Measurement of Fine Spatial Scale Ecohydrologic Gradients in a Pinyon-Juniper Ecosystem

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MEASUREMENT OF FINE SPATIAL SCALE ECOHYDROLOGIC

GRADIENTS IN A PINYON-JUNIPER

ECOSYSTEM

by

Matthew D. Madsen

A thesis submitted in partial fulfillment
of the requirements for the degree

of

MASTER OF SCIENCE

in

Soil Science

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UTAH STATE UNIVERSITY
Logan, Utah

2008
ABSTRACT

Measurement of Fine Spatial Scale Ecohydrologic Gradients in a Pinyon-Juniper Ecosystem

by

Matthew D. Madsen, Master of Science

Utah State University, 2008

Major Professor: Dr. David G. Chandler
Department: Plants, Soils, and Climate

With the dramatic expansion of pinyon-juniper woodlands over the last century, improved understanding of how these woodlands modify infiltration properties is needed, in order for land managers to make informed decisions on how to best manage their specific resources. However, current methods for measuring soil infiltration are often limited by low sample sizes and high experimental error, due to constraints associated with remote, non agricultural settings. This thesis first presents a scheme for automating and calibrating two commercially available infiltrometers, which allows collection of a large number of precise unsaturated infiltration measurements in a relatively short period of time. Secondly, a new method to precisely determine saturated hydraulic conductivity from small intact soil cores collected in the field is demonstrated. This method removes bias due to measurement error using a multiple head linear regression approach. Finally, hundreds of fine spatial scale measurements of soil sorptivity, unsaturated hydraulic
conductivity, saturated hydraulic conductivity, soil water content, and other soil
descriptive measurements along radial line transects extending out from the trunk of
juniper (*Juniperus osteosperma*) and pinyon pine (*Pinus edulis*) trees. Within the
subcanopy of these trees, interactions among litter material, root distributions, and
hydrophobic soil significantly influence ecohydrologic properties by limiting and
redirecting infiltration below the soil surface. Consequently, hydrophobicity appears to be
a mechanism that promotes survival of woody vegetation in arid environments, through
decreasing evaporation rates from the soil surface. We further demonstrate how
differences in unsaturated infiltration and soil water content between the subcanopy and
intercanopy zones are not discrete. Unsaturated infiltration was significantly lower within
the subcanopy than in the intercanopy, and increased by eight-fold across a gradient
extending outward from near the edge of the canopy to approximately two times the
canopy radius. This gradient was not strongly related to soil moisture. In the intercanopy,
increasing structural development of biological soil crust cover beyond this gradient was
positivity correlated with infiltration capacity. Consequently, these results indicate that
the spatial location of the trees should be considered in the assessment and modeling of
woody plant and biological soil crust influence on infiltration capacity in a pinyon-
juniper ecosystem.
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Matthew D. Madsen
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CHAPTER 1
INTRODUCTION

The current extent of pinyon-juniper woodlands in the northern intermountain region is far greater than any other time in recorded history (West et al., 1975; Miller and Wigand, 1994). With woodlands expanding and canopy cover becoming denser (West et al., 1975; Miller and Wigand, 1994) a basic understanding of hydrologic processes in this ecosystem is critical for effective rangeland management (Huxman et al., 2005). However, the effect of these woodlands on rangeland hydrology is not well understood, with research in these lands often returning mixed results (Belsky, 1996). These conflicting results indicate that the ecohydrology and landscape impacts are not uniform across climate or soil types, and may vary with the scale of inquiry. In such complex systems, understanding the controlling physical processes may be the only way for land managers to make informed decisions on how to best manage their specific resources.

Even in a particular climate and soil type setting, the hydrology of pinyon-juniper woodlands is complicated by the heterogeneity of the land cover, which is characterized by distinct canopy and intercanopy regions within the vegetation patch scale (Reid et al., 1999). The distinct soil surface properties of these regions are generally understood to develop in response to the presence or absence of a vascular plant canopy. However, the surface cover of the intercanopy regions can vary greatly with the distribution of cyanobacteria, mosses and lichens (Belnap, 2003). Further, the dependence of soil physical properties such as unsaturated conductivity ($K(h)$) and water repellency is likely to vary seasonally due to the extremes of soil moisture characteristic of a semiarid climate.
The central objective of this study was to quantify the effects of woody vegetation and commonly associated biological soil crust (BSC) in altering the hydrology of the soil surface both underneath the vegetation canopy and extending into the intercanopy region. The approach was to assess several soil physical properties related to infiltration through performing fine scale field measurements. However, current methods for measuring soil infiltration are often limited by low sample sizes and high experimental error, due to constraints associated with remote, non agricultural settings. Consequently, these constraints make it difficult for researchers to determine the ecological significance of their data. Use of the commercially available disk infiltrometer (Mini Disk Infiltrometer (MDI), Decagon Devices, Pullman, WA) overcomes many of these field constraints. However, field experience indicated that the variability in vegetative cover and wide range of environmental conditions among the seasons required extensive replication of measurements in space and time. Additionally, use of the MDI requires the user to visually monitor and manually record drainage from the device while lying on the desert floor. This requires many person-hours of discomfort, which both introduces user error and limits measurement replication. Chapter 2 presents a simple and relatively economical way to automate the MDI as well as some important considerations for calibrating the sensor output and effects of measurement interval and disc diameter on $K(h)$ and sorptivity ($S$) calculations.

Measurement of saturated flow is also important for most hydrological studies and is generally represented as saturated hydraulic conductivity ($K_s$). Determination of $K_s$ through laboratory soil core measurements provides a researcher with greater control over sample saturation and temperature, sample scale, and greater measurement precision.
(Dirksen, 1999). However, despite the general acknowledgement that hydraulic gradients greater than unity are uncommon in the saturated zone and may adversely influence test results, few researchers ensure that gradients less than or equal to unity are maintained within test cores (Mitchell, 1993; Lal and Shukla, 2004). In Chapter 3 a linear regression data analysis approach is developed to test the assumption of linearity, improve measurement resolution, and identify appropriate experimental boundary conditions for laboratory measurements of $K_s$ from saturated soil cores. This multiple head linear regression method allows small soil cores collected in the field, to be used for precise saturated hydraulic conductivity measurements, when bias due to measurement error is separated mathematically.

In Chapter 4 these newly developed hydrologic methods are incorporated along with several other ecohydrologic measurements to investigate whether infiltration in pinyon-juniper woodlands can be described as spatial gradients which correspond to soil hydrophobicity and hydraulic conductivity related to tree position and species. This was accomplished by quantifying: (i) the ecohydrologic properties under the woody vegetation canopies of Utah juniper (*Juniperus osteosperma*) and pinyon pine (*Pinus edulis*); and (ii) how ecohydrologic properties changed with distance from the vegetation canopies and within the biological soil crust (BSC) dominated intercanopy. Hundreds of ecohydrologic measurements were taken to characterize soil properties related to the major controls on unsaturated infiltration, including hydrophobicity, biological soil cover, soil water content, short time infiltration (sorptivity), and steady state soil matrix infiltration (unsaturated hydraulic conductivity) at a high spatial resolution along transects extending from the base of *Pinus* and *Juniperus* trees to several meters within
their respective intercanopy patch. These results may have important implications for surface hydrology, soil moisture dynamics, ecosystem function, and rangeland management.

REFERENCES


CHAPTER 2

AUTOMATION AND USE OF MINI DISK INFILTROMETERS\(^1\)

ABSTRACT

Measurement replication and objectivity of field soil hydraulic properties can be increased through automation. The goals of this study were to test two automated versions of Mini Disk Infiltrometers (AMDI). Both devices were fitted with differential pressure transducers connected to compact data loggers. Instrument design, method of calculation and soil moisture condition all affected measured unsaturated hydraulic conductivity \((K(h))\) and sorptivity \((S)\) at \(h=-2.0\) cm. We found the type of AMDI with a capillary tube head control can be operated at inclination angles up to 25 degrees and returned the least variance in \(K(h)\) and \(S\) if data were not partitioned for calculation. However, changing the initial soil moisture content from 0.07 to 0.26 m\(^3\) m\(^{-3}\) was found to influence \(K(h)\) calculation by up to 50% for a silt loam soil. \(K(h)\) measured by the type of AMDI with a bubble chamber head control and larger disk diameter was less dependent on soil moisture content, but more sensitive to inclination.


INTRODUCTION

Infiltration of water into soil determines hydrologic response and ecological function in many environments and is a critical parameter in understanding land management and ecosystem modeling. Tension infiltrometry is an efficient technique
which can provide information on soil hydraulic properties including; hydraulic conductivity, sorptivity and macroporosity (Reynolds and Elrick, 1991; Zhang, 1997). Several designs for tension infiltrometers have been proposed (e.g. Perroux and White, 1988) and are commercially available. Infiltration rate measurements for large disk tension infiltrometers have been successfully automated with pressure transducers (Ankeny et al., 1988; Casey and Derby, 2002).

Field research in remote, non agricultural settings places several constraints on use of large disk infiltrometers including manual transportation of all research equipment, limited access to water, sloping soil surfaces with locally extreme microtopography and low, dense branch architecture of desert shrubs. We devised a scheme for automating a small, commercially available instrument, a Mini Disk Infiltrometer (MDI), to avoid the tedium and discomfort of lying on the desert floor, to allow instrument placement in visually obscure locations and to increase measurement replications and accuracy.

The purpose of this study is to test the response of two models of the automated Minidisk Infiltrometer (AMDI) to field conditions including use in sloping terrain, variable voltage supply and wet and dry initial soil moisture conditions.

**MATERIALS AND METHODS**

**Design**

The MDI has been manufactured (Decagon Devices, Pullman, WA) in two models, each of which requires slightly different automation schemes. The original MDI design (MDIv1), having a disk diameter of 3.2 cm, is automated by fitting the two ports of a differential pressure transducer (SenSym ASCX01DN Honeywell, Freeport, IL) into
a pair of holes bored in the rubber stopper at the top of the reservoir tube (Fig. 2-1a). A 20 cm section of polyethylene tubing is inserted in the stopper below the transducer pressure port B and extends to about 25 mm above the air inlet tube near the bottom of the infiltrometer. The dimensions of the air inlet tube determine operating tension. We present results of 2.0 cm supply tension for AMDIv1 infiltrometers, since this is recommended as the most broadly applicable model by the manufacturer.

Head control for the current MDI model (MDIv2) is provided by a bubbling chamber on the top of the instrument and necessitates a different setup (Fig. 2-1b). The reservoir of the MDIv2 is connected to a differential pressure transducer with external tubing (\(\frac{1}{16}\) inch diameter Nalgene) and threaded steel fittings (10-32 X .170 barb 5/16 O-Ring fitting straight connector, Pneumadyne, Plymouth, MN) 9.5 cm from the top and 3 cm from the bottom of the infiltrometer. The supply tension for the MDIv2 is determined by the depth of submergence of the suction control tube in the bubble chamber, minus the distance from the bottom of the Marriote tube to the porous disk. We present results of 2.0 cm supply tension for the AMDIv2 infiltrometers. The MDIv2 is filled by removing the elastomer housing which retains the 4.4 cm diameter sintered stainless disk to the bottom of the water reservoir. We fit a small valve near the bottom of the infiltrometer to release pressure during reassembly.

Infiltrometer support stands are fabricated from 10-cm long 3.2-cm diameter PVC tubes. Three support legs, made of 3 mm rod stock are held in grooves in the tubing by a metal hose clamp (Fig 2-1a). A 6-mm by 9-mm wing screw is installed through the side of the stand to suspend the infiltrometer preceding infiltration experiments and support the infiltrometer during measurements.
Transducer output \( (v_o) \) is recorded with a U12 4- External Channel Outdoor/Industrial Hobo data logger (Onset Computer Corporation, Bourne, MA). To avoid out of range signal voltage, transducers are powered with an external 6-V battery. Four transducers, allowing measurements with four infiltrometers at one time, one data logger and the battery are connected with simple wiring manifolds (Fig. 2-2b) and
Fig. 2-2. (a) Infiltrometer stand and (b) schematic wiring diagram to connect four ASCX01 DN pressure transducers (PT1-PT4) with input power ($v_i$) from a 6 VDC battery. Output voltage from the transducers ($v_o$) recorded with a Hobo U12 data logger ($DL$).

Installed in a plastic tool box, which also accommodates the infiltrometers and their support stands for transport and storage.

**Calculations**

Conversion of $v_o$ to volume of water in the infiltrometer ($V$) at time, $t$, is calculated by scaling the maximum $v_o$ ($v_{\text{max}}$) at the beginning of each trial measurement run and the minimum $v_o$ ($v_{\text{min}}$) at the end of each trial to the total volume of water discharged from the infiltrometer during the total measurement period ($V_{\text{tot}}$).

$$V(t) = V_{\text{tot}} \left[ 1 - \frac{v_o(t) - v_{\text{min}}}{v_{\text{max}} - v_{\text{min}}} \right]$$ [1]
The scale on the AMDIs over represents the volume in the reservoir due to displacement of water by internal tubing. \( V_{\text{tot}} \) is calculated as

\[
V_{\text{tot}} = 2 \pi L \left( R^2 - r^2 \right)
\]

[2]

where \( L \) is the distance between the initial water level in the reservoir and the bottom of the poly tube in AMDIv1 or the lower pressure transducer port in AMDIv2, \( R \) is the inside radius of the infiltrometers (1.27 cm) and \( r \) is outside radius of the poly tube (0.32 cm) for the AMDIv1 or the reservoir bubbling tube (0.24 cm) for the AMDIv2. \( S \) and \( K(h) \), are calculated following the approach of Zhang (1997) using the form of the two-term cumulative infiltration equation suggested by Decagon Devices:

\[
I = C_1 t + C_2 t^{1/2}
\]

[3]

where \( I \) is cumulative intake per unit area, \( t \) is time, \( C_1 \) and \( C_2 \) are parameters related to soil \( K(h) \) and \( S \), respectively. \( C_1 \) and \( C_2 \) were obtained by plotting cumulative infiltration vs. the square root of time and fitting the data to a 2nd order polynomial equation using spreadsheet software (e.g. Excel, Microsoft Corp., Seattle, WA). Note, that the first and second order coefficients correspond to \( C_1 \) and \( C_2 \), respectively. This reverses the order of the polynomial terms from equations commonly presented by other authors (e.g. Warrick, 1992; Zhang, 1997; Vandervaere et al., 2000a and b), but facilitates the use of a spreadsheet to fit the constants to the terms. \( K(h) \) and \( S \) were calculated by equations 4 and 5.

\[
K(h) = C_1 / A_2
\]

[4]
\[ S = \frac{C_2}{A_1} \]  

The dimensionless coefficients $A_1$ and $A_2$ were calculated according to Zhang (1997), using his equations 20, 21, and 22. $A_1$ depends on water content and disk diameter. We used 4.17 and 3.97 for dry conditions ($\theta_i = 0.07$) and 3.53 and 3.36 for wet conditions ($\theta_i = 0.26$) for the MDIv1 and MDIv2, respectively. $A_2$ depends on disk diameter and soil texture. For $A_2$ we used 3.33 and 2.48 for the MDIv1 and MDIv2, respectively for the Moab soil 10.88 and 8.09 for Greenville silt loam.

**Calibration and testing**

Measurement procedures followed instructions in the MDI user manuals. In addition, the AMDI was briefly suspended over the soil before running the infiltration experiment to aid in identifying $v_{\text{max}}$ and the start time of a trial during data analysis. Data loggers were programmed to record transducer output voltage at a 1-second interval during testing and calibration and at 3-second intervals in the field in order to conserve data storage space.

Laboratory testing was performed with soil from a field site near Moab, Utah. The soil is described as a Rizon-Rock outcrop complex (loam, mixed (calcareous), mesic, Lithic Ustic Torriorthents). Samples were obtained from the 0-10 cm depth, sorted through a 20.3 cm diameter, 2 mm sieve, mixed thoroughly and hand compacted to a depth of at least 20 cm in a tiled container. We first measured the sensitivity of the AMDI signal response to transducer input voltage, $(v_i)$ by suspending an AMDIv1 filled to 89 ml and varying $v_i$ from 0 to 9 volts. The effect of infiltrometer placement angle on $v_{\text{max}}$ and calculated $K(h)$ was tested for AMDIv1 and AMDIv2 by performing infiltration
experiments with the infiltrometers at six different angles from vertical: $0^\circ$, $5^\circ$, $10^\circ$, $15^\circ$, $20^\circ$ and $25^\circ$. All laboratory trials were conducted at room temperature between 20 and 25 degrees C, at soil moisture content of 0.35 m$^3$ m$^{-3}$, with room temperature water. Most trials were completed within 10 minutes.

Additional tests were conducted to assess the comparability of results from the AMDIv1 and AMDIv2 and what opportunities automation might provide for improvement over manual measurement. Two sets of experiments were performed at the Utah Agricultural Experiment Station Greenville Experimental Farm in North Logan on Millville silt loam (coarse-silty, carbonatic, mesic Typic Haploxerolls). To reduce spatial variability among measurements we performed all experiments on a 1 m$^2$ plot of recently tilled soil. A thin layer (~2 mm) of previously washed dry sand was used to improve contact between the disk and soil and to prevent disk clogging. Two successive measurements were performed in each set of experiments. The first test was performed on initially dry soil (~0.07 m$^3$ m$^{-3}$) and the successive experiment was performed in initially wet soil (~0.26 m$^3$ m$^{-3}$). In the first experimental set, three research assistants manually recorded visual observations of changes in infiltrometer reservoir volume in an AMDIv2 while electronic data were recorded by the automated technique. The infiltrometers were supported by hand because use of the support stand obscured the reservoir scale from sight. Automated $v_o$ measurements were made at one second intervals, and manual $V(t)$ records were taken at five second intervals for the first minute and 10 second intervals thereafter. Each investigator measured 3 locations for a total of 9 replicates for both dry and wet soil conditions. In the second experiment, the same experimental design was used as above to compare $K(h)$ and $S$ values calculated from
AMDIv1 and AMDIv2 measurements. Seven replicates were made under both wet and dry initial conditions. Support stands were used.

We took two approaches in calculating $K(h)$ and $S$: (1) by fitting the total cumulative infiltration data of each measurement to equation 3 and (2) by partitioning the cumulative infiltration into the nonlinear period dominated by $S$ and the near steady state period dominated by $K(h)$. Measurements were compared with the mean and coefficient of variation along with Pearson correlation coefficient ($R$ value) and statistical significance ($P$ value).

**RESULTS AND DISCUSSION**

In the laboratory tests on Moab soils using an AMDIv1 with a full water reservoir we found transducer $v_o$ is linear to $v_i$ from 8.9 V, at which point the maximum data logger input voltage (2.5 v) was exceeded, down to 1.5 V. Additionally, $v_{\text{max}}$ decreased with the cosine of the angle of infiltrometer inclination from 0 to 10 degrees for AMDIv1 and AMDIv2 (Fig.2-3a). However, the relationship does not hold for greater angles of inclination due to the more complex system trigonometry and an empirical calibration is required (Equation 1). Infiltration trials on an inclined sand bed demonstrate the capacity of the AMDIv1, when properly calibrated, to make consistent measurements of $K(h)$ at inclinations up to at least 25 degrees from vertical (Fig. 2-3b). For the AMDIv2, both $v_o$ and $h$ are dependent upon angle of inclination and the axial orientation of the infiltrometer. Pressure ports oriented upslope will record lower pressure heads and consequently lower $v_o$ when compared to pressure ports oriented down slope at the same angle. Whereas the effect on $v_o$ could be calibrated for consistent orientation during use
Fig. 2-3. Laboratory results of $v_{max}$ output voltage for full infiltrometers (AMDiV1 and AMDiV2) suspended at different angles and calibrated $K(h)$ values calculated from infiltration measurements with an AMDiV1 performed at 0-25 degrees from vertical.
of the AMDIv2, we observed that inclination also affected $h$ in the bubbling chamber. We calculated a decrease of 0.5 cm from a nominal $h$ of -2.0 cm as the AMDIv2 was inclined for 0 to 25 degrees from vertical. Accurate field measurements of $K(h)$ with the AMDIv2 would require additional effort to adjust the bubble chamber water level to account for inclination and care to maintain consistent orientation of the pressure ports relative to the slope.

The most significant correlations between manual MDIv2 and automated AMDIv2 data collection and between method of calculation were for $K(h)$ of wet soil (partitioned and complete) and $S$ of dry soil (complete), underscoring the control antecedent soil moisture conditions exert on these two aspects of infiltration calculations (Table 2-1). The poorer agreement between $S$ calculated from manual and automated records for partitioned data under dry initial conditions is evidently due to the greater influence of more precise automated data for short time measurements on partitioned calculation of $S$.

We found that experimental condition, instrument design and method of calculation all affected the results for $K(h)$ and $S$. The most consistent results (lowest

<table>
<thead>
<tr>
<th>Comparison</th>
<th>$K(h)$ dry R value</th>
<th>$K(h)$ dry P value</th>
<th>$K(h)$ wet R value</th>
<th>$K(h)$ wet P value</th>
<th>$S$ dry R value</th>
<th>$S$ dry P value</th>
<th>$S$ wet R value</th>
<th>$S$ wet P value</th>
</tr>
</thead>
<tbody>
<tr>
<td>M vs A partitioned</td>
<td>0.676</td>
<td>0.527</td>
<td>0.995</td>
<td>0.063</td>
<td>0.652</td>
<td>0.548</td>
<td>0.885</td>
<td>0.308</td>
</tr>
<tr>
<td>M vs A complete</td>
<td>0.201</td>
<td>0.871</td>
<td>1</td>
<td>0.006</td>
<td>1</td>
<td>0.011</td>
<td>0.739</td>
<td>0.47</td>
</tr>
</tbody>
</table>
coefficients of variation) for $K(h)$ for both wet and dry initial conditions and for $S$ for dry conditions were obtained by using the complete data records from the AMDIv1 for calculation (Table 2-2). However the difference between $K(h)$ results for wet and dry conditions from the complete records from AMDIv1 was nearly 50%, indicating a sensitivity of this approach to initial soil moisture. On the other hand $K(h)$ sensitivity to soil moisture was least for the AMDIv2 using complete records, but the coefficient of variation for dry conditions was three times that of the AMDIv1. We expect these differences are due to the change in disk diameter between the two devices. For similar reservoir volumes a larger disk, and hence greater measurement volume, would require proportionately greater allocation of water to $S$. However the subsequent flux to the wetting front under dry conditions would be less influenced by the suction at the wetting front.

We chose the AMDIv1 for our field studies (Lebron et al., 2007) because they were simple to operate, obviated any concern about placement angle and gave consistent results. The AMDIv1 proved reliable in the field under a variety of weather conditions and levels of operator expertise. For example quadratic equations fit to infiltration data vs time from one field campaign at the Moab site with 100 measurements resulted in

Table 2-2. Results of $K(h)$ and $S$ mean and coefficient of variation (CV) and comparisons between AMDIv1 (v1) and AMDIv2 (v2), for $h=-2.0$ cm.

<table>
<thead>
<tr>
<th>Comparison</th>
<th>$K(h)$ (x10^{-6} \text{ms}^{-1})</th>
<th>CV (%)</th>
<th>$S$ (x10^{-4} \text{ms}^{-0.5})</th>
<th>CV (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>v1 complete</td>
<td>dry 15.1 10.2 24.3 9.7 5.7 0.8</td>
<td>19.1 88.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>v1 partitioned</td>
<td>dry 12.2 10.1 32.5 20.5 5.2 2.2</td>
<td>28.8 27</td>
<td></td>
<td></td>
</tr>
<tr>
<td>v2 complete</td>
<td>dry 9 9.4 74.6 44.2 7.4 2.2</td>
<td>30.2 24.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>v2 partitioned</td>
<td>dry 6.8 8.8 81.4 41.3 6.7 2.5</td>
<td>39.2 23.6</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
average $r^2$ of 0.995. Also, the number of measurements a person was able to perform in a day increased at least four fold, up to 80-100 infiltration measurements per person-day. Automation and use of the tripod stands also allowed infiltrometer placement in low visibility locations and minimized disturbance to the study area.

Preliminary field application of the AMDI at the Moab site clarified the importance of scaling $V_{tot}$ to the voltage difference over the course of each measurement. The differences between $v_{max}$ and $v_{min}$ for a single test varied up to 50% among infiltrometers and up to 20% for a single infiltrometer over the course of a day, despite nearly constant infiltrometer recharge volumes. Likely causes for the variability include battery strain, infiltrometer placement angle, large changes in ambient air temperature and differences in cable resistance. Occasional failure of the automated instruments was caused by wetting the pressure transducers or forcing water into the port A tubing when filling the infiltrometers.

**CONCLUSIONS**

Automation of mini disk infiltrometers is a relatively simple and inexpensive technique and allows detailed in situ characterization of soil surface hydraulic properties in remote and difficult experimental settings. Measurement quality and number are enhanced through the use of a tripod support stand. Variability in transducer output voltage, $v_o$, can arise from a number of factors, including battery drain and infiltrometer inclination, but can be accounted for with a simple calibration. We found the original instrument (AMDIv1) provided consistent $K(h)$ results at moderate angles of inclination using our simple calibration approach. Use of the AMDIv2 on an incline would require
additional effort to maintain a constant tension in the supply reservoir and operation at a consistent axial orientation to the slope. The AMDIv1 also returned more consistent values of $K(h)$ than the AMDIv2, perhaps due to the smaller volume of water required to reach steady state by the AMDIv1 in the tested soils. Greater accuracy in $S$ and $K(h)$ can be achieved by taking sequential measurements on initially dry soil to account for the control exerted by the initial soil moisture content on these measurements.

This study did not address other factors that contribute to variability in infiltration, which include diurnal and seasonal temperature fluctuations, hydrophobicity, and secondary porosity. We encourage field researchers using this technique, however, to make complementary measurements of these properties to gain a comprehensive understanding of spatial and temporal variability in infiltration.

**REFERENCES**


CHAPTER 3

ACCOUNTING FOR BIAS AND BOUNDARY CONDITION EFFECTS ON MEASUREMENTS OF SATURATED CORE HYDRAULIC CONDUCTIVITY

ABSTRACT

Most hydrological studies require knowledge of saturated soil hydraulic conductivity, $K_s$. This parameter is often measured using saturated soil cores and a constant applied hydraulic head device. A standard approach to reduce uncertainty in the result is to conduct replicate tests at a single hydraulic head gradient. Low permeability soils are often tested at large hydraulic head gradients to decrease measurement time. Our objective was to test the common assumptions implicit in calculating $K_s$ from constant head laboratory tests; i.e. the theoretical linear relationship between head gradient and flux density exists in experimental data; and the relationship passes through the origin. In this paper we use linear regression analysis to test these assumptions and determine $K_s$ from a broad range of head gradients for: 4.5 cm (diameter) x 10 cm (height) intact cores of sandy loam soil; 10 cm (diameter) x 10 cm (height) intact cores of clay loam soil; and repacked sand columns of various sizes. We found non-linear relationships between hydraulic head gradient ($i$) and flux density ($q$) for tests conducted on intact cores of both soils, especially for head gradients greater than unity. When we calculated $K_s$ by linear regression of data from intact cores, we found average values approximately one third greater than the “standard” method of averaging several replicate tests at a single hydraulic head. The difference between the regression and standard analyses was attributed to experimental bias, which is removed by the linear regression.
Although no consistent $i$ or $q$ “thresholds” were identified to predict the onset of non-linearity in $i$ vs. $q$ data, the intact core results imply that $i<1$ and $q<5\times10^{-3}$ cm s$^{-1}$ may be advisable.

1Authored by Matthew D. Madsen, David G. Chandler and W. Daniel Reynolds

**INTRODUCTION**

Most hydrological studies require knowledge of soil permeability, which is most simply represented as saturated hydraulic conductivity, $K_s$ (cm s$^{-1}$). In particular, distributed hydrologic models require information on how $K_s$ is affected by soil texture and land cover, at the scale appropriate to the model (Gerke, 2006; Vanderborght et al., 2006). Ideally, $K_s$ would be measured at the scale of the representative element volume (REV) (Bear, 1972). However the REV is often difficult to determine, since it is sensitive to soil texture, soil structure and the support length scale of the experimental technique (Schulz et al., 2006).

The validity of the $K_s$ measurement also depends on instrument resolution and scale, experimental state and boundary conditions, and correctness of simplifying assumptions (Dane, 1980; Wessolek et al., 1984; Basile et al., 2003). Some advantages of laboratory measurement over field measurement include better control over sample saturation and temperature (Dirksen, 1999), use of well defined sample sizes (Fuentes and Flury, 2005), and greater measurement precision. Despite the general acknowledgement that hydraulic head gradients greater than unity are uncommon in the saturated zone and may adversely influence test results, few researchers ensure that
gradients \leq \text{unity} \text{ are maintained within test cores (Mitchell, 1993; Lal and Shukla, 2004). Linearity of the relationship between flux density and hydraulic head gradient is commonly cited as a test for the applicability of Darcy’s Law (Mitchell, 1993; Hillel, 1998; Lal and Shukla, 2004) but this test is not standard practice. In this paper we develop a simple linear regression approach which tests the assumption of linearity, improves measurement resolution, and identifies appropriate experimental boundary conditions for laboratory measurement of \( K_s \) on saturated soil cores. We apply this approach to soil cores of different spatial scales and textures to relate several cases of apparent flow threshold phenomena to experimental design, sample treatment and flow conditions.}

**Standard method of determining \( K_s \)**

The analysis of \( K_s \) from constant head measurements (see Figure 3-1) is based on Darcy’s Law (Darcy, 1856), which can be written

\[
K_s = \frac{Q}{A \Delta h} \quad [1]
\]

where \( Q \) is volumetric flow rate (\( \text{cm}^3 \ \text{s}^{-1} \)), \( A \) is cross sectional area perpendicular to flow (\( \text{cm}^2 \)), \( \Delta h \) is hydraulic head loss (cm), and \( L \) is distance between the points where \( \Delta h \) is determined (cm).

Laboratory determination of \( K_s \) depends on careful measurement of \( \Delta h \) and \( Q \). Head measurement may be made external to the core (e.g. Reynolds and Elrick, 2002; ASTM, 2003) or within the core (e.g. Bootlink and Bouma, 2002). Flow rate may be
determined manually or automatically by recording outflow into a graduated cylinder or onto an electronic balance (Borcher et al., 1987), or by recording inflow from a supply (Troyer and Skopp, 2002). The wide variety of methodologies used to determine $K_s$ in different laboratories raises the question of measurement parity, since each methodology may be subject to different degrees of measurement bias associated with sample collection conditions, head losses external to the soil, manometer meniscus readings, and flow rate measurement.

The standard method to obtain $K_s$ from constant head laboratory tests is to apply equation 1 to data from individual experiments (i.e. one measurement of $\Delta h$ and the corresponding $Q$) and then calculate a “point” value of $K_s$. Precision is assessed by averaging replicate measurements, made at a single value of $\Delta h$ or within a range conducive to accurate $Q$ measurement (Reynolds and Elrick, 2002; ASTM, 2003). It is common practice to attribute up to a two fold difference in $K_s$ from replicate samples to natural variability. This approach implicitly assumes: 1) a linear relationship between hydraulic head gradient and flux density; and 2) the relationship passes through the origin. Evidence of the failure of both of these assumptions is common in the literature under conditions that give rise to “non-Darcy” behavior.

Non-linear responses between head gradient and flux density have been attributed to non-laminar flow for high pore water velocities in sands (Hubbert, 1956) and apparent “threshold gradients” for initiation of water flux in clays. Proposed explanations for threshold gradients include non-Newtonian behavior of fluids (Miller and Low; 1963), particle migration and pore blockage (Mitchell and Younger, 1967), temperature related activation energy (Swartzendruber, 1968; Zhang et al., 2003), experimental artifacts such
as core end-cap effects (Gupta and Swartzendruber, 1962; Olsen, 1965; Mitchell and Younger, 1967; Chan and Kenney, 1973), inflections in non-linear gradients at low fluxes (Hansbo, 1960; 2003), and soil consolidation due to swelling (Smiles and Rosenthal, 1968). However, measurement bias and apparent ‘non-Darcy’ behavior would not be identified using Eq. [1] and the standard “point calculation” approach to $K_s$ determination.

**Linear regression method of determining $K_s$**

Darcy’s law can be written to more explicitly represent $K_s$ as the proportionality constant between flux density, $q$ and hydraulic head gradient, $i$ (e.g. Hillel, 1998).

$$K_s = \frac{q}{i} \quad [2]$$

where

$$q = \frac{Q}{A} \quad [3]$$

and

$$i = \frac{\Delta h}{L} \quad [4]$$

$K_s$ can thus be determined as the slope of the linear regression of $q$ versus $i$, over a range of $\Delta h$ and $Q$. In this approach, the regression constant $b$ can be used to determine $i$ for $q = 0$, henceforth referred to as the apparent “threshold” hydraulic head gradient $i_t$:

$$i_t = -\frac{b}{K_s} \quad [5]$$
For $i_t \geq 0$,

$$q = K_s(i - i_t), \quad q > 0 \text{ for } i > i_t \quad [6]$$

Alternatively, $q$ may be used as the independent variable and the regression equation becomes

$$i = \frac{1}{K_s} q + c \quad [7]$$

where $c$ is the regression constant, and

$$i_t = c = \Delta h_t / L_a \quad [8]$$

where $\Delta h_t$ is the apparent “threshold” hydraulic head loss ($\Delta h_t \geq 0$) and $L_a$ is the apparent sample length. This approach removes the dependence of $i_t$ on $K_s$ (Eq. [5]) and allows assessment of experimental uncertainty due to $\Delta h_t > 0$. A second advantage is that $q$ can be measured more accurately and precisely than $i$, and thus standard regression analysis (which minimizes the y-axis deviations but not the x-axis deviations) is more accurate. Although the concept of using multiple hydraulic head gradients and regression analysis to determine $K_s$ is not new, the regression fitting of Eqs. [7] and [8] to identify non-linearity in $i$ vs. $q$ data and remove experimental bias from the $K_s$ calculation is not evident in standard methods or other published literature.
We propose that $K_s$ calculation from laboratory core tests are commonly subject to error from systematic measurement bias and inappropriate experimental conditions. Hence, the objectives of this study were to: 1) demonstrate the determination of $K_s$ by regression fitting of Eqs. [7] and [8] to $i$ vs. $q$ data; and 2) identify the extent and causes of bias using the regression and standard “point calculation” approaches for a range of soil textures. Our hypothesis is that apparent thresholds to flux are artifacts caused by experimental bias and changes in the soil fabric arising from experimental boundary conditions foreign to typical in situ conditions outside of riparian zones and flood-irrigated fields. The implications for investigations of scale effects are briefly addressed for simple media through comparison of small repacked sand columns, a reanalysis of Darcy’s (1856) data from large sand columns, and undisturbed silt loam and clay loam cores of varying lengths.

MATERIALS AND METHODS

Undisturbed soil cores were collected from a study site near Moab, UT, where soils have developed from sandstone to form the Rizono-Rock out crop complex (loam, mixed (calcareous), mesic, Lithic Ustic Torriorthents). The hydrometer method (Gee and Or, 2002) was used to determine an average soil particle size distribution of 87% sand, 11 % silt, and 2 % clay. Soil cores were collected manually from bare soil and across the tire tracks of a seismic exploration vehicle in 4.5 cm diameter thin wall PVC soil core sampling cylinders (Geoprobe, Inc, Salina, KS). To minimize disturbance during sampling, the soil surface was prewetted by misting. The sampling cylinders were then hand pressed into the soil a few cm and filled with 15 to 20 ml of water. One hour after
the water had infiltrated, the sampling cylinders were hand pressed 10 to 15 cm into the soil and extracted. Excess soil was shaved from the bottom end of each cylinder with a sharp knife, resulting in cores 8 to 12 cm long. The sampling cylinders extended several cm above the top of the soil cores. The cores were capped for transportation and stored in the laboratory for two weeks at ambient summer conditions in UT (20-35 °C). The range of temperatures and moisture contents in the samples were within the ranges typical of field conditions and were therefore not expected to alter the soil within the cores. Soil bulk density within the site ranged from 1.2-1.6 g cm\(^{-3}\) (Lebron et al., 2007). Mean organic carbon by the Walkley-Black method for 115 samples from 0-10 cm soil depth at a nearby site was 9.4 g kg\(^{-1}\) and was not tested for the Moab samples in this study.

The first set of tests was performed on repacked columns of construction sand, washed construction sand, disturbed Moab site soil, and undisturbed samples from the Moab site. A 4 cm extension was affixed to the bottom of each sample to support the soil core and attach water inlet tubing to the sample column (Figure 3-1). The extension was filled level with 6 mm diameter gravel, separated from the soil with 0.85 mm wire mesh and sealed to the soil core tube with duct tape. Bored rubber stoppers were used to attach the apparatus to a water inlet and outlet manifold constructed of 6.4 mm interior diameter plastic tubing and four plastic clamps. The manifold could be configured for upward or downward flow though the sample and facilitated sample wetting. Prior to the experiment, the soil cores were slowly saturated with tap water from the bottom up, over a period of 24 hrs or until water began to pond on the surface of the soil. Municipal water in Logan, UT has been extensively characterized in cross laboratory comparisons and is unlikely to deflocculate soil aggregates due to high pH (7.84) and water chemistry.
Figure 3-1. Experimental setup for laboratory testing of sand samples, includes (1) Mariotte bottle and (2) test column containing soil core and graduated cylinder, shown in downflow configuration. Head gradient ($\Delta h$) measurements were made for all samples between the air inlet for the Mariotte bottle and the center of the outlet tube, $\Delta h_{m-o}$, additional measurements were performed by installing manometers ($m_1$, $m_2$, $m_3$, $m_4$) to measure $\Delta h$ within the test column and within the soil core.
dominance of Ca$^+$, Mg$^{++}$ and CO$_3^-$.

A subset of samples were saturated over a period of 4 days, as suggested by Reynolds and Elrick (2002), however the resulting $K_s$ values were similar regardless of saturation method.

Five constant head tests were conducted on each sample over a range of $\Delta h$ from 2.0 to 12.0 cm. This was achieved by sequentially raising the air inlet tube of a Mariotte device in 2 cm increments between tests (Figure 3-1). To avoid measurement uncertainty associated with light refraction in the Mariotte bottle, the applied head, $\Delta h_{m-o}$, was calculated by subtracting the length of the Mariotte air tube from the distance between the top of the air tube and the center of the outflow tube (Figure 3-1). We estimate the measurement precision was 2mm, however measurement accuracy may have been lower due to offsets caused by surface tension effects at the Mariotte air inlet and the outflow dripper. Outflow was measured manually at 10 min intervals using a graduated cylinder. $K_s$ was calculated as a point value for each $\Delta h$ using Eq. 1, and as the inverse slope of the linear regression of Eq. 7 using all measured $(q, \Delta h)$ data pairs for each sample. Four additional manometers were fitted to an undisturbed sand core to determine head loss effects on $K_s$ throughout the apparatus (Figure 3-1). The manometers were constructed by affixing 5 mm inside diameter plastic tubing to small barbed fittings which had been sealed into holes drilled one cm above ($m_1$) and below the soil core ends ($m_4$), and at a spacing of 5.5 cm within the soil core ($m_2$ and $m_3$) and suspending then vertically parallel to the soil core.

The second set of tests was performed on intact soil cores of Brookston clay loam soil (fine, loamy, mixed, mesic, Typic Argiaquoll) obtained from two depths in an experimental plot in Ontario, Canada under continuous corn production since 1959.
Three 10 cm diameter x 10 cm long cores were taken from the 10-20 cm depth, and the 20-30 cm depth. Average composition for the soil samples was 28% sand, 35% silt, and 37% clay; pH was 6.1-6.5. Organic carbon content was 14-18 g kg\(^{-1}\), as determined via dry combustion (Skjemstad and Baldock, 2007) in a LECO CN-2000 Carbon Analyzer (LECO Corp., St. Joseph, MI). Average soil bulk density was 1.5 g cm\(^{-3}\) for the 10-20 cm samples and from 1.6 g cm\(^{-3}\) for the 20-30 cm samples. The soil structure is medium-coarse subangular blocky with shrinkage cracks, root channels and worm holes. Three sets of 3 replicate constant head tests were performed on each core within a range of \(\Delta h\) from 4.1 to 111 cm, using the saturated tank constant head method (Reynolds and Elrick, 2002).

Hydraulic conductivity tests on clay samples have previously been compared to straight lines on graph paper (Mitchell and Younger, 1967), but to our knowledge linear regression has not previously been used as a tool for data analysis. To provide context of scale, we applied the technique to data from Darcy’s original experiments on large columns (\(L=0.6-1.7\) m, \(A=0.1\) m\(^2\)), which included tests on unwashed sand (Experiment 1, series 1 and 2) and washed sand (Experiment 1, series 3 and 4) from the Seine River, France (Darcy, 1856).

**RESULTS AND DISCUSSION**

**Identification of experimental bias**

At large scale, Darcy’s (1856) tests with repacked sand columns were conducted over a wide range of hydraulic head gradients and exhibit very linear responses (Figure 3-2, Table 3-1). However, one exception to linearity in Darcy’s trials appears in series 1.
Figure 3-2. Data from Darcy’s experiment 1, series 1-4 with linear regression trend lines (solid) and confidence intervals (dotted). Inset shows trend lines and confidence intervals near the origin.
Table 3-1. Saturated hydraulic conductivity, $K_s$, calculated as the slope of the linear regression (Slope of Line) and average of several point calculations (Point) for constant head measurements made at different head gradients, $i$. The apparent threshold head gradient, $i_t$, is presented for the linear regression results. $P$ gives the statistical significance level for $K_s$ and $i_t$.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Slope of Line</th>
<th>Point</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$K_s$</td>
<td>$P$</td>
</tr>
<tr>
<td></td>
<td>cm s$^{-1}$</td>
<td>$\alpha=0.05$</td>
</tr>
<tr>
<td>Large columns (Darcy)</td>
<td></td>
<td>$i_t$</td>
</tr>
<tr>
<td>river sand series1</td>
<td>2.7 x10$^{-2}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>river sand series2</td>
<td>1.6 x10$^{-2}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>river sand series3+4</td>
<td>2.1 x10$^{-2}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>Small sand columns</td>
<td></td>
<td></td>
</tr>
<tr>
<td>repacked loamy sand</td>
<td>2.3 x10$^{-2}$</td>
<td>0.0027</td>
</tr>
<tr>
<td>construction sand</td>
<td>1.8 x10$^{-2}$</td>
<td>0.0011</td>
</tr>
<tr>
<td>washed sand</td>
<td>2.9 x10$^{-2}$</td>
<td>0.0039</td>
</tr>
<tr>
<td>Clay loam columns</td>
<td></td>
<td></td>
</tr>
<tr>
<td>core 174, 10-20cm</td>
<td>5.6 x10$^{-3}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>core 215, 10-20cm</td>
<td>2.0 x10$^{-3}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>core 322, 10-20cm</td>
<td>4.8 x10$^{-3}$</td>
<td>&lt;0.0001</td>
</tr>
<tr>
<td>core 217, 20-30cm</td>
<td>3.6 x10$^{-3}$</td>
<td>1.63</td>
</tr>
<tr>
<td>core 352, 20-30cm</td>
<td>4.4 x10$^{-4}$</td>
<td>0.53</td>
</tr>
<tr>
<td>core 241, 20-30cm</td>
<td>1.8 x10$^{-4}$</td>
<td>2.08</td>
</tr>
</tbody>
</table>
The test result for $i=18.8$ falls outside of the 95% confidence interval for the regression of the remaining data, perhaps as evidence of divergence from Darcy flow at this high hydraulic gradient (Figure 3-2). Regression of Darcy’s data shows a consistent small negative bias in $i_t$ (Table 3-1). At this large experimental scale such bias is insignificant, and the regression confidence intervals ($\alpha=0.05$) bracket the origin for all experiments except for series 1 (Figure 3-2, inset). The assumptions of a linear relationship between applied head and flux that intersects the origin are essentially satisfied for Darcy’s experiments.

At small scale, our tests with repacked sand columns were made at hydraulic head gradients and flux densities an order of magnitude less than Darcy’s tests (Figure 3-3) and measurement error is significant. In particular, the relative magnitude of $i_t$ (Table 3-1) is greater for this experimental scale than for Darcy’s experiments and the emergent bias ($i_t \sim 0.2$) is significantly greater than zero ($\alpha=0.05$). Nevertheless the range of $K_s$ as calculated by linear regression is similar to those from Darcy’s experiments with comparable material, despite the order of magnitude difference in experimental scale (Table 3-1).

The source of the emergent bias in the small sand column experiments was investigated through tests with a core of undisturbed soil, using the approach of Olsen (1965). Results from measurements of manometers above and below the ends of the soil core (m4-m1) and within the soil core (m3-m2) were nearly identical, but the difference in $i_t$ between the results from m4-m1 and m3-m2, and measurements made between the Mariotte bottle and the drip tube outlet (m-o) are approximately equal to the emergent bias (Figure 3-4). We consequently attribute the bulk of the differences in $i_t$ between the
system measurement and the internal manometers to measurement uncertainty, such as drip formation and detachment at the outflow dripper, and head loss within the instrument. However, the bias in all manometer pairs changed when the flow direction was reversed from downward (down) to upward (up) (Figure 3-4). Similar “hysteretic” behavior was previously attributed to head loss across an air bubble in a capillary tube manometer (Olsen, 1965) and may indicate incomplete sample saturation in our soil core, but could conceivably also be caused by directional particle straining, presence of bacteria, or adsorbed water effects (Mitchell, 1993). In any case, the slope of the lines was not affected by the bias or the change in flow direction for any pair of manometers (Figure 3-4), and $K_s$ calculated by linear regression remained between $3.31 \times 10^{-3}$ and
Figure 3-4. Simultaneous measurements of $\Delta h$ between the Mariotte bottle air inlet and the soil core dripper outlet were made by manometers above and below the ends of soil core (m4-m1) and within the soil core (m3-m2). The effects of flow direction (up or down) and placement of manometers on $i$, were measured for a single undisturbed core ($L=8.5$ cm, $A=13.9$ cm$^2$).

$3.40 \times 10^{-3}$ cm s$^{-1}$ for all manometer pairs and test conditions, which was nearly an order of magnitude less than the $K_s$ obtained from the repacked column of soil from the same site (Table 3-1, repacked site soil).

We therefore contend that $i$, associated with systematic measurement error is inconsequential to the calculation of $K_s$ by the linear regression approach presented here. On the other hand, we found measurement bias influenced point $K_s$ calculation for all repacked columns (Table 3-1) and the extent of this influence, relative to regression $K_s$, increased for decreasing values of $i$ (Table 3-1). Similarly, the relative impact of
experimental bias would be the greatest on cores with low $K_s$ (Eq. 5). One might conclude such errors could be reduced by increased applied head or more precise measurement of $\Delta h$, to allow similar values of $K_s$ to be determined from point calculations and linear regression. However, a simpler approach might be to simply increase the number of $(q, \Delta h)$ measurements per sample.

**Boundary condition effects**

The influence of test boundary conditions on apparent deviations from Darcy flow are investigated for several undisturbed soil samples from the Moab sand and Brookston clay loam. These two soils are extremely different in origin, texture, and disturbance history and should bracket most flow phenomena encountered in undisturbed samples. The sand cores were taken as part of an ecohydrology study within a pinyon-juniper woodland (Lebron et al., 2007) where spatial variability in $K_s$ was hypothesized to be predominantly horizontal and related to the effects of disturbance and vegetation. The clay loam cores were taken in a long term agronomic study where, within a given treatment, vertical changes are expected to dominate variability in $K_s$ due to changes in secondary porosity.

For the Moab sand we present results from intact cores collected across the path made by seismic exploration vehicles (n=17), and near an ephedra bush (n=8) and bunch grasses (n=3). The data from across the vehicle track are presented in Figure 3-5 as open diamonds for sample points within a tire track and closed boxes for samples points clearly outside of a tire track. Each sample is labeled by position in cm along a linear transect. The $i$ vs. $q$ relationships for the trafficked area tended to be more variable and of greater slope (lower $K_s$) than for outside the trafficked area. The relationship between $i$
Figure 3-5. Results from intact sand cores collected across the path made by seismic exploration vehicles at the Moab field site. Each sample is labeled by position in cm along a linear transect and identified by data marker type as either within a tire track (open diamonds) or clearly outside of a tire track (closed boxes). Data series were separated into (a) highly linear and (b) nonlinear with a criterion of $R^2 = 0.995$ to discriminate for linearity among the 7 samples identified as nonlinear, 6 demonstrated an increase in slope of $i$ vs. $q$ for increasing $i$, especially for $i>1$. 
and $q$ is highly linear for most samples and we used $R^2<0.995$ to identify non-linearity. Within the more linear results (Figure 3-5a) the average $K_s$ for the trafficked area ($3.3 \times 10^{-3}$ cm s$^{-1}$) was about one half that of the undisturbed area tracks ($7.0 \times 10^{-3}$ cm s$^{-1}$) and significantly different ($P<0.001$). Among the 7 samples identified as non linear, 6 demonstrated an increasing slope for increasing $i$, especially for $i>1$ (Figure 3-5b). This response is consistent with the hypothesis that particle straining is a common cause for non-Darcy flow as proposed for compacted clays (Mitchell, 1993), but occurs at substantially lower $i$ for intact sand samples, as proposed by Lal and Shukla (2004). Exclusion of data for $i>1$ from the non linear samples left core 360 as the only sample with a non-linear regression. Core 360 was obtained from the middle of a well traveled footpath and as a result had dramatically lower $K_s$ ($1.0-1.4 \times 10^{-3}$ cm s$^{-1}$). Regrouping the $K_s$ results for $i<1$ for the samples in Figure 3-5b (except core 360) with the results from the highly linear samples (Figure 3-5a) increased the mean $K_s$ for the trafficked ($3.8 \times 10^{-3}$ cm s$^{-1}$) and undisturbed ($7.3 \times 10^{-3}$ cm s$^{-1}$) areas slightly but the difference remained significant ($P<0.001$).

The results for samples taken near vascular plants at the Moab site were mixed. Cores from under bunchgrass all show highly non linear relationships between $i$ and $q$, but with a decrease in slope for increasing $i$ (Figure 3-6a), which could be attributed to a greater relative flow through the secondary porosity at $i>1$, or by progressive dissolution of entrapped air for sequential tests from low to high head gradients (Sato et al., 2005). For $i<1$, average $K_s$ ($4.0 \times 10^{-4}$ cm s$^{-1}$) for these samples is an order of magnitude lower than the tire compacted soils and about one third that of the footpath. This low $K_s$ is supported by the following observations we made for these samples: 1) Soil texture under
Figure 3-6. The results for samples taken from under bunch grass (a), and from near an ephedra bush (b). The bunch grass samples show a decrease in slope of $i$ vs. $q$ for increasing $i$, especially for $i>1$. The ephedra results are highly linear (all $R^2>0.995$) up to $i>2$. 
bunchgrass is somewhat finer than the texture of the other samples; 2) Soil water repellency is common in association with vegetation at the site; and 3) The bunchgrass samples were difficult to saturate and retained an uncertain proportion of entrapped air during the test. Results from samples taken near ephedra are highly linear (all $R^2 > 0.995$) up to $i > 2$ (Figure 3-6b). Six of seven tests group tightly (Figure 3-6b) with an average $K_s$ of $2.3 \times 10^{-3} \text{ cm s}^{-1}$, once again lower than the soils in the vehicle tracks. The $K_s$ for the remaining sample was nearly three times greater ($7.1 \times 10^{-3} \text{ cm s}^{-1}$), and similar to the undisturbed soils in Figure 3-5a.

The undisturbed clay loam cores have a wider range of permeability than the undisturbed sand cores and required a similarly greater range in $\Delta h$ for testing (Figure 3-7). The samples from 10-20 cm depth (174, 215, 322) had greater secondary soil structure (e.g. worm holes, root channels, interpedal cracks) than samples from 20-30 cm depth (217, 241, 352); and as a result, sufficient flux for the test was achieved at much lower $i$ for the shallow samples than for the deep samples (Figure 3-7a,b). Whereas 2 of 3 of the less permeable samples (from 20-30 cm depth) show a significant positive $i_t$, the more permeable samples (from 10-20 cm depth) do not (Table 3-1). This result does not support the common premise that increasing $\Delta h$ in low permeability samples is an effective means to improve the accuracy of the point method for calculating $K_s$.

Regardless of statistical significance, $K_s$ calculated using the point method underestimates the slope method (Eq. [7]) when the gradient-axis intercept (Eq. [8]) is positive (this also occurred with the sand samples) and it overestimates the slope method when the gradient-axis intercept is negative. Physical interactions between sample and test conditions can be inferred from the undisturbed cores of loamy sand and clay loam.
Figure 3-7. Results from undisturbed clay loam soil cores. Samples from 10-20 cm depth (a) had greater secondary soil structure than samples from 20-30 cm depth (b), and thus a much lower $i$ was required to achieve adequate flux for the shallow depth than for the deep depth. Sediment was observed in the outflow from two samples (c) which also showed a stepwise reduction in the linear slope of $i$ vs. $q$ for increasing $i$. 

soils. For example, two tests on the clay loam soil (cores 41 and 65) gave “piecewise linear” \( i \) vs. \( q \) relationships with two distinct regions - one “high slope” region at “low” flux density, and one “low slope” region at “high” flux density (Figure 3-7c). In addition, sediment was observed in the outflow at the high flux densities. Given that the cores were visibly structured and thus had "bimodal" pore size distributions, we interpret these results as indicating that flow was concentrated in large pores for the high flux tests, and that the appearance of sediment in the outflow was due to progressive erosion of the large pores. Hence, the high flux tests appeared to be modifying the soil fabric by producing pore water velocities in large pores which exceeded the soil’s shear strength. Together, data from cores of loamy sand soil from the disturbed track (Figure 3-5b) and from under bunchgrass are non-linear over an order of magnitude in \( q \) (Figure 3-6a). There is no clear threshold in \( q \) to non-linear response, but the trends appear linear for both sets of data for \( i < 1 \). Separate calculation of \( K_s \) for data taken at relatively high and low \( q \) for the bunchgrass samples and for clay loam cores 41 and 65 generally differ by greater than two fold (Table 3-2). On the other hand several of the undisturbed samples did not exhibit non-linear response at any of the test conditions. This variability in non linear response does not support tests using high outflow rates, but it does support tests conducted at multiple values of \( i < 1 \).

**CONCLUSIONS**

In this paper we have demonstrated that linear regression of hydraulic head gradient against flux density is a robust technique to determine \( K_s \) independently of systematic error. For repacked columns, on the scale of Darcy’s original work (Darcy,
45

Table 3-2. Saturated hydraulic conductivity, \( K_s \), and apparent threshold head gradient, \( i_t \), calculated by linear regression for fluxes, \( q \), above (high \( q \)) and below (low \( q \)) inflections in the laboratory test data trends.

<table>
<thead>
<tr>
<th>Sample</th>
<th>( K_s ) (cm s(^{-1}))</th>
<th>( i_t )</th>
<th>( K_s ) (cm s(^{-1}))</th>
<th>( i_t )</th>
</tr>
</thead>
<tbody>
<tr>
<td>loamy sand bunchgrass A</td>
<td>1.9 \times 10^{-3}</td>
<td>0.87</td>
<td>4.4 \times 10^{-4}</td>
<td>0.07</td>
</tr>
<tr>
<td>loamy sand bunchgrass B</td>
<td>1.2 \times 10^{-3}</td>
<td>0.76</td>
<td>2.9 \times 10^{-4}</td>
<td>0.06</td>
</tr>
<tr>
<td>loamy sand bunchgrass C</td>
<td>1.2 \times 10^{-3}</td>
<td>0.81</td>
<td>2.2 \times 10^{-4}</td>
<td>-0.11</td>
</tr>
<tr>
<td>clay loam core 41</td>
<td>1.6 \times 10^{-2}</td>
<td>6.0</td>
<td>5.8 \times 10^{-4}</td>
<td>2.06</td>
</tr>
<tr>
<td>clay loam core 65</td>
<td>7.2 \times 10^{-3}</td>
<td>4.67</td>
<td>5.1 \times 10^{-4}</td>
<td>1.54</td>
</tr>
</tbody>
</table>

1856) this is not an important consideration, but the error component becomes significant at the scale of intact soil cores typical of contemporary studies. We propose that separating error or bias in \( \Delta h \) into the regression constant \( b \) allows highly precise measurement of \( K_s \) from small cores. The degree of linearity indicated by the regression correlation coefficient is also a useful indicator of data quality and allows the investigator to identify samples or head gradients for which the test is invalid.

We found non-linear \( i \) vs. \( q \) relationships for tests conducted on both sandy loam and clay loam soils. We infer that increasing slope of \( i \) vs. \( q \) with increasing \( i \) for disturbed or compacted soil samples is the result of flow restriction caused by progressive blocking of pores by migrating soil particles, especially at \( i > \) unity. On the other hand, we infer that locally high flow rates can progressively erode low resistance conduits (large pores), causing a decrease in the slope of \( i \) vs. \( q \) with increasing \( i \). Although no consistent \( i \) or \( q \) “threshold” values were identified to predict the onset of non-linearity in \( i \) vs. \( q \) measurements, the results from the intact sandy loam and clay loam cores imply that \( i < 1 \) and \( q < 5 \times 10^{-3} \) cm s\(^{-1}\) may be advisable. When used within these boundary
condition guidelines, the linear regression approach to calculating $K_s$ should allow sufficiently precise determination of $K_s$ to identify significant differences in bulk hydraulic properties within the generally accepted range of “natural variability”.

The advantage of the regression approach based on Eqs. [7] and [8] lies in its ability to readily identify experimental artifacts (non-linearity in $i$ vs. $q$) and correct for bias (threshold hydraulic gradients), which would otherwise remain hidden when using the standard “point calculation” method. Hence, adoption of the regression approach will serve to improve the quality and inter-comparability of $K_s$ values determined with different apparatuses and at different scales. This will be an important step toward determining the REV relevant to saturated flow through heterogeneous geologic porous media.

REFERENCES


CHAPTER 4

SPATIAL GRADIENTS IN ECOHYDROLOGIC PROPERTIES WITHIN A PINYON-JUNIPER ECOSYSTEM

ABSTRACT

The influence of woody vegetation and commonly associated biological soil crust (BSC) on infiltration capacity is one of several uncertainties associated with the ecohydrologic effects of woody plant encroachment into arid and semi-arid land systems. The objective of this study was to quantify the effects of Utah juniper (Juniperus osteosperma) and pinyon pine (Pinus edulis) in altering ecohydrologic properties under the tree canopies and into the intercanopy space. To meet this objective we established sampling points at 30 cm intervals along radial line transects extending from the trunk of seven Juniperus and seven Pinus trees into the center of the intercanopy space between the trees. For each sampling point we measured soil sorptivity, unsaturated hydraulic conductivity, soil water content, and soil hydrophobicity. Comparisons between Juniperus and Pinus were statistically similar. Within the subcanopy, interactions among hydrophobic soils, litter material and root distributions significantly influenced ecohydrologic properties. Results indicate that for an established woody plant, hydrophobic soil may be ecological advantages by channeling precipitation inputs deeper into the soil through a few select wet patches, thus minimizing evaporative losses at the soil surface. This study also shows that the influence of these species on hydrologic properties extends significantly into the intercanopy region. Within the subcanopy unsaturated infiltration was significantly lower than the intercanopy, but also increased by eight-fold across a gradient extending outward.
from near the edge of the canopy, to approximately two times the canopy radius. We hypothesize that this gradient is related to the trees’ influence on subcritical water repellency and soil microclimate, both of which diminish with distance into the intercanopy. Analysis of BSC cover within the intercanopy showed that increasing structural development was associated with increased infiltration capacity beyond the transitional zone. Furthermore, these results indicate that distance from the canopy and direction should be considered in the assessment and modeling of woody plant and BSC influence on infiltration capacity.

INTRODUCTION

The expansion of pinyon-juniper woodlands into arid grassland ecosystems over the last 130 years (West, 1999) is having a pronounced effect on ecohydrologic processes (Huxman et al., 2005). However, differences in studies among pinyon-juniper woodlands have made it difficult to apply information from one region to another (Gottfried and Pieper, 2000) and have resulted in much debate into how these expanding ecosystems should best be managed (West et al., 1975; Belsky, 1996). In such complex systems, understanding the controlling physical processes may be the only way land managers can make informed decisions on how to manage their specific resources. Understanding these systems requires knowledge of both how these woody plants modify ecohydrological processes directly beneath them and in the intercanopy space between the trees (Breshears, 2006).
The development of hydrophobic soil has been shown to be a key process affecting hydrologic properties under the canopy of many woody species (Scholl, 1971; Doerr et al., 2000; Pierson et al., 2003; Mataix-Solera et al., 2007). However, the impact of hydrophobicity is scale dependent. At small scales, hydrophobic soils reduce soil wetting considerably and enhance runoff responses (Doerr et al., 2000). On the other hand, rainfall simulation tests by Roundy et al. (1978) showed significantly greater infiltration capacity under well-developed *Juniperus* and *Pinus* mounds than in the intercanopy space. Because hydrophobicity can extend several centimeters into the soil profile (Jaramillo et al., 2000), infiltration and redistribution of soil moisture throughout the soil matrix depends on wet patch conduits. These uneven wetting patterns associated with hydrophobic soil (Dekker and Ritsema, 2000) have been hypothesized to be a water conservation mechanism for established woody plants, as these surface hydrophobic zones would minimize evaporation losses from the soil surface (Scott, 1992; Moore and Blackwell, 1998; Jaramillo et al., 2000).

Despite the potential for significant ecohydrologic impacts from hydrophobic soil, detailed analysis of its effects are rarely considered in typical non-fire related studies. In general, results of various types of saturated or near saturated infiltration measurements performed on pinyon-juniper woodland soils simply point toward higher infiltration rates under the canopies when compared to the intercanopy region (e.g. Blackburn and Skau, 1974; Wilcox et al., 2003) with little evidence showing how hydrophobicity affects infiltration. In these studies the hydrophobic layer, if evident, may have been overcome in field infiltration studies because hydrophobicity decreases with increased soil moisture (Scholl, 1971; Doerr et al., 2000). In addition, many conventional infiltration and
saturated hydraulic conductivity measurements, aimed at reaching steady state, are performed with the soils at or near saturation. Understanding the infiltration capacity of the soil surface during the initial stages of an infiltration event, when the soil is still unsaturated and potentially hydrophobic, is critical in many arid regions, as a significant portion of the rain events are characterized by high intensity short duration storms (Reid et al., 1999). Furthermore, since hydrophobic soils may never reach saturated conditions throughout the soil matrix, during a precipitation event, where the soil previously was under dry antecedent moisture conditions (Kobayashi and Shimizu, 2007), methods which measure hydraulic properties under saturated and or steady state conditions may insufficiently characterize infiltration in arid grassland and pinyon-juniper ecosystems.

Understanding how woody plants modify the ecological processes within the intercanopy region is also critical in understanding the effects of pinyon-juniper encroachment. One approach has focused on dividing the landscape into distinct scale-dependent hydrologic functional units (HFUs) (Wilcox et al., 2003) to better quantify landscape patterns and processes (Ludwig et al., 2005). While insight has come about as a result, this approach does not address how woody vegetation modifies ecohydrologic properties beyond the canopy edge. For example, improved infiltration capacity could extend beyond the canopy edge and thus have a significant role in patch to landscape scale infiltration processes (Lugwig et al., 2005). Recent studies have shown woody plants can affect soil microclimate well into the intercanopy region through shading and interception, while affecting below ground resources through root uptake (Breshears, 2006; Lebron et al., 2007). In a transect study with a Pinus tree and a Juniperus tree as endpoints, Lebron et al. (2007) attributed a transition zone in soil moisture between the
canopy and intercanopy HFU’s to canopy interception and modified the existing HFU model to address meteorologically-driven directional gradients in ecohydrological processes. They observed a similar gradient in unsaturated hydraulic conductivity away from the *Juniperus* canopy into the intercanopy, but it was not clear whether the result was solely dependent on soil moisture.

In this paper we assess whether infiltration in pinyon-juniper woodlands can be described as spatial gradients which correspond to soil hydrophobicity and hydraulic conductivity related to tree position and species. This was accomplished by quantifying: (i) the ecohydrologic properties under the woody vegetation canopies of Utah juniper (*Juniperus osteosperma*) and pinyon pine (*Pinus edulis*); and (ii) how ecohydrologic properties changed with distance from the vegetation canopies and within the biological soil crust (BSC) dominated intercanopy. Hundreds of ecohydrologic measurements were taken to characterize soil properties related to the major controls on unsaturated infiltration, including hydrophobicity, biological soil cover, soil water content, short time infiltration (sorptivity), and steady state soil matrix infiltration (unsaturated hydraulic conductivity) at a high spatial resolution along transects extending from the base of *Pinus* and *Juniperus* trees to several meters within their respective intercanopy patch. These results may have important implications for surface hydrology, soil moisture dynamics, ecosystem function, and rangeland management.

**MATERIALS AND METHODS**

**Study Area**

The study site is located on the Colorado Plateau near Moab, Utah, southwest of
Dead Horse Point State Park, (38° 32’38.02 N, 109° 44’01.73 W, at an elevation of 1829 m. The climate is semi-arid, with a mean annual precipitation of 21.7 cm (Davidson et al., 2002). The hydrology of this region appears more sensitive to the timing and characteristics of individual events than to seasonal trends in precipitation and temperature (Spence, 2001). Soils are developed from sandstone parent material, to form the Rizono-Rock out crop complex (loam, mixed (calcareous), mesic, Lithic Ustic Torriorthents) and are generally thin and poorly developed. Texture of the soils was found to be a loamy sand with similar texture within both the subcanopy soil and intercanopy (analysis of particle size is reported in Lebron et al. (2007). Soil organic matter was higher under vascular plant canopies, where rich organic O and slight A1 horizons develop, than the intercanopy zone. The vegetation structure of the study location is described well by West’s (1999), i.e., a mixed woodland of large Pinus and Juniperus tree’s, with the intercanopy soil surface covered with well developed BSC.

Study Design

Studies were conducted March 14-20, 2006, following snowfall of approximately 13 cm. Within a three hectare area, we chose seven Juniperus and Pinus trees that were of typical size and that had various degrees of BSC development within the intercanopy area. To assess the distribution of important hydrological parameters for soil wetness patterns developed under natural conditions, measurements were done just after the snow melted. The variable timing of melt due to surface shading by the canopies allowed us to make several replicate measurements under similar conditions by following the melt throughout the site. For each tree we measured soil physical properties related to infiltration every 30 cm along one radial transect, extending out from the base of the
woody vegetation to approximately the center of the intercanopy area. Transects were oriented to provide a reasonably even representation of cardinal orientation and to avoid intersection of obstacles such as fallen logs, disturbed areas, bare rock, and erosion features. Distance from the trunk to canopy edge was similar for *Juniperus* and *Pinus*, with a mean distance of 1.80 m (+/- 0.4 m). Distances from the base of the tree to approximately the center of the intercanopy were also similar, with a mean distance of 4.45 m (+/- 0.65 m). Within the subcanopy, we repeated measurements of soil hydrologic properties following soil saturation and drainage for three of the seven *Juniperus* and *Pinus* line transects. The soil was wetted by inserting 10 cm diameter, 15 cm deep sheet metal cylinders through the litter and slightly into the underlying mineral soil. The cylinders were filled with water and allowed to drain. This procedure was repeated until no water repellency was detected. This study design gave us a total of 274 sampling points, with 234 line transect measurement points within natural antecedent soil moisture conditions and 40 on subcanopy soil under pre-saturated conditions.

**Measurements**

At each measurement point along the radial line transects, soil sorptivity (*S*), unsaturated hydraulic conductivity (*K*(h)), and volumetric soil water content (*SWC*) were measured and BSC cover was assessed. Sorptivity and *K*(h) were measured *in situ* with 15.9 mm Mini-disk Infiltrometers (Decagon Devices Pullman, WA) at a head of -2.0 cm suction. Automation methods, measurement procedures, and calculations were performed according to Madsen and Chandler (2007). All measurements were corrected to standard temperature (20 C), by the viscosity ratio approach of Constantz (1982). Sorptivity was calculated from the first 30 ml drained from the infiltrometer, excluding
the initial 4 ml of water, which was disregarded to minimize the influence of contact sand on the result. Unsaturated hydraulic conductivity was calculated from infiltration at approximately steady state. Rosetta Lite (Schaap et al., 2001) was also used to predict K(h) from soil texture. Soil water content was measured from the top 6 cm of soil with an ML2x Theta Probe (Delta-T Devices, Cambridge, England). Voltage from the Theta probe was converted to apparent permittivity and consecutively to SWC using the relationship given in Blonquist et al. (2005). Data were discarded from 10 S, one K(h) and one SWC measurements due to measurement errors. At each transect point categories of BSC development were recorded as: 1) light crust (soil light in appearance, with a low biomass of cyanobacteria and low pinnacles); 2) dark crust (typically soil covered with cyanobacteria and lichens such as Placidium spp., Fulgensia spp., Psora decipiens, Aspicillia spp., and Toninia spp and Psora cerebraforum); and 3) moss crust (primarily Syntrichia caninervis).

Within the subcanopy, the depth of the litter was measured and then it was removed down to the mineral soil surface. Soil surface hydrophobicity was measured using the water drop penetration time test (WDPT) (Krammes and DeBano, 1965). With this same method, we also measured the depth of water repellency in a pit excavated midway between each Juniperus or Pinus trunk and canopy edge. We measured water repellency at 1.0 cm depth increments until WDPT was less then five seconds.

A second survey of BSC composition was undertaken to quantify our observation of a directional difference in the development of BSC around trees. Observations of BSC were made within 45 degree sections that covered a radial distance of 0.6 m, beginning at
the canopy edge and extending 1.8 m into the intercanopy. Replicate surveys were performed on twelve individual *Juniperus* and *Pinus* trees.

**Data analysis**

Because the total transect lengths were variable, measurements taken at 30 cm intervals are presented at their normalized distance with respect to their canopy radii, by dividing the radii of the respective tree with distance from the trees trunk. Using these normalized canopy radius (CR) distances, data were grouped within quartiles (CR intervals of 0.25) of the respective canopy radii, to a distance of three times the canopy radii. For example, CR 0.50 is the point midway between the tree trunk and the canopy edge and quartile (Q) 0.50 is the spatial unit representing data points from CR 0.25 to 0.50. Canopy radius distances and Q greater than 1.00 were assigned to the intercanopy region. With both *Juniperus* and *Pinus* line transects combined, there are an average of 22 measurements per Q for dry antecedent soil moisture conditions and 9 measurements on hydrophobic soil, which was made wettable by saturating the tell hydrophobicity was overcome.

Secondly, we analyzed litter depth, BSC composition, SWC, WDPT, K(h) and S within patch and unit HFU scales (Wilcox et al., 2003). At the patch scale, we separated the landscape into subcanopy and intercanopy regions at the canopy edge CR 1.0), and by CR 0-0.75, 0.75-1.75 and 1.75-3.0. At the unit scale, the subcanopy patch was grouped and analyzed by tree species (*Pinus* or *Juniperus*) and the intercanopy patch measurements were grouped and analyzed by BSC development categories, using the grouping light, dark and moss BSC.
Statistical analysis was performed with Sigma Stat 3.1 (Systat Software, Inc. Richmond, CA). For all comparisons a significance level of $P < 0.05$ was used. For each measurement type, cumulative data sets were assembled and tested for normality by Kolmogorov-Smirnov test. Litter depth and SWC were found to be non-normal and non-transformable through standard transformation functions. Therefore, we compared these data using nonparametric methods and descriptive statistics are presented graphically through the median and 25th and 75th percentiles. For these parameters, differences between median Q values with distance were analyzed through the Kruskal-Wallis ANOVA on ranks test, in association with the Dunn’s test to determine significant differences between the variables. Both $K(h)$ and $S$ were not normally distributed and were normalized by square root transformation prior to Q level comparison. Quartile mean values of $K(h)$ and $S$ have been displayed graphically with standard error bars, with significance determined by a one way ANOVA test and multiple comparisons among the different combinations of $Q$’s determined through the Holm-Sidak method. For comparison of the multiple intercanopy BSC types, a one way ANOVA test was also used in association with the Holm-Sidak method. Correlation among populations of $K(h)$, $S$, and SWC were analyzed within larger CR segments at the patch and unit level by the Mann-Whitney Rank Sum Test. Inter-comparison of SWC, litter depth, $K(h)$ and $S$ were performed by CR segments by Spearman Rank Correlation test, with results presented as correlation coefficients $r$ and significant $P$ values.
RESULTS

Soil surface cover

The soil surface was covered primarily with BSC in the intercanopy and leaf litter under the plant canopy. In the intercanopy region, we found that whereas dark BSC occurred at approximately 60% of the transect measurement locations beyond the canopy, moss occurrence declined from 20 to 10% between CR 1.5 and 2.5, with a complementary increase in light BSC cover (Fig. 4.1). A subsequent detailed survey of light and dark BSC from a separate set of 7 trees clarified the radial dependence of cyanobacterial BSC development in this environment (Fig. 4.2). Light BSC occurred most frequently on the southern side of the trees (from 100 to 200 degrees from N) and dark BSC occurred most frequently on the northern side of the trees (0 ± 50 degrees from N).

![Figure 4.1. Percent cover of light, dark and moss biological soil crusts (BSC), within normalized distance quartiles, starting from the canopy edge.](image-url)
Figure 4.2. Spatial distribution of light and dark biological soil crust (BSC) by radial distance and direction away from tree canopies. Percent cover for each category is represented graphically by the size of the dark circles, with larger circles representing a greater frequency of occurrence.
Median litter depths were similar for Juniperus and Pinus along the line transects. Leaf litter depth decreased radially from near the base of the tree (3-5 cm) to only a trace, within Q 1.00 (Fig. 3A).

**Soil water content**

Upon arrival at the study site, we observed snow cover of approximately 13 cm depth. It appeared that the canopy had generally intercepted the majority of the snowfall and while the snow was distributed relatively uniformly in the intercanopy, within the subcanopy from the trees trunk to approximately CR 0.75, the presence of snow was less prominent and more variable due to canopy interception. After snowmelt, we did not find significant differences between radial patterns or directionality in SWC among the Pinus and Juniperus transects, and thus combined these data for subsequent analysis (Fig. 4.3B). At the patch HFU scale, the median intercanopy SWC (0.19 m$^3$m$^{-3}$) was significantly higher ($P < 0.001$) than that of the subcanopy (0.08 m$^3$m$^{-3}$). However, comparison of SWC by Q’s did not show a distinct boundary between HFU patches at the canopy edge (Fig. 4.3B). In the subcanopy area, Q 0.25 and 0.50 showed the lowest median SWC values (0.05 m$^3$m$^{-3}$), which are approximately one quarter of the intercanopy median. A clear gradient in SWC emerges between Q 0.50, and Q 1.25, beyond which SWC remained similar, with median SWC ranging between 0.17-0.19 m$^3$m$^{-3}$.

In areas where the snow had been blown under the canopy and melted, the litter and duff material on the soil surface were completely saturated following snowmelt.
Figure 4.3. A. Median litter depths and their associated 25th and 75th percentiles from canopy to intercanopy, with data normalized with their respective canopy radii distances and combined within quartiles at distance intervals of 25% of the canopy radii. The soil surface is represented at 0 on the y axis. The gray box below the soil surface represents hydrophobicity mean depth and standard errors, measured at the center of the vegetation canopies, and hydrophobicity distance, measured at the soil surface from the base of the vegetation outwards. B. Soil water content (SWC) out to 2.5 times the canopies radius. Data is presented at the patch scale with median (horizontal solid line) and associated 25th and 75th percentiles (horizontal dotted lines) for both subcanopy and intercanopy hydrologic scales. Within these patches, median and associated 25th and 75th percentiles are also shown for data presented at the quartile scale. Statistically distinct values are denoted by different letters.
However, the underlying soil was air dry except for a few small wet patch conduits typically associated with fine roots growing close to the soil surface. From our observations it appeared that the litter and duff material were retaining water from running off of the copious subcanopy mounds, while allowing drainage to be concentrated through the soil profile within a few select locations.

**Hydrophobicity**

Water drop penetration time tests of the soil at the interface of the mineral soil and organic material showed hydrophobicity extended from near the trees trunk to CR 0.75, on average, for both *Juniperus* and for *Pinus* (Fig. 4.3A). Within this region, hydrophobicity at the interface between soil and litter material was so severe (WDPT> 15 min) that the degree of hydrophobicity could only be logistically determined at each sampling point. Within Q 1.00, the incidence of hydrophobicity was more variable, and in the intercanopy, the soil was not hydrophobic (Fig. 4.3A). Soil pit excavations at CR 0.5 showed hydrophobic soils extended to as deep as 17.5 cm and 16 cm for *Juniperus* and *Pinus* respectively, with an average depth of 12.9±1.41 cm for *Juniperus* and 10.0±1.70 cm for *Pinus*.

**Infiltration**

Spatial gradients in both K(h) and S blurred the lines between the subcanopy and intercanopy, with Q’s similar for both *Juniperus* and *Pinus*. Figure 4.4 shows the means and associated standard errors for K(h) and S measurements presented at the Q and HFU scale with data from both *Pinus* and *Juniperus* combined. Quartile average values for K(h) and S from Q 0.25 to 0.50 are less than the subcanopy patch average
Figure 4.4. A. Unsaturated hydraulic conductivity (K(h)) and B. sorptivity (S) along radial transects extending from the base of the tree to the center of the intercanopy. Data are presented at the patch scale with average (horizontal solid line) and associated standard errors (horizontal dotted lines) for both subcanopy and intercanopy hydrologic scales. Mean and associated standard errors is also shown for data presented at the quartile scale by normalizing the data with their respective canopy radii and combining point measurements within quartiles, at distance intervals of 25% of the canopy radii. Statistically distinct values are denoted by different letters.
A positive gradient in $K(h)$ originated near the tree trunk, but is most clear from Q 0.75 (1.91±0.46 cm hr$^{-1}$) to Q 2.0 (16.18 ± 2.01 cm hr$^{-1}$) (Fig. 4.4). Average $K(h)$ values for Q 1.00 (5.75 ± 0.64 cm hr$^{-1}$) and Q 1.25 (7.34 ± 0.90 cm hr$^{-1}$) are between the subcanopy and intercanopy (13.29±0.79 cm hr$^{-1}$) averages. Quartile 2.0 and 2.5 average $K(h)$ values (16.18±2.01 and 16.23±1.59 cm hr$^{-1}$, respectively) were significantly greater than the intercanopy average. For $S$ a shorter gradient emerged between the lowest value at Q 0.50 (0.063 mm s$^{-1/2}$) and Q 1.25, beyond which, Q’s remained similar to the intercanopy average value (0.37±0.02 mm s$^{-1/2}$) and ranged between 0.26 and 0.44 mm s$^{-1/2}$. The subcanopy gradients in $S$ and $K(h)$ did not show a significant directional bias. However the greatest values of $K(h)$ tended to occur on measurement transects between 225 and 45 degrees from north and at CR greater than 1.75.

We tested the radial dependence of $K(h)$ and $S$ on BSC cover type within the intercanopy (Fig 5). For Q 1.25, mean $K(h)$ values for light BSC, dark BSC, and moss BSC types were all significantly lower than the intercanopy mean value (13.3 cm hr$^{-1}$). Between CR 1.25 and 2.0, $K(h)$ increased by two-to three-fold for dark BSC (7.89±1.45 to 15.14±2.24 cm hr$^{-1}$) and light BSC (4.98±0.51 to 16.72±3.15 cm hr$^{-1}$). Beyond CR 2.0, moss BSCs were infrequent (n=2 for each quartile) and thus were not included in the analysis. However, the difference in $K(h)$ between light and dark BSC for Q 2.25 (11.39±1.67 and 19.74±3.34 cm hr$^{-1}$) and Q 2.50 (11.17±2.41 and 19.62±1.95 cm hr$^{-1}$) indicates that BSC cover type influences $K(h)$ far from the trees. Furthermore, at this distance, $K(h)$ for dark BSC was greater than both the intercanopy mean and modeled
Figure 4.5. Quartiles within the intercanopy divided into light, dark, and moss BSC for unsaturated hydraulic conductivity (K(h)), and sorptivity (S). Significant differences in K(h) were found between light and dark BSC at CR 2.50 as denoted by lower case letter “a”. Horizontal lines running the length of the figures represent intercanopy mean (solid line) and standard errors (dashed lines).

Table 4.1. Unsaturated hydraulic conductivity (K(h)) and sorptivity (S) measurements performed within the subcanopy with the soil hydrophobic and wettable (through saturating the soil tell hydrophobicity was overcome). Descriptive statistics show mean and standard error, for each hydrophobic and hydrophilic comparison.

<table>
<thead>
<tr>
<th>Normalized Distance</th>
<th>wettable</th>
<th>hydrophobic</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>K(h) (cm hr⁻¹)</td>
<td>S (mm s⁻¹/²) x 10⁻¹</td>
</tr>
<tr>
<td>0.25</td>
<td>mean</td>
<td>std.error</td>
</tr>
<tr>
<td></td>
<td>16.66</td>
<td>2.18</td>
</tr>
<tr>
<td>0.50</td>
<td>10.55</td>
<td>1.12</td>
</tr>
<tr>
<td>0.75</td>
<td>11.40</td>
<td>2.11</td>
</tr>
<tr>
<td>1.00</td>
<td>10.96</td>
<td>3.17</td>
</tr>
</tbody>
</table>
value (15 cm hr\(^{-1}\)). Within the intercanopy S did not vary systematically with either BSC composition or distance from the canopy.

When hydrophobicity of the subcanopy soil was overcome by prewetting, K(h) and S increased by several fold relative to the initial measurements (Table 4.1). Within Q 0.25, K(h) increased from 0.45±0.11 cm hr\(^{-1}\) to 16.66±2.18 cm hr\(^{-1}\), and S increased from 0.08±0.02 mm s\(^{-1/2}\) to 0.47±0.09 mm s\(^{-1/2}\) following artificial wetting and drainage for points previously measured on hydrophobic soil. For the remainder of the subcanopy, K(h) values for artificially prewetted subcanopy soils were not significantly different from snowmelt-wetted intercanopy soils beyond CR 2.0 and 1.25, respectively. The results of pre-wetting from near the canopy edge were less dramatic due to greater K(h) under natural condition. For example, pre-wetted mean K(h) for Q 1.0 (6.63 cm hr\(^{-1}\)) was at least 3 times greater than Q 0.25, 0.50 and 0.75, respectively, but reached a similar rate following artificial wetting.

**Hydrologic functional distributions**

We separated the data into two patch level HFUs, subcanopy and intercanopy, at CR 1.0, and then into three HFU patches: the hydrophobic subcanopy (CR 0-0.75), the transitional gradient in K(h) (CR 0.75-1.75) and the unaffected intercanopy past CR 1.75 (Table 4.2). Within each of these schemes, we tested for differences at the plant cover or unit level. For the global data sets, SWC had a bimodal normal distribution, whereas K(h) and S were better represented by a Poisson distribution. When these data were divided by HFU, S and SWC were normally distributed beyond CR 0.75, and all data were normally distributed beyond CR 1.75. Subcanopy data were normally distributed only for prewetted conditions (Table 4.2). To simplify comparison of normal and highly
skewed populations, statistical tests of these data are presented for median values, and we also present mean and standard error as descriptive statistics (Table 4.2).

Within the subcanopy, median values of $K(h)$, $S$ and SWC were not significantly different ($P= 0.143$, 0.516, 0.168, respectively) between CR 0-1.0 and CR 0-0.75.

Similarly, unit level analysis comparing the differences between *Juniperus* and *Pinus* subcanopies show they are statistically similar for SWC, $K(h)$ and $S$, regardless of whether the subcanopy patch was defined as the CR1.0 ($P$ values = 0.214, 0.542, and 0.536 for SWC, $K(h)$ and $S$, respectively) or CR 0.75 ($P$ values = 0.214, 0.542, and 0.536 for SWC, $K(h)$ and $S$, respectively). However, average $K(h)$ values are

Table 4.2. Descriptive statistics for unsaturated hydraulic conductivity ($K(h)$), sorptivity ($S$) and soil water content measurements. Sample populations (n) were selected by subdividing the global data set (all) by normalized distance from the base of trees and by cover type (unit) or treatment (e.g. prewetted) within that distance. Median, mean, and standard error (SE) are presented and normally distributed populations are indicated (*).

<table>
<thead>
<tr>
<th>normalized distance</th>
<th>patch/ unit</th>
<th>$K(h)$ (cm hr$^{-1}$)</th>
<th>$S$ (mm s$^{-1/2}$) x 10$^{-1}$</th>
<th>SWC (m$^3$ m$^{-3}$) x 10$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n median</td>
<td>mean  S.E.</td>
<td>n median</td>
<td>mean  S.E.</td>
</tr>
<tr>
<td>0-3.0</td>
<td>global</td>
<td>233 6.11 8.63 0.56</td>
<td>224 1.94 2.56 0.15</td>
<td>233 1.61 1.54 0.05</td>
</tr>
<tr>
<td></td>
<td>subcanopy</td>
<td>83 0.64 1.73 0.29</td>
<td>82 0.679 1.02 0.12</td>
<td>83 0.63 0.86 0.06</td>
</tr>
<tr>
<td></td>
<td>all pre-wet</td>
<td>40 17.12 18.44 1.68</td>
<td>40 4.67 5.64 0.63</td>
<td>40 0.94 0.94 0.09</td>
</tr>
<tr>
<td></td>
<td>pinyon</td>
<td>42 0.69 1.66 0.34</td>
<td>41 0.647 0.87 0.14</td>
<td>42 0.94 0.94 0.09</td>
</tr>
<tr>
<td></td>
<td>juniper</td>
<td>41 0.60 1.79 0.49</td>
<td>41 0.714 1.17 0.19</td>
<td>41 0.57 0.78 0.09</td>
</tr>
<tr>
<td>0-0.75</td>
<td>subcanopy</td>
<td>70 0.36 1.00 0.18</td>
<td>69 0.493 0.88 0.12</td>
<td>70 0.55 0.71 0.05</td>
</tr>
<tr>
<td></td>
<td>pinyon</td>
<td>35 0.39 1.00 0.23</td>
<td>34 0.531 0.80 0.14</td>
<td>35 0.63 0.79 0.08</td>
</tr>
<tr>
<td></td>
<td>juniper</td>
<td>36 0.29 1.14 0.30</td>
<td>36 0.475 0.93 0.18</td>
<td>36 0.55 0.68 0.08</td>
</tr>
<tr>
<td>1.0-3.0</td>
<td>intercanopy</td>
<td>150 9.35 12.51 0.68</td>
<td>142 3.03 3.44 0.20</td>
<td>150 1.86 1.91 0.04</td>
</tr>
<tr>
<td></td>
<td>transition</td>
<td>84 7.56 8.94 0.69</td>
<td>82 2.67 3.11 0.24</td>
<td>84 1.76 1.82 0.05</td>
</tr>
<tr>
<td>1.75-3.0</td>
<td>intercanopy</td>
<td>79 13.34 15.18 1.00</td>
<td>73 3.38 3.52 0.29</td>
<td>79 1.93 1.98 0.06</td>
</tr>
<tr>
<td></td>
<td>light BSC</td>
<td>24 11.45 13.80 1.49</td>
<td>23 3.38 3.29 0.43</td>
<td>24 1.94 1.92 0.12</td>
</tr>
<tr>
<td></td>
<td>dark BSC</td>
<td>43 17.59 16.84 1.30</td>
<td>41 3.76 3.60 0.42</td>
<td>43 1.83 1.95 0.08</td>
</tr>
<tr>
<td></td>
<td>moss</td>
<td>12 7.80 12.00 3.51</td>
<td>9 2.68 3.76 0.96</td>
<td>12 2.44 2.19 0.14</td>
</tr>
</tbody>
</table>
significantly different between CR 0-1.0 and CR 0-0.75 (1.73±0.29 and 1.00±0.18 cm hr$^{-1}$, respectively).

After pre-wetting to remove hydrophobicity, median subcanopy K(h) changed significantly from prewetted conditions ($P<0.001$) and was significantly different from all HFU divisions except CR 1.75-3.0 ($P=0.13$). The populations of K(h), S and SWC in the intercanopy and transition zones are all significantly different from the subcanopy ($P<0.001$), but significant differences between the transition zone and intercanopy occur only for K(h). Further, K(h) is significantly different among CR 0.75-1.75 (8.9 ± 0.7 cm hr$^{-1}$), CR 1.0-3.0 (12.5 ± 0.72 cm hr$^{-1}$), CR 1.75-3.0 (15.22 ± 1.0 cm hr$^{-1}$). Spearman rank order correlations among K(h), S and SWC showed significant trends for CR 0-1.0 and CR 0-0.75 ($P<0.004$), but no significant trends for CR 1.0-3.0, CR 0.75-1.75, and CR 1.75-3.0 (Table 4.3).

These patterns in K(h), S and SWC correspond with litter depth within the area of the subcanopy, where the soil is predominantly hydrophobic. Thus, it follows that Spearman correlation tests are negative between litter thickness in the subcanopy and K(h), S and SWC (Table 4.4). However, we found S was not as strongly correlated to litter thickness as were K(h) and SWC at either the Q or single measurement level. This finding supports our field observation that variability in wetness of the soil O-horizon affected short time infiltration, S, independently of long time infiltration, K(h), which was representative of the underlying mineral soil.
Table 4.3. Spearman rank order correlation coefficient ($r$), $P$ value, and number of samples ($n$) for comparisons among unsaturated hydraulic conductivity ($K(h)$), sorptivity ($S$) and soil water content (SWC) populations within patch level hydrological functional units as defined by different distances of canopy radius (CR) from the tree trunk. There are no significant relationships for $P > 0.050$.

<table>
<thead>
<tr>
<th></th>
<th>CR 0-1.0</th>
<th>CR 0-0.75</th>
<th>CR 1.0-3.0</th>
<th>CR 0.75-1.75</th>
<th>CR 1.75-3.0</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$S$</td>
<td>SWC</td>
<td>$S$</td>
<td>SWC</td>
<td>$S$</td>
</tr>
<tr>
<td>$K(h)$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$r$</td>
<td>0.40</td>
<td>0.56</td>
<td>0.38</td>
<td>0.38</td>
<td>0.00</td>
</tr>
<tr>
<td>$P$</td>
<td>$&lt;0.001$</td>
<td>$&lt;0.001$</td>
<td>$0.001$</td>
<td>$0.001$</td>
<td>0.993</td>
</tr>
<tr>
<td>$n$</td>
<td>82</td>
<td>83</td>
<td>69</td>
<td>70</td>
<td>142</td>
</tr>
<tr>
<td>$S$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$r$</td>
<td>0.31</td>
<td>0.26</td>
<td>-0.01</td>
<td>0.06</td>
<td>-0.06</td>
</tr>
<tr>
<td>$P$</td>
<td>$0.004$</td>
<td>$0.029$</td>
<td>$0.901$</td>
<td>$0.590$</td>
<td>$0.590$</td>
</tr>
<tr>
<td>$n$</td>
<td>82</td>
<td>69</td>
<td>142</td>
<td>82</td>
<td>84</td>
</tr>
</tbody>
</table>

Table 4.4. Spearman rank order correlation coefficient ($r$), $P$ value, and number of samples ($n$) for comparisons with litter depth to unsaturated hydraulic conductivity ($K(h)$), sorptivity ($S$) and soil water content (SWC). Significance was determined at $P < 0.050$.

<table>
<thead>
<tr>
<th>litter depth</th>
<th>$K(h)$</th>
<th>$S$</th>
<th>SWC</th>
</tr>
</thead>
<tbody>
<tr>
<td>$r$</td>
<td>-0.64</td>
<td>-0.26</td>
<td>-0.57</td>
</tr>
<tr>
<td>$P$</td>
<td>$&lt;0.001$</td>
<td>$0.001$</td>
<td>$&lt;0.001$</td>
</tr>
<tr>
<td>$n$</td>
<td>83</td>
<td>82</td>
<td>83</td>
</tr>
</tbody>
</table>

DISCUSSION

The most apparent influences of large woody vegetation on hydrology have previously been associated with the interception of precipitation and solar radiation (Breshears, 2006), enhanced infiltration under the canopy, and root uptake of soil water (Seyfried, 1991; Pierson et al., 1994; Wilcox et al., 2003). The results of this study confirm the importance of interception by *Pinus* and *Juniperus* trees. However, the
finding that $K(h)$ was universally less in the subcanopy than the intercanopy would appear to contradict previous infiltration studies. We contend that the difference in our findings is due to methodological differences which can be used to clarify the processes by which woody plants modify their environment.

In semiarid environments, two important features of the subcanopy environment of woody plants have not been well documented: extensive water repellency of soils and moisture storage within the litter layer. When measured under ambient antecedent moisture conditions, severe hydrophobicity in $Q_{0.25, 0.50}$ and $0.75$ (Fig. 4.3) inhibited water entry into the soil fabric, with a concomitant reduction in $K(h)$ relative the intercanopy (Fig. 4.4). Thus, assuming the presence of hydrophobicity, our results differed from those of Seyfried (1991) and Wilcox et al. (2003), who used measurement techniques that either removed or were not sensitive to soil water repellency. When hydrophobicity was removed in our study by prewetting to saturation, subcanopy $K(h)$ was similar to the intercanopy $K(h)$, as found by Seyfried (1991) and Wilcox et al. (2003). We attribute the difference in prewetted $K(h)$ between $Q_{0.25, 16.66 \text{ cm hr}^{-1}}$ and the other prewetted subcanopy quartile averages ($10.55$ to $11.44 \text{ cm hr}^{-1}$) to modification of the soil structure by the rooting system of the plant or faunal activity under the canopy (Table 4.1).

Our observations that litter mounds effectively retain moisture at near saturation supports the previous findings that there is lower runoff from the subcanopy than there is from the intercanopy (Pierson et al., 1994), as runoff can be reduced by both storage at the soil surface and infiltration into the mineral soil. The additional observation of litter routing water preferentially to wetted spots in the underlying soil, and the frequent
association of these wet spots with fine roots, further clarifies the apparent contradiction that despite extensive hydrophobicity, runoff is seldom observed from under *Pinus* and *Juniperus* canopies during precipitation inputs. We expect that low intensity precipitation inputs such as snow melt or long duration, low intensity rainfall events, typical of winter frontal storms, may allow for the development of wet patches or complete removal of hydrophobicity. This would subsequently allow infiltration to occur at preferential locations under the canopy. However, high intensity monsoonal storms, typical of summer events, may deliver a sufficient depth of precipitation to exceed the storage capacity of the canopy and litter material. Further precipitation at a rate greater than subcanopy soil S under hydrophobic conditions would generate runoff as observed by Reid et al. (1999) from runoff catchment plots.

In the intercanopy, we previously found that canopy interception of rain causes a transitional soil moisture zone east of *Juniperus* trees, due to prevailing winds from the west (Lebron et al., 2007). However, the radial measurements of K(h) were only made in one direction and the trends were complicated by the location of a heavy machinery track through the study site. In this study, soil moisture did not vary significantly with distance from the canopy beyond CR 1.25, and although this pattern was reflected in the patterns of WDPT and S, a clear gradient in K(h) emerged from within the canopy to approximately twice the CR in all directions. Beyond CR 0.75, K(h) was not correlated with SWC or S. We conclude that the continual increase in K(h) in all directions into the intercanopy indicates that the influence of woody vegetation on soil physical and chemical properties extends well beyond the canopy edge.
We hypothesize that subcritical hydrophobicity from litter dispersal, runoff, and mycorrhizal root associations contributes to the observed gradient in $K(h)$ in the intercanopy. Soil hydrophobicity develops as organic compounds on the surface of soil particles dry out, leaving a water repellent coating surrounding the mineral soil (Doerr et al., 2000). Wind-blown litter from the canopy and runoff from the elevated litter mounds into the intercanopy may result in lateral transport of hydrophobic compounds and subcanopy litter into the intercanopy area. On the other hand, organic compounds in fine roots and associated mycorrhizae have previously been shown to result in hydrophobicity in dry soils (Hallett et al., 2003). Our results agree with those of Hallett et al. (2003), who hypothesized that low levels of water repellency would be responsible for high spatial variability in water transport, particularly under conditions of tension infiltration when macropores are inactive. While the extent of hydrophobic compounds within the transitional zone may not be great enough to limit soil wettability, the occurrence of organic coatings in the soil fabric may be sufficient to reduce infiltration in unsaturated conditions. We reason that the influence of hydrophobicity on $K(h)$ would decrease with distance from the canopy, regardless of whether the organic compounds originated from the canopy, roots, or associated mycorrhizae. Although spatially distributed data on root distributions for these species are difficult to obtain, as they can vary widely from site to site, we consider root related water repellency and soil structural effects the likely causes of the observed gradient in $K(h)$ where hydrophobicity was not detected, due to the radial symmetry of the observations (CR 0.75-2.0 in Fig. 4.6).

In addition to the influence of litter and roots on soil physical properties, $K(h)$ may also be indirectly affected by the influence of the tree canopy on soil microclimate.
Pinyon-juniper woodlands often have wide interspaces among the trees, where the intercanopy soils experience more variable temperatures than subcanopy soils (Breshears et al., 1998). Although the shelter effect of the canopy may vary due to factors such as leaf area, density, and height and spacing of the woody plants (Martens et al., 2000; Breshears, 2006), the shelter provided by the canopy likely decreases from the tree trunk outwards to the canopy edge. Differences in the number of freeze-thaw cycles at the soil surface can also affect patterns of K(h) by seasonally decreasing soil bulk density (Lampurlanes and Cantero-Martinez, 2003). Since this study was conducted at the end of winter after the soil had experienced several freezing and thawing events, gradients in K(h) could be due to variations in freeze-thaw effects on soil bulk density, especially within the regions beyond the reach of litter fall and roots (Fig. 4.6). The difference in K(h) beyond CR 1.75 may also be attributable to morphological variations among BSC types, which is also associated with freeze-thaw cycles (Belnap, 2003).

We found the light (least well developed) BSC mostly on the south of the trees, indicating that BSC type is more strongly dependent on canopy interception of radiation than precipitation at this site (Fig. 4.2). Similar to other infiltration studies (e.g. Wilcox et al., 2003; Eldridge et al., 1997; Williams et al., 1999), K(h) and S were not significantly different among BSC types at the intercanopy HFU scale. However, near the canopy edge K(h) was significantly lower than the intercanopy average for all BSC types and an increasing trend in K(h) was observed between CR 1.0-2.0, with BSC types similar within the same Q (Fig. 4.5). From Q 2.0-2.5, K(h) increased for dark BSC and was almost double that of light BSC at Q 2.5 (light BSC = 11.1 cm hr\(^{-1}\), dark BSC = 19.6 cm hr\(^{-1}\)). Coupling this result with the findings that well developed BSC occurs more
frequently on the north side of the trees, may indicate that the woody plant canopy has a positive influence on $K(h)$ for CR 2.0-2.5 through its influence on BSC development (Fig. 4.6). Although the canopy also exerts a control on BSC development from CR 1.0-2.0, we found no evidence that BSC composition influenced $K(h)$ in this transitional zone between subcanopy and intercanopy HFUs.

The observed transformations may confer several advantages in water and nutrient capture, and retention to woody plants in resource limited environments. Biota growing in sandy soils within dryland regions such as our study site can be limited by high evaporation and low water holding capacity of the soil. It is generally believed that one of the reasons woody vegetation can persist in these systems, is through their ability

![Figure 4.6. Schematic diagram a of woody plant canopy and hypothetical root structure, litter mound and spatial distribution of dark and light biological soil crust (BSC).](image-url)
to extract deep soil moisture resources via tap roots. Woody plant roots have also been shown to transfer soil moisture deeper and broader within the soil profile through hydraulic redistribution. This can result in water conservation by minimizing evaporation losses at the soil surface and reducing the rate of uptake (Ryel et al., 2003). Our results indicate that within coarse textured soils, patchy hydrophobicity may be another means by which woody plants modify the environment, resulting in greater soil moisture availability. This patchiness may channel soil moisture through a few wet root-filled patches that have a higher infiltration capacity than the surrounding soil. The predominantly dry hydrophobic soil, with low S and an overlying mulch layer, would also restrict upward movement towards the soil surface, thus minimizing evaporative losses from under the canopy (Scott, 1992; Moore and Blackwell 1998; Jaramillo et al., 2000). The increased frequency of wet litter on the mounds under the trees, due to an underlying hydrophobic soil, may also accelerate decomposition of the litter, increasing nutrient and carbon flow to the underlying tree roots. Whereas soil hydrophobicity may be a water conservation mechanism for established plant species, it may also impede germination and establishment of other species in the dry hydrophobic layer at the soil surface (Osborn et al., 1967). Although the mechanism by which Juniperus and Pinus reduce K(h) beyond the canopy is not clear, the effect of increased canopy cover would be a progressive decrease in the average infiltration capacity of the intercanopy. This process may also be advantageous to the woody plants if it results in more focused recharge to shallow groundwater stores at the expense of more evenly distributed soil moisture.
CONCLUSIONS

Results from this study demonstrate the influence of woody vegetation in modifying soil hydrologic properties through the development of a hydrophobic layer within the subcanopy. For an established woody plant, hydrophobic soil may be ecological advantages by channeling precipitation inputs deeper into the soil through a few select wet patches, thus minimizing evaporative losses at the soil surface.

Hydrologic studies within pinyon-juniper woodlands typically analyze the landscape either by partitioning it into canopy and intercanopy patches or through random placement of points disregarding patch structure (e.g. Blackburn and Skau, 1973; Reid et al., 1999; Wilcox et al., 2003; Pierson et al., 2007). Our results demonstrate that infiltration is not evenly distributed for a soil/cover type, which would support measuring landscapes by HFU’s. However, this study further shows that whereas differences in soil state variables such as hydrophobicity may be discrete between the subcanopy and intercanopy, differences in K(h) are not discrete, but depend on the distance from the canopy edge. We have hypothesized that changes in infiltration properties with distance from the canopy are the result of subcritical water repellency, and or diminishing influences of the canopy on the soils microclimate, which limited the degree of freeze-thaw to the soil surface. Both of these hypotheses may have a significant influence on ecohydrologic properties, from vegetation establishment to runoff and erosion. For a more accurate assessment of infiltration properties, distance from the base of the tree and canopy edge needs to be considered.
REFERENCES


CHAPTER 5
SUMMARY

This research improved methodology for measurements of soil surface hydraulic properties in wildland systems. Hydrologic measurement methods included 1) design and use of automated Mini Disk Infiltrometers, and 2) development of a linear regression approach for calculating hydraulic conductivity from saturated soil cores. Application of this new technology was then applied within a pinyon-juniper woodland to develop a detailed model of how Utah juniper (Juniperus osteoperma) and pinyon pine (Pinus edulis) modify soil state variables spatially within the woodland.

Automation of Mini Disk Infiltrometers is a relatively simple and inexpensive technique which allows for detailed “in-situ” characterization of soil surface hydraulic properties, in remote and difficult experimental settings. The proposed method gives the user the ability to capture quick changes in initial infiltration rates and eliminate error associated with manually recording volume changes in the infiltrometers reservoir with time. Additionally, automated Mini Disk Infiltrometers also decreases the spatial and temporal variability associated with a measurement campaign, by increasing the possibility of obtaining a large sample of precise infiltration records, during a relatively short field campaign.

Previous automated infiltrometers have based infiltration calculations on the linear relationship between transducer voltage output and pressure head difference over the height of the column of water in the infiltrometers. However, neither vertical placement of the infiltrometers, nor a constant input voltage is assured in field application. By scaling the total volume of water used in the infiltrometers to the
difference between the maximum output voltage at the beginning of the trial and the
minimum output voltage at the end of the trial, we can implicitly correct for errors due to
non-vertical infiltrometer placement and transducer signal drift.

Comparison between the two different types of Mini Disk Infiltrometers showed
the original instrument (AMDi\textit{v1}) produced constant infiltration results at moderate
angles of inclination using our simple calibration approach. Use of the current style Mini
Disk Infiltrometer (AMDi\textit{v2}) showed infiltration measurements on an incline would
require additional effort to maintain a constant tension in the supply reservoir and
operation at a consistent axial orientation to the slope. Additionally, the AMDi\textit{v1} returned
more consistent values of unsaturated hydraulic conductivity than the AMDi\textit{v2}, perhaps
due to the smaller volume of water required to reach steady state by the AMDi\textit{v1} in the
tested soils. Greater accuracy in sorptivity and unsaturated hydraulic conductivity can be
achieved by taking sequential measurements on initially dry soil to account for the
control exerted by the initial soil moisture content on these measurements.

Standard calculation methods of saturated hydraulic conductivity ($K_s$)
measurements performed at one head gradient are subject to implicit inclusion of
systematic bias. However, the nature of the data dictate that one cannot assume the trend
line between flux density and hydraulic gradient will pass through the origin. For our
laboratory tests of field samples, we found non linear relationships between flux density
and hydraulic gradient for tests conducted on both sandy loam and clay loam soils.

Results from this study demonstrated that linear regression of hydraulic head
gradient against flux density is a robust technique to determine $K_s$, independently of
systematic error. This is not an important consideration for experiments on the scale of
Darcy’s original work, but the error component becomes significant at the scale of contemporary soil cores. We propose that separating error or bias in due to changes in head, into the regression constant $b$ allows highly precise measurement of $K_s$ from small cores. The degree of linearity indicated by the regression correlation coefficient is also a useful indicator of data quality and allows the investigator to identify samples or head gradients for which the test is invalid.

Hydrologic studies within a pinyon-juniper woodland demonstrate that infiltration is not evenly distributed throughout the woodland, due to the woody vegetation alterations to soil state variables. Within the subcanopy, interactions among hydrophobic soils, litter material and root distributions significantly influenced ecohydrologic properties by both limiting, and potentially redirecting, infiltration below the soil surface. For an established woody plant, hydrophobic soil may be physiologically advantageous to the trees if precipitation inputs are channeled deeper into the soil through a few wet patches. Potential benefits to the tree include minimizing evaporative losses at the soil surface and increase competitive advantage over annual plants.

This study further shows that differences in unsaturated infiltration are not discrete between the subcanopy and intercanopy zones. Within the subcanopy unsaturated infiltration was significantly lower than the intercanopy, but also increased by eight-fold across a gradient extending outward from near the edge of the canopy to approximately two times the canopy radius. We hypothesize that this gradient is related to the trees’ diminishing regulation over water repellence and soil microclimate with distance into the intercanopy.
Analysis of biological soil crust cover within the intercanopy showed that increasing structural development was associated with increased infiltration capacity beyond the transitional zone. Furthermore, these results indicate that distance from the canopy and direction should be considered in the assessment and modeling of woody plant and biological soil crust influence on infiltration capacity.
APPENDICES
APPENDIX A

CALIBRATION OF UNSATURATED HYDRAULIC
CONDUCTIVITY WITH TEMPERATURE
INTRODUCTION

The influence of temperature on soil water infiltration is rarely considered in rangeland infiltration experiments, as a result of high levels of spatial variability across the surface of the soil, and low sample sizes, due to constraints associated with remote, non agricultural settings. However, with increased sample sizes through automation of the Mini Disk Infiltrometer (MDI) (Decagon Devices Pullman, WA) (Chapter I) coupled with the tension infiltrometers ability to remove macropore drainage from the experiment (Ankeny et al. (1988), soil hydraulic properties which typically are considered miniscule, become more evident. Temperature is one such factor which is rarely considered yet has been shown to influence soil hydraulic properties (Jaynes 1990). In rangeland systems diurnal and seasonal temperature fluctuations, coupled with spatial variability associated from vegetation cover may have a significant influence on soil infiltration rates. By adjusting infiltration values measured over the course of a field campaign to a standard temperature (e.g. 20° C), researchers can more accurately describe spatial patterns in infiltration within the study area.

In this study we examine the effects of soil and test water temperature on viscosity and surface tension of infiltrated water. We also show measurements collected within a MDI infiltrometer can be calibrated to 20° C.

Theory

Change in temperature of a substance can be calculated through a calorimetric approach through the specific heat equation
\[ q = vc\Delta T \]  

where:

\( q \) = heat change in system (J)

\( v \) = volume occupied

\( c \) = specific heat of system component (J \text{ m}^{-3} \text{C}^{-1})

\( T \) = temperature change of the substance (C)

Assuming a system composed of soil, soil water, air and infiltrating water; and that the heat capacity of air and the heat flux exchange between the measurement volume and the bulk soil are negligible at short time steps,

\[ v_{sw}c_w(T_f - T_{sw}) + v_sc_s(T_f - T_s) = v_{iw}c_w(T_f - T_{iw}) \]  

where \( v_{sw}, v_s \) and \( v_{iw} \) are the volumes occupied and by the soil water, soil and infiltrometer water, respectively (\text{m}^3); \( c_w \) and \( c_s \) are the volumetric specific heats of water and soil (J \text{ m}^{-3} \text{C}^{-1}); and \( T_f, T_s \) and \( T_{iw} \) are the temperatures of the soil and soil water at the end and beginning of the experiment, and temperature of water in the infiltrometer, respectively. For simplicity, we assume:

\[ T_{sw} = T_s \]  

At the end of the experiment, the volume of influence for the heat balance is approximated by the ratio of the volume of infiltrated water to the available pore space at the applied head,

\[ V = \frac{v_{iw}}{\theta(h) - \theta_i} \]
where $\theta(h)$ and $\theta_i$ are the volumetric water contents at the tension applied by the infiltrometer and prior to the experiment. The volume of ‘old’ soil water and soil at the end of the experiment are

$$v_{sw} = \theta V \quad \text{(5)}$$

$$v_s = (1 - \phi)V \quad \text{(6)}$$

By substitution, of equations 5 and 6 equation 2 becomes

$$T_f = \frac{VT_s (c_w \theta_i + c_s (1 - \phi)) + T_{sw} c_w v_{sw}}{V (c_w \theta_i + c_s (1 - \phi)) + c_w v_{sw}} \quad \text{(7)}$$

The calculation becomes volume independent by further substitution of equation 4.

$$T_f = \frac{T_s (c_w \theta_i + c_s (1 - \phi)) + T_{sw} c_w (\theta(h) - \theta_i)}{c_w \theta_i + c_s (1 - \phi) + c_w (\theta(h) - \theta_i)} \quad \text{(8)}$$

For ponded infiltration it has been shown that once $T_f$ is known $K(h)$ values may be adjusted by a simple viscosity adjustments. (Constantz and Murphy, 1991). We applied this approach with our $K(h)$ measurements by calculating the dynamic viscosity according to Weast and Astle (1982) and then multiplying our $K(h)$ results by its associated dynamic viscosity at that temperature, to correct to a reference of 20° C.

Changes in soil temperature influences infiltration differently on dry soils, requiring sequential calibration for both viscosity and surface tension effects. Surface tension effects can be corrected to a reference ($r$) of 20° C by assuming

$$\frac{K(h_r)}{K(h)} = \frac{\Psi(T_r)}{\Psi(T)} \quad \text{(9)}$$
where the capillary pressure $\Psi$ of the meniscus was determined using the approach of Bachman et al. (2002).

**MATERIALS AND METHODS**

Soil used in the analysis was described as a Rizono-Rock out crop complex (loam, mixed (calcareous), mesic, Lithic Ustic Torriorthents) which were obtained near Dead Horse point near Moab, UT. Laboratory textural analysis through the hydrometer method (Gee and Or 2002) showed the soil to be a loamy sand.

In the laboratory, Soil from the Moab field site was placed through a #10 sieve, mixed thoroughly and hand compacted to a depth of 20 cm in a tiled container, saturated from the bottom up, and allowed to drain to 38% volumetric water content prior to each set of tests. Afterwards 4 stations 20 cm apart with 3 mm of contact sand were place over the soil to insure good contact.

$K(h)$ was measured with 15.9 mm Mini-disk Infiltrometers (Decagon Devices Pullman, WA) at a head of -2.0 cm suction. Automation methods, measurement procedures and calculations of $K(h)$ were performed according to (Madsen and Chandler, 2007).

The effect of temperature on infiltration was measured with the soil near saturation (soil moisture 38 %) and under dry conditions (soil moisture 5 %). For both soil moisture regimes, five replications were performed near 5, 13, 20, 30 and 40° C, with near isothermal conditions between the soil and water in the infiltrometer. Soil temperature was adjusted for the measurements performed near saturation by running hot or cold water through the soil until the desired temperature was reached. Soil
temperature under dry conditions was controlled with high wattage light bulbs, which distance from the soil and timing was adjusted manually, according to readings from the temperature sensors. Heat from light bulbs was also used to dry the soil in-between replications of infiltration measurements at various temperatures. We also performed five infiltration measurements under non isothermal conditions by varying the temperature of the soil and water in the infiltrometer separately, to test the accuracy of equation 8 in predicting the equilibrium temperature of the soil water.

RESULTS AND DISCUSSION

Accuracy of the calorimetric approach in predicting soil temperature decreases as the difference between $T_i$ and $T_w$ increases (Table 1).

Table 1. Measurements of initial soil temperature ($T_i$), temperature of the water in the infiltrometer ($T_w$) and the absolute difference between the two is shown with measurements of the final soil temperature ($T_f$) predicted value of final soil temperature ($T_{f p}$) and the absolute difference between the final and predicted soil temperatures.

<table>
<thead>
<tr>
<th>$T_i$</th>
<th>$T_w$</th>
<th>absolute difference</th>
<th>$T_f$</th>
<th>$T_{f p}$</th>
<th>absolute difference</th>
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<tbody>
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<td></td>
<td></td>
<td>$T_w-T_i$</td>
<td>$T_f$</td>
<td>$T_{f p}$</td>
<td>$T_f - T_{f p}$</td>
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Varying both $T_i$ and $T_f$ had a substantial effect on $K(h)$. $K(h)$ values from measurements made near 40° C were almost double that at 5° C (Fig. 1A).

Measurements under isothermal conditions at 35% soil moisture showed that by
Fig 1. Results for unsaturated hydraulic conductivity measurements (K(h)) taken with the soil near saturation (soil moisture 38 %) and under dry conditions (soil moisture 5 %). For both soil moisture regimes, five replications were performed near 5, 13, 20, 30 and 40° C, with near isothermal conditions between the soil and water in the infiltrometer. A. shows mean and standard error of uncorrected K(h) values. B. K(h) values after corrected to a reference of 20° C and C. sequential calibration of K(h) for viscosity and surface tension effects of temperature, resulting in no significant difference in K(h) within each initial soil moisture content group.
correcting the calculated K(h) value for changes in the dynamic viscosity, effectively calibrated K(h) values at 20° C (Fig 2A).

Variability between replications of isothermal measurements at 5 % soil moisture made it difficult to assess calibration with dynamic viscosity at temperatures above 20° C (Fig 2B). Variability in K(h) measurements on dry soil may be due to wetting hysteresis effects and soil compaction during the sequence of wetting and drying by heating in repeated experiments. However, with $\beta_0 = 192$, sequential calibration for viscosity and surface tension effects of temperature resulted in no significant difference in K(h) within each initial soil moisture content group (Fig. 1C).

CONCLUSIONS

Results indicate that the accuracy of the calorimetric approach in predicting $T_f$ is dependent on the difference between $T_i$ and $T_w$. However, measurements performed at different temperature with both $T_i$ and $T_w$ under isothermal conditions can be accurately calibrated for temperature change by multiplying the calculated K(h) by its associated dynamic viscosity. Therefore these results indicate that when calibrating K(h) measurements with temperature, care should be given to minimize the difference between $T_i$ and $T_w$. Furthermore, accuracy of results can further be improved upon through determining $T_f$ through direct measurement. Under dry soil moisture conditions sequential calibration for viscosity and surface tension effects of temperature can result in effective calibration of K(h) to a reference of 20° C.
REFERENCES


APPENDIX B

PERMISSION-TO-REPRINT LETTER
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