Hydrologic connectivity and implications for ecosystem processes - Lessons from naked watersheds

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Abstract

Hydrologic connectivity has received great attention recently as our conceptual models of watersheds and water quality have evolved in the past several decades. However, the structural complexity of most temperate watersheds (i.e. connections among shallow soils, deep aquifers, the atmosphere and streams) and the dynamic seasonal changes that occur within them (i.e., plant senescence which impacts evapotranspiration) create significant challenges to characterizing or quantifying hydrologic connectivity. The McMurdo Dry Valleys, a polar desert in Antarctica, provide a unique opportunity to study hydrologic connectivity because there is no vegetative cover (and therefore no transpiration), and no deep aquifers connected to surface soils or streams. Glacier melt provides stream flow to well-established channels and closed-basin, ice-covered lakes on the valley floor. Streams are also connected to shallow hyporheic zones along their lengths, which are bounded at ~75 cm depth by ice-cemented permafrost. These hydrologic features and connections provide water for and underpin biological communities. Hence, exchange of water among them provides a vector for exchange of energy and dissolved solutes. Connectivity is dynamic on timescales of a day to a flow season (6-12 weeks), as streamflow varies over these timescales. The timescales over which these connections occur is also dynamic. Exchanges between streams and hyporheic zones, for example, have been estimated to be as short as hours to as long as several weeks. These exchanges have significant implications for the biogeochemistry of these systems and the biotic communities in each feature. Here we evaluate the lessons we can learn about hydrologic connectivity in the MDV watersheds that are simplified in the context of processes occurring and water reservoirs included in the landscape, yet are sensitive to climate controls and contain substantial physical

38 heterogeneity. We specifically explore several metrics that are simple and/or commonly
39 employed in hydrologic analyses and interpret them in the context of connectivity between
40 and among hydrologic features.

1.0 Introduction

Hydrologic connectivity is the concept that hydrologic processes (i.e., the natural movement of water in the environment) provide a vector for transfer of mass and energy between (or even among) different reservoirs (water bodies, aquifers) and/or locations (soils, atmosphere, vegetation) across a landscape (Bracken and Croke, 2007; Pringle, 2003). Whereas the connectivity may be described in general fluxes, e.g., the atmosphere is *connected* to aquifers by infiltration of precipitation/snow melt through soils, the application of this concept is often more distinctly focused, for example on whether (or when) discrete parts of a landscape or water bodies may have flow paths (surface or subsurface) that connect it to a nearby stream (Jencso et al., 2010, 2009). Topographic gradients often provide a hint that such flow paths are possible if water bodies are in the same basin (i.e., no significant topographic or geologic divides between them). Connections may be very dynamic through time as conditions in the water bodies and the land in between them change. Hydrologic connections may only occur, for example, during seasonal snowmelt conditions, or only during rare wet conditions (e.g., Bunn et al., 2006). Perhaps the most significant reason to quantify the spatial and temporal dynamics of hydrologic connectivity is that they may have a substantial influence on the water quality or ecosystem conditions of one of the bodies of water (Jaeger et al., 2014). Hydrologic connectivity is also a framework that is receiving attention from policy makers seeking to determine how water bodies may be connected (especially those without obvious surface connections) and the implications of connections for water quality. In the United States, a recent evaluation of state of the science on physical, chemical, and biological connectivity among water bodies was conducted by the Environmental Protection Agency (US EPA,

2015), in part to determine whether and how the jurisdiction of the Clean Water Act could be revised.

In most watersheds, teasing apart the spatial and temporal connections among water bodies and/or parts of the landscape can be a daunting challenge because of the potential to have many dynamic flowpaths converging at specific locations (particularly when considering stream flow signals). Many connections are in the subsurface, invisible to the naked eye without the use of tracers and hydrometric data collection. Several studies have explored these past efforts, and some have provided trajectories for future research and progress toward better quantification of hydrologic connectivity (e.g., Bracken et al., 2013). In this paper we review several fundamental concepts of hydrologic connectivity and demonstrate how we can evaluate them in the polar desert landscape of the McMurdo Dry Valleys (MDVs), Antarctica – an apparently simplified watershed setting. The MDVs are underlain by permafrost, so in general, surface hydrology is not connected to deep aquifers (though potential deep brine aquifers exist based on recent findings by Mikuki et al. (2015); in addition, Don Juan Pond is a highly saline water body that is connected to an aquifer of ~20 m depth). The MDV surface hydrology is similar to many tundra watersheds across the Arctic, also underlain by continuous permafrost, however, most Arctic tundra watersheds (e.g., north slope of Alaska) flow through landscapes with extensive vegetation and much of the Arctic experiences rainfall in the summer months. The dry (3-50 mm SWE; Fountain et al., 2010) and cold (mean annual air temperature of -18°C) conditions do not support vascular plants, so there is no transpiration that occurs. Thus the hydrologic cycle of the MDVs is simplified with fewer processes occurring than in temperate and even Arctic watersheds. However, the MDV landscape hosts substantial

heterogeneity in composition (geomorphic form, aeolian transport and deposition) and substrate (soil/sediment distributions and influence of paleolakes in the valley floors). Thus, one can view the MDVs as a natural laboratory for investigating hydrologic processes (that occur) and their impacts on biogeochemical cycling and/or ecosystem processes.

Previous studies have explored the hydrologic connectivity of the MDVs from several angles. Gooseff et al. (2011) focus on the different processes that provide hydrologic connectivity in the MDVs, and Wlostowski et al. (2015a) provide a synthesis of streamflow dynamics from 20+ years of record in the context of dynamic connectivity on seasonal and interannual timescales. Here we provide an analysis of the different metrics that can be used to define hydrologic connectivity across this system which may be a useful example for application to temperate systems.

2.0 Regional Setting

The MDVs are located on the western edge of the Ross Sea, an end of the Transantarctic Mountains (Fig. 1) and represent the largest ice-free portion of Antarctica, 4500 km² (Levy, 2013). The landscape is dominated by mountain and piedmont glaciers, exposed soils, stream channels, and ice-covered, closed basin lakes on the valley floors. In the case of the Lake Fryxell basin in Taylor Valley, stream flow is generated by alpine glaciers that have source areas in the Asgard Mountains or the Kukri Hills. The two largest glaciers extend down to and across part of the valley floor (Canada and Commonwealth Glaciers). Streams connect glaciers and closed-basin lakes (except for Commonwealth Stream, which flows to the ocean). Streams are generally 1st or 2nd order and all occupy a single-channel (no braided channels). Because all streamflow is glacial melt, contributing

areas to the streams are glacial, and dynamic on daily and sub-daily basis, depending on solar aspect, air temperature (adiabatic lapse rates), etc. (McKnight et al., 1999).

Fig. 1. Inset location map of the McMurdo Dry Valleys at 78°S, 163°E, and map of Taylor Valley with the three major lake basins. Streams are labeled: A = Aiken Creek, C = Canada Stream, Cr = Crescent Stream, D = Delta Stream, G = Green Creek, H = Huey Creek, LS = Lost Seal Stream, VG = Von Guerard Stream, Com = Commonwealth Stream, Hou = House Stream, and And = Andersen Stream.

Stream flow records from streams in Taylor Valley began in the early 1990s and have been collected continuously since (the one exception is the 1992-93 season in which there was no deployment of the research team, and therefore no records collected). Stream control structures were constructed with the assistance of US Geological Survey hydrologists. Stream gauges measure stream stage, electrical conductivity, and temperature every 15 minutes. Most are outfitted with H-flumes to efficiently pass water and provide a simple cross section in which to make stage measurements. Stream discharges are measured regularly through the austral summer (through late January) to build, maintain, and adjust (if needed) rating curves for these gauges using standard US Geological Survey field and data processing protocols.

2.1 Hydrologic Processes in the McMurdo Dry Valleys

Precipitation is all in the form of snow, and very little falls annually (Fountain et al., 2010). Snow collects in small drifts and patches across the landscape, ablating during the austral summer (Eveland et al., 2013a, b). In the austral summer, snow events may blanket the landscape to a depth of a few cm, but the snow quickly ablates (mostly sublimates) within hours of the end of the storm. Hence, snow is not a significant contributor to stream flow in this system. However, melting snow patches can provide some runoff that moves downslope, often generating water tracks (zero order conduits of mostly sub-surface water moving downslope) (Levy et al., 2011). These water tracks serve to connect some locations of annual snow patch accumulation with downslope locations, in some cases terminating into lakes. Water tracks transport water mostly through the active layer (less than fully saturated through the full depth) with little if any water at the surface (i.e., no real channelized flow at the surface). Water tracks have not been observed to intersect many streams. Stream flow is generated from melt on glaciers. In fact, we have observed that snow events tend to decrease meltwater production because the new snow, though not particularly deep, has a higher albedo than the underlying glacier ice. Hence, the resumption of stream flow generally occurs after the snow has ablated from the glaciers.

Stream flow has two typical temporal patterns – seasonal and daily. Over an austral summer season, stream flow generally starts with low flows for a few weeks, and then ramps up to high flows during the middle of the summer (typically around the new year), and toward the end of the season stream flow decreases again (Fig. 2 A, B). Over the course of a day, stream flow increases and decreases as the sun angle on the glaciers changes causing a daily flood, which has fairly consistent timing (Fig. 2 C, D). The example

in hydrographs from Delta and Canada Streams from the 2008-09 season in Figure 2 demonstrate that stream flow quantities are similar in both short (Canada, 1.5 km) and long (Delta, 8.0 km) streams, which suggests that the maximum transfer of mass and energy from glaciers to streams to lakes in a given season may be similar for both long and short streams. However, the strength of connection (indicated by the magnitude of discharge) varies within a day and across the season in similar ways for all streams. The magnitude and timing of meltwater generation and hydraulic transport from glaciers to stream gauges is related to source glacier aspect, melt area contribution on the glacier, and channel hydraulic characteristics (Conovitz et al., 1998).

Fig. 2. Stream discharge and electrical conductivity records observed at the Canada (A and C) and Delta (B and D) stream gauges (where C and D are subsets of the time series displayed in A and B). Delta stream is 8 km long and Canada Stream is 1.5 km long.

Most MDV streams terminate in closed basin lakes on the valley floors, increasing lake water volumes and levels. Lake levels (Figure 3) are maintained by these seasonal inputs, which are balanced by ice-cover ablation and evaporation from the open waters that form around the edge of the lakes (i.e., ‘moats’) during the austral summer. During the austral winter ice covers thicken from liquid water at the top of the lake freezing on to the bottom of the ice cover. Increasing lake levels since 1972 suggest a net positive energy balance that is producing more melt during the summer than is lost to sublimation of the ice-cover (year-round) and evaporation of open-water moats (during the summer). The decline in lake levels from the early 1990s to the early 2000s is consistent with the cooling trend in mean summer air temperatures observed during that time period (Doran et al., 2002).

Fig 3. Lake level changes since 1972 for closed-basin, ice-covered Lakes Fryxell, Hoare and Bonney (in Taylor Valley), and inset of Lake Hoare from 2006 to 2008. Manual survey measurements have been made 1-2 times per austral summer since 1972, and more recently, automatic data collection methods (pressure transducers anchored to the lake bed, connected to data loggers) have been deployed to collect lake level data. Figure from Gooseff et al. (2011).

Connections of landscapes, streams, and lakes to the atmosphere are manifest in the evaporation liquid water (from soils, streams, and lakes when available during the austral summer) and sublimation of ice and snow year-round. These fluxes have not been regularly measured. However, evidence is provided in isotopic enrichment from riparian soil water (Northcott et al., 2009), and from mass balance estimates of the lakes (Dugan et al., 2013).

3.0 Hydrologic Connectivity in a Polar Desert

The stream gauges are located near the mouths of the streams to enhance our ability to conduct mass balance estimates for the lakes. Hence, observed streamflow at a gauge indicates a connection from the glacier to the entire length of the stream channel, and on to the lake as well. We further expect that when stream flow occurs in the channel, the streams are connected to their surrounding hyporheic zones as well. In any given flow season, as meltwater is produced, and supplied to the channel, the thawed sediments that make up the hyporheic zones around the streams provide an accommodation space that must be filled for water to remain in and moving down the channel. In particularly cool or cloudy years, when little glacial melt water is produced, stream flow may occur in the upper reaches of the stream but not reach the stream gauge. Hence, the glacier and part of the stream are connected, but the stream is not connected to the lake.

We propose the use of several simple metrics of hydrologic record analysis to characterize hydrologic connectivity between glaciers and streams, and streams and lakes in the MDVs. Analyses of flow volumes and their variability from year to year provide an assessment of the strength of connection as this is representative of the total mass

transferred from glaciers to streams to lakes (and the water mass includes dissolved constituents and heat). Frequency analyses provide an assessment of mass flux characteristics (max, median, min) and the ability to compare across specific connection pathways (i.e., specific glacier to a specific stream to the lake). Simple no-flow analyses provide a characterization of how often connections are broken for at least part of the glacier-stream-lake continuum.

The simplest analysis of connectivity is an evaluation of the annual volumes of stream flow that have passed the stream gauge for each stream (Fig. 4A). These magnitudes and their comparison among streams in a given year, or across different seasons for the same stream are indicative of the magnitude of connection between the glacier, as the source of water, and the stream, and because the stream gauges are located near the lakes, the magnitude of the connectivity of the streams to the lakes is also provided with these annual flow volumes. In the Taylor Valley, the 2001-02 flow season was the highest on record and the 1994-95 was the lowest. Hence, hydrologic connectivity of the glaciers-streams-lakes systems was strongest in 2001-02 austral summer and weakest in the 1994-95 flow season. Two additional high flow seasons occurred recently (2008-09 and 2010-11) with relatively similar flow volumes. It is also evident that the magnitude of glacial melt has increased in the decade since the 2001-02 high flow season (compared to the decade prior), indicating a temporal trend of greater hydrologic connectivity in this polar desert system.

Fig. 4. Annual flow volumes for (A) each gauged stream that flows into Lake Fryxell, plus Commonwealth Stream, which flows into McMurdo Sound, but shares a common source with several of the Fryxell basin streams (the Commonwealth Glacier), and (B) the same data grouped by source for each year.

Another approach to evaluating connectivity is to consider the sources of meltwater generation to streams and lakes. Viewed this way for the annual streamflow volumes (Fig. 4B), Commonwealth Glacier is typically the greatest source, with Canada Glacier and the Kukri Hills glaciers contributing less and the Asgard Range glaciers (mainly the source for Huey Creek), contributing the least. Figure 5 presents an ordering of stream flow volume magnitudes for each flow season, coded by the source glacier. Each glacier may serve as

the source for multiple streams, but it is rare that streams have multiple glacier sources (exceptions include Aiken Creek and Crescent Stream). What is evident from this analysis is that the Commonwealth Glacier is often the leading source of stream water (also evident from Fig. 4B), with Canada Glacier streams similarly producing a lot of stream flow, but the small glaciers in the Kukri Hills tend to produce less meltwater. In addition, it is the shorter streams that transfer the largest volumes, and longer streams tend to have the lowest volumes of stream flow. Further, most of the glaciers are connected to streams (and therefore lakes) every season. Thus, the patterns of hydrologic connectivity are fairly consistent year to year. However, there is little information about length of time of connectivity in this analysis.

Fig. 5. Analysis of relative sources of stream flow to Lake Fryxell where source glaciers are color-coded and each cell of the matrix represents a different stream (A = Aiken, C = Canada, Cr = Crescent, D = Delta, G = Green, H = Huey, LS = Lost Seal, VG = Von Guerard, and Com = Commonwealth). The magnitudes of flow volumes for each flow season are arranged from most (top) to least (bottom). Names that are in bold font are long streams (>2 km), regular font is a short stream (<2 km), red font indicates no flow for the season,

and the hatched cells indicate no record due to gauge malfunction for a substantial portion of the season, but flow was observed to have occurred.

The streams and their sources can be further analyzed by calculating the runoff produced per unit area of the contributing glacier (Fig. 6). In most flow seasons, the contributing area of glaciers is somewhat dynamic, based on the air temperatures and the adiabatic lapse rate (Doran et al., 2008). We have used a standardized area to compute the contributing areas below 800 m elevation. From the Canada Glacier and Asgard Mountain source for Huey Creek, in general Green Creek derives more meltwater generation per unit area than Canada Stream (Fig. 6A). The source of Huey Creek is inconsistent, and very substantial in a few of the high flow seasons. This is likely due to its relatively high elevation; compared to the Commonwealth and Canada Glaciers, which reach the valley floor, the source of Huey Creek is perched several hundred meters above the valley floor. From the Commonwealth Glacier (Fig. 6B), Commonwealth Stream derives the most meltwater generation in each season. From the Kukri Hills Glaciers, (Fig. 6C), Delta Stream typically has the greatest runoff yield, though the annual meltwater volumes from these glaciers are sometimes under-represented by stream discharge at the gauge because all of these are long streams and likely lose meltwater to filling the hyporheic zone (early in the flow season), and evaporation (Gooseff and Lyons, 2007; Gooseff et al., 2003). Overall the Canada and Commonwealth Glaciers contribute the greatest runoff to streams in the Fryxell basin.

Fig. 6. Area normalized discharge for each stream for each flow season organized by sources (A) Canada Glacier and Asgard Mountains source, (B) Commonwealth Glacier, and (C) Kukri Hills Glaciers. Harnish Stream was left out of this analysis because it receives discharge from more than one Kukri Hills glacier.

Another simple and common hydrological analysis that can shed light on connectivity is that of flow exceedance probability. Discharge exceedance probability curves (or flow duration curves) quantify the proportion of time discharge equals or exceeds a given value as a cumulative exceedance probability. Here we present exceedance probabilities of our daily average flow records for the entire period of record at each stream gauge (Fig. 7), similar to the analysis of Wlostowski et al. (2016b). High (0.1 probability), median (0.5 probability), and low flows (0.9 probability) are, on average, greater in short streams than long streams (2.89 vs 2.13 m³/s; 0.27 vs. 0.15 m³/s; 0.02 vs. 0.01 m³/s, respectively). In addition more of the long streams have exceedance probabilities that are <1 for the lowest flows observed (~3 L/s), indicating greater zero flow periods. This analysis corroborates those presented in the annual volumetric flow and related source analyses (Figs 4 and 5), but at a finer temporal resolution. The exceedance probability analyses indicate that for any given moment, discharge is likely to be greater in shorter streams than longer streams, enhancing the potential for transfer of solute mass and energy carried by stream flow. However, shorter streams, by definition, have shorter hyporheic zones with which they interact. Hence there is an argument to be made for less overall spatial connectivity in short streams compared to long streams, despite higher flows and therefore greater potential vectors of connectivity in short streams. However, in long streams, there is ultimately a range of connectivity to the source glaciers. In any given year, upstream sections of the stream channel are more likely and more consistently receiving meltwater from the glacier than downstream locations. This is particularly true for low-flow years in which melt water may flow part way down the channel, but not reach the stream gauge or the lake. The stream gauge data, however, only provide information

about that single location in space, not about flow dynamics of specific segments above the gauge.

Fig. 6. Exceedance probability curves for A) long (> 2km) and B) short (<2 km) streams for flows recorded during the flow season only (winter dates have been omitted from analysis) over each stream gauge’s entire record.

Finally, another simple analysis that can be performed with the 15-minute flow data is a zero-flow analysis. While all of these streams are intermittent, flowing only for 6-12 weeks per year, the flow consistency during the typical flow season (late November – mid-February) is an indicator of harshness for the stream ecosystem – i.e., more inconsistent flow would create harsher conditions as stream algal mats and related diatom communities face alternating wetting and drying (Esposito et al., 2006), forcing biological functions to start and stop frequently, and reduced or interrupted nutrient delivery to mats. Furthermore, an analysis of the relationships between mat biomass and hydrologic regime in 16 stream reaches over a 20-yr period found that the number of days without flow during a flow season, referred to as “zero” days, was a significant predictor for the biomass of two of the three major mat types occurring in these streams (Kohler et al., 2015), and

that the sign of the relationship differed for reaches with abundant mats and those with sparse mats. In our analyses of zero flow periods we present an example from a short stream (Canada Stream, Fig. 8A) and a long stream (Von Gureard Stream, Fig. 8B). Canada Stream tends to initiate flow between mid November and early December, and then has fairly consistent flow until it ceases anywhere from late January to mid-February. In 9 of the flow seasons presented, there are no partial days of flow for Canada. Von Guerard Stream typically starts later than Canada Stream, and ends earlier too. In 8 of the flow seasons on Von Guerard Stream, there are no reported partial days of flow. However, flow seasons are generally shorter, a few no-flow seasons occur that did not occur on Canada Stream (1999-00 and 2000-01), and there are more periods of no flow that occur through the record. Hence, we can interpret these records as an indication of less persistent and weaker connectivity from glacier to stream to lake for long streams than for short streams.

Fig. 8. Zero flow analysis for all years of record for A) Canada Stream (0.7 km) and B) Von Guerard Stream (4.7 km). Each rectangle represents a single day and the color for each day indicates the amount of the day that had zero flow occur. Note that no flow data was collected for the 1992-93 season due to no field deployment (i.e., not necessarily a zero-flow season), but in the 1999-00 and 2000-01 flow seasons, no flow was observed at the gauge on Von Guerard Stream (i.e., zero flow seasons).

4.0 Stream Connection to Hyporheic Zones

A consequence of the uniqueness of the McMurdo Dry Valley landscape is that the streams are connected to shallow hyporheic zones under and adjacent to stream channels, without further connection to deeper aquifers or hillslope aquifers. Thaw depths in hyporheic zones reach ~75 cm (Conovitz et al., 2006; Northcott et al., 2009), which is greater than the active layer depths across most of the dry soils, which tend to be <50 cm (Bockheim et al., 2007). Streams actively exchange water between the channel and hyporheic zones, creating the potential for biogeochemical reactions in the streambed sediments. One consequence of exchanging dilute glacial meltwater (typically with electrical conductivity values of <20 $\mu\text{S}/\text{cm}$) through deposits of poorly weathered materials is that very high chemical weathering rates occur (Gooseff et al., 2002; Lyons et al., 1998; Nezat et al., 2001). In fact, the McMurdo Dry Valley streams have some of the highest silicate weathering rates reported (Fig. 9), despite the low temperatures of the system.

Fig. 9. Comparison of silicate weathering rates from temperate watersheds and McMurdo Dry Valley streams. These data are presented as a table in Gooseff et al. (2002).

Weathering of the hyporheic sediments yields increased ion concentrations in the stream water that returns to the channel. Hence, synoptic sampling indicates that streams generally increase concentrations of solutes downstream (Gooseff et al., 2011). The result of this hydrologic connection driving geochemical reactions is that the observed electrical conductivity records in longer streams have generally higher values than the short streams (Fig. 2). This provides an opportunity to develop end-member mixing models of flow contribution at the stream as glacier melt water has a very low electrical conductivity and hyporheic waters appear to have very high electrical conductivities (can be $>300 \mu\text{S}/\text{cm}$). Wlostowski et al. (2015b) have developed models for determining continuous hyporheic influence on streamflow, and we present one example of this modeling here (Fig. 10). Comparing a short stream (Fig. 10A) and a long stream (Fig. 10B), it is clear that there is less hyporehic influence in the shorter stream, because there is less overall hyporheic volume with which stream water may interact than in long streams. When considering these hyporheic interactions across seasons, Wlosowski et al. (2015b) found that longer streams generally had a greater proportion of streamflow go through hyporheic zones than short streams, for the period of record.

Fig. 10. Results of a simple 2-end member mixing analysis for A) Canada Stream, and B) Delta Stream, in which glacial meltwater (one end member) is simulated to have an EC of 15 $\mu\text{S}/\text{cm}$ and hyporheic waters (the other end member) are simulated to have the 314.1 $\mu\text{S}/\text{cm}$, which is the average 1% exceedance probability electrical conductivity value for all streams > 2km. These two end members are mixed to yield the observed EC timeseries at the stream gauge.(see Figure 2 for those data). Hyporheic proportion is an indication of how much of the hyporheic end member (i.e., water from the hyporheic zone or with the hyporheic signature) is present in the stream water at the gauge, given this simple mixing analysis through time.

The hydrologic connectivity between streams and hyporheic zones also facilitates transformation of dissolved nutrients in stream water in the MDVs. Combining surveys of stream and hyporheic waters, McKnight et al. (2004) demonstrated that algal mats in the streams and microbial demands in the hyporheic zone control inorganic nitrogen and phosphorous concentrations in MDV streams. Building on that, Gooseff et al. (2004) separated the nitrate reduction processes between the channel mats and the hyporheic zone and found that the hyporheic zone has less demand than the mats. The connection of

streams to hyporheic zones provides for some channel nitrate and phosphorous removal as water travels from glaciers to lakes.

Hyporheic exchange also buffers stream temperatures in MDV streams. Cozzetto et al. (2006) note that hyporheic exchange accounted for up to 21% of the non-radiative heat losses from streams. Further, thermal dynamics of both stream water and the hyporheic waters have been found to influence the rate of hyporehic exchange in MDV streams. Solute transport modeling from stream tracer injections conducted during times of the day when stream/hyporheic waters were cool and warm indicate that hyporheic exchange occurred more extensively when the water was cool than when it was warm (Cozzetto et al., 2013). This is a bit opposite of the expectations that warmer water would have reduced viscosity and therefore result in greater exchange than cold water. The implication of these findings is that the streamwater conditions (temperature in this case) have some control over the strength of connectivity between streams and hyporheic zones. The temporal dynamics of hyporheic exchange are a primary control on the potential for biogeochemical cycling in hyporheic zones.

Despite the simplicity of the shallow alluvial aquifers, hyporheic exchange occurs over multiple timescales in MDV streams. At the greatest extreme, Gooseff et al. (2003) used natural abundances of δD and $\delta^{18}O$ to track turnover of hyporheic water and found that residence times were on the order of weeks to months in distal hyporheic locations. However, solute transport modeling from stream tracer experiments results in mean residence times that are on the order of hours to days (Gooseff et al., 2004; McKnight et al., 2004; Runkel et al., 1998). During some stream tracer experiments, a few hyporheic sampling sites had very rapid arrival of stream water compared to others at the same

distance from the channel, indicating that preferential flow paths are prevalent in MDV hyporheic zones (Cozzetto et al., 2013). These combined results suggest that MDV hyporheic zones had a set of flow paths that were fairly 'fast' and another set that are much slower. Hence, the timescales of connection between streams and hyporheic zones are multiple as well, thereby influencing the timing of when the effects of hyporheic exchange are observed in the channel.

Given the dynamic nature of the MDV stream flow regimes (Fig. 2), it would be expected that perhaps the hydrodynamic connection of streams to hyporheic zones may also vary on a sub-daily basis. Koch et al. (2010) used end-member mixing models to characterize the temporal change in the signature of hyporheic return flows to an MDV stream during and after a stream tracer experiment that included nitrate and ammonium amendment. They found that nitrogen uptake varied greatly over a full cycle of discharge, indicating that connectivity and possibly reactivity varied throughout the day, coincident with varying discharge and return flow fluxes throughout the day. Groundwater flow modeling of hyporheic exchange during diel stream flow variation indicated that exchange flows varied from 0 to 190 L/hr/m length of channel (Koch et al., 2011), which is clear evidence of the variability in connectivity of streams and hyporheic zones over a diel flow cycle. The MDV stream-hyporheic systems, despite their simplicity (i.e., being disconnected from deep and/or hillslope aquifers), display natural heterogeneity in their make up and process, which provides opportunity to study the interactions of hydrologic exchange and biogeochemical reaction.

5.0 Conclusion and Prospects

The hydrology of the MDVs is simplified compared to temperate and even Arctic tundra catchments, and it is this simplicity that can be leveraged in this system to clearly identify and quantify hydrologic connectivity and the implications of these connections. While it will remain difficult to identify and characterize multiple dynamic flowpaths in most hydrologic systems, and therefore their separate influence on biogeochemical cycling for example, the MDVs provide the opportunity to reduce the competing sources and flowpaths so that one can tease out hydrologic connections and their implications. Here we demonstrated that fairly simple hydrologic analyses can be interpreted in several ways to characterize dynamic connectivity in these glacier-stream-lake systems. In addition, an analysis of the array of hyporheic research in MDV streams demonstrates the biogeochemical importance of this fundamental connection that may be overlooked in many settings.

One purpose of this paper is to provide an example analysis of hydrologic connectivity in a simple system with the goal of inspiring others to conduct similar analyses or seek similar interpretations in their systems. Other desert streams may be a closer corollary to MDV streams than forested watersheds. However, simple analyses of hydrologic records can be conducted wherever they are collected to characterize or quantify how dynamic connections to and from the stream may be. Additional analyses of dissolved constituents in stream water may provide enhanced ability to quantify how sources are dynamically contributing to stream discharge. Whereas such analyses are not foreign to many hydrologic studies, they are not always considered as evidence of connectivity, which provides the opportunity to further evolve our conceptual models and understanding of watershed hydrology, regardless of the system in question.

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