University of Nevada

Reno

Numerical Analysis of Infiltration and Near-Surface Percolation in Relation to Yucca Mountain, Nevada

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Hydrology and Hydrogeology

by

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ABSTRACT

Numerical analysis of infiltration into soil and bedrock at Yucca Mountain was useful in delineating parameter control within the flow system. Antecedent moisture conditions, precipitation duration and intensity, and potential evapotranspiration varied between two season-related scenarios. Soil depth to bedrock variability was constrained by interaction of bedrock with percolating water. Summer conditions effectively inhibited percolation within the soil. Even a nighttime winter potential evapotranspiration rate of 0.02 cm/day strongly influenced moisture movement above shallow bedrock surfaces. A shallow rock depth of 7 cm produced a maximum soil-rock interface pressure head of -5.2 cm under winter conditions. Fracture fillings of Touchet silt loam and fine sand produced limited fracture flow for times considered, while Cecil loamy sand allowed significant flow. Fractures spaced 10 cm apart showed little restriction of flux due to spacing, reflecting an almost linear increase in flux with decreasing fracture spacing on a per area basis.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>ii</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td>iii</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td>1</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>3</td>
</tr>
<tr>
<td>CHAPTER 1: INTRODUCTION</td>
<td>4</td>
</tr>
<tr>
<td>Background</td>
<td>4</td>
</tr>
<tr>
<td>Purpose and Scope</td>
<td>4</td>
</tr>
<tr>
<td>Previous work</td>
<td>5</td>
</tr>
<tr>
<td>Setting</td>
<td>8</td>
</tr>
<tr>
<td>Location</td>
<td>8</td>
</tr>
<tr>
<td>Regional climate</td>
<td>11</td>
</tr>
<tr>
<td>Vegetation</td>
<td>11</td>
</tr>
<tr>
<td>Geology</td>
<td>11</td>
</tr>
<tr>
<td>CHAPTER 2: NUMERICAL ANALYSIS</td>
<td>14</td>
</tr>
<tr>
<td>CHAPTER 3: APPLICATION</td>
<td>17</td>
</tr>
<tr>
<td>Analytic procedure</td>
<td>17</td>
</tr>
<tr>
<td>Data requirements</td>
<td>20</td>
</tr>
<tr>
<td>Potential Evapotranspiration</td>
<td>21</td>
</tr>
<tr>
<td>Infiltration Flux</td>
<td>25</td>
</tr>
<tr>
<td>Soil moisture parameters</td>
<td>27</td>
</tr>
<tr>
<td>Rock matrix moisture parameters</td>
<td>27</td>
</tr>
<tr>
<td>Rock-fracture moisture parameters</td>
<td>30</td>
</tr>
</tbody>
</table>
Non-filled fractures ................................................................. 30
Filled fractures ........................................................................ 34
Initial moisture and lower boundary conditions ...................... 36

CHAPTER 4: RESULTS .................................................................. 37
Stage 1 .......................................................................................... 37
Stage 2 .......................................................................................... 39
Stage 3 .......................................................................................... 42
Non-filled fractures ................................................................. 42
Filled fractures ........................................................................ 45

CHAPTER 5: CONCLUSIONS .................................................... 52
REFERENCES ................................................................................ 56
APPENDIX A ............................................................................. 61
LIST OF FIGURES

1.1 Location of Yucca Mountain. ................................................................. 9
1.2 Generalized regional map of the Yucca Mountain proposed repository
    block and vicinity. .................................................................................... 10
3.1 Schematic of Stage 2 scenario. ................................................................. 18
3.2 Schematic of Stage 3 scenario with a filled fracture. ......................... 19
3.3 Daytime PET distributions. .................................................................. 23
3.4 Nighttime PET distributions. ................................................................. 23
3.5 Winter precipitation, duration vs. intensity. ......................................... 26
3.6 Summer precipitation, duration vs. intensity. .................................... 26
3.7 Rock Valley soil moisture retention curve. ......................................... 28
3.8 Effective saturation vs. log relative hydraulic conductivity for the
    Rock Valley soil. ..................................................................................... 28
3.9 Tiva Canyon Member moisture retention curve. .............................. 29
3.10 Effective saturation vs. log relative hydraulic conductivity for
    Tiva Canyon Member for ε=6.8. ......................................................... 29
3.11 Equivalent porous medium hydraulic conductivity curve showing
    rock with fractures, portion (2), and rock only, portion (1). .............. 33
3.12 Equivalent porous medium moisture retention curve showing rock with
    fractures, portion (2), and rock only, portion (1). ............................ 33
3.13 Moisture retention curves for 3 fracture fillings and the Rock Valley
    soil. ......................................................................................................... 35
3.14 Effective saturation vs. log hydraulic conductivity for 3 fracture fillings
    and the Rock Valley soil. ................................................................. 35
4.1 Summer period equivalent moisture depth and drying front depth vs. time. ................................................................. 38
4.2 Winter period equivalent moisture depth and drying front depth vs. time. ................................................................. 38
4.3 Summer period pressure heads at the soil-rock interface for 5 and 10 cm soil depths. ......................................................... 40
4.4 Winter period pressure heads at the soil-rock interface for 7 and 10 cm soil depths ......................................................... 41
4.5 Equivalent porous medium moisture retention curves for rock with 3 different fracture apertures and for rock alone. All curves coincide. ........ 44
4.6 Equivalent porous medium hydraulic conductivity curves for rock with 3 different fracture apertures and for rock alone. ......................... 44
4.7 Downward Darcy velocities for the uppermost fracture element; s = 80 cm, filling is Cecil loamy sand. ......................................................... 47
4.8 Downward Darcy velocities at the top fracture element for three different fracture spacings. ......................................................... 47
4.9 Fracture wetting pulse equivalent moisture depth vs. fracture spacing for 3 specified times. ......................................................... 48
4.10 Soil zone velocity vectors for s = 10 cm. ................................................................. 49
4.11 Soil zone velocity vectors for s = 40 cm. ................................................................. 50
4.12 Soil zone velocity vectors for s = 80 cm. ................................................................. 50
### LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>Time periods and PET statistics.</td>
<td>24</td>
</tr>
<tr>
<td>3.2</td>
<td>Rock Valley soil functional and flow parameters.</td>
<td>27</td>
</tr>
<tr>
<td>3.3</td>
<td>Tiva Canyon Member functional and flow parameters.</td>
<td>30</td>
</tr>
<tr>
<td>3.4</td>
<td>Functional and flow parameters for fracture fillings.</td>
<td>34</td>
</tr>
<tr>
<td>4.1</td>
<td>Stage 1 summer scenario simulation results.</td>
<td>40</td>
</tr>
<tr>
<td>4.2</td>
<td>Stage 1 winter scenario simulation results.</td>
<td>41</td>
</tr>
<tr>
<td>4.3</td>
<td>Equivalent porous media functional and flow parameters.</td>
<td>43</td>
</tr>
</tbody>
</table>
BACKGROUND

Yucca Mountain, located in southern Nevada, is presently being evaluated as a possible high level nuclear waste repository. Water percolating through rocks surrounding the repository poses a threat as a transmitter of stored radioactive material and therefore presents a contamination hazard to groundwater. With the repository located in tuffs at moisture contents lower than saturation, the source of percolating water is infiltration of precipitation at the surface of Yucca Mountain and in adjacent washes.

The entry of water into the hydrogeologic cycle at the surface of Yucca Mountain is ultimately associated with flow into fractured bedrock. This may occur either as direct infiltration into exposed outcrop, or as infiltration into a soil covering first with subsequent percolation into the underlying fractured bedrock. Both possible occurrences of water entry are affected by numerous factors including evapotranspiration, precipitation intensity and duration, soil depth to bedrock, and fracture aperture, spacing, and filling.

PURPOSE AND SCOPE

The purpose of this study was to examine the various hydrologic, atmospheric, and morphologic parameters involved in the infiltration and near-surface percolation process at Yucca Mountain, Nevada. The desired outcome was to be able to determine what combination of parameters would produce a maximum infiltration/percolation scenario. Delineation of a maximum infiltration and percolation setting is crucial with respect to recharge and the implications of moisture movement in and around a waste repository. Quantitative evaluation of the flow regime was accomplished numerically using a two dimensional finite element computer code which solves the variably saturated flow equation with respect to specified boundary and initial conditions. Due to limited access to Yucca Mountain for data gathering, analysis was limited to the use of published representative values specific to Yucca Mountain in the case of rock moisture retention.
properties, fracture data, and meteorological data, and to the use of published values obtained for settings similar to Yucca Mountain in the case of soil properties and vegetative parameters.

**PREVIOUS WORK**

Although not an uncommon method of analysis for the infiltration problem, numerical simulation of infiltration at Yucca Mountain has not been previously reported in the literature. To date, no direct work on infiltration, whether theoretical or experimental, site specific to Yucca Mountain, has been previously published although much is planned (U.S. Department of Energy, 1987). However, several indirect approaches have been used to estimate infiltration rates including water budget studies, geothermal heat flux studies, pore velocity measurements, and numerical simulation of liquid-water flow within the unsaturated zone. Of these, the first two address the problem on a regional scale whereas the last two examine the problem on a local scale. Since it was not the purpose of this study to estimate values of infiltration but instead to evaluate the mechanisms governing the infiltration process, the above methods will only be briefly mentioned so as to identify the techniques used along with the resulting infiltration estimates.

Infiltration estimates reported from indirect approaches range from 0.0 centimeters per year (cm/yr) to 0.45 cm/yr. Rice (1984), applying a sensitivity analysis to a water balance study, estimated recharge at Yucca Mountain to be 0.0 cm/yr. But, as it was pointed out by Rice, the estimation approach used was at too gross of a scale to give accurate results. A method developed by Eakin et al. (1951) and Malmberg and Eakin (1962) uses regional relationships that were established among recharge, altitude zone, and precipitation. Czarnecki (1984), applying this method together with results from Rush (1970), estimated recharge to be about 0.05 cm/yr in a zone which includes Yucca Mountain. Waddell et al. (1984) estimated a flux of about 0.05 cm/yr through alluvium in Yucca Flat about 40 km northeast of Yucca Mountain. Winograd and Thordarson (1975) estimated recharge to the carbonate aquifer underlying the Nevada Test Site as three percent of precipitation falling on upland outcrop areas. By using an average yearly precipitation of 15.0 cm/yr (Montazer and Wilson, 1984), three percent translates
to 0.45 cm/yr. The concern expressed by Rice with respect to accuracy should be applied to all of these estimates since all use some form of averaging or extrapolating which tend to ignore any possibly significant short term effects taking place at the surface of Yucca Mountain. Also, as pointed out by Montazer and Wilson (1984), estimation of recharge assuming equilibrium between discharge and net infiltration may show effects of paleoclimatic conditions due to the large and variable travel times of water through the unsaturated zone.

Sass and Lachenbruch (1982) estimated vertical fluxes of 0.1 - 1.0 cm/yr in the combined saturated and unsaturated zones at Yucca Mountain, using geothermal data from various boreholes. Problems with this approach arise when interpreting the data taken from boreholes in the saturated zone due to borehole completion methods and horizontal water flux. It was again pointed out by Montazer and Wilson (1984) that these results must be interpreted with respect to paleoclimatic conditions and large travel times in the unsaturated zone.

Two infiltration estimates have been published resulting from numerical simulation of liquid-water flow within the unsaturated zone. Rulon et al. (1986) and Peters et al. (1985) both predicted that steady net infiltration rates exceeding values in the range from 0.05 to 0.1 cm/yr would produce matrix saturation of presently unsaturated zones within Yucca Mountain. These two estimates, drawing upon average properties of limited data for calculation, strongly indicate the upper bound of infiltration to be about 0.1 cm/yr which encompasses most of the estimates just given.

Direct measurement or calculation of infiltration at Yucca Mountain has not yet been published, although work has been done dealing with the problem of infiltration into fractured bedrock. Most notable of these was field work done by Kilbury et al. (1986). Using a fractured rock infiltrometer, measurements were made of vertical infiltration rates versus time, fracture permeabilities, and apparent hydraulic apertures. Using these data along with measurements of fracture length traces for a specified area, estimates of mesoscale intake rates were determined. Applying this information to a stochastic rainfall model coupled to a deterministic runoff/infiltration model, estimates of
infiltration as a percentage of precipitation were calculated.

Analysis of a system most closely resembling the system existing at Yucca Mountain was performed by Germann (1986). Taking into consideration vegetative cover, soil conditions, surface runoff, daily precipitation, daily drainage, and daily soil moisture content, Germann investigated seasonal variation in the frequency of rapid drainage responses with respect to the occurrence of macropores at the 2.4 meter level in the soil. Results indicated that water sorption by the soil matrix was the controlling factor over drainage at the 2.4 meter level. Correlation between the quarterly totals of evaporation and precipitation and the number of drainage responses was low due to lack of correlation between evapotranspiration and precipitation and the matrix sorbance capacity prior to an individual storm. Furthermore, Germann pointed out that higher intensity storms were required for a rapid drainage response during the summer when drier antecedent soil moisture conditions occurred than during the winter when antecedent soil moisture conditions were wetter.

Germann (1985), Germann and Beven (1981a, 1981b, 1985, 1986), and Beven and Germann (1981, 1982) reported attempts at analytical modeling of the flow of water within macropores (with specific reference here to the macropore subset of fractures) with respect to a surface infiltration setting using a kinematic wave approach. While taking into consideration the interaction of the macropores and the surrounding matrix, saturated flow through fractures was assumed to be essentially parallel plate flow with a single scaling factor used to allow for any deviations in flow geometry. The relationship used to describe partially saturated flow within the macropores was empirically derived and allowed for movement along the walls of the fractures. Other work on partially saturated fracture flow has been reported by Evans and Huang (1983) who present a model of unsaturated fracture flow whereby capillary theory can be applied in order to approximate the air-entry potential for different size fractures. Estimates were also made regarding the occurrence of film flow on the walls of a drained fracture.

The influence of geometry on flow in a fracture has been noted by Witherspoon and Long (1987). They delineate four possible flow regimes that could occur depending upon
the state of stress present in the rock. For a fracture which is sufficiently open, flow will
be essentially that of flow between parallel plates. With partial closure of the fracture,
points of contact will occur. For contact areas not well distributed, the flow may be
nonlinear with flux not proportional to the gradient. For contact areas finely distrib­
uted, flow would be similar to that of a porous medium where potential theory is appli­
cable. Finally, they point out that for sufficiently closed fractures, flow will occur in tor­
tuous channels; essentially a two dimensional network of one dimensional connectors.
These considerations suggest that application of the cubic law is valid only under certain
conditions. The question of the validity of the cubic law has also been addressed by
Neuman (1987) who points out that only at low stresses is the cubic law for fracture flow
regarded usable for natural fractures that tend to have rough walls. Neuman further
indicates that given the high frequency and amplitude variation of apertures, attempts to
model the effect of aperture variability on flow through a fracture using the cubic law
over finite segments, as suggested by Iwai (1976), Neuzil and Tracy (1981), and Tsang
and Witherspoon (1985), may not be valid as well.

SETTING

Location

Yucca Mountain is located west of and adjacent to the southwestern portion of the
Nevada Test Site (Figure 1.1), about 105 km northwest of Las Vegas, Nevada. The
approximate portion of Yucca Mountain being considered as a repository is called the
central block (Figure 1.2). Located in the Great Basin, the northernmost subprovince of
the Basin and Range physiographic province, Yucca Mountain reaches a maximum eleva­
tion of 1,509 meters above mean sea level, about 460 meters above Jackass Flats to the
east, and about 180 meters above Solitario Canyon to the west.
Figure 1.1. Location of Yucca Mountain (modified from U.S. Department of Energy, 1986).
Figure 1.2. Generalized regional map of the Yucca Mountain proposed repository block and vicinity (modified from Scott and Castellanos, 1984).
Regional Climate

The climate at Yucca Mountain is classified as M mid-latitude arid (Houghton et al., 1975). Climatic features of this area include: large diurnal and seasonal ranges in temperature, low relative humidity, strong insolation, rapid radiation loss, and generally light wind velocities (Eglinton and Dreicer, 1984). Two basic storm patterns typify the area: (1) those resulting from winter cyclonic activity; and (2) those caused by intense summer convection. Approximately 73 percent of the annual precipitation falls during the cool season running from October to April (Quiring, 1983).

Vegetation

Vegetation is sparse desert scrub indicative of the Mojave Desert and varies in species with elevation (Spaulding, 1983). Dominant species occurring in a mixed transition zone, as seen on the crest of Yucca Mountain, can be classified as a Grayia-Ephedra vegetation association (O'Farrell and Collins, 1983). The percent total vegetative ground cover is approximately twenty-two percent (O'Farrell and Collins, 1983).

Geology

Yucca Mountain is structurally typical of the Basin and Range province. The mountain consists of a series of north-trending fault-block ridges underlain by volcanic rocks dipping 5° to 10° to the east. The volcanic rocks consist of separate formations of silicic volcanic tuffs of Miocene age. Stratigraphically, these are distinguished by their petrographic characteristics. Within each stratigraphic unit, the variation of physical properties is due largely to the degree of welding within each unit.

Three types of regional Cenozoic structures exist at Yucca Mountain: 1) Basin and Range style faults; 2) major strike slip fault zones; and 3) volcano-tectonic structures. Several major north-northeast striking and generally westward dipping Basin and Range style normal faults are seen to exist, with tens to hundreds of meters of vertical displacement (Scott et al., 1983). The west side of the central block of Yucca Mountain is bounded by a major north-striking normal fault. The eastern and southeastern
boundary of the central block is formed by a zone of imbricate normal faults which are west dipping with about 2 to 5 meters of vertical offset. The northeast boundary consists of a strike-slip shear zone underlying Drill Hole Wash (Montazer and Wilson, 1984; see Figure 1.2).

In general, two types of tuffs exist at Yucca Mountain delineated by the degree of welding. One type consists of densely welded, relatively non-porous, highly fractured ash flow tuffs (Scott et al., 1983). The second type of tuff consists of highly porous, nonwelded, relatively unfractured tuff. The fractures in the welded tuff fall into three main categories based on their mode of formation (Barton, 1984). The first type consists of fractures formed during cooling of the tuff. The second and third types were formed from regional and local stress fields acting on the tuff after it cooled (U.S. Department of Energy, 1987). The strikes of fractures on the surface pavements of Yucca Mountain are variable and the dips are generally greater than 65 degrees (Barton and Larsen, 1985). A fourth occurrence of fractures does exist but accounts for only 5-10 percent of all the fractures where they occur. These fractures form parallel to eutaxitic structures which tend to be subhorizontal (Spengler and Rosenbaum, 1980). Montazer and Wilson (1984) separate the rocks at Yucca Mountain into hydrogeologic units based on the degree of welding.

Within the boundaries of the central block, the only geologic unit exposed at the surface is the Tiva Canyon Member of the Paintbrush Tuff. The thickness ranges from 90 to 140 meters and can generally be described as a multiple-flow compound cooling unit of a compositionally-zoned rhyolitic to quartz latitic ash flow tuff (Scott and Bonk, 1984). Numerous zones and subzones exist within the Tiva Canyon unit, as outlined by Scott and Bonk (1984). In general, the units are moderately to densely welded tuffs, with localized zones of less welded tuff with each unit exhibiting its respective degree of fracturing.

Of the two general categories of topography in the area, piedmont slopes and upland landforms, Yucca Mountain falls into the latter. The upland land forms can be further broken down into three types: ridge crests, valley bottoms, and intervening
hillslopes. The crest of Yucca Mountain consists of a gently sloping caprock-protected dip-slope. The surface consists of either locally thin veneers of colluvial or eolian deposits or exposed bedrock. It is the crest area which was the primary focus of this study. Valley morphology ranges from shallow, straight, steeply sloping gullies and ravines to relatively deep, bifurcating, gently sloping valleys and canyons containing varying thicknesses of alluvium which locally thicken. The hillslopes may contain free faces and bedrock outcrops on the upper slopes. Intermediate slopes may be mantled with talus or poorly sorted colluvium which in turn may be cut by gullies and ravines. The lower slopes are typically mantled with varying thicknesses of wash deposits (U.S. Department of Energy, 1987).
CHAPTER 2:
NUMERICAL ANALYSIS

A numerical approach was used to obtain approximate solutions to the governing flow equations. The numerical code used, SATURN: A Two-Dimensional Finite-Element Model for Saturated- Unsaturated Flow and Contaminant Transport, Version 1.4 (Huyakorn et al., 1986), is a code developed to simulate moisture movement and groundwater flow in variably saturated and fully saturated porous media, respectively. The governing equation used in SATURN for water flow in a variably saturated soil is given as

\[ \frac{\partial}{\partial x_1} \left[ K_{ij} k_{w} \left( \frac{\partial \psi}{\partial x_j} + c_j \right) \right] = \left( S_{w} S_{e} + \phi \frac{dS_{w}}{d\psi} \right) \frac{\partial \psi}{\partial t} - q \]  \hspace{1cm} 2.1

where \( \psi \) is the pressure head in centimeters, \( K_{ij} \) is the saturated hydraulic conductivity tensor in centimeters per day, \( k_{w} \) is the relative permeability with respect to the liquid water phase, \( x_i \) \((i=1,2)\) and \( z_j \) \((j=1,2)\) are spatial coordinates in centimeters, \( t \) is time in days, \( c_j \) is the unit vector assumed to be vertically upward, \( S_{w} \) is water phase saturation, \( \phi \) is the effective porosity, and \( q \) is the volumetric flow rate via sources or sinks per unit volume of the porous medium in units of 1/day. \( S_{e} \), the specific storage in units of 1/centimeter, is given as

\[ S_{e} = \rho_{w} g \left( \phi \beta' + \alpha' \right) \]  \hspace{1cm} 2.2

where \( \rho_{w} \) is the density of water in grams per cubic centimeter, \( g \) is the gravitational acceleration in centimeters per day squared, and \( \alpha' \) and \( \beta' \) are coefficients of compressibility of the porous medium and water, respectively, in units of centimeter squared per dyne (Huyakorn et al., 1986).

The initial and boundary conditions used in SATURN are

\[ \psi(x,0) = \psi_0(x_i), \]  \hspace{1cm} 2.3

\[ \psi(x_i,t) = \bar{\psi} \text{ on } B_i, \]  \hspace{1cm} 2.4

and
\[ V_i n_i = -V_n \text{ on } B_2 \]

where \( \psi_0 \) is the initial head value, \( B_1 \) is the portion of the flow boundary where \( \psi \) is prescribed as constant head, \( \bar{\psi} \), \( B_2 \) is the portion of the flow boundary where the outward normal velocity is prescribed as constant flux, \(-V_n\), \( n_i \) is the outward unit normal vector, and \( V_i \) are Darcy velocity vector components, in centimeters per day, calculated from

\[ V_i = -K_{ij} k_{rm} \left( \frac{\partial \bar{\psi}}{\partial x_j} + \epsilon_j \right) \]

Atmospheric boundary conditions include soil-air interfaces where evaporation or infiltration occurs. The upper boundary condition is allowed to change from the Dirichlet to the Neumann type in response to evaporation changing from the constant-rate stage to the falling-rate stage, respectively. A numerical solution must be obtained by maximizing the absolute value of the flux subject to the following requirements:

\[ |V_i| \leq |E^*| \]

and

\[ \psi_L \leq \psi \leq 0 \]

where \( E^* \) is the maximum potential surface flux under the prevailing atmospheric conditions, and \( \psi_L \) is the minimum pressure head allowed under the prevailing soil conditions.

Also required are relationships of water phase saturation versus relative conductivity and pressure head versus water phase saturation. Relative conductivity is calculated by the Brooks-Corey formulation (Brooks and Corey, 1966) given by

\[ k_{rm} = \frac{S_w - S_{wr}}{1 - S_{wr}} \]

where \( k_{rm} \) is the relative conductivity, \( \epsilon \) is an empirical parameter, \( S_w \) is the water phase saturation, and \( S_{wr} \) is the residual water phase saturation. \( S_w \) and \( S_{wr} \) are represented by

\[ S_w = \frac{\epsilon}{\epsilon_S} \]

and
respectively, where $\theta$ is the observed moisture content, $\theta_r$ is the residual moisture content, and $\theta_s$ is the saturated moisture content. The water phase saturation, $S_w$, from Mualem (1976), is calculated from

$$S_w = \frac{\theta_r}{\theta_s}$$

Equation 2.11

where $\alpha$, $\beta$, and $\gamma$ are empirical parameters, $\psi_s$ is the air entry pressure head value, and the left side of Equation 2.9 is the effective saturation, $S_{ef}$. Equations 2.1 is solved by the Galerkin finite element method subject to the initial and boundary conditions given above. A combination of central and backward difference time-stepping schemes is used when solving the time integration. Allowable element shapes are either triangular or rectangular. A non-linear system of algebraic equations is obtained for each time step with the non-linearity being treated using either the Picard or the Newton-Raphson iterative technique.

Assumptions inherent in the SATURN code include (Huyakorn et al., 1986):

1. flow of the fluid phase is isothermal and governed by Darcy's law;

2. the fluid is slightly compressible and homogeneous;

3. hysteresis effects in the constitutive relationships of relative permeability versus saturation, and saturation versus capillary pressure head, are assumed to be negligible; and

4. only single-phase flow is considered and flow of a second phase, such as air, is ignored.
CHAPTER 3:
APPLICATION

ANALYTIC PROCEDURE

The system analysis was performed in three stages with the results from one stage used to formulate following stages. The results of Stage 1 were used to determine the maximum depth a bedrock surface could be placed while still interacting with infiltrated water. Stage 2 was used to evaluate the pressure head distributions at the maximum bedrock depths determined in Stage 1, and at shallower depths. Stage 3 was used to evaluate flux into fractures for the bedrock surface depths evaluated in Stage 2. For each stage, two separate scenarios were considered, one representing summer atmospheric and antecedent moisture conditions and one representing winter conditions.

Stage 1 involved simulating, for each scenario, a semi-infinite soil horizon to determine the maximum depth of the infiltrated water and to determine the transient nature of the wetting and drying fronts. With this information it was then determined to what depth evaporation affects moisture movement within the soil, and, therefore, the maximum depth at which a bedrock surface would be considered in this analysis. Assuming vertical flow and an isotropic, homogeneous porous medium, an essentially one-dimensional grid was developed in which one column of elements, 2 meters deep and 1 centimeter wide, was used. The upper boundary was specified as an atmospheric boundary condition, the bottom boundary was specified as a constant head boundary condition, and the sides of the column were designated no flow boundaries.

Stage 2 of the analysis involved simulations using a non-fractured bedrock surface at the various depths determined from Stage 1, representing one end member of the rock matrix/fracture system hydraulic conductivity where \( K_T \), the total hydraulic conductivity, was equal only to \( K_r \), the rock-matrix hydraulic conductivity. Head values at the soil-rock interface were monitored over time to determine pressure head distributions. The horizontal node spacing, vertical column length, and types of boundary conditions were identical to those used in Stage 1 and are given in Figure 3.1.
In Stage 3, two different fracture flow analyses were carried out, one in which fractures were represented by parallel plates with hydraulic conductivity proportional to the square of the fracture aperture, and one in which fractures were represented by parallel plates filled with a porous medium of hydraulic conductivity different from that of the surrounding rock and overlying soil. In the case of filled fractures, the fractures would wet up gradually as specified by the fracture filling material properties. Although fractures may not be open and filled with a porous medium uniformly to any significant depth at the surface of Yucca Mountain, there is good possibility they will have soil in them at the opening to the soil zone. The mode of water movement into the fracture could therefore be dependent, at least at first, on soil-moisture flow properties. The use of soil filled fractures allowed comparison of flow at different fracture densities and also enabled an analysis of flow within a soil zone to an underlying line sink.

The grid pattern used for the case in which hydraulic conductivity was proportional to the square of the fracture aperture in Stage 3 was essentially that of Stage 2 with the fracture hydraulic conductivities being averaged over the entire rock surface. For the case where the fractures were filled with a porous medium, it was feasible to model a
fracture explicitly within a two-dimensional grid. The fracture was modeled as a vertical series of elements half as wide as the fracture aperture, b, placed at the left edge of that portion of the grid representing the rock matrix, as shown in Figure 3.2. The left boundary was made an impermeable boundary since the other half of the flow regime would be a mirror image of that shown here. The right boundary of the grid was also an impermeable boundary representing the water divide which would occur between two fractures spaced twice the distance of the horizontal grid dimension (i.e. one-half the fracture spacing, s). Upper and lower boundaries had the same boundary condition designation as Stages 1 and 2 and the vertical grid dimension was kept at 2 meters. With this system, three material properties could be specified in a single data set, one for the overlying soil, one for the rock-matrix, and one for the fracture filling.

Figure 3.2. Schematic of Stage 3 scenario with a filled fracture.
Assumptions inherent in all of the above flow analyses include:

1. isotropic, homogeneous porous media;
2. only vertical, single-phase, isothermal flow was considered;
3. hysteresis effects were negligible;
4. soil and bedrock surfaces were horizontal;
5. runoff did not occur; i.e., all of the precipitation was assumed to infiltrate;
6. ponding did not occur;
7. fracture geometry was represented by smooth-walled parallel plates; and
8. film flow was negligible within the open fractures.

It was noted in Chapter 1 that the dip of the ridge of Yucca Mountain was only 5° to 10° to the east. This serves as the basis for the assumption of horizontal surfaces. Shallow dip, along with high soil hydraulic conductivities, would allow for little runoff which was also indicated by Rice (1984). The effects of ponding were considered negligible in relation to precipitation intensities and hydraulic conductivities presented in the following sections. The simplified fracture geometry was used since actual fracture geometries were not available and SATURN was not specifically designed to evaluate fracture flow. Film flow was considered to be quite small in comparison to saturated fracture flow (Evans and Huang, 1983).

DATA REQUIREMENTS

Specific input parameters required in the above analysis for use in SATURN included: potential evapotranspiration (PET) estimates, infiltration flux, soil moisture parameters, rock-matrix moisture parameters, rock-fracture moisture parameters, initial moisture conditions, and specification of the lower boundary condition.
Potential Evapotranspiration

Meteorological data from a 10-meter meteorological tower system on the ridge of Yucca Mountain were obtained on magnetic tape from D. L. Freeman, Desert Research Institute, Atmospheric Sciences Center, Reno, Nevada (Church et al. 1985, 1986) allowed for the use of a combination equation for calculation of PET. As derived by Penman (1948), PET is given by

\[ PET = \frac{1}{L} \frac{H(\Delta) + L(e_s - e_d)f(u)}{\frac{\Delta}{\gamma} + 1} \]

where \( H \) is the net solar radiation in calories per square centimeter per day, \( e_s \) is the saturated vapor pressure, in millimeters (mm), at the temperature 3 meters above the ground, and \( e_d \) is the actual vapor pressure in millimeters, 3 meters above the ground. Values for \( H \) were calculated using the empirical equation from Jensen (1973), given by

\[ H = 0.64R_s - 10.8 \]

where \( R_s \) is the incoming solar radiation obtained from the meteorological tower data in watts per centimeter squared, and the values 0.64 and -10.8 were empirically derived for a desert with low scattered vegetation (Jensen, 1973, Table B1). Values for \( e_s \) were calculated using the relationship from Jensen (1973):

\[ e_s = \frac{r}{0.622} P \]

where \( P \) is the atmospheric pressure which was taken as a constant 850 millibars since the Yucca Mountain ridge atmospheric pressure deviates very little from this value and the calculation is insensitive to minor changes in atmospheric pressure. The parameter \( r \), or mixing ratio, is temperature dependent and was calculated by the empirical relationship

\[ r = 4.7406 e^{0.0447T} \]

where \( T \) is the measured temperature 3 meters above the ground. The actual vapor pressure, \( e_d \), was calculated from
\( e_d = \frac{r_h}{100} e_s \)

3.5

where \( r_h \) is the measured relative humidity. The wind velocity function, \( f(u) \), is given by (Penman, 1948)

\[ f(u) = 1.0 + 0.5263u \]

3.6

where \( u \) is the measured wind velocity in meters per second. The ratio \( \frac{A}{\gamma} \) is the ratio of the slope of the saturation vapor pressure versus temperature curve at mean air temperature to the psychrometric constant, and values for \( \frac{A}{\gamma} \) were obtained from tables given by van Bavel (1966). The parameter \( L \) is the latent heat of vaporization and is equal to 580.0 cal/gm.

Values for \( PET \) were calculated at hourly intervals over 4 months, December through March, which comprised the winter period and over 3 months, June through August, which comprised the summer period. Data for the summer period was taken from both 1983 and 1984. Since data were available for January, 1983, through October, 1984, and January through March, 1984, was relatively dry, the months January through March and December, 1983, comprised the winter period. The daytime and nighttime intervals were delineated by the times at which incoming solar radiation went from positive to zero values and from zero to positive values (i.e. in general, sunset and sunrise, respectively). Any time a negative value was calculated for \( PET \), that value was taken as zero. Mean values for both day and night, winter and summer, were then calculated for use in the numerical simulations. Frequency histograms for winter and summer daytime and winter and summer nighttime \( PET \) rates for class intervals of 0.1 centimeters per day (cm/day) are given in Figures 3.3 and 3.4, respectively. These values, along with the respective time periods and other statistical information, are given in Table 3.1.

In order to determine the response of the soil to atmospheric conditions (i.e. Stage 1, above), daytime evaporation rates were specified following infiltration for both the summer and winter scenarios. For Stages 2 and 3, an alternating day/night sequence was followed. From inspection it was noted that summer-type precipitation events occurred mostly during the late afternoon hours; therefore, the summer precipitation
Figure 3.3. Daytime PET distributions.

Figure 3.4. Nighttime PET distributions.
event was assumed to represent a late afternoon thunderstorm. Following this event, a daytime \( \text{PET} \) rate was specified for 0.08 days, then alternating night and day \( \text{PET} \) rates for 0.42 and 0.58 days, were specified, respectively. From inspection of the winter precipitation data, winter storm events were seen to occur randomly during the day and night.

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Length of Time Period, days</th>
<th>Mean ( \text{PET} ), cm/day</th>
<th>Number of Data, Values</th>
<th>Min/Max, cm/day</th>
</tr>
</thead>
<tbody>
<tr>
<td>winter</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>day</td>
<td>0.46</td>
<td>0.37</td>
<td>529</td>
<td>0.00/2.05</td>
</tr>
<tr>
<td>night</td>
<td>0.54</td>
<td>0.02</td>
<td>563</td>
<td>0.00/0.50</td>
</tr>
<tr>
<td>summer</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>day</td>
<td>0.58</td>
<td>1.30</td>
<td>964</td>
<td>0.00/2.95</td>
</tr>
<tr>
<td>night</td>
<td>0.42</td>
<td>0.16</td>
<td>602</td>
<td>0.00/0.72</td>
</tr>
</tbody>
</table>

In order to allow for a worst case infiltration scenario, winter precipitation was assumed to end at the beginning of a 0.54 day nighttime sequence, thus allowing for maximum percolation following the precipitation event. Alternating periods of day and night \( \text{PET} \) rates for 0.46 and 0.54 days, respectively, were then simulated.

It should be noted here that transpiration was not considered in this analysis. Actual evapotranspiration, \( \text{ET}_a \), can be expressed as

\[
\text{ET}_a = T + EV
\]

where \( T \) represents water loss due to transpiration from plants and \( EV \) represents that due to evaporation from bare soil (Hillel, 1982). Values for \( \text{ET}_a \) and \( T \) were computed by Lane et al. (1984) for perennial vegetation at Rock Valley, Nevada, which is vegetation similar in type and percent cover (about 25 percent) to that found at Yucca Mountain (O'Farrell and Collins, 1983). Their work indicated that annual transpiration rates varied from 15 to 37 percent of \( \text{ET}_a \) for the years 1968 to 1976. Also, Szarek (1979) showed that the growing season transpiration rate is roughly one-half the annual rate, with the growing season assumed to coincide with leaf budding and plant dormancy, estimated to run from February thru June (Ackerman et al., 1980). From these considerations it was assumed that transpiration is roughly equal in magnitude to bare soil
evaporation on a per area basis and that the time of occurrence of the most significant transpiration period does not coincide with the winter and summer periods considered in this study. On this basis transpiration was not considered in this study and $ET_*$ was assumed to be equal to $EV$ only.

**Infiltration Flux**

Hourly precipitation data were also available from the 10-meter meteorological tower system on the ridge of Yucca Mountain (Church et al., 1985, 1986). Precipitation data were broken down into two useful parameters: average precipitation intensity per storm and individual storm duration. Depths of precipitation for a given hour presented in the data were read as an intensity in millimeters per hour, then converted to centimeters per day for use in the simulation. Correlation between precipitation intensity and duration was maintained by averaging intensities over the duration of a storm event, with the averages being used to represent the storm precipitation intensity.

Precipitation duration and intensity data were extracted from the raw data in the following manner. For storms of high intensity over several hours, preceded and/or proceeded by a much smaller intensity value, the storm duration was limited to the time of higher values with the smaller value added in without being given any weight. This was done so that small amounts of precipitation collected during only a fraction of an hour would not adversely affect the precipitation intensity average; yet these values would not be left out completely. If, during a storm, the intensity either increased or decreased to a new fairly constant intensity, then the storm was considered bimodal, consisting of two storms of different duration and intensity. If a zero occurred intermittently within a storm, it was averaged in with the rest of the storm.

Plots of duration versus intensity are shown in Figures 3.5 and 3.6 for winter and summer, respectively. As can be seen immediately, portions of the winter events fall into the low intensity-high duration range whereas portions of the summer events fall into the high intensity-low duration range; however, there also exists a zone of overlap in the low intensity-low duration range. A value 4.8 cm/day for 0.291 days was chosen to represent
Figure 3.5. Winter precipitation, duration vs. intensity.

Figure 3.6. Summer precipitation, duration vs. intensity.
the winter period and a value of 24.0 cm/day for 0.042 days was chosen to represent the summer period.

Soil Moisture Parameters

Since access to Yucca Mountain was restricted, soil moisture parameters required for use in SATURN were obtained from Mehuys et al. (1975) for Rock Valley gravelly loamy sand which was assumed to represent the colluvium one might encounter on the crest of Yucca Mountain. All functional parameters, except \( \epsilon \), for the Rock Valley soil, shown in Table 3.2, were calculated from a simultaneous curve fit of \( \psi-\theta \) and \( \psi-K \) data taken from Mehuys et al. (1975) using the van Genuchten curve fitting routine (van Genuchten, unpublished manuscript). A saturated hydraulic conductivity of 890.3 cm/d was calculated by the van Genuchten curve fitting routine using a specified saturated hydraulic conductivity of 900.0 cm/d, which is roughly a median value for a clean sand (Freeze and Cherry, 1979). Porosity, \( \eta \), was calculated as 0.315 using a bulk density of 1.815 gm/cm\(^3\) and a grain density of 2.65 gm/cm\(^3\) (Mehuys et al., 1975). A porosity of \( \eta=0.3154 \), obtained from the van Genuchten curve fit, was used in SATURN. The parameter \( \epsilon \), required by SATURN, was computed from Equation 2.8 using \( \psi-\theta \) values generated by the curve fitting routine. Plots of observed and fitted \( \psi-\theta \) and \( S_{eff}-k_{rw} \) data are shown in Figures 3.7 and 3.8, respectively.

| Table 3.2. Rock Valley soil functional and flow parameters. |
|---|---|---|---|---|---|
| \( \alpha, \text{cm}^{-1} \) | \( \beta \) | \( \gamma \) | \( \epsilon \) | \( S_r \) | \( K_{sat}, \text{cm/day} \) |
| 0.1042 | 2.7186 | 0.0728 | 13.1 | 0.0000 | 890.3 |

Rock Matrix Moisture Parameters

Moisture retention parameters and the exponent \( \epsilon \) for the rock matrix were taken from Klavetter and Peters (1986) for the Tiva Canyon Member of the Paintbrush Tuff. All functional and flow parameters used in SATURN for the Tiva Canyon Member are
Figure 3.7. Rock Valley soil moisture retention curve.

Figure 3.8. Effective saturation vs. log relative hydraulic conductivity for the Rock Valley soil.
Figure 3.9. Tiva Canyon Member moisture retention curve.

Figure 3.10. Effective saturation vs. log relative hydraulic conductivity for the Tiva Canyon Member for ɛ = 6.8.
shown in Table 3.3, and $\psi-\theta$ and $S_{\text{w}}-k_{\text{w}}$ plots are shown in Figures 3.9 and 3.10, respectively.

Table 3.3. Tiva Canyon Member functional and flow parameters.

<table>
<thead>
<tr>
<th>$\alpha$, cm$^{-1}$</th>
<th>$\beta$</th>
<th>$\gamma$</th>
<th>$\epsilon$</th>
<th>$S_r$</th>
<th>$K_{\text{rad}}$, cm/day</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.8210</td>
<td>1.5580</td>
<td>.3582</td>
<td>6.8</td>
<td>0.0020</td>
<td>0.0000838</td>
<td>0.0800</td>
</tr>
</tbody>
</table>

Rock-Fracture Moisture Parameters

NON-FILLED FRACTURES. For simulations assuming parallel plate fractures not containing any porous material, fracture conductivity values were calculated for 3 fracture apertures using the cubic law derived from application of the Navier-Stokes equation to viscous flow in fractures (Iwai, 1976). For an individual fracture,

$$K_f = \frac{\gamma b^2}{12\mu}$$  \hspace{1cm} (3.8)

where $b$ is the fracture aperture, $\gamma$ is the specific weight of water, and $\mu$ is the dynamic viscosity of water. For a system with a fracture spacing $s$, the system hydraulic conductivity, $K_s$, is given by (Iwai, 1976)

$$K_s = \frac{b}{s} K_f = \frac{\gamma b^3}{12\mu s}.$$

The assumption is made here that the flow of the system is the sum of the flow of the individual fractures, and that the fractures have smooth parallel walls (Iwai, 1976).

Fracture spacings were calculated from fracture frequencies and densities reported in the literature. Two methods of reporting fracture frequencies were evident: one method reported the number of fractures per unit length of core or surface traverse; the second method reported the number of fractures per cubic meter of rock, or fracture density. Due to fractures intersecting cores or traverses at angles less than 90°, the first method represents an apparent fracture frequency per unit length whereas the second method employs a correction which transforms the apparent frequency into a true frequency as a
first step to calculating the density. This corrected, or true, frequency is that which would occur if all of the fracture strikes were at right angles to the core or traverse (Scott et al., 1983). The corrected frequencies are always larger than the apparent frequencies.

Apparent frequencies measured along traverses on the crest of Yucca Mountain, reported by Spengler and Chornack (1984), range from 5.0 to 8.5 fractures per meter. This translates to spacings of 20.0 to 12.0 cm in between fractures, respectively. Several authors have reported fracture densities in the literature. Scott et al. (1983) gives values ranging from 1.2 to 8.2 fractures/m³ from traverses on the crest of Yucca Mountain. Scott and Castellanos (1984) and Spengler and Chornack (1984) report values ranging from 22.0 to 41.3 fractures/m³ measured from core samples taken from the tuff of the Tiva Canyon Member. After converting these densities to corrected frequencies (see Scott et al., 1983), fracture spacings ranging from 3.5 to 83.3 cm were calculated. From this, spacings of 5, 10, 40, 80 cm were originally considered for analysis. For the non-filled fracture flow analysis of Stage 3, only a fracture spacing of 5.0 cm was used since it was determined without further analysis that the equivalent porous medium method would not work. For the filled fractures, spacings of 10, 40, and 80 cm were used in the computer simulations. A spacing of 5 cm was not used for the filled fractures since simulations first done for the larger spacings gave adequate results.

Once the fracture system hydraulic conductivity was calculated, it was then added to the rock matrix conductivity for potentials greater than the wetting potential for the fracture aperture being considered. At potentials below the wetting potential the fracture conductivity was assumed to be zero and the total conductivity, $K_T$, was equal to the matrix conductivity. This can be expressed as

$$K_T = \begin{cases} K_r + K_r & \text{for } \psi \geq \psi_{fw} \\ K_r & \text{for } \psi < \psi_{fw} \end{cases}$$

where $K_r$ is the rock matrix conductivity and $\psi_{fw}$ is the fracture wetting potential. Since hysteresis effects were not considered in this analysis, the wetting potential was taken as the air entry potential, given here as (Evans and Huang, 1983)
\[ \psi_{f_w} = -\frac{2\eta \cos \omega}{b \gamma} \] 3.11

where \( \eta \) is the surface tension of water, taken here to be 71.9 gm/s\(^2\), \( \gamma \) and \( b \) are the same as above, and \( \omega \), the wetting angle, is equal to zero in this case.

By also knowing the moisture content above and below the fracture wetting potential, a \( S_\text{eff} - k_\text{w} \) curve was constructed containing effective saturation-conductivity relationships pertaining to potentials below and above the fracture wetting potential, shown as parts '1' and '2' in Figure 3.11, respectively. Two separate values of \( \epsilon \), for use in SATURN, were then obtained by solving for \( \epsilon \) in equation 2.8 using the two separate data sets representing the two portions of the curve, with \( \epsilon \) already known for part 1, being that of the tuff of the Tiva Canyon Member.

Similarly, two sets of data were used to derive functional parameters used in the \( \psi-\theta \) relationship of Equation 2.9. One set will be that of the equivalent porous medium system for potentials above \( \psi_{f_w} \), indicated by '2' in Figure 3.12, and the other set that of the tuff of the Tiva Canyon Member for potentials below \( \psi_{f_w} \), indicated by '1' in Figure 3.12. As can be seen in Figure 3.12, the addition of the fractures to the overall moisture content above the wetting potential is quite small with the Tiva Canyon Member curve and the equivalent porous medium curve essentially indistinguishable.

In order to accommodate the jump from part 1 to part 2 during simulation, at the time at which the fracture wetting potential was reached during wetting indicating saturation of the fractures, the computer simulation was restarted with the set of parameters representing fracture flow. If the pressure head at the fracture opening dropped below \( \psi_{f_w} \), then the simulation was restarted using only the matrix properties until \( \psi_{f_w} \) was reached again, at which time the properties were switched, and so on.

Three different scenarios were examined in this manner, representing three different fracture apertures which were determined by the results of the various summer and winter scenarios simulated in Stage 2. Fracture apertures were chosen so as to saturate at a potential 15 cm below the maximum potential reached at the soil-rock interface determined in Stage 2 so as to ensure availability of water to the fracture after it saturated.
Figure 3.11. Equivalent porous medium hydraulic conductivity curve showing rock with fractures, portion (2), and rock only, portion (1).

Figure 3.12. Equivalent porous medium moisture retention curve showing rock with fractures, portion (2), and rock only, portion (1).
FILLED FRACTURES. Simulations using filled fractures were carried out with the bedrock at 7 cm, fracture spacings at 10.0, 40.0, and 80.0 cm, and a fracture aperture of 0.2 cm. This size fracture was large enough to model directly without first converting to an equivalent porous medium and also large enough to contain the largest particle fractions of all the porous media used. Three different types of porous media were used to represent fracture filling: a Touchet silt loam (Brooks and Corey, 1964; data from Mualem, 1976); a fine sand (Vachaud et al., 1974; data from Case et al., 1983); and a Cecil loamy sand (Bruce, 1972; data from Case et al., 1983). Soil-moisture retention parameters were obtained from the van Genuchten curve fitting routine (van Genuchten, unpublished manuscript); the fine sand was a simultaneous fit of $\psi-\theta$ and $\theta-K$ data whereas the Touchet silt loam and Cecil loamy sand were fit using $\psi-\theta$ data only. The saturated hydraulic conductivity values, $K_{sat}$, were obtained from the respective literature sources. The parameter $\epsilon$, used in equation 3.7, was determined in the manner described above for the Rock Valley soil using $\theta-K$ data generated from the curve fit for the fine sand and original data provided by Case et al. (1983) for the Cecil loamy sand. For the Touchet silt loam, $\epsilon$ was taken from Brooks and Corey (1964).

All functional and flow parameters for the Touchet silt loam, the fine sand, and the Cecil loamy sand are given in Table 3.4. Plots of $\psi-\theta$ and $S_{eff}-K_{rw}$ curves are shown in Figures 3.13 and 3.14, respectively. The curves for all 3 porous media, plus the curve for the Rock Valley soil, were placed on one plot to provide for easy comparison. Plots of the individual data sets and their respective fitted curves are provided in Appendix A.

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>$K_{sat}$, cm/day</th>
<th>$S_r$</th>
<th>$n$</th>
<th>$\alpha$, cm$^{-1}$</th>
<th>$\beta$</th>
<th>$\gamma$</th>
<th>$\epsilon$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Touchet silt loam</td>
<td>47.23</td>
<td>0.3013</td>
<td>0.469</td>
<td>0.0065</td>
<td>20.9873</td>
<td>0.1004</td>
<td>4.67</td>
</tr>
<tr>
<td>Fine sand</td>
<td>228.00</td>
<td>0.0</td>
<td>0.34</td>
<td>0.0251</td>
<td>2.1400</td>
<td>0.6811</td>
<td>5.0</td>
</tr>
<tr>
<td>Cecil loamy sand</td>
<td>238.0</td>
<td>0.0</td>
<td>0.40</td>
<td>0.2519</td>
<td>1.3406</td>
<td>0.2215</td>
<td>9.0</td>
</tr>
</tbody>
</table>
Figure 3.13. Moisture retention curves for 3 fracture fillings and the Rock Valley soil.

Figure 3.14. Effective saturation vs. log hydraulic conductivity for 3 fracture fillings and the Rock Valley soil.
Initial Moisture and Lower Boundary Conditions

Soil moisture pressure head measurements were not yet available for Yucca Mountain at the time of this study, but this type of data has been extensively collected for specific sites in Rock Valley for the years 1971 through 1975 (Turner, 1972, 1973, 1974, 1975, 1976). Rock Valley was selected as representative of the Mojave Desert in the US/IBP Desert Biome studies conducted by Turner and is located about 20 miles southeast of Yucca Mountain. The requirements for initial conditions specified for the summer and winter scenarios were that they fall within the range of values for each season providing a representative value for each time period, and yet be towards the wet end to allow for a worst case infiltration scenario.

Summer pressure heads varied from about -15,000 cm to -80,000 cm within the upper 45 cm of soil, with the soil generally drier towards the surface. Taking this into consideration, for initial conditions the upper boundary was set at -50,000 cm and the lower boundary at -30,000 cm with the remaining values in between being specified assuming a uniform gradient between the two boundaries. The lower boundary condition was specified at -30,000 cm.

Winter values varied from about -100 cm to -10,000 cm within the upper 45 cm of soil with the soil often wetter towards the surface, but with no set moisture distribution evident. From these considerations, a value of -2000 cm was chosen as the uniform initial condition for the winter scenario. The moisture content corresponding to this pressure head value is 0.110 which, at a porosity of 0.315, is equivalent to 22.0 cm of water in the 2 meter soil column. This value is high when compared to the estimated average annual precipitation of 15.0 cm (Montazer and Wilson, 1984) but was thought useful when considering a worst case infiltration scenario. The lower boundary condition was specified at -2000.0 cm.
Stage 1

Plots of depth to the drying front and moisture content of the downward moving wetting pulse in excess of the antecedent moisture conditions for both the summer and winter scenarios of Stage 1 are shown in Figures 4.1 and 4.2, respectively. For a depth of drying front at a given time, there is an effective moisture content, expressed as an equivalent moisture depth in centimeters, in the wetting pulse at that time below the depth of the drying front.

In the case of the summer simulation, with a total precipitation of 1.0 cm over 0.042 days and a PET of 1.30 cm/day, the downward movement of the drying front begins to slow down after about 1 day. The switch from a constant evaporation flux to a constant head boundary condition at the upper boundary of the column, representing the change from constant-rate to falling-rate stage evaporation, occurs at 0.150 days after the start of precipitation. Similarly, the decrease in the wetting pulse moisture content also slows at comparable times to that of the depth to drying front. From Figure 4.1, it was decided to position the deepest bedrock surface at 10 cm, which corresponds to an equivalent moisture depth of about 0.16 cm in the wetting pulse below this depth. A depth of 10 cm represents the position of a transient zero flux plane which, at the time of its occurrence, has 0.16 cm of water moving downward below it. This is a very small amount of water left in the wetting pulse; consequently, if the maximum bedrock surface was much deeper than 10 cm, interaction between the bedrock and the infiltrating water would be precluded; moreover, assessment of this interaction was one of the purposes of this study. A simulation was also carried out with the bedrock surface positioned at 5 cm.

The corresponding winter simulation was conducted with precipitation totaling 1.40 cm over 0.291 days, and a PET of 0.37 cm/day. The switch from a constant-rate to a falling-rate stage of evaporation occurred 2.05 days after the start of precipitation. As
Figure 4.1. Summer period equivalent moisture depth and drying front depth vs. time.

Figure 4.2. Winter period equivalent moisture depth and drying front depth vs. time.
shown in Figure 4.2, the drying front does not flatten out as in the summer simulation. This is most likely due to the higher antecedent moisture conditions encountered in the winter simulation, providing more water for evaporation and not constraining the depth of the drying front by the amount of moisture available. The wetting pulse equivalent moisture depth curve does begin to flatten out after about 3 days. Since the drying front depth curve does not flatten out, the maximum bedrock depth for the winter scenario was determined by choosing the depth at which the wetting pulse equivalent moisture depth had reached a value close to that occurring for the summer scenario maximum bedrock depth. From Figure 4.2, a depth of 65 cm was chosen, representing a wetting pulse equivalent moisture depth of approximately 0.13 cm. Two other depths to bedrock were also simulated, one at 10 cm for comparison to the summer simulation, and one at 7 cm which provides for conditions approaching saturation by the end of the precipitation event.

Stage 2

As stated above, 2 scenarios with a non-fractured bedrock surface at 5 and 10 cm depths were simulated using summer atmospheric and initial conditions and 3 scenarios with bedrock at 7, 10, and 65 cm depths were simulated under winter conditions. The sequence of events and time periods considered were specified in Chapter 3.

Values of pressure head at the soil-bedrock interface versus time are plotted in Figure 4.3 for both summer simulations. Table 4.1 gives values of five useful measurements resulting from each simulation. Most notable are the maximum head values reached at the soil-rock interface for each simulation with the 5 cm depth run reaching a pressure head of -30.8 cm and the 10 cm depth run reaching a pressure head of -591.4 cm. For the 5 cm depth run, this translates to saturation of a maximum fracture aperture of $4.76 \times 10^{-3}$ cm which corresponds to a single fracture hydraulic conductivity of 159.4 cm/day. For the 10 cm depth run this translates to the saturation of a fracture aperture of only $2.48 \times 10^{-4}$ cm which corresponds to a single fracture hydraulic conductivity of 0.4 cm/day.
Figure 4.3. Summer period pressure heads at the soil—rock interface for 5 and 10 cm soil depths.

Table 4.1. Stage 1 summer scenario simulation results.

<table>
<thead>
<tr>
<th>Depth to Bedrock</th>
<th>Maximum Head</th>
<th>Maximum Soil Sat</th>
<th>Time of Flux Reversal at Interface</th>
<th>Maximum Rock Sat</th>
<th>Time of Change in Evaporation BC</th>
</tr>
</thead>
<tbody>
<tr>
<td>5 cm</td>
<td>-30.8 cm</td>
<td>.7914</td>
<td>.0421 d</td>
<td>.0060</td>
<td>.9448 d</td>
</tr>
<tr>
<td>10 cm</td>
<td>-591.4 cm</td>
<td>.4424</td>
<td>.1132 d</td>
<td>.0059</td>
<td>.6252 d</td>
</tr>
</tbody>
</table>

For the three winter simulations, the initial conditions and precipitation were also identical for each simulation. Values for pressure head at the soil-rock interface versus time are shown in Figure 4.4 for the 7 and 10 cm soil depth to bedrock scenarios. Values measured for the three winter simulations are reported in Table 4.2. The simulation with the soil-rock interface at the 7 cm depth was the most effective in producing the highest pressure head values which would allow saturation of the largest fractures. At the maximum head value of -5.18 cm, a fracture aperture of 0.028 cm could be saturated.
reflecting a hydraulic conductivity of $5.6 \times 10^3$ cm/day. The pressure head at the soil-rock interface for the 65 cm depth bedrock surface only reached a pressure head value of -1088.0 cm. The individual fracture hydraulic conductivity corresponding to this head value is 0.13 cm/day, much smaller than the fracture hydraulic conductivities possible for the 7 and 10 cm simulations.

Table 4.2. Stage 1 winter scenario simulation results.

<table>
<thead>
<tr>
<th>Depth to Bedrock</th>
<th>Maximum Head</th>
<th>Maximum Soil Sat</th>
<th>Time of Flux Reversal at Interface</th>
<th>Maximum Rock Sat</th>
<th>Time of Change in Evaporation BC</th>
</tr>
</thead>
<tbody>
<tr>
<td>7cm</td>
<td>-5.2cm</td>
<td>.9873</td>
<td>.292d</td>
<td>.0220</td>
<td>&gt;3d</td>
</tr>
<tr>
<td>10cm</td>
<td>-31.4cm</td>
<td>.7886</td>
<td>.297d</td>
<td>.0220</td>
<td>&gt;3d</td>
</tr>
<tr>
<td>65cm</td>
<td>-1088.0cm</td>
<td>.3921</td>
<td>4.707d</td>
<td>.0197</td>
<td>?</td>
</tr>
</tbody>
</table>
From Tables 4.1 and 4.2 it can be seen that the effective saturation of the rock only slightly increases above the residual saturation and that flow into the rock is negligible over the short time periods which it is in contact with high moisture contents and pressure heads. It can also be seen from these two tables that even though the time of flux reversal at the element above the soil-rock interface is later for deeper bedrock surfaces, indicating water is moving down for a longer period of time in these cases, the resultant effective saturation values are smaller. This means that evaporation has had more time to deplete the wetting pulse of water and that shallower soil-to-bedrock depths would allow for maximum wetting of fractures. Also evident from Tables 4.1 and 4.2 and Figures 4.3 and 4.4 is that the winter scenario will also produce more favorable conditions for infiltration into the soil and percolation into fractures than will the summer scenario.

Stage 3

NON-FILLED FRACTURES. Fracture apertures considered for analysis were calculated using the maximum pressure heads obtained as shown in Tables 4.1 and 4.2. Since the best chance for infiltration into fractures occurred for the 7 cm bedrock depth for the winter scenario, this was the first scenario considered for this stage of the analysis and only winter conditions were used for the remainder of the analysis. Simulations were planned for the 7 cm bedrock depth for three fracture apertures corresponding to the three maximum pressure heads specified in Table 4.2. Two simulations were then to be performed for the 10 cm bedrock depth for the two fracture apertures corresponding to the maximum pressure heads reached at the 10 and 65 cm bedrock depths. Finally, only one simulation was planned for the 65 cm bedrock depth since larger fractures, corresponding to the larger pressure heads reached for the other surfaces, would not be saturated. This procedure would have allowed comparison of both flow into fractures of different apertures at similar depths and under similar conditions, and flow into fractures of similar aperture but at different depths. As will be explained presently, this method of numerical analysis proved unfeasible.

As stated in Chapter 3, the pressure heads used to determine the fracture apertures to be modeled were 15 cm below the maximum pressure head determined in Stage 2. It
was thought if the fractures were chosen so as to saturate at the maximum pressure head, they would only transmit water for a short period before "turning themselves off" by lowering the head in the immediate vicinity. Pressure heads thus chosen were -20.0 cm, -46.0 cm, and -606.0 cm, corresponding to fracture apertures of 7.34x10^{-3} cm, 3.19x10^{-3} cm, and 2.42x10^{-4} cm, respectively. Functional and flow parameters are given in Table 4.3 for the three apertures considered for a fracture spacing of 5 cm. Plots of $\psi - \theta$ and $S_{ef} - k_{mf}$ curves are plotted in Figures 4.5 and 4.6, respectively. All three curves, referring to the three apertures under consideration, were plotted together along with the curve for the tuff of the Tiva Canyon Member so as to provide for easy comparison.

Table 4.3. Equivalent porous media functional and flow parameters.

<table>
<thead>
<tr>
<th>$b$, cm</th>
<th>$\psi_{fe}$, cm</th>
<th>$K_{sat}$, cm$^2$ day$^{-1}$</th>
<th>$S_r$</th>
<th>$n$</th>
<th>$a$, cm$^{-1}$</th>
<th>$\beta$</th>
<th>$\gamma$</th>
<th>$\epsilon$</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.34x10^{-3}</td>
<td>-20.0</td>
<td>0.554084</td>
<td>0.0020</td>
<td>0.0812</td>
<td>0.0001</td>
<td>0.7779</td>
<td>1.5405</td>
<td>0.3726</td>
</tr>
<tr>
<td>3.19x10^{-3}</td>
<td>-46.0</td>
<td>0.046084</td>
<td>0.0020</td>
<td>0.0806</td>
<td>0.0011</td>
<td>0.8201</td>
<td>1.5550</td>
<td>0.3536</td>
</tr>
<tr>
<td>2.42x10^{-4}</td>
<td>-606.0</td>
<td>0.000104</td>
<td>0.0020</td>
<td>0.0801</td>
<td>1.0000</td>
<td>0.7991</td>
<td>1.4425</td>
<td>0.3930</td>
</tr>
</tbody>
</table>

This attempt at modeling fractures as parallel plates, which were converted to an equivalent porous medium, proved unfeasible. Due to the large gradients encountered when SATURN was restarted with the fracture parameters, significant jumps in the velocities at the fracture openings occurred. These high velocities, an increase of 10 orders of magnitude for small fractures and 15 orders of magnitude for the larger fractures, drastically and immediately lowered the pressure head at the fracture opening enough to effectively "turn off" the fractures. The pressure head was then allowed to increase back to $\psi_{fe}$, turning the fractures on again, with only a very small amount of water entering the fractures before they turned off again. This process had every indication of continuing indefinitely for both the large and small fractures even though the total hydraulic
Figure 4.5. Equivalent porous medium moisture retention curves for rock with 3 different fracture apertures and rock alone. All curves coincide.

Figure 4.6. Equivalent porous medium hydraulic conductivity curves for rock with 3 different fracture apertures and rock alone.
conductivity value using the smallest fractures was nearly the same as that for the rock itself. The inability of flow through fractured bedrock to be analyzed in this way was attributed to the simplified fracture flow theory and the method of its application and not to the numerical solution technique itself. As stated in Chapter 3, simulations were also to be done at spacings of 5, 10, 40, and 80 cm. Due to the inability of the fractures to be modeled in this manner, simulations were only done for a 5.0 cm spacing before it was determined this approach was not useful.

FILLED FRACTURES. Of the three porous media used as fracture fillings, two essentially did not transmit any water. The depth of water movement into the fractures in the case of both the fine sand and the Touchet silt loam was less than 1 cm over a period of three days. The reason for this is evident from examination of Figures 3.13 and 3.14. From Figure 3.14 it is evident that the hydraulic conductivities at lower effective saturations are higher for the Touchet silt loam and the fine sand than for the Cecil loamy sand. From this it would seem that the Touchet silt loam and the fine sand should be transmitting more water than the Cecil loamy sand, at low effective saturations. But from inspection of Figure 3.13 it can be seen that the moisture content, and thus the saturation, of the fine sand does not increase appreciably until pressure heads of about -150 cm are reached. At first it appears that this is not true for the Touchet silt loam which shows high moisture content with low pressure head. By referring to Table 3.4, it can be noted that the Touchet silt loam has a residual saturation of 0.3013. Using the definition of the effective saturation presented in Chapter 2, it becomes clear that only at a pressure head of around -300 cm does the effective saturation begin to show a significant increase. Restated, the hydraulic conductivity - effective saturation relationship for the Touchet silt loam and fine sand represents a very narrow pressure head range.

On the other hand, the Cecil loamy sand and the Rock Valley gravelly loamy sand both show higher moisture contents at lower pressure heads. This increase in moisture content at low pressure heads, which is directly related to the effective saturation since both soils have a residual moisture content of 0.0, permits the hydraulic conductivity to increase earlier in the wetting process than what is seen for the Touchet silt loam and
the fine sand. If the Touchet silt loam and the fine sand had not been constrained by the short duration of the infiltration event, the fracture column would eventually wet with the result of increasing hydraulic conductivities and an increase in water movement into the fracture. The remaining analysis deals with the Cecil loamy sand only.

The Darcy velocities of the top fracture element plotted versus time for the 80 cm fracture spacing case are shown in Figure 4.7 for the Cecil loamy sand. Immediately evident from this plot are the severe velocity fluctuations occurring from just after 0.3 days up to 1.0 days. In order to obtain a stable solution for times when water is flowing in the fracture material, node spacings had to be set at 0.1 and 0.2 cm for the upper part of the fracture (see Appendix 1). Since there was an array maximum of 1666 nodes allowed in SATURN, larger node spacings had to be used below the 17 cm depth, 10 cm below the soil-rock interface. At 0.3245 days the wetting pulse in the fracture is at this depth and is beginning to wet elements 1 cm high. For an element this size, it takes longer for the effective saturation to increase over the whole element; therefore it takes longer for the hydraulic conductivity to increase. Since the flow in the wetting element is not keeping up with the flow in the elements above it, the water in the upper part of the fracture is backed up and flow begins to slow down. When the hydraulic conductivity finally increases along with the effective saturation in the wetting element, high gradients formed due to the build up of water above the wetting element causes the velocity to increase rapidly and the water surges downward into the wetting element. This process was entirely an artifact of the finite element code for fracture vertical node spacings larger than 0.2 cm. As can be seen in Figure 4.7, this process is repeated again and again as each element wets. For this reason, only the results up to 0.3245 days were considered usable and the remaining runs were limited to this time period.

Figure 4.8 shows the velocities of the uppermost element in the fracture for the three simulations done at 10, 40, and 80 cm fracture spacings. As can be seen, the velocities appear relatively unaffected for the 40 and 80 cm spacings but are somewhat diminished for the 10 cm spacing. Overall, the velocities increase rapidly after about 0.2 days until infiltration is stopped, then they immediately begin to decrease up to the time the simulation ended. The shape of the curve up to where precipitation ends is probably
Figure 4.7. Downward Darcy velocities for the uppermost fracture element; s=80 cm, filling is Cecil loamy sand.

Figure 4.8. Downward Darcy velocities at the top fracture element for three different fracture spacings.
related to the thickness of the overlying soil layer and the velocity of the wetting front. The 10 cm fracture spacing velocity shows a more rapid decrease following termination of infiltration than do the rest.

The effect of spacing on the amount of water in the wetting pulse within the fracture can be seen in Figure 4.9. Plots of the equivalent moisture depth within the wetting pulse for each of three time periods for different fracture spacings are shown. The lower and upper time periods are roughly equally spaced before and after the end of precipitation at 0.291 days which is the end of infiltration and the beginning of evaporation. As can be seen, with time the 10 cm spacing shows slightly lower fracture equivalent moisture depth within the wetting pulse than do the others. The fracture wetting pulse equivalent moisture depths for the 40 and 80 cm fracture spacings are fairly close at 0.3245 days indicating that fracture spacing has an effect only at lower spacings and even then the effect appears minimal. Had the simulations been run longer with valid results, flux into the fractures spaced 10 cm apart may have been more restricted with time, whereas flux into wider spaced fractures may have continued at a relatively higher rate for a longer period of time.

![Figure 4.9. Fracture wetting pulse moisture depth vs. fracture spacing for 3 specified times.](image-url)
Figure 4.9 also indicates that the amount of water in the wetting pulse within the fracture shows a larger increase after precipitation is stopped. Since the depth of the wetting pulse within the fracture increases at a higher rate than the equivalent moisture depth, the actual average moisture content within the wetting pulse shows a larger increase from 0.257 to 0.291 days than from 0.291 to 0.3245 days.

Figures 4.10, 4.11, and 4.12 show the velocity field for the 10, 40, and 80 cm spacing simulations, respectively, at 0.3245 days. These three plots indicate that the drying front has moved down only about 0.2 cm at this time whereas the drying front reached the bedrock surface almost immediately after evaporation began for the case where there were no fractures (see Table 4.2). For the case with fractures this phenomenon was attributed to the downward velocity component of water created by flow into the fractures being larger than the upward evaporation-induced velocity component. In the case without fractures the water was essentially not moving as the soil neared saturation; therefore, the application of an evaporation flux, even though very small, caused the

![Figure 4.10. Soil zone velocity vectors for s=10 cm.](image-url)
Figure 4.11. Soil zone velocity vectors for $s=40$ cm.

Figure 4.12. Soil zone velocity vectors for $s=80$ cm.
water to immediately respond by moving upward. In addition, the overall velocity field evident in Figure 4.10 differs markedly from the velocity fields shown in Figures 4.11 and 4.12 in that there is a larger downward vertical velocity component present in Figure 4.10. This is probably related to the effects of a limited water supply for a narrower fracture spacing.

Finally, even though the fractures spaced 10 cm apart show decreased flux into the fractures due to their proximity to other fractures, the effect is quite small during the time interval considered here. If the amounts of percolation were compared on an equal area basis, the system with fractures spaced 10 cm apart had percolation rates 7.3 times that for the system with fractures spaced 80 cm apart and 3.7 times that for the system with fractures spaced 40 cm apart. This indicates there is an almost linear increase in infiltration with a decrease in fracture spacing.
CHAPTER 5: CONCLUSIONS

Numerical simulation of infiltration into soils and percolation into fractured bedrock proved valuable in analyzing parameter control over flow within the system. Soil and rock-matrix flow properties were assumed constant while initial, or antecedent, moisture conditions, precipitation duration and intensity, and potential evapotranspiration were varied between two general season-related scenarios. Soil depth to bedrock variability was constrained by interaction of the bedrock with the infiltrating water. With respect to fracture flow, the simplified equivalent porous medium approach proved unfeasible due to fluctuations in the local pressure heads at the time of fracture wetting. Explicit modeling of fractures filled with a porous medium allowed analysis of the affects of different fracture fillings on fracture flow in addition to fracture spacing effects.

The numerical solution technique utilized by SATURN performed well, especially under extreme conditions of dry soil and instances where two adjacent materials differed greatly in their material properties. Simulation of flow into filled fractures went smoothly provided the vertical node spacing was at a minimum (0.1 - 0.2 cm) which was generally true for all of the simulations. Attempts at analyzing infiltration into fractured rock using an equivalent porous medium approach were not successful, but the performance of the numerical code was not responsible for this; instead, failure was attributed to the simplified fracture flow theory used here which contained nuances which could not be realistically handled by any code.

Seasonal variation of PET and antecedent moisture conditions had a profound effect on infiltration and percolation. The combination of high PET and low antecedent moisture conditions on the order of 1.37 cm/day and 6.0 percent, respectively, as seen in the summer, effectively inhibited percolation within the soil. Even small winter PET rates of 0.02 cm/day greatly influenced water movement within the soil above a rock surface. From this it could be expected that flow into rock with relatively small fractures, or fractures with only smaller conducting segments, and thus smaller downward velocities, would be influenced by PET more than rock with larger fractures.
Soil depth to bedrock variability indirectly had a strong influence on the pressure head distribution at the soil-rock interface. For the winter scenario, the pressure head at the rock surface positioned at 65 cm reached only -1088.0 cm while a pressure head of -5.2 cm was attained with the rock surface positioned at 7 cm. This difference was due to the longer period of time PET had to withdraw water from the wetting pulse as it percolated downward towards the 65 cm depth rock surface. For the summer scenario, the bedrock surfaces had to be much shallower in order for significantly high pressure heads to be reached partly because total precipitation was less for the summer scenario, but largely because PET was approximately one order of magnitude larger for the summer scenario than for the winter scenario.

The time dependency of moisture flux played a critical role in potential distributions above a bedrock surface. Higher precipitation intensity on the order of 24 cm/day, or longer precipitation duration on the order of 0.291 days, especially coupled with shallow bedrock depths, provided for higher pressure heads at the soil bedrock interface which, theoretically, would allow saturation of larger fractures and increased fracture percolation, while moisture movement into the rock matrix was negligible even at higher pressure heads. Since the saturation of fractures was dependent on pressure head and the pressure head at the soil-rock interface was in turn time dependent, fractures of varying aperture saturated at different times with flux into the fractures proportional to the cube of the aperture for the case of non-filled fractures.

The duration of a precipitation event was also seen to affect the flow of water into individual fractures. For the winter scenario considered, Darcy velocities within the fractures were seen to increase rapidly after approximately 0.2 days. For times shorter than this, flow into the fracture was insignificant. For longer times, flow increased sharply up until the time infiltration ceased at 0.291 days at which time the fluxes into the fractures dropped off markedly, with fluxes decreasing more rapidly for fractures spaced 10 cm apart. This rapid flux decrease suggests that maximum flow into the fractures probably occurs at the moment before precipitation ceases for the precipitation duration considered here. In general, for short precipitation events, even of high intensity, the event may end before the velocity in the fracture has reached the steepest portion of the
velocity curve effectively cutting short any possibility of significant flux into the fracture.

Another factor exerting an influence on the flux of water into fractures was the type of soil within the fracture. The conductivity of the soil was dependent on the effective saturation while its related property, moisture content, was dependent on the pressure head. The Touchet silt loam and the fine sand exhibited low conductivities at the pressure head distributions achieved. The Cecil loamy sand, on the other hand, had higher effective saturations at the same pressure heads, thus allowing higher conductivities at earlier times in the wetting phase. The combined effect of low conductivities, and consequently low velocities, as seen with the Touchet silt loam and the fine sand, along with the strong influence of PET especially on water with slow downward velocities, could effectively prevent water from entering the fracture even if the soil was only at the opening of or for a short distance within the fracture. This effect may control flux into smaller fractures since only smaller soil particle fractions can migrate into them. Since porous media composed of smaller particle fractions such as silts and clays generally have lower conductivities, fractures with smaller apertures may inhibit flow leaving fractures with larger apertures as the primary flow conduits.

With the limited time considered for the soil-filled fracture simulations, the decrease in fracture spacing had little effect on flow into a single fracture in relation to the influence of neighboring fractures. For a time of 0.3245 days following initiation of infiltration for a bedrock depth of 7 cm, fractures spaced 10 cm apart had wetting pulse equivalent moisture depths 7.3 times and 3.7 times that of fractures spaced 80 cm and 40 cm apart, respectively, on a per area basis. Consequently, flux into a fractured surface was seen to increase almost linearly with increasing fracture frequency. The trend over time was that of a greater decrease of infiltration into the more closely spaced fractures. Had the simulations been able to run longer, flux into fractures spaced 10 cm apart may have been more restricted with time, whereas flux into wider spaced fractures may have continued at a relatively higher rate for a longer period of time.

Overall, taking into account the assumptions applied to the flow analysis in this study, a generalized scenario allowing for maximum infiltration and percolation can be
formulated. Atmospheric conditions and antecedent moisture conditions most favorable for infiltration and percolation occurred during the winter when PET was low, antecedent moisture conditions were high, and precipitation, although not of as high intensity as in the summer, was of significant duration to minimize the effects of PET. This does not preclude infiltration during the summer; the probability of favorable infiltration conditions would not be as high though. Spatially, areas exhibiting thinner soil coverings and higher fracture frequencies would be more conducive to maximum percolation. Longer precipitation events would favor saturation of larger fractures and also favor wetting of fractures filled with soil of low initial-saturation hydraulic conductivities, ultimately producing larger fluxes.

With the preceding discussion in mind, several suggestions for further investigations can be made. The geometry of a fracture, both at its opening and within the fracture itself, and the respective flow properties, are areas which need to be addressed and would have to be considered a primary concern for analysis of any fractured rock infiltration study. Slope is important as well and may have a significant effect on pressure head distributions at the soil-rock interface which could effect flow into fractures, especially with respect to fracture orientation. Knowledge of runoff characteristics in relation to slope, soil-moisture flow properties, and soil presence or absence would also be required if a complete water budget analysis were to be carried out. The effect of an anisotropic porous medium overlying the bedrock should also be considered with respect to the ability of the soil to transmit water laterally toward fractures. Finally, to obtain a thorough picture of infiltration and percolation in relation to a fractured rock setting, the coupling of water flow with air, vapor, and heat flow would have to be considered.


APPENDIX A
Figure A.1. Moisture retention curve for the Cecil loamy sand, with data.

Figure A.2. Effective saturation vs. log relative hydraulic conductivity for Cecil loamy sand data and curve fit.
Figure A.3. Moisture retention curve for the fine sand, with data.

Figure A.4. Effective saturation vs. log relative hydraulic conductivity for the fine sand data and curve fit.
Figure A.5. Moisture retention curve for Touchet silt loam, with data.

Figure A.6. Effective saturation vs. log relative hydraulic conductivity for Touchet silt loam data and curve fit.