NEAR SURFACE INFILTRATION MEASUREMENTS AND THE
IMPLICATIONS FOR ARTIFICIAL RECHARGE

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DEDICATION

This thesis and all of the hours spent working to complete this project is first and foremost dedicated to my family. From my grandmother Doris Tinsley who knew everything about everything and always inspired me to keep learning, to my granddad Lou Tinsley and my parents Walter and Lou Anne who set an example of success through hard work, to my sister Lindsey and cousins Pat and Bridget who were constant reminders to be a good role model. And finally, this is dedicated to the teachers, coaches, friends, and anyone else that instructed me, supported me, and guided me to this point.
ABSTRACT OF THE THESIS

Near Surface Infiltration Measurements and the Implications for Artificial Recharge

by

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Double-ring infiltrometer tests were done at eight sites in alluvial deposits near Joshua Tree, California to measure surface infiltration rates and aid in evaluating the study area for construction of infiltration ponds for groundwater recharge. Surface infiltration rates ranged from $3.61 \times 10^{-3}$ ft$^3$/s within the wash north of Pinto Mountain Fault, to $6.68 \times 10^{-4}$ ft$^3$/s south of the fault outside of an active wash. Hydraulic conductivity of alluvium was estimated from infiltration data using the computer program VS2D. Hydraulic conductivities estimated using VS2D ranged from $1.74 \times 10^{-4}$ ft/s to $4.63 \times 10^{-5}$ ft/s and were within the range of fine sand to silty sand. Hydraulic conductivities also were measured by laboratory analysis of surface cores from each site and ranged from $7.59 \times 10^{-5}$ ft/s to $4.33 \times 10^{-5}$ ft/s. Hydraulic conductivities computed from particle size data using the pedotransfer function (PTF) based computer program Rosetta ranged from $6.71 \times 10^{-5}$ ft/s to $1.73 \times 10^{-5}$ ft/s. Laboratory hydraulic conductivities matched the field data better than hydraulic conductivities calculated using Rosetta, but neither method correctly predicted infiltration rates measured using the double-ring infiltrometer. Evidence for a surficial low-permeability layer was seen in the data for six of the eight test sites, suggesting the formation of runoff depositional crusts and sieving crusts during intermittent wash flow. Surficial crusts can be removed before pond infiltration, but will distort infiltration measurements and should be accounted for. Simulation results using VS2D were sensitive to low-permeability layers at shallow depths, 0 to 4 feet below land surface, within the alluvium. Results were not sensitive to the thickness of the layer. The length of the infiltrometer test should be long enough to ensure detection of any subsurface layering. VS2D was used to simulate infiltration from one-acre spreading ponds on either side of the Pinto Mountain Fault to provide an estimated range of recharge times for the study area. The time required before the wetting front reaches the water table 400 feet below land surface was 14 – 60 days, implying that the area may be suitable for recharge through surface infiltration. Model simulations do not account for low-permeability layers within the alluvium or decreases in hydraulic conductivity with depth caused by compaction, consolidation, or changes in subsurface geology. As a consequence actual infiltration rates will probably be less than simulated rates. A more extensive survey of the subsurface geology and hydraulic properties of the study area including test drilling would help to assess subsurface conditions that would inhibit infiltration and recharge.
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CHAPTER 1
INTRODUCTION

In arid environments where surface water is intermittent or non-existent, communities often rely on groundwater as their sole source of water supply. Although the quantity of water in storage within aquifers in arid areas may often be large, negligible rainfall in arid areas means that aquifer recharge is limited. To complicate the issue, in areas where groundwater pumping exceeds recharge, water levels in aquifers may be declining. As populations and water needs grow, increased pumping can further decrease water supply.

Artificial recharge is being increasingly used to supplement groundwater supplies in arid areas of the southwestern United States (Flint and Ellett, 2004; Izbicki et al., 2007). Techniques for artificial recharge can be separated into three groups: direct injection, vadose zone infiltration, or surface infiltration from ponds (ASCE, 2001; Bouwer, 2002; Reddy, 2008; Topper et al., 2004; WDNR, 2006). Artificially recharged water may be locally derived runoff, treated wastewater, or imported to the area from sources outside the immediate area. Direct injection techniques involve wells drilled to depths below the water table where water can be injected directly to the saturated zone of the aquifer. Confined aquifers or areas with low permeability surficial materials are well suited for direct injection techniques (Bouwer, 2002; Reddy, 2008). Wells screened within the unsaturated zone and/or infiltration pits/shafts are used for vadose zone infiltration. Vadose zone infiltration is used for unconfined aquifers where surficial material is impermeable or has a low permeability (Bouwer, 2002; Reddy, 2008). Surface infiltration techniques are suitable for moderately to highly transmissive, unconfined aquifers overlain by surficial materials having high infiltration rates are high (Bouwer, 2002; Reddy, 2008). High surface infiltration rates enable rapid movement of water into the ground, and in the absence of impermeable subsurface layers, continued rapid movement through the unsaturated zone. Aquifers with moderate to high transmissivity will allow lateral movement of the water once it reaches the water table; preventing mounding and the associated decrease in rate of vertical water movement.
Constructed ponds have been widely used for surface infiltration (Asano, 2006; Bouwer, 2002; Ferris et al., 1954; Kimrey, 1989; Lee et al., 1992; Reddy, 2008; Sophocleous, 2004). There are numerous successful recharge ponds within southern California (Lee et al., 1992; Moreland, 1970; Reddy, 2008; Stamos et al., 2001; YVWD, 2000). Effectiveness, treatment of the water to remove bacterial and other contaminants during movement through the unsaturated zone, and the minimal need for maintenance are positive features of ponds (Asano, 2006; Ferris et al., 1954; Kimrey, 1989). Clogging from fine particle material or bacteria is one drawback to use of ponds, but filtering of the water prior to infiltration, and periodic drying the basin of the pond with removal of the clogged surface layer can be done to mitigate this problem (Baveye et al., 1998; Bouwer, 2002; Reddy, 2008).

Before determining if recharge of water through the unsaturated zone for storage in the underlying aquifer is possible, studies on prospective sites are commonly done to measure the likely infiltration rate. Hydrogeologic properties of the unsaturated zone can be determined from laboratory analysis of core material and other samples of unsaturated deposits overlying aquifers. Infiltration rates can be directly measured using an infiltrometer (Bouwer, 2002; Izbicki et al., 2000). For arid to semi-arid environments where the unsaturated zone can be hundreds of meters thick, studies have shown that low permeability caliche and clay layers can impede downward movement of water by causing lateral spreading (Izbicki et al., 2000; Izbicki et al., 2000b; Izbicki et al., 2002; Nimmo et al., 2002). Similarly, subsurface geology can impede infiltration of water if older, more consolidated units are present within the unsaturated zone. Areas where natural infiltration is higher such as intermittent streams/washes may provide suitable locations for artificial recharge due to the lack of low permeability caliche layers and a higher moisture content within the alluvium (Flint and Ellett 2004; Izbicki et al. 1998, 2000b; Kulongsoski and Izbicki, 2008). Lateral spreading of infiltrated water through the unsaturated zone results in decreased recharge rates and slower movement through the unsaturated zone (Nimmo et al., 2002; Izbicki et al., 2002). For artificial recharge to be successful, a complete hydrologic assessment of the site is helpful so that the recharge rate and any impediments to the downward movement of water, as well as impediments to the movement of the water in the subsurface on are known.
Information about the hydrology of an area at a large scale can be inferred from geologic maps, and if available, geophysical and hydrologic studies. Field measurements of infiltration rates, laboratory analysis of surface cores, pedotransfer functions, and ultimately, numerical modeling of flow through the unsaturated zone can be helpful in assessing the hydrology of a proposed recharge site at a scale appropriate for siting recharge ponds. Comparison of measured surface infiltration rates to modeled values can provide a better estimate of recharge through the unsaturated zone to storage in the aquifer (Kulongoski and Izbicki, 2008). Surface infiltration data can be input into a computer program such as VS2DH (Healy and Ronan, 1996; Lappala et al., 1987) to estimate hydraulic properties of the unsaturated zone and the feasibility of artificial recharge. The program utilizes a finite-difference method to solve Richards equation (Healy and Ronan, 1996; Lappala et al., 1987) that describes fluid flow through the unsaturated zone. Comparison of modeled hydraulic conductivity values with laboratory derived values from surface cores can give an understanding the suitability of a proposed site for artificial recharge.

**MOVEMENT OF WATER THROUGH THE UNSATURATED ZONE**

An equation to describe the movement of water through porous media was first defined by Henry Darcy in 1856 (Jury et al., 1991). Darcy derived his equation from the results of multiple tests measuring the volume of water flowing through saturated sand columns under hydrostatic pressure gradients (Darcy, 1856). Darcy’s Law, \( Q = KsAi \), describes the relationship between discharge, \( Q \), the saturated hydraulic conductivity, \( Ks \), the cross sectional area through which fluid is moving, \( A \), and the hydraulic gradient, \( i \) (\( i = \frac{\Delta h}{L} \)), where \( \Delta h \) is the change in head between a distance \( L \). The hydraulic conductivity is a constant dependent on soil properties and governs the velocity of the flow. Written in the flux form \( Jw = Q/A \), Darcy’s law is,

Vertical flow: \( Jw = -Ks(H_2 - H_1)/(z_2 - z_1) \)

Horizontal flow: \( Jw = -Ks(h_{p2} - h_{p1})/(x_2 - x_1) \)

where \( H \) is the hydraulic head, \( z \) is depth, and \( h_p \) is the pressure head. The hydraulic head is the combination of the pressure head and the elevation head. Hydraulic head is used in the vertical flow equation because gravity becomes a driving force for vertical flow in the presence of an elevation head gradient.
Darcy’s Law is useful for explaining flow under saturated conditions; however, in the unsaturated zone the presence of an air phase must be accounted for (Jury et al., 1991). Surface tension of the air-water interface and a decrease in moisture content leads to water pressures lower than the reference liquid pressures, confining the flow to narrower channels and decreasing the effective hydraulic conductivity (Jury et al., 1991). Derived in 1907, the Buckingham-Darcy flux law is a modification of Darcy’s Law that describes flow through the unsaturated zone (Jury et al., 1991). The two assumptions of the Buckingham-Darcy flux law are:

1. The driving force for water flow in unsaturated soil is the sum of the matric and gravitational potentials.
2. The hydraulic conductivity of unsaturated soil is a function of the water content or matric potential.

For vertical flow, the Buckingham-Darcy flux law can be expressed as \( J_w = -K(h) \frac{\partial H}{\partial z} \), where \( H \) is the total hydraulic head, \( K(h) \) is the unsaturated hydraulic conductivity, and \( J_w \) is the flow per unit cross-sectional area per unit time (Buckingham, 1907). Flow will reach a steady state when a matric potential difference is maintained across a soil column of length \( L \), and the Buckingham-Darcy flux law may be written as \( J_w = -K(h) (\partial h/\partial z + 1) \) (Buckingham, 1907). Steady-state downward water flow can be used as an approximation of scenarios such as subsurface drainage under frequent rainfall or ponding. Under such conditions, matric potential gradients \( \frac{\partial h}{\partial z} \) approach zero and infiltration is driven solely by gravity. Therefore the Buckingham-Darcy flux law can be approximated as \( J_w \approx -K(h) \) for steady-state downward flow in areas where the water table is far below the surface (Buckingham, 1907).

When combined with the water conservation equation, volume of water in is equal to volume of water out plus volume added to storage, the Buckingham-Darcy flux equation produces an equation to predict water content or matric potential during transient flow. This new equation is termed the Richards equation (Jury et al., 1991). Two forms of the Richards equations are as follows:

Water Content Form: \( \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left( D_w(\theta) \frac{\partial \theta}{\partial z} \right) + \frac{\partial K(\theta)}{\partial z} \)

Matric Potential Form: \( C_w(h) \frac{\partial h}{\partial t} = \frac{\partial}{\partial z} \left[ K(h) \left( \frac{\partial h}{\partial z} + 1 \right) \right] \)
where $D_w(\theta) = K(\theta) \frac{dh}{d\theta}$ is the soil water diffusivity and $C_w(h) = \frac{d\theta}{dh}$ is the water capacity function.

Infiltration refers to the movement of liquid into a soil through the boundary. Vertical infiltration of liquid at the surface is generally implied. As described by the Buckingham-Darcy flux law for vertical flow, infiltration rates are governed by the matric and gravitational potentials as well as hydraulic conductivity of the soil. Infiltration rates are greatest initially and slow as the wetting front moves away from the surface. Matric potential gradients drive infiltration at the near surface, transitioning to gravity driven infiltration with increasing depth of the wetting front. Assuming gravity driven flow and a unit gradient, the rate of infiltration will reach a constant nearly equal to that of the saturated hydraulic conductivity of the soil (Jury et al., 1991). The time necessary to reach a constant infiltration rate is dependent upon the physical properties of the soil. Soils with lower hydraulic conductivities will reach a constant infiltration rate at greater times as slower movement of water requires more time for the wetting front to reach a depth where gravitational forces predominate. Empirical (Horton, 1933; Kostiakov, 1932) and mechanistic (Green and Ampt, 1911) models have been created to explain infiltration.

In 1932, Kostiakov (Jury et al., 1991) proposed an infiltration equation that allowed infiltration to be evaluated by fitting the model, $i = \alpha \gamma t \alpha - 1$, where $i$ is the infiltration rate for a period of time $t$ and $\gamma$ and $\alpha$ are constants, to experimental data. Horton (1933, 1939) discovered that his field data from infiltration tests displayed an almost exponential decrease in infiltration rate with time and that the rate eventually reached a constant value that was slightly smaller than the saturated hydraulic conductivity of the soil. To describe this rate decrease, Horton (1939) created the equation $i = if + (i0 - if) \exp(-\beta t)$, where $if$ is the final constant infiltration rate reached at large times and $\beta$ is a parameter that describes the rate of decrease of infiltration.

By making simplifying assumptions about the movement of water during infiltration, Green and Ampt (1911) derived an approximate mechanistic model that can solve for the infiltration rate exactly. The assumptions made by Green and Ampt are:

1. There is a discontinuous change in water content at the wetting front.
2. The soil in the wetted region has constant properties
3. The matric potential head at the moving front is constant and equal to the matric potential of the wetting front, $h_F$. 

The Green-Ampt equation for infiltration into a flooded soil is: 
\[ i = K \left( H_w + \frac{L_f}{L_f} \right) / L_f \]
where K is the hydraulic conductivity of the wetted zone, \( H_w \) is the water depth above the soil, \( L_f \) is the depth of the wetting front, and \( h_{\text{we}} \) is the negative pressure head at the wetting front (Bouwer, 1978). At initiation of infiltration \( L_f \) is small and infiltration rate is high, but as the wetting front advances the infiltration rate approaches a value equal to the hydraulic conductivity of the wetted zone (Bouwer, 2002). Entrapped air causes K to be less than the saturated hydraulic conductivity (Bouwer, 1978).

J. R. Philip (1957) using the Richards equation formulated a numerical technique to solve the flow equation exactly for infiltration into an infinitely deep homogeneous porous medium at uniform initial water content. Philip’s equation for horizontal infiltration is, 
\[ i = \frac{1}{2} S t^{-1/2} \]
where S is the sorptivity. Sorptivity increases as the boundary water content increases, and is equal to the slope of \( I \), cumulative infiltration, versus \( t^{1/2} \). The Philip model uses separate equations for vertical infiltration at short and long times. The Philip equation for short time vertical infiltration is 
\[ i = \frac{1}{2} S t^{1/2} + A \]
where A is a constant that depends on the initial and boundary water contents. For long time vertical infiltration Philip (1957) concluded:

1. The infiltration rate approaches a constant equal to the unsaturated hydraulic conductivity \( K(\theta_0) \).
2. The wetting front advances without changing shape.
3. The velocity of the advancing wetting front approaches a constant:
\[ V_F = \frac{(K(\theta_0) - K(\theta_i))}{(\theta_0 - \theta_i)} \]

Using an the equation \( \frac{\theta - \theta_r}{\theta_s - \theta_r} = \left[ 1 + (\alpha h)^n \right]^{-m} \)
where \( \theta \) is the water content, \( \theta_s \) is the saturation soil moisture, \( \theta_r \) is the residual soil moisture, h is the pressure head, \( \alpha \) and n are constants, and \( m = 1-\ln^{-1} \), to describe a continuous soil-water retention curve with continuous slope, van Genuchten (1980) derived a closed-form analytical expressions for relative hydraulic conductivity expressed in terms of pressure head:
\[ K_r(h) = \frac{(1-(\alpha h)^{n-1}) \left[1+(\alpha h)^n\right]^{m}}{[1+(\alpha h)^n]^{m/2}} \quad m = 1-1/n \]
Using knowledge of soil moisture characteristics, the relative hydraulic conductivity of a soil can be determined using van Genuchten’s expression.
INFILTROMETER THEORY AND APPLICATION

According to Bouwer (2002), infiltration tests in the field can help estimate the land area needed for desired volumetric recharge rates that is possible with a certain area. Knowledge of recharge values is helpful in designing ponds. Infiltration test methods have been discussed (Bouwer, 2002; Chowdary et al., 2006; Tricker, 1978; Youngs, 1991) and several types are available; including: test ponds, tension infiltrometers, single-ring infiltrometers, and double-ring infiltrometers. Test ponds best simulate the scenario of a recharge pond (Bouwer, 2002), but they are time consuming and labor intensive to build, and a large amount of water is needed. Tension infiltrometers provide a measurement of sorptivity and hydraulic conductivity by providing a source of water at negative pressure at the surface (Youngs, 1991). Tension infiltrometers are useful for studying the effect of soil disturbance or measuring hydraulic properties in areas where macropores may influence infiltration (White et al., 1983) because the negative pressures prevent larger pores from filling with water and becoming preferred flow paths (Youngs, 1991). Ring infiltrometers measure a volumetric infiltration rate. Infiltration through a single-ring infiltrometer is a combination of vertical flow and lateral spreading (Tricker, 1978) and can over estimate the rate at which water will move vertically through the unsaturated zone toward the aquifer. Over estimation of the vertical flow rate is reduced with a double-ring infiltrometer, as the flow from the inner ring is constrained by flow from the outer ring and is therefore assumed to be vertical (Youngs, 1991).

PURPOSE AND SCOPE

The purpose of this study is to evaluate the potential for infiltration through surface deposits at a proposed recharge site in Joshua Tree, California. Scope of the work includes field data collection at selected sites using a double-ring infiltrometer, laboratory analysis of surficial deposits, and calculation of hydraulic properties of surficial alluvial deposits. The scope of the study also included use of the computer program Rosetta to estimate hydraulic properties of unsaturated alluvium, as well as interpretation of infiltration data using the computer program VS2D. The computer program VS2D was used to estimate the time required for water from ponds to reach the water table and recharge the underlying aquifer,
assuming uniform properties estimated from infiltrometer data. Results of the study will be used to provide the basis for decisions on locations of proposed recharge ponds.
CHAPTER 2

DESCRIPTION OF THE STUDY AREA

The study area is the Joshua Tree Groundwater subbasin in the southern Mojave Desert, approximately 120 miles east of Los Angeles, California, near the town of Joshua Tree, California (see Figure 1). The Joshua Tree ground-water subbasin covers 53.8 square miles (DWR, 2003) and is bounded by the Little San Bernardino Mountains to the south, the Yucca Barrier and Warren Valley groundwater subbasin to the west, the Pinto Mountain Fault and the Copper Mountain ground-water subbasin to the north, and the Twentynine Palms ground-water subbasin to the east.

Figure 1. Location map.
Climate in the area is arid with hot dry summers and cool wet winters. Average annual precipitation is 4 in., with most of the precipitation falling between November and March (NCDC). Surface drainage is through Quail Wash draining the Little San Bernardino Mountains to the south and Yuca Wash draining the Warren subbasin areas to the west toward Coyote (dry) Lake. There are no perennial streams in the area. Streamflow occurs for brief periods after summer and winter rains. Summer flows can be large and are often highly destructive.

The population of Joshua Tree, California has increased from 550 in 1947 (Joshua Tree Chamber of Commerce) to 3,898 in 1990 and 4,207 in 2000 (U.S. Census). The increase in population has led to an increase in groundwater pumping. According to Nishikawa and others (2004), between 1958-2001 the cumulative volume of water pumped from the subbasin was 42,210 acre-feet, 99% pumped from groundwater storage and only 207 acre-ft/yr from natural recharge. As a result of pumping in excess of recharge, the water table has declined by as much as 35 feet (Nishikawa et al., 2004). Although pumping and recharge are small relative to the volume of water in storage (750,000 – 2,540,000 acre-feet), without a means to supplement groundwater supplies, continued pumping in the Joshua Tree area may deplete the local groundwater supply.

**GEOLOGY**

Geology of the study area is characterized by thick alluvium overlying thick volcanic deposits, with a deep basement complex. Abundant faulting in the study area influences movement and storage of water in the subsurface by creating barriers to water movement and influencing topography.

**Stratigraphy**

Stratigraphy of the study area is comprised of three generalized units (Nishikawa et al., 2004). The deepest unit is a basement complex of pre-Tertiary granitic and metamorphic rocks. Overlying the basement complex is a unit of Tertiary sedimentary and volcanic deposits with a maximum thickness of 2000 feet. The upper-most unit consists of Quaternary alluvial deposits. The thickness of the upper unit ranges from a few feet, near the basin margin, to almost 1000 feet. The quaternary alluvial deposits are poorly sorted sands and gravels with interbedded silt and clay. These alluvial deposits are unconsolidated at the
surface, but become more consolidated with depth. The most permeable deposits occur beneath active washes (Nishikawa et al., 2004).

Faults and Barriers to Flow

Pinto Mountain Fault is the most prominent fault in the study area. The trend of Pinto Mountain Fault is east-west, creating low permeability zones that act as a barrier to flow between the Joshua Tree ground-water subbasin and the Copper Mountain ground-water subbasin (Nishikawa et al., 2004). Water levels are as much as 70 to 125 feet higher in the Joshua Tree ground-water subbasin (~ 375 ft) than in the adjacent Copper Mountain subbasin (~300 ft) (Mendez and Christensen, 1997; Nishikawa et al., 2004; Whitt and Jonker, 1998). Another barrier to flow is Yucca Barrier, a series of parallel, unnamed, north-south trending faults that restrict flow between the Warren Valley and Joshua Tree groundwater subbasins (Nishikawa et al., 2004).

Aquifer System

Groundwater in the Joshua Tree and Copper Mountain groundwater subbasins is unconfined and occurs in unconsolidated basin fill and alluvial fan deposits of interbedded gravels, conglomerates, and silts (Schaffer, 1978). The shallow Quaternary alluvial deposits are divided into two aquifers, upper and middle aquifers. The deeper Tertiary sedimentary and volcanic deposits constitute a single aquifer called the lower aquifer (Nishikawa et al., 2004). The upper aquifer is comprised of sand and gravel. Thickness of the upper aquifer ranges from 300 feet in the Joshua Tree groundwater subbasin to less than 175 feet in the Copper Mountain groundwater subbasin. Transmissivity of the upper aquifer is 580 to 56,000 ft²/d, with a mean of 6,200 ft²/d, and specific yield is 0.15 in the Joshua Tree groundwater subbasin and 0.14 in the Copper Mountain groundwater subbasin (Lewis, 1972). The middle aquifer contains sand, silt, and clay, with gravel layers. Thickness of the middle aquifer is about 450 feet and transmissivity ranges from 38000 to 98000 ft²/d (Nishikawa et al., 2004). The lower aquifer can be as much as 1500 feet thick and has a maximum transmissivity of 750 ft²/d (Nishikawa et al., 2004). Groundwater levels for 1996 were 490 – 120 feet below land surface within the Joshua Tree groundwater subbasin and 520 – 220 feet below land surface within the Copper Mountain groundwater subbasin (Nishikawa et al., 2004).
Estimates of groundwater storage for the Joshua Tree groundwater subbasin range from 750,000 acre-feet (Krieger and Stewart, 1996) to 1,150,000 acre-feet (Whitt and Jonker, 1998) to 2,540,000 acre-feet (DWR, 2003). A minimum of 940,000 acre-feet of groundwater is estimated to be in storage within the Copper Mountain groundwater subbasin (Whitt and Jonker, 1998).

**GROUNDWATER RECHARGE AND DISCHARGE**

Recharge to the Joshua Tree ground-water subbasin occurs primarily through infiltration of runoff from intermittent streams draining the Little San Bernardino Mountains (Lewis, 1972) and from ground-water underflow from the Warren Valley groundwater subbasin (Nishikawa et al., 2004). Estimated recharge by ground-water underflow from Warren Valley groundwater subbasin ranges from less than 200 acre-ft/yr (Lewis, 1972) to almost 85 acre-ft/yr (Nishikawa et al., 2003) depending on method. Recharge of the Copper Mountain ground-water subbasin occurs by ground-water underflow from the Joshua Tree ground-water subbasin across the Pinto Mountain Fault and by infiltration of intermittent streamflow from surrounding uplands (Nishikawa et al., 2004). Simulated annual recharge from intermittent streamflow in the Joshua Tree and Copper Mountain groundwater subbasins is 158 acre-ft/yr (Nishikawa et al., 2004).

An increasingly important source of recharge in the study area is septage recharge. Assuming 73% of pumped ground-water is returned by septage, the rate of recharge by septage between 1977 and 2001 was 660 acre-ft/yr (Nishikawa et al., 2004). The travel time for septage to reach the water table is 1.4 – 20 years (Nishikawa et al., 2004).

The primary form of discharge from the Joshua Tree ground-water subbasin is ground-water pumping by the Joshua Basin Water District, JBWD. Wells in the Joshua Tree ground-water subbasin yield from 40 to 2,200 gal/min (DWR, 2003). Between 1958 and 2001 almost 42000 acre-ft of ground-water was pumped from the subbasin as annual pumpage increased from 135 acre-ft/yr in 1958 to a maximum of 1700 acre-ft/yr in 1990 (Nishikawa et al., 2004).
CHAPTER 3

METHODS

Data collection included field methods and laboratory methods for measuring physical and hydraulic properties of the alluvium within the study area. Several computer modeling methods were used to interpret the field and laboratory data.

FIELD METHODS

Surface infiltration data were collected using a 4-foot diameter double-ring infiltrometer, having an inner-ring with a diameter of 2 feet (see Figure 2). Study sites were selected based on possible locations for future recharge ponds (see Figure 3). A total of eight infiltrometer tests were completed; four to the north of the Pinto Mountain Fault (Sites 1, 2, 7, 8) and four to the south of the fault (Sites 3 – 6). Two study sites were located within Yucca Wash (Sites 1 and 2).
At each site, the double-ring infiltrometer was driven a few inches into undisturbed surficial material. Where necessary, soil was mounded around the outside of the infiltrometer to prevent leaking. One, 50 gallon, calibrated container was connected to each ring of the infiltrometer via hoses and filled. Flow from the containers was restricted until start of the test. Both rings of the infiltrometer were filled with water to a desired height and flow from the containers commenced. Care was taken to limit disturbance of the surficial material during filling of the infiltrometer in order to prevent creation of artificial preferential
flow paths. Flow from the calibrated containers was regularly adjusted to maintain the initial 
water level within the infiltrometer. Time was recorded for every five gallons of water 
infiltrated through the infiltrometer.

The cumulative infiltration volume was plotted against time and the slope of the best-fit 
line was the ponded infiltration rate. The ponded infiltration rate calculated from the 
infiltrometer data was ultimately compared with modeled data to obtain a hydraulic 
conductivity of the near-surface unsaturated zone.

**LABORATORY METHODS**

Surface (0 – 1 ft bslsd) cores were taken at each test site and analyzed at the USGS 
California Water Science Center (CAWSC) Hydrologic Research Laboratory, in Sacramento, 
CA. Analysis included saturated hydraulic conductivity, physical properties (including bulk 
density, volumetric water content, porosity, effective porosity, saturation, effective 
saturation, residual water content, particle density), and particle-size distribution analysis. 
Physical property data and the estimated saturated hydraulic conductivity data were used as 
input data in the VS2D model. Particle size analysis data was used as input data in the 
Rosetta model.

**INTERPRETATION OF DATA**

Data from field and laboratory tests provided values for the surficial hydraulic and 
physical properties. Models were then used to extrapolate the information gathered at the 
surface to the deep subsurface. Rosetta was used to create van Genuchten retention 
parameters, as well as hydraulic conductivity values, and VS2D modeled the movement of 
the water through the subsurface; also providing hydraulic conductivities.

**Rosetta**

The Rosetta computer program uses five hierarchical pedotransfer functions (PTFs) to 
estimate water retention, saturated hydraulic conductivity, and unsaturated hydraulic 
conductivity based on textural classes or more complex soil data (Schaap et al., 2001). A 
pedotransfer function is a predictive function that attempts to translate basic information into 
more useful information through empirical regression. Particle size data obtained from the 
laboratory analysis of the surface cores was used as input data for the Rosetta model which
was used to estimate saturated hydraulic conductivity and the van Genuchten retention parameters. The van Genuchten parameters estimated by the Rosetta model were used in the VS2D program for all models. The estimated saturated hydraulic conductivities were compared to the field data from the double-ring infiltrometer and to laboratory data.

**VS2D**

The numerical modeling program VS2D was used to simulate infiltration and recharge. The program uses a finite-difference approximation of Richard’s equation to describe the movement of water through the unsaturated zone by combining the conservation of mass equation with equations for fluid flux and storage (Lapalla et al., 1987). VS2D can be used to create models in either rectangular or cylindrical coordinate systems. By retaining volume and area terms in the program’s nonlinear flow equation, radial or rectangular gridding can be used in the block-centered regular finite-difference scheme (Lapalla et al., 1987). Models were used in this study to estimate hydraulic conductivity, test the sensitivity of infiltrometer and model to layers of low permeability, and to model recharge at test sites.
CHAPTER 4

RESULTS

Results show higher infiltration rates within Yucca Wash than outside the wash. A greater sand content, as measured through laboratory analysis of surface cores, is also seen in the wash samples and agrees with the higher infiltration rates. Laboratory measured hydraulic conductivity values do not match field measured infiltration rates, but this may be due to sampling size differences.

INFILTROMETER DATA

Infiltrometer field test data (see Table 1) were used to calculate the ponded infiltration rate for each test site (see Figure 4). Sites 1 and 2, located within Yucca Wash, had the highest infiltration rates of $2.3 \times 10^{-3}$ ft$^3$/s and $3.4 \times 10^{-3}$ ft$^3$/s respectively (see Table 2). Infiltration rates for the near wash sites, 7 and 8, are $1.6 \times 10^{-3}$ ft$^3$/s and $1.5 \times 10^{-3}$ ft$^3$/s respectively. Test sites 3, 4, 5, and 6, located off of Yucca Wash and south of the Pinto Mountain Fault, display the lowest infiltration rates with respective values of $8.1 \times 10^{-4}$ ft$^3$/s, $7.3 \times 10^{-4}$ ft$^3$/s, $9.5 \times 10^{-4}$ ft$^3$/s, and $1.1 \times 10^{-3}$ ft$^3$/s. Most of the sites display a linear trend in the infiltrometer data suggesting that the constant infiltration rate has been reached at very early times and possibly due to the high hydraulic conductivities at these sites, the expected exponential decrease in infiltration rate is not observed. Sites 3 and 6 display some non-linearity and evidence of a near exponential decrease in infiltration rate with time.

LABORATORY DATA

Particle size analysis of the surface cores showed that all of the test sites are composed mostly of sand (60 – 83%) (see Table 3). Sites 1 and 2, within Yucca Wash, had the greatest percentages of sand, 82% and 83% respectively, and the lowest percentages of silt, 3% and 4% respectively. The location of sites 1 and 2 within Yucca Wash and fluvial sorting of the sediment during streamflow can explain the low percentage of fine-grained material. The high fraction of sand in the wash sites is consistent with the higher ponded infiltration rates at those sites. Site 7, near the edge of Yucca Wash, contains a relatively
Table 1. Infiltrometer Field Data, Joshua Tree, California, Jan. 2008

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<th>Site 3</th>
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Figure 4. Field infiltration data, near Joshua Tree, California, Jan. 2008.
Table 2. Calculated Infiltration Rates from Joshua Tree, California, Jan. 2008

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<th>Site ID</th>
<th>Calculated Infiltration Rate, in ft³/s</th>
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<tr>
<td>Site 1</td>
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<td>Site 2</td>
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<td>Site 6</td>
<td>1.34x10⁻³</td>
</tr>
<tr>
<td>Site 7</td>
<td>1.74x10⁻³</td>
</tr>
<tr>
<td>Site 8</td>
<td>1.47x10⁻³</td>
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Table 3. Particle Size Analysis from Joshua Tree, California, Jan. 2008

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<tr>
<th>Site ID</th>
<th>Overall Sample Percentages</th>
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<td>Site 7</td>
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<td>Site 8</td>
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</table>

A high percentage of gravel, 27%. High streamflow within Yucca Wash could transport larger gravel size particles to this location while simultaneously washing away the finer sand and silt particles.

Bulk density and porosity ranged from 1.65 g/cm³ to 1.80 g/cm³, and .325 to .390, respectively and were relatively constant for all of the test sites (see Table 4). Volumetric water content, saturation, and residual water content were low, but not atypical for surficial materials in desert settings. The low values may result from drying of the surface material due to evaporation. The lower water content values at sites 1 and 2 may be explained by the greater percentage of sands at these locations, as sands typically have lower water contents for a given matric potential then siltier and clayier materials.

Laboratory hydraulic conductivity data ranged from 4.33x10⁻⁵ ft/s to 7.59x10⁻⁵ ft/s (see Table 4). These values of hydraulic conductivities are within the range for fine sands to silty sand (see Table 5). Hydraulic conductivity values were expected to be within the range for medium to fine sands based on visual inspection of the surficial materials at the study area. Lower calculated values may be due to compaction of the core material during collection or the presence of a fine-grain dominated surficial crust. Unlike the infiltrometer data, the hydraulic conductivities do not show good correlation with location to Yucca Wash.
Table 4. Laboratory Hydraulic Properties Data from Joshua Tree, California, Jan. 2008

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<tr>
<th>Site ID</th>
<th>Bulk Density (g/cm³)</th>
<th>Porosity (m³/m³)</th>
<th>Volumetric Water Content (m³/m³)</th>
<th>Saturation (m³/m³)</th>
<th>Residual Water Content (m³/m³)</th>
<th>Hydraulic Conductivity (ft/s)</th>
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Table 5. Hydraulic Conductivities of Typical Soil Types

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<tr>
<th>Soil Type</th>
<th>Saturated Hydraulic Conductivity (ft/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravel</td>
<td>1.0 – 1x10⁻³</td>
</tr>
<tr>
<td>Sand</td>
<td>1x10⁻² – 1x10⁻³</td>
</tr>
<tr>
<td>Silty Sand</td>
<td>1x10⁻⁴ – 1x10⁻⁶</td>
</tr>
<tr>
<td>Silt, Loess</td>
<td>1x10⁻⁵ – 1x10⁻⁹</td>
</tr>
<tr>
<td>Glacial Till</td>
<td>1x10⁻⁸ – 1x10⁻¹²</td>
</tr>
<tr>
<td>Unweathered Marine Clay</td>
<td>1x10⁻⁹ – 1x10⁻¹²</td>
</tr>
</tbody>
</table>

The differences between lab and field data may be a result of differences in the scale of the measurements. A greater volume of material was tested using the double-ring infiltrometer than in the laboratory hydraulic conductivity measurement.
CHAPTER 5

MODEL SIMULATIONS

The computer modeling programs Rosetta and VS2D were used in conjunction to extrapolate field measured surficial data to the subsurface alluvium. Rosetta was used to estimate the van Genuchten water retention parameters for input into VS2D for estimation of hydraulic conductivity of the subsurface alluvium. Rosetta was also used as a standalone method for estimating hydraulic conductivity of the subsurface alluvium. Hydraulic conductivities estimated using VS2D were within the range of fine sand to silty sand. Laboratory hydraulic conductivities matched the field data better than hydraulic conductivities calculated using Rosetta, but neither method correctly predicted infiltration rates measured using the double-ring infiltrometer. Simulation results using VS2D were sensitive to low-permeability layers at shallow depths within the alluvium and evidence for a surficial low-permeability layer was seen in the data for six of the eight test sites.

PARAMETER ESTIMATION

The Rosetta computer program was primarily used in this study to estimate the van Genuchten parameters, $\alpha$ and N (see Table 6). The soil profile for all sites was assumed to be homogeneous and isotropic; therefore, the use of particle size analysis data from surface cores to estimate the water retention and the shape of the wetting front at depth is possible. The van Genuchten parameters, $\alpha$ and N, estimated by Rosetta show strong correlations to the Rosetta calculated hydraulic conductivities, with higher conductivity sites having low alpha values and high values for N and lower conductivity sites having high alpha values and low values for N (see Figure 5).

The saturated hydraulic conductivities estimated by Rosetta ranged from $1.73 \times 10^{-5}$ ft/s to $6.71 \times 10^{-5}$ ft/s (see Table 6). These values are within an order of magnitude of the lab measured hydraulic conductivity values, and also fall within the range of hydraulic conductivity values for fine sands to silty sands (see Table 4 and 5). Similarity between the laboratory hydraulic conductivity and the Rosetta hydraulic conductivity are greatest for Site
Table 6. Rosetta Calculated van Genuchten Parameters, Rosetta Calculated Hydraulic Conductivity Values, Laboratory Calculated Hydraulic Conductivity Values, and Percent Difference from Joshua Tree, California, Jan. 2008

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Rosetta Calculated van Genuchten alpha, ft⁻¹</th>
<th>Rosetta Calculated van Genuchten N</th>
<th>Rosetta Calculated Hydraulic Conductivity (ft/s)</th>
<th>Laboratory Hydraulic Conductivity (ft/s)</th>
<th>Percent Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site 1</td>
<td>0.033</td>
<td>2.38</td>
<td>5.87x10⁻⁵</td>
<td>6.65x10⁻⁵</td>
<td>12.5</td>
</tr>
<tr>
<td>Site 2</td>
<td>0.035</td>
<td>2.46</td>
<td>6.71x10⁻⁵</td>
<td>5.27x10⁻⁵</td>
<td>24.1</td>
</tr>
<tr>
<td>Site 3</td>
<td>0.044</td>
<td>1.63</td>
<td>1.80x10⁻⁵</td>
<td>5.62x10⁻⁵</td>
<td>103</td>
</tr>
<tr>
<td>Site 4</td>
<td>0.040</td>
<td>1.94</td>
<td>4.05x10⁻⁵</td>
<td>7.54x10⁻⁵</td>
<td>60.2</td>
</tr>
<tr>
<td>Site 5</td>
<td>0.041</td>
<td>1.80</td>
<td>2.36x10⁻⁵</td>
<td>6.47x10⁻⁵</td>
<td>93.2</td>
</tr>
<tr>
<td>Site 6</td>
<td>0.043</td>
<td>1.69</td>
<td>1.73x10⁻⁵</td>
<td>6.76x10⁻⁵</td>
<td>119</td>
</tr>
<tr>
<td>Site 7</td>
<td>0.038</td>
<td>1.65</td>
<td>1.90x10⁻⁵</td>
<td>4.33x10⁻⁵</td>
<td>78.2</td>
</tr>
<tr>
<td>Site 8</td>
<td>0.037</td>
<td>1.94</td>
<td>4.13x10⁻⁵</td>
<td>7.59x10⁻⁵</td>
<td>59.0</td>
</tr>
</tbody>
</table>

Figure 5. Rosetta estimated van Genuchten parameters compared to Rosetta calculated hydraulic conductivities, from data collected near Joshua Tree California, Jan. 2008.

1 and Site 2. This similarity may be due to the constant reworking of the surface sediments creating well ordered near-surface structure. Excluding Site 2, the Rosetta model underestimates the hydraulic conductivity values determined by laboratory analysis by as much as 74%.
It was unexpected that hydraulic conductivity values estimated using Rosetta do not match the laboratory hydraulic conductivity values more closely considering the Rosetta input data comes directly from the lab analysis. Limited data, particle size and porosity, were input into the model. Also, pedotransfer functions, such as the ones Rosetta uses, however are based on laboratory derived data using general soil types, limiting their accuracy to replicate field conditions in geologic material or alluvium that does not have extensive soil development. Rosetta calculated hydraulic conductivity values can be useful in providing an estimate, within an order of magnitude, of expected values.

**Simulation of Infiltration Data**

Infiltration data was simulated using the computer program VS2D in an attempt to estimate hydraulic conductivity of the subsurface alluvium through matching infiltration rate curves. The modeling program was also used to test the infiltrometer method for sensitivity to subsurface layering.

**VS2Dh Model Development**

The numerical modeling program VS2D was used to estimate hydraulic properties based on infiltrometer data and create infiltration rate curves using laboratory and Rosetta calculated hydraulic properties for comparison with infiltrometer data. Using an upscaled model, VS2D also provided a rough estimate of wetting front arrival time for surface infiltration from a pond to the water table 400 feet below land surface.

**Setup and Solver**

The program VS2D uses a finite-difference approximation of Richard’s equation to describe the movement of water through the unsaturated zone by combining the conservation of mass equation with equations for fluid flux and storage (Lapalla *et al.*, 1987). Model units are seconds and feet. A two-dimensional axially symmetric grid represents one-half of an infiltrometer test site with the infiltrometer located in the upper most boundary of the axis. Transport, evaporation, and transpiration are not simulated. To simulate a thick unsaturated zone, the initial hydraulic condition is specified as an equilibrium profile with a water table at 400 feet below land surface and the minimum pressure head set at land surface with a value of -400 feet. Hydraulic characteristic functions are represented by the van Genuchten model.
The curve that correlates saturation to pressure potential is referred to as the moisture-characteristic curve. The van Genuchten equation for representing the moisture-characteristic curve is: 

\[ s_e = \left[ \frac{1}{1 + \left( \frac{h}{\alpha'} \right)^{N'}} \right]^\gamma, \]

where \( s_e \) is the effective saturation, \( h \) is the pressure potential, and \( \alpha', N', \) and \( \gamma \) are constants related by the equations \( \alpha' = \alpha / \left[ (21/\gamma - 1)1 - \gamma \right] \) and \( N' = 1 / (1 - \gamma) \). The slope of the moisture curve describes the specific moisture capacity and can be integrated to show the relationship between relative hydraulic conductivity and pressure potential (Lapalla et al., 1987). Specific moisture capacity is the change in saturation due to a change in pressure potential and is calculated using the van Genuchten equation:

\[
\begin{align*}
\text{cm} (h) &= \left( -\gamma N' (\Phi_0) \right) \left[ \frac{(h/(\alpha'))^{(N'-1)}}{(\alpha' \left[ 1 + \left( \frac{h}{\alpha'} \right)^{(N')} \right]^\gamma + 1)} \right] , \quad h \leq 0 \\
\text{cm} (h) &= 0 , \quad h > 0
\end{align*}
\]

Relative hydraulic conductivity or the ratio of unsaturated to saturated hydraulic conductivity is defined by van Genuchten using the equation:

\[
K_r = \left\{ 1 - \left[ \frac{h/(\alpha')}^{(N'-1)} \right] \left[ 1 + \left( \frac{h/(\alpha')}^{(N')} \right)^\gamma \right]^{(-\gamma)} \right\}^2 \left[ 1 + \left( \frac{h/(\alpha')}^{(N')} \right)^\gamma \right]^{(\gamma/2)}. 
\]

Intercell relative hydraulic conductivity is computed by upstream weighting. Geometric mean averages provide the most accurate estimate, but it has been shown that to prevent numerical oscillations it is sometimes necessary to use the upstream weighted arithmetic mean method, including the case of a sharp wetting front advancing into a uniform dry medium (Brutsaert, 1971). The option for printing out fluxes through individual boundary faces was activated to measure simulated infiltration rate through the inner ring only.

**DOMAIN**

The model domain is 15 feet in the radial, \( r \), direction and 25 feet in the vertical, \( z \), direction (see Figure 6). The upper-left corner of the model domain is the location of the center of the double-ring infiltrometer. The two rings of the infiltrometer are separated and interflow is restricted. The separation between rings, and the outer wall of the infiltrometer, is shown in the model by a 0.04 inch gap in the domain extending from the surface to a depth of 1 foot.

**GRID**

Grid spacing in the radial direction is a constant value of 0.04 inches for the area underneath the infiltrometer (0 – 2 ft.) (see Figure 6). Grid spacing should be small in the
Figure 6. VS2D model domain and grid used to simulate flow through unsaturated alluvium near Joshua Tree, California.
dimension perpendicular to the boundary to accurately represent a specified boundary potential (Lapalla et al., 1987). A 0.04 inch spacing also assures that the cells describing the infiltrometer walls can be designated as no-flow boundaries. With increasing radial distance (2 – 15 ft.), spacing increases from 0.04 inches by a factor of 1.02 to a maximum spacing of 0.296 inches. Grid spacing in the vertical direction is a constant value of 0.25 feet.

**Textural Distribution**

The ratio of vertical hydraulic conductivity to horizontal hydraulic conductivity was one for all models as isotropic conditions assumed. Three models with homogeneous soil profiles were constructed for each site using the laboratory derived hydraulic conductivity, the hydraulic conductivity calculated by Rosetta, and a hydraulic conductivity that was adjusted to produce an infiltration rate curve matching the infiltrometer data. All models at each site used the laboratory calculated residual moisture content and porosity, as well as the van Genuchten constants calculated by Rosetta. Specific storage was set at zero.

**Recharge Period**

Recharge was simulated as a constant head for each test. A constant one foot stage height within the infiltrometer was simulated at the surface boundary representing the inner ring of the infiltrometer and the surface boundary representing the outer ring of the infiltrometer (see Figure 6). All other boundaries were set as no-flow boundaries. The initial time step was 0.05 seconds with a time-step multiplier of 1.7 and a time-step reduction factor of 0.1. The maximum and minimum time steps were 100 seconds and 0.01 seconds respectively. The duration of recharge differed at each site and was adjusted to match the duration of infiltration measured in the field data.

**Model Results**

Matching infiltration rate curves produced by VS2D with curves from field data allowed for the estimation of the subsurface hydraulic conductivity of the study sites. Sites with higher infiltration rates generally had higher estimated hydraulic conductivity values. Hydraulic conductivity values estimated by VS2D were compared to the laboratory derived values and the Rosetta estimated values; however, agreement between the values was not
A sensitivity to near-surface heterogeneity and the presence of surficial crusts in the wash sediments were evident.

**Estimation of Hydraulic Conductivity**

Field hydraulic conductivity values were estimated by fitting a modeled infiltration rate curve to the infiltration rate curve produced from the infiltrometer data (see Figure 7). Hydraulic conductivity input into the model was adjusted, causing a change in the slope of the modeled infiltration rate curve, until the measured and model simulated slopes matched. Resultant values for hydraulic conductivity are within the range for fine sand to silty sand and are within one order of magnitude of the laboratory values and the Rosetta values (see Table 7). There is a general agreement between the hydraulic conductivities and infiltration rates; higher infiltration rates matched with larger hydraulic conductivity values. Exceptions to this agreement are site 1 which has a higher infiltration rate but a smaller hydraulic conductivity value than site 7, and site 5 which has the lowest modeled hydraulic conductivity but a higher infiltration rate than sites 3 and 4. Discrepancies between hydraulic conductivity and infiltration rate are small. A good correlation is evident between hydraulic conductivity and the location of the site with respect to the wash, with higher hydraulic conductivities found at the wash/near-wash sites.

**Sensitivity Analysis**

A sensitivity analysis was done to determine which variables may or may not influence the model output data. Sensitivity to subsurface layers is important because if the model can sense changes in hydraulic conductivities at depth, these possible impediments to flow can be considered during site evaluation. Effects of a surficial, low permeability layer was seen in infiltration data; therefore, scenarios in which these layers are present were tested to verify the possibility of their existence. Sensitivity to varying initial hydraulic and physical parameters of the alluvium was also tested.

**Sensitivity to Subsurface Layers**

The existence of soil layers with different textures or permeabilities in the subsurface will cause an impediment to water flow. It is useful to locate any impediments to flow so
Figure 7. Comparison of field measured infiltration data and modeled data, near Joshua Tree, California.
Table 7. Modeled Hydraulic Conductivity Values, Laboratory Hydraulic Conductivity Values, and Percent Difference from Joshua Tree, California, Jan. 2008

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Modeled Field Hydraulic Conductivity (ft/s)</th>
<th>Laboratory Hydraulic Conductivity (ft/s)</th>
<th>Percent Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site 1</td>
<td>9.85x10^{-5}</td>
<td>6.65x10^{-5}</td>
<td>38.7</td>
</tr>
<tr>
<td>Site 2</td>
<td>1.74x10^{-4}</td>
<td>5.27x10^{-5}</td>
<td>107.1</td>
</tr>
<tr>
<td>Site 3</td>
<td>6.11x10^{-5}</td>
<td>5.62x10^{-5}</td>
<td>8.22</td>
</tr>
<tr>
<td>Site 4</td>
<td>4.65x10^{-5}</td>
<td>7.54x10^{-5}</td>
<td>47.5</td>
</tr>
<tr>
<td>Site 5</td>
<td>4.63x10^{-5}</td>
<td>6.47x10^{-5}</td>
<td>33.2</td>
</tr>
<tr>
<td>Site 6</td>
<td>8.15x10^{-5}</td>
<td>6.76x10^{-5}</td>
<td>18.6</td>
</tr>
<tr>
<td>Site 7</td>
<td>1.20x10^{-4}</td>
<td>4.33x10^{-5}</td>
<td>93.6</td>
</tr>
<tr>
<td>Site 8</td>
<td>9.37x10^{-5}</td>
<td>7.59x10^{-5}</td>
<td>21.0</td>
</tr>
</tbody>
</table>

that recharge success can be assessed. Removal of near-surface low permeability layers is possible, but is limited by cost and time necessary for removal. Modeled infiltration data was analyzed to test the ability to detect near-surface variations in hydraulic conductivity using surface data. Layers of silt (K = 2.28x10^{-6} ft/s) and clay (K = 5.56x10^{-7} ft/s) with thicknesses of 3 inches and 6 inches at depths of 2.5, 4, 5, and 10 feet below land surface were modeled. Data show a detectable decrease in infiltration rate due to the presence of a layer of lower hydraulic conductivity in the near-surface (0 – ~4 ft.) and that this decrease is evident in early time (30 – 60 min.) data (see Figure 8). At depths greater than almost four feet the detection of an infiltration rate decrease is not significant. Depth, not thickness or contrast in hydraulic conductivity, seems to be the controlling factor with respect to infiltration rate. The difference in infiltration rate decrease caused by layers of differing thickness and differing hydraulic conductivity is undetectable if the layers are at equal depth below land surface. The Green-Ampt equation for infiltration, \( i = K \left( \frac{H_w + L(f-h_{we})}{L_f} \right) \), can be used to explain why depth of the low permeability layer controls the rate of infiltration. Assuming complete saturation of the overlying unit before infiltration into the layer, increasing the depth of the low permeability layer causes not only an increase in the depth of the wetting front, \( L_f \), but also an increase in the depth of water above the infiltration surface, \( H_w \). Therefore, as the depth of the low permeability layer increases, the infiltration rate into, and eventually through, the layer also increases. Model results were not stable for the scenarios with a 6 inch clay layer.
Figure 8. Modeled infiltration rate curves showing the effects of a 6 inch silt layer, 3 inch silt layer, and 3 inch clay layer at depths of 2.5, 5, and 10 feet.
Simulation of the Effects of Surface Crust

The assumption of a homogeneous subsurface is not entirely accurate based on the results. A discrepancy between the hydraulic conductivity values from the laboratory analysis of the surface cores and the modeled hydraulic conductivity suggests that a surface crust or near-surface layer of low permeability exists for most of the sites. Formation of a crust can affect the hydraulic conductivity of the surficial materials due to particle migration (washed fine particles comprising clay and silt), compaction of the surface, and a change in the moisture characteristics of the soil (Issa et al., 2004). Model simulations were run for sites where the hydraulic conductivity of the surface layer is less than the modeled value (1-3, 6-8), to test a 2-layered scenario consisting of a surface crust overlying homogeneous alluvium (see Figure 9). The laboratory hydraulic conductivity values were used for the crust, while the modeled values were assumed to represent an average value for the entire wetted volume. The hydraulic conductivity of the subsurface underlying the surface crust was calculated using the geometric mean equation, \( K_{\text{Modeled}} = e^{\frac{1}{2} (\ln K_{\text{Lab}} + \ln K_{\text{Subsurface}})} \), and solving for \( K_{\text{Subsurface}} \) to give, \( K_{\text{Subsurface}} = e^{2\ln K_{\text{Modeled}} - K_{\text{Lab}}} \). The thickness of the surface crust was varied so that the modeled infiltration curve best matched the infiltrometer data.

Infiltration rate curves for the homogeneous and layered scenarios at sites 3, 6, and 8 are congruent. The percent differences between the laboratory and modeled hydraulic conductivities for these sites were low, < 21, and therefore it is not surprising that the infiltration rate curves are similar; however, this may be a limitation to the modeling program. Curves for both scenarios at sites 3 and 6 compare to the field data.

Infiltration rate curves for the wash sites 1 and 2, as well near-wash site 7, show that the layered model matches the field data better than the homogeneous model. The layered models for these sites created infiltration rate curves with slopes similar to the homogeneous model, but better match the field data with respect to total volume of water infiltrated and final infiltration rate. The ability of these models to predict infiltration seems to be due to the lower infiltration rate at the early times. The presence of a low permeability surface layer leads to a low initial infiltration rate and a smaller volume of infiltrated water during the beginning of the test. As infiltration depth increases and a greater volume of the lower, more permeable layer is wetted, the average hydraulic conductivity value is skewed to a greater
Figure 9. VS2D modeled infiltration rate curves from 2-layer scenarios near Joshua Tree, California.
degree towards the value of the lower layer and the rate of infiltration begins to match that of the homogeneous model. The correlation of the layered model to the infiltration data from the sites within, or within close approximation to, the wash suggests that the presence of a surface crust/low permeability layer is more prevalent in a wash setting. Flow through the wash will initially destroy most structure. Structural degradation of the soil surface causes a reduction in infiltration capacity (Issa et al., 2004). The crusts encountered at the study area are most likely a combination of sieving and runoff depositional crusts, as described by Valentin and Bresson (1992) (see Figure 10). Sieving crusts occur when fine grained sediment is transported below the surface, forming a three micro-layered structure with an upper layer of loose coarse grains, a middle layer of fine, densely packed grains, and a lower plasmic layer of fine grains with reduced porosity (Valentin and Bresson, 1992). The packing and layering along with the contrast in porosity between the micro-layers leads to reduced infiltration capacity. Runoff depositional crusts, compact containing a microbedded layer (Boiffin and Bresson, 1987), are formed by translocation and deposition of fine grained particles, and most often overlay a structural crust (Valentin and Bresson, 1992).

Crusts in the study area wash sites would initially form runoff depositional crusts during intermittent stream flow. Sieving crusts would simultaneously form beneath the depositional crusts as fine-grained particles are translocated to deeper depths. Near wash site 7 would also form crusts under similar processes. At site 7 the layered model was a better fit than the homogeneous model for the early time data, but overestimated the infiltration rate for the late time data. This could be a case where the 2 layer model correctly describes the near-surface, but additional layering below the near-surface causes a departure of the simulated data from the measured data.

The propagation and shape of the wetting front was also affected by the presence of a surface crust (see Figure 11). Sites that did not did not have a significant difference between the hydraulic conductivity values of the surface crust and the underlying layer, sites 6 and 8, did not show a noticeable difference in the shape or depth of the wetting front. For the wash sites 1 and 2, where the surface crust was much less permeable than the underlying layer, there was complete saturation of the surface crust/layer and a decrease in moisture content in the underlying sediment.
Figure 10. Time-sequence soil crusts in loamy and sandy soils.
Sensitivity to Physical and Hydrologic Parameters

A sensitivity analysis was done to assess the effect of variations in the applied pressure head, porosity of the sediment, and residual moisture content (see Figure 12). Model setup, domain, and grid were unchanged during sensitivity analysis. Field values for hydraulic conductivity ($K = 5.0E-5$ ft/s) and the van Genuchten constants ($\alpha = 0.035$, $N = 1.5$) estimated by Rosetta were used in sensitivity analysis, and the subsurface was assumed
Figure 12. Sensitivity analysis.
to be homogeneous and isotropic. The recharge period for the sensitivity analysis models was extended to 21,600 seconds, or 6 hours.

Sensitivity of the technique to applied pressure head was tested to identify optimal infiltrometer test conditions for future studies. Values of 0.5, 1, 1.5, and 1.75 feet of pressure head were used. The results show a direct correlation between final infiltration rate and applied pressure head, but there appears to be a limit on the infiltration rate. The limit on infiltration rate is most likely imposed by the hydraulic conductivity of the sediment. There also appears to be an inverse relationship between applied pressure head and infiltration rate in the early time data (0-60 sec.), and by 3,600 seconds (1 hr) there is a direct correlation between applied pressure head and infiltration rate for the remainder of the test.

Modeled porosity values simulated during sensitivity analysis were 0.25, 0.3, 0.4, and 0.5. Results show an asymptotic decrease in the infiltration rate for all modeled porosity values however rates did not reach a constant (see Figure 12). Simulated infiltration rate and cumulative infiltration increased with porosity. An increase in infiltration rate due to an increase in porosity is expected. More open space in the volume of sediment would suggest larger channels and a greater likelihood that channels connect, allowing for less restricted flow and a higher rate. If considerable, compaction of the sediments with depth can be a major impediment to recharge.

Residual moisture content is the measure of water held within the sediment at highly negative pressures after gravity drainage ceases or slows to a negligible rate (van Genuchten, 1980). The sensitivity of the model results to changes in residual moisture content values of 0.015, 0.025, 0.05, 0.1, and 0.2 was evaluated. Model sensitivity analysis shows infiltration rates increase to a maximum at $\theta_r = 0.025$, and then decrease (see Figure 13).

**Model Limitations**

The ability of the VS2D unsaturated zone flow model to predict infiltration is limited by the available data in the deeper subsurface alluvium. Hydraulic and physical properties data from laboratory analysis of core material aided in creating accurate models. Modeled infiltration rates consistently overestimated infiltration for the early time data. This could be due to an overestimation of the matric potential forces involved in driving early infiltration. Assumption of a homogeneous, isotropic unsaturated zone was applicable for the Joshua
Figure 13. Modeled infiltration rate as a function of residual moisture content.

Tree study area; however if layering is present, knowledge of the location, extent, and hydraulic properties of the layers are needed for accurate prediction of infiltration. The model is limited to detection of shallow impermeable layers. Sensitivity analysis of impermeable layers showed that the model did not detect layering below depths of almost 4 feet. Data showed that the model is limited in its ability to distinguish between a homogeneous unsaturated zone, without surficial layering, and a homogeneous unsaturated zone overlain by a less permeable surficial layer when the hydraulic conductivity of the surficial layer and underlying alluvium are similar in magnitude (< ~21 percent difference). Similarity in hydraulic conductivity between the surficial layer and alluvium would suggest that any decrease in infiltration rate caused by the surficial layer would be minor, but any deviation in estimation of infiltration rates could be magnified at the scale of pond infiltration.
Simulation of Water Movement Through a Thick Unsaturated Zone

In arid to semi-arid areas where the unsaturated zone can be hundreds of feet thick, and mounding is not an issue, the ability of water to infiltrate into the subsurface and to ultimately reach the water table within a timeframe suitable for management purposes is the deciding factor as to whether a recharge project is successful or unsuccessful. As an initial estimation of the time necessary for infiltrating water from a surface infiltration pond to reach the water table, models were created to simulate this scenario using the modeled data from the site with the highest and lowest infiltration rates, sites 2 and 4 respectively.

The Vs2D model domain was enlarged to include a one square-acre spreading pond at the surface as well as 400 feet of unsaturated zone overlying the aquifer. Although a two-layered model has been shown to best simulate surface infiltration rates for Site 2, a one-layered model having uniform hydraulic properties was used for this simulation. The unit of time for these models was increased from seconds to hours to allow for faster model runs.

Model Development

Model dimensions were created to simulate a radial cross-section through a circular spreading pond and a large area surrounding the pond. Infiltration of water within the pond is regulated by the head within the pond and estimated and laboratory derived hydraulic properties. Interaction of the wetting front with model boundaries was avoided.

DOMAIN

The model domain is 4500 feet in the radial, r, direction and 405 feet in the vertical, z, direction. A circular, one-acre spreading pond is simulated at the surface by active cells extending 117.75 feet from the top left corner of the model. The center of the pond is situated in the upper-left corner of the model domain. A pond depth of three feet is simulated by a constant pressure head of three feet applied to the active pond cells. A seepage face is located at the bottom most boundary to allow movement of water into the aquifer. The vertically oriented boundary farthest from center of the pond was designated as a no flow boundary, and was positioned at a great distance from the center of the pond so that there was no interaction with the volume of infiltrating water.
Grid

Grid spacing in the radial direction is a constant value of 0.08 feet (~ 1 inch) for the area underneath the spreading pond (0 – 117.75 ft.). With increasing radial distance (117.75 – 4500 ft.), spacing increases from 0.79 feet by a factor of 1.05 to a maximum spacing of 22.6 feet. Grid spacing in the vertical direction is a constant value of one foot.

Recharge Period

The recharge period for the model was 8760 hours (365 days). The initial time step was 0.001 hours, with a time step multiplier of 1.5 hours, and a maximum and minimum time step of 24 hours and 0.001 hours respectively.

Model Results

Model results for the Site 2 pond show infiltration rates of 21 ft/d and arrival of the wetting front at the water table 400 feet below land surface at 336 hr (14 days) (see Figure 13). Site 4 pond results show an infiltration rate of 5 ft/d and arrival of the wetting front at 1080 hr (45 days) (see Figure 14). These infiltration rate values are higher than what could be expected because the model assumes uniform hydraulic properties and does not account for compaction, consolidation, or changes in subsurface geology.

Figure 14. Simulated position of the wetting front below infiltration ponds at infiltrometer tests sites 2 and 4, Joshua Tree, California.
CONCLUSIONS AND DISCUSSION

Double-ring infiltrometer tests were done to measure surface infiltration rates to assess the potential for recharge through surface infiltration from ponds in the Joshua Tree groundwater subbasin. Infiltrometer tests showed infiltration rates as high as within the wash. Lower rates were measured in non-wash sites. Infiltration rates for most sites in this study reached a constant value within an hour. This is consistent with the predominantly sandy lithology of the alluvium and differs from infiltration data for silts or clays which have lower infiltration rates and require more time to reach a constant value.

Hydraulic conductivity of the near-surface alluvium was calculated from infiltrometer results using the unsaturated-zone flow model VS2D. Hydraulic conductivity values ranged from $4.63 \times 10^{-5}$ ft/d to $1.74 \times 10^{-4}$ ft/d, and were in the range of fine sand to silty sand. Modeled results were compared to laboratory values of hydraulic conductivity and values of hydraulic conductivity estimated from particle-size data using the computer program Rosetta. Results show a generally homogeneous, sandy subsurface, overlain by a low-permeability surficial layer within the wash. Good agreement between modeled hydraulic conductivity values and the laboratory hydraulic conductivity values suggests that surface measurements using a double-ring infiltrometer are useful in providing an estimate of the rate of water movement through the subsurface. The pedotransfer function program, Rosetta, proved to be a useful tool in estimating the van Genuchten constants based on basic particle size data, but was not an accurate method of calculating hydraulic conductivity. The ability of Rosetta to calculate hydraulic conductivity is limited by the small extent of input data.

Discrepancies between laboratory and modeled hydraulic conductivities may be a result of scaling. Hydraulic conductivity values measured from surface cores may be assumed to represent the surficial material, while modeled hydraulic conductivity values based on infiltrometer data can be assumed to be an average value for the volume of material tested. Using the two values and the equation for finding a geometric mean, a value for the volume of material underlying the surficial material can be calculated. Two-layer models
created using this method showed a better correlation to the infiltrometer data collected at the wash sites than models assuming a homogeneous subsurface.

Model results showed that infiltration data measured by a double-ring infiltrometer are sensitive to the presence of near-surface low permeability layers. The detection of layers of lower permeability than the surrounding alluvium is controlled more by the proximity of the layer to the surface than the thickness of the layer. Knowledge of the presence of near-surface low permeability layers can be used in planning pond design and construction. To ensure detection of low permeability layer, infiltrometer tests may need to run for long times. If near-surface low permeability layers are present it may be necessary to excavate those layers prior to large-scale infiltration of water from ponds. The model was not sensitive to low permeability layers at depths greater than about 4 feet. Unsaturated zone flow models of the wash sites showed evidence of a shallow low permeability layer, most likely a surface crust formed due to intermittent flow. The presence of a surface crust will decrease infiltration rates and depth of infiltration, and if not accounted for, recharge rates will may be overestimated. Reduction of infiltration rates caused by surface crusts encountered at this study area would not significantly reduce the time for aquifer recharge and could possibly be removed if desired. More precise description of the subsurface from test drilling data and instrumented boreholes (Izbicki et al., 2000, 2007) will help in creating a better estimate of water movement and recharge at greater depth.

Assuming uniform hydraulic properties estimated from field and laboratory data, model simulations show that water infiltrated from a one-acre pond would recharge the water table 400 feet below land surface within 14 to 45 days. A pond constructed north of the Pinto Mountain fault would require less time to reach the water table because of the higher hydraulic conductivity of the overlying deposits. However, the Pinto Mountain fault would constrain infiltration and recharge within the Copper Mountain subbasin. The homogeneous model does not account for varying hydraulic properties within the alluvium or decreases in hydraulic conductivity with depth caused by compaction, consolidation, or changes in subsurface geology and can only provide an estimate of the fastest expected recharge. For an actual test pond, it would be expected that infiltration rates would be slower, and recharge times would be greater, than the modeled results. Results from a pond in Victorville, CA (Izbicki et al., 2007) show recharge rates from 3 to 0.5 ft/d and a recharge time of 6 years.
Based on infiltrometer data, laboratory data, and numerical modeling, the study site in Joshua Tree, California is suitable for artificial recharge. Although recharge times can be expected to be greater than the results from the uniform hydraulic properties model, times should still be comparable to other ponds in the area. A more extensive survey of the subsurface geology and hydraulic properties of the study area including test drilling would help to assess subsurface conditions that would inhibit infiltration and recharge.
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