ASSESSING THE TIMING AND MAGNITUDE OF PRECIPITATION-INDUCED SEEPAGE INTO TUNNELS BORED THROUGH FRACTURED ROCK

by

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ABSTRACT

Seepage into tunnels bored through fractured rock is a common occurrence that can cause significant problems for the construction process, tunnel longevity, and the regional hydrogeology. Predictions of seepage using analytical solutions are often inaccurate due to the inherent assumptions and volumetric averaging of fractures. A conceptual model is first developed for this research by using the factors shown by previous studies to have control on net infiltration and seepage. An integrated hydrologic model, ParFlow is then used to investigate the control exhibited by these factors that include: climatic forcing; vegetation; soil type and depth; bedrock type; fracture spacing; and tunnel depth on the timing and magnitude of seepage into tunnels. A fracture continuuum is generated for bedrock using FRACK, which maps discrete fracture networks to a finite difference grid with heterogeneous, anisotropic permeability fields. Simulations are run using hourly meteorological forcing over a two-year period. Surface and subsurface properties are varied individually to investigate the change in seepage response. Results show that fracture spacing, bedrock type, and overburden are particularly important pieces in obtaining reliable seepage estimates. Higher fracture spacing causes higher total seepage at a more constant rate than a lower spacing that exhibits a much larger range of fluctuation in seepage volumes. More permeable and porous bedrock actually increases lag times and seepage amounts that are observed to be relatively constant over time. Thicker and less conductive soils both increase lag times and reduce seepage magnitude.
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CHAPTER 1
INTRODUCTION

Tunneling dates back to prehistoric times due to its wide range of applicability in mining, water conveyance, sewers, transportation, and other purposes. The expansion of urban areas and population creates the challenge of building adequate infrastructure to maintain standards of living. As tunnels serve many purposes, can be excavated below existing infrastructure, and generate routes otherwise unattainable, they are an attractive method to cope with population growth, urbanization, and commerce.

Designing tunnels requires creative applications of geotechnical engineering to predict the responses of soil, rock, and water. Whereas the responses of soil and rock to tunneling are well understood, the way in which water flows is difficult to predict. Groundwater inflows to tunnels can pose significant problems for excavation speed, project cost, environmental damage, and human health risk, and can determine whether a project succeeds or fails. Accurate estimates of these inflows are critical in determining how a tunnel is constructed, and whether it is drained, grouted, or lined.

Most of the existing literature on predicting seepage into tunnels focuses on those bored beneath the water table. For example, Heuer (1995; 2005; 2012) created a relatively robust field method by which the hydraulic conductivity statistics from packer tests are used to estimate possible amounts of seepage. Analytical solutions have also been derived for these tunnels in a homogeneous, isotropic, porous medium that make many assumptions to simplify a very complex system. There also exist a number of numerical models that simulate seepage in the saturated zone. Many of these models however, do not include the influence of the unsaturated zone, climatic forcing, vegetation, overburden, and fractures, which are all crucial pieces in obtaining reliable estimates.

The research presented here includes these crucial pieces and focuses on estimating how they affect the timing and magnitude of precipitation-induced seepage to unlined tunnels bored through fractured rock by using an integrated hydrologic model. This is accomplished by generating heterogeneous, anisotropic permeability fields that represent soils overlying a fracture network in bedrock, for use with the hydrologic model. The varied modeling parameters are chosen according to those determined in the literature to exhibit control, the software then obtained and compiled, a conceptual model developed, literature referenced for model inputs, simulations run, and results analyzed. Conclusions are both qualitative and quantitative and serve the tunneling, water resources, and environmental communities.
CHAPTER 2
BACKGROUND

Seepage into underground tunnels, cavities, and drifts gained increasing attention during the past few decades as a result of its potential impacts on the construction process (Chen & Tolon, 2012; Huang et al., 2011; Fernandez & Moon, 2010a; Li et al., 2009; Cesano et al. 2000; Kitterod et al., 2000), water resources and environment (Vincenzi et al., 2009; Gleeson et al., 2009; Gargini et al., 2008; Scanlon, et al., 2006; Kim et al., 2005; Seyfried, et al., 2005; Rademacher et al., 2003; Scanlon et al., 2002; Shimojima et al., 1993), and the quality of nuclear waste repositories (Reeves et al., 2008; Bagtzoglou & Cesano, 2007; Trautz & Wang, 2002; Liu & Bodvarsson, 2001; Philip & Knight, 1989). Various processes have been identified that govern the spatial and temporal variability of net infiltration, recharge, and seepage into soil and rock excavations and cavities. These include land-atmosphere interactions (Maxwell, 2010), soil and vegetation (Scanlon et al., 2006), subsurface heterogeneity (Kollet, 2009: Maxwell, 2010), and preferential flow in soil and fractures (Maxwell, 2010; Pruess, 1998; Shimojima et al., 1993). The first step in developing a model that accurately predicts the factors exerting significant control over seepage and inflows to tunnels, is understanding these processes.

2.1 Analytical Solutions

Intensive research into tunnel seepage began decades ago when the need to assess tunnel design, safety, and longevity arose. A large number of analytical solutions have since been derived to estimate inflows under a variety of groundwater scenarios and boundary conditions. The majority of these analytical solutions have significant simplifying assumptions.

Philip and Knight (1989) derived the analytical solutions for exclusion of unsaturated seepage into underground openings in homogeneous, isotropic, porous media to aid in the design of nuclear waste storage facilities and drains intercepting the unsaturated zone. In nuclear waste repositories it is important to quantify even the smallest amounts of seepage as it can induce migration of radionuclides to the water table (Finsterle et al, 2003). Philip and Knight (1989) claim that seepage velocity at depth is similar to the rate at which storage changes annually, equal to yearly precipitation minus evapotranspiration and runoff. Shallow tunnels, however, may be too close to the surface to dampen the effects of spatially and temporally variable precipitation and thus cannot be approximated with an average velocity. Philip and Knight (1989) and Youngs (2004) identify tunnel shape as an important feature of
exclusion, due to the capillary barrier effect, as a respectively wider tunnel is more likely to admit water than a narrower, more parabolic tunnel.

Perrochet (2007) presents the analytical solutions for a homogeneous aquifer of finite thickness that is radially infinite, and transient flow into a tunnel being drilled beneath the water table at variable speed through a heterogeneous formation. The formation however, is only heterogeneous in a layered sense and the equation does not account for recharge.

### 2.2 Empirical and Observational Methods

In addition to analytical solutions, empirical and observational methods have been utilized in tunnels bored in different fractured formations of igneous, metamorphic, and sedimentary rock in various climates. This extensive body of research returns a number of different conclusions about the controlling factors of net infiltration and seepage including:

- the control that structural heterogeneity of fractures and soils, and their orientation exhibits on net infiltration and seepage in fast and slow flow as a response to spatially and temporally variable precipitation events (Shimojima et al., 1993; Wang et al., 1999; Flint et al., 2001; Zhou et al., 2006; Gleeson et al., 2009; Maxwell, 2010; Dobson et al., 2012);
- resulting fingering, preferential flow, wetting patterns, and varying behavior at fracture intersections (Olofsson, 1994; Pruess, 1998; Zhou et al., 2006);
- more steeply dipping fractures carrying more water than those with a shallower dip (Wang et al., 1999; Cesano et al., 2000; Gleeson et al., 2009);
- a lag time between precipitation and seepage that is negatively correlated to the magnitude of the event and positively correlated to the cavity depth (Shimojima et al., 1993; Rademacher et al., 2003; Dobson et al., 2012);
- lateral flow which may occur and impact seepage from as far away as one half to three times the cavity depth (Wang et al., 1999; Flint et al., 2001; Rademacher et al., 2003);
- a complex response controlled by climate and wetting front conditions which varies the hydrologic properties of soil and fractured rock with time (Shimojima et al., 1993; Rademacher et al., 2003; Dobson et al., 2012);
- support of the capillary barrier model causing accumulation of water at the tunnel ceiling before seepage (Wang et al., 1999; Trautz & Wang, 2002; Zhou et al., 2006);
- the ability of the formation to laterally transmit water significantly affecting the capillary barrier effect of the tunnel (Wang et al., 1999; Trautz & Wang, 2002);
the observation that lower permeability and porosity of matrix, and lower capillarity all increase seepage (Wang et al., 1999; Trautz & Wang, 2002).

Of these observations, the underlying theme is that infiltration and subsequent seepage most commonly occur where a fracture network acting as a flow conduit is in connection with an overlying source of water, be it a groundwater reservoir in permeable soil, a horizon of permeable conductive soil, or a location of direct recharge (Shimojima et al., 1993; Olofsson, 1994; Rademacher et al., 2003; Seyfried et al., 2005; Scanlon et al., 2005; Scanlon et al., 2006; Gargini et al., 2008; Vincenzi et al., 2009; Gleeson et al., 2009; Moon & Jeong, 2011). Tunnels bored in the unsaturated zone or through fractured crystalline rock are more commonly left without liners than tunnels that are beneath the water table and so are more susceptible to seepage (Vincenzi et al., 2009). The overburden was found to be the source of major leakage in some scenarios and for this reason, successful estimates must treat the fractured rock mass and overburden as an interrelated system (Cesano et al., 2000). This research treats the entire system as interrelated through the use of an integrated hydrologic model, and investigates the control that the above observations have on the timing and magnitude of seepage induced by precipitation into a tunnel; an analysis not yet documented in the literature.

2.3 Land-Atmosphere

The transfer of energy and water from land to atmosphere and vice versa are the primary means by which the land and atmosphere communicate. Exchange of these fluxes is dependent on the spatial and temporal variations of both atmospheric forcing and soil moisture. Characterization of water and energy fluxes between the atmosphere and subsurface is crucial to quantify for an accurate model in estimating seepage.

Maxwell and Kollet (2008a), Kollet and Maxwell (2008), and Rihani et al. (2010) show that groundwater dynamics have a significant effect on the energy balance at the land surface. Lateral flow between topographic highs and lows strongly affects groundwater depth and the vulnerability of an area to changing precipitation and climate. By including feedbacks from the ground surface to the atmosphere, they are able to characterize water table depth, recharge, and soil moisture response. At a water table depth between two to seven meters, they find a Critical Zone in which there is a strong correlation between recharge and latent heat flux (evapotranspiration). Land-surface and subsurface processes are most tightly coupled when this is the case. They also find that in deep unsaturated zones, the surface becomes disconnected to any groundwater response and the land-surface becomes equilibrated with the atmosphere. Heterogeneity is also an important feature to capture because it controls the soil
moisture distribution and the water available for evapotranspiration; this is further complicated by depth to bedrock and fractures.

Maxwell (2010) furthered these studies to show how land-atmosphere interactions, most importantly evapotranspiration and latent heat flux, are significantly affected by soil moisture distribution. As stated above, soil moisture distribution is notably influenced by heterogeneity in hydraulic conductivity and in turn affects the water and energy fluxes. These fluxes become more complicated in arid and semiarid environments where precipitation may be intense and episodic, soil moisture is generally minimal, and potential recharge is low. He also showed for his scenario that ‘wet’ years contribute drastically more to recharge, at a level of 33% of precipitation, than ‘dry’ years, which contribute 10% of precipitation. A strong nonlinearity is exhibited, which shows the importance of above-average precipitation and its contribution to recharge, which will affect seepage. He also found that overland flow was not generated, implying that all precipitation causes a change in subsurface storage, is evapotranspired, or is stored in shallow soil and is re-transpired. For this reason, modeling in this study will neglect the influence of overland flow into or out of the domain.

2.4 Land-Surface

It is clear that soil type and vegetation impact the mechanisms via which the land-surface communicates with the atmosphere and subsurface. Much research has been conducted studying the effects of these land-surface variables on recharge in water-limited areas, namely those with arid and semiarid climates, where the vadose zone may extend to considerable depths. The mechanisms exhibiting the strongest control on recharge are highlighted below.

Soil type and heterogeneity significantly affect rates of recharge and seepage. Coarse-grained soils usually provide more recharge than fine-grained soils (Olofsson, 1994; Scanlon et al., 2002; McCulley et al., 2004). Studies in Australia for arid and semiarid climates also indicate a negative correlation between percent clay and drainage (Wohling et al., 2011). Atchley and Maxwell (2011) show that subsurface patterns of hydraulic conductivity govern the distribution of soil-moisture at the hillslope scale. They illustrate that heterogeneity in soil can decrease infiltration due to ponding of water in low conductivity regions, increasing surface run-off and keeping more water available near-surface for evapotranspiration. Their results suggest that soil moisture variation after precipitation is initially determined by patterns in vertical hydraulic conductivity. At later times during dry down, while the water is being evapotranspired, soil moisture variation is controlled by vegetation. Maxwell and Kollet (2008b) and Meyerhoff and Maxwell (2011) also show that increasing subsurface heterogeneity affects the generation of overland flow due to shallow perched water tables.
Ecosystems also play an important role in water-limited systems as the types of plants present significantly affect the ways in which the land surface and atmosphere biophysically interact and thus exert control on overland flow and recharge (Newman et al., 2006; Scanlon et al., 2006; Sandvig & Phillips, 2006). Therefore, it is critical for a robust recharge or seepage model to include vegetation, its location, and how it might affect overland flow (Scanlon et al., 2002; Scanlon et al., 2005; Newman et al., 2006). Rates of recharge below non-vegetated areas tend to be much higher, as much as 50% of precipitation (Sandvig & Phillips, 2006), than those beneath vegetated areas. This is due to the inability of evaporation to remove water that infiltrates deep beneath the upper surface layer (Scanlon et al., 2002; Scanlon et al., 2005; Seyfried et al., 2005; Scanlon et al., 2006).

Not only does the presence of vegetation affect rates of recharge, but the type of vegetation also has a large impact. Recharge is most often greater where annual crops are grown instead of shrubs and trees due to the shallower rooting depth exhibited by crops such as grass (Scanlon et al., 2002; Scanlon et al., 2005; Seyfried et al., 2005; Sandvig & Phillips, 2006). More specifically, soil moisture fluxes are seen to increase across the same region when transitioning from creosote (which shows practically no infiltration) and xeric vegetation (shrubs) to grass and on to juniper (Scanlon et al., 2005; Sandvig & Phillips, 2006). This is because fluxes at the base of the root zone control the return of moisture to vadose zones and water tables. Deep root zones enable the plant to maintain negative soil tension and withdraw water flowing through preferential paths back to the matrix (Sandvig & Phillips, 2006). Since vegetation is such an important component of recharge, appropriate scales in space and time are necessary to quantify recharge that is governed by the availability of water at the surface, of which climate and geomorphology are in strong control (Scanlon et al., 2002; Newman et al., 2006).

Some studies have also shown that topography can have a significant effect on recharge. In arid climates recharge is often focused on topographic lows such as ephemeral streams (Pruess, 1998; Scanlon et al., 2002; Sandvig & Phillips, 2006). However, other studies have shown that soil properties and heterogeneity in arid and semiarid regions exert more control than topography (Famiglietti et al., 1998; Atchley & Maxwell, 2011).

Recharge rates in worldwide climates show that land use and cover exert significant control on net infiltration and that the majority of recharge occurs along channels in fractured mountain blocks where high precipitation, thin soils, and high permeability of underlying fractures exists (Shimojima et al., 1993; Rademacher et al., 2003; Scanlon et al., 2005; Scanlon et al., 2006; Gleeson et al., 2009). Seyfried et al. (2005) also suggest that regions where deep drainage may occur are associated with xeric shrubs on soil less than 1.5m thick overlies fractured bedrock, and that recharge in semiarid and arid climates can exhibit large lag times compared to those in humid and mesic environments.
2.5 Localized Flow in Soils

Evidence from a number of studies suggests the presence of preferential flow paths in both unsaturated and saturated soil by means of macropores, variance in hydraulic conductivity, and root channels. A few different features of preferential flow have been identified as indicators; Liu et al. (2005) highlight the mechanisms by which preferential flow in soil can occur. These mechanisms are fast flow along macropores for structured soils, and fingering that can occur for relatively homogeneous, structureless or spatially heterogeneous soils due to the non-linearity of infiltration. Their studies show that dye penetration of soil exhibits typical power-law relationships characteristic of fractal behavior.

Atchley and Maxwell (2011) simulated preferential flow due to heterogeneity in soil hydraulic conductivity and soil moisture content. McCulley et al. (2004) propose that when transpiration is low and there is a large gradient difference between soil layers, preferential flow along channels of deeply rooted plants could be a reason for extra soil moisture at depth. They claim that two significant controls are the intermittency of precipitation and the grain size of the soil. It has also been suggested that the seasonal change in density of woodland roots shuts down some preferential flow by changing the temporal hydraulic conductivity of soils and fractures (Dobson et al., 2012). Scanlan et al. (2006) emphasize the importance of including preferential flow in water resources recharge estimates.

2.6 Localized Flow in Fractures

Pruess (1998) provides substantial evidence from literature and his own personal communication that is indicative of preferential flow through fractured unsaturated media from a number of field sites. At Yucca Mountain when using an averaged hydraulic conductivity, seepage is predicted to take thousands of years to arrive underground. However, this seepage was observed to occur in only decades with only a few fractures carrying water. He reports on a mine in Arizona at several hundred meters depth in partially saturated rock with seepage in response to precipitation that exhibited a lag time of a few months. In Israel, tracers were observed to migrate rapidly in unsaturated fractured chalk (Nativ et al., 1995).

Shimojima et al. (1993) used geochemical tracers through fractured sandstone and chert in Japan and observed non-uniform spatial discharge in response to precipitation with fast and slow flow components. They propose that hydraulic properties such as soil moisture and relative hydraulic conductivity are not stable with time and a substantial change might be caused by heavy rainfall. Rademacher et al. (2003) also used geochemical tracers through tilted sedimentary rocks and observed a lag time of a few months between precipitation and seepage, to which 20 percent of the water flowing in a water resources tunnel was attributed.
Wang et al. (1999) used dyes to study the capillary effects of a tunnel. Flow paths of the dyes corresponded mainly to a few almost vertical fractures of a network. Zhou et al. (2006) noticed a positive correlation between fracture density and infiltration, but found a lack of correlation between normalized seepage and density due to water diversion caused by preferential flow and by the capillary barrier. Maxwell (2010) observed significant differences in recharge when comparing a heterogeneous fractured model to a homogeneous rock in which the majority of saturation was observed in fractures with very little saturation in matrix material. More recently, Dobson et al. (2012) studied seepage into a mine in Chihuahua, Mexico and also found fast and slow arrival times indicative of preferential flow.

Examples of seepage into transportation and water resources tunnels via localized flow in fractures are also found in literature from Europe. Vincenzi et al. (2009) showed that a 15 km long tunnel for high speed rail bored in fractured turbidite marls and sandstones, once considered aquitards, drained such large volumes of water that streams at the surface ran dry. This significantly affected the hydrologic and ecologic systems as very little groundwater remained at the surface for vegetation, streams, and lakes. Five major inflow events were observed while tunneling, all of which were related to extensional tectonic fractures, but only two of these extensional features were identified while surveying. Gargini et al. (2008) showed through the use of tracer tests that after initial drainage, 85 percent of the discharge into the same tunnel was attributed to direct recharge from precipitation. Preferential flow through the fractures was indicated by a large variance in dispersivity of tracers.

Olofsson (1994) and Cesano et al. (2000) also observed flow to an 80 km long water resources tunnel through crystalline rock in Sweden within extensional fractures. Many of these tectonic features were located in topographically low areas that had been filled with glaciofluvial deposits so that much of the drainage and drawdown occurred there. Olofsson (1994) found preferential flow and a complex drainage pattern indicated by variability in chemistry, noting that gently dipping fractures served as connections to the steeper dipping networks. He claimed that the heterogeneity in overburden and fracture networks must be statistically quantified. Cesano et al. (2000) then furthered this study by performing a systematic statistical analysis to identify which of the parameters were most important in regulating inflows. They found minor leakage into tunnels to be associated with drainage of the rock mass, and major leakage to be associated with the overburden above bedrock. This is because the rock mass generally has low porosity and the majority of drainage travels through fractures in contact with overburden. They also found deeper tunnels to be more susceptible to major inflows due to the increased likelihood of encountering an extensional fracture. These studies all highlight the need to incorporate the complexities of fracture flow and the associated heterogeneity into three-dimensional models.
ParFlow, the Common Land Model (CLM), and FRACK were used for modeling. The governing equations used within each model are listed in the following sections. For more information, refer to the literature.

### 3.1 ParFlow

ParFlow is a coupled, fully integrated, physical hydrologic model that predicts three-dimensional variably saturated subsurface flow and two-dimensional shallow overland flow. It discretizes in time using an implicit backward Euler difference scheme and in space using a cell-centered finite difference grid, and solves using a Newton-Krylov-multigrid preconditioned solver (Ashby & Falgout, 1996; Jones & Woodward, 2001). It is designed to run for efficient computation on massively parallel computer resources (Kollet et al., 2010; Maxwell, 2012). A brief description of the governing equations is provided below as these have already been described extensively in the literature. For more detailed explanations of ParFlow, refer to Ashby and Falgout (1996), and Kollet and Maxwell (2006).

Three-dimensional variably saturated subsurface flow is solved using Richards’ (1931) equation:

\[
\Delta S_s S_w \left( \frac{\delta \psi}{\delta t} \right) + \phi \left( \frac{\delta S_w(\psi_p)}{\delta t} \right) = \nabla \cdot \mathbf{q} + q_s
\]

in which \( \mathbf{q} \) is given by:

\[
\mathbf{q} = -K_s(x)k_r(\psi_p)\nabla(\psi_p - z)
\]

and \( S_s \) is the specific storage coefficient \( [L^{-1}] \), \( S_w \) is the degree of saturation, \( \psi_p \) is pressure head \( [L] \), \( t \) is time \([T]\), \( \phi \) is the porosity \([-] \), \( q_s \) is a source sink term \([T^{-1}]\), \( K_s(x) \) is saturated hydraulic conductivity \([LT^{-1}]\), \( k_r \) is relative permeability \([-] \) and is a function of pressure head, \( \psi_p \), given by the van Genuchten (1980) relationships where:

\[
k_r(\psi_p) = \frac{\left(1-(\alpha \psi_p)^n\right)^{1-m} \left(1+(\alpha \psi_p)^n\right)^m}{\left(1+(\alpha \psi_p)^n\right)^m}
\]
in which \( \alpha \,[\text{L}^{-1}] \) is inversely related to the air-entry pressure of the medium, \( n = (1 - m)^{-1} \,[-] \) and is related to the distribution of pore size; these values are determined empirically.

Soil moisture is also a function of pressure head, \( \psi_p \), and is calculated with:

\[
S_w(\psi_p) = \frac{S_{\text{sat}} - S_{\text{res}}}{\left(1 + (\alpha \psi_p)^m \right)^m} + S_{\text{res}}
\]  

(3.4)

in which \( S_{\text{sat}} \) is relative saturated water content \([-]\) and \( S_{\text{res}} \) is the relative residual saturation.

Two-dimensional shallow overland flow is integrated with subsurface flow by solving the kinematic wave approximation (can use diffusive and dynamic wave approximations also). This is then input to the overland flow boundary condition while preserving continuity conditions of pressure and flux at the land-surface boundary:

\[
q = -K_s(x)k_f(\psi_p)\nabla (h_s - z) = \left( \frac{\delta \parallel \psi_s \parallel}{\delta t} \right) - \nabla \cdot (\nabla \parallel \psi_s \parallel) + q_r(x)
\]

(3.5)

where \( \psi_s \,[\text{L}] \) is the surface ponding depth and is assumed to be \( \psi_p \,[\text{L}] \) at the saturated ground surface \( \) (Kollet & Maxwell, 2006), \( v \) is depth-averaged surface water velocity vector \([\text{LT}^{-1}]\), and \( q_r \) is a source/sink term \([\text{LT}^{-1}]\) (e.g. precipitation). This assumes that:

\[
S_{f,i} = S_{o,i}
\]

(3.6)

in which \( S_{o,i} \) is the bed slope (gravity forcing), and is equal to \( S_{f,i} \), the friction slope, and so Manning’s equation can be used to approximate flow with a depth-discharge relationship:

\[
v = \frac{S_{f,i}^{1/2} \psi_s^{2/3}}{\eta}
\]

(3.7)

in which \( \eta \,[\text{LT}^{-1/3}] \) is Manning’s coefficient (Kollet & Maxwell, 2006).
3.2 Common Land Model (CLM)

ParFlow is coupled to the Common Land Model (CLM) (Dai et al., 2003; Maxwell & Miller, 2005; Kollet & Maxwell, 2008), a land surface model that requires inputs including land-surface characteristics, soil and vegetation properties, atmospheric forcing, and geographic coordinates. The land-surface is characterized with land cover type, soil texture, and soil color. Vegetation inputs include morphological, physiological, and optical properties. Atmospheric forcing uses a time series of solar radiation, temperature, pressure, wind, precipitation rate, and humidity (Dai et al., 2003). CLM is coupled to ParFlow across the ten soil layers in CLM, with the top cell in ParFlow acting as the first cell beneath the ground surface of CLM. The models communicate over the root zone in these ten layers where soil moisture is calculated by ParFlow and is passed to CLM, which in turn simulates precipitation, evapotranspiration, and root flux. With a photosynthesis-based approach that incorporates the vegetation inputs, transpiration is calculated. Bare soil evaporation is calculated at the land surface. These fluxes are then passed back to ParFlow and treated as water fluxes into or out of the model (Maxwell & Miller, 2005). At each time step, mass and energy balance error are calculated to ensure mass and energy conservation. A brief description of some of the governing equations is once again provided (Kollet & Maxwell, 2008). More detailed explanations of the coupling can be found in the above citations.

Mass and energy balance can be calculated at the surface with:

\[ R_n(\theta) = H(\theta) + LE(\theta) + G(\theta) \]  

(3.8)

in which \( R_n \) is net radiation [Wm\(^{-2}\)], H is sensible heat [Wm\(^{-2}\)], LE is latent heat [Wm\(^{-2}\)], and G is ground heat flux [Wm\(^{-2}\)], all of which are functions of soil moisture \( \theta \). H takes into account sensible heat fluxes from vegetation and the ground. L is latent heat of evaporation and E is the sum of vegetative and terrestrial evaporation. The ground heat flux G is calculated with the heat conduction equation. Net radiation is calculated by:

\[ R_n = S_{n,c} + S_{n,g} + L_d - L_u \]  

(3.9)

in which \( S_{n,c} \) is solar radiation absorbed by the canopy [Wm\(^{-2}\)], \( S_{n,g} \) is solar radiation absorbed by the ground [Wm\(^{-2}\)], \( L_d \) is incoming atmospheric long wave radiation [Wm\(^{-2}\)], and \( L_u \) is outgoing long wave radiation [Wm\(^{-2}\)].

A source/sink term is expressed close to the land surface as:
\[ q_s = LE(\theta) + q_g(\theta) \]  

(3.10)

where \( q_s \) is flux of infiltrating water at the land surface.

### 3.3 FRACK

Fractures are generated using FRACK (Reeves et al., 2008), software that maps discrete fractures to a regularly spaced finite-difference grid. Using a grid enables the solutions of matrix and fracture flow when input to ParFlow by assigning contrasting permeability values between grid cells containing fractures and those without. Random fractures are generated in FRACK using fracture density, height to length ratio, and probability distributions of strike and dip, length, and aperture. Anisotropic hydraulic conductivity is then calculated in each grid cell according to the degree of fracture occupation. A brief description of the governing equations for the generation of a fracture network is provided (Botros et al., 2008).

Probability distribution of the strike and dip of each fracture is calculated using a fisher distribution as follows:

\[ f(\phi) = \frac{\kappa \sin(\phi)e^{\kappa \cos(\phi)}}{e^{\kappa} - e^{-\kappa}} \]  

in which \( \phi \) [-] is deviation from mean, and \( \kappa \) is the dispersion factor. A smaller value of \( \kappa \) implies a wider distribution around the mean value.

Length is determined with:

\[ f(l, l_m, a) = \frac{(a-1)l_m^{a-1}}{l^a} \]  

(3.12)

in which \( l \geq l_m \), the minimum fracture length, and \( a \) is the power law exponent. When \( a \) is between one and three, both large and small fractures contribute to the network connectivity.

Hydraulic conductivity of fractures is then solved with:

\[ K = \frac{\rho gb^2}{12\mu} \]  

(3.13)

in which \( \rho \) is fluid density [\( \text{ML}^{-3} \)], \( g \) is gravity, \( b \) is hydraulic aperture (lognormal distribution), and \( \mu \) is dynamic viscosity [\( \text{ML}^{-1}T^{-1} \)].
CHAPTER 4
CONCEPTUAL MODEL

The conceptual model used here is adapted from Maxwell (2010) and is relatively straightforward with a control volume that captures vegetated overburden above fractured bedrock (Figure 4.1). Fluxes into and out of the domain are precipitation ($Q_{\text{PREcip}}$), and evapotranspiration ($Q_{\text{ET}}$). The fluxes are forced by the climatic inputs for CLM and for these simulations were obtained for Denver, CO, a continental semiarid climate.

The equation for flux into the tunnel then becomes:

$$Q_{\text{TUNNEL}} = Q_{\text{PREcip}} - Q_{\text{ET}} - \Delta S_{\text{SURFACE}} - \Delta S_{\text{SUBSURFACE}}$$  \hspace{1cm} (4.1)

In which $\Delta S_{\text{SURFACE}}$ and $\Delta S_{\text{SUBSURFACE}}$ are the change in storage of the surface and subsurface respectively.

Figure 4.1: Conceptual model of the domain and water balance. Not to scale. The lateral fluxes into and out of the domain are assumed to be negligible as the domain is of appropriate width to capture the lateral flow along fractures. Note that runoff will not be considered as a flux into or out of this domain either.
As the conceptual model being explored for this research is hypothetical and not site specific, the parameters used for modeling are based upon those described in the literature. The parameters that need to be considered are discussed in section two of this study, and are those, which appear to exhibit the most significant controls on net infiltration and subsequent seepage. Their use in the model is discussed through the following sections. Each simulation varied one parameter at a time from the base case, defined as thin, barren, highly conductive soils overlying a densely fractured crystalline bedrock with a tunnel at ten meters depth. Varying one parameter at a time from the base case simulation enables the control that parameter exhibits on seepage to be separated. A schematic of the various cases performed for this research is shown on Figure 5.1.

Figure 5.1: Seven simulations that were run for which the parameters are discussed in detail through section five. a) base case, b) different soil type, c) thicker overburden, d) vegetated, e) different bedrock, f) higher fracture spacing, g) deeper tunnel.
5.1 Domain Size

Discretization of the domains is shown in Table 5.1. The only reason there are differences between the domain sizes is that the lateral extent was increased for the simulation with a deeper tunnel. The lateral extent is varied because the quantity of water capable of reaching the tunnel is dependent on tunnel depth and fracture dip. As has been shown by Wang et al. (1999), Rademacher et al. (2003), and Flint et al. (2001) precipitation as far away laterally as three times the tunnel depth can cause seepage. This would most likely occur if the fracture dip was as low as 18 degrees; an orientation that is possible in these simulations, but serves mainly to connect steep fractures not cause major seepage. As the main fracture orientation for this research is specified as steeply dipping, the domain extends outward laterally from the center of the tunnel up to two times the depth to bedrock.

The division of cells into regularly spaced finite-difference grids of 0.5 m and 1 m is to maintain a computational domain that is efficient for ParFlow and FRACK, while still capturing and representing behavior of heterogeneous subsurface flow. It is understood that the cell size should be less than the fracture density (Reeves and Parashar, personal communication) and that this may skew the results of the simulation with a cell size of 1 m and a fracture density of less than a meter. However, exponentially more time is required to generate denser domains and run simulations able to cope with such high grid resolution even when they are performed in parallel. No slope is applied to the top of the domain as Maxwell (2010) showed all overland flow, in a very similar scenario, contributing to run-on, and therefore none leaves the domain.

Table 5.1: Domain size and configuration according to tunnel depth.

<table>
<thead>
<tr>
<th>Domain Size</th>
<th>10 m</th>
<th>25 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tunnel Depth</td>
<td>10 m</td>
<td>25 m</td>
</tr>
<tr>
<td>X (m)</td>
<td>40</td>
<td>100</td>
</tr>
<tr>
<td>Y (m)</td>
<td>40</td>
<td>100</td>
</tr>
<tr>
<td>Z (m)</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>dx=dy=dz (m)</td>
<td>0.5</td>
<td>1</td>
</tr>
<tr>
<td>nx</td>
<td>80</td>
<td>100</td>
</tr>
<tr>
<td>ny</td>
<td>80</td>
<td>100</td>
</tr>
<tr>
<td>nz</td>
<td>100</td>
<td>50</td>
</tr>
<tr>
<td>Total Cells</td>
<td>640,000</td>
<td>500,000</td>
</tr>
</tbody>
</table>
5.2 Climatic Forcing

Many of the observed inflows to tunnels through fractured crystalline rock occur in climates or geographic locations where there is a deep vadose zone in rock of good quality designation (RQD) that is considered relatively impermeable. These climates have been documented as semiarid (Pruess, 1998), and Mediterranean (Vincenzi et al., 2009; Gargini et al., 2008; Rademacher et al., 2003). Denver is a good example of a continental semiarid climate and is used for simulations. To define the bounds of climatic regions in which these conditions may exist, Ponce et al. (2000) use the aridity index, defined as the ratio of potential evaporation to precipitation. The bounding numbers of this ratio for an arid climate are between 12 and 5, for semiarid between 5 and 2, and for subhumid between 2 and 0.75. The Koppen-Geiger climate classification (Figure 5.2), used widely, draws similar distinctions between mean annual precipitation, its time of occurrence (winter or summer), and the mean annual air temperature (Peel et al., 2007) to divide climatic regions.

Forcing data including shortwave and longwave radiation, temperature, pressure, wind speed and direction, precipitation rate, and humidity were obtained from the North American Land Data Assimilation System (NLDAS) dataset for Denver, CO, a continental semiarid climate (BSk). Two total years of hourly meteorological forcing were downloaded and used in simulations: one ‘average’ water year (1995-1996), followed by a ‘wet’ El Nino Southern Oscillation (ENSO) year (1996-1997) that was one of the strongest in the past 60 years (NOAA, 2012). This was to investigate the effects of non-linear response shown by Maxwell (2010) on magnitude of seepage, and to determine the control exhibited by residual soil moisture that affects the timing and magnitude of seepage. Note that precipitation in Denver over the course of this ‘wet’ year only increased by approximately 3% compared to the previous year and by 10% in comparison with the mean, as is normal for that region during ENSO. Precipitation and temperature for the period of interest are shown on Figure 5.3, and compared to historical means in Table 5.2. Using an hourly time series for two years significantly increases simulation time, but maintains resolution and enables comparison to response times observed for varying magnitudes of precipitation by Shimojima et al. (1993) and Dobson et al. (2012).

Walvoord et al. (2004), Maxwell and Kollet (2008), and Kollet and Maxwell (2008) found that in semiarid climates, the sensitivity of near surface water to climate is high and deep response is slow. They and Walvoord et al. (2002) claim that it can take on the order of thousands of years for the subsurface to equilibrate with surface conditions. These systems therefore, don’t really need spin-up since they are under the influence of the last climatic period and so none was performed. As each case begins with the same initial conditions, a better comparison of fluxes and changes in storage across cases is possible.
Figure 5.2: Climatic regions of the western United States according to the updated Koppen-Geiger climate classification (Peel, Finlayson, & McMahon, 2007). Only those of importance to this study are described on the Legend.

Table 5.2: Temperature, precipitation, and the historical mean for each.

<table>
<thead>
<tr>
<th>Denver Climatic Forcing Information</th>
<th>Latitude</th>
<th>N395107</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Longitude</td>
<td>W1044039</td>
</tr>
<tr>
<td></td>
<td>Elevation (ft)</td>
<td>5280</td>
</tr>
<tr>
<td>Total Precipitation (mm)</td>
<td>Year 1</td>
<td>434</td>
</tr>
<tr>
<td></td>
<td>Year 2</td>
<td>447</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td>396</td>
</tr>
<tr>
<td>Temperature (°C)</td>
<td>Year 1</td>
<td>10.5</td>
</tr>
<tr>
<td></td>
<td>Year 2</td>
<td>10.2</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td>10.2</td>
</tr>
</tbody>
</table>
Figure 5.3: Precipitation and temperature values for Denver over two water years.
5.3 Overburden

The dependency of soil type on parent material, previous climate, and genetic processes makes it difficult to identify correlations between climate and soil type for the hypothetical domain. Scanlon et al. (2002), McCulley et al. (2004), and Wohling et al. (2011) all found coarse-grained soils to favor recharge over fine-grained soils. As the fine-grained content of the soil increases, the saturated hydraulic conductivity of the soil decreases. For this reason, the soils used for modeling are sand and sandy loam. These soil types and hydraulic conductivities are plotted on the United States Department of Agriculture (USDA) soil texture triangle (Figure 5.4).

![Soil texture and hydraulic conductivity for two soils. Hydraulic conductivity values are from Schaap and Leij (1998).](image)

Soil hydraulic conductivities and van Genuchten parameters of these soils for model inputs are taken from the values experimentally determined by Schaap and Leij (1998). The dependency on geologic
and climatic processes for the formation of soils also induces spatial variability of hydraulic conductivity and soil moisture that changes with scale (Seyfried M., 1998; Green et al., 2009). Seyfried (1998) claims that at some scale, spatial data for soil moisture is random and may be portrayed with mean and variance. Green et al. (2009) provide evidence that below 5 km scales, variability is statistically stationary and that at hillslope scales (< 200 m) infiltration may be higher than expected over larger scales.

Seyfried and Wilcox (1995) claim that this spatial variability at small scales can be represented with stochastic methods. For these reasons, the Turning Bands approach (Tompson, 1989) is used to simulate hydraulic conductivity fields. This approach is common for stochastic hydrogeology (Rubin, 2003) and represents the spatial variation of saturated hydraulic conductivity as a continuous stationary Gaussian random field in which:

$$\ln K_s(x) = F + f(x)$$  \hspace{1cm} (5.1)

where \( \{ln K_s(x)\} = F, \{f(x)\} = 0, \{f(x)^2\} = \sigma_F^2 = \sigma_{\ln k}^2 \), and \( K_g = \exp(F) \). Here \((x)\) implies the mean of \(x\), \( K_s \) is the mean hydraulic conductivity, and \( \sigma^2 \) is its variance. A correlation structure, which affects the degree of continuity in hydraulic conductivity, must also be specified. The correlation structure is given by:

$$R_{ff}(\xi) = \sigma_F^2 \exp \left( -\frac{\xi}{\lambda} \right)$$ \hspace{1cm} (5.2)

where \( \xi \) [L] is the distance separating two hydraulic conductivity values, \( \lambda \) is the correlation length [L]. The values used for modeling are those suggested by Rubin (2003) for alluvial soil where \( \lambda_x = \lambda_y = 15 \text{ m}, \lambda_z = 2 \text{ m} \) (Maxwell, 2010). Rubin (2003) suggests using \( \sigma_F^2 = 0.81 \), but for these simulations variance is increased to 1 to account for the lack of topography. Numerical turning bands parameters were specified with the number of lines equal to 360 used with the grid spacing of 0.5 m. Normalized frequency increment of 0.2 was used with a \( K_{\text{max}} \) of 100. The principal direction of heterogeneity follows the slope of the land surface. These values were also used in the application of turning bands to bedrock. Soil thicknesses of 2 m and 15 m were applied in the model to account, respectively, for thin soils overlying fractured rock, and a potential reservoir in overburden overlying fractured rock. Soil properties are summarized in Table 5.3.
Table 5.3: Soil properties used in simulations.

<table>
<thead>
<tr>
<th>Soil Properties</th>
<th>USDA Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sand</td>
</tr>
<tr>
<td>Porosity</td>
<td>0.337</td>
</tr>
<tr>
<td>$\theta_r$</td>
<td>0.058</td>
</tr>
<tr>
<td>% Sand</td>
<td>92.7</td>
</tr>
<tr>
<td>% Clay</td>
<td>2.9</td>
</tr>
<tr>
<td>Ksat (m/h)</td>
<td>0.426</td>
</tr>
<tr>
<td>$\alpha$ (m$^1$)</td>
<td>4.79</td>
</tr>
<tr>
<td>n [-]</td>
<td>3.47</td>
</tr>
<tr>
<td>$\lambda_x = \lambda_y$ (m)</td>
<td>15</td>
</tr>
<tr>
<td>$\lambda_z$ (m)</td>
<td>2</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>1</td>
</tr>
</tbody>
</table>

5.4 Vegetation

Vegetation was input according to percent cover and type. Percent cover was shown by Seyfried and Wilcox (1995) to be at a maximum of 50 when soils were greater than 1.5 meters thick. When soils are thinner than that, vegetative cover is reduced and approaches zero. The International Geosphere-Biosphere Program (IGBP) standards are used within CLM. Barren land (IGBP #16) is run for most simulations, as recharge was seen to amount to as much as 50% of precipitation (Seyfried & Wilcox, 1995). Shrubland (IGBP #7) was used in a different run to investigate the effect of transitioning from barren soils to more vegetated soils.

5.5 Bedrock

As seepage to unlined tunnels has been shown to occur in fractured igneous, fractured metamorphic, and fractured sedimentary rocks, bedrock was specified accordingly. The values for hydraulic conductivity used were within the range for those listed in Domenico and Schwartz (1990) for unfractured igneous and metamorphic rock, and low permeability porous rock. Porosity values were taken from Heath (1983) for granite (corresponding to unfractured igneous or metamorphic rock), and sandstone and basalt (for low permeability porous rock). Turning bands was once again used to simulate heterogeneity in bedrock, the values for variance and correlation length are the same as those used by Maxwell (2010), and all other variables are the same as those specified above. The values used in modeling are shown in Table 5.4.
Table 5.4: Rock properties used in modeling.

<table>
<thead>
<tr>
<th>Rock Properties</th>
<th>Model Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Crystalline</td>
</tr>
<tr>
<td>Porosity</td>
<td>0.001</td>
</tr>
<tr>
<td>Specific Storage (m⁻¹)</td>
<td>0.058</td>
</tr>
<tr>
<td>Ksat (m/h)</td>
<td>9.00E⁻⁰⁹</td>
</tr>
<tr>
<td>α (m⁻¹)</td>
<td>4.79</td>
</tr>
<tr>
<td>n</td>
<td>3.47</td>
</tr>
<tr>
<td>λx = λy (m)</td>
<td>50</td>
</tr>
<tr>
<td>λz (m)</td>
<td>2</td>
</tr>
<tr>
<td>σ</td>
<td>1</td>
</tr>
</tbody>
</table>

5.6 Fractures

Modeling fractured rock is often done in a volume averaging fashion in an attempt to homogenize significantly heterogeneous matrix and fracture permeabilities (Pruess, 1998; Pruess et al., 1999). This also averages infiltration rates applied at the upper boundary and significantly underestimates wetting front arrival times as permeability of fractures is orders of magnitude greater than matrix. Due to the complexities in behavior, including film flow in fractures, it is very difficult to accurately model using this technique (Pruess et al., 1999). The most accurate techniques for modeling fracture networks are with Discrete Fracture Network (DFN) models, Stochastic Continuum (SC) models, and a combination of the two (Liu & Bodvarsson, 2001; Ohman & Niemi, 2003; Huang et al., 2011); all of which are becoming common approaches.

Here a stochastic continuum approach was taken, similar to that by Reeves et al. (2008), Botros et al. (2008), and Maxwell (2010), in which discrete fracture networks are generated and mapped to a finite difference grid. This creates a saturated hydraulic conductivity field that is heterogeneous and anisotropic, in which the fracture network is represented as a porous medium with hydraulic conductivity and other physical properties determined by random functions that are variable in space (Parashar & Reeves, 2008). In most simulations a fracture spacing of 0.8 m was specified, although one simulation had a spacing of 4 m to see how this affected seepage.

There is a wide range of fracture permeabilities dependent on the properties of the fracture such as asperity contacts, increasing aperture away from these contacts, roughness of fracture walls, and spatial correlation. Pruess et al. (1999) claim that turning bands can be used to capture the important features
listed above. It was implemented in the model by converting the permeabilities generated by FRACK to tensors according to mean hydraulic conductivity of matrix. This mean hydraulic conductivity was then varied using turning bands. The parameters for fractures in the stochastic continuum model are varied according to those used by Maxwell (2010) and are very similar to those used by Kitterod et al. (2000), Finsterle et al. (2003), and Birkholzer et al. (2004), shown in Table 5.5.

<table>
<thead>
<tr>
<th>Fracture Properties</th>
<th>Primary</th>
<th>Secondary</th>
</tr>
</thead>
<tbody>
<tr>
<td>Probability of Occurrence</td>
<td>0.66</td>
<td>0.34</td>
</tr>
<tr>
<td>Minimum Fracture Length (m)</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>Maximum Fracture Length (m)</td>
<td>10,000</td>
<td>10,000</td>
</tr>
<tr>
<td>A (Truncated Pareto)</td>
<td>1.5</td>
<td>1.5</td>
</tr>
<tr>
<td>Dip Mean (degrees)</td>
<td>120</td>
<td>30</td>
</tr>
<tr>
<td>Dip κ (Fisher Distribution)</td>
<td>10</td>
<td>40</td>
</tr>
<tr>
<td>Strike Mean (degrees)</td>
<td>135</td>
<td>135</td>
</tr>
<tr>
<td>Strike Kappa (Fisher Distribution)</td>
<td>20</td>
<td>10</td>
</tr>
<tr>
<td>Aperture Mean (microns)</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>Aperture Standard Deviation (lognormal)</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>α (van Genuchten) (m⁻¹)</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>n (van Genuchten) [-]</td>
<td>3</td>
<td>3</td>
</tr>
</tbody>
</table>

As many of the observed inflows in the literature are related to steeply dipping fractures, a primary mean dip of 120 degrees was specified. The secondary mean was input as 30 degrees. These values were chosen because they are commonly observed angles generated by extensional and compressional tectonics respectively (Girty, 2009). The main strike was specified as 135 degrees with a wide variance. This is to simulate some fractures that are perpendicular to the tunnel with others that run more parallel.

5.7 Tunnel

The tunnel was specified with a 6 m x 6 m cross-sectional area along the length of the domain. To account for the volume of a more circular or parabolic shape that tunnels are usually constructed with, porosity was specified as 0.785. This is equal to the ratio of a circle’s area inscribed in a square. The one-meter surrounding the tunnel in all directions had a permeability reduction that reduced the value to 65% of the initial. This was to be consistent with results found by Fernandez and Moon (2010) for
permeability reduction due to boring in stiff rock. Two different tunnel depths were investigated in modeling, one at 10 m and the other at 25 m.

5.8 Simulations

All simulations were run on the Colorado School of Mines supercomputer Ra. Each of the seven simulations was distributed across 32 processors and was given four days to complete. Depending on the computational difficulty of the simulations some solved after a couple of days while some did not complete during this time. Those that failed to complete initially were resubmitted with penultimate pressure output file for initial conditions and re-run. Some issues were also encountered with Ra during simulations that caused them to stall shortly after beginning. Total computation time for the simulations was approximately 500 hours.
CHAPTER 6
RESULTS AND DISCUSSION

Realizations of hydraulic conductivity are shown in Figures 6.1, 6.2, 6.3, and 6.4 for four of the seven simulations. These realizations show differing tunnel depth, matrix type, and fracture density. The effect of turning bands is evident in all four images. From this point forward, I will refer to the simulation with 2 m of barren sand overlying a 10 m deep tunnel bored through crystalline rock with 0.8 m fracture spacing, pictured in top left, as the ‘base case’. This is because all other simulations have the same specified parameters as the base case except for one. By varying one parameter on each simulation, the effects of that parameter alone are identifiable.

Figure 6.1: Base case hydraulic conductivity.
Figure 6.2: Larger domain and hydraulic conductivity realization for the 25 m deep tunnel.

Figure 6.3: Realization of hydraulic conductivity for the low permeability porous bedrock.
The total volumetric seepage per meter into tunnels for all seven simulations is shown below on Figure 6.5 and the volumetric seepage per hour of simulation is shown on Figure 6.6. Note that the noise on all subsequent plots is due to difficulty in solving. Lag time to seepage can be clearly seen on the plots below, particularly where measurable seepage begins for each curve on Figure 6.6. The difference in lag times between models is useful for general comparison and a quantitative analysis. The lack of spin up for the soil moisture, and the domain being mostly dry at the outset, is evident in that the time to lag is exaggerated. The first arrival, as predicted, occurs in the “base case” tunnel with barren, thin, highly conductive soils overlying densely fractured bedrock. This is then closely followed by the vegetated case. Interestingly the next arrival is at the deeper tunnel, exhibiting a lag time slightly shorter than a tunnel bored through lower fracture density. Seepage is next observed in the tunnel bored through a more porous and permeable medium. Changing from sand to sandy loam causes a significant increase in wetting front arrival time. The last appearance of moisture to tunnel occurs where there is a much thicker overburden.

Figures 6.5 and 6.6 also give insight into the magnitude of seepage and behavior that occurs between precipitation events. The last two months, after the largest precipitation event, were excluded from the plot for the base case and deeper tunnel because the no flow boundary condition at the base of
Figure 6.5: Precipitation and total seepage into tunnel for the different runs.
Figure 6.6: Volumetric seepage into tunnel per hour per meter for different simulations. Note this is a semi-log plot.
the domain caused the tunnel to fill from beneath and distorted the results. Even without this data, however, some clear distinctions can be seen between simulations. Sustained seepage and peak seepage are two serious concerns here and the differences between these are discussed in more detail over the next few sections that look at the influence of the different factors discussed in the literature.

6.1 Climatic Forcing

Magnitude of precipitation events clearly exerts control on the timing and magnitude of seepage to tunnels. Figure 6.7 shows a year of precipitation and seepage for the base case only. As observed by Dobson et al. (2012), Gleeson et al. (2009), Rademacher et al. (2003), and Shimojima et al. (1993), there is a complex response exhibited that is controlled by climate, temporally variable precipitation, and the subsequent wetting front conditions.

![Seepage into 'Base Case' Tunnel Over One Year](image)

Figure 6.7: Base case seepage in response to precipitation.

A clear seasonality is exhibited over this cycle in which the massive spring and summer precipitation events are causing peak seepage into the tunnel. These large precipitation events are surrounded by other moderate events that appear to be increasing the soil moisture content, increasing the hydraulic conductivity of soils and fractures, and thus reducing the lag time to seepage. The fact that
there are a number of relatively large events also explains the magnitude of seepage being much greater than is observed during the fall and winter months. The complexities of wetting front conditions are further highlighted by the precipitation occurring at the end of July and the end of August. The event at the end of July does not cause an increase in seepage, only slows the rate at which it is decreasing; however, the smaller event in at the end of August causes an increase. This is most likely because the larger event is preceded by a large storm that dwarfs its effects. After the large events have ended, during the fall and winter months, seepage returns to a more consistent rate and lower magnitude. The minor events are dampened out and return to more of a volume-averaged approximation.

Figure 6.8 takes a closer look at the lag and magnitude of induced seepage for the last three months of the above plot. It is quite interesting to note that the moderate precipitation occurring during the beginning of April causes a minor signal in the tunnel; only the event towards the end of the month of approximately 5 mm can be seen. This is most likely due to snowmelt events that are increasing the saturation of the soil that has become relatively dry during the winter, and thus increasing the hydraulic conductivity of the soil. Between the end of May and early June, more precipitation causes considerable seepage that can be seen to arrive 20 days after the first event.

![Figure 6.8](image)

Figure 6.8: Base case precipitation and seepage over three months.
There is clearly a strong nonlinearity exhibited between seepage and precipitation. This is consistent with the findings of Maxwell (2010) in which above-average precipitation contributes exponentially more to recharge than small amounts of precipitation. The seepage response for the base case to precipitation events of different magnitude is explored in more detail and summarized below in Table 6.1. These results are plotted below on Figures 6.9 and 6.10.

Table 6.1: Magnitude of precipitation and time to seepage for the base case simulation.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
<th>Time to Peak Seepage (days)</th>
<th>Magnitude of Peak Seepage (L/hr/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.7</td>
<td>10</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>9.9</td>
<td>6</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>17.3</td>
<td>4</td>
<td>17</td>
<td>0.36</td>
</tr>
<tr>
<td>22.2</td>
<td>2</td>
<td>10</td>
<td>0.43</td>
</tr>
<tr>
<td>23.7</td>
<td>1.75</td>
<td>7</td>
<td>0.89</td>
</tr>
<tr>
<td>23.8</td>
<td>2.9</td>
<td>16</td>
<td>0.28</td>
</tr>
<tr>
<td>25.8</td>
<td>0.5</td>
<td>7</td>
<td>0.93</td>
</tr>
<tr>
<td>29.1</td>
<td>1.5</td>
<td>9</td>
<td>1.03</td>
</tr>
<tr>
<td>47.1</td>
<td>1.5</td>
<td>7</td>
<td>1.28</td>
</tr>
</tbody>
</table>

Figure 6.9: Lag time to seepage for base case tunnel and precipitation event size. Note the line is plotted to show a general trend.
Figure 6.10: Magnitude of precipitation and the associated seepage. Again the line is plotted for general trend.

Table 6.1 and Figures 6.9 and 6.10 show some interesting behavior that is relatively consistent. There appears to be a precipitation threshold of approximately 10 mm per 48 hours, beneath which background seepage does not change and above which lag time to seepage decreases in a power-law fashion. The fastest arrival of seepage is seen to occur after 12 hours, which peaks seven days later at approximately 37 L/h. The magnitude of seepage is also dependent upon the precipitation threshold, beneath which background seepage stays relatively constant and above which seepage increases with precipitation in a non-linear fashion.

6.2 Overburden

Soil type exerts significant control on seepage. As shown below on Figure 6.11, lag time for a sandy loam is significantly higher than for sand while the volume of seepage is much smaller. The last portion of this plot shows the first significant arrival of seepage through the sandy loam. This is probably due to the lack of moisture in the sandy loam significantly reducing the hydraulic conductivity for the first year of simulation. After this first year, the soil becomes wet enough to permit the flow of water at a much higher rate. The subdued signal from the 29 mm/48 hr storm that occurred during April can be seen
to arrive at approximately 57 days after it is seen for sand. The other event observed, is that of 23.8 mm/48 hr that arrives with a lag of 20 days, 17 days after it is seen in the base case tunnel.

The findings are consistent with those observed by Wohling et al. (2011), McCulley et al. (2004), Scanlon et al. (2002), and Olofsson (1994) in which coarse-grained soils are providing more recharge than finer grained soils. Not only does the fine-grained content of soil decrease the rate of seepage, it also affects the volume. This is in part due to the reduction of hydraulic conductivity keeping water near the surface and available for evapotranspiration, similar to the findings of Atchley and Maxwell (2011). This observation is illustrated in more detail in Figure 6.12, which shows the total storage contained in soil, and in Figure 6.13, which shows evapotranspiration for both simulations.

Figure 6.11: Seepage into tunnel for sand and sandy loam. Note this is towards the end of the simulation when boundary conditions became a problem for the base case, because of this the last month for the base case is not shown on this plot.
Figure 6.12: Storage in overburden for sand and sandy loam.

Figure 6.13: Total precipitation and evapotranspiration for sand and sandy loam.
Time to seepage and magnitude of seepage in relation to precipitation are shown below in Table 6.2, but are not plotted. The significance of antecedent conditions is clearly shown in that the time to seepage arrival for the larger event is actually longer than the smaller. Again, little influence can be seen from the precipitation events beneath 10 mm/48 hrs.

Table 6.2: Time to seepage arrival for tunnel beneath sandy loam.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>29</td>
<td>58</td>
</tr>
<tr>
<td>23.8</td>
<td>20</td>
</tr>
</tbody>
</table>

Soil thickness also exerts control over the timing of seepage but less on the magnitude. This can be seen by comparing the total storage in soil and the percent of that which arrives in bedrock that is contained within the tunnel shown on Figures 6.14 and 6.15.

Figure 6.14: Total storage in 2 m and 15 m of sand.

Given the response seen after the first year of simulation, it is expected that the percent of bedrock storage within the tunnel for both soil types would become similar. This is due to the reservoir developing in the 15 m of sand as shown in Figure 6.14. The lag to seepage in this scenario is
significantly greater, the first appearance of seepage to the tunnel beneath 15 m of overburden being approximately 1 year after that with 2 m of overburden. This is most likely due to the domain being dry when simulations start.

![Percent of Bedrock Storage Within Tunnel for 2 m Sand and 15 m Sand](image)

**Figure 6.15:** Percent of bedrock storage in tunnel for 2 m and 15 m overburden.

### 6.3 Vegetation

The impact of vegetation on seepage is not nearly as pronounced as expected and is shown below on Figure 6.16. When changing from barren soils to shrubs, seepage is seen to decrease slightly. This decrease is consistent with that reported by Sandvig and Phillips (2006), Seyfried et al. (2005), and Scanlon et al. (2006), but is not nearly as apparent. This is most likely due to the fast drainage of sands and similar assigned properties such as rooting depth in the land-surface model CLM. The change in storage of bedrock (net infiltration) however, is approximately 50% of precipitation for the barren simulation and is consistent with that reported by Sandvig and Phillips (2006). This is also consistent with that reported by Atchley and Maxwell (2011) in which soil moisture after precipitation is initially controlled by vertical hydraulic conductivity and until later times, during drying periods, experiences little influence from vegetation.
The impact of vegetation can also be investigated through the amount of water transpired. Unexpected behavior is shown on Figure 6.17 in which the amount of transpiration is actually greater for the barren domain than for the shrubs. This means that the reduction in seepage magnitude is most likely due to interception.

A closer look is taken at the magnitude of precipitation and the corresponding seepage for the vegetated case in Table 6.3 and Figures 6.18 and 6.19. Once again there is a precipitation threshold of approximately 10 mm/48 hrs beneath which no increase in seepage is observed. As expected, the magnitude of peak seepage increases with increasing precipitation and peaks at approximately 1.225 L/hr/m seven days after the 47 mm event.

Table 6.3: Precipitation events, and seepage magnitude and timing for vegetated domain.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
<th>Time to Peak Seepage (days)</th>
<th>Magnitude of Peak Seepage (L/hr/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9.7</td>
<td>4</td>
<td>8</td>
<td>0.875</td>
</tr>
<tr>
<td>23.7</td>
<td>3</td>
<td>11</td>
<td>0.97</td>
</tr>
<tr>
<td>29.1</td>
<td>2.5</td>
<td>11</td>
<td>0.97</td>
</tr>
<tr>
<td>47.1</td>
<td>2</td>
<td>7</td>
<td>1.225</td>
</tr>
</tbody>
</table>
Figure 6.17: Transpiration for barren and vegetated simulations.

Figure 6.18: Lag time to seepage for vegetated simulation. Note the trendline is for visualization.
6.4 Bedrock

Tunneling through fractured igneous or metamorphic rock as opposed to fractured sandstone or basalt significantly increases the magnitude of seepage. This is shown on Figure 6.20 and is more clearly emphasized by Figure 6.21 in which the percent of subsurface storage contained within the tunnel is shown. Although there is a lot of noise in Figure 6.20, it appears that seepage in the low permeability porous simulation is occurring in a more volume averaged fashion and this is probably due to exchange of water between fractures and matrix of higher conductivity than the igneous/metamorphic simulation. Both simulations show a similar background seepage of approximately 0.1 L/hr/m, but the base case exhibits much more extreme variations in seepage in response to precipitation. Figure 6.21 really confirms the observations made by Trautz and Wang (2002), and Wang et al. (1999) that lower permeability, porosity, and capillarity of matrix all increase seepage to tunnels. It can be seen that storage in the tunnel for the base case simulation amounts to 22% of net infiltration, while the low permeability porous has only 5%.
Figure 6.20: Seepage into tunnels bored through different bedrock types.

Figure 6.21: Percent storage within tunnel for two different rock types.
The events seen to increase seepage volume are included in Table 6.4. It is clear that seepage into a tunnel bored through low permeability porous rock takes longer to arrive for all precipitation events and the magnitude of each is much less compared to the base case. The precipitation threshold for increased seepage is also larger in this simulation and no events below 20 mm/48 hrs are seen to have any significant impact.

Table 6.4: Precipitation events and seepage timing and magnitude for the low permeability porous bedrock.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
<th>Time to Peak Seepage (days)</th>
<th>Magnitude of Peak Seepage (L/hr/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>23.7</td>
<td>5</td>
<td>19</td>
<td>0.037</td>
</tr>
<tr>
<td>25.8</td>
<td>4</td>
<td>30</td>
<td>0.158</td>
</tr>
<tr>
<td>29.1</td>
<td>4</td>
<td>16</td>
<td>0.06</td>
</tr>
<tr>
<td>47.1</td>
<td>6</td>
<td>14</td>
<td>0.017</td>
</tr>
</tbody>
</table>

As can be seen below in Figures 6.22 and 6.23, trends between precipitation and seepage are not present. A fast arrival or seepage increase doesn’t necessarily occur with a larger precipitation event.

Figure 6.22: Lag time to seepage for the low permeability porous simulation.
6.5 Fractures

The fracture density is seen to have almost a counter-intuitive result in which the more densely fractured rock results in less seepage than the simulation with a less dense fracture network (Figure 6.24). This is an interesting finding and can be explained by the ability of these fractures to transmit water. The more densely fractured network creates a number of possible flow paths around the tunnel through which the diversion of water can occur. However, in the less densely fractured network, when a large fracture happens to intersect the surface and the tunnel there is less chance that this will be redirected by other fractures. This finding is consistent with the observations of Zhou et al. (2006) that a positive correlation between fracture spacing and infiltration exists, but a lack of correlation is seen between normalized seepage and fracture spacing. It is also interesting that the highly fractured network causes more peak seepage with slightly slower response to precipitation than the less dense network. This peak seepage can be attributed to a fast arrival of the wetting front through the well-connected network, which dissipates shortly after its appearance. There also appears to be a seasonal cycle in seepage for both simulations with the higher density network showing more extreme highs and lows, while the lower density is near the same order of magnitude.

Figure 6.23: Magnitude of seepage for low permeability porous simulation.
Figure 6.24: Seepage into tunnels bored through different fracture densities. Note this is a semi-log plot.

Figure 6.25 illustrates the effects of fracture density on the volume of seepage. It shows almost 45% of net infiltration to be stored in the tunnel for the 4 m fracture density. This is quite substantial, given that the tunnel width is only 15% of the domain width, and it’s volume even less. This high storage percentage is likely a result of lateral flow impacting seepage, as seen by Wang et al. (1999), Flint et al. (2001), and Rademacher et al. (2003). The more highly fractured network has approximately half the storage, which again suggests that the fractures are more connected and able to divert the water from the tunnel. These observations highlight the need for appropriate site exploration and characterization to identify large extensional fractures, which can have profound effects on tunnel seepage.

Precipitation events and their corresponding seepage magnitudes are shown in Table 6.5. Again a precipitation threshold exists at approximately 10 mm/48 hrs, beneath which, increased seepage is not seen. It is interesting to note that this seepage is occurring at a much more constant rate than any of the other simulations. This is because the overburden is becoming more saturated in response to these precipitation events because the matrix conductivity is so low and there are fewer high conductivity fractures to flow through. Those fractures, which are connected to the reservoir forming in the overburden and network, are also becoming saturated and are therefore permitting flow of water at a rate more similar to the saturated hydraulic conductivity. Figure 6.26 illustrates this observation.
Table 6.5: Precipitation events and the corresponding seepage for a 4 m fracture spacing.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
<th>Time to Peak Seepage (days)</th>
<th>Magnitude of Peak Seepage (L/hr/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.7</td>
<td>11</td>
<td>18</td>
<td>-</td>
</tr>
<tr>
<td>17.3</td>
<td>2</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>22.2</td>
<td>1</td>
<td>6</td>
<td>0.7</td>
</tr>
<tr>
<td>23.6</td>
<td>1</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>23.8</td>
<td>0.5</td>
<td>11</td>
<td>0.58</td>
</tr>
<tr>
<td>25.8</td>
<td>0.5</td>
<td>14</td>
<td>0.88</td>
</tr>
<tr>
<td>29.1</td>
<td>0.75</td>
<td>16</td>
<td>0.66</td>
</tr>
<tr>
<td>47.1</td>
<td>2.75</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Figures 6.27 and 6.28 once again show the correlation between precipitation event size, and the timing and magnitude of seepage. Again the larger events are arriving at the tunnel much faster. There is little correlation however, between precipitation and seepage volume.
Figure 6.26: Domain slicing that shows saturation after the 29 mm precipitation event for 0.8 m spacing and 4 m spacing. Note the reservoir forming at the top of the domain with 4 m fracture spacing and the higher saturation of those fractures.

Figure 6.27: Lag time to seepage for 4 m fracture spacing. Note the trendline is for visualization.
6.6 Tunnel Depth

The magnitude of seepage into tunnels for the two different depths modeled is shown on Figure 6.29 and follows a different result than expected and predicted by Cesano et al. (2000). The deeper tunnel is actually receiving less seepage than a shallower tunnel. This effect is most likely because the higher fracture density is diverting more water than would be expected for one large extensional feature that is more likely to be encountered with increasing depth. The timing of this seepage is predictable in that the lag time is greater for a deeper tunnel. Interestingly, this is the only simulation in which seepage to the tunnel returns to zero.

Table 6.6 shows the notable precipitation events seen to increase seepage. The precipitation threshold is once again present, but for this simulation appears to be 20 mm/48 hrs. This could be an explanation for the seepage volume returning to zero during the winter months. Figures 6.30 and 6.31 show the trends visible for this data. Again, higher magnitude events are arriving faster than the smaller ones. The magnitude of the event appears to increase the amount of seepage, this behavior is clearer in Figure 6.29. This is possibly due to large early events wetting the soil and fractures, and subsequent
events appearing to have less lag time with larger magnitude due to the increased hydraulic conductivity and addition of the events.

Table 6.6: Precipitation events and the corresponding seepage for the 25 m deep tunnel.

<table>
<thead>
<tr>
<th>Precipitation (mm/48hrs)</th>
<th>Time to Seepage Arrival (days)</th>
<th>Time to Peak Seepage (days)</th>
<th>Magnitude of Peak Seepage (L/hr/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>22.2</td>
<td>10</td>
<td>12</td>
<td>0.27</td>
</tr>
<tr>
<td>23.7</td>
<td>5</td>
<td>12</td>
<td>0.52</td>
</tr>
<tr>
<td>23.8</td>
<td>20</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>25.8</td>
<td>4.75</td>
<td>16</td>
<td>0.5</td>
</tr>
<tr>
<td>29.1</td>
<td>4.9</td>
<td>12</td>
<td>0.53</td>
</tr>
<tr>
<td>47.1</td>
<td>4</td>
<td>14</td>
<td>0.41</td>
</tr>
</tbody>
</table>

Figure 6.29: Seepage into tunnels bored at different depths.
Figure 6.30: Lag time to seepage for the 25 m deep tunnel.

Figure 6.31: Magnitude of seepage for the 25 m deep tunnel.
6.7 Results Summary

An illustration of each simulation is shown again on Figure 6.32. The total precipitation over two days and lag to seepage arrival for each simulation is plotted on Figure 6.33. On this plot, it is evident that the fastest appearance of seepage occurs in the case with 4 m fracture spacing due to the saturation of fractures staying relatively high and the formation of a reservoir. The base case comes next and is closely followed by the vegetated simulation. The 25 m deep tunnel and low permeability porous rock have larger and similar lag times to seepage, but the sandy loam is significantly higher. None of the precipitation events is seen to arrive in less that 12 hours, some take as longer than ten days to be observed.

Figure 6.34 shows the precipitation and associated peak seepage for the four simulations in which it could be separated. The low permeability porous simulation has significantly lower peak seepage than the 4 m fracture spacing. The two highest values for peak seepage are observed in the base case and vegetated simulations.

Figure 6.32: Illustrations of all seven simulations as follows: a) base case, b) sandy loam, c) 15 m overburden, d) vegetated, e) low permeability porous, f) 4 m fracture spacing, g) 25 m tunnel depth.
Figure 6.33: Magnitude of precipitation and lag to seepage arrival for all cases except the 15 m overburden. Note the semi-log plot.

Figure 6.34: Magnitude of precipitation and associated seepage for the four of seven simulations in which there was an observable increase.
CHAPTER 7
CONCLUSIONS

This research uses a fully integrated model with heterogeneous, anisotropic permeability fields that is forced by real meteorological data for Denver, CO to simulate precipitation induced seepage into a tunnel for seven different scenarios. These seven scenarios are hypothetical and represent a tunnel bored through a three-dimensional fracture network in bedrock. The hypothetical domain was constructed by basing the conceptual model on the parameters determined in the literature to be important and geostatistics was then used to generate probable realizations of hydraulic conductivity. Simulations were analyzed by investigating the magnitude of precipitation events that triggered seepage to the tunnels, the timing at which that seepage occurred and the magnitude of seepage.

The effect of each parameter varied on the timing and magnitude of precipitation arrival is summarized in Table 7.1. There are a number of features identified throughout the results, which are important for tunnel design; mainly the arrival time of sustained or varying seepage magnitudes. The importance of sustained versus fluctuating seepage and the timing of its arrival is dependent on the tunnel purpose.

Table 7.1: Summary of the timing and magnitude response for each parameter varied.

<table>
<thead>
<tr>
<th>Parameter Varied: From – To</th>
<th>Lag Timing</th>
<th>Reason</th>
<th>Peak Magnitude</th>
<th>Reason</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand – Sandy Loam</td>
<td>↑</td>
<td>Reduced K slows wetting front advancement</td>
<td>↓</td>
<td>Reduced K keeps water available at surface for evapotranspiration</td>
</tr>
<tr>
<td>2 m – 15 m Soil</td>
<td>↑</td>
<td>Thicker overburden for wetting front to move through</td>
<td>↓</td>
<td>More soil storage</td>
</tr>
<tr>
<td>Barren – Vegetated</td>
<td>↑</td>
<td>Evapotranspiration reduces soil saturation during dry down and thus reduces K</td>
<td>↓</td>
<td>Interception of vegetation</td>
</tr>
<tr>
<td>Igneous/Metamorphic – Sandstone/Basalt</td>
<td>↑</td>
<td>Higher conductivity of bedrock reduces effect of fractures</td>
<td>↓</td>
<td>Higher storage and conductivity of bedrock</td>
</tr>
<tr>
<td>0.8 m – 4 m Fracture Spacing</td>
<td>↓</td>
<td>Fractures remain saturated and therefore have higher K</td>
<td>↓</td>
<td>Reservoir forms in overburden which keeps seepage more constant</td>
</tr>
<tr>
<td>10 m – 25 m Tunnel Depth in Bedrock</td>
<td>↑</td>
<td>Greater distance to travel</td>
<td>↓</td>
<td>Increased diversion away from tunnel with depth</td>
</tr>
</tbody>
</table>
A number of different conclusions can be drawn from the results of the simulations, which reinforce the observations and general trends of seepage shown throughout the literature. These observations that are made using a wide range of different techniques are all shown to occur as an interrelated system through the use of more modern, computationally efficient hydrologic simulations.

The study shows that the timing and magnitude of seepage into tunnels is highly dependent on all of the climatic, ecologic, geologic, and hydrogeologic variables that are present. A clear non-linearity is shown in the relationship between the magnitude of temporally variable precipitation and seepage that is subject to the event, and the antecedent conditions. A precipitation threshold is shown to exist across a number of simulations, below which the background seepage is consistent or decreasing. Above this precipitation threshold, seepage arrival times tend to scale exponentially with event size. Seepage magnitude is found to be highly dependent on the hydrogeologic environment of each simulation. Some of which show a steady rate of seepage that is less influenced by temporally variable precipitation, others that show a high range of extreme fluctuations between storms. The difference between this steady rate seepage and extreme fluctuations is very important to consider in tunnel design.

This study gives insight for those tunneling through fractured rock as to the conditions under which seepage is most likely to occur and the range of volumetric extremes that can be expected. It also highlights the need to appropriately characterize the regional geology in an appropriate manner if one is to obtain reliable estimates when conducting a preliminary investigation. The potential for using tunnels for water resources in these environments, although not explicitly stated, is also a possible outcome.

Coupled, fully integrated hydrologic models are powerful tools that can be used in conjunction with available data to estimate hydrologic response. Future work should focus on seepage responses to precipitation in other climates, with other combinations of the parameters varied for this research, in tunnels bored beneath the water table through fractured media, and using Monte Carlo simulation on a real site to determine the capability of these methods to produce accurate estimates and the associated uncertainty.
REFERENCES CITED


Maxwell, R.M. (2012). A terrain-following grid transform and preconditioner for parallel, large-scale, integrated hydrologic modeling. *Advances in Water Resources, 10*


