

COMMENTARY

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Key Points:

- Nutrient processing in streambed sediments is facilitated by hyporheic exchange across submerged bedforms
- New evidence suggests that stream turbulence plays an outsized but underappreciated role in this process
- Incorporating stream turbulence into stream network models should greatly improve local-to-global assessments of nutrient pollution

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Modeling the Effects of Turbulence on Hyporheic Exchange and Local-to-Global Nutrient Processing in Streams

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Abstract New experimental techniques are allowing, for the first time, direct visualization of mass and momentum transport across the sediment-water interface in streams. These experimental insights are catalyzing a renaissance in our understanding of the role stream turbulence plays in a host of critical ecosystem services, including nutrient cycling. In this commentary, we briefly review the nature of stream turbulence and its role in hyporheic exchange and nutrient cycling in streams. A simple process-based model, borrowed from biochemical engineering, provides the link between empirical relationships for grain-scale turbulent mixing and nutrient processing at reach, catchment, continental, and global scales.

Plain Language Summary Streams transport excess nitrogen and phosphorus from point and nonpoint sources in a watershed to downstream receiving waters. But streams are not pipes. Microorganisms living in streambed sediments catalyze a broad range of redox reactions that reduce the impacts of nutrient pollution or in some cases exacerbate it. In this commentary we discuss recent advances in our understanding of how turbulence influences the transport and biogeochemical processing of nutrients in streambed sediments and explore how these concepts might be incorporated into stream network models of nutrient fate and transport at local-to-global scales.

1. Introduction

Much of the nutrient processing carried out by streams, especially headwater streams, occurs in the hyporheic zone, defined as the region of the streambed where surface water and groundwater mix (Boano et al., 2014; Harvey & Gooseff, 2015; Hester & Gooseff, 2010; Krause et al., 2017). Static and dynamic pressure variations over the streambed surface (together with stream turbulence) drive the bidirectional exchange of water, oxygen, nutrients, and energy between the main stream channel and its surrounding sediments, a process known as hyporheic exchange (Hester et al., 2017). Nutrients (and other solutes) spiral downstream, alternating between relatively fast moving (often turbulent) water in the stream and relatively slow (often laminar) exchange within the hyporheic zone (Boano et al., 2014; Ensign & Doyle, 2006; Mulholland & DeAngelis, 2000). The hyporheic zone—with its unique ecotone, long residence times, steep redox gradients, and biofilm-associated microbial populations—is a biocatalytic filter for streamborne nutrients and pollutants (Azizian et al., 2015; Battin et al., 2008; Boano et al., 2014; Dahm et al., 2002; Krause et al., 2017).

Hyporheic exchange occurs over a broad range of flow paths (and transit times), ranging from >10 km (> 1 year) for lateral exchange through river bars and meander banks to <1 m (<1 hr) for vertical exchange across submerged bedforms such as ripples and dunes (Boano et al., 2014; Gomez-Velez & Harvey, 2014). This vast variation in scale presents a type of Goldilocks dilemma: The hyporheic zone's ability to process nutrients will be compromised if the flow paths and transit times are too long or too short. Harvey et al. (2013) formulated the relevant trade-offs into a reaction significance factor (RSF):

$$RSF = \left(\frac{\tau_{transit}}{\tau_{reaction}} \right) \times \left(\frac{L_{reach}}{L_{turnover}} \right) \quad (1)$$

Here $\tau_{transit}$ and $\tau_{reaction}$ (units of hours) represent time scales for transit and reaction within the hyporheic zone, respectively; $L_{turnover}$ (units of meters) is the distance a water parcel travels downstream before entering the hyporheic zone; and L_{reach} is a reference distance (taken by Harvey et al. as $L_{reach} = 1$ km). The

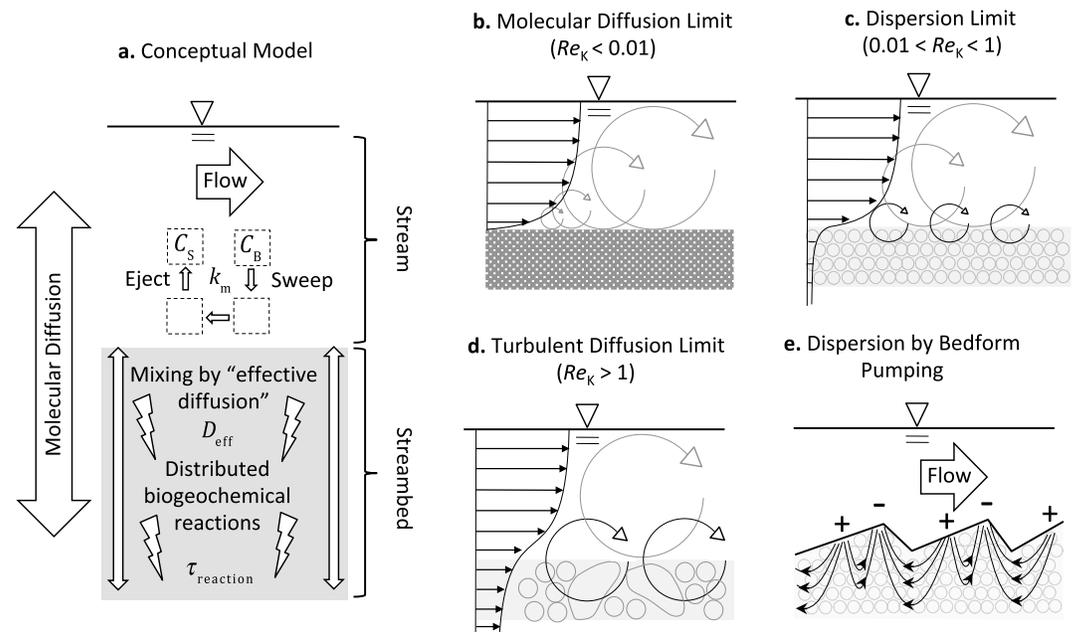


Figure 1. Conceptual model of how stream turbulence affects the vertical exchange and biogeochemical processing of nutrients: (a) nutrient transport from the bulk stream to the sediment bed occurs by fluid sweep and eject motions associated with coherent turbulence; mixing and reaction within the bed is modeled with an effective diffusivity (D_{eff}) and a first-order reaction time scale ($\tau_{reaction}$). The variables C_B and C_S represent nutrient concentrations in the bulk stream and at the surface of the stream bed, respectively. Transport mechanisms contributing to D_{eff} depend on the magnitude of the permeability Reynolds number (Re_k) and include (b) molecular diffusion modified by the tortuosity of the sediments, (c) dispersive mixing associated with laminarization of overlying turbulence, and (d) turbulent diffusive mixing associated with inertial penetration of stream turbulence into the streambed. For each range of permeability Reynolds number, a typical time-averaged velocity profile is shown (horizontal arrows). (e) Static and dynamic pressure variations over the sediment-water interface (illustrated here by + and – symbols over the top of a bedform) also drive hyporheic exchange and contribute vertical dispersive mixing. Molecular diffusion acts to smooth out concentration gradients both above and below the sediment-water interface.

Goldilocks trade-off arises because frequent hyporheic excursions ($L_{reach}/L_{turnover}$ large) are generally associated with short transit times ($\tau_{transit}/\tau_{reaction}$ small). The *just right* condition occurs when the RSF is maximized.

An important question is whether short flow paths (associated with vertical exchange across submerged bedforms) or long flow paths (associated with lateral exchange across river bars and meander banks) dominate nutrient cycling in streams. Based on numerical simulations of aerobic respiration and denitrification in the Mississippi River Network, RSFs for vertical exchange are consistently larger than those for lateral exchange, and the difference increases with river size (Gomez-Velez et al., 2015). Put another way, vertical exchange across submerged bedforms, such as ripples and dunes, has the just right combination of frequent hyporheic excursions and travel times well matched to the time scales of aerobic respiration and denitrification (assumed by the authors to be $\tau_{reaction} \approx 1$ and 10 hr, respectively). These authors also observed reach-to-reach variability (or patchiness) in the simulated RSFs, consistent with the existence of biogeochemical hot spots (and hot moments) along river corridors (Krause et al., 2017; McClain et al., 2003; O'Connor et al., 2012).

The stream network simulations described above do not explicitly account for stream turbulence. Stream turbulence can influence vertical exchange in at least two ways, by controlling (1) the rate nutrients are delivered from the bulk stream to the streambed (or vice versa) and (2) mixing and transport of nutrients within the streambed (Figure 1a). In both cases, the vertical transport of mass (and momentum) is facilitated by eddies in the stream that are spatially coherent; that is, the positive or negative velocity fluctuations are spatially correlated (Grant & Marusic, 2011; Sinha et al., 2017; Vollmer et al., 2002; Zhong et al., 2016). At the sediment-water interface, coherent turbulence manifests as sweep and ejection events, in

which water parcels in the bulk stream with nutrient concentration C_B (units of moles per cubic meter) sweep down to the interface, while water parcels near the interface with interfacial nutrient concentration C_S (units of moles per cubic meter) are ejected into the bulk stream. This turbulent *stirring* of solutes increases the surface area across which concentration gradients develop, accelerating irreversible mixing by molecular diffusion at a rate controlled by the molecular diffusion coefficient D_m (units of meters squared per second). In short, turbulent bursts vertically stir the solute, which is then irreversible mixed by molecular diffusion (Hester et al., 2017).

Sweep and eject events (together with molecular diffusion) result in a net vertical transport of mass to (or from) the sediment-water interface that can be modeled with surface renewal theory (SRT; Danckwartz, 1951; O'Connor & Hondo, 2008; Pinczewski & Sideman, 1974). As applied to the present problem, SRT predicts that the flux of nutrients to the bed J_{bed} (units of moles per square meter per second) is equal to a transfer coefficient k_m (units of meters per second) multiplied by the difference in nutrient concentration in the bulk stream and at the streambed surface (equation (2a)):

$$J_{bed} = k_m(C_B - C_S) \quad (2a)$$

$$k_m = au_*Sc^{-b} \quad (2b)$$

$$u_* \equiv \sqrt{\tau_b/\rho} \quad (2c)$$

$$Sc \equiv \nu/D_m \quad (2d)$$

The mass transfer coefficient k_m is proportional to the shear velocity u_* (units of meters per second; equation (2b)), which, in turn, is a rescaled form of the bed shear stress τ_b (units of kilograms per meter per square second) where ρ is the density of water (units of kilograms per cubic meter; equation (2c)). The Schmidt number Sc (unitless) represents the relative rates of momentum diffusion (as represented by the kinematic viscosity ν , units of square meters per second) and mass diffusion (as represented by the molecular diffusion coefficient, D_m) in the periods between turbulent bursts (equation (2d); see O'Connor and Hondzo for a detailed derivation of equation (2b)). The unitless constants a and b appearing in equation (2b) vary over a narrow range, depending on the conceptual model adopted for mass transfer across the turbulent boundary layer (Grant & Marusic, 2011; Lorke & Peeters, 2006). However, it is important to note that the applicability of equation (2b) in the limit where the interface itself is turbulent (i.e., at high permeability Reynolds numbers, see below) remains untested. Indeed, under such conditions it may be appropriate to set the Schmidt number to unity, reflecting the fact that turbulence (rather than molecular diffusion) dominates both stirring and mixing at the interface in such cases (Voermans et al., 2018).

Empirical evidence supports the application of SRT to the vertical turbulent mixing of nutrients from stream to streambed. For example, O'Connor and Hondzo (2008) reported that SRT correctly predicts the flux of oxygen from the bulk stream to organic-rich sediments given a choice of constants in equation (2b) of $a = 0.17$ and $b = 2/3$. Grant et al. (2018) recently extended O'Connor and Hondzo's laboratory results to the field, showing that the SRT mass transfer coefficient k_m sets an upper bound on measured rates of nitrate removal by assimilation and denitrification in 69 headwater streams across the United States (included as part of the second Lotic Intersite Nitrogen eXperiment; Mulholland et al., 2008).

Turbulence also transports nutrients within the streambed, depending on the magnitude of the permeability Reynolds number: $Re_K = u_*\sqrt{K}/\nu$ where the new variable K (units of square meters) is the streambed permeability. In flume studies employing refractive index-matching particle tracking velocimetry, Voermans et al. (2017, 2018) directly observed turbulent momentum (and by analogy mass) transport below the streambed surface. These researchers found that transport within the streambed can be divided into three regimes depending on the magnitude of the permeability Reynolds number: (1) $Re_K \leq 0.01$, mixing in the streambed occurs primarily by molecular diffusion (Figure 1b); (2) $0.01 < Re_K < 1$, turbulent eddies *laminarize* as they strike the sediment-water interface and mixing in the streambed occurs primarily by dispersion (a pattern also observed by Roche et al., 2018; Figure 1c);

and (3) $Re_K > 1$, turbulent eddies penetrate into the sediment and mixing in the streambed occurs primarily by turbulent diffusion (Figure 1d). While not investigated by Voermans et al. pressure variations over the sediment-water interface (e.g., induced by turbulent flow over submerged bedforms; Boano et al., 2014; Cardenas & Wilson, 2007; Elliott & Brooks, 1997; Thibodeaux & Boyle, 1987) also drives hyporheic exchange, a process known as bedform pumping (Figure 1e). Viewed in the context of vertical exchange, bedform pumping contributes additional dispersive mixing in streambed sediments (Grant et al., 2012, 2014; Hester et al., 2017; O'Connor & Harvey, 2008).

The concepts outlined above can be unified into a simple (steady state and one-dimensional) process-based model for the vertical exchange and biogeochemical cycling of nutrients in the hyporheic zone of streams. The model assumes that vertical exchange occurs in two sequential steps: nutrients are first transported from the bulk stream to the sediment-water interface by sweep and ejection events (top portion of Figure 1a) and from there into the streambed (bottom portion of Figure 1a) by the various mixing mechanisms outlined above (Figures 1b through 1e). While dividing mass transport across the sediment-water interface into two steps may seem unnecessary (after all, a turbulent burst will simultaneously stir mass above and below the sediment-water interface), it is a useful mathematical construct because most nutrient transformations occur in the second step (by microorganisms present at or below the sediment-water interface) and additional quasi-steady processes, such as bedform pumping, contribute to mass transport below the sediment-water interface. For the first step we adopt the SRT model. For the second step we assume that nutrient transport and biogeochemical reaction can be modeled with an effective diffusivity (D_{eff} , units of square meters per second) and a first-order rate constant (with characteristic time constant τ_{reaction}), respectively. For the sake of simplicity (and consistent with experimental estimates of the effective diffusivity presented later), we assume that D_{eff} does not vary with depth into the bed. Because the influence of free stream turbulence on pore-scale mixing does, in fact, decline with depth into the bed (Roche et al., 2018; Vollmer et al., 2002), D_{eff} can be thought of as a depth-averaged mixing coefficient.

Given this simple conceptualization, a closed-form steady state solution is easily derived for the nutrient uptake velocity v_f (units of meters per second), one of several stream spiraling metrics routinely used to quantify nutrient removal in streams and rivers (Mulholland et al., 2009; Wollheim et al., 2006; derivation in the supporting information of Grant et al. (2018)):

$$v_f \equiv \frac{J_{\text{bed}}}{C_B} = \alpha k_m, \quad 0 \leq \alpha \leq 1, \quad k_m > 0 \quad (3a)$$

$$\alpha = 1 - \frac{1}{\psi + 1}, \quad \psi \geq 0 \quad (3b)$$

$$\psi = \frac{\theta \sqrt{D_{\text{eff}}/\tau_{\text{reaction}}}}{k_m} \quad (3c)$$

The removal efficiency α (unitless) represents the fraction of nutrient molecules swept to the sediment-water interface that are removed by the streambed; in turn, the removal efficiency depends on a dimensionless parameter (ψ) that represents the balance of nutrient removal by the streambed and turbulent transport to the streambed, where the new variable θ represents streambed porosity (equations (3b) and (3c)). This model is closely related to film and diffusion mass transfer resistance models developed by biochemical engineers, and thus, the dimensionless number ψ incorporates the dimensionless Thiele Modulus and Biot numbers (see Bailey & Ollis, 1986, p. 219). When $\psi \gg 1$, the removal efficiency is 100% ($\alpha = 1$) and the mass transfer coefficient sets an upper limit on the uptake velocity ($v_f \approx k_m$), a prediction borne out by Grant et al.'s analysis of nitrate uptake velocities measured during the second Lotic Intersite Nitrogen eXperiment field campaign (see earlier). In the opposite limit ($\alpha \approx \psi \ll 1$) the model predicts that the uptake velocity is controlled exclusively by nutrient transport and processing within the streambed ($v_f \approx \theta \sqrt{D_{\text{eff}}/\tau_{\text{reaction}}}$). While other conceptualizations of nutrient fate and transport in the hyporheic zone are possible (e.g., Azizian et al., 2017, 2015; Gomez-Velez & Harvey, 2014; Grant et al., 2014; McCluskey et al., 2016; Trauth et al., 2014), to our knowledge equation (3a) is unique in its simplicity and explicit accounting of turbulence-induced mixing above and below the bed (as well as bedform pumping, see next).

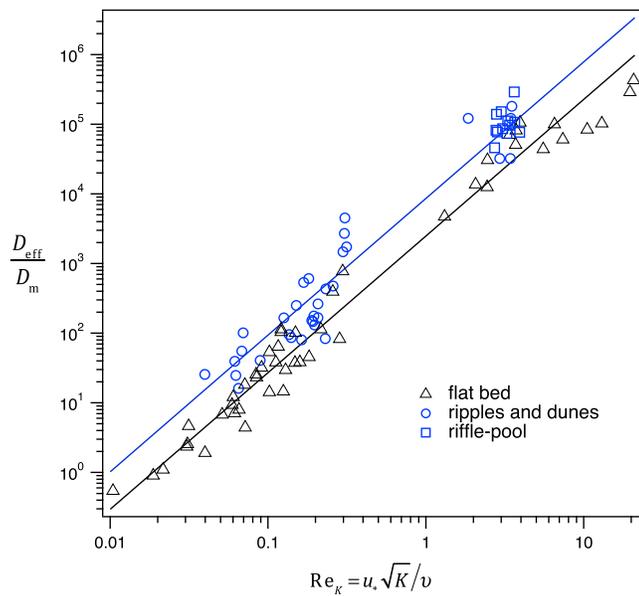


Figure 2. Observed scaling relationship between dimensionless turbulence diffusivity D_{eff} and the dimensionless permeability Reynolds number Re_K , where D_m is molecular diffusion coefficient of the tracer in water. Effective diffusivities computed according to the method of Grant et al. (2012). Primary data are from Richardson and Parr (1988), Nagaoka and Ohgaki (1990), Lai et al. (1994), Elliott and Brooks (1997), Packman et al. (2000), Marion et al. (2002), Packman and MacKay (2003), Packman et al. (2004), Ren and Packman (2004), Rehg et al. (2005), and Tonina and Buffington (2007).

Richardson & Parr, 1988). Remarkably, the same scaling law applies to flume experiments conducted with submerged bedforms and riffle-pool sequences (Figure 2). In the log-log format adopted in this figure, the regression model for D_{eff} appears as two parallel lines with a common slope (consistent with $\sim Re_K^2$) but different intercepts: one for submerged bedforms and riffle-pool sequences (blue line in the figure) and one for flat streambeds (black line). On average, the presence of bedforms and riffle-pool sequences increases the effective diffusivity above that predicted for flat beds by a factor of about 3.5; a small difference compared to the six-decade variability in D_{eff} associated with changes in the permeability Reynolds number. The fact that bedforms of all sizes (from ripples to riffle-pool sequences) increase the effective diffusivity by a constant factor of 3.5 over that predicted for a flat bed is surprising and may warrant further study, for example, by numerical simulations along the lines of Sinha et al. (2017). As alluded to earlier, a number of assumptions were employed to estimate the effective diffusivities appearing in this figure, including a depth-invariant mixing coefficient D_{eff} and homogeneous, isotropic, and infinitely deep sediment bed; see Grant et al. (2012) for a detailed discussion of the data reduction approach.

The two-step model described above for turbulent mixing across the sediment-water interface can be scaled up, for example, by incorporation into the Networks with Exchange and Subsurface Storage platform (Gomez-Velez et al., 2015). The current version of Networks with Exchange and Subsurface Storage assumes vertical exchange occurs only by bedform pumping; future versions will include estimates of vertical exchange based on equation (3a) and the scaling relationships for k_m and D_{eff} described here. Key to this approach is the availability of high-quality continental-scale hydrogeomorphic data, including reach-scale information on average grain size, stream slope, water depth, and streambed permeability (all needed to estimate k_m , D_{eff} and θ). Scaling relationships for τ_{reaction} will also need to be developed that capture the influence of stream metabolism (ecosystem respiration and gross primary production) and stream chemistry (e.g., nitrate concentration) on biogeochemical cycling of nutrients in the hyporheic zone (Hall & Tank, 2003; Mulholland et al., 2008, 2009). Finally, it will be important to validate the scaling relationships derived from flume experiments (lines in Figure 2) with comparable data from field systems. As these challenges are resolved, the inclusion of grain-scale turbulent transport into stream network models should spawn a new

Practical application of equation (3a) requires reach-specific estimates of four parameters: k_m , θ , τ_{reaction} , and D_{eff} . These are discussed in turn. Referring to equation (2b), the mass transfer coefficient depends on the Schmidt number (which, for stream nutrients, will be approximately 10^3 , depending on stream temperature) and the shear velocity, which can be estimated by force balance from the slope (S , unitless) and depth (h , units of meters) of a stream: $u_* = \sqrt{ghS}$ where $g = 9.81 \text{ m}^2/\text{s}$ is gravitational acceleration (Bagherimiyab & Lemmin, 2013). Streambed permeability (needed to estimate effective diffusivity, see below) can be estimated from median grain size with the aid of the Kozeny-Carman equation (Freeze & Cherry, 1979).

In the RSF simulations of the Mississippi River Network described earlier, Gomez-Velez et al. (2015) adopted values of $\tau_{\text{reaction}} = 1$ and 10 hr for aerobic respiration and denitrification, respectively. In general, biogeochemical reaction time scales are likely to be reach specific (Krause et al., 2017; McClain et al., 2003) and depend on, among other things, ambient stream nitrate concentrations (Mulholland et al., 2008), stream ecosystem respiration and gross primary production (e.g., Hall et al., 2009; Hall & Tank, 2003; Mulholland et al., 2009), and the availability of electron donor and anoxic conditions in the sediments (Tomasek et al., 2018; Zarnetske et al., 2011).

For flat beds and when the permeability Reynolds number satisfies the inequality $Re_K > 0.01$ (see Figures 1d and 1e), Voermans et al. (2018) reported that the effective diffusivity increases as the square of the permeability Reynolds number ($D_{\text{eff}} \sim Re_K^2$). This scaling law is consistent with previous reports of turbulent mass transfer across flat streambeds (e.g.,

generation of *grain-to-globe* simulation tools in support of nutrient management at local, continental, and global scales.

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