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Key Points:

- The magnitude of the exchange intensity was found to be dominated by the streambed-induced hydraulic head across stream order
- The spatial fragmentation of upwelling zones at the streambed interface was strongly affected by catchment-scale groundwater circulation
- Streambed-induced hydraulic head depended on the hydrostatic damping factor, dynamic amplitude factor, and depth of the hyporheic zone

Supporting Information:

- Supporting Information S1

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Fragmentation of the Hyporheic Zone Due to Regional Groundwater Circulation

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Abstract By use of numerical modeling and field observations, this work quantified the effects of catchment-scale upwelling groundwater on the hyporheic (below stream) fluxes over a wide range of spatial scales. A groundwater flow model was developed that specifically accounted for the hydrostatic and dynamic head fluctuations induced by the streambed topography. Although the magnitudes and relative importance of these streambed-induced fluxes were found to be highly sensitive to site-specific hydromorphological properties, we showed that streambed topographic structures exert a predominant control on the magnitude of hyporheic exchange fluxes in a Swedish boreal catchment. The magnitude of the exchange intensity evaluated at the streambed interface was found to be dominated by the streambed-induced hydraulic head across stream order. However, the catchment-scale groundwater flow field substantially affected the distribution of groundwater discharge points and thus decreased the fragmentation of the hyporheic zone, specifically by shifting the cumulative density function toward larger areas of coherent upwelling at the streambed interface. This work highlights the spectrum of spatial scales affecting the surface water-groundwater exchange patterns and resolves the roles of key mechanisms in controlling the fragmentation of the hyporheic zone.

1. Introduction

Rivers and groundwater have historically been considered two separate hydrological systems with different hydrodynamic and chemical characteristics (Boulton et al., 1998; Brunke & Gonser, 1997). However, during the last decades, alternatives to the decoupled approach have emerged and led to the conceptual view of a continuous exchange of water, energy, and mass between surface water and groundwater (Cardenas, 2015; Stanford & Ward, 1993). The interaction between surface water and groundwater plays an important role in water quality and water resources management (Edwards, 1998; Fleckenstein et al., 2010; Stanford & Ward, 1988; Winter et al., 1998). Moreover, the surface water-groundwater exchange controls diverse biological activities and the distribution of aerobic and anaerobic conditions within the streambed sediment (Harvey et al., 2013; Marzadri et al., 2011; Zarnetske et al., 2011), as well as the fate and transport of water-borne substances such as nutrients and organic carbon (Mulholland et al., 1997; Stonedahl et al., 2012). Therefore, a detailed understanding of this exchange occurring in the so-called hyporheic zone is an essential step in describing contaminant fate and transport (Benner et al., 1995; Fuller & Harvey, 2000; McKnight et al., 2001; Medina et al., 2002) and nutrient turnover in rivers (Birgand et al., 2007; Gomez-Velez et al., 2015; Jones & Mulholland, 2000).

Surface water-groundwater interactions are controlled by the hydromorphology of the system, occur at a wide range of spatial scales, and drive vertical and lateral hyporheic fluxes (Bencala & Walters, 1983; Boano et al., 2007; Stonedahl et al., 2013). A large number of studies have previously been performed to investigate the effect of distinct scales on the hyporheic exchange. On the stream reach scale, various stream bedforms from ripples to dunes have been shown to significantly modify the local pressure and shear stress fields that in turn drive the hyporheic flux (Bottacin-Busolin & Marion, 2010; Cardenas & Wilson, 2007a, 2007b). In other studies, larger morphological features have been considered such as sediment bars (Boano et al., 2010; Marzadri et al., 2010; Tonina & Buffington, 2007), pool-riffle structures (Harvey & Bencala, 1993; Tonina & Buffington, 2009), and stream channel meandering (Boano et al., 2006; Gomez et al., 2012; Revelli et al., 2008). On the regional scale, several experimental and numerical modeling studies have focused on the effect of groundwater flow on the hyporheic fluxes (Boano et al., 2009; Brunner et al.,

2017; Stonedahl et al., 2012). Previous investigations of flow fields induced by multiple spatial scales have estimated the total hydraulic head by combining individual topographical features to create a nested system of flow cells, in which the streambed-scale topography and the reach-scale bathymetry cause dynamic and hydrostatic head variations, respectively (Boano et al., 2009; Caruso et al., 2016; Marzadri et al., 2014). Somewhat loosely defined, the dynamic head is associated with the stagnation pressure of the flow velocity at the streambed, and the hydrostatic head reflects the elevation of the free stream surface. However, previously, the effect of hydrostatic head variations due to stream surface fluctuations have often been assumed to be negligible (Boano et al., 2009; Caruso et al., 2016; Marzadri et al., 2014), which in some cases might be a limiting assumption affecting the estimates of flow fields and the spatial extent of the hyporheic zone.

Groundwater fluxes and flow paths show strong spatial and temporal variations at different scales (Ellis et al., 2007; Krause et al., 2007). One of the important factors for estimating the groundwater flow field is the characterization of the water table. At large scales (regional to continental), water tables are generally classified as either “topography controlled” or “recharge controlled” depending on the region’s climate characteristics and aquifer properties (Gleeson et al., 2011; Haitjema & Mitchell-Bruker, 2005). Furthermore, the groundwater flow field is partly controlled by the heterogeneity of the aquifer (Conant, 2004; Storey et al., 2003; Wondzell & Swanson, 1996; Wroblicky et al., 1998). Hence, in addition to topography, the spatial heterogeneity of subsurface hydraulic conductivity plays an important role in the distribution of flow paths, especially in the location and magnitude of groundwater discharge zones across the streambed as well as the size of the hyporheic zone (Boano et al., 2008; Cardenas & Wilson, 2007c). In particular, the subsurface spatial heterogeneity affects the fragmentation of the hyporheic flow, that is, the patchiness of upwelling and downwelling regions (Salehin et al., 2004; Sawyer & Cardenas, 2009). Spatial heterogeneities of groundwater fluxes through the streambed influence stream ecosystems (Brunke & Gonser, 1997) and impact the fate and transport of contaminants between aquifers and streams (Chapman et al., 2007; Conant et al., 2004; Kalbus et al., 2007).

The aim of the present study was to determine the principal effects of catchment-scale groundwater flow on the hyporheic fluxes considering circulation over a wide range of spatial scales. In contrast to the methodology proposed by Marzadri et al. (2014) and Caruso et al. (2016), here we implemented an approach where the topography is subdivided into two spatial scales, a catchment scale and a streambed scale, which facilitates a specific emphasis on the interaction of the flow between the different ranges of spatial scale. Hence, an overarching objective of this study was to explore such interaction in terms of controlling factors for the ratio of exchange velocities arising on the two ranges of spatial scale and the associated areal fragmentation of upwelling and downwelling along the streambed. The work built on the exact spectral solution to the groundwater flow field derived by Wörman et al. (2006, 2007), which was extended in this study to include both hydrostatic and dynamic heads as drivers of the hyporheic flow. From a modeling point of view, the entire catchment topography was first used in a numerical simulation to quantify the catchment-scale groundwater flow field considering the geological heterogeneity. Second, the streambed topography and hydraulic properties were used to parameterize an analytical model that described the contribution of the hydrostatic and dynamic heads at the streambed to the hyporheic flow. Superimposing the two flow fields provided the means to analyze their combined flow field and the relative magnitudes of hyporheic fluxes, constraining hydromorphological conditions, as well as the fragmentation of hyporheic flow cells caused by upwelling and downwelling zones at the streambed interface.

2. Methodology

2.1. Material

The study was conducted with data from the 68-km² Krycklan Research Catchment (64°14′N, 19°46′E), which is located approximately 50 km northwest of Umeå in northern Sweden (Laudon et al., 2013). There are 15 observation stations in the catchment that are monitored for hydrological parameters including the discharge values of the streams (Karlsen et al., 2016). The channel morphology of the stream network has been measured and classified into 13 groups (Laudon et al., 2013). The flow velocity, depth, and width of each stream reach have been estimated using observation station data and the stream channel morphology. High-resolution topographical data exist for the entire catchment, where

Figure 1. Conceptual sketch of the total hydraulic head (H) components: streambed-induced, $H_S(x)$, and catchment-scale, $H_C(x)$, hydraulic head fluctuations and the trend in the catchment-scale water surface level, $H_T(x)$.

the resolutions for the horizontal and vertical directions are 2 m and 1 cm, respectively. The Krycklan Catchment ranges in elevation from 114 m above sea level at the lowest to 405 m above sea level at the highest point. The Quaternary deposits consist mostly of till and sand with depths up to 40 m beneath the bed surface (Jutebring Sterte et al., 2018). The mean annual precipitation in the Krycklan Catchment (from 1981 to 2007) is 614 mm/year, and the evaporation rate is half of this amount (Haei et al., 2010). In addition, approximately 35–50% of precipitation comes from snowfall (Oni et al., 2013). The Krycklan Catchment includes seven different soil types (supporting information Figure S1), and hydraulic conductivity values have been estimated for each soil type (Jutebring Sterte et al., 2018).

2.2. Modeling Framework

This study includes a numerical modeling study covering multiple spatial scales of the groundwater flow below streams as well as statistical methods for relating the results to controlling factors. The flow problem was divided into two scales, a catchment scale and a streambed scale, in order to facilitate a comparison of their relative importance. Comsol Multiphysics modeling software was used to numerically solve the 3-D groundwater equation on the catchment scale, while a 3-D exact solution was utilized at the streambed scale. Here we present a general outline of the main methodological steps, while additional details are given in Appendix A.

The governing equation for 3-D groundwater flow under steady state condition is

$$\frac{\partial}{\partial x} \left(-K_x \frac{\partial H}{\partial x} \right) + \frac{\partial}{\partial y} \left(-K_y \frac{\partial H}{\partial y} \right) + \frac{\partial}{\partial z} \left(-K_z \frac{\partial H}{\partial z} \right) = 0 \quad (1)$$

where H (m) is the total hydraulic head and K_x (m/s), K_y (m/s), and K_z (m/s) are hydraulic conductivities in the x , y , and z directions, respectively. Here x (m) and y (m) are defined on the horizontal plane, and z (m) is the vertical direction (positive upward). Subsurface flow in saturated porous media can be evaluated using Darcy's law: $q = [U, V, W] = -K \nabla H$, where q (m/s) is the Darcy velocity vector; U (m/s), V (m/s), and W (m/s) are the velocities in x , y , and z directions, respectively; and ∇ is the nabla operator. Due to the linearity of equation (1), the principle of superposition applies, and the total hydraulic head $H(x,y,z)$ (m) can be separated into three different terms according to equation (2):

$$H(x, y, z) = H_S(x, y, z) + H_C(x, y, z) + H_T(x, y) \quad (2)$$

where $H_S(x,y,z)$ (m) is the streambed-induced hydraulic head fluctuation, $H_C(x,y,z)$ (m) is the catchment-scale hydraulic head fluctuation, and $H_T(x,y)$ (m) is the catchment-scale trend in water surface elevation (Figure 1).

Hence, equations (1) and (2) allow evaluating the different hydraulic head components that govern the groundwater flow field separately and on different spatial scales. In the multiscale approach applied here,

we separate the topographic domain into a catchment scale, $H_T + H_C$, that drives the catchment-scale groundwater flow field (section 2.3) and a streambed scale, H_S , that represents the streambed-induced flow field (section 2.4).

2.3. Catchment-Scale Groundwater Flow Modeling

As part of defining the top boundary condition for the catchment-scale groundwater flow model, specifically to represent the connection between the shape of the groundwater table and the landscape topography, we use the dimensionless water table ratio suggested by Haitjema and Mitchell-Bruker (2005) according to

$$WTR = \frac{\Delta h}{d} = \frac{RL^2}{mKd} \quad (3)$$

where Δh (m) is the groundwater mounding; d (m) is the maximum terrain rise from the average surface water level; R (m/s) is the average recharge rate; L (m) is the average distance between hydrological boundaries, m (—) is a dimensionless factor of 8 or 16, depending on whether the flow is one dimensional or radial symmetric, respectively; K (m/s) is the average horizontal hydraulic conductivity; and D (m) is the aquifer thickness. For $WTR = \frac{\Delta h}{d} \geq 1$, the water table can be classified as topography controlled, whereas for $WTR = \frac{\Delta h}{d} < 1$, it can be classified as recharge controlled. The evaluated water table ratio for the Krycklan Catchment was found to be greater than unity (supporting information Table S1), which implies that the groundwater table follows the topography. Hence, the top boundary condition was defined by the landscape elevation.

In the present study, the groundwater flow simulations were performed for heterogeneous and isotropic hydraulic conductivities according to the soil type and previously estimated values of hydraulic conductivities in the study domain (Jutebring Sterte et al., 2018). Furthermore, the hydraulic conductivity decays exponentially with depth for most geological features, especially in bedrock (Saar & Manga, 2004). Hence, the catchment was considered to be stratified into two horizontal layers representing the Quaternary deposits and the bedrock. The hydraulic conductivity $K(z)$ for each individual layer is described according to

$$K_i(z) = K_{(S \text{ or } D), \text{top}} e^{\frac{(z - z_{\text{top}, i})}{\delta_i}} \quad (4)$$

where the subscript i denotes the layers, $z_{\text{top}, i}$ (m) is the elevation coordinate at the top of layer i ; $K_{S, \text{top}}$ (m/s) and $K_{D, \text{top}}$ (m/s) are the hydraulic conductivity at the top surface of Quaternary deposit and bedrock layer, respectively, and δ (m) is the characteristic depth scale (skin depth). Saar and Manga (2004) showed that the skin depth varies in the range from 200 to 300 m down to a depth $z < 800$ m. In this study, we adopted $\delta = 250$ m.

The catchment-scale groundwater flow field was quantified using a 2-m resolution digital elevation model file representing the topographic surface. A calculation mesh with nonuniform size that increased in resolution toward the groundwater surface ranging between 17 and 403 m was used to reduce the computational cost of the analysis. A flat no-flow horizontal plane at $z = -D_T$ was considered the bottom surface boundary condition. Moreover, D_T was defined as the sum of Quaternary deposit and bedrock thicknesses that vary spatially within the domain (D_T varies in a range from 114 to 405 m). The boundary conditions were stated as $H(x, y, z = 0) = Z_B(x, y)$ and $\frac{\partial H}{\partial z} \Big|_{z = -D_T} = 0$, where $H(x, y, z = 0)$ (m) and $Z_B(x, y)$ (m) are the hydraulic head at the phreatic surface and the ground surface elevation, respectively, at any position in the x - y plane. The hydrostatic hydraulic head was assumed to prevail as the top boundary condition over the entire catchment. A constant hydraulic head was also assumed at the vertical side ($H(x, y, z) \Big|_{x=0, L_x \text{ and } y=0, L_y} = Z_B(x, y)$, where L_x and L_y are the catchment length in the x and y directions, respectively).

Within the catchment, the vertical flow velocities at the streambed were evaluated in four different stream orders, 1, 2, 3, and 5 (Figure 2: stream order 1 [green line], 2 [yellow line], 3 [purple line], and 5 [brown line]). Furthermore, 1 to 5 stream reaches were selected for each stream order, and each individual stream reach was subsequently subdivided into 5-m $\times$ 5-m regions for which the vertical flow velocities were analyzed. Thus, in total, the catchment-scale flow analysis was evaluated for 130 first-order, 110 second-order,

Figure 2. Map showing the Krycklan catchment boundaries as well as topography, selected stream orders (colored lines), and the sequential $5 \times 5\text{-m}^2$ regions that form a representative stream reach (inset) used to evaluate the catchment-scale flow (the colored lines indicate the selected streams; however, it should be noted that only subsections (reaches) of these streams were considered in the analysis). Moreover, the position of the selected $100 \times 100\text{-m}^2$ regions for rescaling of the streambed topography (red squares) and available discharge stations (green stars) used in the streambed-induced model are also shown.

110 third-order, and 50 fifth-order stream regions (for an overview of the combinatorial sampling procedure, see Figure 2 and Table 1).

2.4. Streambed-Induced Flow Modeling

The streambed topography was not available for the selected study site and was estimated by fractal rescaling of the surrounding landscape topography (Tóth, 1963; Wörman et al., 2007; Zijl, 1999). This rescaling was based on the digital elevation model where 20 different regions ($100\text{-m} \times 100\text{-m}$ regions surrounding the stream) from the entire Krycklan Catchment were chosen randomly (Figure 2) to provide an acceptable

Table 1
Schematic Description of the Combinatorial Sampling Method

Stream order	Reach	Number of $5 \times 5\text{-m}^2$ regions in the catchment-scale model	Number of $5 \times 5\text{-m}^2$ regions in the streambed-scale model	Total number of $5 \times 5\text{-m}^2$ regions for each stream reach	Total number of $5 \times 5\text{-m}^2$ regions for each stream order	Total number of $5 \times 5\text{-m}^2$ regions for the whole analysis
First	No. 1	50	20	1,000	2,600	8,000
	No. 2	20	20	400		
	No. 3	20	20	400		
	No. 4	20	20	400		
	No. 5	20	20	400		
Second	No. 1	50	20	1,000	2,200	
	No. 2	20	20	400		
	No. 3	20	20	400		
	No. 4	20	20	400		
Third	No. 1	50	20	1,000	2,200	
	No. 2	20	20	400		
	No. 3	20	20	400		
	No. 4	20	20	400		
Fifth	No. 1	50	20	1,000	1,000	

degree of representativity for the local topography. Thus, 20 rescaled streambed topographies (for details, see Appendix A) were characterized by the coefficients a and b in a power law relationship between amplitude, A (m), and wavelength, λ (m), in the form $A = a \times \lambda^b$ (Table S2 in supporting information).

In streams, the shape and configuration of topographical structures (i.e., different bedforms) induce fluctuations in the hydraulic head along the streambed boundary (Cardenas, 2008; Elliott & Brooks, 1997a; Thibodeaux & Boyle, 1987). As shown in the following sections, both the dynamic and hydrostatic components of the hydraulic head can be derived from the streambed topography. The hydrostatic pressure in a fully stagnant flow is due to the gravity effect over the entire depth of water, but for flowing water, the separation of hydrostatic and dynamic components of pressure at the bed surface is based on an operational definition. Hence, the hydrostatic head component is defined as the stream water surface elevation, whereas the dynamic head component is given as a product between the velocity head and the degree of stagnation (i.e., translation of the velocity head into pressure head). The hydraulic driver in the vertical direction across the streambed interface is further represented by the fluctuations in the hydrostatic head in relation to a trend value and the dynamic head difference between upstream and downstream sides of bedforms. The fluctuations in the streambed-induced total hydraulic head, H_S (m), in relation to a longitudinal trend value are given by

$$H_S(x, y, z = 0) = H_{S,S}(x, y, z = 0) + H_{S,D}(x, y, z = 0) \quad (5)$$

where $H_{S,S}$ (m) and $H_{S,D}$ (m) are the fluctuations in the hydrostatic and dynamic components of the streambed-induced hydraulic head distribution, respectively. The hydrostatic head fluctuations are represented by fluctuations in the stream water elevation, which are different from the fluctuations in the streambed topography. By introducing a damping of the bedform fluctuations for estimating the stream water fluctuations, the expression of streambed-induced hydrostatic head, $H_{S,S}$, can be stated as

$$H_{S,S}(x, y, z = 0) = \sum_{j=1}^N \sum_{i=1}^N C_{\text{damp}}(Fr, \lambda_i) (A_{i,j})_{\text{BS}} \sin(k_i x) \cos(k_j y) \quad (6)$$

where $(A_{i,j})_{\text{BS}}$ (m) are the amplitude coefficients determined from the streambed topography (Appendix A) and $C_{\text{damp}}(Fr, \lambda)$ (–) is the ratio of the surface water fluctuation, Z_{WS} (m), to the streambed topography fluctuation, Z_{BS} (m), and ranges from zero (complete damping, Z_{WS} has no fluctuations) to one (no damping, Z_{WS} is an exact replica of Z_{BS}). Thus, the amplitude coefficients for the water surface fluctuation are defined by $(A_{i,j})_{\text{WS}} = C_{\text{damp}}(Fr, \lambda_i) (A_{i,j})_{\text{BS}}$. Subsequently, we argue that C_{damp} depends on both the wavelength, λ , and the hydraulic regime represented by the Froude number, $Fr = \frac{v_f}{\sqrt{gD_w}}$, where v_f (m/s) is the stream flow velocity, g (m/s²) is the gravitational acceleration, and D_w (m) is the water depth. Thus, the damping factor can be separated (or factorized) into the following form:

$$C_{\text{damp}}(Fr, \lambda) = \alpha(Fr) \beta(\lambda) C_{\text{damp}}(Fr_{\text{ref}}, \lambda_{\text{ref}}) \quad (7)$$

where the coefficients $\alpha(Fr)$ and $\beta(\lambda)$ (–) account for the alterations in flow regime (–) and wavelengths (m), respectively, and the subscript ref indicates a reference region (–).

Morén et al. (2017) performed an investigation of C_{damp} in a small Swedish stream (Tullstorps Brook), and their results were considered here as a reference. They measured streambed topography and water surface elevation for a 500-m stream reach, and based on their data, the ratios of power spectra of the water surface and bed topography elevations, R_p , which is equal to the square of the amplitude coefficient ratio, that is, $R_p = P_{\text{WS}}/P_{\text{BS}} = (A_{i,j})_{\text{WS}}^2 / (A_{i,j})_{\text{BS}}^2 = C_{\text{damp}}^2(Fr_{\text{ref}}, \lambda_{\text{ref}})^2$, were determined. The observations showed that for the longest wavelengths, R_p was close to 1, indicating that the water surface fluctuations closely followed the fluctuations in the streambed topography. Moreover, the results displayed a consistent decrease in R_p (i.e., increase in damping of the surface water fluctuations compared to the bed surface fluctuations) with decreasing wavelength. However, for $\lambda < 24$ m, the Morén et al. (2017) data set shows significant noise, and therefore, an exponential function was used to extrapolate values of R_p for $\lambda < 24$ m. The function was fitted to the observed data in the range $24 < \lambda < 660$ and bounded by the fact that the water surface perturbations are zero at $\lambda = 0$ m (black curve in Figure 3, denoted

Figure 3. Ratio of water surface power spectrum to bedform power spectrum, $R_p = P_{WS}/P_{BS}$. Colored dots represent observed values from Morén et al. (2017), and the black line is the fitted value as a function of wavelength. (a) Fitted exponential function to the observed data. (b) Four exponential curves describing R_p , in which the dashed black line is the extrapolated exponential trend of R_p of the observed data as a function of wavelength. The colored dashed lines are the three additional exponential curves representing R_p as functions of wavelength used to investigate the sensitivity of the hydrostatic head damping factor to the streambed-induced flow field.

$R_{P,3}(\lambda)$). Moreover, to account for the uncertainty in the determination of the damping factor, three additional exponential curves describing R_p as a function of wavelength were used to investigate the sensitivity of C_{damp} to the streambed-induced flow field (colored dashed lines in Figure 3b, denoted $R_{P,1}(\lambda)$, $R_{P,2}(\lambda)$, and $R_{P,4}(\lambda)$). These curves were constrained by the values of $R_{P,1}(\lambda = 5) = 0.001$, $R_{P,2}(\lambda = 5) = 0.01$, $R_{P,4}(\lambda = 5) = 0.5$, and $R_{P,1-4}(\lambda = 0) = 0$. The scenario $R_{P,4}(\lambda)$ reflected the upper boundary value for the damping factor equal to $R_{P,3}(\lambda = 24)$ from the observations in Morén et al. (2017), while $R_{P,1}(\lambda)$ and $R_{P,2}(\lambda)$ reflected values of R_p at $\lambda = 5$ m one and two orders of magnitude lower than the observations. Furthermore, from the theory of open channel flow during steady flow conditions, $dD_{WS}/dZ_{BS} = (dD_{WS}/dx)/(dZ_{BS}/dx) = 1/(Fr^2 - 1)$, which implies decreasing water surface (D_{WS}) fluctuations in comparison to bed surface (Z_{BS}) fluctuations with decreasing Froude number. One can assume that $C_{damp} \propto dD_{WS}/dZ_{BS}$, which leads to $\alpha(Fr) = \frac{C_{damp}(Fr, \lambda)}{C_{damp}(Fr_{ref}, \lambda)} = \frac{Fr_{ref}^2 - 1}{Fr^2 - 1}$. Fr_{ref} was determined from a stream tracer test performed in connection with the investigation of surface water and streambed fluctuations (Morén et al., 2017) and was found to be $\frac{0.055}{\sqrt{9.81 \times 0.24}} = 0.036$. Thus, this value leads to the following approximation of the damping factor:

$$C_{damp}(Fr, \lambda) = \left(\frac{-0.9987}{Fr^2 - 1} \right) \left(\sqrt{R_p(Fr_{ref}, \lambda_{ref})} \right) \quad (8)$$

The dynamic head distribution, which is the head difference between the upstream and downstream sides of streambed-scale morphology such as ripple- or dune-shaped bedforms, can be represented by a sinusoidal pressure distribution in space (Elliott & Brooks, 1997a, 1997b). The streambed induced dynamic head, $H_{S,D}$, can be derived from the rescaled bedform oscillations according to

$$H_{S,D}(x, y, z = 0) = \frac{h_m}{\sigma_{S,B} \sqrt{2}} \sum_{j=1}^N \sum_{i=1}^N (A_{ij})_{BS} \sin(k_i x) \cos(k_j y) \quad (9)$$

where $\sigma_{S,B}$ (m) is the standard deviation of the streambed elevation and h_m (m) is the amplitude of the velocity head variation. Furthermore, the equation used to estimate the amplitude of the head variation for a single bedform can be stated as (Fehlman, 1985):

$$h_m = 0.28 \left(\frac{v_f^2}{2g} \right) \begin{cases} \left(\frac{Z_{BM}/D_w}{0.34} \right)^{3/8} & Z_{BM}/D_w \leq 0.34 \\ \left(\frac{Z_{BM}/D_w}{0.34} \right)^{3/2} & Z_{BM}/D_w \geq 0.34 \end{cases} \quad (10)$$

where Z_{BM} (m) is the bedform height. The bedform height is defined from the standard deviation of streambed elevation measurements, $2\sigma_{S,B}$ (Elliott & Brooks, 1997a, 1997b), and for a perfectly sinusoidal topography, it becomes (Stonedahl et al., 2010): $Z_{BM} = 2\sqrt{2}\sigma_{S,B}$. The flow velocities, v_f , and flow depth, D_w , for the analyzed stream reaches were obtained from the measured and estimated stream discharge values in the study domain (Figure 2). The hydrostatic damping and dynamic amplitude factors differed between the stream reaches due to the variations in flow velocity, flow depth and Froude number (for values representing the different reaches, see Table S3 in supporting information). Therefore, the total hydraulic head distribution at the streambed interface induced by the streambed (equation (5)) can be rewritten as

$$H_S(x, y, z = 0) = \sum_{j=1}^N \sum_{i=1}^N \left[\left(\frac{h_m}{\sigma_{S,B}\sqrt{2}} \right) + C_{damp}(Fr, \lambda_i) \right] (A_{ij})_{BS} \sin(k_i x) \cos(k_j y) \quad (11)$$

Following the work of Wörman et al. (2006) and assuming a homogenous subsurface and a no-flow condition at the depth ε (m), the streambed-induced hydraulic head can be expressed as a Fourier series of the streambed fluctuations modulated by the hydrostatic damping factor and dynamic head amplitude factor according to

$$H_S(x, y, z) = \sum_{j=1}^N \sum_{i=1}^N \left[\left(\frac{h_m}{\sigma_{S,B}\sqrt{2}} \right) + C_{damp}(Fr, \lambda_i) \right] (A_{ij})_{BS} \sin(k_i x) \cos(k_j y) \times \frac{e^{\left(\sqrt{(k_i^2 + k_j^2)}z\right)} + e^{\left(\sqrt{(k_i^2 + k_j^2)}(-2\varepsilon - z)\right)}}{1 + e^{\left(-2\sqrt{(k_i^2 + k_j^2)}\varepsilon\right)}} \quad (12)$$

In equation (12), the depth to the no-flow boundary, ε , confines deeper flow paths and thus reduces the effects of larger topographical scales on the flow field, an effect referred to as shielding. In our study, two different depths ($\varepsilon = 4$ m and $\varepsilon = 0.1$ m) were considered to investigate its influence on the streambed-induced flow field. Finally, equation (12) was used in Darcy's law to calculate the streambed-induced flow field assuming an isotropic streambed-scale hydraulic conductivity of 10^{-4} (m/s). This value lies in the range previously reported for streambed sediment (Wörman et al., 2002; Morén et al., 2017).

2.5. Spatial Representation Using a Combinatorial Sampling Approach

A combinatorial sampling procedure for vertical fluxes across the streambed was used to represent the catchment-scale, streambed-induced, and superimposed models of the subsurface flow field. The purpose of this procedure was to provide a statistically representative sampling of the resulting hyporheic flow field in different stream orders throughout the catchment. Hence, 14 stream reaches representing first-, second-, third-, and fifth-order streams were selected from the catchment-scale model. Each stream reach contained 50 or 20 regions with an area of 5×5 m² (Figure 2) containing 2,500 values of vertical velocity at the streambed interface, $W(z=0)$ (i.e., with the model resolution of 0.1×0.1 m²). Moreover, 20 different rescaled topographies were used to calculate 20 statistical representations (or realizations) of the streambed-induced (local) flow field that were combined with the catchment-scale flow field in all 400 regions (5×5 m²) along each of the 14 stream reaches (see section 2.4 and Table S2 in supporting information). In this process, the vertical velocity at the streambed interface, $W_S(z=0)$, of each streambed model was superimposed on the vertical velocity of the catchment-scale model, $W_C(z=0)$. Consequently, in this study, the vertical velocities at the streambed interface from $400 \times 20 = 8,000$ combinations of the catchment- and streambed-scale models were used to generalize the results (Table 1). Due to this combinatorial sampling, different averaging methods of the vertical velocity at the streambed interface were introduced to facilitate the analyses and the presentation of the results on different spatial scales (Figure 4). The areal average of the absolute value

Figure 4. Schematic sketch of symbol list for different types of averages of vertical velocity at the streambed interface, W . The blue rectangles represent $5 \times 5\text{-m}^2$ regions with a resolution of $0.1 \times 0.1\text{ m}^2$; that is, each region contains 2,500 individual values of W .

of the vertical flow velocity at the streambed interface $\overline{|W|}$ is a measure of the exchange intensity and thus is a key metric defining surface water-groundwater interactions. The exchange intensity was expressed as stream reach averages, $\widehat{|W|}$, and subsequently also as stream order averages $\langle |W| \rangle$ to facilitate comparisons over larger spatial scales.

2.6. Statistical Generalization of Results

To determine the principal effects of catchment-scale groundwater flow on the hyporheic fluxes, we performed multivariate statistical analyses to identify key parameters controlling the subsurface flow field. Thus, we aimed to link catchment characteristics to the groundwater discharge pattern, which would support a generalization of the results. In this study, six independent parameters were used to assess the variations in the catchment-scale groundwater flow field: stream order, SO (–); average local slope, S (–); average landscape elevation, e (m); Euclidean distance from the center of each $5 \times 5\text{-m}^2$ region to the watershed outlet, D_S (m); hydraulic conductivity of the soil, K (m/s); and Quaternary deposit thickness, Z_Q (m). All independent parameters were determined for the 8,000 regions ($5 \times 5\text{ m}^2$) and averaged over their associated stream reach. Principal component analysis (PCA) was used to assess the statistical independence (and covariance) of the 14 stream reach averaged independent parameters. Moreover, the PCA was applied on non-normalized and normalized (with $\frac{C_i - \mu_i}{\sigma_i}$, and with $\frac{C_i}{\mu_i}$) forms of the selected independent parameters, where C_i is an independent parameter and μ_i and σ_i are the mean value and standard deviation of each independent parameter i for all the selected reaches, respectively. Previous studies suggested that two to three watershed parameters stand for the dominant loadings to the principal components (Onesti & Miller, 1974; Sharma et al., 2015). The investigation here adds subsurface parameters, such as hydraulic conductivity and Quaternary deposit thickness, which are believed to be important for surface water-groundwater interaction.

Moreover, in an attempt to quantify the variation in the hyporheic exchange intensity averaged over each stream order analysis, we applied a regression model based on the independent parameters from the PCA

analysis. A power law statistical model was used to explain the dependent parameter, $\frac{\widehat{|W_C|}}{\widehat{|W_S|}}$, by the independent parameters C_1 – C_n according to

$$\frac{\widehat{|W_C|}}{\widehat{|W_S|}} = \alpha_1 C_1^{\alpha_2} C_2^{\alpha_3} \dots C_n^{\alpha_{n+1}} \quad (13)$$

where independent parameters C_1 – C_n are subsequently referred to characteristic parameters and α (–) is the coefficient determined from a multiple linear regression of the logarithmic transform of equation (13). In

Figure 5. Autocorrelation plots for the areal mean value of the vertical velocity, $\overline{W}_C$, in the individual stream reaches (dashed lines) as well as average values for the different stream orders (solid lines). It should be noted that study area contained only one fifth-order stream reach, preventing averaged values for this stream order.

addition, the streambed-induced vertical velocity, W_S , was determined for the case of $R_{P,3}$ (see Figure 3b), and the hyporheic depth was equal to 0.1 m.

2.7. Fragmentation of the Hyporheic Zone

In addition to the exchange intensity, $|\overline{W}|$, a key metric defining the surface water-groundwater exchange is the spatial distribution of upwelling and downwelling zones at the streambed interface. This is referred to as the fragmentation of the hyporheic zone, defined as the distribution of coherent upwelling or downwelling areas. Coherent in this context is defined as an area where all flow is only upward or downward. In the numerical simulation, this behavior was assessed on an orthogonal mesh where flow velocity values were compared only in orthogonal directions, not diagonal. Based on this definition, the 2,500 values of W were organized into upwelling and downwelling areas that have a binary distribution in each $5 \times 5\text{-m}^2$ regions. These values can be used to calculate the frequency of occurrence for different sizes of surface areas with coherent upwelling and downwelling, here represented by the cumulative distribution function (CDF). Hence, based on the ensemble of identified distributions of both coherent upwelling and downwelling areas, we could numerically derive two CDFs for the two classes of coherent areas. Although not theoretically inferred, the structure of the geomorphologic characteristics driving the subsurface flow field generally results in an

inverse relationship between the CDFs of upwelling and downwelling areas; that is, the more fragmented the areas of upwelling are, the less fragmented the areas of downwelling. This feature is a new metric used in hyporheic characterization that may be important for stream ecology as well as behavior of, for example., nutrient decomposition in the hyporheic zone.

3. Results

3.1. Groundwater Flow Field

The areal mean values of vertical velocity at the streambed interface originating from the catchment-scale groundwater model, $\overline{W}_C$, were evaluated for all $5 \times 5\text{-m}^2$ regions along the selected stream reaches (see Figure 2). The spatial autocorrelation of $\overline{W}_C$ as a function of the lag distance for the investigated stream reaches indicated that the distances of statistically correlated $\overline{W}_C$ were considerably larger than the cell size ($5 \text{ m} \times 5 \text{ m}$; Figure 5). The spatial autocorrelation was calculated for all the stream reaches (Figure 5, dashed lines) together with stream order averages (Figure 5, solid lines). It should be noted that study area contained only one fifth-order stream reach, preventing averaged values for this stream order. The cyclic behavior in the autocorrelation graph reflected the general existence of both upwelling ($\overline{W}_C > 0$) and downwelling ($\overline{W}_C < 0$) regions in all single stream reaches. One can define the correlation length as the lag distance when the autocorrelation reaches zero for the first time. The fifth-order stream had the longest correlation distance, which may indicate a particularly high influence from the structured large-scale groundwater flow in the stream. The correlation lengths for first-, second-, and third-order streams were found to be much shorter than that for the fifth-order stream, reflecting more spatially variable patterns of upwelling and downwelling.

The catchment-scale model showed that the values or the areal mean (based on the $5 \text{ m} \times 5\text{-m}$ regions) of the absolute value of the vertical velocity at the streambed interface, $|\overline{W}_C|$, differed both within and between stream orders (Figure 6). If averaged over stream order $|\widehat{\overline{W}_C}|$, the results indicated almost one order of magnitude larger velocities in the first-order streams than in higher-order streams (Figure 6, black dashed line). This higher intensity in groundwater exchange in the headwater streams reflected both their relatively higher elevations and steeper landscape gradients compared to those of lowland streams. The results showed no clear difference in $|\widehat{\overline{W}_C}|$ for the second-, third- and fifth-order streams, but it should be noted that the

Figure 6. Box and whisker plot of the absolute values of the areal mean values of the vertical component of catchment-scale groundwater velocities at the streambed interface $|\overline{W_C}|$ in $5 \times 5\text{-m}^2$ areas characterized by different stream orders.

median value of $|\overline{W_C}|$ increased from second- to fifth-order streams. The reason for the difference in mean and median pattern was due to differences in the distribution of $|\overline{W_C}|$. Although the entire range of $|\overline{W_C}|$ covered almost 3 orders of magnitude for first- through third-order streams, the interquartile range of $|\overline{W_C}|$ was found to be smaller for that of the first-order streams. This result indicated a lower statistical dispersion in $|\overline{W_C}|$ for the headwater streams compared to the second- and third-order streams and may reflect a greater homogeneity in controlling factors. It should be noted that the number of reaches evaluated in the study differs between stream orders, especially for the fifth-order stream that consisted of only one stream reach.

3.2. Locally Induced Flow Field

The streambed-induced flow fields were quantified for the investigated stream reaches using the 20 statistical representations of streambed topography and evaluated using the areal mean values of the absolute vertical flow velocity at the streambed interface, $|\overline{W_S}|$. Moreover, the relative importance of the dynamic and hydrostatic head gradients for the streambed-induced flow fields varied not only with stream reach due to differences in streamflow velocity, stream water depth, and Froude number (Table S3 in supporting information) but also with assumed damping factors and shielding effects caused by the depth of the hyporheic zone (Figure 7). In particular, for the case of $R_{P,4}$ (see Figure 3), the hydrostatic component dominated the total streambed-induced vertical velocity $|\widehat{W_S}|$. Thus, during these conditions, $|\widehat{W_S}|$ values were found to be approximately constant for all stream orders (Figure 7, bottom panels) since the hydrostatic head was based on the same 20 representations of the streambed topography for all stream reaches. On the other hand, $R_{P,2}$, which represents a higher damping of the surface water fluctuations, led to reduced contributions from the hydrostatic vertical velocity, $|\widehat{W_S}|_{\text{Hydrostatic}}$, and thus, a higher relative contribution of the dynamic vertical velocity, $|\widehat{W_S}|_{\text{Dynamic}}$, to the total streambed-induced vertical velocity $|\widehat{W_S}| = |\widehat{W_S}|_{\text{Dynamic}} + |\widehat{W_S}|_{\text{Hydrostatic}}$ (Figure 7, upper panels). $R_{P,2}$ had a specifically large impact on the first-order stream reaches. In these reaches, with high stream velocities and small stream water depths (Table S3 in supporting information), the contribution of $|\widehat{W_S}|_{\text{Dynamic}}$ exceeded that of $|\widehat{W_S}|_{\text{Hydrostatic}}$ and dominated the exchange intensity. $R_{P,1}$, representing an even higher damping of the surface water fluctuations, led to $|\widehat{W_S}|_{\text{Dynamic}} \geq |\widehat{W_S}|_{\text{Hydrostatic}}$.

Figure 7. Box and whisker plots of the areal mean values of the vertical components of streambed-scale induced flow velocities in $5 \times 5\text{-m}^2$ areas from different stream orders assuming four different scenarios of the damping of the surface water fluctuations and depth of the hyporheic zone: (a) $R_{P,2}$ and $\varepsilon = 0.1$ m, (b) $R_{P,2}$ and $\varepsilon = 4$ m, (c) $R_{P,4}$ and $\varepsilon = 0.1$ m, and (d) $R_{P,4}$ and $\varepsilon = 4$ m.

thus highlighting that during these conditions, both the hydrostatic and dynamic head were highly important for the exchange intensity of the streambed-induced flow field (Figure S3, upper panels in supporting information). The damping factor based on the Morén et al. (2017) experiment, $R_{P,3}$, displayed equal contributions of hydrostatic and dynamic head in first-order stream reaches, whereas the contribution of hydrostatic head was greater than those of dynamic heads in other stream order reaches (Figure S3, bottom panels in supporting information). Furthermore, a reduced depth of the hyporheic zone led to an increased shielding effect, which resulted in decreased values of both the hydrostatic and dynamic heads and therefore also reduced exchange intensities (Figure 7).

3.3. Statistical Assessment of Characteristic Parameters Controlling Groundwater Flow Fields

PCA was used to identify the internal covariance among the independent parameters, and it was found that two principal components explained 79% and 87% of the variation in nonnormalized and normalized (with

Figure 8. Principal component analysis results using stream reach average for each characteristic parameter for (a) the nonnormalized independent parameters and (b) the independent parameters normalized by its mean and standard deviation $(C_i - \mu_i)/\sigma_i$.

$(C_i - \mu_i)/\sigma_i$ independent parameters, respectively (Figure S4 in supporting information). When the analyses were done using nonnormalized parameters, the variation in the parameters could be explained using only two independent parameters, which formed the first and second principal components: distance to the outlet, D_s , and elevation, e (Figure 8a). Since the independent parameters had substantially different parameter ranges, the PCA was also carried out using normalized independent parameters to improve data integrity (Figure 8b and Figure S5 in supporting information). The results using normalized independent parameters displayed a more complex internal covariance pattern and suggested that all six parameters were needed to explain the internal variance among the different characteristic parameters (Table S4 in supporting information).

The results from the parametric statistical model (equation (13)) suggested that the best power law statistical model describing the variation in $|\widehat{W}_c|/|\widehat{W}_s|$ was obtained by using two independent parameters of the six selected typical parameters: the stream order, SO, and elevation, e , (see Figure S6 and Table S5 in supporting information). The regression constants connected to these parameters were found to be negative in the entire confidence interval using a 0.15 significance level), hence suggesting a slight negative dependence

of $|\widehat{W}_c|/|\widehat{W}_s|$ on both stream order and elevation. However, the fitted power law trend was found to be relatively weak ($R^2 = 0.26$) and associated with low confidence (P value = 0.19; Table 2). Including more independent parameters in the regression analysis led to higher R^2 values but even lower confidence (higher P value) in the power law trend (see Table S5 in supporting information).

3.4. The Effect of the Catchment-Scale Groundwater Flow on the Hyporheic Zone

Due to the linearity in the governing equations (equation (1)), the catchment-scale groundwater flow field can be superimposed on the streambed-induced flow field, which facilitates a study of the effect of the catchment-scale groundwater flow on the hyporheic fluxes. Two parameters were used to quantify the effect of such superposition: (i) the magnitude of the exchange intensity at the streambed interface, $|\widehat{W}|$, and (ii) the fragmentation of areas with coherent upwelling and downwelling zones at the streambed interface. The latter describes the breaching of

Table 2
Results From the Power Law Regression Analysis Using Only Stream Order and Elevation as the Characteristic Parameters (Code 12 in Table S5 in Supporting Information)

$ \widehat{W}_c / \widehat{W}_s $		Confidence intervals (significance level = 0.15)		
Coefficient	Definition	Value	Min	Max
$\log(\alpha_1)$		30.5	1.88	59.11
α_2	Power of stream order (SO)	−3.05	−5.49	−0.61
α_3	Power of elevation (e)	−6.02	−11.03	−1.02
R^2		0.26		
P value		0.19

Figure 9. Box and whisker plots of the superimposed means of the areal mean values of the absolute vertical velocity in $5 \times 5\text{-m}^2$ areas from different stream orders assuming the following: (a) $R_{P,2}$ and $\varepsilon = 0.1\text{ m}$, (b) $R_{P,2}$ and $\varepsilon = 4\text{ m}$, (c) $R_{P,4}$ and $\varepsilon = 0.1\text{ m}$, and (d) $R_{P,4}$ and $\varepsilon = 4\text{ m}$.

spatially extended hyporheic exchange patterns and is defined as a shift of the CDF of the area, $A\text{ (m}^2\text{)}$, of coherent upwelling and downwelling zones ($W > 0$ and $W < 0$) toward smaller areas. The analyses were made on all $5 \times 5\text{-m}^2$ regions (Table 1) and revealed that the stream order averaged velocity, $|\bar{W}|$, of the groundwater flow field was small compared to the streambed-induced fluxes for all evaluated stream orders, regardless of the assumed value of the damping factor (Figure 9 and Figure S7 in supporting information). However, for the cases of $R_{P,1}$ and $R_{P,2}$, reflecting suppressed surface water fluctuations, the analyses displayed reduced locally (streambed)-induced hydrostatic heads and lower $|\bar{W}_S|$ and thus higher relative contributions of the catchment-scale groundwater flow. This result was evident as partly overlapping distributions of $|\bar{W}_S|$ and $|\bar{W}_C|$, especially when the depth of the hyporheic zone was large (Figures 9a and 9b and Figures S7a and S7b in supporting information), which highlights the existence of

Figure 10. Cumulative distribution function of the size distribution of spatially coherent upwelling and downwelling areas for the fifth-order stream assuming $R_{P,3}$ and $\varepsilon = 4$ m.

regions where groundwater upwelling dominated the hyporheic flux. Increasing the hyporheic depth resulted in lower $|W_s|$ and therefore increased relative effect of the groundwater flow field (Figure 9).

The CDF of the sizes of coherent upwelling and downwelling areas was computed for all 5×5 -m² regions in each stream order to indicate the effect of catchment-scale groundwater on fragmentation of the hyporheic zones. The distribution function was normalized with the total area of the region, A/A_0 , where the total area A_0 (m²) of each region is 25 m² and A (m²) is the area of the upwelling or downwelling zone. Here increasing fragmentation was defined as a shift toward smaller regions in the size distribution of coherent areas with upward or downward directed flow. It should be noted that this definition does not regard the magnitude of the flow.

The fragmentation of upwelling and downwelling areas was highly affected by the catchment-scale groundwater flow field. The superimposed flow field resulted in upwelling zones that were always less fragmented than the downwelling zone. In particular, when analyzing the superimposed flow field assuming a hyporheic depth of 4 m, 28% of the fifth-order stream regions had coherent upwelling areas that were larger than 7.75 m² (31% of the total area, A_0). In contrast, none of the regions had downwelling areas larger than 7.75 m² (Figure 10).

The most fragmented upwelling hyporheic zone was found to be the streambed-induced flow field, which was obtained without superimposing the catchment-scale groundwater flow field and can be seen as the delimiting case without any catchment groundwater flow (Figure 11, black line). When the catchment-scale groundwater flow was superimposed on the streambed-induced flow field, the CDFs of upwelling zones were shifted rightward for all stream orders, which indicates a decreased fragmentation of the hyporheic zone and prevalence of larger coherent upwelling zones induced by the catchment-scale groundwater flow field. The fragmentation of upwelling zones varied with stream order, with less fragmentation for the first- and fifth-order streams, while the CDFs of third- and especially second-order streams were found to resemble the CDF of the streambed-induced flow field (Figure 11). Moreover, the results showed that the fragmentation of upwelling zones was highly affected by the depth of the hyporheic zone and by the damping factor involving the streambed-induced hydrostatic head contribution. A deeper hyporheic zone, which reflects a reduced streambed-induced flux at the streambed interface and thus a higher relative contribution of catchment-scale groundwater fluxes, resulted in substantially less fragmentation of upwelling zones compared to more surficial hyporheic depths (Figures 11b and 11d). In addition, C_{damp} highly affected the fragmentation of upwelling zones, where decreased damping (i.e., increased value of R_P in Figure 3) increased the streambed-induced flux, decreased the relative effect of catchment-scale groundwater, and led to larger fragmentation of the upwelling zones (Figure 11 and Figure S8 in supporting information). For the case of $R_{P,4}$ (i.e., minimum damping of the surface water fluctuations) with $\varepsilon = 0.1$ m, the result displayed highly fragmented upwelling areas across stream orders, in which more than 97% of the regions had upwelling areas smaller than 1.25 m² (5% of the total area, A_0 ; Figure 11c). Due to the binary distribution of upwelling and downwelling areas, an inverse relationship between the fragmentation of upwelling and downwelling areas was observed. Therefore, an opposite order of increasing fragmentation was displayed for the stream orders when evaluating the CDFs of areas with downwelling compared to those of upwelling areas. In this case, the fifth- and first-order streams had the most fragmented areas of downwelling, whereas the second- and third-order streams were the least fragmented (Figure S9 in supporting information).

4. Discussion

A detailed understanding of the effects of catchment-scale groundwater flow on the hyporheic fluxes in streams is essential for quantifying surface water-groundwater interactions and thus also for understanding the ecology of streambeds. Here we use a combined model framework that includes a wide range of spatial scales together with site-specific geomorphological data to estimate the vertical flow velocity at

Figure 11. Cumulative distribution functions of the size distributions of spatially coherent upwelling areas for different stream orders assuming (a) $R_{p,2}$ and $\varepsilon = 0.1$ m, (b) $R_{p,2}$ and $\varepsilon = 4$ m, (c) $R_{p,4}$ and $\varepsilon = 0.1$ m, and (d) $R_{p,4}$ and $\varepsilon = 4$ m.

the streambed interface in a boreal stream network. The exchange flux determines whether gaining or losing conditions prevail on spatial scales that are difficult to observe using currently available sampling strategies (see, e.g., Ploum et al., 2018; Schmadel et al., 2010). The spatial distribution of the exchange fluxes was evaluated through the magnitudes of the exchange intensities (absolute value of vertical velocities) at the streambed interface (Figure 9) and the fragmentation of coherent areas of upwelling or downwelling (Figure 11 and Figures S8 and S9 in supporting information). The study highlighted the relevance of considering all spatial scales governing the hyporheic exchange—from the large-scale groundwater circulation to the local streambed-induced flow field—when estimating hyporheic exchange properties. Particularly for the fragmentation of upwelling areas, the groundwater flow field increased the sizes of coherent areas substantially (Figure 11), whereas the fragmentation of downwelling areas was only slightly affected (Figure S9 in supporting information). The difference was due to the predominance of upwelling fluxes at the streambed interface for the catchment-scale

groundwater flow field (Figure S10 in supporting information). In particular, for the first- and fifth-order streams, the decreasing fragmentation of upwelling areas caused by the catchment-scale groundwater flow field was evident as a shift in the CDF toward larger coherent areas of upwelling. Conversely, the prevalence of smaller and detached gaining areas reflecting more hydrologically fragmented hyporheic zones was most evident in second- and third-order stream reaches. We anticipate that the degree of hyporheic fragmentation can be important for the establishment, growth, and migration of biological (benthic) communities (McClain et al., 2003). Moreover, since deep groundwater generally has temperature and chemical composition different from those of stream water (Kalbus et al., 2009), the fragmentation of the hyporheic zone potentially implies fragmentation in both temperature and aquatic chemistry.

Previous findings by, for example, Morén et al. (2017) show that an increased depth of the hyporheic zone leads to an increase in the flow residence time but a decrease in the hyporheic exchange. Our results based on the flow fields induced by multiple spatial scales confirmed this conclusion and displayed that an increasing depth of streambed-induced flow field leads to a decrease in the magnitude of the streambed-induced exchange intensity (Figure 9) and consequently to an increased hyporheic residence time. In addition to the depth of the hyporheic zone, the damping of the propagation of the streambed topography to the surface water profile (equation (12)) was found to strongly affect the streambed-induced hydrostatic hydraulic head distribution and consequently the relative importance of the catchment-scale groundwater flow field. The damping of surface water fluctuations compared to the bed topography is hypothesized to be a function of wavelength and Froude number according to equation (8). The Froude number is positively correlated with the value of C_{damp} , which implies that higher streamflow velocity and lower stream depth yield lower damping (i.e., higher value of C_{damp}). The observed Froude numbers for the different stream reaches were small (Table S3 in supporting information), and all corresponded to subcritical flow ($Fr < 1$). Thus, in this study, the role of the flow regime on the damping was small, but it can be substantial for higher Froude numbers. This result highlights the need for additional research on the factors influencing the propagation of fluctuations between the streambed and the surface water, (i.e., hydraulic regime and spatial scale), due to its strong control on the hyporheic fragmentation where decreased damping (i.e., high C_{damp}) results in a lower contribution from the catchment-scale flow field and thus an increased fragmentation of the coherent upwelling areas. However, it should be noted that the relative distributions of fragmentation of upwelling zones between different stream orders were found to be constant across the wide range of values for C_{damp} and ε assumed in this study, with first- and fifth-order streams displaying less fragmentation compared to second- and third-order streams (Figure 11).

Previous studies to investigate the effect of watershed topography on hyporheic exchange fluxes have primarily addressed the dynamic head component of the streambed-induced exchange (e.g., Caruso et al., 2016; Marzadri et al., 2014). However, here we show that neglecting the hydrostatic hydraulic head originating from fluctuations in the stream water surface may lead to underestimation of the streambed-induced hydraulic head, which may also influence the role of the regional groundwater on the hyporheic zone. The fractal nature of topography (Jerolmack & Mohrig, 2005; Nikora et al., 1997; Wörman et al., 2007) allowed us to rescale the surrounding landscape topography to generate high-resolution data for stream topography. By utilizing the rescaled streambed topography and introducing a hydrostatic damping factor and a dynamic amplitude coefficient, we could separately define the hydrostatic and dynamic components of the streambed-induced hydraulic head distribution. Our results indicated that the contribution of the hydrostatic head component to the total streambed-induced hydraulic head was higher than or equal to the dynamic head contribution, depending on the assumed hydrostatic damping factor. Although our assumptions of the damping factor were based on observations in a single reach under specific hydrological conditions, four different values were considered as a part of sensitivity analysis to cover the possible range of damping factors (Figure 3b), providing important information about the possible ratio between different components and different spatial scales affecting the surface water groundwater exchange.

A combination of PCA and a power law statistical model was applied to find the best set of parameters required to explain the covariance of the six selected independent parameters and the spatial variation of the vertical velocity at the streambed interface. When these typical parameters were not normalized, we found that the variations in the parameter values could be explained by using only landscape elevation (e) and distance to the outlet (D_s ; Figure 8a). However, when the independent parameters were normalized,

more than two parameters were required (Table S4 in supporting information). The PCA results supported, to some extent, previous findings that stressed the importance of not only the watershed slope but also the drainage density and the shape of the watershed shape for the interaction between surface water and groundwater (Sharma et al., 2015). However, this study also identified subsurface hydraulic conductivity (soil type) as important both as an independent parameter and as an explanation for the degree of interaction between deep groundwater and the hyporheic zone. Normalization only introduced a constant term in the power law regression model, and regardless of normalization, the dependent variable in the power law model would be still nonnormalized. Hence, nonnormalized independent parameters were used in the power law model to provide such an explanation. The best fit trend among all the different sets ($R^2 = 0.24$) was obtained by using stream order (SO), distance to the outlet (D_s), and elevation (e). However, there was a relatively low degree of confidence (P value = 0.4) in the power law statistical model (Table S5 in supporting information). In comparison, using only stream order (SO) and elevation (e) in the power law model resulted in a lower R^2 statistical measure ($R^2 = 0.21$) and a slightly higher confidence (P value = 0.27; Table 2). Using elevation (e) and distance to the outlet (D_s) in the power law statistical model resulted in a very weak fitted trend ($R^2 = 0.02$) with a low confidence (P value = 0.89; Table S5 in supporting information).

5. Conclusions

This study quantified the effect of catchment-scale groundwater flow on the exchange velocities at the streambed interface and the fragmentation of the hyporheic zone. The key findings are as follows:

- The streambed-induced hydraulic head distribution strongly depended on the hydrostatic head due to water surface fluctuations, the dynamic head generated at the streambed surface, and the shielding caused by the depth limitation of the streambed-induced flow field. The often neglected hydrostatic head was found to be an important component in the streambed-induced hydraulic head, although highly sensitive to the damping of fluctuations in the surface water profile compared to fluctuations in the streambed topography.
- The mean absolute value of the vertical velocity at the streambed interface, that is, the magnitude of the exchange intensity, was dominated by the streambed-induced hydraulic head across stream orders. The relative contribution from the streambed scale exceeded that from the watershed scale by up to 2 orders of magnitude.
- Fragmentation of the hyporheic zone was defined herein as a shift in the CDF toward smaller coherent areas with upwelling or downwelling flux at the streambed interface. Regional groundwater flows had negligible effects on the CDF of downwelling areas but led to decreasing fragmentation of upwelling zones. Specifically, in the first- and fifth-order streams, the catchment-scale groundwater flow created large, coherent areas of upwelling, while larger fragmentation was found in second- and third-order streams. Moreover, an increasing depth of the streambed-induced flow field increased the relative contribution of catchment-scale groundwater flow and resulted in less fragmented areas of upwelling.
- The hyporheic fluxes and its fragmentation were primarily controlled by the spectrum of topographic scales. Additional factors that were investigated to generalize the results were found to be of significantly lesser importance. These factors included distance to the watershed outlet, landscape elevation, depth of Quaternary deposits, local stream slope, stream order, and hydraulic conductivity. However, to some extent, the stream order was an additional important factor for generalization.

Appendix A: Rescaling Methodology

The observed 2-D topography is represented by a Fourier series spectrum:

$$Z_{BS}(x, y) = \sum_{j=1}^N \sum_{i=1}^N (A_{ij})_{BS} \sin(k_i x) \cos(k_j y) \quad (A1)$$

where $Z_{BS}(x, y)$ (m) refers to the bed topography fluctuation deviating from the local trend, $N(-)$ is the number of wavelengths, A_{BS} (m) is the amplitude coefficient of the bedform topography, $k = 2\pi/\lambda$ (m^{-1}) is the wavenumber, and λ (m) is the wavelength. Equation (A1) decomposes the topography into a sum of harmonic functions with specified wavelengths. The correlation between amplitude and wavelength was obtained for each 20 regions by a power law equation given by equation (A2) (Marklund & Wörman, 2011)

$$A_{BS} = a\lambda^b \quad (A2)$$

where a and b are coefficients (Table S2 in supporting information). The smallest wavelength considered in the analysis was 4 m, which was 2 times the data resolution in the x and y directions, whereas the maximum wavelength was 100 m, which equaled the length of the study area in the x and y directions.

The power law relationship between A_{BS} and λ is assumed to hold also for shorter wavelengths. Using equation (A2) and the latter assumption makes it possible to rescale the estimated amplitude coefficients for wavelengths in the range from 0.1 to 5 m that represents the streambed scale. Finally, the rescaled amplitude coefficients (i.e., the amplitude coefficients for the streambed scale) from the 20 regions were used in equation (A1) to obtain the simulated rescaled topography for 20 streambed domains (Figure S2 in supporting information).

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