measure shallow

![Conceptual diagram of the diffusion-wave MODFLOW model](https://example.com/diffusion-wave-model-diagram.png)

**Figure 1.** Conceptual diagram of the diffusion-wave MODFLOW model [based on Niswonger et al., 2008] showing the surface and subsurface discretization for flow calculation, where $A$ is channel area ($L^2$) and $Q$ is flow ($L^3 T^{-1}$).
groundwater head and surface water level. At these sites (Sites 1 and 2; Figure 2), groundwater head was measured in drive-point piezometers installed at 1.5 m below ground level in the center of the channel, assumed to be the depth of the streambed. All surface water levels were monitored at the channel thalweg. Pressure transducers (In-Situ Inc LevelTROLL® 300 Series, ±0.2% accuracy, and In-Situ Inc. BaroTROLL®, ±0.05% accuracy) were used to record stream depth and groundwater heads at 1-min intervals. Only the data for the period when flow commenced at the upstream boundary until flow reached the downstream boundary is of relevance to the model calibration.

Flow into the study reach was determined by manual flow gauging at the upstream boundary of Segment 1. A Marsh-McBirney Flo-Mate was used to estimate flows following the area-velocity method [Hipolito and Lour-eiro, 1988]. Due to the transience of the flow as the flood wave passed, the flow gauging began with one measurement of flow in the channel and increased to 12 lateral measurements as the rate of change in surface water level decreased. This constrained error and minimized the time interval between gaugings. The time between subsequent gaugings was between 2 and 12 min depending on the number of velocity measurements. The flood front progression down the channel was tracked using handheld Garmin GPS units.

Streambed sediment properties for the study reach were determined from samples collected at Sites 1 and 2. Site 1 was sand, which appeared to extend downstream to Segment 4. At Site 2, the streambed sediments had greater clay content, which remained present in Segments 5–11. Sediment samples from the top 0.4 m of the streambed were collected and analyzed using the filter paper method [Fawcett and Collis-
George, 1967; Greacen et al., 1989) and hanging water column to estimate the water retention curves and the Brooks and Corey coefficient [Brooks and Corey, 1964]. The channel sediment moisture profile was assumed to be initially at equilibrium; for example, where the perched water table was at 1.5 m, the surface matric potential was assumed to be −1.5 m. The initial \( \theta_i \) was then derived from the water retention curve. The residual \( \theta_r \) and saturated \( \theta_s \) sediment water content was also estimated using the water retention curves (see Table 1). No evaporative losses were considered due to the short duration of flow (2.28 h travel time) within the channel.

3.2. Modeling
The field data provided a framework to construct a numerical model of the captured flow event. The objective of the model calibration was to quantify the spatial variability of the streambed hydraulic conductivity \( K_s \) and thus seepage flux. The sensitivity of the calibrated model to each of the input parameters was investigated, as was the uncertainty in total seepage flux for the reach (see section 3.2.2).

The 15 channel segments provided the surface discretization for the model. A site survey determined the channel cross-section geometry at the upstream and downstream end of each segment (Table 2). Channel surface water hydraulic properties (wetted perimeter, channel width, and hydraulic radius) for each segment were calculated for 10 stream stage depths (0.01–0.7 m). Finer discretization of the hydraulic properties at shallow stream stage depths was necessary to account for vertical variations in friction, which was particularly important in this study as flow depth varied from zero to full event depth (0.5 m) and resistance values are strongly depth, and hence flow dependent [Morvan et al., 2008]. The Manning’s \( n \) roughness coefficient was estimated by using an equivalent roughness parameter "\( k \)" described in USACE [1991] and Brunner [2010]. An appropriate \( k \) value was determined by setting \( n \) as 0.024 at a surface water depth of 0.5 m, which is within the range expected for an excavated earthen channel [0.018–0.025; Chow, 1959]. All \( n \) values decreased from 0.1 at a depth of 0.01 m, to 0.024 at 0.5 m depth. The upper Manning’s \( n \) roughness coefficient was selected as a reasonable estimate of the channel roughness at shallow depth.

The downstream boundary condition is a critical depth condition at which the Froude number is equal to one. To ensure this boundary did not influence the surface water solution, the length of the last segment in the model was increased by 100 m to prevent the flood front reaching the downstream boundary within the calibration period. Time step length was assigned according to the time for the flood front to progress from one segment boundary to the next and ranged from 270 to 961 s. The timing of full channel width flow was also used in the calibration for comparison. In this study, the small difference between the flood

![Table 1. Description of Model Parameters](image-url)
front and full channel width flow coupled with the observed shape of the front suggested that flood front timing was an appropriate measure of the flood front velocity.

Initial groundwater head \( h_g \) below the channel, obtained from groundwater monitoring Sites 1 (Segment 2) and 2 (Segment 10), were 1.41 and 0.37 m, respectively. Additional groundwater head data from a drive-point located at the downstream end of the channel was used as the groundwater head for the last segment. The groundwater head for each segment was interpolated between the measured groundwater heads from these three sites. Only locations where both surface water levels and groundwater head data were collected were used in the model calibration. Specific yield (\( Sy \)) was estimated from the water retention curve and the initial.

Discretization in the direction of flow was approximately 100 m, equal to the segment lengths (Table 2). The groundwater model boundary conditions were set a sufficient distance away (4 km) from the channel, to limit their influence on the solution. The cells beneath and directly adjacent to the channel had a cell width (y-direction, perpendicular to channel) of 25 m. The cell width increased to 500 m at a perpendicular distance of 100 m from the channel edge. Cell depth was set to 2.5 m. Aquifer hydraulic conductivity (\( K_a \)) was assumed to be equal to streambed hydraulic conductivity (\( K_s \)).

### 3.2.1. Model Calibration

Model calibration was undertaken using the parameter estimation program PEST [Doherty, 2010] which uses a least squares fitting routine and a Gauss Marquardt Levenberg method to minimize a user-defined objective function:

\[
\phi_k = \sum_{k=1}^{N_{obs}} (w_k x^k_m - w_k x^k_s)^2
\]

where \( \phi \) is the objective function value (units dependent on \( x \)), \( w_k \) is the weight applied to the difference between the measured (\( x^k_m \)) and simulated (\( x^k_s \)) parameter of the same type \( k \), and \( N_{obs} \) is the total number of measured parameter values of the same type. The objective function is calculated for each data type. During the optimization process, PEST solves global objective function which incorporates the different types of data. We impose a weighting on each data type by the following equation:

\[
w_k = \frac{1}{\sigma \sqrt{N_{obs}}}
\]

where \( \sigma \) was the standard deviation in observation data of the same type, this normalized the weightings according to the different data type measurement values and number of measurements [McCallum et al., 2012]. Model parameters are defined in Table 2.
Three different scenarios were used to investigate the impact of including different data types (namely flood front timing, surface water level and groundwater head) on model calibration. In the first scenario, only flood front timing data were used in the objective function, whereas in the second and third scenarios, all three data types were used (Table 3). In all three cases, streambed saturated hydraulic conductivity ($K_s$) was the parameter to be estimated, plus in the third scenario, Manning’s roughness coefficients ($n$) were also estimated parameters.

A simulated flow of 0.05 m$^3$ s$^{-1}$ into each segment was the criteria used to identify the flood front arrival time at each segment boundary; the streambed was considered dry for flow values below 0.05 m$^3$ s$^{-1}$. This represents the flow at a stage depth of 0.01 m as determined from the manual gauging at Segment 1.

The models were run for a period of 22 h in order to simulate the observed surface water level data and groundwater head data up to a time when the flow had stabilized. The recession of flow was not included due to lack of inflow data; this would prove interesting for future studies.

### 3.2.2. Sensitivity Analysis

A modified Monte Carlo analysis was performed to investigate the sensitivity of the total seepage flux, estimated using the calibrated model, to the input parameters. Latin hypercube sampling (LHS) was used to generate the multivariate parameter distributions used for the Monte Carlo simulations. LHS is a form of stratified random sampling that reduces the number of simulations necessary for the Monte Carlo analysis. Twenty-seven parameters were considered in the Monte Carlo simulations (Table 1) and a sample size of $N = 10,000$ was chosen [Iman and Helton, 1988]. Values for the 27 input parameters were selected from a normal distribution with standard deviation equal to the standard measurement error (see Table 1). Streambed saturated hydraulic conductivities ($K_s$) were assumed to be log-normally distributed for the purpose of the Monte Carlo analysis. A reasonable error was used for the $K_s$ based on observed sediment types within the channel. Model runs that did not converge within 10 times the average run time were observed to produce unrealistic output data and were discarded. The calibrated model from Scenario D2 was used as the base case in the sensitivity analysis.

The model output was assessed in terms of different objective functions: flood front flow ($ff$ flow), surface water level ($\phi_s$), and groundwater head ($\phi_g$). Total simulated seepage along the channel was also compared, defined as the sum of the seepage flux from each of the segments over the total simulation period (i.e., 22 h). The $ff$ flow $\phi$ represents the deviation of the flow output from the observed flow at each segment boundary, i.e., flood front arrival indicated by a flow of 0.05 m$^3$ s$^{-1}$.

### 4. Results

#### 4.1. Experimental Results

Flow into the study reach commenced at 12:28 pm on 19 May 2011. The maximum flow into the reach was 2.6 m$^3$ s$^{-1}$, occurring 25 h after the onset of flow (Figure 3). A distinctive step pattern was observed in the rising limb of the flow, which was also present in the surface water level at the upstream boundary, most likely indicating the flow is utilizing storage in the system upstream of the study reach at this point. Total time for the flood front to travel the 1387 m to the end of the study reach was 2.28 h (flow initiation) and 2.31 h (full channel inundation) (Figure 4). For the majority of segments, full channel flow occurs within minutes of the initial flood front passing. However, there is a greater delay for Segments 3–7 on the order of 9 min. This increased lag can be attributed to the presence of ponded surface water within the channel resulting in an acceleration of the flood front (i.e., piston flow within the ponded water).
The maximum surface water level of 0.5 m at the upstream boundary was reached 14.2 h after the onset of flow. The highest stage observed along the reach occurred at the same time in Segment 10 (0.6 m). The streambed thalweg greatly influenced the shape of the wave, as seen in the large positive (downstream) streambed gradient at Segment 10 and the negative (upstream) streambed gradients in the previous segments (Figure 4). The gradient in the surface water clearly decreased for Segments 8 and 9.

4.2. Model Results

Of the three scenarios used to calibrate the model (D1–D3, Table 3) Scenario D2 provided a marginally better fit to the observed surface water level and groundwater head data (Figure 5). The error in flood front timing for Scenario D2 was only slightly higher than for Scenario D1 where flood front timing was the only data type used in the objective function. Comparing the results of Scenario D3 with those of Scenario D2 suggests the treatment of Manning’s $n$ as an estimated parameter in the model optimization did not improve the calibration for flood front timing, and actually worsened the calibration for surface water level and groundwater head.

The spatial variability in estimated seepage flux along the channel for the three scenarios displays similar trends (Figure 6). The scenarios exhibit a general trend from relatively high values in the upper segments ($10^{-4}$ to $10^{-3}$ m$^3$ d$^{-1}$ m$^{-2}$) to much lower values in the lower segments ($10^{-7}$ to $10^{-4}$ m$^3$ d$^{-1}$ m$^{-2}$), although these fluxes are extremely variable from one segment to the next. Variation in estimated seepage flux between the three scenarios is more pronounced in the downstream segments, in particular, those with a lower seepage flux estimate. The low seepage fluxes observed in the downstream segment may be attributed to the clay layer observed beneath the channel. This clearly demonstrates the

Figure 3. Inflow flood hydrograph determined by manual gauging at the upstream boundary of the WRF reach. The area-velocity technique was applied to determine flow. Error bars are calculated from a 5% measurement error with an additional error calculated from the change in flow during the period of each flow measurement. The total estimated error ranged from 15% at the beginning of the event to 5% for the steady state.

Figure 4. Timing and shape of the flood wave at each segment boundary (gray diamond and labeled S2, S4, etc.) determined from pressure transducer data and site survey at each boundary. The circles and dashed line (unfilled) show the cumulative travel time of the flood front for both the initial flow (filled circles) and flow at full width inundation (open circles).
value of including both surface water level data and groundwater head data as calibration constraints when modeling seepage flux.

The simulated surface water and groundwater response for Site 1 is consistent with the measured data (Figure 7). During the initial 4 h of the simulation the groundwater response is overestimated by ∼0.2 m, subsequent to this the simulated groundwater head fits well to the observed response. At Site 2, accumulated errors in the timing of the flood front arrival resulted in preemptive increase in the both the surface water and groundwater response (Figure 7), e.g., the simulated rise in groundwater head occurred 19 min before the observed response. In the long term (post 4 h), the surface water response was consistently underestimated by up to 0.06 m; this response is not observed in the simulated groundwater head.

Toward the end of the simulation, when conditions are assumed to be near steady state, the simulated surface water level was below the measured value by ∼0.08 m for Site 1 and ∼0.13 m for Site 2. For both sites, the surface water level remained essentially constant after 3.65 h. The simulated groundwater head after 22 h was in close agreement with measured values at both sites, providing further confidence in the estimated seepage flux to the water table. The calibrated model from Scenario D2 is used for all future analysis and comparisons.

4.3. Monte Carlo Sensitivity Analysis Results

Quantifying uncertainty in total seepage from the calibrated model (Scenario D2) will be strongly influenced by the weighting, within and between each observation type (equation (9)). For example, if the weight applied to the groundwater head data were higher than for the flood front flow (ff flow) and surface water stage, the total seepage would be larger (darker colors in Figure 8). Likewise, if groundwater head data weight was reduced or excluded, total seepage along the channel would be smaller. Therefore, selecting the appropriate weights for the different data sets is vital during the calibration process. The weighting given in equation (9) means that, for example, when ϕ = 0.3 the model run output captured 70% of the variation in the observation data sets. For ease of discussion, an acceptable calibration result was selected as those model runs with a ϕ ≤ 0.3 (see Table 4).

The calibrated model used all three data sets during the calibration process and provides a reasonable fit to ff flow, surface water levels, and groundwater head data. This is shown in Figure 8 for the six most
informative parameters. Both the ff flow and surface water levels Φ are close to the optimal value (lowest Φ value), whereas the groundwater head Φ for the calibrated model was not the optimal objective value obtained during the uncertainty analysis. This may be a result of the influence of Sy; the calibration will restrict Sy to close to the mean value (0 on the x axis) because the ff flow Φ will increase by more than an order of magnitude beyond this narrow range (Figure 8j). The model is unable to simultaneously optimize both objective functions (ff flow and gwh). Better groundwater head calibration results in a higher total seepage estimate (Figure 9), whereas a better ff flow calibration results in a lower total seepage (7.09 ± 2.03 m³), the calibrated model is thus a compromise.

Approximately 90% of all Monte Carlo runs produced a groundwater head Φ ≤ 0.3. This suggests that most of the modeled groundwater responses fit the observed data reasonably accurately. The relationship between groundwater head Φ and total seepage along the channel suggested that those model runs with a smaller error in groundwater head (e.g., lower groundwater head Φ) tended to have higher total seepage (8.28 ± 2.43 m³, Table 4). In comparison, 66% of all the surface water level Φ were less than 0.2. The total seepage for surface water level Φ ≤ 0.2 was 7.09 ± 2.03 m³, lower than the total seepage for groundwater head Φ ≤ 0.2. Flood front data has previously been used to constrain this type of model [Shanafield et al., 2012]; if the flood front data were the only data set available for this study, total seepage would be estimated as 4.09 ± 0.04 m³ (Table 4).

The average objective functions of the three output data sets for the six parameters are shown in Figure 10. The ideal result would be a localized area with a low objective function value (white/yellow zone) surrounded by higher objective function values. This would allow both the uncertainty in the parameter value and total seepage to be constrained. However, many of the individual parameters are not well constrained (width of the white/yellow zone), with a range of ±3σ (e.g., Figure 10a). Only the range in specific yield values were constrained; ±0.2σ (Figure 10d). Based on the criterion that Φ ≤ 0.3 represents a reasonable model, the range in total seepage was ~3.49–5.49 m³ along the study reach, which compare favorably with the calibrated model total seepage of 3.8 m³.

5. Discussion

The longitudinal changes in seepage flux were estimated using a recently developed diffusion-wave model, capable of simulating streamflow events in channels with mild slope and backwater effects. The availability of surface water level and groundwater head data provided insight into the models capacity to simulate and use this information to constrain the streambed hydraulic conductivity and seepage flux estimates along the reach. The inclusion of groundwater and surface water data estimated higher total seepage.
Figure 8. Six examples showing the impacts of changes in parameter value, expressed as given in standard deviations relative to the calibration model value, to the flood front flow ($ff$ flow), surface water stage ($sws$), and groundwater head ($gwh$). Modeled parameter errors and values are provided in Tables 1 and 2. For example, $LnK^s$ equals $-8.52 \text{ m s}^{-1}$ ± 1; therefore, when the number of $r$ equals 2, 0.5, and 1, $LnK^s$ equals $-10.52$, $-9.02$ and $-7.52$ respectively ((m), (n), and (o)). The color represents the total seepage flux ($m^3$) along the channel. Stars represent the calibrated model $\theta$ and total seepage flux.
volumes than if the flood front data were used alone. The resultant calibrated model indicated longitudinal heterogeneity within the streambed sediments along the reach, enabling clear definition of areas of high and low seepage flux. A benefit of this approach is it results in a spatially integrated estimate of seepage flux rather than the need to upscale point measurements thereby capturing the sediment characteristic of the entire channel.

The seepage fluxes along the study reach vary by three orders of magnitude. A similar range of streambed conductivity—and by inference seepage flux—was observed by Kennedy et al. [2008] and Genereux et al. [2008] along 50–200 m stream sections in North Carolina. A recent study by Hatch et al. [2010] also identified a temporal variability of streambed seepage fluxes, as a result of deposition of sediment load during flow recession. This suggests that the simulated streambed seepage fluxes for the May 2011 WRF flow event are unlikely to be exactly replicated in any subsequent flow event. Similarly, the reduced energy conditions associated with negative bed slopes would encourage deposition of finer, lower permeability sediments therefore bed slope will influence the spatial characteristics of streambed sediments.

The wet channel antecedent conditions have likely altered the influence that both surface water level and groundwater head will have on the model calibration (Table 2). The model uses the initial wetting up phase of the unsaturated zone to estimate streambed properties ($K_s$). Therefore, the initially wet channel conditions in this study would have limited the period in which the streambed properties could be estimated, thereby having a limited impact on the surface water level and a fast response in groundwater head [Niswonger et al., 2008; Shanafield et al., 2012]. The relative importance of unsaturated zone processes will increase where antecedent moisture conditions are drier and the depth to groundwater is greater.

The inclusion of Manning’s $n$ (Scenario D3) did not improve the calibration suggesting the chosen distribution was an adequate approximation of the channel friction. Niswonger et al. [2008] showed that for the MODFLOW-SFR2 model the flood front timing is very sensitive to $K_s$ where $K_s$ values are greater than 1.75 m d$^{-1}$, but more sensitive to Manning’s $n$ when $K_s$ is less than 0.01 m d$^{-1}$. This suggests that segments with low seepage fluxes (associated with low $K_s$) are most sensitive to Manning’s $n$.

The estimation of Manning’s $n$ is important because it helps control the flood front movement along the channel and the surface water level. For this study, the influence of Manning’s $n$ on total seepage appears to be limited; however, previous studies have identified the difficulty and importance in the estimation of Manning’s $n$ [Khatibi et al., 1997]. Because channel roughness has a strong control on shallow flow, the choice of Manning’s $n$ has significant impact on calibrating surface

<table>
<thead>
<tr>
<th>Total Seepage (m³)</th>
<th>Mean</th>
<th>Variance</th>
</tr>
</thead>
<tbody>
<tr>
<td>All $\phi$</td>
<td>7.92</td>
<td>7.54</td>
</tr>
<tr>
<td>$ff$ $\phi \leq 0.3$</td>
<td>4.09</td>
<td>0.04</td>
</tr>
<tr>
<td>$sws$ $\phi \leq 0.3$</td>
<td>7.09</td>
<td>4.10</td>
</tr>
<tr>
<td>$gwh$ $\phi \leq 0.3$</td>
<td>8.28</td>
<td>5.90</td>
</tr>
</tbody>
</table>
In this study, the same Manning’s \( n \) distribution was applied to all of the 15 segments, negating any differences between segments; therefore, additional errors may have been incurred due to the spatial variability in Manning’s \( n \) between segments. The presence of two bends in the study reach (Figure 2) and a slight increase in channel vegetation in the downstream segments would result in additional variability in the channel friction that was not captured [Jarrett, 1985]. Further investigation should include this variability though to what extent this impacts model calibration is as yet unknown.

In the study channel, the depth to groundwater was shallow and the transition from unsaturated, disconnected to saturated, connected occurred rapidly; limiting the amount of information available with which to constrain the hydraulic conductivity of the streambed. However, where the depth to groundwater is greater, this transition would be slower and the groundwater response would assist in constraining the model. A possible limitation to the use of groundwater head data in this and future studies may also be the influence of localized heterogeneity in streambed sediments on the groundwater response. This could result in a localized groundwater response not representative of the average properties of the channel segment [Kennedy et al., 2010; Irvine et al., 2012].

To date, this model has only been applied to an artificial unlined channel; however, it has the potential to be applied to natural systems. The availability of LiDAR and remote sensing to characterize channel geometry [Passalacqua et al., 2012], in combination with data loggers that can be deployed to monitor surface flood movement and groundwater heads may enable catchment scale investigations. The main challenge associated with the use of this model in a natural system would be the adequate representation of the channel geometry and hydraulic properties, in particular, the estimation of appropriate values for Manning’s \( n \) roughness coefficient and the influence of vegetation on flow within the channel.

6. Conclusions

The spatial variability of seepage flux beneath a 1387 m reach of artificial stream channel was successfully estimated using a numerical model that accounts for flood wave propagation at the surface, infiltration through

![Figure 10. Uncertainty in total seepage flux (m³) from the study reach associated with changes in six parameter value (given in standard deviations, \( \sigma \)) determined from the modified Monte Carlo analysis. The color represents the average \( \phi \) of the three individual \( \phi \) for each data set shown in Figure 8. The star represents the total seepage flux and \( \phi \) of the calibrated model.](image-url)
the streambed and groundwater flow. We found that the method, which has previously only used flood front timing as a calibration constraint, provides more reliable seepage flux estimates when surface water level and groundwater head data and used in the calibration. For this study, surface water level and groundwater head data provided complementary information; lower total seepage estimates would have been estimated if this data were excluded. The total seepage for the study reach was estimated as 4.2 m^2 with an uncertainty of ± 0.6 m^2, and the specific yield of the aquifer was shown to be the most influential parameter. Though a number of limitations have been highlighted in this study, the method offers promising opportunities to determine the longitudinal variability in seepage fluxes from ephemeral and intermittent streams.

Acknowledgments
Funding for this research was provided by the National Centre for Groundwater Research and Training, an Australian Government initiative, supported by the Australian Research Council and the National Water Commission, and the Goyder Institute for Water Research of South Australia. The authors acknowledge Australian Blue Gums for allowing access and use of their land. They also acknowledge the anonymous reviewers for comments.

References

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