Assessment of vadose zone solute transport under a potato field by a 19 month time-lapse cross-hole resistivity imaging survey

By

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B. Applied Physics, Jilin University, China, 2013

A Thesis Submitted in Partial Fulfillment of
The Requirements for the Degree of

Master of Science in Engineering

in the Graduate Academic Unit of Electrical and Computer Engineering

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This thesis is submitted to the
Dean of Graduate Studies

THE UNIVERSITY OF NEW BRUNSWICK

April, 2017

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Abstract
Nitrate is a necessary nutrient for crops, however high concentrations of nitrate in surface water and groundwater can negatively affect aquatic ecosystems and human health. 3D cross-hole Electrical Resistivity Imaging (ERI) has been used to investigate the percolation of a conductive tracer (KCl) through a 17 m thick vadose zone as a proxy for the transport of nitrate under natural recharge conditions.

Post-tracer surveys indicate that tracer movement has slowed significantly by early May, 2015 (about one month after tracer application), at the end of snow melt. The shallow conductivity anomaly produced by the tracer diminished significantly over the winter and spring of 2016, but showed little evidence of bulk matrix flow below approximately 6 m depth (top of the bedrock). It is speculated that fractures in the bedrock, too thin to be resolved by the ERI survey, conveyed tracer downward. After 18.5 months, there is no ERI evidence of tracer migrating through the matrix below approximately 6 m.
Acknowledgements

First and foremost, I would like to thank my supervisors, Dr. Brent Petersen, Dr. Karl Butler, and Dr. Serban Danielescu, for their continual supply of patience, advice, and support during this project. I would like to thank my family and friends for their support and encouragement that they provided throughout the duration my studies.

I would also like to give thanks to those who generously provided support during field and lab activities, Eric Mott (UNB), Mark Grimmett (AAFC), and Keenan Lamb (UNB). I would like to thank the staff of the Agriculture and Agri-Food Canada Harrington Research Farm for their assistance with field activities. And I would also like to thank University of New Brunswick for providing me the opportunity to study here, and benefit from the use of the facilities. Funding for this research was provided by Agri-Food and Agriculture Canada through the Growing Forward 2 project lead by Dr. Bernie Zebarth, and Canadian Water Network (Origin, Occurrence and Fate of Nitrate in Sedimentary Bedrock Groundwater in the Maritimes) lead by Dr. Cathy Ryan.
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List of Symbols, Nomenclature, and Abbreviations

$\rho_a$ = apparent resistivity (Ωm)

$\Delta V$ = potential difference between potential electrodes (V)

$I$ = injected current (A)

$|AM|, |BM|, |BN|$, and $|AN|$ = distances between injection and potential electrodes (m)

$k$ = geometric factor (m)

$\varphi$ = porosity (-)

$\rho_{\text{bulk}}$ = the bulk resistivity of the rock (Ωm)

$\rho_{\text{fluid}}$ = the resistivity of the pore fluid (Ωm)

$\rho_t$ = the measured resistivity at temperature $t$ (Ωm)

$\rho_{25}$ = the resistivity at $25^\circ$C (Ωm)

$F$ = electrical formation factor (-)

$S$ = water saturation (-)

$a$, $m$, and $n$ = empirically derived constants in Archie’s Law (-)

$\alpha$ = temperature coefficient of resistivity (0.0187/°C)

$t$ = ambient temperature in Celsius (°C)

$i$ = the $i^{th}$ element in the data

$\Phi$ = global objective function

$\Phi_m$ = model objective function

$\Phi_d$ = data misfit function

$\theta$ = trade-off parameter that controls the relative importance of the model smoothness through the model objective function and the data misfit function (-)
\( m_i = \) model element, resistivity of the \( i^{\text{th}} \) cell in the resistivity model (-)

\( \hat{d}_i = \) predicted data (-)

\( d_i = \) observed data (-)

\( e_i = \) data error attributed to noise (-)

\( p = \) number of data

\( r = \) vector indicates the cell location in a mesh

\( V = \) the integrating range

\( w_s = \) user-given generalized weighting function of the smallest model

\( w_x, w_y, \) and \( w_z = \) user-given generalized weighting functions in three spatial directions

\( \alpha_s, \alpha_x, \alpha_y \) and \( \alpha_z = \) coefficients that affect the relative importance of different components in the objective function in each axis

\( m = \) is the recovered model

\( m_0 = \) the reference model.

\( l_z = \) the fraction of total current flows below a certain depth, \( z, \) in a homogenous half space

\( l_z' = \) the fraction of total current flows beyond a depth \( z \) and \(-z, \) in a homogenous full space

\( L = \) the separation between the two current electrodes

\( \text{TDS} = \) total dissolved solids

\( \text{ERI} = \) electrical resistivity imaging

\( \text{DOI} = \) depth of investigation

\( \text{SWB} = \) soil water balance
1. Introduction

1.1 Background

Prince Edward Island has a total land area of 0.57 million hectares, with approximately 0.24 million hectares cleared as agricultural land, and potatoes are the single largest agricultural commodity (Government of Prince Edward Island, 2014). PEI accounts for only 0.5% of agricultural cropland in Canada, and yet in 2011, it accounted for approximately 24% of the land in potato production (Statistics Canada, 2013). The potato industry contributes more than 10% of PEI's gross domestic product (BDO and Cammac Economics Limited, 2011). There are three specific markets for Prince Edward Island potatoes: table potatoes, processing potatoes (e.g. French fries), and seed potatoes. Approximately 60% of Prince Edward Island potatoes are destined for processing (The Government of Prince Edward Island, 2015), because of their suitability for French fries (Desroches et al., 2008). Processing potatoes require a significant amount of fertilizer to grow, but are poor at absorbing fertilizer. As a consequence, after fertilizer is applied, some of the contained nitrate will be leached (Zebardh et al., 2014) below the root zone by infiltrating precipitation and snow melt and carried to the groundwater zone, from where it may be drawn towards pumping wells or discharged to surface waters.

Almost nine million Canadians rely on groundwater as their domestic fresh water supply (Environment and Climate Change Canada, 2014). In PEI, groundwater is the only source for drinking water and contributes up to 80% of water in streams and rivers (Zebardh et al., 2014). This means the islanders are at a risk for nitrate contamination in their drinking water from agricultural runoff. In some regions of PEI,
about 20% of domestic wells tested exceeded the maximum acceptable concentration of 10 mg NO₃-N/L (Health Canada, 2012) for drinking water (Liao et al., 2015). Novaczek et al. (2008) stated that “Nitrates in drinking water can be a health hazard when present at relatively high concentrations (over 10 ppm). At much lower concentrations, nitrate interacts with other pollutants such as trace mixtures of pesticides to produce health impacts”. The spatial distribution of elevated groundwater nitrate concentrations is consistent with intensive potato production systems as the primary source of groundwater nitrate contamination (Zebarth et al., 2014). In addition, nitrate contamination can also negatively affect the aquatic environment (Galloway et al., 2004), by leading to eutrophication and hypertrophic conditions, that in turn and lead to fish kills and algal blooms. In PEI an increased number of anoxic events in the nearshore coastal waters have been reported in the recent past (Government of Prince Edward Island, 2016). It is, therefore, important to understand the factors affecting nitrate transport and fate in the subsurface.

In this research, a tracer test was monitored during an 18.5-month period, to study nitrate transportation in the subsurface. The infiltration of water containing a conductive tracer causes changes in ground resistivity distribution. Thus, monitoring the resistivity changes over time allows one to infer water pathways and their rate of movement. Since nitrate is highly soluble, it is readily leached from soils and is mobile in groundwater, so the movement of nitrate should follow the same path as the infiltration of water, although its progress may be delayed by chemical reactions, such as denitrification, which in PEI is considered to be significant only in very localized areas (Zebarth et al., 2014). By learning
tracer transport, we will be able to understand the nitrate transport, and use the knowledge to improve groundwater quality.

1.2 Background knowledge of electrical resistivity imaging

Electrical resistivity is a physical property representing a material's intrinsic resistance to electrical current flow. Multi-electrode electrical resistivity imaging (ERI) has been commercially available for more than two decades. Apart from its use in mapping spatial variations in subsurface geological material and pore water content, it has also proven effective for monitoring dynamic processes such as groundwater infiltration/percolation (Daily and Ramirez, 1992), movement of tracer (Binley et al., 2002), and salt water intrusion (Barlow and Reichard, 2009). In sediments and sedimentary rocks, electrical current moves primarily by the flow of ions through water in pores and fractures. There are four main factors that determine the material electrical resistivities: porosity, water-saturation, pore water chemistry, and clay content. Pore water chemistry is an important factor because the concentrations and mobilities of ions in solution determine the resistivity of the pore water. The presence of clay lowers the bulk resistivity of soil or rock electrical resistivity because clay minerals have an electrically active surface layer, which provides high surface conductance, and it has a relatively high porosity (Bai et al., 2013). The effects of porosity and pore water salinity on bulk resistivity are addressed quantitatively in Chapter 3. There are two main lithologies in PEI: sandstone and shale. Shales typically exhibit lower resistivities than sandstone, as a consequence of the high clay content.
To explore the resistivity of shallow ground, electrodes have often been placed on the surface. However, to obtain better resolution and sensitivity at depth, electrodes may be placed in boreholes (Loke and Barker, 1995). The region of interest in this research is the deep vadose zone below a potato field in PEI, extending between the root zone (at about 1 metre depth) and the water table (at about 17 m depth in the summer). Thus, 3D cross-hole ERI has been employed in this research, as the electrical resistivity changes can reveal the groundwater transport in the subsurface.

1.3 Literature review

Electrical surveys sensitive to resistivity variations have been widely used for the detection of structure or dynamic processes in the subsurface for more than two decades. In a pioneering study, Daily and Ramirez (1992) used two-dimensional cross-hole ERI to monitor point and line source tracer infiltration through a sand and gravel-dominated vadose zone that was approximately 20 m thick. The ERI inversions showed spatial and temporal variations in resistivity very similar to those predicted by the flow models that were simulated for the two cases. They concluded that electrical resistivity imaging was a promising method of monitoring infiltration/percolation through the vadose zone.

Zhou and Greenhalgh (2000) investigated the advantages and disadvantages of pole-pole, pole-dipole, dipole-pole and dipole-dipole arrays, by examination of a dipping conductive strip model and a dislocated fault model. They showed that the dipole-dipole configuration, compared with the pole-pole array, may eliminate the effect of remote electrodes, and yield satisfactory images even with 20% noise-contaminated data. They also showed that the configurations which have either both current electrodes or both
potential electrodes in the same borehole, may cause a singularity problem in data
acquisition, which can cause low readings of the potential or potential difference in cross-
hole surveying, which may cause data to be obscured by background noise.

Binley et al. (2002) highlighted ERI as being effective in investigating dynamic changes in
the vadose zone because the bulk electrical resistivity of subsurface is sensitive to some
key hydrogeophysical parameters, i.e., porosity, moisture content, pore water resistivity,
and tortuosity (detailed explanation will be provided in Chapter 3). ERI allows for
numerous configurations (e.g. borehole and surface arrays) that can be catered
specifically to the requirements of the survey (i.e. greater electrode spacing allows for
greater depth of investigation, but reduced resolution).

Wilkinson et al. (2010) installed vertical electrode arrays in boreholes at a contaminated
land-site to monitor the changes of groundwater quality, as the resistivity imaging can act
as a surrogate monitoring technology for tracking changes in contaminant concentrations
at much higher spatial and temporal resolution than manual intrusive investigations.
They utilized dipole-dipole measurements, in which current flow and potential flow were
both conducted cross-hole, as these provide better signal-to-noise characteristics and
greater image resolution than configurations with in-hole current flow and potential
measurements (Bing and Greenhalgh, 2000). Wilkinson et al. (2010), like LaBrecque et
al. (1996) also suggested that the aspect ratio of the panel (borehole spacing/depth)
should be smaller than 0.75 to provide sufficient image resolution in cross-hole ERI. The
capabilities of geoelectrical monitoring were demonstrated by a saline tracer test
monitored in near real-time.
A sequence of ERI models need to be obtained at different times for observing dynamic changes in the subsurface. Hayley et al. (2011) introduced time-lapsed method as a powerful tool as it offers the potential for achieving inversion results with increased fidelity through the inclusion of complementary information from multiple time-steps. This inclusion of complementary information can reduce the need for spatial smoothing, without adding inversion artifacts to the resulting images. In this thesis, the dynamic changes in the subsurface are observed by a cascaded time-lapsed method, which takes the first inversion result as the reference model and the starting model in the inversion of the second dataset using the same code as used for the independent inversions (Hayley et al., 2011).

Hayley et al. (2010) stated that temperature variations during time-lapse ERI surveys introduce changes in electrical conductivity. Since the goal of the time-lapse ERI survey is usually to image changes in electrical conductivity due to changes in saturation or pore water salinity, they introduced a method to compensate the effect of temperature variations in time-lapsed ERI models. They firstly inverted the uncompensated ERI data, then adjusted the inversion model to a standard temperature image (Hayley et al., 2007). Then they performed forward simulations using the uncompensated inversion and the standard temperature equivalent model. The temperature-compensated simulated resistance data are subtracted from the uncompensated simulated resistance data, forming data correction terms. The data correction terms then are subtracted from the measured data to yield temperature-compensated data. They concluded that by performing the correction to the ERI data as these steps, the temperature effect can be effectively removed from time-lapsed ERI inversions.
1.4 Research objectives

In this research, a conductive tracer has been used as a proxy for nitrate to enhance our ability to track solute movement electrically. The objective of this research was to investigate the solute transport in vadose zone using electrical resistivity methods, with a focus on answering the following questions:

- Does solute percolate homogeneously as a plug, or is there evidence of hydraulic heterogeneity giving rise to preferential flow paths or to perched water table conditions?
- Can both matrix and fracture-dominated flow be evidenced, and which of the two is the dominant form of transport?
- How long does it take, under natural recharge conditions, for solute applied on surface to reach the water table?

Understanding the extent of hydraulic heterogeneity and the relative proportion of fracture versus matrix flow is important for interpreting discretely sampled data (e.g. geochemistry, moisture content, perched water levels) from boreholes and groundwater monitoring instrumentation, and for designing monitoring programs and developing solute transport models. The travel time information is crucial to understand how long it would take changes in farming practices at the surface to impact nitrate loading to the underlying aquifer.
2. Study Site Description

2.1 Study location

The study site is located at the 330 acre Harrington Research Station (Figure 2-1), an experimental farm operated by Agriculture and Agri-Food Canada, approximately 11 km north of Charlottetown airport. The study was carried out in Subsection A of Field 355, at the northwestern corner of the field (Figure 2-2), where topography was relatively flat, thereby favoring vertical infiltration over surface run-off or lateral flow.

Figure 2-1: Satellite photo of Prince Edward Island (Google Earth, 2016). The City of Charlottetown and the location of the Agriculture and Agri-Food Canada Harrington Research Farm are shown. Harrington is located 11 km northwest of Charlottetown airport.

The cropping system in Field 355 generally follows a standard three year rotation: potato-barley-clover. Potatoes were planted in Subsection A in 2014, however there was no crop planted (and no fertilizer applied) in the study area during the 2015 and 2016 growing seasons while monitoring was ongoing. Table 2-1 shows the crop rotations, applied fertilizer, and applied nitrogen concentration for 2011-2014 in
Subsection A. The fertilizer usually contains nitrogen (N), phosphorus (P), and potassium (K), labeled in xx-xx-xx format, indicating the element percentage, in the fertilizer. Generally, the nitrogen is applied as ammonium, or urea, phosphorus as phosphorus pentoxide (P₂O₅), and potassium is in form of potassium chloride (KCl).

Figure 2-2: Topographic map of Field 355 in Harrington Research Farm. Contour lines at intervals of 0.25 m are shown in white. The red box indicates the study area.

Table 2-1: Field 355 Subsection A crop rotations

<table>
<thead>
<tr>
<th>Year</th>
<th>Crop</th>
<th>Applied N (kg/ha)</th>
<th>Applied N (kg/m²)</th>
<th>Applied fertilizer(N-P-K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2011</td>
<td>Potatoes</td>
<td>229</td>
<td>0.0229</td>
<td>17-17-17</td>
</tr>
<tr>
<td>2012</td>
<td>Barley</td>
<td>51</td>
<td>0.0051</td>
<td>34-0-0</td>
</tr>
<tr>
<td>2013</td>
<td>Red Clover</td>
<td>0</td>
<td>0.0000</td>
<td>/</td>
</tr>
<tr>
<td>2014</td>
<td>Potatoes</td>
<td>191</td>
<td>0.0191</td>
<td>17-17-17</td>
</tr>
</tbody>
</table>

2.2 Hydrological settings of study site

The substrate of PEI can be divided into soil, till, and bedrock. The results of the particle size analysis performed by Zebarth et al. (2014) are presented in Table 2-2. The soil is observed to be approximately 1 m deep and sandy loam in texture, with three distinguishable horizons. The till deposits reflect the composition of the underlying sandstone, although heterogeneity is observed in the overburden from the core samples.
retrieved by split spoon at Harrington including discontinuous thin layers of clayey material (Zebarth et al., 2014). Intense horizontal to subvertical fracturing at various depths, are also observed in the cores, which confirm the presence of fractures in the till (Zebarth et al., 2014). van der Poll (1981) stated that heterogeneity was also shown in bedrock, with fluvial cycles of red conglomerate, sandstone and siltstone occurring in the cores in repeated fining-upward sequences.

Table 2-2: Physical characterization of the substrate at Harrington Experimental Farm

<table>
<thead>
<tr>
<th>Substrate Layer</th>
<th>Material/Texture</th>
<th>Depth</th>
<th>Sand</th>
<th>Silt</th>
<th>Clay</th>
<th>Saturated Hydraulic Conductivity</th>
<th>Porosity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>(m)</td>
<td>(g/kg)</td>
<td>(g/kg)</td>
<td>(g/kg)</td>
<td>(m/s)</td>
<td>%</td>
</tr>
<tr>
<td>Soil, A horizon</td>
<td>Sandy loam</td>
<td>0–0.25</td>
<td>592</td>
<td>295</td>
<td>113</td>
<td>$2.8 \times 10^{-5}$</td>
<td>49.5</td>
</tr>
<tr>
<td>Soil, B horizon</td>
<td>Sandy loam</td>
<td>0.25-0.50</td>
<td>630</td>
<td>290</td>
<td>80</td>
<td>$1.7 \times 10^{-5}$</td>
<td>50.1</td>
</tr>
<tr>
<td>Soil, C horizon</td>
<td>Sandy loam</td>
<td>0.50-1.00</td>
<td>600</td>
<td>290</td>
<td>110</td>
<td>$4.9 \times 10^{-6}$</td>
<td>41.0</td>
</tr>
<tr>
<td>Till</td>
<td>Argillaceous sandstone</td>
<td>1-8</td>
<td>591</td>
<td>379</td>
<td>30</td>
<td>$5.4 \times 10^{-5}$</td>
<td>34.9</td>
</tr>
<tr>
<td>Bedrock</td>
<td>Sandstone</td>
<td>&gt;8</td>
<td>698</td>
<td>280</td>
<td>22</td>
<td>$10^{-3} - 10^{-7}$</td>
<td>17.0</td>
</tr>
</tbody>
</table>
3. Methodology

To better understand water percolation pathways and travel times, a variety of instruments were employed in this study: a cross-hole electrical resistivity system, soil moisture/temperature/conductivity sensors, and water level/temperature/conductivity loggers in the nearby monitoring wells.

3.1 3D cross-hole electrical resistivity imaging (ERI) inversion theory

ERI surveys involve the acquisition of multiple apparent resistivity measurements, each involving the injection of a known current between two electrodes (the current dipole), and measurement of the potential drop across other two electrodes (potential dipole).

The apparent resistivity $\rho_a$ (Ωm) can be calculated from the equation below:

$$\rho_a = \frac{4 \pi k \Delta V}{I},$$  \hspace{1cm} \text{Equation 3-1}

with: $k = \frac{1}{\frac{1}{|AM|} + \frac{1}{|BM|} + \frac{1}{|BN|} + \frac{1}{|AN|}}$, \hspace{1cm} \text{Equation 3-2}

where,

$k$ is the geometric factor,

$|AM|$, $|BM|$, $|BN|$, $|AN|$ are the absolute distances between electrodes A, B, M, and N,

$\Delta V$ is the potential drop measured across the potential dipole, and

$I$ is the injected current source.

The measured apparent resistivity is a complicated weighted average or composite resistivity for a region of influence dependent on the electrode geometry and the resistivity structure of the subsurface. Therefore, inversion of many such measurements
are required to yield a best estimate of true resistivity variations in the subsurface. In this research, 4896 apparent resistivities (including the reciprocals) were measured for each survey, and subsequently inverted to obtain one ERI model. The apparent resistivity values obtained from the ERI inversions were not readily convertible to the pore water solute concentrations, since the apparent resistivity calculated by ERI inversions were associated with bulk resistivities. The relationship between pore fluid resistivity and bulk resistivity depends on the porosity of the rock, the tortuosity of the pore space, and the possible presence of surface conductivity, and when the surface conductivity is negligible can commonly be well approximated by an equation in the form of Archie’s Law.

\[ \rho_{\text{bulk}} = F \rho_{\text{fluid}}, \]

Equation 3-3

\( \rho_{\text{bulk}} \) is the bulk resistivity of the rock (\( \Omega \text{m} \)),

\( \rho_{\text{fluid}} \) is the resistivity of the pore fluid (\( \Omega \text{m} \)),

\( F = a\phi^m S^n \) is the electrical formation factor (\(-\)),

\( \phi \) is the porosity (\(-\)),

\( S \) is water saturation (\(-\)),

\( a \) is tortuosity factor, and

\( m, \) and \( n \) are empirically derived constants (\(-\)).

With the resistivity of the pore fluid playing a significant role in the apparent resistivity of the sample, it is important to compensate for the influence of temperature on the resistivity of water. In this research, the changes in water salinity have a far greater effect than changes in temperature, but temperature may be important at very shallow depths. When the temperature of water increases, viscosity decreases, making the ions more
The increase in ion mobility results in a decrease in resistivity. This temperature/resistivity relationship can be well approximated using the following formula (Hayashi, 2004):

\[ \rho_t = \frac{\rho_{25}}{1 + \alpha(t - 25)} \]

where,

- \( \rho_t \) is the measured resistivity at temperature \( t \) (Ωm),
- \( \rho_{25} \) is the resistivity at 25°C (Ωm),
- \( \alpha \) is the temperature coefficient of resistivity (0.0187/°C),
- \( t \) is the ambient temperature in Celsius.

Equation 3-4 holds to within 2.4% for temperatures between 0 and 30 °C.

In a homogenous half space, the apparent resistivity equals the true resistivity, however, in reality, the resistivity of the subsurface varies spatially. Thus, the measured apparent resistivities need to be inverted to obtain the true resistivities. The ERI inversion software employed in this study was A Program Library for Forward Modelling and Inversion of DC/IP Data over 3D Structures (DCIP3D)—a commercially available program produced by University of British Columbia, made freely available for academic use. As described in the DCIP3D Manual (UBC GIF, 2014), DCIP3D produces models for the true subsurface resistivity distribution using an iterative smoothness-constrained least-squares method which attempts to minimize the root mean squared (RMS) error or misfit between the measured apparent resistivities and the calculated apparent resistivities for a given model while also minimizing a model norm so as to maximize model smoothness. Therefore, the inverse problem is formulated as an optimization problem where a global objective function, \( \Phi \), is minimized. The global objective function consists two components: a
model objective function, $\Phi_m$, and a data misfit function, $\Phi_d$. The global objective function is

$$\min \Phi = \beta \Phi_m + \Phi_d,$$

Equation 3-5

where,

$\beta$ is a trade-off parameter that controls the relative importance of the model smoothness through the model objective function and the data misfit function.

Since the data contains noise, a model that fits the data perfectly would be incorrect, as some features observed in the constructed model would assuredly be artifacts of the noise. DCIP3D assumes that the contaminating noise on the data is independent and Gaussian with zero mean, and the data misfit function becomes

$$\Phi_d = \sum_{i=1}^{p} \left( \frac{\hat{d}_i - d_i}{e_i} \right)^2$$

Equation 3-6

where,

$p$ is the number of data,

$i$ means the $i^{th}$ element in the data,

$\hat{d}_i$ is predicted data,

$d_i$ is the observed data, and

$e_i$ is the data error attributed to noise.

Thus, the optimal misfit would have a value equal to the number of data points.

The objective function gives the flexibility to incorporate as little or as much information as possible. At the very minimum, this function drives the solution towards a “best guess” or reference model at a user-determined extent, and requires that the model be relatively
smooth in the three spatial directions. The general form of the model objective function is

\[ \Phi_m(m) = \alpha_s \int_V \{w_s(r)[m(r) - m_0]\}^2 \text{d}v + \alpha_x \int_V w_x(r)\left[\frac{\partial[m(r) - m_0]}{\partial x}\right]^2 \text{d}v + \]

\[ \alpha_y \int_V w_y(r)\left[\frac{\partial[m(r) - m_0]}{\partial y}\right]^2 \text{d}v + \alpha_z \int_V w_z(r)\left[\frac{\partial[m(r) - m_0]}{\partial z}\right]^2 \text{d}v, \]

Equation 3-7

where,

\( \Phi_m \) is the objective function to be minimized,

\( r \) is a vector that indicates the cell location in a mesh,

\( V \) is the volume of the model,

\( w_s \) is the user-given generalized weighting function of the smallest model,

\( w_x, w_y, \) and \( w_z \) are user-given generalized weighting functions in three spatial directions,

\( \alpha_s, \alpha_x, \alpha_y, \) and \( \alpha_z \) are coefficients that affect the relative importance of different components in the objective function in each axis,

\( m(r) \) is the recovered model, and

\( m_0 \) is the reference model (which may also be spatially variable, i.e. \( m_0(r) \)).

The purpose of the generalized weighting functions is to place emphasis throughout the model to utilize prior information (UBC GIF, 2014). In this study, the smallest model component, \( \alpha_s \), was set to be 0.0001 times the value of the coefficient for the derivative in the two horizontal and vertical directions (\( \alpha_x, \alpha_y, \) and \( \alpha_z \)), so the inversion solution is not biased to the reference model. By assigning different values to \( \alpha_x, \alpha_y, \) and \( \alpha_z \), the horizontal and vertical smoothness can be weighted differently.

The elements of the true resistivity models obtained from the ERI inversions are bulk resistivities. While changes in bulk resistivities over time may be indicative of changing
pore water salinity due to solute transport they are also affected by temporal changes in water saturation (moisture) according to Archie’s Law. Thus, in the vadose zone, independent information on temporal changes in water saturation would be required to quantitatively relate changes in resistivity to changes in solute concentrations at any point in the subsurface. As discussed below, soil moisture sensors were installed in the boreholes in an effort to collect such information. The effect of seasonal changes in temperature on resistivity should also be considered; it is shown later that the expected changes due to temperature were small compared to the changes in resistivity observed during the 18.5 month tracer test.

3.2 Field method

3.2.1 Monitoring array installation

Field work started in March, 2014 when three vertical 16.3 m deep boreholes (Figure 3-1), each with a diameter of 6.35 cm, were drilled in the pattern of an equilateral triangle with sides approximately 9 m long. In this project, the panel aspect ratio, which is borehole separation/the distance between the shallowest and deepest electrode, is approximately 0.57.
Electrodes were installed in the boreholes in July 2014. The electrodes were 5 cm long by 1 cm in diameter and were constructed using cylindrical stainless steel rods. Each electrode was attached to its own wire lead using a blow torch and silver solder. The connection was sealed using a heat shrink gun and tubing to prevent corrosion. The wire used was 16–18 AWG 19-strand hard rolled copper weld wire clad in a tough, blue, high density polyethylene jacket (resistivity/IP survey wire sold by Walcer Geophysics). Each electrode was packed inside a cotton sock filled with bentonite; the high electrical conductivity of bentonite and its tendency to retain water would enhance galvanic contact between the electrodes and subsurface. A total of 24 electrodes were installed in each borehole by taping them (and their lead wires) to a ½” diameter 16 m long flexible PVC pipe with a spacing of 0.68 m (Figure 3-2 and Figure 3-3) before lowering the entire assembly into the borehole.

Besides the electrodes, there were 11 soil moisture/temperature/conductivity sensors (Decagon 5TE) attached to the PVC pipe (Figure 3-2) for each borehole. Their depths are given in Figure 3-3 and Table 3-1.
After the borehole monitoring array was installed, the boreholes were backfilled with coarse sand, with layered granular bentonite pellets with an average thickness of 0.2 m at different depths. The layered bentonite was used to lower the contact resistance between the electrodes and subsurface, and prevent the borehole from acting as vertical drain. The coarse sand was used to fill the void between bentonite layers and thus, to prevent conductive anomalies. In retrospect, the choice of coarse sand was not ideal because it could act like fracture pathways, and allow the groundwater to drain directly through the boreholes, and its high resistivity could provide high contact resistance.

![Figure 3-2 Borehole monitoring array. Each electrode was put into a sock filled with bentonite.](image1)

![Figure 3-3: Borehole monitoring array installation.](image2)
Additional electrodes were installed in trenches dug 40–50 cm deep along the three sides of the triangular array in August, 2014 (Figure 3-4); there were eight electrodes evenly spaced every one metre in each trench. The trenches were backfilled with the original soil, leaving the trench electrodes buried at least 40 cm below the surface. The entire array was buried below the maximum plowing depth so that it could remain in place during farming operations. Figure 3-5 shows the study site after the ERI array installation. The ERI monitoring array, and general layout are shown in Figure 3-6.

Table 3-1: Depth of 5TE sensors in three boreholes

<table>
<thead>
<tr>
<th>5TE number</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
</tr>
</thead>
<tbody>
<tr>
<td>S (m)</td>
<td>1.8</td>
<td>2.4</td>
<td>3.1</td>
<td>3.8</td>
<td>4.4</td>
<td>6.4</td>
<td>8.4</td>
<td>10.1</td>
<td>12.0</td>
<td>14.1</td>
<td>16.1</td>
</tr>
<tr>
<td>E (m)</td>
<td>0.3</td>
<td>0.9</td>
<td>1.6</td>
<td>2.3</td>
<td>3.5</td>
<td>5.0</td>
<td>7.7</td>
<td>9.8</td>
<td>11.8</td>
<td>14.3</td>
<td>15.9</td>
</tr>
<tr>
<td>W (m)</td>
<td>0.4</td>
<td>1.0</td>
<td>1.5</td>
<td>2.3</td>
<td>3.9</td>
<td>5.8</td>
<td>7.8</td>
<td>9.8</td>
<td>11.9</td>
<td>14.6</td>
<td>16</td>
</tr>
</tbody>
</table>

Figure 3-4: Trench monitoring array installation.
Figure 3-5 View of study site after monitoring array installation.

Figure 3-6: ERI monitoring array field set up. Red dashed lines indicate the routing of electrode lead wires that were enclosed in PVC conduit buried approximately 40 cm below surface.

During winter, a resistor network (five 20 Ohm resistors connected in series to the 12 V marine battery that was charged by a solar panel) was built and attached to the back of...
the resistivity meter to provide heat, and a dry environment to ensure the resistivity meter functioned properly.

3.2.2 ERI sequence design

The ERI measurement sequence consisted of a total of 4896 different four-electrode arrays that were loaded into the resistivity meter and measured automatically once to three times per day. The time to collect one survey is typically about 2.5 hours, with the instrument running at 5 Hz. Half of the measurements were so-called “forward” measurements while the other half were reciprocals, made by exchanging the current and potential dipoles in each of the previously measured 4-electrode arrays. In the absence of noise the apparent resistivities measured by the forward and reciprocal array configurations should be identical. In reality, ERI measurements contain errors owing to a variety of sources, including poor electrode contact, random device errors and external effects. An accurate assessment of these errors is critical to the efficiency of the inversion process. Binley et al. (1995) have shown that a good estimate of data error is achieved by considering the reciprocal error, as a comparison of the reciprocal measurements can reveal inconsistency with reciprocal behaviour expected under ideal conditions.

Different kinds of dipole-dipole electrode array configurations were used to provide satisfactory resolution, and to cover as much volume as possible at the site. For a dipole-dipole array in conventional resistivity surveying on surface, potential electrodes are located to one side of the current electrodes (Figure 3-7). The depth of penetration (DOI) is increased by increasing the separation between the two dipoles and, or by increasing the unit electrode spacing \( a \). Of the common array types, the dipole-
dipole array offers relatively high sensitivity to lateral changes (Furman et al., 2003; Loke, 2014). As mentioned in Chapter 1.3, a cross-hole implementation of the dipole-dipole array was recommended by Zhou and Greenhalgh (2000) over other possible electrode configurations. The geometric factor for the dipole-dipole array is \( k = an(n+1)(n+2)\pi \) (for a homogenous half-space), where \( n \) is the number of unit electrode spacings between the current dipole and the potential dipole. The depth of investigation (DOI) increases with both \( a \) and \( n \). With \( n = 1 \), the DOI is 0.42\( a \), while at \( n = 6 \), the DOI is 1.73\( a \) (Douglas et al., 1999). The areas of highest sensitivity are beneath each dipole.

Figure 3-7: Schematic of the dipole-dipole array. C1 and C2 are the current electrodes, and P1 – P5 are potential electrodes.

In this study, the ERI sequence involves three main kinds of 3D dipole-dipole arrays:

1. the four quadrapole electrodes are all borehole electrodes (borehole-borehole);
2. four quadrapole electrodes are all trench electrodes (trench-trench);
3. the current dipole and potential both utilize one borehole electrode and one trench electrode (borehole-trench).

The boreholes were labeled E, W and S, indicating their relative positions, and the trenches were labeled accordingly as WE, WS, and ES.

Borehole-borehole configurations from the survey sequence are illustrated in Figure 3-8, including dual panel cross-hole measurement, and single panel cross-hole measurement.

Shown in Figure 3-9 are the borehole-trench configurations. We classify borehole-trench measurements as “single panel hole to trench”, “cross volume hole to trench”, and “cross
volume trench to two holes”. There are measurements only involving trench electrodes, to improve the ERI array resolution at shallow surface. A more detailed sequence description can be found in Appendix D.

Figure 3-8: Borehole-borehole configurations. The black dots depict electrodes, the green lines depict the current injection, and the red lines depict the potential measurement.

Figure 3-9: Borehole-trench configurations. The black dots depict electrodes, the green lines depict the current injection, and the red lines depict the potential measurement.
3.2.3 Tracer test

The tracer was applied on March 27\textsuperscript{th}, 2015, on a day when the air temperature varied between -1° C to 1° C. 100 kg of granular potassium chloride (KCl) fertilizer (N:P:K 0:0:60, meaning this fertilizer contains 60% potassium by weight) was evenly distributed in a 15.2 m\textsuperscript{2} hexagonal area with 2.12 m sides that was positioned symmetrically inside the borehole triangle. Snow cover was removed from the hexagonal area to allow the granules to be spread directly on the ground surface (Figure 3-10). The removed snow was then immediately replaced back on top of the area to the original thickness of 1.1 m, so the tracer movement would be monitored under natural conditions. Four snow tubes were acquired at the site and allowed to melt, revealing that the snow cover was equivalent to 350 mm of water. If all of the snow had melted at once and dissolved all of the KCl tracer, the resulting salinity of the aqueous solution would have been about 18,518 mg/L TDS, which is about half the salinity of sea water.

Figure 3-10: Tracer application on March, 27\textsuperscript{th}, 2015. Red area was the applied KCl.
The tracer application area was a hexagonal area that was designed to be as large as possible within the confines of the triangular survey area, while reducing the chances of tracer migration into the boreholes or trenches. The hexagon sides were 2.12 m long, and approximately 0.5 m away from the nearest side of the triangle. A significant amount of KCl was used in an effort to ensure that the electrical signal would be strong enough to be detected along its entire path to the water table at about 17 m depth. We acknowledge however, that it may have resulted in initial solute concentrations high enough to be influenced by density-driven flow. Kemna et al. (2002) stated that when the electrical conductivity is more than 10 times greater than the background conductivity it can potentially cause gravitational sinking of the injected tracer slug.

3.3 Hydrogeological monitoring

3.3.1 Decagon 5TE sensors

Decagon 5TE sensors were being used to monitor the volumetric moisture content, temperature, and bulk electrical conductivity of the subsurface. The sensors can monitor in a range of 1% to 80% in moisture content, 0 to 23 dS/m in electrical conductivity, and -40 °C to 60 °C in temperature (Decagon Devices, 2016). They use an oscillator running at 70 MHz, to measure the dielectric permittivity of soil to determine the water content; a thermistor in thermal contact with the sensor prongs determine soil temperature; a two-sensor array to measure the electrical conductivity. The electrical conductivity measurements are normalized to 25 °C. In this research, the sensors were set to take measurements every 30 minutes, and were powered from a 12 V external battery.
3.3.2 Monitoring wells

Water level, temperature, and water conductivity were continuously monitored using CTD Divers (vanEssen Instruments, 2017) deployed in two neighbouring wells. Well #1, and Nest A deep well (NAD). Well #1 is 22.8 m upgradient from borehole S, and NAD is 15 m downgradient from borehole S (Figure 3-11).

![Figure 3-11: Map of north-west corner of field 355, indicating the well locations. The red dash triangle shows the rough location of the borehole triangle. The deep well in Nest A is labeled as NADW in the figure.](image)

3.4 Inversion of 3D resistivity surveys

For purposes of 3D modelling, the earth surrounding the boreholes was discretized using a mesh with elements measuring 40 cm by 40 cm by 23 cm high in the central part of a domain measuring 25.6 m by 24.80 m by 21.55 m. Within the core mesh, the size of the cells should be comparable with the spacing of the data; in this work, the vertical extent of the core cells is approximately 1/3rd of the 68 cm borehole electrode separation, and the horizontal extent of the core cells is nearly twice of the vertical. The top of the mesh is arbitrarily assigned an elevation of 0 m, and the southwest corner is at -8.4 m easting and -9.2 m northing.
The mesh was a 44 x 42 x 80 cell domain (147,840 cells), including a 34 x 32 x 75 central domain, 5 padding cells at the outer edges of the model in east, west, north, and south directions, and 5 padding cells at the bottom. Figure 3-12 and Table 3-2 depict the mesh discretization. In this thesis, all the DCIP3D models will be displayed without the padding cells.

**Table 3-2: DCIP3D model mesh discretization. X direction is eastward, Y direction is northward, and Z direction is from bottom to top.**

<table>
<thead>
<tr>
<th>Direction</th>
<th>Padding Cells (m)</th>
<th>Central domain (m)</th>
<th>Padding Cells (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>X(W-E)</td>
<td>2 1.5 1.2 0.8 0.5</td>
<td>34 X 0.4</td>
<td>0.5 0.8 1.2 1.5 2</td>
</tr>
<tr>
<td>Y(S-N)</td>
<td>2 1.5 1.2 0.8 0.5</td>
<td>32 X 0.4</td>
<td>0.5 0.8 1.2 1.5 2</td>
</tr>
<tr>
<td>Z(T-B)</td>
<td>/ / / / /</td>
<td>75 X 0.23</td>
<td>0.3 0.5 0.8 1.2 1.5</td>
</tr>
</tbody>
</table>

**Figure 3-12: DCIP3D model mesh discretization. The red dots depict the locations of the electrodes. Size and numbers of cells in the center domain are shown.**

It was important to set the outer limits of the mesh sufficiently far away from the electrodes because modelled current flow was constrained to flow within the mesh. The
outer limits were selected by considering the proportion of current flowing below a certain depth \( z \) for the case of two current electrodes separated by a distance \( L \) at the surface of a homogeneous half space (Telford et al., 1982).

Obviously, it is necessary to increase the current electrodes separation to increase the depth of exploration. In a homogenous half space, the fraction of total current, \( I_z \), below a certain depth, \( z \), will be:

\[
I_z = 1 - \frac{2}{\pi} \arctan \frac{2z}{L}.
\]  
Equation 3-8

Figure 3-13: Fraction of current flowing below depth \( z \) for a current dipole of length \( L \) at the surface of a homogeneous half space (Telford et al., 1982).

Figure 3-13 shows that there is about 32% of the current flows below a depth equal to the electrode spacing. Conversely, 68% of current flows between surface and that depth. When considering a certain depth \( z \), the larger the current electrode spacing is, the more current flows below \( z \) (Telford, et al., 1982).

In a homogenous full space, the current fraction \( I'_z \) that would flow beyond a distance \( z \) to one side of the current dipole is:
\[ I_z = \frac{1}{2} \left( 1 - \frac{2}{\pi} \arctan \frac{2z}{L} \right), \]  

Equation 3-9

In this study, the average borehole current dipole is about 10 m long, and the distance from that borehole triangle panel (panel connecting borehole W and E) to the nearest vertical edge of the mesh is 8.4 m. Thus, in a uniform whole space, about 83% of the current would flow within the region that was meshed. However, once tracer was applied, the presence of conductive tracer would have drawn current into the triangular prism.

3.4.1 Input parameters for DCIP3D

The XYZ positions of the uppermost electrode in each borehole and of the trench electrodes were determined using RTK (real time kinematic) GPS positioning with cm-precision. The positions of the remaining borehole electrodes were determined using the known spacing of 68 cm and the inclination angle of each borehole as measured by the stick-up of steel casing (prior to its removal) at surface. The borehole inclinations and the electrode positions are listed in Appendix C.

Apparent resistivity measurements are most sensitive to the regions near the electrodes. Cell weighting functions \(w_i(r), w_x(r), w_y(r),\) and \(w_z(r),\) (Equation 3-5) were therefore used to bias the inversion routine against fitting the data by imposing large gradients in resistivity close the boreholes and trenches. Cells within 0.5 m from electrodes were weighted four times more heavily than those further than 1 m from the electrodes; cells between 0.5 m to 1.0 m from electrodes are weighted two times more heavily than those further than 1 m from the electrodes. Figure 3-14 shows a visualisation of the weighting matrix.

For an initial inversion of apparent resistivity measurements collected prior to tracer application, we further biased the solution towards a horizontally layered structure, in keeping with the expected near-surface geology \((\alpha_x = \alpha_y = 5\alpha_z).\) The first post-tracer
resistivity model was inverted by utilizing the resistivity model obtained a few hours earlier (before tracer application) as both a starting model and a reference model. In running all the post-tracer time-lapsed inversions, since they are constrained by using the previous model as the reference model, the horizontal smoothness was set same as the vertical smoothness ($\alpha_x = \alpha_y = \alpha_z$). The smallest model component, $\alpha_s$, was set to be 0.0001 times of the coefficient for the derivatives in the easting, northing, and vertical direction ($\alpha_x, \alpha_y$, and $\alpha_z$, respectively).

When inverting the post-tracer data set from the spring of 2015 (when the conductivity anomaly caused by tracer was highly conductive), a hard constraint was applied on the inverted model, with an upper bound of 5000 Ohm-m, and a lower bound of 1 Ohm-m. And the data errors were assumed to be 2.5% for all inversions.

Figure 3-15 is an example of the control file for DCIP3D (UBC GIF, 2014). Inversion parameters including, for example, the maximum number of Gauss-Newton iterations to be performed, the hard conductivity limits (i.e. maximum and minimum conductivities allowed for any cell), and the relationship of $\alpha_x, \alpha_y$, and $\alpha_z$, etc., are set in this file. We know that the true data misfit should be achieved near the number of data, and DCIP3D would perform a line search to choose the appropriate trade-off parameter $\beta$ iteratively, and in the end, a best-fitting mode would be calculated with a reasonable misfit. In this study, the maximum allowed iterations was set to 40, and it usually took less than 10 iterations to converge to the target misfit.

The time typically required by DCIP3D to complete a cross-hole inversion, when running on a Windows 2012 workstation, having dual 6-core 2.1 GHz Xeon processors (model ES-
2620 v2), and 64 GB RAM was between 1 to 3 hours, depending on the number of iterations.

Figure 3-14: Weighting matrix/function visualisation for biasing the cells close to the electrodes different to the cells far away from the electrodes. White nodes are used to illustrate electrodes. Pink cells are those within 0.5 m from electrodes, and are weighted 4 times more than those further than 1 m from the electrodes; green cells are those between 0.5 m and 1.0 m from electrodes, and are weighted 2 times more than those further than 1 m from the electrodes.

Figure 3-15: Control file of DCIP3D.

3.4.2 Filtering system

Following the field surveys, several quality control filters were applied to the survey data prior to inverse modelling with DCIP3D. The detailed filtering code is given in Appendix
A. The filtering code first eliminated measurements associated with electrodes that were out of contact (as evidenced by extremely high contact resistances and an inability to inject current through them); by May, 2016, four of the 96 electrodes that were out of contact. The filtering code then removed measurements considered to be excessively noisy—exhibiting inadequate repeatability, defined as measurements for which the standard deviation of three consecutive voltage drop exceeded 2.5% of the average. It also eliminated measurements with reciprocal errors > 5%, and measurements made with very small currents (< 1 mA). Poor reciprocals typically accounted for the greatest number of rejected measurements, and typically, there were about 50 measurements that failed reciprocal check. The reciprocal check filtered out the maximum number of measurements (about 150) during the first few days after the tracer was applied. The minimum number of reciprocal check failures for any one survey was less than 10. These filters were applied to assure quality data for inversion in DCIP3D. Contact resistances were typically below 10 kΩ at the time of survey.

3.5 Infiltration modelling

Infiltration of snowmelt and precipitation at the study site was estimated daily using the Soil Water Balance (SWB) model (Salazar and McNutt, 2010). SWB provided a method of performing soil-water balance calculations in order to determine a time series for infiltration. The simulation spanned the time interval from March, 2015 to October, 2016. The parameters used for the simulation are given in Table 3-3. The SWB model uses a standard lookup table for assigning parameters for crop and soil characterization. Daily
temperature and precipitation data of Charlottetown airport were retrieved from the Environment Canada Historical Climate Data archive (Environment Canada, 2016).

**Table 3-3: SWB model options and input parameters**

<table>
<thead>
<tr>
<th>SWB Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land Use (-)</td>
<td>Other Row Crops</td>
</tr>
<tr>
<td>Soil Hydrologic Group (-)</td>
<td>B</td>
</tr>
<tr>
<td>Evapotranspiration Method</td>
<td>Thornthwaite-Mather</td>
</tr>
<tr>
<td>Latitude (°)</td>
<td>46</td>
</tr>
<tr>
<td>Base Soil Water Capacity (cm/m)</td>
<td>13.3</td>
</tr>
<tr>
<td>Initial Soil Moisture (%)</td>
<td>50</td>
</tr>
<tr>
<td>Initial Snow Cover (cm)</td>
<td>110</td>
</tr>
</tbody>
</table>

In the SWB model, the land use type and soil hydrologic group parameters are used to specify crop interception, base soil water capacity, maximum infiltration rate, soil conservation service (SCS) curve numbers (5 for soil group in study site), and depth of root zone (Salazar and McNutt, 2010). Soil groups can be assigned to four categories (A-D), based on their infiltration capacity, "A" soils have a high infiltration capacity and subsequently a low overland flow potential, while "D" soils have a very low infiltration capacity and subsequently a high overland flow potential. The soil at the study site fell in soil group B, with an infiltration capacity within 0.38 to 0.76 cm/h. The values used for the model were selected based on knowledge of the crop and soil conditions at Field 355. The Thornthwaite-Mather method (Steenhuis and van der Molen, 1986) was used for calculating evapotranspiration. The initial soil moisture term expresses the percent saturation of the root zone and this value was set to 50 % because the simulations began during the wet season. The initial snow cover was set to 110 cm based on the measurement taken on the study site on March 27th, 2015.
4. Results and Discussion

4.1 Forward modelling

Several forward modelling exercises were carried out to explore the expected sensitivity and resolution of the ERI array. The approach involved calculating expected or “synthetic” apparent resistivity data for various simplified synthetic resistivity models and then inverting those synthetic data using the DCIP3D algorithm and inversion parameters like those used with the real data. For display purposes, all inverted resistivity models and their corresponding synthetic models are plotted using the same colour scale. Also, unless otherwise indicated, results for all models are shown in one vertical cross-section which cuts through the borehole triangle 0.4 m inside the EW trench.

4.1.1 Two layer structures to explore ERI array sensitivity above the trench electrodes

Since the trench electrodes were installed at approximately 0.4 m depth, the ERI system’s ability to correctly resolve a layer above was in question. Two simple two-layer synthetic models were used to explore that issue. Synthetic Model 1 (Figure 4-1(a)) has a 0.46 m thick top layer with a resistivity of 100 Ohm-m overlying a 1000 Ohm-m half-space. Synthetic Model 2 (Figure 4-2 (a)) has the same geometry, but the upper layer is more resistive (2000 Ohm-m) than the underlying 1000 Ohm-m half-space. A half-space is either of the two parts into which a plane divides the three dimensional Euclidean space.
Figure 4-1: (a) Synthetic Model 1 of a two layer structure with upper, conductive layer (100 Ohm-m) overlying a much more resistive 1000 Ohm-m homogeneous half-space; (b) Inversion of synthetic apparent resistivity measurements obtained by DCIP3D. White dots represent the electrodes. The colour scale on the right, ranging from 100 to 1000 Ohm-m, is used for both models. The original resistivity range of the inversion was from 56 to 3060 Ohm-m.

Figure 4-2: (a) Synthetic resistivity model of a two layer structure with upper layer resistive (2000 Ohm-m) overlying a much more conductive 1000 Ohm-m homogeneous half-space; (b) Inversion of synthetic apparent resistivity measurements obtained by DCIP3D. The colour scale on the right, ranging from 1000 to 2000 Ohm-m, is used for both models. The original resistivity range of the inversion is from 779 to 2070 Ohm-m.
The two cases above show that a conductive (Synthetic Model 1) or resistive (Synthetic Model 2) near-surface layer above all of the electrodes can be recovered by inversion (Figure 4-1 (b), and Figure 4-2 (b)), although the full resistivity contrast is only recovered in close proximity to the trench electrodes and the upper layer’s resistivity is quite variable. In case 1, the recovered resistivity value of the lower layer is reduced near its contact with the overlying conductive layer. Neither of these two inversions resolve the very sharp edge of the upper layer, since DCIP3D is a smoothness-constrained code.

4.1.2 Simulation of two different tracer application areas

In this section, two synthetic models are used to show how a plan was developed for tracer application.

Model 3 (Figure 4-3) simulates tracer present in two 0.4 m wide strips, extending to a depth of 2.07 m and having a resistivity of 100 Ohm-m in a background of 1000 Ohm-m. The depth of 2.07 m was chosen to investigate our ability to visual the early stage of tracer infiltration. The two strips join inside the borehole triangle, at about 0.8 m away from borehole W. The resistivity range recovered by inversion increased to 25 to 3700 Ohm-m. The main shortcoming is that the conductive strips recovered by inversion do not extend as deeply as they do in the synthetic model, reaching a maximum depth of about 1.6 m. The shape of the two strips is blurry at the surface (not shown), but becomes distinct at the depth of 1 m, with a colour change from pink to blue. Given that the simulated tracer was not well resolved numerically or spatially, it was decided that the strips were likely too thin and applying a strip-shaped tracer inside the borehole triangle was not an appropriate plan.
Figure 4-3: (a) Synthetic Model 3 of a two-strip conductive anomaly in a 1500 Ohm-m homogeneous half-space; (b) Inversion of Synthetic Model 3; note that the top view is shown at a depth of 1.15 m. Color scale of the inversion is constrained to the same resistivity range as that of the synthetic model (100 to 1500) Ohm-m. The original resistivity range of the inversion is from 25 to 3700 Ohm-m.

Model 4 simulates a triangular tracer plume with 5.2 m-long sides, present from surface to 5 m depth in a two layer background, and positioned symmetrically within the borehole triangle (Figure 4-4 (a)). The resistivity range of the recovered model 2 (Figure 4-4 (b)) is from 70 to 2100 Ohm-m. In the inverted model, the tracer resistivity anomaly appears at surface and extends to a depth of approximately 4.6 m (bottom of yellow, with a resistivity of approximately 200 Ohm-m). With the help of the trench electrodes, the triangular shape of tracer can be well imaged at the surface, however, with a slightly higher resistivity of 200 Ohm-m. Since DCIP3D generates a smoothness-constrained model, the tracer at the surface is constrained by the resistive top layer, so it is shown in a smaller region compared to the synthetic model, and the tracer plume recovered by the inversion at depth has smooth edges.
Figure 4-4: (a) Synthetic Model 4 of 5 m-long triangle tracer in a two layer background; (b) Inversion of Synthetic Model 4, with a view of cutting at northing 5.6 m (0.8 m further into the borehole triangle). Color scale of the inversion is constrained same as the resistivity range of forward model, from 100 to 1500 Ohm-m. The original resistivity range of the inversion is from 70 to 2120 Ohm-m.

Forward model 5 was designed to explore the model ability to visualize a vertical 0.4 m wide, 3 m long conductive feature crossing the third layer of a five layer structure (Figure 4-5 (a)). This could represent a vertical fracture along which tracer has penetrated into the adjoining matrix. A resistive upper layer, 6 m thick with a resistivity of 1500 Ohm-m, overlies two relatively conductive layers—the first one being about 4 m thick, starting at 6 m depth, and the second one about 2 m thick, starting at 13 m deep. The resistivity of these two layers is 70 Ohm-m, and the layer in between, shown in yellow, is 200 Ohm-m. The inverse model (Figure 4-5 (b)) reveals 5 distinguishable layers, and all the edges of the layers are well-defined, except in the vicinity of the simulated fracture. The fracture cannot be resolved but it does have the effect of partially drawing the two conductive layers together. Due to the smoothness constraint on the model, our ability to visualize a vertical fracture is very limited. In reality, the expected fractures are much thinner than
the one simulated here, so it is not likely that we are able to visualize any fractures with the ERI system.

Figure 4-5: (a) Synthetic Model 5 of a 0.4 m wide, 3 m long fracture in a 5 layer structure; (b) Inversion of Synthetic Model 5. Color scale of the inversion is constrained same as the resistivity range of forward model, from 100 to 1500 Ohm-m. The original resistivity range of the inversion is from 16 to 7600 Ohm-m.

4.2 Pre-tracer ERI imaging

The pre-tracer surveys conducted during the fall of 2014, on September 5th, October 10th, November 7th, and December 10th, were used to determine the background resistivity structure of the study site. All the pre-tracer models were inverted using the same inversion parameters including a homogeneous half space for the reference model, as well as the same filtering system, and weighting matrices. In each case, the global smoothness weighting factors $\alpha_x, \alpha_y$ were set to be 5 times larger than $\alpha_z$. This had the effect of emphasizing horizontal smoothing five times more than vertical smoothing, in keeping with our expectation that the geology is predominantly horizontally layered.

Inversions of the three pre-tracer surveys yielded very similar resistivity models, with
minor differences attributed to noise, as well as changes in moisture content and temperature in the soil at different times.

The resistivity model obtained from the survey of December 10th, 2014 is given as an example in Figure 4-6, the color scale of this model was from 69.4 to 1480 Ohm-m. Pink represents area that is more conductive, and blue represents area that is more resistive. It reveals five sub-horizontal layers of alternating resistivity in general agreement with the expected geology of overburden overlying interbedded sandstone to shaley sandstone layers. Two of these layers are more electrically conductive (orange-red hues) and are thought to represent shale-rich units which could impede water infiltration and give rise to perched water table conditions. The water table in the study site is at about 17 m deep in summer, which is usually below our ERI array. It only raises up to cover the last one or two electrodes in the boreholes in late April and May. Thus, in this model, the water table is not imaged.

Figure 4-6: Resistivity model obtained by inversion of the Dec. 10th, 2014 ERI survey, showing five distinct layers. White dots represent the electrodes. The model is shown in its original color scale, from 70 to 1480 Ohm-m.
4.3 Tracer test results

4.3.1 Cross-hole 3D surveys

Hundreds of post-tracer 3D ERI surveys have been conducted since March 27\textsuperscript{th}, 2015. The resistivity meter was left at the study site to conduct automatic surveys at 8 hour intervals starting with April 1\textsuperscript{st}, 2015. During spring, 2015, three surveys were done for each day, at 08:00 hrs, 16:00 hrs, and 24:00 hrs local time each day. Then the monitoring rate was reduced to two surveys (at 12:00 hrs and at 24:00 hrs) starting on June 20\textsuperscript{th}, 2015, due to the slow movement of tracer in dry seasons, and then one survey (at 24:00 hrs) per day starting on September 5\textsuperscript{th}, 2015.

The ERI data were subsequently filtered and inverted to yield time-lapse images (Figure 4-7) showing migration of conductive tracer through the triangular study area over a monitoring period of 18.5 months. The time-lapse inversions were obtained by using the cascaded inversion method, which uses the inversion result of the previous dataset as the starting and reference model for the inversion of the subsequent datasets (Hayley et al., 2011). Ratio images were also produced-dividing each post-tracer conductivity model by the background conductivity model of March 27\textsuperscript{th}, 2015, so as to emphasize resistivity changes caused by tracer percolation (Figure 4-8). Since the tracer is far more conductive than natural infiltration, its presence should be associated with ratios greater than 1. Thus, the colour scale of the ratio images were constrained with a lower boundary of 1. While ratios less than 1 can certainly occur, as a consequence of seasonal reductions in water saturation or temperature for example, such changes are comparatively very small.
The vertical scale of all the inversions and ratio images from cross-hole surveys are from 0 to 17.25 m.

Figure 4-7: Evolution of tracer infiltration over 18.5 months (March, 2015, to October, 2016) in the study site at irregular time intervals. White dots indicate the electrode positions, and black hexagon on surface indicates the tracer application area. All the colour scales of ERI models in this figure were constrained from 50 to 1500 Ohm-m
Figure 4-8: Conductivity ratio images of time-lapsed ERI models with constrained colour scale. Each ratio image was obtained by using the current ERI model divided by the background ERI model (model of March 27th, 2015). Pink indicates area that is getting more conductive compared to the background model. The white dots represent the electrodes and the three white lines in each image are at 3 m, 6 m, and 9 m depth, respectively. Note that the ratio image of model obtained on November 9th, 2015, was plotted twice, one in 1-1000 colour scale, the other in 1-150 colour scale.
During the 18.5 months of monitoring, the ERI models revealed that preferential percolation paths have been observed during April, 2015, indicating heterogeneity within overburden, and tracer had percolated to depths of 1 m by the 4th day after tracer application. Its movement slowed significantly by early May, 2015, with the end of snow melt. Tracer spread laterally very slowly through the summer and early fall, 2015, but has remained within the triangular array. The shallow conductivity anomaly produced by the tracer diminished significantly over the winter and spring of 2016. After 18.5 months, there is no ERI evidence of tracer migrating deeper than approximately 6 m (top of the middle conductive layer) depth by matrix flow.

The infiltration rate varies with seasons, and the most rapid tracer movement is expected to happen during the spring thaw, and relatively wet fall season, whereas, in summer growing season, the infiltration rate should be slowed down by the active evapotranspiration. This is borne out by the results of Soil Water Balance (SWB) infiltration simulation presented in Figure 4-9, which shows the highest infiltration rate are expected to have occurred during the springs of 2015 and 2016, followed by the fall of 2015. The SWB simulation predicts there was no infiltration during the summers (July–September) of 2015 and 2016 and negligible infiltration during the winter (January–March) of 2016. More detailed ERI model of different seasons during the monitoring period are shown in Figure 4-10, Figure 4-12, and Figure 4-13.
Figure 4-9: Bar chart depicting the simulated monthly infiltration versus time (from April, 2015, to October, 2016).

Figure 4-10: Sequence of conductivity ratio images respect to the model obtained on March 27th, 2015, at irregular intervals during the first two months following tracer application, starting with the first post-tracer model acquired on March 31st, 2015, ending at May 25th, 2015. White dots represent the electrodes. The colour scale of these ratio images has been constrained to 1 to 1000; the full ratio range is shown below each model.
Figure 4-10 revealed the tracer migration for the first 59 days following the tracer application. For this period, the tracer was considered to be moving most rapidly during the entire 18.5 month monitoring period. The ratio range of these images was constrained from 1 to 1000, whereas conductivity ratios estimated in the heart of the tracer plume were actually as high as 10000 more conductive than the pre-tracer model, in the early stages of tracer development. At this point, we interpret that the pink (1000 times more conductive than the pre-tracer model) to light blue (5 times more conductive than the pre-tracer model) region indicates the extent of the tracer plume. On that basis, the tracer is interpreted to have moved below 1 m depth by the fourth day after tracer application. The high rate of movement for this period suggests that the ground cannot have been solidly frozen. Indeed, the shallowest 5TE sensors in the boreholes, at depths of 0.3 and 0.4 m recorded temperatures of 0 to 3.5 °C in late March. Furthermore, the high salt concentration would have lowered the freezing point of surficial soil water below 0 °C.

Between April 12th (Day 16) and April 22th (Day 30), 2015, the tracer plume was preferentially moving towards borehole W, which indicates heterogeneity within the overburden, though not necessarily fractures. We note that some distortions appeared in the middle region between Days 16 and 26, suggesting that the distortion may be a spurious artefact of the extremely high resistivity contrasts that were associated with the tracer plume developed above this depth. Lateral flow of tracer towards borehole W on surface, becomes evident around days 16-26, and is likely a consequence of the surface micro-topography, which was relatively low near borehole W, as confirmed by the topographic survey conducted (discussed later in this section).
The water table in this area is positioned at about 17 m depth in summer, and typically remains below the deepest borehole electrode most of the year, except during April and early May, when it may rise above the lowest one or two electrodes. Thus, the higher conductive ratio showing at 14 m depth during April and early May might be a raised water table (Figure 4-10 (b) and (c)).

Tracer movement slowed significantly by early-May with the end of the snow melt. Historical climate (Environment Canada, 2016) showed that there was no snow left on the ground as of May 2nd, 2015.

Figure 4-11: Topographic map of the deep vadose site. The blue dots represent the 88 level points, and the black dots descript the borehole locations, the hexagon area indicates the tracer application area. The level point spacing of the first line (Y = 0 m) is 0.5 m, and roughly 1 m for the remaining 10 lines. Level range of the survey area is from 99.12 m to 99.29 m, i.e. 88 to 71 cm lower than the top of a nearby well casing (#110) that was arbitrarily assigned to an elevation of 100 m.
A topographic survey was conducted on May 12th, 2015, to confirm whether the pooling of tracer on the surface near borehole W was caused by the difference of topography of the study area. This topographic survey involved using an optical level and a stadia rod. Level points were taken from 11 lines, with points spaced 0.5 m or 1 m apart as shown in Figure 4-14. The topographic survey shows that the ground elevation near borehole W was 8 cm lower than the average level of the surveyed area (Figure 4-11), explaining why tracer was pooling near borehole W, as being dissolved by precipitation.

Figure 4-12: Sequence of conductive ratio images of time-lapsed ERI models obtained during the tracer test from May 25th to September 5th, 2015 - a period of time when little infiltration was expected. White dots represent the electrodes. The color scale of these ratio images has been constrained to 1 to 1000. Panels (a) through (d) show ratio images at irregular intervals from Day 59 to Day 163 as labeled.
Time lapse ERI conductivity ratio images from the summer (Figure 4-12) suggest that tracer percolation slowed dramatically over the summer of 2015, and this could be due to the high evapotranspiration during this period of time. ERI data from September 5th, 2015 (Figure 4-12 (d)), yielded a resistivity model similar to that obtained on May 25th (Figure 4-10 (f)), except that the tracer plume showed minor lateral spreading at about 3 m depth in the later model. During the entire summer, the vertical matrix migration of the resistivity anomaly has not been significant, however the conductivity ratio of the surface was about three times higher in August and September than in the end of July. During the field trip conducted at the end of July 2015, it was observed that KCl tracer granules were still present on the ground surface, thus, the high conductivity ratio in August and September at the surface could be a result of summer rainfall events continuing to dissolve the applied KCl into the subsurface. Infiltration rate in summer is expected to be very low, since most of the precipitation is taken up by plants or evapotranspirated. The ratio images suggested that the tracer has remained within the triangular array during summer and early fall.

Figure 4-13 illustrates the tracer migration from December 14th, 2015 (Day 260), to October 13th, 2016 (Day 567). For this period, the major movement of tracer occurred during the spring of 2016 (Day 282 to 404), while no apparent movement was observed during summer of 2016, which is expected to be the low infiltration period. The weather records (Environment Canada, 2016) indicate that there was no snow on the ground starting with April 17th, 2016. The total precipitation between December 15th, 2015, and May 6th, 2016, was 453.1 mm. The shallow conductivity anomaly produced by the tracer diminished significantly over the winter and spring of 2016, but showed little evidence of
bulk matrix flow below about 6 m depth (Figure 4-13). The bottom of the conductive anomaly stayed at the depth of about 6 m (top of the middle conductive layer), but the reduction of the conductivity from surface to 3.5 m depth, between December 14th, 2015, to October 13th, 2016, suggests that fracture flow may have carried the tracer into the bedrock below about 6 m depth during this time period. After 18.5 months after the tracer was applied, there is no ERI evidence of tracer migrating outside of the borehole triangle.

Figure 4-13: Sequence of conductivity ratio images of time-lapsed ERI models from December 14th, 2015 to October 13th, 2016. The white dots represent the electrodes. The color scale of these ratio images has been constrained to 1-150. Panels (a) through (d) show ratio images at irregular intervals from Day 260 to Day 537 as labeled.
The simulated SWB generated a total amount of 915 mm infiltration, from April, 2015 to October, 2016. Considering that the tracer migration was driven by this 915 mm of water, to a depth of 6 m, the inferred porosity would be about 15.3 %, if all of the tracer movement was caused by piston-like flow.

4.3.2 Surface 3D surveys

Since the cross-hole ERI array does not have good sensitivity on area outside of the borehole triangular prism, it would probably not image tracer moving outside the borehole triangle, thus two one-time surface surveys were conducted in the borehole triangle area, on May 13th, 2015, and July 14th, 2016, to confirm that no tracer had moved into the boreholes or trenches.

The first surface survey made use of 48 electrodes, 10 in each first four lines, and 8 in the fifth line. The in-line spacing was 2 m and the between-line spacing was 2.5 m. All the electrodes were temporarily installed at the surface, and were removed after the survey. The resistivity instrumentation used for this investigation was the same 10 watt battery-powered Lippmann Geophysikalische Messgeräte “4point light 10W” resistivity meter used for the pre-tracer and post-tracer surveys.

This shallow survey model was a 54 x 38 x 31 cell domain (63, 612 cells); this equated to model dimensions (length x width x depth) of 31.4 m x 31.4 m x 11.75 m. The individual finite volume dimensions (in-line length x cross-line width x height) ranged from 0.5 x 0.5 x 0.25 m immediately beneath the electrode array to padding cells at the outer edges of the model as large as 2 m x 2 m x 1.5 m. Figure 4-14 (a) and (b) depict the mesh discretization in cross-line and in-line sections respectively.
The resistivity model obtained from this survey reveals that tracer had not moved significantly outside of the borehole triangle (Figure 4-15). A cross-section view of the model (Figure 4-16) suggests that the tracer had moved only approximately 2.5 m (orange region, with resistivity of approximately 50 Ohm-m) below the surface, which is less than the estimate of approximately 3.5 m as we obtained from the cross-hole post-tracer surveys. The cross-hole estimate is considered more accurate because it is able to
maintain higher resolution at depth than a survey geometry composed entirely of surface electrodes.

Figure 4-15: Surface survey resistivity model showing the tracer remained mostly contained within the borehole triangle by May 13th, 2015. The white dots represent the 48 electrodes used in this survey, the red triangle indicates the borehole triangle, and the black hexagon is the tracer application area. The resistivity range of this model is from 17.7 to 2300 Ohm-m.

Figure 4-16: cross-section view of the resistivity model of the surface survey.

On July 14th, 2016, a second surface ERI survey was conducted on a large area of 60 m by 10.8 m in the study site. The borehole triangle was sitting symmetrically inside the surface array rectangle. The surveys utilized a sequence containing in-line and off-line
Wenner-Schlumberger array configurations (Figure 4-17), which made use of 48 electrodes, 16 in each three lines. The in-line spacing was 4 m and the between-line spacing was 5.4 m. The off-line configurations provided increased sensitivity to areas between survey lines. Effective depths of exploration were varied by using "a-spacings" of 1-4 dipole length (s) and "n values" of 1-6. Additional sensitivity to regions between survey lines was provided by a series of cross-line measurements using the dipole-dipole array with parallel current and potential dipoles set 4, 8 and 12 m apart. All the electrodes were temporarily installed at the surface, and were removed after the survey.

![Diagram](image)

**Figure 4-17: General geometry for a Wenner-Schlumberger array, shown in an off-line configuration.** A, B and M, N represent current and potential electrodes respectively.

The ERI model was an 82 x 26 x 34 cell domain; this equated to model dimensions (length x width x depth) of 61.9 m x 101.0 m x 62.0 m. The individual finite volume dimensions (in-line length x cross-line width x height) ranged from 1.00 m x 1.35 m x 0.75 m immediately beneath the electrode array to padding cells at the outer edges of the model as large as 5 m x 7 m x 7 m. Figure 4-18 (a) and (b) depict the mesh discretization in cross-line and in-line sections, respectively.
Figure 4-18: (a) Plan view mesh discretization; (b) In-line cross-sectional mesh discretization. The black dots depict the locations of the electrodes.

Figure 4-19: Inversion of surface ERI survey conducted on a scale of 60 m by 10.8 m. White dots represent the temporary surface electrodes, and the black triangle indicates the borehole area. The black dash line indicates the conductivity interface in the subsurface. Full resistivity range is shown, from 31.7 to 12600 Ohm-m.
Figure 4-19 shows a vertical slice through the centre of the borehole triangle, and since the depth of exploration of this survey is approximately 11 m, only 0 m to a depth of 12.4 m is shown. Based on the results, there is again no evidence to suggest that the tracer has moved outside of the borehole triangle. The cross-sectional view of the model suggests that the bottom of the tracer plume had percolated to approximately 8 m (bottom of the orange area, with a resistivity of approximately 80 Ohm-m) below the surface, which is in the middle of the 4 m-thick conductive layer identified in the resistivity models obtained from the cross-hole surveys. This depth estimate is comparable to, but considered less accurate than the estimate of 6 m derived the cross-hole imaging surveys which offer higher resolution at depth. The fourth electrode of the three lines were positioned in the middle of the road, and the last three electrodes of the three lines were in the forest on the west side of the study site. It is interesting to note the anomalously low resistivity below the road, the anomalously high resistivity below the forested area, and the increased depth to higher conductivities outside of the field. One hypothesis is that the fertilizer accumulated over years before this research began, slowly infiltrated into the subsurface, has increased the conductivity of the subsurface, whereas, the subsurface outside the crop field, was not affected by the fertilizer.

4.4 5TE sensor results

The 5TE sensors have been used for monitoring bulk electrical conductivity, in addition to volumetric water content and soil temperature. Good contact between 5TE sensor probes and the ground is required to get precise conductivity and moisture content
readings. However, the 5TE sensors were attached to a PVC pipe, and were lowered into the boreholes, which were later backfilled with coarse sand and bentonite. Because the contact between the probes and the backfilled material was relatively poor, the electrical conductivity data collected by the 5TE sensors is not reliable in a quantitative sense. Low moisture contents in the boreholes, causing contact resistances to be relatively high, also degraded the accuracy of conductivity readings.

Eleven 5TE sensors were installed in each borehole together with electrodes at different depth (Figure 3-3 and Table 3-1) to monitor the moisture content, electrical conductivity and temperature variations. The sensor at 11.9 m depth in borehole W stopped working on October 26th, 2015 (7 months after the tracer has been applied), and the sensor at 1.5 m depth in borehole W stopped working since the tracer has been applied. Moisture content, temperature, and soil conductivity variations with depth and over time from March 27th, 2015 (Day 0), to July 13th, 2016 (Day 567), are plotted in Figure 4-20, Figure 4-21, and Figure 4-22. The 5TE data from July 14th, 2016 (Day 476), to August 23rd 2016 (Day 516), were not available.

Two extra 5TE sensors were installed in the center of the tracer application area on April 8th, 2015 (Day 12), one at 10 cm depth, and the other at 20 cm depth. The temperature, conductivity and moisture content of these two sensors, from April 8th, 2015, to October 13th, 2016, along with the daily precipitation are shown in Figure 4-23. The 5TE data of the two shallow sensors, from July 14th, 2016 (Day 476), to August 23rd 2016 (Day 516), were not available.
Figure 4-20: Moisture content of three boreholes, from March 27th, 2015, to October 13th, 2016. X axis is time from day 0 to day 567, Y axis is depth from 0 m to about 16 m, and color scale shows the moisture content of all three plots are constrained, from 0 to 0.46. The original moisture content range of borehole E, W, S, is 0 to 0.44, 0.42, and 0.46, respectively.
Figure 4-21: Temperature of three boreholes, from March 27th, 2015, to October 13th, 2016. X axis is time from day 0 to day 567, Y axis is depth from 0 m to about 16 m, and constrained color scale from 0 to 22 °C, shows the temperature of each depth at different time, with red indicates high temperature. The full temperature range of borehole E, W, S, is 0 °C to 22 °C, 21 °C, and 17 °C, respectively.
Figure 4-22: Soil conductivity of three boreholes, from March 27th, 2015, to October 13th, 2016. X axis is time from day 0 to day 567, Y axis is depth from 0 m to about 16 m, and constrained color scale from 0 to 0.23 mS/cm, shows the conductivity of each depth at different time, with red indicates high conductivity. The full conductivity range of borehole E, W, S, is 0.16 mS/cm, 0.21 mS/cm, and 0.23 mS/cm, respectively.
With reference to Figure 4-20, the moisture contents of three boreholes from March 27th, 2015, to October 13th, 2016, are roughly the same. Moisture contents ranged from 0% to about 46%, but the readings may not be truly representative of moisture content in the adjacent natural soil/sediment. It is therefore more meaningful to focus on temporal changes, rather than absolute readings of the moisture content. In borehole W and borehole E, the moisture content is about 18% at shallow surface during almost the entire monitoring period. This compares to a minimum expected residual water content of 7% for the sandy loam soils at the site. The same is likely true in borehole S but we cannot be certain as the shallowest sensor in that hole is at 1.8 m depth, whereas the shallowest STE sensors in boreholes W and E are at 0.4 m and 0.3 m respectively.

The major pattern evident in the moisture data is a series of pulses of elevated moisture content that affect the entire vadose zone in all three boreholes during the times of most recharge (spring 2015, fall, 2015, and spring 2016). The abrupt onset-and the similarly abrupt decrease in moisture content at most depths-suggests that water may be flowing down the boreholes from surface, and possibly also from fractures intersecting the boreholes. This interpretation is supported by the equally abrupt temperature anomalies that appear and disappear in all three boreholes at the same times as the moisture pulses. Thus, it would appear that the layers of coated bentonite pellets inserted into the boreholes during backfilling had not hydrated sufficiently to stop water from flowing preferentially in the boreholes even by the spring of 2016, almost two years after the backfilling had been completed in August, 2014. The effect of these borehole moisture pulses dominates the plots in Figure 4-20, so strongly that it is not possible to identify any other more subtle seasonal trends in moisture content versus depth. In light of this
observation, the soil moisture data in the boreholes is, unfortunately, not considered to be a very reliable indicator of in-situ moisture contents in the vadose zone.

In all three boreholes, moisture content of the bottom of the boreholes raised up to about 40%, meaning fully or nearly fully saturated, in the spring thaw season of 2015. According to the hydrogeological data recorded by the near wells (Figure 4-24), the water table at the study usually stays below the deepest electrode and the deepest 5TE sensors, but it would raise and cover the last 5TE sensors in each borehole around day 25 (April 21\textsuperscript{st}, 2015) to day 55 (May 21\textsuperscript{st}, 2015). Thus, the highly saturated zone at the bottom of each borehole shown in the spring, 2015, could be a result of the raised water table, or the flushed water flowing down the boreholes, and pooled at the bottom of the holes. No significant moisture content changes were observed from surface to the bottom of the boreholes during the summer and fall of 2015, and as mentioned previously, this has been connected to the low infiltration rate during this period. In the early winter of 2015/16, and the spring of 2016, most of the moisture content increase in three boreholes occurred at the same time as in the previous year, with some of the events showing rapid downward flow of water through the boreholes, while other events indicate the presence of perched water conditions. These events were a result of precipitation in the form of rainfall and reduced evaporation, as most of these events happened on the days with high amounts of precipitation (Figure 4-24). However, it should be noted that precipitation events do not necessarily result in increased moisture in the boreholes.

No raised water table was noticed in spring, 2016, this could be caused by the lower precipitation, given that the total precipitation from January 1\textsuperscript{st}, 2015, to April 30\textsuperscript{th}, 2015 was 558 cm, and from January 1\textsuperscript{st}, 2016, to April 30\textsuperscript{th}, 2016 was 319 cm; the water table
in the well #1, and Nest A deep well in spring, 2016, is deeper than it in spring, 2015. The moisture content in all three boreholes did not show significant changes, in the summer of 2016, similar to the summer of 2015.

The temperature variation plots of three boreholes indicate that the shallow subsurface was mostly above zero during the winters of 2014/15, and 2015/16. In the spring of 2015, temperature increased gradually beginning on May 11th, 2015 (Day 45), and then started to decrease as the fall came around October 13th, 2015 (Day 200), until it started to gradually increase again around April 29th, 2016 (Day 400). The annual thermal amplitude at about 0.3 m depth was 22 °C, at about 1 m depth was 10 °C, at about 3 m depth was 5 °C, and for the zone below 8 m deep was less than 2 °C. In all three boreholes, the temperature decreased dramatically on May 4th, 2015 (Day 38), then returned to the normal background trend on May 8th, 2015 (Day 42). Considering that the snow has all melted by May 4th, 2015, this result suggests that cold meltwater was bypassing the vadose zone via boreholes; this would also explain the strong increases in water content observed in all three boreholes on the same day; similar events could be seen in winter, 2015, and spring, 2016.

The temperature effect on subsurface resistivity is predominant at or near 0 °C, below which the resistivity increases sharply due to the freezing of pore water which conducts current. In our study, the top soil may be frozen for a short period of time during spring, 2015, but for most of the monitoring period, ground surface was above the freezing point. In spring, 2015, the tracer application area became much more conductive (over 1000 times more conductive than before the tracer was applied), suggesting that the tracer concentration was the dominant factor affecting the top soil resistivity.
Within 0 to 30 °C, the water resistivity in the subsurface decreases by 1.87% for every 1 °C increase in the temperature (Equation 3-4). Thus, at a constant water saturation rate, the conductivity of water-bearing sediment would be expected to change in the same way. At 1 m depth, the temperature varied by about 10 °C annually (from about 8 °C to 18 °C), while at 3 m depth, the annual temperature range decreased to about 5 °C. The post-tracer ERI models revealed that the resistivity of the subsurface decreased more than 1000 times at the ground surface, and a few hundred times at a deeper region, compared to the pre-tracer models. Thus, the temperature effect on resistivity can be neglected in comparison to the effect of the tracer concentration.

Conductivity variations of in three boreholes (Figure 4-21) did not provide much valuable information. Sudden significant increases in the conductivity readings, were attributed to water bypassing the vadose zone though the boreholes. The conductivity anomalies observed in spring, 2015, and 2016 are believed to be a result of water, rather than tracer, draining into boreholes, as conductivity changes would be more dramatic if caused by tracer.

4.5 Tracer dissolution inferred from very shallow soil conductivity sensors

Figure 4-23 presents the conductivity, moisture content, and temperature of the sensors in the center of the tracer area at 10 cm and 20 cm depth, along with the daily precipitation from day 12 (April 8th, 2015) to day 567 (October 13th, 2016), the data from July 14th, 2015, to October 13th, 2015 were not available. In general, the readings of these sensors are considered more accurate than those of the borehole sensors, as the prongs of the shallow sensors were pushed into the soil to provide better contact between the...
sensors and the soil particles. The temperatures of the two sensors are almost the same, and dropped below 0 °C only for the first few days of the monitoring period. The conductivity and moisture content of the sensor at 10 cm depth show more rapid variations than those of the sensor at 20 cm depth.

In general, following a short period (Day 12 to Day 35) of more intense dissolving of the granular KCl, the conductivity of this area gradually decreases over time. The conductivity measured by the sensor at 10 cm shows a spike on April 23rd, 2015 (Day 28), which coincides with a spike in moisture content (reaching 40%), suggesting that heavy infiltration occurred, and significant amounts of tracer were dissolved and started to migrate downward in the soil profile. However, the conductivity of the deeper sensor did not react as dramatic, thus, not all the dissolved KCl was flushed to 20 cm depth. The conductivity of the shallow sensors started to decrease after April 30th, 2015 (Day 35) and reached a relative minimum on May 2nd, 2015 (Day 38), which was the first day with no snow left on the ground (Environment Canada, 2015). Thus, it is believed that the majority of the tracer had been dissolved, and flushed into the subsurface during the spring of 2015. Some minor spikes were observed during the fall of 2015 and the winter of 2016, indicating that there was still undissolved KCl on the ground during that time.

The moisture content of the two shallow sensors showed the same trend, but the sensor at 20 cm depth is generally moister than the sensor at 10 cm depth, except when heavy infiltration occurred. The shallow sensor show a more rapid and significant response to changes in moisture. Also, as expected, the shallower surface sensor showed lower moisture content during summer.
Figure 4-23: (a) Conductivity of shallow 5TE sensors in logarithm scale; (b) Temperature of shallow 5TE sensors; (c) Moisture of shallow 5TE sensors; (d) Daily precipitation. The two shallow 5TE sensors are installed in the center of the tracer application area, at 10 cm and 20 cm depth, respectively, and the time period shown in this figure are from April 8th (Day 12), to October 13th, 2016 (Day 567). Data from sensor at 10 cm depth is shown in orange, and data from sensor at 20 cm depth is shown in blue.
4.6 Neighbouring well results

Groundwater temperature, conductivity, and elevation were monitored in two wells near the borehole triangle (Figure 3-11).

From March 27th, 2015 (Day 0), to October 13th, 2016 (Day 567) the water table elevation in well #1 was typically about 50 cm higher than that observed in NAD, and no abrupt temporal variations were observed in the two wells during the monitoring period. The water table was highest in April in both 2015, and 2016, and it was higher in 2015 than in 2016, which explains why no obvious water table was observed in the ERI models in the spring of 2016. Water temperatures averaged about 7.25 °C in well #1, and 7.65 °C in NAD, with a total variation of less than 0.3 °C in either well. No rapid variation of water conductivity was observed in NAD, suggesting no tracer had reached the water table as of October 13th, 2016 (Day 576). The three spikes, although small in magnitude, in water conductivity observed in well #1, on April 20th, 2015 (Day 25), December 24th, 2015 (Day 271) and August 19th, 2016 (Day 481) were speculated not related to the tracer applied to the study site on Day 0. This interpretation is supported by the fact that similar spikes appeared in the previous year, before the tracer had been applied.
Figure 4-24: (a) water elevation; (b) temperature; (c) water conductivity of well #1, and nest A deep well (NAD) from day 0 (March 27th, 2015) to day 567 (October 13th, 2016). Data of well #1 are donated in blue, and the data of NAD are donated in orange.
5. Conclusions

5.1 Thesis contribution

This research utilized a surface-applied tracer test with the help of 3D cross-hole ERI and hydrogeological monitoring methods to investigate the solute transport in the vadose zone. The study was designed to improve the understanding of nitrate transport after leaching from the root zone of agricultural crops in Prince Edward Island.

In this research, a 3-hole ERI array with 4896 measurements was developed, and has successfully provided satisfactory resolution of the area of interest, which is a triangular prism starting at ground surface and extending downward 16 m. The instrumentation was able to collect data intensively, relative to the tracer movement rate to monitor its transport in the subsurface.

Geological settings of the study site were determined by the pre-tracer ERI model. Tracer pathways and travel time were monitored by the ERI array, and also involved data collected by 5TE sensors installed at various depths in the ground and data retrieved from the neighboring wells. The results of this research will provide knowledge to understand contamination of groundwater with nitrate and further assist future groundwater quality improvement measures.

A background ERI survey revealed five sub-horizontal layers of alternating resistivity in general agreement with the geology of approximately 6 m soil and glacial till overburden overlying interbedded sandstone and shaley sandstone layers.

Preferential percolation paths have been observed during April, 2015 – within one month after the tracer was applied at the surface of the ground-indicating heterogeneity within
overburden. Post-tracer surveys indicate that tracer had percolated to depths of 1 m, 2.5 m, 3.0 m, 4.5 m and 6.0 m by the 4\textsuperscript{th}, 26\textsuperscript{th}, 32\textsuperscript{nd}, 59\textsuperscript{th}, and 226\textsuperscript{th} days after tracer application. Its movement slowed significantly by early May, 2015, once the snowmelt was completed. Tracer spread laterally very slowly through the summer and early fall, 2015, but remained within the triangular array. The shallow conductivity anomaly produced by the tracer diminished significantly over the winter and spring of 2016, but showed little evidence of bulk matrix flow below approximately 6 m depth. It is speculated that fractures in the bedrock, too thin to be resolved by the ERI survey, conveyed tracer downward. After 18.5 months, there was no ERI evidence of tracer migrating through the matrix deeper than approximately 6 m (top of the middle conductive layer) depth—a finding corroborated by the absence of any significant anomalies in a water conductivity logs measured about 15 m downgradient of the tracer test site.

The majority of tracer had dissolved and infiltrated into subsurface in the spring of 2015, although undissolved granular KCl was observed on the surface of the ground in the summer of 2015. The 5TE measurements suggest that water may rapidly drain through the boreholes, thus bypassing the natural recharge regime in the vadose zone. It is believed that no tracer has moved into the borehole, since the presence of the tracer in the boreholes would cause a significant increase in the conductivity readings. In April and early May, 2015, the raised water table was visualized with the help of the 5TE sensors at the bottom of each borehole, and was confirmed by the ERI models for the respective period.
5.2 Recommendations

The ERI and 5TE monitoring should be continued at the study site, and the water electrical conductivity of the neighboring wells should be monitored, as well. A vertical borehole in the center of the tracer hexagon could be drilled, and the sediments and bedrock should be sampled for $K^+$ and $Cl^-$ concentrations. In this study, it was not possible to rely on the electrical conductivity measured with the 5TE sensors installed in the boreholes, as it is suspected that these sensors were not in good contact with the coarse sand used to backfill the boreholes. If the experiment was to be repeated, a more appropriate moisture/conductivity/temperature sensor could be chosen, and the boreholes should be filled with material more similar to the in-situ geology which would retain moisture better and provide lower contact resistance. The bentonite pellets filled in the boreholes were supposed to stop water draining, however, they did not work as expected. Bentonite slurry would act as a better barrier than bentonite pellets. The research indicated that the tracer was carried downward by the fractures in the bedrock as suggested by the reduction with time of the conductive anomaly observed at shallow depth; if a tracer could be injected into a hole below the top of the bedrock, the monitored tracer movement might allow for a more straightforward interpretation.
6. References


Loke M. H. 2014. "Electrical imaging surveys for environmental and engineering: A practical guide to 2D and 3D surveys."


Appendix A: Filtering System

MATLAB code was used to filter out measurements with reciprocal error higher than 5% (if measurement one uses electrode A and B as current electrodes, and M and N as potential electrodes, measurement two uses electrode M and N as current electrodes, and A and B as potential electrodes, then measurement one and two are reciprocal measurements), standard deviation of repeat voltage measurements better than 3%, and current smaller than 0.1 mA. It also filtered out measurements associated with electrodes that were no longer in contact. In the end, it writes the filtered quadrupole measurements, along with an error estimate (set to 5% of the measured voltage drop) to an ASCII text file in the data file format required by the DCIP3D inversion code.

% This MATLAB code requires two input parameters: an ‘INPUT’ parameter contains the 4896 measurements of one survey, and for each measurement, there are 10 required information, A, B, M, N, I, U, dU, U90, dU90, Tx(transmitter voltage); an ‘topo’ parameter contains the 96 electrodes’ coordinates.

[nrows,ncols] = size(INPUT);

% Filter out measurements associated with electrodes that are no longer in contact (electrode #58, #60, #63, #64)
for k = 1:nrows
    if INPUT(k,1) == 60 || INPUT(k,2) == 60 || INPUT(k,3) == 60 || INPUT(k,4) == 60
        INPUT (k,11) = 1;
    end
end

for a = 1:nrows
    if INPUT(a,1) == 63 || INPUT(a,2) == 63 || INPUT(a,3) == 63 || INPUT(a,4) == 63
        INPUT (a,11) = 1;
    end
end

if INPUT(a,1) == 64 || INPUT(a,2) == 64 || INPUT(a,3) == 64 || INPUT(a,4) == 64
    INPUT (a,11) = 1;
end
if INPUT(a,1) == 58 || INPUT(a,2) == 58 || INPUT(a,3) == 58 || INPUT(a,4) == 58
    INPUT (a,11) = 1;
end
end
indice_without60 = find (INPUT(:,11) == 0);
input = INPUT(indice_without60, [1,2,3,4,5,6,7,8,9,10]);
R=size(input);
p=0;%counter
output(1,1)=0;
for m=1:R;
    for n=1:R;
        if (input(n,1)==input(m,3));
            if (input(n,2)==input(m,4));
                if (input(n,3)==input(m,1));
                    if (input(n,4)==input(m,2));
                        if input(n,1)<input(m,1)
                            p=p+1;
                            output(p,1)=input(n,1);
                            output(p,2)=input(n,2);
                            output(p,3)=input(n,3);
                            output(p,4)=input(n,4); % get all forward measurements
                            output(p,5)=input(n,5);
                            output(p,6)=input(n,7);
                            output(p,7)=input(n,8);
                            output(p,8)=input(m,5);
                            output(p,9)=input(m,7);
                            output(p,10)=input(n,8);
                        else
                            i=5;
                        end
                    else
                        i=3;
                    end
                else
                    i=2;
                end
            else
                i=4;
            end
        else
            i=5;
        end
    end
end
r=size(output); %cals voltage normalized to voltage and reciprocal error
for q=1:r;
    output(q,11)= (output(q,5)/output(q,6));
    output(q,12)= (output(q,8)/output(q,9));
    output(q,13)= abs(output(q,11)-output(q,12))/output(q,11)*100;
end
REC=5; %set Reciprocal error here
p=0; %counter
for l=1:r;
    if output(l,13)<REC;
        p=p+1;
        Output(p,1)=output(l,1);
        Output(p,2)=output(l,2);
        Output(p,3)=output(l,3);
        Output(p,4)=output(l,4);
        Output(p,5)=output(l,5);
        Output(p,6)=output(l,6);
        Output(p,7)=output(l,7);
        Output(p,8)=output(l,8);
        Output(p,9)=output(l,9);
        Output(p,10)=output(l,10);
        Output(p,11)=output(l,11);
        Output(p,12)=output(l,12);
        Output(p,13)=output(l,13);
    else
        i=6;
    end
end
recerror=size(Output);
UERR=3; %set voltage error filter
p=0; %counter
for l=1:recerror;
    if Output(l,7)<UERR;
        p=p+1;
        OUTput(p,1)=Output(l,1);
        OUTput(p,2)=Output(l,2);
        OUTput(p,3)=Output(l,3);
        OUTput(p,4)=Output(l,4);
        OUTput(p,5)=Output(l,5);
        OUTput(p,6)=Output(l,6);
        OUTput(p,7)=Output(l,7);
        OUTput(p,8)=Output(l,8);
        OUTput(p,9)=Output(l,9);
        OUTput(p,10)=Output(l,10);
        OUTput(p,11)=Output(l,11);
        OUTput(p,12)=Output(l,12);
        OUTput(p,13)=Output(l,13);
    else
        i=6;
    end
end
uerr=size(OUTPUT);
p=0; %counter
for l=1:uerr;
    if OUTPUT(l,5)>-9999; %filters out negative voltages
        p=p+1;
        OUtput(p,1)=OUTPUT(l,1);
    end
end
OUTput(p,2)=OUtput(l,2);
OUTput(p,3)=OUtput(l,3);
OUTput(p,4)=OUtput(l,4);
OUTput(p,5)=OUtput(l,5);
OUTput(p,6)=OUtput(l,6);
OUTput(p,7)=OUtput(l,7);
OUTput(p,8)=OUtput(l,8);
OUTput(p,9)=OUtput(l,9);
OUTput(p,10)=OUtput(l,10);
OUTput(p,11)=OUtput(l,11);
OUTput(p,12)=OUtput(l,12);
OUTput(p,13)=OUtput(l,13);

else
    i=6;
end
end
NV=size(OUTput);
p=0; %counter
for l=1:NV;
    if OUTput(l,6)>=.05; %filters out currents smaller than 0.1 mA
        p=p+1;
        OUTPut(p,1)=OUTput(l,1);
        OUTPut(p,2)=OUTput(l,2);
        OUTPut(p,3)=OUTput(l,3);
        OUTPut(p,4)=OUTput(l,4);
        OUTPut(p,5)=OUTput(l,5);
        OUTPut(p,6)=OUTput(l,6);
        OUTPut(p,7)=OUTput(l,7);
        OUTPut(p,8)=OUTput(l,8);
        OUTPut(p,9)=OUTput(l,9);
        OUTPut(p,10)=OUTput(l,10);
        OUTPut(p,11)=OUTput(l,11);
        OUTPut(p,12)=OUTput(l,12);
        OUTPut(p,13)=OUTput(l,13);
        OUTPut(p,14)=abs(OUTput(l,11)*0.05); %Error for DCIP3D
    else
        i=6;
    end
end
[nrows, ncols] = size(OUTPut);

for a = 1:nrows
    if OUTPut(a,1) == 84 || OUTPut(a,1) == 84 || OUTPut(a,1) == 84 || OUTPut(a,1) == 84
        OUTPut(a,14) = abs(OUTPut(a,11))*0.5;
    end
    if OUTPut(a,1) == 4 || OUTPut(a,1) == 4 || OUTPut(a,1) == 4 || OUTPut(a,1) == 4
OUTPut(a,14) = abs(OUTPut(a,11))*0.5;
end

if OUTPut (a,1) == 28 | OUTPut (a,1) == 28 | OUTPut (a,1) == 28 | OUTPut (a,1) == 28
  OUTPut(a,14) = abs(OUTPut(a,11))*0.5;
end

if OUTPut (a,1) == 76 | OUTPut (a,1) == 76 | OUTPut (a,1) == 76 | OUTPut (a,1) == 76
  OUTPut(a,14) = abs(OUTPut(a,11))*0.5;
end

if OUTPut (a,1) == 92 | OUTPut (a,1) == 92 | OUTPut (a,1) == 92 | OUTPut (a,1) == 92
  OUTPut(a,14) = abs(OUTPut(a,11))*0.5;
end
end

SC=size(OUTPut);
% AxAyA2BxBxBzMzMzNyNyNz format with voltage, and voltage error.
p=0;
for m=1:SC
  OUTPUT(m,1)=topo(OUTPut(m,1),1);
  OUTPUT(m,2)=topo(OUTPut(m,1),2);
  OUTPUT(m,3)=topo(OUTPut(m,1),3);
  OUTPUT(m,4)=topo(OUTPut(m,2),1);
  OUTPUT(m,5)=topo(OUTPut(m,2),2);
  OUTPUT(m,6)=topo(OUTPut(m,2),3);
  OUTPUT(m,7)=topo(OUTPut(m,3),1);
  OUTPUT(m,8)=topo(OUTPut(m,3),2);
  OUTPUT(m,9)=topo(OUTPut(m,3),3);
  OUTPUT(m,10)=topo(OUTPut(m,4),1);
  OUTPUT(m,11)=topo(OUTPut(m,4),2);
  OUTPUT(m,12)=topo(OUTPut(m,4),3);
  OUTPUT(m,13)=OUTPut(m,11);
  OUTPUT(m,14)=OUTPut(m,14);
end
for m=1:SC; %DCIP3D format
  OUTPUT(3*m-2,1)=OUTPut(m,1);
  OUTPUT(3*m-2,2)=OUTPut(m,2);
  OUTPUT(3*m-2,3)=OUTPut(m,3);
  OUTPUT(3*m-2,4)=OUTPut(m,4);
  OUTPUT(3*m-2,5)=OUTPut(m,5);
  OUTPUT(3*m-2,6)=OUTPut(m,6);
  OUTPUT(3*m-2,7)=1;
  OUTPUT(3*m-2,8)=999;
  OUTPUT(3*m-1,1)=OUTPut(m,7);
  OUTPUT(3*m-1,2)=OUTPut(m,8);
  OUTPUT(3*m-1,3)=OUTPut(m,9);
  OUTPUT(3*m-1,4)=OUTPut(m,10);
  OUTPUT(3*m-1,5)=OUTPut(m,11);
  OUTPUT(3*m-1,6)=OUTPut(m,12);
OUTPUT(3*m-1,7)=OUTPUT(m,13);
OUTPUT(3*m-1,8)=OUTPUT(m,14);
OUTPUT(3*m,1)=999;
OUTPUT(3*m,2)=999;
OUTPUT(3*m,3)=999;
OUTPUT(3*m,4)=999;
OUTPUT(3*m,5)=999;
OUTPUT(3*m,6)=999;
OUTPUT(3*m,7)=999;
OUTPUT(3*m,8)=999;
end
Appendix B: ERI Measurements Summary

Table B1 is a summary of post-tracer ERI measurements during the study period, from March 27\textsuperscript{th}, 2015 (Day 0), to October 13\textsuperscript{th}, 2016 (Day 567). There were 639 surveys collected in the monitoring period, including both Lippmann and Geotest surveys.

<table>
<thead>
<tr>
<th>Days</th>
<th>Dates</th>
<th>Number of Surveys</th>
<th>Measurement times</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>2015-03-27</td>
<td>1</td>
<td>One Geotest survey (5 Hz)</td>
<td>Field visit by Serban and Shuang for tracer application; Tracer applied; Overcast</td>
</tr>
<tr>
<td>1-3</td>
<td>2015-03-28 to 2015-03-30</td>
<td>0</td>
<td>No measurements</td>
<td>Crew returned to Fredericton, no snow melting had been expected</td>
</tr>
<tr>
<td>4</td>
<td>2015-03-31</td>
<td>3</td>
<td>Three Geotest surveys (Two at 5 Hz, one at 10 Hz)</td>
<td>Field visit by Shuang for data collecting; Lippmann left to record one overnight survey</td>
</tr>
<tr>
<td>5</td>
<td>2015-04-01</td>
<td>2</td>
<td>One Geotest survey (5 Hz); One-time Lippmann survey at 0:00 am</td>
<td>Field visit by Shuang for data collecting;</td>
</tr>
<tr>
<td>6-11</td>
<td>2015-04-02 to 2015-04-07</td>
<td>18</td>
<td>One Geotest survey (10 Hz); Lippmann survey three times daily, starting at 2015-04-02 16:00 pm</td>
<td>Field visit by Shuang for data collecting; Lippmann left to record three times per day at 16:00, 00:00, and 08:00 hrs local time</td>
</tr>
<tr>
<td>12</td>
<td>2015-04-08</td>
<td>2</td>
<td>Two Geotest surveys (5 Hz &amp; 10 Hz)</td>
<td>Field visit by Shuang for data collecting;</td>
</tr>
<tr>
<td>13</td>
<td>2015-04-09</td>
<td>2</td>
<td>Two Geotest surveys (5 Hz &amp; 10 Hz)</td>
<td>Lippmann left to record 2 overnight surveys</td>
</tr>
<tr>
<td>14-21</td>
<td>2015-04-10 to 2015-04-16</td>
<td>20</td>
<td>Two Lippmann surveys at 0:00 am and 04:30 am on 2015-04-10; One Geotest survey (10 Hz) on 2015-04-10; Lippmann survey three times daily, starting at 2015-04-10 16:00 pm</td>
<td>Lippmann left to record three times per day at 16:00, 00:00, and 08:00 hrs local time</td>
</tr>
<tr>
<td>22</td>
<td>2015-04-17</td>
<td>2</td>
<td>Two Geotest surveys (5 Hz &amp; 10 Hz)</td>
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<td>2015-04-18 to 2015-04-29</td>
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<td>Lippmann left to record three times per day at 16:00, 00:00, and 08:00 hrs local time</td>
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<td>One Geotest survey (10 Hz) on 2015-04-30; Lippmann survey three times daily, starting at 2015-04-30 16:00 pm; One Geotest survey (5 Hz) on 2015-05-12;</td>
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<td>One Geotest surface survey (5 Hz) on 2015-05-26; Lippmann survey three times daily starting at 2015-05-26 16:00 am</td>
<td>Field visit by Shuang and Keenan for data collecting surface survey; Lippmann left to record three times per day at 00:00, 08:00, and 16:00 hrs local time; Data from this period of time was erroneous</td>
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<td>119- 127</td>
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<td>Lippmann survey once daily starting at 2016-09-25 0:00 am;</td>
<td>Lippman left to record once per day at 00:00 hrs local time</td>
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Appendix C: Borehole and Electrode Geometry

The three borehole locations and the trench electrode positions were determined using RTK (real time kinematic) GPS positioning with a base station set up over the nearby well in Nest A (Nest A deep well). The UTM (Universal Transverse Mercator) grid coordinates and elevations above mean sea level for the uppermost electrode in each hole are given in Table C1. For ease of handling, local grid, aligned with UTM, was established with its origin at UTM coordinates (388277.4651, 699623.7094) and elevation at 27.883 m, corresponding to a point on surface near the top of borehole W. Borehole inclination angles (Table C2), measured from steel casing stick-ups of about 30-40 cm, and the electrode separation of 68 cm were then used to extrapolate the borehole electrode coordinates from the measured location of the uppermost electrode in each hole. The local XYZ coordinates of all electrodes are listed in Table C3. The borehole electrodes (E1, W1, S1, etc) were numbered from 1 to 24 from shallowest to deepest. The trench electrodes were labeled from 1 to 8 in each trench; for example, electrode EW1 was the electrode in trench EW, close to borehole E.

Table C1: UTM Grid Coordinates and Elevations

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<tr>
<th>Uppermost Electrode in Borehole X</th>
<th>Easting</th>
<th>Northing</th>
<th>Elevation above Sea Level (m)</th>
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<td>388281.9271</td>
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<td>Borehole W</td>
<td>388275.5613</td>
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<td>Borehole E</td>
<td>388284.1748</td>
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Table C2: Borehole Inclinations

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<th>Borehole</th>
<th>Borehole W (direction top of casing tilted towards)</th>
<th>Borehole E (WNW)</th>
<th>Borehole S (NE)</th>
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<td>285°</td>
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Table C3: Electrode Coordinates

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<th>Electrode #</th>
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Appendix D: Deep Vadose ERI Monitoring Sequence

There are 4896 apparent resistivity measurements (Table D1), including the reciprocals, measured for each survey, to yield a desirable resistivity model for the study site. The sequence is listed below. The electrodes are labeled from 1 to 96; electrodes 1 to 24 were installed in borehole W; electrodes 25 to 48 were installed in borehole E; electrodes 49 to 72 were installed in borehole S; and electrodes 73 to 96 were installed in the trenches.

A and B are the electrodes being used as current dipole, and M and N are the electrodes being used as potential dipole.

Table D1: Deep Vadose ERI monitoring sequence.

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Vitae

Candidate’s full name: Shuang Wang

Universities attended: Jilin University, China

B. Applied Physics. 2013

Conference Presentations:

1. Shuang Wang, Karl Butler, Serban Danielescu, Mark Grimmett and Brent Petersen, “Dynamic solute transport, as a proxy of nitrate transport, through the vadose zone under a potato field by a 19- month tracer test and cross-hole resistivity image”, American Geophysics Union Fall Meeting, (San Francisco, US), December, 2016

2. Shuang Wang, Karl Butler, Serban Danielescu and Brent Petersen, “Dynamics of solute transport through the deep vadose zone under a potato field as assessed by a year-long tracer test and cross-hole resistivity imaging”. Annual Scientific Meeting of the Canadian Geophysical Union, (Fredericton, NB, Canada), May 30-June 2, 2016.

3. Shuang Wang, Karl Butler, Serban Danielescu and Brent Petersen, “Monitoring percolation of a conductive tracer by cross-hole electrical resistivity imaging at Harrington Research Farm, PEI”. Agriculture and Agri-Food Canada (AAFC), Scientific Seminars, (Fredericton, NB, Canada), May 4, 2016.

4. Shuang Wang, Karl Butler, Serban Danielescu and Brent Petersen, “Monitoring water and conductive tracer percolation rates and processes through deep vadose zone at Harrington Research Farm in PEI”. Agriculture and Agri-Food Canada (AAFC), Program Meeting, Growing Forward 2 Program, (Charlottetown, PE, Canada), June 18-19, 2015.