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Hyporheic Exchange in Mountain Rivers I: Mechanics and Environmental Effects

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Abstract

Hyporheic exchange is the mixing of surface and shallow subsurface water through porous sediment surrounding a river and is driven by spatial and temporal variations in channel characteristics (streambed pressure, bed mobility, alluvial volume and hydraulic conductivity). The significance of hyporheic exchange in linking fluvial geomorphology, groundwater, and riverine habitat for aquatic and terrestrial organisms has emerged in recent decades as an important component of conserving, managing, and restoring riverine ecosystems. Here, we review the causes and environmental effects of hyporheic exchange, and provide a simple mathematical framework for examining the mechanics of exchange. A companion paper explores the potential effects of channel morphology on exchange processes and the hyporheic environments that may result in mountain basins (Buffington and Tonina 2009).

1 Introduction

Geographic maps of a region can reveal a complex mosaic of landscapes, each characterized by its own topography, vegetation, geology, climate and human impact. It is well known that river networks connect these diverse regions, allowing the development of complex, integrated ecosystems and the exchange of water, sediment, organisms, and nutrients from mountain headwaters to the ocean (Bertuzzo et al. 2007; Schumm 1977; Vannote et al. 1980). Because the sediment surrounding alluvial rivers is porous, there is a less obvious but equally important coupling of the river with (1) the subsurface groundwater system, (2) off-channel water bodies (lakes, ponds, and abandoned side channels), and (3) floodplain soils and vegetation (Amoros and Bornette 2002; Duke et al. 2007; Poole et al. 2004; Stanford and Ward 1993; Stanford et al. 2005; Whiting and Pomeranets 1997; Woessner 2000). Alluvial rivers are coupled to these environments via the hyporheic zone (Figure 1). The term hyporheic is from the Greek words 'hypo-' (υπο, under) and 'rhe-' (ρε, flow) (Dahm et al. 2006; Orghidan 1959)

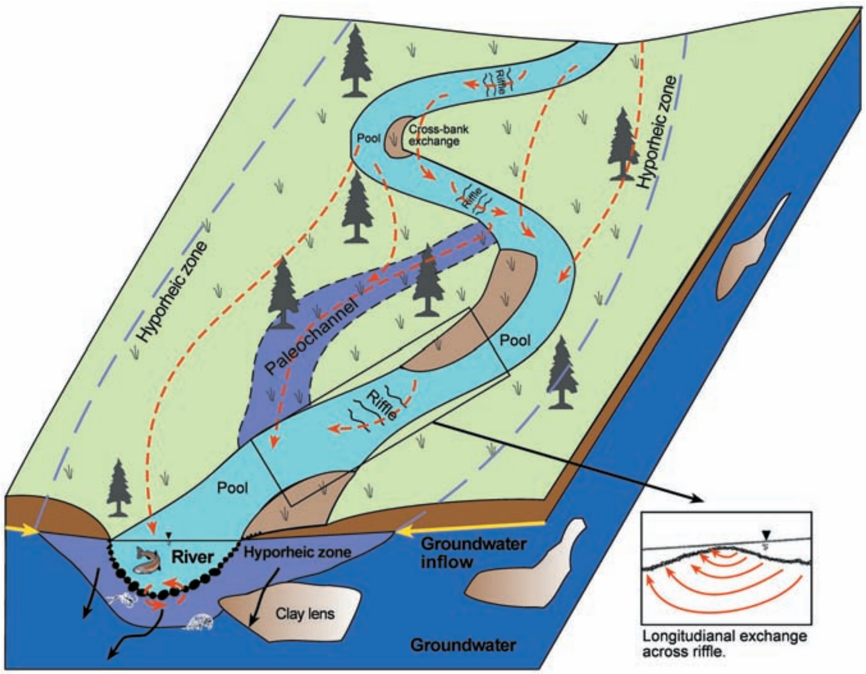

Fig. 1. Surface, hyporheic, and groundwater flow domains in an alluvial valley. The hyporheic zone is defined by the extent of surface-subsurface mixing (hyporheic exchange) through the porous sediment surrounding a river, and is characterized by circulation cells that move surface water into the surrounding alluvium and back to the river again (dashed red arrows). Hyporheic exchange is distinguished from far-field inflow of groundwater (yellow arrows) and from one way outflow of river water (i.e. flow paths that do not circulate water from the river and back again). Modified from drawings by White (1993), Harvey et al. (1996), Edwards (1998), and Stanford (2006). Macroinvertebrates redrawn from Gibert et al. (1994).

and indicates shallow groundwater flow beneath a river and in the surrounding floodplain. The hyporheic zone delineates a volume of saturated sediment that surrounds a river, where mixing of river water and shallow groundwater occurs, and constitutes a transitional area (ecotone) between the surface and groundwater hydrologic systems and between aquatic and terrestrial habitats in the riparian zone (Boulton et al. 1998; Dahm et al. 2006; Edwards 1998; Jones and Mulholland 2000; NRC 2002).

Hyporheic mixing occurs via circulation cells that move river water into the alluvium (downwelling) and back again (upwelling) (Figures 1–2), and is referred to as hyporheic exchange (or exchange flow) (Bencala 2005; Vaux 1962). Hyporheic exchange is distinguished from far-field inflow of groundwater to the river or vice versa (sensu Harvey et al. 1996) (Figure 1). The spatial and temporal scales of hyporheic mixing depend on the extent of the hyporheic zone, which may be limited in small rivers that are confined by hillslopes, or may be expansive in larger floodplain rivers (Stanford 2006). Furthermore, the hyporheic zone may expand and

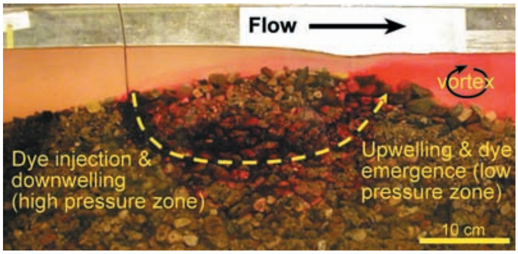

Fig. 2. Hyporheic exchange generated by pressures differences across a dune-like bedform constructed in a teaching flume. The exchange path is visualized by red dye injected at the upstream edge of the bedform, where it downwells into the sediment and follows an arcuate path that emerges into the surface flow just downstream of the bedform crest.

contract seasonally with changes in surface and subsurface flows (Malard et al. 2002). Consequently, the hyporheic zone is a spatially and temporally variable feature (Boulton et al. 1998; Stanford 2006; Stanford et al. 2005).

The significance of hyporheic exchange in linking fluvial geomorphology, groundwater, and riverine habitat for aquatic and terrestrial organisms has emerged in recent decades as an important component of conserving, managing, and restoring riverine ecosystems (Edwards 1998; Jones and Mulholland 2000; NRC 2002; Smith 2005; Stanford and Ward 1993; Stanford et al. 1996; Triska et al. 1993b). Although hyporheic studies have been conducted since the 1920s, process-based hyporheic investigations did not begin until the 1960s, and have increased exponentially since then (Dahm et al. 2006; Valett et al. 1993). In this article, we provide a process-based examination of hyporheic exchange in mountain basins, starting with a review of the environmental significance of hyporheic exchange, followed by a discussion of methods for delineating the hyporheic zone. We then present a simple mathematical model for examining the primary mechanics of hyporheic exchange and discuss the components of the model. A companion paper explores the potential effects of channel morphology and associated fluvial processes on the mechanics of exchange and the hyporheic environments that may result (Buffington and Tonina 2009).

2 Environmental Effects of Hyporheic Exchange

Hyporheic exchange has multiple effects on riverine ecosystems. Downwelling flows transfer solutes and surface water advectively into the sediment, modifying subsurface solute concentrations (Bencala and Walters 1983), thereby influencing the type and quality of subsurface habitat available to different organisms (Kim et al. 1992). Organisms living in the streambed further modify solute concentrations, taking up a portion of the delivered solutes and releasing waste products that are carried away by the flow (Bott et al. 1984; Triska et al. 1993a,b). These life forms include microorganisms such as bacteria, fungi, and protozoa that live on sediment surfaces, as well as macroinvertebrates, fish, and other organisms that live some portion of their lives within the hyporheic zone (Edwards 1998).

Downwelling flow also delivers water to the subsurface sediment that is rich in dissolved oxygen, creating aerobic environments within the shallow hyporheic zone that aquatic animals may colonize. However, decomposition of organic matter within the hyporheic zone consumes oxygen, creating concentration gradients that vary depending on reaction time, initial oxygen concentration, flow velocity, and length of hyporheic flow paths (Findlay et al. 1993). In contrast, upwelling flow brings reduced elements and transformed solutes from the hyporheic environment into the river (Nagaoka and Ohgaki 1990; Triska et al. 1989a,b, 1993b). Upwelling hyporheic flow can also be a source of nutrients and algal cells to streams that are recovering after flood events (Valett et al. 1994). Similarly, riparian vegetation is responsive to patterns of hyporheic exchange and nutrient-rich upwelling areas (Harner and Stanford 2003; NRC 2002). Riparian processes, in turn, influence hyporheic characteristics by creating local upwelling areas where tree roots draw in hyporheic water (Constantz et al. 1994; Duke et al. 2007), and by modifying hyporheic chemistry due to seasonal expansion and inundation of floodplain soils by hyporheic waters (Edwards 1998).

Additional factors, such as the composition of the sediment and the rate of mixing between surface, hyporheic, and deeper ground waters will influence temperature, pH, electrical conductivity, and dissolved element concentrations of hyporheic waters, producing physicochemical gradients to which fish, macroinvertebrates, and other organisms are sensitive. The gradients that develop within the hyporheic zone produce a complex environment, allowing for species diversity and the expression of different biological niches (Boulton et al. 1998; Brunke and Gonser 1997; Dole-Olivier et al. 1997; Malard et al. 2002). In some cases, hyporheic characteristics and physicochemical gradients are driven by surface–subsurface exchange, while in other cases the deeper groundwater system controls, or strongly modifies, hyporheic characteristics (Harvey and Bencala 1993; Malcolm et al. 2005; Storey et al. 2003; White 1993). For example, the downstream gradient of river valleys creates a downvalley underflow of deeper groundwater, the velocity of which influences hyporheic circulation; strong underflow velocities compress the vertical extent of the hyporheic zone (Storey et al. 2003; Tonina 2005), altering the length of hyporheic flow paths and resultant physicochemical gradients and reaction times. Similarly, the lateral extent of the hyporheic zone may be influenced by the seasonal strength of regional groundwater flow to the channel (Boano et al. 2008; Storey et al. 2003), and the net flux of water into the channel (net upwelling) may be composed of waters derived from both the hyporheic zone and the far-field groundwater system (Harvey and Bencala 1993; Malcolm et al. 2005; White 1993). In some cases, regional groundwater inflow may dominate observed magnitudes and patterns of water temperature in both the hyporheic zone and the river (Bilby 1984; Creuzé des

Châtelliers and Reygrobellet 1990; Shepherd et al. 1986; White et al. 1987). Conversely, riverine pressure gradients that drive hyporheic exchange can locally modify regional groundwater flow fields (Harvey and Bencala 1993). Hence, it is important to be able to identify the relative controls (surface vs. regional groundwater) on the hyporheic zone and its biophysical characteristics.

3 Delineating the Hyporheic Zone

Depending on the extent of alluvium within a river valley and the strength of processes driving hyporheic flow, the hyporheic zone may extend vertically up to tens of meters and horizontally hundreds of meters to more than a kilometer (Stanford and Ward 1988). In floodplain rivers, the hyporheic zone can be subdivided into different biophysical zones (biotopes), representing different types, rates and magnitudes of physical, biological, and chemical processes (Boulton et al. 1992; Stanford et al. 2005). For example, the hyporheic zone can be divided into fluvial (riverbed and banks), parafluvial (saturated sediments under exposed bars), and floodplain environments (Edwards 1998; Stanford 2006). Because hyporheic exchange acts over multiple spatial and temporal scales (Buffington and Tonina 2009), no single definition for the hyporheic zone exists that will suit all applications and study goals (Bencala 2000); consequently, one's definition should be tailored to the problem at hand. Nevertheless, three primary methods are used for delineating the hyporheic zone, reflecting the historic discipline-specific interests in hyporheic processes: geochemical, hydraulic, and biological.

The geochemical method defines the hyporheic zone as the volume of sediment containing a specified percentage of surface water (e.g. $\geq 10\%$, Triska et al. 1989b). The relative abundance of surface versus subsurface constituents can be determined by measuring concentrations of conservative tracers added to the surface flow (fluorescein, rhodamine, and various types of chlorides) (Bencala and Walters 1983; Castro and Hornberger 1991; Jonsson et al. 2004; Triska et al. 1989a,b; Wondzell and Swanson 1996) or by measuring natural properties of the surface and subsurface water, such as pH, electrical conductivity, and temperature (Hendricks and White 1991; White et al. 1987). These characteristics are measured with an array of sampling devices placed within the channel and surrounding floodplain from which the three-dimensional boundaries of the hyporheic zone can be derived by constructing contour maps of the spatial variation of tracer concentration (e.g. Harvey et al. 1996).

The hydraulic method defines the hyporheic zone in terms of groundwater models that trace the trajectories of hyporheic exchange through the sediment as a function of streambed pressure gradients and hydraulic conductivity (Figure 3). In this method, all flow paths of hyporheic exchange that begin and end at the streambed and banks within the study reach collectively define the hyporheic volume (Cardenas et al. 2004; Tonina and Buffington 2007).

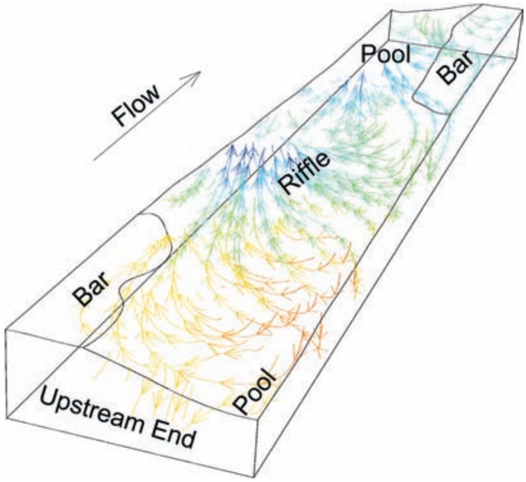

Fig. 3. Predicted hyporheic flow paths in a gravel pool-riffle channel. All pathlines originate from the surface and are colored by total pressure (Pascals, increasing from cold to warm colors). Modified from Tonina and Buffington (2007).

The biological method defines the hyporheic zone in terms of the presence of hyporheic fauna (Gibert et al. 1994; Stanford and Ward 1988). It requires knowing species types and their habitats, and can be defined in terms of occasional users of the hyporheic zone (epigean fauna), or permanent hyporheic dwellers (hypogean fauna) (Edwards 1998). Sampling of these organisms usually involves obtaining a series of sediment cores (Williams and Hynes 1974) or installing arrays of piezometers (standpipes) from which subsurface water is pumped (Malard et al. 2002) or in which multi-level pit traps are installed (Danielopol and Niederreiter 1987).

Each of these methods can yield different results and may depend on the sampling effort, which is typically of limited spatial extent and resolution due to the cost of installing piezometers or other measurement devices (Edwards 1998). Furthermore, temporal changes in discharge and other properties of the surface and groundwater flow may have different effects on each method. For instance, changes in dissolved oxygen concentrations or water temperature may cause hypogean fauna to shrink closer to the riverbed or expand deeper into the sediment, altering the boundaries of the hyporheic zone as defined by the biological method, but these changes would not alter the boundaries of the hyporheic zone as defined by the hydraulic method (all else being equal).

4 Mechanics of Hyporheic Exchange

Here, we develop a simple model for quantifying the primary mechanisms of hyporheic exchange, building from the approach of Vaux (1962, 1968).

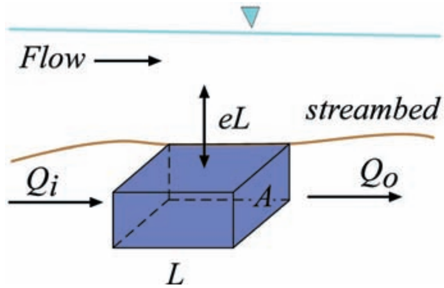

Fig. 4. Flow through an element of the streambed. L is length, A is cross-sectional area, Q_i and Q_o are the subsurface inflow and outflow, respectively, and e is the upwelling or downwelling hyporheic flux per unit length (L) of riverbed. After Vaux (1962).

Consider a sediment volume within the riverbed that has a cross-sectional area A and a longitudinal dimension L (Figure 4). The direction of subsurface flow through this volume is assumed parallel to its sides, such that no lateral fluxes occur. Furthermore, it is assumed that the bottom of the volume is impermeable (extends to underlying bedrock or to a low-permeability layer) and its upper surface coincides with the sediment–water interface. Hence, the change in the amount of water in this volume over time (dV_w/dt) is equal to the subsurface inflow (Q_i) minus the outflow (Q_o), plus the surface–pore water exchange per unit length of riverbed (e, the upwelling or downwelling hyporheic flux)

$$\frac{dV_W}{dt} = Q_i - Q_o + e \cdot L \qquad (1)$$

For steady-state conditions, V_w does not change with time ($dV_w/dt = 0$), allowing solution of the hyporheic flux as

$$e = \frac{Q_o - Q_i}{L} = \frac{\Delta Q}{\Delta l} = \frac{dQ}{dl} \qquad (2)$$

where l indicates an infinitesimal streambed length. We define the subsurface discharge (Q) from the Darcy (1856) equation

$$Q = uA = -K\frac{dh}{dl}A \qquad (3)$$

where u is the subsurface flow velocity, K is the sediment hydraulic conductivity (its ability to transmit subsurface flow as a function of porosity and connectivity of pores), and dh/dl is the spatial gradient of the energy head (h, total streambed pressure, expressed in meters of water) (Darcy 1856; Freeze and Cherry 1979). Substituting (3) into (2) and allowing all of the terms in the Darcy equation to vary spatially yields

$$e = \frac{d\left(-KA\frac{dh}{dl}\right)}{dl} = -KA\frac{d^2h}{dl^2} - K\frac{dA}{dl}\frac{dh}{dl} - A\frac{dK}{dl}\frac{dh}{dl} \qquad (4)$$

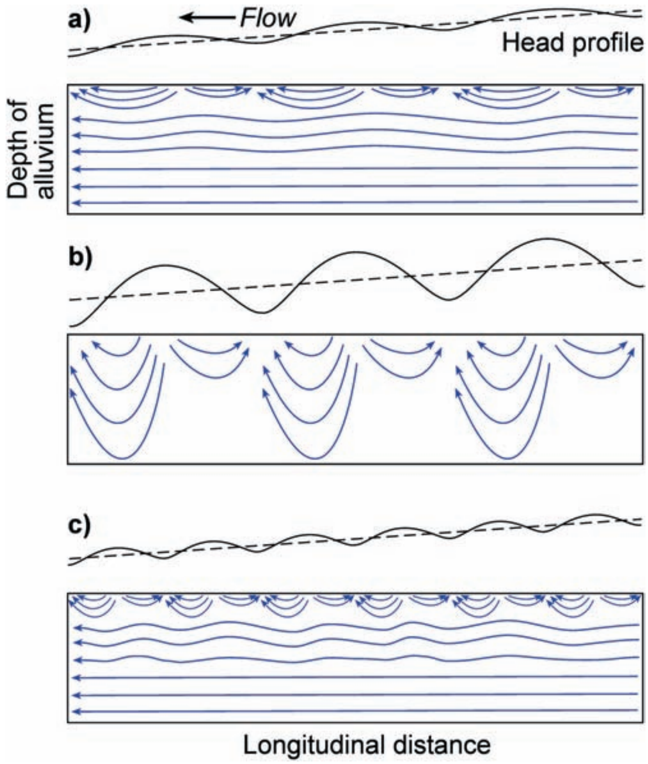

Fig. 5. Depth and spatial pattern of hyporheic exchange (vectors) as a function of the amplitude and wavelength of the energy-head profile (a-c). The gradient of each head profile (dashed line) is the same in each case. The depth of alluvium limits the depth of exchange in b). After Tóth (1963).

This equation states that hyporheic exchange is driven by spatial changes in the energy head gradient (d^2h/dl^2), spatial changes in the cross-sectional area of alluvium (dA/dl), and spatial changes in hydraulic conductivity (dK/dl). Each of these mechanisms can individually generate hyporheic exchange and is examined in turn in the following sections.

4.1 Energy head (total streambed pressure)

The first right-hand term of equation (4) states that hyporheic exchange depends on the curvature of the energy head surface. Concave curvature ($+d^2h/dl^2$) indicates upwelling flow, and convex curvature ($-d^2h/dl^2$) indicates downwelling flow (Vaux 1962) (Figure 5). Furthermore, the depth and spatial pattern of the exchange are controlled by the amplitude and wavelength of the head surface (Figure 5) (Tóth 1963). Assuming that the depth of alluvuim is not limiting, higher-amplitude head variations drive deeper hyporheic flow (Figure 5a and b), while shorter-wavelength head variations create more circulation cells with reduced path lengths and exchange times (Figure 5a and c).

When the water surface approximates the energy grade line, upwelling and downwelling areas can be identified visually by observing the water-surface shape, and whether it is convex or concave (Figure 6a) (Anderson et al. 2005; Harvey and Bencala 1993; Vaux 1962). This approximation requires conditions of gradually varying streambed topography and flow, which may be difficult to satisfy in natural channels. Nevertheless, the water-surface profile may be a first-order approximation of the energy grade line and consequent hyporheic exchange.

In rivers, variations in head are caused by flow–boundary interactions (Raudkivi 1963; Savant et al. 1987; Thibodeaux and Boyle 1987; Vanoni and Hwang 1967; Vittal et al. 1977). Flow obstructions (boulders, logs, and bedforms) are a common cause for head variations in mountain rivers, creating a zone of high pressure upstream of the obstruction and a downstream zone of low pressure in its wake, driving hyporheic circulation under the obstruction (Figure 6a) (Cooper 1965; Elliott and Brooks 1997a,b; Savant et al. 1987; Thibodeaux and Boyle 1987). This mechanism is known as pumping exchange (Elliott and Brooks 1997b; Rutherfurd et al. 1995) or bedform-induced convection (Figure 2; Packman et al. 2004; Thibodeaux and Boyle 1987).

Regions of high and low pressure can be due to variations in (1) the elevation head (changes in bed elevation, z), (2) the static pressure head (changes in flow depth, h_p), and (3) the dynamic pressure head (changes in flow velocity and momentum, $CU^2/2g$, where U is mean flow velocity, g is gravitational acceleration, and C is a generic loss coefficient representing changes in momentum due to form drag or channel contraction/expansion; Thibodeaux and Boyle 1987; Vittal et al. 1977), which collectively define the total streambed pressure

$$h = z + h_p + C\frac{U^2}{2g} \quad (5)$$

The relative importance of each of these pressure components changes with discharge, channel hydraulics, and bed topography. For example, bedforms exert a strong influence on the flow when they are partially exposed or only shallowly submerged, as is the case in Figure 2. This interaction between the bedforms and the flow creates spatial variations of the water-surface profile that enhance spatial variations in depth, static pressure gradients, and hyporheic exchange compared to a more uniform water-surface profile (Figure 7) (Buffington and Tonina 2008; Harvey and Bencala 1993; Tonina and Buffington 2007; Vaux 1962). Critical or unstable flow conditions may also develop during low submergence, creating spatial variations in velocity and dynamic pressure-head gradients that further enhance hyporheic exchange (Figures 2 and 7a) (Tonina and Buffington 2007; Vaux 1962). As flow increases and submerges bed topography, the bedforms exert less influence on the water-surface profile, which becomes

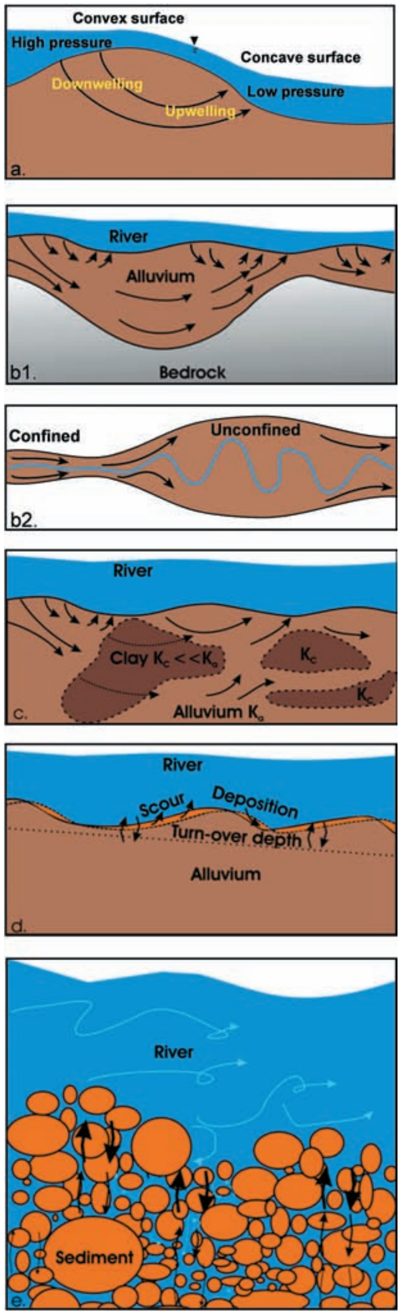

Fig. 6. Mechanisms causing hyporheic exchange: (a) spatial variation in near-bed pressure head, (b) spatial variation in alluvial area (resulting from either (b1) vertical or (b2) lateral changes in alluvial extent), (c) spatial variation in hydraulic conductivity, (d) bedload transport (turnover) and (e) near-bed turbulence.

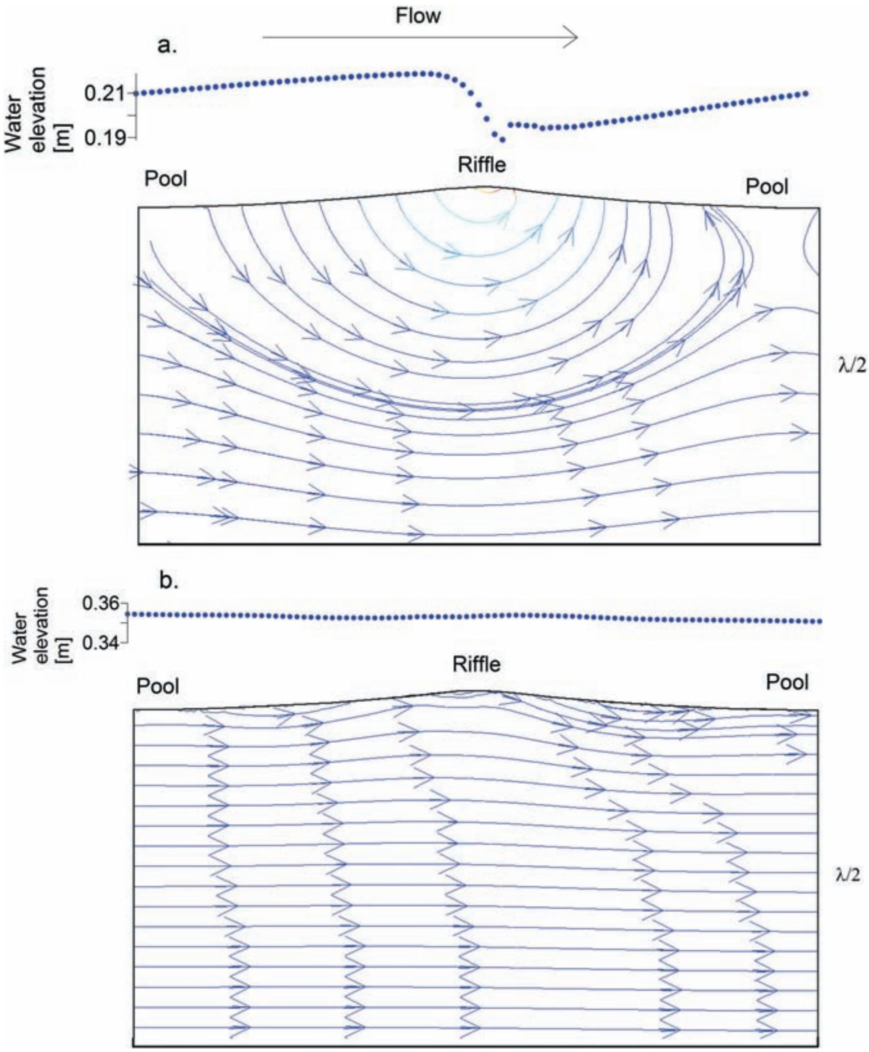

Fig. 7. Simulated hyporheic pathlines for a synthetic two-dimensional pool-riffle channel with (a) low discharge (8% bankfull flow) and (b) high discharge (100% bankfull flow). Predicted water surface profiles are plotted separately above each panel at exaggerated vertical scales to emphasize their differences. Hyporheic exchange is shown over a depth equal to half the wavelength of pool-riffle topography (λ/2). Modified from Buffington and Tonina (2008).

more uniform (Figure 7b), reducing the spatial variation of static pressure and decreasing hyporheic exchange (Buffington and Tonina 2008; Tonina and Buffington 2007). In contrast, dynamic pressure-head variations may increase or decrease with discharge, depending on bedform shape and whether flow separation (turbulence) occurs in the lee of the bedform (Figure 2). For example, dynamic pressure-head variations decrease with discharge and bedform submergence in channels with smoothly varying topography that minimizes flow separation (Tonina and Buffington 2007), while dynamic pressure-head variations increase with discharge in channels with sharp-edged bedforms that promote flow separation (Middleton and Southard 1984; Raudkivi 1963) and increased hyporheic exchange (Elliott and Brooks 1997b; Thibodeaux and Boyle 1987).

Logs and other woody debris that fall into channels in forested basins create flow obstructions that have particularly strong controls on channel form and hydraulics (e.g. Buffington and Montgomery 1999a; Buffington et al. 2002; Keller and Swanson 1979; Marston 1982; Montgomery et al. 2003). Although the effects of wood debris on streambed pressure distributions and hyporheic exchange have not been widely studied, several recent investigations highlight its potential importance. Mutz et al. (2007) found that wood debris caused a significant increase in the depth and volume of hyporheic exchange in a laboratory study of sand-bed rivers, and Lautz et al. (2006) observed that debris dams in a sand-bed river produced exchange comparable to that caused by meander bends. Furthermore, in step-pool channels of the Oregon Coast Range, Wondzell (2006) found that large, infrequent steps created by log jams produced greater hyporheic exchange than smaller, more frequent steps caused by individual logs or boulder clusters. The importance of wood debris in modifying local head gradients and generating hyporheic exchange likely depends on its frequency, size, and orientation within the channel.

Flow obstructions can also be created by aquatic organisms. For example, White (1990) proposed that hyporheic circulation is developed below beaver dams, lamprey nests, and *Chara* hummocks (macrophytic algae), each of which modify bed topography, local pressure distributions, and nutrient fluxes (White and Hendricks 2000). Similarly, shrimp mounds, mussel clusters, and crabs induce pressure variations and hyporheic circulation in ocean environments (Huettel and Gust 1992; Ziebis et al. 1996), while salmon and trout nests do the same in riverine and lacustrine environments (Carling et al. 2006; Tonina 2005). In general, hyporheic circulation caused by aquatic organisms enhances subsurface velocity and the flow of oxygen and nutrients in ways that are beneficial to the organism or its progeny (Tonina 2005; White 1990; White and Hendricks 2000).

Other sources of pressure-head variations in rivers may include channel sinuosity, divided or braided channels, sudden changes in channel width or slope, turbulent mixing at tributary junctions, steady and unsteady waves, and rapid changes in reach morphology (Boano et al. 2006, 2007; Gooseff et al. 2006; Harvey and Bencala 1993; Kasahara and Wondzell 2003; Qian et al. 2008; Wondzell 2006).

4.2 Alluvial area

The second right-hand term of equation (4) states that spatial changes in alluvial area (dA/dl) can produce hyporheic exchange. A reduction of alluvial area results in a smaller subsurface volume, expelling pore water into the river and creating an upwelling flux, while expansion of the alluvial deposit increases the subsurface volume, drawing river water into the substrate and producing a downwelling flux (Figure 6b) (Vaux 1968), particularly during dry seasons (Malard et al. 2002; Stanford and Ward 1988). Spatial variations in alluvial area can result from downstream changes in either alluvial depth (Figure 6b1) or valley width (degree of confinement by hillslopes and bedrock) (Figure 6b2). For example, irregular bedrock topography that is covered by shallow alluvium can create strong spatial variations in alluvial depth that drive hyporheic exchange at reach and sub-reach scales. At larger spatial scales, the frequency of peaks and valleys in the underlying bedrock topography can drive valley-scale circulation of hyporheic flow that may be superimposed on local, smaller-scale hyporheic exchange caused by bed topography (Figure 6b1) (Baxter and Hauer 2000). The geologic and geomorphic history of an area are particularly important at these larger scales. For example, glaciers can leave long-lasting legacies on hyporheic exchange, influencing the underlying bedrock topography, valley geometry, and the volume and composition of alluvial fill in mountain basins (Baxter and Hauer 2000). Similarly, volcanic eruptions that generate ashflows and mudflows can dramatically alter the volume and characteristics of alluvial fill (Hayes et al. 2002; Lipman and Mullineaux 1981; Newhall and Punongbayan 1996) and consequent hyporheic processes.

Alluvial volume generally increases downslope in mountain basins because of the combined effects of declining valley gradients, decreasing channel competence (size of sediment that can be carried by a river), and increasing sediment supply (Montgomery and Buffington 1997), but local geologic structures (faults, dikes) and changes in rock type can reset this downstream trend, causing spatial variations in the volume of alluvium. Spatial variations in the river profile and volume of alluvium can also result from tributary debris fans that partially block the river and force upstream sedimentation (Benda et al. 2003a,b, 2004).

The composition of the alluvium may also be important. Extensive fine-grained deposits embedded at depth within the alluvial fill can create impervious layers that reduce the available depth of alluvium and extent of hyporheic exchange (Figure 6c) (Vaux 1968). Hence, it is the depth of the near-surface permeable sediment, rather than the entire depth of alluvium, that actually structures hyporheic exchange (Storey et al. 2003; Tonina 2005). This highlights the importance of quantifying alluvial stratigraphy and sediment composition for accurate modeling of hyporheic exchange (Acworth and Dasey 2003; Cardenas et al. 2004). It is also important to note that hyporheic exchange does not necessarily extend to the bedrock boundary even in homogeneous alluvial deposits. Rather, the hyporheic volume depends on the available alluvial volume, strength of head variations driving hyporheic flow (Figure 5) (Tóth 1963), and compression of the hyporheic volume by valley slope and consequent velocity of the groundwater underflow (Storey et al. 2003; Tonina 2005). Furthermore, changes in the depth of alluvium will have no effect on the pattern and depth of hyporheic fluxes (Storey et al. 2003) unless hyporheic flow paths are limited by the alluvial depth (Figure 5b).

4.3 HYDRAULIC CONDUCTIVITY

The last term of equation (4) states that spatial variations in hydraulic conductivity (dK/dl) can also cause hyporheic exchange. A reduction in hydraulic conductivity impedes subsurface flow, causing the excess to be expelled into the stream (upwelling), whereas an increase of sediment permeability (higher K) increases subsurface velocities and draws water into the sediment (downwelling) (Figure 6c) (Vaux 1968). Hydraulic conductivity depends on sediment porosity and connectivity of pores, which are functions of the type of sediment (size distribution and percentage of fine material) and its condition (degree of consolidation and presence of organic material) (Freeze and Cherry 1979).

Alluvial valleys are composed of a mosaic of sediment patches and hydraulic conductivities, reflecting the depositional and erosional history of the valley (e.g. river migration, glaciation, mass wasting). Furthermore, streambeds commonly exhibit textural patches (grain-size facies) that vary both horizontally and vertically (Buffington and Montgomery 1999a,b; Dietrich et al. 1989, 2006; Lisle et al. 1993; Rice 1994; Wolcott and Church 1991), increasing the spatial variability of hydraulic conductivity and consequent hyporheic exchange (Malard et al. 2002).

Few hyporheic studies have explicitly quantified the effects of spatial variations in sediment stratigraphy and hydraulic conductivity. Recent work by Salehin et al. (2004) showed that sediment heterogeneity increases the spatial variability and magnitude of hyporheic exchange in sand-bed channels with planar or dune-like topography, but Cardenas et al. (2004) found that effects of sediment heterogeneity may be secondary to pressure head variations in meandering sand-bed rivers (also see Sophocleous 1991).

Biotic activity can also influence hydraulic conductivity. For example, spawning fish sort and mix streambed sediments (Chapman 1988; Hassan et al. 2008), as do benthic organisms as they move and forage along and within the streambed (Statzner et al. 1996, 2000). Animal secretions may also bind sediments (Statzner et al. 1999) and fill pore spaces, further altering hydraulic conductivity.

4.4 ADDITIONAL MECHANISMS

Two additional mechanisms for hyporheic exchange not included in equation (4) are bedload transport and turbulent fluctuations at the bed.

4.4.1 Bedload Transport

Sediment transport in rivers can cause hyporheic exchange through mechanical mixing of the bed material. For example, in sand-bed rivers, dune and ripple bedforms migrate downstream by eroding sediment from the upstream face of the bedform and depositing it on the downstream side, causing mechanical mixing that releases pore water from the eroding side of the bedform and traps river water on the depositing side (Figure 6d). This type of hyporheic exchange is called turnover, and works as a plug flow, with river water penetrating the sediment to the depth of scour (Elliott and Brooks 1997a,b). The amount of hyporheic exchange produced by this process depends on the size of the bedform and its rate of downstream translation (Packman and Bencala 2000). Furthermore, the size and frequency of bedforms change with discharge in sand-bed rivers (Gilbert 1914; Middleton and Southard 1984; Shields 1936), potentially altering the volume of hyporheic exchange that occurs with bedform movement. In rivers with mobile bedforms (i.e. sand-bed channels), turnover and pumping models can be coupled to predict hyporheic exchange (Packman and Brooks 2001). In general, turnover reduces the time available for a parcel of water to move through the subsurface via pumping exchange, reducing the magnitude of exchange that would otherwise occur for stationary bedforms (Packman and Bencala 2000).

4.4.2 Turbulence

The permeability of riverbeds allows turbulent exchange of momentum between the river and near-surface pore water (Ho and Gelhar 1973; Mendoza and Zhou 1992; Ruff and Gelhar 1972), resulting in hyporheic exchange via turbulent diffusion (Nagaoka and Ohgaki 1990; Packman et al. 2004; Shimizu et al. 1990) (Figure 6e). Near-bed turbulence is generated by a variety of factors, including large-scale flow structures caused by flow–boundary interactions (Buffin-Bélanger et al. 2000a), meso-scale turbulent wakes behind large flow obstructions (bedforms, wood, boulders) (Buffington et al. 2002; Middleton and Southard 1984), and small-scale wakes of local obstructions on the bed (protruding grains, particle clusters) (Brayshaw et al. 1983; Buffin-Bélanger et al. 2000b).

In order for turbulent fluctuations to be effective, they must be able to propagate into the sediment, suggesting that this mechanism is more important for well-sorted coarse sediments, which tend to be more porous than poorly sorted sediments. Nevertheless, turbulent diffusion into the streambed is a shallow phenomenon due to rapid damping of turbulent energy by the bed sediment (Detert et al. 2007; Ruff and Gelhar 1972; Vollmer et al. 2002). Turbulent fluctuations may only penetrate the bed to depths of about 2–10 times the mean diameter of the bed material (Detert et al. 2007; Packman et al. 2004; Shimizu et al. 1990; Vollmer et al. 2002), affecting only a fraction of typical alluvial depths.

Turbulent diffusion in the near-surface sediment also causes non-Darcy flow due to the occurrence of inertial forces and a nonlinear dependence of the subsurface flow on interfacial drag (nonlinear K) (Shimizu et al. 1990). In this case, Forchheimer's law can replace the Darcy equation to better describe this nonlinear behavior (Hassanizadeh and Gray 1987; Shimizu et al. 1990; Venkataraman and Rao 1998).

Head gradients are also modified by the presence of permeable boundaries and the coupling of surface and subsurface flows (Ho and Gelhar 1973; Mendoza and Zhou 1992; Nagaoka and Ohgaki 1990; Packman et al. 2004; Ruff and Gelhar 1972; Shimizu et al. 1990). The turbulent exchange of momentum between surface and subsurface flow across the permeable bed surface creates a slip velocity (river flow does not go to zero at the streambed, but rather declines exponentially with depth into the bed; Detert et al. 2007; Ruff and Gelhar 1972; Vollmer et al. 2002). These interactions modify the head profile and cause a downstream shift of the head distribution and consequent patterns of hyporheic exchange compared to that of an impermeable bed (Ho and Gelhar 1973). The modulation of head gradients by subsurface flow increases with bed permeability and the degree of surface–subsurface hydraulic coupling (Ho and Gelhar 1973; Mendoza and Zhou 1992).

5 Measuring Hyporheic Exchange

A variety or methods exist for measuring hyporheic exchange, ranging from direct observations to numerical predictions. A common field technique for assessing local exchange is to measure the vertical head gradient (VHG) at a given depth within the sediment using a piezometer. The VHG is the difference in water elevation between the piezometer and the stream surface (Δh) divided by the depth of measurement below the streambed (Z), with positive/negative values indicating upwelling/downwelling, respectively (Anderson et al. 2005; Lee and Cherry 1978; Terhune 1958; Valett et al. 1994). Point measures of VHG can be contoured to determine reach-scale patterns of hyporheic exchange (Baxter and Hauer 2000), but the magnitude and direction of exchange (upwelling vs. downwelling) will be sensitive to the depth of measurement; a factor that typically has not been accounted for in prior studies.

At reach and subreach scales, passive tracers (salt, rhodamine, and various chlorides) can be used to quantify exchange rates and magnitudes (Bencala and Walters 1983; Gooseff et al. 2003, 2007; Harvey and Bencala 1993; Tonina and Buffington 2007; Triska et al. 1989b, 1993a,b; Zarnetske et al. 2008). In this approach, characteristics of the exchange are interpreted from the shape of the breakthrough curve (a plot of concentration versus time) (Bencala and Walters 1983; Castro and Hornberger 1991; Gooseff et al. 2003, 2007; Harvey et al. 1996; Packman et al. 2004). Natural tracers (e.g. organic carbon, nitrogen) also have been used to investigate geochemical and biochemical reactions within the hyporheic zone (Hendricks and White 1991; Triska et al. 1989a,b, 1993a,b), however these tracers are reactive, rather than passive. Water temperature is a particularly useful natural tracer because it also has biological relevance for aquatic organisms (Constantz and Thomas 1997; Torgersen et al. 1999; White et al. 1987).

Hyporheic exchange can also be predicted through numerical modeling. For example, the transient storage model (Bencala and Walters 1983) is a one-dimensional advection-dispersion approach that treats exchange in terms of fixed-volume storage elements along the river (Bencala 2000; Harvey and Gorelick 2000; Runkel 1998). However, the model does not separate storage caused by hyporheic exchange versus that due to 'dead zones' within the river (backwater eddies, stagnant water at the bottom of pools, flow through vegetation) (Bencala and Walters 1983). The approach also treats hyporheic exchange abstractly (i.e. the storage elements are 'black boxes') and it may not correctly describe spatial and temporal rates of exchange, particularly the longer exchange paths (tail of the residence time distribution) that may have important biochemical implications (Bencala 2005; Gooseff et al. 2003; Harvey et al. 1996; Wondzell 2006; Zaramella et al. 2003). Nevertheless, it is a useful first-order approximation that captures the bulk of the hyporheic exchange (that which occurs in the shallow, near-surface portion of the hyporheic zone, characterized by rapid exchange along short flow paths that are strongly coupled to surface hydraulics; Harvey and Wagner 2000; Tonina and Buffington 2007; Zaramella et al. 2003).

An alternative modeling approach is to treat hyporheic exchange as groundwater flow within porous media. This method requires measuring or predicting the pressure distribution along the streambed and in the floodplain and determining the hydraulic conductivity of the sediment. Elliott and Brooks (1997a,b) developed a Darcy-type model for predicting hyporheic exchange in sand-bedded channels with two-dimensional dune-like bedforms (Marion et al. 2002; Packman and Bencala 2000; Packman et al. 2000; Salehin, et al. 2004; Zaramella et al. 2003), but three-dimensional solutions for any streambed pressure distribution can be obtained by coupling the known pressure field with numerical groundwater models, such as MODFLOW (Cardenas et al. 2004; Gooseff et al. 2006; Kasahara and Wondzell 2003; Storey et al. 2003; Wondzell and Swanson 1996; Wroblicky et al. 1998) or FLUENT (Tonina and Buffington 2007). For example, Tonina and Buffington (2007) extended the Elliott and Brooks model to gravel-bed rivers with pool-riffle bedforms, showing the importance of accounting for the three-dimensional structure of streambed pressures and the effects of changing discharge in these channels.

6 Conclusion

Hyporheic exchange results from a complex interaction among in-stream flow, bed topography, alluvial characteristics and groundwater. Streambed permeability allows this exchange and strongly influences flux intensity. The factors controlling hyporheic exchange (streambed pressure, bed mobility, alluvial volume and hydraulic conductivity) vary along the stream network, but may vary systematically with stream type, leading to distinct hyporheic environments (scales and rates of exchange) (Buffington and Tonina 2009).

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