DETERMINATION OF SUBSURFACE WATER MOVEMENT USING SELF-POTENTIAL MEASUREMENTS

by

Emily Voytek
A thesis submitted to the Faculty and the Board of Trustees of the Colorado School of Mines in partial fulfillment of the requirements for the degree of Doctor of Philosophy (Hydrology).

Golden, Colorado

Date ______________________

Signed: _______________________________

Emily Voytek

Signed: _______________________________

Dr. Kamini Singha
Thesis Advisor

Golden, Colorado

Date ______________________

Signed: _______________________________

Dr. Terri Hogue
Professor and Director of
Hydrologic Science and Engineering Program
ABSTRACT

Accurate quantification of water movement, both in magnitude and direction, is a necessary component of evaluating any hydrologic system. Groundwater flow patterns are usually determined using a network of piezometers or wells, which can be limited due to logistical or regulatory constraints. In the unsaturated zone, tensiometers can be used to determine unsaturated flow, but require knowledge of site-specific soil moisture. In either case, piezometers and tensiometers measure potentials from which flow is inferred rather than directly measuring water movement.

An emerging alternative is to measure small currents that are generated as water moves through earth material. These currents generate small voltage differences detectable at the ground surface. Measurement of these voltage differences is the basis of the self-potential (SP) method. Signals can be measured using only two electrodes, or through installation of an array of electrodes. Here we present the results of multiple SP surveys designed to help address open hydrologic questions at multiple temporal (single snapshot, monthly surveys and continuous measurements) scales. In the first project, SP surveys are used to map complex flow patterns contributing to preferential hillslope drainages in an area of continuous permafrost. In the second, a subsurface electrode array is used to measure small changes in vertical and horizontal unsaturated flow rates induced by tree transpiration. Finally, through the example of repeat SP surveys collected in a remote sub-alpine meadow, we demonstrate how additional field data sets and coupled fluid flow and electrical models can constrain interpretations of SP data.
TABLE OF CONTENTS

ABSTRACT ........................................................................................................................................ iii
LIST OF FIGURES .......................................................................................................................... vii
LIST OF TABLES ........................................................................................................................... x
ACKNOWLEDGEMENTS ............................................................................................................ xi

CHAPTER 1  INTRODUCTION ............................................................................................1

CHAPTER 2  BRIEF HISTORY OF THE SELF-POTENTIAL METHOD ...........................5

CHAPTER 3  IDENTIFYING HYDROLOGIC FLOWPATHS ON ARCTIC HILLSLOPES USING ELECTRICAL RESISTIVITY AND SELF-POTENTIAL ...........................................................................................................................7
  3.1 Introduction ........................................................................................................7
  3.2 Field Site Description ......................................................................................10
  3.3 Methods............................................................................................................12
    3.3.1 Electrical Resistivity ......................................................................12
    3.3.2 ER Inversion ..................................................................................14
    3.3.3 Self-potential ..................................................................................15
  3.4 Results and Discussion ....................................................................................19
    3.4.1 Individual ER Profiles....................................................................19
    3.4.2 Self-potential Data .........................................................................22
  3.5 Conclusions ......................................................................................................24

CHAPTER 4  PROPAGATION OF DIEL TRANSPIRATION SIGNALS IN THE SUBSURFACE OBSERVED USING THE SELF-POTENTIAL METHOD .................................................................................................................25
  4.1 Introduction ......................................................................................................25
  4.2 Background ......................................................................................................27
    4.2.1 Vegetation-hydrology interactions.................................................27
    4.2.2 Geophysics to Map Ecohydrologic Processes ...............................28
4.2.3  Self-potential Background .............................................................30
4.3  Field Site .........................................................................................40
4.4  Methods ......................................................................................................41
  4.4.1  Self-potential and corroboratory measurements ........................................42
  4.4.2  Coupled Soil Water Flow and Electrical Modeling .......................................44
4.5  Results and Discussion ....................................................................................50
  4.5.1  Precipitation ...........................................................................50
  4.5.2  Tree Transpiration ..........................................................................50
  4.5.3  Soil Moisture ........................................................................51
  4.5.4  Matric Potential ........................................................................52
  4.5.5  Field SP Data ........................................................................53
    4.5.5.1  Seasonal variability in SP ............................................54
    4.5.5.2  Diel fluctuations in SP .................................................56
4.5.6  Modeling Results ...........................................................................57
  4.5.6.1  Seasonal Variations ......................................................58
  4.5.6.2  Diel Fluctuations ..........................................................59
4.6  Conclusions ......................................................................................................60

CHAPTER 5  EVALUATION OF SELF-POTENTIAL SIGNAL SOURCES
ON FIELD DATA FOR HYDROLOGIC STUDIES ..................................................61

5.1  Introduction ..............................................................................................61
5.2  Background ..............................................................................................62
  5.2.1  Self Potential Method .............................................................................62
    5.2.1.1  Streaming Potential ........................................................................67
    5.2.1.2  Diffusion Potential ..........................................................................69
    5.2.1.3  Thermoelectric Potential ................................................................71
    5.2.1.4  Redox Potential .............................................................................72
5.2.2 Other sources of SP voltages ................................................................. 73
  5.2.2.1 Anthropogenic and environmental noise .............................. 73
  5.2.2.2 Temperature effect on electrodes ........................................ 73

5.3 Methods ........................................................................................................ 74
  5.3.1 Field Site ......................................................................................... 75
  5.3.2 SP data collection ........................................................................... 76
  5.3.3 Auxiliary data .................................................................................. 78
    5.3.3.1 Stream Levels ........................................................................ 79
    5.3.3.2 Air Temperature ...................................................................... 80
    5.3.3.3 Electrical Resistivity .............................................................. 80

5.4 Results and Discussion .............................................................................. 83
  5.4.1 Evaluation of Electrochemical Potential ..................................... 83
  5.4.2 Evaluation of Thermoelectric Potential ......................................... 84
  5.4.3 Evaluation of Streaming Potential ................................................. 86

5.5 Conclusions ............................................................................................... 90

CHAPTER 6 CONCLUSIONS ............................................................................. 91

REFERENCES CITED ....................................................................................... 94

APPENDIX A MAKING IRIS COMPATIBLE RESISTIVITY CABLES ................ 108
APPENDIX B BASIC SP FIELD GUIDE .......................................................... 117
APPENDIX C GETTING STARTED MODELING SP IN COMSOL 5.2a .............. 124
APPENDIX D MATLAB THREE-POINT PROBLEM SCRIPT ....................... 138
LIST OF FIGURES

Figure 3.1  Regional map showing position of Water Tracks 1 and 6 .............................. 11
Figure 3.2  Photo of Water Track 6 looking upslope ........................................................ 13
Figure 3.3  ER profiles from Water Track 1 ................................................................. 16
Figure 3.4  ER profiles from Water Track 6 ................................................................. 16
Figure 3.5  Plots of measured self-potential voltages .................................................... 18
Figure 3.6  Comparison of four types of data evaluated at Water Track 1 ........................ 21
Figure 4.1  a) Schematic of electrical double layer (EDL) formed on the exterior of a mineral grain when in contact with water. b) Plot of relative charges in the pore water with relative distance from the mineral surface. Modified from Revil and Jardani (2013) .................................. 33
Figure 4.2  Cross-sectional view of sensor array relative to selected tree. ...................... 36
Figure 4.3  Measured monthly field data from sensors in Figure 4.2. .............................. 37
Figure 4.4  Measured daily field data from sensors shown in Figure 4.2. ........................ 38
Figure 4.5  a) Saturation-dependent effective excess charge, $Q_v$, $sw$, and effective conductivity accounting for saturation, $\sigma_{sw}$. b) Ratio of $Q_v$, $sw$ to $\sigma_{sw}$, indicative of SP signal strength at given saturation. The same fluid velocity will produce a stronger signal at low saturation. ................................................................. 39
Figure 4.6  Conceptual model used to investigate SP signals ........................................ 43
Figure 4.7  a) Measured monthly tree transpiration and precipitation used as boundary conditions in the model, b) modeled Darcy velocity, c) soil moisture, d) matric head, and e) SP voltages between subsurface electrodes. (w/o = without root-water uptake). ........................................................................ 45
Figure 4.8  a) Measured daily tree transpiration and precipitation used as boundary conditions in the model, b) modeled Darcy velocity, c) soil moisture, d) matric head, and e) SP voltages between subsurface electrodes. (w/o = without root-water uptake) ......................................................... 46
Figure 5.1  The four most common sources of electrical potential ................................. 66
Figure 5.2  a) Streaming potential coupling coefficient as a function of fluid conductivity from Equation 5.7 and b) bounds of thermoelectric coupling coefficient bounds as a function of fluid conductivity from Equations 5.10 and 5.11. ........................................... 69

Figure 5.3  Regional map of the field site ........................................................................ 75

Figure 5.4  Interpolated plots of SP data from Andrews Meadow ................................... 77

Figure 5.5  a) Elevation of water level in Andrews Creek and Andrews Spring ....................... 79

Figure 5.6  ER profiles shown as bulk conductivity values. Locations of individual lines shown in Figure 5.3. ....................................................................... 81

Figure 5.7  Thermoelectric potential generation from a) horizontal and b) vertical temperature gradients ........................................................................................................ 85

Figure 5.8  Site specific coupling coefficient ...................................................................... 88

Figure 5.9  SP data converted to maps of hydraulic head using coupling coefficient of -132 mV/m from Equation 5.7. ................................................................. 89

Figure A.1  Cutting and assembly instructions .................................................................. 111

Figure A.2  Pin diagram of a 22-55 connector ................................................................. 112

Figure A.3  Wire preparation ........................................................................................... 113

Figure A.4  Tool preparation ......................................................................................... 113

Figure A.5  Assemble connector ..................................................................................... 114

Figure A.6  Crimp connection ....................................................................................... 114

Figure A.7  Insert pins .................................................................................................... 115

Figure A.8  Pin removal .................................................................................................. 115

Figure A.9  Strain relief ................................................................................................ 116

Figure B.1  The SP field equipment I used in the field ..................................................... 119

Figure B.2  Close up of the field equipment and wire connections .................................. 121

Figure B.3  Demonstration of a tip-to-tip reference measurement .................................. 123

Figure C.1  Starting a COMSOL model ........................................................................ 125
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>C.2</td>
<td>COMSOL workspace</td>
<td>126</td>
</tr>
<tr>
<td>C.3</td>
<td>Model building menus within COMSOL</td>
<td>128</td>
</tr>
<tr>
<td>C.4</td>
<td>Preparing the groundwater flow model</td>
<td>130</td>
</tr>
<tr>
<td>C.5</td>
<td>Preparing the groundwater flow problem</td>
<td>131</td>
</tr>
<tr>
<td>C.6</td>
<td>Adding the electrical module to the COMSOL model</td>
<td>133</td>
</tr>
<tr>
<td>C.7</td>
<td>Forward modeled streaming potentials results, in Volts</td>
<td>134</td>
</tr>
<tr>
<td>D.1</td>
<td>Data plot produced by sample data provided in three_point_scaled.m</td>
<td>139</td>
</tr>
<tr>
<td>Table</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>---------</td>
<td>------------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>4.1</td>
<td>Parameters used in coupled fluid flow and electrical models</td>
<td>48</td>
</tr>
<tr>
<td>5.1</td>
<td>Parameters used in data processing and numerical modeling</td>
<td>82</td>
</tr>
<tr>
<td>A.1</td>
<td>Specifications of the resistivity cable</td>
<td>108</td>
</tr>
<tr>
<td>A.2</td>
<td>Parts and tools required and recommended to build cables</td>
<td>110</td>
</tr>
<tr>
<td>A.3</td>
<td>Pin addressing for the two connections on the IRIS</td>
<td>112</td>
</tr>
<tr>
<td>B.1</td>
<td>Items required for conducting an SP field survey</td>
<td>118</td>
</tr>
<tr>
<td>B.2</td>
<td>Sample field form for recording SP data</td>
<td>122</td>
</tr>
</tbody>
</table>
ACKNOWLEDGEMENTS

I sincerely thank my advisor, Dr. Kamini Singha, for all of her guidance and support throughout this process. This would not have been possible without your patience and the incredible independence you have allowed me to follow my research interests. I thank you for all of the time and energy you have invested.

I extend deep gratitude to Dr. John Lane and the entire U.S.G.S. Branch of Geophysics in Storrs, CT. You introduced me to the field of hydrogeophysics and have always pushed me one step further, including out the door when the time came to pursue a PhD. For that, and so much more, I am thankful. I’ve also received much support from other colleagues from the U.S.G.S. including Dr. Fred Day-Lewis, Dr. Burke Minsley and Dr. Michelle Walvoord.

There are many folks from Mines to whom I am grateful: Jennifer Jefferson, who started this journey with me and has continued to support me until the end; my crew of Ryans, who, along with difficulty distinguishing you to others, have each brought a unique and important perspective to this process; Carolyne Bocovich, who’s helped hold me accountable to my daily and longer-term goals; Jackie Randell, whose organization skills and field knowledge have been invaluable; and the many members of the Singha Research Group and the larger Mines Hydrologic Science and Engineering community who have helped me during my time here.

Beyond the Mines community, thanks are due to Dr. André Revil for sparking my interest in SP and answering many, many questions on the subject, and to Dr. Damien Jougnot for continued discussions about SP and research in general.

Financial support for this work was provided by the Department of Defense National Defense Sciences and Engineering Graduate Fellowship (NDSEG), Air Force Office of Scientific Research, and numerous grants from the School of Mines Department of Geology and Geological Engineering. Field work in Chapter 3 was supported by National Science Foundation.
Office of Polar Programs Award Number 1259930. Logistical support was provided by CH2M Hill Polar Services and the staff of Toolik Field Station. Terrestrial LiDAR support was provided by UNAVCO. Field work in Chapter 4 was supported by the American Geophysical Union Horton Research Grant and the National Science Foundation Hydrologic Sciences Program EAR 1446161.
CHAPTER 1
INTRODUCTION

Groundwater is often the most difficult piece of the hydrologic cycle to constrain, because data are limited. Typically, investigations to characterize and monitor groundwater rely on well networks. However, logistical and regulatory constraints can make the use of wells impractical or impossible (e.g. Clow et al., 2003). In these cases, low-cost, non-intrusive methods of determining where groundwater is flowing are essential.

As water moves in the subsurface, it generates small electrical currents. These currents can be large enough to produce measureable voltage differences at the ground surface. The self-potential (SP) method uses measurement of these voltage differences to unobtrusively evaluate where water is flowing in the subsurface. SP data collection requires only two electrodes, a handheld voltmeter and some wire, so it’s relatively easy to add the technique to ongoing hydrologic studies. Additionally, it is particularly well suited for remote field sites where weight and volume of equipment may be limited.

Despite SP’s long history (Fox, 1830), the method has seen relatively little use in hydrologic studies until recent years (e.g. Revil and Jardani, 2009). One reason was the lack of a comprehensive set of equations to describe the how water movement generates currents until the 1990s (Pride, 1994). Another limitation is the (real or perceived) difficulty of separating confounding signal sources during data interpretation. In addition to groundwater movement, other subsurface process such as chemical and temperature gradients can generate voltage differences, and the multiple signals sources can be difficult to separate. This challenge led the SP method to acquire the label of the “ugly duckling of environmental geophysics” (Nyquist and Corry, 2002). Recent advances in expanded governing equations and data processing techniques
have led to more quantitative interpretations of SP measurements. Given these advances, it is
time to reconsider whether the ugly duckling is finally transforming into a “beautiful swan”
(Revil and Jardani, 2009). I outline the history of SP in Chapter 2.

In the following research, I seek to expand the use of the SP method by applying it to
multiple hydrologic processes spanning geographic settings (i.e. continuous permafrost in
Alaska, an alpine meadow in the Colorado Rockies, and a forested hillslope in Oregon) and
differing time scales (single snapshot, monthly data sets and continuous measurements). In
addition to advancing use of the SP method in hydrology, each of the projects included here
sought to answer questions associated with specific hydrologic processes at spatial and temporal
resolutions not previously captured with other tools.

In Chapter 3 I use SP and another geophysical tool, electrical resistivity (ER), to analyze
groundwater flow patterns around water tracks, drainage features unique to hillslopes underlain
by permafrost in Alaska. Water tracks have been previously analyzed on the watershed and
hillslope scale (McNamara et al., 1998), but until now the smaller-scale patterns of flow and their
relation of the underlying permafrost boundary have not been described. This type of detailed
knowledge of flow patterns is required for accurate quantification of nutrient cycling (e.g.
Fischer et al., 2004), permafrost thaw (e.g. Sjoberg et al., 2016) and runoff generation (e.g. Freer
et al., 2002). In this project, I collected SP measurements in a dense (2 m x 2 m) grid across two
water tracks in northern Alaska, and used these data to determine relative flow magnitudes and
direction around these drainage features. These results lay the groundwork for future
investigations into the seasonal dynamics, hydrologic connectivity, and climate sensitivity of
arctic hillslopes.
My second project, in Chapter 4, pushes the use of SP into the realm of continuous measurement to test developing ideas of how tree transpiration impacts the distribution of water around tree roots, with the ultimate goal of determining how these signals propagate to soil water, groundwater, and streams downslope at a site in Oregon. Current methods of quantifying these impacts include isotopic and fluid potential measurements. However, isotopic measurements do not directly measure water movement and are often space- and time-limited. Fluid potential measurements in the unsaturated zone are notoriously difficult to acquire, and require knowledge of relationships between to pressure and soil moisture to quantify soil water movement. In this work, I use SP measurements to observe the small changes in soil-water movement generated when trees ‘turn on’ during the day, and ‘turn off’ at night. To do this, I installed an array of SP electrodes in the ground, and continuously measured the differential voltages between pairs of electrodes over 5 months. I was able to reproduce the SP signals I observed through numerical modeling to confirm that they were caused by tree water uptake.

My third project, in Chapter 5, builds on the success of the work in Alaska and Oregon and uses SP to evaluate seasonal changes of groundwater flow within an alpine meadow in Rocky Mountain National Park, Colorado. Geochemical analyses have identified seasonal changes in the sources of water to Andrews Creek, a mountain headwater stream bordering the meadow (Wilson, 2015). Quantifying the seasonal changes in sources of water to the stream (i.e. snowmelt, rain water and groundwater) will help predict how stream chemistry will be altered by expected changes in precipitation patterns with a changing climate (Clow, 2010). Andrews Meadow, the site of the geochemical analysis, is designated Wilderness and regulatory constraints prevent the installation of groundwater wells to support the findings from water geochemistry. Here SP measurements were collected in a grid across the meadow seven times
over five months to document changing patterns of groundwater flow in response to hydrologic conditions at the ground surface. I use these data and coupled modeling, to demonstrate how additional field data can constrain interpretations of SP data. Through these examples I seek to inform hydrologists of the best practices for collecting SP measurements as part of ongoing hydrologic studies.

SP can be an effective tool for mapping patterns of shallow groundwater and unsaturated zone flow to analyze near-surface processes at high spatial and temporal resolution. Collectively these three projects advance the use of SP as an important tool in the hydrogeophysicist’s toolbox, and help the method mature from ugly duckling to beautiful swan.
CHAPTER 2
BREIF HISTORY OF THE SELF-POTENTIAL METHOD

SP is one of the oldest geophysical methods, yet it is still infrequently used, especially in hydrogeology. The first use of SP was in ore exploration when Fox (1830) presented observations of ‘electrical action’ in association with sulfide veins in Cornwall, UK. This and other early observations of SP signals around ore bodies formed the basis for the comprehensive model of Sato and Mooney (1960), which associated SP signals to half-cell reactions or a “geo-battery” (Revil and Jardani, 2013). In the “geo-battery” model, SP signals are generated from areas in the Earth with differing redox potentials.

While the early use of SP primarily focused on ore and hydrocarbon exploration, the association between SP measurements and groundwater flow was observed as early as 1894 by Bachmetjew (referenced in Revil and Jardani, 2013). A physical explanation emerged when Nourbehecht (1963) first derived generalized equations for electrokinetic coupling, current flow resulting from fluid flow, using Darcy and Ohm’s laws. These relationships were refined by Pride (1994) through derivation from first principles. In the past decade, these equations have been expanded to include more complex hydrologic conditions including unsaturated flow (Revil et al., 2007) and high Reynolds numbers (Bolève et al., 2007).

With improvements in the physical description of signal generation and non-polarizing electrodes required for measurement (Petiau, 2000), use of SP in hydrology has increased in recent years. The method has been used to determine groundwater flow paths in the subsurface (e.g. Revil et al., 2005), monitor hydrologic pumping tests (Jardani et al., 2009; Rizzo et al., 2004; Soueid Ahmed et al., 2014), identify flow through earthen dams (Ikard et al., 2014) and...
determine vertical infiltration rates (Doussan et al., 2002; Suski et al., 2006). In periglacial environments, SP has been used to investigate potential seepage through an ice-cored moraine (Moore et al., 2011) and to estimate sub-glacial drainage (Kulessa et al., 2006). Despite the variety of settings in which SP has proved useful, its use is still small compared with other geophysical methods. Limited use of the method may be because of the multiple sources of electrical noise that can interfere with measurements and difficulty in data interpretation due to multiple signal sources (Nyquist and Corry, 2002). A lot of recent SP work has been conducted in the laboratory to isolate sources of signal generation such as temperature (Ikard and Revil, 2014) and water chemistry (Ikard et al., 2012).

Historically, interpretation of SP data has been qualitative. Recently, more quantitative interpretations have resulted from inversion of SP data. Minsley et al. (2007) inverted synthetic data to determine the distribution of current point sources and presented an example of inverting field data from a groundwater pumping test conducted by Bogoslovsky and Ogilvy (1973). Bolève et al. (2009) used a coupled SP inversion (using electrical and hydrologic data) to determine preferential flow pathways through an earthen dam.

To become a go-to geophysical tool, the usefulness of SP in real-world applications must continue to be proven in diverse settings and for diverse processes. In each of the three projects included here, SP data were collected in remote field sites where the level of control achieved in the laboratory (porosity, water chemistry, and temperature) is not possible. Instead, alternative means of interpretation, including coupled numerical modeling, are necessary. Beyond verification of the SP method, each project evaluates a particular hydrologic process in a time- or spatial-scale not previously considered due to data limitations.
CHAPTER 3
IDENTIFYING HYDROLOGIC FLOWPATHS ON ARCTIC HILLSLOPES USING
ELECTRICAL RESISTIVITY AND SELF POTENTIAL

A paper published in Geophysics

Reprinted with permission from the Society of Exploration Geophysicists

Emily B. Voytek*, Caitlin R. Rushlow, Sarah E. Godsey and Kamini Singha

3.1 Introduction

Water flow through the saturated soils of the shallow subsurface is often a dominant
process controlling runoff generation (e.g. Hewlett and Hibbert, 1967), soil development (e.g.
Lohse and Dietrich, 2005), and solute transport (e.g. McGlynn and McDonnell, 2003) in
watersheds. Despite their mechanistic importance, locating shallow subsurface flowpaths
remains challenging (Nippgen et al., 2015). Traditional methods for mapping shallow subsurface
flowpaths include direct observation through labor-intensive soil surveys and water content
monitoring schemes (e.g. Tromp-van Meerveld and McDonnell, 2006; James and Roulet, 2007;
Ali et al., 2011), or indirect predictions using terrain-based modeling (e.g. Jencso and McGlynn,
2011) or chemical or isotopic tracers (e.g. Tetzlaff et al., 2014). Both direct and indirect methods
have major drawbacks: tracer and modeling techniques are data intensive, and manual surveying
is unsuitable for environments where soil properties vary strongly through space or time.

*Primary researcher and author

1http://dx.doi.org/10.1190/geo2015-0172.1


3Department of Geoscience, Idaho State University, Pocatello, Idaho.

4Department of Geology and Geological Engineering, Colorado School of Mines, Golden, Colorado
In arctic systems, subsurface flowpaths are limited in depth by the frozen boundary of permafrost (Woo, 2012). The soil profile is fully frozen in the winter, but higher energy inputs in the summer cause the progressive downward growth of a thawed subsurface region called the active layer, before the soils freeze again in the fall (e.g. Harris et al., 1988; Kane et al., 1991). The subsurface topography at the boundary between the thawed and frozen ground, which controls the location of shallow subsurface flowpaths, is difficult to assess with direct observations over large regions, at a fine resolution, or through time (Nelson et al., 1998). Instead, subsurface topography is generally assumed to be a function of the surface topography (Stieglitz et al., 2003).

Geophysical techniques show promise for testing this assumption and mapping the active layer and subsurface flowpaths on arctic hillslopes. Several techniques have been applied to permafrost systems, including electrical resistivity (ER) tomography, ground-penetrating radar (GPR), and electromagnetic (EM) methods (see reviews by Scott et al. (1990), Kneisel et al. (2008), and Hauck, (2013)). Electrical methods are ideal for work in permafrost regions because bulk electrical resistivity depends on the temperature and phase of water (Ananyan, 1958; Hayley et al., 2007). Frozen ground is more resistive to electrical current than unfrozen ground, with typical resistivity values above 1000 ohm-m (Hauck and Kneisel 2008), although the exact value varies depending on soil material and the proportional frozen/liquid water content (Ananyan, 1958). Recent theoretical calculations of resistivity changes due to temperature and water contents have been based on variations of Archie’s law which relates bulk electrical conductivity to the fluid, through porosity and empirically derived parameters (Hauck et al., 2011; Minsley et al., 2015). These relationships are the basis of using ER to identify areas of partially frozen ground in the subsurface. ER has previously been used to identify the frozen
ground boundary in many settings in both continuous (Overduin et al., 2012) and discontinuous (Lewkowicz et al., 2011; McClymont et al., 2013) permafrost zones. Technological advances have also made long-term monitoring of permafrost boundaries possible, resulting in quantitative understanding of annual freeze-thaw cycles in mountain permafrost (e.g. Hauck, 2002; Hilbich et al., 2008; Krautblatter et al., 2010).

In addition to ER and other active electromagnetic methods, the self-potential (SP) method can help inform our knowledge of subsurface processes in permafrost environments. SP is a passive electrical method that is sensitive to the small currents generated as water moves through soils. Voltage differences resulting from these currents are measured at the ground surface and analyzed to determine groundwater flowpaths in the subsurface (e.g. Revil et al., 2005). The method has been used to successfully monitor hydrologic pumping tests (Jardani et al., 2009; Rizzo et al., 2004; Soueid Ahmed et al., 2014), identify flow through earthen dams (Ikard et al., 2014) and determine infiltration rates (Doussan et al., 2002; Suski et al., 2006). In periglacial environments, it has been used to investigate potential seepage through an ice-cored moraine (Moore et al., 2011). However, SP has not previously been used to identify subsurface flowpaths in arctic hillslopes.

In this study, we pair ER and SP to investigate groundwater flowpaths in and around common drainage features of arctic hillslopes called water tracks. ER images the basal boundary of the active layer, which controls shallow subsurface flowpaths beneath water tracks on permafrost-underlain hillslopes, while SP maps the direction of groundwater flow through these features. We explore the direction and magnitude of flow beneath the primary channel of two water track features, and investigate whether the subsurface topography mimics the surface topography along transects crossing the water tracks. Our study sets the stage for quantifying the
seasonal growth, hydrologic connectivity, and climate sensitivity of spatially distributed flowpath networks on arctic hillslopes.

3.2  **Field Site Description**

In August 2014, a series of ER profiles and SP data were collected in the Kuparuk River watershed of northern Alaska to characterize thickness of active layer thaw (Figure 3.1). The watershed is underlain by continuous permafrost and defined locally by massive and gently sloping moraines of the Sagavanirktok River Glaciation (Hamilton, 1986). The ecology and hydrology of the Kuparuk River has been studied since the mid-1980s, and the river is currently part of the Arctic Long-Term Ecologic Research Site (e.g., Bowden et al., 2014). Six water tracks that drain into the Kuparuk River are extensively monitored as part of ongoing research. This study focuses on two water tracks, Water Tracks 1 and 6 (Figure 3.1), which were selected for their proximity to roads and to minimize interference with other experiments. High-resolution topography from ground-based LiDAR collected during the same field campaign suggests that Water Track 1 drains 0.09 km$^2$ of hillslope at its weir on the western side of the river. Water Track 6 is about 1/3 the size of Water Track 1, draining 0.03 km$^2$ of the hillslope on the eastern side of the river. Both water tracks occur on hillslopes with moist, acidic tundra vegetation, but the dominant emergent vegetation along the study reach of Water Track 1 is sedge, while Water Track 6 is characterized by abundant dwarf willows and birches. At both sites, a peaty organic soil horizon covers deeper mineral soils. Twenty-four soil cores collected at Water Track 1 in
Figure 3.1 Regional map showing position of Water Tracks 1 and 6 relative to the Kuparuk River and the Dalton Highway (AK 11) in northern Alaska. Site maps of Water Tracks 1 and 6 showing the position of ER transects relative to previously installed weirs, and two-meter elevation contour intervals. The shaded areas represent the area contributing to flow at the monitoring weirs as calculated from ground-based LiDAR data.
July and August 2014, twelve inside the water track feature and twelve outside on the non-track hillslope, revealed that organic soil horizon thickness was variable, but generally thicker inside the water track. Organic layer thickness inside Water Track 1 ranged from 23 cm to more than 88 cm, while organic layer thickness outside the water track was usually less than 23 cm, ranging from only 3 cm to 43 cm in thickness. Some glacial erratics are visible on the surface at both sites, but organic soils cover >99% of the drainage area (Figure 3.2).

3.3 Methods

3.3.1 Electrical Resistivity

Electrical resistivity (ER) methods work by injecting electrical current into the ground using two electrodes, and measuring the resulting voltage distribution at other electrode locations. From the known amount of injected current and measured voltage differences, a resistance can be calculated for each quadripole (combination of four electrodes). Multiple resistance measurements from different quadripole spacings and offsets can be combined through the process of inversion to produce a profile of subsurface resistivity. These methods have been used for delineating subsurface lithology and hydrologic units in a variety of settings (Parsekian et al., 2015; Pellerin, 2002; Robinson et al., 2008). As discussed above, ER has been used successfully in numerous permafrost settings to identify frozen ground extent.

ER data were collected using an IRIS Syscal Pro and stainless steel electrodes. Electrodes were inserted into the ground until good contact was made in the mineral soil or competent organic material, which was typically 10-15 cm below land surface. Given the generally moist conditions, contact resistances between electrodes were low, with a median value of 4.1 kohm and 6.5 kohm between electrodes at Water Tracks 1 and 6 respectively. At both water tracks, ER
Figure 3.2 Photo of Water Track 6 looking upslope from approximately 0-meter ER profile. The subtle topographic changes within the water track are highlighted by differences in vegetation. The weir, flume, and site-access boardwalks are in the foreground. The nine shallow groundwater wells are visible along the water track axis and on the adjacent non-track hillslope.
data were collected in a series of parallel transects, approximately centered on, and perpendicular to, the primary drainage of the water track (Figure 3.1). One transect at each site was collected below a plywood weir installed for flow monitoring, while the remaining transects were collected upstream of the weir.

At Water Track 1, six parallel resistivity lines were collected approximately 20 meters apart (Figure 3.1, Water Track 1). Each survey consisted of 96 electrodes at 0.5-m spacing, for a total transect length of 47.5 m. At Water Track 6, a similar collection scheme was used with a total of eight transects spaced 10 m apart (Figure 3.1, Water Track 6). Water Track 6 is narrower than Water Track 1, and therefore each ER profile consisted of only 48 electrodes at 0.5 m spacing for a total of 23.5 m. Such fine spacing (0.5 m) was used to better capture the thaw boundary, which was estimated to be between 0.5 and 1 m below the ground surface from frost probing during August when the geophysical data were collected. In August, the active layer thickness is nearly maximized for the year, as the ground begins freezing again in September or early October.

3.3.2 ER Inversion

In ER surveys, any combination of electrodes can be used for current injection and voltage measurements, but certain sequences have emerged that balance sensitivity and collection time. A dipole-dipole sequence was used for data collection at both sites due to its speed and ability to detect lateral variations (Barker, 1979). At Water Track 1, the collection sequence for the 96-electrode transects included 1050 quadripoles, while 1159 quadripoles were used for the 48-electrode transects at Water Track 6. The collection sequence at Water Track 1 had fewer quadripoles despite the longer transect length because quadripoles sensitive to depths
well below the permafrost boundary were not collected. Each quadripole was measured twice during a 500 ms current injection, and the relative error between the two measurements was used as quality control and to weight the data during processing (median error = 0.1%).

The inversion code R2 was used to process the field data with a 0.125-m cell size; details of the code can be found in Binley and Kemna (2005). Each inversion converged in 4-6 iterations with an average root mean square error of 1.1 relative to the measured noise calculated by the stacking errors. The depth of investigation (DOI) method of Oldenberg and Li (1999) was used to determine what portion of the inverted profile was informed by the data, rather than the inversion parameters. This method involves processing the same data set at least two times, each regularized to different background resistivity. Areas that are informed by the data result in similar resistivity values in each of the two inversions; areas beyond the DOI change as the background model is adjusted. These areas of lower data sensitivity are removed from plotting (Figure 3.3 and Figure 3.4) to reduce the possibility of over-interpretation.

3.3.3 Self-potential

SP is a passive geophysical technique which relies on measurements of voltage differences at the land surface. These voltage differences are created by naturally occurring electrical currents in the subsurface. One source of the electrical current is water flow through porous material or “streaming potential”. These currents initiate from the electrical double layer formed at the fluid-grain boundary (Revil and Leroy, 2001). As water molecules move past charged grain surfaces, a small electrical current is produced. The depth of investigation for a
Figure 3.3 ER profiles from Water Track 1 plotted on a local coordinate system. Elevations are shown in meters above sea level. Manual thaw probe data is indicated by black xs. Vertical exaggeration is 2x.

Figure 3.4 ER profiles from Water Track 6 plotted on a local coordinate system. Elevations are shown in meters above sea level. Manual thaw probe data is indicated by black xs. No vertical exaggeration.
passive technique, such as SP, is difficult to determine since it is dependent on the strength of the source, which is unknown.

At the water tracks, voltage differences between two Petiau-type (Petiau, 2000) non-polarizing electrodes were measured with a Fluke 87V handheld voltmeter. The reference electrode was buried ~20 cm to reduce drift caused by internal temperature variation. Temperature drift of Petiau-type electrodes is small, 0.22 mV/°C, relative to other non-polarizing electrodes, which can exceed 2 mV/°C (Revil and Jardani, 2013). Individual measurements were collected in a grid using the roving electrode, typically placed on the surface or 0-3 cm into the ground when sediment allowed. The voltage difference between the reference and roving electrodes was checked every 60 – 120 measurements. The limited amount of drift between these electrodes observed during a survey (< 3 mV), was assumed to be due to temperature variations between the two electrodes, and was distributed evenly over the intervening measurements during drift correction.

At Water Track 1, SP measurements were made in a 2-m grid across a 60 x 74m area resulting in 1178 measurements. The measurement spacing at Water Track 6 was modified so that measurements were more densely spaced surrounding the primary channel (1-m spacing), and more widely spaced on the surrounding hillslopes (up to 4-m spacing) resulting in 828 measurements in a 60 m x 70 m area. Due to spatial constraints, the footprint of the ER and SP measurements are not identical at each of the sites, and their relative positions are shown in Figure 3.1

To better evaluate the trends across each water track, a continuous surface was interpolated from the individual SP measurements (Figure 3.5). The surface was created by solving an inverse problem that minimizes the curvature of the surface while respecting the
confidence of the measurements. Error was estimated by comparison of duplicate measurements made during the course of the SP surveys. The median error between repeat measurements at Water Track 1 and Water Track 6 was 1.5 mV. Since self-potential measurements are relative to the reference electrode, rather than absolute values, measured voltages at each water track have been shifted so that the minimum value in each plot is zero.

3.3.4 Additional Datasets

To compare the ER-derived subsurface topography with surface topography, a 0.5-m digital elevation model for each site was generated using BCAL LiDAR Tools.
(https://bcal.boisestate.edu/tools/lidar) from point cloud data collected using a Riegl VZ-1000 scanner and retro-reflective targets that were georeferenced with survey-grade GPS units. The precision of the point cloud collection and georeferencing was 3 cm or better. The LiDAR surveys were conducted over the same week in August 2014 as the geophysical investigations. Ground-surface elevation along the ER profiles was also surveyed using an automatic level and stadia rod. At each survey location, active layer thickness (thaw depth) was measured as the refusal depth of a 3/8 inch, hex-shaped, insulated steel frost probe inserted into the ground. Finally, at each site, a set of nine fully-screened shallow groundwater wells outfitted with pressure transducers (Onset U20 sensors with a 3.7m range) and referenced to an atmospheric pressure logger were used to monitor the depth to the water table. Three wells were installed along the channel of each water track, and a pair of wells were installed on either side of each water track well on the non-track hillslope.

3.4 Results and Discussion

3.4.1 Individual ER Profiles

The ER profiles at both water tracks have a thin, low-resistivity layer above a more resistive unit (Figure 3.3 and Figure 3.4). This is the expected resistivity structure for a thin thawed layer over frozen ground. At both sites the low-resistivity layer is thicker within the water track compared with the non-track hillslope. The thickening of the lower resistivity in the water track is due to enhanced thaw of the permafrost below the primary channel, which is also observed in the manual frost-probe data. Enhanced thaw results from increased thermal conductivity due to higher water content (Hinzman et al., 1991; Yoshikawa et al., 2002) and snow insulation (Walker et al., 1999). Snow trapped in the topographic low within the water
tracks leads to greater insulation from cold over-winter air temperatures and warmer soils. These processes also support the occurrence of the thickest low-resistivity zone directly upstream of the weir where surface water ponding is continually present (weir position shown in Figure 3.1, thaw visible in 0-m profile in Figure 3.3 and 10-m profile in Figure 3.4).

The absolute resistivity values within these channels are lower than that of the lower resistivity values of the surrounding hillslopes. This trend is likely due to differences in water content, as increased water content would lower the bulk resistivity. Ponded water occurs on the surface upstream of the weirs at both water tracks. The shallow groundwater wells corroborate these general observations at their particular locations. The water table remained 4-8 cm above the ground surface at all three water track wells at Water Track 1 over the study period, while five of the six non-track wells had water tables at depths ranging from 0-20 cm below the surface, with three wells with water tables at least 10 cm below ground. Similar conditions were observed at Water Track 6, where two of the three water track wells had water tables above the ground surface.

Manual frost probe measurements made every two meters along each ER transect suggest that the geometry of the base of the active layer is similar to that of the surface elevation. There is a slight deepening within the water tracks, relative to the surrounding non-track hillslopes (Figures 3.3 and 3.4), although the contrast is not as great as in the ER data. Frost probe measurements were at times limited by the presence of boulders buried in the subsurface that prevented accurate identification of the base of the active layer. In these cases, measurements were made up to 0.5 m off the survey-line, possibly resulting in the differences between frost probe and ER data observed in Figure 3.6. The correlation between the ER and frost probe boundaries is best outside of the main flowpath (black xs in Figures 3.3 and 3.4). A comparison
of all four data types is shown in Figure 3.6. The ER profiles also reveal areas of possible water movement that would not be detectable from frost-probing alone. For example, it is impossible to accurately measure thaw depth below rocks using a frost probe. The ER data revealed a large area of thaw beneath a rock that might facilitate water movement (Figure 3.3, 80 m transect at ~5 m across the transect). Rocks may transfer heat more effectively to the subsurface, especially during the snowmelt when their dark, lower albedo surfaces are exposed, enhancing permafrost thaw below. A second previously unidentified area of thaw is in the low-resistivity bulb that occurs below the main flowpath of the water track, particularly at Water Track 1, where the zone of low resistivity extends significantly deeper than is detected by the thaw probe. If the bulb area were entirely thawed, it would be penetrable by a frost probe. If frozen, the resistivity would be
similar to the surrounding frozen areas. Instead, the presence of a lower resistivity zone beyond the probe-observed frost boundary may indicate an area of partial thaw, in which the unfrozen water content is greater than surrounding areas at the depth. Individual ER profiles cannot identify the water movement through the subsurface, but the lower resistivities suggest that liquid water may be present in these regions. It was not possible to sample these materials to make a conclusive interpretation of the origin of the lower resistivity materials. These data suggest that estimates of active layer thaw from manual frost-probe measurements underestimate the extent of potential flow in the subsurface and therefore the degree of hydrologic exchange between surface and subsurface water of arctic hillslopes.

3.4.2 Self-potential Data

SP signals observed at the surface are a 2-D rendering of complex 3-D patterns in the subsurface. In purely horizontal flow, measured self-potential voltages should increase in the direction of flow (Revil and Jardani, 2013); flowpaths down the hillslope should lead to increasing voltages downstream. In reality, observed signals are complex due to the superposition of multiple signal sources including horizontal and vertical components of flowpaths and changes in ground resistivity. In Water Track 6 (Figure 3.5), values increase from 4 mV to 14 mV (yellow-green to blue) down the length of the primary channel; this pattern is present but less obvious in Water Track 1. However, the SP signal does not increase monotonically in either data set. Instead local maxima are present, likely due to groundwater contributions from the adjacent non-track hillslope watershed into the primary channel. Since upward flow, as well as horizontal flow, can result in positive SP signals (Richards et al., 2010) these local maxima may also represent areas of local upwelling. This type of local maxima have
been observed in SP measurements in association with preferential flowpaths through earthen
dams (Bolève et al., 2009) and along faults (Richards et al., 2010). The inverse, local minima,
have been observed associated with pumping wells (Rizzo et al., 2004).

We interpret the troughs in voltage on either side of the water tracks as hydrologic
divides from which water is flowing away. This is supported by increasing voltage values in the
lower left and right corners of Figure 3.5 (Water Track 1 and 6), suggesting that flow at these
locations is moving toward the neighboring water track on the hillslope. These divides
correspond approximately to the highs in both the surface topography from LiDAR (thick dashed
lines in Figure 3.5), suggesting that at least on a broad scale subsurface flowpaths approximate
surface patterns. Given the inter-profile spacing (10 – 20 m), it is not possible to accurately
compare divides derived from ER data. However, the local maxima occur in two locations where
local upwelling might be expected: around the weir and in areas where the observed low
resistivity layer thins, forcing water through a thinner active layer. Around the weir, horizontal
flow is blocked by an impermeable barrier and therefore vertical water movement is expected.

An alternative source of voltage differences in SP data are changes in subsurface
resistivity; however, in this case, the ground resistivity decreases from the flanks to the primary
flowpath. This direction of increasing resistivity would dampen, rather than enhance, the strength
of the observed voltage differences (Revil and Jardani, 2013). Therefore, this suggests that the
distribution of voltages we observe are primarily due to groundwater flow rather than changes in
resistivity.
3.5 Conclusions

We compared measurements of active layer thaw beneath common arctic hillslope drainage features called water tracks using ER and depth-to-refusal frost probing. Frost probe and ER data compared well in the areas of moist acidic tundra on the hillslope outside the water tracks, but in the water tracks, zones of lower electrical resistivity extend deeper than the frost probe measurements. These areas below the main water track flowpath may represent partially frozen, saturated soil, an extension of the flowpath network that is not be identifiable with traditional methods. Future work should focus on using ER to better resolve the interannual active layer thaw response to variable energy and water inputs. In particular, ER could be a valuable tool for long-term mapping and prediction of the response of terrestrial flowpath networks in permafrost regions to climate change.

The SP data indicate that on a hillslope scale, flowpaths generally follow surface topography. However, on a smaller scale, the SP measurements do not increase monotonically as would be expected from purely lateral flow. SP voltages increase along and toward the primary channel, suggesting flow toward the channel and downhill; however, local maxima are present within the water track, which could result from local upwelling. Incorporation of this type of data could lead to more targeted sampling to identify potential hotspots for biogeochemical transformation. SP could also be used to investigate how water flow changes through time; together with ER, these data provide insight on how flowpaths and the frost boundary interrelate.
CHAPTER 4

PROPAGATION OF DIESEL TRANSPERSION SIGNALS IN THE SUBSURFACE

OBSERVED USING THE SELF-POTENTIAL METHOD

A paper to be submitted to Hydrological Processes

Emily Voytek*, Holly Barnard2, Damien Jougnor3, and Kamini Singha1+

4.1 Introduction

Quantifying the interconnection between evapotranspiration (ET) and groundwater is one of the most difficult challenges facing hydrologists today (National Research Council, 2012; Rodriguez-Itrube, 2000). The two-way link between hydrology and plant ecology is particularly complex in the vadose zone, where soil moisture availability controls vegetation distribution, and the vegetation affects the soil moisture distribution (e.g. D’Odorico et al., 2007; Moore et al., 2011; Swetnam et al., 2017; Tromp-van Meerveld and McDonnell, 2006). Vadose zone processes within hillslopes, such as plant water uptake, mediate groundwater discharge to streams (Asbjornsen et al., 2011). At the small-catchment scale, in temperate climates, vegetation transpiration has been implicated in producing diel fluctuations in streamflow, particularly during baseflow conditions when transpiration rates are high relative to total streamflow (e.g., Bond et al., 2002; Burt, 1979; Graham et al., 2013; Lundquist and Cayan, 2002). The two most dominant explanations for the transpiration-driven origin of these diel

*Primary researcher and author
1Hydrologic Science and Engineering Program, Colorado School of Mines, Golden, Colorado.
2Institute of Arctic and Alpine Research, Department of Geography, University of Colorado Boulder, Boulder, CO 80309, USA
3Sorbonne Universités, UPMC Univ Paris 06, CNRS, EPHE, 75005 Paris, France
fluctuations are: (1) uptake of water from lateral subsurface flowpaths linking hillslopes to streams (Bren, 1997), (2) removal of hyporheic water from stream-side aquifers (Bond et al., 2002), or a combination of both. The relative contribution and timing of these processes leading to diel streamflow generation, particularly under differing hydrologic regimes, remains poorly described with existing data (Bond et al., 2002; Graham et al., 2013; Newman et al., 2006; Voltz et al., 2013; Wondzell et al., 2010). In particular, the mechanisms which propagate signals from upslope processes, including transpiration, through hillslopes to streams remains unverified (Ali et al., 2011; McGuire and McDonnell, 2010). To better describe subsurface processes that link transpiration and streamflow and quantify flowpaths within catchments, development of new sensors capable of measuring in situ water movement beyond the point scale (i.e. integrative measurements) are needed.

Here, we explore the use of the self-potential (SP) method, a passive electrical geophysical tool that is sensitive to saturated and unsaturated water flow, to quantify subsurface processes occurring in response to plant transpiration. SP has been used to determine unsaturated flow rates in one dimension (Doussan et al., 2002; Thony et al., 1997) and to measure sapflow within a tree trunk (Gibert et al., 2006), but not previously used to measure vegetation-induced water movement in the vadose zone. The objective of this work is therefore to: 1) evaluate the effect of soil moisture on propagation of transpiration signals at the single-tree scale at the H.J. Andrews Experimental Forest (HJA), Oregon, USA and 2) capitalize on the sensitivity of SP to explore the propagation of diel transpiration signals into the unsaturated subsurface.
4.2 **Background**

4.2.1 **Vegetation-hydrology interactions**

At the small-catchment scale, vegetation has been implicated in producing diel fluctuations in streamflows since the 1930s (Blaney et al., 1933, 1930; Troxell, 1936; White, 1932). Many studies, reviewed in Hewlett and Hibbert (1967), Bosch and Hewlett (1982), and Gribovszki et al. (2010), suggest that diel patterns of transpiration are the prominent driver of streamflow fluctuations in forested systems. These processes have been tested through comparison studies of catchments with and without vegetation (e.g. Dunford and Fletcher, 1947; Rothacher, 1965) and time-series analysis of transpiration and resultant streamflow (e.g. Bond et al., 2002). However the mechanisms by which the transpiration signal are propagated from trees, through the hillslope, to streams are complex and difficult to evaluate.

Despite strong coupling observed between transpiration and streamflow time series, isotopic work has found that stream water and water used by near-stream plants are not from the same source (Brooks et al., 2010). Isotopic data suggest that during dry seasons, when diel fluctuations in streamflow are highest, trees and other vegetation are using water that is disconnected from the deeper groundwater discharging to the streams (Brooks et al., 2010; Dawson and Ehleringer, 1991; Evaristo et al., 2015). However, this hypothesis remains unproven in the light of other data (McCutcheon et al., 2017). Explaining the connection between diel transpiration and stream fluctuations while allowing for differing isotopic signals between plant xylem and stream water would solve a conundrum: if the water that vegetation and streams are sourcing is different water, how is the diel signal propagated from plant to stream in the subsurface?
Methods used to quantitatively determine which hillslope process(es) are occurring in the subsurface generally include catchment-averaged or point-scale measurements, with few support volumes in between. Catchment-averaged measurements include analysis of streamflow volume or stream-water chemistry. Point-scale measurements include soil moisture content and matric potential. However, due to sensor sensitivity, changes in soil moisture associated with plant-water use may be limited to periods when soil moisture is high and resulting plant uptake is large (Musters et al., 2000). Geophysical methods provide spatially integrative measurements, which can be useful in evaluating hillslope scale processes related to plant-water use and the impact on subsurface water flow.

4.2.2 Geophysics to Map Ecohydrologic Processes

Soil moisture content affects the bulk electrical resistivity (or its reciprocal, bulk electrical conductivity) of a soil; consequently, methods such as electrical resistivity (ER) imaging and electrical magnetic imaging (EMI) can be used to observe changes in soil moisture, controlling for temperature and salinity changes. Numerous examples exist of mapping changing soil moisture in response to infiltration with ER (e.g. Binley et al., 2002; Daily et al., 1992; French and Binley, 2004; Nimmo et al., 2009; Travelletti et al., 2012). ER has also been used to image ecohydrologic processes such as root-water uptake (e.g. Jayawickreme et al., 2010, 2008), hydraulic redistribution (e.g. Robinson et al., 2012) and tree-water dynamics within trunks (al Hagrey, 2006; Guyot et al., 2013; Mares et al., 2016). EMI has been used in soil studies to map soil type and water content (e.g. Doolittle and Brevik, 2014). However, neither ER nor EMI is capable of directly measuring water flow.
Self-potential (SP) is a passive geophysical method that is sensitive to water flow in the vadose or saturated zone. Unlike other geophysical methods that are sensitive to static or state variables (e.g. water content, lithology), SP is sensitive to dynamic processes (e.g. water flow, ionic fluxes, electron transfer). For this reason, use of SP in hydrologic studies has increased in recent years (e.g. Darnet and Marquis, 2004; Jougnot et al., 2015; Linde et al., 2011; Maineult et al., 2008). Thony et al. (1997) first used SP to monitor movement of newly infiltrated water within a soil column from a single precipitation event, and observed a linear trend (defined by a coupling coefficient) between unsaturated flow rates and measured SP signal. Doussan et al. (2002) built on this work by comparing measurements of SP and infiltration for multiple precipitation events in two soil types. They also observed changes in SP measurements related to rainwater infiltration, but found that a single coupling coefficient could not be used to relate the two processes as was done by Thony et al. (1997); properties of the soil affected the magnitude of SP measurements and needed to be included in the coupling coefficient. Without a better description of the sources and magnitude of the variation in the coupling coefficient, SP could not be used to quantitatively determine water flow patterns.

By including a saturation-dependent coefficient when modeling SP responses to unsaturated flow, rather than purely linear response, Darnet and Marquis (2004) better fit the observed data of Thony et al. (1997) and Doussan et al. (2002). Through a series of modeling exercises using the saturation-dependent coefficient, they showed that SP is capable of effectively estimating water flux in the vertical direction on the scales of decimeters (resolution is electrode-spacing dependent). Note that in ecohydrology, where the overall quantity of water in the soil is important, soils are often described by their soil moisture, or volumetric water content (VWC). However, in discussion of the physics of SP, including in the following
background sections, the percentage of pore space that is filled is more pertinent, so saturation, \( S_w \), is used. Saturation is related to VWC [-] by porosity, \( n \): 

\[
\text{VWC} = S_w n .
\]  

(4.1)

The theorized saturation dependence of the signal strength of SP has since been confirmed and expanded on in many additional theoretical and laboratory studies (Guichet et al., 2003; Revil and Cerepi, 2004; Vinogradov and Jackson, 2011). In this work, we build on the existing use of SP to evaluate soil moisture movement in response to infiltration and evaporation (e.g. Sailhac et al., 2004) to include movement associated with tree root-water uptake.

4.2.3 Self-potential Background

SP relies on measurements of electrical potential differences (in mV) generated by natural currents in the ground. The amplitude of measured SP voltages depend on the magnitude of generated currents and the electrical conductivity of the ground material as described by a generalized version of Ohm’s law in the framework proposed by Sill (1983):

\[
J = \sigma E + J_s
\]  

(4.2)

where \( J \) is the macroscopic current density [A m\(^{-2}\)], \( \sigma \) is the electrical conductivity of the ground [S m\(^{-1}\)], \( E \) is the electrical field [V m\(^{-1}\)] and \( J_s \) is the source current density [A m\(^{-2}\)].

The first right-hand side term, \( \sigma E \), describes the conduction current density, or how electrical signals propagate through the material, while \( J_s \) describes the distribution of currents generated by water movement or other possible current sources, discussed below. The electrical field is further defined by:
\[ E = -\nabla V \]  \hspace{1cm} (4.3)

where \( V \) is the electrical potential [V], measured in the SP method. The SP signal is the electrical potential difference between a reference and a potential electrode:

\[ \text{SP}_i = V_i - V_{\text{ref}}. \]  \hspace{1cm} (4.4)

To solve Equation 4.2 for the SP voltages, the above constitutive equations describing current density must be combined with a charge conservation equation, which describes the current quantity. At the quasi-static limit of Maxwell’s equations, the conservation of charge is described by:

\[ \nabla \cdot \mathbf{j} = 0. \]  \hspace{1cm} (4.5)

Combining Equations 4.2, 4.4 and 4.5 produces the field equation describing the complete electrical problem:

\[ \nabla \cdot (\sigma \nabla V) = \nabla \cdot \mathbf{j}_S , \]  \hspace{1cm} (4.6)

Multiple processes, including ground- or soil-water movement, thermal and chemical diffusion (e.g. Leinov and Jackson, 2014) and under specific conditions, redox gradients (Hubbard et al., 2011), can generate electrical currents and contribute to \( \mathbf{j}_S \) (Revil and Jardani, 2013).

To make meaningful SP measurements, a few corrections must be made to the data. For example, SP measurements are affected by temperature variations between the reference and measuring electrode. Despite improved temperature stability of Petiau-type electrodes (Petiau, 2000) over earlier electrodes, measurements must be corrected for temperature differences between the two electrodes:

\[ \text{SP}_{i}^T = \alpha (T_i - T_{\text{ref}}) \]  \hspace{1cm} (4.7)
where $SP^T_i$ is the temperature correction to be applied to the measurement electrode to the reference electrode, $\alpha$ is the electrode-type correction factor (0.2 mV/°C for Petiau-type; Petiau, 2000), and $T_i$ and $T_{ref}$ are temperatures at the measurement and reference electrodes respectively. SP measurements are also sensitive to temporally variable electrode drift caused by electrode age and changing chemistry in the immediate vicinity of the electrode (Jougnot and Linde, 2013). In surface SP surveys, this electrode drift is corrected by measuring, through time, the voltage difference between the two electrodes by placing them in direct contact with each other. In long-term measurements where the electrodes are buried into the ground, this type of correction is not possible, but assumed to be insignificant in this work based on manufacturer’s data (0.2 mV/month, SDEC, Reignac sur Indre, France).

The primary signal of interest in this work results from the current generated by the movement of water in the vadose zone (i.e. the electrokinetic phenomenon). Water movement produces electrical current due the existence of the electrical double layer (EDL) at the pore water-mineral interface. The EDL develops when a mineral surface is in contact with water, which alters the surface charge of the mineral and the surrounding water chemistry. For example, under near-neutral pH conditions, 5-8, the surface charge of a silica grain is negative (Revil and Jardani, 2013). The charged surface attracts ions of the opposing charge (positive in the case of silica) from the bulk pore water, the so-called counter-ions. These counter ions sorb directly onto the mineral surface, forming the less-mobile Stern layer on the surface of the grain, while additional counter-ions exist in the more-mobile diffuse layer (Figure 4.1). The diffuse layer exists as ions are simultaneously attracted to the excess surface charge of the mineral and repelled by the enriched concentration of like charges (Revil and Jardani, 2013). These two layers are separated by the shear plane, separating fixed water molecules and flowing ones.
Under saturated conditions, the net charge of the EDL, which comprises both the Stern layer and diffuse layer, is positive relative to the surrounding pore water and is called the total excess charge density, \( Q_v \) [C m\(^{-3}\)]. \( Q_v \) is dependent on soil properties (mineralogy, pH, and pore-water chemistry) and is related to the cation-exchange capacity (CEC, [meq L\(^{-1}\)]) through:

\[
Q_v = \rho_g \left( \frac{1-n}{n} \right) \text{CEC}
\]

(4.8)

where \( \rho_g \) is the grain density [kg m\(^{-3}\)] and \( n \) is porosity [-]. (Waxman and Smits, 1968). CEC measures all of the excess charges in the EDL, mobile and immobile; however, current is only generated by moving charges. Consequently, only the excess charge in the mobile, diffuse layer in the EDL contributes to generation of SP signals. This value, the effective excess charge density, \( \tilde{Q}_v \) [C m\(^{-3}\)], can be 3-4 orders of magnitude smaller than total excess charge (Jougnot et al., 2012; Leroy and Revil, 2009). Determining \( \tilde{Q}_v \) from \( Q_v \) is difficult; therefore, constitutive
relationships between \( \hat{Q}_v \) and other measurable parameters have been developed. For example, Jardani et al. (2007) presents an empirical relationship between \( \hat{Q}_v \) and permeability, \( k \ [m^2] \) defined for saturated conditions:

\[
\log(\hat{Q}_v) = -9.2349 - 0.8219 \log(k). \tag{4.9}
\]

In unsaturated conditions, \( \hat{Q}_v \) is dependent on saturation. Waxman and Smits (1968) observed a relationship of increasing \( Q_v \) as water was replaced by oil in a two-phase system. To account for the effect of saturation on electrokinetic coupling, Linde et al. (2007) and Revil et al. (2007) proposed a theoretical framework using a volume-averaging upscaling procedure. This upscaling procedure proposes describes the effect of water saturation on the corresponding effective excess charge density:

\[
\hat{Q}_v(S_w) = \frac{\hat{Q}_{v,\text{sat}}}{S_w} \tag{4.10}
\]

where \( \hat{Q}_v(S_w) \ [C \ m^{-3}] \) is the effective excess charge as a function of saturation. As the saturation in the medium decreases while surface charges remain constant, the excess charge density increases. Although the volume-averaging approach of Linde et al. (2007) provides a first-order approximation \( \hat{Q}_v \) in homogeneous media at high saturation (e.g. clean sand in Linde et al. (2007); Mboh et al. (2012); and Jougnot and Linde, (2013)), it has been shown that it could not reproduce SP signal amplitudes at lower saturation in more complex soils (Jougnot et al., 2012; 2015; Allègre et al. 2014). To better fit soils at low saturations, Jougnot et al. (2012) propose an upscaling approach that they labeled “flux-averaging” (i.e., an averaging procedure based to the water flux distribution in the medium). Here, the soil is treated as a bundle of capillaries that were either fully saturated or dry depending on their capillary radius, \( R \), and the applied matric
potential. For each capillary radius, \( R \), an equivalent effective excess charge density \( \hat{Q}^R_v(R) \) [C m\(^{-3}\)] is computed from the distribution of excess charge and pore water velocity in the capillary. This parameter is then correlated to an equivalent pore-size distribution to obtain an effective excess charge function with respect to the saturation at the REV scale:

\[
\hat{Q}_v(S_w) = \frac{\int_{R_{min}}^{R_{max}} \hat{Q}^R_v(R)v_R f_D(R)dR}{\int_{R_{min}}^{R_{max}} v_R f_D(R)dR}
\]  

(4.11)

where \( f_D[-] \) is the equivalent pore-size distribution function inferred from hydrodynamic parameters (i.e. van Genuchten parameters, permeability \( k \)) and \( v_R [m \, s^{-1}] \) is the mean pore water velocity for a given capillary radius. This approach is more complex than Equation 4.10 but better describes SP amplitudes in natural media over a large range of saturations (Jougnot et al., 2012; 2015). Given the large variation in saturation observed during the period of data collection in this work (Figures 4.3c and 4.4c), we use the Jougnot et al. (2012) flux-averaging approach in this work (Figure 4.5).

Once \( \hat{Q}_v(S_w) \) is defined, the streaming current term of water movement in variable saturated conditions is:

\[
J_{\text{streaming}} = \hat{Q}_v(S_w)U
\]  

(4.12)

where \( U \) is the soil moisture velocity [m s\(^{-1}\)]. Soil moisture velocity can be solved according to Richards equation:

\[
-\nabla U = \nabla \cdot [K(S_w)\nabla h]
\]  

(4.13)
Figure 4.2 Cross-sectional view of sensor array relative to selected tree. Four SP electrodes were installed in a grid at 0.3 and 0.8 m depth, and 0.1 m and 0.9 m away from the tree. Two tensiometers, two soil moisture sensors and two thermocouples were installed 0.3 m downslope from the lower SP electrodes.
Figure 4.3 Measured monthly field data from sensors in Figure 4.2. a) Tree transpiration calculated from measured sap flux within the tree and precipitation recorded at the HJA PriMet station, where July 9-14\textsuperscript{th} sap flow data were infilled from measurements on Douglas fir trees due to equipment failure, b) ground temperature, c) soil moisture, d) matric potential and e) measured SP voltage differences between subsurface electrodes.
Figure 4.4 Measured daily field data from sensors shown in Figure 4.2. a) Tree transpiration calculated from measured sap flux within the tree and precipitation recorded at the HJA PriMet station, b) ground temperature, c) soil moisture, d) matric potential and e) measured SP voltage differences between subsurface electrodes. Mid-day day removed due to noise added by solar charger.
where $K(S_w)$ is saturation-dependent hydraulic conductivity [m/s], and $h$ is the total matric head [m].

In addition to the saturation dependence of $\bar{Q}_v$, ground electrical conductivity, $\sigma$ is a saturation-dependent parameter that increases with added soil moisture. Under certain conditions this relationship should obey Archie’s law (Archie, 1942), but this equation does not account for surface conductivity present in many natural soils. Instead, a modification of Archie’s law obtained through volume averaging and that includes a surface conductivity term not scaled by saturation as in the Waxman and Smits model (1968) (Linde et al., 2006; based on Pride, 1994), has been found to best predict electrical conductivity of loamy soils (Laloy et al., 2011):

$$\sigma_{sw} = \frac{1}{F} [S_w^{b} \sigma_f + (F - 1) \sigma_s]$$  \hspace{1cm} (4.14)
where $\sigma_{sw}$ is the effective conductivity with varying saturation [S m$^{-1}$]; $b$ is Archie’s second, or saturation, exponent [-] (e.g. Waxman and Smits, 1968); $F$ is the formation factor [-]; $\sigma_f$ is the fluid conductivity [Sm$^{-1}$]; and $\sigma_s$ is the surface conductivity [Sm$^{-1}$].

In variably saturated conditions, Equations 4.6 and 4.12 can be expanded to:

$$\nabla \cdot (\sigma_{sw} \nabla V) = \nabla \cdot (\hat{Q}_{v,sw} \mathbf{U}).$$  \hspace{2cm} (4.15)

Using this equation, we can use measured SP values to evaluate relative unsaturated flow rates through time and therefore examine the role of trees in subsurface water redistribution.

### 4.3 Field Site

Our work is conducted in Watershed 10 (WS10), a 0.1-km$^2$ watershed of the H.J. Andrews (HJA) Experimental Forest in the western Cascade Mountains. The steep catchment (27-48° slopes) ranges in elevation from 480 m to 565 m (McGuire et al., 2007) and is underlain by highly weathered andesitic tuffs and coarse breccias. The soils are residual and colluvial deposits (mesic Andic Humudepts; Soil Survey Staff, accessed 04/15/2017) with an average depth of 1.3 m. Soil textures in the upper meter are gravelly, silty clay loams to very gravelly clay loams, with slightly blockier textures below 0.70 m (Harr, 1977). Mean porosity for the soils, measured from extensive soil sampling, ranges from 60-64% (Harr, 1977). The soils cover weathered saprolite that can be up to 7 m thick (3.6 m average) (Harr and McCorison, 1979). The mean precipitation at the HJA is 2220 mm, with approximately 80% falling between October and April (McGuire et al., 2007). As part of early studies on the effects of logging, the catchment was clear cut in 1975 to evaluate the effects of local forestry practices on the catchment (Harr,
Vegetation is now dominated by second-growth (~40 year old) Douglas fir trees. We selected one of these second-growth Douglas fir trees to focus our SP and corroboratory measurements on; selection of the tree was based on ease of access and proximity to other experiments.

### 4.4 Methods

SP data were collected in a two-dimensional subsurface array at the base of a Douglas fir tree between June and November 2016. It was not possible to exclude the effects of other vegetation from the measurements as root zones of Douglas fir are known to extend 3 - 4 times maximum canopy thickness (e.g. Eis, 1987) and smaller vegetation, including western swordfern (Polystichum munitum), deer fern (Blechnum spicant), and Oregon grape (Berberis nervosa) are present on the hillslope. While isolating the effects of a single tree in this in the densely vegetated catchment is impossible, we selected a relatively isolated Douglas fir tree (23.3 cm diameter at breast height, dbh). The nearest neighboring trees were two other Douglas fir trees located 3.2 m downslope (30.7 cm dbh) and 4.3 m upslope (23 cm dbh). The stream draining the catchment was approximately 11 m downslope from the selected tree. In addition to data collected specifically for this project, precipitation data were used from the PriMet station run by HJA.
4.4.1 Self-potential and corroboratory measurements

The SP array consisted of four Petiau-type non-polarizing electrodes (Petiau, 2000) on the downslope side of the tree. The array includes two electrodes at 0.3 m and two electrodes at 0.8 m depth (measured from the porous tip of the electrode, Figure 4.2). This configuration provided for measurements of SP in both the vertical and horizontal directions. Electrodes were installed by hand augering to the desired depths, emplacing the electrodes, and refilling the hole with the native soil. Prior to installation, the electrodes were checked using a handheld Fluke 87V voltmeter (accuracy 0.1 mV) to ensure the tip-to-tip differential voltage between all pairs was <1 mV. Tip-to-tip differences greater than 1 mV can indicate electrode failure such as poor internal wire connections, or drying out of the internal solution. To ensure good contact between the electrodes and the ground and to reduce the drying out of the electrodes, a small amount of bentonite (~0.5 L) was added into the base of hand-augered holes prior to emplacement of the electrodes. The electrodes were left in place for the duration of data collection, reducing position uncertainty. Voltages were recorded between four pairs of electrodes (Figure 4.6) with a resolution of 0.67 mV using a Campbell Scientific CR1000 logger at 15-min intervals. The data were then corrected for the known temperature drift of 0.2 mV/°C in the Petiau-type electrodes (Petiau, 2000; Equation 4.7) based on ground temperature measurements, described below.

For data correction and comparison purposes, soil moisture content, temperature and matric potential were collected at 15-min intervals in the immediate vicinity of the SP array, and recorded on the same Campbell Scientific CR1000 used for SP data collection. Soil moisture content was measured at 0.3 m and 0.8 m depths using two Decagon EC-5 sensors (accuracy: 0.03 [m³ m⁻³]) and an assumed porosity of 60% (Harr, 1977). No soil-specific calibrations were used for the EC-5 sensors because only relative changes in soil moisture were needed and soil
Figure 4.6 Conceptual model used to investigate SP signals associated with root-water uptake and description of hydrologic and electrical boundary conditions. Arrows indicate dipole orientation: reference electrode is located at the flat end and measurement electrode at arrowhead.
properties at the depths investigated were considered relatively homogenous based on previous soil surveys at the site (Harr, 1977). Soil temperature was recorded using three Type T thermocouples (accuracy: ±1 °C). Soil matric potentials were measured at two locations, 0.3 m and 0.8 m, using ecoTech Tensiomark sensors (ecoTech, Bonn, Germany). These heat-pulse sensors have an accuracy of 5% pF, where pF = log10(hPa), resulting in decreased accuracy at higher matric potentials. The matric potential sensors are independent of soil texture because they measure heat dissipation in the incorporated ceramic tip; consequently, we did not use a site-specific soil calibration. The relative positions of all sensors were determined using a Leica total station.

Sapflow measurements were collected on the focal tree nearest the SP array for comparison with the measured SP data. Heat-pulse sapflow sensors (Burgess et al., 2001) were built and installed at 1.8 and 3.5 cm depth into the tree and measured at 30-min resolution. Sapflow velocity was converted to tree transpiration rate using the methods of Dawson et al. (2007) and Hu et al. (2010).

4.4.2 Coupled Soil Water Flow and Electrical Modeling

A 2-D coupled fluid flow and electrical model was used to test the impact of root uptake on expected SP data. The radially symmetric model was built and solved in COMSOL 5.2a. The model simulates a single tree in the middle of homogenous ground material, extending beyond the zone of influence of the tree (Figure 4.6). The fluid flow and electrical models were run sequentially, and the details of each are discussed in that order below. The model was run with 30-min time steps to coincide with the interval of the sap flow data. Through evaluation of modeled SP signals, including models with and without root-water uptake, we can test the
Figure 4.7 a) Measured monthly tree transpiration and precipitation used as boundary conditions in the model, b) modeled Darcy velocity, c) soil moisture, d) matric head, and e) SP voltages between subsurface electrodes. (w/o = without root-water uptake).
Figure 4.8 a) Measured daily tree transpiration and precipitation used as boundary conditions in the model, b) modeled Darcy velocity, c) soil moisture, d) matric head, and e) SP voltages between subsurface electrodes. (w/o = without root-water uptake).
hypothesized transpiration-driven origin of the diel fluctuations in the field SP data.

Additionally, unlike field data, in which the velocity of water movement in the subsurface is unknown, in the coupled model we can evaluate changes in flow velocities (Figures 4.7b and 4.8b) along with the SP signals that they generate (Figure 4.7e and 4.8e).

Unsaturated water flow patterns were solved according to Richards equation (Equation 4.13). The boundary conditions are based on observed field conditions: the basal boundary is a constant head, based on observed groundwater levels (3.5 m below ground surface); the ground surface is a prescribed flux, linked to the measured precipitation at the field site (range: 0 – 54.1 mm/day (Figures 4.7a and 4.8a) and the outside edge is a no-flow boundary, where the tree behavior is assumed to have no influence. The porosity, hydraulic conductivity, and soil moisture retention parameters are based on measured values from Harr (1977) and Carsel and Parrish (1988), and are shown in Table 4.1.

Root-water uptake was implemented as a distributed source/sink in the center of the domain following the radially symmetric model of Vrugt et al. (2001) and made up of two terms. The first, $\beta$, describes the geometric distribution of root-water uptake under unstressed conditions (Raats, 1974):

$$
\beta(r,z) = \left[ \left( 1 - \frac{z}{z_m} \right) \right] \left[ \left( 1 - \frac{r}{r_m} \right) \right] e^{-\frac{p_z|z^* - z| + p_r|r^* - r|}{r_m}}
$$

(4.16)

where $r [\text{m}]$ and $z [\text{m}]$ are the position relative to the base of the tree, $r_m [\text{m}]$ and $z_m [\text{m}]$ are the maximum radial rooting length and rooting depth, and $p_z [-]$, $z^* [\text{m}]$, $p_r [-]$, and $r^*[\text{m}]$ are empirical parameters describing the distribution of uptake.

To account for changing moisture conditions, $\beta$ is combined with a soil-stress equation, defined by $\gamma(r,z,h) [-]$, from van Genuchten (1987):
Table 4.1 Parameters used in coupled fluid flow and electrical models.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Value</th>
<th>Unit</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Hydrodynamic Parameters</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$K_s$</td>
<td>8.30E-06</td>
<td>[m/s]</td>
<td>Saturated hydraulic conductivity</td>
</tr>
<tr>
<td>$\theta_r$</td>
<td>0.089</td>
<td>[-/-]</td>
<td>Residual moisture content</td>
</tr>
<tr>
<td>$n$</td>
<td>0.43</td>
<td>[-/-]</td>
<td>Porosity</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>0.93</td>
<td>[1/m]</td>
<td>van Genuchten alpha parameter</td>
</tr>
<tr>
<td>$N$</td>
<td>1.58</td>
<td>[-]</td>
<td>van Genuchten N parameter</td>
</tr>
<tr>
<td><strong>Vrugt et al., Root-water uptake model parameters</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$z_m$</td>
<td>1</td>
<td>[m]</td>
<td>Maximum rooting depth in soil profile</td>
</tr>
<tr>
<td>$r_m$</td>
<td>5</td>
<td>[m]</td>
<td>Maximum rooting radius in soil profile</td>
</tr>
<tr>
<td>$z^*$</td>
<td>0</td>
<td>[-]</td>
<td>Empirical parameter</td>
</tr>
<tr>
<td>$r^*$</td>
<td>0</td>
<td>[-]</td>
<td>Empirical parameter</td>
</tr>
<tr>
<td>$p_z$</td>
<td>0</td>
<td>[-]</td>
<td>Empirical parameter</td>
</tr>
<tr>
<td>$p_r$</td>
<td>0</td>
<td>[-]</td>
<td>Empirical parameter</td>
</tr>
<tr>
<td>$h_{50}$</td>
<td>-204</td>
<td>[m]</td>
<td>Soil water pressure at which root water uptake reduced by 50%</td>
</tr>
<tr>
<td><strong>Electrical Parameters</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\sigma_f$</td>
<td>0.0022</td>
<td>[S/m]</td>
<td>Fluid electrical conductivity</td>
</tr>
<tr>
<td>$\sigma_s$</td>
<td>0.0006</td>
<td>[S/m]</td>
<td>Surface conductivity</td>
</tr>
<tr>
<td>$\hat{Q}_e$</td>
<td>See Figure 6</td>
<td>[C/m³]</td>
<td>Effective excess charge, after Jougnot et al. (2012)</td>
</tr>
</tbody>
</table>
\[ \gamma(r, z, h, t) = \frac{1}{1 + \left( \frac{h(r, z, t)}{h_{50}} \right)^{1/50}} \]  

(4.17)

where \( h \) is the total matric head [m] at a location \((r, z)\), \( h_{50} \) [m] is the soil-water pressure head at which root-water uptake is reduced by 50% (defined by mean cavitation pressure \([P]\) in tree physiology literature; Table 4.1). The values of each of the parameters used are shown in Table 4.1. For our model, an experimentally derived \( h_{50} \) value of -204 m (-2 MPa; Sperry and Ikeda, 1997) was used. A maximum rooting depth of 1 m was also used, corresponding to the thickness of soil at the site and based on studies such as Curt et al. (2001), who found little Douglas fir root biomass in the substratum below soil, including in areas of fractured substratum, similar to the system at HJA.

The localized source/sink applied to our axially symmetric model can be determined normalizing the product of the two terms, \( \beta \) and \( \gamma \):

\[
Q(r, z, h, t) = \frac{\pi r_m^2 \beta(r, z) \gamma(r, z, h, t)}{2\pi \int_0^{r_m} \int_0^{z_m} \beta(r, z) \gamma(r, z, h, t) \, dr \, dz} T_{\text{sap}}
\]

(4.18)

where \( T_{\text{sap}} \) is total transpiration \([m^3 \, s^{-1}]\) calculated from sapflow measurements (see Section 4.2), and the denominator is the volume integral of \( \beta \gamma \).

Once the fluid flow problem is solved at each time step, the electrical problem is solved using Ohm’s law (Equation 4.4). The model has insulation conditions at all boundaries, combined with an external current density linked to calculated flow velocities from the fluid flow model (Equation 4.8). In the electrical simulation, both the excess charge density, \( \tilde{Q}_v \), and bulk conductivity are saturation dependent as described in Equations (4.7) and (4.9) (Table 4.1 and Figure 4.5). Only voltage differences, rather than absolute measurements, will be analyzed; placement of the reference \((V = 0 \, V)\) consequently does not matter. From the resulting voltage
distribution, voltage differences comparable to the electrode placement in the field are extracted (Figure 4.3).

4.5 Results and Discussion

4.5.1 Precipitation

During 2016, HJA received a total of 2091 mm of precipitation, of which 474 mm fell during the period analyzed in this work, July 1 to November 8 (Figures 4.3a and 4.4a). This rainfall is comparable with the mean annual precipitation for the site, 2200 mm (McGuire et al., 2007). A total of 88.1 mm fell during the dry period between July 1 and September 29. During this period, most precipitation occurred during three discrete periods (July 6 - 9, August 1 - 6, and August 16 - 22). In contrast to the relatively dry conditions during the summer, 388 mm of precipitation occurred between September 30 and November 8.

4.5.2 Tree Transpiration

Daily total tree transpiration rates were greatest in July and decreased throughout the summer (Figure 4.3a). Diel fluctuations reached a maximum during midday, and return to near-zero flow overnight (Figure 4.4a). Negligible transpiration occurs on July 8th and July 10th during precipitation events due to cloudiness. Due to technical malfunctions, sapflow measurements from the selected Douglas fir are missing for July 9 - 14, 2016. To drive the coupled model, the data gap was filled using average sapflow rates from eight nearby (within ~100m) sapflow sensors in other Douglas fir trees, which had similar patterns outside of this data gap to the sensor in our tree, and were normalized to the sap flow measured at our tree. Transpiration was
assumed to be negligible after the sensors were removed on November 4, 2016 due to persistent cloud cover and low vapor pressure deficit.

4.5.3 Soil Moisture

As with precipitation, the soil moisture records can also be divided into a dry and wet period (Figure 4.3c). During the dry period from July through September, when precipitation was low, the soil moisture at 0.3 m and 0.8 m steadily decreased. Soil moisture at 0.8 m dropped from 13% at the beginning of July to 9% percent at the end of September. During the dry period, the soil moisture at the 0.3-m sensor was consistently ~3% lower than the soil moisture at 0.8 m. Slight increases in soil moisture were visible at the 0.3-m sensor following discrete precipitation events in early July, suggesting water from these precipitation events infiltrated to at least 0.3 m. Increases in soil moisture at 0.8 m following the same precipitation events are much smaller or not detectable, suggesting that the pulse of soil moisture from precipitation is either evaporated or transpired prior to reaching the greater depths.

The soil moisture at 0.3 m abruptly increased from 6% to 17% on October 5 after three days of rain, marking the transition from the dry to wet period. Comparable increases in soil moisture at 0.8 m were observed nine days later on October 14. From October 14 until November 8, soil moisture at both depths remains above 17%. During this wet period, soil moisture at both depths increased in response to precipitation events, with a distinct peak followed by a gradual decrease in soil moisture. The greatest variability occurred at 0.3 m, with more muted responses at depth as the infiltration front diffused. A maximum soil moisture content of 22% occurs at 0.3 m on October 20.
No diel variations in soil moisture are visible in the soil moisture records. Lack of diel variations in data could be due to low sensor sensitivity (accuracy: 0.03 \([m^3 m^{-3}]\)). Barnard et al. (2010) previously observed diel fluctuations of soil moisture of up to 10% within WS10, but fluctuations were greatest at 0.0 - 0.3 m depth (Barnard et al. 2010, Figure 3), which is shallower than the sensors used in this work. Diel fluctuations at greater depth in this previous study were more muted, as observed here.

4.5.4 Matric Potential

The transition from wet to dry periods was also present in the matric potential data as a rapid decrease in pressure following the onset of storm events (Figure 4.3d). Matric potential at 0.3 m depth shows the greatest diel variations. Maximum matric potential occurs coincident with the maximum of sap flow. Our intention was to use data from these tensiometers to calculate hydraulic gradients and compare with flow patterns interpreted from the measured SP data. However, these tensiometers, TensioMark (EcoTech, Bonn, Germany), like many others, could not measure accurate matric potentials at the field conditions encountered in this work. The manufacturer-reported accuracy of these sensors scales as 5% of the measured value (i.e. logarithm of pressure), with a minimum value of ±3000 Pa or 0.3 m of pressure head. Consequently, as matric potential increases, there is increasing error as the soil dries out. At the maximum pressure recorded in our study, 1.3 MPa, this error is equivalent to greater than 1 m of pressure head and greater than our sensor spacing. This error is typical for this type ceramic tip heat dissipation matric potential sensor (Phene et al., 1971), and likely the reason that hydraulic gradients calculated from tensiometers at this spacing are rarely reported in the literature.
4.5.5 Field SP Data

As noted earlier, multiple processes, including ground- or soil-water movement, thermal and chemical diffusion, and redox gradients can generate electrical currents and contribute to measured SP signals. The low ionic concentrations of precipitation and stream water at the field site are not expected to generate the large concentration gradients necessary for an electrochemical source current. Also, conditions necessary for redox gradients to generate SP potentials, namely an electronic conductor (i.e. a metallic body or conductive biofilms; Hubbard et al., 2011), are not met at this site, so redox can be excluded as a possible signal source. We additionally have corrected for known temperature-induced voltage differences using Equation 4.7. Therefore, the most likely source of the remnant SP signal at our site is water movement.

Rather than use a single reference electrode in the interpretation of this SP data, we evaluated the voltage difference between four pairs of electrodes to evaluate flow at specific depths and distances around the tree to better look at flow patterns associated with transpiration. These pairs will be referred to as inner (0.1 m from the tree), outer (0.9 m from the tree), upper (0.3 m below the ground surface) and lower (0.8 m below the ground surface) pair corresponding to their positions relative to the tree and ground surface (Figure 4.6). For the vertical electrode pairs, inner and outer, the lower electrode is taken to be the reference such that positive values indicate upward movement. For the horizontal pairs, upper and lower, the reference is the outer electrode such that positive values indicate inward movement (towards tree axis; Figure 4.6).
4.5.5.1 Seasonal variability in SP

Temperature-corrected SP data plotted through time for the four data pairs reveal two distinct periods, distinguished primarily by signal amplitude, which increases in early October coinciding with the onset of the wet period as recorded by precipitation, saturation and matric potential. During the dry period, measured SP amplitudes are smaller (< ±10 mV) than during the wet period (> ±40 mV). This increase in amplitude during the wet period is due to increases in water movement, as more water moving through the system moves more charges.

The voltage difference between the upper pair is consistently positive, indicating inward (towards the tree) flow at 0.3 m soil depth, while the outer pair at 0.9 m distance is consistently negative, suggesting continuous downward flow at 0.9 m from the axis of the tree. As noted above, the voltages measured between the inner and lower pairs are smaller in magnitude during the dry season than both the upper and outer pairs; additionally, the voltage differences change sign in these two datasets during the dry season, suggesting a possible change in flow direction. The voltage difference of the lower pair starts positive, but becomes increasingly negative throughout the dry period, suggesting that water is moving towards the tree during the early part of the dry season, but gradually transitions to downward flow as the conditions dry out. This may be indication of disconnection between transpiration and stream flow theorized by Bond et al. (2002), which occurs as soil moisture decreases during rain free periods. Once the wet season begins, downward flow persists, but increases in magnitude, as expected with increasing infiltration due to precipitation.

In addition to the large changes in SP amplitude associated with transition to the wet season, the SP signal responds to precipitation events during the dry season. Increases in voltage magnitude can be generated by faster movement of the existing quantity of water, or increasing
the quantity of water moving (i.e. increased saturation). Precipitation events increase the saturation (Figure 4.3c), resulting in increased voltage differences. In SP data, this appears as a marked increase in voltage, followed by steady decline back to the baseline voltage as the pulse of water diffuses into the subsurface. This signal has been observed in association with precipitation events by Doussan et al. (2002), and modeled by Darnet and Marquis (2004). This signal is visible in our data as an increase that begins on July 17 and September 22, a few days after precipitation events on July 8-10 and September 17, 18 and 20. This delay is indicative of the time it takes for the precipitation signal to propagate to the depth of the electrodes.

In contrast to the voltage increases followed by more gradual decreases that occur 3-7 days after the precipitation events, a rapid voltage increase occurs on July 10th, during the precipitation event. The increase is most evident in the inner, upper, and lower pairs, but less evident in the outer pair. This sudden increase could be an indication of preferential flow into the subsurface, unlike diffuse matrix flow, which likely takes days to propagate to the depth of the electrodes. SP signals generated specifically by preferential flow has not been previously studied, but this interpretation is supported by increases in moisture content at depth, including at 0.8 m, on the same day as the precipitation event (Figure 4.3c). A similar jump occurs on October 5th at the start of the rainy season. However, not all observed precipitation events produce sudden jumps in the SP data. A discrete precipitation event (4 mm in 30 min) on August 17th resulted in rapid voltage decreases (8 mV in 12 hours in the outer pair, Figure 4.3e). Polarity changes in SP data have been observed in association with the onset of pumping tests (Malama, 2014). Precipitation events, which initiate quick changes in water content, may have similar effect.

Finally, a jump of 53 mV occurs on October 29 in only the inner and lower electrode pairs, with no observable change in the outer and upper pairs (Figure 4.3e). The features of this
jump, including a change of exactly the same voltage in two pairs sharing a common electrode (Figure 4.6) and no changes in the any of the surrounding pairs, suggest that it is caused due to an electrode malfunction. The differing polarity of the jump—one positive and one negative—is due to the relative orientation of the shared electrode in the pair (Figure 4.6). One possibility is oxidation along the wire connecting the electrode to the data logger, perhaps even at the wire tip inserted in the data logger, or poor electrode-soil contact, though this typically occurs under dry conditions (e.g., Doussan et al., 2002), rather than during wet conditions as observed here. With the jump removed (Figure 4.3e), the recorded voltages trend similar to the upper and outer pairs.

4.5.5.2 Diel fluctuations in SP

Diel fluctuations in voltage are visible from the start of the SP record in July, with a maximum occurring around 12:00 h (Figure 4.4). They are most evident in the outer pair, where diel fluctuations of ~1 mV are present in July. The true amplitude of the diel signals may be larger than these values as some mid-day data were removed due to unexpected electrical noise generated by an attached solar panel. The amplitude of diel signal increases through the summer and fluctuations of up to 2 mV are visible in late August. As the strength of transpiration decreases through September and August, the diel signal decreases. During precipitation events of October, the signal from increased unsaturated flow is greater than any observed diel transpiration signals. Fluctuations in both horizontal pairs are smaller than the fluctuations observed in the vertical pairs, implying that vertical flow dominates horizontal flow, consistent with previous findings from the site (Harr, 1977; McGuire and McDonnell, 2010).
Diel variations in SP associated with tree transpiration have been recorded before. Gibert et al. (2006) continuously measured SP in electrodes installed into a tree trunk and roots, relative to a single ground electrode, to quantify distribution of sapflow within a tree. Diel variations in measured potentials occurred on most of their electrodes during spring and summer months, with less distinct diel signals during the winter months. In addition to the multiple trunk and root electrodes, Gibert et al. (2006) measured the potential between the single reference electrode and a second ground electrode (called a soil “soil chemical electrode” in the paper) closer to the tree. The potential differences measured between the two ground electrodes are attributed to changes in soil chemistry, but no corroboratory data were presented. In the absence of other data, these signals could be due to transpiration. The potential differences are positive and increase during transpiration, as with our data, suggesting movement towards the tree controlled by transpiration.

4.5.6 Modeling Results

Using a coupled fluid flow and electrical model we were able to generate voltage differences similar to those observed in the field (Figures 4.7 and 4.8) and confirm that the signals observed in the field could be a result of precipitation and root-water uptake. The goal of the modeling exercise was to explore the mechanisms controlling the observed signals rather than to duplicate signals exactly; therefore, the magnitudes of the SP signals are not identical to the field data. Even though the modeled velocities and matric potentials may differ from the field conditions due to uncertainty of soil moisture and porosity parameters, they are presented to provide an indication of the scale of water movement producing these SP signals.
4.5.6.1 Seasonal Variations

When measured precipitation and root-water uptake are used to drive the vadose zone flow model, the modeled SP signals (Figure 4.7e) are similar to those observed (Figure 4.3) at the seasonal scale. Voltages measured on all pairs increase in amplitude from the dry to rainy period. The modeled voltages between the upper and outer pairs have similar magnitude increases as the measured data between dry and wet periods; however, the transition between dry and wet is not as great between the modeled inner and lower pairs data as in the measured data. Additionally, the modeled inner and lower pairs deviate in opposite direction from their measured counterparts during the wet season. Instead of becoming more negative during the wet period, suggesting movement away from the tree as in the measured data, the modeled SP signal of the lower pair becomes slightly more positive indicating movement towards the tree. One explanation is that our modeled system does not include the 37° hillslope present at the field site. In the modeled system, new soil moisture from precipitation moves towards the tree to fill in the low soil moisture area created by root-water uptake. In contrast, at the field, the electrodes are on the downslope side of the tree and replenishment of the low soil moisture is likely to occur in part from topographically driven downslope flow and result in a negative contribution to the measured SP signal.

The modeled SP signals in all pairs increase in magnitude in response to rain in early July, and late September, and then return to background levels, although more slowly than in the measured data. The modeled data are additionally smoother than the measured data, and do not contain some of the jumps observed in the measured data. In Section 4.5.1 we hypothesized that two of these jumps, on July 10th and October 5th might be due to preferential flow paths, moving soil moisture to depth quickly. Indeed, these jumps observed are not reproduced by the model,
which only incorporates matrix flow and does not include any mechanism for preferential flow. No sudden voltage jump occurs in the modeled data on October 29, further supporting the interpretation of a technical malfunction, rather than environmental origin. When the root-water uptake is excluded from the numerical model, the greatest difference occurs in the outer pair, which is on average 1.8 mV more negative, suggesting greater downward movement of water in the absence of root-water uptake, and the upper pair, which is on average -0.5 mV less positive, suggesting less movement towards the tree (Figure 4.7e).

4.5.6.2 Diel Fluctuations

When evaluated on the daily time-scale, the modeled SP signals contain diel fluctuations (Figure 4.8e) similar to those observed in the measured data (Figure 4.4e). As with the seasonal change in magnitude, the strongest diel signals are observed in the outer and upper pairs. In the outer pair, the voltages, which are consistently negative (indicative of downward flow), became less negative during the day (indicative of decreased downward flow), and more negative (indicative of increased downward flow) during the night hours. The decrease in amplitudes of measured voltages coincides coincide with the peak in transpiration suggesting they originate from root-water uptake.

Additionally, diel variations in matric head are observed at 0.3 m within the model as with the measured data. However, due to uncertainty in soil the soil moisture curve data, the magnitude of matric heads differ. As the matric head at 0.3 m decreases, the vertical head gradients also decrease, reducing the rate of vertical flow. Diel voltage fluctuations disappear when root-water uptake is excluded from the model (Figure 4.8e). When root-water uptake is
excluded, the diel fluctuations in SP and pressure disappear, confirming their plant-induced origin.

4.6 Conclusions

Here, we collected the first direct measurements of two-dimensional water movement in association with root-water uptake. In particular, we investigated sensitivity of SP measurements to evaluate small changes in subsurface water flow induced by transpiration, and the role of soil moisture on mediating the transpiration signal. Continuous soil moisture, temperature and SP measurements, in combination with a coupled fluid flow and electrical modeling, were used to investigate the propagation of transpiration and precipitation signals within the root zone of a tree over five months.

In this study, diel variations are induced by root-water uptake and were evident under dry conditions, while gravity-driven flow dominated under wetter conditions. Periods of high transpiration reduced the rate of downward fluid flow in the soil. Using SP, we were able to evaluate smaller changes in water flow, and at a smaller spatial scale, than was possible with tensiometers. The transpiration origin of the SP signals were confirmed through coupled fluid flow and electrical model, which only generated diel fluctuations similar to those observed in the field, when root-water uptake was included.
CHAPTER 5
EVALUATION OF SELF-POTENTIAL SIGNAL SOURCES ON FIELD DATA FOR HYDROLOGIC STUDIES

Prepared for submission to *Journal of Applied Geophysics*

Emily B. Voytek $^{1*}$ and Kamini Singha $^{1,2}$

5.1 Introduction

Self potential (SP) is a passive geophysical method that relies on measurements of naturally occurring electrical potentials in the ground. Electrical potentials are generated by any current-inducing phenomena, including groundwater flow, electron movement due to temperature and chemical gradients, and oxidation-reduction (redox) potentials, among others. SP is the only geophysical method directly sensitive to the movement of water, and therefore is a promising tool for mapping groundwater flow (e.g. Jouniaux et al., 2009; Revil and Jardani, 2013). The method is especially appealing for remote locations, because the amount of equipment required is small and data collection is relatively straightforward. Improvements in the physical description of signal sources (e.g. Linde et al., 2007; Revil et al., 2007) and of the non-polarizing electrodes required for measurement (e.g. Petiau, 2000) has led to increased use of SP in recent years (e.g. Revil and Jardani, 2009). However, the use of SP by hydrologists is still limited, in part due to the perceived difficulty of separating confounding signal sources such as water movement, concentration gradients, thermal gradients and redox potentials (e.g.

---

*Primary researcher and author
$^{1}$Hydrologic Science and Engineering Program, Colorado School of Mines, Golden, CO
$^{2}$Department of Geology and Geological Engineering, Colorado School of Mines, Golden, CO
Recent laboratory work has improved our understanding of each of these processes by isolating a signal source, such as temperature gradients (e.g. Leinov and Jackson, 2014; Revil et al., 2016) or water chemistry (e.g. Leinov and Jackson, 2014; Maineult et al., 2004) on the SP signal. However, with each advance in process-based understanding the complexity of descriptive equations has increased such that the method remains underutilized by hydrologists.

Here, we provide an overview of the SP method and discuss the natural processes that can generate electrical currents, which produce voltage differences measured by SP. In discussion of each of the current-generating processes we provide an overview of relevant equations, and references to extended discussion should the reader be interested in greater detail. Through analysis of a field data set from a sub-alpine meadow in Rocky Mountain National Park, in Colorado, we evaluate the potential magnitude each process could have on field data, and discuss the best methods to quantify their influence on data interpretation. This is achieved through analysis of field data in Matlab. Through these examples we seek to aid hydrologists considering use of SP determine which processes may control SP signals at their field sites, and what auxiliary field data should be collected to improve confidence in SP-derived interpretations.

5.2 Background

5.2.1 Self Potential Method

Groundwater is one of the most difficult pieces of the hydrologic cycle to constrain, because available data are usually quite limited and the subsurface is highly heterogeneous. Groundwater studies typically rely on data from wells or piezometers to determine direction and magnitude of flow. However, there are often few wells at a field site, and under some conditions,
none may be available due to logistical or regulatory constraints. Logistical constraints can include difficult or remote site access, or practical considerations such as drilling through blocky talus (Clow et al., 2003). In these cases, and even in the absence of such constraints, hydrologists may wish to know more about flow patterns than is available from limited well data. SP requires little equipment relative to other geophysical techniques, and so it has also been used to monitor flow in hard to access locations such as glaciers (e.g. Kulessa et al., 2003), permafrost environments (Voytek et al., 2016) and remote ice-cored moraines (Moore et al., 2011).

SP measurements allow users to fill in information about subsurface hydrologic processes between other existing data (e.g. hydraulic head from wells). For example, SP has been used to monitor spatially heterogeneous infiltration from drainage ditches (e.g. Suski et al., 2006) and the distribution of hydraulic heads resulting from pumping tests (e.g. Jardani et al., 2009; Maineult et al., 2008; Rizzo et al., 2004; Soueid Ahmed et al., 2014). The method has also seen extensive use in identifying flow through earthen dams (e.g. Bolève et al., 2009; Ikard et al., 2014, 2012; Rittgers et al., 2015) where even qualitative interpretations of SP data can aid in the identification of preferential flow paths, which can be detrimental to the integrity of dams. SP measurements have been used to monitor vertical unsaturated zone flow associated with precipitation events (Darnet and Marquis, 2004; Doussan et al., 2002; Suski et al., 2006). SP measurements are a versatile addition to hydrologic studies because they are inexpensive and can be collected in situ to monitor processes at a variety of spatial scales, and from the land surface.

SP is a passive geophysical method, meaning that measurements are made of naturally occurring potentials rather than through the introduction of an artificial signal as is done in other geophysical methods such as electrical resistivity (ER) imaging or seismic reflection or refraction. Users therefore are attempting to solve for both signal source and propagation. The
added challenge of an unknown signal source has not prevented widespread use of gravity, another passive geophysical method. SP data are the electrical potential differences, V [V], between a reference electrode and a potential electrode:

$$\text{SP}_i = V_i - V_{\text{ref}}.$$ \hspace{1cm} (5.1)

Voltages are generated by any process that moves electrons (i.e. produces current). The strength of measured voltages is a function of the magnitude of the current producing the signal, and the electrical conductivity of the ground. These terms are described by a generalized version of Ohm’s law:

$$\mathbf{J} = \mathbf{J}_S + \sigma \mathbf{E}$$ \hspace{1cm} (5.2)

where \(\mathbf{J}\) is the macroscopic current density [A m\(^{-2}\)], \(\mathbf{J}_S\) is the source current density [A m\(^{-2}\)], \(\sigma\) is the electrical conductivity of the ground [S m\(^{-1}\)], and \(\mathbf{E}\) is the electrical field [V m\(^{-1}\)], further defined by:

$$\mathbf{E} = -\nabla V$$ \hspace{1cm} (5.3)

where \(\nabla V\) is the gradient of the voltage field [V], which is measured at discrete locations in Equation 5.1. The first right-hand side term of Equation 5.2, \(\sigma \mathbf{E}\) describes how currents propagate through the material, while \(\mathbf{J}_S\) describes the distribution of current sources. A source current is generated by any mechanism producing movement of change. In natural environments these include water movement (streaming potential), chemical gradients (diffusion potential), temperature gradients (thermoelectric potential) and, at times, redox conditions (redox potential). These effects are additive and contribute to \(\mathbf{J}_S\) (Revil and Jardani, 2013):

$$\mathbf{J}_S = \mathbf{J}_{\text{streaming}} + \mathbf{J}_{\text{chemical}} + \mathbf{J}_{\text{redox}} + \mathbf{J}_{\text{thermo}}.$$ \hspace{1cm} (5.4)

A basic knowledge of how each of these processes generate current, such as those presented in section 5.1.1-5.1.5, can ensure that they are properly evaluated as potential signal
sources in field data, and lead to better constrained interpretation of SP data. Once the source of SP signal is known, there are two ways of predicting SP potentials: coupling coefficients, which can approximate SP potentials from individual or even coupled processes, and physically based, electrical flow models, which numerically solve equations describing the physics (e.g. groundwater flow, diffusion) to determine voltage distributions.

Coupling coefficients relate changes in measured voltages to changes in another parameter (e.g. hydraulic head or chemical concentration). Coupling coefficients are calculated from field or laboratory measurements (e.g. Jardani et al., 2006; Revil et al., 2016), and can also be used if more than one current-producing process is occurring at a site. Because measured SP signals are the sum total of voltage differences generated by individual processes, if the distribution of one parameter is known (e.g. chemical concentration) the contribution of that parameter to the SP signal can be removed using one coupling coefficient, and the remaining signal analyzed to determine the distribution of the unknown parameter (e.g. hydraulic head) using another coupling coefficient (e.g. Naudet et al., 2003).

In contrast to coupling coefficients, which can provide a first-order estimate of SP signals, physically based, numerical modeling solves both the underlying physicochemical process and the resultant electrical problem, which is more computationally demanding. For example, SP2DINV is a Matlab-based code that has been developed specifically for modeling SP signals resulting from groundwater flow (Soueid Ahmed et al., 2013). Alternatively, multiphysics solvers, such as COMSOL, can be used to model coupled processes (e.g. Boleve et al., 2007; Vasconcelos et al., 2014). Fully coupled models allow for incorporation of any number
of current-generating processes. For example, Jougnot and Linde (2013) show the importance of including changing water chemistry caused by diffusion of ions across the porous tip of non-polarizing electrodes in laboratory-scale SP experiments using a coupled numerical model that includes water and electrical flow and chemical diffusion.

In each section below, we review how to estimate coupling coefficients for the four most common current sources in environmental applications—water movement, chemical and thermal gradients and redox potential (Figure 5.1). Finally, additional sources of voltages differences (i.e. anthropogenic noise and temperature effect on electrodes), which can contaminate SP signals from natural sources, will then be discussed to help users collect the best possible SP data.
5.2.1.1 Streaming Potential

Voltage differences resulting from the movement of water are called streaming potentials. Streaming potentials are generated by the small electrical currents produced by movement of electrons when water moves through porous materials. Movement of electrically neutral pore water does not generate current. Instead currents are generated by the advection of excess charges in the mobile layer of the electrical double layer (EDL) of the surface of mineral grains. The EDL develops on mineral surfaces when contact with water results in the surface becoming slightly charged. The polarity of the charge is dependent on mineral chemistry and pH of the surrounding water. Under typical field conditions (i.e., pH 5-8), the surface charge of a silica grain is negative (Revil and Jardani, 2013). The negative surface charge of the mineral attracts positive ions, resulting in an EDL that is enriched in positive ions relative to the bulk pore water. Some of these excess charges sorb directly to the mineral surface and are immobile, while other opposing charges in the ionically enriched EDL remain more mobile (5.1a). The total net charge of the EDL is called excess charge density, $Q_v$ [$C m^{-3}$] and the net charge of the mobile portion is called the effective excess charge density, $\tilde{Q}_v$ [$C m^{-3}$]. As $\tilde{Q}_v$ is the pertinent term to generating currents measured with the SP method, constitutive relationships between $\tilde{Q}_v$ and other measureable parameters have been developed, for example, $k$, the intrinsic permeability [$m^2$] (e.g., Jardani et al., 2007):

$$\log(\tilde{Q}_v) = -9.2349 - 0.8219\log(k). \quad (5.5)$$

Once $\tilde{Q}_v$ is defined, the resultant SP signals can be numerically estimated from measured or modeled groundwater flow patterns using Equation 5.2 with the following description of current flow generated by water movement:

$$J_{\text{streaming}} = \tilde{Q}_v U \quad (5.6)$$
where $U$ is the Darcy velocity (m/s).

Rather than determining $\dot{Q}_v$ and explicitly modeling the movement of charges, coupling coefficients can be used to approximate SP signals from the distribution of other known values. In the case of streaming potentials, the coupling coefficient, $C_{\text{head}}$ [mV/m] takes the place of several terms in the physically based equations, specifically electrical conductivity, $\sigma$, in Equation 5.4 and hydraulic conductivity, $K$ [m/s], from the Darcy velocity term in Equation 5.6, to relate SP measurements to changes in hydraulic head. Sill (1983) first formulated a streaming-potential coupling coefficient that equated voltages to changes in fluid pressure [V Pa$^{-1}$]. However, this formulation did not account for the observed dependence of measured voltages on soil parameters (i.e. $K$; Jouniaux and Pozzi, 1995). Additionally, it was not possible to extend this pressure-dependent coefficient to unsaturated conditions where saturation-dependent $K$ leads to different flow rates for the same pressure difference (e.g. Guichet et al., 2003; Perrier and Morat, 2000; Revil and Cerepi, 2004). To address the shortcomings of the streaming potential coupling coefficients linked to changes in pressure, Revil and Leroy (2004) and Revil et al. (2005b) developed a streaming potential coupling coefficient that related voltages to water velocity to head [V m$^{-1}$]. Streaming-potential coupling coefficients can be determined for a given soil through laboratory experiments where a known hydraulic gradient is applied, and the resultant voltage differences are measured, as in Jardani et al. (2006) and Suski et al. (2006); however, coefficients estimated from field data may be preferred for estimating site-specific characteristics. If site-specific coupling coefficient determination is not possible, Revil et al. (2017) derived a constitutive equation for a coupling coefficient based on fluid conductivity $\sigma_f$ [S m$^{-1}$]:

$$\log(C_{\text{head}}) = a + b \log_{10}(\sigma_f) + c \log_{10}(\sigma_f)^2$$  \hspace{1cm} (5.7)
where a, b and c are fitting parameters of -0.895, -1.319 and -0.1227 respectively (Figure 5.2a). This equation for a streaming-potential coupling coefficient is appropriate for most salinities observed in groundwater, but does not hold under very high salinities, such as seawater. The reader is referred to Vinogradov et al. (2010) for extended discussion of streaming potential coefficients in high salinity environments.

5.2.1.2 Diffusion Potential

Diffusion potentials are voltages differences generated by chemical concentration gradients. Diffusion moves ions from areas of high concentration to low concentration due to the increased collision rate of ions in other ions in the high concentration area, which result in a net
movement away from the concentration center (Figure 5.1b). Chemical diffusion generates electrical potentials due to the differential diffusion rates of dissolved species. Not all ions move at the same rate; ions with smaller hydrated radii will diffuse more quickly than those with larger radii. Electrical current is generated when one type of changed ion (positive or negative) moves more quickly than the other, generating current and resulting in potential differences.

A diffusion-potential coupling coefficient, \( C_{\text{diffusion}} \), relates voltage changes to relative changes in concentration. The equations of Revil (1999) for \( C_{\text{diffusion}} \) rely on changes in fluid concentration, or the related value of fluid electrical conductivity \( \sigma_f \), relative to a reference point:

\[
\Delta V = C_{\text{diffusion}} \ln \left( \frac{c_i}{c_{\text{ref}}} \right) \approx C_{\text{diffusion}} \ln \left( \frac{\sigma_{f,i}}{\sigma_{f,\text{ref}}} \right) \quad (5.8)
\]

\[
C_{\text{diffusion}} = n \frac{RT}{Ae} \frac{u_{\text{Cl}} - u_{\text{Na}}}{u_{\text{Cl}} + u_{\text{Na}}} \quad (5.9)
\]

where \( c_i \) and \( c_{\text{ref}} \) are the salt concentration at a given point, \( i \), and the reference point [mol L\(^{-1}\)], \( \sigma_{f,i} \) and \( \sigma_{f,\text{ref}} \) are fluid conductivity at a given point and the reference point [m/S], \( n \) is the porosity [-], \( R \) is molar gas constant [\( \approx 8.314 \text{ J mol}^{-1} \text{ K}^{-1} \)], \( T \) is the absolute temperature [K], \( A \) is Avogadro’s number [\( \approx 6.022 \text{ 10}^{23} \text{ mol}^{-1} \)], \( e \) is the elementary charge [\( \approx 1.602 \text{ 10}^{-19} \text{ C} \)], and \( u_{\text{Cl}} \) and \( u_{\text{Na}} \) are ionic mobilities of \( \text{Cl}^- \) and \( \text{Na}^+ \) [m\(^2\) s\(^{-1}\) V\(^{-1}\)] or other species (Maineult et al., 2005). The ionic mobilities describe how quickly one species moves relative to another. The relative mobilities of Na and Cl are 0.38-0.40, and 0.60-0.62 [-] respectively, resulting in the ratio \((u_{\text{Cl}} - u_{\text{Na}})/(u_{\text{Cl}} + u_{\text{Na}})\) in Equation 5.9 equal to 0.21 [-] (Maineult et al., 2005). While the value of coupling coefficients are influenced by temperature, the ratio of ionic mobilities is relatively independent of temperature (Revil, 1999).
5.2.1.3 Thermoelectric Potential

Voltage differences generated by temperature gradients are called thermoelectric potentials. The currents driving thermoelectric potentials are generated in much the same way as diffusion currents; in concentration-driven diffusion potentials ion movement is generated by increased collisions due to higher concentrations, whereas thermal gradients are driven by increased collision rates due to higher activities at increased temperatures (Revil et al., 2016; Figure 5.1c). As with chemical diffusion, the smaller ions will diffuse more quickly and the differential movement generates a current.

The thermoelectric coupling coefficient, $C_{thermal}$ [mV/°C], relates changes in voltages to temperature gradients (Leinov and Jackson, 2014; Revil et al., 2016). The thermoelectric coupling coefficient is equal to the negative Seebeck coefficient used in electrochemistry ($C_{thermal} = -S_e$). The polarity of ions moved by thermoelectric diffusion, and thus the value of the thermoelectric coupling coefficient, depends on whether surface conductivity or pore water conductivity dominates in a given combination of porous material and fluid concentration. This ratio is given by the dimensionless Dukhin number. In situations in which pore water conductivity dominates over the contribution of surface conductivity (low Dukhin number), the thickness of the diffuse layer of excess charge in the electrical double layer is small as compared to the size of the pore. Therefore, thermal diffusion typically results in preferential movement of smaller ions of the bulk pore water (Figure 4.1b). Whereas conditions with a high surface conductivity contribution (high Dukhin number), the diffuse layer of excess charge is large relative to the pore volume and diffusion causes movement of the excess charges in the diffuse layer (similar to advection in Figure 5.1a).
Revil et al. (2016) developed equations that describe these two bounds. Equation 5.10 describes behavior when fluid conductivity dominates (low Dukhin) and Equation 5.11 that describes behavior when surface conductivity dominates (high Dukhin):

\[ S_e = \frac{1}{eN} \left[ t_{(+)}s_{(+)}^* - (1 - t_{(+)}s_{(-)}^*) \right] - \frac{2.303k_b}{e} \left( 2t_{(+)} - 1 \right) \log C_f \]  
(5.10)

\[ S_e = \frac{1}{eN} s_{(+)}^* - \frac{2.303k_b}{e} \log C_f \]  
(5.11)

where \( t_{(+)} \) is the microscopic Hittorf number [-] describing the fraction of electrical current carried by the cations in the pore water, \( s_{(\pm)}^* \) is the effective ionic partial entropies of the solute [J K\(^{-1}\)], \( k_b \) is the Boltzmann constant [1.381 x 10\(^{-23}\) J K\(^{-1}\)] and \( C_f \) is the concentration of ions [M].

Figure 5.2b shows these theoretical thermoelectric coupling coefficients vary between -1.5 and 0.1 mV/K over a range of fluid conductivities.

5.2.1.4 Redox Potential

The final common source of electric potentials in the subsurface is redox gradients. Redox gradients can be formed due to biologic or metallic sources. In order for redox potentials to produce measurable SP voltages, two conditions must be present: (1) a redox gradient and (2) a conductive connection (Hubbard et al., 2011). In order for redox potentials to generate large (>100mV) SP signals, an effective electronic conductor is needed. A metallic body, biofilm, or iron precipitates have been suggested (Revil and Jardani, 2009).

Laboratory experiments have calculated redox coupling coefficients, \( C_{\text{redox}} \) [mV/mV], at between 0.2 and 0.55 mV/mV (Naudet et al., 2004; Naudet and Revil, 2005). In field examples, redox coupling coefficients have been derived from SP data by first removing the effect of streaming potentials, and relating the remaining potentials to measured redox potential (Naudet
et al., 2004, 2003). Because redox potential can effect SP measurements only under very specific conditions, they will not be evaluated further in this work. Instead, those interested should to look to existing work surrounding redox potentials (e.g. Naudet et al., 2003; Naudet and Revil, 2005; Risgaard-Petersen et al., 2012; Rittgers et al., 2013).

5.2.2 Other sources of SP voltages

5.2.2.1 Anthropogenic and environmental noise

In addition to processes in the subsurface that generate electrical current, SP measurements are sensitive to ambient electrical noise, such as telluric currents generated from current flow in the ionosphere, electrical storms, and power lines (Revil and Jardani, 2013). The effects of these sources can be reduced by avoiding data collection during solar storms, during thunderstorms, and near electrical infrastructure. If impossible to avoid, high frequency noise can be overcome by taking repeat measurements, or using a voltmeter that averages measurements over a certain period of time.

5.2.2.2 Temperature effect on electrodes

In addition to subsurface temperature gradients resulting in thermoelectric potentials, temperature differences between measurement electrodes can result in drift, a voltage difference due to the equipment and not subsurface processes. If the drift is not corrected for, incorrect data interpretations are possible. Of the three common types of non-polarizing electrodes used for SP measurements, Ag/AgCl, Cu/CuSO4 and Pb/PbCl2 (or Petiau-type electrode), the Petiau-type electrode is the most temperature stable with only a 0.2 mV/°K dependence (Petiau, 2000).
In surface SP surveys, in which the same two electrodes are used to collect data for the entire data set, temperature drift can be reduced through proper field procedures including minimizing exposure of the electrodes to heat sources such as hands or sunshine. Additionally, this difference can be corrected for by conducting regular measurements between the reference and measurement electrodes (i.e. tip-to-tip measurements; Revil and Jardani, 2013). Through this process, any changes in voltage difference between the reference and measurement electrode due solely to changes within the equipment can be removed prior to interpretation.

For longer-term studies, where electrodes are buried into the ground, tip-to-tip measurements are not possible. Instead, SP measurements can be corrected for temperature differences between the two electrodes:

\[
SP_i^T = \alpha(T_i - T_{\text{ref}}),
\]

\[
SP_{\text{corrected}} = SP_i + SP_i^T.
\]

where \(SP_i^T\) is the temperature correction to be applied to the measurement electrode to the reference electrode, \(\alpha\) is the electrode-type correction factor (0.2 mV/°K for Petiau-type as noted above), and \(T_i\) and \(T_{\text{ref}}\) [°C] are temperatures at the measurement and reference electrodes.

5.3 Methods

To evaluate the potential influences of these processes, we collect an SP data set at a sub-alpine meadow within the Loch Vale watershed in the Colorado Rockies. These data were needed to help solve an open question regarding how groundwater moved within a meadow (Dave Clow, personal communication), but regulatory regulations prevented drilling at the site.
The limited equipment required for the SP method—two electrodes, a voltmeter and wire—was ideal for the site which is requires a 6-km hike, with 400-m elevation gain, from the nearest trailhead, to access.

5.3.1 Field Site

Loch Vale is a subalpine catchment of Rocky Mountain National Park (Figure 5.3) that was established as a long-term research site in 1981 (Baron, 1992). The 6.6 km² watershed ranges in elevation from 3050 to 4026 m at the Continental Divide on the east side of the watershed (Campbell et al., 1995). The bedrock in the Loch Vale watershed is predominantly Precambrian biotite schist with isolated outcrops of younger Precambrian granite (Baron and Mast, 1992; Clow et al., 2003). Exposed bedrock makes up 53% of the basin area; the remaining surficial geology is dominated by glacial deposits with limited vegetation (Clow et al., 2003). Only 6% of the Loch Vale watershed is covered by forested or meadow soils, which are often
located in valley floors near surface water bodies. The soils were developed after the retreat of glaciers and are generally coarse textured (Baron et al., 1992). Despite their limited area, the high reactivity of the organic soil materials in the meadows have a large influence the surface water quality (Baron et al., 1992; Cosby et al., 1985). Additional information about the Loch Vale catchment can be found in Baron and Mast (1992).

Andrews Meadow is one of two areas of alluvial soil in the watershed, which when combined account for only 1% of the watershed (Clow et al., 2003). The meadow is approximately 50 x 60 m and located adjacent to Andrews Creek at 3213 m elevation, just below local tree line. The meadow is bound to the south by Andrews Creek, and to the east by a smaller tributary, Andrews Spring (Figure 5.4). The northern boundary of the meadow grades into a talus slope made from rockfall. Seismic refraction surveys indicate the meadow material is 0.3 to 2.7 m thick (Clow et al., 2003). Despite the known influence of the meadow soils on the surface water chemistry, rocky substrate and regulatory constraints prevent installation of piezometer networks to quantify groundwater flow with the Loch Vale catchment (Clow et al., 2003).

5.3.2 SP data collection

Seven SP surveys were conducted between June 17, 2015 and October 2, 2015. SP measurements were made using two non-polarizing electrodes and a Fluke 87 V handheld voltmeter with high internal impedance (>10 MOhm). All SP measurements were made relative to a reference electrode buried in a shaded area outside of the meadow to minimize temperature variation. The position of the reference electrode remained constant between each of the surveys. Petiau-type electrodes were used because they experience the least temperature-dependent drift of commercially available non-polarizing electrodes (Petiau, 2000). The voltage difference
Figure 5.4 Interpolated plots of SP data from Andrews Meadow. Orientation map shows position of some permanent location markers in the meadow, relative to the streams. All interpolations have been cropped to the same extent, except for June 17th, when the meadow remained partially covered by snow. Increased gradients, indicated by more densely spaced contours, suggest increasing flow rates.
between the two electrodes was measured before and after each survey, as well as between each 20 measurements to track the equipment drift. During post-processing the voltage difference between consecutive drift measurements was distributed evenly between the intervening measurements.

Each survey consisted of ~120 points in a grid with approximately 5-m spacing. Three SP measurements were made at each point and used to estimate the uncertainty of each measurement. The median uncertainty of measurements throughout all of the surveys was 1.3 mV. Permitting constraints precluded the installation of long-term plot markers. Therefore, the location of each SP measurement was recorded using a Trimble Geo7x GPS with a Zephyr antenna. After post-processing the horizontal accuracy of points was ~10 cm. Geo-referenced data were used to create contour plots that cover the same spatial extent for each survey, except the first visit on June 17th when snow remained on the northern and western portions of the meadow (Figure 5.4). The contoured surface was created through a least squares minimization, which respects the confidence of the individual measurements.

5.3.3 Auxiliary data

Regulatory constraints prevented us from collecting auxiliary subsurface data, including installing piezometers to measure hydraulic head values within the meadow. Instead, to evaluate the influence of other current sources on our SP measurements, we used existing data and previously published interpretations.
5.3.3.1 Stream Levels

We relied on stage measurements from two USGS-maintained stream gages at Andrews Creek and Andrews Spring to the south and east of the meadow (Figure 5.4) to estimate hydraulic gradients within the meadow. We used the water levels in the streams to estimate hydraulic head at these two positions. Additionally, there was an area on the northern edge of the meadow that was persistently saturated at the surface, we assumed this was a point of constant head. From these three points a hydraulic gradient was calculated (Figure 5.5; Appendix D).
5.3.3.2 Air Temperature

Mean daily air temperature for May 1 – September 15, 2015 were available from a USGS operated weather station in the southeast corner of the meadow (Figures 5.4). The weather station was relocated on September 15. No ground or stream temperature from this time period was available.

5.3.3.3 Electrical Resistivity

A series of six electrical resistivity (ER) profiles, running approximately west to east, were collected in early August to determine the subsurface electrical conductivity structure of Andrews Meadow (Figure 5.4). ER data were collected using an IRIS Syscal Pro and stainless-steel electrodes. Each survey consisted of 48 1-m spaced electrodes. Contact resistances between electrodes were low, with a median value of 2.2 kOhm. A dipole-dipole survey (1159 quadrupoles) was used to optimize the multi-channel capabilities of the IRIS Syscal Pro, and maximize the number of surveys that could be completed in one day. Measurement errors were based on stacking, the difference between two resistance measurements of the same quadripole. The mean stacking error was 0.1%. The raw resistance measurements were inverted into electrical conductivity profiles using the R2 inversion code (Figure 5.6; Binley and Kemna, 2005). A rectilinear mesh, with a minimum spacing of 0.25 m was used. The depth of investigation was determined using the method of Oldenburg and Li (1999).
Figure 5.6 ER profiles shown as bulk conductivity values. Locations of individual lines shown in Figure 5.3.
The meadow material was assumed to have a hydraulic conductivity between $4.4 \times 10^{-6}$ and $9.5 \times 10^{-4}$ m/s (intrinsic permeability $4.6 \times 10^{-13}$ to $1 \times 10^{-10}$ m$^2$ at 10 °C) based on constant-head permeameter testing at another meadow within the Loch Vale watershed by Bachmann (1994, Table 5.1). The porosity was assumed to be 0.2 (Clow et al., 2003). The average fluid conductivity of the pore water within the meadow was assumed to be $4.9 \times 10^{-4}$ S/m (4.9 uS/cm) based on average soil water chemical concentrations provided in Wilson (2015). The average stream water conductivity was assumed to be $9.8 \times 10^{-4}$ S/m (9.8 uS/cm), also based on Wilson (2015).

### Table 5.1 Parameters used in data processing and numerical modeling.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Unit</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>hydraulic conductivity, $K^*$</td>
<td>[m/s]</td>
<td>$4.4 \times 10^{-6}$ to $9.5 \times 10^{-4}$</td>
</tr>
<tr>
<td>porosity</td>
<td>[-]</td>
<td>0.2</td>
</tr>
<tr>
<td>pore-water fluid conductivity, $\sigma^-$</td>
<td>[S/m]</td>
<td>$4.9 \times 10^{-4}$</td>
</tr>
<tr>
<td>stream water conductivity</td>
<td>[S/m]</td>
<td>$9.8 \times 10^{-4}$</td>
</tr>
<tr>
<td>density, $\rho$</td>
<td>[kg/m³]</td>
<td>999.7</td>
</tr>
<tr>
<td>gravity, $g$</td>
<td>[m/s²]</td>
<td>9.81</td>
</tr>
<tr>
<td>dynamic viscosity of water, $\mu$ at 10°$^\wedge$</td>
<td>[kg/m*s]</td>
<td>$1.04 \times 10^{-3}$</td>
</tr>
</tbody>
</table>

* Bachman, 1994
\~ Calculated from Wilson, 2015
$^\wedge$ Crittenden et al., 2012

### 5.3.3.4 Hydraulic conductivity and fluid electrical conductivity

The meadow material was assumed to have a hydraulic conductivity between $4.4 \times 10^{-6}$ and $9.5 \times 10^{-4}$ m/s (intrinsic permeability $4.6 \times 10^{-13}$ to $1 \times 10^{-10}$ m$^2$ at 10 °C) based on constant-head permeameter testing at another meadow within the Loch Vale watershed by Bachmann (1994, Table 5.1). The porosity was assumed to be 0.2 (Clow et al., 2003). The average fluid conductivity of the pore water within the meadow was assumed to be $4.9 \times 10^{-4}$ S/m (4.9 uS/cm) based on average soil water chemical concentrations provided in Wilson (2015). The average stream water conductivity was assumed to be $9.8 \times 10^{-4}$ S/m (9.8 uS/cm), also based on Wilson (2015).
5.4 Results and Discussion

The primary goal of SP data collection at this site was to determine the seasonal patterns of groundwater flow through the meadow. However, to do that, we must first evaluate the potential effects of other current generating processes, and account for them in the data processing as necessary. Below, we walk through a framework for exploring field SP data that might be useful for other similar studies.

5.4.1 Evaluation of Electrochemical Potential

To determine whether chemical gradients must be considered in evaluation of the data set, we first estimate the electrokinetic potential generated by maximum expected chemical gradient across the meadow. We assume the maximum possible concentration difference occurs between the pore water fluid conductivity and stream water conductivity, based on values from Wilson (2015) (Table 5.1). We use Equation 5.9, with an assumed temperature of 1 °C, the simulated mean annual air temperature for Andrews Meadow from Clow et al. (2003), to determine an electrochemical coupling coefficient of 1.49 mV/-. Then, using the fluid conductivity values from Table 5.1, we estimate the potential difference generated from this chemical gradient using Equation 5.8. Either the pore water or stream water concentration can be used as the reference; either produces the same result. If the higher conductivity value is used as the reference in Equation 5.8, $\Delta V$ will be negative; if, instead, the lower conductivity value is used, $\Delta V$ will be positive and voltages will increase towards the higher concentration. The ratio of stream to pore-water conductivities is 2, resulting in a log difference of 0.3. This, combined with the diffusion coupling coefficient produces a maximum diffusion potential of 0.5 mV, a
very small contribution to measured SP potentials. The potential contribution of electrochemical potentials is smaller than the mean uncertainty of the individual measurements (1.3 mV). While the expected influence of electrochemical gradients is small at this field site, Equation 5.9 shows that the same concentration gradient could produce a greater potential difference with increased porosity or temperature.

5.4.2 Evaluation of Thermoelectric Potential

Similar to the process of accounting for chemical gradients, temperature gradients can be approximated. Figure 5.2b shows the range of thermoelectric coupling coefficients that can be expected over a range of fluid conductivities. The effect of temperature is greatest under low fluid conductivity conditions, such as found at Andrews Meadow. For the estimated pore water fluid conductivity of Andrews Meadow, 4.9 x 10^{-4} S/m, the expected thermoelectric coupling coefficient (from Equations 5.10 and 5.11) is between -1.2 and 0.03 mV/K. The value could be constrained in laboratory settings in most cases; however, here we were unable to remove and soil or groundwater from the site. Given the very low fluid conductivity at the site and the high organic material, we can assume that the Dukhin number is likely high (surface conductivity dominates), resulting in a thermoelectric coupling coefficient close to the value determined by Equation 5.11: -1.2 mV/K.

While this number indicates that a temperature difference of 5 °C would result in a potential difference of -6.2 mV, it is also important to consider the distribution of the thermal gradient. In the case of the data from Andrews Meadow, increases in ground temperature likely predominantly propagate vertically throughout the meadow. If we assume that heat propagates
Figure 5.7 Thermoelectric potential generation from a) horizontal and b) vertical temperature gradients.

\( C_{\text{temp}} = 1 \text{mV/K} \)
from the surface only such that any temperature gradients generated are in the vertical direction, no change will be observed in voltages measured at the ground surface (Figure 5.7a). If instead, horizontal thermal gradients were observed, they would need to be considered, because they would result in measurable thermoelectric potentials measureable from the ground surface (Figure 5.7b).

Here we show that even in the absence of temperature data, we can still make educated estimates of how much, or little, various processes will affect SP measurements. From the expected orientation of the temperature gradients, and our surface-measurement only survey design, we rule out a large influence of temperature gradients on the measured potentials. It is recommended that SP surveys include measurement of temperature in multiple locations, especially if the survey is to take place across many months such as this. For the remainder of this work, we will assume that thermoelectric potentials in this data set are negligible.

5.4.3 Evaluation of Streaming Potential

Once the effects of other current-generating processes have been evaluated, and either excluded as possible current sources or corrected for as needed, the remaining signal can be analyzed to provide information about groundwater flow patterns in the subsurface. As a first-order interpretation, SP gradients can be used to qualitatively evaluate fluid flow. Voltages increase in the direction of flow, therefore direction of fluid flow to be interpreted as perpendicular to iso-potential lines. In the data sets from each visit to Andrews Meadow there is a trend of increasing voltage towards the south-southeast, indicating that flow is occurring across the meadow from the north to the south-southeast. The gradient increase between June 17 and
June 29 suggests that flow rates increased between these two dates. This increase is particularly evident in the SP data from the northwest corner of the meadow between June 17 and June 29.

June 17 was the first date we were able to collect an SP survey due to snow at the high elevation site. Even during the June 17th survey an isolated snow patch remained on the western side of the meadow, and we were unable to access the northern portion of the meadow due to snow cover. On June 17, the SP gradient north of the isolated snow patch was smaller than the gradient in the middle of the meadow. A shallower gradient suggests less flow occurred in this area, perhaps because the ground beneath the snow patch remained frozen, thus reducing the hydraulic conductivity.

By June 29, the isolated snow patch had melted, and the gradients increased, as would be expected from enhanced flow resulting from ground thaw. The stream gage data from both Andrews Creek and Andrews Spring, which increased rapidly during this time (Figure 5.5.5), support the interpretation of changing hydrologic conditions. Stage at both Andrews Creek and Andrews Meadow increase at the beginning of June. The level of Andrews Creek increased again around June 10, with the onset of warmer surface temperatures.

The measured SP on July 14, August 18 and August 31 have similar gradients to one another, suggesting that flow through the meadow did not change much during the summer season. The gradient then decreased through October 2. On June 29 and July 14, the voltage increases were perpendicular to Andrews Creek, suggesting that flow occurred towards the creek. However, in the data from August 31, September 14 and October 2, the data in southwest corner of the meadow are as much as 5 mV less than the values father downstream. This indicates that groundwater may have been entering the meadow from the west, which is the upstream direction of Andrews Creek.
To convert the map of SP to a map of hydraulic head a coupling coefficient is required. One method is to use Equation 5.7 and the estimated fluid conductivity for the soil water (Table 5.1), which results in a streaming coupling coefficient is -132 mV/m (Figure 5.2). Alternatively, coupling coefficients can be calculated in laboratory experiments (e.g. Boleve et al., 2007; Ikard and Revil, 2014; Malama and Revil, 2014) or from well data at the field site (e.g. Jardani et al., 2009; Rizzo et al., 2004; Suski et al., 2006). A laboratory-derived coupling coefficient only measures voltages generated by water movement through soil, but does not include the influence on the electrical conductivity of the surrounding geologic units. In field settings, the same quantity of groundwater flow, through the same soil, can produce different SP voltages at the surface depending on the conductivity of the underlying material, even if no flow is occurring in the underlying material. Site-specific coupling coefficients should be developed if possible to account for the effects of the surrounding electrical conductivity in field settings.

Figure 5.8 Site specific coupling coefficient determined from the empirical relationship between interpolated SP values and measured heads at three locations shown in Figure 5.4.
We attempted to develop a local coupling coefficient for Andrews Meadow using the stream gage data described in Section 5.3.3.1. To do this, we differenced the stage elevation at two points, and compare that to the differenced SP values at the same locations. Because stage was known at three points, and SP values could be interpolated for these locations, this ratio was calculated between three pairs of locations for each survey visit (i.e. meadow to Andrews Spring, meadow to Andrews Creek and Andrews Spring to Andrews Creek), resulting in a total of 21 points (Figure 5.8; 3 pairs at each of 7 visits). The spread of these points is great and fit of a linear trend line poor ($R^2$ value = 0.08). In fact, the slope of the linear fit is opposite the expected direction (negative). This suggests that the assumptions we were forced to make were not appropriate. Similar methods have produced better fitting coupling coefficients at field areas with more data availability (e.g. Jardani et al., 2009; Rizzo et al., 2004; Suski et al., 2006).

Regardless of how it is determined, this streaming current coupling coefficient is then applied to an entire survey to produce a spatially extensive estimate of head (Figure 5.9). The streaming potential coupling coefficient applies a linear change, therefore the qualitative interpretations from the previous section remain unchanged.

Figure 5.9 SP data converted to maps of hydraulic head using coupling coefficient of -132 mV/m from Equation 5.7.
5.5 Conclusions

SP is a promising tool for hydrologic investigations. However effective use of the tool requires knowledge underlying processes that generate the potentials which are measured. Here we outline the multiple natural processes that generate current and are measured by SP, detail the auxiliary data that should be considered when conducting an SP survey to constrain interpretations, and described processing SP data using coupling coefficients. Through this we outline the controls of SP data, and highlight how these surveys can help in data-limited hydrologic studies.

In this work we used SP to evaluate groundwater flow through time at a hard-to-access site in Rocky Mountain National Park, in Colorado. Due to regulatory constraints we were unable to use wells or piezometers to determine groundwater flow patterns and SP provided another means of evaluating this important component of the hydrologic cycle. Before interpreting the measured SP data for groundwater flow patterns, we had to consider other possible sources of electrical potential, including chemical gradients, thermal gradients and redox potentials.

We used existing data sets to determine that these other sources of potentials would not result in a large contribution to the measured voltages at this site. The contribution from chemical gradients was determined to be less than the average error on individual measurements. Making assumptions about the directionality of temperature gradients within the meadow, we determined that potentials generated from thermal gradients would not affect measurements made from the meadow surface. In doing so, we were able to explore changes in groundwater movement from our data. Flow through the meadow appears to increase as the meadow thaws, and then decrease again at the end of the summer when the quantity of meltwater decreases.
CHAPTER 6

CONCLUSIONS

As the study of hydrologic processes progresses, new tools are needed that can measure groundwater flow in remote and hard-to-access locations at high temporal and spatial scales. In the work presented here I have shown that the self-potential (SP) method can be an effective tool for mapping patterns of shallow groundwater and unsaturated zone flow at high spatial and temporal resolution. The data presented in these three chapters come from disparate locations, but taken together they demonstrate the effectiveness, as well as some of the potential difficulties, of incorporating SP into ongoing hydrologic studies.

In Chapter 3, a combined geophysical approach of SP and electrical resistivity (ER) was used to map shallow subsurface flow paths in and around water tracks in Alaska, drainage features common to arctic hillslopes. Qualitative interpretation of the SP data, in combination with knowledge of the geometry of the subsurface thaw boundary from ER data, provided the first detailed look at groundwater flow patterns that occur within these unique and remote environments. We identify flow patterns wherein hillslope flow occurs both downslope and towards the primary water track ‘channels’. Additionally, areas of localized upwelling were identified in the SP data. These areas of localized upwelling occur in areas of shallower frozen ground boundary as observed in the ER transects, suggesting that the shape of the underlying permafrost boundary affects patterns of groundwater flow. The data presented in this chapter are only from one snapshot of time within a very temporally dynamic system (i.e. depth to frozen boundary changes during season). Additionally, each of the geophysical data sets were evaluated independently in this work. Future work could include the collection of time-lapse data sets of
SP and ER, and joint data processing, to evaluate the interconnectedness of flow paths and thaw patterns. Resistivity structures observed in ER profiles can be used to constrain the interpretation of the SP data through either forward or inverse modeling.

In Chapter 4, continuously measured SP was used to answer open questions about how transpiration signals propagate from trees to streams at a long-term research site in Oregon. Using an array of SP electrodes installed in the ground, we collected the first direct measurements of two-dimensional water movement in association with root-water uptake. Unlike tensiometers, which measure matric pressure at two locations to determine direction of flow, SP signals are generated by the dynamic movement of water. The continuous data collection, spanning wet and dry seasons within the forest, yielded information about the relative influence of root-water uptake and gravitational flow. This work could be expanded to evaluate propagation of transpiration signals from trees at different positions within a catchment, including upslope trees like the one measured here, and trees closer to surface water sources to help quantify the role of transpiration in the generation of diel streamflow fluctuations. As with the work in the Arctic, expansion of this work could also incorporate joint processing of SP and ER data to constrain some of the uncertainty in each data type independently (e.g. depth of SP signal).

Finally, Chapter 5 documents the use of SP to evaluate groundwater flow, through time, in another hard-to-access site in Rocky Mountain National Park, in Colorado. We began with discussion of the various processes that generate current in the subsurface and can influence SP measurements and reviewed the use of coupling coefficients to account for these potential sources. We evaluated the potential effect of each of these sources on the field data. Using existing data sets, we determined that the contribution of electrochemical and thermoelectric
potentials were small relative to the potentials measured at the site, and then used the SP signals to determine changes in groundwater flow. We demonstrate how other auxiliary data can assist in interpretation of SP data, and describe what corroboratory measurements are needed to advance the use of field SP measurements in the greater hydrologic community.

Through the three projects contained herein, we have shown that SP can be an effective tool for mapping patterns of shallow groundwater and unsaturated zone flow to analyze near-surface processes at high spatial and temporal resolution. While each piece exists in a unique location, taken together these three projects exhibit the effectiveness of SP in quantifying a variety of hydrologic processes. SP is still emerging as a tool for use in hydrologic studies, and the work here showcases only a fraction of its potential (and also, potentially, the challenges) of using a tool that is sensitive to many processes. While some of the challenges such as non-uniqueness and confounding signal sources may prevent SP from ever independently becoming a beautiful swan, it is certainly no longer an ugly duckling; perhaps it is now more appropriately described as a perfectly imperfect bird coming-of-age, one in a flock of tools available to hydrologists.


97
Harr, R.D., McCorison, F.M., 1979. Initial effects of clearcut logging on size and timing of peak
doi:10.1029/WR015i001p00090

Counc. Canada.

Hauck, C., 2013. New Concepts in Geophysical Surveying and Data Interpretation for

Lett. 29. doi:10.1029/2002GL014995

Hauck, C., Böttcher, M., Maurer, H., 2011. A new model for estimating subsurface ice content
based on combined electrical and seismic data sets. Cryosph. 5, 453–468.


Hewlett, J.D., Hibbert, A.R., 1967. Factors affecting the response of small watershed to
precipitation in humid areas. For. Hydrol. 275–290.

Hilbich, C., Hauck, C., Hoelzle, M., Scherler, M., Schudel, L., Völksch, I., Vonder Mühll, D.,
Mäusbacher, R., 2008. Monitoring mountain permafrost evolution using electrical
resistivity tomography: A 7-year study of seasonal, annual, and long-term variations at


tree carbon assimilation rate using observed transpiration rates and needle sugar carbon

In search of experimental evidence for the biogoebattery. J. Geophys. Res.

Ikard, S.J., Revil, A., 2014. Self-potential monitoring of a thermal pulse advecting through a
preferential flow path. J. Hydrol. 519, 34–49. doi:10.1016/j.jhydrol.2014.07.001


APPENDIX A

MAKING IRIS COMPATIBLE RESISTIVITY CABLES

Field work in remote locations requires detailed consideration of equipment weight. I designed these lightweight, small electrode spacing cables for use with the IRIS Syscal Pro electrical resistivity control unit for the field work in Alaska described in Chapter 3, where all equipment was flown in by helicopter. The same cables were used for the field work described in Chapter 5, which required hiking in all equipment to the field site 4 miles from the trailhead and with 1300’ elevation gain. This tutorial walks through the process of making a cable to the specifications in Table A.1.

The cable is intended for use with 0.5-m spaced electrodes, but includes an extra 0.05m length between electrode takeouts for slack to work around trees, rocks and other often-encountered environmental obstacles. Cable lengths can be modified as needed to create a custom resistivity cable. A complete parts and tools list is included in Table A.2. Resistivity

Table A.1 Specifications of the resistivity cable detailed in these instructions. Lengths and spacing can be adjusted as needed, but will require recalculation of materials needed.

<table>
<thead>
<tr>
<th>Description</th>
<th>Specification</th>
</tr>
</thead>
<tbody>
<tr>
<td>ER transect length [m]:</td>
<td>23.5</td>
</tr>
<tr>
<td>Electrodes:</td>
<td>48</td>
</tr>
<tr>
<td>Spacing [m]:</td>
<td>0.55</td>
</tr>
<tr>
<td>Distance to first take out [m]:</td>
<td>10</td>
</tr>
<tr>
<td>Wires inside cable casing:</td>
<td>8</td>
</tr>
<tr>
<td>Required 75 m (250 ft) lengths of cable:</td>
<td>3</td>
</tr>
<tr>
<td>Maximum weight [kg]:</td>
<td>10.4</td>
</tr>
</tbody>
</table>
cables consist of a connector head compatible with the IRIS and a bundle of wires with enough length to reach the electrodes. The first step is to prepare the bundle of wires according to Figure A.1. Once the wires are ready, the next step is to attach a compatible connector head (Figures A.2 - A.9). The connectors on the IRIS Syscal Pro are part of the MIL-DTL-26482 Series 1, sometimes called MIL-C-26482 Series 1. The pin arrangement is called 22-55 because there are fifty-five connections for twenty-two gauge wire (Figure A.2). While each connector has fifty-five connections (pin or socket), only forty-eight are used by the IRIS Syscal Pro. The center pins BB – GG are unused (Table A.3). The two connections on the IRIS Syscal Pro (1-48 and 49-96) are pinned in reverse order to accommodate the double-ended cables typically used with the system.

Always double check pin connections before collecting data. One method is to hook the cable up to the IRIS and to run a resistance check. The resistance between the wires intended for the two electrodes in the resistance check should be almost zero when the two wire tips are touching.
Table A.2 Parts and tools required and recommended to build cables. Part numbers valid as of June 2014.

<table>
<thead>
<tr>
<th>Required Parts</th>
<th>Vendor</th>
<th>Part Number</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Required Parts</td>
<td>Newark*</td>
<td>85C4145</td>
<td>Connector head with crimp connection pins</td>
</tr>
<tr>
<td>Required Tools</td>
<td>Any</td>
<td>Any</td>
<td>Wire strippers</td>
</tr>
<tr>
<td>Required Tools</td>
<td>DMC~</td>
<td>AF8-TH1A</td>
<td>Crimp tool with TH1A Turret Head</td>
</tr>
<tr>
<td>Required Tools</td>
<td>DMC</td>
<td>DAK20B</td>
<td>Installing tool</td>
</tr>
<tr>
<td>Required Tools</td>
<td>DMC</td>
<td>DRK20B</td>
<td>Removal tool</td>
</tr>
<tr>
<td>Optional Parts</td>
<td>Newark</td>
<td>85C4146</td>
<td>Connector head with socket connection</td>
</tr>
<tr>
<td>Optional Parts</td>
<td>Newark</td>
<td>97B1571 or 61R9103</td>
<td>Spare pins (Red/Yellow/Black color bands)</td>
</tr>
<tr>
<td>Optional Parts</td>
<td>Newark</td>
<td>unknown</td>
<td>Spare socket (Red/Green/White color bands)</td>
</tr>
<tr>
<td>Optional Parts</td>
<td>Newark</td>
<td>24C5108</td>
<td>Metal connector cap (special order, 2+ required)</td>
</tr>
<tr>
<td>Additional Information</td>
<td>Any</td>
<td>Any</td>
<td>Shrink tubing to protect cables</td>
</tr>
</tbody>
</table>

* Newark, www.newark.com
^ McMaster-Carr, www.mcmaster.com
~ Daniels Manufacturing Corporation (DMC), www.dmctools.com

**Note:** I purchased high quality tools directly from Daniels Manufacturing Corporation (DMC, www.dmctools.com). These are also available from resellers such as Granger and Newark, but they are significantly more expensive. Cheaper versions are also available, but I broke them on my first pin. Helpful link for additional resources: http://www.dmctools.com/store/browser.asp
a) Cut cable to the following lengths:

- 36.85 m
- 32.45 m
- 28.05 m
- 23.65 m
- 19.25 m
- 14.85 m

b) From one end of each cable expose 4 m of wire using the Kevlar fiber inside the casing.

c) Clip exposed wires by the following amounts:

- 0 m white
- 0.55 m red
- 1.1 m orange
- 1.65 m yellow
- 2.2 m green
- 2.75 m blue
- 3.3 m brown
- 3.85 m black

d) Secure cables and exposed colored wires with zip ties.

Figure A.1 Cutting and assembly instructions for 23.5 m ER cable with 48 take outs at 0.55 m spacing. Full specifications provided in Table A.1, and complete materials list in Table A.2.
Figure A.2 Pin diagram of a 22-55 connector. This is as viewed from the face of the connector. The sockets arrangement on the IRIS Syscal Pro is a mirror image in order to accept the connector pins.

Table A.3 Pin addressing for the two connections on the IRIS. Note that the second connector is reverse order of the first to accommodate the double ended cables sold by IRIS

<table>
<thead>
<tr>
<th>Electrode</th>
<th>Pin</th>
<th>Electrode</th>
<th>Pin</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>AA</td>
<td>49</td>
<td>A</td>
</tr>
<tr>
<td>2</td>
<td>z</td>
<td>50</td>
<td>B</td>
</tr>
<tr>
<td>3</td>
<td>y</td>
<td>51</td>
<td>C</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>46</td>
<td>C</td>
<td>94</td>
<td>y</td>
</tr>
<tr>
<td>47</td>
<td>B</td>
<td>95</td>
<td>z</td>
</tr>
<tr>
<td>48</td>
<td>A</td>
<td>96</td>
<td>AA</td>
</tr>
</tbody>
</table>
Figure A.3 Wire preparation. Determine and cut the required length of wire for each electrode. If multi-wire cable is used, remove 2-3” of the outer casing (grey in photograph), without damaging the individual wire insulation. Strip each wire ¼” from the end (top). There should be no exposed wire once the stripped wire is inserted into the pin (bottom). When inserted, the wire should be visible through the small hole on the side of the pin.

Figure A.4 Tool preparation. Ensure the crimp tool is correctly adjusted for the connectors you are using. Select the connector size (red 20, if using listed parts), by releasing and turning the turret. Select the gauge of the wire to be crimped using the black knob on the right. Correctly setting this will ensure a good connection.
Figure A.5 Assemble connector. Ensure all pieces are included and in the right orientation on the wires before starting to insert pins in the head. Disassembly appears to be impossible. This photo does not have any wires connected, but they will need to be fed through the strain relief washer (not pictured), both bayonet parts, and the clear plastic alignment ring before assembly. Screw the pin housing into the bayonet base.

Figure A.6 Crimp connection. Place pin and wire into small hole on the back of the crimp tool. Top of pin (gold) should be flush with tool (silver). Squeeze handle of crimp tool completely. Once started, the locking mechanism will not release until the tool has fully cycled (crimped). The pin should now be crimped to the wire. Remove, and test connection by pulling on the wire.
Figure A.7 Insert pins. Grab the top of the crimped pin with the insertion tool, and push into the correct hole from the back of the connector. It will take some force, and you will hear/feel a few clicks. Refer to pin diagram (Figure A.1) and Table A.1 for exact placement. Work from the inside out to prevent a tangle mess, but remember positions BB – GG are not used (empty).

** The lighter color piece with flanges shown here will be installed later. **

Figure A.8 Pin removal. If necessary, pins can be removed after insertion using the extraction tool (red tool with spring). Simply place the extraction tool around the pin on the face of the connector. Once it is inserted as far as it will go, push down on the spring loaded collar (black) with some force. Remove the pin by pulling it out the rest of the way from the back of the connection.
Figure A.9 Strain relief. Clamp the washer to the flanges of the bayonet base using the provided hardware to reduce strain on individual wire/pins. Heat shrink tubing can be added prior to strain relief for additional protection. Attach optional metal protective cap.
APPENDIX B

BASIC SP FIELD GUIDE

Self potential (SP) measurements are not difficult to collect, but require some important details to collect quality measurements. The following document provides an overview of best practices when it comes to collecting SP data in the field. Additional information about survey design and drift correction can be found in The Self-Potential Method by Revil and Jardani (2013). Required and recommended equipment are outline in Table B.1.

SP data collection requires two non-polarizing electrodes, a wire to connect them, and a voltmeter somewhere along the length (Figures B.1 and B.2). One electrode is the reference and left stationary during the survey and the other is moved around to take measurements. An example field form is provided in Table B.2.

To begin, select a location for the reference electrode. It should be located in the most hydrologically and temperature-stable location possible. For example, if you are interested in the effects of river fluctuations, it is not advisable to locate the reference electrode directly adjacent to the river. A shady spot is preferred over direct sun, because the electrodes are sensitive to temperature variations. Once a reference location is selected, use a blub digger or small trowel to dig a hole the slightly larger than the diameter of the electrode to further protect the electrode from temperature variations. If repeat surveys will be collected, consider installing a short length of PVC and cap to mark the location and keep the ground from artificially drying out (Figure B.3).

At the start of a survey, connect all equipment as shown in Figure B.1. Measure and record the voltage difference (mV setting) and contact resistance (Ω setting) between the two electrodes, which is known as a tip-to-tip measurement (Figure B.3). Ideally the voltage
Table B.1 Items required for conducting an SP field survey.

<table>
<thead>
<tr>
<th>Required Equipment</th>
<th>Quantity</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Voltmeter</td>
<td>1</td>
<td>High internal impedance required. Fluke 87V is a common model.</td>
</tr>
<tr>
<td>Non-polarizing electrodes</td>
<td>2</td>
<td>Additional spare electrodes recommended</td>
</tr>
<tr>
<td>Reel with long wire to connect reference and roving electrodes</td>
<td>1</td>
<td>Length required depends on project. I used ~450 m of 22 gage wire, because that's what fit on my spool.</td>
</tr>
<tr>
<td>Banana plug junction (female/female)</td>
<td>1</td>
<td>Used to connect reference electrode to wire</td>
</tr>
<tr>
<td>Banana jumper wire (male/male, 60&quot; or 72&quot;)</td>
<td>1</td>
<td>Used to connect wire spool to voltmeter</td>
</tr>
<tr>
<td>GPS</td>
<td>1</td>
<td>Accuracy required depends on project needs.</td>
</tr>
<tr>
<td>Bulb digger or small trowel</td>
<td>1</td>
<td>Can be helpful to dig past surface for measurements</td>
</tr>
<tr>
<td>Field sheets and writing implements</td>
<td>Lots!</td>
<td>See example</td>
</tr>
<tr>
<td>Small toolkit</td>
<td>1</td>
<td>See recommended items below.</td>
</tr>
</tbody>
</table>

**Recommended toolkit items:** wire strippers, small screw driver (to open volt meter), replacement fuses, spare banana plugs, spare banana jumper wire, spare banana plug junctions, wire screw caps (wire nuts), electrical tape, spare 9V battery (volt meter), spare AA (GPS), pens, spare volt meter, short length of PVC with screw cap (for long-term reference station installation)
Figure B.1 The SP field equipment I used in the field.
difference between the reference and measurement electrodes is 0 mV, but anything smaller than $\pm 2$ mV is acceptable. A potential difference greater than 2 mV suggests that something is wrong with an electrode, and it should not be used. Tip-to-tip measurements should be repeated again at regular intervals (~20 measurements or ~1 hour, depending how much measurements are drifting) to correct for drift as described in Revil and Jardani (2013). Once completed, place the reference electrode snuggly in the hole you created earlier. Consider including some sort of strain relief on the wire (e.g. tie to tree, loosely tape off some wire loops), so that the reference electrode is not accidently pulled out of the ground while collecting measurements. It will need to be removed periodically for tip-to-tip measurements.

Collect at least three measurements at each location to estimate the measurement error, or use a voltmeter with an averaging feature. Make sure to collect a tip-to-tip measurement prior to disconnecting the equipment at the end of the survey.
Figure B.2 Close up of the field equipment and wire connections. a) Voltmeter and wire reel, b) close up of banana-banana jumper wire plugged into the banana socket, which was installed and connected to the interior end of the wire prior to spooling, c) the reference cable plugged into the COM port of the voltmeter and voltmeter turned to mV for measurement, and d) a socket-socket connection that connects the banana plug of the reference electrode to the banana plug of the exposed wire end.
Table B.2 Sample field form for recording SP data, including time and location. Modify as needed for specific projects.

<table>
<thead>
<tr>
<th>Location:</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Collected by:</td>
<td></td>
</tr>
<tr>
<td>Date:</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Time (hh:mm)</th>
<th>Location (m)</th>
<th>Normal Voltage (mV)</th>
<th>Reverse Volt (mv)</th>
<th>Resis. (Kohm)</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>used for drift correction</td>
<td>every measurement</td>
<td>every measurement</td>
<td>every 10-23</td>
<td>every 10-23</td>
<td>anything of note: obstructions, changes, etc.</td>
</tr>
</tbody>
</table>
Figure B.3 Demonstration of a tip-to-tip reference measurement taken between the reference and roving electrode. Additionally, a semi-permanent reference station has been set up in the bottom half of the photo using a 10” length of PVC and sealing cap.
APPENDIX C

GETTING STARTED MODELING SP IN COMSOL 5.2a

The following tutorial is written using COMSOL 5.2a, and requires the Subsurface Flow and ACDC modules. Forward modeling of streaming potentials in COMSOL requires two parts: modeling water movement using Subsurface Flow Module and modeling the electrical problem using the ACDC Module. This tutorial is not exhaustive, but demonstrates the basics of setting up a 2-D, 10 x 3 m, steady-state SP model in COMSOL. The values included here are for reference only, use values that best fit your problem. **Bold font** indicates setting, buttons, or other text within COMSOL.

1. Begin by opening COMSOL 5.2a and select **Model Wizard** (Figure C.1a). Select **2D** model (Figure C.1b), and then add the **Darcy’s Law (dl)** and **Electric Currents (ec)** physics modules and continue by pressing **Study** (Figure C.1c). Select **Stationary** study type and click **Done** (Figure C.1d). A new project untitled.mph file will open with at least three nested windows: **Model Builder, Settings** and **Graphics** (Figure C.2). Additional windows may open, depending on how the layout when the program was last closed.

2. The **Model Builder** tree is a list of all of the objects, boundary conditions, actions, solvers, etc. included in your COMSOL model (Figure C.2a). Eventually, when you press **Compute** the model will run consecutively through each item in the **Model Builder** tree. Therefore, with some exceptions, to reference something (e.g. variable, geometry), the definition must occur higher in the **Model Builder** tree.
Figure C.1 Starting a COMSOL model: a) Model building method selection, b) spatial dimension selection, c) physics modules selection and d) study time domain selection.
Figure C.2 COMSOL workspace consisting of a) the Model Builder tree indicating the order of operations, b) settings window providing more options for each of the objects in the Model Builder and c) the Workspace which allows for manipulation and viewing of objects in the Model Builder.
3. For example, to calculate SP voltages, we must first calculate flow velocities and then couple those to the electrical problem. Therefore, **Darcy’s Law (dl)** must be above **Electric Currents (ec)** in the **Model Builder** tree (Figure C.2a). If it is not, click and drag **Darcy’s Law (dl)** above **Electric Currents (ec)** in the tree. The characters in the parentheses of list items describes the object’s handle, or short hand code. We will use the handles to link the two sets of physics later in this example.

4. Now we build the model, starting with the geometry. Make a rectangle by right-clicking **Geometry** and selecting **Rectangle** (Figure C.3a). The size of the rectangle can be modified in the **Settings** window to 10 m x 3 m (Figure C.3b).

5. After the geometry is described, it must be assigned a material, which is a collection of physical parameters such as density, porosity and electrical conductivity permeability (Figure C.3c). You can or create your own (not used in this tutorial):

   **Right click Materials → Blank Material**

   or select from a library of materials:

   **Materials → Add Material Materials → Built-In Materials → Water, liquid**

6. If needed, assign the rectangular model domain a **Material** by selecting **Water, liquid (mat1)** in the tree and then clicking on the domain it in the **Graphics** window, or selecting from **Geometric Entity Selection** in the **Settings** menu. Small red stop signs indicate that three of the material properties are lacking and will need to be added before the model can
Figure C.3 Model building menus within COMSOL including a) one method of building model geometry, b) further defining geometry parameters, c) adding built-in material properties and d) assigning additional material parameters.
Compute. Fill in the remaining parameters (Figure C.3d): Porosity (“0.3”), Permeability (“10^-6[m^2]”) and Relative permittivity (“1”). Include units by using square brackets. Including units is recommended, as COMSOL will highlight with yellow-brown text units that don’t work out properly. NOTE: Incorrect units will not prevent your model from running.

7. Permeability is the physical property defined by a Material. If you prefer to use hydraulic conductivity, it is hidden deep within menus (Figures C.4a and C.4b):

Darcy’s law (dl) → Fluid and Matric Properties (Settings) → Matrix Properties → Permeability model (drop down) → Hydraulic Conductivity → K → 2.94e-4[m/s]

8. Now that the geometry and material properties are assigned, we define the boundary conditions of the groundwater flow model. To begin, disable the Electric Currents (ec) module by selecting the item in the Model Builder and pressing F3 or selecting Disable in the right-click menu.

9. Right-click Darcy’s Law (dl) header to show all of the boundary conditions available in the module (Figure C.4c). Select Hydraulic Head. Assign the desired value of hydraulic head (0.1 m) for the boundary in the Settings window (Figure C.5a). Select the desired domain boundary (left) by clicking on it in the Graphics window (Figure C.5a). Repeat the procedure to create a second Hydraulic Head boundary for the right boundary (0.0 m).
Figure C.4 Preparing the groundwater flow model. a) Defining fluid and matrix properties, including b) hydraulic conductivity and c) hydraulic head boundary conditions.
Figure C.5 Preparing the groundwater flow problem. a) Assigning hydraulic head boundary conditions, and b) the default plot of the groundwater flow problem solution.
10. Once the model geometry, material properties and boundary conditions are defined, we can

**Compute** the solution using F5, or the button in the top menu ribbon **Home → Compute**.

The default result, pressure in Pa, will automatically display (Figure C.5b). There are many ways to modify the graphics, which are not covered here.

11. Now we can use the velocities calculated in the **Darcy’s Law (dl)** module as a contributing boundary condition in the electrical problem. First, re-enable the **Electric Currents (ec)** module using **F4 or Enable** in the right-click menu.

12. Streaming potential is generated by currents generated as water moves, described by:

\[
J = \tilde{Q}_v U
\]  

(C.1)

where \(J\) is the distribution of current sources, \(\tilde{Q}_v\) is the effective excess charge [C m\(^{-3}\)] and \(U\) is the Darcy velocity in [m/s].

13. Begin by defining a new parameter called \(Qv\). Right click on **Global Definitions** header, and select the **Parameter** subheader. Create a **Parameter** with the **Name** “\(Qv\)”, **Expression** “0.1 [C/m\(^3\)]” and **Description** “Effective Excess charge” (Figure C.6a)

14. The distribution of source currents is defined as an **External Current Density** condition, added under the **Electrical Currents (ec)** header (Figure C6.b). Once the boundary condition has been added, make sure the rectangular domain is selected under **Settings → Domain Selection**. Just because a boundary condition, or any other feature, appears in the **Model Builder**, does not necessarily mean it has been applied to a domain.
Figure C.6 Adding the electrical module to the COMSOL model. a) Defining a global parameter, b) creating a spatial condition, and c) linking the spatial condition to the solutions of the earlier groundwater flow problem.
16. To access the Darcy velocity, we use the handle for the Darcy’s Law module “dl” combined with the handles for the directional components of Darcy velocity “u” (horizontal) and “v” (vertical): dl.u and dl.v. Qv was assigned in step 12, and can be referenced by the name we assigned it: Qv. Type Qv*dl.u and Qv*dl.v into the External current density text boxes (Figure C.6c).

17. Finally, a fixed voltage must be assigned. This can be accomplished using a point boundary condition, accessed from Electrical Currents (ec) → Points → Ground. Select one point, upper left, to be the ground. All voltages will be calculated relative to this point. You must have a fixed value somewhere in the model. If not, you will receive a “Failed to find a solution” error. While there are many reason you could receive this message, ensure that you have a Ground defined.

18. Now you can Compute your model (Figure C.7)

19. The COMSOL model described here can be exported to produce the following Matlab script:
function out = model
  
  % AppendixC.m
  
  % Model exported on Jun 18 2017, 13:30 by COMSOL 5.2.1.262.

  import com.comsol.model.*
  import com.comsol.model.util.*

  model = ModelUtil.create('Model');
  model.modelPath('C:\Users\evoytek\Dropbox\Dissertation\Appendix');
  model.comments(['Untitled
  
  ']);

  model.param.set('Qv', '0.1 [C/m^3]', 'Effective excess charge');

  model.modelNode.create('comp1');

  model.geom.create('geom1', 2);
  model.mesh.create('mesh1', 'geom1');

  model.geom('geom1').create('r1', 'Rectangle');
  model.geom('geom1').feature('r1').set('size', ['10' '3']);
  model.geom('geom1').feature('r1').set('pos', ['1' '1']);
  model.geom('geom1').run;

  model.material.create('mat1', 'Common', 'comp1');
  model.material('mat1').propertyGroup('def').func.create('eta', 'Piecewise');
  model.material('mat1').propertyGroup('def').func.create('Cp', 'Piecewise');
  model.material('mat1').propertyGroup('def').func.create('rho', 'Piecewise');
  model.material('mat1').propertyGroup('def').func.create('k', 'Piecewise');
  model.material('mat1').propertyGroup('def').func.create('cs', 'Interpolation');

  model.physics.create('dl', 'DarcysLaw', 'geom1');
  model.physics('dl').create('hh1', 'HydraulicHead', 1);
  model.physics('dl').feature('hh1').selection.set([1]);
  model.physics('dl').create('hh2', 'HydraulicHead', 1);
  model.physics('dl').feature('hh2').selection.set([4]);
  model.physics.create('ec', 'ConductiveMedia', 'geom1');
  model.physics('ec').create('ecd1', 'ExternalCurrentDensity', 2);
  model.physics('ec').feature('ecd1').selection.set([1]);

  model.view('view1').axis.set('abstractviewxscale', '0.012687427923083305');
  model.view('view1').axis.set('abstractviewtratio', '1.346021056175232');
  model.view('view1').axis.set('abstractviewlratio', '-0.05000000074505806');
  model.view('view1').axis.set('abstractviewyscale', '0.012687429785728455');
  model.view('view1').axis.set('abstractviewrratio', '0.05000000074505806');
  model.view('view1').axis.set('abstractviewbratio', '-1.3460208177566528');
  model.view('view1').axis.set('ymin', '25.933595657348633');
  model.view('view1').axis.set('xmax', '11.25');
  model.view('view1').axis.set('ymin', '-20.93359375');
  model.view('view1').axis.set('xmin', '0.7499995231628418');
model.material('mat1').label('Water, liquid');
model.material('mat1').set('family', 'water');
model.material('mat1').propertyGroup('def').func('eta').set('pieces', ['273.15' '413.15' '1.3799566804-0.021224019151*T^1+1.3604562827E-4*T^2-4.6454090319E-7*T^3+8.9042735735E-10*T^4-9.0790692686E-13*T^5+3.8457331488E-16*T^6'; '413.15' '553.75' '0.0040123578-2.10746715E-5*T^1+3.85772275E-8*T^2-5.33180564E-11*T^3+3.38796247E-14*T^4-6.68015684E-17*T^5+2.122401915E-19*T^6']);
model.material('mat1').propertyGroup('def').func('eta').set('arg', 'T');
model.material('mat1').propertyGroup('def').func('Cp').set('pieces', ['273.15' '553.75' '12100.1471-80.4072879*T^1+0.309866854*T^2-5.38186884E-4*T^3+3.62536437E-7*T^4']);
model.material('mat1').propertyGroup('def').func('Cp').set('arg', 'T');
model.material('mat1').propertyGroup('def').func('rho').set('pieces', ['273.15' '553.75' '838.466135+1.40050603*T^1-0.0030112376*T^2+3.71822313E-7*T^3']);
model.material('mat1').propertyGroup('def').func('rho').set('arg', 'T');
model.material('mat1').propertyGroup('def').func('k').set('pieces', ['273.15' '553.75' '0.869083936+0.00894880345*T^1-1.58366345E-5*T^2+7.97543259E-9*T^3']);
model.material('mat1').propertyGroup('def').func('k').set('arg', 'T');
model.material('mat1').propertyGroup('def').func('cs').set('interp', 'piecewisecubic');
model.material('mat1').propertyGroup('def').func('cs').set('table', ['273' '1403'; 278' '1427'; 283' '1447'; 293' '1481'; 303' '1507'; 313' '1526'; 323' '1541'; 333' '1552'; 343' '1555'; 353' '1555'; 363' '1550'; 373' '1543']);
model.material('mat1').propertyGroup('def').set('dynamicviscosity', 'eta(T[K])[Pa*s]');
model.material('mat1').propertyGroup('def').set('ratioofspecificheat', '1.0');
model.material('mat1').propertyGroup('def').set('electricconductivity', '{5.5e-6[S/m] 0 0 0 5.5e-6[S/m] 0 0 0 5.5e-6[S/m]}');
model.material('mat1').propertyGroup('def').set('heatcapacity', 'Cp(T[K])[J/(kg*K)]');
model.material('mat1').propertyGroup('def').set('density', 'rho(T[K])[kg/m^3]');
model.material('mat1').propertyGroup('def').set('thermalconductivity', '{k(T[K])[W/(m*K)] 0 0 0 k(T[K])[W/(m*K)] 0 0 0 k(T[K])[W/(m*K)]}');
model.material('mat1').propertyGroup('def').set('soundspeed', 'cs(T[K])[m/s]');
model.material('mat1').propertyGroup('def').set('porosity', '.3');
model.material('mat1').propertyGroup('def').set('hydraulicpermeability', '{10^-6[m^2] 0 0 10^-6[m^2] 0 0 10^-6[m^2]}');
model.material('mat1').propertyGroup('def').set('relpermittivity', '{1 0 0 1 0 0 0 1}');
model.material('mat1').propertyGroup('def').addInput('temperature');
model.physics('dl').feature('dlm1').set('ktype', 'conductivity');
model.physics('dl').feature('hh1').set('H0', '0.1');
model.physics('ec').feature('ecdl1').set('Je', '{Qv*dl.u; Qv*dl.v; 0}');
model.mesh('mesh1').run;
model.study.create('std1');
model.sol.create('sol1');
model.sol('sol1').study('std1');
model.sol('sol1').attach('std1');
model.sol('sol1').create('st1', 'StudyStep');
model.sol('sol1').create('v1', 'Variables');
model.sol('sol1').create('s1', 'Stationary');
model.sol('sol1').feature('s1').create('fc1', 'FullyCoupled');
model.sol('sol1').feature('s1').feature.remove('fcDef');

model.result.create('pg1', 'PlotGroup2D');
model.result.create('pg2', 'PlotGroup2D');
model.result('pg1').create('surf1', 'Surface');
model.result('pg2').create('surf1', 'Surface');

model.sol('sol1').attach('std1');
model.sol('sol1').runAll;

model.result('pg1').label('Pressure (dl)');
model.result('pg1').feature('surf1').label('Surface');
model.result('pg1').feature('surf1').set('resolution', 'normal');
model.result('pg2').label('Electric Potential (ec)');
model.result('pg2').set('frametype', 'spatial');
model.result('pg2').feature('surf1').set('descr', 'Electric potential');
model.result('pg2').feature('surf1').set('unit', 'V');
model.result('pg2').feature('surf1').set('expr', 'V');
model.result('pg2').feature('surf1').set('resolution', 'normal');

out = model;
APPENDIX D

MATLAB THREE-POINT PROBLEM SCRIPT

The following code will produce a plot (Figure D.1) with the direction, and relative magnitudes of gradients calculated from a three-point problem. It can be used for transducer or tensiometer data. If used with SP data (in units of mV), set elevations to 0. Code was created in Matlab 2016b:

```matlab
function three_point_scaled(xlocs,ylocs,elevations, n)
  
  % Function for plotting angle and relative slope of gradient from three know elevation points
  
  % INPUT: % xloc - x location of three measurement points in meters, 1x3 matrix
  % yloc - y location of three measurement points in meters, 1x3 matrix
  % elev - elev or head through time at three points described by % xloc and yloc, nx3 matrix, where n is the number of data points
  % n - scaling factor for plotting, increase or decrease as needed
  
  example
  % xloc = [2 3 4];
  % yloc = [1,0,5];
  % elev = [cos(0:.02:1);.7:.01:1.2;1.2*ones(1,51)];
  % n = 3;
  %
  % plot_circle_tensio_grads(xloc,yloc,elev, 3)
  
  output
  % Figure(101) - A plot of relative gradients and diretion for
  % Created by Emily Voytek, 5/20/2017, with help from:
  % http://serc.carleton.edu/quantskills/activities/three_point.html

  % Declare figure and clear
  figure(101), clf

  % Create Colormap that is the length of the data file
  cmap =colormap((parula(size(elevations,1))));

  % Plot three locations
  plot(xloc,yloc,'ro'), axis equal, hold on

  % Determine geometric center of the three points, and approximate scale
  avg_xloc = mean(xloc);
  avg_yloc = mean(yloc);
  std_xloc = std(xloc);
  std_yloc=std(yloc);
```

138
% Plot center point
plot(avg_xloc,avg_yloc);

A = [xlocs',ylocs',ones(3,1)];

% Calculate direction and relative magnitude for each time step
for i = 1:size(elevations,1)
    z = [elevations(i,:)/10]';
    u = A\z;
    hold on
    deg = atan2d(-u(1),-u(2));
    dip = atan(sqrt((u(1)^2)+(u(2)^2)))*(180/pi);
    scale = (dip/90)*mean([std_xloc,std_yloc])*n;
    plot([0, scale*sind(deg)/2]+avg_xloc,...
         [0,scale*cosd(deg)/2]+avg_yloc,...
         'Color',cmap(i,:))
end
xlabel('distance [m]'), ylabel('distance [m]')
caxis([1,size(elevations,1)])
c = colorbar('northoutside');
c.Label.String = 'data point';
end

Figure D.1 Data plot produced by sample data provided in three_point_scaled.m