Three-dimensional Subsurface Controls on Stream Intermittency in a Semi-arid Landscape:

A Case Study in Gibson Jack Creek, Pocatello, ID

by

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A thesis

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To the Graduate Faculty:

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## Dedication

To my family and friends, for their support, and putting up with my constant babbling about streams.

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Thesis Abstract – Idaho State University (2018)

To understand subsurface mechanisms controlling flow permanence, we assessed how lateral, vertical, and longitudinal shallow subsurface flow paths control the expression of surface flow during low flow periods, including intermittency at Gibson Jack Creek, in Pocatello, Idaho, USA. Water-table measurements, hydraulic gradients, and hyporheic exchange were monitored throughout hydrograph recession in WY2018. Our findings indicate: 1) shallow subsurface flow paths are dominant controls on surface flow permanence and 2) deviations from the drainage-area discharge relationship may predict intermittent locations in a stream network. Our results suggest that accurate subsurface characterization is critical to understanding the drivers of expansion and contraction cycles in intermittent streams and their likely responses to changes in climate and land use.

Keywords: Flow permanence, intermittency, drying, intermittent streams, shallow subsurface flow paths, headwaters, discharge drainage-area, hydrology, Idaho

# Chapter 1: Intermittency and Connectivity

## 1.1 Introduction

When people think about streams, they likely imagine a large flowing river. Merriam-Webster even defines a stream as ‘a body of running water.’ This definition is quite broad, and yet it is limiting in scope because headwater streams are the most prevalent stream type in length and number (Leopold, Wolman, and Miller, 1964; Nadeau and Rains, 2007) and headwaters are prone to contraction from their most upstream reaches. In these cases, the streambed is often seasonally above the water table (Horton, 1945), though drying can occur throughout the stream network. Intermittent systems are characterized by alternating wet and dry periods (Figure 1-1; Acuña, Hunter, and Ruhi, 2017). Although stream intermittency is characteristic of the semi-arid and arid regions of the western United States (Levick et al., 2008), intermittency is a global phenomenon, observed in both arid and humid climates (e.g. Nadeau and Rains, 2007; Jensen, McGuire and Prince, 2017; Zimmer and McGlynn 2017a, 2017b, 2018). Intermittent streams may constitute 30 % of the global river network (Tooth, 2000) or >50% of the global river network with the inclusion of low-order streams (Datry, Larned, and Tockner, 2014), including headwaters which are often under-represented in maps (Brooks and Colburn, 2011). Below 60 ºN, 69 % of first-order streams are estimated to be temporary (Raymond et al., 2013).

Figure 1-1. Surface flow alternates between being present or absent in intermittent streams, like Gibson Jack creek, Pocatello, ID, USA. These images were taken during the summer (left) and fall (right) 2018 during the seasonal recession discussed in detail in Chapter 2.

The number of intermittent streams is expected to increase due to changes in land use and climate (Larned et al., 2010; Acuña et al., 2014). As the world population continues to grow, increased agriculture, timber harvesting, groundwater withdrawal, urbanization, surface water diversion, and channelization (Levick et al., 2008) are expected to increase the frequency of stream intermittency. Climate models predict increased temperatures, more intense droughts, more variable precipitation, and more rain-on-snow events (Reynolds, Shafroth and Poff, 2015). These are expected to lead to changes in streamflow patterns in both space and time (NASA Earth Observatory, 2005). Peak streamflow timing has already shifted earlier and is likely to continue along this trend under future climate predictions (Stewart, Cayan, and Dettinger, 2005). Studies predict decreasing minimum flows and increasing numbers of zero-flow days will occur in the future (Jaeger et al. 2014). This means that intermittent streams will remain dry for longer periods and that some perennial streams may become intermittent. Despite these predictions, several key gaps remain in our understanding of intermittent streams and rivers. Here, I review five of these key areas, including what is currently known and what challenges remain in our understanding: 1) the value of intermittent streams, 2) the regulation of intermittent streams, 3) modeling and predicting intermittency, 4) defining connectivity in hydrologic systems and its challenges in intermittent systems, and 5) the distinction and overlap between intermittent streams and hyporheic studies.

### 1.2 Intrinsic value of intermittent streams

A decreased valuation of intermittent streams as compared to perennial streams (Armstrong et al. 2012) has affected our research and understanding of these systems. Historically, most stream studies have been conducted in perennial streams (Datry, Larned, and Tockner, 2014). However, intermittent streams are observed to provide many of the same ecosystem services that perennial streams provide. These include food (Steward et al., 2012), irrigation (Sandor et al., 2007), drainage for agricultural and municipal effluent (Larned et al., 2010), increased biodiversity (Levick et al., 2008; Larned et al., 2010; Datry, Larned, and Tockner, 2014) and high rates of nutrient cycling, both in streams (McIntyre et al., 2009) and in the hyporheic zone when surface flows are not present (Burrows et al., 2017). Humans also exploit headwater streams for beneficial uses; for example, Pennsylvania reported that over 1.5 million people received their drinking water from headwater, intermittent or ephemeral streams (Nadeau and Rains, 2007). Dry streambeds are utilized by humans and animals for a variety of purposes, including cultural events, transportation, terrestrial habitat, wildlife corridors, and storage of organic matter or even egg banks for aquatic biota (Steward et al., 2012).

Discussions of intermittent streams have generally focused on the ecological consequences associated with stream drying (Leigh et al., 2015; Costigan et al., 2016), as aquatic habitat availability decreases with decreasing flows. For example, intermittency affects community structure, species survival, resistance, and resilience (Larned et al., 2010; Jaeger et al., 2014; Leigh et al., 2015), as well as stream biogeochemistry during periods of disconnection and reconnection (Larned et al., 2010). Intermittent streams have often been neglected by both aquatic and terrestrial ecologists, perhaps because these streams appear to fall into neither category (Steward et al., 2012). Increasingly, intermittent streams are thought of as coupled aquatic-terrestrial ecosystems (Larned et al., 2010; Datry et al., 2016), or ecosystems of shifting habitat mosaics (Larned et al., 2010) which alternate between terrestrial and aquatic habitats. During terrestrial periods, dry streambeds can act as wildlife and migration corridors as well as serve as locations where vegetation can stabilize banks and provide food for animals (Levick et al., 2008). In arid climates, a dry streambed may have relatively high moisture content compared to the rest of the landscape and as such, support some of the highest biodiversity of flora and fauna on the landscape (Levick et al., 2008).

Intermittent headwater streams serve as important sources of sediment, water, nutrients, seeds, and organic matter (Gomi et al, 2002). They provide critical habitat for species seeking refuge from predators or from poor downstream water quality, and are also important for particular life history phases, such as providing spawning and rearing habitat for salmon (Nadeau and Rains, 2007). However, decreases in flow influence water quality, stream temperature (Webb, Clack and Walling, 2003), habitat availability (McKee et al., 2015), and aquatic food webs (Larned et al., 2010). Accumulation of organic matter is common in dry streambeds, but the low moisture content and microbial activity (Larned, Datry, and Robinson, 2007) render dry reaches relatively inactive biogeochemically. The onset of flow due to precipitation can prompt a rapid pulse of activity; this pattern has led to intermittent streams and rivers being described as 'punctuated biogeochemical reactors' (Larned et al., 2010).

Although intermittent streams are often perceived as deteriorated systems that are inferior to perennial ecosystems (Acuña, Hunter and Ruhi, 2017), intermittent streams can provide the same important functions as perennial systems; these include conducting water, energy, materials and organisms to downstream waters (Nadeau and Rains, 2007; Acuña and Tockner, 2010). Through vertical, lateral and longitudinal hydrologic connections, these streams contribute to the functional integrity of their downstream waters, and flow pulses in these streams may provide a considerable amount of groundwater recharge (Nadeau and Rains, 2007; Zimmer and McGlynn, 2017b).

### 1.3 Regulation of intermittent streams

Regulatory challenges associated with intermittent streams are related to a lack of understanding about these systems. Flow intermittence may be the natural flow regime (sensu Poff et al., 1997) in many streams and is not necessarily an indication of an impaired stream (Larned et al., 2010). In intermittent streams, local flora and fauna have evolved life history traits in response to the natural flow regime (Bunn and Arthington, 2002). Despite the wealth of evidence of the critical contributions of intermittent streams (Gomi et al, 2002; Nadeau and Rains, 2007; Levick et al., 2008; Larned et al., 2010; Datry, Larned, and Tockner, 2014), the lack of aesthetic appeal negatively impacts water quality concern for intermittent water bodies (Armstrong et al., 2012). Landowner surveys have indicated that those with perennial streams on their property perceived the streams to be more important and demonstrated more concern for water quality than those with intermittent streams on their land (Armstrong et al., 2012). This suggests that headwater and other intermittent streams may be perceived as already degraded systems and be neglected when it comes to water management and protection.

Other countries are currently trying to regulate intermittent streams with varying success. In the European Union's Water Framework Directive, the classification of intermittent streams as waterbodies requiring regulation is complicated by the non-uniform authority of that designation as well as different classification methods (Acuña et al., 2014). In Australia, where intermittent streams and rivers are prevalent, these watercourses are often included in management plans and receive protection. This includes 'water provisioning' in areas of Queensland where extensive no-flow periods may put even drought tolerant biota at risk (Acuña et al., 2014).

It was not until 2006 when the United States Supreme Court's Rapanos decision gave federal agencies jurisdiction over intermittent waterways, which include many headwater streams that impact the functional integrity of downstream waters (Nadeau and Rains, 2007; Acuña et al., 2014). Under this decision, agencies may now include seasonal water features (flowing or standing water for three months annually) in their water protection plans (Doyle and Bernhardt, 2011). However, their protection is complicated by inaccurate delineations of stream networks (Fritz et al. 2013) and variability in the definitions of what constitutes a 'significant nexus' to 'navigable waters' (Alexander, 2015). These inconsistencies as well as the under-representation of headwater reaches in USGS topographic maps (Brooks and Colburn, 2011) likely result in significant underestimation of the number of intermittent streams in the United States. Understanding which chemical, biological, and physical characteristics distinguish a stream from other hydrological features is essential for effective regulation of intermittent and perennial streams (Doyle and Bernhardt, 2011).

**1.4 Modelling and predicting intermittency**

Lack of data on the extent and drivers of intermittency are a challenge for modeling efforts. Meteorology, geology and land cover are predicted to be broad controls on the frequency of intermittency (Costigan et al., 2016), but the relative influence of each factor or how these factors interact is not well understood. Some studies have evaluated long-term flow records and used remote sensing techniques in order to develop models of stream intermittency (Fritz, Johnson and Walters, 2008; Snelder et al. 2013; Reynolds, Shafroth, and Poff, 2015). In these studies, physical variables (e.g. precipitation, soils, slope, valley width, canopy cover) were used to predict the flow regime of various streams, as either intermittent or perennial, but the processes governing flow permanence were not directly studied. Ward, Schmadel and Wondzell (2018) developed a model of flow permanence, which accurately predicted when the stream dried, but poorly described where it dried (Figure 1-2). Others have worked at small scales in intermittent systems, field mapping the active drainage network (e.g., Godsey and Kirchner, 2014; Shaw, 2016; Lovill, Hahm and Dietrich, 2018). Investigating flow permanence at fine scales may improve our ability to identify causes of drying. Using this information to refine the controls on stream drying may improve our ability to leverage remote sensing to predict regional-scale change on both a spatial and temporal scale.

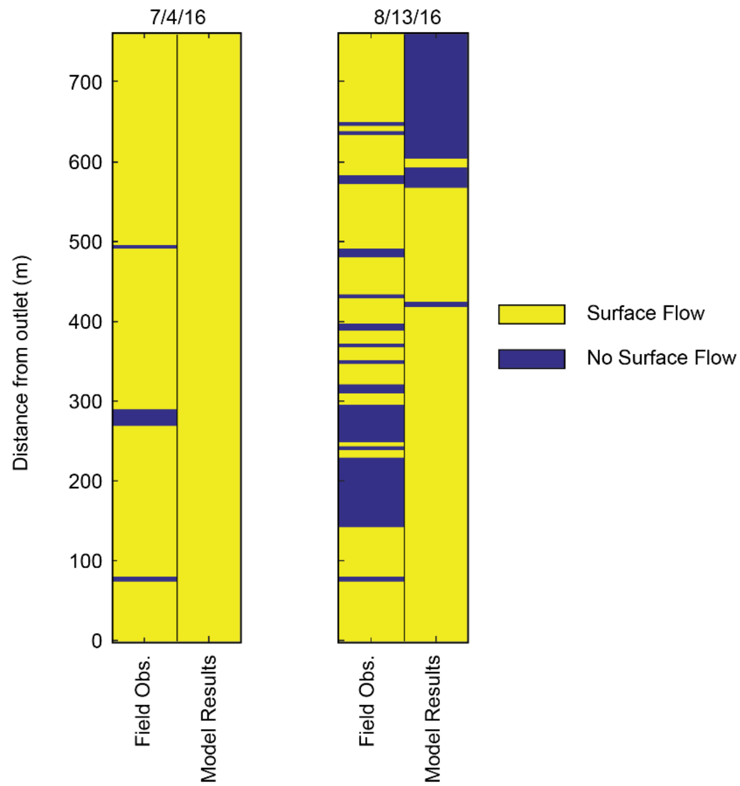


Figure 1-2. Stream drying in 2016 in the study reach from the outlet at 0m to the headwaters at 750m as observed in the field (left column of each panel), and as predicted by a reduced complexity model (right column of each panel), during two surveys on July 4th (left) and August 13th (right) (adapted from Ward et al., 2018). As flows decrease and the stream dries, the model largely predicts contraction from the headwaters whereas field observations indicate that surface flow became discontinuous along the length of the stream.

## 1.5 Defining connectivity in hydrologic settings

Connectivity often describes the function and structure of biological and physical dynamics in a watershed, but the term is used inconsistently throughout hydrologic literature (Bracken and Croke, 2007). Inconsistencies arise in part due to varying scales of consideration, where connectivity on the reach scale may be characterized differently than connectivity on the landscape scale. Connectivity may also be conceptualized as either static or dynamic in space and time. More recently, the concept of hydrologic connectivity has been separated into two categories: 'structural connectivity' and 'functional connectivity' (Turnbull, Wainwright, and Brazier, 2008). Structural connectivity describes the physical connectivity of landscape units (e.g. adjacent hillslope, riparian zone and stream). One measure of structural connectivity is the directional connectivity index, which describes the linearity of patches in one direction over multiple scales to infer transport and dispersal between these patches (Larsen et al., 2012). Functional connectivity, on the other hand, describes process-based connectivity such as the development of a shallow water table between the aforementioned landscape units. Heterogeneities in structural connectivity can influence the functional connectivity of a system (Fleckenstein et al., 2006), and similarly, functional connectivity can modify structural connectivity over time (e.g. weathering and erosion). The feedback between structural and functional connectivity can be characterized as dynamic connectivity (Turnbull, Wainwright, and Brazier 2008).

Particularly in research focusing on the expansion and contraction of streams, connectivity is often limited to the surficial observations directly along the stream network, or the active drainage network (Godsey and Kirchner, 2014). These surficial disconnections can arise from a seasonal lowering of the water table, resulting in stream contraction from their uppermost segments, or from landslides and debris dams causing transient disconnections. The locations of flow disappearance and emergence provide some insights into the structure of the subsurface (Godsey and Kirchner, 2014), but do not clearly illustrate the subsurface flow paths or lack thereof, leading to the observed active drainage network. For example, the location of flow emergence may result from an increased capacity of the subsurface to accommodate flow, or there may be a preferential flow path increasing input beyond the threshold of the subsurface; however, surficial observations cannot disentangle these two processes. Although surface flow connectivity is important for many ecological and geological processes, it does not reflect all stream connections. As succinctly stated by Bencala, Gooseff, and Kimball (2011), 'the stream is not a pipe,' and accordingly, stream connections are not limited to the surface. Subsurface flow can be a substantial component of total down valley streamflow.

As flows decrease, the actively contributing watershed area decreases over time (Bergstrom et al., 2016). Low flows are thus generally sourced from subsurface storage in the riparian zone or valley bottom (Jencso et al., 2010) or from groundwater (Blumstock et al., 2015). Antecedent moisture conditions are key components in determining runoff mechanisms and relative source contributions (James and Roulet, 2009). As the water table recedes, preferential subsurface flow paths shift as some flow paths are cutoff completely, and subsurface water must be rerouted through other active pathways. For example, during hydrograph recession two different flow paths through a channel unit were observed visually using Rhodamine Water Tracing (RWT) dye (Scordo and Moore, 2009). Earlier during the recession period, two consecutive injections followed the same flow path, with RWT travelling laterally around a boulder in the riparian zone before returning to the stream at the downstream end of the step within 5-10 minutes. Later in the recession period, RWT injected at the same location travelled vertically and returned to the stream at the start of the pool, and RWT was visible within 9 minutes, and persisted for over an hour. As discharge decreased, subsurface flow paths shifted, and residence time increased. These processes may shift discharge away from reaches that ultimately disconnect from the drainage network, while zones of consistent shallow subsurface discharge during base flows may be less likely to dry out.

**1.6 Hyporheic flow**

Early conceptual models of stream ecosystems included longitudinal and lateral water connections (e.g. River Continuum Concept and Flood Pulse Concept), but did not explicitly incorporate the subsurface vertical connection with the hyporheic zone until much later (Vannote et al., 1980; Junk, Bayley and Sparks, 1989). It is now better understood that the hyporheic zone is an important component of stream ecosystems, and encompasses a much larger area than originally thought (Boulton et al., 2010). Besides extending vertically below the stream, it also extends laterally into sediments beside the stream, particularly in floodplain rivers (Hancock, 2002).

Hyporheic flow paths occur simultaneously at multiple scales and influence the spatial distribution of gaining and losing conditions throughout the stream network. At the sediment scale, most hyporheic processes are determined by the size, shape, and composition of the sediment as well as the distribution of organic matter. These factors influence the hydraulic gradient and porosity of the hyporheic zone. Since the hyporheic zone is heterogeneous in nature and flows may even be turbulent, there are preferential flow paths as well as anoxic zones (Boulton et al., 1998). Reach-scale geomorphology influences the locations of upwelling and downwelling. These regions of flow convergence and divergence result from discontinuities in slope and depth of riffle-pool sequences, channel morphology, sediment roughness and permeability. These factors determine flow path length and thus the locations of upwelling and downwelling water throughout the stream network (Boulton et al. 1998). For example, gross hydrologic losses from the stream are influenced by local channel topography and head gradients that promote loss of surface water to valley bottom flow paths (Bergstrom et al., 2016). Two-dimensional groundwater flow models (MODFLOW) coupled with particle tracking models have been used to simulate vertical flow exchanges in mountainous catchments (Gooseff et al., 2006). Stream size, slope, and channel unit sequence control flow path lengths through the hyporheic zone. In these controlled modeling simulations, subsurface conditions were homogeneous and isotropic, which led to systematic changes in hyporheic properties, but in reality, channel unit sequences are much more complex. In contrast to these modeling results, surveyed stream reaches had multiple nested scales of upwelling and downwelling hyporheic flow paths, suggesting that heterogeneity in channel morphology influences hyporheic flow paths (Gooseff et al., 2006). Locations of upwelling and downwelling may change as discharge decreases and head gradients shift from gaining to losing conditions, influencing the spatial distribution of surface flow during low flow periods.

### 1.7 Challenges in intermittent stream research

There are a number of obstacles and opportunities for improving our understanding of intermittent streams. In this chapter, I highlighted some key areas that would benefit from focused research efforts. Specifically, our understanding of intermittent streams and rivers would improve by:

1. Detailing the processes and extent of intermittent streams that affect valuation of these unique systems;

2. Applying process-based understanding of intermittency to regulation of intermittent streams;
3. Improving spatial agreement of modeled and observed drying patterns;
4. Including both spatial and temporal information in connectivity and intermittency metrics; and
5. Applying understanding from hyporheic studies to intermittency models, and resolving associated scaling issues.

In this thesis, I focus on challenges 3, 4, and 5, and hope my work will eventually influence larger-scale studies that can better address challenges 1 and 2.

## 1.8 References


Acuña, V., Datry, T., Marshall, J., Barcelo, D., Dahm, C.N., Ginebreda, A., McGregor, G., Sabater, S., Tockner, K., and Palmer, M.A., 2014, Why should we care about temporary waterways? Science, v. 343, p. 1080–1081, doi: 10.1126/science.1246666.

Acuña, V., Hunter, M., and Ruhí, A., 2017, Managing temporary streams and rivers as unique rather than second-class ecosystems: Biological Conservation, v. 211, p. 12–19, doi: 10.1016/j.biocon.2016.12.025.

Acuña, V., and Tockner, K., 2010, The effects of alterations in temperature and flow regime on organic carbon dynamics in Mediterranean river networks: Global Change Biology, v. 16, p. 2638–2650, doi: 10.1111/j.1365-2486.2010.02170.x.

Alexander, L.C., 2015, Science at the boundaries : scientific support for the Clean Water Rule: Freshwater Science, v. 34, p. 1588–1594, doi: 10.1086/684076.

Armstrong, A., Stedman, R.C., Bishop, J.A., and Sullivan, P.J., 2012, What's a stream without water? Disproportionality in headwater regions impacting water quality: Environmental Management, v. 50, p. 849–860, doi: 10.1007/s00267-012-9928-0.

Bencala, K.E., Gooseff, M.N., and Kimball, B.A., 2011, Rethinking hyporheic flow and transient storage to advance understanding of stream-catchment connections: Water Resources Research, v. 47, p. 1–9, doi: 10.1029/2010WR010066.

Bergstrom, A., Jencso, K., and McGlynn, B., 2016, Spatiotemporal processes that contribute to hydrologic exchange between hillslopes, valley bottoms, and streams: Water Resources Research, p. 1–18, doi: 10.1002/2015WR017972.

Blumstock, M., Tetzlaff, D., Malcolm, I.A., Nuetzmann, G., and Soulsby, C., 2015, Baseflow dynamics: Multi-tracer surveys to assess variable groundwater contributions to montane streams under low flows: Journal of Hydrology, v. 527, p. 1021–1033, doi: 10.1016/j.jhydrol.2015.05.019.

Boulton, A.J., Datry, T., Kasahara, T., Mutz, M., and Stanford, J. a., 2010, Ecology and management of the hyporheic zone: stream–groundwater interactions of running waters and their floodplains: Journal of the North American Benthological Society, v. 29, p. 26–40, doi: 10.1899/08-017.1.

Boulton, A.J., Findlay, S., Marmonier, P., Stanley, E.H., Boulton, A.J., Findlay, S., Marmonier, P., Stanley, E.H., and Valett, H.M., 1998, The functional significance of the hyporheic zone in streams and rivers: Annual Review of Ecology and Systematics, v. 29, p. 59–81.

Bracken, L.J., and Croke, J., 2007, The concept of hydrological connectivity and its contribution to understanding runoff-dominated geomorphic systems: Hydrological Processes, v. 21, p. 1749–1763, doi: 10.1002/hyp.6313.

Bracken, L.J., Wainwright, J., Ali, G.A., Tetzlaff, D., Smith, M.W., Reaney, S.M., and Roy, A.G., 2013, Concepts of hydrological connectivity: research approaches, pathways and future agendas: Earth Science Reviews, v. 119, p. 17–34, doi: 10.1063/1.2756072.

Brooks, R.T., and Colburn, E.A., 2011, Extent and channel morphology of unmapped headwater stream segments of the Quabbin watershed, Massachusetts: Journal of the American Water Resources Association, v. 47, p. 158–168, doi: 10.1111/j.1752-1688.2010.00499.x.

Bunn, S.E., and Arthington, A.H., 2002, Basic principles and ecological consequences of altered flow regimes for aquatic biodiversity: Environmental Management, v. 30, p. 492–507, doi: 10.1007/s00267-002-2737-0.

Burrows, R.M., Rutlidge, H., Bond, N.R., Eberhard, S.M., Auhl, A., Andersen, M.S., Valdez, D.G., and Kennard, M.J., 2017, High rates of organic carbon processing in the hyporheic zone of intermittent streams: Scientific Reports, v. 7, p. 1–11, doi: 10.1038/s41598-017-12957-5.

Costigan, K.H., Jaeger, K.L., Goss, C.W., Fritz, K.M., and Goebel, P.C., 2016, Understanding controls on flow permanence in intermittent rivers to aid ecological research: Integrating meteorology, geology and land cover: Ecohydrology, doi: 10.1002/eco.1712.

Covino, T., McGlynn, B., and Mallard, J., 2011, Stream-groundwater exchange and hydrologic turnover at the network scale: Water Resources Research, v. 47, p. 1–11, doi: 10.1029/2011WR010942.

Datry, T., Larned, S.T., and Tockner, K., 2014, Intermittent rivers: A challenge for freshwater ecology: BioScience, v. 64, p. 229–235, doi: 10.1093/biosci/bit027.

Doyle, M.W., and Bernhardt, E.S., 2011, What is a stream? Environmental Science & Technology, v. 45, p. 354–359, doi: 10.1021/es101273f

Fleckenstein, J.H., Niswonger, R.G., and Fogg, G.E., 2006, River-aquifer interactions, geologic heterogeneity, and low-flow management: Ground Water, v. 44, p. 837–852, doi: 10.1111/j.1745-6584.2006.00190.x.

Fritz, K.M., Johnson, B.R., and Walters, D.M., 2008, Physical indicators of hydrologic permanence in forested headwater streams: Journal of North American Benthological Society, v. 27, p. 690–704, doi: 10.1899/07-117.1.

Fritz, K.M., Hagenbuch, E., D’Amico, E., Reif, M., Wigington Jr., P.J., Leibowitz, S.G., Comeleo, R.L., Ebersole, J.L., and Nadeau, T.-L., 2013, Comparing the extent and permanence of headwater streams from two field surveys to values from hydrographic databases and maps: Journal of the American Water Resources Association, v. 49, p. 867–882, doi: 10.1111/jawr.12040.

Godsey, S.E., and Kirchner, J.W., 2014, Dynamic, discontinuous stream networks: Hydrologically driven variations in active drainage density, flowing channels and stream order: Hydrological Processes, v. 28, p. 5791–5803, doi: 10.1002/hyp.10310.

Gomi, T., Sidle, R.C., and Richardson, J.S., 2002, Understanding processes and downstream linkages of headwater systems: BioScience, v. 52, p. 905.

Gooseff, M.N., Anderson, J.K., Wondzell, S.M., LaNier, J., and Haggerty, R., 2006, A modelling study of hyporheic exchange pattern and the sequence, size, and spacing of stream bedforms in mountain stream networks, Oregon, USA: Hydrological Processes, v. 20, p. 2443–2457, doi: 10.1002/hyp.6349.

Hancock, P.J., 2002, Human impacts on the stream-groundwater exchange zone: Environmental Management, v. 29, p. 763–781, doi: 10.1007/s00267-001-0064-5.

Horton, R.E., 1945, Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology: Bulletin of the Geological Society of America, v.56, p. 275-370. doi: 10.1177/030913339501900406

Jaeger, K.L., Olden, J.D., and Pelland, N.A., 2014, Climate change poised to threaten hydrologic connectivity and endemic fishes in dryland streams: Proceedings of the National Academy of Sciences of the United States of America, v. 111, p. 1–6, doi: 10.1073/pnas.1320890111.

James, A.L., and Roulet, N.T., 2009, Antecedent moisture conditions and catchment morphology as controls on spatial patterns of runoff generation in small forest catchments: Journal of Hydrology, v. 377, p. 351–366, doi: 10.1016/j.jhydrol.2009.08.039.

Jencso, K.G., McGlynn, B.L., Gooseff, M.N., Bencala, K.E., and Wondzell, S.M., 2010, Hillslope hydrologic connectivity controls riparian groundwater turnover: Implications of catchment structure for riparian buffering and stream water sources: Water Resources Research, v. 46, p. 1–18, doi: 10.1029/2009WR008818.

Jencso, K.G., McGlynn, B.L., Gooseff, M.N., Wondzell, S.M., Bencala, K.E., and Marshall, L.A., 2009, Hydrologic connectivity between landscapes and streams: Transferring reach- and plot-scale understanding to the catchment scale: Water Resources Research, v. 45, p. 1–16, doi: 10.1029/2008WR007225.

Jensen, C.K., McGuire, K.J., and Prince, P.S., 2017, Headwater stream length dynamics across four physiographic provinces of the Appalachian Highlands: Hydrological Processes, v. 31, p. 3350–3363, doi: 10.1002/hyp.11259.

Junk, W.J., Bayley, P.B., and Sparks, R.E., 1989, The Flood Pulse Concept in River-Floodplain Systems: Canadian Journal of Fisheries and Aquatic Sciences, v. 106, p. 110–127.

Larned, S.T., Datry, T., Arscott, D.B., and Tockner, K., 2010, Emerging concepts in temporary-river ecology: Freshwater Biology, v. 55, p. 717–738, doi: 10.1111/j.1365-2427.2009.02322.x.

Larned, S.T., Datry, T., and Robinson, C.T., 2007, Invertebrate and microbial responses to inundation in an ephemeral river reach in New Zealand: Effects of preceding dry periods: Aquatic Sciences, v. 69, p. 554–567, doi: 10.1007/s00027-007-0930-1.

Larsen, L.G., Choi, J., Nungesser, M.K., and Harvey, J.W., 2012, Directional connectivity in hydrology and ecology: Ecological Applications, v. 22, p. 2204–2220, doi: 10.1890/11-1948.1.

Leigh, C., Boulton, A.J., Courtwright, J.L., Fritz, K., May, C.L., Walker, R.H., and Datry, T., 2015, Ecological research and management of intermittent rivers: An historical review and future directions: Freshwater Biology, v. 61, p. 1181–1199, doi: 10.1111/fwb.12646.

Leopold, L.P., Wolman, M.G., and Miller, J.P., 1964, Fluvial processes in geomorphology: New York, Dover Publications, 522 p.

Levick, L.R., Goodrich, D.C., Hernandez, M., Fonseca, J., Semmens, D.J., Stromberg, J., Tluczek, M., Leidy, R.A., Scianni, M., Guertin, P.D., and Kepner, W.G., 2008, The ecological and hydrological significance of ephemeral and intermittent streams in the arid and semi-arid American Southwest: US Environmental Protection Agency, Office of Research and Development., p. 1–116.

Lovill, S.M., Hahm, W.J., and Dietrich, W.E., 2017, Drainage from the critical zone: lithologic controls on the persistence and spatial extent of wetted channels during the summer dry season: Water Resources Research, doi: 10.1029/2017WR021903.

McIntyre, R.E.S., Adams, M.A., Ford, D.J., and Grierson, P.F., 2009, Rewetting and litter addition influence mineralisation and microbial communities in soils from a semi-arid intermittent stream: Soil Biology and Biochemistry, v. 41, p. 92–101, doi: 10.1016/j.soilbio.2008.09.021.

McKee, C.J., Stewart, K.M., Sedinger, J.S., Bush, A.P., Darby, N.W., Hughson, D.L., and Bleich, V.C., 2015, Spatial distributions and resource selection by mule deer in an arid environment: Responses to provision of water: Journal of Arid Environments, v. 122, p. 76–84, doi: 10.1016/j.jaridenv.2015.06.008.

Nadeau, T.-L., and Rains, M.C., 2007, Hydrological connectivity between headwater streams and downstream waters: How science can inform policy: Journal of the American Water Resources Association, v. 43, p. 118–133, doi: 10.1111/j.1752-1688.2007.00010.x.

NASA Earth Observatory, 2005, The Water Cycle and Climate Change: https://earthobservatory.nasa.gov/Features/Water/page3.php (accessed October 2018).

Poff, N.L., Allan, J.D., Bain, M.B., Karr, J.R., Presetegaard, K.L., Richter, B.D., Sparks, R.E., and Stromberg, J.C., 1997, The Natural Flow Regime: A paradigm for river conservation and restoration: BioScience, v. 47, p. 769–784, doi: 10.2307/1313099.

Raymond, P.A., Hartmann, J., Lauerwald, R., Sobek, S., McDonald, C., Hoover, M., Butman, D., Striegl, R., Mayorga, E., Humborg, C., Kortelainen, P., Dürr, H., Meybeck, M., Ciais, P., and Guth, P., 2013, Global carbon dioxide emissions from inland waters: Nature, v. 503, p. 355–359, doi: 10.1038/nature12760.

Reynolds, L. V., Shafroth, P.B., and Poff, N.L., 2015, Modeled intermittency risk for small streams in the Upper Colorado River Basin under climate change: Journal of Hydrology, v. 523, p. 768–780, doi: 10.1016/j.jhydrol.2015.02.025.

Scordo, E.B., and Moore, R.D., 2009, Transient storage processes in a steep headwater stream: Hydrological Processes, v. 23, p. 2671–2685, doi: 10.1002/hyp.7345.

Shaw, S.B., 2016, Investigating the linkage between streamflow recession rates and channel network contraction in a mesoscale catchment in New York state: Hydrological Processes, v. 30, p. 479–492, doi: 10.1002/hyp.10626.

Snelder, T.H., Datry, T., Lamouroux, N., Larned, S.T., Sauquet, E., Pella, H., and Catalogne, C., 2013, Regionalization of patterns of flow intermittence from gauging station records: Hydrology and Earth System Sciences, v. 17, p. 2685–2699, doi: 10.5194/hess-17-2685-2013.

Steward, A.L., von Schiller, D., Tockner, K., Marshall, J.C., and Bunn, S.E., 2012, When the river runs dry: Human and ecological values of dry riverbeds: Frontiers in Ecology and the Environment, v. 10, p. 202–209, doi: 10.1890/110136.

Stewart, I.T., Cayan, D.R., and Dettinger, M.D., 2005, Changes toward earlier streamflow timing across western North America: Journal of Climate, v. 18, p. 1136–1155, doi: 10.1175/JCLI3321.1.

Stonedahl, S.H., Harvey, J.W., Wörman, A., Salehin, M., and Packman, A.I., 2010, A multiscale model for integrating hyporheic exchange from ripples to meanders: Water Resources Research, v. 46, p. 1–14, doi: 10.1029/2009WR008865.

Tooth, S., 2000, Process, form and change in dryland rivers: A review of recent research: Earth Science Reviews, v. 51, p. 67–107, doi: 10.1016/S0012-8252(00)00014-3.

Turnbull, L., Wainwright, J., and Brazier, R.E., 2010, A conceptual framework for understanding semi-arid land degradation: ecohydrological interactions across multiple-space and time scales: Ecohydrology, v. 1, p. 126–130, doi: 10.1002/eco.

Vannote, R.L., Minshall, W.G., Cummins, K.W., Sedell, J.R., and Cushing, C.E., 1980, The river continuum concept: Canadian Journal of Fisheries and Aquatic Sciences, v. 37, p. 130–137, doi: 10.1139/f80-017.

Wainwright, J., Turnbull, L., Ibrahim, T.G., Lexartza-Artza, I., Thornton, S.F., and Brazier, R.E., 2011, Linking environmental régimes, space and time: Interpretations of structural and functional connectivity: Geomorphology, v. 126, p. 387–404, doi: 10.1016/j.geomorph.2010.07.027.

Ward, A.S., Schmadel, N.M., and Wondzell, S.M., 2018, Simulation of dynamic expansion, contraction, and connectivity in a mountain stream network: Advances in Water Resources, v. 114, p. 64–82, doi: 10.1016/j.advwatres.2018.01.018.

Webb, B.W., Clack, P.D., and Walling, D.E., 2003, Water-air temperature relationships in a Devon river system and the role of flow: Hydrological Processes, v. 17, p. 3069–3084, doi: 10.1002/hyp.1280.

Zimmer, M.A., and McGlynn, B.L., 2017a, Ephemeral and intermittent runoff generation processes in a low relief, highly weathered catchment: Water Resources Research, v. 53, p. 7055–7077, doi: 10.1002/2016WR019742.

Zimmer, M.A., and McGlynn, B.L., 2017b, Bidirectional stream–groundwater flow in response to ephemeral and intermittent streamflow and groundwater seasonality: Hydrological Processes, v. 31, p. 3871–3880, doi: 10.1002/hyp.11301.

Zimmer, M.A., and McGlynn, B.L., 2018, Lateral, vertical, and longitudinal source area connectivity drive runoff and carbon export across watershed scales: Water Resources Research, p. 1576–1598, doi: 10.1002/2017WR021718.

## Chapter 2: Three-dimensional subsurface flow path controls on flow permanence

### 2.1 Abstract

Intermittent streams currently constitute an estimated ~30 to >50 % of the global river network and the number of intermittent streams is expected to increase due to changes in land use and climate. These streams provide important ecosystem services, such as water for irrigation, increased biodiversity, and high rates of nutrient cycling. Past hydrological studies have focused on mapping current intermittent flow regimes, which indicate streams not only expand and contract from their headwaters, but also disconnect and reconnect throughout the stream network. Other studies have evaluated long-term flow records to isolate physical metrics that have a high spatial correlation with intermittent segments. These studies describe spatiotemporal observations of stream intermittency, but few studies have investigated the underlying causes of intermittency.

Here we focus on intermittent streams where the disconnection and reconnection of surface flow may reflect the capacity of the subsurface to accommodate flow. We assess how lateral, vertical, and longitudinal subsurface flow paths control local surface flows across varying flow conditions, including intermittency. Water table dynamics were monitored across an intermittent section of Gibson Jack Creek in southeastern Idaho from April to September 2018. Four transects were delineated with groundwater wells located in the hillslope, riparian zone, and stream. Hillslope-riparian-stream connectivity in the shallow subsurface was more frequent in transects spanning perennially flowing stream reaches than intermittent reaches. During low flow periods, falling head tests suggest that losing vertical hydraulic gradients were larger in the intermittent reach than in perennial reaches. A longitudinal array of electrical conductivity (EC) loggers measured higher background EC levels and lower temperatures immediately below the

intermittent reach, consistent with an increased groundwater contribution to the stream. Measurements of longitudinal changes in losses and gains of flow from adjacent reaches also show significant gains in one perennial reach. Drainage area only weakly predicted discharge in this basin with large discrepancies in typical area-discharge relationships in the intermittent reach, consistent with heterogeneous subsurface flow paths not always reflected by surface topography. Thus, deviations from this relationship may be useful as a predictive tool for assessing reaches more or less susceptible to drying. Our findings suggest that accurate characterization of subsurface function is critical to understanding the drivers of drying cycles in intermittent streams, and their likely responses to changes in climate and land use.

## 2.2 Introduction

Intermittent streams are found across arid and humid landscapes (Nadeau and Rains, 2007) observed in first-order headwater streams to fifth-order rivers (Raymond et al., 2013). Although the frequency of stream expansion and contraction is expected to increase globally (Larned et al., 2010; Acuña et al., 2014), we currently lack a complete mechanistic understanding of the processes that control flow permanence, or whether surface flow is perennial, or intermittent (Zimmer and McGlynn, 2017a). A process-based understanding is important for directing future land-use practices to minimize negative impacts on water resources, and for predicting how climate change will impact flow patterns. Meteorology, geology, and land cover are broadly recognized as controls on the presence of surface flow (Buttle et al., 2012; Godsey and Kirchner 2014; Costigan et al., 2016; Ward, Schmadel, and Wondzell, 2018), however, the relative importance of these controls and their interactions remain unclear. They all affect water delivery to the stream network, and ultimately flow permanence, in the stream network.

Surface flow expansion and contraction, as well as disconnection and reconnection, reflect the capacity of the subsurface to accommodate flow; surface flow emerges where the subsurface can no longer accommodate flow and disappears where subsurface capacity is greater than flow (Godsey and Kirchner 2014). Thus, both water fluxes and the ability of the subsurface to transmit water control flow permanence. In porous media, Darcy's law states that discharge is proportional to the hydraulic conductivity of the subsurface. Thus, all else equal, decreases in hydraulic conductivity would limit subsurface flows, leading to more reliable surface flows. In contrast, increases in hydraulic conductivity would result in more subsurface discharge, and less reliable surface flows. Darcy's law also suggests that the hydraulic gradient controls subsurface flow if porous media flow dominates (as expected in the stream corridor).

Discussion of stream expansion and contraction has often relied heavily on the concept of the "variable source area" (VSA; Hewlett and Hibbert, 1967), focusing on the lateral connectivity with hillslopes, driven primarily by seasonal water table fluctuations. The VSA concept predicts stream network expansion and contraction in the headwaters, but it does not explain observed disconnections within the stream network without coupling it with other controls, such as subsurface topology and hydraulic conductivity (Day, 1978; Godsey and Kirchner 2014). Nonetheless, lateral connectivity is often strongly linked to the VSA or topographic contributing area and is a predominant driver of streamflow (Jencso and McGlynn, 2009; Detty and McGuire, 2010). However, lateral connectivity is not the only relevant possible control; here we review controls due to vertical and longitudinal connectivity as well.

Vertical connectivity of streams with the shallow subsurface and deeper groundwater is driven by seasonal water table fluctuations. Local gradients between surface and subsurface water direct flow toward or away from the stream (Zimmer and McGlynn, 2017b). Longitudinal connectivity along the stream corridor has often been studied in the context of hyporheic flow, where upwelling and downwelling locations result from water exchange between the surface and subsurface in the downstream direction and contribute to streamflow dynamics (Boulton et al., 1998). Thus, lateral, vertical, and longitudinal connectivity have all been observed to influence flows within the stream corridor.

The importance of three-dimensional subsurface flow has only recently been proposed as a key control on stream permanence (Ward et al., 2018; Zimmer and McGlynn, 2018). Ward et al. (2018) simulated flow permanence with a reduced-complexity model that incorporated three-dimensional flow. Their model matched observed flow recession patterns, but observed locations of drying were inaccurately modeled during low-flow periods. They attributed the mismatch to model resolution as well as their assumption of spatially homogeneous subsurface conductivity. Other modeling studies suggest that small changes in subsurface sediment heterogeneity can significantly alter surface flows (Fleckenstein et al. 2006) because of spatially variable hydraulic conductivity of streambed sediments.

Many streamflow models assume lateral inflows are proportional to upslope accumulated area (UAA) (e.g., Rodriguez-Iturbe and Rinaldo, 1997; Galster, 2007; Godsey and Kirchner, 2014; Bergstrom, Jencso and McGlynn, 2016; Ward et al., 2018). This assumption is consistent with classic studies showing increased subsurface flow accumulation with increasing contributing area (Hewlett and Hibbert, 1967; Dunne and Black, 1970; Freeze, 1972). There is a strong positive relationship between UAA and the annual duration of shallow subsurface connectivity (defined as observed saturation above the depth of bedrock in the hillslope and riparian area, as well as surface flows in the stream, Jencso and McGlynn, 2009). However, the relationship between streamflow and UAA varies in time, with a weaker relationship during base flow recession when the influence of subsurface structure on stream discharge increases (Payn et al. 2012). The streamflow-UAA may become noisy at low flows because spatially variable subsurface runoff generation mechanisms dominate at low storage stages (Zimmer and McGlynn, 2018), and the actual contributing area is much smaller than the total watershed area (Bergstrom, Jencso, and McGlynn, 2016). A discharge measurement at any given point in a stream network integrates all flow path connections with the stream at that point in space and time (Bergstrom, Jencso and McGlynn, 2016). Thus, the scale of a discharge measurement influences our interpretations of internal catchment processes. Investigations of hydrologic loss and gain in headwater streams have suggested that reach-scale discharge is closely linked to local hillslopes and runoff generation processes, but discharge becomes increasingly decoupled from lateral inflows moving from headwaters to the watershed outlet (Covino, McGlynn, and Mallard, 2011). If controls on discharge change moving downstream through the stream network, then by extension, it is likely that controls on the presence or absence of surface flow also change with scale, becoming increasingly less dependent on local processes (e.g. a given rate of discharge loss would have more impact on surface flows in a small headwater stream than in a large river). We hypothesize that these relationships likely break down further in intermittent streams, where disconnections throughout the stream network suggest that surface flow can sometimes decrease with increased UAA, and thus local-scale subsurface flow path variability in space and time may predict flow permanence.

For this study, we quantify reach-scale variability in shallow subsurface flow in three categories: 1) lateral flow from the hillslopes, 2) vertical flow exchange between the stream and streambed, and 3) longitudinal hyporheic flow along an intermittent reach that dries and leads to a surface disconnection in the stream network. We compare 3D, shallow subsurface connectivity in intermittent and perennial reaches over a seasonal recession to assess how connectivity differences shift with the presence and absence of surface flow in both space and time. Specifically, in perennial reaches we hypothesize that there will be high lateral connectivity with hillslopes, vertical gradients predominantly towards the stream, and longitudinal flow paths that produce consistently gaining conditions. These would all increase the likelihood of flow permanence in the stream. By contrast, in the intermittent reaches we hypothesize little to no lateral connectivity with adjacent hillslopes, vertical gradients away from the stream, and longitudinal flow paths that produce predominantly losing conditions (Figure 2-1). These flowpaths would all decrease the likelihood of flow permanence in the stream. We expect that changes in any flow path could cause intermittency under the sufficiently dry conditions, but that combinations and/or interactions among the categories will complicate predictions of intermittency. If 3D subsurface connectivity varies predictably between intermittent and perennial reaches, we expect that including these flow paths in modelling efforts will improve predictions of both when and where streams dry.

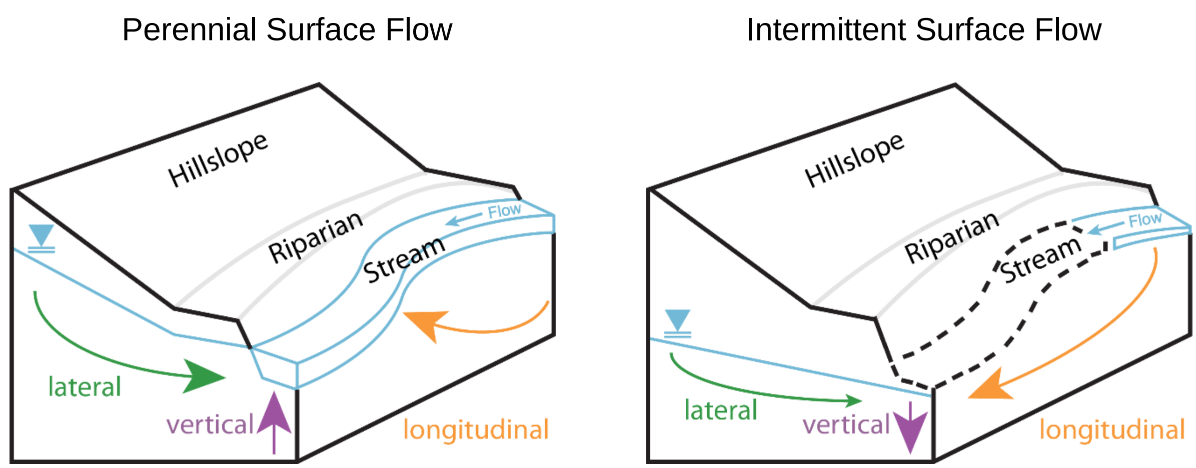


Figure 2-1. Simplified conceptual model of 3D subsurface flow paths influencing flow permanence in streams. The size of the arrowheads indicates relative influence on surface flows. Left: In perennial reaches with the most stable surface flow, we expect high lateral connectivity with hillslopes, vertical gradients predominantly towards the stream, and consistently gaining longitudinal conditions. Right: In intermittent reaches with the most unreliable surface flow, we expect little to no lateral connectivity with adjacent hillslopes, vertical gradients away from the stream, and predominantly losing longitudinal conditions. Other combinations of these flow paths are possible that may lead to intermediate flow permanence conditions. Figure adapted from Ward et al. (2013).

## 2.3 Site description

Gibson Jack Creek is a headwater stream located in the Caribou-Targhee National Forest (42°47’09.0” N, 112°26’37.2” W), near Pocatello, Idaho (Figure 2-2) with elevations ranging from ~1500 m to ~2100 m. Originating in the Bannock Range of Idaho, Gibson Jack is a tributary to the Portneuf River, which ultimately flows into the Snake River, the largest tributary of the Columbia River in the Pacific Northwest region of the USA. The climate is classified as semiarid steppe, with an average annual precipitation of 635 mm (Kline, 1978) and a trend of increasing precipitation with elevation ranging from 380 mm/yr at lower elevations to 760 mm/yr (Welhan, 2006) at the highest elevations of the watershed. Temperatures range from sub-zero in winter to +30 ºC in the summer (National Weather Service, accessed October 2018). The vegetation is mostly sagebrush, grass, juniper and other dry open-slope species, with the exception of the north-facing slopes, which are dominated by coniferous trees (Kline, 1978), mostly Douglas fir. The soil textures are primarily silt loam and fine sandy clay loam, and also contain gravel (Davidson, 1977). Four National Resources Conservation Service (NRCS) soil types converge along the lower mainstem of Gibson Jack, including the Lanoak-Hades complex, Lanoak family, Farlow family, and the Sparky-Jedediah family (NRCS Web Soil Survey, accessed October 2018).

The Gibson Jack watershed is characterized by steep (~20º) hillslopes of high (~500 m) relief (Kline, 1978). Bedrock in the northern portion predominantly consists of quartzite and shale, while the southern portion mostly consists of limestone (Idaho Geological Survey, accessed October 2018). The bedrock is moderately fractured where quartzite and limestone occur (Kline, 1978). The Pocatello region has many bedrock faults and although most faults are north-south trending, one fault in the watershed has its axis along the lower reaches of Gibson

Jack Creek. There is limited grazing in the headwaters of the Gibson Jack watershed and several trails, small bridges and culverts accommodate recreation from off-highway vehicles, hikers, and mountain bikers.

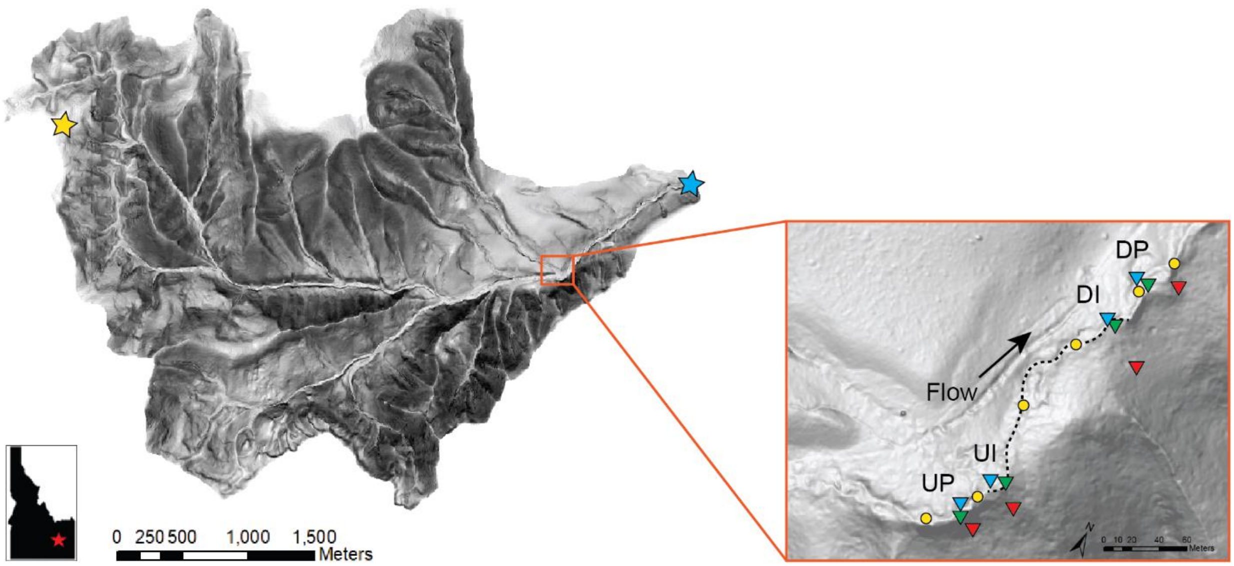


Figure 2-2. A 3-m DEM of the Gibson Jack watershed located in southeastern Idaho. Precipitation data were collected at a weather station (marked with a yellow star) located near the upper boundary of the watershed and streamflow was measured at the watershed outlet (marked with a blue star). The orange box highlights the ~200-m study site. Instrumentation in the study site is depicted in the inset. Four transects of shallow groundwater wells are represented by triangles with wells located in the hillslope (red), riparian zone (green), and stream (blue). Transects are designated DP (downstream perennial), DI (downstream intermittent), UI (upstream intermittent) and UP (upstream perennial). Circles (yellow) represent electrical conductivity (EC) loggers. The furthest downstream EC logger is located at 1630 m above the gaged outlet to Gibson Jack, and the furthest upstream EC logger is located at 1920 m above the outlet. The dashed line along the stream in the inset represents the intermittent section.

The focus of this study is a ~200 m perennial-to-intermittent-to-perennial site along the mainstem of Gibson Jack Creek, which was chosen not only because it occasionally dries, but also for accessibility and tractability of a mechanistic study. This section was observed to dry in November 2016; that month's average discharge in the Portneuf River (to which Gibson Jack Creek is a tributary) was approximately 185 $ft^3/s$, while the average November discharge (WY1966-2018) is typically 250 $ft^3/s$ (USGS National Water Information System, site 13075500). The previous winter's peak snow water equivalent (SWE) at the nearest SNOTEL site (NRCS Wildhorse Divide, site 867) was about 140 cm, ~94 % of the long-term average (WY1982-2018) peak SWE. Discharge was above average in WY2017 consistent with peak SWE that was ~120 % of the long-term average (WY1982-2018), and no drying was observed in the study site. In WY 2018, peak SWE was only ~56 % of the long-term average and drying was again observed in the study site. Data described in the next section were typically collected from May 2017 through September 2018; we focus here primarily on records from April to September 2018, during the seasonal recession, in order to capture subsurface patterns prior to the onset of drying and during the dry period.

## 2.4 Methods

### *2.4.1 Field data collection*

Meteorological data were collected at 10-min intervals within the watershed at a ~3-m weather station (Figure 2-2) located at 2150 m elevation with a 0.01" tipping bucket precipitation sensor (Texas Electronics Rain Gauge, 8" orifice). From November to May each year, the rainfall sensor was equipped with a snow adapter (CS705) filled with an ethanol/antifreeze combination and topped with a thin layer of mineral oil to minimize evaporation. In addition, snow depths were verified with an acoustic depth sensor (SR50a) to validate the precipitation/snow accumulation data. All data were locally stored on a Campbell Scientific CR-1000 datalogger and manually downloaded at least twice a year during seasonal station maintenance.

A handheld acoustic doppler velocimeter (FlowTracker, SonTek) was used to manually measure discharge over various flow conditions from May 2017 through June 2018 at the outlet. Stream stage was continuously measured at 15-minute increments with an in-stream pressure transducer (ONSET Hobologger U20-001-04) corrected for local changes in barometric pressure measured at the same frequency with an identical unit installed in the atmosphere near the in-stream logger. A stage-discharge relationship was developed ($R^2$=0.986) to estimate continuous streamflow from the stage record.

Four groundwater well transects were installed along the study site perpendicular to the stream (Figure 2-2). Transects were positioned above and below the points at which surface flow disappeared or reappeared longitudinally within the stream channel in November 2016. Specifically, the downstream perennial (DP) transect was located below flow reappearance, the downstream intermittent (DI) and upstream intermittent (UI) transects were in the intermittent reach, and the upstream perennial (UP) transect was positioned above flow disappearance. Along each transect, a groundwater well was installed in the north-facing hillslope, in the riparian zone below the hillslope, and in the streambed itself (following Jencso and McGlynn, 2009), for a total of 12 groundwater wells. The riparian zone was distinguished from the hillslope by a transition from steep slopes dominated by conifers to relatively shallow slopes dominated by willow or sedge. Based on their location, wells were designated as hillslope (H), riparian (R) or stream (S) within each transect.

Hillslope and riparian wells were drilled manually to the depth of refusal, which ranged from 0.60-2.59 m (2" AMS Sediment Sampler). PVC well screens (Schedule 40, 1" diameter, 0.01" slot width, fully screened) were inserted into these drilled wells and cut to extend ~20 cm above the ground surface. A filter pack of fine-grained silica sand was used to backfill the area between the PVC well screen and the surrounding soil from the bottom of the well up to ~15 cm below the ground surface. Filter pack sediment size was selected to be slightly larger than the PVC slot width to decrease entry of fine particles into the wells. A bentonite seal was used to cap the wells, preventing direct seepage of water into the wells from surface runoff (Figure 2-3). Fully screened drive-point piezometers (Water Source stainless steel well point, 1.25" diameter, 80 mesh screen, screen length 36") were installed to 29-43.5 cm below the bed surface in the stream locations to ensure measurement of stream conditions. These were not back-filled with a filter pack or bentonite seal. We did not actively develop these wells, instead allowing natural streamflow to develop them. Pressure transducers (ONSET Hobologger U20-001-01 or U20-001-04) were suspended in each well from a coated wire rope attached to the cap of the well to ensure consistent measurements, recording pressure at 15-minute intervals from March 2018 through September 2018. These data were corrected with local barometric pressure measurements taken with an identical unit placed above well DP-H to determine shallow groundwater level.

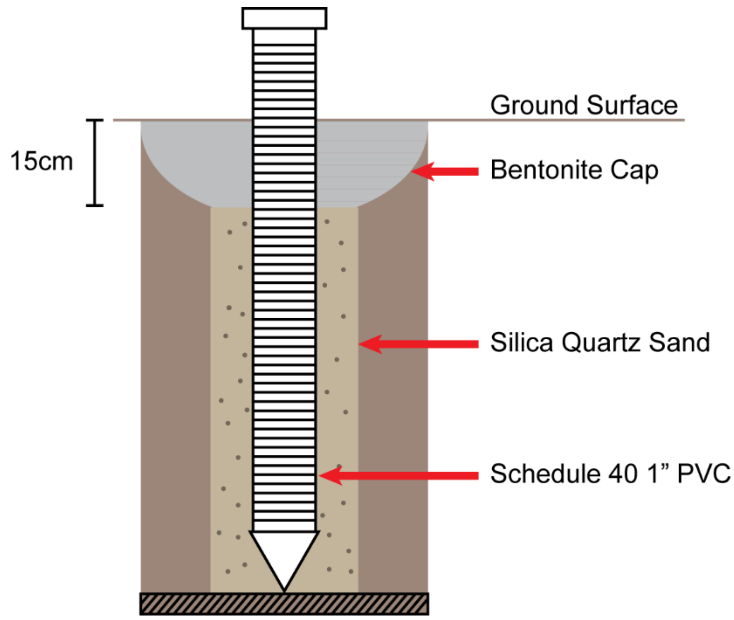


Figure 2-3. Schematic of well installation. Hillslope and riparian wells were drilled to depth of refusal, ranging from 0.42m-2.59m. Fully screened schedule 40 1" PVC was inserted into the drilled holes, and wells were backfilled with silica quartz sand from the bottom of the well up to ~15cm below the ground surface. A bentonite cap was installed on top of the sand extending to the ground surface.

We measured relative well elevations with an optical auto level (Topcon AT-G4) and determined horizontal distance between wells using a GPS and receiver (Trimble R6, GNSS System). Within each transect, hydraulic gradient was calculated between 1) the hillslope and riparian wells, and 2) the riparian and in-stream wells whenever the water level in each pair exceeded zero.

Falling head tests were conducted in the hillslope and riparian wells to estimate hydraulic conductivity. Pressure transducers in the wells (ONSET Hobologger U20-001-01) were adjusted to measure at one-second intervals during the tests; tests began when water was added to a well, and the tests concluded when the water level returned to background levels (Fetter, 2013). Long-term water-level records during these falling head tests were manually corrected afterwards. Falling head tests were also used to estimate hydraulic conductivity of the streambed; however, because water was present in the stream during these tests, we could not accurately measure water level recession in the installed, fully screened in-stream piezometers. Instead, partially screened minipiezometers (4.3-cm radius, 24-cm screened length) were temporarily installed ~30 cm below the streambed adjacent to the permanent in-stream wells and were developed using a pump (Geotech Geopump Peristaltic Pump). After equilibration, falling head tests were conducted in triplicate as previously described (except using an In-Situ Level TROLL 500 Data Logger because its narrower diameter fit the minipiezometers).

The in-stream vertical hydraulic gradient (VHG) was measured during low flow periods to quantify directional exchange between the stream and groundwater. Once the minipiezometers had equilibrated, we measured the water level in the subsurface of the stream using the partially screened minipiezometers and compared this level to the elevation of the stream surface adjacent to the minipiezometer (following Baxter, Hauer and Woessner, 2003).

Freshwater electrical conductivity (EC) dataloggers (ONSET Hobologger, U-24) were installed in the thalweg along the study site (Figure 2-2). Six EC dataloggers were installed in April 2018, spaced such that two dataloggers were installed below flow reappearance, two were installed within the intermittent reach, and two dataloggers were installed above flow disappearance (Figure 2-2). EC dataloggers were named based on their distance in meters from the watershed outlet, which was defined by the location of the outlet stage measurement. The EC dataloggers served three purposes: 1) monitoring relative EC and temperature at fifteen-minute intervals, 2) measuring EC during salt dilution gauging for determination of discharge and hydrologic exchange, and 3) monitoring presence or absence of flow in the stream.

Seven dilution gauging experiments were conducted following methods from Payn et al. (2009) over flows decreasing from 0.296 to 0.054 $m^3/s$ at the watershed outlet. We chose dilution gauging because it is typically more accurate than velocity gauging in irregular, high-gradient mountain headwater streams and because flows were often too shallow in this segment for reliable ADV-based velocity measurements (Day, 1978). Each experiment included three injections of a known, pre-dissolved mass of salt (NaCl). Injections occurred at the same three locations for each experiment and proceeded from downstream to upstream. Injection one occurred at ~1690 m, injection two at ~1820 m, and injection three at ~1950 m. We conducted one injection per day over three days for each experiment to ensure that all tracer was removed from the reach. Full mixing of the tracer prior to the first measurement location was validated periodically with real-time *in situ* EC measurements (YSI Pro 30 with T and EC probe). EC loggers downstream of the designated injection point recorded EC at two-second intervals in order to capture the tracer concentration breakthrough curves. EC loggers were removed from the stream in September 2018; Hobo Pendant dataloggers (UA-002-64) modified to measure EC

(Chapin et al., 2014) were installed in the intermittent section to monitor presence and absence of flow at a finer spatial resolution.

*2.4.2 Data analysis*

The Gibson Jack watershed was delineated in ArcGIS 10 from 1-m LiDAR (vertical RMSE = 0.036 m; InPort, accessed October 2018) using the Spatial Analyst toolbox. We delineated the contributing topographic area for each discharge measurement location, which included the weir and all EC dataloggers, using the Snap Pour Point tool in ArcGIS 10. Although this watershed delineation describes surficial runoff patterns, it may not accurately describe subsurface runoff patterns (Payn et al., 2012). Log-transformed discharges at each EC logger and the outlet were plotted against drainage area and fit with a power-law relationship to assess the UAA-flow relationship.

*2.4.2.1 Lateral connectivity*

Lateral connectivity was defined as periods where water level in wells showed shallow subsurface connection between the hillslopes, riparian zones, and the stream, hereafter referred to as 'HRS connectivity' (after Jencso et al., 2009). Because the intermittent reach dried seasonally, we specifically define HRS connectivity as periods when water was simultaneously measured in both the hillslope and riparian wells, and a water level above the streambed was observed in the in-stream wells. To be consistent with this definition, hillslope and riparian well data are reported as water height above the base of the well, whereas in-stream well data are reported as water above the streambed surface to clearly demonstrate periods when the streambed is dry.

Data quality control for all water level loggers revealed that approximately 0.04% of the data were missing or recorded unrealistic jumps or drops. These errors often occurred after the logger was removed for downloading in which case the gaps were filled by linear interpolation. Rarely, these jumps or drops occurred without any known human disturbance; because nearly all of these shifts occurred in the in-stream wells, we attributed the jumps or drops to periods of debris buildup or removal. If the shifts were <1 hour in duration, we simply removed the points and linearly interpolated to fill the gaps. In the few remaining cases, re-installation of the logger resulted in an immediate and persistent shift to shallower water depths, which we believe resulted from the instrument or wire rope being caught on the well screen, so that the logger registered artificially shallow water depths. For these instances, data were linearly shifted to reflect the water level of the previously undisturbed well. None of these adjustments altered our interpretation of water presence or absence in the wells, which was our primary interest. We also note that specified manufacturer instrument errors were ±0.3 % of the range, translating to ~3.5-4 cm of uncertainty in the barometrically corrected water levels. Excluding data below these error thresholds would have removed most observed periods of HRS connectivity from our record. Field observations suggest water was present in the wells during these periods. Instead of using manufacturer specifications, we calibrated the loggers in our laboratory up to a depth of 20 cm and found an average error of ~0.5 cm. Thus, we used 0.5 cm as the minimum threshold for water presence in the wells. We performed traverse surveying in a loop to determine well elevations used for hydraulic gradient calculations and used the difference in elevation between the initial and final elevation measurement of the same point as our elevation uncertainty. The horizontal uncertainty threshold was set at 10 cm for distance measurements between the wells.

*2.4.2.2 Vertical connectivity*

We determined two key parameters relevant to vertical connectivity in the study reach. First, we calculated the hydraulic conductivity of the surrounding soils and sediments around each well and minipiezometers using falling head test data and the Hvorslev method. This method is appropriate for wells that do not fully penetrate the aquifer and assumes that the length of the piezometer is more than 8 times the radius of the screened interval (Fetter, 2013). We solved for hydraulic conductivity, *K* (cm/s), using Equation 1:

$$K = \frac{r^2 \ln(L_e/R)}{2L_e t_{37}} \qquad Eq. (1)$$

where *r* is the radius of the well casing (cm), *R* is the radius of the perforated interval (cm), $L_e$ is the length of the perforated interval (cm), and $t_{37}$ is the time (s) required for the water level to reach 37% of the initial change. To determine $t_{37}$, we calculated $h/h_0$ (cm/cm), where *h* (cm) is the change in water level and $h_0$ (cm) is the static water level. These data were plotted on a semi-logarithmic graph of log($h/h_o$) vs time and fitted with an exponential function. The uncertainty in hydraulic conductivity depends on the uncertainty in piezometer dimensions and the calculated value for $t_{37}$. We assumed an uncertainty of 0.5cm for all length measurements. We followed methods by Rushlow and Godsey (2017) to determine the slope and intercept of the best-fit line of the semi-log plot and their uncertainties, and to propagate those uncertainties in the length through to the estimate of $t_{37}$ (as detailed in Appendix A).

For the in-stream wells, VHG (cm/cm) was also calculated using Equation 2,

$$VHG = \frac{\Delta h}{\Delta l} \qquad Eq. (2)$$

where $\Delta h$ (cm) is the difference in head between the water level in the piezometer and the water level in the stream and $\Delta l$ (cm) is the distance from the streambed surface to the top of the perforated interval (Baxter, Hauer and Woessner, 2003). Positive VHG values indicate upwelling conditions, while negative VHG values indicate downwelling conditions.

*2.4.2.3 Longitudinal connectivity*

Following the methods of Payn et al. (2009), we calculated hydrologic gain and loss throughout the study reach. Because the discharge and hydrologic gain/loss calculations rely on paired upstream and downstream tracer injections, paired consecutive salt injections were deemed the 'downstream' and 'upstream' injections based on their relative locations for a given calculation. For example, the first and second injections, as well as second and third injections, were paired as 'downstream' and 'upstream' injections, respectively; the second injection served as the 'upstream' injection in the 1st pair and the 'downstream' injection in the 2nd pair. Furthermore, to simplify calculations for $Q_{LOSS}$ and $Q_{GAIN}$ in each segment, we paired spatially sequential loggers as 'upstream' and 'downstream' measurement points. Therefore, during subsequent salt injections, most measurement points act as both upstream and downstream measurement points, with the exclusion of the lowermost and uppermost measurement points (Figure 2-4). For example, the tracer was first injected above EC 1660, while the second injection occurred above EC 1790. Tracer concentration breakthrough curves were measured at EC 1660 during both the first and second injections. During the first injection, EC 1660 could act as a downstream logger to the subsequent upstream logger (EC 1730), and during the second injection, EC 1660 could act as an upstream logger to the subsequent downstream logger (EC 1630). EC 1660 was not used as an upstream logger to EC 1630 during the first injection because we did not have an independent measurement of discharge at the downstream measurement point. Similarly, flow gains and losses were not determined for the reach between EC 1920 and EC 1880 because we had no independent estimate of discharge at EC 1880. Although it would be theoretically possible to add an additional injection point further upstream or downstream, instrumentation and logistics constraints limited the scope of the experiment because flows needed to remain as constant as possible for all injections and breakthrough curve passage.

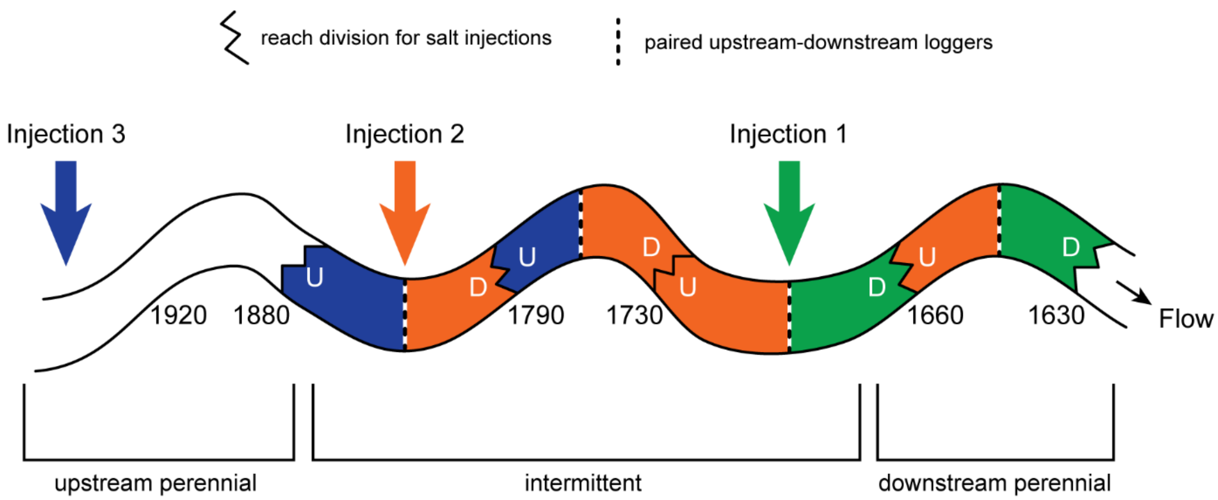


Figure 2-4. Pairing of EC loggers with upstream and downstream injections. Labels for each EC logger are below the stream, with the number indicating meters upstream from the outlet. Each injection is color-coded, with green for injection one, orange for injection 2, and blue for injection 3. Within each reach between sequential loggers, there is a 'U' designating upstream, and 'D' designating downstream. The color of the stream segment indicates which injection is being used for estimates of discharge for each upstream and downstream logger (e.g. EC 1660 uses injection 2 for the upstream discharge measurement for reach 1660-1630m, but uses injection 1 for the downstream discharge measurement for reach 1730-1660m).

From the ‘downstream’ injection in any given pair of injections, we calculated downstream discharge ($Q_D$) using Equation 3,

$$Q_D = \frac{M_D}{\int_0^t C_D(t)dt} \qquad Eq. (3)$$

where $M_D$ is the mass of injected salt at the downstream injection point, and $\int_0^t C_D(t)dt$ is the integrated tracer concentration breakthrough curve at the downstream measurement location. Rather than measuring the breakthrough curve at just one measurement point as in Payn et al. (2009), we measured the breakthrough curve at multiple measurement locations for each injection. This allowed us to minimize the number of injections to attempt to maintain constant discharge while estimating flow and hydrologic exchange at as many points as possible. Concentration breakthrough curves were truncated when values returned to background. When concentrations did not return to background, we truncated breakthrough curves after 3-e folding times. This was only necessary during the final experiment in late July, when flows were the lowest. This suggests that the duration of measurement may not have been long enough to capture all flow paths during low flows or that there were larger mass losses to longer flow paths than measurable with our sampling design. $Q_D$ was estimated at all downstream measurement points. Based on the first and second injections, $Q_D$ was estimated at EC 1630 and EC 1660. Based on the second and third injections, $Q_D$ was estimated at EC 1790 and EC 1730 (Figure 2-4).

From the ‘upstream’ injection in any given pair of injections, we calculated upstream discharge, $Q_U$, based on Equation 4,

$$Q_U = \frac{M_U}{\int_0^t C_U(t)dt} \qquad Eq. (4)$$

where $M_U$ is the mass of injected salt at the upstream injection point, and $\int_0^t C_U(t)dt$ is the integrated tracer concentration breakthrough curve at the upstream measurement location. Based on the first and second injections, $Q_U$ was estimated at EC 1730 and EC 1660. Based on the second and third injections, $Q_U$ was estimated at EC 1920, EC 1880 and EC 1790 (Figure 2-4). Although EC 1920 was not used for estimations of hydrologic exchange, the value of $Q_U$ was used to represent discharge at this location.

The change in discharge, $\Delta Q$, between each sequential measurement location was calculated according to Equation 5:

$$\Delta Q = Q_D - Q_U \qquad Eq. (5)$$

Mass recovery was also calculated at each 'downstream' measurement point based on

Equation 6:

$$M_{REC=}Q_D \int_0^t C_{UD}(t)dt \qquad Eq. (6)$$

where $M_{REC}$ is the mass recovered at the 'downstream' measurement point, and $\int_0^t C_{UD}(t)dt$ is the sum of the tracer concentration breakthrough curve at the 'downstream' measurement point resulting from the 'upstream' injection ($M_U$). For this calculation, it is imperative to have an independent estimate of $Q_D$, which we acquired during the downstream salt injection.

The tracer mass lost in transport over the reach, $M_{LOSS}$ (referred to simply as mass loss hereafter) was estimated between sequential measurement points, using Equation 7,

$$M_{LOSS} = M_{REC} - M_U \qquad Eq.(7)$$

$M_{LOSS}$ is negative when mass has been lost from the reach (more mass injected than recovered). If a positive mass loss was calculated in a net gaining segment, we assumed $Q_{LOSS}$ to be zero (following Ward et al., 2013).

The spatial pattern of dilution can significantly change the salt concentration in the stream at each measurement point. If dilution ($Q_{GAIN}$) occurs prior to $Q_{LOSS}$ within a reach, the mass of salt leaving the stream will be less than if dilution occurs after $Q_{LOSS}$. The calculation of $Q_{LOSS}$ depends on the estimated mass loss, and thus $Q_{LOSS}$ is strongly influenced by the location of any $Q_{GAIN}$. For this reason, minimum and maximum hydrologic gains and losses are determined. $Q_{LOSS,MIN}$ assumes minimum dilution before losses, whereas $Q_{LOSS,\ MAX}$ assumes maximum dilution before loss. We estimate the maximum and minimum discharge loss with Equation 8:

$$Q_{LOSS,MIN} = \frac{M_{LOSS}}{\int_0^t C_U(t)dt}, Q_{LOSS,MAX} = \frac{M_{LOSS}}{\int_0^t C_{UD}(t)dt} \qquad Eq.(8)$$

where $Q_{LOSS,MIN}$ and $Q_{LOSS,\ MAX}$ are the minimum and maximum discharge losses over the reach, $\int_0^t C_U(t)dt$ is the sum of the tracer concentration breakthrough curve at the 'upstream' measurement point from the 'upstream' injection, and $\int_0^t C_{UD}(t)dt$ is the sum of the tracer concentration breakthrough curve at the 'downstream' measurement point resulting from the 'upstream' injection ($M_U$).

Finally, the minimum and maximum discharge loss over the reach, $Q_{GAIN,MIN}$ and $Q_{GAIN,MAX}$, respectively, were calculated according to Equation 9:

$$Q_{GAIN,MIN} = \Delta Q - Q_{LOSS,MIN}, Q_{GAIN,MAX} = \Delta Q - Q_{LOSS,MAX} \qquad Eq.(9)$$

## 2.5 Results

*2.5.1 Lateral flow paths*

*2.5.1.1 Water levels and hillslope-riparian-stream connectivity*

We hypothesized that there would be increased lateral connectivity in the perennial stream reaches compared to the intermittent reach. Indeed, HRS connectivity was more common in the perennial transects (DP and UP) than in the intermittent transects (DI and UI) during the study period (Figure 2-5). Hillslopes in the UP transect were connected through the riparian zone to the stream for 37.5 % of the study period, the most of all transects. The DP and UI transects had moderate HRS connectivity of 11.9 % and 9.1 % of the study period, respectively. Hillslopes in the DI transect were regularly disconnected from the stream, with HRS connectivity only 0.3% of the study period. The timing of HRS connectivity varied among transects, and connections were often not immediate responses to rainfall. For example, in the UI and UP transects, 81.3 % and 57.0 % of HRS connections occurred in April and May when rainfall was relatively high. By contrast, connections were more common along the DP transect in July, and at the DI transect, HRS connectivity was only observed in July when rainfall was low.

Diel water-level fluctuations were observed in all in-stream and riparian wells, and also in hillslope wells, whenever water was present for at least 24 hours (Figure 2-6). The amplitude of variation was ~0.6-0.8 cm at all sites from March to the end of June. In early July, the amplitude began increasing in the in-stream wells in the UI and DI intermittent transects, but remained stable in the riparian and hillslope wells. The amplitude of diel water-level fluctuations remained stable in all wells in the UP and DP transects throughout the entire study period. Daily minimum water levels were correlated with the peak daily temperature. Previous work in the Gibson Jack watershed shows peak evapotraspiration rates occur at roughly the time during the day that we observed minimum water levels (Whiting, 2015). During high flows, water levels in the in-stream wells along the DP, DI and UI transects followed similar patterns to the outlet discharge, while the water level in the UP in-stream well remained elevated longer than in the remaining in-stream wells, and then dropped more rapidly in early May (Figure 2-5).

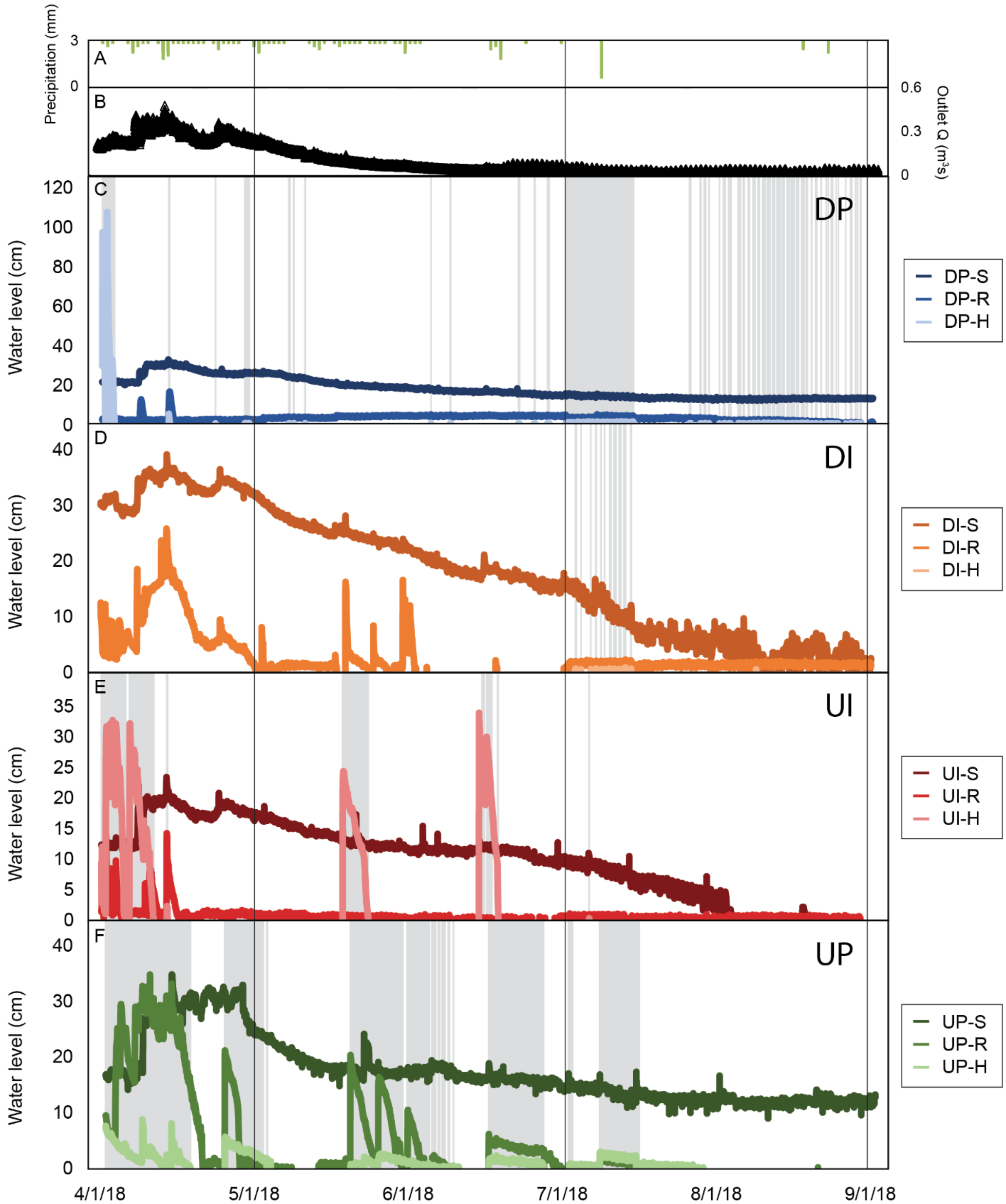


Figure 2-5. (A) Hyetograph and (B) hydrograph at the weather station and outlet, respectively (see locations in Figure 2-2). Water levels in the hillslope and riparian wells above the base of each well, and water level above the surface in the stream in each transect (DP, DI, UI, UP are plotted in panels C, D, E and F, respectively). Gray shading in panels C-F indicates periods of HRS connectivity for each transect, defined as periods when water is present in the hillslope and riparian wells, and there is surface flow in the stream.

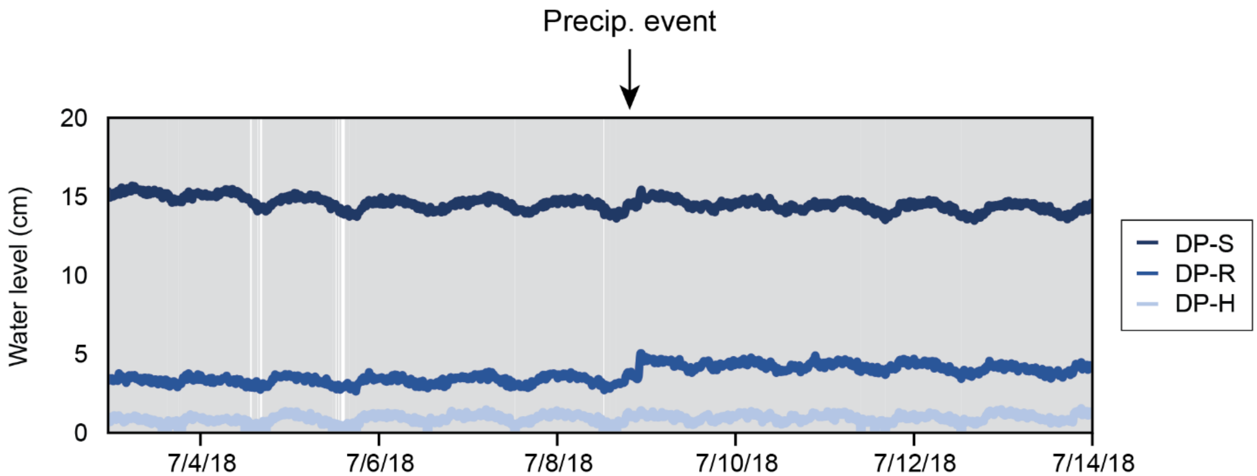


Figure 2-6. Diel-water level fluctuations observed in hillslope, riparian, and in-stream wells over a 10-day period in July 2018. Gray shading indicates HRS connectivity as in Figure 2-5, with only a few short periods during this 10-day period when the hillslope well completely dries, interrupting HRS connectivity. The lowest water levels occur at ~3-7PM each day during this period, consistent with peak air temperatures.

*2.5.1.2 Hydraulic gradients, conductivity, and inferred transit times*

The hydraulic gradients between the hillslope wells and riparian wells in each transect were relatively stable through time and flow conditions, controlled primarily by the large elevation difference between the hillslope and riparian wells (Figure 2-7a). These gradients were measured over a horizontal distance of ~20 m. The hydraulic gradients in the DP and DI transects were 0.28 m/m and 0.32 m/m, respectively, towards the riparian zone, while in the UI and UP transects, the hydraulic gradients were approximately 0.76 m/m and 0.89 m/m, respectively. The higher hydraulic gradients in the UI and UP transects primarily reflect steeper topographic slopes in these transects.

The hydraulic gradients between the riparian wells and in-stream wells were more dynamic than between the hillslope and riparian wells, with the exception of the DP transect. These gradients were measured over horizontal distances of 3-10 m. Like the hillslope-riparian well pair for the DP transect, its riparian-stream gradient was relatively stable throughout the season with a mean of 0.066 m/m towards the stream (Figure 2-7, blue symbols). The DI transect also had an average hydraulic gradient of 0.066 m/m, but the gradient toward the stream increased from high to low flows, likely due to decreasing water level in the stream (Figure 2-7b, orange symbols). The UP transect (green symbols) had the highest hydraulic gradient towards the stream, fluctuating between 0.26 m/m and 0.38 m/m. In contrast to all other sites, the riparian-stream hydraulic gradient at the UI transect (red symbols) was negative throughout all flow conditions, indicating a hydraulic gradient away from the stream and toward the riparian zone.

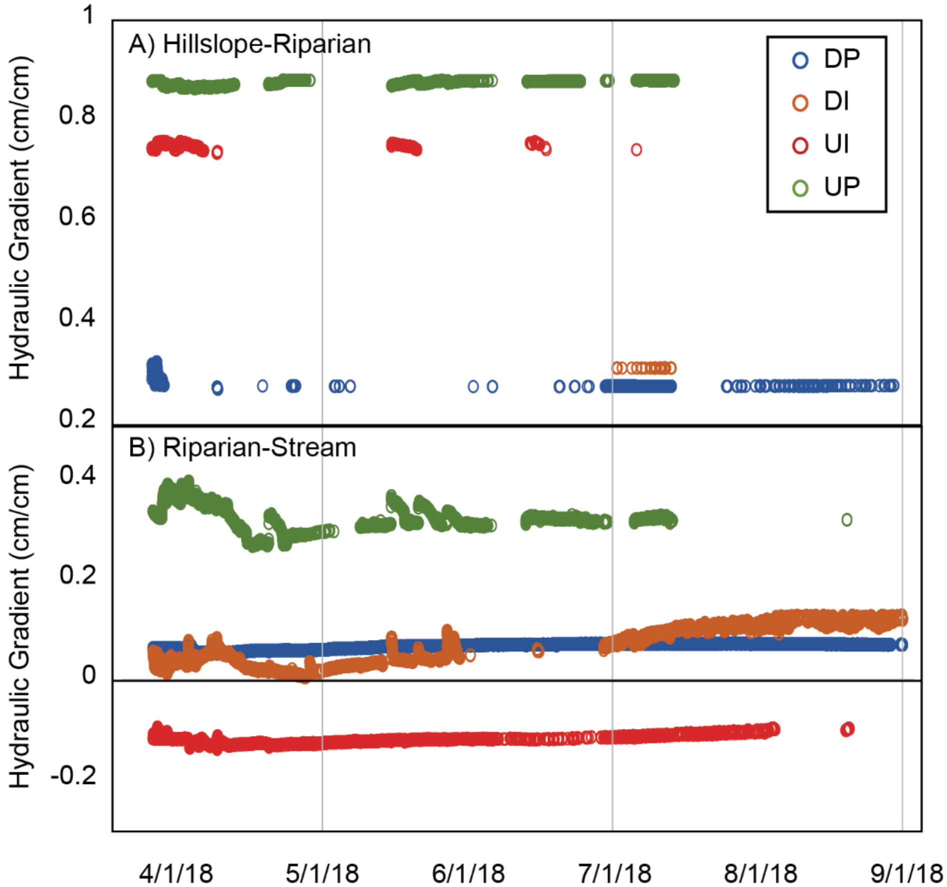


Figure 2-7. Hydraulic gradient between A) the hillslope and riparian zone and B) the riparian zone and the stream at each transect. Because the hillslope wells were often dry, there are many gaps in the hillslope-riparian hydraulic gradient time series, but when there is water in each well, the gradient remains relatively stable (A). In contrast, the riparian-stream hydraulic gradient (B) varied more. The UI transect had a consistently negative hydraulic gradient between the stream and riparian zone throughout the study period, indicating flow away from the stream and toward the riparian zone.

Saturated hydraulic conductivity in the hillslope and riparian wells for the DP, DI, and UI transects was consistent with typical values for silt, sandy silt, or diamicton (Fetter, 2013). The UP transect had a much higher hydraulic conductivity correlative with those of well-sorted gravel, well-sorted sands, or glacial outwash (Table 2-1). These results agree with field observations while installing the wells, with the UP transect being significantly more gravel-rich than the other transects. We calculated the travel time for a parcel of water to travel through each transect from the hillslope to the stream by dividing the distance between wells by the measured hydraulic conductivity. The fastest mean travel time was calculated for the UP transect, with hillslope water reaching the stream after 5.0 days with over half of that time in the riparian zone. Travel times in transects DP, DI, and UI were two orders of magnitude higher, with 6-12 months between the hillslope and riparian zones, and just a few days or weeks in the riparian zone (Table 2-2).

Table 2-1. Saturated hydraulic conductivity and its standard error (s.e.) in hillslope and riparian wells. Riparian $K_{sat}$ was higher and varied less than the hillslope $K_{sat}$, excluding well UP-H. $K_{sat}$ in well UP-H was two orders of magnitude higher than other hillslope wells, and an order of magnitude greater than the riparian wells. Well locations are described in Figure 2-2.

| Site | $K_{sat}$ (mm/hr) | s.e. (mm/hr) | description |
|---|---|---|---|
| DP-R | 16.83 | 2.85 | silt, sandy silts, clayey sands, till |
| DI-R | 40.22 | 55.76 | silty sands, find sands - well sorted sands, glacial outwash |
| UI-R | 11.08 | 26.60 | silt, sandy silts, clayey sands, till |
| UP-R | 40.70 | 4.43 | silty sands, fine sands - well sorted sands, glacial outwash |
| | | | |
| DP-H | 2.84 | 22.68 | silt, sandy silts, clayey sands, till |
| DI-H | 2.53 | 1.14 | silt, sandy silts, clayey sands, till |
| UI-H | 5.09 | 100.48 | silt, sandy silts, clayey sands, till |
| UP-H | 552.62 | 488.95 | well sorted gravel, well-sorted sands, glacial outwash |

Table 2-2. Conservative estimates of subsurface mean travel times derived from measured $K_{sat}$ and the shortest possible straight-line travel distance between wells. Estimated hillslope-riparian travel times were determined with the hillslope $K_{sat}$ in each transect, and riparian-stream travel times were determined with the riparian $K_{sat}$ in each transect. Travel times are longer in the DP and DI transects, and the UP transect has the shortest travel time by two orders of magnitude.

| Site | Estimated mean travel time *(d)* | Estimated mean total travel time *(d)* |
|---|---|---|
| DP transect | | |
| hillslope-riparian | 321.4 | 346.5 |
| riparian-stream | 25.1 | |
| DI transect | | |
| hillslope-riparian | 399.4 | 402.1 |
| riparian-stream | 2.7 | |
| UI transect | | |
| hillslope-riparian | 150.6 | 171.1 |
| riparian-stream | 20.5 | |
| UP transect | | |
| hillslope-riparian | 1.4 | 5.0 |
| riparian-stream | 3.6 | |

*2.5.2 Vertical flow paths*

Vertical hydraulic gradients (VHGs) and the saturated hydraulic conductivity ($K_{sat}$) of streambed sediments varied predictably with the observed patterns of flow permanence. Downward VHGs were measured at all stream locations in August 2018, and both VHGs and saturated hydraulic conductivities were higher in the intermittent reach. At UI-S, in the intermittent reach, the VHG was 1-2 orders of magnitude higher than the other locations, promoting significant losses from the stream. The smallest VHG occurred at DP-S, suggesting minimal losses from the stream in this perennial reach. $K_{sat}$ values were similar at DP-S, DI-S and UI-S, but $K_{sat}$ was 3-4x lower in UP-S, upstream of the point of flow disappearance (Table 2-3). Although all locations were found to be hydrologically losing, the hydraulic conductivity controls the rate at which water is lost from the stream. Therefore, lower hydraulic conductivity in UP-S delays stream drying compared to downstream locations.

Table 2-3. In-stream saturated hydraulic conductivity and vertical hydraulic gradient. $K_{sat}$ was lowest at UP-S and higher in perennial sites than intermittent sites. All sites had downward VHGs, but sites in the intermittent reach (DI-S and UI-S) had higher VHGs than sites in the perennial reaches. The VHG at UI-S was 1-2 orders of magnitude higher than VHGs at all other sites.

| Site | $K_{sat}$ (mm/hr) | error (mm/hr) | VHG (cm/cm) | error (cm/cm) |
|---|---|---|---|---|
| DP-S | 723.25 | 438.59 | -0.04 | -0.22 |
| DI-S | 868.10 | 555.86 | -0.33 | -0.19 |
| UI-S | 1062.70 | 646.72 | -1.12 | -0.19 |
| UP-S | 231.55 | 137.27 | -0.23 | -0.19 |

The absolute errors in VHG were similar across sites, ranging from 0.19-0.22 (cm/cm), but ranged from 17-565 % of calculated VHGs. When the VHG was low, the uncertainty was proportionally much greater than when the VHG was larger. For example, at DP-S the uncertainty in VHG is one order of magnitude higher than the calculated VHG, whereas at UI-S the uncertainty in VHG is an order of magnitude lower than the calculated VHG.

*2.5.3 Longitudinal flow paths*

In calculating hydrologic exchange, we made assumptions of 1) constant discharge, 2) full mass recovery, and 3) complete mixing of tracer during each 3-day experimental period. Thus, if the discharge increased between paired injections due to a storm, we might overestimate the change in discharge over the reach and hydrologic gains would be artificially high. We would also underestimate the mass recovered due to an artificially low calculation of $Q_D$, and overestimate hydrologic losses. We assume full mass recovery by attributing any mass loss to a loss of discharge to the subsurface over the reach. However, if the duration of the test is not long enough, any tracer that is merely delayed in transient storage rather than lost from the reach will be mischaracterized as mass loss – that is, tracer that has flowed in the subsurface past the measurement point (Payn et al., 2009; Ward et al., 2013). Finally, if complete mixing does not occur, local electrical conductivity measurements may be artificially low or high, and we risk inferring that there is more or less discharge lost, respectively, to the subsurface than in reality. To meet assumptions 1-3 to the best of our abilities, we 1) injected over a short 3-day period and avoided storms, 2) measured concentrations for 10 hours from the time of injection to increase our window of tracer mass detection, and 3) verified full mixing with a YSI Pro 30 with a T and EC probe that allowed us to adjust our injection/measurement points if needed.

Seven salt injection experiments occurred over a range of flows during the recession period (Figure 2-8). Repeated measurements of $Q_{GAIN}$ and $Q_{LOSS}$ along four reaches revealed one consistently gaining reach between 1660 and 1730 m above the outlet, directly downstream of where flow reappears. Throughout all flow conditions, this was a gaining reach, with average net discharge gains of 40-60 L/s. (Figure 2-9; Table 2- 4). Estimates of $Q_{LOSS}$ and $Q_{GAIN}$ did not indicate significant differences between the intermittent and remaining perennial reaches, except for this consistently gaining reach. In all other reaches, a net $Q_{GAIN}$ was measured during high flow periods from April through the end of May, while a net $Q_{LOSS}$ was measured in June and July (Figure 2-9). Uncertainties in hydrologic losses and gains were often quite large, mostly due to the large uncertainty in the integrated tracer concentration breakthrough curves.

Flow contributions between 1730 m and 1660 m were observed visually, especially during low flow conditions when multiple points of upwelling along the bed and bank of the stream were visible. Thermal camera (FLIR One) images taken in early June and mid-September show relative stream surface temperatures near one of these visually identified upwelling locations. In June, we observed a core of warmer water in the middle of the stream, with colder water present on both sides for 10 m along both stream banks where upwelling appeared to be active (Figure 2-11). The core of warmer water was the same temperature as upstream surface waters measured above the upwelling locations. In September, when upstream surface flows ceased, images taken at the same location show spatially homogeneous stream temperatures. Continuous stream temperatures measured at the same locations as the EC measurements show that temperatures were stable at and below EC 1660 (Figure 2-10), especially during low-flow periods.

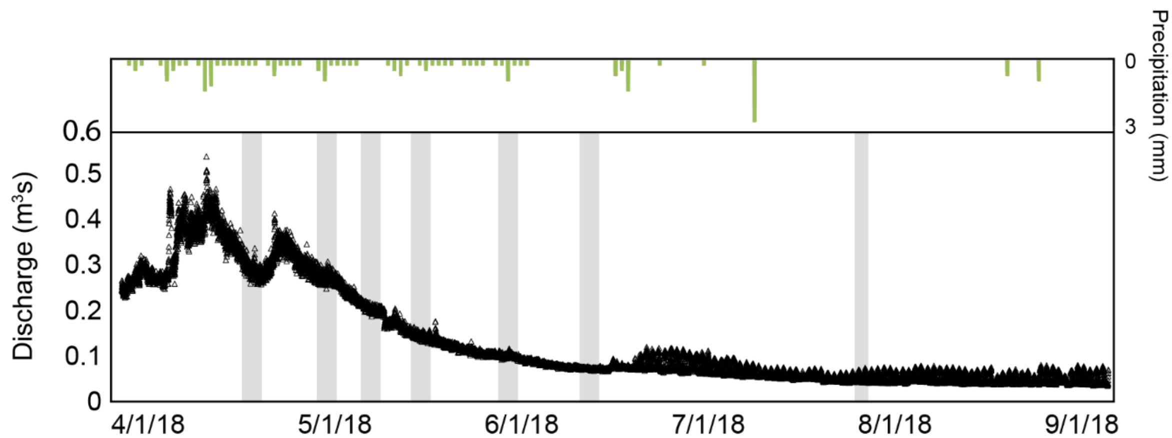


Figure 2-8. Detailed hyetograph and hydrograph at the outlet of Gibson Jack. Grey shading indicates the timing of salt dilution experiments conducted to determine longitudinal gains and losses along the stream. Precipitation data were measured at a local weather station (Figure 2-2).

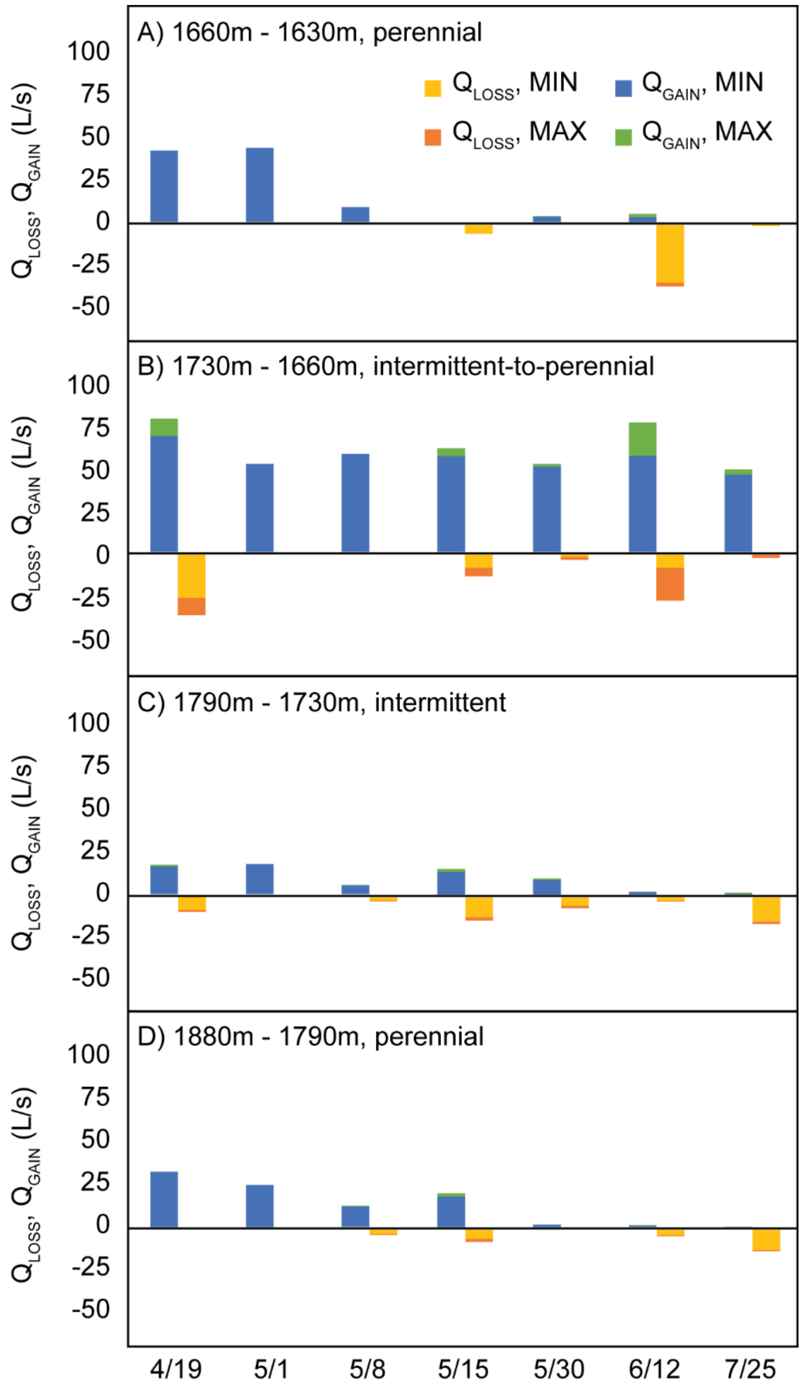


Figure 2-9. The maximum and minimum hydrologic loss and gain for each reach from April-July 2018. In the early season, the perennial sites (A and D) display net gaining conditions, but starting on 5/30, losses become more prevalent. Both losses and gains are visible throughout the season in the intermittent reach (C), with decreasing gains and increasing losses as the season progresses. A large persistent $Q_{GAIN}$ was observed in the intermittent-to-perennial reach (B).

Table 2-4. Tracer masses, integrated breakthrough curve areas, estimated discharges, and hydrologic exchange at four reaches in Gibson Jack Creek from high to low flow states during the seasonal recession from April to July 2018.

| Date | Reach | $M_D$ (g) | $C_D(t)dt$ (mg·s/L) | $Q_D$ (L/s) | $M_U$ (g) | $C_U(t)dt$ (mg·s/L) | $Q_U$ (L/s) | $C_{UD}(t)dt$ (mg·s/L) | $\Delta Q$ (L/s) | $M_{LOSS}$ (g) | $Q_{LOSS,MIN}$ (L/s) | $Q_{LOSS,MAX}$ (L/s) | $Q_{GAIN,MIN}$ (L/s) | $Q_{GAIN,MAX}$ (L/s) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 4/19/2018 | 1660-1630m | 1,393.40 | 4.20 | 331.37 | 1,028.92 | 3.57 | 288.50 | 3.59 | 42.87 | 0.00 | 0.00 | 0.00 | 42.87 | 42.87 |
| | 1730-1660m | 1,393.40 | 5.53 | 252.20 | 1,028.92 | 4.92 | 209.08 | 3.57 | 43.12 | -129.48 | -26.31 | -36.31 | 69.43 | 79.42 |
| | 1790-1730m | 1,028.92 | 4.92 | 209.08 | 857.52 | 4.29 | 199.71 | 3.93 | 9.37 | -35.74 | -8.32 | -9.09 | 17.70 | 18.47 |
| | 1880-1790m | 1,028.92 | 4.71 | 218.36 | 857.52 | 4.61 | 186.13 | 4.29 | 32.23 | 0.00 | 0.00 | 0.00 | 32.23 | 32.23 |
| 5/1/2018 | 1660-1630m | 808.77 | 2.95 | 274.36 | 764.55 | 3.32 | 230.10 | 2.98 | 44.26 | 0.00 | 0.00 | 0.00 | 44.26 | 44.26 |
| | 1730-1660m | 808.77 | 3.30 | 245.33 | 764.55 | 3.97 | 192.69 | 3.32 | 52.63 | 0.00 | 0.00 | 0.00 | 52.63 | 52.63 |
| | 1790-1730m | 764.55 | 3.97 | 192.69 | 865.32 | 4.97 | 174.21 | 4.60 | 18.48 | 0.00 | 0.00 | 0.00 | 18.48 | 18.48 |
| | 1880-1790m | 764.55 | 3.98 | 191.91 | 865.32 | 5.17 | 167.28 | 4.97 | 24.63 | 0.00 | 0.00 | 0.00 | 24.63 | 24.63 |
| 5/8/2018 | 1660-1630m | 767.44 | 3.77 | 203.83 | 748.49 | 3.87 | 193.38 | 3.80 | 10.45 | 0.00 | 0.00 | 0.00 | 10.45 | 10.45 |
| | 1730-1660m | 767.44 | 3.98 | 192.99 | 748.49 | 5.58 | 134.24 | 3.87 | 58.75 | 0.00 | 0.00 | 0.00 | 58.75 | 58.75 |
| | 1790-1730m | 748.49 | 5.58 | 134.24 | 652.28 | 4.98 | 130.86 | 4.76 | 3.38 | -12.80 | -2.57 | -2.69 | 5.95 | 6.07 |
| | 1880-1790m | 748.49 | 5.92 | 126.38 | 652.28 | 5.51 | 118.44 | 4.98 | 7.94 | -22.32 | -4.05 | -4.48 | 11.99 | 12.41 |
| 5/15/2018 | 1660-1630m | 514.26 | 3.13 | 164.12 | 558.97 | 3.24 | 172.42 | 3.30 | -8.30 | -16.86 | -5.20 | -5.10 | -3.10 | -3.19 |
| | 1730-1660m | 514.26 | 3.23 | 159.10 | 558.97 | 5.08 | 110.07 | 3.24 | 49.03 | -43.17 | -8.50 | -13.32 | 57.53 | 62.34 |
| | 1790-1730m | 558.97 | 5.08 | 110.07 | 488.48 | 4.50 | 108.59 | 3.93 | 1.49 | -55.53 | -12.34 | -14.12 | 13.83 | 15.60 |
| | 1880-1790m | 558.97 | 5.59 | 100.02 | 488.48 | 5.49 | 89.01 | 4.50 | 11.01 | -38.52 | -7.02 | -8.56 | 18.03 | 19.58 |
| 5/30/2018 | 1660-1630m | 460.04 | 4.05 | 113.59 | 468.91 | 4.31 | 108.87 | 4.11 | 4.72 | -1.54 | -0.36 | -0.37 | 5.08 | 5.09 |
| | 1730-1660m | 460.04 | 4.33 | 106.21 | 468.91 | 8.36 | 56.07 | 4.31 | 50.14 | -11.47 | -1.37 | -2.66 | 51.51 | 52.80 |
| | 1790-1730m | 468.91 | 8.36 | 56.07 | 453.72 | 8.57 | 52.92 | 7.19 | 3.15 | -50.47 | -5.89 | -7.02 | 9.04 | 10.17 |
| | 1880-1790m | 468.91 | 8.81 | 53.25 | 453.72 | 8.76 | 51.77 | 8.57 | 1.47 | 0.00 | 0.00 | 0.00 | 1.47 | 1.47 |
| 6/12/2018 | 1660-1630m | 410.54 | 4.93 | 83.30 | 459.64 | 4.08 | 112.62 | 3.87 | -29.32 | -137.68 | -33.73 | -35.62 | 4.41 | 6.30 |
| | 1730-1660m | 410.54 | 4.85 | 84.64 | 459.64 | 13.00 | 35.35 | 4.08 | 49.29 | -114.18 | -8.78 | -27.98 | 58.08 | 77.27 |
| | 1790-1730m | 410.54 | 12.76 | 32.17 | 459.64 | 14.06 | 32.69 | 13.00 | -0.52 | -41.30 | -2.94 | -3.18 | 2.42 | 2.65 |
| | 1880-1790m | 410.54 | 14.76 | 27.82 | 459.64 | 14.71 | 31.24 | 14.06 | -3.43 | -68.56 | -4.66 | -4.88 | 1.23 | 1.45 |
| 7/25/2018 | 1660-1630m | 398.75 | 7.79 | 51.22 | 390.40 | 7.30 | 53.50 | 7.49 | -2.28 | -6.93 | -0.95 | -0.93 | -1.33 | -1.35 |
| | 1730-1660m | 398.75 | 7.92 | 50.34 | 390.40 | 101.26 | 3.86 | 7.30 | 46.48 | -23.06 | -0.23 | -3.16 | 46.71 | 49.64 |
| | 1790-1730m | 390.40 | 101.26 | 3.86 | 399.81 | 22.12 | 18.08 | 20.35 | -14.22 | -321.35 | -14.53 | -15.79 | 0.31 | 1.57 |
| | 1880-1790m | 390.40 | 96.88 | 4.03 | 399.81 | 23.35 | 17.12 | 22.12 | -13.09 | -310.68 | -13.31 | -14.05 | 0.21 | 0.95 |

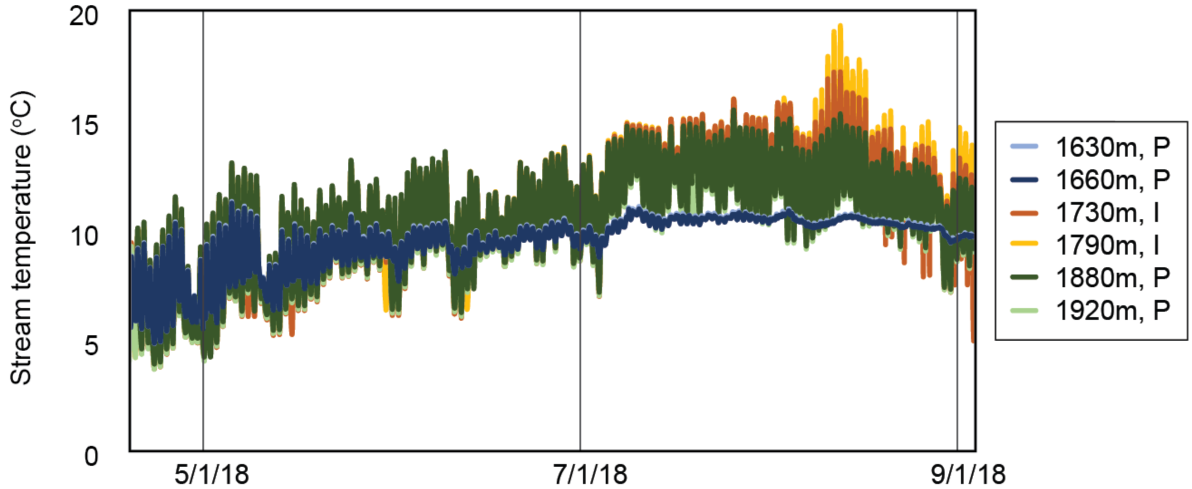


Figure 2-10. Stream temperature varies longitudinally throughout the study reach from April-September 2018. Colors refer to temperatures measured in the stream at the stated distance above the watershed outlet (Figure 2-2), and each site is designated intermittent (I) or perennial (P) based on flow presence/absence observations made throughout the season (see Figure 2-14). The diel amplitude of stream temperature decreases over time at 1660 m and 1630 m, below the consistently gaining reach (Figure 2-11), and continues to decrease through early September, while remaining relatively consistent at all other measured locations throughout the seasonal flow recession.

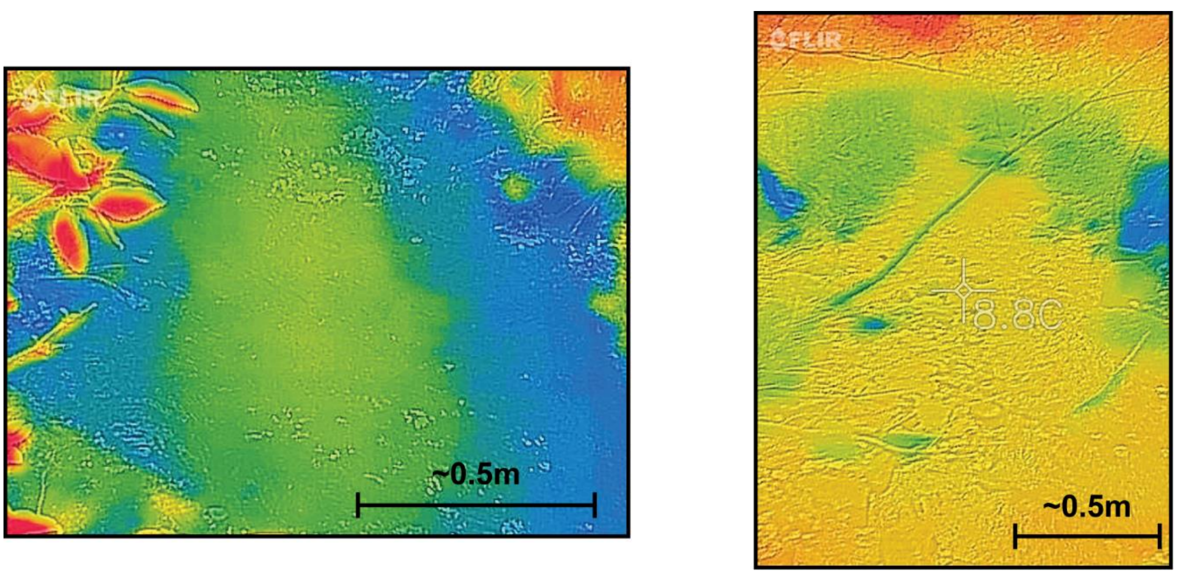


Figure 2-11. Thermal photos of Gibson Jack in the consistently gaining reach between 1730-1660 m. Colors represent relative surface temperatures. Left: Photo taken on June 1st, 2018, when temperatures in the middle of the stream are warmer than temperatures near both stream banks. Right: Photo taken at the same location on September 18th, 2018, when stream surface temperature is homogeneous.

*2.5.4 Integrated observations*

*2.5.4.1 Watershed area scaling relationships*

A power-law relationship (e.g. $Q = kA^c$, where $k$ and $c$ are constants) is often employed to predict discharge from topographic area (e.g., Galster, 2007), but we found that drainage area only weakly predicted discharge in this basin (Figure 2-12). Coefficients of determination ($R^2$) for the power-law area-discharge relationship during each salt injection experiment ranged from 0.04-0.69 over the season. The relationship was usually strongest at highest flows, but the discharge predicted at the intermittent locations (1730 m and 1790 m upstream from the outlet) was consistently higher than observed discharge at those sites as shown by negative residuals in the area-discharge relationship for each of seven sampling dates summarized in Figure 2-12h across outlet flows of 0.054-0.296 $m^3/s$. Two perennial locations (1630 m and 1660 m upstream from the outlet) had consistently higher discharge than predicted by the power-law area-discharge relationship. These patterns persisted through all seven measurements of discharge. Minimal deviations from the area-discharge relationship were associated with other perennial reaches. Differences among these perennial reaches are discussed in the vertical and longitudinal flow sections below.

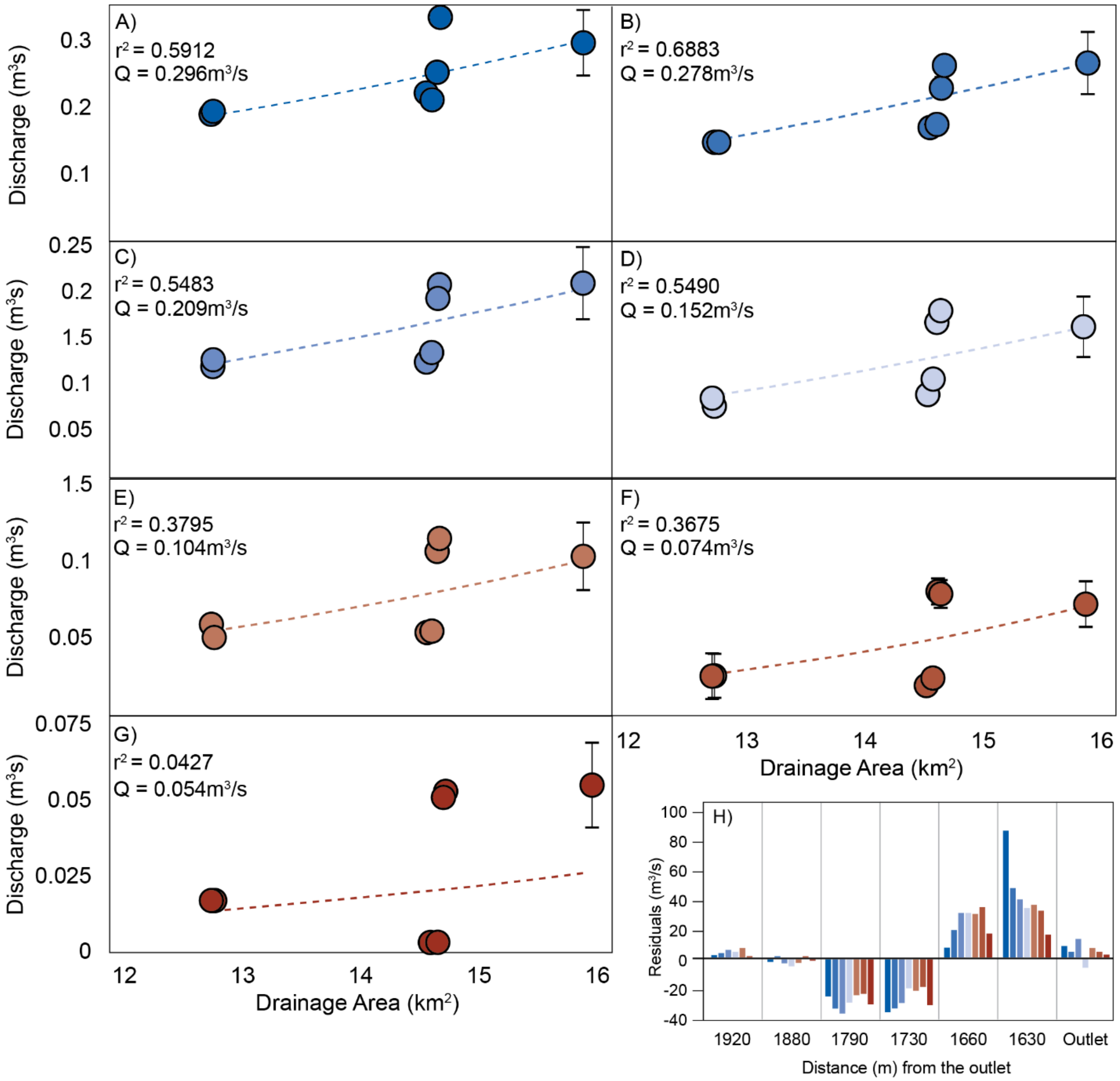


Figure 2-12. The relationship between drainage area and discharge at the outlet and all six EC logger locations during each salt dilution experiment. Mean outlet discharge decreased with the seasonal recession from 0.296 $m^3$/s to 0.054 $m^3$/s in panels A to G. The dashed lines represent the best-fit power-law relationship between drainage area and discharge at each flow condition. H) Residuals from each best-fit line, with colors correlating to plots A-G. Positive residuals are observed in the downstream perennial reach at 1630m and 1660m from the outlet, and negative residuals are observed in the intermittent reach at 1730m and 1790m from the outlet.

*2.5.4.2 Observed drying and rewetting patterns*

In early July, diurnal water-level fluctuations began to increase in the in-stream wells, DI-S and UI-S (Figure 2-13C). The stream began to dry at both DI and UI transects in early August 2018. This is reflected by decreasing water levels in the intermittent wells and much higher diel fluctuations relative to the stable water levels observed in the perennial in-stream wells (DP-S and UP-S) (Figure 2-13C).

Both in-stream wells in the intermittent reach dried in August 2018, but the patterns of drying differed between the two wells (Figure 2-13D). Drying was first observed at UI-S on August 6$^{th}$. A short period of rewetting at UI-S occurred from August 22$^{nd}$-23$^{rd}$, after which the stream was dry at UI-S until a storm on October 9$^{th}$ initiated rewetting. Drying first occurred at DI-S on August 7$^{th}$. Unlike patterns observed at UI-S, drying and rewetting occurred daily or every other day at DI-S until mid-September. The stream then remained dry at DI-S until the October 9$^{th}$ storm.

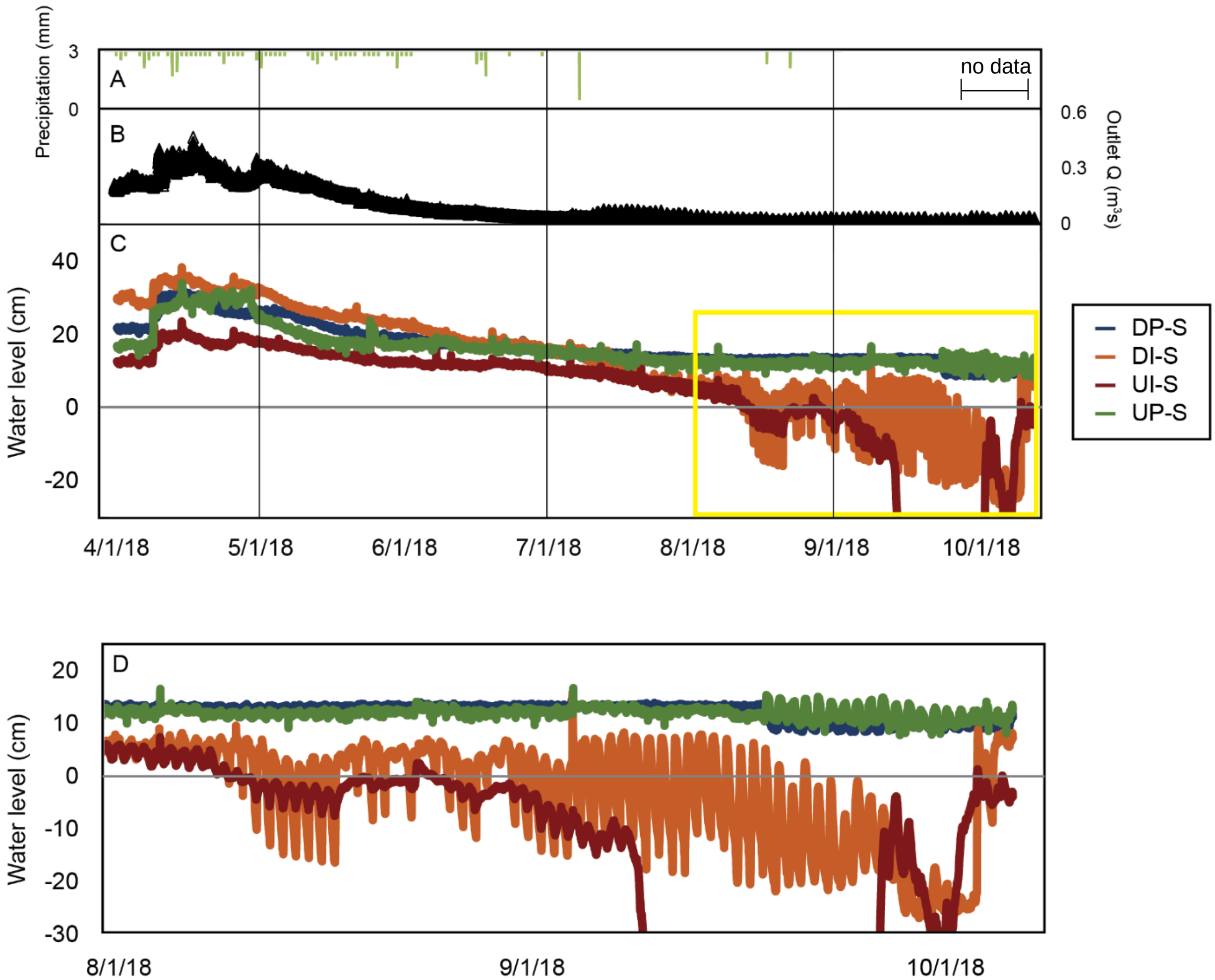


Figure 2-13. Water level in the in-stream wells, with zero indicating the height of the streambed at each location. Points below zero, such as those seen at DI-S and UI-S in Aug-Oct, indicate the disappearance of surface flow. A gap between 9/15 and 10/1 in the UI-S water level indicates that water levels fell more than 30 cm below the stream bed, and the in-stream piezometer remained dry during this period. A) hyetograph, B) outlet discharge, C) water level at DP-S, DI-S, UI-S & UP-S and D) late-season diel dynamics highlighted in yellow in C.

Daily dynamic drying and rewetting was also observed with the EC dataloggers and adapted Hobo Pendant loggers (Chapin, Todd and Zeigler, 2014) at locations throughout the intermittent reach (Figure 2-14). Dynamic drying and rewetting was observed throughout this period, with the exception of EC 1730 and EC 1820. No surface flow was observed at EC 1730 after September 5th. Drying was observed to occur at EC 1820 only on September 8th, but otherwise this location maintained surface flow. EC 1820 was located ~5 m below a tributary confluence with the mainstem of Gibson Jack Creek, which may have sustained flow in this location. Surface flow became more persistent in the intermittent reach in mid-September and early October, but rewetting did not progress sequentially from upstream to downstream or vice versa. At 1690 m, flow became more consistent on October 4th, likely buffered by a pulse of precipitation (~0.4cm), which increased surface runoff. At 1730 m, no rewetting was observed by Oct 7th, the last date for which data are available for this thesis. At 1760 m, rewetting initiated on September 21st, and at 1790 m, rewetting was observed starting on September 13th. In all cases, however, short periods of drying were observed during the re-wetting period, illustrating the highly dynamic processes of drying and rewetting (Figure 2-14).

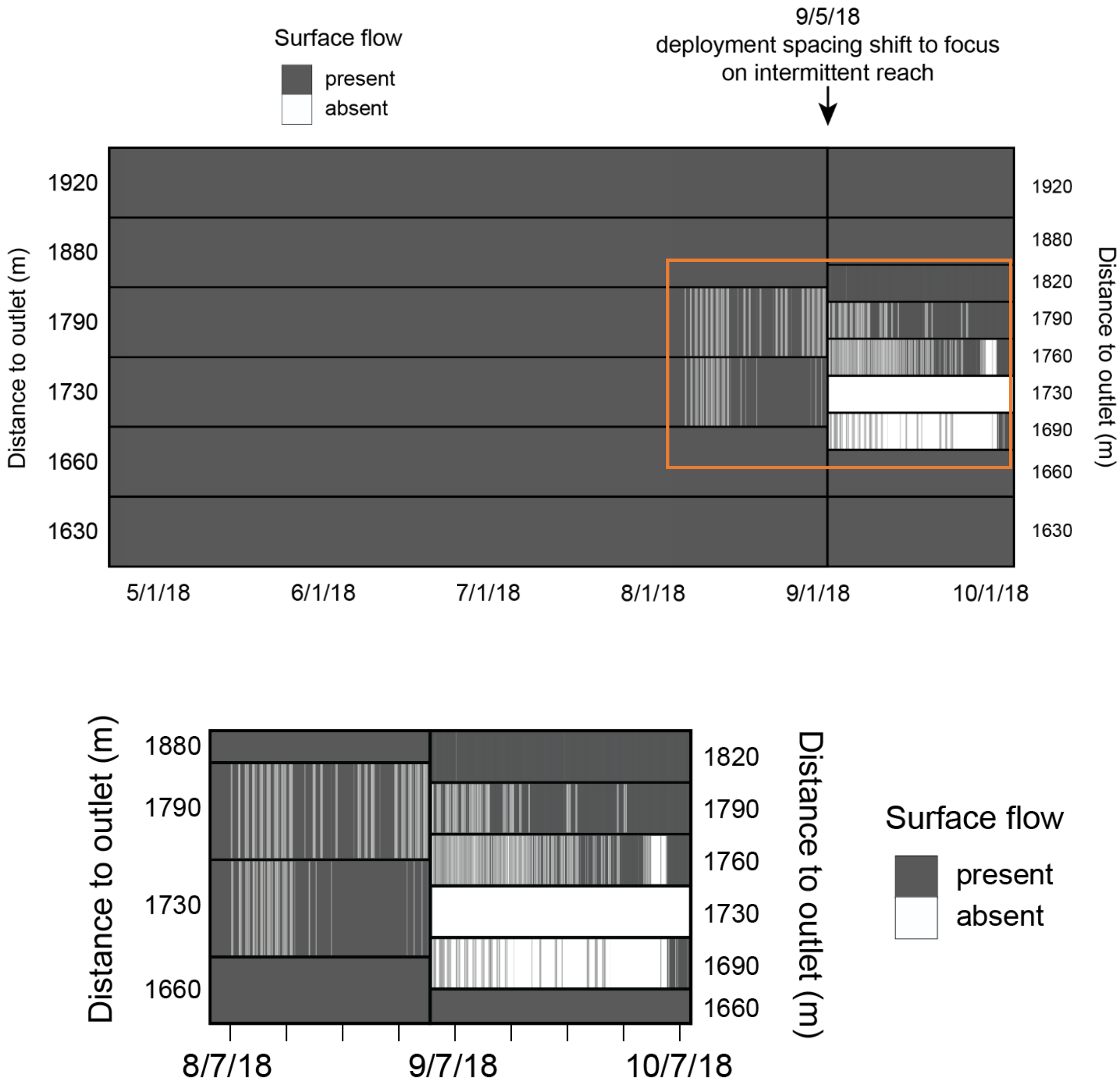


Figure 2-14. Presence and absence of surface flow measured every 15 minutes based on EC and water level dataloggers throughout the study reach. Portions that appear light gray reflect high-frequency variations in surface flow presence/absence, usually at approximately daily timescales. The area surrounded by the orange box is enlarged below, highlighting drying and rewetting patterns observed in the intermittent reach in August-September 2018. On September 5$^{th}$, the logger deployment was shifted to infer surface flow presence/absence dynamics on a finer spatial scale as described in the methods section.

## 2.6 Discussion

*2.6.1 Flow presence and absence resulting from 3D subsurface flow paths*

By integrating the patterns from lateral, vertical, and longitudinal shallow subsurface flow paths, we can understand the drivers of spatial patterns of surface flow in a perennial-to-intermittent reach in Gibson Jack creek. Figure 2-15 summarizes the relative influence of lateral, vertical, and longitudinal flow paths in perennial and intermittent reaches in the study at high, low, and intermittent flow periods. Specifically, consistently large hydrologic gains due to an in-channel spring promoted continuous surface flow in the perennial reach at DP-S. This reach also had moderate lateral water contributions and the smallest losing VHG, which reduced the amount of flow leaving the stream. By contrast, HRS connectivity was highest in the perennial reach above flow disappearance (UP transect), although the $Q_{gain}$ and $Q_{loss}$ data suggested no significant groundwater source at this location. Streambed hydraulic conductivity was an order of magnitude lower in the UP transect than all other locations, and the VHG was relatively small, reducing losses. In the intermittent reach spanning DI-S and UI-S, the VHG was most negative, HRS connectivity was low to moderate, hydraulic conductivity was the highest promoting vertical losses that were not maintained by lateral inputs, and there was a high subsurface capacity for flow. At UI-S, the hydraulic gradient was consistently directed away from the stream toward the riparian zone during all flow conditions.

In general, these observations suggest that during low to intermittent flow periods: 1) persistent lateral subsurface sources sustain surface flow in perennial reaches; 2) in-stream subsurface heterogeneity results in larger and faster rates of vertical losses in intermittent reaches; 3) consistently gaining sections are less likely to dry; and 4) upstream contributions and small vertical losses maintain surface flow presence.

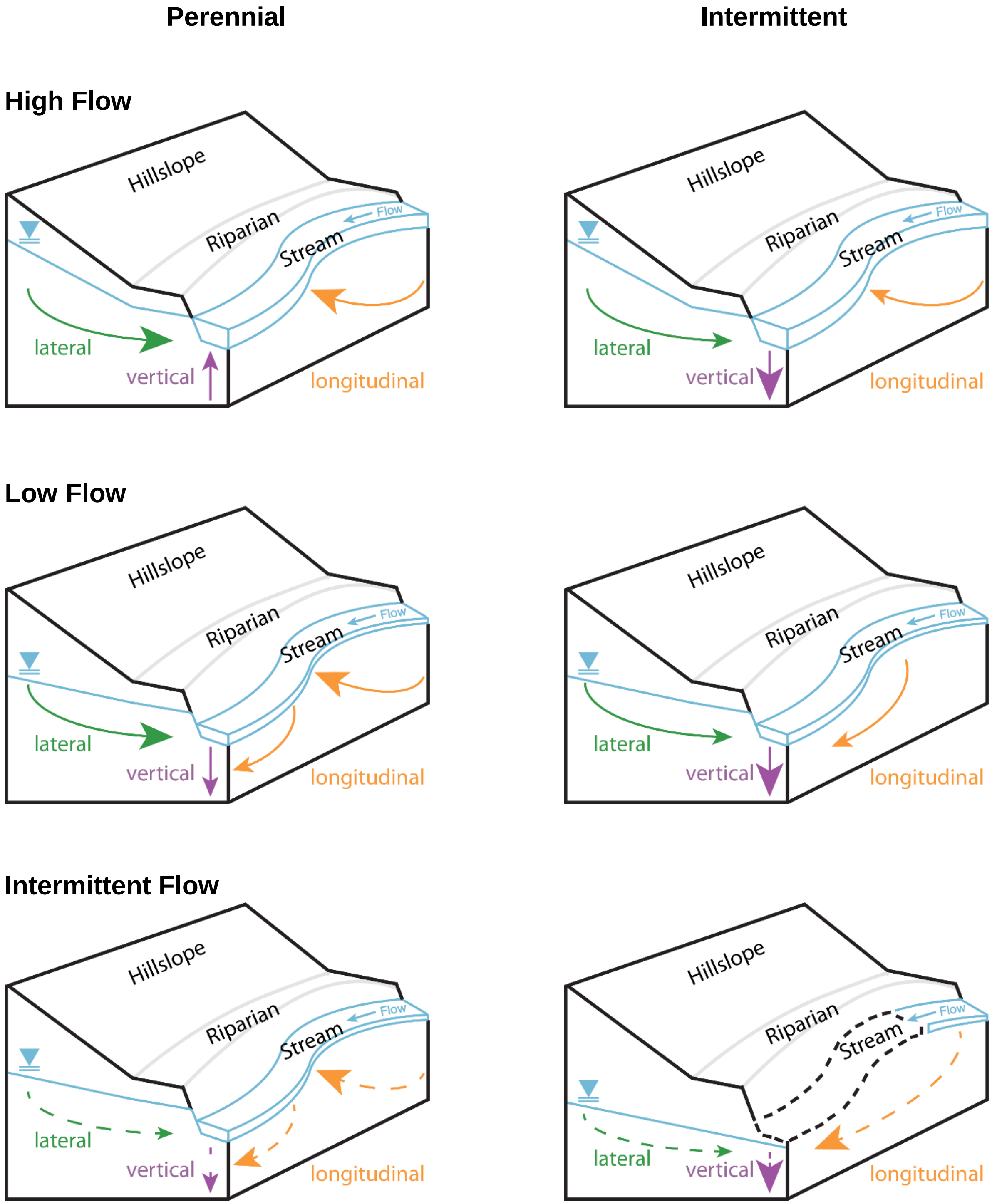


Figure 2-15. Summary of 3D shallow subsurface flow paths in perennial and intermittent reaches during high, low, and intermittent flow scenarios. The size of the arrowhead indicates relative magnitude of that flow path. Flow in each dimension was quantified with different units, and thus comparisons can only be made within a given dimension. Predicted flow paths during intermittent flow are indicated by dashed lines because some of the methods applied cannot be applied to fully dry in-stream conditions.

In contrast to findings from Zimmer and McGlynn (2017b) where bi-directional gradients were observed as flows decreased (head gradients were towards the stream during high storage states and away from the stream during low storage states), we observed consistent directionality through all flow conditions. Site-specific characteristics likely drive these contrasting observations. The Gibson Jack watershed has much greater relief than the study site in the Duke Forest Research Watershed (DFRW) and does not contain clay soils with perched water tables, as in DFRW. At DRFW, expansion and contraction of surface flows is driven by seasonal water table fluctuations as well as event-activated contributions through surface and shallow subsurface paths (Zimmer and McGlynn, 2018). In the DRFW study, a system of nested wells was used to measure water levels in the shallow soil horizons (A/B and B/C) and down to the deeper saprolite and weathered bedrock. During low storage states, when head gradients were away from the stream, shallow subsurface flow path contributions drove streamflow in ephemeral headwaters (Zimmer and McGlynn, 2017b) because impeding soil layers quickly directed flow to the stream during events. At Gibson Jack, we also found that subsurface flow paths control flow permanence, but unlike at DFRW, Gibson Jack does not have any significantly impeding soil layers, and the soils are much shallower. Steep slopes in the Gibson Jack watershed drive lateral flow paths from hillslopes toward the stream, while in DFWR, the impeding soil layers drive these flow paths in a catchment with a gradual slope. Zimmer and McGlynn (2017b) found that dominant flow paths differed between headwaters and lowlands at DFRW. Lowland streams were fed by the regional water table and depended less on dynamic shallow subsurface flow paths. We did not evaluate changes in flow paths across the network at Gibson Jack, but do not expect to see similar changes because shallow perched water tables were never observed at Gibson Jack. Soil depth and type play key roles in subsurface water delivery to streams, and thus subsurface characterization is important for understanding water transport to streams, which affects flow permanence.

*2.6.2 Deviations from drainage-discharge relationship indicate internal catchment properties*

The power-law relationship between drainage area and discharge ($Q = kA^c$) presumes that each unit of upslope area contributes the same amount of discharge to the stream. This is founded on the assumption of spatially homogeneous precipitation and evaporation rates throughout a given catchment. These assumptions are inherent in many models (Rodriguez-Iturbe and Rinaldo, 1997; Galster, 2007; Godsey and Kirchner, 2014; Bergstrom, Jencso and McGlynn, 2016; Ward et al., 2018), though they are unlikely to be true because of spatial heterogeneity in precipitation and evapotranspiration (Gurtz, Baltensweiler and Lang, 1999). Even if precipitation and evaporation rates were equivalent throughout a catchment, subsurface heterogeneity and flow path variability could still lead to unequal contributions of water per unit area.

Although we found that a fitted discharge-drainage area (Q-DA) relationship was not consistently reliable for at-a-point discharge predictions, the Q-DA relationship may still be useful as a predictive tool for assessing reaches more or less susceptible to drying. Perennial reaches were either well-described by the power-law relationship, or flows were under-predicted (e.g. where springs directly fed the stream). Over-predictions of flow were observed only in the intermittent reach. This pattern was consistent throughout all flow conditions, suggesting that evaluating the Q-DA relationship at high flows may be useful for predicting intermittency where topography is a poor predictor of discharge. Payn et al. (2012) similarly found watershed topography poorly predicted discharge under certain conditions. Specifically, they found a decreasing influence of topography (quantified by watershed area) on stream discharge as flows decreased. We found the power-law relationship also degraded with flow state, however, watershed topography was either a consistently good or a consistently poor predictor of flows at each discharge measurement location. That is, the influence of topography was temporally, but not spatially, consistent at each discharge measurement location. Systematic deviations from a fitted power-law relationship between discharge and drainage area may thus reflect flow permanence in the stream. This method could be used to identify intermittent reaches within stream networks or potentially to predict the locations of reaches that are more susceptible to drying in the future. Although this method may not identify every location of intermittency, it may help identify streams with large or spatially frequent disconnections.

*2.6.3 Subsurface capacity and evapotranspiration losses controls surface flow drying and rewetting dynamics*

Godsey and Kirchner (2014) suggested that the emergence of surface flow would occur at points where total flow exceeds the accommodation capacity of the subsurface and the disappearance of flow would occur at points where subsurface capacity is greater than the total flow. Although subsurface capacity is an important control on surface expression, the role of fractures and springs on the persistence of surface flow through hydrologic gains and losses is not explicitly considered in their formulation. Our results identified a significant gaining reach fed by an in-stream spring. This spring was the observed location of surface flow reemergence during the end of the seasonal recession, below which flow was perennial and above which the streambed was dry. Conversely, fractures could significantly increase hyporheic transmissivity if the capacity of the fractures is larger and they lead to conditions poorly represented by porous media models such as a Darcy’s law.

In some cases, losses to the subsurface can be small compared to losses to the atmosphere, and cyclical drying and rewetting patterns associated with diel ET cycles have only recently been predicted in intermittency models (Ward, et al., 2018). We observed diel fluctuations in surface flow in August and September 2018 (Figure 2-13) during the period when discharge fluctuates near the capacity of the subsurface to transport water. Unexpectedly, the amplitude of fluctuations (Figure 2-13) as well as the frequency and duration of drying (Figure 2-14) varied along the intermittent reach in Gibson Jack Creek. For example, diel fluctuations at UI-S were at most ~5 cm in amplitude, while at DI-S they were up to ~25 cm. At 1730 m from the outlet, the streambed remained dry throughout the month of September. Only 30 m away, either downstream or upstream from this location, however, surface flow returned on a daily basis or every few days (Figure 2-14). We hypothesize that this spatial heterogeneity in the drying pattern reflects differences in the amount of evapotranspiration loss, 3D subsurface connectivity, and/or the capacity of the subsurface to accommodate flow in these locations. For example, at 1820 m from the outlet, in the intermittent reach, surface flow was observed consistently from late April through early October, which may result from a smaller subsurface capacity at this location than at 1730 m from the outlet. Alternatively, a tributary confluence enters the study area along the mainstem of Gibson Jack at ~ 1825 m above the outlet. This tributary maintained surface flow throughout the study period, and may have sustained surface flow at 1820 m, even if the subsurface capacity was comparable to that of 1730 m, which regularly dried. Without this tributary input, we might have observed diel fluctuations in surface flow at 1820 m. 3D connectivity also differed between intermittent locations. Although lateral connectivity was more common in the UI transect than the DI transect, vertical losses were also greater. These flow paths affect the volume of water in the stream at a given point in time, and might explain differences in diel fluctuations at intermittent locations.

In summary, drying and wetting of the drainage network are dynamic processes, reflecting inflows, outflows, and storage. Local characteristics are critically important at the reach-scale, and at the network scale, the spatial heterogeneity and topology of those local characteristics control whether drying and rewetting lead to network contraction or disconnections along the network.

**2.7 Conclusion**

We found that three-dimensional shallow subsurface flow paths were the primary controls on flow permanence in a discontinuous stream reach. Perennial and intermittent flow can arise from a variety of subsurface flow path relationships and the integration of these flow paths determines the presence or absence of surface flow (Figure 2-15). Intermittency is a dynamic process with high spatiotemporal variability in drying and rewetting patterns. Measurements of watershed-scale subsurface dynamics are needed to extend this work from the reach to the network scale, and we suggest that incorporating geophysical data, local topographic analysis and network topology into modelling efforts would likely improve spatial predictions of flow permanence and highlight the scales at which key controls on intermittency dominate.

Negative deviations from power-law drainage-area relationship predicted locations of intermittency even during high flow states. This method should be tested to determine its applicability to other intermittent watersheds. If it is transferable, water managers could use this tool to predict current and future intermittent reaches within stream networks with a focus on sites that are particularly susceptible to drying. Overall, refinement of these predictive and mechanistic models will improve predictions of stream sensitivity to future changes in land use and climate.

## References


Acuña, V., Datry, T., Marshall, J., Barcelo, D., Dahm, C.N., Ginebreda, A., McGregor, G., Sabater, S., Tockner, K., and Palmer, M.A., 2014, Why should we care about temporary waterways? Science, v. 343, p. 1080–1081, doi: 10.1126/science.1246666.

Baxter, C. V., and Hauer, F.R., 2003, Measuring groundwater – stream water exchange : New techniques for installing minipiezometers and estimating hydraulic conductivity: Transactions of the American Fisheries Society, v. 132, p. 493–502, doi: 10.1577/1548-8659(2003)132.

Bergstrom, A., Jencso, K., and McGlynn, B., 2016, Spatiotemporal processes that contribute to hydrologic exchange between hillslopes, valley bottoms, and streams: Water Resources Research, p. 1–18, doi: 10.1002/2015WR017972.

Bencala, K.E., Gooseff, M.N., and Kimball, B.A., 2011, Rethinking hyporheic flow and transient storage to advance understanding of stream-catchment connections: Water Resources Research, v. 47, p. 1–9, doi: 10.1029/2010WR010066.

Boulton, A.J., Findlay, S., Marmonier, P., Stanley, E.H., Boulton, A.J., Findlay, S., Marmonier, P., Stanley, E.H., and Valett, H.M., 1998, The functional significance of the hyporheic zone in streams and rivers: Annual Review of Ecology and Systematics, v. 29, p. 59–81.

Blumstock, M., Tetzlaff, D., Malcolm, I.A., Nuetzmann, G., and Soulsby, C., 2015, Baseflow dynamics: Multi-tracer surveys to assess variable groundwater contributions to montane streams under low flows: Journal of Hydrology, v. 527, p. 1021–1033, doi: 10.1016/j.jhydrol.2015.05.019.

Buttle, J.M., Boon, S., Peters, D.L., Spence, C., (Ilja) van Meerveld, H.J., and Whitfield, P.H., 2012, An overview of temporary stream hydrology in Canada: Canadian Water Resources Journal, v. 37, p. 279–310, doi: 10.4296/cwrj2011-903.

Chapin, T.P., Todd, A.S., and Zeigler, M.P., 2014, Robust, low-cost data loggers for stream temperature, flow intermittency, and relative conductivity monitoring: Water Resources Research, p. 6542–6548, doi: 10.1002/2013WR015158.

Costigan, K.H., Jaeger, K.L., Goss, C.W., Fritz, K.M., and Goebel, P.C., 2016, Understanding controls on flow permanence in intermittent rivers to aid ecological research: Integrating meteorology, geology and land cover: Ecohydrology, doi: 10.1002/eco.1712.

Day, D.G., 1978, Drainage density and streamflow: Earth Surface Processes, v. 3, p. 319–326.

Detty, J.M., and McGuire, K.J., 2010, Topographic controls on shallow groundwater dynamics: Implications of hydrologic connectivity between hillslopes and riparian zones in a till mantled catchment: Hydrological Processes, v. 24, p. 2222–2236, doi: 10.1002/hyp.7656.

Dunne, T., and Black, R.D., 1970, Partial area contributions to storm runoff in a small New England watershed: Water Resources Research, v. 6, p. 1296–1311

Fetter, C.W., 2013, Applied Hydrogeology: Pearson New International Edition, Fourth Edition, Pearson Education Limited, 612 p.

Fleckenstein, J.H., Niswonger, R.G., and Fogg, G.E., 2006, River-aquifer interactions, geologic heterogeneity, and low-flow management: Ground Water, v. 44, p. 837–852, doi: 10.1111/j.1745-6584.2006.00190.x.

Freeze, A.R., 1972, Role of subsurface flow in generating surface runoff 1. Base flow contributions to channel flow: Water Resources Research, v. 8, p. 609–623.

Galster, J.C., 2007, Natural and anthropogenic influences on the scaling of discharge with drainage area for multiple watersheds: Geosphere, v. 3, p. 260–271, doi: 10.1130/GES00065.1.

Godsey, S.E., and Kirchner, J.W., 2014, Dynamic, discontinuous stream networks: Hydrologically driven variations in active drainage density, flowing channels and stream order: Hydrological Processes, v. 28, p. 5791–5803, doi: 10.1002/hyp.10310.

Gurtz, J., Baltensweiler, A., and Lang, H., 1999, Spatially distributed hydrotope-based modelling of evapotranspiration and runoff in mountainous basins: Hydrological Processes, v. 13, p. 2751–2768.

Hewlett, J.D., and Hibbert, A.R., 1966, Factors affecting the response of small watersheds to precipitation in humid areas: Forest Hydrology, p. 275-279, doi: 10.1177/0309133309338118.

InPort. OCM Partners, 2015 OLC FEMA Lidar: Snake River, ID. https://inport.nmfs.noaa.gov/inport/item/49758. (accessed 2017)

Jencso, K.G., McGlynn, B.L., Gooseff, M.N., Wondzell, S.M., Bencala, K.E., and Marshall, L.A., 2009, Hydrologic connectivity between landscapes and streams: Transferring reach- and plot-scale understanding to the catchment scale: Water Resources Research, v. 45, p. 1–16, doi: 10.1029/2008WR007225.

Kline, R.P., 1978, Municipal watershed management plan for West Mink and Gibson Jack creeks of the Pocatello Ranger District: Caribou National Forest Open-File Report, 122 p.

Larned, S.T., Datry, T., Arscott, D.B., and Tockner, K., 2010, Emerging concepts in temporary-river ecology: Freshwater Biology, v. 55, p. 717–738, doi: 10.1111/j.1365-2427.2009.02322.x.

Nadeau, T.-L., and Rains, M.C., 2007, Hydrological connectivity between headwater streams and downstream waters: How science can inform policy: Journal of the American Water Resources Association, v. 43, p. 118–133, doi: 10.1111/j.1752-1688.2007.00010.x.

NOAA National Weather Service, Pocatello, ID. https://www.wrh.noaa.gov/pih/climate/descrip.php (accessed October 2018).

Payn, R.A., Gooseff, M.N., McGlynn, B.L., Bencala, K.E., and Wondzell, S.M., 2012, Exploring changes in the spatial distribution of stream baseflow generation during a seasonal recession: Water Resources Re, v. 48, p. 1–15, doi: 10.1029/2011WR011552.

Payn, R.A., Gooseff, M.N., McGlynn, B.L., Bencala, K.E., and Wondzell, S.M., 2009, Channel water balance and exchange with subsurface flow along a mountain headwater stream in Montana, United States: Water Resources Research, v. 45, doi: 10.1029/2008WR007644.

Raymond, P.A., Hartmann, J., Lauerwald, R., Sobek, S., McDonald, C., Hoover, M., Butman, D., Striegl, R., Mayorga, E., Humborg, C., Kortelainen, P., Dürr, H., Meybeck, M., Ciais, P., and Guth, P., 2013, Global carbon dioxide emissions from inland waters: Nature, v. 503, p. 355–359, doi: 10.1038/nature12760.

Rodriguez-Iturbe, I., and Rinaldo, A., 1997, Fractal river basins: Change and self-organization: Cambridge University Press, 570 p.

Rushlow, C.R., and Godsey, S.E., 2017, Rainfall–runoff responses on Arctic hillslopes underlain by continuous permafrost, North Slope, Alaska, USA: Hydrological Processes, v. 31, p. 4092–4106, doi: 10.1002/hyp.11294.

U.S. Department of Agriculture. Natural Resources Conservation Service Soils: Web soil survey: https://www.nrcs.usda.gov/wps/portal/nrcs/detail/soils/survey/tools/?cid=nrcseprd1407030 (accessed October 2018).

U.S. Geological Survey. NWIS Site Information for Idaho. https://nwis.waterdata.usgs.gov/id/nwis/inventory/?site_no=13075500&agency_cd=USGS. (accessed February 2017)

Ward, A.S., Gooseff, M.N., Voltz, T.J., Fitzgerald, M., Singha, K., and Zarnetske, J.P., 2013, How does rapidly changing discharge during storm events affect transient storage and channel water balance in a headwater mountain stream? Water Resources Research, v. 49, p. 5473–5486, doi: 10.1002/wrcr.20434.

Ward, A.S., Schmadel, N.M., and Wondzell, S.M., 2018, Simulation of dynamic expansion, contraction, and connectivity in a mountain stream network: Advances in Water Resources, v. 114, p. 64–82, doi: 10.1016/j.advwatres.2018.01.018.

Welhan, J.A., 2006, Water Balance and Pumping Capacity of the Lower Portneuf River Valley Aquifer, Bannock County, Idaho

Whiting, J., 2015, Understanding Tree Water Use Across the Snow-Rain Transition in Idaho's Mountain Watersheds: Interactions Between Stream Networks, Transpiration, and Basin Geomorphology. [M.S. Thesis] Pocatello, Idaho State University, 122 p.

Winter, T.C., Harvey, J.W., Franke, O.L., and Alley, W.M., 1998, Ground water and surface water: A single resource: USGS Publications, p. 79, doi: 10.3389/fpsyg.2012.00044.

Zimmer, M.A., and McGlynn, B.L., 2017a, Ephemeral and intermittent runoff generation processes in a low relief, highly weathered catchment: Water Resources Research, v. 53, p. 7055–7077, doi: 10.1002/2016WR019742.

Zimmer, M.A., and McGlynn, B.L., 2017b, Bidirectional stream–groundwater flow in response to ephemeral and intermittent streamflow and groundwater seasonality: Hydrological Processes, v. 31, p. 3871–3880, doi: 10.1002/hyp.11301.

Zimmer, M.A., and McGlynn, B.L., 2018, Lateral, vertical, and longitudinal source area connectivity drive runoff and carbon export across watershed scales: Water Resources Research, p. 1576–1598, doi: 10.1002/2017WR021718.

## Chapter 3: How do we monitor and model intermittent systems and how can we improve?

### 3.1 Coordinated interdisciplinary projects in intermittent networks

Incorporation of intermittent rivers and ephemeral stream (IRES) into river science is hindered due to lack of data about these systems (Datry et al., 2016). Most conceptual developments in the study of rivers are based on perennial rivers, which may not be applicable to IRES systems. We are learning more about the ecological, economic and social values associated with IRES, but still know very little about their sensitivity to land use and climate change (Steward et al., 2012; Acuña et al., 2014; Datry et al., 2014). Over the last decade, a number of collaborative groups have started working to further knowledge of intermittent river networks. These collaborative groups have similar goals, which include improving estimates of the spatial extent of IRES, understanding of how they function, and using this data to improve water management.

The 1000 Intermittent Rivers Project (1000IRP) covers a global scale, with researchers in 28 countries that sample IRES and share their data (The 1000 Intermittent Rivers project, accessed October 2018). Most of the sampling for the 1000IRP has been conducted in North America, Australia, and Europe, with very few samples from Africa and Asia. The 1000IRP have taken a broad approach, hoping "to build an international network of researchers dedicated to IRES, to support, complement, and federate current and future international projects on IRES…as well as global science network initiatives" (Datry et al. 2016). In general, the 1000IRP chooses a topic for participants to investigate, and communicates the data that is needed and common sampling methods. For example, in 2015-2016, 1000IRP focused their efforts on the role of river networks in the global carbon cycle, requesting quantification of organic material in streambeds, biodegradability of the organic material, and reactivity of rewetted streambed, as well as compounding environmental variables such as climate and riparian cover (Datry et al., 2016). This large-scale approach used by the 1000IRP encourages unified research and promotes IRES-focused collaborations at smaller scales.

The Science and Management of Intermittent Rivers and Ephemeral Streams (SMIRES) project was funded by the European Cooperation in Science & Technology in 2015 (SMIRES, accessed October 2018). The project's 4-year plan encompasses fourteen European countries. It outlines four major objectives associated with IRES research: 1) locate, map and predict IRES in river networks and distinguish between natural and anthropogenic IRES, 2) understand how alterations in flow regimes impair IRES biodiversity, functions and services, and how to define environmental flows in IRES, 3) model carbon and nutrient dynamics at the catchment scale, and 4) disentangle the effect of flow intermittence on river communities from other stressors (Datry et al., 2017). The SMIRES team is developing a conceptual framework for IRES, identifying risk areas for increased flow intermittence, and linking researchers with stakeholders to improve management and conservation (Datry et al., 2017).

The Intermittent River Biodiversity Analysis and Synthesis (IRBAS) project involves a collaborative research team studying intermittent river ecology, hydrology and modelling (IRBAS, accessed October 2018). They analyze available data to estimate abundance and spatial extent of intermittent rivers and analyze temporal trends to investigate relationships between flow regime, habitat dynamics and biodiversity in order to predict future responses to increased intermittency. Ultimately, IRBAS' objective is to help implement the management and restoration of intermittent rivers, and to raise awareness about their importance.

The DryRiversRCN is a new (as of July 2018) NSF-funded project that aims to synthesize current knowledge on hydrology and ecology of stream and rivers (National Science

Foundation, Award Abstract #1754389, accessed October 2018). Workgroups will be formed each year from 2019-2021 to create general frameworks about how intermittent systems work. The goals of this project are to characterize and predict intermittent flow regimes, better understand structure and function of intermittent systems, and integrate hydrologic controls into current frameworks built from perennial river networks.

TRivers is a project initiated by the University of Barcelona, with an emphasis on providing tools to water managers (Life TRivers, accessed October 2018). They been developing a software tool TREHS (Temporary Rivers' Ecological and Hydrological Status) to help implement the Water Framework Directive in temporary rivers. Their mobile app can be used by members of the general public to document hydrological and ecological dynamics of intermittent rivers for the benefit of water managers. With more data, the predictive capacity of the tool will be able to detect if the river is naturally intermittent, or if anthropogenic factors caused intermittency. This predictive modeling will improve the determination of stream ecological status by targeting appropriate sampling dates and methodology, rather than relying on models from perennial streams and rivers.

Two new citizen science projects active since 2017, CrowdWater and Stream Tracker, also involve using a mobile app to collect data on streamflow (Kampf et al., 2018). Stream Tracker is funded by NASA and relies on the general public to help collect data on the presence or absence of flow. This app is limited to four categories: flow, no flow (dry or damp channel), standing water (ponded or pooled disconnections), and channel covered (Stream Tracker, accessed October 2018). CrowdWater is funded by the Swiss National Science Foundation in order to improve hydrologic forecasting by using citizen scientists to create times-series of data that include water level, streamflow, soil moisture and flow condition. Rather than simply collecting data on the presence or absence of flow, the degree of intermittency is broken down into six categories that include flowing water, trickling water, standing water, isolated pools, wet streambed and dry streambed (CrowdWater, accessed October 2018).

The numerous recent collaborative research efforts focusing on intermittent streams reflect the increasing recognition of the prevalence and importance of these ecosystems. Intermittency is a global phenomenon (Acuña et al., 2014), and thus collaborative projects and crowd-sourced data will be critical to tackling these issues on a large scale.

### 3.2 Recommendations for improving intermittent research

Because intermittent streams are unique systems that may not fit into the current conceptual models of stream hydrology and ecology (Steward et al., 2012), new conceptual models from these efforts are critically important. For example, results from this thesis emphasize the importance of subsurface characterization in modeling intermittency. With these results in mind, I suggest focusing future research on the following topics.

I propose incorporating geophysical information into hydrologic models of stream drying. For example, more data on the thickness of the alluvium, soil depths, and hydraulic conductivity would provide valuable information on storage space and potentially preferential subsurface flow paths. While this approach is labor intensive and can be expensive, it would provide the most detailed characterization of subsurface heterogeneity. I suggest targeting known intermittent sites that already have geophysical data, such as some long-term ecological research sites, Critical Zone Observatories or experimental forests, and incorporating this data into current modeling efforts. By incorporating subsurface heterogeneity into models, hydrologists may better constrain locations of stream drying on a larger scale.

Differentiating between shallow and deep groundwater flow paths would also be useful for a process-based understanding of intermittency. Although my research highlighted the importance of shallow flow paths for streamflow permanence, this may not be the case in all environments. Future studies might monitor water levels in the shallow subsurface in conjunction with deep groundwater to ascertain how these dynamics interplay to produce surface flow expression in different environments. Water age dating using tracers could also provide insight into relative water contributions from deep groundwater and shallow surface water.

Although flow paths and water sourcing are key next steps for furthering a mechanistic understanding of intermittency, it is clear that some agencies have a purely predictive interest in stream drying. For example, water resource managers may find that a process-based understanding is secondary to predicting and identifying intermittent reaches. This thesis suggests that comparing discharge and drainage area might be a powerful predictive tool for locating stream reaches more or less prone to drying. Further studies of this relationship are needed to determine if patterns are consistent for streams of different sizes and in different climatic and geologic settings, where dominant controls on intermittency may be different. This would require spatially distributed discharge measurements throughout a given stream network, as well as accurate estimates of upslope accumulated area at each discharge measurement point. If comparative studies validate these findings, this approach could be used by water managers to identify streams prone to drying and may ultimately affect water-use regulations.

More research in these areas will advance our understanding of intermittency as well as enhance our ability to predict intermittency. In particular, including these data in coordinated, collaborative research endeavors will inform both water conservation and water management entities on the timing and magnitude of intermittency in their region. For cases of human-caused intermittency, the next steps will be assessing risks associated with intermittency from an ecological and ecosystem service perspective and developing site-specific strategies to protect and conserve these dynamic ecosystems.

### 3.3 References


Acuña, V., Datry, T., Marshall, J., Barcelo, D., Dahm, C.N., Ginebreda, A., McGregor, G., Sabater, S., Tockner, K., and Palmer, M.A., 2014, Why Should We Care About Temporary Waterways? Science, v. 343, p. 1080–1081, doi: 10.1126/science.1246666.

CrowdWater: https://crowdwater.ch/en/welcome-to-crowdwater/. (accessed October 2018).

Datry, T., Larned, S.T., and Tockner, K., 2014, Intermittent rivers: A challenge for freshwater ecology: BioScience, v. 64, p. 229–235, doi: 10.1093/biosci/bit027.

Datry, T., Corti, R., Foulquier, A., von Schiller, D., and Tockner, K., 2016, One for all, all for one: A global river research network, Eos, 97, https://doi.org/10.1029/2016EO053587. (accessed October 15, 2018).

Datry, T., Singer, G., Sauquet, E., Jorda-Capdevilla, D., von Schiller, D., Stubbington, R., Magand, C., Pařil, P., Miliša, M., Acuña, V., Alves, M., Augeard, B., Brunke, M., Cid, N., Csabai, Z., England, J., Froebrich, J., Koundouri, P., Lamouroux, N., Martí, E., Morais, M., Munné, A., Mutz, M., Pesic, V., Previšić, A., Reynaud, A., Robinson, C., Sadler, J., Skoulikidis, N., Terrier, B., Tockner, K., Vesely, D., and Zoppini, A., 2017, Science and management of intermittent rivers and ephemeral streams (SMIRES). Research Ideas and Outcomes 3: e21774. https://doi.org/10.3897/rio.3.e21774 (accessed October 16, 2018).

IRBAS: http://irbas.cesab.org/irbas. (accessed October 2018).

Kampf, S., Strobl, B., Hammond, J., Anenberg, A., Etter, S., Martin, C., Puntenney-Desmond, K., Seibert, J., and van Meerveld, I., 2018, Testing the waters: Mobile apps for crowdsourced streamflow data, *Eos, 99,* doi: 10.1029/2018EO096355.

Life TRivers: http://www.lifetrivers.eu. (accessed October 2018)

National Science Foundation: Awards: Intermittent river research coordination network (IRRCN): Integrating intermittent river ecology and hydrology: https://www.nsf.gov/awardsearch/showAward?AWD_ID=1754389&HistoricalAwards=false (accessed October 2018).

SMIRES: Science and management of intermittent rivers and ephemeral streams: http://www.smires.eu. (accessed October 2018).

StreamTracker: Community Powered Stream Monitoring: https://www.streamtracker.org. (accessed October 2018).

Steward, A.L., von Schiller, D., Tockner, K., Marshall, J.C., and Bunn, S.E., 2012, When the river runs dry: Human and ecological values of dry riverbeds: Frontiers in Ecology and the Environment, v. 10, p. 202–209, doi: 10.1890/110136.

The 1000 intermittent rivers project: https://1000_intermittent_rivers_project.irstea.fr. (accessed October 2018).

## Appendix A: Uncertainty calculations for the Hvorslev method

Following Rushlow and Godsey (2017), we first calculated uncertainty in the slope of the exponential fit,

$$s_\beta = \frac{\beta}{\sqrt{n-2}} \times \sqrt{\frac{1}{\rho^2} - 1} \qquad (Eq. A1)$$

where $s_\beta$ is the standard error of the slope of the exponential regression $\beta$, $n$ is the number of water level measurements, and $\rho$ is the Pearson correlation coefficient. The maximum and minimum intercepts can then be calculated,

$$\alpha_{max} = \alpha + (\beta - s_\beta) \times \bar{t}, \alpha_{min} = \alpha + (\beta + s_\beta) \times \bar{t} \qquad (Eq. A2)$$

where $\alpha_{max}$ and $\alpha_{min}$ are the maximum and minimum intercepts, $\alpha$ is the intercept as given by the fitted exponential regression, and $\bar{t}$ is the mean natural log of the time in seconds. Uncertainty in the intercept $a$ can then be determined,

$$s_\alpha = \frac{[(\alpha_{max} - \alpha) + (\alpha_{min} - \alpha)]}{2} \qquad (Eq. A3)$$

where $s_\alpha$ is uncertainty in the intercept $\alpha$ of the exponential regression. Gaussian error propagation was then employed to calculate the uncertainty in $t$, which was purposely selected to be $t_{37}$ in accordance with the Hvorslev method. The Gaussian error propagation rule is used for functions with uncorrelated variables. For example, $t_{37} = f(\alpha, \beta)$, and error is propagated as,

$$s_{t37} = \sqrt{\left(\frac{\partial t_{37}}{\partial \alpha} x\, s_a\right)^2 + \left(\frac{\partial t_{37}}{\partial \beta} x\, s_\beta\right)^2} \qquad (Eq. A4)$$

where $s_{t37}$ is the uncertainty in $t_{37}$, $\frac{\partial t_{37}}{\partial \alpha}$ is the partial derivative of $t_{37}$ with respect to α, and $\frac{\partial t_{37}}{\partial \beta}$ is the partial derivative of $t_{37}$ with respect to β.

## Appendix B: Love letter to a future graduate student

Here I have compiled a list of things I wish I had known in the beginning of my graduate school journey. Some items are project-specific, while others apply more broadly.

### Getting a head start

***Dive into reading papers.*** Your schedule is going to be busy, particularly during the first year when you take most of your classes. Be sure to set aside specific time each week to read papers related to your work, otherwise it likely won't happen. You may already know a great deal about hydrology in a broad sense, but your project is going to be very specific. Get acquainted with the current research. Note how others have formulated their questions and the methods they used to tackle those questions. Find some way to organize your thoughts about these papers that works for you, whether it be an annotated bibliography, notes in Mendeley (or some other reference manager), or color-coded highlighting. Take useful notes that you will easily understand months later. Re-read papers you find most important.

***Start thinking about specific questions you want to answer early on.*** Draw simple figures of your anticipated results, decide if they are meaningful, and determine how to collect the data needed to assess the results. I found this to be particularly challenging, but also the most helpful in formulating my project. Expect to iterate on this. Do some preliminary field work to see if your methods are feasible and practical at your site.

***Learn what software will best suit the data you'll be collecting.*** If you already are familiar with this software, great! If not, get a jump-start playing around in it, potentially even before you collect any of your own data. This will save you a great deal of time later on when you're doing analysis.

## Field Work

***Prep your field gear ahead of time.*** If I was heading out to the field early in the morning, I would lay out my gear the night before. This greatly reduced my risk of forgetting equipment in the rush of getting out to the field in the morning. Check the gear; make sure things are charged and your waders don't have too many holes. If you're frequently going to the field to perform routine tasks (e.g. downloading data, running a specific test), make a list of the equipment needed for each task so you can quickly and completely assemble your gear. Importantly, if you share the lab with any other students who are currently doing fieldwork, make sure to communicate the gear you want to use in the field, particularly if you're using something out of the ordinary. You may have to compromise and adjust your schedule if you both need a particular tool. Side note: Keep the sed lab organized so others can use the space.

***Make a field kit.*** For my fieldwork, I had a collection of items I regularly used that basically lived in my field bag. For my field kit, I included things like duct tape, scissors, screwdrivers, pliers, a wire brush, my field notebook, flagging, multiple pencils, a hobo shuttle and adapters. This will vary depending upon your work.

***Things take longer than you think in the field.*** Don't procrastinate field work. Many times I thought my tasks for the field would only take a half-day, but ended up being a full day. Prioritize your field tasks so you accomplish what is necessary.

***Winterize your site while it's easy, BEFORE things freeze.*** I had dataloggers in my study stream and waited too long to remove them both winters during my project. (If working in Gibson Jack, remove equipment before November!) Thankfully the loggers didn't freeze, but the surface of the stream was frozen thick enough in places that I couldn't break through without extra equipment. If this happens, I recommend using a backpacking stove to heat water to melt the ice, and an ice axe to break through. Be patient. In my experience, I was able to retrieve all of the loggers, but it took some time. I know it’s tempting to leave the loggers running for as long as possible to get more data, but the reality is that much of that data may not be useful. Water level and electrical conductivity measurements may not be accurate when there’s a solid layer of ice on top of the stream. Avoid the unnecessary stress.

***Use a standard naming system for installed equipment.*** I had to rename locations for my equipment partway through my research to help them be more explanatory. For example, don’t name loggers 1,2,3 etc., and instead use the distance from your designated outlet (e.g. 100m). I initially named by transects 1-4, but found it was confusing to audiences, and thus named them based on location and flow status. The switch can be difficult when you already have one naming system burned into your brain, so do it right the first time!

***Write out protocols.*** In particular, I did this for calibration methods. This was especially useful when I had interns. I could send them the protocol document and they could replicate my methods. In order for this to work, however, you must be very explicit in your descriptions. This was also helpful later on if I had to recalibrate something and didn’t quite remember the steps.

***Be aware of wildlife at your field site.*** I had a number of spooky run-ins with moose in the Gibson Jack watershed. Know what wildlife you may encounter at your field site, and when they are more active or likely to be more aggressive. Be prepared accordingly.

***Listen to locals and share your research with them.*** People who live around your field site or have worked there for years have a wealth of knowledge to share. They might tell you that your stream used to have a ton of beaver dams (I learned that from a local on the trail who had done his master’s degree at my field site back in the day). They also might tell you that you’re stream doesn’t dry, even though you’ve seen it do so with your own eyes. Keep an open mind, but take what they say with a grain of salt. Particularly if you're working on public land, people will be wondering why you're hauling waders and buckets and other equipment along the trail. You may have a lot of work to do that day, but be patient and take the time to talk to them and tell them what you're doing. In my experience, most people are genuinely curious and supportive of your work.

**Data Management**

***Backup your data.*** Put it on an external hard drive, save it to google drive, anything to have multiple copies of your data. Although I was lucky and didn't have any issues with data loss, there are plenty of horror stories floating around the department about computers crashing before students backed up their data (ask Sarah if you don't believe me). Don't be another story.

***Use a standard naming system for downloaded data.*** This will make it much easier to assess at a moment's notice, such as during a meeting. I used the site name and date downloaded. Organize the data into folders. Folders are your friends.

***Update figures, tables, time series etc. as you go.*** Don't wait until the end to start analyzing your data. Updating your visuals each time you download data allows you to detect if something isn't working and needs to be replaced. Also, you may see some patterns that you want to explore more, and may lead you to change locations of or install more instrumentation.

***Keep a running list of the data you have.*** Maybe you initially collected data for one purpose, but later realize you can apply it in another way with other data that you're collecting. This was the case with my salt dilution gauging, where I initially just wanted data on hydrologic loss and gain. Later, I realized that within this data I had spatially distributed discharge measurements throughout my reach, and could compare this with drainage area. I don't think I would have thought to look into this if I didn't have my data summed up clearly in a list. In summary, keep a list of your data, and leave yourself time to play around with it, without a specific goal in mind. You might find something interesting!

**Interns**

Working with interns is a great opportunity to develop your mentorship skills. It will teach you a lot about your own working style and how you work with other people. Mentorship experience will be a great asset on your resume later on, so take advantage of this opportunity.

***Be friendly and accessible.*** In some cases, your interns may have had no experience in hydrology or even in the field. Although you are in a supervisory position, you are also a student and they may feel more comfortable asking you questions than Sarah. Be supportive and encourage them to ask any and all questions. They will likely do a better job when they have a better understanding of their work and why you're doing your project.

***Communicate with your interns often.*** Ask about their projects often, and any challenges they're facing. It encourages them to think critically about their work, and summarize what they're doing. Help them with their projects. Ask for their help when you need it. Coordinate your field tasks to be sure that neither of you are unknowingly impacting the others work (e.g. someone is standing downstream during a salt injection).

***Remember your interns are adults.*** This may sound silly, but I've seen some mentors be condescending to interns. You will be more familiar with your field site and probably some of the methods they'll be using, but be encouraging and don't talk down to them. Encourage them to try and solve their own problems, while also letting them know that you're happy to help them. I would often provide suggestions for how they should approach something or be more efficient, but I did not micro-manage them and let them learn on their own. I think setting explicit goals and expectations early on is a good way to do this. If they procrastinate fieldwork in the beginning, they'll just have to put in some long days and weekends later on to get the job done.

***Listen to your interns.*** They have great ideas. Having a new perspective on your work can be very helpful, especially when you've become accustomed to doing things a particular way. I loved troubleshooting problems with my interns, and just having someone to bounce ideas off of.

**General suggestions for making your life easier**

**Prepare for weekly meetings with your advisor.** Don't walk into that meeting without a plan. You likely have a ton of things to discuss, and your advisor is busy. Use the hour you have efficiently. Prioritize things you want to discuss, because that hour will fly by.

***Use your resources.*** Don't struggle endlessly when something doesn't work. Talk to other students, especially in your lab. They may provide some useful insight. Don't be afraid to ask other faculty for help. Initially I felt uncomfortable asking other faculty for equipment or advice since I didn't know them well. The entire geoscience department is friendly and supportive, and in my experience, more than willing to help any student. If you're having issues with funding or travel, go ask the front office. Use the resources you have available.

***Take breaks, or risk burning out.*** You can't work 24/7 for the next 2-4 years. Work hard and take breaks. Go to the farmer's market, first Friday Art Walk, contra dance, hot springs, or the used bookstore downtown. Take advantage of free movies at Bengal theater. Go running and mountain biking on the trails, go skiing at Pebble, or go to open climb. We're close to so many wonderful public lands. Go to Yellowstone, Glacier, and the Sawtooths. Go to colloquium, and go to lunch with guest speakers. Attend department events. Go on seminar trips when you can, you'll see and learn incredible things. When other students need help with their research, volunteer. Have potlucks with the other graduate students.

Being a graduate student has been difficult and rewarding, and I've had some of the best times of my life here. Don't miss out.