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Hydrogeologic controls on disconnection between surface water and groundwater

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[1] Understanding how changes in the groundwater table affect surface water resources is of fundamental importance in quantitative hydrology. If the groundwater table below a stream is sufficiently deep, changes in the groundwater table position effectively do not alter the infiltration rate. This is referred to as a disconnected system. Previous authors noted that a low-conductivity layer below the surface water body is a necessary but not sufficient criterion for disconnection to occur. We develop a precise criterion that allows an assessment of whether surface water–groundwater systems can disconnect or not. We further demonstrate that a disconnected system can be conceptualized by a saturated groundwater mound and the development of a capillary zone above this mound. This conceptualization is used to determine the critical water table position at the point where full disconnection is reached. A comparison of this calculated critical water table position with a measurement of the water table depth in a borehole allows the assessment of the disconnection status. A sensitivity analysis of this critical water table showed that for a given aquifer thickness and river width, the depth to groundwater where the system disconnects is approximately proportional to the stream depth and the hydraulic conductivity of the streambed sediments and inversely proportional to the thickness of these sediments and the hydraulic conductivity of the aquifer. The conceptualization also allows the disconnection problem to be analyzed using both variably saturated and fully saturated groundwater models and provides guidance for numerical and analytical approaches.

1. Introduction

[2] There are numerous examples where lowering the groundwater table has caused reductions in streamflow and drying of wetlands [e.g., *Sophocleous*, 2000]. Clearly, changes in the groundwater table can affect the infiltration from surface water bodies. However, different states of connection between surface water and groundwater exist. To what extent a change of the groundwater table alters the infiltration rate of the overlying surface water body strongly depends on the state of connection between the two compartments. Determining whether streams are hydraulically connected or disconnected to the groundwater is thus crucial for water resources management. Although surface water–groundwater interaction has been studied extensively, the subject of disconnection remains poorly understood.

[3] In principle, two fundamentally different flow regimes between surface water and groundwater are possible: the surface water body drains the aquifer (gaining stream) or recharges the aquifer (losing stream). The head difference between the two compartments determines whether the stream is gaining water from or losing water to the aquifer. Consider a perennial stream which is gaining water from the aquifer. If we lower the water table, the rate of groundwater discharge to the stream will decrease, until a point is reached where the heads are equal. If the groundwater table is lowered further, the stream will begin to lose water to the aquifer. As the groundwater table drops, the infiltration flux increases. If the groundwater table is lowered sufficiently, an unsaturated zone will sometimes develop underneath the streambed. In the case that an unsaturated zone develops and the groundwater table is lowered further, the infiltration rate asymptotically approaches a constant value. (The implications of this asymptotic behavior are discussed later in the paper.) When further reductions in the groundwater table no longer significantly affect the infiltration rate, the stream is said to be disconnected (Figure 1). The state between initial desaturation and the point where the infiltration rate no longer changes in response to a decline of the water table is called the transition state. Disconnected systems can be induced by groundwater pumping close to the stream, but they can also occur naturally. In either case, further lowering the groundwater table in an already disconnected system will not significantly increase the infiltration rate where the

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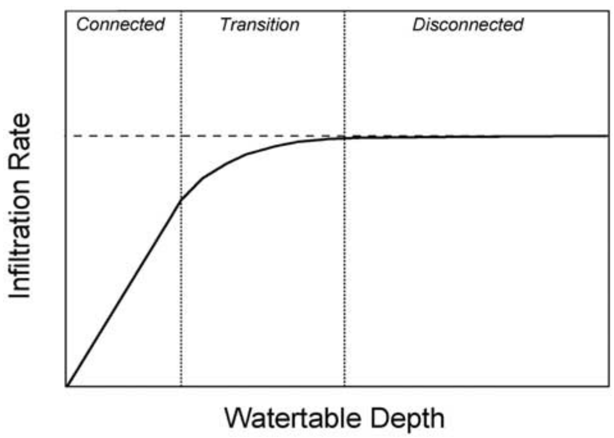

Figure 1. Changes in infiltration rate from surface water to groundwater as the water table drops. Three different flow regimes can be identified. For small head differences, the infiltration rate between the surface water body and the groundwater is proportional to the head difference. In the transition zone, the flow rate is no longer a linear function of the head difference. If the water table is lowered further, the infiltration rate approximates a constant value, and changes in the water table no longer significantly affect the infiltration rate. In this final regime the surface water and the groundwater are disconnected.

stream is disconnected, but is expected to increase the length of stream over which disconnection occurs.

[4] Previous studies suggested disconnection can only occur if the permeability of the stream sediments is small compared to the permeability of the aquifer [e.g., *Fox and Durnford*, 2003; *Bruen and Osman*, 2004]. However, as we will show, this is a necessary but not sufficient criterion for disconnection to occur. We explicitly formulate a hydrogeologic criterion for when disconnection can occur. Furthermore, there are a number of points that require clarification. The use of the term "disconnection" may incorrectly suggest no flow between surface water and groundwater, but in fact the infiltration flux is greatest when disconnection occurs. It has also been suggested that the disconnection status can be assessed simply by determining the ratio of the river width to the groundwater depth [*Sophocleous*, 2002], or by comparing the depth to groundwater and the stream width and stream depth [*Environment Canterbury*, 2001] (accessed June 2008). However, these criteria neglect many of the important hydrogeological variables and do not clearly define where the depth to groundwater is measured. Different classifications will therefore arise if the location where the groundwater depth is measured is changed and so these criteria are not useful.

[5] The principal aim of this paper is to explore the physics of disconnection phenomena and to develop quantitative criteria to predict their occurrence. We establish a sound theoretical framework for assessing the state of disconnection of losing streams and determine which hydrogeological parameters influence the connection status and how variations in these parameters influence the status of connectivity. Such a framework could provide a basis for field assessments of stream-groundwater connectivity and may be useful in spatially and temporally extrapolating local results.

2. Previous Studies

[6] A large number of papers have examined stream depletion induced by groundwater pumping [e.g., *Kollet and Zlotnik*, 2003; *Spalding and Khaleel*, 1991]. However, most of these papers only consider fully saturated flow regimes and represent the semipervious layer beneath the streambed as an increased resistance to flow. Since these approaches implicitly assume fully saturated conditions always exist, they do not allow for disconnection to occur and will therefore not accurately predict stream depletion in cases where disconnection does occur.

[7] Only a few studies have explicitly taken unsaturated flow into account. These studies have pointed out that disconnection can only occur when the hydraulic conductivity of the streambed sediments is smaller than that of the aquifer. From herein, we refer to the low-conductivity layer as the clogging layer. However, this is not a sufficient condition for desaturation and it is still unclear under what conditions a surface water body will become disconnected from groundwater. *Osman and Bruen* [2002] stress the importance of taking an unsaturated zone into account and present a modified version of MODFLOW (called MOBFLOW) which accounts for unsaturated conditions beneath a surface water body. On the basis of this model, *Bruen and Osman* [2004] evaluated the sensitivity of stream-aquifer seepage to the saturated hydraulic conductivity of the aquifer. *Desilets et al.* [2008] studied effects of stream-aquifer disconnection on local flow patterns using a numerical model. *Su et al.* [2007] analyzed unsaturated flow regions induced by pumping close to a perennial stream using a code capable of simulating unsaturated flow. *Fox and Durnford* [2003] discuss the development of unsaturated flow under a surface water body and suggest a semianalytical solution to estimate aquifer drawdown and stream depletion. The analytical component of their approach describing the saturated flow is based on a solution suggested by *Hunt* [1999], which relies on the Dupuit assumption and assumes a constant transmissivity. The papers cited above recognize the importance of disconnected losing streams. However, they focus on case studies or the development of numerical codes rather than on the development of a general theoretical framework describing disconnection.

[8] A theoretical study carried out by *Bouwer* [1969] on the basis of fully saturated conditions has erroneously been cited in the context of disconnected losing streams. *Bouwer* [1969] showed that for fully saturated losing systems, the rate of loss between the surface water body and the aquifer is related to the head difference and that this relationship need not be linear. He concluded that the rate of increase in flow with increasing head gradient decreases as the groundwater table falls. The curve relating infiltration rate to head difference presented by Bouwer is superficially similar to Figure 1. Bouwer concluded that the infiltration flux stabilizes if the depth to groundwater is greater than twice the width of the stream. The qualitative similarity of Bouwer's work to Figure 1 may have led to Bouwer's work being misinterpreted as an illustration of disconnection [*Sophocleous*, 2002]. The concept of using groundwater

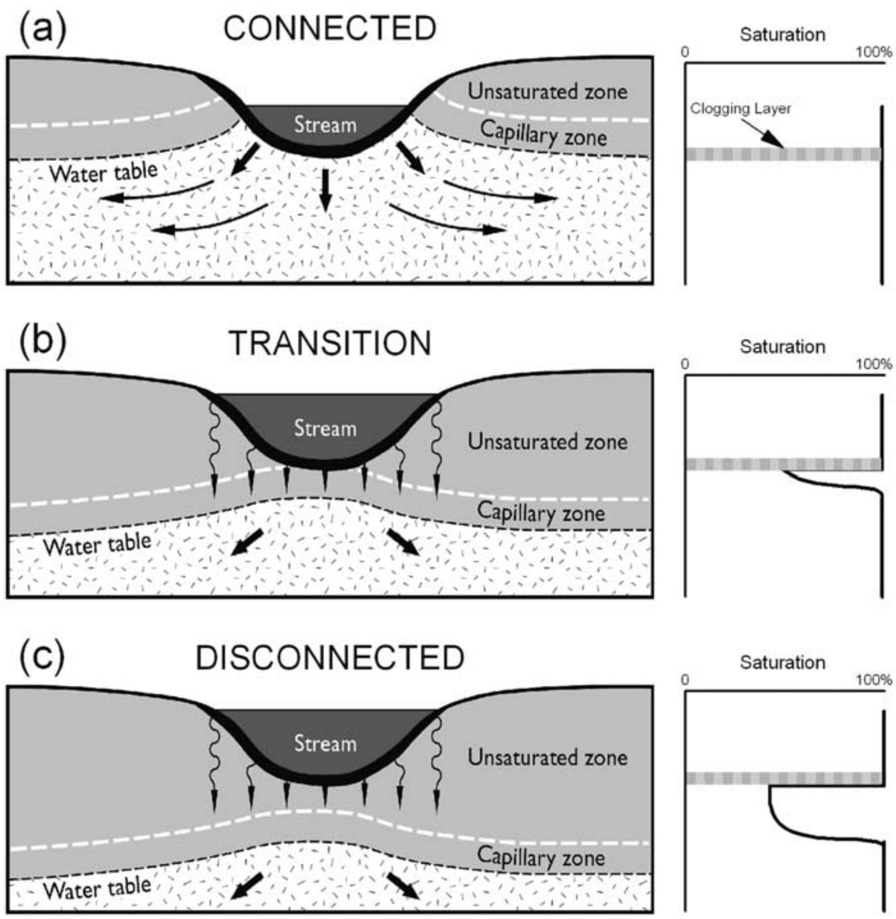

Figure 2. Different flow regimes between surface water and groundwater. (a) Fully connected flow, (b) transition flow, and (c) disconnection are shown. A profile of saturation beneath the base of the stream is indicated in each case. For the disconnected case, the saturation is constant between the base of the clogging layer and the top of the capillary zone.

head data near a stream to assess the state of disconnection is useful. Nevertheless, it is important to note that Bouwer's analysis was for a completely saturated regime, where the relationship between head and infiltration flux is fundamentally different to an unsaturated regime. Furthermore, the location at which the depth to groundwater is measured and Bouwer's criterion should be applied is never stated. Consequently, the theory presented in his approach cannot be used for an assessment of disconnection.

[9] There are two principal ways in which we can assess the status of connection or disconnection for a stream.

[10] 1. The direct observation of disconnection: The only direct method involves observing an unsaturated zone beneath the middle of the stream. We are not aware of a single case in the literature where this has been reported for a perennial stream. Installing tensiometers, piezometers and other similar devices beneath perennial streams is not straightforward. Furthermore, this only establishes that the stream is not connected, and does not distinguish between transition and disconnected states. Similarly, observing that the groundwater level beneath the stream is below the streambed sediments only determines that the system is not fully connected.

[11] 2. Besides a direct observation, disconnection can be assessed indirectly: Indirect methods to determine disconnection include (1) establishing that the rate of infiltration does not significantly change when the position of the water table changes [*Moore and Jenkins*, 1966] and (2) measurement of groundwater depth in bores adjacent to the stream, which must be interpreted with a groundwater model [e.g., *Su et al.*, 2007]. Both of these indirect approaches would benefit from a more solid theoretical framework than currently exists.

3. Problem Conceptualization

[12] Figure 2 depicts the position of the water table and capillary zone beneath connected, disconnected and transitional streams, together with the saturation profile beneath the base of the stream. Our definition of the term capillary zone will be discussed later. The presence of a permeable clogging layer is critical in our analysis. Without a clogging layer, further lowering of the water table will produce an increasing stream loss as water table depth increases. In the first type of interaction shown in Figure 2a, the flow between the surface water body and the aquifer is connected and therefore fully saturated. In the case of a full disconnection (Figure 2c), an unsaturated zone has developed below the clogging layer. Saturation is effectively constant between the streambed sediments (clogging layer) and the

capillary zone above the groundwater. Figure 2b represents the transition between Figures 2a and 2c. This occurs when the capillary zone intersects the clogging layer but the water table is below it. In this case, saturation remains a function of depth between the groundwater surface and the clogging layer.

[13] We will demonstrate in section 5 that the disconnection problem can be conceptualized in terms of the buildup of a saturated groundwater mound under a surface water body and the development of a capillary zone above the water table. The development of the capillary zone above the water table is essentially a 1-D problem. However, the buildup of the groundwater mound is either 2-D or 3-D. Analyzing the buildup of a groundwater mound allows us to assess the state of disconnection in terms of the position of the water table at a bore adjacent to the stream. Consider the case of an observation bore adjacent to a stream with a head of 2 m below that of the stream. Clearly, the stream is losing, but is this groundwater–surface water system connected, disconnected or transitional?

[14] In this paper we only consider the case of the buildup of a mound underneath a straight infiltrating stream where all groundwater flow lines are normal to the stream. This reduces our flow analysis to the simpler 2-D case. We assume the buildup of the groundwater mound is superimposed on an otherwise flat groundwater table: flow geometry perturbations, for example due to regional flow or pumping from bores adjacent to the stream, are thus not considered in this study. We also assume homogeneous and isotropic hydraulic conductivities and restrict our analysis to steady state conditions. Although the analysis is simplified, these simplifications are required to develop a fundamental understanding of the key variables and processes inherent to the physics of disconnection phenomena.

[15] A connected (fully saturated) flow regime will only occur if the pressure beneath the clogging layer in the middle of the stream is greater than zero. If a negative pressure occurs, it will occur at the interface below the base of the clogging layer and the aquifer. To identify the conditions under which negative pressure at the base of the clogging layer can develop, it is necessary to express the pressure head at this point as a function of the hydraulic head at the water surface and at the groundwater table. In order to identify these conditions and develop a general criterion for when disconnection can occur, a 1-D analysis is required. The 1-D analysis is also useful for describing the capillary zone and hence the sharpness of the transition between fully connected and fully disconnected systems.

4. One-Dimensional Analysis

[16] Using a 1-D analysis we can (1) formulate a criterion to identify systems for which disconnection can occur, (2) calculate the infiltration flux that will occur for a disconnected system, and (3) calculate the height of the capillary zone, which determines the magnitude of the transition zone between connected and disconnected status. Table 1 provides an overview of the variables and their units used for these calculations.

4.1. Condition for Disconnection

[17] We consider steady state flow through a two-layer soil column, above which surface water with a constant water level of depth d is ponded (Figure 3). The pressure is set to zero at zero elevation. This choice of boundary condition corresponds to a soil column whose lower end is open to the atmosphere. In a 2-D analog, this lower boundary condition would correspond to the water table below a disconnected infiltration zone. This is because the water table is defined as a surface with a pressure that equals zero.

Table 1. Notation and Units[a]

Symbol	Description	Units
	1-D and 2-D Models	
d	depth of ponded water	m
h_a	thickness of saturated/unsaturated aquifer	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}
k_r	relative hydraulic conductivity at a given pressure	–
q	vertical flow rate through the clogging layer	m d^{-1}
q_{max}	limiting value of q as h_a becomes large; equivalent to the infiltration flux of disconnected systems	m d^{-1}
γ_p	pressure head at base of clogging layer	m
γ^*_p	limiting value of γ_p as h_a becomes large	m
	Additional Parameters Required for 2-D Model	
B	thickness of aquifer	m
ΔH_d	value of ΔH where full disconnection occurs	m
h_0	constant head boundary at $x = L$	m
L	distance of lateral model boundary from center of water body	m
w	width of stream	m
ΔH	distance between base of clogging layer and water table at boundary of the model ($x = L$); ($\Delta H = B - h_0$)	

[a]Van Genuchten parameters α and β are also used and are defined by *van Genuchten* [1980].

[18] For fully saturated conditions, the average hydraulic conductivity (K) of the two-layer system is given by [*Bear*, 1979]

$$K = \left(\frac{1}{h_c + h_a}\left(\frac{h_c}{K_c} + \frac{h_a}{K_a}\right)\right)^{-1}, \tag{1}$$

where h_c and K_c are the thickness and hydraulic conductivity of the clogging layer, and h_a and K_a are the thickness and hydraulic conductivity of the aquifer strata. Under conditions of steady state flow, the saturated infiltration flux q over the entire domain is given by the average hydraulic conductivity multiplied by the hydraulic gradient, which reduces to

$$q = (h_a + h_c + d)\left(\frac{h_c}{K_c} + \frac{h_a}{K_a}\right)^{-1}. \tag{2}$$

The continuity equation requires that q throughout the entire domain must equal the flux through the first layer:

$$q = (h_a + h_c + d)\left(\frac{h_c}{K_c} + \frac{h_a}{K_a}\right)^{-1} = K_c \frac{h_c - \gamma_p + d}{h_c}. \tag{3}$$

Solving for γ_p yields

$$\gamma_p = \frac{h_a(dK_c + h_cK_c - h_cK_a)}{(h_cK_a + h_aK_c)}. \tag{4}$$

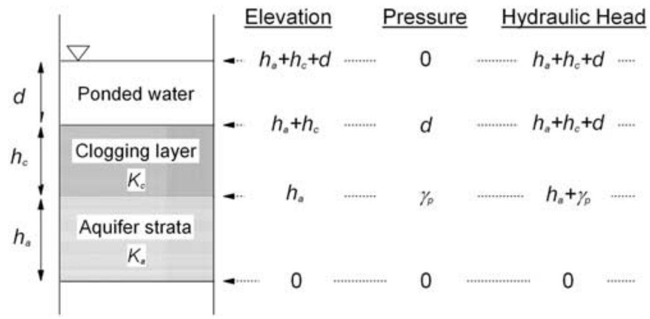

Figure 3. Overview of the 1-D setup: d is the ponded water depth, h_c is the thickness of the clogging layer and K_c is its saturated hydraulic conductivity, and h_a and K_a are the thickness and saturated hydraulic conductivity of the underlying aquifer. Relative elevations, pressure potentials, and hydraulic heads at the top and bottom of each layer are expressed in terms of layer thicknesses and the pressure potential at the base of the clogging layer, γ_p.

The pressure at the interface between the two layers remains negative provided that the numerator in equation (4) is less than zero and hence that

$$\frac{K_c}{K_a} \leq \frac{h_c}{d + h_c}. \tag{5}$$

This condition can also be derived from equations presented by [*Zaslavsky*, 1963], who studied flow through layered soils, although the link to disconnection was not made by this author.

[19] If the condition in equation (5) is not fulfilled, the soil column beneath the clogging layer will never desaturate, and the pressure below the clogging layer will remain positive independent of the head gradient through the system and irrespective of the value of h_a (provided $h_a > 0$). Whether the system can potentially become disconnected or not is therefore only a function of the depth of ponded water, the thickness of the clogging layer and the ratio of the hydraulic conductivities of the clogging layer to the aquifer.

4.2. Infiltration Flux for Disconnected Systems

[20] Provided that unsaturated flow can occur, it will occur when h_a becomes sufficiently large for the negative pressure at the interface to be less than the air entry value of the aquifer strata. At this point, flow in the aquifer strata is unsaturated. As the water table drops further, the pressure at the interface becomes more negative, increasing the gradient (and therefore also the flux) through the clogging layer which remains saturated. At the same time, the decreasing negative pressure reduces the hydraulic conductivity of the aquifer strata. However, although the hydraulic conductivity decreases, the hydraulic gradient through the aquifer increases. This allows the flux to increase while the hydraulic conductivity decreases. A continuous lowering of the water table therefore causes the infiltration rate to increase and approximate a constant, maximum value, as illustrated in Figure 1. It is important to note, however, that in a strict mathematical sense, a constant infiltration rate is never reached and the flux is always head-dependent. This is because as the groundwater table is lowered, the pressure at the base of the clogging layer asymptotically approaches a minimum value (which is associated with this constant, maximum flux). A cutoff value must therefore be defined at which the pressure and hence infiltration rate is considered constant. The maximum infiltration rate will be equal to gravitational flow through the aquifer strata [*Osman and Bruen*, 2002]. On the basis of the continuity equation, we can write

$$K_c \frac{\left(h_c + d - \gamma_p^*\right)}{h_c} = K_a^* k_r\left(\gamma_p^*\right), \tag{6}$$

where γ^*_p is the pressure head that occurs at the interface between the clogging layer and the aquifer strata under gravity drainage, and which is approached as the water table drops. $k_r(\gamma^*_p)$ is the relative hydraulic conductivity at this pressure ($0 < k_r(\gamma^*_p) < 1$). The left hand side of equation (6) is the flow through the clogging layer, and the right hand side is the flow rate through the unsaturated aquifer strata. Equation (6) allows us to calculate the pressure head that will occur in full disconnection, and thus also the maximum flux. To solve the equation, however, a relationship between pressure and hydraulic conductivity must be assumed. Several parameterizations for the relation between pressure-saturation-relative hydraulic conductivity k_r exist, although in this paper we use the formulation suggested by *van Genuchten* [1980]. Equation (6) is then solved numerically.

4.3. Height of the Capillary Zone in Case of Disconnection

[21] To examine the relationship between infiltration rate and h_a, numerical simulations were carried out using the software package HydroGeoSphere (HGS) [*Therrien et al.*, 2006] which simulates both saturated and unsaturated flow. The simulations use typical van Genuchten parameters for a sand (α = 14.5 m^{-1} and β = 2.68, K_a = 100 m d^{-1}) and a loam (α = 3.6 m^{-1} and β = 1.56, K_a = 12.5 m d^{-1}) [*Carsel and Parrish*, 1988]. Figure 4 plots simulation results for the flux as a function of h_a. (The parameters have been chosen so that the condition in equation (5) is fulfilled, and so $\gamma_p < 0$ for $h_a > 0$.)

[22] In Figure 4 the maximum values of the exchange flux for the two soil types, calculated using equation (6) are also shown. It is clear that simulated fluxes approach these values for large values of h_a. The system is disconnected when the infiltration rate approximates the maximum value. For the case of sand, this is around h_a = 0.2 m and for the loam around h_a = 0.8 m. Since the van Genuchten equations do not have an explicit air entry value, the system is in transition for $0 < h_a < 0.2$ m and $0 < h_a < 0.8$ m for the sand and loam cases, respectively. The extent of the transition zone is equal to the height above the water table where pressure and saturation no longer significantly change with height. As apparent from Figure 4, this is greater for loam soil than for sand.

[23] Figure 5 shows the saturation profiles for a loam soil under vertical water fluxes of 0, 0.1 and 0.5 m d^{-1}. This is calculated by discretizing the soil profile above the water table and by solving the Richards equation for the specified flow rate. Where there is no water flux, the water content continues to be reduced as the height above the water table increases. However, under a constant flux, the soil water

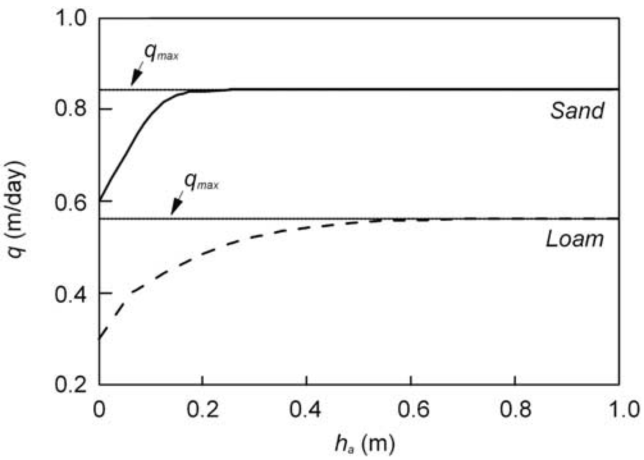

Figure 4. Infiltration rate q as a function of h_a based on numerical simulations. Simulations for the sand use a saturated hydraulic conductivity of K_a = 100 m d^{-1}, van Genuchten parameters α = 14.5 m^{-1} and β = 2.68, and a depth of ponded water of d = 0.2 m. Simulations for the loam soil use K_a = 12.5 m d^{-1}, α = 3.6 m^{-1}, β = 1.56, and d = 0.05 m. For the purpose of this illustration the choice of K_a is not identical to the K_a used in subsequent calculations and Figures 7–10 and A1. (Both simulations use K_c = 0.2 m d^{-1} and h_c = 0.1 m for the clogging layer.) Values of q_{max} calculated using equation (6) are also shown. Numerical simulations approach these values for large values of h_a.

content approximates a constant value (S_{min}) which is dependent upon the magnitude of the flux. As the water flux increases, the minimum saturation value that will occur also increases. Besides saturation, the hydraulic conductivity and the pressure above the water table also approximate a constant value. We define the capillary zone as the zone immediately above the water table, in which saturation, hydraulic conductivity and the pressure decrease with increasing height above the water table. To our knowledge, there is no existing term for this zone. The capillary zone should not be confused with the capillary fringe, which is the zone above the water table in which the pressure is greater than the air entry value [see, e.g., *Hillel*, 1998]. Theoretically, the capillary zone is of infinite extent because the pressure and saturation above the water table approach their minimum values asymptotically. For both practical and comparative purposes, we define the top of the capillary zone as the height above the water table where the pressure is within 0.1% of the minimum possible value (for any given infiltration rate). Likewise, we could have defined a cutoff criterion in terms of saturation, hydraulic conductivity or infiltration rate. We have defined the cutoff value to be very close to the minimum value to ensure that changes in the infiltration rate in response to lowering the groundwater table are very small and therefore effectively no longer influence the infiltration rate. For the example shown in Figure 5, the height of the capillary zone is 1.45 m for a flux of 0.1 m d^{-1}, and 0.88 m for a flux of 0.5 m d^{-1}. The height of the capillary zone defines the sharpness of the transition zone between fully connected and fully disconnected systems.

[24] In the 1-D system considered so far, the surface water body is of infinite lateral extent, and consequently the water table underneath it is flat. Clearly this is not a realistic scenario. For rivers and streams, a 2-D flow system is more appropriate. However, because the flow regime between the clogging layer of a disconnected losing stream and the water table is essentially a 1-D problem, the criterion to identify if two systems can disconnect (equation (5)) is therefore also valid for a 2-D system.

5. Two-Dimensional Analysis

[25] For the 2-D analysis, we consider the buildup of a groundwater mound under a straight infiltrating stream. The height of the mound at any particular observation point is related to the infiltration rate beneath the stream and the transmissivity of the aquifer. Therefore, it should be possible to assess the state of connection/disconnection in terms of the groundwater level defined at some point distant from the surface water body, and not only directly beneath it as in the 1-D case. For example, a borehole adjacent to a stream is much more common than a borehole beneath the stream. Identifying the hydrogeological data that can be used to assess connection is of great practical relevance. Taking these practical implications into account, we aim to formulate our approach in a way that allows the state of disconnection to be expressed in terms of a groundwater level in an observation bore at a given distance from the center of the stream.

[26] As we will illustrate later, the regime beneath a disconnected surface water body may be conceptualized as a groundwater mound and a capillary zone which develops above this mound. The size of the capillary zone was discussed in the previous section and here we include the groundwater mound in the analysis. The buildup of a

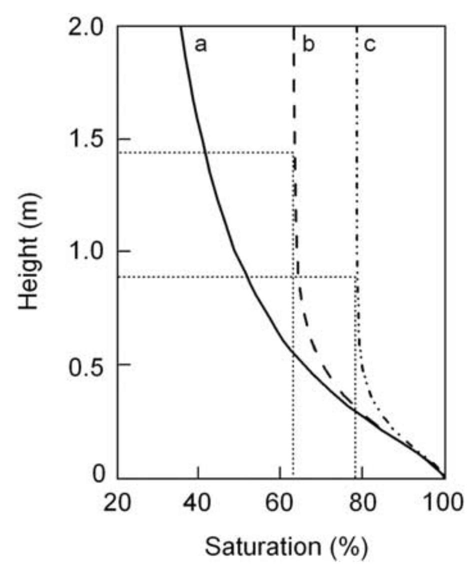

Figure 5. Saturation versus height above the water table for a loam soil under vertical infiltration rates of (a) 0, (b) 0.1, and (c) 0.5 m d^{-1}. The height of the capillary zone is defined as the height above the water table where the pressure is within 0.1% of the minimum possible value. Thus, for a flux of 0.1 m d^{-1}, the height of the capillary zone is 1.45 m, while for a flux of 0.5 m d^{-1} it is 0.88 m.

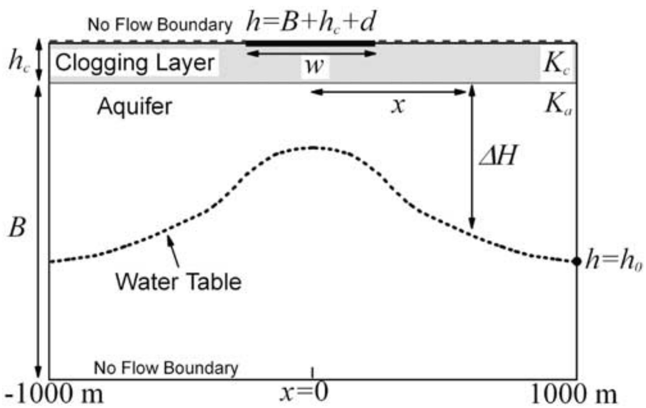

Figure 6. Conceptual model of the 2-D system. The surface water body is represented as a constant head boundary ($h = B + h_c + d$) and is separated from the groundwater by a clogging layer with a thickness h_c and hydraulic conductivity K_c. The underlying aquifer has a vertical thickness B. For the particular case shown in the setup, the difference ΔH is large enough to cause a complete disconnection of the system. An additional increase in ΔH does not increase the exchange flux.

groundwater mound in two dimensions has been studied by several authors. *McWhorter and Nelson* [1979] studied the buildup of a mound under tailings impoundments. *Warner et al.* [1989] provided an overview and discussion of commonly used equations for estimating mound height. The equations often rely on the Dupuit assumption. However, the rise of a groundwater mound in response to steady infiltration may involve steep hydraulic gradients and neglecting vertical fluxes can lead to large errors. Only recently has an analytical solution of the two-dimensional Laplace equation that does not invoke the Dupuit assumption or any other form of linearization been presented [*Schmitz and Edenhofer*, 2000]. However, this solution is for a horizontally infinite aquifer. Because of the inherent limitations of the available analytical approaches, we use numerical models.

5.1. Modeling Approach

[27] We have modeled the disconnection problem in 2-D using the numerical model HydroGeoSphere (HGS [*Therrien et al.*, 2006]). HydroGeoSphere's capability to simulate the buildup of a groundwater mound beneath an infiltration source was tested by numerically reproducing the analytical solution presented by *Schmitz and Edenhofer* [2000]. This comparison is detailed in Appendix A. There is excellent agreement between the two approaches. The principal model setup for subsequent simulations in HGS is shown in Figure 6. A clogging layer with thickness h_c separates the aquifer from the surface water body. The surface water body itself is defined as a constant head boundary in the center of the model domain. In addition to the variables identified in the 1-D system, three additional variables are now important: the width of the surface water body (w), the distance x where the water table is observed and the thickness B of the aquifer. No flow boundaries are used along the top of the model domain outside of the surface water body, and at the base of the aquifer. Constant head conditions ($h = h_0$) were used for the lateral boundaries, at $x = \pm 1000$ m. The choice of 2000 m for the horizontal model domain length ensured that the boundaries did not significantly affect the vertical component of the flow field beneath the mound for the range of simulation parameters employed in this analysis.

[28] The model domain was discretized finely to ensure grid-independent results. Vertical discretization must be high in the region below the clogging layer where unsaturated flow can occur. A vertical discretization of at most 0.05 m was applied for a vertical extent of 2 m immediately below the clogging layer. In most cases, however, an even finer discretization over a larger vertical extent had to be used in order to adequately simulate the development of the capillary zone. The vertical discretization of the clogging layer was at most 0.05 m. The horizontal discretization increases from 1 m in the region of the surface water body up to 100 m close to the boundary at the edge of the model domain.

5.2. Relating Water Table Height and Disconnection Status to Infiltration Rate

[29] Figure 7 illustrates the effect of lowering the head at the lateral boundaries on the infiltration flux and pressure at the base of the clogging layer in the middle of the model domain ($x = 0$). We report the results as a function of ΔH, which is the depth of the water table at the observation point relative to the base of the clogging layer. Clearly, ΔH is a function of the location of the observation point, and can be reported for any location between the surface water body and the point where the boundary condition is defined. Here we have used an observation point at $x = 100$ m. ΔH will increase as the observation point is moved further away from the edge of the surface water body. ΔH_d is the water table depth at the observation point at which the entire

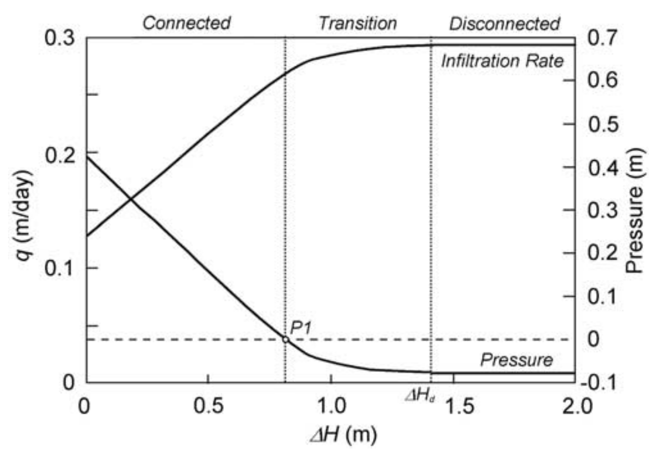

Figure 7. The infiltration rate q between surface water and groundwater and the pressure at the base of the clogging layer (at $x = 0$) plotted as a function of ΔH at $x = 100$ m. Model parameters used to generate Figure 7 are $w = 10$ m, $B = 120$ m, $h_c = 0.3$ m, $K_c = 0.1$ m d^{-1}, $K_a = 5$ m d^{-1}, $d = 0.5$ m, $\alpha = 14.5$ m^{-1}, and $\beta = 2.68$. For $\Delta H > \Delta H_d$, the flux remains constant. At the point P1, the pressure at the base of the clogging layer is zero for this particular parameter set. P1 corresponds to a ΔH of 0.82 m. Values of infiltration and pressure calculated by the numerical model for values of $\Delta H > \Delta H_d$ are identical to those calculated using equations (8) and (9).

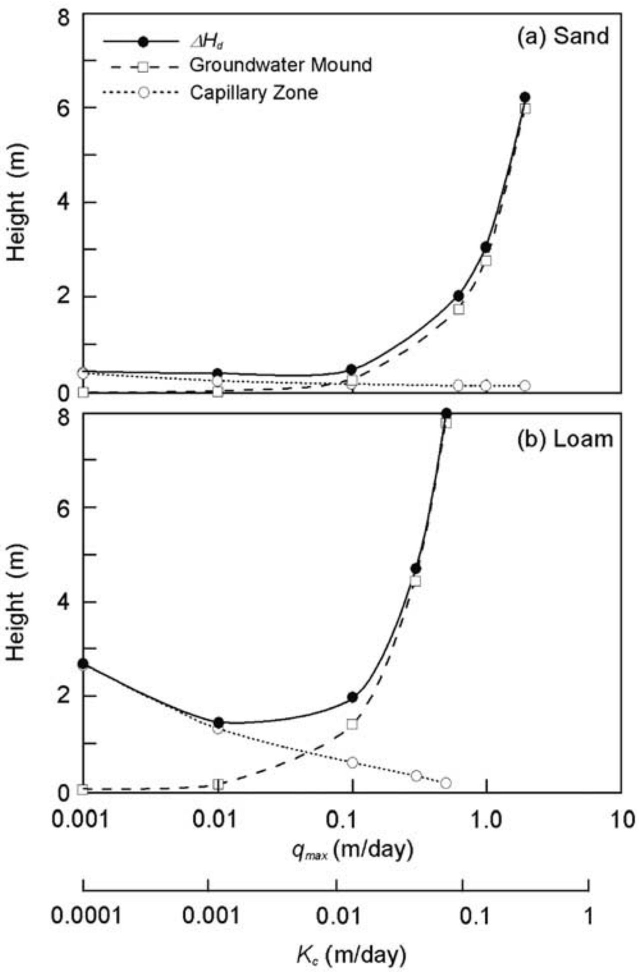

Figure 8. ΔH_d for two different soil types plotted for different values of K_c. ΔH_d is reported 100 m away from the center of the stream (x = 100 m) and was calculated using HGS (closed circles and solid line). Different values of K_c correspond to different infiltration rates at disconnection (q_{max}). Consequently, ΔH_d can be expressed either in terms of K_c or q_{max}. For every plotted value of q_{max} the capillary zone height was calculated using the Richards equation. This is shown as a dotted line. Also, the buildup of a groundwater mound under the infiltration layer was calculated by using MODFLOW (open squares and dashed line). For the sand simulation, the parameters are $K_a = 5$ m d^{-1}, $\alpha = 14.5$ m^{-1}, $\beta = 2.68$, $w = 10$ m, $d = 1$ m, $B = 120$ m, and $h_c = 0.2$ m. The loam simulation differs only in the choice of the van Genuchten parameters ($\alpha = 3.6$ m^{-1} and $\beta = 1.56$) and the saturated hydraulic conductivity ($K_a = 1$ m d^{-1}). Note that the sum of the groundwater mound and the height of the capillary rise is equal to ΔH_d.

surface water body becomes fully disconnected. Because the pressure at the base of the clogging layer approaches the pressure at disconnection asymptotically, for practical purposes a cutoff value must be defined at which the system is considered to be fully disconnected. As illustrated in Figure 5, we have defined this cutoff value to be when the pressure at the base of the clogging layer is within 0.1% of the pressure present in full disconnection (γ^*_p in equation (6)). Thus while the beginning of the transition zone is clearly defined, defining the precise point when full disconnection occurs is arbitrary.

[30] When ΔH is small (i.e., water table at x = 100 m is close to the base of the clogging layer), the pressure at x = 0 is positive. With increasing ΔH, the pressure at the interface is lowered until it is zero. This point (P1) is shown in Figure 7. For the parameters chosen, the ΔH where this pressure is reached is 0.82 m. For $\Delta H < 0.82$ m, the infiltration rate is approximately proportional to the head difference and K_c. For $\Delta H > \Delta H_d$, the system is fully disconnected, and consequently neither the flux nor the pressure change significantly with an increasing ΔH. The point at which $\Delta H = \Delta H_d$ is of considerable interest and will be subsequently called the point of disconnection. If the water table at the point of disconnection is known, it can be used to assess the status of connection by comparing it with a measurement of the depth to groundwater in a borehole. For the model setup shown, ΔH_d = 1.4 m. It should be noted that the change in flux becomes small as ΔH_d is approached.

[31] In order to calculate the point of disconnection for different systems efficiently, the lateral head boundary condition was lowered until the simulated values for pressure at x = 0 at the interface between the clogging layer and aquifer were within 0.1% of the pressure present in full disconnection (γ^*_p) as discussed above. The task was automated using the parameter estimation software PEST [*Doherty*, 2002]. Figure 8 shows the variation in ΔH_d as a function of K_c for an aquifer consisting of sand (Figure 8a) and a loam (Figure 8b). Parameters other than K_c are held constant.

[32] As shown in the 1-D section and illustrated in Figures 4 and 7, the maximum infiltration flux occurs when ΔH_d is reached. Because all parameters except K_c are kept constant, ΔH_d can be expressed either in terms of K_c or q_{max}. This allows the state of disconnection to be assessed on the basis of a model that only simulates saturated flow since at the point ΔH_d is reached, the height of the capillary zone is clearly defined and can be calculated using the Richards equation. This is also shown in Figure 8 for both soil types. Likewise, the shape of the groundwater mound under an infiltration layer with the infiltration rate q_{max} can be calculated. The numerical model MODFLOW 2000 [*Harbaugh et al.*, 2000] has been used to calculate the height of the saturated groundwater mound, and unlike HGS, unsaturated flow was not simulated. Since MODFLOW was used only to simulate the buildup of the saturated groundwater mound in the case of full disconnection, the setup differed slightly from that shown in Figure 6. For this situation, the surface water body and the clogging layer can be replaced by a constant flux boundary. The applied infiltration rate q_{max} was calculated as a function of K_c. The lateral boundary head was varied until the height of the groundwater mound beneath the center of the surface water body was below the clogging layer by an amount equal to the height of the capillary zone. The results are reported as depth to groundwater below the clogging layer at x = 100 m. As can be seen in Figure 8, ΔH_d calculated using HGS is exactly equal to the sum of the groundwater mound height (calculated using MODFLOW) and the height of the capillary zone.

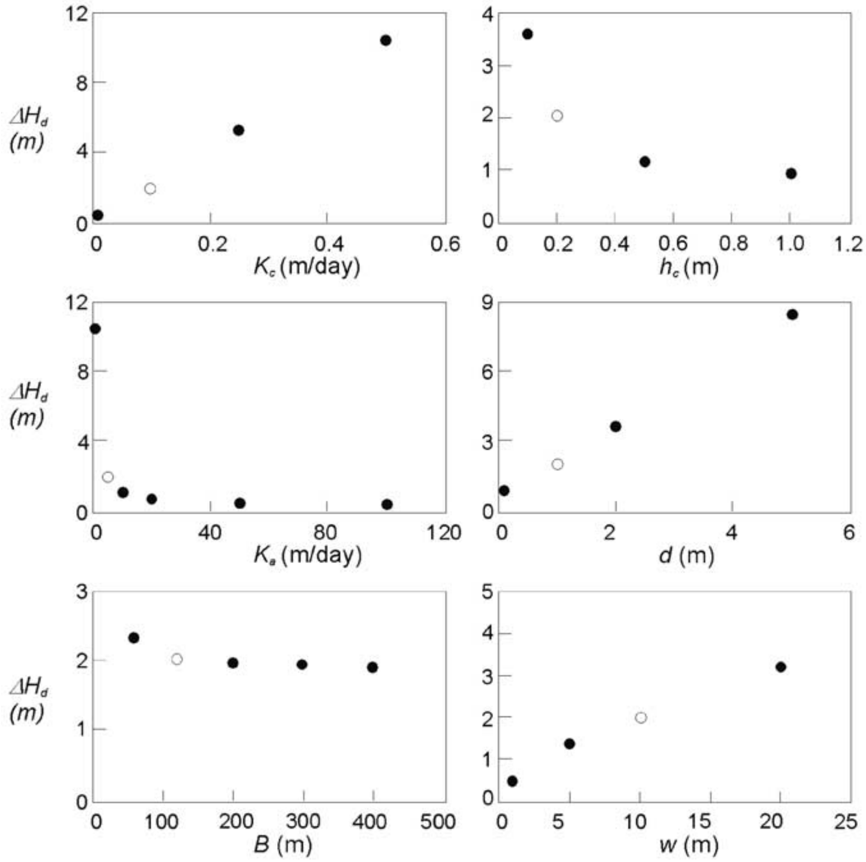

Figure 9. Sensitivity of head difference ΔH_d to various model parameters. In each case, a single parameter is varied while all other parameters remain unchanged, and ΔH_d is expressed relative to the water table at $x = 100$ m. The base case simulations are represented by open symbols and are for $w = 10$ m, $d = 1$ m, $B = 120$ m, $h_c = 0.2$ m, $K_c = 0.1$ m d^{-1}, and $K_a = 5$ m d^{-1}. All simulations use van Genuchten parameters for sand.

[33] For large fluxes shown in Figure 8, the capillary zone is small (Figure 5), and the groundwater mound dominates ΔH_d. Conversely, the extent of the capillary zone increases as the flux is reduced. It is apparent from Figure 8 that for very small fluxes, the capillary zone dominates ΔH_d. For example in a loam soil with small flux, e.g., $q_{max} = 0.001$ m d^{-1}, ΔH_d equals 2.7 m, which is due entirely to the height of the capillary zone.

[34] Figure 8 demonstrates that the height at which full disconnection occurs is equal to the height of the groundwater mound plus the capillary zone, as stated in section 3. This finding is important as it allows one to use the depth to groundwater measured at any distance away from the river to assess the state of disconnection. In theory, this can be done for every system by finding a water table that fulfills the condition $\Delta H = \Delta H_d$. A measurement of the water table in a borehole can then be compared to this water table. If the measured depth to groundwater is greater than ΔH_d the system is disconnected.

5.3. Parameter Sensitivities

[35] Figure 8 demonstrated two different ways to assess the sensitivity of K_c on the point of disconnection at $x = 100$ m. In Figure 9, the sensitivity of all parameters is plotted for the sandy material. The simulations were carried out in HGS using PEST. ΔH_d could also have been determined using the 1-D analysis, the Richards equation and MODFLOW. Figure 9 shows the variation of all parameters around a base case simulation. The base case simulation uses $w = 10$ m, $d = 1$ m, $B = 100$ m, $h_c = 0.2$ m, $K_c = 0.1$ m d^{-1}, $K_a = 5$ m d^{-1} (and van Genuchten parameters for the sand as previously described). All results are reported at $x = 100$ m.

[36] ΔH_d appears approximately proportional to the hydraulic conductivity of the clogging layer, K_c, and also to the depth of ponding, d. Inverse relationships between ΔH_d and K_a as well as h_c are also apparent from Figure 9. The relationships between the head difference and the stream width w and aquifer thickness B are more complex. ΔH_d increases as w increases, but the rate of increase is not constant. The value of ΔH_d decreases as B increases. With an increasing B, however, the rate of decrease is reduced. All proportionalities described above were independent on the location x of the observation point.

[37] It is noteworthy that the relationships between K_c, d and w and ΔH_d shown in Figure 9 do not pass through the coordinate origin. This is because of the influence of the capillary zone. Nevertheless, for the range in parameters shown in these simulations, the size of the capillary zone is relatively small, and so the dependencies between the parameters are largely due to the size of the groundwater mound. The influence of the capillary zone becomes more pronounced for finer textured soils and smaller infiltration rates. Also, in the plots showing relationships to K_c, h_c and d, the size of the capillary zone is not constant, and this

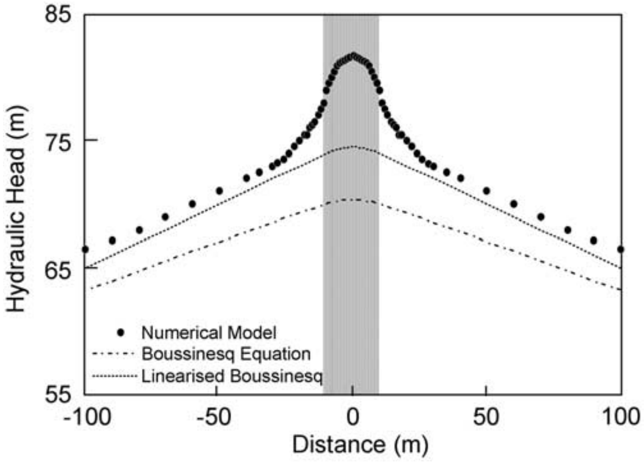

Figure 10. Comparison of different approaches to calculate the position of a groundwater mound under a constant flux boundary. The parameters used are q = 0.5 m d^{-1}, w = 20 m, B = 100 m, K_a = 1 m d^{-1}, L = 250 m, and h_0 = 50 m. The transmissivity for the linearized Boussinesq equation was assumed to be 50 m^2 d^{-1}. The shaped region represents the horizontal extent of the infiltration zone. The numerical solution was obtained using the HydroGeoSphere numerical model. The analytical solutions are given in Appendix B.

contributes to some small deviations from the linear and inverse relationships. The influence of the capillary zone, however, can easily be calculated and subtracted from ΔH_d. This allows an analysis of the groundwater mound independent of the influence of the capillary zone. For the range of parameters we tested to construct Figure 9, the approximate proportionalities identified in Figure 9 become exact if only the buildup of the mound is analyzed. These proportionalities suggest that a dimensionless parameter $(\frac{wq_{\max}}{BK_a})$ identical to that found in the linearized Boussinesq equation may be an appropriate variable to describe system behavior. The sensitivity of ΔH_d to the parameters tested was assessed for various positions x between the surface water body and the lateral boundary condition. In contrast to the absolute values of ΔH_d, the proportionalities were found to be independent of the location x and suggest that this dimensionless parameter holds for all x.

[38] The linearized Boussinesq equation (which incorporates the Dupuit assumption of horizontal flow) is given in Appendix B. It is interesting to note that the dimensionless parameter discussed above is essentially a measure of overall hydraulic gradient in the saturated system (volumetric flow rate divided by transmissivity). However, the linearized Boussinesq equation cannot be used to calculate the height of the groundwater mound accurately because 2-D flow is not considered. The steep gradients of the mound under the infiltration zone require the consideration of 2-D flow components. Simplified analytical expressions that do not take 2-D flow into account are therefore inaccurate and the largest discrepancies occur closest to the infiltration zone. For the disconnection problem, this is the most important region. Figure 10 compares the shape of the groundwater mound beneath an infiltration zone calculated by the numerical model, with that calculated using the Boussinesq equation and the linearized Boussinesq equation. In the linearized form of the Boussinesq equation a constant transmissivity is assumed. The equations are presented in Appendix B. Both of these analytical solutions significantly underestimate the height of the mound in the vicinity of the surface water body where the infiltration flux occurs.

6. Discussion

[39] Accurate estimation of the effect of changes in the groundwater table on infiltration rates from losing streams is crucial for water resource management. However, the effect is fundamentally different for connected and disconnected systems. Furthermore, many of the analytical solutions that are used for quantifying these impacts assume full saturation, and therefore are only applicable to fully connected systems. Identifying the state of connection is therefore critical.

[40] We show that the description of disconnected systems can be broken down into two separate problems (mound and capillary zone), and thus disconnection can potentially be assessed in two different ways: (1) the use of a fully coupled variably saturated flow model and (2) the independent determination of the mound and the additional contribution of the capillary zone. In the latter case, where the mound and the capillary zone are calculated independently the following steps should be performed: The infiltration flux q_{max} is calculated using equation (6). The height of the capillary zone can then be calculated independently as a function of q_{max} by finely discretizing the Richards equation. The development of the mound under a constant flux boundary (with the infiltration rate q_{max}) is determined using a numerical model that simulates saturated flow only.

[41] Alternatively, the disconnection problem can be approached by using a fully coupled, variably saturated flow model. If this approach is used, the point of disconnection can be identified without using the equations outlined in the 1-D analysis. This can be done by lowering the lateral boundary condition until the infiltration flux or the pressure under the clogging layer is essentially constant. However, even if such a numerical model is used, there are two reasons why calculating q_{max} using the 1-D approach is still useful: (1) The highest infiltration rate or lowest possible pressure are the "observations" required if the point of disconnection is assessed using a parameter estimation software such as PEST and (2) calculating q_{max} allows an assessment of the height of the capillary zone before the fully coupled saturated/unsaturated model is set up. This is important information relevant to the vertical discretization under the clogging layer. Because of the highly nonlinear nature of the Richards equation, the vertical discretization must be very fine to adequately describe the unsaturated zone and minimize errors. A priori knowledge of the vertical extent of the unsaturated zone before the construction of the numerical model allows fine vertical discretization to be used only where it is required. Clearly the considerations on the model setup and boundary conditions mentioned in the context of the fully saturated model are also relevant if a saturated/unsaturated model is used.

[42] Some important remarks can now be made. Independent of the chosen approach to disconnection assessment, equation (5) is a very useful condition that determines whether a system can disconnect or not. If no disconnection is possible, it is expected that lowering the water table will continue to increase the infiltration rate. In such cases, the problem can be approached with a model that does not take

unsaturated flow into account. If disconnection is possible, the goal is then to identify the point of disconnection (the water table where $\Delta H = \Delta H_d$). As discussed above, this can be done by using either fully coupled variably saturated or fully saturated flow models. However, to determine the change in infiltration flux over the connected-disconnected transition phase requires a model capable of simulating both saturated and unsaturated flow. If analytical expressions are developed to describe the mound, the errors introduced by assumptions of horizontal flow or any kind of linearization must be assessed and taken into account. Taking the limitations of simplified analytical solutions into account, the use of numerical models may be preferable. If the mound is assessed using a saturated numerical model, the choice of the boundary conditions is critical. *Osman and Bruen* [2002] used a constant head boundary condition for the base of the aquifer and the lateral aquifer boundaries. Such a boundary condition creates a flow through the base of model domain. In our study, we have chosen to use lateral constant head boundaries, and a no flow boundary at the base of the aquifer. We believe that this is more representative of field conditions. The no flow boundary at the bottom represents the base of the aquifer system. However, by defining a constant head at the lateral boundary, we are essentially imposing a horizontal flow condition at this point. It is therefore important that the location of this boundary is far enough away from the infiltration layer to ensure that the vertical flow component under the mound is not affected by it. The influence of this vertical component depends on the infiltration flux as well as the transmissivity of the aquifer.

[43] The critical depth ΔH_d can be used to determine whether any particular system is connected or disconnected by using a measurement of the groundwater table depth in a bore adjacent to the stream. The sensitivity analysis of the hydraulic parameters we carried out covers a wide range of systems for which we have identified the point of disconnection (however, we only presented information at $x = 100$ m). In principle, these data can be used to assess disconnection in the field. In practice, however, this would require accurate measurements of both the thickness h_c and hydraulic conductivity K_c of the clogging layer which, unlike the other parameters, are often difficult to obtain. Streambed conductivity, in particular, can be highly spatially variable and difficult to measure [*Calver*, 2001; *Cey et al.*, 1998]. Of course, our sensitivity analysis is for a highly idealized system: a 2-D horizontal, isotropic, homogeneous aquifer, and a stream of constant depth underlain by a homogeneous isotropic clogging layer of constant thickness. We do not consider regional groundwater flow. The analysis is simplified, but these simplifications are necessary to develop a fundamental understanding of the key variables and processes inherent to disconnection phenomena. More complex flows and geometries, for example due to regional flow or pumping from bores adjacent to the stream, are not considered here. Our analysis is also restricted to steady state conditions. Where stream levels rise or fall rapidly, transient effects will be important. Steady state conditions are particularly unlikely to occur in intermittent and ephemeral streams. Further work is required to examine the effect of heterogeneity, irregular geometries of the clogging layer and aquifer materials and stream morphology on the sensitivity of the parameters and transient disconnection behavior. In addition, a comparison of 2-D (river) and 3-D (lake) disconnection processes will also be important. Despite these limitations, our analysis provides an important first-order conceptual understanding of the major physical processes involved in disconnection phenomena and provides quantitative criteria that may be useful for predicting their occurrence.

[44] This study provides an understanding of the sensitivity of parameters which may also assist in extrapolation of results from local field assessments to the wider basin scale. Some comments regarding local-scale and wider basin-scale disconnection assessments and their implications for water resources management may be made. It is important to note that even if a stream is locally disconnected and therefore an additional lowering of the groundwater table does not increase the infiltration rate at that particular point, the influence of groundwater pumping on a basin scale must be considered. Increased pumping adjacent to a disconnected reach of a river would be expected to increase the length of the disconnected reach. Furthermore, even if a river is disconnected over its entire length, lowering the groundwater table (by groundwater pumping or other mechanisms) will always have an affect on the water balance of the entire basin. Because of the complexity of disconnection processes, strong precaution must therefore be exercised when attempting to account for disconnection phenomena in water resources management. The existence of disconnected systems is therefore not a license to increase groundwater pumping without very careful assessment of both surface water and groundwater balances, and other potential basin-scale water resources and environmental impacts. Further work is still required to examine disconnection processes at a basin scale and its implications for water resources management.

7. Conclusions

[45] In this paper we use a theoretical and modeling approach to assess the hydrogeological parameters that influence the connection status between surface water and groundwater and to quantify how variations in these parameters influence the status of connectivity. The analysis is for straight rivers, and assumes steady state conditions and homogeneous hydraulic conductivities. The main findings are as follows:

[46] 1. Equation (5) is a criterion that can be used to determine whether a given system can potentially become disconnected or not. This criterion helps to choose a correct conceptual model. Equation (5) shows that a clogging layer is a necessary but not sufficient criterion for disconnection.

[47] 2. The most basic physical condition required for disconnection is that the top of the capillary zone does not intersect the base of the clogging layer in the center of the surface water body. If the water table corresponding to the precise point of disconnection (ΔH_d) is known, it can be used to assess the state of connection by comparing ΔH_d with a measured depth to groundwater. If the measured depth to groundwater is greater than ΔH_d, the system is disconnected.

[48] 3. ΔH_d is determined by two quantities: the buildup of a saturated groundwater mound and the development of the capillary zone above the mound. These two quantities

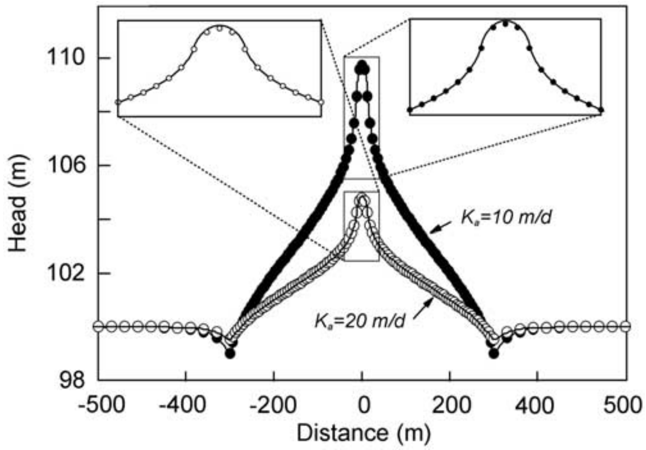

Figure A1. Comparison of the free water surface calculated with the analytical solution with HydroGeoSphere for two hydraulic conductivities (K_a = 10 m d^{-1} and 20 m d^{-1}). The model uses a constant flux boundary (q = 2.48 m d^{-1}) along the upper boundary between -10 m $< x <$ 10 m. The thickness of the model is 1 m. Two constant flux nodes at a height of 90 m at $x = \pm 250$ m are assigned a flux of each $q = -12.4$ m^3 d^{-1}. The insets show a close-up view of the water table in the vicinity of the infiltration flux. The lines represent the analytical solution while the dots represent the solution obtained with HydroGeoSphere.

can be calculated separately and can be superimposed. Therefore, ΔH_d can be estimated using both variably saturated and fully saturated groundwater models, the latter including examples such as MODFLOW. If a fully saturated model is used, the contribution of the capillary fringe to ΔH_d has to be calculated separately using the Richards equation. If the mound is calculated using analytical solutions, the influence of linearization of the flow equation can be significant and must be taken into account. However, considering the limitations of analytical solutions shown in this paper, we recommend the use of numerical approaches in disconnection analyses.

[49] 4. A sensitivity analysis showed that for a given aquifer thickness and stream width, the depth to groundwater where the system disconnects is approximately proportional to both the stream depth and the hydraulic conductivity of the streambed sediments and inversely proportional to both the thickness of these sediments and the hydraulic conductivity of the aquifer. Knowing the relevant parameters and their sensitivities may assist in design of field investigations aimed at assessing disconnection processes status, and may also be useful in spatially and temporally extrapolating results of local field studies to the wider basin scale.

[50] Further theoretical work is required to consider more complicated situations. This will improve the conceptual understanding of disconnection processes and will allow the development of urgently required practical methods to determine the state of disconnection in the field.

Appendix A: Code Verification

[51] Numerical simulations were carried out with HydroGeoSphere [*Therrien et al.*, 2006]. To assess the suitability of this model for simulating the buildup of a groundwater mound, comparisons were made of the free water surface predicted using HydroGeoSphere and the analytical solution presented by *Schmitz and Edenhofer* [2000]. The analytical solution was calculated by using the software package TEXAS2D [*Schmitz et al.*, 2004]. In the example used to verify HydroGeoSphere, the steady state flow between an infiltration zone in the center of the model domain and two pumps at a lateral distance of $x = \pm 250$ m and a height of 90 m is simulated. The simulation of the flow field was carried out in both TEXAS2D as well as HydroGeoSphere. Figure A1 shows a comparison of simulations for two different values of hydraulic conductivity. HydroGeoSphere accurately reproduces the analytical solution.

Appendix B: Analytical Solutions

[52] The shape of the groundwater mound under an infiltration zone can be approximated using analytical solutions based of the Dupuit assumption. Approximate solutions based on the Boussinesq and linearized Boussinesq equations are presented below.

B1. Boussinesq Equation

[53] The general flow equations for a given infiltration flux q are given by (see Figure 6 for illustration)

$$Kh\frac{dh}{dx} = -qx \text{ for } 0 \le x \le w/2 \tag{B1}$$

$$Kh\frac{dh}{dx} = -q\frac{w}{2} \text{ for } w/2 \le x \le L, \tag{B2}$$

where K is the hydraulic conductivity of the aquifer, q is the infiltration rate, w is the width of the infiltration zone and h is the hydraulic head. Boundary conditions are

$$h(L) = h_0 \text{ and } dh/dx = 0 \text{ at } x = 0.$$

L corresponds to the distance from the center of the stream where the boundary condition is defined. The solutions for these equations are

$$h^2(x) = \frac{q}{K}\left(wL - \frac{w^2}{4} - x^2\right) + h_0^2 \text{ for } 0 \le x \le w/2 \tag{B3}$$

$$h^2(x) = \frac{wq}{K}(L - x) + h_0^2 \text{ for } w/2 \le x \le L. \tag{B4}$$

B2. Linearized Boussinesq Equation (Constant Transmissivity)

[54] General equations are

$$T\frac{dh}{dx} = -qx \text{ for } 0 \le x \le w/2, \tag{B5}$$

$$T\frac{dh}{dx} = -q\frac{w}{2} \text{ for } w/2 \le x \le L, \tag{B6}$$

where T is the transmissivity of the aquifer. Boundary conditions are

$$h(L) = h_0 \text{ and } dh/dx = 0 \text{ at } x = 0.$$

Solutions are

$$h(x) = \frac{q}{2T}\left(wL - \frac{w^2}{4} - x^2\right) + h_0 \text{ for } 0 \le x \le w/2 \qquad \text{(B7)}$$

$$h(x) = \frac{qw}{2T}(L - x) + h_0 \text{ for } w/2 \le x \le L. \qquad \text{(B8)}$$

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