Effects of stream-aquifer disconnection on local flow patterns

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[1] Disconnected stream-aquifer systems are becoming increasingly common because of lowering groundwater tables. This work focuses on the pathways and rates of infiltration and seepage as streams transition from fully connected to disconnected conditions. HYDRUS-2D simulations show for a connected stream, water infiltrates vertically then moves laterally below the top of the preflooding capillary fringe height, finally causing an upward displacement of antecedent water into the vadose zone. This contradicts the commonly held assumption that stream water moves laterally for some distance into the stream bank, forming a wedge above the antecedent water. Even for shallow disconnections (<1 m), there is an increase in infiltration losses from the stream, elimination of postflooding seepage, and increased lateral and vertical mixing of stream water and antecedent pore water.


1. Introduction

[2] There is growing recognition of the importance of exchanges of water and solutes between surface water bodies and the subsurface [Choi et al., 2000; Dahm et al., 1998; Gooseff et al., 2002; Gooseff et al., 2007; Packman and Mackay, 2003]. These fluxes are particularly critical in riparian areas within water-stressed regions. For example, recent studies have shown that lowering of water tables due to climate change combined with sustained human withdrawals has led to die off of long-lived riparian species [Cooper et al., 2006; Naumburg et al., 2005; Sarr and Hibbs, 2007]. These studies demonstrated a correlation between water table depth and riparian vegetation morbidity.

[3] Most studies of the interaction of surface water and groundwater in stream banks make use of one of two limiting simplifying assumptions. Either, they assume that the stream is connected to the groundwater system [Huth, 2003; Kondolf et al., 1987; Whitaker, 2000; Whiting and Pomeranets, 1997], or they assume that the water table is far below the stream [Constanitz et al., 2002; Niswonger et al., 2005; Ronan et al., 1998]. The former case is typically conceptualized as lateral flow from the stream to the stream bank during flooding, followed by lateral return flow to the stream, via seepage, during recovery. The latter case is conceptualized as predominantly vertical flow during flooding with no return flow to the stream. Analytical solutions are available for these two end-member conditions [Hunt, 1990; Neuman, 1981; Philip, 1989; Vazquez-Suñe et al., 2007]. We examine in more detail the changes in the rates and patterns of water flow and solute transport that occur as the water table elevation decreases, causing a stream to transition from an initial connected condition to a state of disconnection from the underlying aquifer.

2. Methods

[4] We use a numerical model of water flow and solute transport [Šimůnek et al., 1999] to examine the response to a 6 h step function flooding event followed by 48 h of recovery (Figure 1, right). We focus on the following three features: water loss from the stream during flooding, seepage from the subsurface to the stream following flooding, and the spatial distribution of stream water in the subsurface following flooding and following 2 days of recovery. We consider an initially flat lying water table and examine the impacts of changes in the water table depth below the base of the stream on these three behaviors.

[5] To focus on interactions between the stream bank and seepage face, we model a vertical cross section, with the left boundary located at the interface of the stream bank and stream (Figure 1, left). The domain is 6 m in the vertical and 15 m in the horizontal. The datum elevation is located at the base of the stream. The stream is represented as a vertical line segment along the left boundary extending from 0 to 1 m (the elevation of the ground surface). The left boundary below the stream is set to zero flux for water flow and for solute transport to represent a symmetry boundary; as such, the left side boundary could be considered the midpoint of an infinitely narrow stream. The stream segment of the left boundary is set to a constant hydraulic head of 0.4 m during the 6 h flooding stage (Figure 1, right). Following this, during the seepage and recovery stage, the stream boundary segment has a seepage face condition. In HYDRUS-2D a seepage face condition maintains zero pressure head along the saturated portion and zero flux along the unsaturated portion of the boundary nodes. Water that crosses the boundary is considered to be immediately removed, as by stream flow. The bottom (z = −5 m) and top (z = 1 m) boundaries are zero flux for water and solute. A distant right boundary (x = 15 m) is also zero flux.

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The medium is composed of an isotropic, homogeneous sand with van Genuchten [van Genuchten and Nielsen, 1985] parameters ($\alpha = 15.5$ m$^{-1}$, $n = 2.68$) and a saturated hydraulic conductivity of $K_s = 0.297$ m h$^{-1}$. The maximum initial water table height is $z = 0$ m, representing a stream that is minimally connected to groundwater. The initial condition above the water table is hydrostatic. We then examine increasing stream-aquifer disconnection by increasing the initial water table depths ($d$) to a maximum of 2.0 m.

We use the solute transport modeling capabilities of HYDRUS-2D to track the movement of stream water (assigned a relative concentration of zero) and antecedent pore water (assigned a relative concentration of one throughout the subsurface). We assign longitudinal and transverse dispersivities of $\alpha_L = 0.5$ m and $\alpha_T = 0.1$ m. However, we set the diffusion coefficient to zero to avoid unreasonably large diffusive mass fluxes due to the large difference in relative concentrations in the stream water and antecedent pore water.

### 3. Results and Discussion

Consistent with results published previously [Niswonger et al., 2005], cumulative infiltration during flooding is higher for disconnected streams ($d > 0$) than for connected streams (solid curve in Figure 2). Our results show that the cumulative infiltration transitions nonlinearly, with relatively large changes occurring with the initial disconnection of the stream and aquifer. For the flooding scenario that we examined, if the initial water table depth is at 0 m below the streambed, the cumulative infiltration will be only 34% of the amount that would occur for an infinitely deep water table (i.e., a water table that is so deep that the infiltrating water does not reach it during the simulated period); whereas, if the water table has dropped to 1 m below the stream, it is 88% of that value.

In all cases, infiltrated water causes the water table to rise near the stream and extend laterally as a wedge away from the stream (Figure 3, left). For the conditions that we examined, water table rise near the stream led to nonzero seepage (difference between the solid and dashed curves in Figure 2) for an initial water table depth of 0.25 m or less. The net water flux to the subsurface through flooding and recovery (cumulative infiltration minus cumulative seepage) was highly nonlinear. That is, even a small disconnection between the stream and the underlying aquifer led to a much greater net loss of water from the stream (dashed curve in Figure 2) throughout a flooding event.

The spatial distribution of stream water following flooding ($t_f = 0$ h) and 48 h after the end of the flood ($t_f = 48$ h) shows the subsurface zone that is affected by stream water exchange (Figure 3, right). We show two contour levels. Water within the 0.1 relative concentration contour represents essentially unmixed stream water. Water outside of the 0.9 contour represents essentially unmixed antecedent pore water. The area between these contours represents a zone of mixed stream water and antecedent pore water. If the initial water table depth is located at the base of the stream ($d = 0.0$ m), there is essentially no region of unmixed stream water. The mixing zone extends horizontally $\sim 1.7$ m.
from the stream bank and vertically to a maximum depth of 1.2 m. There is little or no change in the mixing zone during recovery because of the significant water flow toward the seepage face. If the initial water table depth is lowered by 0.25 m ($d = 0.25$ m), there is still no region of unmixed stream water. But, the increased cumulative infiltration extends the mixing zone horizontally ~2.0 m from the stream bank and vertically to a maximum depth of 1.6 m. There is continued downward and rightward expansion of the mixing zone during recovery as the water table mound dissipates (Figure 3b). As the initial water table depth increases to 2 m, there are significant changes in the distributions of stream water and antecedent pore water. Unmixed stream water extends to a lateral distance of 0.7 m and vertically to a depth of 1.1 m. The mixing zone extends vertically in the shallow subsurface, then laterally at and above the initial water table depth. The lateral extent of the mixing zone is much greater (2.1 m after flooding), partly due to the increased cumulative infiltration. Finally, there is significant additional lateral movement of the stream water plume away from the stream during recovery (mixing zone extends to 3.1 m, Figure 3c).

To examine the change in distribution of the antecedent pore water following flooding and 2 days of recovery, we calculated the change in solute mass ($dM$) throughout the domain as

$$dM = (C_f \theta_f - C_i \theta_i) V;$$
where $C$ is concentration, $V$ is the volume per unit length, $\theta$ the water content of each model element, and the subscripts $i$ and $f$ indicate initial and final conditions. The stream water has a concentration of 0.0, and the antecedent pore water throughout the subsurface has a concentration of 1.0. Therefore, a decrease in solute mass represents a decrease in the volume of antecedent water per unit volume of porous medium. Similarly, an increase in solute mass represents an increase in this “antecedent water saturation.” Changes in the saturation of stream water have no effect on the solute mass unless they displace antecedent water.

As expected, there is a large region of infiltrated stream water manifested as decreased solute mass (not shown on Figure 4a for clarity of presentation). Unexpectedly, there is a region of increased mass in the stream bank, above the base of the stream (Figure 4a), indicating an increase in antecedent water in this area. Examination of the flow vectors at the onset of flooding (Figure 4b) shows that the cause of this increase is upward flow of water from beneath the initial water table depth in response to mound- ing beneath the stream. Pressure from the mounded water propagates faster in the saturated zone than in the unsaturated zone and causes antecedent pore water to flow upward into the previously unsaturated stream bank sediments. Conceptually, this means that the leading edge of the wedge of water content increase in the stream bank is composed of antecedent water, not stream water. The stream water moves vertically downward and mixes in higher water content regions of the subsurface. This helps to explain why there is no significant region of unmixed stream water after flooding for shallow initial water table conditions (Figures 3a and 3b, right).

This circulating flow pattern is distinct from the common assumption of lateral flow during flooding of connected stream-aquifer systems and has several implications for solute transport. In particular, the pattern of infiltration that we show results in limited lateral transport of stream water during flooding. This could limit delivery of nutrients to the riparian area from the stream and/or flushing of nutrients from the riparian area to the stream. In addition, water that may have moved vertically through the root zone before a flooding event may be displaced upward, into the root zone, in response to flooding. Finally, the dependence of the persistence of antecedent water in the vadose zone on the water table depth may result in geochemical changes in the root zone as the water table depth increases.

To summarize the changes in the distribution of antecedent pore water above the base of the stream, we calculated the change in mass of the antecedent water compared to the mass before flooding (Figure 4c). For the connected stream ($d = 0.0$), there is a significant increase in the amount of antecedent water present immediately following flooding. However, most of this water drains out as seepage after 2 days of recovery. If the initial water table depth is decreased by 0.25 m, there is a similar increase in antecedent water mass above the base of the stream. However, under these conditions, the mass of antecedent water continues to increase as the water table mound redistributes laterally following flooding. Even this small disconnection significantly limits seepage, so the antecedent water mass that moves upward persists during recovery. However, if the initial water table depth is 50 cm or more.
below the base of the stream, there is a net loss of antecedent water mass due to vertical displacement by stream water.

[15] The numerical investigation presented here is intended to highlight some overlooked aspects of water flow in streams that are minimally connected or disconnected from an underlying aquifer. This study is not exhaustive, and the findings should be tested by field investigations. In particular, the following three factors that were not considered in this initial work could impact the rates and patterns of water flow during flooding and recovery: anisotropy, incomplete drainage before flooding, and the stream bank geometry.

[16] Many streambeds exhibit anisotropy in hydraulic conductivity, particularly as horizontal layering associated with fluvial depositional environments. If the hydraulic conductivities of adjacent layers contrast sharply, flow could be guided laterally through the higher conductivity layer. This structure could exert a primary control on both the rate and pattern of flow. The impact of this structure will likely depend on the pre-flood location of the water table relative to the layer boundaries.

[17] We modeled a hydrostatic initial (preflood) condition. In reality, it is unlikely that the vadose zone will achieve these conditions unless the soil is coarse and it has been a very long time since the last flood. Incomplete drainage, due to either frequent flooding or finer-grained materials, could lead to a thicker zone of near-complete water saturation above the water table. As a result, the water table would have to fall to a lower elevation to achieve disconnection than is reported in this study. However, even for these conditions, water would likely flow laterally within the saturated and near-saturated soils, as it did in the saturated zone for the sandy soil that we examined. Then, we expect that the same process of lateral flow in (near) saturated soils followed by upward flow into the vadose zone would occur. Hysteresis, which would result in higher water contents at each depth following drainage than would be predicted by a nonhysteretic moisture release curve, may have a similar effect.

[18] Finally, the horizontal stream bank geometry used in this study is greatly simplified compared to a natural stream. Analyses of streams with stream bank angles less than 90° from the horizontal (not shown) had infiltration rates and mixing patterns that were very similar to those shown in this study. However, the seepage rate decreased with decreasing stream bank slope.

4. Conclusions

[19] We present a preliminary numerical investigation of the response of a stream-aquifer system due to lowering of the water table below the base of the stream (stream disconnection). There are three primary parameters that occur when the water table elevation drops from the base of the stream to a depth of 0.25 m below the stream. First, there is a significant increase in cumulative infiltration and an associated decrease in seepage return, leading to much larger water loss from the stream over the course of flooding and recovery. Second, there is an increase in the lateral and vertical extents of the zone of mixing of stream water and antecedent pore water. Furthermore, this zone continues to expand during recovery; in contrast, this zone shrinks or remains stable during recovery under connected conditions. Third, there are significant changes in the movement of antecedent pore water. The observed upward flow pattern of antecedent water from the saturated zone to the unsaturated zone, most prominent in the minimally connected case ($d = 0$), decreases rapidly with increasing stream disconnection. However, antecedent water that flows upward persists in the stream bank longer in the disconnected case because of the decrease in seepage flow. As a result, there is more net gain of antecedent water redistributed into the stream bank above the base of the stream.

[20] While the magnitudes and timings of the responses seen in our modeling study will vary for specific site conditions, we expect that the underlying processes will occur to some degree in most systems. Therefore, we recommend that researchers who are interested in processes that rely on water and solute exchange between a stream and the stream bank should consider the processes shown here when designing experiments. In particular, our findings suggest that the assumption of one-dimensional lateral flow into the stream bank for connected and slightly disconnected stream-aquifer systems is incorrect. Rather, infiltrating water flows predominantly downward beneath the stream. This causes lateral flow in the saturated flow, which leads to upward flow of antecedent water into the previously unsaturated zone. This suggests that water content (or pressure) and solute concentration should be measured at multiple depths in the stream bank to fully characterize stream-aquifer interactions.

References


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