ABSTRACT

Hyporheic exchange flow involves the two-way movement of water between the stream channel and the bed and banks. These exchange flows create distinctive habitats and influence biogeochemical processes and water temperatures. This study focused on the characterization of the spatial distribution of subsurface flow pathways and associated travel times through the hyporheic zone within a low-order, high-gradient headwater stream located in the UBC Malcolm Knapp Research Forest, approximately 60 km east of Vancouver, British Columbia. Hyporheic zone processes were examined May to October 2006, at three spatial scales in East Creek: point, channel-unit and reach.

Hydrometric data collected from piezometers installed within the stream channel, along with solute injection tracer experiments, were used to characterize subsurface flow pathways within a 100 m stream section. Stream tracer breakthrough curves were used to model the processes of advection, dispersion, lateral inflow and transient storage within the hyporheic zone using the numerical model OTIS-P. Tracer injections at individual step-pool units were used to identify locations of hyporheic discharge, as well as to estimate separate travel times for hyporheic and surface-water transient storage zones.

Solute transport process varied with discharge at the reach scale. Transient storage area ($A_s$) increased with discharge, while transient exchange coefficient ($\alpha$) remained fairly constant. At the scale of individual pools, transient storage area and residence times were higher than the reach scale estimate, suggesting that pools and back eddies do contribute to transient storage in headwater streams. Water fluxes calculated with Darcy's Law in one channel-unit did not “scale-up” to the reach scale estimate of hyporheic exchange ($\alpha$), and was two orders smaller than the reach scale. Direct measurements of water fluxes into the streambed, including vertical hydraulic gradients and infiltration rates, did not vary systematically with discharge. Hydraulic gradients varied significantly with scaled location within the channel-unit, but not with the downstream step height. Hydraulic conductivity varied with site conditions (upwelling, downwelling and neutral sites), suggesting that channel geometry and hydraulic conductivity control exchange flow. This multiple scale approach highlights the considerable spatial and temporal variability and complexity of hyporheic exchange processes within step-pool streams.
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La mia famiglia
DEDICATION

For Claire and Kees, the Adventurers
CHAPTER ONE: INTRODUCTION

1.1. Hyporheic exchange flow in small streams

Hyporheic exchange flow involves the two-way movement of water between the active stream channel and subsurface sediments in the stream bed and banks. These exchange flows play an important role in the functioning of stream ecosystems. Frequent hyporheic exchange keeps stream water in close contact with chemically or biologically active stream bed sediments, which increases the opportunities for biogeochemical processing. Interactions between the stream channel and subsurface influence water quality by generating gradients of nutrients and dissolved gases (Boulton et al. 1998) and regulating water temperatures (Moore et al. 2005a). The hyporheic zone plays an important role in ecosystem functioning including stream metabolism (Mulholland et al. 1997), nutrient retention and cycling (Triska et al. 1989, Wondzell and Swanson 1996b), habitat for benthic invertebrates (Stanford and Ward 1988), and general ecosystem stability (Valett et al. 1994).

Spatial heterogeneity in exchange flow pathways results in a hydrologically linked region beneath and adjacent to streams and rivers where surface water and subsurface ground water mix (Triska et al. 1989, Stanford and Ward 1993, Findlay 1995). This mixing zone is defined as the “hyporheic zone,” in which the water chemistry reflects a mixture of streamwater and groundwater. Exchange flows through the hyporheic zone link aquatic and terrestrial components of the riparian ecosystem (Wondzell and Swanson 1996b).

Early attempts to define the physical boundaries of the hyporheic zone were based on the distributions of aquatic invertebrates, including hypogeon (groundwater origin) and epigeon (channel origin). Triska et al. (1989) used solute patterns to operationally define the boundaries of the hyporheic zone as the depth to which greater than 10% advected channel water and less than 90% groundwater is present. Beneath the hyporheic zone is the groundwater zone where the water chemistry is not influenced by stream water. Hyporheic zones are linked to a nested series of flow paths that can travel both
laterally and vertically through subsurface flow paths, rather than entering the stream bed in one location as does groundwater (Harvey and Bencala 1993, Harvey et al. 1996, Harvey and Wagner 2000, Kasahara and Wondzell 2003).

Hyporheic exchange creates distinct zones of aquifer discharge (upwelling or outwelling) from the sediments into the stream channel, and recharge (downwelling) from the stream channel into the saturated sediments. Upwelling water can supply stream organisms with the nutrients which influence primary production, while downwelling water can provide dissolved oxygen and organic matter to benthic invertebrates living in the sediments (Boulton et al. 1998). Downwelling water also provides oxygen to fish eggs in the subsurface (Baxter and Hauer 2000).

Small streams (<10 m width) with active exchange between surface and subsurface waters (hyporheic exchange) are thought to facilitate nitrogen-removal and reduce the export of nitrate \( (\text{NO}_3^-) \) downstream (Triska et al. 1989, Jones and Holmes 1996, Duff and Triska 2000). Nitrate is considered a major pollutant of aquatic systems in much of the northern hemisphere (NRC 2000). The competing processes of nutrient retention and hydrological export are expressed in the nutrient spiraling concept (Webster and Patten 1979, Newbold et al. 1981). Valett et al. (1996) proposed a conceptual model suggesting that the solute retention is a product of chemical transformation rates and surface-subsurface interactions which increase residence times. Additional studies support the theory that the extent of subsurface interactions influences solute retention (Mulholland et al. 1997, Hill and Lymburner 1998). Solute retention and residence time in the subsurface depend primarily on hydraulic gradients and hydraulic conductivity.

The hyporheic zone increases the residence time for water within the stream ecosystem and enhances transient storage (Bencala 1984). Transient storage refers to the temporary detainment of solutes in slow moving areas such as side pools or back eddies relative to the faster flowing areas in the main channel. Solutes, such as nitrogen and phosphorus, may also enter the permeable substrate surrounding the stream (i.e. hyporheic zone) and travel at a slower velocity than that of the main channel. Solutes detained within transient storage zones are eventually re-released back into the main channel, but at a slower rate relative to solutes traveling at the advection rate in the main channel (Jones and Mulholland 2000). In the main channel, solutes are transported...
through the hydrological processes of advection and dispersion (Stream Solute Workshop 1990). Advection refers to the downstream transport of solute mass at the mean velocity of the streamwater. Dispersion is the spreading of the solute mass due to shear stress and molecular diffusion in the downstream direction.

Interactions between the stream channel and subsurface have been a focus of research for over two decades (Jones and Mulholland 2000). Natural riffle-pool and step-pool units have been commonly studied as channel morphology that exerts a strong control on hyporheic exchange. Current knowledge of hyporheic zone processes is based largely on studies conducted within small (<10 m width), alluvial, low to mid order gravel-bed, headwater streams with riffle-pool (e.g. Triska et al. 1989, Harvey et al. 1996, Hill et al. 1998, Wagner and Bretschko 2002) and step-pool morphologies (e.g., Harvey and Bencala 1993, Wondzell and Swanson 1996a, Valett et al. 1996, Haggerty et al. 2002, Kasahara and Wondzell 2003, Storey et al. 2003, Anderson et al. 2005), as well as within high-order, low gradient, braided to meandering, river systems (e.g. Stanford and Ward 1988, Boano et al. 2006).

This thesis is concerned with the characterization of the spatial distribution of subsurface flow pathways and associated residence times through the hyporheic zone within a low-order, high-gradient headwater stream. Section 1.2 reviews the current literature on the physical controls of hyporheic exchange to provide the background for the specific research objectives presented in Section 1.3.

1.2. Physical controls on exchange flow

Surface-subsurface interactions, or hyporheic exchange, are driven by variations in hydraulic head gradients as a result of instream structural complexity created from large woody debris (e.g. log jams) and geomorphic features such as step-pool sequences, or breaks in topography (Harvey and Bencala 1993). Hyporheic flow varies considerably over space scales (1 cm – 100 m) and time scales (10 s – 100 days) and at various rates through different types of substrate (Harvey et al. 1996). Depending on local geology and channel morphology, the extent of the hyporheic zone can range in length from centimeters to hundreds of meters (Stanford and Ward 1993). Local channel features
including sediment composition, permeability (hydraulic conductivity) and the bed topography control the lateral extent of the hyporheic zone below the saturated stream channel (Triska et al. 1993). Exchange flows occur at different spatial scales including those of (1) individual bedforms, (2) channel units and (3) reaches. The following sections will further examine the exchange flow processes occurring at the different spatial scales.

1.2.1. Bedform scale

Topographic features known as bedforms (riffles and dunes) develop due to streamflow over a loose sediment bed. Obstructions or irregularities in the streambed such as sand riffles (Johnson 1980) or even fish redds (Tonina and Buffington 2005) create a high-pressure zone upstream of the obstruction and a low pressure region downstream. Flume studies conducted at this scale have shown that flow is induced by pressure imbalances generated from gradients of temperature, density and hydrostatic head (Thibodeaux and Boyle 1987, Elliott and Brooks 1997). The process of solutes and water flowing between high-pressure and low-pressure zones in the bed is referred to as “advective pumping exchange” (Savant et al. 1987, Thibodeaux and Boyle 1987). Water also moves through the sediments through the process of “turnover”. Turnover occurs as moving bedforms trap and release interstitial fluid.

Studies in laboratory flumes have indicated that hyporheic exchange rates increase with discharge, specifically with stream flow velocity, due to an increase in the pressure difference between high and low pressure regions (Thibodeaux and Boyle 1987, Elliott and Brooks 1997). Wondzell (2005) speculated that interactions between streamflow and channel bedforms must drive exchange flow in headwater streams; unfortunately field studies have not been able to incorporate finer scale effects (as studied in flumes) into a cohesive framework for exchange in headwater streams. As a result, the influence of individual bedforms on hyporheic exchange flow has not been investigated within a field setting.
1.2.2. **Channel-unit scale**

At the scale of individual channel units, morphological features in the stream channel, such as large woody debris, create head gradients that drive advection of stream water through the hyporheic zone. Several authors have shown that longitudinal gradients in step-pool and riffle-pool sequences drive small scale exchange flow both vertically and laterally (Harvey and Bencala 1993, Hill et al. 1998, Storey et al. 2003, Anderson et al. 2005, Gooseff et al. 2006).

Hyporheic exchange has been described conceptually as short pathways that enter the subsurface and return to the stream channel at multiple locations (Harvey and Bencala 1993). Stream water flowing through well-defined flow pathways in the alluvium may enter a streambed at the top of a riffle or step and then return to the stream a short distance downstream in the bottom of pool (Figure 1.1).

Flow begins when the total head in the surface channel is greater than that in the subsurface, resulting in a negative vertical hydraulic gradient (VHG), which drives stream water down into the subsurface sediments. Surface water may mix with or displace groundwater and eventually return to the surface where the total head in the subsurface is greater than the stream channel. Hyporheic exchange creates distinct zones of aquifer discharge (upwelling or outwelling) from the sediments into the stream channel, and recharge (downwelling) from the stream channel into the saturated sediments. Flow paths may travel vertically between the channel to the subsurface or laterally from the adjacent riparian zone (White 1993, Findlay 1995).

Wondzell and Swanson (1996) along with Kasahara and Wondzell (2003) further identified channel-unit features which drive hyporheic exchange, such as side channels, meander bends, gravel bars and boulder or log-steps. Kasahara and Wondzell (2003) found that steps accounted for approximately 50% of the exchange flows in second and fifth-order stream reaches based on the results of a sensitivity analysis using groundwater flow models. Channel morphology has also been documented as a significant control for lateral hyporheic exchanges (Vervier et al. 1993, Morrice et al. 1997, Storey et al. 2003). These studies suggest that channel-unit form and geometry are significant controls on hyporheic exchange flow.
Studies examining hyporheic exchange flow at the channel-unit scale typically employ a hydrometric approach. This approach requires an extensive network of piezometers and/or wells to measure hydraulic gradients and hydraulic conductivity in order to characterize and map exchange flows. Studies are therefore limited to a small spatial area (e.g. Harvey and Bencala 1993), and inter-reach comparisons are challenging.

As a result, Bencala (2000) expressed the need to identify the physical and hydrometric properties of the stream system that contribute to solute transport within the hyporheic zone, and that can be routinely measured or map across spatial scales. Recent studies have begun to examine channel-unit spacing in stream longitudinal profiles to predict the spacing between zones of upwelling and downwelling in step-pool and pool-riffle morphologies using a hydrometric (Anderson et al. 2005) or modelling approach (Gooseff et al. 2006).

In both these studies, channel-unit spacing, size and sequence were considered important controls in determining exchange patterns of upwelling and downwelling. These results suggest that a scaling relationship to identify zones of upwelling and downwelling based on channel-unit geometry would be a useful tool for characterizing
and predicting exchange flow in step-pool streams. An objective of this thesis is to develop a geometric scaling relationship relating vertical hydraulic gradients to channel-unit geometry including pool length and downstream step height, in order to determine if hyporheic discharge and recharge zones are a function of stream channel location.

While it is known that structural complexity from large woody debris and geomorphic features such as step-pool sequences drive exchange flow, there are still uncertainties regarding the spatial patterns of hyporheic flow, and the locations where hyporheic water discharges back into the stream. Recent studies of hyporheic exchange in steep, headwater streams in the Lookout Creek basin (Oregon, USA) have generally not observed coherent upwelling of hyporheic water below steps, as described by the typical flow pathway, despite predictions from groundwater flow models that upwelling should occur (Anderson et al. 2005, Gooseff et al. 2005, Wondzell 2005).

These results suggest that hyporheic discharge occurs under different mechanisms, including lateral inflow or outwelling from other locations within the step-pool. For example, Moore et al. (2005b) observed upwelling sites within a concentrated zone of lateral inflow that was consistent with the convergent topography and hillslope of a section of headwater stream in coastal British Columbia. Solute tracer tests indicated that these upwelling sites underwent little to no mixing with water from the stream channel, suggesting that upwelling sites are a result of lateral inflow. In this scenario it is hypothesized that flow pathways could include an interaction between groundwater or lateral inflow and hyporheic exchange pathways.

An objective of this thesis is to determine which model best conceptualizes exchange flow within steep, step-pool streams. Subsurface flow pathways that create zones of discharge (i.e. upwelling or outwelling) and recharge (i.e. downwelling) within step-pool units can be described using three different conceptual models:

1. **Model 1a** – Represents a typical flow pathway in which downwelling of water occurs at the top of a riffle or step and returns to the stream channel a short distance downstream in the bottom of pool. Flow is aligned with the channel creating upwelling in the pool. Models 1b and 2 represent possible alternative hypotheses to the typical flow pathway and could explain the lack of observed upwelling in previous studies.
2. **Model 1b** – Hyporheic flow is aligned with the channel and driven by vertical hydraulic gradients as per Model 1a, but flow creates upwelling or outwelling sites at another location within the step-pool, such as directly below the step. Models 1a and 1b do not include a lateral inflow component, which is represented by Model 2.

3. **Model 2** – Hyporheic exchange flow includes a lateral inflow component in which zones of upwelling are a result of lateral inflow from the riparian zone and adjacent hillslope due to convergent topography. As well, flow pathways could include an interaction between groundwater or lateral inflow and hyporheic exchange pathways. For example, hyporheic water could flow laterally into the riparian zone after infiltrating in a step, then flow laterally into the channel.

### 1.2.3. Reach scale - Transient storage processes

Hyporheic exchange processes are typically studied at the reach scale using a transient storage model (TSM) consisting of a one-dimensional advection-dispersion equation with an additional term for transient storage (Bencala and Walters 1983, Runkel 1998). The TSM provides reach-scale estimates of the solute transport processes of advection, dispersion, transient storage and lateral inflow, by simulating the breakthrough curves (BTC) generated from stream tracer injections (e.g. D’Angelo et al. 1993, Harvey and Bencala 1993, Morrice et al. 1997, Mulholland et al. 1997, Gooseff et al. 2003). Additionally, TSMs have been widely applied to link hydrological transport parameters to biological processes (e.g. Valett et al. 1996, Mulholland et al. 1997, Gooseff et al. 2004).

Transient storage includes (1) in-channel storage in “dead zones” (i.e. side pools, back eddies) and (2) storage in the hyporheic zone. Current TSMs, such as the One Dimensional Transport with Inflow and Storage (OTIS-P; Runkel 1998) model, lump both transient storage zones together into a single model-derived estimate of the cross-sectional area of the transient storage zone (A<sub>S</sub>). Exchange between the main channel and the transient storage zone is controlled by a transient exchange coefficient (\(a\)), which is considered an estimate of hyporheic exchange. Typically, the model-derived transient
storage area ($A_s$) parameter is assumed to represent storage in the streambed, or the
hyporheic zone (Bencala et al. 1993), due to the inability to separate in-channel storage
from storage in the hyporheic zone. However, previous studies have suggested that both
transient storage mechanisms occur at the reach scale (Harvey and Bencala 1993,
in the transient storage residence-times estimated from stream tracer experiments and
hydrometric well breakthrough dynamics, and concluded that at the reach-scale solute
transport processes are sensitive to both transient storage mechanisms.

Attempts to distinguish between in-channel storage and hyporheic transient
storage have resulted in the development of TSMs with two storage or multiple
exponential residence time distributions (Castro and Hornberger 1991, Choi et al. 2000,
Gooseff et al. 2004). However, these models are often difficult to correctly parameterize
using current field techniques (Runkel 2002). As a result, there is a need to better
understand processes at the point and channel-unit scale for interpretation of reach scale
tracer tests, particularly in terms of the relative roles of hyporheic exchange and transient
storage in pools.

Solute detained within dead zones are eventually re-released back into the main
channel, but at a slower rate than the main solute pulse (Jones and Mulholland 2000).
This slow release is manifested by a long tail in the breakthrough curves (Bencala and
Choi et al. 2000). The OTIS-P transient storage model assumes that the late-time
residence times of solutes within these dead zones are exponentially distributed (Runkel
1998). However, recent workers have shown residence time distributions that are better
characterized using a lognormal distribution (Wörman et al. 2002), or displayed scale-
invariant behavior, with the underlying residence time following a power law distribution
(Haggerty et al. 2002).

Gooseff et al. (2003) characterized exchange processes in three reaches at
Lookout Creek basin using two solute transport models with different residence time
distributions (exponential and power-law). Transient storage within a bedrock reach
followed an exponential distribution with a mean residence time of three hours. In
contrast, transient storage in an alluvial reach followed a power law distribution with a
mean residence time of >100 hours. Although this study makes the assumption that the alluvial reach represents both in-stream and transient storage processes, whereas the bedrock reach represents only in-channel transient storage due to the absence of a hyporheic zone, these results show that in-channel features such as pools and back eddies do contribute to transient storage in small, headwater streams.

Harvey et al. (1996) assumed that the residence-time for in-channel transient storage to be very short and is therefore accounted in the dispersion coefficient rather than the transient storage coefficient. However, Gooseff et al. (2003) highlights the variability in residence times and late-time distributions that can occur depending on the dominant transient storage mechanism at the reach scale. Although it is sometimes possible to eliminate one of the transient storage mechanisms based on morphology (i.e. alluvial versus bedrock reach), studies have not been able to isolate the relative contribution from each transient storage zone in stream reaches where both processes occur together. As well, the contribution from in-channel storage has only been examined at the reach scale, and not at the channel-unit scale within individual pools using a stream tracer approach. In stream reaches where pools are located, no previous study has specifically treated an individual pool as a “stream reach” to test the assumptions of the transient storage model. This application of the TSM may help to characterize the spatial extent and residence times of transient storage as measured from stream tracer experiments. A key objective of this thesis is to quantify transient storage and estimate separate travel times for hyporheic and surface-water transient storage zones.

Several studies have sought to identify the factors that control transient storage and hyporheic exchange, including channel morphology (Harvey and Bencala 1993, D’Angelo et al. 1993, Kasahara and Wondzell 2003; Gooseff et al. 2003, Wondzell 2005), discharge (Harvey et al. 1996, Wondzell 2005, Zarnetske et al. 2007), parent material (Morrice et al. 1997), groundwater inflows (Harvey and Bencala 1993, Wroblicky et al. 1998) and stream complexity (Patschke 1999, Gooseff et al. 2007).

D’Angelo et al. (1993) used the stream tracer approach to investigate the influence of geomorphological constraint on transient storage within a fifth-order stream in the Lookout Creek basin. Transient storage was lower in constrained stream reaches (n = 5) as compared to reaches that were unconstrained (n = 2). This study also reported a
decrease in transient storage with stream order (first-order to fifth-order), indicating that transient storage is variable with spatial scale.

Morrice et al. (1997) compared three headwater streams with different parent lithologies and showed that the extent of hyporheic exchange increased with increased hydraulic conductivity. This study also suggested that conductivity and infiltration capacity influences the amount of water that infiltrates the stream bed and enters the hyporheic zone. Wondzell (2005) conducted stream tracer experiments in two small, steep-mountain streams in the H.J. Andrews Experimental Forest (Oregon, USA) at low baseflow and high baseflow conditions. This study found that transient storage varied with channel morphology and to a lesser extent with discharge.

An increase in stream discharge is hypothesized to result in a greater wetted area, which provides for enhanced hyporheic exchange. In headwater streams, the influence of discharge on hyporheic exchange and spatial extent will vary with catchment wetness, hillslope connectivity and the strength of hydraulic gradients to the stream channel (Wondzell and Swanson 1996a). Wroblicky et al. (1998) applied a 2-D groundwater flow model to study the surface-subsurface interactions within two first-order headwater streams in New Mexico (USA), and found that the spatial extent of the hyporheic zone decreased by approximately 50% during higher flows. A sensitivity analysis indicated that exchange flows were governed by the interplay of extrinsic (e.g. variability in precipitation and recharge) and intrinsic geomorphological and hydrological factors (e.g. hydraulic conductivity).

Stream tracer experiments have confirmed that solute transport processes vary with discharge at the reach scale (Legrand-Marcq and Laudelout 1985, D’Angelo et al. 1993, Harvey et al. 1996, Morrice et al. 1997, Hart et al. 1999, Patschke 1999, Wondzell 2005); although conflicting responses have been documented.

Morrice et al. (1997) compared transport processes during 4 tracer releases within one headwater stream reach, and observed a decrease in transient storage ($A_S$) while transient exchange ($\alpha$) increased with increasing discharge. Similarly, D’Angelo et al. (1993) found that the transient exchange coefficient increased with discharge and that the ratio of storage zone to cross-sectional area ($A_S/A$) decreased with discharge. Additional studies have reported similar trends (Harvey et al. 1996, Hart et al. 1999, Wondzell 2005).
2005), however trends that are in contradiction to those studies have also been

Patschke (1999) observed an increase in $A_S$ with discharge in reaches with a
greater degree of stream complexity. Hart et al. (1999) reported that transient storage
remained constant with discharge, but transient exchange increased as discharge
increased. Tracer tests by Legrand-Marcq and Laudelout (1985) showed a rapid decrease
in $A_S$ as discharge increased to a threshold value of 2 L/s, after which transient storage
remained constant to a value of 12 L/s. Legrand-Marcq and Laudelout (1985) and
Patschke (1999) observed that the transient exchange coefficient remained constant with
discharge.

These conflicting results in the literature highlight the uncertainty surrounding the
response of transient storage area and exchange parameters to discharge in small
headwaters streams. Part of this uncertainty is attributed to the TSM, which is sensitive to
experimental design (Wagner and Harvey 1997) and parameter fitting routines (Runkel
1998). Harvey et al. (1996) explained that because stream velocity increases with
discharge, a majority of the observed differences in parameters for intra-reach or inter-
reach comparisons will result from a switch in dominance of advective flow over
transient storage, and not from a change in physical processes. For example, Wondzell
(2005) found that the results of TSM simulations were at odds to hydrometric data, which
showed little change in the location and extent of the hyporheic zone despite a four-fold
increase in discharge (1.0 – 11.5 L/s). These results are slightly confounded by high
uncertainty in parameter values as indicated by Damkohler values which were >2 (ranged
from 2.4 to 21.1) for 5 of the 9 simulated tracer experiments. However, Wondzell (2005)
cautioned that TSM comparisons should be restricted to stream tracer experiments
performed within different reaches of a single stream under comparable flow conditions.

1.2.4. Study objective and scale-dependent questions

The broad objective of this study is to examine how hyporheic exchange
processes vary spatially and temporally within a low-order, high-gradient headwater
stream under a range of flow conditions. This study focused on the characterization of the
spatial distribution of subsurface flow pathways and associated travel times through the hyporheic and surface-water transient storage zones using both a hydrometric and stream tracer approach.

A significant challenge of hyporheic zone research has been to scale up small scale physical measurements to the results of reach-scale stream tracer injections. As a result, specific questions are applied to the spatial scale of interest including the reach scale, the channel-unit scale and the local or point scale. A multiple scale approach to examining hyporheic exchange processes has not been explicitly applied in previous research. It is expected that the variability in solute transport processes observed at the reach scale, specifically transient storage area \( (A_s) \) and the transient exchange coefficient \( (\alpha) \), can be explained using observations made at smaller spatial scales.

The following research questions will be addressed in this thesis:

1. How do solute transport processes at the reach-scale, specifically transient storage area and transient exchange, vary over space and time (i.e. with discharge)? Of specific interest is the contribution of hyporheic and surface-water transient storage zones to transient storage area \( (A_s) \).

2. Can separate residence times for hyporheic and surface-water transient storage zones be quantified using a stream tracer approach at the channel-unit scale?

3. Where are zones of hyporheic discharge and recharge located within the channel-unit, and do exchange flows vary with position in the channel-unit and downstream step height?

4. How do water fluxes at the point scale vary, and can these be quantified and “scaled up” to reach scale estimates of hyporheic exchange?

The following four chapters examine the research question in detail. Chapter 2 describes the study area, field and laboratory methods, and data analysis. Results are presented in Chapter 3 and are discussed in Chapter 4. The main conclusions of the study and recommendations for future research are summarized in Chapter 5.
CHAPTER TWO: METHODS

This chapter provides a detailed description of the study location and study reaches. An overview of the study design, along with the methods used for field sampling, laboratory analysis and data analysis are also described.

2.1. Study location

This study was conducted in the University of British Columbia (UBC) Malcolm Knapp Research Forest (49° 16' N, 122° 34' W), located in the temperate Fraser Valley foothills of the Coast Mountains, approximately 60 km east of Vancouver, British Columbia (Figure 2.1).

The research forest is located within the Coastal Western Hemlock (Tsuga heterophylla) biogeoclimatic zone (Coupe et al. 1991). This zone is characteristic of the maritime climate with wet, mild winters and warm, dry summers. Mean annual precipitation at the UBC Research Forest headquarters (147 m elevation) is approximately 2184 mm (Environment Canada 1993), of which 70% falls primarily as rain between October and April due to Pacific frontal systems. Snowfall comprises only 5% of the total annual precipitation at the headquarters. The higher elevation areas are snow covered for approximately four months of the year. Streamflow is typically very “flashy” in response to rainfall. Periods of low flow dominate during the summer months (base flow <1 L/s). Periods of peak flow correspond to fall and winter storms. Runoff generation processes are dominated by subsurface flow and saturation overland flow (Hutchinson and Moore 2000).

Soils within the watershed are highly permeable, and are dominantly shallow podzols (averaging about 1 m deep) formed from glacial till. Soil parent material is glaciofluvial in origin (Klinka and Krajina 1986). Soils are typically underlain by basal till over granitic bedrock (Klinka and Krajina 1986). In some locations, the soils directly overlie bedrock. Geological conditions have contributed to the observed low background concentrations of nitrogen, phosphorus and low conductivity in surface water.
Chemically the study stream is nearly pristine with low concentrations of major ions (Feller and Kimmins 1984).

Vegetative cover along East Creek is dominated by coniferous forest consisting of Douglas-fir (*Pseudotsuga menziesii*) and western cedar (*Thuja plicata*), that are approximately 130 years old. Red alder (*Alnus rubra*) occurs along the stream banks in patches. Selective forest harvesting occurred adjacent to the stream in the 1970’s, but instream large woody debris was retained in the stream.

The study was conducted within the East Creek watershed, a 4 km long second-order stream that drains approximately 100 ha at the study reach (Figure 2.2). East Creek is located upstream from the confluence of Spring Creek. Research was conducted from May to October, 2006. Hyporheic exchange processes were examined within a 100 m long section of East Creek, which extends downstream from a culvert crossing at road M to the stream’s confluence at Spring Creek (Figure 2.3). This stream was selected due to its high structural complexity and variety of geomorphic features, which include approximately 7-10 pool-cascade sub-units within the reach (Patschke 1999). Structural complexity from in-stream large woody debris and heterogeneous substrate deposition has contributed to distinct pool-cascade sequences with both boulder and log steps.

![Figure 2.1. Site map showing location of UBC Research Forest](image)
Figure 2.2. Site map showing East Creek drainage basin
Figure 2.3. East Creek study reach with upper and lower sub-reaches and stream tracer experiment injection and sampling locations
The reach has an elevation drop of approximately 20 m (Figure 2.4), and varies in morphology and substrate along the study length (Figure 2.5, Figure 2.6). The channel width for the reach is 2.5 – 3 m wide. A large sediment dam exists at mid reach, which divides the reach into two sections. Above the sediment dam, the top 45 m of the reach has a gradient of approximately 4%. Stream morphology consists of low gradient riffle-run and pool sequences within natural log steps, and the substrate is composed of fine to coarse gravel and medium size cobbles. During periods of low flow, the stream goes subsurface through preferential flow pathways in the sediment dam (at approximately 42 m downstream from the culvert at M road). Directly below the log jam, the stream is braided for approximately 5-10 m. At the beginning of the study period, seepage from a cutbank was observed at this location.

The bottom 45 m section has a gradient of approximately 12% and contains pool-cascade sequences with boulder and log steps. Fine to medium gravels in deposition areas behind steps and large anchor cobbles and boulders contribute to a spatially heterogeneous substrate. During periods of high flow (>100 L/s), a secondary channel at approximately 55 m downstream from the culvert becomes active. Sediment transport processes in East Creek vary seasonally with discharge. Since 2003, approximately 10 to 15 sediment mobilizing events per year have been observed, of which 3 to 4 peak rainfall events during the fall and winter months of each year move the majority of sediment (J. Caulkins, pers. comm.). These events result in the greatest geomorphic changes to the channel.
2.2. Study design

This study employed a combination of hydrometric data and solute injection experiments to characterize subsurface flow pathways. Tracer experiments were designed: (1) to examine the extent of stream-subsurface interactions at the reach scale (Stream Solute Workshop 1990), (2) to measure the relative connectivity of the subsurface to the stream channel, (3) to quantify storage in individual pools during reach scale tracer injections, and (4) to provide reach scale estimates of the processes of advection, dispersion, lateral exchange and transient storage.

Stream tracer experiments were conducted at a range of flow conditions to examine how transport processes vary with discharge. Tracer injections at individual step-pool units were used to identify locations of hyporheic discharge, as well as to estimate separate travel times for hyporheic and surface-water transient storage zones. Hydrometric data included streamflow, hydraulic head, hydraulic conductivity measured using piezometers, along with direct measurements of infiltration into the stream bed.
Figure 2.5. Downstream view of log steps and piezometers in upper reach of East Creek

Figure 2.6. Upstream view of lower reach of East Creek showing steps and pool complexes
2.3. Stream discharge and geometry

2.3.1. Discharge measurements

Streamflow, or discharge, was measured using the constant-rate salt injection method (Moore 2004a). This method involves injecting a conservative solute tracer solution of known concentration into a stream at a constant rate and then measuring the stream water electrical conductivity (EC) as it becomes uniformly mixed across the stream some distance below the injection point. Downstream mixing was monitored using a conductivity probe to verify lateral and vertical mixing at a downstream location.

Tracer injections were performed according to procedures outlined by Moore (2004a). Sodium chloride (NaCl) or common table salt was used as a tracer because it is inexpensive, readily available and environmentally benign for the concentrations and durations involved in discharge measurement (Moore 2004a). Chloride is a conservative ion that occurs in low background concentrations (0.6-1.2 mg/L) in East Creek (Feller and Kimmins 1984). Sodium chloride was injected into East Creek to raise the concentration above background until the concentration reached plateau (approximately 2-3 mg/L). Solutions of NaCl and water were mixed in the laboratory using pre-weighed bags of salt. In the field, solutions were vigorously shaken to ensure that the added salt was completely dissolved. Solutions were prepared at concentrations below the saturation point of 238 g/L (Webster and Ehrman 1996).

For injection trials prior to August 11, solutions were added at a constant rate using a Mariotte bottle (Story et al. 2003, Moore 2004b). The Mariotte bottle consisted of a 30 L carboy sealed at the top with a rubber stopper. The release rate was controlled using a pipette connected to a spigot at the base of the Mariotte bottle. Release rates were measured for each tracer experiment both pre-injection and post-injection since changes in barometric pressure can affect the release rates (Webster and Ehrman 1996).

Beginning on August 11, a battery-operated Solinst Model 410 peristaltic pump connected to a 30 L drum was used to conduct longer duration injections during base flow conditions. A voltage regulator attached to the 12 volt battery was used to prevent fluctuations in the release rate due to changes in the battery voltage.
Electrical conductivity was measured as a surrogate for tracer concentration during stream tracer tests. EC increases linearly with salt concentration and is considered an economical alternative to direct measurements of chloride (Gooseff and McGlynn 2005, Wondzell 2005). EC was measured at the downstream end of the stream reach during discharge measurements using either a WTW™ LF340, 340i or 350i conductivity meter. A non-linear calibration built into the WTW™ meter was used to correct automatically the EC values to a standard temperature of 25°C. Downstream changes in conductivity were recorded every 30 seconds until a steady-state plateau was reached (Stream Solute Workshop 1990). The LF340 and 340i WTW™ conductivity meters were attached to a Campbell Scientific CR510 data logger to record measurements. EC values measured using the WTW™ 350i meter were recorded to the meter’s internal memory. Once steady-state was achieved, as determined as a constant conductivity reading for more than 10 minutes at the farthest downstream location, the injection was stopped, and conductivity recordings continued until the stream water returned to background levels to characterize the “falling limb” of the BTC.

2.3.2. Discharge calculations

Discharge was calculated as:

\[ Q = \frac{q}{k \cdot (EC_{SS} - EC_{BG})} \]  \hspace{1cm} (2.1)

where \( Q \) is stream discharge (L/s), \( q \) is the injection rate of tracer solution (L/s), \( k \) is the calibration coefficient, \( EC_{SS} \) is steady state conductivity (\( \mu \)S/cm), and \( EC_{BG} \) is the background or pre-release conductivity (\( \mu \)S/cm). The calibration coefficient \( (k) \) represents the slope of the relation between relative concentration (L/L) and electrical conductivity (\( \mu \)S/cm). Electrical conductivity is linearly related to relative concentration (RC) for dilute solutions. As a result, the relative concentration at steady state \( (RC_{SS}) \) can be determined from EC measurements:

\[ RC_{SS} = (EC_{SS} - EC_{BG}) \cdot k \]  \hspace{1cm} (2.2)
Secondary and calibration solutions were created in the field for each tracer experiment in order to determine $k$. To perform the calibration, two 1-L volumes of stream water were first measured using a glass volumetric flask and poured into sample bottles. The secondary solution consisted of 10 mL of the primary injection solution, measured using a glass pipette, mixed with 1 L of stream water. The secondary solution was then added in 10 mL increments into a separate 1 L of stream water to create the calibration solution. Electrical conductivity was recorded with each addition of secondary solution into the calibration solution. Relative concentration was calculated as:

$$RC = S_A \frac{RC_{sec}}{V_T}$$

(2.3)

where $S_A$ is the volume of secondary solution added (mL), $RC_{sec}$ is the relative concentration of the secondary solution ($10 \text{ mL}/1010 \text{ mL}$), and $V_T$ is the total volume of the calibration solution.

If assumptions of the constant rate injection method (i.e. complete mixing, steady-state plateau) are met, the method can measure flows within ± 5% uncertainty (Johnstone 1988). However, the potential error associated with the discharge measurements was calculated using an error analysis (Story 2002):

$$\frac{\delta Q}{Q} = \sqrt{\left(\frac{\delta Q_i}{Q_i}\right)^2 + \left(\frac{\delta k}{k}\right)^2 + \left(\frac{\delta (EC_{ss} - EC_{BG})}{EC_{ss} - EC_{BG}}\right)^2}$$

(2.4)

where:

$$\delta (EC_{ss} - EC_{BG}) = \sqrt{(\delta EC_{ss})^2 + (\delta EC_{BG})^2}$$

$$\frac{\delta Q_i}{Q_i} = \sqrt{\left(\frac{\delta}{r}\right)^2 + \left(\frac{\delta t}{t}\right)^2}$$

where sources of error included injection rates ($Q_i$), electrical conductivity measurements and the slope of the regression ($k$) relating electrical conductivity to the relative concentration. The error associated with the injection rates ($\delta Q_i/Q_i$) was estimated based on the error associated with measuring the injection rate ($\delta r$) and time ($\delta t$). The uncertainty with reading the graduated cylinder ($\delta r$) was estimated as ± 0.5 mL; the uncertainty with measuring time ($\delta t$) was ± 2 s (1 s on either of the timing). The average injection rate
was 74 mL/min (n = 14). Error associated with the measurement of the electrical conductivity at plateau (ECSS) and background (ECBG) were 0.1 μS/cm for ECSS <200 μS/cm and 1 μS/cm for ECSS >200 μS/cm. Standard error of the slope (k) from the regression was used to calculate the error in k.

2.3.3. Characterizing lateral exchanges

Constant-rate injection tracer experiments also provided estimates of the rate of lateral inflow. Lateral inflow rates were estimated as the difference in discharge measurements (Equation 2.1) as measured at upstream and downstream EC sampling locations within the reach (Figure 2.3) divided by the reach length:

\[ Q_L = \frac{Q_{ds} - Q_u}{L} \]  

where \( Q_L \) is the net lateral inflow rate (Ls\(^{-1}\))m\(^{-1}\), \( Q_{ds} \) and \( Q_u \) are streamflow (L/s) measured at the downstream and upstream locations within the reach respectively, and \( L \) is the reach length (m).

2.3.4. Cross-section measurements

Cross sections were measured at 10 m intervals along the stream reach. At each location, the depth, measured at 20 cm intervals across the stream, and the wetted channel width were recorded. Measurements of depth and channel width were used to calculate the stream cross-sectional area. The reach averaged cross-sectional area was used as an input variable for hyporheic zone modelling at the reach scale.

2.4. Stream tracer experiments – Reach scale

2.4.1. Method of injection

Stream tracer tests were conducted using the constant-rate salt injection method in order to examine the extent of surface-subsurface interactions at the reach scale. Wagner
and Harvey (1997) determined that the constant injection method with sampling through the concentration rise, plateau and fall provides more reliable parameter estimates than other sampling designs (e.g. slug injections).

Tracer injection points and sample locations were the same for all flow conditions (Figure 2.3). Tracer injections were initially conducted by separating the study stream into two approximately 50 m sections (i.e. lower and upper reach) and injecting above the upper boundary for each reach. Tracer experiments were conducted on the same day or within 1-2 days depending on conditions. Full reach tracer experiments were also conducted for ease of field work. Stream tracer was injected above the upper boundary of the upper reach (approximately 10 m downstream from the culvert at M road) and the breakthrough curves were measured at the lower boundaries of the reaches. During base flow conditions ($Q < 2$ L/s), the reach was again separated into two sub-reaches to conduct longer duration tracer experiments.

2.4.2. Quantifying pool storage and residence time

During two reach-scale tracer injections, electrical conductivity meters were placed at the inflow and outflow location of three pool sub-units in order to quantify pool storage and residence time. Breakthrough curves from individual pools were simulated using OTIS-P. Simulations were conducted for two pools located in the upper reach (September 29) and one pool located in the lower reach (September 30). Each pool was simulated as a distinct reach.

2.4.3. Hydraulic parameters (OTIS-P)

Transport and transient storage mechanisms in the stream reach were analyzed by fitting a numerical model to breakthrough curves generated from stream tracer experiments. The processes of advection, dispersion, lateral inflow and transient storage within the hyporheic zone were modeled using the OTIS-P code, which was developed by the US Geological Survey (Runkel 1998). OTIS-P, a modified version of OTIS (One-dimensional Transport with Inflow and Storage), numerically solves finite difference
approximations to Equation 2.6 with the Crank-Nicolson Method, and uses a nonlinear least squares method to optimize parameters by minimizing the sum of squared errors. The model uses a modification of a one-dimensional advection-dispersion model with additional terms for groundwater inputs and transient storage (Bencala and Walters 1983, Stream Solute Workshop 1990). The following set of differential equations (Runkel and Broshears 1991) is used by OTIS-P to solve for solute mass balance between the main channel (Equation 2.6) and the transient storage zones (Equation 2.7):

\[
\frac{dC}{dt} = -\frac{Q}{A} \frac{\partial C}{\partial x} + \frac{1}{A} \frac{\partial}{\partial x} \left( AD \frac{\partial C}{\partial x} \right) + \frac{q_{\text{lin}}}{A} (C_L - C) + \alpha (C_s - C) \tag{2.6}
\]

\[
\frac{dC_s}{dt} = -\alpha \frac{A}{A_s} (C_s - C_s) \tag{2.7}
\]

where \( A \) is main channel cross-sectional area (\( L^2 \)); \( A_s \) is storage zone cross-sectional area (\( L^2 \)); \( C \) is main channel solute concentration (\( M/L^3 \)); \( C_L \) is lateral inflow solute concentration (\( M/L^3 \)); \( C_s \) is storage zone solute concentration (\( M/L^3 \)); \( D \) is dispersion coefficient (\( L^2/T \)); \( Q \) is volumetric flow rate (\( L^3/T \)); \( q_{\text{lin}} \) is lateral inflow rate (\( L^3/T-L \)); \( t \) is time (\( T \)) and \( x \) is distance (\( L \)); \( \alpha \) is storage zone exchange coefficient (1/T). Cross-sectional area (\( A, m^2 \)) as entered into the model is calculated as mean of the summed products of mean wetted depth and channel width for each cross-section. The model estimates a reach-averaged \( A \) to match observed solute concentration data. Average stream velocity (\( u, m/s \)) is calculated as \( Q/A \), using the model-estimated \( A \) value. Dispersion (\( D, m^2/s \)), reach-averaged transient storage zone cross-sectional area (\( A_s, m^2 \)), and transient storage exchange coefficient (\( \alpha, s^{-1} \)) are estimated using the model. Transient storage refers to any zone within the channel, such as an eddy, pool, or hyporheic zone, where some water is temporarily detained relative to the faster moving water near the center of the channel.

Model parameters solved using OTIS-P were used to derive the following quantities: (1) hydraulic uptake length (Mulholland et al. 1994), (2) hydraulic residence time (or contact time) in the storage zone and stream (Mulholland et al. 1994), (3) the hydraulic retention factor (Morrice et al. 1997), and (4) the standardized storage zone area (Stream Solute Workshop 1990, D’Angelo et al. 1993). Derived quantities are often
used to compare solute transport processes between stream reaches (Stream Solute Workshop 1990).

The hydraulic residence time, $T_{stor}$ (s), in the storage zone can be calculated as:

$$T_{stor} = \frac{Az}{A \alpha} \quad (2.8)$$

whereas residence time in the stream ($T_{str}$) is calculated as:

$$T_{str} = \frac{1}{\alpha} \quad (2.9)$$

The hydraulic uptake length ($S_{hyd}$) is the average distance a water molecule travels downstream before entering the storage zone:

$$S_{hyd} = \frac{Q}{A \alpha} \quad (2.10)$$

The hydraulic retention factor ($R_h$) is a measure of the storage zone residence time per unit of stream reach traveled:

$$R_h = \frac{T_{stor}}{S_{hyd}} \quad (2.11)$$

The standardized storage zone area ($A_s/A$) is the ratio of storage cross-sectional area to stream cross-sectional area and is the mathematical equivalent of storage zone residence time to stream residence time.

### 2.4.4. Evaluation of parameter uncertainty

Uncertainty surrounding the parameter estimates was examined using two indices: the experimental Damköhler number ($Dal$) and the uncertainty ratio. The Dal number was calculated as:

$$Dal = \frac{\alpha(1 + A/A_z)L}{u} \quad (2.12)$$
where L is length of the stream reach over (m). Wagner and Harvey (1997) showed that when the relationship between the exchange rate and advection deviates from 1.0, the uncertainty in the modelled parameters increases. Parameter uncertainty is minimized when Dal is close to 1.0. High values may occur because exchange with the streambed is relatively fast compared to the water velocity or the reach length may be too long. A Dal less than 1.0 indicates that only a small amount of the stream tracer is exchanging with the storage zone. Small Dal numbers (<0.1) could result from: (1) high stream velocity, (2) long exchange time scale as indicated by a low α and A_s/A ratio, and/or (3) short reach length.

The uncertainty ratio for each estimated parameter (D, A, A_s, α) is equal to the parameter estimate divided by its standard deviation. A low ratio indicates that the parameter estimate is highly uncertain.

2.5. Stream tracer experiments – Channel-unit scale

2.5.1. Qualitative observations of hyporheic discharge

Tracer injections of the dye tracers Brilliant Blue FCF (C.I. 42090) and Rhodamine WT (RWT) were conducted at three locations (II, 14 and 15; Figure 2.8) using infiltrometers installed within the stream bed as described in Section 2.5.6. Injections were used to identify qualitatively the location of hyporheic discharge within individual channel units. Approximately 1-5 g of RWT was injected, and less than 50 mL of brilliant blue was injected at one time. Experiments using RWT were conducted over six dates (July 11, August 14, August 21, September 27/28, October 6) in the lower reach at two locations (I-4, I-5). Tracer injection experiments were also conducted in the upper reach at one location (I-1) on four dates (July 12, August 23, and September 28, October 6).
2.5.2. Quantifying residence times

Sodium chloride was also used as a tracer to quantify solute residence time within one boulder step-pool unit at the lower reach. A solution of 1 L deionized water and 20 g of salt was added directly to the infiltrometer 15 (Figure 2.8) installed within the streambed. Electrical conductivity was measured at the base of the step (or pool inflow) and at the pool outflow location to quantify residence time within the hyporheic zone and pools respectively. Calibrations were conducted using the same procedure as described in section 2.3.2. Two channel-unit experiments were conducted over the study period (September 25 and October 5, 2006).

2.5.3. Modelling mean residence time

A simple modeling approach was used to model the mean residence time of water within the transient storage zones. The system was modeled using linear reservoir theory assuming that the pool unit behaves as a continuously stirred tank reactor (CSTR) (Chapra 1997). Complete mixing of the solute and steady-state water flow for the duration of the tracer injection experiment was assumed. A mass balance for the pool can be expressed as:

$$\frac{dM(t)}{dt} = q_{in}(t) - q_{out}(t) \quad (2.13)$$

where $M(t)$ is the mass of tracer in pool (kg), $q_{in}(t)$ is the input of tracer (kg/s) and $q_{out}(t)$ is the output (kg/s). Mass can be related to tracer concentration using the equations:

$$M(t) = V \cdot c(t) \quad (2.14)$$

$$q_{out}(t) = Q \cdot c_{out}(t) \quad (2.15)$$

where $V$ is the volume of water in pool (m$^3$), $c(t)$ is the concentration of tracer in the pool at time $t$ (kg/m$^3$), $Q$ is the water discharge at the outlet of the pool (m$^3$/s), and $c_{out}(t)$ is the tracer concentration at the outlet. Using CSTR theory, the mean residence time (MRT) of solutes within the system can be expressed as:
where $k$ is a first order exchange coefficient ($s^{-1}$). Toward the end of the experiment, tracer discharge to the pool will become small, and the relationship between concentration and time can be modelled as a first-order reaction:

$$\frac{dc(t)}{dt} = -k \cdot c(t)$$  \hspace{1cm} (2.17)

Integrating Equation 2.17, subject to the initial condition $c(t_0) = c_0$, yields:

$$c(t) = c_0 e^{-kt}$$  \hspace{1cm} (2.18)

where $c_0$ is the concentration at time $= t_0$, which is an arbitrarily selected time. This equation specifies an exponential depletion of the tracer concentration over time. Equation 2.18 can be transformed by taking logarithms of both sides to yield:

$$\ln[c(t)] = \ln[c_0] - k(t - t_0)$$  \hspace{1cm} (2.19)

If a plot of the logarithm of concentration against time yields a straight line, Equation 2.15 holds true, and $k$ can be calculated as the slope of a straight line fitted to the linear portion of the log-transformed breakthrough curve.

### 2.6. Subsurface flow measurements – Point scale

#### 2.6.1. Piezometer design and installation

Piezometers were installed in a dense network near the center of the channel in six step-pools to provide a high spatial resolution of subsurface flow pathways (Figure 2.7). A total of 68 piezometers were installed in the stream during early summer (May to June, 2006). Two types of piezometer designs were used: (1) aluminum piezometers for taking water quality samples ($n = 28$), and (2) plastic piezometers to measure hydraulic gradients and measure hydraulic conductivity ($n = 40$).
Aluminum piezometers were constructed using an approximately 60 cm length of 1 cm internal diameter aluminum tube with a slot zone of 5 cm. Aluminum piezometers were driven into the streambed using a sledgehammer. Hydraulic gradients were also measured for the aluminum piezometers.

Plastic piezometers were approximately 60 cm in length and constructed using 0.7 cm internal diameter polyvinyl chloride (PVC) pipe with a 5 cm slot zone. The slot zones were screened with nylon mesh to prevent clogging with fine sediments. Plastic piezometers were installed into the streambed by first driving in a steel rod, contained within a metal sleeve. The metal rod was then removed from the sleeve and the plastic piezometer was placed inside. Once the metal sleeve and rod were removed, the piezometer was left imbedded in the subsurface. The piezometers were installed at multiple depths (0-15, 15-30, 30-50 cm) both longitudinal and perpendicular to the wetted stream channel. The location and elevation of all piezometers were surveyed using a Leica Geosystems© total station.

2.6.2. Hydraulic head

Vertical hydraulic gradients (VHG) in the streambed were used to locate areas of upwelling and downwelling within the hyporheic zone. VHG was calculated as:

$$ VHG = \frac{\Delta h}{\Delta l} $$

where \( \Delta h \) is the elevation of the water in the piezometer minus the elevation of the stream water surface (cm), and \( \Delta l \) is the distance between the surface of the stream bed and the middle of the slot zone (cm). Positive VHG indicates upwelling hyporheic or groundwater flow (flow potential from the bed towards the channel); negative VHG indicates downwelling flow (flow potential is directed from the channel into the bed). Reversal of exchange potential was defined as a shift between positive and negative VHG. Neutral piezometers were defined as having hydraulic gradients between -0.05 to 0.05 cm/cm, which lie within the uncertainty of VHG measurements (Guenther 2007). Hydraulic heads were measured using a water level sensor. The sensor consisted of two electrical wires attached to a rod and connected to a battery and buzzer. The rod was
lowered into the piezometer. When the wires contact water, the electrical circuit is closed and a buzzer sounds. The water depth was measured using a measuring tape, resulting in an accuracy of ± 0.05 cm.

2.6.3. Hydraulic conductivity

Piezometers were also used to measure saturated hydraulic conductivity of the bed sediments using a Hvorslev test (also known as falling head slug test; Freeze and Cherry 1979). Water was added to the piezometer using a syringe connected to a tube and inserted to the top of the piezometer. The tube was disconnected, and the time required for the water to return to a specified level on the piezometer was measured using a stopwatch. Hydraulic conductivity \( K \) was computed based on the empirical equation of Hvorslev (1951) as modified by Baxter et al. (2003) for closed-bottom perforated piezometers:

\[
K = SF \cdot \frac{\ln(H_1 / H_2)}{(t_2 - t_1)}
\]

(2.21)

where:

\[
SF = \frac{d^2 \cdot \ln \left( \frac{L}{D} \right) + \sqrt{1 + \left( \frac{L}{D} \right)^2}}{8 \cdot L}
\]

where \( H_1 \) and \( H_2 \) are the head in the piezometer (cm) at time \( t_1 \) and \( t_2 \) (s), \( SF \) is the shape factor for the piezometer intake (m), \( L \) is the length of the perforations (m), \( D \) is the diameter of the perforated intake (m) and \( d \) is the inside diameter of the piezometer (m).

Hydraulic conductivity measurements of stream sediments are typically reported to be positively skewed or not normally distributed (Ryan and Boufadel 2007). As a result, the geometric mean (rather than arithmetic mean) was calculated. The mean was computed from of all falling head tests taken at each site (\( n = 4 \)).
Figure 2.7. Map of sampling locations, morphology and thalweg profile for both reaches.
2.6.4. Relating discharge and recharge zones to stream geometry

Calculated vertical hydraulic gradients (Equation 2.20) were used to map zones of hyporheic discharge (upwelling) and recharge (downwelling) along the stream profile in ArcGIS version 9.1. A geometric scaling relationship was used to describe vertical hydraulic gradients as a function of location in the stream channel based on mapping results. The general form of the relation is:

$$ \frac{dh}{dl} = f\left(\frac{X}{L}, SH\right) $$

(2.22)

where $X$ is the distance from the upstream end of the pool to the piezometer (m), $L$ is the distance from the upstream end of the pool to the edge of the step downstream of the pool (m), and $SH$ is the height of the downstream step (m), as defined by Zimmermann and Church (2001). It is hypothesized that $dh/dl$ should be negatively related to $X/L$, with positive values for values near 0, and increasingly negative values as $X/L$ approaches 1. Furthermore, hydraulic gradients in the downwelling zone should be negatively associated with $SH$. That is, higher steps should exhibit stronger downwelling gradients.

A total of 7 channel-units were used for the analysis.

The average water flux for one step-pool unit (Pool 1) was calculated under baseflow (September 29) and high flow conditions (June 19) using Darcy’s law:

$$ q = -K \frac{dh}{dl} A $$

(2.23)

where $q$ is the volume of water (L/s), $K$ is the average hydraulic conductivity (m/s), $\frac{dh}{dl}$ is the average hydraulic gradient (cm/cm), and $A$ is the area of stream channel (m$^2$) as determined by the $X/L$ category. The area above and below the step was segmented into spatial units based on the $X/L$ category. The average flux for each segment was computed and totaled for each channel sub-unit. The total channel-unit flux was divided by the wetted channel volume (L) of the reach in order to obtain a scaled-up estimate of hyporheic exchange (s$^{-1}$). This estimate was compared to the reach-scale estimates of
transient exchange obtained by fitting the OTIS-P model to tracer injection breakthrough curves.

2.6.5. **Stream bed infiltrometers**

Direct measurements of stream bed infiltration rates were conducted using a constant-head stream bed infiltrometer (Figure 2.8). Infiltrometers were constructed using an approximately 20-30 cm long cylindrical, open-ended, PVC pipe with a 7.5 cm internal diameter. A hole drilled into the PVC pipe at mid length was used to connect the pipe to a Mariotte cylinder using a piece of tygon tubing. The PVC pipe was installed into the stream bed at a depth of 5-10 cm such that the mid-length opening to the PVC pipe was level with the stream bed. The Mariotte tube was installed vertically in the stream using a piece of rebar hammered into the stream bed. The Mariotte tube was filled with water prior to conducting measurements. The basic premise for the Mariotte tube is that when water within the PVC pipe “infiltrates” into the bed, the Mariotte tube supplies additional water to maintain a constant water level. The Mariotte tube consisted of a 30 cm length of plastic tube with both a plastic stopper at both ends. The bottom end of the tube attached to the PVC pipe installed in the stream, while the stopper at the top was used to hold a smaller diameter tube in place.

![Figure 2.8: Streambed infiltrometer. Adapted from Martin (1996).](image-url)
Infiltrometers were installed into the stream bed in five downwelling locations within the reach, specifically above the step within a step-pool unit (Figure 2.7) to measure the amount of water directly infiltrating into the hyporheic zone. Infiltration measurements were taken periodically during the study period at a range of flows. The infiltration rate (IR) was calculated as:

\[
IR = \frac{\Delta h}{\Delta t} \frac{\pi (r_1^2 - r_2^2)}{\pi (r_p^2)}
\]  

(2.24)

where \(\Delta h\) is the change in water level in the Mariotte tube (cm) over an interval, \(\Delta t\) (s), \(r_1\), \(r_2\) and \(r_p\) are the inside radius of the Mariotte reservoir (cm), the outside radius of the bubbler tube in the Mariotte reservoir (cm), and the inside radius of the evaporation pan (cm), respectively.

The uncertainty associated with the infiltration measurements was calculated using an error analysis:

\[
\frac{\delta I}{I} = \left( \frac{\delta (\Delta h)}{\Delta h} \right)^2 + \left( \frac{\delta (\Delta t)}{\Delta t} \right)^2 + \left( \frac{2 (r_1 \cdot \delta r_1 + r_2 \cdot \delta r_2)}{r_1^2 - r_2^2} \right)^2 + \left( \frac{2 r_p \cdot \delta r_p}{r_p^2} \right)^2 \right)^{1/2}
\]  

(2.25)

where \(\delta (\Delta h)\) is the error associated with measuring the change in water level in the Mariotte tube, estimated at \(\pm 0.2\) cm; \(\delta (\Delta t)\) is the uncertainty with measuring time, estimated at \(\pm 2\) s (1 s on either of the timing); \(\delta r_1\), \(\delta r_2\) and \(\delta r_p\) are the uncertainty in measuring the inside radius of the Mariotte reservoir, the outside radius of the bubbler tube in the Mariotte reservoir, and the inside radius of the evaporation pan, respectively.

Infiltrometers were installed adjacent to piezometers (approximately 10 cm) in order to back-calculate hydraulic conductivity \((K_i)\):

\[
K_i = \frac{IR}{VGH}
\]  

(2.26)

where IR is the measured infiltration rate (cm/s) and VGH is the vertical hydraulic gradient (cm/cm) measured using Equation 2.20. Infiltrometer measurements of hydraulic conductivity were then compared to measurements of conductivity as measured in the piezometers using the falling-head test.
2.6.6. Subsurface relative connectivity

Subsurface water samples were collected from the aluminum piezometers to measure the relative connectivity of the subsurface to the stream channel during reach scale tracer injections. Piezometers were initially purged and allowed to refill prior to sampling. A polypropylene syringe (60 mL) attached to tygon tubing (3-mm inside diameter) was used to withdraw water samples from the piezometers. Small volumes of water (approximately 15-30 mL) were removed for purging. A volume of approximately 25-30 mL was removed to analyze for electrical conductivity. Samples were removed in small quantities to minimize the influence on the subsurface flow system. The syringe, tubing and sample bottles were deionized. Water samples were taken prior to tracer release and again once the stream concentration reached plateau. The electrical conductivity of the water samples was measured to calculate relative connectivity. A two-component mixing equation using the electrical conductivity of stream water and piezometer porewater at stream tracer initiation (t = 0) and steady state were used to calculate the fraction of stream water present in the subsurface (Gooseff and McGlynn 2005). The percent stream water in the hyporheic zone was calculated as:

\[ \chi = \frac{EC_{p(t)} - EC_{p(0)}}{EC_{s(t)} - EC_{s(0)}} \times 100\% \]  

(2.27)

where \( EC_{p(t)} \), \( EC_{p(0)} \), \( EC_{s(t)} \) and \( EC_{s(0)} \) refer to the conductivity in the piezometers and stream at steady state (t) and background (time 0) respectively. The mixing ratio within the hyporheic zone was calculated for all piezometers. It is assumed that no stream water has exchanged with the subsurface when \( \chi \) is equal to 0; when \( \chi \) is equal to 1, complete replacement of the hyporheic zone water with tracer labeled stream water has occurred.
2.7. Statistical Analysis

Variability of exchange flow was assessed by comparing changes in VHG and infiltration rates over the range of flow conditions observed. Spearman’s rank correlation analysis was used to explore the relationships between discharge, vertical hydraulic gradient and infiltration at each point location. In addition, VHG as a function of the downstream step height was also tested using this approach. A value of 1 or -1 indicates a strong positive (or strong negative) correlation between two nonparametric variables (Kutner et al. 2004).

A linear mixed-effects model (Maindonald and Braun 2007) was used to examine the in-site variability in hydraulic gradients and hydraulic conductivity using the LMER function in R 2.5.1 for unbalanced experimental designs (R Development Core Team 2007). To model the variability in hydraulic gradients, channel-unit (n = 7) was considered a random effect while downstream step height (7 levels) and position within the channel (as defined by X/L; 5 levels) were fixed effects. Three separate linear models were created including (1) a base model with only channel-unit, (2) channel-unit with step height and (3) channel-unit, step height and XL factors. A sequential analysis of variance was used to determine the significance of the fixed effects on hydraulic gradients by comparing all three models using the ANOVA function in R 2.5.1. Data were weighted by the inverse of X/L to account for the heteroscedasticity of the residuals (Kutner et al. 1994).

This statistical method was also used to examine the variability in hydraulic conductivity due to site condition (i.e. upwelling, downwelling or neutral) and reach location (i.e upper or lower reach). Reach (n = 2) was considered a random effect while site condition was a fixed effect. Data were log transformed to account for the heteroscedasticity of the residuals (Kutner et al. 1994). A significance level of 0.05 was used for all analyses.
CHAPTER THREE: RESULTS

This chapter presents the results of field observations in East Creek between May and October, 2006. The chapter starts with an overview of the study period conditions (section 3.1) and discussion of the data quality (section 3.2). Results are then presented for each scale of investigation beginning with the reach scale (Section 3.3), followed by the channel-unit scale (section 3.4), and the individual point scale (section 3.5). The chapter concludes with a summary of the key findings (section 3.6).

3.1. Study period conditions

Based on data from the Haney-UBC Research Forest Admin climate station (Environment Canada), located within the UBC Research Forest (49° 16.2'N, 122° 34.2'W), summer temperature and precipitation conditions were warmer and drier than the 30 year norm (Table 3.1). Figure 3.1 provides an overview of flow conditions in East Creek along with daily precipitation and temperature in the Research Forest.

Stream tracer experiments were conducted from May 31 to October 20, 2006, over a range of flow conditions to examine hyporheic exchange processes over time. A majority of the tracer injections were conducted during baseflow conditions (Q < 5 Ls⁻¹) when the contribution of the hyporheic zone is hypothetically maximized. Streamflows ranged from 0.21 L/s on September 13 to 30.6 L/s on May 31 over the study period. Discharge levels were too high by early November (Q > 200 L/s) to continue the study.

Table 3.1: Comparison of 2006 mean daily temperature and monthly precipitation to 30 year climate normal (1961-1990) as measured at the Haney-UBC Research Forest Admin climate station (Environment Canada).

<table>
<thead>
<tr>
<th>Month</th>
<th>Normals</th>
<th>2006</th>
<th>Difference</th>
<th>Normals</th>
<th>2006</th>
<th>% of Normal</th>
</tr>
</thead>
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<tr>
<td>May</td>
<td>11.8</td>
<td>12.7</td>
<td>+0.9</td>
<td>114</td>
<td>121.8</td>
<td>106.8</td>
</tr>
<tr>
<td>Jun</td>
<td>14.6</td>
<td>16.1</td>
<td>+1.5</td>
<td>93.1</td>
<td>55.2</td>
<td>59.3</td>
</tr>
<tr>
<td>Jul</td>
<td>16.8</td>
<td>18.7</td>
<td>+1.9</td>
<td>80.9</td>
<td>27.2</td>
<td>33.6</td>
</tr>
<tr>
<td>Aug</td>
<td>17.0</td>
<td>17.8</td>
<td>+0.8</td>
<td>74.3</td>
<td>17.0</td>
<td>22.9</td>
</tr>
<tr>
<td>Sep</td>
<td>14.5</td>
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<td>+1.2</td>
<td>119.7</td>
<td>90.8</td>
<td>75.9</td>
</tr>
<tr>
<td>Oct</td>
<td>9.9</td>
<td>11.6</td>
<td>+1.7</td>
<td>223.8</td>
<td>32.6</td>
<td>14.6</td>
</tr>
</tbody>
</table>
Figure 3.1: Daily precipitation, maximum and minimum daily temperatures, measured discharge and net lateral inflow from tracer injections conducted during the study period of May to October 2006. Discharge values represent streamflow measured at the lower reach boundary. Lateral inflow measured as the difference between upstream and downstream streamflow measurements. Note log scale for Q and Q_L. Climate data recorded at the Haney-UBC Research Forest Admin climate station (Environment Canada).
3.2. Data quality

Error rates varied between 3-7% for all flow conditions (Table 3.2). These error rates are roughly consistent with the cited uncertainty of ±5% (Johnstone 1988).

During three tracer injections conducted during low flow conditions (July 20, August 13 and 31), the lower reach boundary did not achieve plateau. Streamflow at the lower reach boundary was then estimated as a percentage of streamflow as measured at the upper reach boundary. This was based on the relationship between the measured discharge at the upper and lower reach boundaries for full reach tracer injections that reached plateau (n = 4). The increase in discharge was estimated at 20.5% of the upstream discharge. Model simulations were only conducted for tracer injections where the upper and lower reach boundary concentrations reached plateau (n = 10).

<table>
<thead>
<tr>
<th>Date</th>
<th>Q (L/s)</th>
<th>R (mm/mi)</th>
<th>k (10^-6)</th>
<th>SE (k x 10^-9)</th>
<th>EC_bkg (μS/cm)</th>
<th>EC_plat (μS/cm)</th>
<th>Probable Error (%)</th>
<th>Reach</th>
</tr>
</thead>
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<tr>
<td>May 31*</td>
<td>30.6</td>
<td>71</td>
<td>2.8</td>
<td>3.0</td>
<td>20.0</td>
<td>34.0</td>
<td>3.6</td>
<td>LR</td>
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<tr>
<td>June 7</td>
<td>24.9</td>
<td>34</td>
<td>5.4</td>
<td>6.8</td>
<td>20.4</td>
<td>24.6</td>
<td>5.0</td>
<td>LR</td>
</tr>
<tr>
<td>June 19*</td>
<td>15.4</td>
<td>33</td>
<td>1.7</td>
<td>14</td>
<td>22.6</td>
<td>43.2</td>
<td>3.8</td>
<td>UR</td>
</tr>
<tr>
<td>June 19*</td>
<td>17.2</td>
<td>23</td>
<td>1.7</td>
<td>14</td>
<td>22.3</td>
<td>27.8</td>
<td>4.8</td>
<td>LR</td>
</tr>
<tr>
<td>June 27*</td>
<td>5.5</td>
<td>59</td>
<td>3.3</td>
<td>13</td>
<td>23.8</td>
<td>79.0</td>
<td>3.5</td>
<td>UR/LR</td>
</tr>
<tr>
<td>July 20</td>
<td>2.1</td>
<td>87</td>
<td>3.3</td>
<td>13</td>
<td>29.0</td>
<td>270.0</td>
<td>3.4</td>
<td>UR/LR</td>
</tr>
<tr>
<td>Aug 13</td>
<td>0.53</td>
<td>40</td>
<td>9.2</td>
<td>12</td>
<td>29.6</td>
<td>174.0</td>
<td>3.6</td>
<td>UR</td>
</tr>
<tr>
<td>Aug 31</td>
<td>0.33</td>
<td>116</td>
<td>130</td>
<td>3.0</td>
<td>38.8</td>
<td>84.0</td>
<td>4.0</td>
<td>LR</td>
</tr>
<tr>
<td>Sept 13</td>
<td>0.21</td>
<td>70</td>
<td>52</td>
<td>3300</td>
<td>35.4</td>
<td>143.0</td>
<td>7.1</td>
<td>LR</td>
</tr>
<tr>
<td>Sept 21*</td>
<td>9.9</td>
<td>109</td>
<td>6.6</td>
<td>8.3</td>
<td>31.1</td>
<td>58.9</td>
<td>3.4</td>
<td>UR/LR</td>
</tr>
<tr>
<td>Sept 29*</td>
<td>1.1</td>
<td>92</td>
<td>21</td>
<td>60</td>
<td>33.0</td>
<td>102.0</td>
<td>3.7</td>
<td>UR</td>
</tr>
<tr>
<td>Sept 30*</td>
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<td>92</td>
<td>21</td>
<td>60</td>
<td>32.4</td>
<td>71.9</td>
<td>3.4</td>
<td>LR</td>
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<tr>
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<td>104</td>
<td>4.5</td>
<td>140</td>
<td>30.2</td>
<td>74.6</td>
<td>4.5</td>
<td>UR/LR</td>
</tr>
</tbody>
</table>

* Tracer injections simulated by OTIS-P
3.3. Solute transport model analysis – Reach scale

3.3.1. Summary of OTIS-P simulations

Best fit model parameters for six tracer experiments conducted in the lower reach and four tracer experiments conducted in the upper reach are summarized in Table 3.3. All tracer experiments doubled or quadrupled the background electrical conductivity of stream water to a plateau concentration. Figure 3.2 shows an example of simulated and observed tracer concentrations for the June 19 tracer injection conducted in the upper and lower reach. Additional breakthrough curves and model simulations are provided in Appendix A.

Table 3.3: Summary of best fit model parameters for solute releases including stream discharge (Q), dispersion coefficient (D), stream cross-sectional area (A), cross-sectional area of storage zone (As), storage zone exchange coefficient (a), net lateral inflow (QL), the Damkohler number (Dal).

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Reach</th>
<th>Q (10^-3 m^3 s^-1)</th>
<th>D (10^-2 m^2 s^-1)</th>
<th>A (10^-2 m^2)</th>
<th>As (10^-2 m^2)</th>
<th>a (10^4 s^-1)</th>
<th>QL (10^-6 m^3 s^-1 m^-1)</th>
<th>Dal</th>
</tr>
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<tr>
<td>Upper</td>
<td>June 19</td>
<td>15.9</td>
<td>31.6</td>
<td>11.6</td>
<td>5.83</td>
<td>2.1</td>
<td>26.8</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>Sept 21</td>
<td>9.8</td>
<td>18.3</td>
<td>6.73</td>
<td>1.45</td>
<td>2.7</td>
<td>8.5</td>
<td>0.22</td>
</tr>
<tr>
<td></td>
<td>Sept 29</td>
<td>1.0</td>
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<td>2.82</td>
<td>0.6</td>
<td>3.0</td>
<td>3.3</td>
<td>0.50</td>
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<td>Oct 20</td>
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<td>4.0</td>
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<td>1.46</td>
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<td>10.5</td>
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<td>1.4</td>
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<td>17.2</td>
<td>29.5</td>
<td>19.2</td>
<td>5.9</td>
<td>1.9</td>
<td>20.4</td>
<td>0.14</td>
</tr>
<tr>
<td></td>
<td>June 27</td>
<td>5.5</td>
<td>1.9</td>
<td>13.7</td>
<td>4.1</td>
<td>2.7</td>
<td>5.6</td>
<td>0.44</td>
</tr>
<tr>
<td></td>
<td>Sept 21</td>
<td>9.9</td>
<td>8.4</td>
<td>14.3</td>
<td>2.8</td>
<td>2.1</td>
<td>6.9</td>
<td>0.36</td>
</tr>
<tr>
<td></td>
<td>Sept 30</td>
<td>1.9</td>
<td>4.6</td>
<td>6.5</td>
<td>1.3</td>
<td>0.8</td>
<td>6.9</td>
<td>0.17</td>
</tr>
<tr>
<td></td>
<td>Oct 20</td>
<td>11.9</td>
<td>5.0</td>
<td>19.2</td>
<td>2.0</td>
<td>1.6</td>
<td>9.4</td>
<td>0.28</td>
</tr>
</tbody>
</table>
Figure 3.2. Model simulations using OTIS-P for June 19 for the upper reach (a) and the lower reach (b)
3.3.2. Variability of fitted parameters

Solute transport processes varied both temporally (with variations in discharge) and to a lesser extent, spatially (i.e. between reaches). Dispersion rates, transient storage area and stream cross-sectional area increased with discharge (Figure 3.3). This general trend is observed for both the upper and lower reaches; however, for all tracer simulations the stream cross-sectional area of the lower reach is greater than the upper reach. Increased channel complexity and pool storage in the lower reach may explain this difference. Transient exchange rates between the main channel and the storage zone did not vary systematically with discharge or between reaches (Figure 3.3).

Figure 3.3. Simulated model parameters for solute releases in the upper and lower stream reach. Dispersion coefficient (D), stream cross-sectional area (A), cross-sectional area of storage zone (A_s), storage zone exchange coefficient (a) versus stream discharge (Q). Error bars represent ± 1 standard deviation.
3.3.3. Parameter uncertainty

A general decreasing trend in Dal numbers was observed as discharge increased (Figure 3.4). Damkohler numbers ranged from 0.08 on May 31 (30.6 L/s) to 0.5 on September 29 (1.04 L/s). The tracer injection with the best parameter estimates, as indicated by a Dal number closest to one, was on September 29.

![Figure 3.4. Experimental Damkohler number (Dal) versus stream discharge (Q)](image)

The uncertainty ratio for each estimated parameter (D, A, A_s, α), calculated as the parameter estimate divided by its standard deviation, varied over the range of flow conditions (Table 3.4). No trend between discharge and the uncertainty ratio was observed (Figure 3.5).
Table 3.4. Summary of uncertainty ratios for the parameter estimates of dispersion coefficient (D), stream cross-sectional area (A), cross-sectional area of storage zone ($A_s$), and the storage zone exchange coefficient ($\alpha$).

<table>
<thead>
<tr>
<th>Reach</th>
<th>D</th>
<th>A</th>
<th>$A_s$</th>
<th>A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>June 19</td>
<td>1.9</td>
<td>211.1</td>
<td>8.1</td>
<td>20.7</td>
</tr>
<tr>
<td>Sept 21</td>
<td>5.1</td>
<td>69.5</td>
<td>2.7</td>
<td>8.2</td>
</tr>
<tr>
<td>Sept 29</td>
<td>5.5</td>
<td>56.5</td>
<td>14.2</td>
<td>5.2</td>
</tr>
<tr>
<td>Oct 20</td>
<td>1.2</td>
<td>35.2</td>
<td>6.2</td>
<td>2.1</td>
</tr>
<tr>
<td>Lower</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>May 31</td>
<td>9.8</td>
<td>.76</td>
<td>3.1</td>
<td>4.8</td>
</tr>
<tr>
<td>June 19</td>
<td>7.6</td>
<td>60.2</td>
<td>0.9</td>
<td>1.4</td>
</tr>
<tr>
<td>June 27</td>
<td>4.2</td>
<td>78.7</td>
<td>25.8</td>
<td>8.6</td>
</tr>
<tr>
<td>Sept 21</td>
<td>7.9</td>
<td>87.1</td>
<td>11.9</td>
<td>5.2</td>
</tr>
<tr>
<td>Sept 30</td>
<td>13.7</td>
<td>99.1</td>
<td>8.7</td>
<td>9.3</td>
</tr>
<tr>
<td>Oct 20</td>
<td>5.6</td>
<td>51.4</td>
<td>6.0</td>
<td>2.5</td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper</td>
<td>3.4</td>
<td>93.1</td>
<td>7.8</td>
<td>9.1</td>
</tr>
<tr>
<td>Lower</td>
<td>8.1</td>
<td>75.5</td>
<td>9.4</td>
<td>5.3</td>
</tr>
</tbody>
</table>

Figure 3.5. Uncertainty ratio (UR) for the simulated model parameters of dispersion (D), stream cross-sectional area (A), cross-sectional area of storage zone ($A_s$), storage zone exchange coefficient ($\alpha$) versus stream discharge ($Q$)
3.3.4. Derived quantities

Derived quantities including hydraulic residence time and retention and uptake lengths are summarized in Table 3.5. Hydraulic residence time in the stream was higher than in the storage zone for all tracer injections and did not appear to vary with discharge for either reach (Figure 3.6). The average stream residence time was higher in the lower reach (109 min) compared to the upper reach (49.8 min, Figure 3.6A). Storage zone residence time was also higher in the lower reach (31.0 min) compared to the upper reach (10.2 min, Figure 3.6B).

Table 3.5. Summary of derived quantities including stream velocity (u), hydraulic residence time for the stream (T_{s}^{r}) and storage zone (T_{s}^{r}), hydraulic uptake length (S_{hyd}), hydraulic retention factor (R_{h}) and the standardized storage zone coefficient (A_{s}/A).

<table>
<thead>
<tr>
<th>Reach</th>
<th>Q (10^{-3} m^{3} s^{-1})</th>
<th>u (m s^{-1})</th>
<th>T_{s}^{r} (min)</th>
<th>T_{s}^{r} (min)</th>
<th>S_{hyd} (m)</th>
<th>R_{h} (sm^{-1})</th>
<th>A_{s}/A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>June 19</td>
<td>15.9</td>
<td>0.13</td>
<td>80.7</td>
<td>40.4</td>
<td>648</td>
<td>3.7</td>
<td>0.50</td>
</tr>
<tr>
<td>Sept 21</td>
<td>9.8</td>
<td>0.15</td>
<td>61.8</td>
<td>13.3</td>
<td>542</td>
<td>1.5</td>
<td>0.22</td>
</tr>
<tr>
<td>Sept 29</td>
<td>1.0</td>
<td>0.04</td>
<td>55.2</td>
<td>11.6</td>
<td>122</td>
<td>5.7</td>
<td>0.21</td>
</tr>
<tr>
<td>Oct 20</td>
<td>9.7</td>
<td>0.10</td>
<td>48.3</td>
<td>7.6</td>
<td>302</td>
<td>1.5</td>
<td>0.16</td>
</tr>
<tr>
<td>Lower</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>May 31</td>
<td>30.6</td>
<td>0.14</td>
<td>121.5</td>
<td>75.7</td>
<td>1029</td>
<td>4.4</td>
<td>0.62</td>
</tr>
<tr>
<td>June 19</td>
<td>1.7</td>
<td>0.09</td>
<td>86.2</td>
<td>26.7</td>
<td>465</td>
<td>3.4</td>
<td>0.31</td>
</tr>
<tr>
<td>June 27</td>
<td>5.5</td>
<td>0.04</td>
<td>61.0</td>
<td>18.1</td>
<td>146</td>
<td>7.4</td>
<td>0.30</td>
</tr>
<tr>
<td>Sept 21</td>
<td>9.9</td>
<td>0.07</td>
<td>80.2</td>
<td>16.0</td>
<td>334</td>
<td>2.9</td>
<td>0.20</td>
</tr>
<tr>
<td>Sept 30</td>
<td>1.9</td>
<td>0.03</td>
<td>196.5</td>
<td>38.9</td>
<td>343</td>
<td>6.8</td>
<td>0.20</td>
</tr>
<tr>
<td>Oct 20</td>
<td>11.9</td>
<td>0.06</td>
<td>106.3</td>
<td>11.0</td>
<td>397</td>
<td>1.7</td>
<td>0.10</td>
</tr>
</tbody>
</table>

Stream velocity increased with discharge (Figure 3.7A). Generally, velocity was higher in the upper reach than the lower reach. The hydraulic uptake length increased as discharge increased for both the upper and lower reach (Figure 3.7B). Values ranged from 122 m (1.04 L/s) in the upper reach to 1028 m (30.6 L/s) in the lower reach. On average the uptake length was higher in the upper reach (318 m) than the lower reach (305 m). However, during two tracer injections (June 19, September 21), uptake length was greater in the upper reach than the lower reach, and less than the lower reach during two additional injections (September 29, October 20).
The hydraulic retention factor, a measure of the storage zone residence time per unit of stream reach traveled, showed no clear trend with discharge (Figure 3.7C). During low flow conditions, retention was highest, ranging from 5.7 s/m at 1.0 L/s to 7.4 s/m at 5.5 L/s. At mid to high stream flow rates (10 L/s to 30 L/s) a slight increasing trend in retention was observed; however, retention factors did not reach low flow values. The standardized storage zone area also did not show a clear trend with discharge (Figure 3.7D). A majority of the values ranged from approximately 0.10 to 0.30, with the exception of the May 31 injection at a value of 0.62.

![Figure 3.6. Hydraulic residence time of solutes in the stream (A) and storage zone (B) versus stream discharge (Q) for the upper and lower reach.](image-url)
Figure 3.7. Stream velocity (A), hydraulic uptake length (B), hydraulic retention factor (C) and the standardized storage zone coefficient (D) versus stream discharge (Q) for the upper and lower reach.

3.3.5. Lateral inflow rates

Lateral inflow rates increased with discharge (Figure 3.8). Inflow rates ranged from $3.26 \times 10^{-5}$ m$^3$ s$^{-1}$ m$^{-1}$ at low flow conditions to $3.18 \times 10^{-5}$ m$^3$ s$^{-1}$ m$^{-1}$ at high flow conditions. Net inflow represents less than 1% gain of streamflow. Lateral inflow rates did not appear to vary significantly between reaches.
3.3.6. Quantifying pool storage and residence times

Breakthrough curves from two pools located in the upper reach and one pool located in the lower reach were simulated using OTIS-P. Results are presented for one injection on September 30 (Figure 3.9). Additional model simulations are presented in Appendix A. For all pool simulations, the transient storage area was greater than channel area, resulting in a larger $A_s/A$ ratio than the reach simulation (Table 3.6). High uncertainty values were observed for parameter estimates associated with dispersion ($D$) and storage area ($A_s$) for two pool simulations (Figure 3.10). The simulations for pool 2 and pool 3 had the greatest uncertainty based on the Damkohler number. The Dal numbers were 0.09 and 2.8, respectively, compared to a Dal number of 0.8 in pool 1.

Storage zone residence times were higher than for the entire reach for both pools located in the upper reach (Figure 3.11). Residence times were also greater than the reach scale estimated values. However, this pattern was not supported in the pool located in the lower reach (Pool 3).
Table 3.6. Summary of the simulated parameter estimates of dispersion coefficient (D), stream cross-sectional area (A), cross-sectional area of storage zone (As), and the storage zone exchange coefficient (α), stream velocity (u), the standardized storage zone coefficient (As/A) and the Damkohler number (Dal).

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Reach</th>
<th>Q (10⁻³m³s⁻¹)</th>
<th>D (10⁻²m²s⁻¹)</th>
<th>A (10²m²)</th>
<th>As (10⁵m²)</th>
<th>α (10⁴s⁻¹)</th>
<th>U (ms⁻¹)</th>
<th>As/A</th>
<th>Dal</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept 29</td>
<td>Reach 1</td>
<td>1.0</td>
<td>2.9</td>
<td>2.8</td>
<td>0.6</td>
<td>3.0</td>
<td>0.04</td>
<td>0.21</td>
<td>0.50</td>
</tr>
<tr>
<td></td>
<td>Pool 1</td>
<td>2.5</td>
<td>6.3</td>
<td>60.7</td>
<td>3.1</td>
<td>0.02</td>
<td>9.61</td>
<td>0.80</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pool 2</td>
<td>1.8</td>
<td>0.8</td>
<td>6.1</td>
<td>3.4</td>
<td>0.13</td>
<td>7.40</td>
<td>0.09</td>
<td></td>
</tr>
<tr>
<td>Sept 30</td>
<td>Reach 2</td>
<td>1.9</td>
<td>4.6</td>
<td>6.5</td>
<td>1.3</td>
<td>0.85</td>
<td>0.03</td>
<td>0.20</td>
<td>0.17</td>
</tr>
<tr>
<td></td>
<td>Pool 3</td>
<td>8.4</td>
<td>2.5</td>
<td>5.9</td>
<td>145</td>
<td>0.07</td>
<td>2.33</td>
<td>2.60</td>
<td></td>
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</tbody>
</table>

Figure 3.9. Model simulations using OTIS-P for September 30. Results are from one pool location in the lower reach.
Figure 3.10. Simulated model parameters for solute releases in the upper and lower stream reach during September 29 and 30. Pools 1 and 2 were located in the upper reach. Pool 3 was located in the lower reach. Dispersion coefficient (D), stream cross-sectional area (A), cross-sectional area of storage zone ($A_s$), storage zone exchange coefficient (α) versus location. Error bars represent ± 1 standard deviation.
3.3.7. **Subsurface relative connectivity**

A total of 5 to 8 observations of mixing ratios (Equation 2.27) were made for each piezometer, except for piezometers P56, P18, and P48, where only three observations were conducted due to the drying out of the stream channel over the study period. The relative connectivity varied between reaches (Figure 3.12). The upper reach appeared to have a higher percentage of piezometers where the tracer concentration in the piezometer decreased during the tracer injection, as indicated by a negative value of $\chi$ (e.g. P16, P20, P22, and P21). A negative mixing ratio could be a result of stratification of water chemistry in the streambed. In general, the tracer concentration for piezometers located within the lower reach increased during tracer injections as indicated by a positive value of $\chi$. This suggests that piezometers located within the lower reach have a higher degree of connectivity to water in the stream channel. There was no clear distinction between upwelling and downwelling zones in terms of relative connectivity. In the upper reach,
hydraulic gradients were not measured in piezometers 16, 18 and 19. All three sites were located at the upstream end of a riffle habitat.

Figure 3.12. Relative connectivity (RC) as measured using a non-dimensional mixing ratio (x) for piezometers sampled during tracer injection experiments in the upper and lower reach

3.4. Solute injection experiments – Channel-unit scale

3.4.1. Qualitative observations of hyporheic discharge

Two different flow pathways were observed within one channel-unit in the lower reach (Figure 3.13). The following observations were made during experiments at the infiltrometer I-5 location:

- **July 11** - RWT tracer water infiltrated the bed, traveled laterally around a large anchor boulder in the riparian zone and returned to the stream at the base of the step within 5-10 minutes post-injection.
• **August 14** – RWT followed the same flow pathway as July 11. RWT was visible at the start of the pool within four minutes post-injection and was visible within the pool until approximately 18 minutes post-injection.

• **September 27** - Discharging hyporheic water was again observed at this location; however, during this experiment RWT infiltrated the streambed, travelled vertically through the step and returned to the stream at the start of the pool. RWT was visible in the pool within 9 minutes, and was still slightly visible within the pool after one hour.

Similar observations were made at infiltrometer I-4:

- **August 21** - No RWT was visible within the pool after 4 hours of observation. However, an increase in fluorescence (as measured using a fluorometer) was observed. Fluorescence returned to background in approximately 24 hours, suggesting the presence of hyporheic flow at this location.

- **September 28** – RWT infiltrated the streambed, traveled vertically through the step and returned to the stream at the start of the pool.

- **October 6** - Discharging hyporheic water was observed at the base of the pool (similar location as the September 27 and 28 injections). RWT was observed 13 minutes after injection and was still visible in the pool after 1 hr 24 min. No tracer was observed at 1 hr 48 min after injection.

- Injections of Brilliant Blue FCF were also conducted; however, discharging dye was not observed at any locations within the lower reach.

Observations were also made at infiltrometer I-1 installed in a log-step in the upper reach; however, discharging of RWT or Brilliant Blue FCF was not observed during any tracer injection experiments at this location (July 12, August 23, and September 28, October 6), except for an initial trial on June 27. Hyporheic discharge of Brilliant Blue FCF was observed within the log-step channel-unit during the trial on June 27. Tracer infiltrated the stream bed and was observed upwelling from the sediments at the upstream end of the pool.
3.4.2. Quantifying residence times

The pool behaved like a continuously stirred tank reactor (CSTR), as indicated by a straightening of the late time period of the breakthrough curve when concentration is plotted on a logarithmic scale (Figure 3.14B, Figure 3.15B). MRT varied over the experiments possibly due to variability in streamflow and hydraulic gradients (Table 3.7).

Electrical conductivity increased rapidly to a peak concentration within the step and pool sub-units. During both experiments, the maximum electrical conductivity was greater in the step than pool. Additional minor peaks in EC are visible on the rising and recession limb of the breakthrough curves for each experiment. These peaks could indicate separate flow pathways with different residence times within the channel-unit.

Table 3.7. Mean residence times for hyporheic zone (step) and pool storage zones in Pool 4. Vertical hydraulic gradients (VHG) measured using piezometer 61.

<table>
<thead>
<tr>
<th>Date</th>
<th>Q (Ls⁻¹)</th>
<th>VHG (cm/cm)</th>
<th>Mean Residence Time (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept 25</td>
<td>2.4</td>
<td>-0.96</td>
<td>23.8</td>
</tr>
<tr>
<td>Oct 5</td>
<td>1.4</td>
<td>-0.89</td>
<td>68.8</td>
</tr>
</tbody>
</table>
Figure 3.14. Step-pool residence time experiment conducted on September 25, 2006

Figure 3.15. Step-pool residence time experiment conducted on October 5, 2006
3.5. Subsurface flow – Point scale

3.5.1. Hydraulic gradients

Considerable spatial and temporal variability in hydraulic gradients was observed at all locations. Strong negative hydraulic gradients typically occurred at the start of a step or riffle indicating infiltration into the stream bed (Figure 3.16). Consistent negative gradients were observed in piezometers 1 to 6 within the log-step channel-unit, although gradients varied temporally (Figure 3.17). In general, gradients tended to get weaker as stream flow decreased during low flow conditions in mid to late summer. Positive hydraulic gradients were consistently observed downstream from obstructions in the stream channel, such as the log step located in the upper reach 13-15 m downstream. Piezometers 10, 11, 13, 14 and 15 showed consistent positive hydraulic gradients over the study period (Figure 3.18); however, hydraulic gradients varied over short spatial scales. For example, piezometers 10 and 13 were located approximately 20 cm apart and were installed at similar depths; however, the average gradient in piezometer 13 was 1.06 cm/cm over the study period compared to 0.03 cm/cm in piezometer 10.

A similar pattern of hydraulic gradients was observed in a boulder-step channel-unit located in the upper reach 20 m downstream from the culvert crossing. Piezometers 21 to 23 showed a consistent strong downwelling response (Figure 3.19A). The magnitude of response increased as the distance to the step increased with piezometer 23 showing the strongest response (average = -0.6 cm/cm). Piezometers 24 to 26 all showed consistent upwelling and were located near the head of the pool (Figure 3.19B).
Figure 3.16. Vertical hydraulic gradients measured in the upper (A) and lower (B) reaches. Symbols indicate study period means.
Figure 3.17. Vertical hydraulic gradients measured over the study period in piezometers 1-6 in the upper reach.

Figure 3.18. Vertical hydraulic gradients measured over the study period in piezometers 10,11,13-15 in the upper reach.
Figure 3.19. Vertical hydraulic gradients measured over the study period in the piezometers 21-23 and piezometers 24-26 in the upper reach

Upwelling sites were also observed within the lower reach, although hydraulic gradients were not as strong as in the upper reach. Values for upwelling sites tended to be within measurement error (± 0.05 cm/cm). Reversal of hydraulic head gradients was also more common within the lower reach, with piezometer response switching between neutral, downwelling and upwelling over the course of the study period. For example, in one boulder step channel-unit, piezometers installed at the downstream end of a large step (60, 61 and 62) showed consistent downwelling (average -1.3, -0.88, -1.1 cm/cm respectively), whereas the gradients measured in piezometers located within the pool (67 and 68) fluctuated from neutral to slight upwelling and back to downwelling during the late summer (Figure 3.20). Positive VHG's observed in piezometers 67 and 68 during mid to late September correspond to qualitative observations made at the channel-unit, where RWT injected into the streambed traveled vertically through the step and returned to the pool at the start of the pool during an injection on September 28. Prior to this, exchange flow was only observed flowing laterally around a large anchor boulder located within the channel-unit.
Of the 66 piezometers tested, only four locations had correlations between discharge and VHG using a non-parametric Spearman's rank correlation that are significant at $\alpha = 0.05$ (Table 3.8). An upwelling (P25) and downwelling site (P2) had negative correlations with discharge, while two downwelling sites had positive correlations (P27, P47) with discharge. These results suggest that stream discharge does not control vertical hydraulic gradients in East Creek.

Table 3.8. Spearman correlation coefficient ($r_s$), associated p-values, number of observations (n) and vertical hydraulic gradients (VHG) for each piezometer indicating a significant correlation with discharge.

<table>
<thead>
<tr>
<th>Piezometer</th>
<th>$r_s$</th>
<th>P-value</th>
<th>N</th>
<th>VHG (cm/cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>-0.87</td>
<td>0.002</td>
<td>9</td>
<td>-0.01</td>
</tr>
<tr>
<td>25</td>
<td>-0.72</td>
<td>0.02</td>
<td>10</td>
<td>0.03</td>
</tr>
<tr>
<td>27</td>
<td>0.84</td>
<td>0.02</td>
<td>7</td>
<td>-0.12</td>
</tr>
<tr>
<td>47</td>
<td>0.62</td>
<td>0.04</td>
<td>11</td>
<td>-0.03</td>
</tr>
</tbody>
</table>
3.5.2. VHG and scaled location within channel units

Zones of hyporheic discharge and recharge appear to be a function of the scaled location in the channel unit (Figure 3.21). Zones of hyporheic discharge, or upwelling (as indicated by a positive VHG), were generally confined to the upper portion of the channel units (X/L = 0.0 to 0.4). For X/L > 0.2, there is a trend to increasingly negative hydraulic gradients with increasing distance from the head of the channel-unit. The height of the downstream step does not appear to control hydraulic gradients. A Spearman’s rank correlation analysis indicated that hydraulic gradients were not correlated with step height (r_s = -0.186, p = 0.17). In addition, a sequential analysis of variance using three linear mixed-effects models confirmed that position within the stream channel (as defined by a category of X/L) significantly contributed to the observed variability in hydraulic gradients (χ^2 = 40.9, p < 0.001, Table 3.9). The addition of downstream step-height to the base model (i.e. channel unit as a random effect) did not significantly contribute to the observed variability in hydraulic gradients (χ^2 = 4.6, p = 0.09, Table 3.9).

Table 3.9. Analysis of variance table comparing three linear mixed-effects models for vertical hydraulic gradients including a base model with only channel-unit, a second model with channel-unit and step height (SH), and a third model with channel-unit, step-height and channel position (X/L). A chi-square (χ^2) statistic was used to test for significance.

<table>
<thead>
<tr>
<th>Model</th>
<th>DF</th>
<th>χ^2</th>
<th>DFχ^2</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unit + SH</td>
<td>4</td>
<td>4.6</td>
<td>2</td>
<td>0.09</td>
</tr>
<tr>
<td>Unit + SH + XL</td>
<td>8</td>
<td>40.9</td>
<td>4</td>
<td>&lt;0.001</td>
</tr>
</tbody>
</table>
Figure 3.21. Vertical hydraulic gradient (cm/cm) versus scaled location within the channel-unit. Hydraulic gradients are averaged over the entire study period.

Figure 3.22. Vertical hydraulic gradient (cm/cm) versus step height (m) as a function of scaled location within the channel-unit (X/L). Hydraulic gradients are averaged over the entire study period.
3.5.3. Hydraulic conductivity

The geometric means of conductivities for the lower and upper reaches were $2.54 \times 10^{-4}$ (n = 24) and $2.37 \times 10^{-4}$ m/s (n = 16) respectively. Conductivities appeared to be higher at neutral and upwelling sites than at downwelling sites in the lower reach (Figure 3.23); however, only three sites in the analysis were considered upwelling sites, compared to downwelling (n = 21) and neutral sites (n = 17). A sequential analysis of variance using two linear mixed-effects models confirmed that site condition significantly contributed to the observed variability in hydraulic conductivity ($\chi^2 = 6.7$, p = 0.01, Table 3.10). Hydraulic conductivity also did not appear to vary systematically with installation depth (Figure 3.24).

Table 3.10. Analysis of variance table comparing two linear mixed-effects models for hydraulic conductivity (log transformed), including a base model with only reach as a factor and a second model with reach and site condition (upwelling, neutral and downwelling).

<table>
<thead>
<tr>
<th>Model</th>
<th>DF</th>
<th>$\chi^2$</th>
<th>DF $\chi^2$</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reach</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reach + Site Condition</td>
<td>3</td>
<td>6.7</td>
<td>1</td>
<td>0.01</td>
</tr>
</tbody>
</table>
Figure 3.23. Hydraulic conductivity (K) for downwelling (D), neutral (N) and upwelling (U) sites located in the lower (n = 24) and upper reach (n = 17). Note log scale.

Figure 3.24. Hydraulic conductivity (K) with depth of piezometer installation for the upper and lower reaches. Note log scale.
3.5.4. **Streambed infiltration rates**

Temporal variation in infiltration rates was observed over the study period (Figure 3.25). Rates were difficult to measure with the constant head infiltrometer during low flow conditions, resulting in a lack of observations during base flow conditions (mid to late August). The probable errors for infiltration measurements were almost ± 60% of the measured value.

The relationship between discharge and infiltration rates was tested using a non-parametric Spearman’s rank correlation. In two locations, infiltration rates were significantly correlated with discharge (Table 3.11), including the sediment-step (I-3) and boulder step (I-5). The boulder step (I-5) location also had the strongest mean VHG (-1.1 cm/cm).

Hydraulic conductivity estimates based on streambed infiltrometers (Equation 2.26) were higher than estimates from falling head tests (Figure 3.26). This result suggests that bed infiltration computed from piezometer data alone may underestimate actual infiltration rates.

![Figure 3.25. Infiltration rates over the study period. Error bars represent probable errors based on Equation 2.25.](image)
Table 3.11. Spearman correlation coefficient ($r_s$), associated p-values and number of observations (n) for infiltration rates versus discharge at each infiltrometer location.

<table>
<thead>
<tr>
<th>Location</th>
<th>$r_s$</th>
<th>p-value</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>Log step (1-1)</td>
<td>0.37</td>
<td>0.21</td>
<td>13</td>
</tr>
<tr>
<td>Sediment step (1-3)</td>
<td>0.92</td>
<td>0.01</td>
<td>5</td>
</tr>
<tr>
<td>Boulder step (1-4)</td>
<td>0.51</td>
<td>0.25</td>
<td>6</td>
</tr>
<tr>
<td>Boulder step (1-5)</td>
<td>0.83</td>
<td>0.01</td>
<td>7</td>
</tr>
</tbody>
</table>

Figure 3.26. Hydraulic conductivity calculated using infiltration rates and slug-tests for five locations. Values represent the geometric mean ± standard error.

3.5.5. **Streambed water fluxes computed from Darcy’s law**

Streambed water fluxes within one step-pool unit in the upper reach (Pool 1) varied with flow conditions. The total computed flux into the bed was larger during high flow conditions on June 19 ($Q = 15.4$ L/s) than during low flow conditions on September 29 ($Q = 1.1$ L/s), as calculated using Equation 2.23. The fluxes into the bed were also larger than fluxes out of the bed as summarized in Table 3.12. The average fluxes out of
the bed, as categorized where $X/L < 0.4$, did not change significantly with flow (Figure 3.27). However, the fluxes into the area above the step, where $X/L > 0.6$, did increase slightly with flow, specifically where $X/L \approx 0.8$. The reach scale estimate of hyporheic exchange was two-orders of magnitude greater than the scaled-up estimate of exchange for both flow conditions (Table 3.12).

![Figure 3.27. Water fluxes calculated using Darcy’s Law for each $X/L$ category within one step-pool channel-unit during low flow ($Q = 1.1 \, L/s$) and high flow ($Q = 15.4 \, L/s$)](image)

Table 3.12. Water fluxes within one step-pool unit (Pool 1) along with scaled-up and reach-scale estimates of hyporheic exchange ($s^{-1}$).

<table>
<thead>
<tr>
<th>Date</th>
<th>$Q$ ($L/s$)</th>
<th>$q_{in}$</th>
<th>$q_{out}$</th>
<th>Scaled-up ($10^6 , s^{-1}$)</th>
<th>Reach scale $\alpha$ ($10^4 , s^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>June 19</td>
<td>15.4</td>
<td>0.78</td>
<td>-0.07</td>
<td>4.1</td>
<td>2.1</td>
</tr>
<tr>
<td>September 29</td>
<td>1.1</td>
<td>0.46</td>
<td>-0.02</td>
<td>2.6</td>
<td>3.0</td>
</tr>
</tbody>
</table>
CHAPTER FOUR: DISCUSSION

This chapter discusses the results presented in Chapter 3. Sections 4.1 to 4.3 discuss the research objectives outlined in Chapter 1. The last section integrates observations from the different scales of interest.

4.1. Reach scale

4.1.1. Modelled parameter uncertainty

Wagner and Harvey (1997) explained that parameter uncertainty is minimized when \( \text{Dal} = 1.0 \). Parameter uncertainty, especially with respect to the estimates of transient storage and exchange, increases in experiments with very high or low \( \text{Dal} \) values. In East Creek, \( \text{Dal} \) values were less than 1.0 for all experiments and ranged from 0.1 to 0.5, which is consistent with Wagner and Harvey’s (1997) conclusion that “well estimated” parameters are likely to be obtained when the Damkohler number is on the order of 0.1-1.0. Modelled parameter estimates in East Creek were also fairly consistent with the typical range of parameter values reported in the literature for high gradient streams (1-15%) before 1997 (Wagner and Harvey 1997, Table 4.1).

Under conditions when \( \text{Dal} < 1.0 \), parameter uncertainty is high because only a small amount of tracer interacts with the storage zone (Wagner and Harvey 1997). This may occur because: (1) stream velocity is high, (2) exchange timescales are short, as indicated by low values of \( \alpha \) or (3) the reach length is short. Parameter uncertainty was greatest during periods of higher flows, possibly due to the slow rates of transient exchange relative to the stream water velocity. The transient exchange coefficient (\( \alpha \)) was estimated at 0.00014 \( \text{s}^{-1} \) under high flow conditions on May 31 (\( Q = 30.6 \text{ L/s, } u = 0.14 \text{ m/s} \)), which was half the average \( \alpha \) for all simulations (0.00022 \( \text{s}^{-1} \)). Harvey et al. (1996) concluded that the stream tracer method does not reliably characterize exchange at higher flows, which complicates the efforts of studies examining the influence of discharge on transient storage processes over a range of flow conditions (e.g. Hart et al. 1999).
The Damkohler number was highest in pool 3 (Dal = 2.8) compared to values of 0.8 and 0.09 in pools 1 and 2 respectively, indicating higher uncertainty in the parameter estimates associated with transient storage. However, parameter uncertainty was lower in pool 1 than at the reach scale (Dal = 0.5). The low Dal numbers within pool 1 and 2 may be attributed to the length of the pool (4 m) compared to 50 m at the reach-scale. Wagner and Harvey (1997) indicated that the length of the study reach should be adjusted to maintain a balance between advective transport and transient storage. This suggests that reducing the "reach-length" to the scale of individual pools may have contributed to parameter uncertainty. The high Dal number in pool 3 indicates that exchange rates were fast relative to the stream water velocity and that all the solute was exchanged with the storage zone over the reach length (Wagner and Harvey 1997).

The uncertainty ratio was also used to quantify parameter uncertainty for reach scale and channel-unit scale simulations. Ratios were lowest for the reach-scale dispersion and transient exchange parameters (Table 3.4), which indicates a higher degree of uncertainty. Low uncertainty ratios were also associated with the dispersion and transient storage parameters for two of the three pool simulations. Uncertainty ratios were within the ranges reported by a previous study examining the broad heterogeneity of hyporheic zone processes across seven small streams in western Washington (Reidy 2004). However, that study encompassed a wider range of flow conditions (0.7 to 216 \(10^3\) m\(^3\)/s) and channel morphologies than studied in East Creek.

Table 4.1. Range of parameter values reported for high-gradient streams (Wagner and Harvey 1997) compared to modelled parameter values in East Creek.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Wagner and Harvey (1997)</th>
<th>East Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q (m(^3)/s)</td>
<td>0.005 - 0.2</td>
<td>0.002 - 0.03</td>
</tr>
<tr>
<td>Q(_L) (m(^3)/s/m)</td>
<td>0 - 0.0001</td>
<td>0 - 0.00003</td>
</tr>
<tr>
<td>A (m(^2))</td>
<td>0.02 - 0.6</td>
<td>0.06 - 0.2</td>
</tr>
<tr>
<td>D (m(^2)/s)</td>
<td>0.025 - 0.8</td>
<td>0.02 - 0.5</td>
</tr>
<tr>
<td>A(_S) (m(^2))</td>
<td>0.01 - 2.0</td>
<td>0.006 - 0.1</td>
</tr>
<tr>
<td>(\alpha) (s(^{-1}))</td>
<td>(5.0 \times 10^{-6}) - 0.001</td>
<td>(8.0 \times 10^{-5}) - 0.0001</td>
</tr>
</tbody>
</table>
4.1.2. **Solute transport parameters and discharge**

Solute transport parameters varied with discharge at the reach scale. The results of the OTIS-P simulations were in partial agreement with previous studies which reported an increase in dispersion (D) and the channel cross-sectional area (A) with discharge (Legrand-Marcq and Laudelout 1985, D'Angelo et al. 1993, Harvey et al. 1996, Morrice et al. 1997, Hart et al. 1999, Wondzell 2005). However, those studies reported that transient storage area (A_s) decreased with discharge while transient exchange (a) increased with increasing discharge, with the exception of Legrand-Marcq and Laudelout (1985) who reported that transient exchange remained constant with discharge. In East Creek, on the other hand, exchange rates remained fairly constant with discharge while the storage area increased. A previous study in the upper and lower reaches of East Creek observed similar trends in the response of a and A_s (Patschke 1999), as did Hart et al. (1999). A recent study in a fourth-order stream in central Michigan (USA) found that the size of the transient storage zone increased with discharge from 1.9 m² at baseflow (2.5 m³/s) to 7.3 m² at a flow of 19.1 m³/s (Phanikumar et al. 2007).

Studies have attributed a constricting of the transient storage area during periods of higher discharge to increased groundwater discharge and catchment wetness (Boulton et al. 1998, White 1993). As the catchment wetness increases, hydraulic gradients to the stream from the riparian zone are stronger and can overwhelm the influence of channel morphology (Harvey and Bencala 1993, Wondzell and Swanson 1996a), resulting in a decrease in the extent of the hyporheic zone.

Channel complexity is cited as a possible explanation for the observed differences in the response of the transient storage area (A_s) to discharge in East Creek (Patschke 1999). Storage area increased with discharge in two reaches associated with a relatively high degree of stream complexity (i.e. upper and lower reach) and remained constant with discharge in two less complex reaches located upstream from the current study location (Patschke 1999). This study attributed the difference to variability in stream complexity and suggested that as discharge increased, the contribution from storage in pools and back eddies may have contributed to the increase in storage area. Story (2002)
also attributed the variability in storage zone cross-sectional area within three sub-reaches to differences in channel complexity.

The ratio $A_s/A$ increased with discharge in both reaches. This trend is in contradiction to results presented by Harvey et al. (1996) and Morrice et al. (1997) in which $A_s/A$ decreased with discharge. Patschke (1999) reported a similar trend for the lower reach; however, a decreasing trend in $A_s/A$ was observed for the upper reach. D'Angelo et al. (1993) and Morrice et al. (1997) reported transient storage zones that were as large as or larger than the surface water area. Mulholland et al. (1997) determined that metabolic rates and nutrient uptake were significantly greater in streams with a larger transient storage zone relative to the channel cross-sectional area ($A_s > A$). The transient storage area ($A_s$) in East Creek was consistently smaller (on average 70%) than the surface water area ($A$). Estimates of $A_s/A$ ranged from 0.1 to 0.6 in both reaches. At higher discharges, low-lying areas adjacent to the main channel in East Creek were filled with stagnant water which may have provided additional surface storage. As well, at higher flows, a side channel at mid-reach (approximately 45 – 50 m) downstream from the culvert at M road was activated. Wondzell and Swanson (1996a) observed significant exchange flow between primary and secondary stream channels. At higher discharges more wetted channel area may also be available for hyporheic exchange to occur.

The transient exchange coefficient remained fairly constant with discharge. Some evidence of a threshold response could be suggested, as transient exchange plateau to a value of $2.7 \times 10^4$ s$^{-1}$ at 5.5 L/s and then declined with discharge in the lower reach. Morrice et al. (1997) also reported a threshold response in which $\alpha$ increased then decreased with discharge in a first-order stream. Additional studies have reported either a steady increase (D'Angelo et al. 1993, Harvey et al. 1996, Hart et al. 1999, Wondzell 2005) or a decrease in transient exchange with discharge (Legrand-Marcq and Laudelout 1985, Hart et al. 1999, Patschke 1999).
4.1.3. Residence times and retention

Residence times were higher in the stream channel than in the transient storage zone for all tracer injections. Residence times also varied spatially (i.e. between reaches), but did not vary temporally with discharge (Figure 3.6). Between reaches, stream channel and transient storage residence times were consistently higher in the lower reach. This variation may be attributed to stream complexity. Wondzell (2005) observed that the storage zone residence times and the hydraulic retention factor ($R_h$) were greater in reaches with log jams than in a companion pool-step reach. The $R_h$ factor was also consistently greater in unconstrained reaches than constrained reaches. Retention factors were fairly comparable between the two reaches in East Creek (Table 3.5), despite the upper reach having a greater number of log steps and a higher degree of incision. The lower reach was generally less confined than the upper reach, which may explain the longer residence times observed for transient storage.

Reach-scale estimates of residence times and retention are not consistent with estimates for individual pools during two tracer injections. For both pools located in the upper reach (Pools 1 and 2), storage zone residence times were greater than in the stream channel. Storage zone residence times were also greater than reach scale estimated values. Harvey et al. (1996) assumed that the residence time for the in-channel transient storage is very short and is therefore accounted in the dispersion coefficient rather than the transient storage coefficient. Results from TSM show that residence times within the pool are higher than the main channel. This suggests that it may be invalid to assume that the residence times within the in-channel storage zones (i.e. pools or back eddies) are negligible. However, this pattern was not completely supported in the pool located in the lower reach (Pool 3). Residence times in both the stream channel and storage zones were between 1 to 2 min and were lower than the reach-scale estimates of 196 and 39 min, respectively.
4.1.4. In-channel transient storage

A major criticism of current transient storage models is the inability to separate in-channel transient storage from storage within the hyporheic zone (Harvey et al. 1996). Based on the assumption that the dominant transient storage processes within the pools are from in-stream transient storage, TSM simulations conducted at the scale of individual pools were used to separate in-channel storage from hyporheic exchange. Transient storage area ($A_s$) was generally higher within the pools, resulting in a higher $A_s/A$ ratio than at the reach scale (Table 3.6). The transient storage area parameter ($A_s$) is assumed to incorporate both storage processes despite the inability to separate in-channel transient storage from hyporheic zone storage (Harvey et al. 1996). The results from the individual pool simulations suggest that this may be a valid assumption.

4.2. Channel-unit scale

4.2.1. Variability in exchange flow pathways

Qualitative and quantitative observations made at the channel-unit scale highlight the temporal variability in exchange flows. Within one channel-unit (Pool 4), two separate flow pathways were observed (Figure 3.11). On July 11 and August 14, RWT tracer water infiltrated the bed, traveled laterally due to the deflection from a large anchor boulder in the riparian zone and returned to the stream at the base of the step. This confirms that exchange flow pathways can include a lateral flow component, as described by model 2 in Section 1.2.2. However, in this channel-unit, lateral flow was a result of the step morphology, and therefore does not confirm whether zones of upwelling are the result of lateral inflow or return flow from the riparian zone and adjacent hillslope. On September 27, RWT infiltrated the streambed, traveled vertically through the step and returned to the stream at the start of the pool. This is described as model 1a, or the typical flow pathway, in Section 1.2.2.

Additional injections in a different location within the channel-unit (location I-4) and in the upper reach (location I-1) confirmed Model 1a. Discharging Brilliant Blue FCF
tracer water was only observed on one occasion within the upper reach (June 27). Even though RWT tracer was injected in a comparable quantity at this location (location 1-1), discharging tracer water was not observed with the log-step in the upper reach. This observation may be due to Rhodamine WT absorbing to the sediments or organics in the streambed. Rhodamine has been observed to absorb to sediments in laboratory experiments, especially finer sediments (Munn and Meyer 1988).

Exchange flows are induced by the hydraulics of streamflow over an irregular streambed in the process known as “advective pumping exchange” as described by Savant et al. (1987) and Thibodeaux and Boyle (1987). These exchange flows are controlled by the spatial and temporal variability in hydraulic heads along the stream boundary, resulting in a distribution of travel times through the hyporheic zone. Observations made at the channel-unit scale highlight the variability in residence times associated with transient storage. Solute residence times within an individual pool were consistently greater than those in the step (or hyporheic zone) based on modelling the mean residence time of solutes within both storage zones using linear reservoir theory. Mean residence times varied over between experiments possibly due to variability in streamflow and hydraulic gradients (Table 3.7). Previous studies have reported a wide range of residence times in the hyporheic zone; for example, Wörman et al. (2002) reported residence times in the range of 160 to 800 minutes.

Examining the breakthrough curves from these stream tracer experiments also highlights the variability in exchange flow pathways observed within a small spatial area (Figure 3.14, 3.15). Electrical conductivity increased rapidly to a peak concentration within the step and pool sub-units. The maximum EC was also greater in the step than pool. This could be an effect of dispersion or possibly indicate a dilution in concentration within the pool. However, additional minor peaks in EC are visible on the rising and recessional limb of the breakthrough curves for each experiment, which could indicate separate flow pathways with different residence times within the channel-unit.
4.2.2. Transient storage modelling

Modelling residence times within the pool using linear reservoir theory indicated that the pool behaved like a continuously stirred tank reactor (CSTR). Residence times within the pool were higher than the hyporheic zone. This suggests that it may be an invalid assumption to assume that the residence times within pools are negligible (Harvey et al. 1996). Results from OTISP-P simulations in Pool 1 and 2 also support this hypothesis.

Residence times within both transient storage zones (i.e. step and pool sub-units) followed an exponential distribution. These results support the use of an exponential residence time distribution to model transient storage. Late-time solute residence times or “tailing” of solute concentration after the main solute pulse, are represented as an exponential probability density in current TSM’s (Bencala and Walters 1983). Additional studies have suggested alternative approaches to represent the timescale of transient exchange including TSM’s with multiple storage zones (Choi et al. 2000). However, the results from this study suggest that it may be valid to use one transient storage zone to represent residences times using an exponential distribution.

4.3. Point scale

4.3.1. Interpretation of flow pathways

Repeated observations of vertical hydraulic gradients measured from piezometers installed within the stream channel confirmed the currently accepted conceptual model of exchange flow within step-pool streams, which involves infiltration into the bed upstream of a step and subsequent discharge a short distance downstream in the bottom of pool as described by Harvey and Bencala (1993).

In East Creek, downwelling flow was observed upstream of obstructions in the channel (i.e. steps and logs) with upwelling occurring at the start of the pool below a step (Figure 3.5). Patschke (1999) similarly observed strong downwelling in piezometers located upstream from steps in East Creek. In contrast to the observations of upwelling
observed in East Creek and by Moore et al. (2005a) at another stream in Malcolm Knapp Research Forest, Wondzell (2005) failed to locate areas of upwelling within a step-pool reach, despite predictions from groundwater flow models that upwelling should occur. This may be partially explained by that author’s difficulty with installing piezometers in the streambed without significantly disturbing the bed sediments surrounding the piezometer (S. Wondzell, pers. comm.). Additional studies in the Lookout Creek basin (Oregon, USA), have also not observed coherent hyporheic discharge below steps, despite predictions from groundwater flow models that upwelling should occur (Anderson et al. 2005, Gooseff et al. 2005).

Although upwelling was observed in locations where it was expected in East Creek, the hydraulic gradients were substantially weaker than in the downstream sections of the step-pool units (i.e., X/L > 0.6). Given that hyporheic exchange should be approximately at a steady state during extended periods of baseflow, upwelling and downwelling fluxes should balance. The fact that downwelling gradients were generally stronger and more spatially widespread than upwelling gradients could be explained by one or more of the following: (1) greater hydraulic conductivities in upwelling zones, (2) discharge being concentrated along preferred pathways, and (3) the presence of lateral flow paths. All three are possibilities at East Creek. The presence of higher conductivities in upwelling zones would be consistent with the observation that penetration of fine sediments into the bed (resulting in clogging of larger pore spaces) is greater in downwelling areas (Schalchli 1992, Packman and MacKay 2003). Furthermore, there is weak empirical evidence supporting the contrast in hydraulic conductivities between upwelling and downwelling zones (Figure 3.23). There is some support for the presence of preferred pathways from the unit-scale tracer experiments, although it is difficult to quantify how much discharge occurs via these preferred pathways (section 3.4). The presence of a lateral component to hyporheic exchange is possible, particularly at some areas along the reach, but the study design did not allow the assessment of lateral hyporheic exchange because horizontal gradients were not measured.

Zones of hyporheic discharge and recharge varied with position in the stream channel (as defined by X/L, Table 3.9). Generally, for X/L > 0.2, negative hydraulic gradients increased with distance to the step; however, the actual step height does not
appear to control VGH (Table 3.9). Where X/L > 0.8, there appears to be a slight negative trend with increasing step height (Figure 3.22); however, this relationship was not significant. These results suggest that the relation between VHG and X/L could be a useful tool for characterizing and predicting hydraulic gradients in step-pool streams, a need expressed by Bencala (2000). Channel-unit spacing, size and sequence are also documented as significant controls on exchange flow in previous studies (Anderson et al. 2005, Gooseff et al. 2006). However, the influence of step-height on exchange flow does not appear to have been directly quantified in previous studies.

4.3.2. Interactions between hyporheic flow and lateral inflow

Increased lateral or groundwater inflows from adjacent hillslopes can reverse head gradients along the stream margin, leading to a reduction in hyporheic zone extent and the degree of hyporheic exchange (Harvey and Bencala 1993, Wroblicky et al. 1998, Storey et al. 2003, Wondzell 2005). Hydraulic head reversals (fluctuations between upwelling, downwelling and neutral) were common in the lower reach (Figure 3.7).

Groundwater discharge from the stream banks could have contributed to the observed reversal of hydraulic gradients. This observation appears to be consistent with documented climatic conditions, as hydraulic head reversals were observed following precipitation events during September and October. This may indicate that exchange flows are controlled by seasonal variability in lateral inflow from the hillslope, further suggesting that areas of upwelling could be lateral inflow from the riparian zone.

Lateral inflow steadily declined during the early part of the study period (May to August) due to the atypical dry summer conditions and general increased with increasing discharge. Streamflow measurements also confirmed that East Creek is a gaining reach; however, net lateral inflow remained less than 1% of the total streamflow. Wondzell and Swanson (1996a) found that the strength of hydraulic gradients towards the stream varied with season. During the summer months, hydraulic gradients towards the stream channel were weaker than during the winter months. A reduction in groundwater discharge during the summer months due to less rainfall and drier soil conditions may have contributed to the variability in gradients observed in this study.
Moore et al. (2005b) suggested that the upwelling sites may be connected to lateral inflows from the hillslope, particularly those with convergent topography based, on observations that upwelling sites underwent little to no mixing ($\chi = 0$) with stream water during tracer injections in one coastal headwater stream. In the lower reach, upwelling and downwelling sites appeared to be highly connected to water in the stream channel, based on consistently positive mixing ratios during reach scale tracer injections. These ratios indicate a partial replacement of hyporheic zone water with tracer-labeled stream water. In the upper reach, negative mixing ratios were observed and the median mixing ratio for most sites was near 0. These findings could indicate that water was being drawn from different depths of the hyporheic zone and thus represent water with a different chemical signature or residence time.

### 4.3.3. Water fluxes and discharge

Observations of vertical hydraulic gradients from piezometers installed within the streambed ($n = 66$) showed considerable spatial and temporal variability, suggesting that hyporheic exchange or water fluxes into the stream also vary. The high within-site variability demonstrated is not uncommon (Thibodeaux and Boyle 1987, Baxter and Hauer 2000). Physical processes, such as discharge or local bedform characteristics, are considered possible controls on temporal and spatial patterns of exchange flow.

Considerable temporal variation in water fluxes including infiltration rates and hydraulic gradients were observed over the study period. However, stream discharge was not a significant physical control on water fluxes into the bed, based on the Spearman’s rank correlation analysis relating hydraulic gradients and infiltration rates to discharge (Tables 3.3, 3.4). These findings suggest that additional mechanisms contribute to the observed temporal variability in water fluxes, including lateral inflow from the riparian.

Hydraulic conductivity was also hypothesized to be a significant physical control on the spatial variability of water fluxes into the streambed. The results of a sequential analysis of variance determined that hydraulic conductivity varied with site condition (i.e. upwelling, downwelling and neutral gradients), but not between reaches (Table 3.10). Hydraulic conductivity did not appear to vary with depth of installation in the subsurface.
(Figure 3.24). However, hydraulic conductivity as calculated from streambed infiltrometers (depth ≈ 10 cm), was higher than estimates from falling head tests (Figure 3.10), suggesting that bed infiltration computed from piezometer data alone may underestimate actual infiltration rates. These results are in contradiction with previous studies, which observed an increase in conductivity with depth (Larkin and Sharp 1992, Conrad and Beljin 1996). Those studies attributed the difference in conductivity at the streambed to the settling of silt, clay and organic materials on the surface in the process referred to as colmation (Brunke and Gonser 1997). This process can also reduce the degree of hyporheic exchange. In contrast, supplementary studies examining the conductivity of porous streambed sediments support the results observed at East Creek (Landon et al. 2001, Song et al. 2007). The streambed conductivity was typically greater than the sediments directly below this layer (≈ 30 cm) for both studies. Song et al. (2007) hypothesized that hyporheic exchange flows formed localized pathways which increased the sediment pore size resulting in an increase in hydraulic conductivity.

The observed spatial variation in conductivity could be caused by the imprecision of each measurement method. The probable error associated with infiltration measurements was almost ± 60% of the measured value. In addition, the higher conductivity value at the streambed interface could be a result of the larger sampling area of the infiltrometer (diameter ≈ 6 cm) compared to the piezometer (diameter ≈ 1 cm), such that the infiltrometers are more likely to capture the effects of infiltration via preferred pathways, consistent with the results of Song et al. (2007). Saturated hydraulic conductivity has been widely documented to increase with the volume of porous medium under consideration (Freeze and Cherry 1979).

4.3.4. Scaling streambed water fluxes

The streambed water fluxes computed from Darcy's Law within one channel-unit in the upper reach (Pool 1) were significantly greater for infiltration (where X/L > 0.6) than for discharge (where X/L < 0.4). This indicates that hyporheic discharge was a lower proportion of the total flux. Fluxes into the bed also increased with discharge, as observed on June 19 (Q = 15.4 L/s), compared to low flow conditions on September 29 (Q = 1.1).
L/s). The area of the channel-unit that contributed to the highest proportion to the total flux was where X/L > 0.8, with a rate of over 0.5 mL/s on June 19. Previous studies have suggested that at higher discharges, the hydraulic potential for downwelling into the bed increases, thus increasing the potential for hyporheic exchange flow (e.g. Wondzell 2005). Although water fluxes within this channel-unit appeared to vary with discharge, hydraulic gradients and infiltration rates measured at the point scale did not vary significantly with discharge. As a result, it is difficult to conclude whether discharge was a first order control on water fluxes within East Creek.

Attempts to “scale-up” the total flux to the reach scale estimate of hyporheic exchange (s⁻¹) indicated that the reach-scale exchange coefficient was two orders larger than the scaled-up estimate of hyporheic exchange (Table 3.12). Several processes may explain this result. Firstly, using Darcy’s Law to calculate water fluxes may underestimate the amount of exchange, due to the tremendous spatial variation in hydraulic conductivity. At least part of this bias may result from underestimation of K by the slug tests. If the infiltrometer measurements are accurate, they suggest that the estimated values of K may be an order of magnitude too low. Secondly, lateral fluxes or horizontal exchange flow may have contributed to a portion of the hyporheic exchange flow that was not quantified at the channel-unit scale. Thirdly, the TSM may overestimate the amount of exchange, possibly due to transient exchange in pools. However, this analysis was only conducted within one channel-unit, and should be extended to additional channel-units.
CHAPTER FIVE: CONCLUSIONS

The final chapter summarizes the main results of the thesis research and concludes with areas of future research.

5.1. **Summary of main results**

Hyporheic zone processes were examined at three spatial scales during the period of May to October 2006 in East Creek: reach scale, channel-unit scale and point scale.

At the reach scale, the breakthrough curves from a total of 10 stream tracer injection experiments were simulated using OTIS-P. Solute transport processes varied both temporally (with variations in discharge) and to a lesser extent, spatially (i.e. between reaches). Dispersion rates (D), channel area (A) and transient storage area (As) showed an increasing trend with discharge, while the transient exchange coefficient (α) remained fairly constant with discharge in both reaches. The ratio As/A increased with discharge. Stream and storage zone hydraulic residence times did not vary with discharge at the reach scale. Retention was highest during low flow conditions. Model parameter uncertainty was greatest during periods of high flows, possibly confounding the ability to examine transient storage processes over a range of flow conditions.

During two tracer injections (September 29 and 30), breakthrough curves from individual pools were simulated in order to quantify pool storage and residence times. Residence times within the transient storage zone of the pool (assumed to be in-channel storage) were higher than the residence time in for the entire reach. The transient storage area (As) was also generally higher within the pools, resulting in a higher As/A ratio than at the reach scale. These results suggest that it may be valid to assume that the transient storage area adequately incorporates both storage zone processes at the reach scale.

Two different flow pathways were observed during stream tracer experiments conducted at the channel-unit scale. One flow pathway was aligned with the stream channel, while a second flow pathway included a lateral component within the riparian zone. This second pathway was associated with flow around a large boulder.
Observations made at this spatial scale highlight the temporal and spatial variability in exchange flows.

At the channel-unit scale, stream tracer experiments were used to determine the mean residence time of solutes in both transient storage zones, specifically the hyporheic zone versus in-channel storage in pools. Using continuously stirred tank reactor theory, mean residence times were found to be greater within the pool than the step (i.e. hyporheic zone) during two stream tracer experiments. These results, along with the results from OTIS-P simulations within two individual pools, suggest that it may be invalid to assume that residence times within pools are negligible. Model results also confirm that the residence time distributions within the step and pool sub-units follow an exponential distribution, suggesting that current TSM (e.g. OTIS-P) do accurately represent the late-time solute residence times using an exponential probability density with one transient storage zone.

At the point scale, direct measurements of water fluxes into the stream bed, including vertical hydraulic gradients and infiltration rates, showed considerable temporal and spatial variability. However, these water fluxes did not vary statistically with discharge, suggesting that other processes contribute to the observed variability. Vertical hydraulic gradients varied systematically with the scaled location within the channel-unit, indicating that stream geometry is a significant control on water fluxes. Repeated observations of VHG, as measured in piezometers installed within the streambed, indicated a strong downwelling of water upstream from obstructions in the streambed such as boulders and logs, corresponding to where X/L > 0.8. Zones of upwelling occurred downstream at the base of pools, and corresponded to areas in the stream channel where X/L < 0.2. Step height was not a significant control on hydraulic gradients. Upwelling sites were located within both reaches, although gradients were stronger within the upper reach. Reversal of hydraulic gradients was also more common within the lower reach, possibly due to hillslope discharge.

Hydraulic conductivity measurements were spatially heterogeneous, but were within the same order of magnitude ($10^{-4}$ m/s) in both reaches. However, hydraulic conductivity estimates based on streambed infiltrometers were higher than estimates from falling head tests. This result suggests that bed infiltration computed from piezometer
data alone may underestimate actual infiltration rates. A sequential analysis of variance indicated that hydraulic conductivity varied with site condition (i.e. upwelling, downwelling and neutral sites), suggesting that conductivity is an additional control on exchange flows. This finding is also consistent with the notion that downwelling zones should be more influenced by the clogging of pore space by infiltration of fine sediment.

Channel-unit water fluxes calculated with Darcy’s Law did not “scale-up” to the reach scale estimate of hyporheic exchange (α), which were two orders of magnitude lower than the reach. Additional processes such as lateral inflow or transient storage in pools could have resulted in the observed differences, in addition to bias in the hydraulic conductivity measurements.

5.2. Areas for future research

This research contributes to a body of work examining the physical properties of the stream channel that influence solute transport and retention in small, headwater streams. A hydrometric and stream tracer approach was used to characterize the spatial distribution and the associated residence times through the hyporheic and surface-water transient storage zone at three spatial scales of interest including the reach, channel-unit and local or point scale. A multiple scale approach to examine hyporheic exchange has not been explicitly applied in previous research, and highlights the considerable spatial and temporal variability and complexity of hyporheic exchange processes within step-pool streams. This study shows that channel-unit spacing is a dominant control on the magnitude of exchange flow and the extent of the hyporheic zone. In addition, this research suggests that a scaling relationship based on the channel-unit geometry could be a useful and practical tool for characterizing and predicting exchange flow in step-pool streams. Continued research should focus on examining the longitudinal patterns in fluvial geomorphology, such as channel unit spacing, in order to characterize and apply the hyporheic exchange flow processes across a broad range of stream size and scales.

Studies attempting to link biological and geochemical processes to physical characteristics of the hyporheic zone have highlighted the importance of the hyporheic zone and in-channel “dead” zones for nutrient uptake and temporary retention of surface
water nutrients. The results from this study show that in-channel features such as pools and back eddies do contribute to transient storage in headwater streams. Observations made at the channel-unit scale demonstrate that separate residence times for hyporheic and surface-water transient storage zones can be quantified using a stream tracer approach. These observations provided insight into the residence time distribution of water in the hyporheic zone. However, spatial and temporal replication was limited in this study, and future studies should apply the approach to multiple channel-units within a single reach, over a range of flow conditions. In addition, a metric relating in-channel residence times to pool geometry would be a valuable contribution towards understanding the interplay between channel morphology and solute residence times. A continuing challenge in the area of hyporheic zone reach will be to "scale-up" small scale physical measurements to reach-scale observations of solute transport processes. Continued efforts to quantify among-channel-unit variability in small scale water fluxes would be beneficial in order to compare hydrometric estimates of exchange rates to the results of the transient storage model values.
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APPENDIX A: MODEL SIMULATIONS

Figure A.1. Model simulations using OTIS-P for May 31 for the lower reach

Figure A.2. Model simulations using OTIS-P for June 27 for the lower reach
Figure A.3. Model simulations using OTIS-P for Sept 21 for the upper reach (a) and the lower reach (b)
Figure A.4. Model simulations using OTIS-P for Sept 29 for the upper reach

Figure A.5. Model simulations using OTIS-P for Sept 30 for the lower reach
Figure A.6. Model simulations using OTIS-P for October 20 for the upper reach (a) and the lower reach (b)
Figure A.7. Model simulations using OTIS-P for September 29. Results are from two pools located in the upper reach.