The Pennsylvania State University
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CHARACTERIZING SOLUTE TRANSPORT IN COUPLED STREAM-HYPORHEIC SYSTEMS USING ELECTRICAL RESISTIVITY IMAGING

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by

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ABSTRACT

Hyporheic zones contribute a number of ecological services to streams, including buffering of stream temperatures, carbon and nutrient cycling, and providing habitat used by a variety of species. The hyporheic zone has been referred to as the stream’s liver because of its ability to process pollutants in a zone of extensive biogeochemical activity. Despite the clear relationship between hyporheic exchange and numerous ecosystem services, the distribution of these services (a function of the hyporheic flowpath network itself) remains unknown. Common techniques to quantify hyporheic exchange rely on solute tracers, where downstream observations are used to infer transport processes occurring upstream. This type of inverse modeling suffers from a lack of physically meaningful parameters, lack of spatial resolution (i.e., providing only reach-average parameter sets), and results that suffer from a window of detection problem (i.e., the inability to quantify flowpaths at a spatial or temporal scale greater than some threshold, which is itself a function of the experimental design). Direct observation in subsurface monitoring wells provides spatially discrete subsurface measurements, but lacks an extensive spatial coverage. Monitoring wells characterize only those flowpaths that directly intersect the well. Three-dimensional modeling of flow and transport is possible, but model validation and calibration are data intensive, and may not capture heterogeneity that exists in natural systems.

This thesis presents work using electrical geophysical methods to overcome the limitations of traditional characterization methods. The first study provides a proof-of-concept in the use of electrical resistivity imaging and electrically conductive tracers to quantify hyporheic exchange. This study demonstrates the ability to quantify physical characteristics of the hyporheic zone through time and discusses the uncertainty in the method. Results demonstrate that not only does the spatial distribution of tracer concentration change in the hyporheic zone through time, but the physical size of the hyporheic zone is temporally variable. The second
study demonstrates the use of geophysical imaging to characterize subsurface flowpaths that might be otherwise undetected using traditional geophysical study. The third study includes four replicate solute injections completed during baseflow recession in a headwater stream. This study suggests hyporheic exchange is highly variable as a function of catchment wetness, expressed by baseflow. The fourth study presents a numerical study developed to test the use of temporal moments in characterizing subsurface solute transport processes. Results of this study suggest temporal moments of both solute concentration and electrical resistivity can be used to characterize solute transport processes. Finally, the fifth study applies temporal moment analysis to time-lapse electrical resistivity images collected in the field. This study demonstrates the distributed analysis of physical transport processes in the subsurface. This body of work establishes electrical resistivity imaging as a tool to quantify stream-hyporheic interactions in the field, and represents a substantial advance in our ability to observe coupled surface-subsurface processes in-situ.
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For Melissa, whose infinite support and patience made this possible.
We must, in fact, not divorce the stream from its valley in our thoughts at any time. If we do we lose touch with reality.

H.B.N. Hynes, 1975

*Edgardo Baldi memorial lecture: The stream and its valley*
Chapter 1

State of the science: Characterizing hyporheic flow and transport

1.1 Background

Hyporheic zones (i.e., locations in the near-stream aquifer that are highly connected with streams over relatively short spatial and temporal scales) have been identified as important locations of biogeochemical cycling [Baker et al., 1999; Triska et al., 1989], habitats for macroinvertebrates [Stanford and Ward, 1988; Williams and Hynes, 1974] and buffers for both stream water temperatures [Arrigoni et al., 2008; Burkholder et al., 2008] and a variety of pollutants [D'Angelo et al., 1993; Fuller and Harvey, 2000; Schwarzenbach et al., 1983]. Hyporheic exchange (i.e., the movement of water between the active channel and hyporheic zone) is an integral part of denitrification in headwater streams, due to residence times and redox gradients [Dahm et al., 1998]. Human-altered reaches, such as much of the nations’ drainage infrastructure, minimize the connectivity between streams and riparian ground waters, thereby restricting the beneficial impacts of hyporheic exchange [Dahm et al., 1998]. Despite the clear relationship between hyporheic exchange and numerous ecosystem services, the distribution of these services (a function of the hyporheic flowpath network itself) remains unknown.

Substrate permeability and vertical hydraulic gradients are the primary drivers of hyporheic exchange [Crispell and Endreny, 2009]. The effect of natural in-stream geomorphologic features on hyporheic exchange has been the topic of much research [Goosseff et al., 2006; Harvey and Bencala, 1993; Hester and Doyle, 2008]. Vertical hydraulic gradients, which drive hyporheic exchange, have been found to increase with surface concavity [Anderson
Streambed permeability or hydraulic conductivity exerts primary control on intragavel flow, and has received much attention by modern researchers. Packman and Salehin [2003] found sedimentary conditions may control the extent of hyporheic exchange in detailed flume studies. Even small-scale heterogeneity in substrate hydraulic conductivity results in complex hyporheic exchange patterns [Salehin et al., 2004; Woessner, 2000].

The geologic setting of a stream reach provides a first-order control on hyporheic exchange, setting the template upon which exchange may occur [Wondzell and Gooseff, In Press]. Vallet et al. [1996] demonstrated the relationship between parent lithology and hyporheic exchange with solute tracer studies in three different geologic settings. Valley constraint is a control on hyporheic exchange, with larger transient storage zones in unconstrained reaches [e.g., solute tracer studies by D'Angelo et al., 1993]. Valley constraint was found to reduce hyporheic residence time in a paired study of two headwater streams in paired catchments [Wondzell, 2006], and was identified as a control in larger floodplain systems [Wright et al., 2005]. Larkin and Sharp [1992] demonstrated that stream-aquifer interactions are a function of channel slope, sinuosity, incision, and aspect ratio, as well as the depositional system. While their study considers the interaction of valley morphology and stream-groundwater interactions at a macroscale, few studies have considered how this regional framework acts as a boundary condition to smaller-scale exchange processes occurring in the valley bottom. Field studies have confirmed that longitudinal valley gradient is a control that influences the pattern of subsurface flows [Wondzell and Swanson, 1996].

Nested within this larger-scale context, riparian hydrology exists as a buffer between boundaries set by large-scale, relatively slow-moving processes at the hillslope, catchment, or regional setting and more dynamic processes occurring within the stream channel itself. Burt [2005] notes “the riparian zone is perhaps the most important element of the hydrological
landscape given that it can decouple the linkage between the major landscape elements, hillslope and channel.” Left out of this perspective is hyporheic hydrology, in which some portion of the riparian zone is comprised of stream water traveling along hyporheic flowpaths. Several studies have focused on riparian hydrology, linking hillslopes to channels [e.g., Vidon and Hill, 2004]. Gilbert et al. [1990] proposed a classification system linking hydrologic regimes with hyporheic ecotones, based on the relative dominance of surface water and groundwater dynamics, two end-members of the spectrum of hydrologic boundary conditions. Such conceptual models do not accommodate hyporheic hydrology as a separate process; rather they treat the riparian zone as a location that is grossly influenced by boundaries (i.e., head gradients between the stream and hillslopes) [Duval and Hill, 2006].

The view of riparian hydrology as a function of its boundaries is prevalent in the literature. Integrated studies of coupled stream and groundwater controls on exchange are not common; the two have traditionally been investigated independently seeking to characterize the control of either boundary on exchange dynamics. The stream has been studied as a boundary condition, including the individual controls of velocity and discharge. Theoretical and flume studies of pumping exchange for streams with planar beds and triangular, repeating bedforms [Elliott and Brooks, 1997a; b] demonstrate hyporheic exchange is proportional to velocity squared [Packman and Salehin, 2003]. Solute tracer studies conducted at the reach-scale generally agree that the ratio of storage zone area to stream area (commonly \(A_S/A\)) decreases with increasing flow [Butturini and Sabater, 1999; Fabian et al., 2010; Karwan and Saiers, 2009; Morrice et al., 1997; Schmid et al., 2010; Zarnetske et al., 2007], though other studies have found little correlation [Hart et al., 1999; Schmid et al., 2010]. Increasing discharge may increase hyporheic exchange rate [Fabian et al., 2010; Hart et al., 1999]. Field-based studies have also shown no relationship between subsurface residence time and discharge [Butturini and Sabater,
1999], while others have shown increased mean residence time during baseflow recession due to rapid changes in in-channel storage areas [Legrand-Marcq and Laudelout, 1985]. The stream represents an integration of catchment processes, where local-scale variability due to channel morphology interacts with discharge.

The regional groundwater setting, the second end-member of the spectrum considered by Gilbert et al. [1990], represents a hydrodynamic boundary that is spatiotemporally variable on timescales that may be longer than residence times of individual hyporheic flowpaths. D’Angelo et al. [1993] found that lateral inflows, as a percentage of flow in a reach, were greatest in headwater streams, suggesting these locations might be most sensitive to fluctuations in groundwater discharge. Indeed, a number of reviews and conceptual manuscripts hypothesized interactions between regional groundwater tables and hyporheic exchange [e.g., Hakenkamp et al., 1993; Hynes, 1983; Meyer et al., 2008; Palmer, 1993; Vervier et al., 1992; White, 1993].

Numerical studies of pumping exchange due to triangular bedforms demonstrate a contraction of hyporheic flowpaths due to ambient gaining conditions [Boano et al., 2008; Cardenas and Wilson, 2007b]. Field studies of exchange, however, have yielded contradictory results. Several studies report contraction of hyporheic flowpaths during periods of increased groundwater discharge to streams [Harvey and Bencala, 1993; Storey et al., 2003; Williams, 1993; Wondzell and Swanson, 1996; Wroblicky et al., 1998]. In contrast, Wondzell [2006] completed tracer studies in steep headwater catchments in the H.J. Andrews Experimental Forest under different baseflow conditions (4.5 and 1 L s$^{-1}$ in WS1, 10 and 3 L s$^{-1}$ in WS3). The study found hyporheic extent, evaluated as tracer arrival in a monitoring well network, was unchanged across baseflow conditions at 4 and 35 L s$^{-1}$.

In a recent opinion paper, Bencala et al. [2011] suggest that “a stream is a dynamic expression of local groundwater conditions, where exchanges of water between the catchment
and the channel are continuously changing in response to heterogeneous temporal and spatial water table dynamics.” Indeed, several recent studies have begun to consider coupled groundwater and stream processes. Wondzell [2006] reported hyporheic exchange and extent as a function of valley setting and baseflow recession, using stream flow as a proxy for the hillslope-riparian-hyporheic-stream continuum. Stream velocity has been linked with catchment evapotranspiration signals as a control on the movement of water and solutes through riparian zones [Wondzell et al., 2007]. A subsequent study by the same authors demonstrated that hyporheic response is a function of both in-channel and catchment controls operating across a range of spatial and temporal scales [Wondzell et al., 2010]. Covino and McGlynn [2007] linked stream gains and losses with valley aquifer storage and stream water chemistry. Further studies demonstrate the role of hillslope-riparian-stream connections as a control on the magnitude and timing of both flow rates and solute transport observed at a catchment’s outlet [Jencso et al., 2010]. Still, current conceptual models are only beginning to consider stream-catchment connections and their role in determining solute transport and transformation within a catchment [Bencala et al., 2011].

While a comprehensive understanding of how catchments and streams control hyporheic exchange is lacking, it is recognized that hyporheic hydrodynamics are the template upon which biogeochemical cycling in the subsurface occurs [Argerich et al., In Review; Battin, 1999; 2000]. Recent numerical and field studies have demonstrated zonation of biogeochemical processes in the subsurface as a function of flowpath residence time [Boano et al., 2010; Zarnetske et al., 2011]. Groundwater flow models have been widely applied to study hyporheic flowpaths and residence times, yet such models all suffer from equifinality and an inability to validate modeling based on spatially discrete measurements [e.g., Beven, 2006]. After a detailed modeling study of flow and transport in a headwater stream, Wondzell et al. [2009] conclude “for detailed analysis...
of solute transport pathways and breakthroughs, intensive sampling of the subsurface may be necessary.” Hanrahan [2008] concluded that local-scale measurements may not be sufficient to explain interactions between streams and hyporheic zones. Despite an extensive body of field studies and a range of modeling efforts (from simple box-models to three-dimensional models of flow and transport), we are not yet able to predict the spatial and temporal distribution of hyporheic flowpaths, nor their relationship to controlling hydrodynamics. Previous studies suggest the limiting factor in our scientific understanding of hyporheic dynamics is the toolkit commonly applied to such problems [Bencala et al., 2011; Wondzell, 2006; Wondzell et al., 2009].

1.2 Characterizing hyporheic exchange and transport

Increased hyporheic exchange holds the potential to improve the physical, chemical, and biological services provided by stream habitats, and thereby benefit stream ecosystems. The hyporheic zone includes source, sink, and refuge functions that affect the biological, chemical, and physical integrity of headwater streams [Leibowitz et al., 2007]. In order to manage or design for hyporheic exchange, the ability to quantify the spatial and temporal dynamics of hyporheic exchange is necessary (i.e., where do hyporheic flowpaths occur?).

Traditional characterization of solute transport through coupled stream-hyporheic systems uses in-stream solute tracers and observations made in the stream channel to characterize exchange [Stream Solute Workshop, 1990]. This method generally uses in-stream observations to make inferences about transient storage of solute tracers. An observed downstream solute breakthrough curve is an integration of all of the upstream transport processes that occur within the study’s window of detection [Harvey and Wagner, 2000]. Simplified transport equations and
numerical solutions are widely implemented [e.g., Bencala and Walters, 1983; Runkel, 1998; Thackston and Schnelle, 1970], but known to lack physical meaning [Marion et al., 2003; Wondzell, 2006]. Such simplified models lack the ability to fit observations of late-time behavior [Zaramella et al., 2003], possibly owing to the assumed exponential residence time distribution [Gooseff et al., 2003; Haggerty et al., 2002]. Finally, this method overlooks down-valley groundwater transport processes, which have been identified as important in some settings [Castro and Hornberger, 1991; Jackman et al., 1984; Kennedy et al., 1984].

Given the inability of solute tracers alone to characterize subsurface transport of solutes, shallow monitoring wells have been widely installed to monitor for stream tracers [e.g., Wondzell, 2006] and characterize hydraulic conductivity of the subsurface [Baxter and Hauer, 2000; Nowinski et al., 2011]. Sampling pore water from these wells during tracer studies is problematic because the withdrawal of water from the well field modifies the flow field itself and provides a spatially discrete measurement. Finally, well observations are limited by the ability to install wells using hand-driven wells where equipment access is limited [Baxter et al., 2003; Burkholder et al., 2008]. Direct investigation of the subsurface by digging of soil pits and trenches, or manual probing of the subsurface to identify bedrock interfaces is also common, albeit invasive [e.g., Jones et al., 2008; Nyquist et al., 2008; Nyquist et al., 2010; Wondzell et al., 2009]. Widely applied methods are summarized in Table 1-1, along with our objective of a high spatial and temporal coverage that is able to observe processes in-situ.
In contrast to direct probing or sampling of the subsurface, hydrogeophysical methods can be used to characterize subsurface structure and hydrological parameters, and to monitor transport processes [Binley et al., 2010]. Two types of subsurface characterizations that are possible using hydrogeophysical tools: process areas, or geologic properties that do not change over time, and process dynamics, or characteristics that change through time, are inherently linked to the movement of fluids (primarily water) in the subsurface [Binley et al., 2010; Koch et al., 2009]. Indeed, geophysical investigation to characterize process areas for study of hyporheic exchange is growing in popularity, with demonstrated application of electrical resistivity [Crook et al., 2008], ground penetrating radar [Brosten et al., 2006; Jones et al., 2008; Naegeli et al., 1996], and seismic refraction and reflection [Anderson et al., 2005; Jones et al., 2008]. These techniques are used to characterize subsurface architecture based on changes in the physical properties of the subsurface, and commonly produce continuous spatial distributions of such properties.

<table>
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<tr>
<th>Method</th>
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<tr>
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<td>High</td>
<td>Low</td>
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</tr>
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1.3 Research objectives

Hyporheic exchange near headwater streams is widely recognized as an important ecological process, but we lack an understanding of where and when transport along hyporheic flowpaths occurs. Controls on hyporheic exchange are readily identified by considering Darcy’s law,

\[ q = K \frac{dh}{dl} \quad (1) \]

a simplification of the groundwater flow equation describing area-averaged groundwater flow, where Darcy flux \( q \) (units L/T) is a function of the distribution of hydrogeologic parameters in the subsurface \( K \) (the hydraulic conductivity, units L/T) and the spatiotemporally variable hydraulic gradient \( \frac{dh}{dl} \) (units L/L). In the hyporheic literature, these are often described as subsurface architecture (i.e., the heterogeneous distribution of hydrogeologic properties in the subsurface), and riparian hydraulic gradients (in vertical, lateral, and/or down-valley components).

Despite a rich body of literature describing groundwater flow and identifying the independent variables necessary to describe subsurface flowpaths, we lack an understanding of how these variables control hyporheic exchange near headwater streams. The overarching goal of this study is to address the fundamental questions:

- Where do hyporheic flowpaths exist?
- How do solutes move along hyporheic flowpaths?

Each of these questions explores a dependent variable (hyporheic extent, hyporheic solute transport) as a function of the independent variables spatiotemporally dynamic hydrologic boundary conditions (i.e., hydraulic gradients) and subsurface architecture.
1.3.1 Spatiotemporally dynamic hydraulic gradients as a control on hyporheic exchange

Although it is widely recognized that hydraulic gradients are a control on hyporheic exchange, many researchers consider only vertical hydraulic gradients and/or cross-valley gradients (i.e., those perpendicular to a stream, assume flow only along lateral 2-D planes). Hydraulic gradients between the stream and larger-scale groundwater setting (in the vertical, cross-valley and down-valley) are reported as controls hyporheic exchange, yet field observations characterizing hyporheic response to changing boundary conditions are lacking. This study seeks to answer the fundamental questions:

• How does hyporheic extent respond to changing boundary conditions (i.e., dynamic hydraulic gradients in the stream and riparian zone)?
• How do spatiotemporally dynamic hydraulic gradients control transport along hyporheic flowpaths?

1.3.2 Subsurface architecture as a control on hyporheic exchange

While the spatially heterogeneous distribution of hydrogeologic parameters (e.g., porosity, permeability) in the subsurface is recognized as a control on hyporheic exchange, past studies have relied on spatially lumped down-gradient observations to infer upstream behavior. Detailed numerical modeling efforts are similarly limited by an inability to characterize subsurface architecture based on these limited observations. In order to better inform a process-based understanding of hyporheic exchange, measurements of exchange that are both spatially distributed and observed in-situ are necessary. To that end, this study uses electrical resistivity imaging to inform both spatial extent of hyporheic flowpaths and transport of solutes along these
flowpaths. To open the “Pandora’s box” of subsurface heterogeneity, this study seeks to answer the fundamental questions:

- How does subsurface architecture control hyporheic extent?
- How does subsurface architecture influence hyporheic transport processes, and responses due to changing boundary conditions?

### 1.4 Summary of research efforts

The research presented in this thesis advances the ability to quantify the spatiotemporally variable controls on hyporheic flowpaths in headwater streams using field and numerical studies of (1) hydraulic gradients and (2) subsurface architecture as the primary controls on hyporheic exchange.

The first study (Chapter 2) serves as a proof-of-concept in using electrical resistivity imaging to characterize solute transport along hyporheic flowpaths. The time-lapse imaging of tracer transport allows distributed observation of flowpaths (2D along transects perpendicular to the stream channel). The procedures for quantifying hyporheic area from inverted models are established in this manuscript, and distributed hyporheic dynamics were observed in-situ for the first time in published literature.

The second study (Chapter 3) explores the use of different subsurface investigation techniques to characterize subsurface architecture. Data, including surface observations, direct probing of the subsurface, seismic refraction, and electrical resistivity (ER) imaging during a solute tracer study are used to characterize increasingly complex models. The study demonstrates that while geophysical techniques may be used to characterize subsurface structure, only those studies sensitive to a solute tracer provide information about transport dynamics. It is also shown
that the distribution of geophysical parameters (i.e., electrical resistivity) in the subsurface is not necessarily correlated with the distribution of hydrogeologic properties.

The third study (Chapter 4) investigates how hyporheic extent changes during baseflow recession in a small mountain stream. Replicate studies conducted during a three-month period characterize in-stream flow, vertical hydraulic gradients, and valley-bottom hydraulic gradients as controls on hyporheic exchange. ER imaging of hyporheic transport is analyzed for hyporheic cross-sectional area, and spatial moments are used to track changes in plume centroid location and variance in both the vertical and lateral dimensions. While published studies suggest simple relationships, this study demonstrates spatially variable controls on hyporheic exchange.

The fourth study (Chapter 5) uses a numerical model to explore the use of temporal moments in characterizing flow and transport processes. Simulations of near-surface ER methods and conservative solute transport were used to compare simulated temporal trends in ER and solute transport data. Results show that temporal moments based on ER data are adequate to characterize transport in complex systems, particularly for areas where advection is not the dominant solute transport process. The study demonstrates the usefulness of ER data in the time domain to increase our understanding of the fate and transport of subsurface solutes.

The fifth study (Chapter 6) applies the temporal moment techniques presented in Chapter 5 to data collected during four replicate solute tracer studies in a headwater stream. The analysis demonstrates that temporal moment analysis of ER images is a useful analysis, even with the noise and uncertainty inherent in field data. Distributions of pixels are used to generate probability density functions for each normalized temporal moment (i.e., mean arrival time, variance, and skewness) at each transect during each injection. The study uses both qualitative and quantitative analyses of transport during the baseflow recession period to characterize the spatiotemporal distribution of solute transport processes.
The unifying goal of this research is to characterize spatiotemporally variable controls on hyporheic exchange using novel geophysical methods that provide spatially distributed characterization of flowpaths in-situ. Data collected in headwater streams were coupled with numerical modeling to provide both empirical observations and generalizable modeling results. Chapter 7 provides a synthesis of the studies included in this dissertation, summarizing the promise of geophysical monitoring of subsurface flows in advancing our understanding of hyporheic exchange. An improved understanding of how solutes move through coupled stream-hyporheic systems is the first step in management and restoration of these ecologically important flowpaths. Results will also allow for development of more accurate and better-constrained models of flow and transport through stream networks.
Chapter 2

Imaging hyporheic zone solute transport using electrical resistivity

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2.1 Abstract

Traditional characterization of hyporheic processes relies upon modeling observed in-stream and subsurface breakthrough curves to estimate hyporheic zone size and infer exchange rates. Solute data integrate upstream behavior and lack spatial coverage, limiting our ability to accurately quantify spatially heterogeneous exchange dynamics. Here, we demonstrate the application of near-surface electrical resistivity imaging (ERI) methods, coupled with experiments using an electrically conductive stream tracer (dissolved NaCl), to provide in-situ imaging of spatial and temporal dynamics of hyporheic exchange. Tracer-labeled water in the stream enters the hyporheic zone thereby reducing electrical resistivity in the subsurface (to which subsurface ERI is sensitive). Comparison of background measurements with those recording tracer presence provides distributed characterization of hyporheic area (in this application, ~0.5 m$^2$). Results demonstrate the first application of ERI for two-dimensional imaging of stream-aquifer exchange and hyporheic extent. Future application of this technique will greatly enhance our ability to quantify processes controlling solute transport and fate in hyporheic zones and provide data necessary to inform more complete numerical models.
2.2 Introduction

The exchange of stream water with near-stream aquifers and the associated physical, chemical, and biological processes have been shown to provide a number of ecosystem services [Brunke and Gonser, 1997]. The quantification of these hyporheic processes, however, has been limited by our ability to understand the magnitude, extent, and spatial variability of exchange between the stream channel and the hyporheic zone. Improved characterization of these exchange processes in space and time is, therefore, critical to our understanding of the role of the hyporheic zone in stream ecosystem and water quality function.

Stream tracer studies are frequently used to estimate the interaction of the mobile stream domain with the less mobile hyporheic zone [Stream Solute Workshop, 1990]. Such studies produce in-stream breakthrough curves that are a product of integrated transport processes over a reach (e.g., advection, dispersion), and exhibit tailing behavior commonly attributed to solute exchange between the stream and near-stream aquifers or in-stream zones of low velocity [Harvey et al., 1996]. Stream solute transport modeling often yields estimates of the reach-representative extent of lumped mobile and immobile zones and exchange rates [e.g., Wagner and Harvey, 1997], which rely on observations in the surface mobile domain to characterize less mobile domains. Such results have been identified as not physically meaningful [Marion et al., 2003; Wondzell, 2006]. Further limiting our understanding of the hyporheic zone from these models are the problems of tracer experiment sensitivity to more than storage zone processes [Harvey et al., 1996] and the difficulty in accurately estimating precise storage zone characteristics. These discrepancies are due in part to the representation of transient storage as a single, well-mixed zone and are sensitive to monitoring location with respect to the flowpaths. Installation of extensive and invasive monitoring well networks provides point-verification of
exchange, but ultimately lack spatial resolution and representation of extensive spatial coverage of the solute transport processes. Data from monitoring wells rarely agree with predictions based on in-stream solute transport modeling \cite{Harvey1996, Wondzell2006}.

Understanding solute transport in the subsurface ultimately requires quantification beyond classical advection and dispersion processes. Observations of tailing behavior that cannot be described by these processes alone have led to the conceptualization of mass transfer among a spectrum of domains ranging from very mobile (i.e., stream) to immobile (i.e., dead-end pore space) \cite{Goltz1986}. The subsurface domain itself, often viewed by stream solute models as a lumped domain that is completely immobile, is instead conceptualized as including a mobile domain (e.g., connected pore space modeled with advection and dispersion) and immobile domain (e.g., fluid in poorly connected pore space). Local rate-limited mass transfer among domains is conceived as controlling the tailing behavior observed for both the subsurface mobile domain and stream solute transport. Direct sampling in monitoring wells provides only an assessment of mobile water in the subsurface. Complete quantification of solute transport must, then, attempt to quantify solute presence in subsurface immobile domains as well as those domains more traditionally sampled (i.e., the total distribution of tracer in the subsurface).

To overcome the limitations inherent in stream solute transport modeling and current hyporheic zone assessment methods, we propose the use of electrical resistivity imaging (ERI) to estimate the spatial distribution of introduced ionic tracers in the hyporheic zone. ERI is a direct-current (or low-frequency alternating-current) method that can be used to estimate the spatial and temporal distribution of subsurface electrical resistivity (the reciprocal of electrical conductivity). The introduction of an electrically conductive fluid decreases the bulk electrical resistivity of the soil-water matrix. ERI has been used to successfully image the exchange of seawater with a freshwater sand aquifer in a tide-water stream \cite{Acworth2003} and groundwater
discharge into streams [Nyquist et al., 2008]. Crook et al. [2008] used cross-borehole ERI and in-stream longitudinal profiles to assess the subsurface sediment deposits in two streams, informing predictions of hyporheic exchange with estimates of subsurface architecture in streams, primarily focused on the size and distribution of alluvial deposits. ERI allows for the assessment of total concentration distribution within the mobile and immobile zones [Day-Lewis and Singha, 2008]. Singha et al. [2007] demonstrated the use of ERI to directly quantify mass transfer in heterogeneous aquifers during a push-pull tracer test in fractured porous media. More recently, Singha et al. [2008] demonstrated numerically that ERI could be coupled with an analysis of temporal moments to quantify solute exchange between the stream and less mobile domains. None of these studies, however, leverage ERI to assess surface-subsurface exchange dynamics. This is the first study, to the best of our knowledge, quantifying temporal dynamics of surface-subsurface exchange and providing a framework for quantifying hyporheic extent at the field scale.

Here, we apply direct-current geoelectrical techniques to quantify the extent of exchanging tracer from the surface into the subsurface. Because ERI is sensitive to both mobile and immobile zone concentrations, the method more completely represents surface-subsurface exchange mechanisms than point measurements in monitoring wells. Exchange of the tracer-labeled stream water with subsurface fluids reduces the electrical resistivity of the area actively communicating with the stream. Our research objectives are (1) to demonstrate the ability of ERI to quantify hyporheic extent, and (2) to compare in-stream concentration with hyporheic extent. Inverted resistivity tomograms will be used to image, for the first time in a field setting, the extent of hyporheic exchange in the subsurface. Spatially distributed assessment of solute transport in the subsurface provides insight to subsurface heterogeneity and localized processes that would otherwise be averaged over the reach using traditional characterization methods.
2.3 Methods

2.3.1 Site description

We conducted a stream tracer experiment in stream #1 of the Leading Ridge Watershed Research Unit in central Pennsylvania USA. The watershed ranges in elevation from 270 m to 440 m above mean sea level. Leading Ridge stream #1 is a cobble-bedded stream with a gradient of 5% through the study reach, draining a catchment of approximately 0.43 km\(^2\). The stream reach is 1.5-2.0 m wide and 0.1-0.2 m deep, on average.

![Diagram](image)

**Figure 2-1:** A constant rate tracer injection was performed at the Leading Ridge Experimental Watershed near State College, PA, USA. The injection of NaCl to the surface water occurred immediately downstream of a V-notch weir, used to measure discharge. Geophysical data were collected approximately 70 m downstream of the injection, using 12 electrodes set transverse to the stream at 2 m typical spacing. In-stream tracer concentration data were collected at the geophysical transect

2.3.2 Tracer injection

We conducted a constant rate injection (~1 mL/s) of dissolved NaCl (61.9 mS/cm) for 20.8 hr (starting 17:04 on 31 October 2008) immediately downstream of a weir at the head of the
study reach (0 m). Tracer was injected from a single, well-mixed reservoir prepared prior to the injection. Injection rate was constant, with volumetric flow rate of the injection verified before and after the injection period. In-stream conductivity measurements were collected upstream of the injection point and at the geophysical transect (located approximately 70 m downstream of the weir, Figure 2-1). The tracer breakthrough curve presented here is from a replicate experiment completed in mid-November, because of a failure of in-stream data loggers during the October 2008 experiment; however, conditions for the two injections were similar, with an average flow of 0.6 L/s at the upstream weir for both injections. The plateau for the in-stream breakthrough curve presented was extended by 1.3 hr to account for the difference in injection duration between the October and November trials. The in-stream breakthrough curve is presented only as representative of arrival and departure times for the tracer in the stream, as the focus of this study is the interpretation of electrical resistivity data.

2.3.3 Electrical resistivity imaging

ERI data were collected with an IRIS Syscal Pro Resistivity Meter, using a dipole-dipole geometry with 166 quadripoles collected across 12 electrodes set at 2 m spacing (Figure 2-1) to balance depth of penetration and good spatial resolution with high temporal sampling. Background data collection consisted of collecting four sets of pre-injection measurements along the transect. Data collection continued for 22.8 hr after the injection ended to monitor the substrate flushing of the conservative tracer. During collection, replicate measurements were collected and stacked (averaged) twice by the instrument for data within 3% difference of one another. If this quality threshold was not met, a third measurement was included in the average. The average standard deviation between replicate measurements was 0.085 ohm-meters (0.08%).
for the 35,856 individual measurements collected for the transect. Data were collected at approximately 1.5-hour intervals to avoid temporal smearing.

### 2.3.4 Inversion of ERI data

Inverted models of electrical resistivity are smooth representations of the subsurface due to the physics of diffusion and the data inversion process [Day-Lewis et al., 2005]. While these inversions cannot represent the small-scale heterogeneity of the subsurface given the regularization in inversion, they can be used to approximate relative changes in the area of the subsurface where the tracer is present. Because results are sensitive to a particular inversion scheme, model settings, and inherent assumptions in the model code (seeking the smoothest model that the data will accommodate, for example), the same inversion settings were used for each time step. Inverted resistivities were exported from EarthImager2D using inversion block sizes ranging from 0.5 to 1.0 m horizontal and 0.07 to 0.18 m vertical. To parse meaningful changes in bulk resistivity from error in the data collection and modeling scheme, we chose a threshold of 3.5% decrease in bulk resistivity. This value is greater than the error in the data collected and attempts to account for model and structural error, which can be difficult to quantify. Interpretation of results is sensitive to the threshold chosen (see Discussion section below, and Figure 2-3A).

ERI inversion was completed for each transect using EarthImager 2D [Advanced Geosciences, 2008]. Background (pre-injection) data were used as a baseline for comparison of data collected during the injection. Using EarthImager2D’s time-lapse inversion, differences between baseline resistance and subsequent (during injection) data sets were inverted using the background data as a-priori information (the starting model for each time step), allowing faster
convergence and sensitivity to smaller changes in bulk resistivity. Inversion results produced a root mean square error of less than 2% (mean 1.47%) for all time steps.

2.4 Results

Results of the ERI models (as percent difference from the background resistivity) are presented for selected time steps (Figure 2-2, cross-sections transverse to the stream with flow directed out of the page). The differenced models were processed to calculate an approximate hyporheic area communicating with the stream, based on bulk resistivity decreases due to tracer presence. Time-lapse imaging of solute tracer (Figure 2-2) shows solute tracer presence both below the stream (X = 10-12 m) and on the inside of the meander bend (X = 4 m). Hyporheic cross-sectional area ranges from 0.0 to ~0.5 m\(^2\) during the injection for a threshold of 3.5%. We note that the estimated hyporheic area is dependent on this choice of threshold. Because the “true” threshold value is difficult to quantify and likely spatially and temporally variable, we consider a simple sensitivity analysis, which shows while the magnitude changes, the same areal trends are present across a range of threshold values (Figure 2-3A). Bimodal response of the subsurface through time for lower threshold values (2% and 3.5%, Figure 2-3A) suggests that observed in-stream concentration histories alone cannot predict subsurface solute transport. Average horizontal distribution (Figure 2-3B) is based on the ERI inversion grid and shows a secondary emergence of the tracer at X = 4 m (Figure 2-2F), which we interpret as an abandoned streambed from a past alignment, where the cobble bed transports solute more rapidly than the surrounding substrate. We interpret the second mode (Figure 2-3A) as a result of a second “pulse” of tracer past the geophysical transect, moving through slower flowpaths. It is reasonable to hypothesize that the stream migrated laterally in the valley, but maintained its elevation.
Average vertical distribution of the tracer (Figure 2-3C) shows the solute below the streambed and saturated bank areas near the stream.

2.5 Discussion

We develop a 2-D technique to detect the presence of the salt tracer in the subsurface, simultaneously assessing both mobile and immobile domains, which has been previously impossible using in-stream, monitoring well, or modeling techniques. Our results demonstrate a peak hyporheic zone of ~0.5 m² for the 20.8 hr injection for an assumed 3.5% threshold in the change in resistivity. Based on the quality of our data and ability to match those in inversion, we believe that a threshold change of 3.5% is adequate to distinguish meaningful changes in bulk resistivity from those due to data noise and model error, though results are sensitive to the threshold chosen (Figure 2-3A). We acknowledge that calculations of the hyporheic zone area from ERI data are sensitive to the error in the data collection and the resolution of the ERI method. Despite these uncertainties, which are inherent in all geophysical inversions, these data enhance understanding of localized exchange processes. ERI data provide more complete subsurface solute transport data than could be collected with standard measurements.

For future studies, analysis of the threshold chosen should be completed, and the threshold compared to the known error in both data collection and inversion; we selected 3.5% based on these factors, and explored the sensitivity of results to the chosen threshold for interpretation. It is important to note the overall trends in hyporheic area remain constant between chosen thresholds. Application of ERI to hyporheic studies holds the potential to more accurately quantify interactions between streams and hyporheic zones. ERI methods provided a distributed assessment of solute transport in the hyporheic zone, a data set that is impossible to collect using
Figure 2-2: (A) Pre-injection electrical resistivity model. Transects shown are perpendicular to the stream, with flow directed out of the page. We propose that areas of high resistivity at $X = 2–4$ m at an elevation of -3 m local datum are representative of a cobble-bed that was abandoned during lateral migration of the stream (currently located inside a meander bed). Bedrock is observed in the model at $X = 7–18$ m and elevations below -5 m local datum. Low resistivity areas ($2 < X < 6, 4 < Z < 7$) are likely a product of low model sensitivity (Panel B) in this same region. (B) Model sensitivity as a function of physical location, from EarthImager2D. Note the areas of highest sensitivity near the surface, and decreasing sensitivity with depth. (C–G) Time-lapse ERI results, at time elapsed after beginning the conservative solute injection. Resolution of the geophysical image decreases at the edges. Grid cell color indicates the percentage change in bulk resistivity from background conditions. (C and D) Decrease in bulk resistivity in the subsurface due to tracer presence. (E) Tracer is flushed from the subsurface. (F) Second increase in flowpath centered at $X = 4$ m (see 2% threshold curve in panel A). (G) Flushing of tracer from the subsurface. (H) Interpretation of resistivity images. We hypothesize the decrease centered at $X \approx 4$ m is due to an abandoned stream channel, given its elevation equal to that of the stream and location on the inside of a meander bed.
Figure 2-3: (A) Storage zone area versus time for the transect, using a threshold of minimum 3-5% decrease in resistivity. Two to five percentage threshold range is indicated by the shading. Instream concentration (dashed grey) is normalized to maximum concentration and representative of in-stream processes during the geophysical data collection. Note the second mode of tracer concentration, corresponding to a flowpath centered at X = 4 m. (B) Contour plot of the horizontal distribution of resistivity decreases (and thus tracer presence) in the subsurface through time. X = 0 m corresponds to the left-most and X = 22 m the right-most data presented in Figure 2-2. The changes from X = 15 m to X = 22 m are attributed to noise in the data set, as these locations are substantially above the channel elevation and in a poorly resolved area of the transect. (C) Contour plot of the vertical distribution of resistivity decreases (and thus tracer presence) in the subsurface through time. Note the changes remain within 1 m of streambed elevation, at the expected vertical location hyporheic zone streambed elevation, at the expected vertical location hyporheic zone.
traditional and invasive point measurements. We expect that ERI results can be used to characterize transport behavior by illuminating the temporal dynamics of flowpaths and their deviation from mobile in-stream and subsurface observations. In-stream and subsurface tracer concentration are correlated during the injection plateau, but subsurface dynamics produce a different tailing and a secondary mode for some sensitivity thresholds at later times. Thus, while the in-stream tracer data appear to be insensitive to small fluxes of tracer coming back to the stream [after Harvey et al., 1996; Wondzell, 2006], the ERI data provide continued measurement of tracer in the hyporheic zone. Geophysical data collected during this tracer study guided installation of monitoring wells for subsequent tracer studies (Ward, unpublished data), which confirmed the presence of the tracer in monitoring wells at approximately 25, 50, and 65-75 cm below the streambed. Confirmation of tracer presence during subsequent studies suggests ERI accurately identified the tracer location in the subsurface.

Pre-injection resistivity models (Figure 2-2A) consistently identify areas of high resistivity at an elevation of about -6 m (local datum) representing the shallow bedrock below valley soils, consistent with observations of 0.5 to 2.0 m soil depth in the watershed [Lynch and Corbett, 1989]. The area of high resistivity located at X = 3 to 5 m at elevation -3 m (local datum) is at the same horizontal and vertical location as the observed preferential flowpath during the solute injection, supporting our interpretation that the preferential flowpath is an abandoned cobble streambed.

Collection and processing of data during this experiment yielded meaningful results and provides a foundation upon which the technique may be refined and improved. Because electrical current travels in three dimensions, the problem of dimensionality must be considered for a two-dimensional array. Collecting data across and between multiple transects would allow construction of a fully three-dimensional image of hyporheic zones and would eliminate
dimensionality concerns. Future research will correlate subsurface mobile domain concentration with changes in bulk resistivity, which would lend itself to additional interpretation of ERI data and model results. Traditional modeling of solute transport with transient storage assumes a constant value to represent the size of a lumped immobile domain (i.e. lumped transient storage) with a concentration that changes through time. Here, we demonstrate that not only does the spatial distribution of tracer concentration change in the hyporheic zone through time, but the physical size of the hyporheic zone is temporally variable. Furthermore, this method has the potential to inform new models of mobile-immobile solute transport [Zhang et al., 2006] that account for the spatial and temporal distributions of both mobile and immobile solutes. This spatial resolution provides a distinct advantage in predicting fate of water quality constituents over those that fit only mobile solute data.

2.6 Conclusions

We used ERI to characterize the physical extent of the hyporheic zone during a conservative tracer injection to provide a fully distributed characterization of hyporheic solute concentration in both the mobile and immobile domains in the subsurface. For the first time, direct and fully distributed assessment of the arrival of tracer into the hyporheic zone and assessment of subsurface areal trends is possible. The data collection schemes used here were chosen for efficiency; the ability to collect additional data with better spatial or temporal resolution will lead to improved quantification of the dynamics of the hyporheic zone. This work demonstrates the first field-application of ERI to actively image stream tracers to quantify hyporheic extent and exchange dynamics.
The field and modeling techniques of this study provide spatially distributed assessment of solute transport in the hyporheic zone, including the ability to assess mobile and immobile domains. Through time-lapse ERI inversion we are able to understand both spatial and temporal exchange characteristics and interpret the physical extent of the hyporheic zone. Results demonstrate spatial variability and temporal trends that have been otherwise unobserved in hyporheic studies. Our understanding of the complex biogeochemical processes in the hyporheic zone must be founded upon an understanding of hyporheic exchange processes, which we are only beginning to elucidate more accurately with this technique.

This work demonstrates the potential of ERI for studies of hyporheic exchange in small streams. We identified limitations in the interpretation of ERI for quantifying hyporheic extent and demonstrated the need for at least a basic level of sensitivity analysis. The apparent increase in solute concentration at some locations (X = 4 m in Figure 2-2F) despite the decreasing in-stream concentration at this time (Figure 3A) demonstrates the inability of in-stream concentration histories alone to predict solute presence in the subsurface due to dramatically different transport time scales.

2.7 Acknowledgments

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Chapter 3

Geophysical characterization of subsurface heterogeneity to inform models of groundwater flow and transport

This chapter will be submitted for publication with co-authors M.N. Gooseff and K. Singha. See Appendix B for co-author permissions.

3.1 Abstract

We investigated the role of increasingly well-constrained geologic structures in the subsurface (i.e., subsurface architecture) in predicted streambed flux and hyporheic residence time distribution for a headwater stream in central Pennsylvania, U.S.A. Five subsurface realizations with increasingly resolved facies boundaries were simulated in COMSOL to represent our interpretation of the subsurface. Model geometries were based on increasing information about flow and transport using soil and geologic maps, surface observations at the site, probing to depth to refusal, seismic refraction, static electrical resistivity imaging of subsurface architecture, and time-lapse electrical resistivity imaging during a solute tracer study. Particle tracking was used to generate residence time distributions for each model run. We demonstrate how increasingly complex characterization of facies boundaries and calibration of porosity and hydraulic conductivity affect model prediction of hyporheic flow and transport. Temporal moment analysis of time-lapse ER images allowed for order-of-magnitude calibration, with ER observations matching particle arrival times at the ER transects. Calibrated models yield estimates of streambed flux that are three orders of magnitude larger than models using common values for hydraulic conductivity ($10^{-4}$ for calibrated vs. $10^{-7}$ for common published values). Residence time distributions for calibrated models are centered near $10^2$ hr, while published
hydrogeologic values yielded residence time distributions on the order of $10^4$ hr. Furthermore, increasingly complex subsurface architectures yield wider distributions of hyporheic residence time. The use of time-lapse resistivity imaging informs subsurface structure that was not apparent from other techniques, and illustrates the important difference between geophysical properties (i.e., resistivity) and transport properties in the subsurface (i.e., hydraulic conductivity). Results of this study demonstrate the value of geophysical measurements of both structure and transport in groundwater flow modeling.

### 3.2 Introduction

The ecological relevance of exchange flows between streams and their hyporheic zones (i.e., near-stream aquifers) has been widely recognized in the past 20 years [see reviews by Boulton et al., 2010; Brunke and Gonser, 1997; Krause et al., 2010] yet our ability to model hyporheic exchange with high spatial resolution lags behind our understanding of the associated biogeochemical processes occurring along these flowpaths. Wondzell et al. [2009] summarize the growing interest in numerical groundwater flow modeling as a tool to improve our understanding of solute transport in coupled stream-hyporheic systems in four areas: (1) quantifying the location and magnitude of fluxes across the streambed; (2) quantifying residence time distributions of water in hyporheic flowpaths; (3) providing a spatially explicit depiction of flowpaths; and (4) examining the factors that control hyporheic exchange. Indeed, numerical models of hyporheic flow and transport have been widely used in studies of hyporheic exchange [e.g., Gooseff et al., 2006; Kasahara and Wondzell, 2003; Lautz and Siegel, 2006; Saenger et al., 2005; Wondzell et al., 2009].
Despite the promise and utility of numerical modeling to inform distributions of flowpath lengths and reach residence times, accurate representation of such models may be difficult, if not impossible, due to heterogeneity in the subsurface that is both difficult to characterize and serves as a primary control on stream-aquifer interactions [e.g., Packman and Salehin, 2003; Ryan and Boufadel, 2006]. The presence of a complex network of hyporheic flowpaths has been attributed to both small-scale [Sawyer and Cardenas, 2009] and large-scale [Vaux, 1968] heterogeneities in the subsurface, observed in field, flume, and numerical studies. Subsurface heterogeneity in hydrogeologic parameters can limit or enhance hyporheic exchange in the subsurface. Increasing heterogeneity has been linked with increased exchange flux [Cardenas et al., 2004; Salehin et al., 2004]. At a larger scale, the subsurface architecture (i.e., heterogeneities occurring at a macroscopic scale, such as different geologic units, sediment types, etc. as opposed to variability occurring within a given unit) such as valley constraint of lateral and vertical flowpaths or bedrock confining units may limit the extent of hyporheic flowpaths.

The challenge of characterizing heterogeneity in the subsurface is central to the equifinality thesis as applied to hyporheic exchange modeling (i.e., the thesis that multiple realizations of the subsurface may produce equally good fits based on data observed at discrete locations) [e.g., Beven, 2006]. Indeed, a long-standing debate exists in the literature regarding whether such models can ever be verified or validated given the documented equifinality [e.g., Bredehoeft and Konikow, 1993; Hassan, 2004; Konikow and Bredehoeft, 1992; Oreskes et al., 1994]. Most recently, Wondzell et al. [2009] extended these issues to include models of stream-aquifer interactions, finding equally acceptable performance from homogeneous subsurface models with idealized boundaries and models including heterogeneity and more complex boundary topography. Their study finds simplified models acceptable to predict fluxes and
approximate travel times, but they conclude that such simple models are not sufficient to predict
the movement of solutes through the hyporheic zone.

One way to characterize flowpath dynamics and subsurface heterogeneity in the
subsurface is to install monitoring well networks. Shallow monitoring wells have been widely
installed to monitor for stream tracer movement in the subsurface [e.g., Wondzell, 2006] and to
classify hydraulic conductivity of the subsurface [e.g., Baxter and Hauer, 2000; Nowinski et
al., 2011]. Sampling pore water from these wells during tracer studies is problematic because the
withdrawal of water from the well field may modify the flow field itself and provides only a
spatially discrete measurement. The extent of well installations (and therefore observations) is
limited by the ability to install hand-driven wells where equipment access is limited [Baxter et al.,
2003; Burkholder et al., 2008]. Direct investigation of the subsurface by digging of soil pits and
trenches, or manual probing of the subsurface to identify bedrock interfaces, is also common,
 albeit invasive [e.g., Jones et al., 2008; Nyquist et al., 2008; Nyquist et al., 2010; Wondzell et al.,
2009].

In contrast to direct probing or sampling of the subsurface, hydrogeophysical methods
can be used to characterize subsurface structure, hydrological parameters, and monitor transport
processes [Binley et al., 2010]. Two types of subsurface characterizations that are possible using
hydrogeophysical tools: process areas, or geologic properties that do not change over time, and
process dynamics, or characteristics that change through time and are inherently linked to the
movement of fluids (primarily water) in the subsurface [Binley et al., 2010; Koch et al., 2009].
Indeed, geophysical investigation to characterize process areas for study of hyporheic exchange is
growing in popularity, with demonstrated application of electrical resistivity [Crook et al., 2008;
Ward et al., 2010b; Chapter 2], ground penetrating radar [Brosten et al., 2006; Jones et al., 2008;
Naegeli et al., 1996], and seismic refraction and reflection [Anderson et al., 2005; Jones et al.,
These techniques can be used to characterize subsurface architecture based on changes in the physical properties of the subsurface and commonly produce continuous spatial distributions of such properties. Geologic structure is inferred from the spatial distribution of geophysical properties (e.g., electrical resistivity, compressional wave velocity). Linking geophysical properties with hydrologic properties (e.g., correlation of electrical resistivity and hydraulic conductivity) requires development of site-specific relationships, which has had limited success due to high levels of uncertainty [e.g., Lesmes and Friedman, 2005; Pride, 2005; Singha and Moysey, 2006].

The application of geophysical methods to study hyporheic process dynamics in the field has been limited to-date, in contrast to the relatively common study of process areas. Nyquist et al. [2008] used electrical resistivity to characterize locations of groundwater discharge to a stream. Binley et al. [2010] suggest that perturbing a system using a controlled experiment over relatively short time scales (hours to weeks) and rapid collection of geophysical data might be the most informative application of geophysics to characterize subsurface hydrologic properties. Indeed, saline tracers have been used to characterize hyporheic exchange in both field [Nyquist et al., 2008; Ward et al., 2010b; Chapter 2] and numerical studies [Singha et al., 2008; Ward et al., 2010a; Chapter 5]. One example of geophysical monitoring of solute transport is work by Slater et al. [1997], who identified preferential flow through fractures using saline tracers and electrical geophysical monitoring. Electrical resistivity imaging of saline tracers has been used to inform groundwater flow properties [e.g., Bevc and Morrison, 1991; White, 1988] including preferential flowpaths [e.g., Schima et al., 1996; Slater et al., 1997]. Electrical resistivity images have been used to parameterize numerical models of flow and transport in the subsurface [Binley et al., 1996b; Kemna et al., 2002].
The objective of this study is to determine how increasingly complex realizations of subsurface architecture affect predictions of hyporheic residence time distributions and total flux across the streambed. In this study, we use observations of both static features and dynamic solute transport to characterize subsurface heterogeneity in hydraulic conductivity and to calibrate several realizations of groundwater flow models of hyporheic exchange. Specifically, we began with a model parameterized using only site observations, soils maps, and geologic maps of the site. Next, we used depth-to-refusal measurements to parameterize topography between weathered shale and overlying sediment. We used seismic refraction to parameterize the boundary between underlying weathered and unweathered bedrock. Next, electrical resistivity imaging was used to more completely constrain subsurface architecture. Finally, we used coupled solute transport and time-lapse electrical resistivity imaging to calibrate a model’s subsurface architecture (i.e., heterogeneities between geologic units, as opposed to small-scale heterogeneity with a given deposit) and solute transport. Through both qualitative (i.e., pseudo 3-D visualization) and quantitative (i.e., temporal moment analysis) interpretation of a coupled solute tracer and electrical resistivity experiment, a model is calibrated to match observations of subsurface solute transport collected using electrical resistivity imaging. Direct investigation of the subsurface and seismic refraction are integrated with this data set to demonstrate the differing information content of multiple methods. Results of this study inform the role of subsurface architecture in hyporheic exchange and illustrate limitations of poorly constrained models in predicting the extent of hyporheic flowpath networks and their residence time distributions.
3.3 Methods

3.3.1 Site description

Watershed #1 at the Leading Ridge Watershed Research Unit is of a small, forested watershed draining 106 ac. Typical valley slopes are 4-5%, with a southeastern aspect. The site is covered with even-age forests including oak, hickory, and maple trees, uncut since commercial logging in the 1840s and 1900s [Kostelnik et al., 1989]. Watershed #1 is a control watershed, with limited harvesting since the research unit was established in the 1960s [Rishel et al., 1982].

Topography at the research site is relatively flat. The stream draining Watershed #1 is 1.5 to 2 m wide and 0.1 to 0.2 m deep through the study reach, with a gradient of approximately 2% along the centerline in the surveyed reach. The stream and watershed react quickly to storm events, with response lagging precipitation by only about 30 minutes [Lynch and Corbett, 1989]. Runoff response in the watershed is primarily as subsurface flow to the stream, as opposed to overland flow [Lynch and Corbett, 1989]. The research site and instrumentation are shown in Figure 3-1.

Soils at the surface are generally high in organic matter, typical of a hardwood forest floor. Soils in the research site are primarily residual, having developed as a result of in-situ weathering of the underlying strata. The research site is underlain by three primary soil types: Brinkerton silt loam, Edom-Weikert complex, and Newark silt-loam. All three soil series are primarily silt and stony loams that are well drained. Soils are 8 to 200+ cm deep and range from
21.0 to 42.8% clay in content [USDA-NRCS, 2008]. A shallow seismic study by Shields and Sopper [1969] on the upper slopes of the watershed identified weathered shale layers averaging 3.2 m thick, lying below 1.4 m of soil.

The site is located in the Appalachian Mountain Section of Pennsylvania’s Ridge and Valley Province [Sevan, 2000]. Most of the lower slope is underlain by the Rose Hill shale formation (approx. 700-ft thick). The upper portion of the lower- and middle-slopes are underlain by the Castanea sandstone (approx. 500-ft thick), and the ridge top is underlain by the Tuscarora quartzite [Lynch et al., 1986]. Within the Rose-Hill shale formation, layers of sandstone and

Figure 3-1: Geophysical studies and a constant-rate solute injection (20.8 hour drip of NaCl into the stream) were performed at the Leading Ridge Experimental Watershed near State College, PA, USA. Depth to refusal was recorded by driving a 2.54 cm diameter steel rod into the streambed at five locations, and a bedrock outcrop was observed in the stream at the downstream end of the study reach. Electrical resistivity data were collected along six two-dimensional transects of 12 electrodes each (2 m electrode spacing) perpendicular to the stream, with transects spaced approximately 4 m along the stream centerline. Data from one transect were omitted from the study because they were too noisy to produce ER images of the subsurface. Seismic refraction data were collected along a single transect parallel to the stream. Flow at the V-notch weir was 0.6 L s$^{-1}$ during the study.
siltstone were observed at an outcrop along the stream centerline at the downstream end of the study reach.

### 3.3.2 Subsurface characterization

*Direct investigation of the subsurface*

Direct observation of the depth of alluvium and location of subsurface features was completed several months after the tracer study was completed. A 2.54-cm diameter steel rod was driven into the bed at the stream centerline until refusal. This refusal was interpreted as the interface between the weathered shale layer and overlying streambed sediment matrix. We assume an unweathered bedrock layer exists at some depth below the weathered shale, since an outcrop of this bedrock is observed at the downstream end of the study reach. Observation locations and depths are shown in Figure 3-2. Stream topography and locations of measurements were surveyed using a sight level and tape.

*Seismic refraction survey*

Shallow seismic refraction data were collected along the transect labeled in Figure 3-1. Data were recorded on a seismograph manufactured by Geometrics, Inc. (San Jose, CA, U.S.A.), using 24 geophones at 1 m spacing. The seismic transect was located approximately perpendicular to the ER transects and parallel to the stream (offset from the stream by approximately 10 m). Shot offset was 1 m from the geophone array for both the forward and reverse shots, respectively located at the southeast and northwest ends of the transect. Data were collected across a
relatively flat topography, with total elevation changes of 0.57 m along the transect. No terrain correction (time delay) was assigned based on topography.

Visual interpretation of seismograms was used to identify first arrival times at each geophone. All first arrival times were identified by a single interpreter for consistency, and times for each arrival were calculated from graphical output of the seismograph. Assuming that all first arrivals identified were prone to a random (rather than systematic) error, the procedure of using best-fit lines for the data minimizes the error associated with an individual observation.

The data were interpreted as a three-layer system (soil, weathered shale, unweathered bedrock) with plane-bed interfaces, based on available soil and geologic data. Average slopes for

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**Figure 3-2:** Longitudinal profile of the study reach geomorphology based on topographic survey. The longitudinal location of electrical resistivity transects (dashed lines) and depth to refusal measurements (black dots) are shown. The water surface shown was calculated using the surface hydraulics model HEC-RAS [Brunner, 2006] with observed topography and flow rate (at the upstream weir). A bedrock outcrop was observed beginning at X = 61 m and continuing downstream beyond the extent of the study reach.
the forward and reverse shot were used to calculate velocities for each layer. The slope-intercept method was used to estimate thicknesses for the top two layers. Intercept times were taken for the forward shot to provide an initial estimate of layer thicknesses. The program REFRACT [Burger et al., 2006] was used to develop a subsurface model that best fit to the observed seismic data (as minimum RMSE between observed and simulated data). The initial model was based on thicknesses estimated from the forward shot, and layer velocities estimated from the slope and intercept of arrival data. The subsurface architecture of the layers (i.e., layer thicknesses, dip angles) was determined by manually adjusting REFRACT model parameters to achieve a best fit.

Solute tracer study

A constant-rate injection (~1 mL s⁻¹) of dissolved sodium chloride (NaCl, 61.9 mS cm⁻¹) into the stream channel immediately downstream of the V-notch weir was completed for 20.8 hours in October 2008. In-stream flow during the study was 0.6 L s⁻¹, determined at a V-notch weir located immediately upstream of the study reach. The observed in-stream breakthrough curve at T1 is presented in Ward et al. [2010b; Chapter 2].

Electrical resistivity imaging

ER data were collected in a series of six transects oriented perpendicular to the stream channel at distances of 70 to 92 m downstream of the injection location (hereafter T1 through T5, Figure 3-1; located at X-coordinates 21 through 43 m, Figure 3-2), using an Syscal Pro Switch (IRIS Instruments, France). ER data for each transect were collected using 12 electrodes at a 2 m spacing approximately centered on the stream. ER data were collected only along transects (e.g.,
measurements were collected using data from multiple transects). 166 unique quadripoles (measurements consisting of a potential pair and current pair of electrodes) were collected at each transect before, during, and after the solute tracer injection using a dipole-dipole scheme. Two replicate measurements were collected for each quadripole and stacked (i.e., averaged); a third measurement was collected and included in the stacking if there was a greater than 3% difference between the first two measurements. Errors were not quantified using reciprocal measurements to maximize temporal resolution within the data set. Data were collected at each transect approximately every 1.5 hours, and stationarity was assumed during the collection time for each transect (i.e., it was assumed solute tracer distribution did not change during the 15 minute collection time for each transect).

ER data were inverted using EarthImager2D [Advanced Geosciences, 2008], which seeks the smoothest possible subsurface model to explain the subsurface architecture. Horizontal and vertical block sizes ranged from 0.5 to 1.0 m and 0.07 to 0.18 m, respectively. A uniform starting model was used to invert background (i.e., pre-tracer) data. Time-lapse inversion was used to develop subsurface models for ER data collected during the injection, with the previous timestep used as the starting model for subsequent inversions. The background ER model was used as the starting model for the first timestep. The pre-tracer model was subtracted from the time-lapse models, and the percent change from background was calculated for each pixel. The presence of the electrically conductive tracer is recorded when a drop in electrical resistivity (i.e., negative percent change) was predicted in the inverted images. It is important to note that ER imaging is non-exact, and that inversion results (i.e., subsurface models) are non-unique. Results are sensitive to the inversion scheme, regularization of the model, and parameterization of the inversion process [e.g., Day-Lewis et al., 2005; Friedel, 2003] and result in over-smoothed representations of the subsurface due to diffusive physics governing electrical fields and the
inversion scheme [Day-Lewis et al., 2005]. Electrical resistivity data, inversion models, and interpretation for the upstream-most transect are presented in Ward et al. [2010b; Chapter 2].

The resulting two-dimensional inversion models were located in three-dimensional space based on field-surveys locations, and linear interpolation between these 2-D “slices” was used to construct pseudo 3-D images of the hyporheic zone. We chose not to invert the data in 3-D because of the limited data collection scheme (collection only along, and not between, transects), and to minimize the parameters necessary for data inversion. We acknowledge that linear interpolation over distances of 4 to 8 m between transects may not produce data that are representative of heterogeneities at all scales, but we believe there is value in identifying trends between transects where ER data and models are available. This interpolation provides a qualitative tool that helps identify the subsurface architecture at a macro scale (i.e., features on the order of meters). The surface presented is based on a 3.5% change from background, which was selected to parse changes due to solute presence from those due to both noise in the data collection and error in the inversion process [after Ward et al., 2010b; Chapter 2]. This threshold was selected for this data set and site and should be evaluated if this technique is applied at other locations.

*Temporal moment analysis*

Temporal trends in solute transport through the matrix can be characterized by the analysis of individual pixels (i.e., elements) in the inverted images [Binley et al., 1996a; Slater et al., 2000; Slater et al., 2002]. Here, we calculate the $k^{th}$ normalized temporal moment for the change in resistivity in each pixel using the equation
\[ n_k = \int_0^{+\infty} r(t) \cdot t^k dt \]  

(1)

where \( t \) is time, and \( r(t) \) is the normalized change in pixel resistivity, which is calculated as

\[ r(t) = \frac{R(t)}{\int_{-\infty}^{+\infty} R(t) dt} \]  

(2)

where \( R(t) \) is the resistivity change of the pixel. The normalized first moment \( (n_1/n_0) \) is physically interpreted as the mean arrival time for the tracer at the pixel location. For this study, only normalized first temporal moments were calculated from the time-lapse images. Mean arrival time based on ER data was compared to the arrival time of particles at the ER transect to calibrate hydraulic conductivity for each model unit.

### 3.3.3 Numerical modeling

The finite-element model COMSOL [COMSOL, 2008] was applied in a 2-D space along the stream thalweg to solve the steady-state groundwater flow equation. The bottom boundary of the model was treated as a no-flow boundary, fixed at the average slope of the stream topography within the reach. An average subsurface depth of 8 m was simulated. Stream flow over the surveyed streambed topography was simulated using the U.S. Army Corps of Engineers’ 1-D surface flow model HEC-RAS [Brunner, 2006] to calculate a steady-state water surface through the study reach. HEC-RAS flow depths were used as a constant-head boundary along the upper extent of the groundwater flow model. No iteration between up- and downwelling predicted by
COMSOL and HEC-RAS flow rates was completed, as past studies have found this step unnecessary for hydraulic conductivities less than $10^{-2}$ m s$^{-1}$ [Hester and Doyle, 2008]. To condition the inflow to the study reach in the subsurface, we simulated an additional 21 m of stream length (parameterized by observed topography) upstream of the study reach in the HEC-RAS and COMSOL models. Finally, in the COMSOL model we extended the upstream and downstream slopes and constant heads an additional 40 m in both directions to isolate the upstream and downstream boundaries of the COMSOL model from the study reach. Subsurface architecture was extended upstream and downstream in layers parallel to the streambed for the domains outside of the highly studied stream reach.

Flow and transport parameters for the bedrock, less weathered shale, weathered shale, and overlying streambed sediment (from deepest to shallowest) were assumed homogeneous and isotropic; hydrogeologic parameters for each layer are summarized in Table 3-1. Five realizations of subsurface architecture, as informed by a variety of tools as discussed in preceding sections, were simulated. Several increasingly complex subsurface realizations were simulated to represent our interpretation of the subsurface based on increasing information about flow and transport, summarized in Table 3-2. Here, we define complexity as the increasingly well-resolved facies boundaries.

Simulation of flow and transport for each geometry were first run using uncalibrated values of porosity and hydraulic conductivity. Particle arrival times at T4 and T5 were compared to mean arrival times observed in ER data along the stream centerline at these locations. The hydrogeologic properties of the streambed sediment and weathered shale layers were adjusted to calibrate the model. Calibration was determined by producing the best agreement between observed mean arrival times (i.e., first temporal moments of individual pixels in time-lapse ER
images) and simulated particle arrival times at each transect. Assumptions of homogeneity and isotropy within each unit were maintained for the calibration.

Total flux across the streambed was computed using trapezoidal integration of flux magnitudes within the study reach (i.e., $21 < X < 61$ m, Figure 3-2). The integration combines both up- and downwelling fluxes into a single parameter representing gross exchanges of water across the streambed. To characterize the residence times of hyporheic flowpaths we used particle-tracking tools in COMSOL. Particle tracking has been used to quantify upwelling and downwelling locations [e.g., Gooseff et al., 2006] and hyporheic residence time distributions [Wroblicky et al., 1998; Ward et al., In Press]. This technique is only able to quantify timescales of advective transport in the subsurface [Lautz and Siegel, 2006], with effective porosities used to convert Darcian to advective velocities for particle tracking. Particles were released at 2.5 cm spacing along the streambed, and those traveling greater than 5 cm from their initial position were tracked until they exited the subsurface domain. Particle releases were not flux weighted, and particle-tracking timesteps were non-uniform and automatically tuned by COMSOL. Flux calculations and particle tracking methods replicate those used by Ward et al. [In Press] in their study of subsurface structures.
3.4 Results

3.4.1 Subsurface architecture models

Geometry A: Soil and Geology Maps, Surface Observations

Published data (i.e., soil and geology maps from the site) indicate that the site includes shale bedrock, overlain by weathered bedrock, and finally a layer of alluvial sediment. The subsurface architecture derived from this information is presented as Figure 3-3A, which includes a 1.4 m sediment layer and 3.2 m weathered shale layer, underlain by bedrock. The observed bedrock outcrop beginning at X = 61 m and continuing downstream was simulated as a vertical bedrock outcrop that terminated the longitudinal extent of both the soil and bedrock layers.

Geometry B: Depth to refusal

At five locations in the study reach, a steel rod was driven into the bed to refusal, labeled A through E in Figures 3-1 and 3-2. Depth to refusal was interpreted as the depth to the interface between streambed sediment and weathered shale. The interface was linearly interpolated between observations and was assumed to end at the bedrock outcrop. The subsurface architecture model including these observations is presented in Figure 3-3B.

Geometry C: Seismic refraction survey

Data for the seismic transect were visually interpreted as a three-layer system (sediment, weathered shale, and bedrock), with an interface between the bottom two layers dipping from the
southeast toward the northwest. The model for subsurface structure was fit by manually adjusting dip angle and layer thicknesses to minimize the root-mean-square (RMS) error between observed and model data for seismic shot arrival at each geophone. For the final model (Figure 3-4B), average RMS error was calculated by REFRACT [Burger et al., 2006] as 0.96 ms. The dip angle for the best-fit model is 4 degrees. For a surface velocity of 190.5 m/s the surface topographic relief along the transect (0.57 m) could add up to 2.99 ms to a measurement. Both model error and topographic deviation were considered negligible for the initial interpretation of the data. The seismic refraction model was used to modify the subsurface architecture as shown in Figure 3-3C. The bedrock depths at T1 and T5, determined from the seismic study, were used to simulate a planar interface between bedrock and weathered shale layers. The bedrock interface was linearly interpolated downstream to the bedrock outcrop observed at the streambed.

Table 3-1: Hydrogeologic properties of simulated geologic units. Values are within the ranges for the local soils and shale presented by Kuntz [2010] for a nearby site and those for weathered shale values in Mayes et al. [2000]. Calibrated values were based on matching observed mean arrival times (based on first temporal moments of time-lapse ER images) with simulated particle arrival times. Hydrogeologic values for unweathered shale and bedrock were not changed from values reported by Kuntz [2010] and Mayes et al. [2000].

<table>
<thead>
<tr>
<th>Geologic Unit</th>
<th>Uncalibrated Values</th>
<th>Calibrated Values</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Porosity (%)</td>
<td>Hydraulic Conductivity (m s⁻¹)</td>
</tr>
<tr>
<td>Streambed Sediment</td>
<td>14</td>
<td>1.0×10⁻⁷</td>
</tr>
<tr>
<td>Weathered Shale</td>
<td>21.15</td>
<td>8.3×10⁻⁸</td>
</tr>
<tr>
<td>Unweathered Shale</td>
<td>12.33</td>
<td>5.0×10⁻¹¹</td>
</tr>
<tr>
<td>Bedrock</td>
<td>0.35</td>
<td>2.7×10⁻¹⁵</td>
</tr>
</tbody>
</table>
Table 3-2: Summary of total flux across streambed for alternative models of subsurface architecture.

<table>
<thead>
<tr>
<th>Model Run No.</th>
<th>Geometry</th>
<th>Subsurface Characterization</th>
<th>Hydrogeologic Parameters</th>
<th>Total Streambed Flux (^1) (m(^2) s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>A</td>
<td>Soil + Geologic Maps, Surface Observations</td>
<td>Uncalibrated(^2)</td>
<td>1.01(\times)10(^{-6})</td>
</tr>
<tr>
<td>2</td>
<td>B</td>
<td>A + Depth to Refusal</td>
<td>Uncalibrated(^2)</td>
<td>7.44(\times)10(^{-7})</td>
</tr>
<tr>
<td>3</td>
<td>C</td>
<td>B + Seismic Refraction</td>
<td>Uncalibrated(^2)</td>
<td>7.33(\times)10(^{-7})</td>
</tr>
<tr>
<td>4</td>
<td>D</td>
<td>C + Background ER</td>
<td>Uncalibrated(^2)</td>
<td>7.04(\times)10(^{-7})</td>
</tr>
<tr>
<td>5</td>
<td>E</td>
<td>D + Solute Tracer ER</td>
<td>Uncalibrated(^2)</td>
<td>6.62(\times)10(^{-7})</td>
</tr>
<tr>
<td>6</td>
<td>A</td>
<td>Soil + Geologic Maps, Surface Observations</td>
<td>Calibrated(^3)</td>
<td>4.88(\times)10(^{-4})</td>
</tr>
<tr>
<td>7</td>
<td>B</td>
<td>A + Depth to Refusal</td>
<td>Calibrated(^3)</td>
<td>4.49(\times)10(^{-4})</td>
</tr>
<tr>
<td>8</td>
<td>C</td>
<td>B + Seismic Refraction</td>
<td>Calibrated(^3)</td>
<td>3.70(\times)10(^{-4})</td>
</tr>
<tr>
<td>9</td>
<td>D</td>
<td>C + Background ER</td>
<td>Calibrated(^3)</td>
<td>3.73(\times)10(^{-4})</td>
</tr>
<tr>
<td>10</td>
<td>E</td>
<td>D + Solute Tracer ER</td>
<td>Calibrated(^3)</td>
<td>3.53(\times)10(^{-4})</td>
</tr>
</tbody>
</table>

1 – Flux is reported as m s\(^{-1}\) per m of streambed width.
2 – Uncalibrated hydrogeologic parameters are based on published values for nearby sites (see Table 3-1)
3 – Geometry E was used to calibrate hydrogeologic parameters for the soil and weathered shale
**Figure 3-3**: Five conceptual geometries for subsurface architecture, with facies boundaries cumulatively informed by (A) soils and geologic maps, and site observations, (B) depth to refusal measurements, (C) seismic refraction to identify topography of the bedrock interface, (D) electrical resistivity imaging of background conditions, and (E) electrical resistivity imaging during an electrically conductive solute tracer study. Up to four geologic units (sediment, weathered shale, less weathered shale, and shale bedrock) were simulated (each homogeneous and isotropic; hydrogeologic parameters summarized in Tables 3-1 and 3-2).
Geometry D: Background electrical resistivity data

The average standard deviation between stacked measurements of electrical resistivity was 0.085 Ω·m (0.08 %) for the 35,856 measurements collected during the study. Average RMS error in the inverted images relative to the noise for T1, T2, T3, T4, and T5 combined was 1.05%; RMS error ranged from 0.18% to 2.25%. Data for one transect (dashed line in Figure 3-1) did not invert to levels of RMS that were comparable to the remaining transects, possibly owing to poor contact between electrodes and sediment generating noisy data, and has been omitted from this study. Pre-tracer subsurface models based on the ER data generally show a low resistivity layer underlain by a higher resistivity layer (Figure 3-5A-E), interpreted as weathered shale above bedrock based on ER images [Ward et al., 2010b; Chapter 2]. The weathered shale layer is approximately 3 m below the streambed at the upstream end of the study reach (Figure 3-5A, 3-5B), nearly gone in Figure 3-4C, and is not present in Figure 3-4D and 3-4E. The high resistivity location in the shallow subsurface (X = 0 to 6 m, Figure 3-5A) was interpreted as an abandoned streambed by Ward et al. [2010b; Chapter 2], an explanation for which there is additional evidence in the continuity of that feature observed in subsequent transects (Figures 3-5B, 3-5C). Finally, the low resistivity layer interpreted as soil and weathered bedrock (Figures 3-5A, 3-5B) at depths of 0 to 3 m below the streambed appears to pinch-out in Figure 3-5C, and is replaced by a higher resistivity deposit in Figures 3-5D and 3-5E. This is interpreted as a less weathered shale layer and used to define a fourth material in the subsurface (Figure 3-3D).
Geometry E: Electrical resistivity monitoring of solute tracer

Subsurface models based on ER data collected during the peak of the tracer study (collected immediately before the solute injection was ended) show the presence of the tracer as a drop in resistivity (Figures 3-4F through 3-4J). Solute is generally present at the upstream and downstream ends of the study reach at depths of 1 to 1.5 m, with apparent upwelling upstream of...
T3 (Figure 3-4H) and downwelling downstream of that location. The geometry of the less weathered bedrock layer was adjusted such that flowpaths identified by particle tracking match those observed with ER imaging of the solute. Notably, the layer was extended upstream to force shallower flowpaths observed at T1, and thickness was increased to create the up- and downwelling observed at T3 through T5 (Figure 3-3E).

3.4.2 Groundwater flow model calibration using time-lapse electrical resistivity monitoring

Time-lapse ER models during the tracer injection show that solute tracer first appears in the subsurface at T1 and T4 (Figure 3-6). Time-lapse images show the growth of a hyporheic zone of relatively constant width (about 1.5 m) forming throughout the study reach, reaching a relatively constant plateau, and slowly flushing solute tracer. Pixel breakthrough curves were analyzed to determine mean arrival time (i.e., normalized first temporal moment) along the stream centerline at several depths in the subsurface.

Temporal moment analysis was completed for each pixel in the T4 and T5 inverted data sets, with mean arrival times in the top 1.5 m of the subsurface occurring between 24 and 29 hours (Figure 3-7). We omitted T1 through T3 from this analysis because particle arrival times at these locations are dependent upon subsurface characterization upstream of T1, where no geophysical characterization was completed. Mean arrival times (i.e., normalized first temporal moments) based on ER images of solute tracer range from 24 to 29 hours in the shallow subsurface at T4 and T5 based on ER imaging (Figure 3-7). Simulated particle arrival times within 10 cm of this location in the X-direction are orders of magnitude larger for simulations using published values of hydraulic conductivity and porosity (Figure 3-7). We calibrated the
Figure 3-5: (A-E) Background (i.e., pre-tracer) electrical resistivity imaging of subsurface structure. High resistivity regions are interpreted as bedrock. An apparent structural change exists between transects 2 and 5, interpreted as less weathered shale. (F-J) Resistivity change from background (as percent) at the peak of the solute injection along each transect. Decreases in electrical resistivity are interpreted as the presence of the tracer in the subsurface. The bedrock layer appears closer to the surface in transects 4 and 5 based on electrical properties (D-E). These changes in resistivity do not correspond with the location of the solute tracer (I-J), illustrating the difference between electrical properties of the subsurface and hydrogeologic properties of the subsurface. The geophysical observation of process dynamics is an important tool to inform numerical models of groundwater flow and transport, although changes in physical properties may not correspond to changes in hydrologic properties.
Figure 3-6: Time-lapse visualization of solute transport in the hyporheic zone, as linear interpolation between the 2D inversions. In all figures, X is the cross-stream coordinate, Y the coordinate along the stream centerline, and Z is the vertical datum. Flow is from left to right in all images. Isometric (top row), plan (middle row), and profile (bottom row) views of the subsurface imaging are shown (on different scales). The approximate bank locations (middle row) and the stream profile (bottom row) are shown for reference. Solute enters the subsurface from the upstream end of the study reach and by downwelling near Y = -9 m. Imaging shows subsurface architecture causes upwelling near the center of the study reach. Hyporheic width is approximately constant and equal to the width of the stream through the study reach.

hydraulic conductivity and porosity for the soil and weathered shale layers using geometry E (the most highly constrained subsurface model) until particle arrival times were within an order of magnitude of those observed using time-lapse ER imaging in the shallowest 1.5 m of the subsurface. Calibration was achieved by manually adjusting hydraulic conductivity and porosity
of the sediment and weathered shale layers. Parameters were adjusted to achieve the best fit of mean arrival time (based on first temporal moments of ER data) and simulated particle arrival time depth profiles at T4 and T5. Best fit was achieved for both the soil and weathered shale layers set at hydraulic conductivity and effective porosity of $8 \times 10^{-5}$ m s$^{-1}$ and 25%, respectively. In summary, model runs 1-5 are each of the five geometries parameterized using uncalibrated hydrogeologic values; model runs 6-10 are the same five geometries parameterized with calibrated hydrogeologic values (model runs are summarized in Table 3-2).

### 3.4.3 Simulated hyporheic flux and residence times

The gross flux of water across the streambed for alternative subsurface models is summarized in Table 3-2. For both model sets 1-5 (uncalibrated hydrogeologic parameters) and 6-10 (calibrated hydrogeologic parameters), flux is largest for geometry A, and decreases as the subsurface architecture is more completely characterized. Flux is relatively constant for the first five model runs (those using the uncalibrated values for hydraulic conductivity and porosity), with values on the order of $10^{-7}$ m$^2$ s$^{-1}$ (or m s$^{-1}$ per m of streambed width). Model runs 6-10 (those using the calibrated values for hydraulic conductivity and porosity) have flux values on the order of $10^{-4}$ m$^2$ s$^{-1}$. As more complex subsurface architecture is simulated, the predicted streambed flux generally decreases. This is attributed to the decreasing thickness of the soil and weathered shale layers as subsurface architecture is constrained by the bedrock and less-weathered shale layers.

Particle tracklines generally show the “beads on a string” pattern of hyporheic exchange that is expected from pool and riffle morphology (Figure 3-3) [Stanford and Ward, 1993]. Flowpaths simulated for geometry A are generally the deepest. Added topography along the bedrock interface (geometry C) and the location and extent of the less weathered shale deposit
Figure 3-7: Particle arrival times within 10 cm of electrical resistivity transects 4 (left column) and 5 (right column) vs. ER-derived mean arrival times for both uncalibrated (top row) and calibrated hydrogeologic parameters (bottom row). Values for a perfectly calibrated model should plot on a 1:1 line (i.e., a perfect fit between model and observed data). Using published hydrologic values the particle arrival time is 1-4 orders of magnitude larger than observed arrival using ER (top row). Calibrated models produce results that match mean arrival times determined from time-lapse ER imaging within an order of magnitude; further refinement of heterogeneity within each layer (as opposed to only adjusting homogeneous, isotropic parameters for the geologic units included) could further improve model performance. Some disparity between ER and particle tracking data is expected, since ER relies on spatially large electrical field measurements, while particle tracking is finite in space. Still, without ER imaging of solute transport to calibrate the model, residence time estimates would be grossly large. In both calibrated and uncalibrated models, increasingly complex subsurface models produced later particle arrival times.
(geometry D, E) generally produce interactions between the individual “beads” generated by surface topography. A complex network of both short and extensive hyporheic flowpaths is ultimately generated using geometry E, which represents the best-constrained facies boundaries (Figure 3-3). Increasingly complex subsurface architecture generally resulted in longer particle residence times in the subsurface at T4 and T5 (Figure 3-7). Particles arrived at a given location in the subsurface more rapidly for the highly idealized geometry A, while the inclusion of additional complexity resulted in later arrival at a given depth.

Model runs 1-5 have residence time distributions centered near $10^4$-$10^5$ hr, whereas model runs 6-10 have residence time distributions centered near $10^2$ hr (Figure 3-8). For both sets of hydraulic conductivity and porosity modeled, increasing subsurface complexity increases the variance in particle residence times. Additionally, the more complex subsurface architectures generate long residence time flowpaths (greater than $10^8$ hr in models 1-5, and $10^5$ hr in models 6-10), due to the transport of a small number of particles through the low hydraulic conductivity model domains.
Figure 3-8: Particle residence time distributions for model runs using (A) uncalibrated and (B) calibrated values of hydraulic conductivity and porosity. Uncalibrated models (A) produced residence time distributions shifted approximately three orders of magnitude later than those using calibrated values (B). In both sets of runs, increasingly complex subsurface architecture results in a wider distribution of subsurface residence time, ultimately generating some significantly longer flowpaths in the most complex geometries.
3.5 Discussion

3.5.1 Comparison of subsurface architecture characterization methods

The suite of subsurface characterization techniques applied in this study provides differing interpretations of subsurface architecture, which do not all agree. Seismic refraction, background ER, and field observations of bedrock outcropping suggest a confining layer that is generally deeper at the upstream end of the study reach and that outcrops at the downstream end of the study reach. Depth to refusal, background ER, and ER imaging of solute tracer all suggest that some geologic structure exists in the shallow subsurface near T3 causing upwelling, but only ER imaging of solute tracer and depth to refusal characterize features that suggest downwelling downstream of this location.

The shallow subsurface characterization using depth to refusal measurements provided a limited characterization of subsurface architecture, based on agreement with other investigation methods. These observations may be complicated by the presence of large cobbles and boulders in the subsurface. In the best-fit simulations, the streambed sediment and weathered bedrock layers were parameterized identically, suggesting the interface inferred from depth to refusal was erroneous. The two, combined, might accurately represent the actual sediment layer in the field, with the “less weathered bedrock” layer in the final architecture representing the actual extent of the weathered shale. It is also possible that the depth to refusal measurements were artificially shallow due the boulders in the subsurface at this site. Finally, a more detailed study at the site could consider topography along the bedrock layer.

Imaging of process areas (i.e., static features) provides a different characterization of the subsurface than imaging of process dynamics (i.e., solute transport). Time-lapse images of solute
tracer enhanced the interpretation of subsurface architecture by highlighting both temporal trends in tracer arrival and by illustrating downwelling in a location that was not identified in background ER alone. Although background electrical properties in the subsurface appear to change from upstream (T1, T2) to downstream (T3, T4, T5), we conclude these changes are related to electrical properties of the soil-water matrix (i.e., electrical resistivity) rather than hydrogeologic parameters such as permeability or hydraulic conductivity. This conclusion is based on similar observed solute transport both at upstream (T1-2) and downstream (T4-5) ER transects, which had different electrical resistivity values during background (i.e., pre-tracer) conditions. Background ER measurements are a function of electrical properties of the soil-water matrix in the subsurface, but they do not accurately represent the variables that control hyporheic exchange (i.e., head gradients, hydraulic conductivity). The ability of ER to directly monitor the transport of solutes in the subsurface identifies transport phenomena that would have been overlooked using static ER images. The conflicting interpretations of electrical properties and observed transport illustrate the difficulty in assigning hydrogeologic properties based on geophysical studies, a problem which is unique neither to this site nor method [see reviews by Lesmes and Friedman, 2005; Pride, 2005].

3.5.2 The utility of geophysical monitoring of process dynamics to reduce uncertainty in subsurface architecture

The application of geophysical methods to characterize subsurface architecture reduces the uncertainty in numerical models of groundwater flow and transport in two ways. First, geophysical investigation of subsurface architecture (i.e., process areas) is spatially distributed, characterizing the locations of geologic structures in the subsurface. While geophysical methods
rely on simplifying assumptions and/or data inversion to characterize the subsurface, they are able to inform major geologic boundaries and layer thicknesses distributed throughout the domain. Second, time-lapse monitoring of transport dynamics provides a data set that characterizes solute transport with high spatial (relative to monitoring well networks) and temporal resolution. The use of coupled geophysical monitoring and solute tracer studies shows promise for monitoring of transport along flowpaths and allowed calibration of a model that matched the distributed transport properties observed in the field.

Despite the promise of geophysical monitoring of subsurface architecture and transport dynamics, the use of these methods is not sufficient to completely characterize all subsurface transport and hydrogeologic properties at all scales. In our study, the calibrated model was only able to corroborate particle tracking and ER data within an order of magnitude. While this is a clear advance over the model based on published hydrogeologic values, it is not sufficient to make predictions at the scale of individual flowpaths (i.e., individual particles). Heterogeneity within the simulated geologic units would be necessary to achieve a better match between ER and model results. Finally, we note that ER data have large and poorly resolved support volumes; comparison of these data with particle tracking may never yield perfect results due to the disparity of measurement support volumes. The application of geophysical methods demonstrated in this study focuses on the characterization of subsurface architecture (i.e., the changes in geologic properties occurring at scales on the order of $10^{-1}$ m and larger). The exact resolution and uniqueness of results for geophysical studies like those presented in this study are subject to a variety of assumptions made in the collection and interpretation processes [see review by Linde et al., 2006]. Still, these methods provide information about subsurface structure and transport that cannot be readily obtained using traditional solute tracers and sparse monitoring.
well networks, and which has a direct effect on predicted residence time distribution and streambed flux.

Time-lapse electrical resistivity imaging at multiple transects provides a pseudo 3-D picture of solute transport in the hyporheic zone. While interpolation between transects is limited, this tool provides visualization that is useful in developing models of flow and transport. The application of ER imaging of solute tracer to models of hyporheic exchange allows for more complex parameterization of subsurface architecture, and increasingly realistic predictions of residence time distributions and flowpath network extent.

3.6 Conclusions

The objective of this study was to characterize to determine how increasingly better characterized facies boundaries in the subsurface affect predictions of streambed flux and hyporheic residence time distributions. Our work demonstrates that increasingly complex subsurface architectures generate more variable residence time distributions (i.e., wider distributions with lower peaks). Simulated streambed flux was also lower for more complex subsurface architectures, suggesting that highly idealized subsurface characterizations (e.g., homogeneous and isotropic without geologic complexity) could over-estimate hyporheic exchange flux.

We demonstrated the integration of coupled solute injection and ER imaging studies along with other common shallow subsurface investigation techniques to construct and calibrate numerical models of hyporheic exchange. The combined use of multiple geophysical methods allowed better spatial resolution of facies boundaries in the subsurface, and allowed calibration of the model to match transport observed with time-lapse ER monitoring of saline tracer. ER
monitoring of solute tracer studies was used to infer subsurface architecture and transport properties that were not otherwise be observable, leading to better calibrated flow and transport models include macro-scale heterogeneity in the subsurface. Pseudo 3-D time-lapse visualization of hyporheic flowpaths provides a conceptual model from which additional characterization of subsurface architecture can be inferred. Finally, we have demonstrated the promise of geophysical monitoring of transport dynamics as a first step toward reducing equifinality in groundwater flow models. While our objective was to characterize the role of macro-scale heterogeneities (i.e., those on the order of \(10^{-1}\) m and larger), Pollock and Cirpka [2010] demonstrated the coupled inversion of flow and electrical resistivity data collected during a solute tracer study might also be used to characterize smaller-scale hydraulic conductivity distributions in the subsurface.

### 3.7 Acknowledgements

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Chapter 4

Patterns of hyporheic exchange during baseflow recession in headwater mountain streams

This chapter will be submitted for publication with co-authors M. Fitzgerald, M.N. Gooseff, K. Singha, T.J. Voltz, and A.M. Binley. See Appendix B for co-author permissions.

4.1 Abstract

Hyporheic hydrodynamics are a control on ecological functions of hyporheic exchange, yet we lack an understanding of catchment controls on these flowpaths, including valley constraint and hydraulic gradients in the valley bottom. We performed four whole-stream, steady-state solute tracer injections and collected electrical resistivity (ER) imaging to directly characterize the 2-D spatial extent of hyporheic exchange through seasonal baseflow recession in the H.J. Andrews Experimental Forest (Oregon, U.S.A.). ER images provide spatial coverage that is unavailable with solute tracer studies and monitoring wells alone; we quantified the lateral and vertical extent of the hyporheic zone using both images and spatial moment analysis. We found that the common conceptual model of hydraulic gradients toward the stream compressing hyporheic networks was true for cross-valley (i.e., perpendicular to the valley) gradients in locations where lateral constraint (i.e., valley width) due to subsurface units (i.e., bedrock) does not exist. We found increasing hyporheic extent with decreasing vertical gradients away from the stream. Increasing hyporheic extent was observed with both increasing and decreasing down-valley (i.e., parallel to the valley) hydraulic gradients. We conclude that analysis of hydraulic gradients alone is necessary but not sufficient to characterize the likely interaction of groundwater tables as a boundary on hyporheic exchange; interaction of gradients with subsurface architecture
controls exchange processes. Increased knowledge of the controls on hyporheic exchange, temporal dynamics of these exchange flowpaths, and the spatial distribution of these is the first step toward predicting hyporheic exchange at the scale of individual flowpaths.

4.2 Introduction

Linking the transport of solutes along hyporheic flowpaths with dynamic hydrologic processes occurring at the catchment-scale remains a challenge. The physical processes controlling the individual aspects of hyporheic exchange, catchment hydrology, and riparian hydrology and hydraulics, however, have been relatively well studied as discrete processes. Hyporheic exchange (i.e., the movement of water between the stream and near-stream aquifer over relatively short spatial and temporal scales) has been well characterized at the channel unit to sub-reach scales (10s to 100s of m), with primary controls including channel bedforms [e.g., Cardenas and Wilson, 2007a; Cardenas, 2008; Elliott and Brooks, 1997a; b], geomorphology [e.g., D'Angelo et al., 1993; Gooseff et al., 2006; Wondzell, 2006; Zarnetske et al., 2007], and distribution of hydrogeologic properties of the subsurface [e.g., Cardenas et al., 2004; Packman and Salehin, 2003; Ryan and Boufadel, 2006; Salehin et al., 2004; Sawyer and Cardenas, 2009; Valett et al., 1996]. While a comprehensive understanding of how catchments and streams control hyporheic exchange is lacking, it is recognized that hyporheic hydrodynamics are the template upon which biogeochemical cycling in the subsurface occurs [Argerich et al., In Review; Battin, 1999; 2000; Boano et al., 2010; Zarnetske et al., 2011]. To advance our understanding of biogeochemical cycling, we must advance our ability to resolve hydrodynamic subsurface processes [Bencala et al., 2011; Hanrahan, 2008; Wondzell, 2006; Wondzell et al., 2009].
Hyporheic exchange occurs over a range of spatial and temporal scales [e.g., Bencala et al., 2011; Dahl et al., 2007; Stanford and Ward, 1993; Wondzell and Gooseff, In Press]. The geologic setting of a stream reach provides a first-order control on hyporheic exchange, setting the template upon which exchange may occur [Wondzell and Gooseff, In Press]. Valett et al. [1996] demonstrated the relationship between parent lithology and hyporheic exchange with solute tracer studies in three different geologic settings. Valley constraint has been associated with reduced hyporheic residence times and smaller hyporheic zones [D'Angelo et al., 1993; Stanford and Ward, 1993; Wondzell, 2006] and has also been identified as a control in riparian/floodplain systems [Wright et al., 2005]. Larkin and Sharp [1992] demonstrated that stream-groundwater interactions throughout catchments are a function of channel slope, sinuosity, incision, and aspect ratio. Few studies have considered how riparian hydrology controls hyporheic exchange processes occurring in the valley bottom. At smaller scales, both stream geomorphology and hydrogeologic parameter distribution in the subsurface are known to be controls on hyporheic exchange. Streambed geomorphology has been widely studied as a control on hyporheic exchange [Gooseff et al., 2006; Hester and Doyle, 2008; Kasahara and Wondzell, 2003]. Hyporheic flowpath complexity has been attributed to the presence of both small-scale [Sawyer and Cardenas, 2009] and large-scale [Vaux, 1968; Ward et al., In Press] heterogeneities in the subsurface, and? are observed in field, flume, and numerical studies [e.g., Cardenas et al., 2004; Salehin et al., 2004].

In addition to hydrogeologic parameters, riparian hydrology is a control on hyporheic exchange. Riparian hydrology exists as a buffer between boundaries set by large-scale, relatively slow-moving processes at the hillslope, catchment, or regional setting and more dynamic processes occurring in the stream channel. Burt [2005] notes “the riparian zone is perhaps the
most important element of the hydrological landscape given that it can decouple the linkage between the major landscape elements, hillslope and channel”. Left out of this perspective is hyporheic hydrology, in which some portion of the riparian zone is comprised of stream water traveling along hyporheic flowpaths. The view of riparian hydrology as a function of its boundaries is prevalent in the literature [e.g., Vidon and Hill, 2004]. As a boundary condition, increasing stream flow may increase hyporheic exchange rates [e.g., Elliott and Brooks, 1997a; Elliott and Brooks, 1997b; Fabian et al., 2010; Hart et al., 1999; Packman and Salehin, 2003]. Solute tracer studies conducted at the reach-scale generally note that the ratio of storage zone area to stream area (commonly $A_S/A$) decreases with increasing flow [Butturini and Sabater, 1999; Fabian et al., 2010; Karwan and Saiers, 2009; Morrice et al., 1997; Schmid et al., 2010; Zarnetske et al., 2007], though others have found little correlation [Hart et al., 2002; Schmid et al., 2010]. Groundwater discharge to streams has widely been theorized to restrict the spatial extent of hyporheic flowpaths [e.g., Hakenkamp et al., 1993; Hynes, 1983; Meyer et al., 1988; Palmer, 1993; Vervier et al., 1992; White, 1993]. Indeed, numerical and field studies have confirmed this hyporheic contraction due to ambient gaining conditions [e.g., Boano et al., 2008; Cardenas and Wilson, 2007a; D'Angelo et al., 1993; Harvey and Bencala, 1993; Storey et al., 2003; Williams, 1993; Wondzell and Swanson, 1996; Wroblicky et al., 1998]. In contrast, Wondzell [2006] completed replicate tracer studies in steep headwater catchments in the H.J. Andrews Experimental Forest under different baseflow conditions, finding hyporheic extent, evaluated as tracer arrival in a monitoring well network, was not changed during baseflow recession. The limited view of riparian hydrology may be due to the different scales of spatiotemporal variability in both streams and hillslopes and has proven insufficient to characterize riparian hydrology (as evidenced by contradictory results of past studies).
Solute tracer experiments are commonly employed to characterize hyporheic exchange [Stream Solute Workshop, 1990], and results are commonly interpreted with simple solute transport models that simulate time series of stream tracer concentrations [Bencala and Walters, 1983; Briggs et al., 2008; Choi et al., 2000; Runkel, 1998]. Model interpretations provide only spatially lumped characterizations of only a subset of subsurface flowpaths (i.e., those within the study’s window of detection, generally the short timescales) [e.g., Gooseff et al., 2003; Harvey et al., 1996; Harvey and Wagner, 2000]. Further, stream solute transport model results are sensitive to exchange with other slow moving water in streams (i.e., surface transient storage zones [e.g., Briggs et al., 2008]). Shallow monitoring wells have been used to sample hyporheic water, but they provide spatially discrete point measurements of tracer in the subsurface, as an individual well is only sensitive to the suite of flowpaths that intercept its location. Hence there is significant mis-match between well observations and stream solute transport model simulations [e.g., Harvey et al., 1996; Wondzell, 2006]. In response to the need for increased spatial and temporal resolution in monitoring hyporheic exchange, Ward et al. [2010b; Chapter 2] demonstrate the use of electrical resistivity (ER) imaging coupled with electrically conductive solute tracers to image hyporheic flowpaths.

The objective of this study is to assess the roles of changing hydrologic conditions during baseflow recession and valley constraint as controls on hyporheic exchange. We seek to answer the questions: (1) How does hyporheic extent change as a function of valley constraint? and (2) How does hyporheic extent change as a function of changing hydrologic forcing during baseflow recession? Based on our review of the literature, we expect: (1) hyporheic flowpath networks will spatially expand during baseflow recession due to falling gradients from the hillslopes toward the stream; (2) hyporheic extent will be most consistent through time in locations where valley constraint is largest, because subsurface controls (i.e. hydrogeologic properties, confining
units) control extent rather than cross-valley gradients; and (3) hyporheic extent will be largest in locations and during periods where down-valley gradients are largest, because steeper down-valley gradients drive more down-valley flow in the subsurface. To answer these questions we conducted four solute tracer studies in a study reach including a laterally constrained upper reach and a less constrained lower reach during baseflow recession. Tracer concentrations were monitored in-stream, in monitoring wells, and using electrical resistivity imaging. A suite of associated hydrological data was also collected to characterize vertical, cross-valley, and down-valley hydraulic gradients within the riparian zone. These assessment techniques provide superior spatial resolution in the subsurface and temporal resolution through the baseflow recession period, allowing us to link hyporheic exchange with dynamic valley-bottom boundary conditions (i.e., in-stream flow and hydraulic gradients from the hillslopes to the stream).

4.3 Methods

4.3.1 Site description

Field studies were completed at the H.J. Andrews Experimental Forest, located in the western Cascade Mountains of Oregon, USA (48° 10′ N, 122°15′W). Studies were conducted in WS3, a steep headwater catchment draining 101 ha. The watershed is steep, with hillslope gradients greater than 50% constraining narrow valley bottom. The watershed ranges in elevation from 497 to 1070 meters above mean sea level. Soils are generally shallow loams (1-2 m) with high porosities and infiltration rates [Dyrenness, 1969]. Wondzell et al. [2009] report a saturated hydraulic conductivity of $1.7 \times 10^{-5}$ m s$^{-1}$ as the geometric mean of slug tests in the riparian soils of
an adjacent watershed, while Kasahara and Wondzell [2003] report an average value of $7 \times 10^{-5}$ m s$^{-1}$ for WS3 and WS1 (a neighboring headwater catchment).

A highly instrumented study reach of approximately 40 m was established (Figure 4-1A). The second-order stream has a 14% gradient through the study reach. Stream morphology is a sequence of pools, riffles, and steps. An average of 8.4 steps or riffles per 100 m of stream length contributes 54% of elevation change along the stream [Kasahara and Wondzell, 2003].

WS3 is a narrow valley (8.5 m), described as bedrock-constrained by Wondzell [2006]. The upper end of the study reach (ER transects 1-3) is highly constrained, with the stream against the steep bedrock wall on the north side of the valley. A small alluvial deposit is present along the southern valley bottom. In contrast, the lower portion of the study reach (ER transects 5-6) is in a flatter and wider valley bottom. A large alluvial deposit exists across the width of the valley, held in place by deposits of boulders that span the valley width.

**4.3.2 Hydrologic observations**

Flow rate at the outlet of WS3 is gauged using a permanently installed weir maintained by the US Forest Service. WS3 contains a network of 17 shallow monitoring wells and 8 piezometers installed in 1997. Wells and piezometers are hand-driven lengths of PVC, with maximum penetration of 1.7 m (less than 1 m in many locations). Wells were screened over the bottom 50 cm by drilling 0.32 cm diameter holes, with an approximate density of 0.25 cm$^{-2}$. Piezometers were screened with the same size holes and density over the bottom 5 cm only. Additional details about well installation and network layout are provided by Wondzell [2006]. The potentiometric surface was recorded using pressure transducers and water level capacitance.
rods in a sub-set of the well network (13 wells and 8 piezometers) and at 5 locations in the stream channel.

Water surface elevations observed in the stream and monitoring wells were used to construct potentiometric surface maps at 30-minute intervals for the study period by linearly interpolating between measurement locations on a 25 cm grid. The area of analysis was clipped to only include the riparian and hillslope areas bounded by the monitoring well network and on the south side of the stream centerline. For each grid point the cross-valley gradient (i.e., gradient perpendicular to the down-valley axis) and down-valley gradient (i.e., the gradient parallel to the down-valley axis) were calculated. Cross-valley gradient is positive for gradients sloping in the northeast direction, and negative for gradients sloping in the southwest direction. The 25 cm grid data were used to calculate the valley bottom average cross-valley and down-valley gradients. Additionally, observation locations were used to construct a triangular grid consisting of 16 finite elements covering the valley bottom (Figure 4-1B). Cross-valley and down-valley gradients for each element were calculated at 30-minute intervals for the study period. The triangular grid was divided into elements that extended through the stream itself (“through-stream”, or TS), in the riparian area (“stream-area”, SA), and those near the hillslopes (“near-hillslope”, or NH).

Water surface observations in the stream and piezometer network were used to calculate vertical hydraulic gradients at 2 locations. Vertical hydraulic gradient (VHG, m m⁻¹) was calculated as:

\[
VHG = \frac{h_{\text{stream}} - h_{\text{piezometer}}}{e_{\text{streambed}} - e_{\text{tos}}}
\]

(1)

where \(h_{\text{stream}}\) and \(h_{\text{piezometer}}\) are the head in the stream and piezometer, \(e_{\text{streambed}}\) is the streambed elevation, and \(e_{\text{tos}}\) is the elevation of the top of the piezometer screen. [Baxter et al.,]
Gradients are positive in upwelling locations and negative in downwelling locations.

### 4.3.3 Solute tracer studies

Four 48-hr constant-rate injections of sodium chloride (NaCl, a conservative tracer) were completed in each watershed, at the same locations used by Wondzell [2006] (approximately 50 m upstream of the study reach in each watershed). A constant rate of concentrated NaCl solution was injected directly into the stream channel; injections were designed to increase in-stream electrical conductivity (EC) by 100 µS cm$^{-1}$. EC was used as a surrogate for tracer concentration in all measurements [Gooseff and McGlynn, 2005; Payn et al., 2009; Wondzell, 2006]. For all studies, injections began between 13:00 and 14:00.

EC of surface water was recorded in the stream channel using temperature and EC probes manufactured by Campbell Scientific, Inc. (Logan, Utah, United States) at a well-mixed stream reach near the upstream end of the study reach (IS1 in Figure 4-1A). EC was also monitored in the well network during the study. Well sampling procedure was to purge one well volume using a manual siphon and measure EC using either an EcoSense EC300 (YSI, Inc., Yellow Springs, Ohio, United States) or a Model 107 Temperature/Level/Conductivity meter (Solinst, Inc., Georgetown, Ontario, Canada) deployed down the well.
Figure 4-1: Site location and instrumentation maps for WS3 (A) in the H.J. Andrews Experimental Forest, located in the Cascade Range of central Oregon, USA. Piezometers and well transects are organized by letter (from C downstream to I upstream) and number (increasing from North to South across the valley); in-stream observations are identified by the prefix IS and location number. (B) Finite elements used for analysis of riparian hydraulic gradients in WS3, representing the maximum number of data that can be independently obtained from the instrumented well network. Each of the finite elements is classified as through-stream (TS), stream-area (SA) or near-hillslope (NH), and assigned a number to uniquely identify each element.
4.3.4 Electrical resistivity imaging

Data collection

ER data were collected using a 10-channel Syscal Pro Resistivity Meter (IRIS Instruments). Data were collected on a network of electrodes installed in 6 transects oriented perpendicular to the valley (T1-4 during injection 1; T1-6 for injections 2-4), spaced approximately 5 m apart within the study reach. Electrodes were positioned within each transect across the stream using a variable spacing layout where electrodes in the valley bottom and stream portion of the site were positioned with approximately 1 m spacing laterally in the valley bottom; electrode spacing was increased to a maximum spacing of 2 m in the hillslopes (Figure 4-1A). This arrangement was selected to yield high-resolution measurements in the valley bottom, where the greatest changes in electrical conductivity were expected.

Electrodes were manufactured from 1.27-cm diameter PVC pipes approximately 0.75 m in length. Conductive foil tape was wrapped around the pipe about 10 cm from the bottom. An 18-gauge stranded wire connected the foil tape (i.e., contact surface below ground) to the trunk line of wires (solid strand 18-gauge), which connected to the ER switch box. Contact resistance of the electrodes was checked with the Syscal Pro prior to each injection and periodically during the ER monitoring period. Contact resistance values ranged from 1 to 16 kOhms with most (about 90%) lower than 5 kOhms for the entire duration of each monitoring period.

Data were collected at each transect using a mixed dipole-dipole array with a total sequence of 323 measurements collected; the same sequence was used for all transects. The
sequence was selected to maximize coverage in the valley bottom with a minimum number of individual measurements collected (to maximize temporal resolution of the data set). Data collection took approximately 14 minutes per transect, with one full survey of all six transects taking approximately 84 minutes. The electrical resistivity distribution in the subsurface was assumed static during the collection for each transect (i.e., we did not account for changes between the first and last of the 323 quadripoles collected at each transect), and all data were assigned to the time at the mid-point of the data collection for a given transect.

In an effort to maximize temporal resolution of ER data collection, all data were collected along individual transects; no between-transect data were collected. Background data were collected before the tracer injection began to characterize the distribution of background resistivity distribution. ER surveys were collected continuously for the first 12 hours following the start of the tracer injection and for 12 hours immediately following the end of the injection (i.e., when the rate of change of the electrical conductivity in the hyporheic zone would be the greatest). ER surveys were collected every 2-4 hours for the remaining times during the study (during both plateau and tailing conditions, when slower changes were expected). For this study, we limited our analysis to the first 120 hours after the study began because that was the period during which data were collected for each of the four injections. To improve quality, a minimum of two measurements were stacked (i.e., averaged) for each resistance measurement. If the standard deviation for those measures was greater than 2%, an additional two data points were added to the average. The final standard deviation of the measurements was recorded for each quadripole.
Data inversion

ER data inversion was completed using the research code R2 (v2.6, Generalized 2-D Inversion of Resistivity Data, available online at: http://www.es.lancs.ac.uk/people/amb/Freeware/freeware.htm). The inversion code is described in Binley and Kemna [2005]. Data were weighted for the inversion scheme as the reciprocal of the standard deviation between stacked measurements. Pre-injection data were inverted starting with a homogeneous subsurface model. Error model parameters were manually adjusted to yield a root-mean-square error (RMSE) with respect to the noise in the data as close to 1.00 as possible for the background inversion. Data collected during and after the tracer injection were inverted using the final background model as a starting model. Thus, each timestep was inverted independently. This method limited the potential for errors in the inversion process to be propagated through the time series data, although no temporal regularization was utilized [LaBrecque and Yang, 2001; Day-Lewis et al., 2002]. Inversion RMSE throughout the injection and monitoring period was used to determine timesteps that failed to converge on a solution. For transects and timesteps where inversion models did not converge for a majority of timesteps due to high levels of noise in the data collected, the transect or timestep was omitted from the analysis.

Post-processing

The background resistivity image was subtracted from each of the images collected during and after the tracer study to calculate the percent change in resistivity for each pixel. Where salt-labeled stream water entered the subsurface, we expected resistivity to decrease
compared to background. To quantify the hyporheic extent, we applied two filters. First, the resolution matrix for each inversion was used to select only pixels where \( \log_{10}(\text{resolution}) \geq -2 \). This limits analysis to the pixels where the pixels are well resolved, based on the chosen measurement and inversion schemes. Next, we applied a threshold of a minimum decrease in electrical resistivity of 3% to parse meaningful changes due to tracer presence from error in the data collection and inversion [after Ward et al., 2010b; Chapter 2]. The cross-sectional hyporheic area was calculated as the number of pixels meeting both filters multiplied by pixel area. This procedure was completed for each transect at all timesteps. The calculated hyporheic area is sensitive to the thresholds set for both resolution and percent-change.

Finally, we compared the breakthrough curves observed in monitoring wells to pixel breakthrough curves in corresponding locations. Comparison of pixel breakthrough curves with observed point measurements has been successfully applied in the literature to characterize fluid flow and solute transport [e.g., Binley et al., 1996a; Slater et al., 2000; Slater et al., 2002]. The purpose of our comparison is to demonstrate that the ER images are sensitive to the tracer in the subsurface. ER is based on inversion of 3-D field measurements; monitoring wells provide a spatially discrete measure of concentration. It is not expected that a perfect fit will be observed between the two, but this comparison provides qualitative evidence of ER sensitivity to the tracer.

### 4.3.5 Spatial moment analysis

We calculated the first and second spatial moments of the ER data at each 2-D transect to characterize changes in the plume centroid and variance throughout the injection period. The evolution of a solute plume can be described by spatial moments of tracer concentration. In a
two-dimensional space, the \( ij \)th spatial moment \( (M_{ij}) \) is defined as [after Aris, 1956; Freyberg, 1986]:

\[
M_{ij} = \int_\Gamma \Delta C(x, z, t) x^i z^j dx dz
\]

where \( \Gamma \) is the domain, \( x \) and \( z \) are Cartesian coordinates, and \( t \) is time elapsed. The application of spatial moments in this study focuses solely on the spatial location and variance of the plume (the first and second spatial moments), so \( \Delta C \) is taken as the opposite of the change in bulk resistivity (i.e., \( \Delta C = -\Delta \sigma_b \)), where \( \sigma_b \) is the bulk resistivity change, assuming all changes in electrical resistivity are related to concentration. Similar studies [e.g., Hagarty et al., 2010; Singha and Gorelick, 2005] have used Archie’s Law [sensu Archie, 1942] to convert bulk resistivity change to change in fluid tracer concentration; we omit this step (a conversion using a linear relationship) because we are not attempting to estimate tracer masses in the subsurface and focus only on their spatial distribution. The calculations used the post-processed data set, meaning that only pixels that were both well resolved and where solute was observed to cause a minimum decrease of 3% were included. The value of all other pixels was set to zero for the analysis.

The zeroth moment \( (M_{00}) \) is the total mass in the aquifer at time \( t \). The centroid of the plume \( (x_c, z_c) \) is defined by the first moments normalized by the total mass, as [after Freyberg, 1986; Liu et al., 2004]:

\[
x_c = \frac{M_{10}}{M_{00}} \quad \text{(3), and}
\]

\[
z_c = \frac{M_{01}}{M_{00}} \quad \text{(4).}
\]

Finally, the spatial covariance tensor of the plume is defined by the normalized second moments of the plume about the centroid, \( \sigma^2_{ij} \), defined as [after Freyberg, 1986; Liu et al., 2004]:
4.4 Results and discussion

4.4.1 Hydrologic changes during baseflow recession

During the study period flow ranged from 3.7 to 334.1 L s⁻¹, with a seasonal recession from early June to late August (Figure 4-2). Temporal trends in cross-valley and down-valley components of hydraulic gradient within the riparian zone for each finite element. Down-valley and cross-valley gradients throughout the season are presented in Figure 4-3. Down-valley gradient both increased and decreased for some elements of each of the three groupings (NH, SA, and TS). In general, the storm event created a large perturbation that relaxed by mid-June 2011. After that time, down-valley gradients for elements NH5, NH7, SA1-2, and TS3-4 decreased through the study period, while those for NH1-4, NH6, SA1, TS1-2, and TS5 increased through the remainder of the season. The storm event caused some cross-valley gradients to increase in magnitude, though gradients turned both toward and away from the stream in different locations. After the catchment relaxed from the storm event, gradients for elements NH2, NH4-6, SA3, and TS1-5 turned away from the stream (i.e., toward the southwest valley wall, negative in our convention), while those for NH1, NH3, NH7, and SA1-2 turned toward the stream (i.e., toward the northeast valley wall, positive in our convention). Additional analysis of the gradients throughout the study period is presented by Voltz [2011].
Vertical hydraulic gradient was calculated at T3 (piezometer D4) and T5 (piezometer G4), where paired piezometer and in-stream water elevations were collected. At T3, VHG was -0.619, -0.634, -0.701, and -0.679 m m⁻¹ for injections 1-4, respectively. VHG at T5 was -0.436, -0.442, -0.454, and -0.450 m m⁻¹ for injections 1-4, respectively. VHG estimates by Wondzell [2006] at piezometer D4 showed little change under high and low baseflow conditions.

![Figure 4-2: Flow and precipitation in WS3. Flow was gauged at a weir located approximately 100 m downstream of each study reach. Injection periods are shown as shaded bars.](image)

**Figure 4-2**: Flow and precipitation in WS3. Flow was gauged at a weir located approximately 100 m downstream of each study reach. Injection periods are shown as shaded bars.

### 4.4.2 Solute tracer studies

The breakthrough curves for the four replicate tracer studies were logged at the upstream end of the study reach (Figure 4-4). Background tracer concentration was subtracted from each observation set. Plateau EC values range from increases of 60 to 120 µS cm⁻¹ during the injections. Injection flow rates were checked at intervals of 3-4 hours to ensure a constant rate
injection into the stream. Tracer arrival time at the study reach was fastest during the first injection, with first arrival occurring within minutes of the injection start time, and slowed with decreasing flow rate, ultimately taking approximately 20 minutes for first arrival during the final injection (Figure 4-4). Tailing behavior of the tracer was delayed between injections in a similar pattern to arrival time, with more prolonged tailing in each injection (Figure 4-4). This increased tailing suggests an increased influence of hyporheic exchange in retarding the transport of tracer through the study reach.

**Figure 4-3:** Down-valley gradient (i.e., gradient parallel to the valley bottom) and cross-valley gradient (i.e., gradient perpendicular to the valley bottom) for each element throughout the study period, in the left and right columns, respectively. Larger down-valley gradients indicate steeper down-valley components of hydraulic gradients; smaller values represent shallower down-valley gradients. For cross-valley gradient, positive values represent gradients toward the northeast valley wall; negative values represent gradients toward the southwest valley wall. Increased magnitude of cross-valley gradient indicates steeper gradients in that direction.
4.4.3 Electrical resistivity imaging of solute tracer

Both observed fluid EC and electrical resistivity for the co-located wells and ER transects are presented in Figure 4-5. For monitoring wells, the 50 cm screened length intersected up to six pixels in the inversion images; for piezometers the screened section intersected a single pixel. Pixel resistivity trends are generally well correlated with monitoring well observations of EC. Decreases in predicted resistivity of the pixels are temporally aligned with the appearance and disappearance of tracer in the wells. No effort was made to optimize the resistivity models based on these observations.

Figure 4-4: Breakthrough curves at the upstream end of the study reach in WS3, as the change in observed electrical conductivity (EC) of the surface water (a surrogate for the concentration of the sodium chloride tracer) in log-space. The three panels provide increased resolution on the rising (left panel) and tailing (right panel), while the center panel shows the plateau conditions during a majority of the injection. The injections occurred from times 0 to 48 hours. During tracer injection 1 the injection rate approximately doubled after about 39 hours of injection (at approximately 03:00). This change was discovered at approximately 06:00 and the injection flow rate was reset to its initial rate. Arrival time at the study reach was increasingly later as streamflow decreased during the season. Studies during lower flow conditions show increased tailing at late times, suggesting increased transient storage between the injection point and study reach.
Figure 4-5: Examples of observed monitoring well tracer concentration (as change in electrical conductivity) and change in pixel resistivity for piezometer G5 (left column) and well D4 (right column) during each injection. Multiple lines are shown where the well screen intersects multiple pixels in the inversion. Each row corresponds to one of the four constant-rate injections. Tracer presence in wells is temporally correlated with the observation of tracer in the electrical resistivity models, demonstrating that the ER models are sensitive to the solute tracer. Because concentration is a point measurement and pixel resistivity is based on 2-D inversion of a 3-D field measurement, perfect agreement is not expected. However, results demonstrate conclusively that the geophysical inversion is sensitive to the tracer.
Figure 4-6A: Visualization of hyporheic extent 24 hours after the start of each injection. Color relates to the observed drop in pixel resistivity; opacity is based on resolution matrices with less certain data plotted as transparent. Hyporheic location and extent are spatially consistent for T1, T2, and T5 across all injections. Hyporheic extent in T3 and T6 appears to decrease with decreasing flow. Plateau concentration during each injection ranged from 60 to 120 µS cm⁻¹, with the highest plateau during injection 2 and the lowest during injection 4.
Figure 4-6B: After 48 hours, hyporheic areas are larger than those at 24 hours elapsed. Spatial differences are apparent in T2, T3, T5, and T6. See Figure 4-8 for areas through time for each transect and injection.
Figure 4-6C: After 72 hours, hyporheic flowpaths are still labeled with solute tracer, particularly in T1-T3 and T5. At this time the stream returned to background concentration for all injections (based on in-stream logging of electrical conductivity). Electrical resistivity imaging is able to characterize flowpaths that are beyond the window of detection for traditional solute tracer studies. The in-stream plateau achieved was highest during injection 2, which delivered increased solute mass along hyporheic flowpaths. This may contribute to the highly persistent flowpaths observed during injection 2.
on the well data; the comparison is presented only as evidence that the electrical geophysical models were sensitive to the solute tracer.

T4 was omitted from all injections due to a failure to converge on a subsurface model for both background conditions and during the solute tracer injection. Data from this transect were very noisy, and evidence of electrical shorts in the wiring was observed when electrodes were removed from the site after injection 4. T1 was omitted from injection 4 due to a failure to converge in the inversion process, due also to noisy measurements. For those locations where clean data were collected, mean stacking error in the ER data collected ranged from 0.08 to 0.46% for each transect during each injection. Mean RMSE for the inversions ranged from 1.00 to 1.10% for each transect during each injection. Maximum, minimum, and mean data collection (i.e., stacking) and inversion error are summarized in Table 4-1. Throughout the study period, errors in both data collection and inversion remained very low and relatively consistent.

4.4.4 Qualitative trends interpreted from ER images

ER images of solute in the subsurface are presented at 24, 48, and 72 hours after the start of the injection (Figures 4-6A, 4-6B, and 4-6C, respectively). After 24 hours solute tracer appears in the alluvial deposit adjacent to the stream in T1-3, though in a relatively small cross-section. The hyporheic zone appears to pinch-out at T3. Downstream, a large hyporheic zone is observed in T5 and T6 after 24 hours (Figure 4-6A). When the injection is stopped after 48 hours (Figure 4-6B), similar spatial trends are observed, though the hyporheic zone is larger at all cross-sections. After 72 hours (Figure 4-6C) the flowpaths are flushing solute, although a substantial hyporheic zone is still observed. 24 hours after the injection ends (72 hr since it began), the
stream has returned to background; tracer is still present in the monitoring well network, and a substantial amount remains in hyporheic flowpaths.

Both spatial patterns of exchange, as well as the late-time presence of solute along hyporheic flowpaths (hereafter “persistence”) can be assessed qualitatively using time-lapse images of the subsurface, and quantitatively through late-time behavior observed in plots of hyporheic area through time (Figure 4-8). At T1, the pattern of hyporheic exchange is consistent during injections 1 through 3 after 24 and 48 hours, though a more persistent tracer-labeled hyporheic zone appears present during lower flow conditions (injection 3), and the most persistence at 72 hours is observed during injection 2 (i.e., more tracer is present at later times). For T2, tracer presence is spatially similar during all injections, although a greater extent is observed under higher flow conditions. The most persistent flowpaths appeared for injection 2, though lower flow conditions (injections 3 and 4) had more persistent tracer-labeled flowpaths than injection 1 based on solute presence after 72 hours (Figure 4-6C). In general, both hyporheic extent and persistence (24 hours after the injection) are largest during higher flow conditions.

The peak change in pixel resistivity during each injection is presented in Figure 4-7. This time-independent analysis allows comparison of solute presence or absence in the subsurface between injections without the complicating factor of arrival time at a given location. For T1-3, the hyporheic area appears largest for the highest flow-rate conditions. At T5 and T6 the spatially largest hyporheic zones are observed during injection 2, with smaller plumes during injections 3 and 4.
Figure 4-7: Peak changes in pixel resistivity observed during the tracer study. The distribution of the peak change in pixel resistivity provides a time-integrated view of the distribution of solute tracer in the subsurface.
Figure 4-8: Temporal trends in hyporheic area (minimum 3% decrease in resistivity) based on electrical resistivity images. Peak area for T1-T3 (in the steeper, more constrained section of the study reach) generally increases with increasing flow rate. In the flatter, wider valley segment (T5-T6) peak area does not show a consistent relationship with flow.
Figure 4-9: Cross-sectional area of the hyporheic zone interpreted for thresholds ranging from 0.5% to 6.5% decrease in resistivity (increments of 0.5% shown) for T1, injection 1. While the interpreted area is always lesser for increasingly high thresholds, the timing of the solute arrival and peak are consistent across thresholds greater than about 3%.
Figure 4-10: Peak hyporheic area as a function of (A) stream flow, (B) vertical hydraulic gradient, (C) valley-bottom averaged down-valley gradient, and (D) valley-bottom averaged cross-valley gradient. Overall trends show the relationship between peak area and flow is slightly negative. Hyporheic area trends are negatively related to down-valley gradient (i.e., increasingly steep down-valley gradients generate smaller hyporheic zones) and positively related to increasing cross-valley gradient (i.e., increasingly strong gradients toward the stream). Increasingly strong vertical hydraulic gradients away from the stream are related to smaller peak hyporheic area, in opposition to the conceptual model of “compressed” hyporheic zones with stronger gradients toward the stream. In all cases, data are highly variable about the overall relationships, and these relationships break down when considering individual transects.
4.4.5 Quantitative spatial and temporal trends in hyporheic extent

Peak cross-sectional area of the hyporheic zone at each electrical transect was variable during the study period. The temporal patterns in cross-sectional area where resistivity dropped by at least 3% are presented in Figure 4-8. First, individual transects can be examined under different flow conditions to analyze trends as a function of flow rate. At T1, peak area ranged from 19 to 25 m². The largest and smallest areas were observed during injections 3 and 2, respectively, while injection 1 produced an intermediate peak value for area. Hyporheic persistence in time was greatest for intermediate flows (i.e., injections 2 and 3). At T2, peak area ranged from 14 to 23 m², observed during injections 4 and 3, respectively. The hyporheic zone was flushed of tracer most rapidly during the highest flows, and had increased persistence for intermediate and low flows. Peak areas at T3 ranged from about 14 to 25 m², observed during injections 3 and 2, respectively. Solute in the hyporheic zone was most persistent during higher flow conditions. At T5, hyporheic area ranged from 22 to 33 m², observed during injections 2 and 3, respectively. No apparent trend with flow rate was observed for area or persistence. Finally, peak areas at T6 were largest for higher flow rates, with peak areas of 25 and 34 m² at the lowest and highest flow conditions observed at that transect (injections 3 and 4, respectively). No trend in persistence is apparent at this transect.

Our findings show that hyporheic extent, as interpreted from 48-hr stream tracer injection studies, generally increases with increasing stream discharge (Figure 4-10A). This relationship is most clearly apparent for T6. Results at T1-3 show the same overall trend, though cross-sectional areas observed during injections 2 and 3 are higher than those observed during the highest flow rate conditions (injection 1). The behavior of T5 is anomalous in this plot, showing a general trend of decreasing area with increasing flow. We believe increased flow in the channel is an
expression of hydraulic gradients in the valley bottom; the increased flow rate itself may not be responsible for increased hyporheic extent.

Hyporheic extent shows a positive relationship with decreasing VHG at T3 (i.e., stronger losing conditions occur with smaller peak hyporheic areas; Figure 4-10B). At T5, VHG was nearly constant across the flow conditions, though more strongly losing VHGs at this location generally occur with larger hyporheic areas (i.e., increasing losing gradients appear with increasing area). Our finding at T3 is in contrast to the conceptual model of increased hyporheic extent due to losing conditions. The converse (i.e., compressed hyporheic zones under strong gaining conditions) has been reported in both numerical [Boano et al., 2008; Cardenas and Wilson, 2007b; D'Angelo et al., 1993] and field studies [Harvey and Bencala, 1993; Storey et al., 2003; Williams, 1993; Wondzell and Swanson, 1996; Wroblicky et al., 1998]. While increasingly strong gains may compress hyporheic zones, our results do not indicate hyporheic expansion with increasing losing gradients. Observations in our study indicate that the strength of the losing gradient does not necessarily increase hyporheic extent in the subsurface. The flowpath network is likely limited by some other control (i.e., surface morphology, subsurface distribution of hydrogeologic parameters) under the conditions we observed.

Hyporheic area as a function of valley-bottom averaged down-valley gradient is plotted in Figure 4-10C. In the upper reach (i.e., T1-3, steeper and higher confinement), increasingly steep down-valley gradient was weakly related to decreasing hyporheic extent. For T5 (in the flatter and wider section of the study reach), increasingly steep down-valley gradient trended with increasing hyporheic extent, while T6 (the next transect downstream) showed the opposite behavior. T5 is the first transect downstream of an apparent “pinch point” in the hyporheic zones that causes upwelling of hyporheic flowpaths. Downwelling is expected at T5 because increased
subsurface capacity to transport water down-valley is present; steeper gradients more rapidly drive flow into the subsurface yielding larger observed hyporheic areas.

Peak hyporheic extent generally increases with increasingly positive valley-bottom averaged cross-valley gradients (i.e., larger areas when average gradients from the hillslope to the stream are larger; Figure 4-10D). This relationship holds for T1-3 (in the steeper, confined section of the study reach) and is observed for T6. As with down-valley gradient, behavior at T5 is anomalous, showing decreasing hyporheic area with increasingly strong gradients toward the stream. Behavior at T5 fits our conceptual model of gradients toward the stream compressing hyporheic networks, while behavior at T1-3 and T6 is in opposition to this model. Behavior in upstream transects suggests that cross-valley gradients are not the limiting factor in hyporheic extent; confinement due to bedrock limits hyporheic exchange in these locations. Although valley gradients toward the stream are increasing for T1-3, we suspect increased riparian water demand may be lowering riparian water tables that, in turn, create steeper cross-valley gradients from the hillslope to the stream. This demand would also cause a steeper gradient from the stream to the riparian area, allowing hyporheic network expansion. We hypothesize in this case it is not the gradients from the hillslope to the valley bottom that control exchange, but the evapotranspiration demand in the riparian area.

The spatial coverage provided by ER images allows assessment of both vertical and lateral constraint in the subsurface (whereas previous studies have only been able to assess this control at the broad valley scale). In the steeper, confined valley (T1-3), the lateral extent of tracer penetration into the aquifer is relatively consistent under all flow conditions. Vertical penetration is decreased for injections with lower flow-rates. It is likely that lateral confinement limits the tracer along the northeast valley wall (X = 10 and larger). Vertical confinement limits hyporheic extent in higher-flow conditions for these transects (area change in the vertical
dimension indicates lateral confinement is not the control). In contrast, the flatter and less-confined lower portion of the study reach (T5-6) exhibits lateral penetration that is more sensitive to cross-valley gradients. The changing lateral extent of the hyporheic zone in T5 shows the cross-valley control in this section of the study reach. Consistent vertical penetration suggests that vertical confinement is not limiting at T5-6.

Based on the analysis of valley-average gradients and hyporheic area, which showed some trends but variation for individual transects, linear regression slopes for cross- and down-valley gradients at each finite element were plotted against the observed peak hyporheic area at each transect were calculated (Table 4-2). We compare every finite element with every hyporheic area observed because the flowpaths present at each ER transect are a complex function of the gradients in the entire valley bottom, not only those that intersect the transect. This analysis considers gradients at a more local scale than the valley-bottom averages presented in Figure 4-10C and 4-12D. We choose not to present R^2 values, nor do we posit that linear relationships are necessarily expected to provide the best fit; these data are presented only to consider positive and negative relationships between gradients and areas. As such, these data should be interpreted only as a binary indication (i.e., positive or negative) of the relationship between gradients and hyporheic area.

In the flatter and less constrained valley, increasing near-hillslope down-valley gradients are generally directly related to decreasing areas (NH1-4), whereas those in the steeper and more constrained upper reach are generally correlated with increasing hyporheic extent (NH5 and 7, in particular). Steeper down-valley gradients trend with increasing hyporheic area for stream-area finite elements (SA), and a mixed relationship is present in the through-stream (TS) elements. In all down-valley gradient analyses, T5 is anomalous in its behavior compared to the other transects.
With respect to cross-valley gradients, where positive gradients are toward the northeast valley wall, and negative gradients toward the southwest valley wall, NH2 and NH4-6 behave in opposition of our expectation. Increasing gradients toward the stream channel in these elements.

Table 4-1: Summary of error in electrical resistivity data collection and inversion. Error in the data collection is reported as the standard deviation of the stacked measurements as a percent of the mean (i.e., the coefficient of variation). Inversion error is reported as the root-mean-square error (RMSE) for the best-fit subsurface model. Data points listed as “--” represent locations where data were either not collected, or where inversions were not possible due to noisy data.

<table>
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<th>Stacking Error (%)</th>
<th>Inversion RMSE (%)</th>
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<td>T6</td>
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Table 4-2: Linear regression slopes for finite-element gradient vs. area relationships. Linear regression analysis of gradient-area relationships provides insight into general patterns. Shaded cells (positive values) indicate relationships where (A) steeper down-valley gradients are related to increased hyporheic area, and (B) steeper cross-valley gradients toward the northeast valley wall are related to increased hyporheic area. Patterns show that steeper down-valley gradients are related to decreasing hyporheic extent for near-hillslope (NH) elements, and positively related to hyporheic area for observations in both stream-area (SA) and through-stream (TS) elements. Steeper cross-valley gradient toward the stream is associated with increased hyporheic area for both NH and TS elements, and with decreased hyporheic area for SA elements. There is variability within these general patterns for both down-valley and cross-valley gradient.

A. Down-valley Gradient vs. Area

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B. Cross-valley Gradient vs. Area

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trend with increased hyporheic area for these elements. NH1, NH3, and NH7 fit our conceptual model of a negative relationship (i.e., increasing gradients toward the stream compress hyporheic flowpath networks). SA1-2 fit our conceptual model, while SA3 does not. All TS elements show increasing hyporheic extent with increasingly positive cross-valley gradients. Because these elements are informed by data from both banks, it is difficult to interpret these changes as increasingly toward or away from the stream.

We hypothesized that hyporheic flowpaths would spatially expand during baseflow recession due to falling gradients from the hillslopes toward the stream. Our study found falling gradients toward the stream occurred with increased hyporheic area for some elements, but there was no consistent trend. The expected relationship was present in the elements closer to the riparian valley-hillslope transition (i.e., NH1, NH3, NH7, SA1-2), while elements closer to the hillslope show the opposite relationship (i.e., NH2-4, NH6). Results of this study show that hyporheic compression by hillslope gradients is possible, though the location and scale of the gradient measurement is important. Gradients across the hillslope-riparian transition appear to control hyporheic extent.

We hypothesized that the steepest down-valley gradients would drive the largest hyporheic exchange, because such gradients would create the largest down-valley subsurface flows. This hypothesis is neither confirmed nor rejected based on our observations. In near-hillslope, stream area, and through-stream elements, we found both positive and negative relationships between hyporheic area and down-valley gradient. We posit that results are a function of the gradient interacting with sub-surface architecture and the heterogeneous distribution of hydrogeologic parameters (e.g., hydraulic conductivity) in the subsurface.
Next we consider spatial trends during each injection. For injection 1, the largest hyporheic area was observed at T1, while T2 and T3 had similar areas across all injections. Persistence of the solute tracer in the subsurface was largest at T3, the downstream-most transect, possibly due to longer temporal-scale flowpaths slowly transporting solute to T3 from upstream locations. During injection 2 the largest area was observed at T6, followed by T1 and T3, then T5, and finally the smallest area at T2. No spatial trend is apparent in hyporheic area, but tracer persistence is greatest in the upstream transects (T1-3). For injection 3, downstream transects (T5-6) have the largest areas (more than 10 m$^2$ greater than observations at T1-3). Persistence is relatively constant at T1-5, but is greatest at T6. Finally, during injection 4 the largest areas are again observed in the T5-6, while T2-3 have much smaller cross-sectional extents. Spatial trends in hyporheic persistence (as interpreted from late-time images like Figure 4-6C) are not clear; monitoring may have ended before all tracer had passed downstream transects T5 and T6.

All observations presented in Figure 4-8 are based on a threshold of 3% (i.e., only changes of at least 3% are interpreted for cross-sectional area. The effect of changing this threshold is demonstrated for T1, injection 1 in Figure 4-9. While increasing the threshold will always decrease the magnitude of hyporheic area interpreted from the ER images, the same general temporal trends are present for all thresholds greater than 1% for this transect. The analysis of areas presented in this manuscript is based on a constant threshold of 3%. The peak time of the injection occurs 54 hours after the start of the injection (6 hours after it was ended) for all thresholds greater than a 1% change.
4.4.6 Temporal trends in spatial moments

Solute plume movement in groundwater systems is commonly characterized by spatial moment analysis, which can be used to describe the location and spreading of a solute tracer [Aris, 1956; Freyberg, 1986]. While many examples exist using numerical simulations or field observations, relatively few exist where spatial moment analysis is applied to ER images to estimate saturated transport behavior. Singha and Gorelick [2005] demonstrated the interpretation of ER images using spatial moments in three dimensions. They showed that this analysis underrepresented mass, but accurately predicted solute plume centroid and slightly over-estimated temporal spreading. Hagarty et al. [2010] used spatial moments of electrical resistivity images to describe the center of mass and spatial variance of a solute tracer moving through hyporheic flowpaths near a headwater stream.

Cross-valley position of the centroid of the hyporheic zone

The cross-valley position of the centroid of the solute plume through time is plotted for each transect during each injection in Figure 4-11 (left-hand column). For T1 and T2 the plume is centered 3 to 4 m from the stream centerline, in the small alluvial deposit on the south side of the valley. In T3, where the alluvial deposit pinches out, the hyporheic flowpaths are centered more near the stream itself. In T5 and T6 the hyporheic plume is located within 2 m of the channel centerline, between the thalweg and a secondary channel (parallel to and on the south side of the fallen tree pictured in Figure 4-1). In all transects, the plume centroid is generally closer to the stream centerline with decreasing baseflow. Temporal trends show the plume moving slowly away from the stream centerline, then gradually back toward the stream for T1 and T2. For T3,
T5, and T6 the spatial hysteresis is not apparent in all injections, suggesting more complicated flow and transport relationships may exist. Trends are similar to those observed by Hagarty et al. [2010].

Vertical position of the centroid of the hyporheic zone

The vertical position of the centroid of the solute plume through time is plotted for each transect during each injection in Figure 4-11 (right-hand column). The centroid of the plume is generally within 2 m of the channel’s centerline elevation at each transect. In many cases, the plume centroid is located higher in elevation than the channel centerline. This is physically possible for two reasons: (1) the flowpaths leave the stream upgradient of the transect itself, and (2) the bed elevation at the channel centerline is the lowest elevation in the transect. In T1, T2, and T3 the plume is generally deeper in the subsurface during higher flows and shallower during lower flows. Temporal trends in plume location are similar for all injections in T1, T2, and T3. Similar spatial and temporal trends are also observed for injections 3 and 4 in T5 and T6, though the plume behavior during injection 2 has different behavior at these locations. Temporal trends for all transects show the general pattern of the plume appearing deeper in the subsurface and then gradually returning to a shallower location.

Spatial variance of the solute plume

Spatial variance of the solute plume was calculated for each transect at each timestep in both the cross-valley direction (Figure 4-12, left-hand column) and the vertical direction (Figure
4-11, right-hand column). In the cross-valley direction, spatial variance decreases during the injection period, then increases as the solute flushes from the subsurface. A similar decrease in spatial variance during the injection period was also observed in a field study by Hagarty et al. [2010]. In their study, this decrease in variance was attributed to solute entering hyporheic flowpaths focusing electrical current in a different area of the subsurface (i.e., the spatial variance due to background noise was large; tracer presence in hyporheic flowpaths creates a focused, high-change area that decreases spatial variance).

In the vertical direction, spatial variance increases slowly during the study at all transects under all flow conditions, as the penetration of salt-labeled stream water becomes deeper and deeper early in the injection. Spatial variance is generally largest for the injections during the highest flow conditions and decreases with decreasing flow. Thus, hyporheic exchange becomes generally shallower as baseflow recession occurs. This trend agrees with the shallower plume centroids found in the analysis of vertical centroid location.

Summary

Spatial moment analysis showed solute plumes with less lateral and vertical penetration into the aquifer during the tracer injections completed under lower stream flow rates including decreased spreading in the vertical direction with lower flows. While common conceptual models suggest that gradients toward the stream control hyporheic extent, we see mixed results in our study. We studied down- and cross-valley hydraulic gradients at both the valley-bottom averaged and finite-element scales. Results at these scales do not necessarily agree at each transect, and we found variability between transects. Studies in which greater spatial resolution
could be acquired would allow for greater understanding of controls on subsurface riparian hydrology. Such studies would test the spatial scale necessary for gradients to adequately characterize the relationship between hyporheic flowpaths and hydraulic gradients. The interaction of gradients and heterogeneous hydrogeologic parameter distributions is ultimately the control on hyporheic exchange.
Figure 4-11: Plume centroid in the cross-valley (left column) and vertical (right column) directions, for each transect. In all transects, the plume centroid is generally closer to the stream centerline with decreasing baseflow in both directions, indicating less penetration of the plume into the aquifer through baseflow recession. In T1 and T2 the lateral plume movement was centered through the small alluvial deposit, not near the confining valley wall. In the vertical, plume centroids above the stream centerline are not erroneous. Because flowpaths at a given transect entered the subsurface at an upstream (and up-gradient) location, the plume centroid at a transect can be higher in elevation than the thalweg elevation of the stream.
Figure 4-12: Temporal variability of cross-valley ($\sigma_{xx}^2$, left column) and vertical ($\sigma_{zz}^2$, right column) spatial variance of tracer plume in the subsurface. Spatial variance decreases during the early times in the injection and are likely due to the tracer’s presence in the subsurface focusing electrical current in the area of hyporheic flowpaths [Hagarty et al., 2010]. No consistent trends are observed under different flow conditions, though similar temporal trends were observed for each study. Spatial variance in the vertical direction increases slowly during each injection, with the highest values present for higher flow rates. The larger spatial variance of the plume suggests broader penetration of hyporheic flowpaths into the subsurface laterally, $\sigma_{xx}^2$ and vertically, $\sigma_{zz}^2$. 
Table 4-2: Linear regression slopes for finite-element gradient vs. area relationships. Linear regression analysis of gradient-area relationships provides insight into general patterns. Shaded cells (positive values) indicate relationships where (A) steeper down-valley gradients are related to increased hyporheic area, and (B) steeper cross-valley gradients toward the northeast valley wall are related to increased hyporheic area. Patterns show that steeper down-valley gradients are related to decreasing hyporheic extent for near-hillslope (NH) elements, and positively related to hyporheic area for observations in both stream-area (SA) and through-stream (TS) elements. Steeper cross-valley gradient toward the stream is associated with increased hyporheic area for both NH and TS elements, and with decreased hyporheic area for SA elements. There is variability within these general patterns for both down-valley and cross-valley gradient.

A. Down-valley Gradient vs. Area

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B. Cross-valley Gradient vs. Area

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4.5 Conclusions

*Bencala et al.* [2011] suggest that “a stream is a dynamic expression of local groundwater conditions, where exchanges of water between the catchment and the channel are continuously changing in response to heterogeneous temporal and spatial water table dynamics.” Several recent studies have begun to consider coupled groundwater and stream processes. *Wondzell* [2006] reported hyporheic exchange and extent as a function of valley setting and baseflow recession, using stream flow as a proxy for the hillslope-riparian-hyporheic-stream continuum. It has been proposed that hyporheic response to seasonal baseflow recession is a function of both in-channel and catchment controls operating across a range of spatial and temporal scales [*Wondzell et al.*, 2010]. Further studies demonstrate the role of hillslope-riparian-stream connections as a control on the magnitude and timing of both flow rates and solute transport observed at a catchment’s outlet [*Jencso et al.*, 2010]. Still, current conceptual models are only beginning to consider stream-catchment connections and their role in determining solute transport and transformation within a catchment [*Bencala et al.*, 2011].

A commonly held conceptual model in riparian hydrology suggests that increasing gradients from the catchment toward the stream should cause hyporheic zones to contract (i.e., stronger gradients from the stream to hyporheic zone are necessary to overcome this ambient condition) [*e.g., Hakenkamp et al.*, 1993; *Hynes*, 1983; *Meyer et al.*, 2008; *Palmer*, 1993; *Vervier et al.*, 1992; *White*, 1993]. Both numerical and field studies have confirmed this hyporheic contraction due to ambient gaining conditions [*e.g., Boano et al.*, 2008; *Cardenas and Wilson*, 2007a; *D’Angelo et al.*, 1993; *Harvey and Bencala*, 1993; *Storey et al.*, 2003; *Williams*, 1993;
Wondzell and Swanson, 1996; Wroblicky et al., 1998], yet none of these studies was able to observe solute transport along flowpaths in the field. Evidence in these studies was based on either spatially limited hyporheic data or reach-scale tracer studies, and does not consider the spatially complete distribution of flowpaths in the subsurface, nor those that may exist beyond the window of detection of in-stream tracers.

Our results suggest that this commonly held conceptual model of gradients between the stream and groundwater, be it cross-valley gradients toward the hillslope or vertical hydraulic gradients, as the dominant control on hyporheic exchange may be limited in their applicability, particularly in steep valley settings. We show evidence of increasing hyporheic area with both increasing and decreasing valley gradients toward the stream and with decreasing vertical hydraulic gradients from the stream to the subsurface. Based on our study, we conclude that characterizing gradients alone is not sufficient to predict hyporheic flowpath behavior; the interaction of these gradients with physical features (e.g., subsurface confining units, hydrogeologic parameter distributions) controls hyporheic exchange.

A clear contrast exists between the upper (steeper, valley constrained) and lower (flatter, unconstrained) sections of the study reach. Hyporheic flowpaths appear rapidly in the lower valley and are persistent through late-times. In the upper valley, the hyporheic network is constrained by the valley walls and bedrock that underlies the shallow alluvial deposit. In our study, valley constraint is a primary control on hyporheic exchange. The valley setting acts as a template upon which smaller exchange processes are controlled by local gradients, subsurface architecture (i.e., confining bedrock units, buried boulder and logs, etc.), and hydrogeologic parameters.

We expected that decreased cross-valley gradients during baseflow recession would have the least impact on hyporheic area in the constrained upper reach (i.e., T1-3), because control was
due to subsurface structure and hydrogeologic properties rather than gradients. Peak areas in the constrained upper reach (T1-3) had ranges of 6, 10, and 11 m², while the less-constrained lower reach (T5-6) has areas with ranges of 14 m² for both transects.

We expected to find that valley-constrained hyporheic flowpaths were less sensitive to changing hydrologic conditions, which was confirmed by our study. We expected that decreasing cross-valley gradients toward the stream would both contribute to larger hyporheic zones. Our study found that steeper cross-valley gradients were generally observed with larger hyporheic extent, in contrast to numerous studies predicting hyporheic contraction during periods with steeper gradients toward the stream. Finally, we expected that steeper down-valley gradients would be related to increased hyporheic area. We found no consistent trends with respect to down-valley gradients.

ER imaging provides an in-situ analysis of subsurface solute transport with high spatial and temporal resolution. This novel data set shows the movement of tracer through hyporheic pathways in larger spatial and temporal scales than could be characterized with in-stream tracers and/or monitoring wells alone. Our ability to collect this high-resolution data in replicate studies during baseflow recession allowed us to characterize both valley constraint and hydrologic gradients and controls on hyporheic exchange. The ability to consider both lateral and vertical constraint due to subsurface architecture allows consideration of this type of control at a much higher spatial resolution than past methods (i.e., transect-by-transect analysis, as compared to past studies at the valley scale).
4.6 Acknowledgements

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Chapter 5

Characterizing hyporheic transport processes – Interpretation of electrical geophysical data in coupled stream-hyporheic zone systems during solute tracer studies

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5.1 Abstract

Quantifying hyporheic solute dynamics has been limited by our ability to assess the magnitude and extent of stream interactions with multiple domains: mobile subsurface storage (MSS, e.g., freely flowing pore water) and immobile subsurface storage (ISS, e.g., poorly connected pore water). Stream tracer experiments coupled with solute transport modeling are frequently used to characterize lumped MSS and ISS dynamics but are limited by the ability to sample only “mobile” water and by window of detection issues. Here, we couple simulations of near-surface electrical resistivity (ER) methods with conservative solute transport to directly compare solute transport with ER interpretations and to determine the ability of ER to predict spatial and temporal trends of solute distribution and transport in stream-hyporheic systems. Results show that temporal moments from both ER and solute transport data agree for locations where advection is not the dominant solute transport process. Mean arrival time and variance are especially well-predicted by ER interpretation, providing the potential to estimate rate-limited mass transport (i.e. diffusive) parameters from these data in a distributed domain, substantially increasing our knowledge of the fate and transport of subsurface solutes.
5.2 Introduction

The movement of stream water into relatively slow-moving domains increases reach residence time of water and solutes [Morrice et al., 1997; Wondzell and Swanson, 1996]. By slowing the movement of water within a reach, the potential for processing of nutrients and other solutes is increased through longer exposure of stream solutes to microbial communities and biogeochemically active sediment [Boulton et al., 1998; Findlay, 1995; Valett et al., 1996]. Understanding the transport and fate of stream solutes is critical to understanding the temporal and spatial distribution of biogeochemical cycling associated with hyporheic exchange in streams.

Stream solute transport studies frequently yield concentration time series from in-stream and subsurface monitoring locations with long tails that are not explained by the in-stream processes of advection and dispersion alone. The movement of solute from the highly advective mobile domains into less-mobile domains with heterogeneous residence times is a common explanation of the observed tailing behavior in hydrological sciences. This type of tailing behavior has been observed in both flow through porous media [e.g., Haggerty and Gorelick, 1994; Harvey and Gorelick, 2000] and the exchange of stream water with the less-mobile hyporheic zone [e.g., Bencala and Walters, 1983; Haggerty et al., 2002]. Movement of tracer into the slower moving in-stream dead zones and into the hyporheic zone is typically considered responsible for this tailing behavior in stream-hyporheic studies. However, most stream solute studies overlook the range of mobile domains in the hyporheic zone, where domains range from highly advective and mobile (i.e., free pore water) to immobile (i.e., dead-end pore space, bound pore fluid) [Goltz and Roberts, 1986].

Tracer studies are frequently used to estimate interaction between mobile and immobile domains in both streams and porous media, yet there are inherent flaws in their formulation and
interpretation. Stream tracer studies rarely yield a full mass recovery, even with in-stream monitoring at substantial distances downstream, due to flux of labeled water into larger spatial and temporal scale flowpaths than can be observed with common tracer techniques [Payn et al., 2009; Ruehl et al., 2006]. The space and time scales that can be characterized in stream tracer experiments are commonly referred to as a ‘window of detection’ [Harvey and Wagner, 2000]. Payn et al. [2009] investigated, among other characteristics, mass loss along a mountain headwater stream, finding mass loss of over 10% through a 200 m reach that was not explained by groundwater flow out of the system. They identified flowpaths providing tracer loss to deeper, slower moving flowpaths as one potential fate for tracer. Typical solute monitoring methods (i.e. wells), however, are insufficient to characterize these low-concentration, spatially complex flowpaths due to their sensitivity to a limited spatial scale and because well samples only characterize the mobile subsurface (MSS) water in the immediate vicinity of a monitoring well. Consequently, concentration breakthrough curves are largely indicative of the mobile domain; despite this fact, they are used to infer characteristics of less mobile domains, including exchange rates and physical characteristics [Wagner and Harvey, 1997]. Because field observations of mobile domain concentration alone are used to quantify unobserved domains, the results are often not physically meaningful [Marion et al., 2003; Wondzell, 2006]. In stream-hyporheic studies, installation of invasive monitoring well networks provides discrete point observations provide only discrete, point assessment of the temporal or spatial distribution of tracer within the subsurface. Additionally, the stream-tracer approach is limited by a window of detection that is commonly focused on hyporheic flowpaths occurring at short temporal (e.g., seconds to hours) and spatial (millimeters to tens of meters) scales [Harvey and Wagner, 2000]. The relative scales of advection vs. diffusive processes (including transfer between mobile and less-mobile domains known as rate-limited mass transfer (RLMT)) suggest that stream tracer studies considering only
data in-stream, or in shallow monitoring wells, are biased toward short spatial and temporal flowpaths [Gooseff et al., 2003], despite a recognition that exchange occurs on larger scales. Indeed, Harvey et al. [Harvey et al., 1996] found that common transient storage modeling techniques [Bencala and Walters, 1983] provided a good fit for short, rapid exchange with gravel bar sediment but a poor fit for solute transport within slower MSS flowpaths.

Despite the recognition of a bi-continuum of MSS and immobile subsurface (ISS) zones, stream transport modeling has often lumped all immobile domains into one set of physical and chemical coefficients. Only recently has stream solute transport modeling begun to address multiple scales of storage [e.g., surface vs. hyporheic transient storage; Briggs et al., 2008], and no models have yet addressed solute transport processes in the hyporheic-bicontinuum, i.e. simulation of both MSS and ISS. To calibrate more complex model formulations, additional field observations are required. In the past, multiple storage-zone models have relied on observations in monitoring wells and direct sampling from in-stream dead zones to provide the additional data required. A spatially distributed characterization of hyporheic solute transport would require more monitoring wells than can feasibly be installed and monitored during a typical tracer study.

To overcome the data limitations used in solute transport monitoring and modeling, we explore the use of electrical resistivity (ER) measurements to quantify hyporheic solute transport processes. ER methods pass direct current (or low-frequency alternating current) through a soil-water matrix and can be used to estimate the spatial and temporal distribution of electrical resistivity. ER has been used to track saline tracers across a range of scales, including changes in hyporheic salinity for a tidal creek [Acworth and Dasey, 2003], aquifer storage and recovery tests [Singha et al., 2007], and transport through porous media in both laboratory [Binley et al., 1996a] and field-scale two- and three-dimensional experiments monitoring tracer (frequently saline) plumes [Cassiani et al., 2006; Kemna et al., 2002; Slater et al., 1997; Slater et al., 2000; Slater et
Near streams, these data have been successfully used to generate tomograms, or maps of subsurface electrical conductivity, which highlight the spatial distribution of the stream tracer in the subsurface [Ward et al., 2010b; Chapter 2] and to investigate subsurface architecture [Crook et al., 2008]. Despite its demonstrated capabilities, ER is limited by our difficulty in defining the support volume of the measurement [Blöschl and Sivapalan, 1995]. While ER data provides spatially integrated measurements over broad spatial areas with the potential to do so at a high temporal frequency, determining the depth of penetration or the volume averaged by a single measurement is difficult and dependent on the electrode geometry and electrical resistivity of the subsurface [Daily and Ramirez, 1995; Vasco et al., 1997].

Numerical studies by Singha et al. [2008] suggest the potential of ER for hyporheic assessment using conservative, electrically conductive tracers is great. The studies suggest that ER results are useful for calculation of solute transport parameters, including the immobile fractions of pore space and the RLMT coefficient that controls the rate of exchange between the domains. Singha et al. [2008] worked with inverted tomographic results, which are complicated by the assumptions inherent within inversion [Day-Lewis et al., 2005] and are limited by assumptions inherent in the solute transport model [i.e., OTIS; Runkel, 1998]. Limiting assumptions include a well-mixed storage zone with no downstream transport in the subsurface. Here, we seek to address these limitations by working with non-inverted electrical responses and utilizing a single, coupled model to simulate solute transport in three dimensions. This more complex model produces ER data more representative of field studies and thus better enables interpretation of electrical response to assess the distribution of solute within multiple subsurface domains. Without an understanding of ER sensitivity to multiple processes, occurring in a dynamic system, we cannot reliably interpret field data. Our numerical solute transport model accounts for RLMT between MSS and ISS domains while advection and dispersion drive
exchange of the solute between the stream and MSS domains. The model formulation allows us to explore the contributions of solute presence or absence in the highly mobile stream domain, in combination with both mobile and immobile domains within the hyporheic zone to aid in interpretation of ER data.

To the best of our knowledge, the interpretation of ER data (as opposed to tomograms) during a stream-hyporheic tracer study has not been attempted. Here, we focus on the potential to use temporal moment analysis to discern the dominant solute transport processes within a stream-hyporheic system. This study provides a framework for collection and interpretation of ER data during these studies and demonstrates a novel application of ER. We expect that the presence of electrically conductive tracer in different domains (mobile and immobile) and distributed in the subsurface by spatially varied transport processes will exhibit differing responses in the ER data. The objectives of this study are to use a numerical model (1) to investigate the ability of an ER surface array to predict spatial trends and temporal moments of solute concentrations during a stream-hyporheic tracer study; and (2) to use ER and solute transport data to identify the distribution of dominant solute transport processes in the subsurface.

5.3 Numerical simulation

5.3.1 Numerical model formulation

We used COMSOL [COMSOL, 2008], a fully coupled finite-element model, to simulate solute transport within and between domains, in addition to simulation of ER measurements. Model geometry is based on an idealized 3rd order pool-riffle-step morphology [Gooseff et al., 2006], with a channel two meters in width (Figure 5-1). A series of geomorphologic pool-riffle-
step sequences was modeled to drive hyporheic exchange within the model. ER measurements focused on the third riffle from the upstream boundary (Figure 5-1) to avoid model boundary effects and condition inflow and outflow in the area of interest. A floodplain on either side of the stream was modeled, extending 10 m from the stream surface. An active substrate depth of 5 m was included above a no-flow boundary. A stream flow rate of 0.17 m$^3$/s was simulated in the channel. Groundwater flow was modeled with fixed head at the upstream and downstream model extents corresponding to the elevation of the stream surface. Hydraulic conductivity in the entire subsurface was assigned as homogeneous and isotropic, at $1.0 \times 10^{-3}$ m/s. The model domain was bounded by no-flux conditions on the top, bottom, and lateral extents, such that all flow entered the upstream and exited the downstream ends of the domain with no lateral inflow or outflow. No unsaturated zone was included in the model domain. A total of 123,808 finite elements were simulated in the model domain with average element to volume ratios of $2.50 \times 10^{-4}$ for the flow and solute transport domains (Regions “Subsurface” and “Stream” in Figure 5-1B) and $1.68 \times 10^{-5}$ for the outer domain added to minimize edge effects for the ER model (Region “ER Only” in Figure 5-1B).

Solute transport in the stream and MSS domains is simulated by the advection-dispersion equation. Stream flow was modeled as well mixed in the vertical and transverse dimensions by assigning a dispersivity of 1 m in these directions, yielding essentially one-dimensional flow with no in-stream transient storage. Longitudinal dispersion in the channel was set at $3.45 \times 10^{-7}$ m$^2$/s. In the subsurface, an immobile domain was partitioned from the mobile porosity as 50% of the 35% total porosity and assigned a diffusion coefficient of $1.0 \times 10^{-4}$ m$^2$/s. Parameters were selected to be representative of a headwater mountain stream with a subsurface of mixed gravel, sand, and silt. First-order RLMT was parameterized using the following equations for exchange in the solute transport advection-dispersion equation:
Figure 5.1: (A) Profile of model showing steady-state velocity in the subsurface. Model results yield similar velocity profiles to two-dimensional results published by Gooseff and others [37]. A no-flow boundary is enforced at the subsurface limit, with velocity parallel to the bed as shown by the bottom arrow in each column. (B) Isometric view of model geometry. Profile view to show pool-riffle-step geomorphology and highlight the velocity patterns observed in the subsurface.
where $C_{\text{MSS}}$ and $C_{\text{ISS}}$ are the MSS and ISS concentrations [kg/m$^3$], $\alpha$ is the exchange coefficient [1/d], and $\theta_{\text{MSS}}$ and $\theta_{\text{ISS}}$ the MSS and ISS porosities [unitless]. The exchange coefficient was set as $\alpha = 1.0 \times 10^{-3}$ d$^{-1}$ (within the range of values from Singha et al. [2008]). For this study, we conceptualize the subsurface as consisting of either mobile or immobile space, with a single coefficient describing RLMT. In field settings, heterogeneity and scale effects of the exchange coefficient should be considered. Despite this simplification, these models provide insight into hyporheic transport processes that are overlooked by many stream tracer studies.

5.3.2 Simulated tracer injection

To assess the sensitivity of electrical geophysics to the presence of the electrically conductive tracer in the various domains, simulation of a conservative, electrically conductive tracer injection was completed. As a baseline for comparison, deviation is compared to modeled electrical results for the entire system spun-up to a uniform, steady-state background concentration of 1,000 mg/L. A solute injection was modeled as a constant rate injection at the upstream end of the model domain, where stream concentration was increased from 1,000 to 4,000 mg/L for a 3-hour constant rate injection. Density affects due to the increased solute concentration were assumed negligible. A total simulation of one week was completed to capture
the response of the tracer slowly flushing from the subsurface domain. Model data were extracted at 60 s intervals for the first 24 hours of simulation time to resolve rapid changes in concentration through time and at 360 s intervals for the remaining 144 hours of the simulation. ER modeling used the Conductive DC Media package in COMSOL. The electrical model was used to simulate the injection of direct current and measurement of voltage at the electrodes. A 10-m domain was added beyond the flow and solute transport model domains in all directions to avoid boundary effects in the electrical field (Figure 5-1). This added domain remained at initial (background) levels of tracer concentration throughout the simulation. It was added to decrease boundary effects on the electrical model, but assumed to remain at a static concentration to maintain reasonable computational efficiency for the flow and solute transport models.

Electrodes were modeled as points located 0.1 m below the model surface, distributed across the model domain (Figure 5-2A). Points were located below the surface to represent the shallow penetration achieved by a typical surface array, where electrodes are in contact with the ground at shallow subsurface depths. Quadripole data (i.e., measurements including 4 electrodes as a single potential pair and single current pair) were collected using a modified Schlumberger array (i.e., current driven from outer-most electrodes, but potential pair not necessarily centered or symmetric within the potential pair). Electrodes were located at X = 0 (the stream centerline), and X = ±1.1, ±2.1,…,±11.1 m (floodplain on either side of the stream). Current was driven from electrodes at X = ±11.1 m for all measurements. All possible potential pairs were collected. Modeling the electrodes as line or three-dimensional sources from the ground surface to the penetration depth was not attempted because it would have increased computational demand. Because the model’s top surface was a no-flux boundary, we felt placement at this shallow depth would provide more representative current and potential readings, independent of the concentration at the no-flux boundary.
Figure 5-2: (A) Stream and electrode locations within model domain. Tracer concentration at background levels (1,000 mg/L, uniform) is also shown. (B) Arrival of the tracer is first seen in the stream. (C-D) Diffusion of the tracer into the subsurface, and advection in the hyporheic zone below the stream expand the pulse of tracer in the subsurface. (E-F) Tracer flushes from the stream, leaving a large tracer pulse in the hyporheic zone. (G) A second pulse of tracer appears in the hyporheic zone, transported from temporary hyporheic storage in an upstream pool-riffle-step sequence. (H-J) The advection-dominated hyporheic zone flushes of tracer, leaving a “ring” of tracer in areas dominated by diffusive and rate-limited mass transfer processes.
5.4 Theory

5.4.1 Electrical resistivity & tracer concentration

Bulk electrical conductivity for the soil-water matrix of a single subsurface domain was calculated using Archie’s law [sensu Archie, 1942]. Application of Archie’s Law includes the underlying assumptions that there is little to no clay present, and that surface conductivity is substantially less than conductivity through the fluid. Both assumptions are valid for the simulation we completed. For the model domain remaining at background concentration, Archie’s Law is applied as:

\[ \sigma_b = a \sigma_f \theta^q \]  \hspace{1cm} (3)

where \( \sigma_b \) is bulk electrical conductivity [S/m], \( \sigma_f \) is fluid conductivity [S/m], \( \theta \) is total porosity [unitless], \( a \) is a fitting parameter related to tortuosity [unitless], and \( q \) is a fitting parameter related to cementation of the matrix [unitless]. Values of \( a = 1 \) and \( q = 1.3 \) are used as representative values to parameterize the finite element cells, within the ranges commonly accepted in the literature [Keller and Frischknecht, 1966]. Bulk apparent electrical conductivity in the porous media is calculated using a bicontinuum formulation of Archie’s law (Equation 4, 21, 40), which incorporates both the MSS and ISS domains:

\[ \sigma_b = a \left( \theta_{\text{MSS}} + \theta_{\text{ISS}} \right)^{q-1} \left( \theta_{\text{MSS}} \sigma_{\text{MSS}} + \theta_{\text{ISS}} \sigma_{\text{ISS}} \right) \]  \hspace{1cm} (4)

where \( \theta_{\text{MSS}} \) and \( \theta_{\text{ISS}} \) represent the mobile and immobile porosities [--], and \( \sigma_{\text{MSS}} \) and \( \sigma_{\text{ISS}} \) the mobile and immobile electrical conductivities [S/m].
5.4.2 Interpretation of electrical measurements

To assess sensitivity of the measurements to the geometry of the ER measurements, 342 quadripoles were simulated. Electrical geophysical methods rely upon Ohm’s Law \((R=V/I)\) to calculate the electrical resistance of the subsurface, where \(V\) is the potential difference measured across the potential pair \([V]\), \(I\) is the current driven through a circuit \([A]\), and \(R\) is the resistance of the circuit \([\Omega]\). A geometric factor, \(K\) \([m]\), is applied to convert between resistance (a property dependent upon the geometry of the electrodes used in a particular measurement) to resistivity (an intrinsic property, the reciprocal of electrical conductivity, which is independent of measurement geometry). Assuming a homogeneous half-space, the geometric factor can be calculated as:

\[
K = \frac{4\pi}{\frac{1}{AM} + \frac{1}{AN} - \frac{1}{AM_{image}} - \frac{1}{AN_{image}} - \frac{1}{BM} - \frac{1}{BN} + \frac{1}{BM_{image}} + \frac{1}{BN_{image}}}
\]

where \(A\) and \(B\) represent the current pair, \(M\) and \(N\) the potential pair, and \(AM, AN, BM,\) and \(BN\) are the distances between electrode pairs. The geometric factor is used with Ohm’s Law to calculate apparent resistivity \((\rho)\), \([\Omega m]\) as \(\rho = KR\). To work in more familiar units of conductivity (which is directly proportional to solute concentration), we convert apparent resistivity to apparent conductivity \((\sigma)\), where \(\sigma = 1/\rho\).

Using the known quadripole geometry and electrical current, a bulk apparent electrical conductivity was calculated for each measurement through time using the Ohm’s Law and a formulation of Archie’s Law (see the final two paragraphs of section 3.3). Results were converted from bulk apparent electrical conductivity to percent change from background apparent conductivity \((\sigma_{pc})\) to accentuate changes in the electrical signal as follows:

\[
\sigma_{pc} = \frac{\sigma(t) - \sigma(t = 0)}{\sigma(t = 0)}
\]

(6)
Fluid concentration was converted to electrical conductivity using the relationship of 2 mg/L = 1 µS/cm [Keller and Frischknecht, 1966].

5.4.3 Temporal Moment Analysis

The temporal patterns observed inform interpretation of the dominant transport processes. Temporal moments have been used to describe characteristics of solute breakthrough curves including mass arrival, mean arrival time, variance, skewness, and kurtosis of concentration observations through time [Gupta and Cvetkovic, 2000; Harvey and Gorelick, 1995]. Temporal moments have been used to drive routing models for streams with transient storage [Schmid, 2003], characterize mixing and dilution processes [Cirpka and Kitanidis, 2000; Gupta and Cvetkovic, 2000], and predict solute transport parameters related to RLMT [Day-Lewis and Singha, 2008; Singha et al., 2008]. For the remainder of this paper, superscripts “ST” and “ER” are used to denote properties derived from solute transport and electrical resistivity data, respectively.

To highlight the temporal trends in the breakthrough curves observed in both solute transport and ER data we consider analysis of the temporal moments of each data set. An \( n \)th order temporal moment \( (M_n) \) is calculated by:

\[
M_n = \int_0^\infty t^n c(t) dt
\]  

(7)

where \( t \) is time, \( n \) is the order of the central moment, and \( c(t) \) is fluid concentration as a function of time [kg/m³]. In subsequent equations, \( M_0 \) refers to the zeroth temporal moment, \( M_1 \) to the first temporal moment, etc. Here, we consider the zeroth through fourth moments, because they are readily interpreted as descriptions of the physical breakthrough curves observed and
predicted. Physically, the zeroth moment may be interpreted as the total tracer mass having passed by an observation point, after subtracting the background concentration. The first moment, normalized to total mass, is physically interpreted as the mean arrival time of the injected solute at the observation point, calculated as:

\[ \bar{t} = \frac{M_1}{M_0} \]  

(8)

The variance of the pulse describes the spread of the breakthrough curve, which may be due to differences in advective flowpaths, as well as diffusive and RLMT processes. Variance is related to the second and lower temporal moments, and is calculated by the following equation:

\[ \mu_2 = \sigma^2 = \frac{M_2}{M_0} - \left( \frac{M_1}{M_0} \right)^2 \]  

(9)

The skewness of the distribution describes the asymmetry of the breakthrough curve; in solute transport studies we expect a positive skewness to result from both diffusive and RLMT processes. A more positive skewness indicates an observed breakthrough curve with a larger degree of tailing-behavior. Skewness is related to the third and lower temporal moments and is calculated by:

\[ \mu_3 = \frac{M_3}{M_0} - 3\mu_2 \frac{M_1}{M_0} - \left( \frac{M_1}{M_0} \right)^3 \]  

(10)
The kurtosis describes the “peakedness” of the distribution. High values indicate a
distribution with higher central peaks and fewer extreme events. Kurtosis is related to the fourth
and lower temporal moments and is calculated by:

\[
\mu_4 = \frac{M_4}{M_0} - 4\mu_3 \frac{M_1}{M_0} - 6\mu_2 \left( \frac{M_1}{M_0} \right)^2 - \left( \frac{M_1}{M_0} \right)^4
\]  

(11)

The same temporal moment equation and relevant parameters can be applied to ER data,
using Archie’s law to convert from apparent electrical conductivity to fluid conductivity. This
empirical relationship is applied at initial conditions to establish a tortuosity fitting factor (a) for
each quadripole. The forward model was run with \( a = 1 \) within each finite element. Because
measurements are now at a substantially larger scale, it is appropriate to calculate a new fitting
factor. Variability in Archie’s law parameters across spatial and temporal resolution of
measurements may yield misleading estimates of solute concentration [Singha and Gorelick,
2006a]. Here, we use observations during pre-injection conditions (t = 0 s) to calculate a fitting
factor, \( a \), that is appropriate for the spatial scale of each measurement and which yields agreement
between known fluid concentration and bulk apparent electrical conductivity. We calculate an
effective \( a \) for each quadripole (rather than use the value \( a = 1 \) that parameterized the finite
element model) to enforce agreement between the ER and solute transport models at the initial
conditions. At t = 0 s, MSS and ISS concentrations are assumed equal. Assuming constant
porosity and cementation exponents, \( a \) is calculated for each quadripole as:

\[
a = \frac{\sigma_{\text{bulk}}(t = 0)}{\sigma_{\text{fluid}}(t = 0)\theta^q}
\]  

(12)
Given apparent bulk conductivity, fluid conductivity in the MSS and ISS domains can be predicted if the distributions of tracer within the MSS and ISS domains are known. In practice the MSS concentration is measured using shallow monitoring wells. The ISS concentration, however, cannot be measured using existing field techniques. Thus, we interpret the data using two cases that define end-members for solute transport. We define Case 1 as the end-member where ISS concentration is equal to MSS concentration (effectively assuming $\alpha = \infty$, or $\theta_{\text{ISS}} = 0$). Equation 3 is applied to solve for fluid conductivity ($\sigma_{\text{MSS}}$) at each time step for Case 1. We define Case 2 as the end-member where no tracer enters the ISS domain (effectively assuming $\alpha = 0$, thus $\sigma_{\text{ISS}} = \sigma_{\text{Background}}$). Equation 4 is used to solve for fluid conductivity ($\sigma_{\text{MSS}}$) at each time step for Case 2. The case where there is tracer only in the ISS domain is not considered given the short temporal scale of the typical stream tracer study. The case where more concentration is in the ISS domain would only occur for experiments with very low RLMT exchange coefficients, which would allow the ISS to fill with tracer during a very long tracer injection, and the solute in that domain would then be slowly transferred back to the MSS domain at timescales much greater than advective flushing of the MSS domain.

To effectively test the ability of ER data to predict solute transport, zeroth through fourth order temporal moments and the associated physical characteristics were calculated for both solute transport and ER simulations. For solute data, moments were based on mobile-domain observations on a 10 cm grid for the stream and subsurface domain at the geophysical transect. For ER data, bulk electrical conductivity at all potential pairs for the geophysical transect was used to calculate temporal moments.
5.5 Results & discussion

5.5.1 Flow model

Spatial patterns of hyporheic exchange driven by in-stream geomorphologic features produce the characteristic upwelling and downwelling expected from the streambed profile. Results generally match the two-dimensional patterns published by Gooseff et al. [2006]. Velocity along the stream centerline (Figure 5-1A) exhibits downwelling and upwelling due to variations in hydrostatic pressure at the streambed.

5.5.2 Solute transport model

Time-lapse images of solute concentration in the subsurface (Figure 5-2) suggest that different transport processes control different regions of the subsurface domain. Advection dominates in the stream and very near-stream subsurface, as tracer is rapidly transported into and out of this region. Beyond this advection-dominated region, however, we observe the effects of RLMT and diffusion controlling solute behavior. Moving outward from the stream, a region exists in which advective flowpaths still flush the tracer from the region below the stream within 5 hours of the injection ending (Figure 5-2I). Beyond this region of the subsurface, solute exhibits a longer residence time suggesting that RLMT has stored a portion of the tracer temporarily. Continuing outward, a “ring” of tracer has diffused into the hyporheic zone and exchanged with immobile domains. Even when the portion of the hyporheic zone below the stream has returned to background or undetectably low concentrations of tracer, a substantial, distributed mass of tracer continues to diffuse (Figure 5-
2H-I). As the near-stream subsurface concentration is flushed to lower levels than the diffusive “ring”, propagation of the ring into the floodplain stops. Diffusion in both directions continues to control behavior in the subsurface, and tracer is slowly advected down gradient at very low concentrations.

5.5.3 Process-based signatures in the ER and solute models

Visual inspection of both the solute transport and ER data through time identified four distinct temporal “signatures” collected (representative signatures in Figure 5-3, distribution of signatures in Figure 5-4A). These signatures describe the shape of the breakthrough curve, and are indicative of solute transport processes at a given location. The four signatures identified are present in both the ER and solute transport data, and may be classified as:

1 - Background signature. Background signatures are those produced by quadripoles exhibiting minimal response to the tracer. The identifying characteristic of background signatures is their lack of response to the tracer. In this study, a threshold of less than 5% change in electrical conductivity was used to classify background signatures. For field data sets, this threshold should be adjusted to identify signatures that are bounded by the noise in the data set. Background signatures were generally located where potential electrode spacing was greater than 9 m. This is expected based on the solute transport model; these measurements are averaging over very large support volumes that are largely dominated by elements where the tracer is not present. For potential pair spacing less than 9-m, background signatures existed for measurements with larger spatial distances between the stream and potential electrodes (Figure 5-5).
2 - Stream signature. The characteristic stream signature is a response that is highly correlated to the simulated in-stream breakthrough curve. Specifically, the stream signature exhibits a rapid response to in-stream conductivity and zero or near-zero gradient during the in-stream plateau. We characterized stream signatures based on the observed data and not on the physical location of the measurement. Stream signatures generally exist for electrode pairs located very close to the stream with small potential pair spacing (Figure 5-5).

3 - Mobile subsurface (MSS) signature. The mobile subsurface signature is identified by a positive correlation with MSS concentration observed at 50 cm below the streambed, at the stream centerline (typical placement for a monitoring well in field studies). The signal responds to the presence of tracer in advective subsurface flow paths by rising initially and then falling when the advective flowpath being sampled is flushed by unlabeled stream water. MSS signatures generally exist for potential pairs centered within 3-m of the stream and across a range of potential pair spacing (Figure 5-5).

4 - Diffusive signature. Interpretation of ER data identified a fourth characteristic signature, identified as a diffusive signature. The diffusive signature response to the spatially distributed, low concentration tracer that remains after flowpaths dominated by advection have been flushed of the tracer. The diffusive flowpath exhibits peak concentration substantially later in time than the stream or mobile subsurface signature. Diffusive signatures respond to the tracer mass that has been transported away from more advective, near-stream flowpaths and is primarily transported by diffusion, both laterally and vertically, away from the stream. Diffusive signatures generally exist for potential pairs with small spacing, located physically outside of the MSS signature observations (Figure 5-5).

Representative signature types for simulated solute transport are readily calculated for the entire subsurface domain (Figure 5-4A) and lend to interpretation of solute transport results. We
interpret the area defined by the MSS signature as the region typically characterized as the hyporheic zone. Beyond this MSS signature, RLMT slows the movement of the tracer down gradient. This process is not noted in many field studies due to (1) lack of detection ability (i.e., extremely low concentrations are not detectable with field instruments), (2) lack of sufficient monitoring points to characterize this region, (3) time limitations to intensively monitor subsurface wells, and (4) difficulty partitioning RLMT from substrate heterogeneity.

**Figure 5-3:** Representative breakthrough curves for the four identified signatures. The curves shown are representative of the characteristic temporal signatures identified as (1) stream signatures (flat plateau during injection), (2) mobile subsurface signatures (positive correlation to solute transport observations at 50 cm below the streambed), (3) diffusive signatures (negative correlation with solute transport observations at 50 cm below the streambed), and (4) background signatures (overall change of less than 5%). Data plotted are from the solute transport model, though the same signatures types are also identified in electrical resistivity data. The background signatures remains at 0% and is plotted coincident with the X-axis.
5.5.4 Spatial trends of solute transport model

Spatial distribution of temporal moments based on solute transport data is explored to understand the rates of tracer movement in the subsurface and controlling processes in each location (Figure 5-4A). Total mass ($\mu_{0}^{ST}$) trends are as expected (Figure 5-4B), with the largest mass passing through the stream channel itself. The darker area of the plot below the stream channel is due to advection of tracer from the stream into the subsurface. Total mass decreases moving away from this highly advective zone (stream and MSS signatures) due to diffusion of the tracer. Mean arrival time ($\mu_{1}^{ST}$) is highest near the stream and decreases away from the stream (Figure 5-4C).

For solute transport data, increased distance from the stream decreases total mass (Figure 5-4B), delays mean arrival time (Figure 5-4C), and increases variance ($\mu_{2}^{ST}$) (Figure 5-4D). MSS signatures are spatially aligned with the areas of the largest skew ($\mu_{3}^{ST}$), suggesting that these signals are most affected by RLMT and diffusion in the subsurface. Stream signatures exhibit a low skew, suggesting that advection overwhelms diffusive and exchange processes that might lead to tailing behavior. Diffusive signatures exhibit a low skew, as they are largely symmetric because their controlling process of diffusion leads to a nearly symmetric breakthrough curve.

Mean arrival time (Figure 5-4C) exhibits similar trends to total mass, with mean arrival time in the stream and advective subsurface occurring rapidly. Moving away from the stream, the center of mass arrival is delayed due to the slower diffusive processes transporting tracer into the subsurface. Delayed arrival in the subsurface may also be explained by RLMT slowing the movement of tracer as it diffuses away from the stream. Increasing variance in the breakthrough curve farther from the stream suggests that diffusive processes are dominating in these regions of the hyporheic zone. The highly advective stream and subsurface locations exhibit the lowest
variance, because advection moves a majority of the tracer through these areas with little diffusive or RLMT influence.

**Figure 5-4:** (A) Spatial distribution of signatures and (B-F) temporal moment interpretations for solute transport in the MSS domain, based on the solute transport simulation. The advection-dominated hyporheic zone below the streambed exhibits similar behavior to the stream itself based on moment analysis of solute data. Outside of this highly advective zone, diffusion and rate-limited mass transfer slow the movement of tracer, yielding a “ring” of highly skewed measurements with substantial tailing behavior. (A) Signature type exhibits a pattern similar to variance and skew, suggesting diffusive and rate-limited mass transfer processes are dominant in these regions. (B) Total mass is greatest near the stream and in the advection-dominated region of the hyporheic zone. (C) Average arrival time increases away from the stream. Note the rapid arrival time in the highly advective portion of the hyporheic zone, located below the stream bed. (D) Variance is greatest in a “ring” in the subsurface, which is generally bounded on the interior by the highly advective hyporheic zone. (E) The largest values of skew occur where the primary transport mechanism transitions from advection of diffusion and rate-limited mass transfer. (F) Kurtosis exhibits a similar trend as skew, suggesting that the higher values in the “ring” are due to a more distributed, tailed distribution. Lower kurtosis values indicate a distribution with a more flat plateau near the mean.
Skewness of the solute transport is an indicator of the relative controls of advection vs. diffusion and transient storage. Transport with only advection and dispersion would exhibit a distribution with minor skew (data are slightly non-gaussian due to diffusion opposing advection for some particles). By including diffusive and RLMT processes, greater positive skew is expected in the data set. Skewness (Figure 5-4D) is highly correlated with late-time solute behavior (Figure 5-2H-I). The same “ring” pattern exists, suggesting that diffusion into the subsurface and RLMT, which slow the movement of tracer past the transect, are responsible for the tailing behavior observed in the MSS and diffusive signatures.

Physically, we can interpret higher kurtosis ($\mu_4^{ST}$) as an indicator of a single peak of tracer in a given cell. Lower kurtosis values inside the “ring” of higher values (Figure 5-4F) suggest that the variance in the observation is controlled by either multiple peaks or more heavily tailed observations. From a physical-process standpoint, we expect that as tracer moves downstream in our pool-riffle sequence that pulses of tracer will leave the highly advective domain and return at later times. The low kurtosis of the near-stream aquifer indicates that the region receives a relatively more varied loading that the surrounding, higher kurtosis region.

5.5.5 Spatial trends of ER model

Analysis of ER response to the tracer (Figure 5-5, presented as maximum percent change from background levels) suggest that maximum electrical response is found for quadripoles that include (1) a more closely spaced potential pair (Figure 5-5A), (2) a closer proximity of electrodes to the stream (Figure 5-5B), and (3) a measurement that is centered on the stream (Figure 5-5C). Also, analysis of maximum response by signature reveals that stream signatures
are generally observed from quadripoles with small potential pair spacing near the stream. MSS signatures are present for potential pairs still centered near the stream, but with larger spacing, and diffusive signatures are found for quadripoles with still larger potential pair spacing, and located farther from the stream.

**Figure 5-5:** Spatial trends in the response of quadripoles to the stream solute tracer. The largest responses were stream signatures, located with potential pairs tightly spaced near the stream. Increasing spacing and distance from the stream centerline yields mobile subsurface and diffusive signatures, trending away from the stream centerline. The largest response of electrical resistivity to solute presence is for quadripoles with (A) closely spaced potential pairs, (B) potential electrodes located near the stream, and (C) potential pairs centered on the stream.
The spatial trends in ER signatures are indicative of the processes that dominate solute behavior within the domain. Based on the simulation completed, we interpret the controlling processes for each signature. Stream signatures indicate quadripoles are sensitive not only to in-stream processes, but rapid diffusion of solute from the stream into the near-stream. The mobile subsurface signature characterizes small spatial and temporal scale advective flowpaths (typically flowpaths within 2.5 m of the stream with mean arrival times less than 8 hours). Correlation with solute transport data suggests that these quadripoles respond mostly to advection of tracer along short subsurface flowpaths. Diffusive signatures identify a mass transport process that is dominated by diffusion away from the shorter spatial and temporal scale advective flowpath near the stream.

Temporal moments of ER data, for both Case 1 and Case 2, display spatial trends due to the distribution of tracer in the subsurface (Figure 5-6, scatter plot). With increasing distance from the stream, total mass (\( \mu_0^{ER} \)) decreases, while average arrival time (\( \mu_1^{ER} \)) and variance (\( \mu_2^{ER} \)) increase. Skewness (\( \mu_3^{ER} \)) and kurtosis (\( \mu_4^{ER} \)) exhibit trends that peak near 3 m and 5 m from the stream, respectively. These results are expected, as the tracer diffuses slowly away from the stream and highly advective hyporheic flowpaths. Spacing of the potential electrodes directly influences the observation, with closer spacing generally resulting in increased total mass, earlier average arrival times, lower variance, higher skew, and lower kurtosis.

The zeroth temporal moment (total mass) predicted by both Case 1 and Case 2 were substantially lower than the mass flux from the solute transport model (Figure 5-6A, 5-6F). Case 2 (ISS = \( \sigma_{\text{Background}} \)) was more representative for this model, given the low exchange coefficient defined. This is not necessarily the case for all parameter sets; the analysis using both end
members provides insight regarding the behavior of the system. If Case 1 provides a better fit to field observations, a more rapid exchange between MSS and ISS can be inferred.

5.5.6 Comparison of ER and solute transport models

Temporal moments based on ER data exhibit reasonable approximations of temporal moments based on solute transport models (Figure 5-6). The interpretations of Case 1 and Case 2 differ in their prediction of total solute mass (Figure 5-6A and 5-6F), but the properties of the distribution based on higher-order temporal moments are identical. This is expected, given that these quantities are all normalized by the zeroth temporal moment. In addition to agreement of general trends, numerical results are reasonably close between solute transport and ER-derived temporal moments, especially for mean arrival time ($\mu_1$) and variance ($\mu_2$). ER measurements integrate across a three-dimensional support volume in the subsurface, which is itself a function of time as electrically conductive solute moves through the subsurface. While true assignment of the measurement to a single point in space and time is not physically possible, we explore patterns in temporal moments by assigning the data a location at the spatial center of potential pair and comparing ER predictions to values derived from vertically averaged solute transport data. Both ER and solute transport results exhibit decreasing total mass, increasing mean arrival time, and increasing variance as observations move spatially farther away from the stream centerline. ER and solute derived parameters exhibit similar patterns in skew and kurtosis, which are similar to the “ring” pattern of tracer observed in the solute transport time lapse images (Figure 5-2), in that a local maximum exists with lesser values laterally in either direction. Skew and kurtosis peak at approximately 3 m and 4 m from the channel, respectively. The alignment of these thresholds with the general solute transport patterns (Figure 5-2) and representative
Figure 5-6: Spatial trends in temporal moments for both electrical resistivity (ER) data (points) and solute transport (line, vertical average). Point shading represents the spacing of the potential pair (current pair was fixed at ±11.1 m). Darker points indicate electrodes that are closer to one another. A minimum spacing of 1.0 m and maximum of 18.2 m are represented on the plot as the darkest and lightest points, respectively. ER data are assigned a point in space at the center of the potential pair. We note that assigning apparent resistivity data, which is based on a volumetric measure, to a single point is a flawed assumption, but one commonly used within the electrical geophysics community to explore spatial patterns in uninverted data. Because the model is symmetric, only one site of the floodplain is shown for ease of interpretation. The ability to compare the spatial location of moment data provides insight regarding the locations in which ER can be used to accurately estimate temporal moments of solute transport. Cases 1 and 2 represent end-members of tracer distributions between the MSS and ISS domains. (A & F) Total mass is under predicted by ER across the entire domain, with the largest error near the stream. (B & G) ER accurately bounds mean arrival time near the stream within a wide range, and generally predicted later arrival time than observed farther from the stream. More closely spaced potential pairs over-predict solute arrival; pairs with greater spacing over-predict mean arrival time. (C & H) Observed variance is bounded by the ER calculated variance. (D & I) Skew of the observed solute transport data is generally well predicted by ER across the entire domain. Both solute transport and ER moment analysis predict a peak in skew, identifying the location where the propagation of the maximum concentration of solute into the subsurface stops due to flushing of the highly advective hyporheic zone below the stream. (E & J) Kurtosis based on solute transport data is well-predicted by the upper-values of ER calculated kurtosis, and more accurately predicted by quadripoles with larger spacing between the potential pair of electrodes.
signature plot (Figure 5-4A) suggest that ER holds the potential to accurately identify spatial locations where transitions between advection- and diffusion-dominated subsurface regions.

Temporal moments based on ER exhibit smaller zeroth moments (\( \mu_0^{ER} \), Figure 5-6A, 5-6F), later mean arrival times (\( \mu_1^{ER} \), Figure 5-6B, 5-6G), and relatively similar variance (\( \mu_2^{ER} \), Figure 5-6C, 5-6H), skew (\( \mu_3^{ER} \), Figure 5-6D, 5-6I), and kurtosis (\( \mu_4^{ER} \), Figure 5-6E, 5-6J) than MSS-derived moments. The larger zeroth moments from direct observation can be explained by the spatial averaging of ER data, since electrical current flows through subsurface domains of varying concentration and always provides some degree of spatial averaging, yielding apparent petrophysical properties that may vary from small-scale, true values [Singha and Gorelick, 2006b]. Smaller first moments in ER data compared to solute transport data (Figures 5-6B and 5-6G, as variance) are likely due to the three-dimensional nature of electrical current paths. Electrical signals travel out of the transverse plane upstream and downstream. Thus, the electrical signal responds to the tracer more quickly than point measurements of MSS concentration.

5.5.7 Prospects & limitations

The use of ER to provide spatially distributed observations with good temporal resolution is an exciting prospect for hyporheic studies. ER data holds the potential to overcome the “window of detection” issues commonly associated with stream tracer studies. The ability to track distributed, low-concentration tracer across a substantial subsurface domain can more completely capture the range of residence times and flowpaths within a given stream unit than typical stream tracer experiment and transport modeling approaches. ER data may be interpreted
to assess ISS concentrations and exchange rates in the subsurface, independently of those observed between stream and hyporheic zones. Finally, properly designed ER data collection may be automated to collect large amounts of data with minimal operational effort by the researcher, minimizing the additional burden on scientists during field studies. The predictive ability of ER data, especially for properties related to lower-order moment, suggests the potential for ER to inform solute transport modeling based on moments [Schmid, 2003] or to provide distributed estimates of solute transport parameters [Day-Lewis and Singha, 2008].

With these benefits come limitations. We reiterate here that the support volume for ER measurements is both dynamic and unknown for a single measurement, requiring data inversion (and its associated uncertainty) to assign spatial distributions with improved resolution. While important quantities derived from temporal moments from ER data generally bound the true values from the solute transport model, variability in the data set exists as a function of spatial location and geometric arrangement of the quadripole (Figures 5-5, 5-6). Interpretation of these data should be completed with these effects in mind, as the proximity of the quadripole to the tracer affects readings. The spatial locations of ER temporal moments are represented as the center coordinate of the potential pair and are compared to the vertical average of moments from the solute transport model (Figure 5-6). The proximity of the tracer pulse to the electrodes affects the ability of ER to accurately detect the pulse [Singha and Gorelick, 2006a]. One’s ability to collect verification data across the range of domains and temporal scales may be limited by sensitivity of monitoring equipment and subsurface access to sample fluid in the MSS. Hydrologic field techniques do not allow direct assessment of the ISS domain, requiring some degree of interpretation for all data. Finally, we note that ER is limited in accuracy of mass assessment for small, highly concentrated targets and is better suited to assessment of lower, distributed tracer presence [Singha and Gorelick, 2006a]. ER monitoring does not replace the
need for direct observation, but enhances our ability to follow tracer mass beyond spatial and
temporal scales that are typically observed.

In planning a stream-hyporheic tracer study, it is important to consider the strengths and
weaknesses of different techniques. Direct observation in the stream and streambed (i.e., via
wells) is necessary to capture short, rapid flowpaths. Coupling direct observation with an ER
array focused on slower, deeper flowpaths and tracking more diffuse mass transport processes in
the relatively deeper aquifer and floodplains would compliment the observational data set. By
targeting different spatial locations and temporal scales with unique observational methods, a new
data stream can be collected to refine solute transport models. ER extends our ability to track
mass in space and time, and can account for mass that is otherwise lost during traditional studies.

5.6 Conclusions

This work has demonstrated that electrical resistivity measurements during stream-
hyporheic solute transport experiments provide a distributed data set inclusive of both mobile and
immobile subsurface domains. Interpretation of these data allows scientists to parameterize more
complex models that include RLMT (i.e., diffusion between domains of varied mobility) in the
subsurface, a process that has been identified as important but largely overlooked in stream
modeling efforts. Temporal moment analysis has been shown to be a promising tool in
interpretation of ER data and in identification of dominant solute transport processes.

The long-term presence of solute in the subsurface, diffusing laterally and vertically away
from the stream and adjective flowpaths, suggests a potential fate for tracer that is otherwise
“lost” from downstream measurements. Our analysis of numerically simulated spatial patterns
and ER data identifies four representative signatures (stream, MSS, diffusive, and background),
and their spatial distribution in the subsurface. Traditional hyporheic monitoring in-stream and in the streambed is largely focused on stream and MSS signatures, overlooking the larger and slower diffuse signatures. Tracer mass that would otherwise be unaccounted for can be tracked through time and space via minimally invasive ER monitoring. Typical patterns identified here may provide an initial template for planning both ER and direct solute monitoring efforts; observations can be focused in the locations where the method and processes are well matched. For example, direct monitoring of highly advective hyporheic flowpaths with monitoring wells is appropriate, whereas ER provides better data in larger, slower tracer distributions. Thus, ER does not replace the need for traditional in-stream and monitoring well measurements. Rather, ER supplements these data streams by providing a distributed assessment in more slowly changing regions of the subsurface.

Simulation of mass transport by the solute transport model compared to ER-derived quantities demonstrated that estimation of solute concentration by ER is generally lower than solute transport values, yet temporal trends are visually similar. The reduced total mass values are likely because ER is averaging over a spatially complex support volume of heterogeneous concentration, while solute transport observations are point measurements in space and time. In comparison to observed MSS data, ER under predicts fluid concentration, though adequately captures spatial and temporal trends in the model domain.

Numerical simulations of this type can be used to plan for ER monitoring of stream-hyporheic tracer studies at laboratory model (e.g., flume) or field-scale studies. The coupled model allows experimentalists to consider a range of conditions that could be expected in the field. Simulation of several potential ER electrode arrays could help design the most efficient array for the expected tracer transport and could be used to determine if surface or down-well electrodes would be more logical for researchers’ objective for an experiment. Finally, these
simulations allow planning of the appropriate level of solute to be released during a tracer study, ensuring that tracer levels are maintained at as low of levels as feasible to minimize environmental impact.

Future work should attempt to identify the same patterns observed here in field data and correlate ER and solute data in a heterogeneous, unsteady-state setting. The assumptions of a homogeneous, isotropic subsurface are a limitation of this study that should be tested in future work. Differences in hydraulic conductivity and measurement support may also lead to discrepancies between ER and fluid conductivity measurements. Additionally, sensitivity analysis to model parameters (e.g., RLMT coefficient, porosity, etc.) was not completed. Predictions of solute transport parameters would be improved by combining observed ER signals in the field, direct observation in streams and monitoring wells, and a numerical model. A limited number of intensive field or large-scale laboratory studies with comprehensive monitoring would aid in determining more precisely the processes (and their relative speeds) that are well characterized by ER.

5.7 Acknowledgements

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Chapter 6

Spatially distributed characterization of solute advection, dispersion and mobile-immobile exchange along hyporheic flow paths during baseflow recession in a headwater stream

This chapter will be submitted for publication with co-authors M. Fitzgerald, M.N. Gooseff, K. Singha, and T.J. Voltz. See Appendix B for co-author permissions.

6.1 Abstract

The transport of solutes along hyporheic flowpaths is recognized as central to numerous biogeochemical cycles, yet our understanding of how this transport occurs is poor. We conducted four whole-stream, steady-state solute tracer injections coupled with electrical resistivity (ER) imaging to characterize hyporheic flow and transport through seasonal baseflow recession in the H.J. Andrews Experimental Forest (Oregon, U.S.A.). We used temporal moment analysis of pixels from the ER images to compress time-lapse data into descriptive statistics (mean arrival time, temporal variance, and temporal skewness) for each pixel. We observed the spatial distribution of these temporal moments in the subsurface at each of five 2-D transects perpendicular to the stream. We found increasing mean arrival time, variance, and skewness during baseflow recession, suggesting that changes in hydrologic forcing change the relative influence of transport phenomena (advection, dispersion, mobile-immobile exchange) along flowpaths. Pixel data were used to construct probability density functions for each moment at each transect, generalizing results into a form that can readily be used to compare differences between transects and experimental conditions. Increased understanding of the controls on
hyporheic transport and the spatial distribution of transport processes is a significant advance toward predicting hyporheic exchange at the scale of individual flowpaths based on process understanding. We demonstrate the distributed characterization of locations that are dominated by advective versus dispersive and mobile-immobile exchange processes in the subsurface.

### 6.2 Introduction

Despite a broad recognition of the ecological relevance of transport along hyporheic flowpaths [e.g., Boulton et al., 2010; Brunke and Gonser, 1997; Krause et al., 2010], relatively little is known about hyporheic transport processes and their distribution in the subsurface. Recent field and numerical studies have demonstrated that residence time along a flowpath is a primary control on biogeochemical cycling [Boano et al., 2010; Zarnetske et al., 2011]. Indeed, studies by Battin [1999; 2000] demonstrate that hydrodynamics are a first-order control on ecological processes. Transport processes in the subsurface include advection and dispersion along hyporheic flowpaths and dispersion away from individual flowpaths. An improved ability to predict spatiotemporally variable flow and transport patterns in the subsurface is necessary to predict ecologically relevant solute fluxes and to manage or design for restoration of ecosystem services.

Larger-scale hydraulic gradient, channel form, and subsurface composition are controls on hyporheic exchange. Vertical hydraulic gradient and lateral gradients from the hillslopes to the stream have been demonstrated to control hyporheic exchange in many studies [e.g., Boano et al., 2008; Cardenas and Wilson, 2007b; D'Angelo et al., 1993; Harvey and Bencala, 1993; Storey et al., 2003; Williams, 1993; Wondzell and Swanson, 1996; Wroblicky et al., 1998]. The effect of natural in-stream geomorphologic features (i.e., channel form) on hyporheic exchange has also
been the topic of much research [e.g., Gooseff et al., 2006; Harvey and Bencala, 1993; Kasahara and Hill, 2007; Revelli et al., 2008]. Packman and Salehin [2003] found that heterogeneity of the hydraulic conductivity matrix controls the extent of hyporheic exchange in detailed flume studies; even small-scale heterogeneity results in complex hyporheic exchange patterns [e.g., Salehin et al., 2004; Sawyer and Cardenas, 2009; Woessner, 2000], however, detailed study of hydraulic conductivity heterogeneity in-situ and its control on hyporheic flow and transport is lacking. Understanding the spatial distributions of temporal trends in hyporheic transport is a necessary step toward a process-based understanding of biogeochemical cycling in the subsurface.

Because the boundary conditions controlling hyporheic exchange can be dynamic (e.g., hydraulic gradients), we expect that transport processes in the subsurface will respond to spatiotemporally dynamic forcing. At the reach scale, several studies have documented changes in transient storage (as evidence of hyporheic exchange changes) under different flow conditions. Increasing stream flow has been found to increase exchange with bedforms [e.g., Elliott and Brooks, 1997a; Elliott and Brooks, 1997b; Packman and Salehin, 2003]. Field studies using numerical modeling of stream solute transport from tracer injections have found decreasing storage-area ratios with increasing discharge [Butturini and Sabater, 1999; Fabian et al., 2010; Karwan and Saiers, 2009; Morrice et al., 1997; Schmid et al., 2010; Zarnetske et al., 2007], though increasing discharge may increase hyporheic exchange rate [Fabian et al., 2010; Hart et al., 1999]. Hydraulic gradients created by ambient gaining and losing conditions have been studied as a control on hyporheic exchange. Numerical studies of pumping exchange found hyporheic flowpaths contract due to ambient gaining conditions [Boano et al., 2008; Cardenas and Wilson, 2007b]. Several field studies report contraction of hyporheic flowpaths during
periods of increased groundwater discharge to streams [Harvey and Bencala, 1993; Storey et al., 2003; Williams, 1993; Wondzell and Swanson, 1996; Wroblicky et al., 1998].

In contrast, Wondzell [2006] completed tracer studies in steep headwater catchments in the H.J. Andrews Experimental Forest under two different baseflow conditions (4.5 and 1 L s\(^{-1}\) in WS1, 10 and 3 L s\(^{-1}\) in WS3), noting that hyporheic extent, evaluated as tracer arrival in a monitoring well network, was not changed during baseflow recession. In a modeling study of one of the watersheds in their 2006 study, Wondzell et al. [2009] concluded that even complex physical models are insufficient to predict the movement of solute through the hyporheic zone. This shortcoming is attributed to the inability to identify unique model solutions and invalidate all other solutions, a problem which arises because spatially discrete measurements are used to infer upgradient behavior along flowpaths. Indeed, hydrologic modeling efforts attempting to simulate a limited number of point measurements are plagued by problems of equifinality [e.g., Beven, 1993; 2006]; an infinite number of upstream realization can adequately produce the observed data. An ongoing debate about the usefulness of such models in predicting at the scale of individual flowpaths and in predicting solute transport continues in the literature [e.g., Bredehoef\(t\) and Konikow, 1993; Hassan, 2004; Konikow and Bredehoef\(t\), 1992; Oreskes et al., 1994; Poeter, 2007; Wondzell et al., 2009].

Solute tracer studies have been commonly applied to characterize transient storage at the reach scale [Stream Solute Workshop, 1990] and are commonly interpreted with simple, highly idealized models [Bencala and Walters, 1983; Thackston and Schnelle, 1970]. Like more complex groundwater flow models, these models suffer from equifinality problems associated with data limitations. Simplified transport models are known to lack physical meaning [Marion et al., 2003; Wondzell, 2006]. Such simplified models lack the ability to fit observations of late-time behavior [e.g., Zaramella et al., 2003], which may be due to the assumptions of a well-

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mixed storage domain and assumed residence time distributions. The observed downstream solute breakthrough curves to which such models are fit are an integration of all of the upstream transport processes that occur within the study’s window of detection [Harvey and Wagner, 2000] and lack spatial resolution. Finally, tracer studies at the reach-scale interpreted with simple box-models overlook down-valley groundwater transport processes, which have been identified as important in some settings [Castro and Hornberger, 1991; Jackman et al., 1984; Kennedy et al., 1984]. Networks of monitoring wells have been installed to provide point-verification of exchange processes, but they lack spatial resolution (i.e., each well only samples those flowpaths that intersect the well screening). Monitoring well data during solute studies rarely agrees with modeling based on in-stream observations [Harvey et al., 1996; Wondzell, 2006].

Given the limitations of solute tracer studies alone, we look to electrical resistivity imaging to provide spatially distributed information with high temporal frequency. In contrast to spatially lumped tracer studies, and spatially sparse networks of monitoring wells, hydrogeophysical methods can be used to monitor transport processes, characterize subsurface structure, and quantify hydrological parameters in the subsurface with dense spatial distribution [Binley et al., 2010]. Of particular interest is the characterization of process dynamics (i.e., characteristics that change through time and are inherently linked to the movement of fluids in the subsurface) [Binley et al., 2010; Koch et al., 2009]. Indeed, electrical resistivity (ER) imaging of solute tracers in hyporheic zones has been demonstrated in recent field trials [Nyquist et al., 2010; Ward et al., 2010b; Chapter 2; Cardenas and Markowski, 2011]. Such imaging provides high spatial and temporal resolution of solute transport in-situ, overcoming the limitations of reach-scale studies that infer upstream behavior based on discrete downstream observations.

Interpretation of ER images to characterize solute behavior in groundwater systems is becoming commonplace. Individual pixels in a soil column experiment demonstrated
characteristic trends of advection-dispersion behavior of tracer-labeled water moving through the column [Binley et al., 1996a]. In a 2-D laboratory experiment, Slater et al. [2000] interpreted pixel breakthrough curves to quantify first arrival times of tracer at individual pixels and recommended comparison of pore fluid with ER images in future studies. In a 3-D laboratory study, Slater et al. [2002] demonstrated broad agreement between pore fluid conductivity and pixel resistivity. Their study demonstrates the value of a densely sampled, spatially continuous data set to observe greater complexity in transport processes than was possible with their flow and transport modeling alone. Kemna et al. [2002] compared ER image behavior with advection-dispersion modeling of groundwater flow and transport and were able to observe more complexity in flowpath behavior than was present in their numerical simulations of flow and transport. Indeed, the use of ER to characterize groundwater flow and transport processes is common [e.g., Singha and Gorelick, 2006a]. With the promise of ER imaging to characterize solute transport come inherent limitations, including sources of inaccuracy in interpreting solute behavior from ER: (1) differences in support volumes of pore fluid concentration and electrical resistivity data, (2) inaccuracies in petrophysical relationships to convert between ER pixel resistivity and pore fluid concentration, (3) measurement errors associated with ER data, (4) non-uniqueness of ER inversion models, and (5) limitations of ER images applied to time-lapse problems, where timesteps are either inverted independently (limiting consistence between timesteps) or using previous results as the starting model for a given timestep (where errors in a given timestep may be propagated forward) [Slater et al., 2002; Day-Lewis et al., 2005]. Despite inherent uncertainty in inverted ER images, both qualitative and quantitative interpretation of individual pixel or finite element behavior during saline tracer studies has proven useful [Binley et al., 1996a, b; Slater et al., 2000; Slater et al., 2002; Singha and Gorelick, 2005]. Ward [Chapter 3] used time-lapse ER images of saline tracer movement through hyporheic flowpaths as
one tool to parameterize groundwater flow and transport models. Time-lapse ER has also been interpreted to inform subsurface structure and hyporheic extent from ER imaging during a solute tracer study [Chapter 3, Chapter 4].

In this study, we use temporal moment analysis to characterize transport of solutes along hyporheic flowpaths in the subsurface. Temporal moment analysis has been used to describe the transport of solutes in the subsurface in many studies [e.g., Cirpka and Kitanidis, 2000; Day-Lewis and Singha, 2008; Singha et al., 2007] and has also been used to describe flow and transport in coupled stream-aquifer systems [e.g., Gupta and Cvetkovic, 2000; Schmid, 2003]. Cirpka and Kitanidis [2000] interpret temporal moments of simulated tracer breakthrough curves at points in the subsurface to derive apparent seepage velocity and dispersivity in the subsurface. In a numerical study of hyporheic exchange, Singha et al. [2008] calculate temporal moments based on reach-averaged concentrations and suggest moments calculated at a higher spatial resolution may be appropriate if variability is expected across a domain of interest. In a modeling study, Ward et al. [2010a; Chapter 5] interpret the spatial distribution of temporal moments of solute transport and raw ER data (as opposed to images) to infer transport processes in the subsurface. Results of their simulations identified areas where advection dominates dispersion and mobile-immobile exchange in the subsurface. They speculate that a similar analysis could be completed using time-lapse ER images collected during a field study.

The overarching objective of this study is to assess how solute transport processes in the hyporheic zone change during seasonal baseflow recession, a period during which we expect a change in the boundary conditions that control potential for hyporheic exchange to occur. A secondary objective is to demonstrate the ability of temporal moment analysis of ER images to provide spatially distributed solute transport information in the subsurface (to quantify where and when advection, dispersion and mobile-immobile exchange are occurring), similar to the spatially
distributed analysis of numerical simulations by Ward et al. [2010a; Chapter 5]. We expect that changing boundary conditions during baseflow recession will change temporal trends in hyporheic solute transport in the subsurface (for example, later mean arrival times and increased evidence of transient storage and dispersion). To achieve these objectives, we conducted a series of replicate solute tracer studies and monitored tracer movement in the subsurface using electrical resistivity imaging in a small headwater stream. Spatially distributed characterization of temporal trends at each transect provides an assessment of flowpaths that are commonly beyond a tracer study’s window of detection using in-stream tracer recovery or well data. We develop probability density functions describing temporal moments at each transect during each injection to quantitatively compare transport behavior along flowpaths. Ultimately, understanding the distribution of transport processes provides a more complete assessment of transport than has been possible with reach-scale tracer studies and monitoring well networks.

6.3 Methods

6.3.1 Site description

The field studies presented here were conducted in WS3 in the H.J. Andrews Experimental Forest, located in the western Cascade Mountains of Oregon, USA (48° 10’ N, 122°15’W). The study watershed is a steep (hillslope gradients greater than 50%), narrow (8.5 m wide valley bottom) catchment draining 101 ha and ranging in elevation from 497 to 1070 meters above mean sea level. Shallow, highly porous loams are typically 1 to 2 m deep on the site [Dyrness, 1969]. Saturated hydraulic conductivity in the watershed has been reported as $7 \times 10^{-5}$ m s$^{-1}$ in the watershed [Kasahara and Wondzell, 2003], while Wondzell et al. [2009] report a
geometric mean of $1.7 \times 10^{-5}$ m s$^{-1}$ for an adjacent headwater catchment. *Wondzell* [2006] describes the study reach as bedrock constrained. The upper reach (ER transects 1-3) is highly constrained, with bedrock observed to outcrop along the Northeast valley wall. The lower section of the study reach (ER transects 4-5) has a flatter gradient and is wider, with no apparent bedrock outcropping. A large fallen tree and several boulders hold a wedge of alluvium in the valley bottom.

Studies were primarily focused on a short, second-order study reach with a gradient of 14% (Figure 6-1), including several pool, riffle, and step features. *Kasahara and Wondzell* [2003] report an average of 8.4 steps or riffles contributing an average of 54% of the elevation change along each 100 m reach. Flow is gauged at the outlet of WS3 at a permanent weir; flow ranged from 35 L s$^{-1}$ early in the recession period to 4 L s$^{-1}$ by the end of the study period (Figure 6-2). Precipitation data were recorded at a nearby weather station at the H.J. Andrews Experimental Forest headquarters.

### 6.3.2 Solute tracer studies

We conducted four 48-hr constant rate injections of sodium chloride (NaCl, a conservative tracer) in the watershed, with injections at the same locations used by *Wondzell* [2006] (approximately 50 m upstream of the study reach in each watershed). Tracer was injected directly into the stream channel, with each experiment designed to increase in-stream electrical conductivity (EC) by approximately 100 µS cm$^{-1}$. For all measurements, ER was used as a surrogate for tracer concentration [after Gooseff and McGlynn, 2005; Payn et al., 2009; Wondzell, 2006]. All tracer studies began between 13:00 and 14:00 and lasted for exactly 48 hours.
Figure 6-1: Study reach map for WS3 in the H.J. Andrews Experimental Forest, located in the Cascade Range of central Oregon, USA. Flow is from southeast to northwest. Solute tracers were injected into the stream approximately 50 m upstream of the study reach. Wells and piezometers were assigned unique identification values as letter-number combinations. In-stream monitoring locations are labeled “IS”.
Tracer presence in the surface water was recorded at the upstream end of the study reach using a temperature and EC probe manufactured by Campbell Scientific, Inc. (Logan, Utah, United States). EC was monitored in the well network during the study by purging one well volume and making down-well measurements using either an EcoSense EC300 (YSI, Inc., Yellow Springs, Ohio, United States) or a Model 107 Temperature/Level/Conductivity meter (Solinst, Inc., Georgetown, Ontario, Canada). In-stream and down-well EC measurements began prior to the injection and lasted for at least 120 hours after the start of the injection.

Figure 6-2: Flow and precipitation in WS3 during the 2010 study season. Flow was gauged at weirs located approximately 100 m downstream of each study reach. Injection periods for each solute tracer injection are shown as shaded bars.
6.3.3 Electrical resistivity imaging

Data collection

Electrical resistivity data were collected at six transects within the study reach (one of which yielded data too noisy to interpret, and has been omitted from the remainder of this manuscript). Data were collected using a 10-channel Syscal Pro Resistivity Meter (IRIS Instruments). Electrodes were spaced 1 to 2 m apart along a transect, and transects were spaced at approximately 5 m intervals along the valley floor. The transects included at least two electrodes on each hillslope, and were centered on the valley bottom (Figure 6-1). We constructed electrodes by attaching conductive foil tape to 1.27-cm diameter PVC pipes. Foil tape was connected to the solid strand 18-gauge trunk-lines (i.e., connections to the Syscal Pro meter) using 18-gauge stranded wire. Electrodes were driven into the subsurface to approximately 25-cm below the surface.

Before and during the tracer studies we monitored contact resistance using the integrated contact resistance test in the Syscal Pro meter. Contact resistance values ranged from 1 to 16 kOhms during the study, with most (about 90%) lower than 5 kOhms for the entire monitoring period. ER data were collected at each transect using a mixed dipole-dipole array with a total sequence of 323 measurements collected; the same sequence was used for all transects. The sequence was selected to maximize coverage in the valley bottom with a minimum number of individual measurements collected (to maximize temporal resolution of the data set). Data collection took approximately 14 minutes per transect, with one full survey of all six transects taking approximately 84 minutes to complete the full six-transect sequence. We assumed the subsurface was static during each 14-minute collection period (i.e., no corrections were made to
account for changes that may have occurred between the first and last measurements recorded. Data were collected only along (and not between) transects to maximize temporal resolution. Data were stacked (i.e., averaged) in the field during collection to provide a measure of quality control. A minimum of two measurements was stacked for every data point. If the standard deviation between the two measurements was larger than 2%, an additional two measurements were recorded and included in the average.

Data inversion

ER data were inverted using a code described by Binley and Kemna [2005], available for download as the research code R2 (v2.6, Generalized 2-D Inversion of Resistivity Data, available online at: http://www.es.lancs.ac.uk/people/amb/Freeware/freeware.htm). Data were weighted for the inversion scheme as the reciprocal of the standard deviation between stacked measurements. Background (i.e., pre-injection) data were inverted using a homogeneous starting model; data collected during the tracer study were inverted using the inverted background data as a starting model. Error model parameters were adjusted to yield a root mean square error (RMSE) relative to noise in the data as close to 1.00 as possible for the background inversion, and to provide the minimum average RMSE for all inversion timesteps collected during the solute study. Independent inversion of each timestep limited the potential for errors in the inversion process to be propagated through the time series data, although no temporal regularization was utilized [LaBrecque and Yang, 2001; Day-Lewis et al., 2002].
**Post-processing**

Each image collected during and after the tracer study was post-processed by subtracting the background model from the inversion model, and the percent change from background was calculated for each pixel. The presence of the solute tracer in the subsurface was expected to result in decreases compared to background ER. Pixels for analysis were limited to those which were both well resolved (i.e., log_{10}(\text{resolution}) \geq 2.5) and where solute was present to cause a minimum percent change of -3% in resistivity from background [after thresholding by Ward et al., 2010b; Chapter 2]. Results of our analyses are sensitive to the thresholds chosen for both resolution and percent change.

Pixels were analyzed to determine the first arrival of the tracer (i.e., the earliest time that a decrease of greater than or equal to 3% was observed at the transect). Additionally, the latest arrival times at each pixel were tracked, and hyporheic persistence (i.e., the maximum range of time where tracer was present at the transect, characterized as a decrease of 3% or more in pixel resistivity) was calculated. Hyporheic persistence can be interpreted as a proxy for residence time along an individual flowpath; persistence informs the temporal scale over which the 48-hr solute injection is present at a given location in the subsurface. Finally, Ward [Chapter 4] compared the breakthrough curves observed in monitoring wells to pixel breakthrough curves in corresponding locations. Comparison of pixel breakthrough curves with observed point measurements has been successfully applied in the literature to characterize fluid flow and solute transport [e.g., Binley et al., 1996a; Binley et al., 1996b; Slater et al., 2000; Slater et al., 2002]. Ward’s comparison [Chapter 4] demonstrates the sensitivity of ER images to the tracer presence in the subsurface.
6.3.4 Temporal moment analysis

Temporal moments of time-variable data (e.g., solute breakthrough curves) can be used to compress time series concentration histories into descriptive statistics (i.e., mean arrival time, variance, skewness). An \( n \)th order temporal moment \( (M_n) \) is calculated by:

\[
M_n = \int_0^\infty t^n r(t) \, dt
\] (1)

where \( t \) is time, and \( r(t) \) is normalized change in pixel resistivity. Field observations of resistivity, \( R(t) \), are normalized as:

\[
r(t) = \frac{R(t)}{\int_{-\infty}^{\infty} R(\tau) \, d\tau}
\] (2)

such that \( M_0 = 1 \) (i.e., the total area bounded by \( c(t) \) is unity). The first temporal moment (i.e., \( n=1 \)) is interpreted as the mean arrival time of the breakthrough curve. Higher order temporal moments defined about the centroid \( (M_1) \) can be calculated as:

\[
M_2 = \sigma^2 = \int_0^\infty (t - M_1)^2 r(t) \, dt
\] (3)

and

\[
M_3 = \int_0^\infty (t - M_1)^3 r(t) \, dt
\] (4)

where \( M_2 \) (the second central moment) is the variance of the concentration profile, and \( M_3 \) (the third central moment) is the skewness [Gupta and Cvetkovic, 2000; Harvey and Gorelick, 1995]. The first moment, \( M_1 \), is physically interpreted as the mean arrival time of the resistivity change at the pixel. Because we attribute all changes in pixel resistivity to the solute tracer, \( M_1 \) is equivalent to the mean arrival time of the tracer in this study. Variance of the pulse describes the
temporal spreading of the pixel resistivity changes (e.g., solute tracer, by the same logic as $M_1$) due to differences in advective velocities and dispersion of the tracer [Ward et al., 2010a; Chapter 5]. Skewness ($M_3$, again of the pixel resistivity changes, interpreted here as solute tracer) describes the asymmetry of the distribution, with positive values suggesting solute breakthroughs that are weighted toward early times and heavily tailed (i.e., “leaning” left). Hereafter we will refer to the moments by the physical transport mechanisms associated with each moment (i.e., $M_1$ as advective timescale, $M_2$ as dispersion along a flowpath, and $M_3$ as mobile-immobile exchange along a flowpath). Finally, we calculated unique probability density functions (PDFs) for each temporal moment at each transect during each injection. Temporal moments at each pixel were binned with 10 bins per log cycle, and the resulting histogram converted into a PDF.

6.4 Results

6.4.1 Stream solute tracer studies

The breakthrough curves for the four replicate tracer studies in each watershed were logged at the upstream end of the study reach (Figure 6-3). Background tracer concentration was subtracted from each observation set. Plateau concentrations range from increases of approximately 60 to 120 $\mu$S cm$^{-1}$ during the injection, due to differing in-stream and injection flow rates between injections. Injection flow rates were checked at intervals of 3-4 hours to ensure a constant rate injection into the stream. During tracer injection 1 the injection rate approximately doubled after about 39 hours of injection (at approximately 03:00). This change was discovered at approximately 06:00 and the injection flow rate was re-set to its initial rate.
Tracer arrival time at the study reach was fastest during the first injection, with first arrival occurring within minutes of the injection start time. Tracer arrival time increased with decreasing flow rate, ultimately reaching a time of approximately 20 minutes during the final, lowest flow rate injection. During all studies a gradually increasing plateau was reached during the 48-hr injection as progressively longer residence-time flowpaths returned to the stream. Tailing behavior of the tracer after the injection ended was delayed between studies in a similar pattern to arrival time. This increased tailing suggests an increased influence of transient storage on the overall behavior of tracer in the study reach.

6.4.2 Imaging solute transport in the subsurface

Electrical resistivity images were sensitive to the solute tracer, providing distributed assessment of solute presence in the subsurface [Ward, Chapter 4]. The cross-sectional extent of the domain presented in Figures 6-4 through 6-6 represents the maximum extent of a minimum decrease of 3% in resistivity, interpreted as the presence of the tracer at that location during at least one timestep. The peak size of the hyporheic zone during each injection was larger during high-flow conditions for upstream transects (T1-3), while extent was more consistent for downstream transects (T4-5). Ward [Chapter 4] presents a detailed assessment of spatiotemporally variable trends in hyporheic extent as controlled by valley confinement and hydraulic gradients (both vertical and valley-bottom).

First arrival times at each pixel are presented in Figure 6-4. Arrival times are generally fastest near the stream (less than 10 hours in many pixels) and slower at greater distances. First arrival times at T1-3 are later at most pixels for injections that occurred later in the season with baseflow recession. In lower reaches (T4-5), the rapid tracer arrival in the subsurface was over a
much larger spatial area, compared to upstream reaches during all flow conditions. This arrival suggests rapid and extensive downwelling at T4 and T5. Tracer arrival times in the stream at the study reach are less than 0.5 hr (Figure 6-3); the delay in advection along the stream channel is not sufficient to cause the differences observed in Figure 6-4.

Hyporheic persistence (i.e., the time elapsed between the first and last images where pixel resistivity dropped by at least 3%) is presented for all transects and injections in Figure 6-5. Hyporheic persistence generally increased with baseflow recession at all locations, indicating a slower transport along subsurface flowpaths (and therefore a longer flowpath residence time) during lower-flow injections. Downstream transects (T4-5) exhibit larger areas of high-persistence hyporheic flowpaths compared to upstream transects (T1-3). The area of lower persistence near the stream surrounded by higher-persistence pixels (e.g., the ~60-hr persistence

**Figure 6-3:** Breakthrough curves at the upstream end of the study reach in WS3, as the change in observed electrical conductivity (EC) of the surface water (a surrogate for the concentration of the sodium chloride tracer) in normal (A) and log-space (B). Arrival time at the study reach was increasingly later as flow decreased during the season. Studies during lower flow conditions show increased tailing at late times, suggesting increased transient storage between the injection point and study reach.
pixels surrounded by those with 80+ hour persistence in injection 2, T2 in Figure 6-5) is indicative of an advection-dominated region where solute was rapidly flushed from flowpaths, surrounded by a domain where dispersion and rate-limited mass transport have a stronger effect on transport behavior.

6.4.3 Spatial distribution of temporal moments

Advective timescale of flowpaths

The advective timescale (i.e., mean arrival time, $M_1$) of solute in the subsurface is less than 100 hr for most pixels (Figure 6-4). The fastest flowpaths are in the shallow subsurface near the stream channel (timescales of less than 50 hours in many locations), while longer advective timescales are observed farther from the stream. This is consistent with the spatial patterns of mean arrival times reported by Ward et al. [2010a; Chapter 5]. The advective timescale in the upper (steeper, confined) reach appears to generally increase throughout baseflow recession (i.e., tracer takes longer to reach a given location in the baseflow recession). In contrast, the lower reach (less confined alluvial deposit) had widespread rapid arrival under all flow conditions.

For a perfectly advective system (i.e., plug flow), we would expect mean arrival of about 24 hours (since the injection was 48 hours in duration). Arrival in the subsurface generally takes much longer, though some highly advective flowpaths are observed. At T1, mean arrival is less than 60 hr for depths up to 5 m during injection 1. During lower flow conditions, this zone of rapid arrival decreased substantially, with only a shallow lens of rapid arrival for injection 2, and arrivals of greater than 60 hours in this near-stream advection-dominated location during injection
Figure 6-4: Tracer first arrival time for all transects and injections. First arrival is defined here as the earliest time at which a decrease of 3% or more in pixel resistivity was observed. Arrival times are generally faster near the stream and decrease with both lateral and vertical distance. In upstream transects (T1-3), arrival time generally gets later through baseflow recession, indicating slower transport along hyporheic flowpaths. In downstream transects (T4-5) there is a large zone of rapid arrival under all flow conditions, suggesting a consistent, extensive location of tracer downwelling. Tracer arrival times in the stream at the study reach were within 0.5 hours (Figure 6-3) through the recession period, a delay not long enough to create the patterns seen here.
Figure 6-5: Hyporheic persistence, defined here as the maximum range in times where a pixel decrease of 3% or more was observed, for each transect and injection. Persistence of tracer along flowpaths generally increases with decreasing baseflow. Downstream transects (T4-5) exhibit large regions of high persistence; these locations exhibit both rapid arrival and substantial persistence of hyporheic flowpaths. The low-persistence areas near the stream surrounded by zones of higher persistence are indicative of advection-domination near the stream channel; flowpaths are rapidly flushed of solute tracer. The surrounding zone is interpreted as a location where dispersion and rate-limited mass transport are more dominant transport processes.
3. At T2 similar patterns are observed. Notably there are highly advective flowpaths (i.e., short advective timescale) centered near X = 7 m that is present during injections 1 and 2, but disappears for injections 3 and 4. T3 has a similarly shrinking zone dominated by highly advective flow centered near X = 7m. At T4 and T5, rapid arrival of tracer was observed during higher flow conditions (injection 2); arrival time was later during lower flow injections.

Dispersion along flowpaths

Dispersion of solute along flowpaths (i.e., variance of the solute pulse, $M_2$) is generally lowest in locations with low advective timescales (often those nearest the stream channel), and increases with increasing distance from the channel (Figure 6-7). In several transects, the low variance near the stream is surrounded by a “ring” of higher variance followed by lower variance (for example, T3 during injection 2, Figure 6-7). Ward et al. [2010a; Chapter 5] report a similar pattern based on their numerical study, attributing the lower dispersion near the stream to an advection-dominated region of the subsurface. The low dispersion areas near the edges are interpreted as locations where tracer arrived more slowly than the advective timescale by diffusion-dominated transport, or where the inversion process erroneously calculates pixel resistivity changes. Dispersion of the solute pulse along any individual flowpath (i.e., at any pixel) generally increases during baseflow recession. Injection 3 is anomalous in this pattern, showing decreased variance relative to injection 2, but higher variance than injection 4. Solute concentration was highest during injection 2 and lowest during injection 4; changes in in-stream solute concentration do not explain the anomalous behavior observed during injection 3.
Mobile-immobile exchange along individual flowpaths

Mobile-immobile exchange along individual flowpaths (i.e., increasingly positive skewness, $M_3$) generally shows highly tailed positive distributions near the stream channel, indicative of rapid advective transport with dispersion and/or mobiel-immobile exchange (both in the stream delivering the solute signal to downwarding locations and along the flowpaths themselves, Figure 6-8). Mobile-immobile exchange exhibits a similar spatial trend as dispersion, with a “ring” of apparent high transient storage in the subsurface (for example, T2 and T3 during injection 2 in Figure 6-8) surrounding a region near the valley surface with almost no transient storage. This pattern indicates advective transport dominated the “inside” of the ring, while transient storage and dispersion were more important to transport with increasing distance from the stream. Mobile-immobile exchange increased during injections 3 and 4, likely due to longer advective timescales allowing for increased spreading of the solute. Transient storage generally increases during baseflow recession for all transects. As with dispersion, transient storage is anomalously high during injection 3, and decreases across transects during injection 4.

In the flatter and wider section of the study reach (T4-5), transient storage along flowpaths increased with decreasing baseflow (a general trend, despite the anomalously high injection 3). In-stream plateau concentration was highest during injection 2 and lowest during injection 4; solute concentrations in the stream do not explain the anomalous behavior observed during injection 3. This trend, interpreted with the longer advective timescales, suggests that slower advection allows for increased dispersion or rate-limited exchange with less mobile domains (i.e., bound pore-water, dead-end fractures, etc.).
Advective timescales are generally fastest near the stream and during higher flow conditions. To restrict the analysis to only meaningful changes due to the solute tracer, pixels shown in each frame are those where there was both high resolution in the inversion model and where a change of at least -3% in resistivity was observed during the study.
Figure 6-7: Dispersion along flowpaths for each transect and injection. Dispersion is generally low adjacent to the stream, interpreted as an advection-dominated region of the subsurface, and in the boundaries of each cross-section. A “ring” of higher-dispersion exists around the highly advective area of the subsurface (injection 3, T3, for example), interpreted as a region where dispersion and rate-limited mass transport become more dominant in solute transport [Ward et al., 2010a; Chapter 5]. To restrict the analysis to only meaningful changes due to the solute tracer, pixels shown in each frame are those where there was both high resolution in the inversion model and where a change of at least -3% in resistivity was observed during the study.
Figure 6-8: Mobile-immobile exchange along flowpaths for each transect and injection. Mobile-immobile exchange (i.e., positive skewness) is generally lowest adjacent to the stream, interpreted as an advection-dominated region of the subsurface, and in the boundaries of each cross-section. A “ring” of higher skewness exists around the highly advective area of the subsurface (injection 3, T3, for example), interpreted as a region where dispersion and rate-limited mass transport become more dominant in solute transport [Ward et al., 2010a; Chapter 5]. To restrict the analysis to only meaningful changes due to the solute tracer, pixels shown in each frame are those where there was both high resolution in the inversion model and where a change of at least -3% in resistivity was observed during the study.
6.4.4 Probability density functions of tracer arrival and persistence

PDFs were constructed for tracer first arrival and persistence (Figure 6-9). First arrival times were highly variable during baseflow recession. PDFs of first arrival have highest-probability peaks increasing throughout baseflow recession, showing later arrival of the tracer at the transect. No consistent spatial trends are apparent in the PDFs for tracer first arrival. Tracer persistence along flowpaths generally increased through baseflow recession. Hyporheic persistence is also most uniform under high-flows and shows increasing spreading under lower flow conditions. No consistent spatial trends are present in the PDFs of persistence. Both peak probability and median values for first arrival and persistence follow this same general trends, with earlier arrival and shorter persistence during high flow conditions (Figure 6-11A, B, F, and G).

6.4.5 Probability density functions of temporal moments

PDFs of the spatial distribution of the first three temporal moments at each transect were calculated for all four injections (Figure 6-10). The distribution of advective timescales is generally wider during higher flow conditions (injections 1 and 2), and becomes more compact with higher peaks during lower flows. The advective timescale in the subsurface is delayed from in-stream arrival by 25 to 50 hr for all flow conditions, suggesting tracer transport observed in the subsurface occurs at much slower scales that that observed in the stream itself. Mean arrival times in the upstream transects were generally earlier than observations at the downstream transects. Peak probability and median values for the PDFs are generally larger for upstream transects, suggesting slower arrival in the more confined valley segment (i.e., at T1-3; Figure 6-11C and H).
The PDFs of the dispersion are generally larger for lower flows, and these more spread-out tracer signals tend to be more similar distributions with larger peaks (Figure 6-10). Dispersion in the subsurface is greater than the signal observed in the stream for nearly all pixels under all flow conditions (low-probability exceptions exist during injection 2). Downstream transects exhibit more dispersion than upstream transects during lower flow conditions (injections 3 and 4), likely due to the more spatially extensive flowpath networks intersecting the observation transects. Peak probability and median values for the PDFs exhibit no spatial trend, but are generally largest during lower-flow conditions (Figure 6-11 D and I). This trend suggests reduced advective dominance during low-flow periods.

Mobile-immobile exchange PDFs are varied for all injections, and both positive and negative skewness are presented (Figure 6-10). Skewness generally increases in magnitude with decreasing flowrate; maximum skewness magnitude was observed during injection 3 with a slight decrease observed for injection 4. Higher mobile-immobile exchange is observed at downstream transects (i.e., peaks at larger magnitudes), especially during injections 3 and 4. Mobile-immobile exchange PDF peaks generally reduce (i.e., distributions spread out) during lower flow conditions. Mobile-immobile exchange observed in the stream is less than that observed in the ER images. The stream’s skewness shifts from negative to positive between injections 2 and 3, a result of the increased tailing observed (Figure 6-3). Peak probability and median values from the PDFs exhibit increased skewness during lower-flow periods, a line of evidence that advection becomes less dominant in the subsurface during lower flows (Figure 6-11E and J).
6.5 Discussion

6.5.1 Changes in solute transport during baseflow recession

The results of this study demonstrate a fundamental shift in transport processes as seasonal baseflow recession occurs. Throughout the study, the solute tracer arrives more slowly both in the channel (due to slower advective velocities in the stream) and in the subsurface at the geophysical transects. Observed breakthrough curves for inversion pixels generally show increased dispersion and mobile-immobile exchange. Temporal trends in the spatial moment analyses show that the hyporheic zone decreased in both horizontal and vertical penetration into the near-stream aquifer as flow receded. The shrinking areas of advection-dominated transport in the subsurface suggest that changing hydraulic gradients around the channel and catchment wetness (i.e., lateral groundwater gradients beyond the valley floor) substantially control subsurface transport. The increasing spread of both PDFs through baseflow recession suggests increased diversity of hyporheic flowpaths during lower flow condition.
Figure 6-9: Probability density functions for tracer first-arrival (left column) and hyporheic persistence, constructed from analysis of inversion pixels. Through baseflow recession, first arrival times and hyporheic persistence are generally later and longer, respectively. First arrival times are more uniform under high flow-rates, with increased spread of the PDF as baseflow decreased. Hyporheic persistence is also most uniform under high-flows, and shows increasing spreading under lower flow conditions. No consistent spatial trends are present in the PDFs. Non-physical small first arrivals and persistence are a product of error in the inversions.
Figure 6-10: Probability density functions for mean arrival time (i.e., advective timescale; left column), temporal variance (i.e., dispersion; second column), and temporal skewness (i.e., mobile-immobile exchange; negative skewness in third column, positive skewness in right-most column), constructed from analysis of inversion pixels. Temporal moments for solute breakthrough observed in the stream channel are plotted as vertical dashed lines for each experiment. Mean arrival time, variance, and magnitude of skewness all generally increase with decreasing baseflow. In all cases, distributions become more compact and spatially uniform with baseflow recession, suggesting more uniform transport processes in the subsurface. Temporal moments observed in the stream at the upstream end of the study reach are lower than most pixels, confirming that in-stream observations are limited by a window of detection (i.e., a sensitivity to only the shortest timescale processes) [Harvey and Wagner, 2000].
Figure 6-11: Summary of the peak (i.e., most likely, left column) and median (right column) values for the PDFs presented in Figures 6-9 and 6-10 for all injections (by in-stream flow, X-axis) and transects (by linetype in each plot). During higher flow conditions, first arrival times were generally earliest, persistence was shortest, and positive skewness was lowest. These data suggest hyporheic transport was dominated by highly adjective, rapidly flushed flowpaths. Peak and median values track well across both spatial and temporal variability.
In comparison to the observed temporal moments in the stream at the upstream end of the study reach (vertical dashed lines in Figure 6-7), the advective timescale of tracer transport is longer in the subsurface. Dispersion in the subsurface is greater than that observed in the stream channel. Transient storage of in-stream injections is less than many of the subsurface pixels. This comparison demonstrates the ability of ER to characterize transport and specific transport processes (evidenced by changing temporal moments) at longer temporal scales than would be possible using in-stream observations alone. The stream returns to background while many subsurface flowpaths are still labeled with tracer [Ward, Chapter 4].

6.5.2 Comparison of observations with common flow and transport models

Common solute transport models assume exponential or power-law distributions of solute residence times in transient storage at each spatial step and assume a well-mixed transient storage domain [e.g., Bencala and Walters, 1983; Haggerty et al., 2000; Haggerty et al., 2002; Thackston and Schnelle, 1970]. While these numerical models have been demonstrated to fit field observations, they are known to be sensitive to only the shortest spatial and temporal flowpaths (commonly called the “window of detection” problem) [Harvey and Wagner, 2000]. Much of this temporally short transient storage might even be attributed to in-channel dead zones (e.g., recirculating eddies behind in-channel features) [e.g., Choi et al., 2000; Gooseff et al., 2005; Gooseff et al., 2008; Thackston and Schnelle, 1970].

Geophysical monitoring shows subsurface behavior in stark contrast to the assumptions of the numerical models. The flowpath behavior in the subsurface is not well mixed (i.e., non-uniform in space), the hyporheic area is widely varied through time, and substantial down-valley
transport in the subsurface is observed. While reach-scale model fits based on in-stream observations may provide a reasonable routing of in-stream behavior, this study demonstrates that a spatially extensive flowpath network exists in the subsurface exhibiting a wide array of advective timescales and physical transport processes. The repeated solute tracer experiments show a high degree of variability in transport processes through both time and space. Based on the changing spatial distributions of temporal moments (Figures 6-6 through 6-8) and PDFs (Figure 6-10), the subsurface behaves as a unique filter of solute signals during each injection.

In comparing our results to commonly applied numerical models, each geophysical transect represents a single spatial step. The PDF for mean arrival time at each transect, therefore, is a surrogate for the residence time distribution for flowpaths that intersect the geophysical transect. Results clearly demonstrate that the PDF at a transect does not fit common exponential or power-law distribution assumption. While the integration of assumed distributions (i.e., exponential, power law) may provide adequate fits for reach-scale breakthrough curves, the behavior of individual flowpaths at a given location is much more complicated. Analysis of individual pixels provides thousands of individual breakthrough curves in the subsurface; distillation into PDFs allows for rapid synthesis of this extensive data set. PDFs show wide ranges of arrival time, variance, and skewness at individual pixels (each of which represents a spatially averaged suite of flowpaths intersecting the ER transect), indicating that a single, idealized distribution may not be sufficiently complex to explain all behavior.

The geophysical monitoring of solute tracer in the subsurface is sensitive to a much larger spatiotemporal distribution of flowpaths than common transient storage modeling reflects. Even after the stream has returned to background conditions (within 60 hours elapsed), the ER images still show a substantial tracer presence in the subsurface for all injections [Chapter 4]. Because the pixel resolution and peak resistance change are different for each injection and spatial
location, PDFs were constructed from a different spatial coverage in the subsurface. The selection of pixels was designed to only analyze meaningful pixels (i.e., those where tracer was observed and confidence in resolution value is highest), as opposed to choosing a constant spatial domain where pixel resolution and sensitivity to solute transport were not considered. Thus, the PDFs inform the suite of labeled flowpaths that were detected by ER; it is important to recognize that we have chosen to analyze the flowpaths as a whole rather than a uniform spatial domain.

Temporal moment analysis of ER images provides a spatially distributed assessment of solute transport in the subsurface. Analysis of this novel data set allows for both qualitative assessment of spatial patterns in transport, and for quantitative assessment of hyporheic transport patterns (e.g., spatial distributions of temporal moment analysis, first arrival times, and hyporheic persistence). Distributed transport process information could be used to improve existing groundwater flow models. Indeed, numerical experiments demonstrate the potential to derive hydraulic conductivity fields from ER inversions and constrain model formulations for subsurface architecture [Pollock and Cirpka, 2010; Chapter 3]. Changes in spatial moment analysis, such as those presented in Chapter 4, can also be used to parameterize advection-dispersion models of subsurface flow and transport [Cirpka and Kitanidis, 2000].

6.6 Conclusions

A recent numerical study by Ward et al. [2010a; Chapter 5] concluded that future work in spatially distributed characterization of hyporheic transport should focus on implementing similar analyses in field settings, where heterogeneity and sensitivity would be subject to the noise and distribution common to working in natural systems. Results in Chapter 4 demonstrated that the
ER data analyzed here was, in fact, representative of the solute moving through the subsurface based on a comparison with monitoring well data.

ER imaging provides an in-situ analysis of subsurface solute transport with high spatial and temporal resolution. This novel data set shows the movement of tracer through hyporheic pathways at spatial and temporal scales (i.e., higher resolution) than could be characterized with in-stream tracers and/or monitoring wells alone. Our ability to collect this high-resolution data in replicate studies during baseflow recession allowed us to characterize spatial trends in solute transport and their dynamics in response to changing baseflow conditions. This study is the first, to the best of our knowledge, to analyze the spatial distribution of transport processes a hyporheic zone based on ER images monitoring these processes in-situ. PDFs of each temporal moment were constructed to quantify differences between transects under each flow conditions and to compare behavior between injections completed throughout the baseflow recession period. The increasing spread of PDFs for first arrival time and persistence through baseflow recession suggests increased diversity of hyporheic flowpaths during lower flow conditions.

We demonstrated the potential for coupled ER and solute studies to characterize hyporheic transport in-situ. With this advance, we reiterate here the inherent uncertainties in the geophysical methods employed that cannot be ignored: (1) differences in support volumes of pore fluid concentration and electrical resistivity data, (2) inaccuracies in petrophysical relationships to convert between ER pixel resistivity and pore fluid concentration, (3) measurement errors associated with ER data, (4) non-uniqueness of ER inversion models, and (5) limitations of ER images applied to time-lapse problems, where timesteps are either inverted independently (limiting consistence between timesteps) or using previous results as the starting model for a given timestep (where errors in a given timestep may be propagated forward) [after Slater et al., 2002; Day-Lewis et al., 2005]. Despite these limitations, the application of ER techniques allows
for the unprecedented observation and spatial discretization of hyporheic transport processes. As such, investigations of subsurface dispersion, diffusion, and mobile-immobile exchange are now possible. Existing techniques employing only solute tracers and a limiting monitoring well network cannot provide this level of spatial resolution nor characterize the array of behavior observable with geophysical imaging.

6.7 Acknowledgements

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Chapter 7

Synthesis: On the use of electrical geophysics to characterize hyporheic transport

Field and numerical studies presented here (Chapters 2-6) demonstrate the application of coupled solute tracer and electrical resistivity imaging to provide both qualitative and quantitative assessment of hyporheic transport adjacent to headwater streams. The studies are organized into categories of either a focus on spatial patterns of hyporheic exchange (Chapters 2-4) or characterizing hyporheic transport processes (Chapters 5-6). The net result of this work is the establishment of electrical geophysics as a useful tool to characterize spatiotemporal exchange processes in stream-aquifer systems.

Interpretation of electrical resistivity images overcomes the limitations of in-stream tracer studies (limited window of detection, spatially lumped, not based on physical transport processes) and monitoring well networks (spatially sparse). The spatially distributed, high temporal resolution data sets provide unparalleled in-situ assessment of solute transport; studies need not rely on downstream observations to infer upstream behavior. Furthermore, these studies open the “Pandora’s box” of heterogeneity of both the subsurface conditions (i.e., heterogeneous fields of hydrogeologic parameters, subsurface architecture) and solute transport. The advantages gained by ER monitoring of solute transport are not without flaw, however. Images are oversmoothed approximations of the subsurface, and are sensitive to decisions made during both collection and inversion of the data.
7.1 Subsurface architecture as a control on hyporheic exchange

The spatial coverage provided by ER images of hyporheic transport provides the most extensive data set to-date to understand how subsurface heterogeneity and structure act as controls on hyporheic exchange. Whereas past studies have relied on numerical models of flow and transport parameterized by limited data sets (i.e., hydraulic conductivity from sparse monitoring well networks, limited observations of depth to refusal, assumption of homogeneity and isotropy) or destructive sampling of the subsurface during dry conditions [e.g., soil cores by Cardenas et al., 2003], ER images allow direct observation of how hyporheic flowpaths respond to subsurface features interact with hydraulic gradients to control hyporheic networks. In Chapter 2, subsurface structure is interpreted from time-lapse ER images. Features including a confining bedrock layer and suspected preferential flowpath were identified based on both background images and time-lapse images of transport.

The modeling exercise in Chapter 3, constrained by field observations, shows how increasingly complex subsurface architecture, derived from a variety of investigation methods, changes predicted subsurface residence time distributions and streambed flux. Imaging of solute transport allowed more accurate characterization of subsurface architecture than was possible from other methods (including depth to refusal, seismic refraction, and background ER imaging). This is most notable at the downstream end of the study reach where geophysical parameter structure did not appear to occur with hydrogeological parameters observed in ER images. This study illustrates a common limitation of hydrogeophysics: the lack of agreement between spatial distributions of geophysical properties (i.e., electrical resistivity) and hydrogeologic properties (i.e., hydraulic conductivity), and also a potential solution (i.e., coupled solute studies and geophysical monitoring).
Finally, the field study of spatial patterns in hyporheic exchange during baseflow recession (Chapter 4) allows interpretation of subsurface structure in the catchment (e.g., the location of subsurface confining units). ER images and spatial moments show a vertical confinement of hyporheic flowpaths in the upper end of the study reach, whereas this was not observed in the more extensive alluvial deposit in the lower transects of the study reach. Past studies in the reach relied on solute tracers and a monitoring well network, but were only able to infer transport behavior between wells. The application of ER imaging and analysis of the resultant data set provided the resolution necessary to observe, rather than infer, the role of subsurface heterogeneity and architecture in transport behavior.

7.2 Hyporheic response to baseflow recession in a headwater stream

The response of hyporheic flowpath network extent and transport to baseflow recession was explored in Chapters 4 and 6. Hydrologic controls on hyporheic exchange have been studied in numerical models and using solute tracer studies in the field; such studies generally conclude that hyporheic zones contract during periods when hydraulic gradients from the groundwater table to the stream are largest. However, these studies have relied on highly parameterized or idealized models of solute tracer studies to make these observations; few data exist demonstrating this phenomena in-situ. The application of ER imaging in Chapter 4 provides both qualitative images of spatial patterns in flowpath networks and quantitative interpretation of hyporheic cross-sectional area. This work found no apparent trends between hyporheic extent and cross-valley, down-valley, or vertical hydraulic gradients. Geophysical observations demonstrate the widely theorized and reported hyporheic contraction during gaining conditions is not evident at the study site.
Chapter 5 demonstrates the utility in temporal moment analysis to infer spatial distributions of transport processes (e.g., advection, dispersion, transient storage) in the subsurface. In Chapter 6, temporal moment analysis of ER images was applied to field data, allowing us to assess changes in hyporheic transport through the baseflow recession season. Later mean arrival times, increasing dispersion (i.e., spreading of the tracer through time at a given location), and increasing tailing during the recession period due to transient storage along a flowpath demonstrate that changes in dispersion and mobile-immobile exchange occur along hyporheic flowpaths, in addition to spatial changes in the flowpath network itself. The development of probability density functions from the temporal moments distills tens of thousands of pixel observations into a generalized form, allowing comparison both spatially (along the stream reach) and temporally (under varying flow conditions).

7.3 Toward a process-based understanding of riparian hydrology

Taken together, the changes in hyporheic extent and transport patterns demonstrate a highly dynamic hyporheic zone, responding to controls including lateral and vertical hydraulic gradients, subsurface heterogeneity, and head at the stream boundary condition. The studies included in this dissertation provide, for the first time, spatially distributed evidence of hyporheic response to changing hydraulic gradients. These techniques represent one tool in a rapidly expanding suite of technological advances that enhance our understanding of hyporheic processes. Limitations of ER methods are offset by the benefit of spatial completeness. Application of these methods in addition to complimentary techniques (i.e., those whose strengths are spatially accurate measurements, and whose weakness are a lack of spatial coverage) will
enable researchers to better understand hyporheic transport processes. This necessary step is the foundation for an improved management of coupled surface-subsurface hydrological systems.

7.4 Future prospects and implications

With the establishment of electrical resistivity imaging as a method to characterize hyporheic flow and transport near small streams, there are several future directions for research. The increased spatial coverage in the subsurface provides a new data set that could be used to constrain three-dimensional models of hyporheic flow and transport. Such models would allow for assessment of transport along individual flowpaths. This knowledge coupled with sampling for biogeochemically reactive solutes (for example, coupling a $^{15}$N-labeled nutrient study with a conservative tracer) could be used to characterize the distribution of biogeochemical reaction rates distributed in the subsurface. Finally, this data set can be used to verify (or reject) widely applied box-models of flow and transport by comparing multiple methods to assess hyporheic exchange.

One direct application of this method is in stream restoration projects that would focus on the creation of hyporheic exchange [e.g., Ward et al., In Press]. The effectiveness of such designs is directly related to the flux and residence time of solutes in the hyporheic zone, yet quantifying these metrics is difficult with traditional methods. The use of ER imaging and characterization of flowpaths could be used to test effectiveness of structures in the field, and to verify their performance under variable hydrologic conditions. The techniques demonstrated in this manuscript are the simplest means to characterize subsurface solute transport, and would be a useful contribution to post-restoration monitoring.
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You have received this email because you are a co-author on one or more manuscripts that will result from the work included in my dissertation. Penn State University requires that, for published material or that which is shortly expected to be published, I obtain permission from co-authors. I have listed the chapters subject to this requirement below, including their current status and co-authors.

Chapter 2: Imaging hyporheic zone solute transport using electrical resistivity
Status: Published in Hydrological Processes: HPToday (2010)
Co-authors: Mike Gooseff, Kamini Singha

Chapter 3: Geophysical characterization of subsurface heterogeneity to inform models of groundwater flow and transport
Status: To be included in draft form
Co-authors: Mike Gooseff, Kamini Singha

Chapter 4: Patterns of hyporheic exchange during baseflow recession in headwater mountain streams
Status: To be included in draft form
Co-authors: Michael Fitzgerald, Mike Gooseff, Kamini Singha, Tom Voltz, Andrew Binley

Chapter 5: Characterizing hyporheic transport processes - Interpretation of electrical geophysical data in coupled stream-hyporheic zone systems during solute tracer studies
Status: Published in Advances in Water Resources (2010)
Co-authors: Mike Gooseff, Kamini Singha

Chapter 6: Spatially distributed characterization of hyporheic solute transport processes using electrical resistivity imaging and temporal moments at the field scale
Status: To be included in draft form
Co-authors: Mike Gooseff, Kamini Singha, Michael Fitzgerald, Tom Voltz

I am requesting that you reply to this email if you agree that this information can be published in my dissertation, and that I was the primary author for each chapter. Please sign your email with your full name - I will include these responses in an appendix as proof of your agreement.

Thank you for your continued support - I look forward to working with you all to get the next round of manuscripts submitted in the coming months.

Adam

--------------------------------------------------
Adam S. Ward, PE, CFM, LEED-AP
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University Park, PA 16802
http://web.me.com/adamsward
--------------------------------------------------
Hi Adam,

Looks good. Keep me posted for how I can help. I know we'll discuss author order later, but I often prefer to be last on student papers, unless Mike wants that position. We can talk later! In any regard, I agree that you are certainly first author on all below.

Good luck wrapping up.

Cheers,

Kamini Singha

--
Kamini Singha
Assistant Professor
Dept. of Geosciences
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311 Deike Building
University Park, PA 16802
phone: +1 814.863.6649
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web: http://www.geosc.psu.edu/~ksingha
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http://web.me.com/adamscottward
-----------------------------------------------
Hi Adam,

You have my permission to include this work in your dissertation, and I confirm that you were the primary author on the manuscripts that I'm a co-author on. Good luck with finishing up.

cheers,
Michael R. Fitzgerald

Michael Fitzgerald
Postdoctoral Researcher
The Pennsylvania State University
Dept. of Geosciences
University Park, PA 16802
Phone: +1.814.441.1566
Email: mxf218@psu.edu

On May 23, 2011, at 11:33 PM, Adam Ward wrote:

Mike, Kamini, Tom, Michael, and Andy -

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Adam

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University Park, PA 16802
http://web.me.com/adamscottward

*******************************************************************************
Adam,

You have my permission to publish all work aided by my contributions, and I agree that you are the primary author for each chapter listed.

Tom J. Voltz

On Tue, May 24, 2011 at 5:33 AM, Adam Ward <asw178@psu.edu> wrote:

Mike, Kamini, Tom, Michael, and Andy -

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--------------------------------------------------
Hi Adam,

I'm happy with your suggested inclusion of co-authored papers.

I'm looking forward to seeing the thesis is entirety.

Best wishes

Andy

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LA1 4YQ, UK

Room: LEC 5539
Tel: +44 (0)1524 593927
Fax: +44 (0)1524 510217
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---

From: Adam Ward <asw178@psu.edu>
Sent: 24 May 2011 04:34
To: Mike Gooseff; Kamini Singha; Tom Voltz; Michael Fitzgerald; Binley, Andrew
Subject: Co-authorship on Dissertation Chapters

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VITA

Adam S. Ward

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Penn State University, University Park, PA
   Doctorate of Philosophy in Civil Engineering (minor in Geosciences), 2011.

Michigan State University, East Lansing, MI

Michigan Technological University, Houghton, MI
   Masters of Science in Civil Engineering, 2006.

Michigan Technological University, Houghton, MI
   Bachelor of Science in Civil Engineering (minor in Enterprise), 2005. *Summa cum laude.*

Experience

Department of Civil & Environmental Engineering, Penn State University, University Park, PA
   Graduate Research Assistant, Fall 2009-Summer 2011
   Instructor for Fluid Mechanics (CE 360), Summer 2009
   Graduate Teaching Assistant, Fall 2008 – Spring 2009

Spicer Group, Inc., Saginaw & St. Johns, MI
   Design Engineer, 2005-2008

Performance Engineering, Inc., Charlevoix, MI
   Engineering Technician, 2003

Recent Publications

