STREAM SYSTEMS TECHNOLOGY CENTER

STREAM NOTES

To Aid in Securing Favorable Conditions of Water Flows

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Hyporheic Zones in Mountain Streams: Physical Processes and Ecosystem Functions

by Steven M. Wondzell

The hyporheic zone is the area below the streambed and in the unconfined aquifer adjacent to the stream where stream water is found in the subsurface. Hyporheic exchange flow is the movement of stream water from the surface channel into the subsurface and back to the stream over relatively short periods of time (fig. 1), which creates the hyporheic zone. The boundaries of the hyporheic zone are not distinct because the stream-source water mixes with groundwater so that there can be a gradual transition from 100% stream water to 100% groundwater. Triska et al. (1989) set a threshold of 10% stream-source water to define the limits of the hyporheic zone so that regions with <10% stream-source water were defined as groundwater. Alternatively, the extent of the hyporheic zone can be delimited by water residence time, for example, the subsurface zone delineated by hyporheic exchange flows with residence times less than 24 hr (the 24-hr hyporheic zone; Gooseff 2010).

The hyporheic zone was originally described by Orghidan (1959), who observed that many aquatic insects and other macroinvertebrates that characterized surface stream channels could be found some distance into streambed gravels. He surmised that the flow of fresh stream water into those gravels created the environmental conditions that allowed stream-dwelling macroinvertebrates to persist within the substrate. Soon thereafter, Vaux (1962) showed how the mounds of gravels associated with salmon redds (the gravel "nests" in which salmonids lay their eggs) created pressure gradients on the streambed that forced the flow of stream water through the gravels and thereby kept the salmon eggs bathed in highly oxygenated water. However, the hyporheic zone is not just a place where "stream-like" water is found in the subsurface. Rather, the hyporheic zone is a place of strong environmental gradients that are determined by the length of time that stream water remains in the subsurface and the degree to which it mixes with groundwater.

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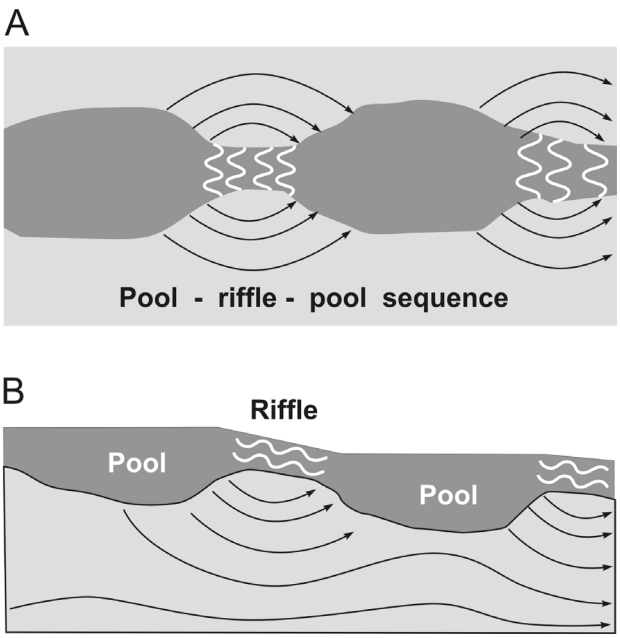

Figure 1. A highly stylized drawing of hyporheic flow paths generated by the change in the longitudinal gradient of a stream's energy profile across a pool-riffle sequence of a gravel-bed mountain stream: (A) plan-view and (B) longitudinal section.

Processes Driving Hyporheic Exchange

A variety of processes can drive hyporheic exchange. These processes can be broadly divided into hydrostatic and hydrodynamic processes. Hydrostatic-driven exchange results from static hydraulic gradients which are primarily determined by changes in water-surface elevation. Geomorphic features of the stream channel and valley floor control the elevation of surface water and can thereby create significant head gradients.

One commonly described factor creating hydrostatic head gradients responsible for hyporheic flow is the change in the longitudinal profile of the stream surface over pool-riffle or step-pool sequences (figs. 1a and 1b). These head gradients generate flow paths of stream-source water through the hyporheic zone that typically follow a nested pattern in which short flow paths are nested inside of successively longer and longer flow paths. As shown here, some of the stream water will enter the streambed at the very crest of the riffle and only flow under a few cobbles before re-entering the stream (fig. 1b). Other water will seep out of the bottom of the pool, traverse the full length of the riffle and eventually re-enter the channel deep in the downstream pool. Pool-riffle sequences also generate lateral flows where stream-source water exits the upstream pool through the streambank, flows in an arcuate path through the adjacent riparian zone, and reenters the stream through the bank of the downstream pool. Like the vertical flow paths, short lateral flow paths are also nested within successively longer and longer lateral flow paths (fig. 1a).

Locations where stream water enters the hyporheic zone are often called "downwelling zones" and locations where stream water re-emerges from the hyporheic zone are often called "upwelling zones". Graphics such as fig. 1b usually have greatly exaggerated vertical scales so that the upward and downward components of flow are easily apparent. In reality, the elevational change in the water surface between successive pools is likely to be quite small and/or the pools are likely to be widely spaced. So, while steep vertical gradients do occur in some locations, in most places hyporheic flow paths tend to flow more or less parallel to the overall valley floor gradient within a stream reach.

Hyporheic exchange flows, such as those shown in fig. 1, tend to persist over a wide range of stream discharges. For example, even when the stream is nearly dry and holding water only in the deep pools, water will continue to flow down valley, flowing out of the downstream side of one pool, through the gravels, and into the upstream edge of the next pool. As discharge increases to the point that continuous surface flow is reestablished, the head gradients through the gravels between the two pools persist, and so does the flow of stream water through the streambed gravels. Only as the abrupt change in the longitudinal gradient over the riffle becomes drowned out in very high flows will the head gradients driving hyporheic flow weaken. These will eventually disappear when the riffle is transformed into a long run of flood waters that entirely hide all traces of the pool-riffle sequence.

Many morphologic features of stream channels and valley floors drive hyporheic exchange. For example, differences in water-surface elevations between the main stream and either back channels or floodplain spring brooks commonly generate cross-valley head gradients and relatively long hyporheic flow paths. Similarly, the elevation

differences between the main channel and secondary channel along different sides of islands or mid-channel gravel bars drive hyporheic exchange flows through those features, as do the elevation differences along the upstream and downstream sides of meander bends and point bars.

Other "static" features of valley floors and alluvial sediment also drive hyporheic exchange. For example, where bedrock-constrained valleys open into wide alluvial valleys, stream water downwells into the subsurface alluvium. This process is reversed where wide alluvial valleys narrow into constrained reaches, pushing subsurface water flowing down the valley back into the stream channel. In fact, anywhere the saturated cross-sectional area of the valley floor alluvium increases, stream water may downwell into the alluvium to feed hyporheic flowpaths and wherever the cross-sectional area decreases, water will be forced back into the channel. Thus, upwelling zones are commonly present where alluvial fills thin over shallow bedrock. For example, springs may be present in normally dry desert stream channels where bedrock ledges span the full width of the valley floor and force subsurface flows to the surface. Changes in the cross-sectional area of the alluvial aquifer can create extensive hyporheic zones in large mountain valleys. The same processes also create small hyporheic zones in the wider sections of small headwater streams.

Hydrodynamic-driven hyporheic exchange results from flowing streamwater interacting with stream bedforms. Flowing water pushes against the upstream faces of ripples or dunes in sand-bedded streams, leading to increased pressure. A zone of low pressure occurs along the downstream face and the pressure differences between the upstream and downstream faces drive hyporheic exchange flow. This process has not been extensively studied in steep mountain streams, but it is the same process described by Vaux (1962) that drives hyporheic exchange through salmon redds. Further, while flow velocities through sand are quite slow, flow velocities through coarse gravels used by spawning salmon can exceed 2 m/hr (Zimmerman and LaPointe 2005), keeping salmon eggs well oxygenated and sweeping away waste materials.

Hyporheic exchange is likely to be more limited in strongly gaining reaches than in neutral reaches because of steep streamward hydraulic gradients surrounding the channel. Similarly, where water is lost to regional aquifers in strongly losing reaches, return flows of stream water back to the stream are likely to be severely restricted and thus also limit the expression of the hyporheic zone. While hyporheic exchange may be limited in gaining or losing reaches, it is seldom entirely eliminated because of the nested structure of subsurface flow paths and because hyporheic exchange can occur at a variety of spatial scales. Thus, the hyporheic zone may persist within an envelope of larger non-hyporheic flow paths.

Any given stream reach will typically contain many of the morphologic features described above, so that the resulting patterns of hyporheic and non-hyporheic flows are likely to be complex (fig. 2). Further, interactions among features may also be important in determining the actual hyporheic exchange in any given stream reach. In some cases, the effects of multiple features could be additive. For example, cross-valley flow paths between main channels and floodplain spring brooks can be accentuated by riffles. However, interactive effects could also cancel, for example, where riffles at the inflection points of meander bends reduce head gradients through point bars. There are relatively few studies that have examined multiple processes concurrently within natural stream channels and attempted to evaluate the net effect of each process on hyporheic exchange. One such study by Kasahara and Wondzell (2003) examined a number of channel morphologic features among streams of different sizes in a mountainous stream network under conditions of summer baseflow discharge. They showed that the shape of the hyporheic flow net in the 2nd-order streams was strongly oriented down valley in response to the steep (13%) longitudinal gradient of the valley floor (fig. 2a). In the 5th-order stream, the hyporheic flow net was strongly controlled by the presence, location, and relative elevation difference between water in the main channel and the back channels (fig. 2b). Despite the dominance of these features in shaping the flow net, the single strongest driver determining the amount of exchange flow occurring within the simulated stream reaches was the change in longitudinal gradient over step-pool sequences in the 2nd-order channel and pool-riffle sequences in

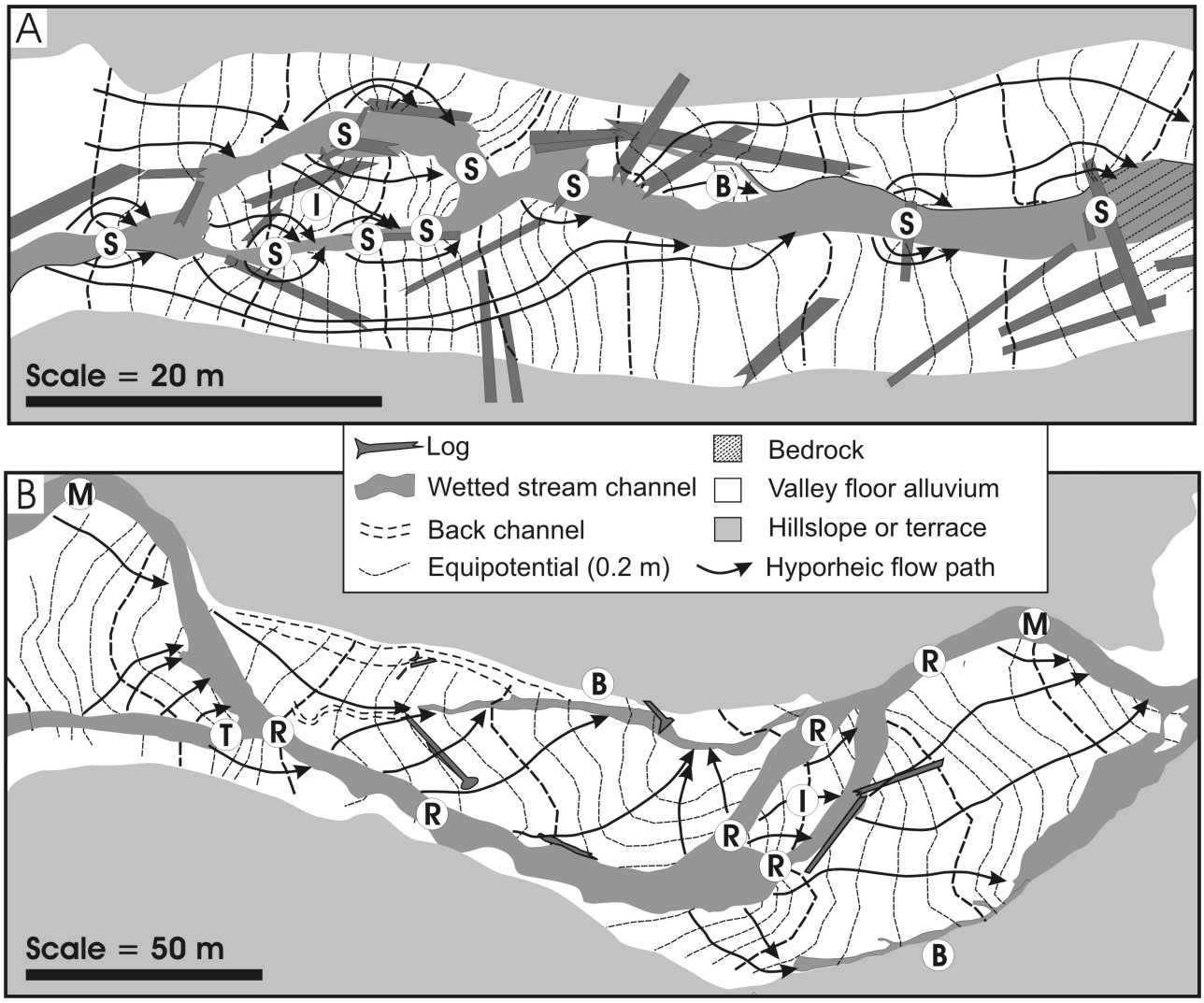

Figure 2. Examples of complex hyporheic flow paths resulting from interactions between channel morphologic features: A) a steep, 2nd-order step-pool channel with abundant large wood, and B) a moderate gradient, 5th-order pool-riffle channel with two major spring brooks. Note the difference in spatial scale between the two stream reaches. Letters indicate morphologic features driving hyporheic exchange: S = steps; R = riffles; M = meander bends; B = back channels/spring brooks; I = islands; and T = a steep riffle at the mouth of a tributary. Equipotential intervals (dashed contour lines) are 0.2 m. Hyporheic flow paths (arrows) are hand drawn to indicate general direction of hyporheic flow through the valley floor. For a more in-depth review of the geomorphic factors that lead to the development of hyporheic zones see Wondzell and Gooseff (in press).

the 5th-order channel (fig. 2).

The Residence Time Distribution of Hyporheic Zones

Hyporheic flow paths are complex and occur at a variety of spatial scales within stream reaches. Thus, hyporheic flow paths span a wide range of lengths, from very short flow paths under a few sediment grains to flow paths that exceed the length of a study reach. The amount of time required for water to travel along each flow path is not only a function of the length of the flow path, but also the head gradient driving flow and the conductivity of the sediment – factors described by Darcy's Law: $q = -K * (\Delta H/\Delta L)$.

The average head gradient ($\Delta H/\Delta L$) that drives the flow of water through the subsurface of the pool-riffle sequence shown in fig. 1 can be calculated as the difference in the elevation of the water between

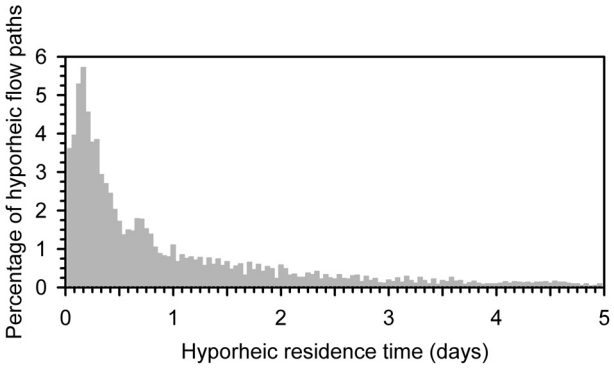

Figure 3. Residence time distribution of hyporheic exchange flows in a 100-m long reach of a 2nd-order mountain stream estimated from a groundwater flow simulation model (MODFLOW) combined with a particle tracking model (MODPATH). The distribution has been truncated at 5 days. The simulation suggested that at least some residence times exceeded 100 days, but very little water would have been flowing along such long-residence time flow paths. For more details see Kasahara and Wondzell (2003).

the two pools, divided by the distance between the pools. The hydraulic conductivity (K) is broadly related to the sediment texture. Coarse-textured sediment has large-diameter interconnected pore spaces so that, for a given head gradient, water can flow quite fast. In finer-textured sediment, or wherever pore spaces in a coarse matrix of streambed gravels become clogged with fine sediment, water is forced to flow in tiny pore spaces, and flow velocities slow dramatically.

The flow velocity of water in the hyporheic zone is usually hundreds to thousands of times slower than that of water flowing in the adjacent stream channel. The hyporheic flow velocity is determined by both the saturated hydraulic conductivity and the head gradient. For example, measurements of tracer movement through gravel/cobble streambed sediment over a steep riffle recorded maximum flow velocities of 5 m/hr. Most measurements, however, suggested flow velocities were more typically in the range of 0.5 to 1.0 m/hr (Kasahara, unpublished data), even in mountain streams where sediment is coarse textured and hydraulic gradients are steep. In lower-gradient streams, similar tracer studies showed flow velocities decreased to as little as 0.2 m/hr – requiring a full day for the tracer to flow the full length of a 4-m long stream-side gravel bar (Zarnetske et al., 2011). In sand-bedded streams, where sediment is much finer textured and head gradients are weaker still, hyporheic flow velocities are even slower, requiring 5 to 10 hours to travel 50 cm through a dune-shaped bedform (Savant et al. 1987). Because of the wide ranges in flow path lengths and because of spatial heterogeneity in both sediment textures and head gradients, there will be a broad range of residence times of stream water in the hyporheic zone within any stream reach ranging from a few seconds to many days (fig. 3). This distribution of residence times tends to be highly skewed, with much more water flowing in short residence time flow paths than in those with long residence times.

Hyporheic Exchange and Stream Temperature

Hyporheic exchange flows are often considered important to stream ecosystems because they can influence stream nutrient cycles and stream thermal regimes. In the latter case they are considered especially important in that they can help cool streams on hot summer days. The actual effect of hyporheic exchange, however, depends on the amount of hyporheic exchange flow and residence time of that water in the subsurface. For example, consider the data on stream and hyporheic temperatures shown in figs. 4a and 4b. In this small mountain stream, the stream cools overnight, reaching a minimum temperature of 15°C around 8:00 A.M. and then warms rapidly reaching maximum temperatures of 18°C in the early afternoon for a diel temperature variation of slightly more than 3°C. Very short residence time hyporheic water would show a similar pattern and thus have little, if any effect, on the stream temperature. As the residence time of water in the hyporheic zone increases, however, the diel variability in temperature decreases and the time of minimum and maximum temperature lags further and further behind the stream. In the example shown here, median time required for water to travel from the stream to individual piezometers was measured in a stream tracer test and water temperatures were recorded hourly with digital recording thermometers. At a median travel time of 8.25 hr (fig. 4a; HZ 8 hr), the diel temperature range is reduced to 1.3^{o}C, peak temperatures are recorded about 2 hr later than in the stream and minimum temperatures occur about 1 hr later. At a median travel time of 13.75 hr (fig. 4a; HZ 14 hr), the diel temperature range is reduced to 0.4^{o}C, and

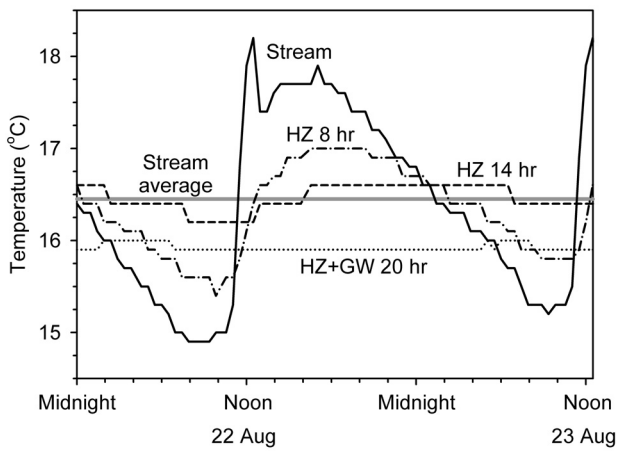

Figure 4a. Comparison of stream and hyporheic zone temperatures in late summer in a 2nd-order mountain stream (Johnson and Wondzell, unpublished data). Labels denote the location of the temperature measurements: HZ denotes hyporheic locations that are nearly 100% stream-source water; HZ+GW denotes a location where some groundwater (GW) is mixed with the stream-source water; 8 hr, 14 hr, and 20 hr denote the median time required for a conservative tracer injected into the stream to travel from the stream to the sampling location. The average stream temperature from 21 to 23 August was 16.45°C and is denoted by the bold grey horizontal line. Average hyporheic temperatures over this same time period were within ± 0.1°C of the average stream temperature; mixing with groundwater decreased the average temperature at HZ+GW 20 hr by 0.52°C.

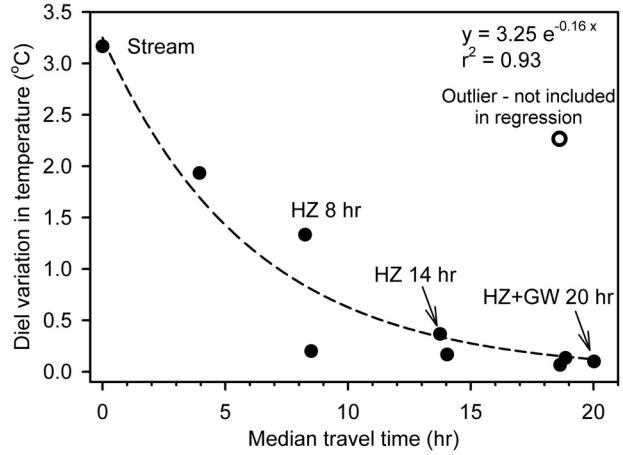

Figure 4b. Relationship between median travel time and the observed diel variability in water temperatures in either the stream channel or in piezometers located in the hyporheic zone of a steep, headwater mountain stream (Johnson and Wondzell, unpublished data). Labels indicate data from the four locations illustrated in fig. 4a.

it becomes difficult to identify discrete times for the maximum and minimum temperatures. In both cases, however, the average temperature of the hyporheic water differs from the stream by less than 0.1°C. These data demonstrate two important attributes: 1) the hyporheic water is cooler than the stream water during the hottest part of the day, but during the night the hyporheic water is actually warmer than the stream water; and 2) with increasing residence time the diel temperature variation is lost because heat exchange between stream-source water and the sediment of the hyporheic zone causes the temperature of the hyporheic zone to stabilize around the average stream temperature (fig. 4b).

So, does the hyporheic zone cool the stream? Upwelling hyporheic water can certainly cool a stream, but only under specific circumstances. First, the stream must have a substantial diel temperature variation. If the diel variation in stream temperature is negligible, then both the stream water and the hyporheic water (regardless of the residence time) will be at uniform temperature close to the average water temperature during the several preceding days so that the hyporheic zone will have no discernible effect on the stream temperature. If, however, there is a strong diel temperature range, then upwelling of hyporheic water from long residence-time exchange flow paths will cool the stream during the hottest part of the day. But the process is reversed at night and upwelling hyporheic water will actually warm the stream.

While these effects of hyporheic exchange flow might be small, they can certainly be important. If daily maximum temperatures are a critical concern in a stream with wide diel variation in temperature, and if there is sufficient hyporheic exchange, it can limit the maximum temperature attained by the stream. Even if there is too little hyporheic exchange flow to significantly reduce the peak temperature of the bulk stream water, upwelling hyporheic water will still be substantially cooler than the stream water throughout the afternoon and can provide cool thermal refugia for stream organisms.

Hyporheic water may also mix with groundwater,

which results in very distinctive thermal signature. The piezometer with the 20-hr median travel time (fig. 4a; HZ+GW 20 hr) shows virtually no diel temperature variation, but the average temperature is more than 0.5°C cooler than the stream. Groundwater temperatures are typically very stable and equal to the long-term average annual air temperature in the groundwater recharge zone, which at this site is approximately 9.2°C. A mixture of 7% groundwater at this temperature and 93% stream-source water at the short-term average temperature of the stream (16.5°C) would account for the observed decrease in average temperature. In temperate climates, groundwater will always be cooler than the stream during the warm season and thus cool the stream throughout the summer. The patterns are reversed in the winter, of course, when groundwater warms the stream, which in arctic and sub-arctic environments can prevent streams from freezing over or prevent anchor ice formation and thereby provide critical winter habitat for a wide variety of stream-dependent organisms.

Hyporheic Exchange and Stream Nitrogen Cycles

The hyporheic zone is a unique stream environment because flow velocities are very slow and water is in intimate contact with biofilms growing on sediment surfaces. Fungi and bacteria forming the biofilms interact with solutes transported with the stream water through the hyporheic zone. Thus, residence time determines how long solutes are exposed to intense biological activity. Fig. 5 from Zarnetske et al. (2011) demonstrates how residence time in a gravel bar of a low gradient stream determines if the hyporheic zone will be a net source or a net sink for nitrate in the stream ecosystem. In the example shown here, nitrate (NO_3^-) labeled with a stable isotope of nitrogen (^{15}N; a rare, naturally-occurring, non-radioactive form of nitrogen containing an extra neutron) was injected into the stream along with a salt tracer ($NaCl^-$). The movement of water could be followed using the salt tracer, from which median travel times could be determined and the fate of the $^{15}NO_3^-$ could be followed by testing samples for the presence of ^{15}N.

Results of this experiment show that aerobic biogeochemical processes dominate biological activity at residence times shorter than 7.5 hours in

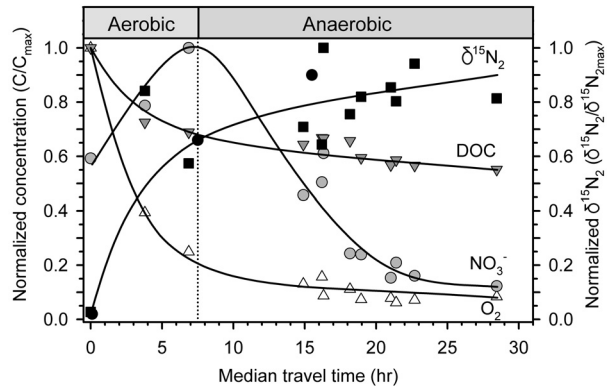

Figure 5. Relationship between median residence time and concentration of constituents related to nitrogen cycling processes in the hyporheic zone. For more information see Zarnetske et al. (2011).

the hyporheic zone, either along short flow paths or at the upstream end of long flow paths (fig. 5). These processes include aerobic metabolism in which dissolved organic carbon (DOC) and fine-particulate organic carbon (FPOC) are consumed, using up the available supply of dissolved oxygen (O_2). At the same time, dissolved organic compounds containing nitrogen (DON) are broken down, and organic nitrogen is first mineralized to ammonium (NH_4^+), and as long as the environment remains aerobic, NH_4^+ is nitrified to NO_3^-. These aerobic processes consume about 20% of the DOC and 80% of the O_2 supplied to the hyporheic zone in the stream water. Over the same time, nitrification nearly doubles the concentration of NO_3^-.

At residence times longer than 7.5 hours in the hyporheic zone, there is no longer sufficient O_2 to support aerobic metabolism and the hyporheic zone shifts toward a dominance of anaerobic biogeochemical processes (fig. 5). Dissolved organic carbon continues to be metabolized, but NO_3^- rather than O_2 is used as a terminal electron receptor in the metabolic pathways, which converts NO_3^- into di-nitrogen gas (N_2) that is subsequently lost to the atmosphere. This process is known as denitrification, and in this example denitrification leads to the loss of approximately 90% of the NO_3^- present in the hyporheic zone, including that originally in the stream-source water as well as that regenerated through mineralization and nitrification from DON .

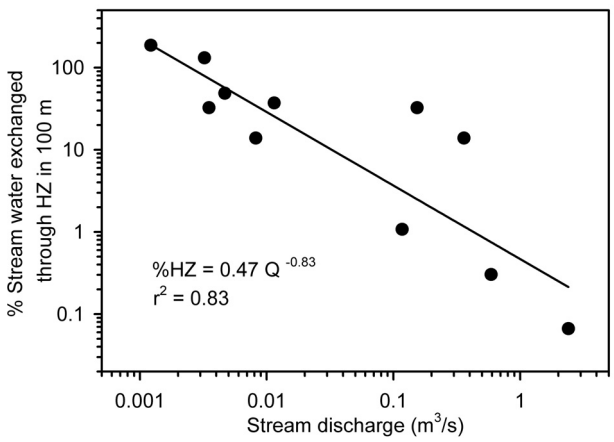

Figure 6. The relationship between stream size and the amount of hyporheic exchange flow that occurs within a 100-m long stream reach, expressed as a percentage of the total stream discharge. The scatter around the regression line is quite large, indicative of the high variability in the amount of hyporheic exchange flow that might occur within any given stream. Despite the high variability, the regression equation provides a rough first estimate of how much hyporheic exchange should be expected in a gravel-bed mountain stream. Data are compiled from several publications that were summarized in Wondzell (2011).

The threshold between aerobic and anaerobic processes is critical to stream ecosystems because it determines the fate of nitrogen in the hyporheic zone. In streams where primary productivity is limited by the supply of readily available nitrogen, the regeneration of NO_3^- from DON in the hyporheic zone will be critical to supporting ecosystem productivity. Conversely, in nitrogen-enriched streams, anaerobic denitrification is ultimately the only process that can permanently remove nitrogen from the aquatic ecosystem by converting it to N_2 gas which is lost to the atmosphere.

The threshold residence time needed to switch from dominance of aerobic processes to anerobic processes is likely to be highly variable. Temperature controls both the saturation concentration of O_2 in stream water as well as the metabolic rate of fungi and bacteria—cold water can hold more O_2 and metabolic rates are slower so that longer residence times will be necessary to use up the available O_2. The amount and composition of DOC will also be important. Hyporheic zones in streams where DOC is recalcitrant (not biologically available) or where little DOC is present may never become anoxic. Conversely, if labile (biologically available) DOC is abundant, O_2 may be rapidly consumed and anaerobic conditions may occur at very short residence times. While there is no simple way to make an a-priori prediction, it is probably reasonable to assume that aerobic conditions will persist long distances into the hyporheic zones of cold, forested mountain streams, and that the hyporheic zone will serve as a net source of nitrate because of the preponderance of relatively short residence time flow paths.

Just How Much Hyporheic Exchange is There?

For hyporheic exchange flows to significantly influence nutrient concentrations in the bulk stream water, or change the temperature of the stream water, the amount of water exchanged through the hyporheic zone must be large relative to stream discharge. Unfortunately, it is not possible to make direct measurements of hyporheic exchange nor is there any simple way to estimate the amount of hyporheic exchange.

The most common method of hyporheic investigation has relied on stream tracer injections and the collection of a breakthrough curve from the bottom of the study reach (Bencala and Walters 1983). A model optimization routine is used to parameterize a one-dimensional advection, dispersion, and transient storage model to simulate the observed breakthrough curve (e.g., Runkle 1998). The transient storage parameters from the model are usually interpreted as an index of the relative size of the hyporheic zone, but these models do not provide a quantitative estimate of the actual hyporheic exchange. Further, stream tracer tests are usually only sensitive to relatively short residence time exchange flows whereas many of the hyporheic processes important to stream ecosystems require relatively long subsurface residence times.

Groundwater flow models have also been used to simulate hyporheic exchange flows. However, these models have substantial data requirements, including fine-scale spatial distributions of saturated hydraulic conductivity (K). Unfortunately, K is also difficult to measure, and point-scale measurements from individual wells

must be interpolated to the entire model domain. Thus, while groundwater flow models do provide quantitative estimates of the amount of hyporheic exchange, those estimates are uncertain. Despite these uncertainties, groundwater flow models are useful for investigating hyporheic zones in mountain streams.

Groundwater flow models were used to simulate hyporheic exchange at five locations within the Lookout Creek watershed, a 62 km^2 mountainous watershed in the western-central Cascade Range of Oregon, USA. These sites included two 2nd-order headwater tributaries, a 4th-order tributary, and two sites on the 5th-order mainstem of Lookout Creek (fig. 6). These study sites are not "pristine", they are all located in forested watersheds where road building and forest harvest have occurred in the past. Large wood was also removed from portions of the stream network, including much of the 5th-order stream channel. Much of the road network remains in place and in use. Thus, these study sites are likely typical of managed forested watersheds. More detailed study site descriptions and the specifics of the model simulations are given in Wondzell (1994), Wondzell and Swanson (1996), and Kasahara and Wondzell (2003).

Estimates of hyporheic exchange flows from these studies suggested that the size of exchange flows, relative to stream discharge, was large only in very small streams at low discharge (fig. 6). At higher flows and in all larger streams, hyporheic exchange flows were small relative to stream discharge. For example, at a stream discharge of 1.5 L/sec (0.0015 m^3/sec) approximately 100% of the stream water is exchanged with the hyporheic zone in 100 m. This is equivalent to a stream turnover length of 100 m, meaning that an amount of water equal to the entire in-channel flow seeps into the hyporheic zone and is replaced by upwelling hyporheic water over a 100-m length of stream channel. Of course, some water makes multiple passages through the hyporheic zone and other water remains in the channel for long distances without ever entering the hyporheic zone.

Turnover length increases rapidly as discharge increases because hyporheic exchange flows become ever smaller relative to the size of the stream. Thus, at a stream discharge of only 10 L/sec, slightly more than 20% of the stream flow is exchanged through the hyporheic zone in 100 m, with a resulting turnover length of 465 m. At a discharge of 100 L/sec, the turnover length already exceeds 3 km. Despite the uncertainties associated in using groundwater flow models to estimate hyporheic exchange flows, the regression line fit to these data (fig. 6) provides a first approximation of the amounts of hyporheic exchange that should be expected in small- to medium-sized gravel-bed streams draining mountainous watersheds.

Does the Hyporheic Zone Matter?

The data in fig. 6 clearly show that only in small streams at low discharge does a large enough proportion of the stream water get exchanged with the hyporheic zone for biogeochemical processes or heat exchange occurring there to substantially influence the stream's solute load and thermal regime. At higher discharges or in larger streams, too little water is exchanged with the hyporheic zone to influence bulk water quality. However, the hyporheic zone may influence stream ecosystems in many other ways. For example, the hyporheic zone represents a unique habitat for some organisms, with patterns and amounts of upwelling and downwelling water determining the physiochemical environment within the hyporheic zone. Similarly, hyporheic exchange creates distinct patches of downwelling and upwelling on the surface of the streambed. Upwelling environments are of special interest, because upwelling water has the potential to be thermally or chemically distinct from stream water. For example, hyporheic exchange can create a diversity of thermal environments (Arrigoni et al. 2008) which can provide thermal refugia for cold-water fishes in streams on late summer days when discharge is low and ambient stream temperature is high (Ebersole et al. 2003). Also, hyporheic upwelling zones can be enriched with nitrate, thus supporting higher algal biomass, and after floods, algal biomass may recover more quickly in upwelling zones than in downwelling zones (Valett et al. 1994). These studies suggest that, even where the proportion of stream water exchanged through the hyporheic zone is too small to measurably change water temperatures or nutrient concentrations of the whole stream, hyporheic exchange can create environmental patches critical to structuring stream ecosystems.

Additional Information

For a more in-depth discussion and complete list of references on these various hyporheic zone topics, please refer to the following website: http://www.fs.fed.us/pnw/lwm/aem/people/wondzell.html.

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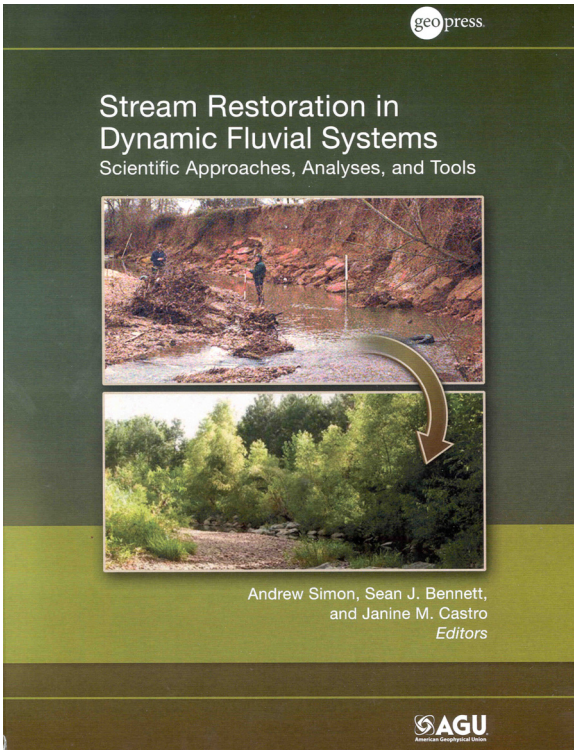
Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses, and Tools

Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses, and Tools is a multi-disciplined collection of papers from experts in both the science and practice of stream restoration. The book provides an interdisciplinary synthesis of process-based approaches, tools, and techniques currently being used in stream restoration. Stream restoration is a general term that describes any modification to the stream and adjacent riparian zone to improve the geomorphic and ecologic function, structure, and integrity of the stream corridor. The diversity of topics presented in the book reflects the ongoing debate in the research and professional communities regarding the approaches, applications, and tools being used in assessing, designing, and implementing a stream restoration project.



The book consists of 27 separately authored chapters organized into seven sections. The first section, "*Introduction*", consists of single chapter that sets the context of the book by briefly discussing the history and evolving science of stream restoration along with issues that are challenging the stream restoration community. The second section, "*General Approaches*", consists of five papers that provide background information on the different conceptual approaches currently being used in stream restoration by practitioners and academics. Section three, "*Stream Hydrology and Hydraulics*", consists of 4 chapters that examine select hydrologic and hydraulic topics that are critical and challenging design considerations in many stream restoration projects. Section four, "*Habitat Essentials*", consists of 4 chapters that discuss various physical and biological indices that are addressed in many stream restoration projects designed to enhance aquatic and floodplain habitat. Section five, "*Sediment Transport Issues*", consists of four chapters that discuss the importance of considering sediment in stream restoration projects as the source, transport, and continuity of sediment strongly influence channel stability, habitat improvement, and water quality issues being addressed in many restoration projects. Section six, "*Structural Approaches*", consists of five chapters that review commonly used structures used to stabilize channels and improve aquatic habitat. Section seven, "*Model Applications*", consists of four chapters that review the wide range of models and technology that are available to assess channel stability, design stream channels, and determine the impacts of restoration projects on flow hydraulics and sediment transport.

Researchers and practitioners involved in assessing, designing, and implementing restoration projects to improve geomorphic and ecologic processes and conditions along the stream corridor will find this book a useful reference. ***Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses, and Tools*** is published by the American Geophysical Union and can be purchased for $120 ($90 members) online at http://www.agu.org/pubs/books/. The citation for this book is: Simon, A.; Bennett, S.J.; Castro, J.M. (Editors). 2011. Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses, and Tools. Geophysical Monograph Series, volume 194. American Geophysical Union. Washington, D.C. 544 pp. doi:10.1029/GM194.

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