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Numerical investigation of leakage in sloping aquifers

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SUMMARY

A sloping aquifer resting on impermeable bedrock is a common paradigm in hydrological modeling. The underlying assumption of this paradigm is examined in this study of leaky hillslope systems. Leakage is simulated with a three-dimensional finite element Richards equation model for a 100 m synthetic hillslope composed of an unconfined and a confined aquifer separated by an aquitard. The simulations examine different configurations of aquifer and aquitard properties (hydraulic conductivity, aquitard thickness), hillslope geometry (uniform, convergent, divergent), hillslope inclination (0.2, 5, and 30%), and boundary conditions (Dirichlet, seepage face), as well as the interplay between leakage, water levels, and outflow. The results show that leakage generally percolates in both directions, with downward (positive) leakage in upslope portions of the aquifer and upward (reverse or negative) leakage in downslope regions. Geometry is found to be a main determinant of the partitioning of leakage along a hillslope, with for instance upward leakage in large portions of convergent slopes but only in a restricted downslope region for divergent slopes. In steep hillslopes, the reverse leakage that occurs downslope as a result of quick upslope drying represents a major component of the water budget. Outflow boundary conditions also exert a major control on the volume and direction of leakage, with the placement and extent of Dirichlet or seepage face nodes along the outflow face being particularly important factors. A dimensional analysis is used to synthesize the main findings and to highlight the differences in response between leaky and non-leaky hillslope conceptualizations. Leakage is also examined for a larger scale aquifer system, in a preliminary assessment of the importance of this exchange process for river basin models that are based on extensions of simple hillslope conceptualizations.

1. Introduction

The gaining recognition of hillslopes as fundamental units in watershed hydrology has led to hillslope-based models being used as building blocks for larger scale river basin models (e.g., Yang et al., 2002; Matonse and Kroll, 2009). At the hillslope scale, models based on the Boussinesq equation are commonly used because of their relative simplicity and low dimensionality, their amenability to analytical solutions and, their adaptability to complex plan and profile morphologies (e.g., Beven, 1981; Troch et al., 2003; Daly and Porporato, 2004; Hilberts et al., 2004; Basha and Maalouf, 2005; Chapman, 2005; Harman and Sivapalan, 2009b). The basic conceptualization for these models is that of an unconfined aquifer or soil mantle resting on sloping impermeable bedrock. The assumption of zero flow across the base of these units, while appropriate for many applications, has evident limitations, especially at larger scales when the interactions between different components of a system (e.g., flow between a soil mantle and a confined aquifer) play a greater role. It is thus important to examine the behavior of hillslope systems when the no-flow assumption is relaxed to allow leakage to an underlying aquifer, and to explore the geological, hydrological, morphological, and climatic conditions under which the leakage process is important.

The concepts of leakage and leaky layers have been extensively studied in classical hydrogeology, particularly in relation to aquifer pumping test analyses. Jacob (1946), Hantush (1949, 1960), and Hantush and Jacob (1954, 1955) published fundamental papers on plane and radial steady and non-steady flow in pumped infinite and finite leaky aquifers. The theory was extended by Neuman and Witherspoon (1969a,b) to take into consideration previously neglected effects of storage in the aquitard and drawdown in the unpumped aquifer. Hemker (1984) presented a combined analytical–numerical solution technique for steady flow in leaky aquifer systems with an arbitrary number of aquifers, applicable to flow problems with more complex boundary conditions. A generalized semi-analytical solution for a leaky and finite aquifer system was presented by Zhou et al. (2009). Leakage has also been investigated in the context of transient multilayer aquifer modeling (Herrera,

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1970; Cheng and Morohunfola, 1993; Gambolati and Teatini, 1996), landfill percolation (Jayawickrama et al., 1988; Foose et al., 2001), urban water management (Karpf and Krebs, 2005) and coastal aquifer tidal response (Jiao and Tang, 1999).

In hillslope hydrology, field evidence for the importance of leakage is documented in a number of studies. At the Coos Bay experimental catchment in Oregon, Montgomery et al. (1997, 2002) found that a significant proportion of storm runoff flows through sediments overlaying shallow fractured bedrock, and Anderson et al. (1997) corroborated these findings by means of tracer experiments. At the Maimai catchment in New Zealand, Graham et al. (2010) demonstrated preferential lateral flow at the soil–bedrock interface, with percolation from the bedrock contributing significantly to the water balance. At the Fudoji site in Japan, Uchida et al. (2003, 2006) concluded that up to 95% of streamflow can be contributed by bedrock groundwater. Tromp-van Meerveld et al. (2007) demonstrated experimentally the influence of leakage to bedrock on the subsurface stormflow response and overall water balance using a sprinkler setup at the Panola Mountain Research Watershed in Georgia.

The field evidence is complemented by a handful of modeling studies that have examined leakage in a hillslope context. For Coos Bay, Ebel et al. (2008) concluded that consideration of bedrock fracture flow would probably improve the simulation of pore water pressure distribution. For the Panola watershed, Tromp-van Meerveld and Weiler (2008) have shown that bedrock leakage was necessary to simulate adequately the long-term subsurface flow response at the small watershed or hillslope scale. Koussis et al. (1998) included a leakage term in a version of Boussinesq's equation that was linearized and converted to a form amenable to Muskingum–Cunge solution techniques. Broda et al. (2008) modified the hillslope-storage Boussinesq model (Troch et al., 2003) to include a leakage term and explored the sensitivity of this model to a range of constant and variable leakage rates. Cloke et al. (2003) found that, in modeling hillslope-river interactions, including recharge to the riparian zone from the bedrock aquifer (with flow towards this aquifer originating at the foot of the hillslope) improved simulated heads significantly. Hopp and McDonnell (2009) ran synthetic three-dimensional simulation experiments based on the Panola catchment and confirmed that bedrock permeability can play a key role in subsurface flow response. For a sloping two-layer system, Ahuja and Ross (1983) assessed analytically the effects on hillslope flow of a constant leakage towards the underlying base material.

These studies underline the importance of considering leakage in hillslope groundwater hydrology and the need to improve our understanding of the factors that influence this flow process. A detailed investigation of leakage is important to the continuing evolution of simple, parsimonious models to simulate flow and transport processes at the hillslope and watershed scales. The objective of this paper is to explore, via detailed numerical simulations, how leakage behaves in complex hillslopes and how it is affected by a variety of factors. Leakage is simulated with a three-dimensional finite element Richards equation model of synthetic sloping aquifer systems composed of an unconfined and a confined aquifer separated by an aquitard. The simulations examine different configurations of aquifer and aquitard properties, hillslope geometry, and boundary conditions, as well as the interplay between leakage, water levels, and outflow. The analysis is intended to provide insight into conditions under which leakage becomes an important component of a hillslope's dynamics and water budget.

Richards equation, in its three-dimensional form and with a storage coefficient that accounts for both the saturated and unsaturated zones, is commonly regarded as the most realistic representation of flow in a 'classical' porous medium (disregarding complexities such as macropores or air phase effects that are anyhow not considered in this study). The finite element solver used allows consideration of heterogeneity, and, in particular for our study, of a layered aquifer system. The Richards equation numerical model is thus an ideal tool for investigating the importance of leakage phenomena in the context of hillslope modeling, where non-leaky conceptualizations are very commonly used.

2. Methodology

The model used in this study is the three-dimensional finite element subsurface flow module (Paniconi and Putti, 1994) of the coupled catchment hydrological model described in Camporese et al. (2010). Only a brief description is provided here. The model solves the three-dimensional Richards equation:

$$\eta(\psi)\frac{\partial\psi}{\partial t} = \nabla\cdot(K_s K_r(\psi)(\nabla\psi + e_z)) \tag{1}$$

where $\eta = S_w S_s + \theta_s(\partial S_w/\partial\psi)$ [L^{-1}] is the general storage term, S_w [–] is the water saturation, S_s [L^{-1}] is the aquifer specific storage coefficient, θ_s [–] is the saturated moisture content, ψ [L] is the pressure head, K_s [LT^{-1}] is the saturated hydraulic conductivity, and $K_r(\psi)$ [–] is the relative hydraulic conductivity and e_z [–] is the vector $(0, 0, 1)^T$. In this study, the Brooks and Corey (1964) relationships were used for the saturation–pressure and conductivity–pressure relationships:

$$\begin{aligned} S_e(\psi) &= (\psi_c/\psi)^{\beta} && \psi < \psi_c \\ S_e(\psi) &= 1 && \psi \geqslant \psi_c \end{aligned} \tag{2}$$

$$\begin{aligned} K_r(\psi) &= (\psi_c/\psi)^{2+3\beta} && \psi < \psi_c \\ K_r(\psi) &= 1 && \psi \geqslant \psi_c \end{aligned} \tag{3}$$

where the effective saturation S_e [–] is defined as $S_e = (\theta - \theta_r)/(\theta_s - \theta_r) = (S_w\theta_s - \theta_r)/(\theta_s - \theta_r)$ and where θ [–], θ_r [–] and θ_s [–] are the volumetric, residual, and saturated moisture contents, respectively. β [–] is a constant representing the pore size distribution index and ψ_c [L] is the capillary fringe height.

The basic configuration for the layered hillslope aquifer simulated in this work is shown in Fig. 1. The total soil or sediment thickness of 10 m consists of an unconfined aquifer of 2 m thickness, an aquitard of uniform thickness between 0.2 m and 0.8 m (for different test cases), and a confined aquifer of uniform thickness between 7.8 m and 7.2 m. Uniform, convergent, and divergent plan shapes are simulated, as well as three different slope angles α representing gentle (0.2%), moderate (5%), and steep (30%) hillslope inclinations.

The 10 m total thickness is discretized into 20 layers and the surface into 201 (x direction, along the length of the hillslope) by 7 (y direction, along the width) nodes, producing a numerical grid of $201 \times 7 \times 21$ = 29,547 node points. The layer thicknesses increase from 0.1 m at the top to 3 m at the bottom. Each layer is homogeneous and isotropic. The hillslope crest, the lateral divides, and the bottom of the discretized domain are set as no-flow boundaries. The outflow face of the hillslope is modeled with two types of boundary conditions: Dirichlet, accounting for instance for seasonal variations in river stage level; and seepage face, representing more rapid variations in water table levels or the response to more intense rain events where the saturated zone intersects the outflow face above the river stage level. In both cases the boundary condition is applied over different extensions of the aquitard and aquifer faces.

The base case used as a reference point for all subsequent test cases has a moderate slope (5%), a hydraulic conductivity of 10^{-5} m s^{-1} for both the unconfined and confined aquifers, a conductivity of 10^{-7} m s^{-1} for the aquitard, and thicknesses of 0.6 m