9.13 Geomorphic Controls on Hyporheic Exchange Across Scales: Watersheds to Particles

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Glossary

Bedform streambed features, such as ripples, sand-waves, or dunes, formed in streambed sediment through the interaction between flowing water and sediment. Bedforms can range in size from a few centimeters (ripples) to many meters (dunes).

Channel unit relatively homogeneous areas of a channel with characteristic substrate, depth, and flow pattern, typically bounded by other channel areas with different features. The length of channel units generally ranges from one to a few times the wetted width of the stream. Pools and riffles are common channel units in small streams.

Hydrodynamic exchange (also known as pumping exchange) exchange of stream water with the subsurface driven by the velocity head component of the total head gradient on the bed surface and exchange induced by momentum gradients across beds and banks.

Hydrostatic exchange exchange of stream water with the subsurface driven by static hydraulic gradients that are determined by changes in water surface elevation, spatial heterogeneity in saturated hydraulic conductivity, or changes in the saturated cross-sectional area of floodplain alluvium through which hyporheic flow occurs.

Hyporheic exchange flow the flow of stream water from the surface stream channel, through the streambed and into the shallow unconfined aquifer beneath or adjacent to the stream channel and back into the surface stream over relatively short spatial (1-100s of meters) and temporal (hours to weeks) scales.

Hyporheic zone the zone of saturated sediments underlying the surface stream or in the floodplain adjacent to the stream perfused with stream-source water that has recently left the stream channel and will return to the stream channel relatively quickly. Stream water in hyporheic flow paths may mix with groundwater so that the relative proportion of stream-source water in the hyporheic zone is highly variable, ranging from 100% stream water to nearly 100% groundwater. The extent of the hyporheic zone can be arbitrarily defined on the basis of the proportion of stream water present in the subsurface (e.g., > 10%) or the residence time of stream water (e.g., areas with residence times 24 h).

Stream reach a length of stream channel in which topographic features and sequences of channel units are relatively homogeneous. The length of stream reaches are...
generally many 10s of times the width of the wetted stream channel.  

**Transient exchange** the temporary movement of stream water into stream banks due to short-term increases in stream stage (i.e., bank storage processes due to changes in hydrostatic head gradients between stream and lateral riparian aquifer).

**Abstract**

We examined the relationship between fluvial geomorphology and hyporheic exchange flows. We use geomorphology as a framework to understand hyporheic process and how these processes change with location within a stream network, and, over time, in response to changes in stream discharge and catchment wetness. We focus primarily on hydrostatic and hydrodynamic processes – the processes where linkages to fluvial geomorphology are most direct. Hydrostatic processes result from morphologic features that create elevational head gradients, whereas hydrodynamic processes result from the interaction between stream flow and channel morphologic features. We provide examples of the specific morphologic features that drive or enable hyporheic exchange and we examine how these processes interact in real stream networks to create complex subsurface flow nets through the hyporheic zone.

**9.13.1 Introduction**

Hyporheic exchange flow (HEF) is the movement of stream water from the surface channel into the subsurface and back to the stream (Figure 1). Stream water in hyporheic flow paths may mix with groundwater so that the relative proportion of stream-source water in the hyporheic zone (HZ) is highly variable, ranging from 100% stream water to nearly 100% groundwater. Also, the residence time distribution of stream water in the hyporheic zone tends to be highly skewed, with most of the stream water moving along short flow paths and thus having short residence times (hours), but some water either moving on long flow paths or encountering relatively immobile regions having very extended residence times (weeks to months, or longer). The boundaries of the hyporheic zone are arbitrary, commonly defined by the amount of stream-source water present in the subsurface. 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 HEFs with residence times less than 24 h (the 24-h hyporheic zone; Gooseff, 2010).

The objective of this chapter is to examine the relation between geomorphology and hyporheic processes. The two primary controls on hyporheic exchange are the gradients in total head established along and across streambeds and the hydraulic conductivity of the streambed and adjacent aquifer, both of which are significantly influenced by geomorphology. Total head (also known as potential) is the sum of pressure head, elevation head, and velocity head. Pressure head represents height of a column of fluid to produce pressure. Velocity head represents the vertical component of HEF flows through the streambed. Elevation head represents the potential energy of a fluid particle in terms of its height from reference datum. Hydrostatic head is referred to as the sum of elevation and pressure head. Groundwater tables in unconfined aquifers represent the spatial gradients in hydrostatic head. A number of processes either drive or enable HEF, several of which are based on changes in head gradients. We follow the organizational structure presented by Käser et al. (2009), who divided these processes into five distinct classes:

1. **Transient exchange** – the temporary movement of stream water into stream banks due to short-term increases in stream stage (i.e., bank storage processes due to changes in hydrostatic head gradients between stream and lateral riparian aquifer).
4. Hydrostatic-driven exchange – exchange driven by static hydraulic gradients that are determined by changes in water surface elevation (Harvey and Bencala, 1993), spatial heterogeneity in saturated hydraulic conductivity, or changes in the saturated cross-sectional area of floodplain alluvium through which hyporheic flow occurs.

5. Hydrodynamic-driven exchange – exchange driven by the velocity head component of the total head gradient on the bed surface (i.e., pumping exchange; Elliott and Brooks, 1997a, 1997b) and exchange induced by momentum gradients across beds and banks.

These classes of HEF processes are coupled to geomorphic processes in many ways. This is most obvious for hydrostatic effects, which are directly dependent on channel and valley-floor morphology and the depositional environment that controls spatial heterogeneity in saturated hydraulic conductivity (K). However, turnover of streambed sediment is also related to fluvial geomorphic processes. Similarly, hydrodynamic effects result from the interaction of flow over stream bedforms. Geomorphic processes build stream bedforms and determine channel morphology, especially longitudinal gradient, bed roughness, and water depth, all of which influence flow velocity. The relationship between geomorphology and the other classes of processes is less direct, but still plays a role in controlling these processes through channel form and the size distribution of sediment that makes up the streambed. This chapter focuses primarily on the hydrostatic and hydrodynamic processes where linkages to geomorphic processes are most direct.

We organize our discussion of the interactions between geomorphology and HEF using a hierarchical scaling framework developed for river networks (Frisse1 et al., 1986; Bisson and Montgomery, 1996), starting at the whole network, through the stream segment, to the stream reach, to the channel unit, and down to the subchannel unit scale. We recognize that describing any given process or related flow path at a single scale is somewhat arbitrary because of the nested structure of the hyporheic flow net and dispersion among HEF flow paths. Despite this, the concept of scale is an important heuristic tool to organize our understanding of hyporheic processes. In many senses, the reach scale is the most informative scale at which to consider HEF. A single reach, by definition, has characteristic channel morphology so that the factors driving HEF within the reach are relatively consistent. However, only a few of the geomorphic factors driving HEF actually operate at this scale. Most of the drivers work at the channel unit or smaller scales. Moreover, to understand the importance of HEF in stream ecosystem processes, the cumulative effects of HEF must be evaluated at scales much larger than a single reach.

### 9.13.2 The Effect of Geomorphology on HEFs

#### 9.13.2.1 The Whole Network to Segment Scale

The geologic setting of the stream network is an important factor determining the likely occurrence of HEF, but there have been few attempts to study HEF at this broad scale. Rather, our expectations are pieced together by drawing comparisons among HZ studies that have been conducted in widely varying geologic settings, at different locations in the stream network, or under widely varying flow conditions. We expect that geomorphic–hyporheic relationships will differ substantially among different geologic settings.

Fluvial geomorphic studies have examined the factors that determine the types of channel morphologies present within stream networks (Montgomery and Buffington, 1997; Wohl and Merritt, 2005; Brardinoni and Hassan, 2007). Montgomery and Buffington (1997) presented one such description of the distribution of channel morphologies typical of many mountainous landscapes. They showed that catchment area and channel longitudinal gradient controlled the development of distinct channel types such that the channel types tended to follow a characteristic sequence within a catchment (Figure 2(a)). In their example, this sequence starts with bedrock and colluvial channels in the steepest, uppermost headwaters. As longitudinal gradients decrease, channels change to cascades, to step–pool, to plane–bed, to pool–riffle, and the largest, lowest-gradient rivers were typified as dune–ripple channels. Along with these changes in channel morphology, the following would be expected: decreased longitudinal gradient and mean grain size of streambed sediment, and increased depth, width, hydraulic radius, and flow velocity (Leopold and Maddock, 1953; Wohl and Merritt, 2008).

In this chapter, we use Montgomery and Buffington’s (1997) description of the sequence of channel types within a catchment as a simple heuristic model to organize our examination of the relative importance of the different processes that drive HEF within stream networks. We recognize that local controlling factors commonly interrupt simple sequencing of channel types. For example, landslides may block large mainstem channels, creating locally steep gradients over the landslide debris and uncharacteristically low gradients in the depositional reach directly upstream (Benda et al., 2003). We also recognize that regional differences in geology and geomorphology will lead to dramatically different spatial organization of channel types (see, e.g., characteristic channel type in glaciated mountainous regions as described by Brardinoni and Hassan (2007)). Our descriptions of the spatial organization of stream types and the resulting HEF processes will have to be modified for any specific landscape.

Most hyporheic exchange results from head gradients pushing water through the streambed. The amount of stream water entering the hyporheic zone is thus a function of the steepness of the head gradient and the saturated hydraulic conductivity of the streambed and underlying aquifer. The head gradients can be induced in many ways, but the two of primary influence are the hydrostatic and hydrodynamic processes. The relative importance of each of these processes is expected to vary among channel types and with longitudinal
gradient. In high-gradient streams, channel forms, such as step–pool sequences or pool–riffle sequences, can create very steep hydrostatic head gradients. Further, because of high bed roughness and relatively shallow water depth, flow velocities tend to be lower in small steep streams than in larger, low-gradient streams (Leopold and Maddock, 1953; Wondzell et al., 2007). By contrast, it is difficult for natural processes to create steep changes in the longitudinal gradient in low-gradient streams. Instead, stream flow interacts with stream bedforms, such as dunes or ripples, such that hydrodynamic forces dominate the development of head gradients through the streambed. Thus, we expect that hydrostatic effects will dominate in high-gradient channels and that hydrodynamic processes will dominate in low-gradient channels (Figure 2). Further, because channel types and longitudinal gradients generally vary systematically within stream networks, we further expect that hydrostatic effects will tend to dominate in the upper portions of stream networks and that the relative importance of hydrodynamic processes will increase down the stream network.

9.13.2.2 The Reach Scale – Setting the Potential for Hyporheic Exchange

The potential for HEF to occur varies within any given stream reach. Roughly speaking, this potential is determined by the factors that generate head differences that drive HEF; the properties of the subsurface alluvium through which HEF occurs, and the potential effect of lateral groundwater inputs from adjacent hillslopes that might limit hyporheic expression.

9.13.2.2.1 Losing and gaining reaches

Hyporheic exchange is likely to be more limited in strongly gaining reaches than in neutral reaches because of steep streamward hydrologic gradients surrounding the channel (Wroblicky et al., 1998; Storey et al., 2003; Malcolm et al., 2003, 2005; Cardenas, 2009). 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 (Cardenas, 2009). These patterns of gains and/or losses are controlled, at some level, by regional groundwater and catchment characteristics interacting with smaller-scale effects. In large gaining rivers, Larkin and Sharp (1992) demonstrated that the relative dominance of cross-valley versus down valley flow paths through valley-floor aquifers varied depending on the longitudinal gradient of the valley floor and the hydraulic conductivity of the valley-floor alluvium. In higher-gradient reaches (>0.004 m m⁻¹) and in areas with coarser substrate, flow was predominantly down valley. Conversely, where valley-floor gradients were shallower or sediment more finely textured, flow tended to be toward the stream. Thus, the way in which lateral inputs influence hyporheic exchange is not solely a function of their magnitude, but also a function of the ability of subsurface water to move down valley (Storey et al., 2003).
The ratio between these two factors—the magnitude of the inputs relative to down-valley flow—determines how hyporheic exchange is affected.

As a first approximation, the potential for down-valley flow can be estimated using the relationships summarized in Darcy’s law—that is, the product of the longitudinal valley gradient, the saturated cross-sectional area of the floodplain perpendicular to the direction of subsurface flow, and the hydraulic conductivity of the alluvium. As lateral inputs increase, several factors may change: (1) water tables may rise, thus increasing the saturated thickness and the cross-sectional area through which water flows allowing the transmission of more water, or (2) flow paths may begin to turn obliquely toward the stream, which also increases the saturated cross-sectional area and may also increase head gradients.

Under dry conditions when lateral inputs are relatively small, the potential extent and magnitude of hyporheic exchange can be fully expressed (Figure 3(a)). As subsurface flows turn toward the channel, they begin to limit the extent of the hyporheic zone with only a minor effect on the HEF (Wondzell and Swanson, 1996). If sufficiently large, lateral inputs can severely limit both the spatial extent and magnitude of hyporheic exchange (Figure 3(b); Harvey and Bengala, 1993; Wroblicky et al., 1998; Storey et al., 2003; Cardenas and Wilson, 2007; Malcolm et al., 2003; Soulsby et al., 2009).

Simple generalizations of where and when lateral inputs will limit HEF are difficult because of the wide range of geomorphic settings in which HEF occurs and because the magnitude of lateral inputs changes with catchment wetness. Lateral inputs are expected to be high when catchments are wet and decrease as catchments dry out. However, lateral inputs are not spatially uniform. In steep mountainous settings, the size of the upslope area draining directly to the valley floor is important, concentrating lateral inputs in zones at the base of hillslope hollows (Jencso et al., 2009). Lateral inputs may persist the entire year at the bases of the largest hillslope hollows. Most hillslope hollows are small, however, so that most of the stream network would be disconnected from lateral inputs except for short periods of time when catchments are very wet, for example, after large storms or during peak snowmelt. We are unaware of similar studies relating topography to spatial patterns of hillslope inputs in areas of low relief with humid climates. However, Storey et al. (2003) reported that an extensive shallow surficial aquifer was present along their lowland, low-gradient study reach and that lateral inputs of groundwater substantially reduced both the extent and the amount of HEFs except during summer base flow. Clearly, the influence of lateral inputs in lowland catchments may be much different from that in steep mountainous catchments.

Changes in lateral inputs to streams do not occur in isolation. Rather, they are likely to be accompanied by corresponding changes in stream stage (and discharge). The change in water-table elevations resulting from changed lateral inputs must be considered relative to the accompanying changes in stream stage. Although the number of studies examining changes in hyporheic flow paths with changing catchment wetness is limited, studies in small mountain streams suggest that water-table elevations in the floodplain increase more than stream stage so that HEF is typically more restricted when catchments are wet (Figures 4(a) and 4(b); Harvey and Bengala, 1993; Wondzell and Swanson, 1996; Stednick and Fernald, 1999). Storey et al. (2003) reported similar results for a lowland, low-gradient river.

In some cases, however, stream stage may change markedly without corresponding changes in precipitation recharge or changes in lateral inputs. Most examples of these processes come from large, lowland rivers because river stage is controlled by processes far upstream. These ‘bank storage’ processes (Pinder and Sauer, 1971) have been recognized as a form of transient hyporheic exchange (Figures 4(c) and 4(d)) that can result from both in-bank or over-bank floods (Bates et al., 2000; Burt et al., 2002). In some situations, increased stream stage may even lead to groundwater ridging in the floodplain, reversing head gradients and limiting lateral groundwater inputs. Similarly, hyporheic exchange through stream banks can result from diel variations in stream stage (and discharge) during snowmelt periods (Loheide and Lundquist, 2009) or from tidally induced changes in water elevations in coastal streams and rivers (Bianchin et al., 2010).

Transient hyporheic exchange may be especially evident in regulated rivers where releases from dams (or other control structures) can result in large and rapid changes in river stage without corresponding local precipitation to recharge floodplain aquifers (e.g., Fritz and Arntzen, 2007; Lewandowski...
et al., 2009; Sawyer et al., 2009; Francis et al., 2010). However, transient hyporheic exchange may not always result from fluctuations in river stage. For example, Hanrahan (2008) studied vertical HEF through the streambed of a large, regulated gravel bed river where stage sometimes changed by nearly 2 m in an hour. For the most part, they did not observe transient hyporheic exchange related to changes in stage. They concluded that hydrostatic and hydrodynamic processes remained the dominant control on HEF. Notably, Hanrahan (2008) did not examine lateral exchanges through the stream banks, which can be more responsive to changes in stage than are locations in the stream channel itself (Storey et al., 2003). Water-table fluctuations in the floodplain at long distances from the stream are not necessarily indicative of extensive HEF because pressure fluctuations can propagate through surficial (unconfined) aquifers much faster than does the actual flow of stream water. This was clearly demonstrated by Lewandowski et al. (2009) who showed that river water penetrated, at most, only 4 m into the stream bank even though water-table fluctuations were observed more than 300 m from the river.

HEF can occur in strongly gaining and losing reaches because of the nested structure of hyporheic flow paths, and because HEF can occur at a variety of spatial scales. Thus, an envelope of the HZ can be set within larger nonhyporheic flow paths (Figure 3(b); Cardenas and Wilson, 2007). Similarly, smaller-scale HEF can occur as a result of smaller-scale geomorphic drivers, even within a reach that is, overall, strongly losing (Payn et al., 2009). Further, because HEF is dominated by relatively near-stream flow paths that are short in length and residence time (Kasahara and Wondzell, 2003), the magnitude of HEF can be substantial, even in strongly gaining reaches where the spatial extent of the hyporheic zone is greatly restricted (Wondzell and Swanson, 1996; Cardenas and Wilson, 2007; Payn et al., 2009).

9.13.2.2 Changes in saturated cross-sectional area
The saturated cross-sectional area of the floodplain (orthogonal to groundwater flow path direction) is one of the factors determining the amount of groundwater transmitted down valley through the valley-floor alluvium. Thus, any change in the cross-sectional area along the length of a stream reach will lead to parallel changes in the down-valley flow of water through the floodplain, thereby driving downwelling from, or upwelling to, the stream (Stanford and Ward, 1993). Downwelling occurs where valley floors increase in width, for example, downstream of bedrock-constrained reaches (Figure 5(b); Poole et al., 2004, 2006; Acuna and Tockner, 2009). Conversely, upwelling occurs where valley floors narrow at the lower end of wide unconstrained reaches (Figure 5(a); Baxter and Hauer, 2000; Acuna and Tockner, 2009). Similarly, variations in the thickness of the surficial aquifer, caused by variations in depth to bedrock or other confining layers, drive similar patterns of upwelling and downwelling. For example, upwelling commonly occurs just upstream of bedrock sills with a subsequent transition to downwelling just downstream of such bedrock sills as the surficial aquifer again thickens (Figure 5(h); Valett, 1993). This is easily observed in streams in arid regions during the dry season, where perennial flow may only occur above bedrock sills, which force the subsurface flow to the surface.
9.13.2.3 The Subreach to Channel-Unit Scale – Hydrostatic Processes

Geomorphic features of the stream channel and valley floor within stream reaches control the elevation of surface water and can thereby create significant head gradients through the valley-floor alluvium, driving HEF. Because these geomorphic features are static on the timescales typical of hyporheic exchange (hours to weeks), they are broadly recognized as ‘hydrostatic processes’.

9.13.2.3.1 Step–pool and pool–riffle sequences

One of the best-studied examples of hydrostatic processes involves the changes in water surface elevation along a step–pool sequence and the resulting head gradients that drive HEF (Figure 1; Harvey and Bencala, 1993). Harvey and Bencala (1993) showed that the change in the longitudinal gradient of the stream channel (which approximates the stream energy profile) drove HEF. They also observed that HEF flow paths tended to be curved – first curving away from the stream above the step or riffle and then curving back to the stream below the step or riffle. Building from their observations, model analyses showed that along an idealized straight channel with homogeneous isotropic porous sediments, hyporheic flow paths around a change in the longitudinal gradient will exploit the full three-dimensional (3D) saturated volume along the channel, thus extending both vertically beneath the streambed and horizontally through the streambanks and near-stream aquifer (Figures 1(a) and 1(b)).

Real streams are substantially more complicated, however, such that changes in hydraulic conductivity of the alluvium, bends in the channel, and the spatial location of lateral groundwater inputs lead to the development of a complicated flow net through the valley floor (e.g., Cardenas and Zlotnik, 2003). Despite these complexities, the steepness of the hydraulic head gradient imposed by the change in the longitudinal gradient and the saturated hydraulic conductivity control the amount of stream water exchanged with the subsurface.

Many factors can modify the effect of steps or riffles on HEF. For example, the height of the step (or steepness of the riffle) determines the head gradient available to drive HEF so that a single very large step has the potential to drive more HEF than if the same amount of elevational change is spread over several smaller steps (Kasahara, 2000). Because of this, large wood can be important in determining the amount of HEF in forest streams. Single logs tend to create frequent, small obstructions that collect and store small amounts of sediment, forming pool–step sequences in which the extent of the hyporheic zone tends to be small (Wondzell, 2006). Although logjams are less common, they can create large obstructions storing sediment in wedges several meters deep and ≥ 10 m in length, and significantly widen constrained stream channels. Consequently, logjams can form extensive hyporheic zones in steep, confined mountain streams (Wondzell, 2006).

Large, channel-spanning logs can wedge into steep narrow channels, forcing the accumulation of sediment in channels, converting bedrock reaches to alluvial reaches with a step–pool morphology (Montgomery et al., 1996), thereby greatly enhancing HEF. Similarly, large wood can force plane-bed channels into a pool–riffle morphology (Montgomery et al., 1996), which should lead to more HEF than would be present in a comparable wood-free channel. Large wood can have the opposite effect in channels that would have a free-formed pool–riffle morphology. In one documented case, accumulations of large wood tended to force a pool–riffle channel toward a step–pool morphology (Wondzell et al., 2009). The channel adjusted to removal of all large, in-stream wood by developing a better-defined pool–riffle structure around meander bends, leading to increased sediment storage. Continued channel adjustment over time, following the removal of large wood, eventually led to substantial increases in HEF.

The size, spacing, and sequence of channel units (e.g., pools and riffles) along the stream longitudinal profile can also affect HEF (Anderson et al., 2005; Gooseff et al., 2006). Anderson et al. (2005) made detailed measurements of channel profiles and patterns of HEF, and showed that channel unit size and spacing increased as did the length of channel characterized by downwelling with increasing drainage area in a mountainous stream catchment. Gooseff et al. (2006) built on these results, examining HEF using 2D groundwater models of idealized longitudinal profiles of mountain streams. The modeling results of Gooseff et al. (2006) confirmed that both channel unit spacing and size were important in determining hyporheic exchange patterns of upwelling and downwelling. Perhaps more surprising, however, was the observation that the sequence of channel units also affected simulated HEF. Gooseff et al. (2006) compared pairs of idealized stream reaches that varied only by the way the longitudinal gradient changed over the pool–riffle sequence – that is, the slope of the riffle was gradual on its upstream end and steepest at its downstream end (described as a pool–riffle–step sequence) versus riffles that were initially steep with the slope decreasing toward the downstream end.

9.13.2.3.2 Meander bends and point bars
A variety of channel and valley-floor morphologic features, in addition to changes in the longitudinal gradient, create head gradients with the potential to drive HEF. These include channel meander bends and associated point bars, back channels or floodplain spring brooks, and islands set between main and secondary channels. In all these cases, differences in the elevational head of surface water between two channels, between different points in a single channel around a meander bend, or between points on opposite sides of an island create head gradients that drive HEF. For example, head gradients through the point bar in a meander bend are steeper than the longitudinal gradient of the stream channel around the point bar (Peterson and Sickbert, 2006) so that stream water infiltrates the upper end of the point bar and is returned to the channel at the lower end of the point bar (Figure 6(a); Vervier and Naiman, 1992). More generally, these exchange flows occur across the full length of meander bends and are influenced by both the change in stream water elevation around the meander bend and the plan-view shape of the meander bend. Highly evolved meander bends support steep head gradients across the mender neck because of the close proximity of the stream channels (Figure 6(b); Boano et al., 2006; Revelli et al., 2008) so that HEF is dominantly located in the meander neck, with much reduced HEF across the remainder of the meander where head gradients are much lower. In other cases, meanders develop a characteristic pattern of alternating pools and riffles, with riffles located at the thalweg crossovers in the inflections between adjacent meanders and pools or low-gradient runs wrapping around the point bar (Figure 6(c)). This combination of channel morphologic features can create complex HEF flow paths within meander bends. The residence times of HEF traversing meander bends can be quite short where meanders are small and saturated hydraulic conductivities are high (Pinay et al., 2009). Conversely, residence times of HEF may be extremely long in meander bends of low-gradient rivers with fine-textured sediment (Boano et al., 2006; Peterson and Sickbert, 2006).

9.13.2.3.3 Back channels and floodplain spring brooks
Channel planforms are generally complex in wide floodplains, including a network of old or abandoned channels. If the upstream ends of these channels are plugged with sediment and if the downstream ends are sufficiently incised to intercept the water table and are connected back to the river at their downstream ends, they will act as drains, imposing head gradients from the stream to the old channel (Figure 7(a); Wondzell and Swanson, 1996; Poole et al., 2006). These channels are also known as floodplain spring brooks because water upwells into the channel, forming a spring at its head. In addition to creating HEF; these channels will capture whatever water is in the surficial aquifer of the floodplain, including down-valley flows from upstream locations, and lateral inputs of groundwater or hillslope water from the valley margin. However, because lateral inputs tend to be small and spatially isolated (Jencso et al., 2009; and as discussed above), floodplain spring brooks will most generally be fed by HEF (Wondzell and Swanson, 1996; Jones et al., 2007).

Abandoned channels can also be plugged at their downstream ends and open to the river at their upstream ends. In this case, stream water can flow into the abandoned channel, infiltrate the channel bed, and raise the water table in the middle of the floodplain, thereby creating head gradients and driving HEF from the abandoned channel back to the main stream channel (Figure 7(b)). More complex situations arise when the longitudinal gradients in either the back channel or mainstem channel are interrupted by steeper riffles or steps. Figure 7(c) shows the interactions between a back channel and riffle. Above the riffle, the elevation of water in the main channel is higher than the back channel so water flows toward the spring brook. Downstream of the riffle, water in the main channel is lower than the back channel so that the back channel loses water over its downstream extent, eventually going dry before reaching the main channel.

The channel planform features that drive HEF can occur over a range of spatial scales, and their influence may change through time as the stage height of water in the main channel.
changes. For example, a small gravel bar may have low points along the stream bank. At high stage, the entire gravel bar may be submerged. As the stage decreases, the center of the bar may become exposed, creating a secondary channel along the bank. As the stage decreases further, flow may become discontinuous through the secondary channel such that it functions as a drain if it is plugged at the upstream end, or functions as a conduit allowing stream water to infiltrate the surface of the gravel bar if it is plugged at its downstream end. Old channels in large floodplains may act similarly, with continuous flow along their full length during floods, but becoming disconnected at intermediate to low stage, or even dry completely during periods of minimum discharge. In large floodplain reaches, these channels can be hundreds of meters to kilometers in length, extending nearly the full length of the stream reach (Poole et al., 2006; Arrigoni et al., 2008).

9.13.2.3.4 Secondary channels and islands
Islands present a special case of back channels in which the channel is continuously connected to the main channel over its full length. The hyporheic hydrology of islands has not been extensively studied. However, we expect that the surface water elevations in channels bounding the island create boundary conditions for total head and control HEF through islands as is generally indicated by the available literature (Dent et al., 2007; Francis et al., 2010). If channels along both sides of the island are parallel and symmetric with constant longitudinal gradient, then flow through the island will parallel the channels and the head gradient driving flow will equal the overall longitudinal gradient of the stream reach (Figure 8(a)). If riffles are present in the channels, the head gradient through the island adjacent to the riffles can be much steeper than the reach averaged longitudinal gradient (Figure 8(b)). Also, if riffles are displaced along the primary and secondary channels surrounding an elongated island such that a riffle is located near the head of the island in one channel and near the tail of the island in the second channel, the resulting head gradients would tend to drive flows laterally through the island, leading to very large cross-sectional areas experiencing HEF, and therefore large amounts of HEF, albeit, with shorter-length flow paths (Figure 8(c)). Although islands may be uncommon in most channel types, they may dominate HEF in braided and anastomosing stream reaches (Ward et al., 1999; Arscott et al., 2001). Given the complexities

![Figure 7](image1.png)

**Figure 7** Idealized conceptual model of the influence of back channels on hyporheic exchange flows. (a) A back channel is incised below the water table, acts as a drain, and creates head gradients from the main channel to the back channel. (b) A back channel is plugged near its downstream end, conducts water onto the floodplain, raises the water table, and creates head gradients toward the main channel. (c) Complex pattern of HEF caused by interactions between a riffle in the main channel and a back channel. Paleochannels (dashed lines) support preferential flow. Legend follows Figure 1.

![Figure 8](image2.png)

**Figure 8** Idealized conceptual model of the influence of mid-stream islands on hyporheic exchange flows. (a) Parallel and smooth longitudinal gradients in the channels on both sides of the island create HEF flow paths that parallel stream flow. (b) Riffles at the head of the island enhance head gradients leading to greater HEF. (c) Offset riffles create strong cross-island head gradients and flow paths, resulting in more HEF but with shorter flow path lengths and residence times. Legend follows Figure 1.
of potential sizes and shapes of islands and patterns in longitudinal gradients in the bounding channels, the resulting flow nets, residence times, and amounts of HEF are likely to vary widely.

9.13.2.3.5 Spatial heterogeneity in saturated hydraulic conductivity

Fluvial processes control the depositional environment on the streambed and across the floodplain creating spatial heterogeneity in the texture of deposited and reworked sediment across a range of scales, from the surface of the streambed to the entire floodplain. Because sediment texture is closely related to saturated hydraulic conductivity ($K$), these processes can substantially influence HEF. However, because of the difficulties in quantifying these patterns at the scales at which they influence HEF, they have been relatively little studied. At fine scales, streambed roughness can control the depositional environment across the streambed (Buffington and Montgomery, 1999), which lead to spatial patterns in the distribution of $K$ within the streambed (Genereux et al., 2008), which in turn can influence both the location and amount of HEF. HEF will be restricted where the streambed is clogged with fine sediment and preferentially located in zones with higher $K$. Experiments in flumes have also shown that HEF can also influence patterns of fine-sediment deposition, with fine sediment preferentially deposited in downwelling zones (Packman and MacKay, 2003; Rehg et al., 2005), which may explain differences in $K$ between upwelling and downwelling zones observed in a steep headwater stream (Scordo et al., 2009).

Spatially heterogeneous patterns in $K$ influence HEF. For example, groundwater flow modeling studies using homogeneous versus heterogeneous $K$ showed that spatial heterogeneity may add substantial complexity to the spatial patterns of the hyporheic flow net (Woessner, 2000). Where relatively high $K$ regions are aligned parallel with head gradients, they create preferential flow pathways (Wagner and Bretschko, 2002) that can increase the total amount of HEF (Cardenas and Zlotnik, 2003; Cardenas et al., 2004). Results from Cardenas et al. (2004) showed that influence of heterogeneity in $K$ was relatively greater in lower-gradient streams and where head gradients driving HEF were reduced. To our knowledge, the influence of fine-grained heterogeneity has not been studied in steeper channels where hydrostatic processes dominate.

Fluvial processes also influence spatial patterns in $K$ at the scale of the entire floodplain. Especially important is the layering of stream and floodplain alluvium. Layering can create strong vertical anisotropy (Chen, 2004), limiting vertical exchange and promoting lateral flows through the streambed and floodplain (Packman et al., 2006; Marion et al., 2008). Overbank deposition can also bury back channels creating ‘paleochannels’ where coarse streambed alluvium is buried under finer floodplain soils (Stanford and Ward, 1993; Stanford et al., 1994; Poole et al., 2004). If these paleochannels intercept the water table, they will function as large preferential-flow pathways that can route water the full length of a floodplain. In this regard, they function much like a subsurface version of back channels or floodplain spring brooks – either acting as drains lowering the water table in the floodplain and imposing head gradients from the stream to the paleochannel, or acting as distributaries routing water into the floodplain and imposing head gradients from the paleochannel to the stream. Locations of paleochannels are sometimes evident from shallow depressions along the floodplain. In other cases, over-bank deposition will have completely filled old channels so that there is no surficial indication on the flat floodplain surface. The influence of paleochannels is difficult to discern because networks of widely spaced wells are unlikely to find and trace the location of these features along the length of the floodplain. As a consequence, their influence on HEF has not been widely studied.

9.13.2.4 The Bedform Scale – Hydrodynamic Processes

Channel hydraulics, and the spatial and temporal distribution of velocity (kinetic energy) across streambeds, are significantly influenced by the form of the channel and the bedforms that occur in channels. The continuous feedback between pressure distribution and shear stress across the bed surface and the potential to erode the bed will cause turnover exchange to occur during times of high flows. During lower flows, when bed sediment is relatively stable, bedforms cause some level of form drag on the flows, inducing pressure distributions across the bedforms, thereby driving HEF at a scale smaller than the bedform (Figure 9). The size of the bedform is set by both the energy regime of the reach and the material that makes up the bed, and the form drag induced on the water column by the bedform is of course partly controlled by its size. Thus, the scale of HEF flowpaths induced by hydrodynamic exchange across the bedforms will scale in part with the size of bedforms present (Cardenas et al., 2004). Finally, the heterogeneity of the bed material that makes up the bedforms will have a distinct control on the flux rate and actual flowpaths through and around the bedforms (Sawyer and Cardenas, 2009).

In sand-bed streams, hydrodynamic HEF has been extensively studied both theoretically and empirically. Typical bedforms in sand-bed streams are dunes and ripples, which have a fairly predictable geometry and spacing, based on bed sediment composition and flow rate. Thibodeaux and Boyle (1987) pioneered investigations of the hydrodynamic pressure distribution across dunes, noting the penetration of channel water into the porous bedforms. Further development of a ‘pumping exchange’ model by Elliott and Brooks (1997a, 1997b) expanded the ability to predict HEF and associated

Figure 9 Idealized longitudinal-section in the center of a straight stream channel with bedforms (triangular dunes) showing the interaction with stream flow that creates regions of low and high pressure on the streambed that drive HEF. Nonhyporheic subsurface flows, known as underflow (dashed arrows), are present beneath the hyporheic zone. Legend follows Figure 1.
solute dynamics in channel-bed systems. Whereas most studies of hydrodynamic exchange processes were generally carried out in or applied to flume studies, there has been at least one application of incorporating the pumping exchange model to tracer transport in field studies. Salehin et al. (2003) studied the transport of tracer along several kilometers of Sava Brook in Sweden and successfully applied a solute transport model to the observed data to explain long residence-time distributions using the pumping exchange model theory. The predictability of dune and ripple sizing and spacing makes the pumping exchange model a useful tool to explore HEF in sand-bed streams and rivers.

In gravel-bed streams, bedform types may be generally predictable (i.e., Montgomery and Buffington, 1997; Wohl and Merritt, 2008; Chin, 2002), but the exact geometry and spacing of bedforms are less predictable, particularly at a scale that will directly influence head distributions across and along the channel. Hence, the velocity distribution in the channel and around the bedform, which contributes to hydrodynamic exchange, is also unpredictable. Tonina and Buffington (2007) conducted careful studies of total pressure distribution across streambeds in flumes that had realistic geometry of a pool–riffle sequence in a gravel-bed channel. Their results indicated that total head distribution (i.e., incorporating velocity head in addition to hydrostatic head) was important to exchange at focused points in the channel where high velocity occurred. Further, they confirmed that, in general, there was little or no contribution of velocity head to parts of the bed that were overlain by deeper, slower flow, and therefore a hydrostatic representation of exchange will likely be more applicable in these locations.

Regardless of the predictability of bedform geometry and spacing, the associated hydrodynamic HEF may induce only limited lengths of exchange in the subsurface because much of the exchange dynamics are expected to be vertical rather than lateral. Exchange lateral to the channel is more likely to be driven by hydrostatic gradients set up across meander bends or bars (as described above). Hydrodynamic HEF will contribute to, but be only one component of, total HEF in natural channels, and its importance will be dictated by both channel hydraulics and, if present, competing hydrostatic factors that can create steeper head gradients.

9.13.2.5 The Particle Scale – Turbulent Diffusion

At the particle scale on streambeds, turbulent diffusion is significantly influenced by the size and arrangement of surface sediment. Because turbulent diffusion is induced by the momentum transfer between the water column and the porous media, HEF due to turbulent diffusion is a function of the decreasing velocity profile within the surface layers of the porous media (Shimizu et al., 1990). Thus, the distribution of sediment at the surface will greatly influence the potential for energy and mass transfer within this zone. Turbulent diffusion HEF is prominent in gravel-bed streams where surface pores are more likely to accommodate such open exchanges of momentum across the bed (Tonina and Buffington, 2009). Beds composed of sand particle sizes and smaller provide too much resistance to the momentum exchange between the water column and the bed. Hence, turbulent diffusion is more likely to be an important component of HEF in low-order, high-gradient streams (Figure 2(b)). Careful theoretical and empirical research on turbulent diffusion has been conducted largely on planar beds (Shimizu et al., 1990; Habel et al., 2002). Therefore, in the complex bed topography of typical gravel channels, turbulent diffusion will be a component of HEF, likely not the singular driver of HEF.

9.13.3 Discussion

9.13.3.1 Multiple Features Acting in Concert

In the examples presented above (Figures 1 and 3–9), we have mostly focused on single types of channel morphologic features that drive or enable hydrostatic and hydrodynamic HEF. However, these features never occur in isolation. Rather, a single stream reach will typically contain many of the morphologic features described above. Interactions among these features are likely to be important in determining the actual HEF in any given stream reach. In some cases, the effects of multiple features could be additive and result in higher HEF than if they did not co-occur. For example, cross-valley flow paths between main channels and floodplain spring brooks can be accentuated by riffles (Figure 7(c)). However, interactive effects could also cancel, for example, where riffles at the inflection points of meander bends reduce head gradients through point bars (Figure 6(c)). The interactions between different processes driving or enabling HEF is complex; and, to some degree, site specific, making it difficult to quantify the effects of these interactions. Because of these difficulties, there are relatively few comparative studies that have examined multiple processes concurrently, within natural stream channels, and attempted to evaluate the net effect of each process on the total HEF within stream reaches.

Sensitivity analyses with groundwater flow models calibrated to simulate HEF in a studied stream reach provide one opportunity to examine the relative importance of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach. For example, Kasahara and Wondzell (2003) examined a number of channel morphologic features on HEF where multiple features are present in a single reach.

9.13.4 Summary

HEF is a rich and complex subject that has been studied for more than 20 years. Significant progress has been made in how to model HEF using hydraulic models that incorporate ground water, riverine flow and the coupled system. Nevertheless, many areas of research remain to be explored, including the importance of this exchange process in channel design, the impact of HEF on habitat, and the effects of climate change on HEF processes.
In low gradient channels, morphologic features can interact with changes in steam stage and lateral groundwater inputs in ways that can substantially influence the amount of HEF over time, across seasons, or within a single storm event. Storey et al. (2003) examined HEF in a pool–riffle sequence at both high- and low-base flow discharge. At high stage, the stream tended to drown the riffle, substantially reducing the change in the longitudinal gradient over the pool–riffle sequence and thus reducing HEF. By contrast, at low stage, the water surface more closely followed the streambed topography, thus creating steeper head gradients that supported more HEF. Storey et al. (2003) also showed that lateral inputs during the wet season were sufficient to eliminate most of the HEF through the riffle. Cardenas and Wilson (2006) showed that low rates of groundwater discharge limited the extent of the HZ formed by the hydrodynamics of stream bedforms, and that high rates of groundwater discharge could completely eliminate HEF.

We know of only one study comparing the relative influence of hydrostatic and hydrodynamic effects. In a flume, Tonina and Buffington (2007) investigated the control of total head (i.e., including dynamic head) in driving hyporheic exchange. Their results suggested that there are specific locations within channels where the velocity head can provide additional potential and thereby influence the pattern of hyporheic exchange.

**9.13.3.2 Change in Processes Driving HEF through the Stream Network**

Hyporheic exchange will vary widely across the sequence of channel types occurring in stream networks (Figure 2: Buffington and Tonina, 2009). Channel networks generally follow a pattern of steep headwaters to low-gradient reaches downstream. In mountain stream networks in particular, gradient changes are expected to be accompanied by channel morphology changes resulting in a sequence of distinct channel morphologies (Figure 2(a)). Obviously, bedrock reaches have negligible hyporheic zones (Gooseff et al., 2005; Wondzell, 2006). We are unaware of any studies of HEF in colluvial and cascade channel morphologies; however, the extremely high longitudinal gradients of these channels likely result in high-velocity underflow, which has been shown to restrict the extent of the hyporheic zone (Storey et al., 2003). In addition, the relatively disorganized structure of the bed sediment prevents development of stepped water surface

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**Figure 10** Examples of complex hyporheic flow paths resulting from interactions between channel morphologic features: (a) a steep, second-order step–pool channel with abundant large wood and (b) a moderate-gradient, fifth-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 HEF: 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 lines) are 0.2 m. Hyporheic flow paths (arrows) are hand drawn to indicate general direction of hyporheic flow through the valley floor.
profiles so that hydrostatically driven exchange due to longitudinal changes in gradient will likely be low. Turbulent diffusion is likely to be a primary driver of HEF in such reaches (Figure 2(b)). Free-formed step–pool channels occur at slightly lower gradients (Figure 2(a)). These channels have well-organized structure with periodic spacing of both steps and pools (Chin, 2002; Wohl and Merritt, 2008) that have been shown to be primary drivers of HEF (Figure 2(b); Kasahara and Wondzell, 2003). The addition of large wood can substantially increase sediment storage (Nakamura and Swanson, 1993; Montgomery et al., 1996), the development of step–pool structure, and the extent, amount, and residence times of HEF in these stream reaches (Wondzell, 2006). Other hydrostatic factors tend to have less dominance on HEF; these reaches have low sinuosity so meander bends are uncommon and steep longitudinal gradients limit the potential for back channels to create lateral HEF flow paths.

We are unaware of any published studies examining HEF in plane–bed channels. However, we expect HEF to be lower than in either step–pool or pool–riffle channels (Figure 2(b)). The streambed tends to be smoothly graded in these channels as suggested by their name, and there is low spatial heterogeneity in surface texture (Buffington and Montgomery, 1999). Pools are widely spaced, and both steps and riffles are rare. Although these channels occur as free-formed morphologies, pool–riffle channels can be converted to plane–bed channels by land-use practices that increase sediment supply and through the direct removal of large wood, with concurrent decreases in HEF.

Lower in the stream network, channels tend to have lower longitudinal gradients (Figure 2(a)), and even in mountainous areas, unconstrained stream reaches become increasingly common. Channel plans or morphologies can be quite complex in these rivers and, as a consequence, a wide array of channel geomorphic features influences HEF. Braided and anastomosing channels may form where sediment loads are high and stream banks are erodible; the complex of channels likely leads to substantial HEF through islands. Meandering channels form under lower sediment loads and where banks are more stable. Meandering channels typically have pool–riffle morphologies, although complexes of secondary channels, back channels, and paleochannels are common, a legacy of past floods, channel avulsions, and overbank deposition. Because most HEF occurs along short, near-stream flow paths, riffles are the dominant feature determining the amount of HEF (Kasahara and Wondzell, 2003). However, the shape of the hyporheic flow net and the residence time distribution of HEF will be strongly influenced by the complex of channel plans or morphologies. Finally, hydrodynamic processes are expected to dominate in streams with relatively mobile streambeds characterized by dune–ripple bedforms. These streams have low longitudinal gradients and therefore channel morphologic features tend not to create steep hydrostatic head gradients (Figure 2(b)).

Other exchange processes are likely to be related to specific conditions. Turnover exchange will only occur when bed material is mobile - a characteristic feature of both anastomosing and dune–ripple channels. Transient exchange will only be appreciable during wet catchment conditions, when channel stage is high and surrounding groundwater tables are comparatively low. However, transient exchange may be a dominant form of HEF in regulated rivers where stage fluctuates over daily cycles due to hydroelectric generation. Turbulent diffusion, on the other hand, is likely to occur in gravel-bed sections of the network, likely with the greatest potential influence in either cascade or plane–bed sections of stream networks where stream velocities are expected to be high.

### 9.13.4 Conclusion

Hyporheic exchange results from distinct processes, and the relations between these processes and geomorphology are well understood from a mechanistic perspective. Thus, geomorphology provides a critical framework to understand hyporheic processes and how they change with location within a stream network, and, over time, in response to changes in stream discharge and catchment wetness. To the degree that these geomorphic patterns are predictable, they provide the foundation for hydrologists to make general predictions of the relative importance of the hyporheic zone at the scale of entire catchments. Reach-to-reach variability is high in stream networks, however, so understanding HEF at the reach scale continues to require detailed study of specific stream reaches. These studies are difficult and current methodological approaches are insufficient to fully examine the full suite of processes that account for patterns of HEF in any specific stream reach. Consequently, hyporheic studies tend to focus on a single factor, or at most a small subset of the factors driving HEF. Hyporheic researchers recognize that such studies are incomplete. Detailed, holistic understanding of the importance of different processes in driving HEF; how the relative importance of these processes changes with location in the stream network, with the specific structure of any given stream reach, and with changes in discharge and lateral groundwater inputs remains elusive.

### References


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Biographical Sketch

Steve Wondzell is a research riparian ecologist in the US Forest Service’s Pacific Northwest Research Station, located in Olympia, Washington, USA. His research broadly focuses on both basic and applied problems in watershed management and riparian and aquatic ecosystems. His basic research focuses on the interactions between hydrological, geomorphological, and ecological processes that create, maintain, or modify aquatic and riparian habitats, and the ways in which these processes either interact with, or are affected by, land-use practices. His applied research focuses on developing models and decision support tools that synthesize the current knowledge of aquatic and riparian systems into forms that can help inform management decisions at large spatial and temporal scales.
Dr. Michael Gooseff is an associate professor in the Department of Civil & Environmental Engineering at Penn State University. He began his research career studying hyporheic zones associated with glacial meltwater streams in the McMurdo Dry Valleys of Antarctica for his PhD work at the University of Colorado. He continued to study hyporheic exchange at broader scales in his postdoctoral research in the HJ Andrews Experimental Forest of central Oregon, USA. He then moved to conducting hyporheic research in tundra streams of Alaska, mountain and valley streams in the intermountain western USA, and in restored and unrestored streams. He continues to have active research programs in both polar regions and continues strong collaborations to develop new techniques to study stream-groundwater interactions and implications.
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