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Vegetation controls on variably saturated processes between surface water and groundwater and their impact on the state of connection

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[1] The vadose zone plays an important role in surface water–groundwater interaction and exerts strong influences on biogeochemical, ecological, and hyporheic processes. It is also the presence of an unsaturated zone that controls the state of connection between surface water and groundwater. Despite recent advances on how hydrogeological variables affect surface water–groundwater interactions, there is limited understanding of the hydroclimatic effects of precipitation and evapotranspiration. More specifically, there is a need for a physically based understanding on the changes that may occur in response to changes in vegetation. While it may seem qualitatively obvious that the presence of vegetation can cause an unsaturated zone to develop underneath a riverbed and alter the state of connection, it has so far not been demonstrated quantitatively. Also, the influence of variables such as root extinction depth, topography, and the influence of land clearance has so far not been explored. In this study, fully coupled, physically based 2-D transient homogeneous models were used to simulate the impact of land clearance and revegetation on the state of connection of a perennial river system. The simulations showed that the presence of vegetation can create an unsaturated zone between a river and an aquifer and affect the state of connection and that the removal of deep-rooted vegetation from a catchment may have a significant impact on the state of connection as well as the condition of the water resource.

1. Introduction

[2] Vadose zone processes play an important role in surface water–groundwater interaction. For example, the presence of an unsaturated zone has a strong influence on biogeochemical processes of river systems [*Bencala*, 1993], the fate of nutrients [*Boulton et al.*, 1998], and various ecological and hyporheic exchange processes [*Brunke and Gonser*, 1997; *Findlay*, 1995]. Also, it is the presence of an unsaturated zone which controls the state of connection between surface water and groundwater. The state of connection has received greater attention in the last decade [*Brunner et al.*, 2009a; *Brunner et al.*, 2011; *Fox and Durnford*, 2003; *Harvey and Wagner*, 2000; *Vazquez-Sune et al.*, 2007], in response to concerns about water scarcity and the sustainable management and allocation of water resources [*Sophocleous*, 2002; *Winter et al.*, 1998].

[3] In natural environments, surface water–groundwater systems are influenced by physical (hydrogeological), topographical (terrain and landform), and hydroclimatic (precipitation and evapotranspiration) variables and are generally classified as connected or disconnected systems. Connected systems are either (1) gaining, where groundwater discharges through the riverbed to contribute to river flow, or (2) losing, where water infiltrates from the river to the groundwater system. Losing systems can sometimes be disconnected. Disconnected systems show flow losses through an unsaturated zone, and as a result changes in the water table do not significantly affect the infiltration rates from the river. The flow regime between connected and disconnected is called transitional and is the state between the initial development of an unsaturated zone and the point where the infiltration rate no longer changes in response to a further decline in the water table. In transitional and disconnected systems, an unsaturated zone under the riverbed is present [*Brunner et al.*, 2009b; *Brunner et al.*, 2011].

[4] In a recent study, *Brunner et al.* [2009b] used a theoretical and modeling approach to examine the most important hydrogeological parameters that influence the state of connection between surface water and groundwater and developed a set of criteria to determine whether a system can become disconnected or not in the presence of a low conductivity streambed. *Reisenhauer* [1963] showed that a disconnection is possible even in the absence of a clogging layer due to capillary effects. However, we are unaware of any field documentation to support such a case, and *Peterson and Wilson* [1988] concluded this is unlikely to occur. Typically, an unsaturated zone (and therefore a disconnection) can develop in the presence of a streambed with a lower hydraulic conductivity than the underlying aquifer.

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Such low conductivity layers (subsequently called a clogging layer) are widespread and have been observed in rivers throughout all climatic zones. Low conductivity layers can develop due to biological clogging [*Treese et al.*, 2009] or sedimentary processes [*Schälchli*, 1992]. *Hatch et al.* [2010] showed using field data that heterogeneity in the streambed can result in the presence of both saturated and unsaturated zones adjacent to one another. *Frei et al.* [2009] reached the same conclusion through simulating the effect of heterogeneity on surface water–groundwater interaction. Other modeling studies [*Niswonger and Fogg*, 2008] have investigated the hydrogeological effects on perched stream aquifer type systems, while *Desilets et al.* [2008] focused on the pathways and rates of infiltration as stream–aquifer systems transition from connected to disconnected and how these effect local groundwater flow patterns. It has also been shown that groundwater pumping adjacent to surface water systems can have a considerable effect on the interaction between surface water and groundwater [*Moore and Jenkins*, 1966; *Spalding and Khaleel*, 1991; *Su et al.*, 2007]. *Fox and Durnford* [2003] showed that groundwater pumping adjacent to a surface water body can induce an unsaturated zone and a disconnection between surface water and groundwater.

[5] Despite these recent advances in knowledge on the relation between hydrogeological variables, the presence of an unsaturated zone and the state of connection, there is limited quantitative understanding of the role of vegetation (i.e., evapotranspiration) in forming an unsaturated zone between surface water and groundwater. It is also unclear, in a precise quantitative cause and effect manner, how land clearance or revegetation affects the state of connection. In Australia, evapotranspiration is an important hydroclimatic process because native vegetation clearance has had considerable impacts on surface water and groundwater salinities, through the mobilization of salt stored in the shallow regolith as well as increased groundwater discharge [*Allison and Forth*, 1982; *Allison et al.*, 1990; *Cook et al.*, 1994; *George et al.*, 1997]. *Williamson et al.* [1987] compared water and salt balances from cleared and uncleared catchments in south-western Australia and showed that in the cleared catchments there was a significant increase in streamflow and exported salt, and that the water and salt balances had not yet reached a new equilibrium.

[6] There is an analogy that trees behave like groundwater pumps and that they are able to remove or intercept a large portion of precipitation input in a catchment water balance. While trees and pumps might have similar effects on the water balance, the way they extract water from the aquifer is fundamentally different. It is not intuitively obvious how a change in the water table relates to a tree's ability to consume groundwater [*Butler et al.*, 2007; *Loheide et al.*, 2005; *Shafroth et al.*, 2005], let alone effect its ability to create an unsaturated zone between a river and an aquifer or alter the state of connection. Several studies have calculated the amount of water certain vegetation types transpire on an annual basis [*Farrington et al.*, 1994; *Salama et al.*, 1994]. Results from these studies found that 11.4–18.0 m^3 of water was transpired per year per eucalyptus tree in Western Australia. *Banks et al.* [2011] hypothesized that there may be important vegetation (evapotranspiration) controls on the state of connection between surface water and groundwater in a pristine catchment on Kangaroo Island, South Australia. Their study suggested that the presence of vegetation was a fundamental difference in controlling the surface water–groundwater interactions compared to adjacent catchments that had been cleared of vegetation and which had significantly more saline surface water [*Henschke et al.*, 2003; *Shand et al.*, 2007]. The results of the study by *Banks et al.* [2011] also showed that the relatively low salinity of the fresh water river system can be maintained in an otherwise saline regional groundwater system by virtue of the dominantly losing state of connection. It was hypothesized that this losing state was created and maintained by vegetation cover in the pristine catchment. Qualitatively, it may seem obvious that the presence of vegetation can alter the state of connection or that a change to the rates of precipitation and evapotranspiration may, under certain conditions, have some effect on the state of connection. However, there is little understanding what the quantitative effects will be and the sensitivity of the state of connection (and associated exchange fluxes) to various controlling physical variables. Such variables include the hydraulic conductivity of the aquifer and clogging layer, catchment slope, and root extinction depth.

[7] The aim of this paper is to explore the hypothesis: can the presence of vegetation create an unsaturated zone or even a disconnection between a river and an aquifer using reasonable and representative vegetation and hydroclimatic variables? We further explore how land clearance can affect the state of connection. In this context we study the following question: is evapotranspiration a plausible mechanism to create an unsaturated zone underneath a riverbed and how does it influence connected gaining and losing, and losing–disconnected type conditions? The physical controls of catchment slope, the hydraulic conductivity of the riverbed clogging layer and aquifer, and the vegetation root extinction depth (transpiration extinction depth) are also examined to determine how these variables influence the state of connection. Understanding how vegetation type and cover (and hence evapotranspiration) relates to an unsaturated zone and affects the state of connection provides valuable information on what the impacts of vegetation clearance, revegetation or changes in land use are likely to be on surface water–groundwater connection and the consequential effects on water quality. Similarly, climate change impacts can be qualitatively inferred from these general cause and effect type relationships.

2. Numerical Modeling

2.1. Conceptualization of Disconnected Systems

[8] Whether a surface water–groundwater system is connected, disconnected, or in transition between the two states has profound implications on how changes in the water table affect the exchange fluxes in the system. In the presence of a clogging layer within the streambed, lowering the water table can result in an unsaturated zone under the clogging layer; the system first enters a transition mode. Further lowering the water table increases the infiltration flux, and a maximum value for the current hydraulic system is approximated. This upper limit of infiltration (corresponding to an upper limit of suction under the clogging layer) can be calculated. However, in reality this upper limit is only approximated and therefore a cut-off value

that separates transition from disconnection has to be defined. In this study, the cut-off between disconnected and transitional type systems was defined at 1% of the upper limit of pressure head (suction; γ_p^*) at full disconnection. The pressure head at full disconnection was determined using the following equation, and it can be solved graphically [*Osman and Bruen*, 2002] or numerically [*Brunner et al.*, 2009b].

$$K_c \frac{\left(h_c + d - \gamma_p^*\right)}{h_c} = K_a k_r \left(\gamma_p^*\right), \tag{1}$$

where γ_p^* is the pressure head that develops at the interface between the clogging layer and the aquifer at full disconnection, K_c is the hydraulic conductivity of the clogging layer, K_a is the hydraulic conductivity of the aquifer, h_c is the thickness of the clogging layer, d is the depth of the river, k_r is the relative hydraulic conductivity of the aquifer, which is derived from a relationship between pressure and hydraulic conductivity according to pressure-saturation curves from the work of *van Genuchten* [1980]. The left-hand side of equation (1) is the infiltration flux through the clogging layer calculated following Darcy's law. For disconnected systems the infiltration under the clogging layer is driven by gravity drainage, therefore, the infiltration flux equals the hydraulic conductivity of the aquifer (right-hand side of the equation; $K_a k_r$). Equation (1) cannot be solved analytically due to the highly nonlinear relations between pressure and relative hydraulic conductivity.

2.2. Numerical Model: HydroGeoSphere

[9] Surface water–groundwater interactions were simulated using the groundwater flow model HydroGeoSphere (HGS) [*Therrien et al.*, 2010]. HydroGeoSphere is a physically based, numerical model describing fully integrated surface and unsaturated and saturated flow in the subsurface. The capability of HGS to model flow in the unsaturated zone using the Richards equation as well as disconnection between surface water and groundwater has significant benefits over modeling codes that do not explicitly consider the unsaturated zone as outlined in the manuscript by *Brunner et al.* [2010]. We limit the description of HGS to the conceptualization of evapotranspiration. For further details on the code and a recent software review the reader is referred to *Therrien et al.*, [2010] and *Brunner and Simmons* [2011].

[10] Evapotranspiration (ET) is modeled as a combination of plant transpiration and evaporation, and affects both the surface and subsurface flow domains. Transpiration from vegetation occurs within the root zone of the subsurface and is a function of the leaf area index (LAI) [dimensionless], nodal water (moisture) content (θ) [dimensionless] and a root distribution function (RDF) over a prescribed extinction depth. Water content is simulated as saturation because it is more stable and always varies between 0 and 1, while in reality moisture content varies from 0 to a value equal to porosity. As we will discuss later in the base case setup, the effects of vegetation clearance and revegetation are simulated by modifying the extinction depth, e.g., a small depth value is used for a change from native vegetation to shallow rooted pasture crops and a zero depth value is used for no vegetation (land clearance). The rate of transpiration (T_p) is estimated using the following relationships [*Kristensen and Jensen*, 1975]:

$$T_p = f_1(\mathrm{LAI}) f_2(\theta) \mathrm{RDF}\left[E_p - E_{\mathrm{can}}\right], \tag{2}$$

where E_p is the reference potential evapotranspiration which may be derived from pan measurements or computed from vegetation and climatic factors [L T^{-1}] and E_{can} is the tree canopy evaporation [L T^{-1}]. E_p can also be described as the amount of water that would be removed through evapotranspiration if the water table was at the ground surface. The value and description of E_p has followed the notation and conceptualization of *Therrien et al.* [2010] and *Kristensen and Jensen* [1975]. The vegetation function (f_1) correlates the transpiration (T_p) with the leaf area index (LAI) in a linear fashion and is expressed as

$$f_1(\mathrm{LAI}) = \max\{0, \min[1, (C_2 + C_1 \mathrm{LAI})]\}. \tag{3}$$

[11] The root zone distribution function (RDF) is defined by the relationship:

$$\mathrm{RDF} = \frac{\int_{c_1}^{c_2} r_F(z) dz}{\int_0^{L_r} r_F(z) dz}. \tag{4}$$

[12] The moisture content (θ) function (f_2) correlates T_p with the moisture state at the roots and is expressed as

$$f_2(\theta) = \begin{cases} 0 & \text{for } 0 \le \theta \le \theta_{wp} \\ f_3 & \text{for } \theta_{wp} \le \theta \le \theta_{fc} \\ 1 & \text{for } \theta_{fc} \le \theta \le \theta_o \\ f_4 & \text{for } \theta_o \le \theta \le \theta_{an} \\ 0 & \text{for } \theta_{an} \le \theta \end{cases}, \tag{5}$$

where:

$$f_3 = 1 - \left[\frac{\theta_{fc} - \theta}{\theta_{fc} - \theta_{wp}}\right]^{\frac{c_3}{E_p}}, \tag{6}$$

$$f_4 = 1 - \left[\frac{\theta_{an} - \theta}{\theta_{an} - \theta_o}\right]^{\frac{c_3}{E_p}}, \tag{7}$$

and where C_1, C_2, and C_3/E_p are dimensionless fitting parameters, L_r is the effective root length [L], z is the depth coordinate from the soil surface [L], θ_{fc} is the moisture content at field capacity, θ_{wp} is the moisture content at the wilting point, θ_o is the moisture content at the oxic limit, θ_{an} is moisture content at the anoxic limit and r_F (z) is the root extraction function [L^3 T^{-1}] which typically varies logarithmically with depth. Below the wilting point moisture content, transpiration is 0; transpiration then increases to a maximum at the field capacity moisture content. This maximum is maintained up to the oxic moisture content, beyond which the transpiration decreases to 0 at the anoxic moisture content. When available moisture is larger than the anoxic moisture content, the roots become inactive due to lack of aeration [*Therrien et al.*, 2010].

[13] In HGS, evaporation from the soil surface and subsurface soil layers is a function of nodal water content and

an evaporation distribution function (EDF) over a prescribed extinction depth. The model assumes that evaporation (E_s) occurs along with transpiration, resulting from energy that penetrates the vegetation cover and is expressed as [*Therrien et al.*, 2010]

$$E_s = \alpha^* \left(E_p - E_{\text{can}}\right)[1 - f_1(\text{LAI})]\text{EDF}. \tag{8}$$

[14] The wetness factor (α^*) is given by

$$\alpha^* = \begin{cases} \dfrac{\theta - \theta_{e2}}{\theta_{e1} - \theta_{e2}} & \text{for} \quad \theta_{e2} \le \theta \le \theta_{e1} \\ 1 & \text{for} \quad \theta \ge \theta_{e1} \\ 0 & \text{for} \quad \theta \ge \theta_{e2} \end{cases}, \tag{9}$$

where θ_{e1} is the moisture content at the end of the energy–limiting stage (above which full evaporation can occur) and θ_{e2} is the limiting moisture content below which evaporation is 0.

2.3. Conceptual Model

[15] The conceptual model employed here was based on the research outcomes of a field-based study which investigated the state of connection between a fresh water river, a perched sedimentary aquifer and a saline fractured rock aquifer system in the pristine Rocky River catchment on Kangaroo Island, South Australia [*Banks et al.*, 2011]. The long-term mean annual precipitation for this catchment is 780 mm yr^{-1} and the mean reference potential annual evapotranspiration is 1400 mm yr^{-1}. Evapotranspiration is greater than precipitation during the summer months while in winter there is potential for groundwater recharge when precipitation exceeds evapotranspiration (Figure 1).

[16] We defined a conceptual model that was used for all simulations (Figure 2). A base case was defined by assigning properties such as hydraulic conductivities, the slope of the catchment and the evaporation and vegetation parameters to the conceptual model. We then varied the hydraulic conductivity of the aquifer and the clogging layer as well as the slope and the evaporation and vegetation extinction depths to understand how these parameters relate to the influence of vegetation. While a number of other parameters could be varied additionally (e.g., porosity, the retention functions of the aquifer or surface properties such as rill storage height), the analysis was limited to these key parameters in order to keep the interpretation focused.

[17] To analyze how vegetation affects the interaction between surface water and groundwater, forcing functions have to be applied to the system. A common way to conceptualize forcing functions in surface water–groundwater systems is to apply head boundaries to the water table [*Bruen and Osman*, 2004; *Brunner et al.*, 2009b; *Osman and Bruen*, 2002]. As pointed out by several authors [*Freeze*, 1974; *Lewandowski et al.*, 2009; *Panday and Huyakorn*, 2004], natural systems are often flux controlled, and therefore hydraulic heads are not themselves the forcing function. Instead, hydraulic heads are the response to changes in fluxes such as recharge, evaporation, transpiration or pumping. In order to study the effect of vegetation on surface water–groundwater interaction, we therefore applied precipitation and evapotranspiration as forcing functions. This approach is conceptually different to the aforementioned modeling approaches by *Bruen and Osman* [2004] and *Osman and Bruen* [2002].

[18] We addressed the questions raised in the introduction with a homogeneous 2-D model (Figure 2) and did not consider any transience in the surface water domain. A simple conceptual model was chosen for a number of reasons: (1) In a more complicated setting (e.g., heterogeneous streambed and aquifer), it would be very difficult to examine and deconvolute the responses in the flow regime to changes in evaporation and transpiration processes; and (2) perennial river

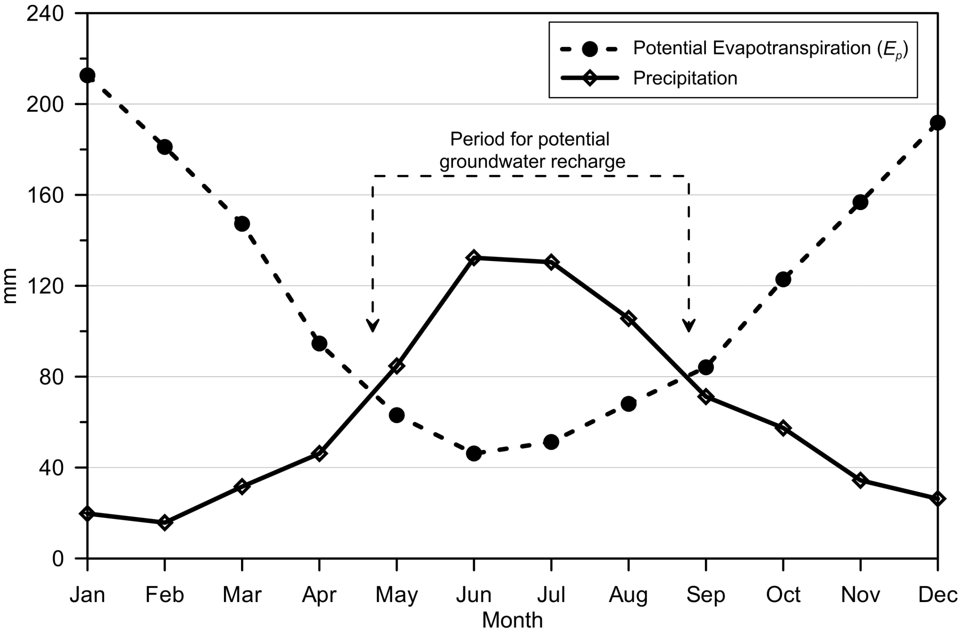

Figure 1. Monthly precipitation and reference potential evapotranspiration (E_p) for Rocky River catchment.

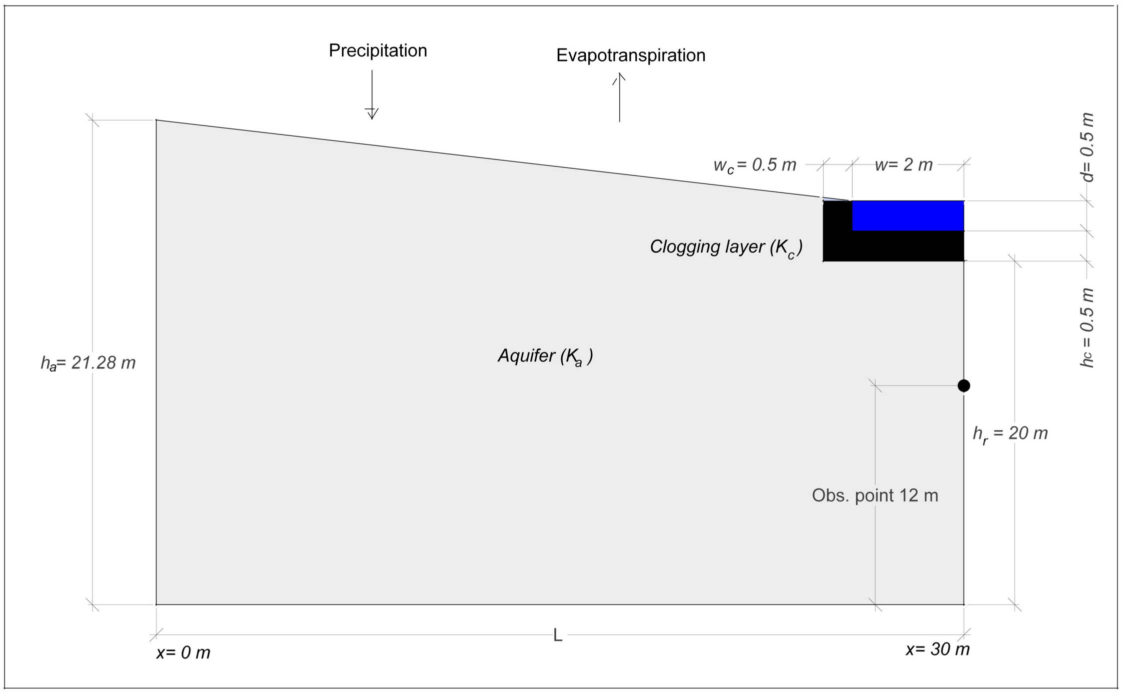

Figure 2. Conceptual model of the 2-D surface water–groundwater system. The river and aquifer are separated by a clogging layer (h_c) that is 0.5 m thick and 0.5 m wide at the river edge (w_c). The clogging layer has a hydraulic conductivity (K_c) that is less than the hydraulic conductivity of the aquifer (K_a). The river is defined by a constant head boundary with a water depth (d) of 0.5 m and has a width (w) of 2 m. L is the length (30 m) of the model in the x-direction. The height of the left-hand side of the model (h_a) at $x = 0$ is 21.28 m and the right-hand side (h_r) at $x = 30$ is 20.0 m. The left- and right-hand side and the base of the model are all no flow boundaries. The observation point directly beneath the center of the river at 12 m elevation is also shown.

conditions (defined by a constant head) maintained 100% saturation in the clogging layer. Simulating the temporal dynamics of ephemeral or intermittent streams and the wetting and drying cycles of the clogging layer undermine a unique and unambiguous interpretation of the systems response to the effects of vegetation and hence evapotranspiration.

[19] The model boundary conditions were designed according to the conceptual model in Figure 2. Because of the symmetry of the conceptual model only the left-hand side of the catchment was represented in the model domain to reduce computational time. The left- and right-hand side and the base of the model were all no flow boundaries. The river was perennial and represented as a constant head boundary. The horizontal extent of the model domain was designed in a way so that there were no significant impacts of the boundary conditions on the near river environment. The model discretization was fine enough to ensure grid-independent results and to provide an appropriate level of vertical detail of the unsaturated zone. To ensure that this was the case, 60 subsurface layers were used. For example, below the river the vertical discretization was set to 0.1 m from the top of the model domain down to 15.5 m elevation. Below 15.5 m elevation to the base of the model domain the layer thickness was increased and ranged from 0.2 up to 4 m in the base layer. The horizontal discretization increased from 0.02 m near the river edge to 1 m at the left boundary of the model domain. The datum elevation (0 m) is located at the base of the aquifer.

[20] The river was represented as a channel 0.5 m deep (vertical direction = 20.5 − 21 m) and 2 m wide (x-direction = 28 − 30 m). A slope between the river and the top left boundary was defined, with values provided in section 2.4. The slope was varied in the other scenarios tested. The clogging layer extended 0.5 m beneath the river channel and 0.5 m upslope from the edge of the river bank. The choice of this thickness is not critical for the following reason: The ratio between the hydraulic conductivity of the clogging layer and its thickness is the first-order control of infiltration flux from the river to the aquifer. Given that the hydraulic conductivity of the clogging layer can vary by many orders of magnitude there is a large degree of freedom in choosing the thickness of the clogging layer.

[21] The identification of the presence of an unsaturated zone is straight forward: if the water table drops below the clogging layer, an unsaturated zone develops. To determine whether the system is disconnected or not the hydraulic head beneath the center of the river needs to be evaluated [*Brunner et al.*, 2009b]. Observation nodes were used to obtain point specific model outputs such as the hydraulic head or the position of the water table. For the purposes of this study, an observation borehole was located 12 m above the reference datum beneath the center of the river. We used the hydraulic head of this borehole to approximate the location of the water table (defined through pressure = 0). This approximation was implemented to accelerate post-processing the large amount of model output data. A

systematic comparison of a representative number of models was carried out and revealed that the largest deviation between the hydraulic head at an elevation of 12 m and the true location of the water table beneath the center of the river was at greatest around a centimeter. We therefore considered the hydraulic head at this borehole location as a sufficiently accurate approximation of the location of the water table.

2.4. Base Case Setup

[22] The base case scenario was a transient model setup for a period of 7304 days (~20 years) using an initial time step of 0.1 days, a maximum time step of 1 day and a maximum time step multiplier of 1.25. The initial conditions of the model were determined numerically from a dynamic steady state under vegetated type conditions (evapotranspiration taking place. Based on these initial conditions the model was run for a period of 10 years (3652 days) at which point the vegetation was removed or modified (through a change of the extinction depth) and the model was run for a further 10 years. The slope of the catchment was 1 cm m^{-1}. As a result, the ground elevation at the edge of the river (x-direction = 28 m) was 21 m and at the left boundary (x-direction = 0) was 21.28 m.

[23] The physical properties of the clogging layer and the more hydraulic conductive homogeneous aquifer were based on representative literature values [*Carsel and Parrish*, 1988; *Freeze and Cherry*, 1979]. The soil moisture retention curve for the aquifer, defined by the van Genuchten parameters α and β, was kept constant (as opposed to the hydraulic conductivity which is varied around the base case). No retention curve needed to be defined for the clogging layer because it remained saturated for the entire simulation due to perennial river conditions (i.e., saturation of the clogging layer occurs when the hydraulic conductivity of the clogging layer is less than the hydraulic conductivity of the aquifer).

[24] Equation (1) was used to determine the critical water table beneath the center of the river that defines the border between transition and disconnection of surface water and groundwater. In the base case, this critical water table was at 18.68 m above the base of the model (reference datum = 0 m), or 1.32 m below base of the clogging layer (i.e., water table depth below a disconnected infiltration zone) for the physical parameters used.

[25] Evapotranspiration was dynamically simulated as a combination of evaporation (equation (8)) and transpiration (equation (2)) processes by removing water from all model cells of the surface and subsurface flow domains within the defined zone of the evaporation and root extinction depths. To simulate evaporation only, the transpiration process was shut down by changing the root extinction depth and LAI to 0 (i.e., from the last day of the first 10 year period [day 3652] to the next day of the following 10 year period [day 3653] there was no transpiration in the base case). The daily reference potential evapotranspiration (E_p) rate (equations (2) and (8)) was based upon the historical average daily reference potential evapotranspiration data for the Rocky River catchment (Figure 1) and transpiration parameters, typical of native vegetation in southern Australia (Table 1). The evapotranspiration processes were simulated for the entire model duration beginning on day 1. For example the E_p for the first day in the simulation is 0.0065 m and represents the average daily potential reference evapotranspiration for 1 January from the weather station from over 30 years of historical data. The E_p value for each day of the year is then repeated again for the next year and so on for the entire model duration.

Table 1. Notation, Units and Selected Model Parameters

Symbol	Description of Conceptual Model[a]	Units
d	depth of river with constant head	m
h_a	thickness of saturated/unsaturated aquifer at model boundary	m
h_r	thickness of saturated/unsaturated aquifer at x= 30	m
h_c	thickness of clogging layer	m
K	hydraulic conductivity	m d^{-1}
K_a	hydraulic conductivity of aquifer	m d^{-1}
K_c	hydraulic conductivity of clogging layer	m d^{-1}
Q	vertical flow rate through the clogging layer	m d^{-1}
L	distance to lateral model boundary from center of river	m
w	width of river from center of river to river edge	m
w_c	width of clogging layer at river edge	m

Model Parameters (Fixed For All Simulations)	Value	Units
Maximum time step	1	days
Minimum time step	0.1	days
Sand porosity	0.25	
Clay porosity	0.38	
Specific storage sand	0.001	
Specific storage clay	0.01	
Residual water content sand (θ)	0.04	
Residual water content clay (θ)	0.04	
Van Genuchten alpha (α) for sand and are defined by *van Genuchten* [1980]	4.0	m^{-1}
Van Genuchten beta (β) for sand and are defined by *van Genuchten* [1980]	1.4	
Reference potential evapotranspiration (E_p)-daily data taken from Rocky River weather station		m
Tree canopy evaporation (E_{can})	0	m
Evaporation extinction depth defined by a quadratic decay evaporation distribution function (*EDF*)	1.0	m
Evaporation limiting saturation sand (min)	0.05	
Evaporation limiting saturation clay (min)	0.25	
Evaporation limiting saturation sand (max)	0.9	
Evaporation limiting saturation clay (max)	0.9	
Transpiration extinction depth defined by a quadratic decay root distribution function (*RDF*)	5.0	m
Leaf area index (LAI)	1.5	$\mathrm{m}^2\ \mathrm{m}^{-2}$
Transpiration fitting parameter (c1)	0.6	
Transpiration fitting parameter (c2)	0.0	
Transpiration fitting parameter (c3)	1.0	
Transpiration limiting saturation (wilting point)	0.05	
Transpiration limiting saturation (field capacity)	0.1	
Transpiration limiting saturation (oxic limit)	0.8	
Transpiration limiting saturation (anoxic limit)	0.95	
Rill storage height	0.01	m

[a]The values are modified for different simulations.

[26] Precipitation was simulated for the entire model duration beginning on day 1. The daily precipitation values used in the model were based on the historical daily average precipitation for the Rocky River catchment. Precipitation was not simulated over the river channel because the river was set with a constant hydraulic head boundary condition and therefore the additional water on top of the

constant head boundary would cause erroneous results in the surface water domain. The stage height elevation of the river was set to 21 m, corresponding to a surface water depth of 0.5 m to provide an unlimited source of water to the aquifer system (i.e., perennial river), to ensure the clogging layer remained saturated and to isolate the effects of evapotranspiration on the state of connection. Surface runoff can only occur if the water table rises above the ground surface and exceeds the assigned rill storage height. The rill storage was sufficiently high (0.01 m) to prevent overland flow.

[27] The aquifer was composed of an isotropic homogeneous sand with a saturated hydraulic conductivity of 1.0 m d^{-1} and van Genuchten [*van Genuchten*, 1980] parameters ($\alpha = 4.0$, $\beta = 1.4$; [*Carsel and Parrish*, 1988]) (Table 1). The clogging layer beneath the river had a saturated hydraulic conductivity of 0.005 m d^{-1}. The extinction depths for evaporation and transpiration were 1 and 5 m, respectively, with both processes modeled over these depths using a quadratic decay function. Transpiration extinction depth (root depth) varies largely between different types of vegetation, ranging from over 60 m in the case of *Boscia albitrunca* and *Acacia erioloba* found in the central Kalahari, Botswana to less than half a meter for shallow rooted cereal crops [*Canadell et al.*, 1996; *Schenk and Jackson*, 2002; *Shah et al.*, 2007]. The extinction depth chosen for the base case scenario (5 m) was selected to represent an average extinction depth of native vegetation that is commonly found in Australia (e.g., eucalyptus and acacia species) [*Robinson et al.*, 2006; *Stone and Kalisz*, 1991]. The leaf area index (LAI) used to describe the transpiration function was set to 1.5 and is typical of native vegetation in southern Australia [*Ellis and Hatton*, 2008]. The limiting saturation constants for evaporation and transpiration and other model parameters are shown in Table 1.

3. Results and Discussion

3.1. Base Case

[28] Here we describe the base case model and the different scenarios using a subset of representative and realistic hydrogeological, hydroclimatic, and vegetation variables (Table 2).

[29] Figure 3 shows the hydraulic head versus time at an observation point at 12 m elevation located directly beneath the center of the river for the transient principal model ($K_a = 1$, $K_c = 0.005$ and slope = 0.01 [1 cm m^{-1}]). The dashed horizontal line shown in Figure 3 (and subsequent figures) describes the position of the water table (i.e., zero pressure head) directly beneath the center of the river at disconnection between the river and the aquifer, and is also the point of maximum flux calculated using equation (1). A total head at an observation point beneath the center of the river and above this horizontal line indicates a system in transition between a connected and disconnected regime. A total head at an observation point beneath the center of the river below this horizontal line is a system that is disconnected and can also be described as a system with a deep water table.

[30] As mentioned in section 2.3, the hydraulic head at this observation point is a very close approximation of the water table of the aquifer. The results showed that the model reached a quasi-steady state between precipitation and evapotranspiration over the first 10 year period and the river and aquifer were in a losing disconnected type regime. The results also showed that the presence of vegetation, through evapotranspiration, was able to cause and maintain an unsaturated zone beneath the river. It is worth noting that the aquifer was still responsive to the seasonal fluctuations in precipitation and evapotranspiration (Figure 3) with higher head levels in winter and lower head levels in summer. At the end of the 10 year period, transpiration was set to zero (i.e., the trees were removed) and the state of connection of the model changed from a losing disconnected type system to one that was connected and seasonally gaining and losing. Recall that evaporation was still simulated after transpiration was turned off. The hydraulic head fluctuated by approximately 1.5 m with a maximum at winter which is associated with gaining type conditions and a minimum at summer which is associated with losing type conditions. The change from disconnected to connected status was quite rapid with the water table rising several meters in a matter of days after the removal of vegetation.

3.2. Scenarios

[31] The model's sensitivity to the parameters of catchment slope, the hydraulic conductivity of the aquifer and the clogging layer, and the transpiration function was examined using several different model scenarios (Table 2) which are shown in Figures 4–6.

[32] To explore the impact of catchment slope, the slope gradient of the base case scenario was increased from 0.01 to 0.02 to 0.08 and up to 0.32 (Figure 4). Increasing the catchment slope increased the thickness of the vadose zone away from the river and limited the availability of soil moisture. Hence, the amount of water removed by evaporation and transpiration was limited to the functions' prescribed extinction depths of 1 and 5 m, respectively. In the model with a catchment slope of 0.08 (Figure 4), the river and aquifer were connected under a vegetated catchment

Table 2. Hydrogeological, Evaporation and Vegetation Variables Used in the Different Model Scenarios

Simulation	Hydraulic Conductivity Aquifer (K_a)(m d^{-1})	Hydraulic Conductivity Clogging Layer (K_c)(m d^{-1})	Catchment Slope (−)	Evaporation Extinction Depth (m)	Transpiration Extinction Depth (m)
Figure 3	1	0.005	0.01	1	5
Figure 4	1	0.005	0.02	1	5
Figure 4	1	0.005	0.08	1	5
Figure 4	1	0.005	0.32	1	5
Figure 5	4	0.005	0.01	1	5
Figure 5	10	0.005	0.01	1	5
Figure 5	1	0.0005	0.01	1	5
Figure 6	1	0.005	0.01	1	1, 2, 7

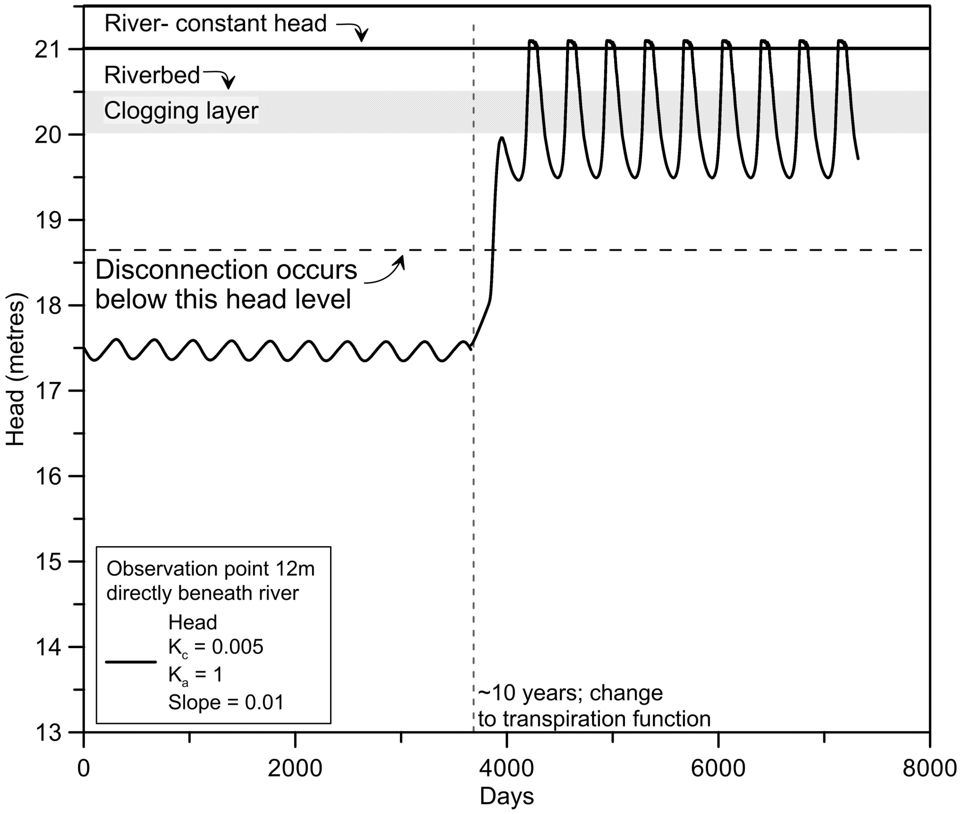

Figure 3. Hydraulic head at an observation point at 12 m elevation directly beneath the center of the river ($x = 30$ m) for the transient model. The catchment slope of the 2-D model is 0.01. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1}.

regime, and there were gaining and losing connected type conditions in response to the seasonal variation in precipitation and evapotranspiration. Once the transpiration was shut down after 10 years (by setting the root extinction depth and LAI to 0), the water table was much closer to the ground surface, but there were still seasonal gaining and losing connected type conditions. Figure 4, with a catchment slope of 0.32, shows that the river and the aquifer were connected. However, the river was losing under a vegetated catchment regime and transitioned to gaining and losing type conditions when transpiration was set to 0. It is also worth noting that the minimum and maximum values in the hydraulic head levels between the seasons were not as great as those observed in the scenario with a catchment slope of 0.08. The increased slope resulted in a wider vertical extent of the unsaturated zone and therefore changed the response in evaporation and transpiration which led to a dampened seasonal response in the hydraulic head (Figure 4).

[33] To examine the system's response to hydraulic conductivity, the conductivity of the aquifer and the clogging layer were varied by an order of magnitude and were shown to have a significant effect on the state of connection between the river and the aquifer (Figure 5). When the conductivity of the clogging layer was kept constant at 0.005 m d^{-1} and the conductivity of the aquifer was modified from 1 to 4 and up to 10 m d^{-1}, the extent that the evaporation and transpiration functions lowered the water table of the aquifer was increased. Under vegetated conditions (first 10 year period) the increase in the hydraulic conductivity of the aquifer still resulted in disconnected conditions. Once transpiration was turned off (set to 0) after the 10 year period the river and aquifer became connected again, however, in the scenario where the hydraulic conductivity of the aquifer was 10 m d^{-1}, there were periods during the summer when evaporation was high and the river and the aquifer became disconnected (Figure 5).

[34] When the hydraulic conductivity of the aquifer was kept constant at 1 m d^{-1} and the conductivity of the clogging layer was modified from 0.005 m d^{-1} (base case scenario) to 0.0005 m d^{-1} there was a significant effect on the state of connection between the river and the aquifer. In the scenario where the clogging layer conductivity was 0.0005 m d^{-1} the initial condition of the water table was considerably lower and once the transpiration was set to 0 after the 10 year period it took a considerable amount of time for the river and aquifer to become connected (Figure 5). The model simulation time was extended for this scenario to show that the model reached a quasi-steady state again. In comparison, when the hydraulic conductivity of the clogging layer was increased above the value used in the base case scenario (0.005 m d^{-1}) the river and aquifer remained connected.

[35] To explore the effect of transpiration on causing an unsaturated zone to develop underneath the river bed and

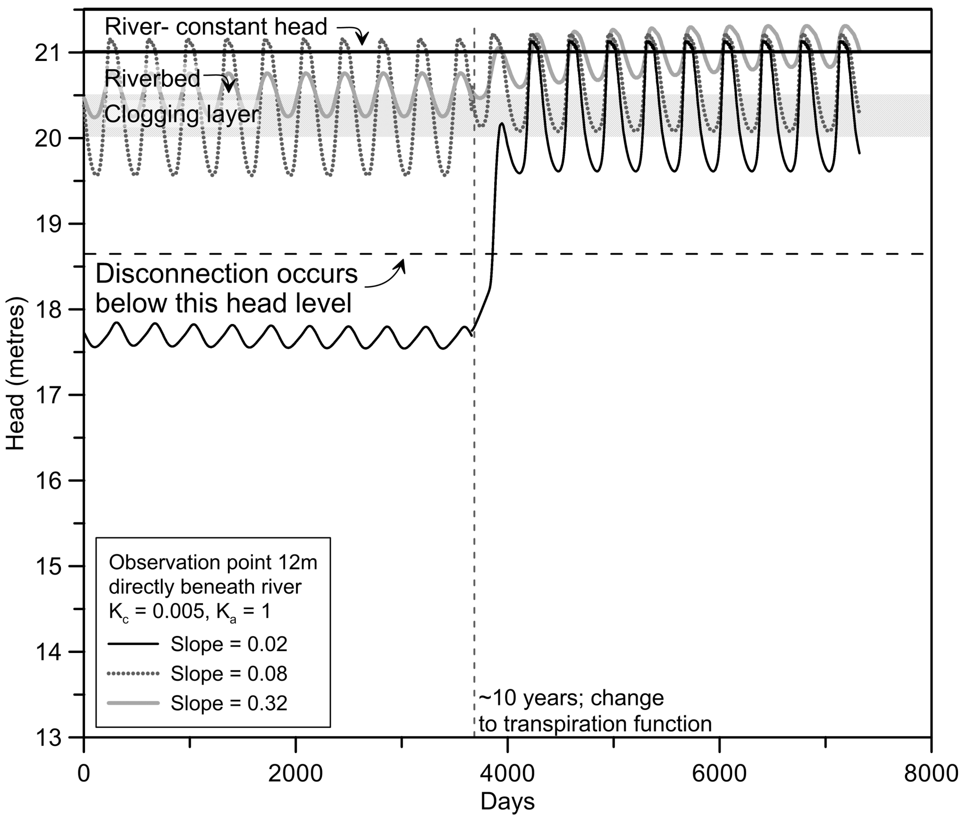

Figure 4. Hydraulic head at an observation point at 12 m elevation directly beneath the center of the river ($x = 30$ m) showing the sensitivity of the catchment slope on the transient 2-D model. Three model scenarios are shown with different values for the catchment slope. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1} for all three models.

alter the state of connection in the base case scenario, the transpiration extinction depth was modified from 0 (i.e., no transpiration) to 1 m, 2 m, and 7 m depth without changing the transpiration function values and LAI (Figure 6). The first 10 year period of the simulation was with evapotranspiration as simulated in the base case scenario (solid dashed line shows transpiration extinction depth 5 m). After the 10 year period different transpiration extinction depths were simulated for another 10 years (shown by four different line weights in Figure 6). The results showed that the greater the extinction depth (i.e., greater depth of the plant root zone) the more water was removed from the aquifer and the more likely that the river and the aquifer would transition to a disconnected type system. The change from a disconnected to connected system (defined by the hydraulic head level at the observation point crossing the determined line of disconnection) was rapid when there was no transpiration (less than 200 days), while changing to a shallower extinction depth there was a time lag (of at least 600 days) before the system reached a new quasi-steady state.

3.3. Effects of Evapotranspiration on the Presence of an Unsaturated Zone and the State of Connection

[36] The modeling presented in this study has shown that evapotranspiration can cause and maintain an unsaturated zone between a perennial river and aquifer system and in some cases a state of disconnection. Removing native deep rooted vegetation and replacing it with shallow rooted vegetation (i.e., modification of the transpiration function) can have a substantial effect on the state of connection and is more likely to change from a disconnected to a connected type system. While the present work attempted to evaluate the effects of evapotranspiration on the development of an unsaturated zone and the state of connection, we only addressed a few of the possible scenarios that may be observed in nature. For example, the temporal and spatial dynamics of ephemeral and intermittent rivers and their contrasting wetting and drying cycles compared to perennial rivers are also likely to have a significant impact on the infiltration flux from the river to the aquifer beneath and the state of connection [*Hatch et al.*, 2010; *Niswonger et al.*, 2008]. In our study, a constant head in the river was used in the conceptual model to maintain 100% saturation in the clogging/streambed layer to reduce system complexity so that we could accurately test our hypothesis.

[37] Ultimately, the processes of evaporation and transpiration were restricted to the prescribed extinction depths in the model domain which was investigated in the scenario models shown in Figure 6. Therefore, the steepness of the catchment slope had a considerable effect on the amount of water that could be removed by evaporation and transpiration processes due to the thickness of the vadose zone increasing with greater distance away from the river. In the base case scenario, the slope of the catchment was small

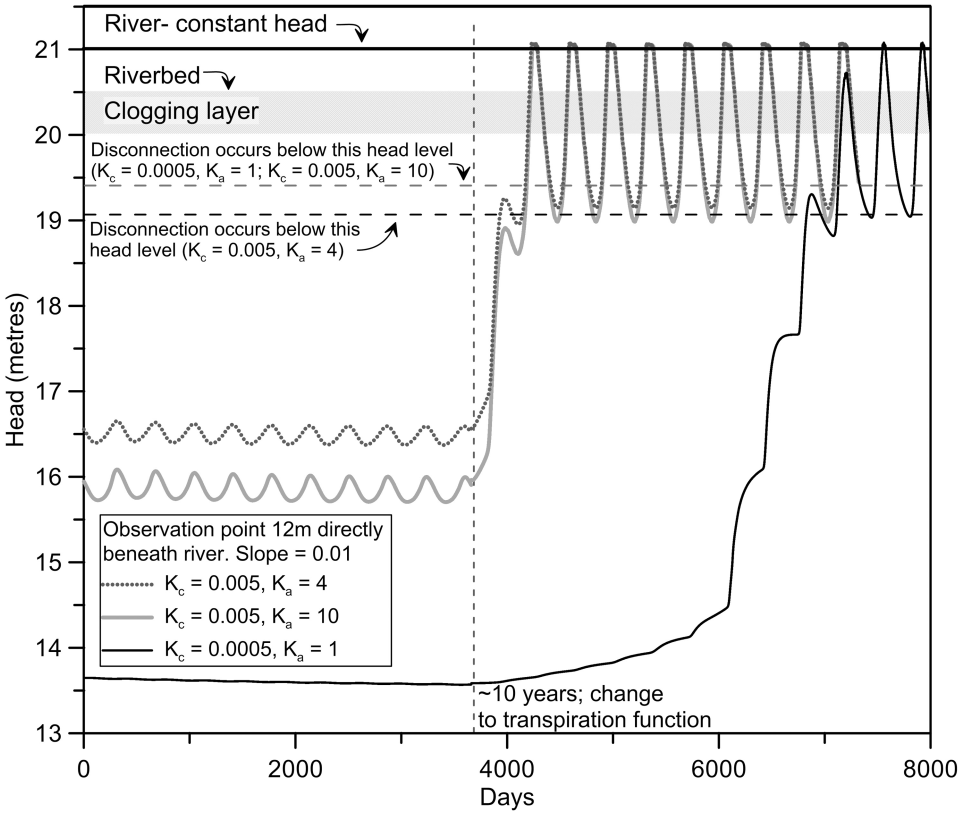

Figure 5. Hydraulic head at an observation point at 12 m elevation directly beneath the center of the river ($x = 30$ m) showing the sensitivity of the hydraulic conductivity (m d^{-1}) of the aquifer (K_a) and clogging layer (K_c) on the transient 2-D model. Three model scenarios are shown with different values for K_a and a K_c. All three models have a catchment slope of 0.01.

(1 cm m^{-1}), and therefore, there was minimal influence of the slope on the evaporation and transpiration functions. Increasing the slope increased the thickness of the vadose zone and decreased the depth of available soil moisture to transpiration and evaporation processes. It is worth noting that in most real systems the transpiration capacity of vegetation communities would also change along the slope in response to the available soil moisture and this has not been addressed in our study. However, the analysis of the catchment slope sensitivity does provide some insight as to where the greatest changes in the state of connection may occur in different types of catchment settings in response to a change in vegetation. For example, there would be a greater impact in catchments that are flat compared to ones that are steep.

[38] In the different model scenarios described here, the initial water table elevation of the aquifer was important in influencing the state of connection. In model scenarios where the initial water table elevation (located at the observation point directly beneath the center of the river) was well below the bottom of the clogging layer and the river, there was complete capacity of the evaporation and transpiration functions to remove water. In comparison, when the initial water table elevation was relatively shallow and close to the ground surface, the transpiration function was severely limited by complete saturation of the vegetation root zone.

[39] The effect of evapotranspiration on the development of an unsaturated zone beneath a riverbed and the state of connection depended largely on the hydraulic conductivity of the clogging layer beneath the river being less than the hydraulic conductivity of the aquifer. When the hydraulic conductivity of the clogging layer was large, the river continuously replenished the aquifer and no rate of evapotranspiration could induce an unsaturated zone below the clogging layer. In comparison, when a smaller hydraulic conductivity value was used, the processes of evapotranspiration were able to create an unsaturated zone below the clogging layer and in some cases resulted in disconnected type conditions. This illustrates the complex interplay between the various controlling variables and processes. The analysis of the described conceptual model was for a homogeneous system with homogeneous hydraulic conductivities. Simplifying the system complexity was necessary in order to remove any of the confounding effects (i.e., heterogeneity within the clogging layer and aquifer) to ensure clarity of the specific hypothesis that was being examined. It is worth noting that heterogeneity within the clogging layer and aquifer can be an important control on river seepage temporally (in response to streambed scouring) as well as its spatial distribution along the channel [*Fleckenstein et al.*, 2006; *Hatch et al.*, 2010; *Niswonger and Fogg*, 2008]. *Frei et al.* [2009] also noted that spatial and temporal heterogeneity within alluvial sediments can cause distinct patterns and dynamics of

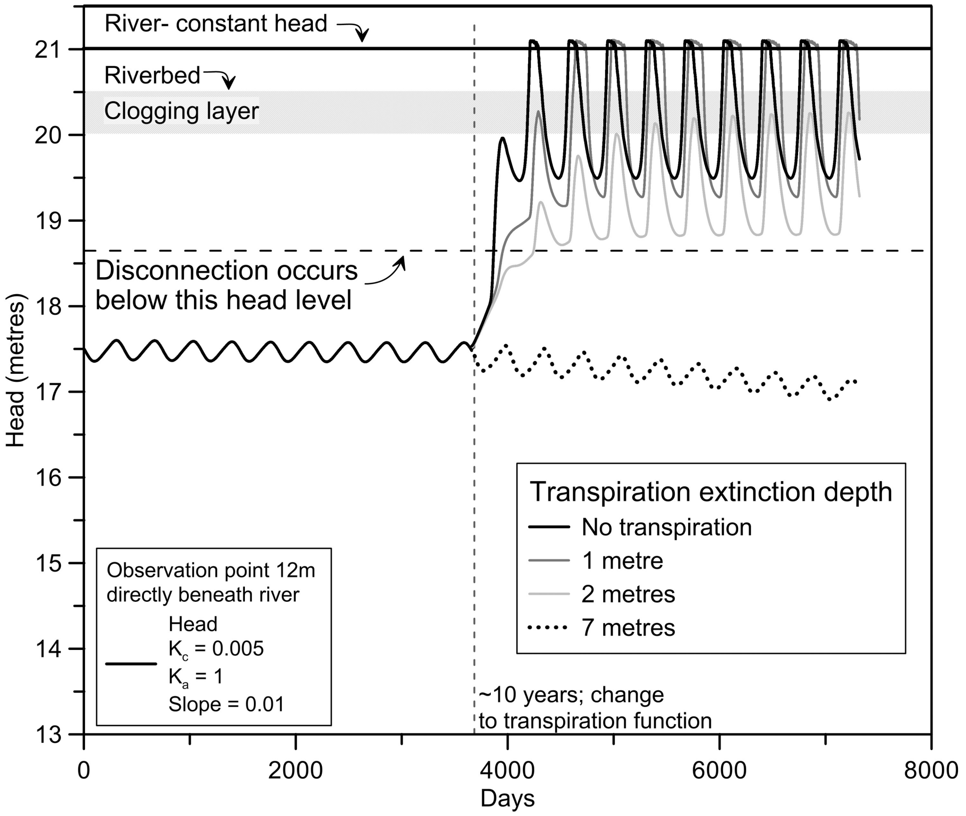

Figure 6. Hydraulic head at an observation point at 12 m elevation directly beneath the center of the river ($x = 30$ m) showing the sensitivity of the transpiration extinction depth function on the state of connection between surface water and groundwater. The hydraulic conductivity of the aquifer (K_a) is 1 m d^{-1} and the hydraulic conductivity of the clogging layer (K_c) is 0.005 m d^{-1}, and the catchment slope is 0.01 for all models.

river seepage in rivers overlying a deep water table (i.e., disconnected systems), and that most seepage occurs along preferential flow zones.

[40] All of the model scenarios described in this study have used the historical average precipitation and reference potential evapotranspiration data for the Rocky River catchment. According to Figure 1 there is a period from mid-April until the end of August where precipitation is greater than evapotranspiration which would be a time when groundwater recharge could occur. The seasonal trends in precipitation and evapotranspiration in Figure 1 are representative of hydroclimatic conditions in many parts of southern Australia. The simulations showed that there was a strong seasonal response in the aquifer to changes in precipitation and evapotranspiration, and it was only when the water table was at a considerable depth (~6.5 m) below the riverbed that the seasonal response was not observed (Figure 5; $K_c =$ 0.0005). The seasonal variations between summer (hydraulic minimum) and winter (hydraulic maximum) were more pronounced after the removal of vegetation when the water table was closer to the ground surface as a result of increased recharge to the aquifer.

3.4. Trees as Groundwater Pumps

[41] The analogy of trees as groundwater pumps and their potential to influence the exchange fluxes between surface water and groundwater has been well established [*Butler et al.*, 2007; *Loheide et al.*, 2005]. However, so far it has not been demonstrated in a quantitative and systematic way if evapotranspiration can cause an unsaturated zone to develop underneath a riverbed and in some instances cause a disconnection. The results of this study support the earlier hypothesis proposed by *Banks et al.* [2011], which suggested that the river system in the Rocky River catchment may be fresher (less saline) than rivers in adjacent catchments in apparently similar geologic and climatic settings. The low salinity in the river is maintained by virtue of the fact that the Rocky River catchment is pristine and covered by native vegetation which creates losing conditions, while the others are cleared and are likely to be gaining systems. The vegetation controls are a plausible explanation for different state of connection and poorer water quality of the adjacent cleared catchments which were studied previously by *Henschke et al.* [2003] and *Shand et al.* [2007].

[42] According to the water balance of the Rocky River catchment, the annual precipitation input volume is 147.4 × 10^6 m^3 yr^{-1} (based on precipitation of 780 mm yr^{-1} and gauged catchment area of 189 km^2) and the annual streamflow discharge is 1.4 × 10^6 m^3 yr^{-1}. Therefore, 146 × 10^6 m^3 yr^{-1} of the catchments' precipitation is lost to evapotranspiration and/or groundwater recharge (assuming that streamflow represents surface runoff only). It can be assumed that there is zero to very little groundwater recharge beneath native vegetation

[*Allison and Hughes*, 1983; *Leaney and Allison*, 1986], and therefore 0.77×10^6 m^3 per square kilometer would be removed via transpiration from the catchment water balance, which would equate to approximately 53,986 trees per km^2 (using the average water use of a eucalyptus tree equal to 14.7 m^3 per tree per year [*Farrington et al.*, 1994]). The high density of the vegetation (and therefore high evapotranspiration) has the potential to maintain the lower elevation of the water table beneath the river and hence a disconnected type system. In comparison, in the adjacent cleared catchments to the pristine Rocky River catchment, a decrease in evapotranspiration through the removal of native vegetation and replacement with shallow rooted vegetation has evidently resulted in a change to a connected type system and salinisation of the water resource as a result of increased recharge and a rising water table. Our results also showed that with the removal of vegetation and a rising water table the seasonal variations between summer (hydraulic minimum) and winter (hydraulic maximum) were more pronounced.

4. Conclusion

[43] By using a simple conceptual model based on realistic and representative parameter values, we have demonstrated that the presence of vegetation is a plausible mechanism for causing an unsaturated zone to develop between a perennial river and an aquifer. Vegetation can therefore also affect the state of connection between surface water and groundwater and in some instances create a disconnection. This may appear intuitively plausible in a qualitative sense; however, it has not been demonstrated quantitatively. Our study therefore suggests that in addition to the well known influences of physical variables such as hydraulic conductivity or topography, the effects of vegetation need to be carefully considered when investigating surface water–groundwater interactions. By examining different conceptual models of catchments with different slopes and vegetation type (i.e., root depth) we provided insights into the conditions where changes to vegetation can affect the flow regime and the presence of an unsaturated zone. Our analysis showed that the flow regime and hydraulic response to the presence of vegetation and subsequent removal can be much greater in flatter catchments than those that are steep.

[44] Given the importance of vegetation on surface water–groundwater interactions, changes in vegetation can have considerable consequences to shifting the state of connection. Such changes can be associated with land clearance, revegetation or climate change. In catchments in southern Australia where the aquifer systems are often saline or there is a significant amount of salt stored in the unsaturated zone, changing from a losing disconnected to gaining type system results in serious water quality issues because saline groundwater discharges to the surface water system. In the longer-term, the change in vegetation in pristine catchments can lead to the salinization of the surface water resource. This link between pristine and cleared catchments, and the resulting state of connection or disconnection, may be important in explaining differences in observed river water quality between catchments in similar geographic, geologic, and climatic locations. The results of this current study also appear to support the hypothesis raised in the earlier *Banks et al.* [2011] study that land clearance may be the key factor to maintain low levels of salinity in the Rocky River, while the other rivers in adjacent catchments are more saline.

[45] The purpose of this study was to demonstrate that vegetation can (under reasonable and representative conditions) create an unsaturated zone and therefore affect the state of connection. Even though we did not develop a generalized theoretical framework on the effects of evapotranspiration on the state of connection between surface water and groundwater, the findings of this study are likely to be applicable to other catchments where land clearance or revegetation occurred or is occurring.

[46] Further work carried out in 3-D would be useful to explore the effects of evapotranspiration on the state of connection along the river (compared to the cross section of the river as discussed in this study) and how the interface between connected and disconnected regimes migrates up and down the river channel in both space and time as a function of vegetation clearance or land use change. 3-D analysis could also provide insights into the influence of more natural meandering river geometries and how this may affect the convergence and divergence of groundwater flow paths near the river. Additional simulations might explore the effect of different vegetation types throughout catchments (e.g., vegetation with increasing extinction depth up slope), but the basic findings of this study are not expected to be significantly altered in those more complex cases.

[47] **Acknowledgments.** We gratefully acknowledge the Department for Water, Adelaide, South Australia (formerly the Department for Water, Land, and Biodiversity Conservation) and staff from the Science, Monitoring and Information division for field support. We thank Andy Love for his contribution to discussions of the modeling and its relevance to the Rocky River catchment. We are also very grateful to the editor John Selker, the associate editor and three anonymous reviewers for their insightful and constructive feedback on this manuscript. The project was financially supported by the National Water Commission through its Raising National Standards Program and by the Swiss National Foundation, Ambizione grant PZ00P2_126415.

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