Aquifer response to surface water transience in disconnected streams

Margaret Shanafield,[1] Peter G. Cook,[1,2] Philip Brunner,[3] James McCallum,[1] and Craig T. Simmons[1]

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[1] Existing analytical solutions to determine aquifer response to a change in stream stage are inappropriate where an unsaturated zone exists beneath the stream, as in the case of disconnected stream-aquifer systems. A better understanding of the relationship between aquifer response and transient stream stage in disconnected systems is therefore required, as this would also aid in the field determination of the status of connection between the stream and aquifer. We use a numerical model to examine transient stream stage and the corresponding water table response. Beneath disconnected streams, the magnitude of head change in the water table level is a balance between the cumulative infiltration during a flow event and the rate at which the water can disperse laterally. Increases in wave duration, stream width, and streambed permeability result in greater infiltrated water volume and therefore a higher peak response at the water table. Conversely, higher aquifer transmissivity and aquifer hydraulic conductivity allow the water to move laterally away from the stream faster, resulting in a smaller head change below the stream. Lower unsaturated storage results in a greater and faster aquifer response because the unsaturated zone can fill more quickly. Under some combinations of parameters, the magnitude of the disconnected head response is more than seven times greater than the change in stream stage driving streambed infiltration; an effect which can never occur beneath a connected stream. The results of this sensitivity analysis are compared to field data from a river in eastern Australia to determine periods of disconnection. Where the change in aquifer head is greater than the change in stream stage, disconnection between the stream and aquifer can be determined.

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1. Introduction

[2] Many studies have considered the response of the aquifer to both steady state and transient surface water infiltration. For connected stream-aquifer systems, both linearized and later, nonlinearized analytical solutions to the Boussinesq equation have been developed to estimate recharge from flood waves (*Workman et al.* [1997], *Serrano and Workman* [1998], *Barlow et al.* [2000], and others). In the connected stream-aquifer system, flux into the streambed is controlled by the aquifer and streambed hydraulic conductivity and the head gradient between the stream and groundwater. If the water level in the stream fluctuates, the aquifer response (as might be observed in a piezometer adjacent to the stream) is controlled by the diffusivity of the system, which is the ratio of transmissivity to storativity. The change in groundwater head cannot be greater than the change in the stream stage.

[3] A stream-aquifer system is considered disconnected if an unsaturated zone is present beneath the stream. In disconnected systems, infiltrating water must travel vertically through the unsaturated zone before reaching the water table and moving horizontally away from the stream. Aquifer response to changes in stream stage is a function of stream stage, stream width, streambed hydraulic conductivity and thickness, and aquifer hydraulic conductivity, transmissivity, and specific yield [*Brunner et al.*, 2009a; *Osman and Bruen*, 2002]. The presence of the unsaturated zone affects both the timing and volume of groundwater recharge from the surface [*Hunt et al.*, 2008], and the common numerical model assumption of constant and instantaneous transfer of infiltration to recharge is not valid in disconnected systems [*Brunner et al.*, 2010].

[4] Although there is a large body of literature discussing groundwater response beneath connected streams, fewer studies have addressed the dynamics of the disconnected

[1]National Centre for Groundwater Research and Training, School of the Environment, Flinders University, Adelaide, South Australia, Australia.
[2]Water for a Healthy Country National Research Flagship, Division of Land and Water, Commonwealth Scientific and Industrial Research Organization, Adelaide, South Australia, Australia.
[3]Centre of Hydrogeology and Geothermics, Neuchâtel, Switzerland.

Corresponding author: M. Shanafield, National Centre for Groundwater Research and Training, School of the Environment, Flinders University, GPO Box 2100, Adelaide, SA 5001, Australia. (mshanafield@yahoo.com)

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stream-aquifer interaction. Early work on disconnected streams investigated changes in the magnitude of flux between the stream and the aquifer as streams change from connected to disconnected. *Moore and Jenkins* [1966] and *Weeks et al.* [1965] concurrently described hydraulically disconnected stream reaches in which lowering water table depths no longer affected infiltration rates. This link between flux and connection was further explored by *Osman and Bruen* [2002] and *Desilets et al.* [2008]. *Fox and Durnford* [2003] showed the importance of including unsaturated flow beneath disconnected streams when considering stream recharge of groundwater during aquifer drawdown. *Brunner et al.* [2009b] analyzed how the state of connection affects system dynamics in terms of infiltration fluxes, and demonstrated that changes in surface water depth in disconnected systems almost immediately result in a new steady state infiltration flux, provided the state of connection does not change in response to the change in flux. In comparison, connected systems show a timelag in reaching a steady infiltration flux following changes in surface water depth. However, although it has been pointed out that stream stage effects infiltration rates in disconnected systems [*Sophocleous*, 2002; *Vázquez-Suñé et al.*, 2007], none of the previous studies have explicitly explored the relationship between transient stream water levels and the aquifer head response.

[5] In the field it is difficult to determine the connection status of a given stream reach. Conventional methods of determining whether a stream is connected or disconnected include showing that the flux through the streambed does not change as the water table is lowered in response to groundwater pumping [*Moore and Jenkins*, 1966] or that an unsaturated zone is present beneath the stream. Discerning disconnection based on the lack of change in stream infiltration in response to aquifer pumping is often impractical because time lags usually exist between groundwater pumping and changes in stream infiltration [*Brunner et al.*, 2011]. Identifying an unsaturated zone beneath the stream can also be complicated, as layers of low permeability (known as "clogging layers") can be "highly site specific and difficult to estimate without extensive drilling" [*Reid and Dreiss*, 1990] and various streamflow and deposition processes can result in differing types and depths of clogging layers [*Brunke*, 1999].

[6] We hypothesize that aquifer response beneath a disconnected stream can be greater than the input signal. Because of the storage in the unsaturated zone, it might intuitively be expected that the head response at the water table would be dampened; however, the magnitude and timing of the aquifer response is a function of the complex interplay between infiltration through the streambed and the rate at which the water can laterally move away from beneath the stream. None of the existing literature describes the factors that control the height of the groundwater mound beneath a disconnected stream for transient stream stage. In this paper we explore the sensitivity of this groundwater head response to stream and aquifer parameters, with the goal of identifying water table responses that are unique to disconnected streams and may be useful in determining connection status. To examine the occurrence of these processes in a natural aquifer setting, stream depth and water table levels from a river in southern Australia are presented, and the data are evaluated with regard to connection status.

2. Modeling Approach

[7] The Hydrus 2-D/3-D software package [*Sejna and Simunek*, 2007; *Simunek et al.*, 2008] was used to simulate water movement beneath a two-dimensional, vertical cross section for half of a hypothetical stream for nine model scenarios (Table 1 and Figure 1). Because the objective was to observe changes in water table directly beneath the stream, a relatively wide half stream width (w) of 60 m was chosen, and the observation point was placed at 30 m (half way across the half stream) to limit the effects of bank infiltration and minimize boundary condition influences in the numerical model. A 0.5 m layer of material with lower permeability (clogging layer) lined the streambed and banks. The model domain extended 120 m below the stream thalweg and 300 m to the right of the stream. Because the model width was relatively small, the absolute water levels to the right of the stream were influenced by the right hand boundary. However, the change in groundwater head beneath the stream was not affected by this constant head boundary.

[8] No flow boundary conditions were applied along the top, left, and bottom boundaries, and the right boundary was assigned a constant, hydrostatic pressure head that

Table 1. Variables Used in Theoretical Model Simulations, Where K_c and K_a Are the Clogging Layer and Aquifer Hydraulic Conductivities, Respectively[a]

	Simulation								
					Aquifer Transmissivity				
Parameter	Base Case	Long Wave	Short Wave	Stream Width	Depth	K_a	K_c	K_c and K_a	Unsaturated Storage
Total Wave Time (day)	30	**2000**	**2**	30	30	30	30	30	30
Stream Width (m)	60	60	60	**15**	60	60	60	60	60
Aquifer Depth (m)	120	120	120	120	**1000**	120	120	120	120
K_c (m d^{-1})	0.1	0.1	0.1	0.1	0.1	0.1	**0.01**	**0.01**	0.1
K_a (m d^{-1})	5.0	5.0	5.0	5.0	5.0	**50.0**	5.0	**0.5**	5.0
Unsaturated Storage ($\theta_s - \theta_r$)	0.39	0.39	0.39	0.39	0.39	0.39	0.39	0.39	**0.19**
Ratio of Head Change (Aquifer: Stream)	4.89	7.76	3.62	1.92	3.89	0.81	0.40	1.60	6.45
Ratio of Peak Water Table Response (Altered: Base Case)		1.6	0.7	0.4	0.8	0.2	0.1	0.3	1.3

[a]The numbers in bold indicate the parameter value that was changed in each simulation, relative to the base case.

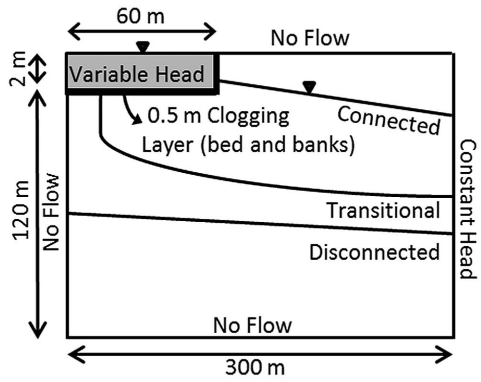

Figure 1. Cross section of two-dimensional Hydrus model used for generic simulations and sensitivity analysis. The clogging layer was applied to the base and bank of the stream.

fixed the water table at a prescribed distance below the stream elevation. Similar to the base case of *Brunner et al.* [2009a], the soil hydraulic properties of the aquifer were typical for sand, with residual and saturated moisture contents (θ_r and θ_s) of 0.045 and 0.43, respectively, and van Genuchten α, n, and m values of 14.5 m^{-1}, 2.68, and 0.5, respectively [*Carsel and Parrish*, 1988]. The saturated hydraulic conductivities of the clogging layer and aquifer (K_c and K_a) were assigned values of 0.1 and 5 m d^{-1}, respectively (Table 1).

[9] To establish the initial head distribution in the system, the model was run to steady state with a constant depth of 1 m in the stream, and with water table depths between 3 and 30 m below the stream elevation, specified at the right hand model boundary (see Figure 1). For each water table depth, the steady state flux across the streambed was documented (Figure 2). With a stream stage of 1 m the flow regime stayed connected until the water table at the right model boundary was lowered 7 m below the elevation of the stream thalweg. As the constant boundary head was lowered further, an unsaturated zone developed beneath the edge of the stream, which we define as a transitional connection. Once the right hand, constant head boundary was lowered 11 m below the stream elevation, there was an unsaturated region below the entire stream. At this point the aquifer was considered disconnected from the stream. Flux through the stream increased as the water table was lowered through the connected and transitional depths, then leveled off once the aquifer became disconnected from the stream (Figure 2).

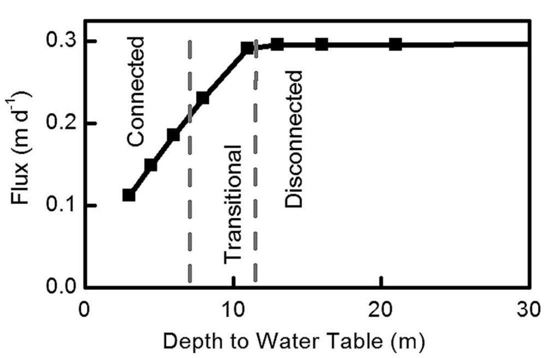

Figure 2. Modeled flux through the stream for water table levels beneath the constant head stream boundary indicating connected, transitional, and disconnected stream-aquifer relationships under steady state conditions with 1 m stream depth.

[10] The steady state head distribution was used as the initial condition for subsequent transient simulations. Based on the relationship in Figure 2 and the observed unsaturated zone development, water table depths (at the constant head boundary) of 3, 15, and 30 m below stream elevation were chosen to demonstrate aquifer response to a stream wave under different connection conditions. Although the 15 m water table depth resulted in a disconnected stream for a constant stream stage of 1 m (Figure 2), the water table rose to intersect the streambed as the stream stage increased during transient simulations, and therefore represented the transitional case for the transient simulations.

[11] For the transient simulations, stream stage was gradually increased (to ease unsaturated model calculations) from 1 to 2 m over a 10 day period, held at 2 m for 10 days, then decreased back to 1 m (see Figure 3 for shape of wave). The change in aquifer head over time was calculated beneath the midpoint of the half stream (i.e., at $w = 30$ m for all scenarios except the scenario specifically testing sensitivity to w). The determination of head directly beneath the stream minimized the attenuation and delay in the wave signal at the water table that would be observed away from the stream, and therefore presents the maximum possible aquifer response. By not placing this observation point directly at the true midpoint of the stream, any numerical

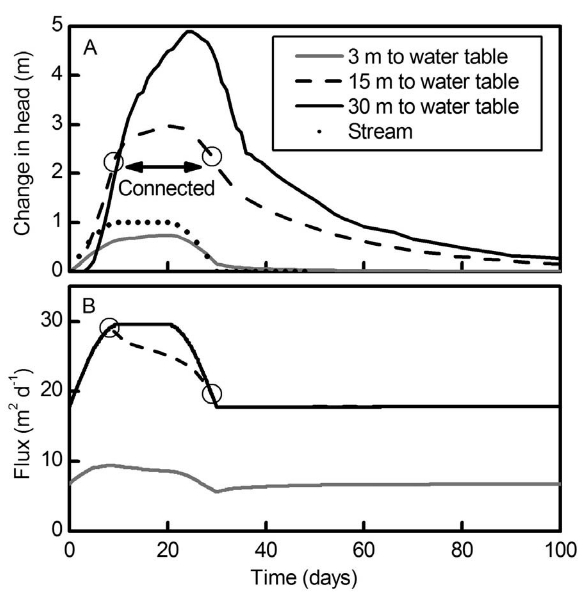

Figure 3. Base case transient simulations for water tables at the constant head boundary (see Figure 1) 3, 15, and 30 m below the stream elevation, respectively. (a) Response in the aquifer, and (b) flux through the stream boundary in response to a pressure wave in the stream. The circles on the 15 m depth to water table curve show the period for which the aquifer was connected.

artifacts of the left (no flow) boundary were avoided. The difference in the water table level between the true stream midpoint and the observation point was less than 0.8 m in all simulations. For the connected and transitional simulations, head was determined by recording the pressure head at an observation node 5 m below the stream, whereas in the disconnected model the depth to water table was noted for each time step. Finally, a series of model simulations was run at the 3 and 30 m water table depths (measured at the right hand boundary) in which the stream width (w), aquifer depth (B), K_c, both K_c and K_a, and unsaturated storage ($\theta_s - \theta_r$) were altered (Table 1). Both the depth and the hydraulic conductivity of the clogging layer control the volume of water that infiltrates through the streambed. However, the clogging layer in this study was kept at a uniform thickness of 0.5 m and its control on infiltration was examined exclusively through hydraulic conductivity.

3. Modeling Results: Base Case and Sensitivity Analysis

3.1. Base Case

[12] In response to the 1 m change in stream stage, head beneath the connected stream increased by 0.74 m, with the peak increase occurring 21 days after the stream stage began to rise (Figure 3a). Flux into the streambed increased for the first 10 days as the stream stage rose, then began to decline as the pressure beneath the stream continued to rise (Figure 3b). In the transitional case (water table at the right hand boundary 15 m below the stream elevation), the stream was initially disconnected from the aquifer, but a 1 m change in stream stage caused the water table to rise and intersect the streambed. The stream and aquifer were connected after 10 days. Flux into the streambed confirmed the switch from disconnected to connected, with the infiltration rate matching the disconnected stream for the first 10 days, then declining suddenly (Figure 3b). Maximum change in aquifer head was almost three times greater than the input signal, but occurred at roughly the same time as in the connected case. After 29 days the aquifer again began to disconnect from the stream. Change in aquifer head was most pronounced in the disconnected stream, which peaked 24 days into the simulation at a level almost five times greater than the input signal. A change of more than 1 m in the water table was observed over 150 m from the stream channel. After peaking at 4.9 m above baseline, the groundwater mound beneath the stream dissipated rapidly for 12 days and then more gradually. The infiltration rate was also greatest for this case, and maintained maximum infiltration for several days while the stream stage was constant at 2 m. During this time, the height of groundwater mound rose rapidly, as water infiltrated to the aquifer beneath the stream faster than it could escape laterally. Small instabilities in the disconnected stream water table level over time are an artifact of interpolation of the water table level between model nodes.

[13] Additional water table depths (i.e., at 6, 11, 13, 35, and 60 m as measured from the constant head boundary) were also tested, but omitted from the graph for clarity. All simulations with water table depths (at the right hand model boundary) less than 7 m below the stream (i.e., connected stream-aquifer conditions) had almost the same head response curves, with peaks within 10% of the one shown in Figure 3a. For losing, connected streams, total head cannot exceed the surface water pressure, and therefore the change in groundwater head cannot be greater than the change in stream stage. Once there is an unsaturated zone beneath the stream, the time to peak response is delayed, as was observed in the transitional case. Beneath this transitional stream, the water table can rise until the unsaturated zone is filled, after which the rate of rise slows. If the water table stays high, the aquifer level slowly approaches the surface water level. However, the initially deep water table allows for a greater magnitude of head change before the stream head is reached. In this case, the water table will continue to rise as long as the infiltration flux exceeds the ability for the aquifer to transmit water laterally. The magnitude of the head increase would therefore be expected to increase as the wave duration and height, stream width, and streambed permeability increase, and as the aquifer transmissivity decreases. The magnitude and timing of the aquifer response beneath a transitional stream is a combination of connected and disconnected responses, and depends on when the wave causes the stream and aquifer to reconnect. For clarity, we focus on the disconnected case in the following sensitivity analysis.

3.2. Wave Duration and Height

[14] Wave duration is a key parameter controlling the volume of infiltrated water and therefore the volume of water available to fill the unsaturated zone and mound beneath the stream. A step function was used to test the equilibrium disconnected aquifer response to a long duration wave. A very short increase in stream stage was simulated to test whether a short flood wave would cause any measurable aquifer response. For the long wave, head in the aquifer was increased by 1 m, as in the base case, but then held at the higher level for the remainder of the simulation (Figure 4a). For the short wave, the stream head was raised

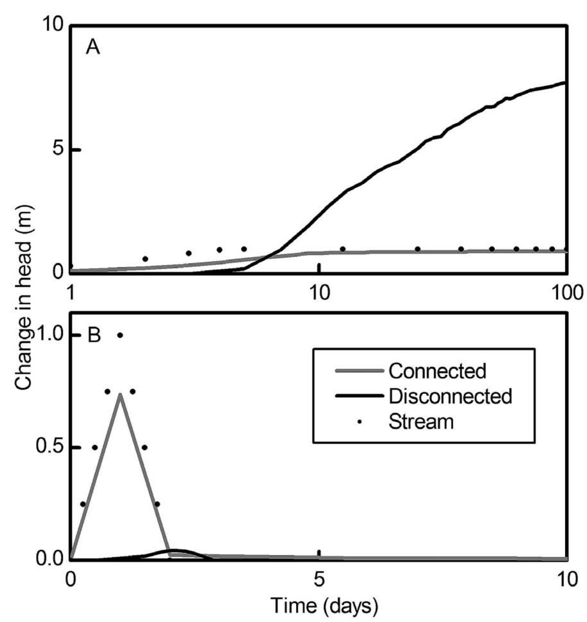

Figure 4. Connected and disconnected aquifer responses to (a) the stream stage permanently raised by 1 m, and (b) a short wave applied over 2 days.

by 1 m after 1 day and then returned to the base stream depth over the following day (Figure 4b). In both cases, head in the connected aquifer responded to the wave immediately, gradually increasing to a steady aquifer head 1 m above base conditions in the case of the long wave, and spiking 0.74 m above base conditions for the short wave (the same magnitude as the base case). In comparison, response in the disconnected aquifer was delayed by several days in both cases. For the long wave scenario, the groundwater level beneath the stream gradually rose almost 8 m, after which the aquifer reached a new, but still disconnected (as evidenced by the unsaturated zone present beneath the stream), steady state level. The short wave produced a minor change in water table level of 0.05 m beneath the disconnected stream, much less than the change in stream stage. Waves of shorter duration did not increase the level of the disconnected water table; the infiltration remained within the capacity of the unsaturated zone storage and did not reach the water table.

[15] Because the infiltration rate is dependent on stream depth, wave height also affects aquifer response. Initial simulations also included doubling and halving the increase in stream stage (i.e., increasing stream stage from 1 to 3 m and 1 to 1.5 m). The response scaled linearly with wave height and this parameter was omitted from Figure 4 for simplicity.

3.3. Stream Width

[16] Stream width is another parameter that controls the flux through the streambed. The stream half-width was reduced from 60 to 15 m to test the importance of this parameter. The observation point remained half way across the half stream; in this case at 7.5 m from the left hand boundary. For the connected system, the reduction in width reduced the peak head response in the aquifer relative to the base case by less than 10% (results not shown). The reduction of stream width did not affect the timing of the aquifer response beneath the disconnected stream relative to the base case, but the magnitude of the head response was 40% of the base case (Figure 5a). The ability of the aquifer to transmit water laterally is not affected by the stream width, and the groundwater mound does not reach the same magnitude beneath the narrower stream because there is less infiltration.

3.4. Aquifer Transmissivity

[17] Aquifer transmissivity was varied by individually changing the thickness of the aquifer and by increasing K_a to change the rate at which the groundwater mound could dissipate laterally. By increasing aquifer thickness from 120 to 1000 m, the peak head response in the aquifer beneath the connected stream increased by less than 10% (results not shown). Of course this change in transmissivity would likely cause a difference in the timing and magnitude of the response in wells located further from the stream. Increasing K_a by an order of magnitude caused the stream to disconnect from the aquifer, and it remained disconnected during an increased stream stage of 2 m.

[18] In the disconnected system the horizontal transmissivity of the aquifer determines how quickly the infiltrated water can move away (horizontally) from underneath the stream, and therefore more groundwater mounding would be expected at lower transmissivities. For the base case

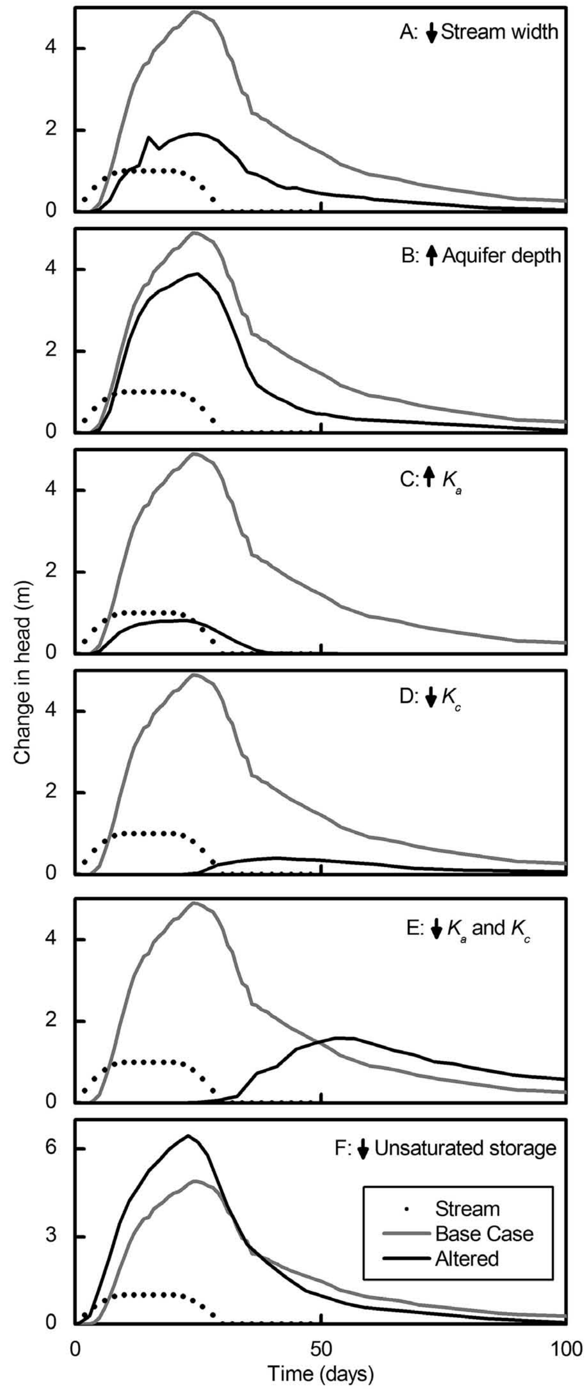

Figure 5. Disconnected aquifer responses to (a) a change in the width of the half stream from 60 to 15 m, (b) an increase in aquifer transmissivity by increasing aquifer depth from 120 to 1000 m, (c) an increase in aquifer transmissivity by increasing the aquifer hydraulic conductivity (K_a) by an order of magnitude, (d) an order of magnitude reduction in the hydraulic conductivity of the clogging layer, (e) an order of magnitude reduction in both the clogging layer and aquifer hydraulic conductivities, and (f) lower unsaturated storage (note change in y axis values).

aquifer depth of 120 m, the change in aquifer head beneath the disconnected stream was almost five times greater than the stream stage change. With a total aquifer depth of 1000 m (depth to water table at the constant head boundary was kept at 30 m), the time to highest groundwater mound was not altered relative to the base case, but the peak response was lowered by 20% (Figure 5b). Due to the importance of vertical head gradients, the effective transmissivity in the 1000 m deep aquifer was not increased in proportion to the increase in aquifer thickness. The magnitude of aquifer response was controlled by the volume of infiltration relative to the amount of unsaturated zone storage that had to be filled before the water table rose. The importance of lateral transmissivity was even more apparent at higher aquifer hydraulic conductivity values. An order of magnitude increase in K_a resulted in a change in water table level of less than 1 m, i.e., less than the input signal. Although the magnitude of change in water table height was not greater than the input signal, the response was still unique to a disconnected system because of the delay in peak water table rise.

3.5. Clogging Layer and Aquifer Conductivities

[19] Both the clogging layer thickness and conductivity (K_c) determine the rate and volume of water infiltrating beneath the stream, while K_a determines how quickly the water flows laterally away from the stream. Therefore, aquifer response beneath both connected and disconnected streams is expected to decrease as K_c is lowered or K_a is increased.

[20] For the connected case, reducing K_c by an order of magnitude reduced the head response below the stream by almost an order of magnitude (head change of 0.01 m). However, lowering both K_c and K_a by the same amount resulted in a head change within 5% of the base case (results not shown). As the ratio of K_c to K_a was maintained, the ratio of volume of infiltrating water (controlled by K_c) to the ability of the aquifer to transmit water away from the stream (controlled by K_a) was also maintained. For the disconnected scenarios in which K_c and both K_c and K_a were lowered, less water was able to flow into the aquifer and the magnitude of aquifer response in the disconnected cases was greatly reduced, as expected (Figures 5c and 5d). Peak groundwater mounds were 8% and 33%, respectively, of those observed for the base case. The reduced volume of infiltration resulted in a longer time to fill unsaturated storage between the stream and water table. Therefore, the delay between stream wave and groundwater response was also significantly greater than for the base case, causing highest groundwater levels 64% and 112% later than in the base case scenario, respectively. In comparison to reducing only K_c, reducing both K_c and K_a but maintaining the ratio between these two parameters increased the time to peak aquifer response but resulted in less attenuation of the peak. Although less water infiltrated through the streambed, the water was also not able to move away laterally as quickly (Figure 5d). At much lower K_c or higher K_a values, no response to the flood wave would be expected.

3.6. Unsaturated Storage

[21] The magnitude and timing of aquifer head response beneath a disconnected stream is highly dependent on the amount of available storage in the unsaturated zone. If the moisture content is high, the aquifer should respond faster and form a higher groundwater mound because there is less unsaturated storage to fill. For the sensitivity analysis, θ_r was set to 0.24, decreasing the unsaturated storage ($\theta_s - \theta_r$) by half. In this case the water table rose to a peak of 6.4 m above baseline (Figure 5e), which represents a 30% increase in head change relative to base case conditions.

3.7. Relative Sensitivity

[22] Sensitivity of the water table to all of the tested parameters is due to parameter control on one of three factors: (1) the volume of water infiltrated through the stream, (2) the volume of water that reaches the water table, or (3) the ability of the aquifer to move water laterally away from the stream (Table 1). The parameters that control the first factor are wavelength, stream width, and K_c. Unsaturated storage determines the second factor and aquifer transmissivity and K_a control the third factor. For the parameter values tested, the magnitude of change in the groundwater level decreased to below that of the input signal for the high K_a, low K_c, and short wave cases (Table 1). The relative sensitivity of the groundwater mound height was highest for these parameters. Disconnected head response was always delayed in comparison to the instantaneous, connected response. Increases in wave duration, stream width, and K_c resulted in greater infiltrated water volume and therefore a higher peak response at the water table. Conversely, higher aquifer transmissivity allowed the water to move laterally away from the stream faster, resulting in a smaller aquifer response below the stream. Aquifer response was greater and more rapid for lower values of specific yield. Furthermore, lower K_c values resulted in greater delays to peak response, because less water infiltrated through the streambed and therefore it took longer for the unsaturated zone to fill.

4. Field Example

[23] The Lachlan River is a tributary to the Darling River in southeastern Australia, with an annual average flow of 40 $m^3\ s^{-1}$. The river is underlain by unconsolidated alluvial sediments, with a heterogeneous mixture of sand, clay, silt, and sandy clay lenses identified in well logs at the Gonowlia Weir study site, 20 km upstream of the town of Hillston.

[24] In 2009, pressure transducers were installed in the river and at a piezometer located 20 m perpendicular to the river bank. The river was thought to be disconnected from the deeper aquifer at that time. However, the streambed is highly heterogeneous and it was not possible to verify the presence of a clogging layer and an unsaturated zone beneath the stream using vertical sediment profiles [*Lamontagne et al.*, 2010]. No stream data was available during the periods from 3 to 18 September 2009 and 1 December 2009 to 19 January 2010. Between 19 January and 15 February 2010, the recorded water level in the stream did not rise above the logger, located approximately 0.84 m above the stream thalweg.

[25] The piezometer data show the response of the perched aquifer to flood events in the river (Figure 6a). The water table varied between 2 and 4.5 m below stream level, indicating that the river was losing, and the steep gradient between the stream stage and water level at the piezometer

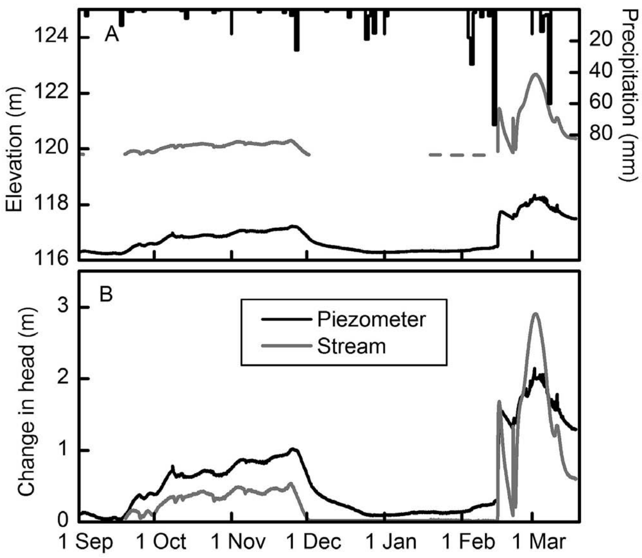

Figure 6. (a) Stream and piezometer data collected in 2009–2010 from the Lachlan River, Australia (reproduced from *Lamontagne et al.* [2010]). No stream levels were available for the periods 3 to 18 September 2009 and 1 December 2009 to 19 January 2010. The dashed line from 19 January to 15 February 2010 represents the period when data were available but water levels were less than the logger elevation. Precipitation recorded at Hillston airport (Government of Australia Bureau of Meteorology site 175,032). (b) Lachlan River change in head above stream and piezometer levels at the start of the study period.

suggests disconnection of the stream and aquifer. Normalizing the stream depth and piezometer level by their base levels at the beginning of the study period highlights the aquifer response to changes in stream stage (Figure 6b).

[26] Connected, transitional, and disconnected stream-aquifer relationships may be present in these data. As seen in Figure 6b, during the long, two-month event between 19 September and 24 November, the aquifer response to increases in the stream stage was approximately twice the magnitude of the change in stream stage. The direct influence of rainfall on piezometer levels is not considered likely during this long event, due to the close correlation between the stream and piezometer variations and the lack of large rainfall events during this period (Figure 6a). Therefore, if the change in the piezometer was solely due to streamflow, the stream and aquifer must have been disconnected at the onset of this event and must have stayed disconnected or transitional through most of the two months of increased stream stage. The rapid decline in aquifer level at the end of November shows rapid infiltration through the relatively shallow unsaturated zone, and may indicate that the system was transitional at this time.

[27] In contrast, the second event is more difficult to interpret. In January 2010, water level in the piezometer was at the same level as that observed before the first event, and therefore the system was clearly disconnected. On 15 February the water level in the piezometer increased 2 m and stayed high throughout the end of the month in response to a sudden, 3 m change in stream stage. The short duration of the second event captured with the stream and piezometer data may not have been sufficient to cause connection between the stream and aquifer during the study period. Infiltration at the study site following local, heavy rainfall on 14 February and 8 March may also have influenced the observed aquifer response. The lack of data for two months previous to the event commencing 15 February, combined with a smaller response in the aquifer as compared to the stream and significant rainfall at the site therefore make it difficult to discern connection status for this event. As seen in the sensitivity analysis, a greater response in the aquifer in comparison to stream stage can only be associated with a disconnected stream, but a comparatively smaller response can be observed in any connection status. Furthermore, heterogeneity in the aquifer and the unknown extent and hydraulic conductivity of the aquitard separating the shallow and deep aquifers complicate interpretation of connection status for both events, but especially where the aquifer response was not greater than stream stage change.

5. Discussion and Conclusions

[28] The timing and magnitude of aquifer head response beneath disconnected streams is a complex interplay between the parameters controlling infiltration and lateral movement of water away from the stream. In all tested scenarios, the disconnected aquifer head response was distinct from the connected response, and analytical methods to determine connected aquifer head response cannot be applied to the disconnected system. Numerical solutions show that for transitional and disconnected streams, change in aquifer water level can exceed the change in stream stage. This effect, and the factors controlling it, has not been specifically discussed in the literature. We found the transient disconnected aquifer response to be sensitive to variability in all tested parameters.

[29] Further parameters that may affect disconnected aquifer response include the soil water retention parameters and heterogeneity in the streambed. Analysis was limited to one soil type in this study; however, differences between soil types were partially addressed through comparison of the effect of specific yield, as the presence and amount of unsaturated zone storage is the key difference between connected and disconnected aquifers. The effects of heterogeneity within the streambed are complex and beyond the scope of the current study. However, the sensitivity of the groundwater response to variations in both K_a and K_c shows that a disconnected or transitional stream with significant heterogeneity in the streambed or shallow aquifer may make it difficult to interpret stream infiltration based on localized groundwater response (as also documented by *Fleckenstein et al.* [2006] and *Irvine et al.* [2012]). The joint presence of both saturated and unsaturated areas in heterogeneous streambeds complicates the determination of the state of connection [*Frei et al.*, 2009; *Irvine et al.*, 2012].

[30] The current study was motivated by the need for a better method of determining stream-aquifer connection. The ability to determine the connection status from commonly available field data such as stream and piezometer levels would allow more accurate determination of storm recharge and the development of water management

strategies based on this information. Through the sensitivity analysis it is apparent that an aquifer response of greater magnitude than the driving change in stream stage can only arise in a disconnected system. This type of response was identified for period of the Lachlan River data, indicating disconnection at that time. For periods where the magnitude of the aquifer response is not greater than variation in stream stage, the magnitude of aquifer response alone cannot be used to determine connection status, and the stream and aquifer may be connected, transitional, or disconnected. Because the magnitude of the disconnected groundwater response does not scale linearly with change in each of the tested parameters, a simple rule of thumb for discerning between connected and disconnected response is not yet possible. The field example demonstrated the current limitations of using stream and piezometer levels to discern connection status when there are other unknown factors. However, this study provides an overview of the range of the magnitude and timing of disconnected aquifer head responses that may be observed under a range of conditions. Although not always of greater magnitude than the connected aquifer response, the disconnected response was fundamentally different under all tested conditions. With further study it may be possible to develop a convenient and reliable method for determining connection status from flood wave and aquifer response data.

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