Groundwater

The Influence of Streambed Heterogeneity on Hyporheic Flow in Gravelly Rivers

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Abstract

Deposits of open-framework gravel occurring in gravelly streambeds can exert a significant influence on hyporheic flow. The influence was quantified using a numerical model of the hyporheic zone. The model included open-framework gravel stratasets represented with commonly observed characteristics including a volume fraction of about one-third of the streambed sediment, a hydraulic conductivity two orders of magnitude greater than other strata present, and a spatial connectivity forming preferential-flow pathways. The influence of open-framework gravel stratasets on hyporheic flow was much greater than the influence of the channel morphology including meanders, point bars, dunes, and ripples. Seventy percent of the total hyporheic exchange occurred across 30% of the channel boundary at locations of open-framework gravel stratasets. The maximum local interfacial flux rates occurred at these locations, and were orders of magnitude greater than those at other locations. The local flux rates varied by six orders of magnitude over the channel boundary. The composite flow rate through the model with open-framework gravel stratsets was an order of magnitude greater than that through an equivalent but homogeneous model.

Introduction

Hyporheic flow is influenced by a variety of factors including channel morphology and streambed heterogeneity. Figure 1 illustrates some morphologic influences at different scales. At a large scale, elevation gradients around channel bends drive flow across the channel boundary. Flow is generally driven from the stream into the subsurface on the upstream side of point bars at the inside of channel bends, and back into the stream on the downstream side. Other important aspects of morphology can include a hierarchy of bedforms such as unit bars, dunes, and ripples (Bridge 2003, 2006). The interaction of streamflow with bedforms creates perturbations in hydraulic head over the channel boundary that add complexity to the subsurface flow paths (e.g., Elliott and Brooks 1997; Cardenas and Wilson 2007). Consequently, the local flow across the channel boundary per area, termed the interfacial flux hereafter, can vary with a complex spatial distribution. Stonedahl et al. (2010) evaluated the combined influence of multiple scales of fluvial morphology on the spatial pattern of interfacial flux using a numerical model that was developed with data derived from physical flume experiments. They concluded that all scales of channel morphology influenced the interfacial flux, and that bedform morphology had the greatest influence. Furthermore, given their observation of complex interactions between the morphologic influences arising at different scales, they concluded that models with simplified representations of channel morphology (e.g., models that are two-dimensional or that represent one scale of influence) may not adequately characterize patterns and rates of hyporheic exchange.

The influence on hyporheic flow of permeability heterogeneity, arising from certain types of streambed stratification, has also been examined (e.g., Cardenas et al. 2004; Conant 2004; Salehin et al. 2004; Fleckenstein et al. 2006; Krause et al. 2007; Sawyer and Cardenas 2009; Niswonger, 2006). This article focuses on permeability

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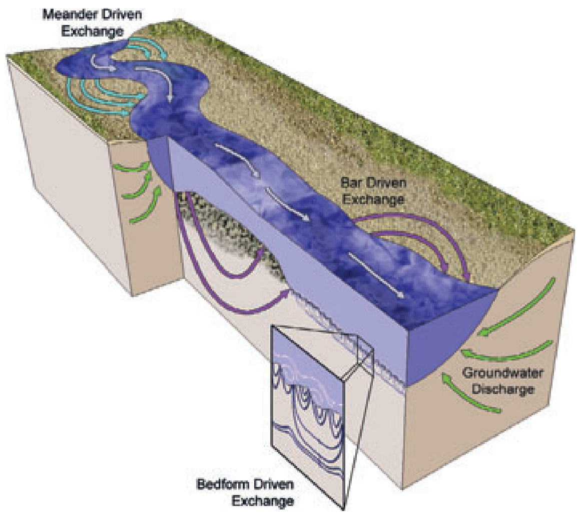

Figure 1. From Stonedahl et al. (2010). Conceptual representation of three scales of channel morphology that can influence hyporheic exchange, including meanders, bars, and bedforms (inset).

Figure 2. From Lunt (2002). A dissected unit bar showing alternating stratasets composed of sandy gravel (white arrow) and open-framework gravel (black arrow). Shovel is 1.3 m long.

heterogeneity arising from stratification commonly found in gravelly streambeds. The sedimentary architecture and related permeability heterogeneity found in gravelly streambeds are well studied (Klingbeil 1998; Kleineidam et al. 1999; Klingbeil et al. 1999; Heinz et al. 2003; Lunt and Bridge 2004; Lunt et al. 2004; Arrigoni et al. 2008; Jones et al. 2008). Modern examples of rivers transporting sandy gravel include the Sagavanirktok and Flathead Rivers (Alaska and Montana, respectively), both of which are fed by glacial meltwater. The hyporheic zones of such river systems can contain strata comprising relatively well-sorted gravel in a self-supporting framework with the space between grains unfilled by sand or finer sediment. Such strata are termed open-framework gravel (Figure 2). The formation and preservation of open-framework gravel was described by Lunt and Bridge (2007). Subsurface networks of connected open-framework gravel stratasets are known to create preferential flow pathways (Lunt 2002; Heinz et al. 2003; Tye et al. 2003; Lunt and Bridge 2004; Lunt et al. 2004; Lunt and Bridge 2007). The influence of open-framework gravel stratasets on spatial patterns of interfacial flux has not been quantitatively examined.

Quantitative field studies (e.g., Heinz et al. 2003; Lunt et al. 2004) have found open-framework gravel stratasets to collectively equal one-quarter to one-third of the volume of streambed sediment, and to individually have a thickness on the order of decimeters, a lateral extent of meters to tens of meters, and an average downstream dip on the order of 10 degrees. The permeability of open-framework gravel stratasets has been found to be orders of magnitude greater than that of surrounding sand and sandy gravel stratasets (e.g., Klingbeil 1998; Klingbeil et al. 1999; Tye et al. 2003; Conrad et al. 2008; Ferreira et al. 2010). Because of their similar hydraulic conductivities, the sand and sandy gravel stratatypes are not differentiated in this study. They are collectively referred to as sandy gravel hereafter.

The goal of this study was to better understand how heterogeneity arising from open-framework gravel strata influences the spatial pattern of interfacial flux over a channel boundary. The objective was to create a three-dimensional model for hyporheic flow and use it to quantify the influence of this heterogeneity relative to morphologic influences. We modified the model presented by Stonedahl et al. (2010) to include streambed heterogeneity and quantified the resulting interfacial flux. We compared these results to those with a homogeneous steambed to ascertain the influence of heterogeneity.

Methodology

Stonedahl et al. (2010) used a three-dimensional finite-difference code (Harbaugh et al. 2000; the MODFLOW-2000 code) to model hyporheic flow and compute steady-state interfacial fluxes. Their grid comprised 96 rows, 676 columns, and 17 layers. The grid cells in the top layer had prescribed-head values representing the spatial distribution of head across a channel boundary. They created this prescribed-head boundary as follows: A meandering channel, approximately 0.4 m wide was created by flow within a flume with dimensions roughly 10 m long × 1 m wide × 0.6 m deep. The topography of the channel boundary was measured using a high-resolution laser profiler. The channel topography (Figure 3) was used to calculate the head fluctuations on the channel boundary with a new approach adapted from Elliott and Brooks (1997) and Wörman et al. (2006). Cells on the two side-boundaries of the finite-difference model parallel to flow and on the bottom boundary all had a prescribed flux of zero. The up- and down-gradient boundary cells had head prescribed with an overall hydraulic gradient of 0.0013 which matched the slope of the energy grade line in the flume experiment.

Although the model presented by Stonedahl et al. (2010) was based on a flume experiment, it is relevant to natural river systems because the relevant bedforms scale with channel size, regardless of channel pattern. These scaling relationships were reviewed by Bridge (2003, 2006). Furthermore, the length-to-width ratio of bedforms

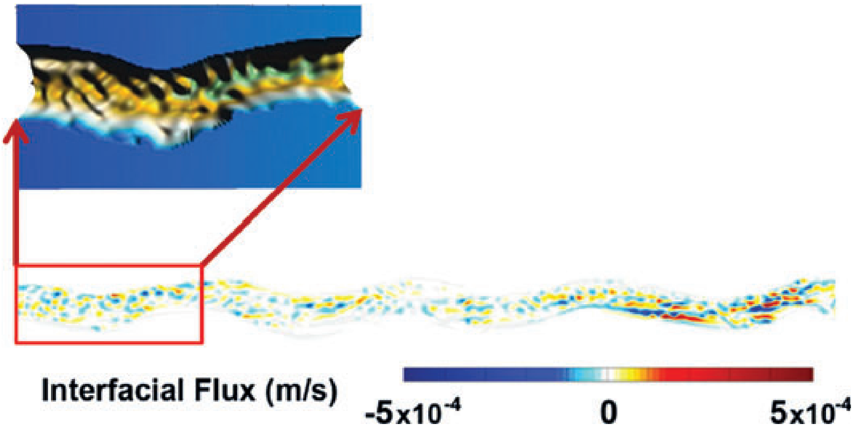

Figure 3. Bottom image is from Stonedahl et al. (2010) and shows their computed interfacial flux (their "full complexity" scenario). Flow was from left to right. Image on top shows the channel topography represented in the first meander of the Stonedahl et al. (2010) model. The average channel width is 0.4 m as used in the Stonedahl et al. (2010) model and is 144 m as used in our model.

and the stratatypes they create within streambed sediment scale together linearly, so the correlation scales of heterogeneity also scale together with the size of the channel and its bedforms (see also Rubin et al. 2006). These scaling relationships are observed in natural fluvial systems with channel belt widths from 600 m to over 15 km, and mean grain sizes from 0.04 mm to over 15 mm (Bridge 2006). Accordingly, the channel in the Stonedahl et al. (2010) model was taken to represent a larger natural river capable of transporting gravel. We used a channel width of 144 m as the basis for renormalizing length dimensions, reflecting an average channel width in the well-studied Sagavanirktok River system (Lunt et al. 2004). The prescribed-head boundary in the top grid layer of the Stonedahl et al. (2010) model had an energy grade line of 0.0013, which is equal to the energy grade line in the Sagavanirktok River system reported by Lunt et al. (2004), and which is preserved in the renormalized model. The prescribed-head boundary in the Stonedahl et al. (2010) model represented a ratio of bedform height to channel depth of the order of 1×10^{-1}. This is the general ratio of the heights of dunes and unit bars to channel depth in the Sagavanirktok River and was preserved in the renormalized model. The prescribed-head boundary in the Stonedahl et al. (2010) model represented the influence of bedforms with amplitudes of the order of 1×10^{-3} m and wavelengths of the order of 1×10^{-2} to 1×10^{-1} m. The renormalized prescribed-head boundary here represents dunes and unit bars with amplitudes of the order of 1 m and wavelengths of the order of 10 to 100 m. The finite-difference grid was made finer to adequately resolve the size and connectivity of dipping stratasets. To keep the number of grid cells less than 5 million, and thus keep the problem computationally tractable, only the first meander (Figure 3) was included in the simulation. The grid comprised 200 rows, 400 columns, and 60 layers of cells; each cell was taken to represent 2 m × 2 m × 0.05 m. This gave an overall model domain representing 400 m × 800 m × 3 m. Prescribed-head values in the top layer were interpolated from renormalized values on the Stonedahl et al. (2010) grid (from 96 rows × 169 columns to 200 rows × 400 columns) by kriging. Figure 4 shows that the interpolation preserved the overall hydraulic gradient (0.0013), and preserved the local variations in head due to channel morphology.

Heterogeneity was added to the model in two steps (Zhou 2012). In the first step, the stratal architecture was represented. In the second step, cells were populated with variable hydraulic conductivity drawn from a statistical distribution for each stratatype. In the first step, a Markov Chain model and indicator simulation with quenching (Carle 1999; the TProGS code) were used to simulate the structure of open-framework gravel stratasets occurring within all other, lower-permeability, sandy gravel stratatypes (referred to as the "background" hereafter). The proportion, mean length, and orientation of the simulated open-framework gravel stratasets were intended to match the field quantified values of Lunt et al. (2004) as summarized above. The Markov Chain model represents the mean and variance in length of open-framework gravel strata in direction i using a transition probability structure having a transition rate of $-1/\bar{l}_i$, where $\bar{l}_i$ is the mean length. The coordinate axes of the Markov Chain model are oriented relative to the axes of the simulation domain to achieve a dip of 10° in the downstream direction. The quenching step helped to ensure that a specified proportion of 28% was honored. The simulation (Figure 5) was created using mean lengths of 5 m in the downstream direction, 2 m in the cross-stream direction, and 0.2 m in the vertical direction.

The second step was to represent heterogeneity at the scale of grid cells within open-framework gravel stratasets and background strata. The heterogeneity among cells within a cluster of open-framework gravel cells was represented by randomly assigning hydraulic conductivity to each cell. These hydraulic conductivity values were drawn from a lognormal distribution defined by an appropriate geometric mean and variance. The

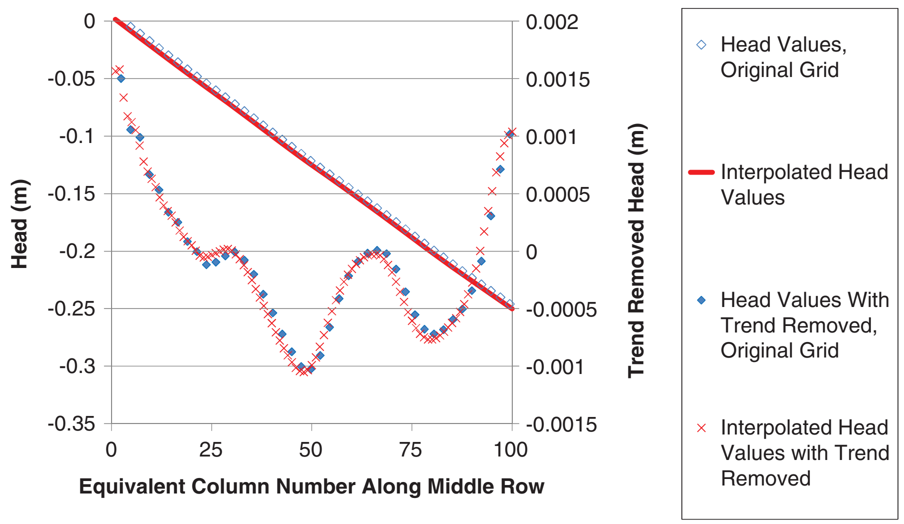

Figure 4. Interpolated, prescribed-head values in the top grid layer along the middle row in columns 1 to 100, compared to head values on the original grid. The kriging interpolation preserves the overall trend in the head values. The comparison of trend-removed values shows that the interpolation preserves the oscillations in head reflecting the influence of channel morphology. Slight differences between interpolated and source values are mainly due to the middle rows of the two grids being centered slightly differently on the *y* coordinate axis.

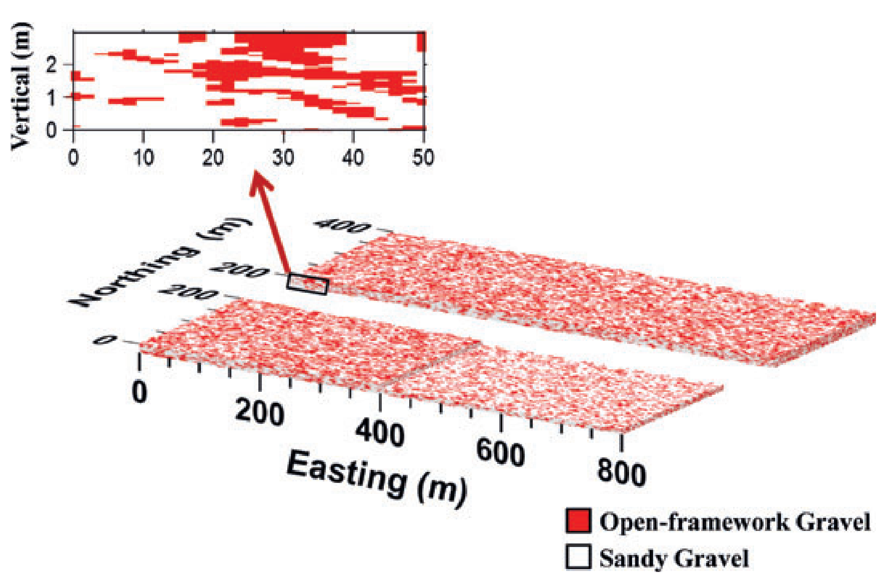

Figure 5. Simulation of stratasets of open-framework gravel and sandy gravel throughout the entire finite difference grid, below and adjacent to the channel (channel not shown). To show subsurface structure, the grid is split along the middle row, and the front upper octant of the grid is removed. A cross section along part of the middle row is enlarged to show the dip of the stratasets in the downstream direction.

heterogeneity among cells within the background was represented by randomly assigning hydraulic conductivity to each cell from a different lognormal distribution, one having a lower geometric mean. The model is intended to represent general hydraulic attributes common to channel-belt deposits composed of sandy gravel, rather than any single site. In this vein, the hydraulic conductivity distributions used to populate the cells of the model were generalized from those quantified in the literature. Some of the best field studies of hydraulic conductivity in these strata types are those by Klingbeil et al. (1999), Kleineidam et al. (1999), and Heinz et al. (2003). As summarized in Klingbeil (1998), the geometric mean, $\overline{K}_G$, for horizontal hydraulic conductivity of open-framework gravel measurements from among these studies is of the order of 0.1 m/s. The $\overline{K}_G$ of horizontal hydraulic conductivity for measurements in the other strata types is of the order of 0.0005 m/s. The vertical hydraulic conductivity measurements are generally one-tenth of the corresponding horizontal measurement. The variance among these measurements is large, and reflects the variation between different sites as well as the variation at a given site. The coefficient of variation (CV, the variance divided by the squared mean) for a given site is indicated to be at least as large as unity. We assumed that random variation with a lognormal distribution and a coefficient of variation of unity represent the variation of permeability within each of the two region types defined by the indicator simulation. Two natural-log normal distributions were used to represent the permeability distributions for each of the two stratal regions (open-framework gravel and background). For the open-framework gravel region, a random number generator was used to generate ln(K) values with a mean of −2.3025 and variance of 1.0, which gave K values with a geometric mean of 0.1 m/s and coefficient of variation of 1.0. The same procedure was followed for the background region, generating ln(K) values with a mean of −7.6009 and variance of 1.0, which gave K values with a geometric mean of 0.0005 m/s and coefficient of variation of 1.0. A vertical-to-horizontal anisotropy ratio of 0.1 was used for both distributions. The combined distributions are shown in Figure 6.

A search algorithm was used to determine how the open-framework gravel is connected in the model for heterogeneity. This algorithm determined that 94.34% of open-framework gravel cells are connected in a single cluster that spans between any pair of opposing domain boundaries, creating a percolating cluster. This high degree of connectivity is consistent with studies by Guin et al. (2010), and is consistent with observations of the hydraulic effect of connected pathways of open-framework gravel strata in nature (e.g., Tye et al. 2003). Huang et al. (2012) illustrated how two-dimensional slices from a three-dimensional model do not have the connectivity of the three-dimensional domain they are extracted from, as per percolation theory (Hunt and Ewing 2009). Indeed, the two-dimensional views on the faces of blocks of the domain shown in Figure 5 do not convey the complex three-dimensional connectivity of open-framework gravel cells. Open-framework gravel cells are differentiated as to whether or not they are

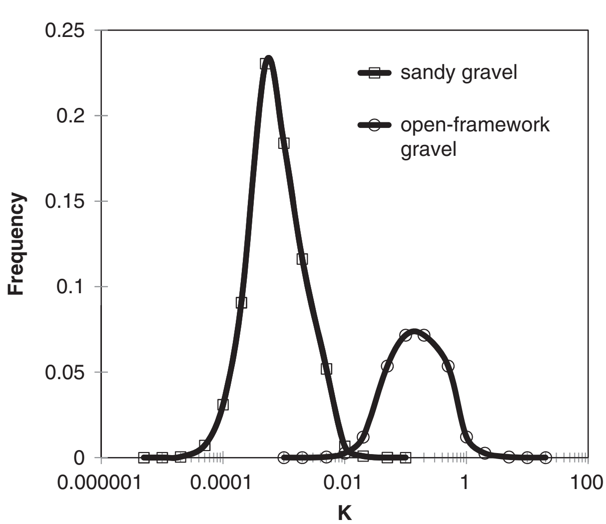

Figure 6. Distributions of hydraulic conductivity, K, in the heterogeneous model (m/s). Note that sandy gravel cells are also referred to as "background" cells in the article.

part of a percolating cluster in the presentation of results below.

For comparison, we also created a homogeneous model with a hydraulic conductivity of 0.002211 m/s, which is the geometric mean of hydraulic conductivity in the heterogeneous model. This value is two times greater than the homogeneous value used in the Stonedahl et al. (2010) model.

The steady-state finite-difference solution gave the total-volume inflow and outflow rates, and also gave the flow rate across the prescribed head boundary for each cell on the boundary (Harbaugh et al. 2000). The boundary-flow rates divided by the cell area on the channel boundary gave the interfacial flux.

Results

The connected open-framework gravel stratsets enable a great deal more water to move through the heterogeneous system relative to the homogeneous system. For both the heterogeneous and homogeneous models, the steady-state total-volume inflow rate equals the total-volume outflow rate. This total-volume rate of flow through the heterogeneous model is 1.1 m^3/s, and is 0.12 m^3/s for the homogeneous model. Thus, the total-volume rate of flow through the heterogeneous model is an order of magnitude higher than that of the homogeneous model. This result is consistent with the observation that the heterogeneous model has preferential pathways through open-framework gravel connecting prescribed-head cells across the boundaries of the flow domain. In flow through systems with higher- and lower-conductivity pathways connected in parallel (rather than in series), the effective permeability is expected to be above the geometric mean permeability (Hunt and Ewing 2009).

The results of computing the interfacial flux in the homogeneous system and the heterogeneous system are compared in Figure 7. Figure 7a shows the interfacial flux in the homogeneous model. Note that this result does not exactly match the result in Stonedahl et al. (2010) because the hydraulic conductivity is different. Relatively large regions with positive (into the subsurface) and negative (out of the subsurface) interfacial flux occur at the margins of the channel. The channel interior has regions periodically varying between positive and negative flux, a pattern reflecting the influence of boundary-head fluctuations related to bedform morphology. The greatest interfacial flux rates in the homogeneous model have magnitudes of 1×10^{-5} m/s. Figure 7b shows the interfacial flux in the heterogeneous model. In the heterogenous model the pattern of interfacial flux is different; the regions of positive and negative flux are smaller and are scattered throughout the channel. The greatest interfacial flux rates in the heterogeneous model have magnitudes of 1×10^{-3} m/s. Of the total positive or total negative flow across the channel boundary in the heterogeneous model, 69% of flow is through open-framework gravel cells that are part of the large connected cluster, 4% is through open-framework gravel cells that are not part

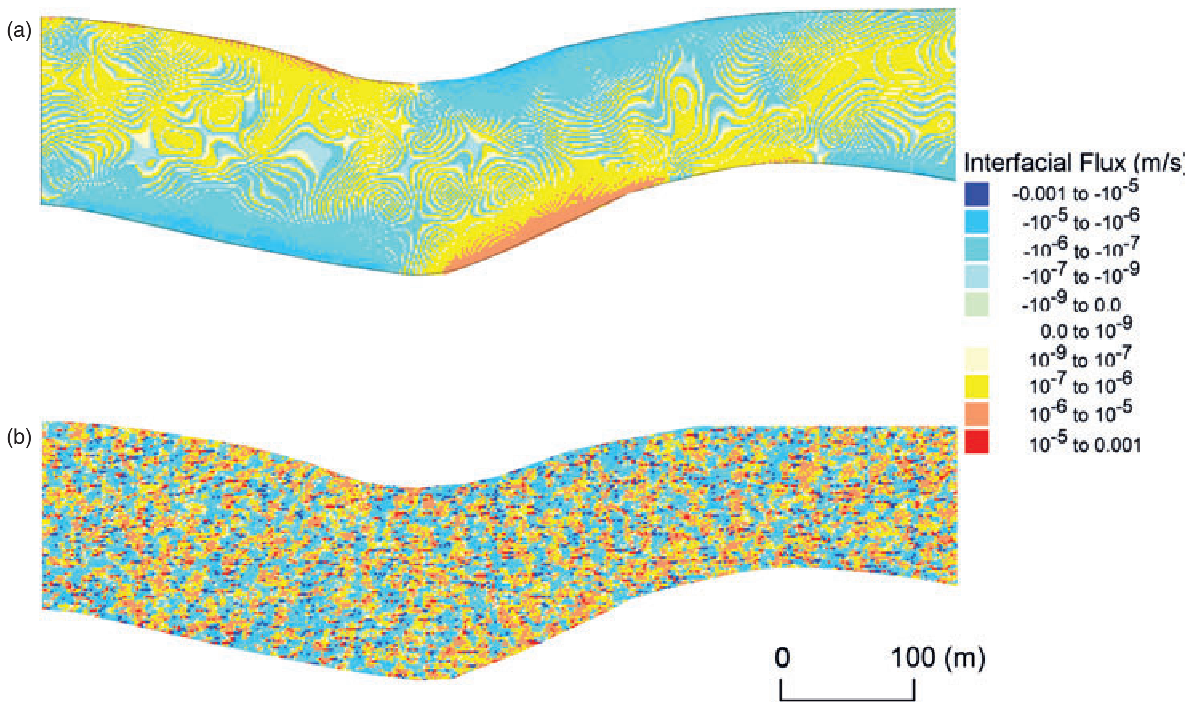

Figure 7. Interfacial flux across the channel boundary for (a) the homogeneous model and (b) the heterogeneous model. Positive flux is into the subsurface and negative flux is out of the subsurface. Lateral mean flow was from left to right.

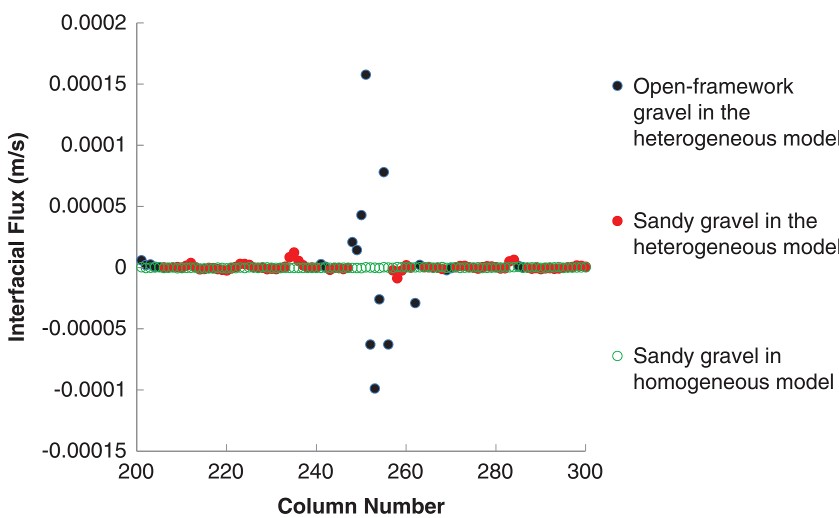

Figure 8. Interfacial flux rates for columns 200 to 300 in row 100 of the grid, shown to emphasize extreme values at cells with connected open-framework gravel. Lateral mean flow was from left to right. Positive flux is into the subsurface and negative flux is out of the subsurface. More information associated with these and other cells is given in Figure 9.

of that cluster, and 27% is through cells that are sandy gravel. The percentage of open-framework gravel cells on the channel boundary is 29%, similar to their percentage throughout the model. Thus, about 70% of the hyporheic exchange is across about 30% of the channel boundary where open-framework gravel occurs.

Figure 8 allows a closer examination of the local variation in interfacial flux within part of the heterogeneous model. This figure shows the interfacial flux rates for columns 200 to 300 in row 100 (middle row) of the grid. An arithmetic ordinate scaling is used to emphasize the extreme values of both positive and negative interfacial flux. The extreme values correspond to open-framework gravel cells. The interfacial flux through those open-framework gravel cells is an order of magnitude greater than flux through other cells. The total-volume flow rate entering into the subsurface over the entire line of these 100 cells is 0.045 m^3/s. Most of that inflow, 0.037 m^3/s, comes into connected open-framework gravel cells. The total flow rate out of the subsurface across the 100 cells is $-0.035\ m^3/s$, with $-0.031\ m^3/s$ out of open-framework gravel cells.

The results reflect a complex flow system as shown in Figure 9. Interfacial flux is plotted in Figure 9 with additional information for cells in all columns across the three middle rows (rows 99 to 101) of the grid. The information includes hydraulic conductivity, whether the open-framework gravel cells percolate, and the channel topography. The information for hydraulic conductivity includes the horizontal hydraulic conductivity in the uppermost active (i.e., head computed, not prescribed) grid layer, and the vertical conductance (a harmonic average, see Harbaugh et al. [2000]) between that layer and the prescribed-head layer above it. The greatest (magnitude) interfacial flux rates are shown to be associated with connected open-framework gravel cells where the vertical conductance is high. However, high flux rates do not always occur at the locations of open-framework gravel cells, nor do they occur exclusively there. The variation in K is over two orders of magnitude within the open-framework gravel stratasets or within the sandy gravel stratasets, and the complex topology of connected open-framework gravel clusters contribute to the complexity of the heterogeneous flow system. For example, in the region of cells from column 250 to 300 shown in Figure 9, the stream channel topography (part f) is rising on the upstream side of the point bar at the inside of the channel bend. The pattern in the result from the homogeneous model (part e, same columns) is closely related to small- and large-scale variations in the bed form topography. The interfacial flux is mostly positive indicating flow into the point bar (larger scale variation), with small and periodically occurring bands of negative flux corresponding to discharge on the downstream side of dunes on the point bar (dunes are more visible in Figure 3; also see trend-removed head values in Figure 4 reflecting dune presence). In the heterogeneous model, the pattern of interfacial flux rates (Figure 9, columns 250 to 300, part d) is quite different and complex. The pattern corresponds more to the spatial variation of vertical conductance (part c) than to the variations in bed form topography (part f). In this region, as across all columns, the extreme values of interfacial flux occur at connected open-framework gravel cells (part a), but not all connected open-framework gravel cells have extreme values, and high flux rates do not occur exclusively at open-framework gravel cells.

Discussion

The results suggest that the presence of open-framework gravel strata can (1) increase the total-volume flow rate through the system by an order of magnitude, (2) control the pattern of interfacial flux rates on the channel boundary, and (3) increase the magnitude of local interfacial flux on small areas of the stream channel by orders of magnitude. Studies of hyporheic exchange at specific field sites often seek to quantify the total hyporheic exchange rate relative to the surface water discharge rate, because the ratio of those two rates influences the potential impact of hyporheic exchange on surface water chemistry and temperature (e.g., Grant et al. 2006). If the total volume exchange rate is estimated using the geometric mean hydraulic conductivity of the streambed, the results presented here show that the exchange rate would be underestimated by an order of magnitude. Thus, estimates which do not represent the influence of open-framework gravel could grossly underestimate the potential impact of hyporheic exchange on water chemistry and temperature in this type of river. In field studies where dye tracers are injected into gravel-dominated point bars, the downstream location at which the dye emerges is commonly unpredictable. The complex patterns of interfacial flux shown in Figure 7 are consistent with such experience. These results also suggest that the location of open-framework gravel strata on the channel boundary could strongly influence nutrient concentrations, redox conditions, and temperature, which could, in turn, influence salmonid species behavior and the ecology of other hyporheos, as suggested in field studies by Stanford and Ward (1993, 1998). Future computational studies modeling temperature and reactive solute transport could be used to quantify the influence of open-framework gravel on water temperature and nutrient concentrations.

As with Sawyer and Cardenas (2009), the heterogeneity examined in this study represents the effect of cross-strata sets on hyporheic exchange. Sawyer and Cardenas (2009) studied the effect created by cross-stratified sands, located below dunes, on various metrics of hyporheic exchange. They concluded that the metrics were relatively insensitive to heterogeneity created by stratasets and that a homogeneous approximation is sufficient to describe first-order patterns of hyporheic fluxes. In contrast to results in Sawyer and Cardenas (2009), flux patterns were quite sensitive to stratasets in this study. Cardenas et al. (2004) proposed a dimensionless ratio to represent the influence of heterogeneity (numerator) relative to the influence of bedform topography (denominator) on hyporheic flow:

$$N = \frac{\sigma^2_{\ln K} l_z \left(z_{HZ}\right)^{-1}}{4A\left(\lambda \left|J_y\right|\right)^{-1}} \tag{1}$$

which can be evaluated for this study as follows. The $|J_y|$ term in Equation 1 is the magnitude of the lateral hydraulic gradient, which was 0.0013 in this study. The $\sigma^2_{\ln K}$ is the variance of log conductivity. The composite $\sigma^2_{\ln K}$ was 6.65 for this study. [Note that $\sigma^2_{\ln K}$ is related to the proportions (p_i), and the mean (m_i) and the variance σ_i^2 of log conductivity in each stratal subpopulation, where $i =$ open-framework gravel (OFG) or sandy gravel (SG), by the relationship: $\sigma^2_{\ln K} = p_{\mathrm{SG}}\,\sigma^2_{\mathrm{SG}} + p_{\mathrm{OFG}}\,\sigma^2_{\mathrm{OFG}} + (m_{\mathrm{SG}} - m_{\mathrm{OFG}})^2 p_{\mathrm{SG}}\,p_{\mathrm{OFG}}$ as per Ritzi et al. (2004).] The parameter l_z in Equation 1 is the spatial-bivariate vertical correlation length of hydraulic conductivity, which is given by $3\bar{l}_{OFG}(1 - p_{OFG})$ where $\bar{l}_{OFG}$ is the mean thickness of OFG stratasets (Ritzi et al. 2004). The value of l_z is 0.432 for this study. The parameter λ in Equation 1 is the bedform wavelength and A is the amplitude. The bedforms in this study have λ of the order of 10 to 100 m, and have A of the order of 0.5 to 1 m. The parameter z_{HZ}

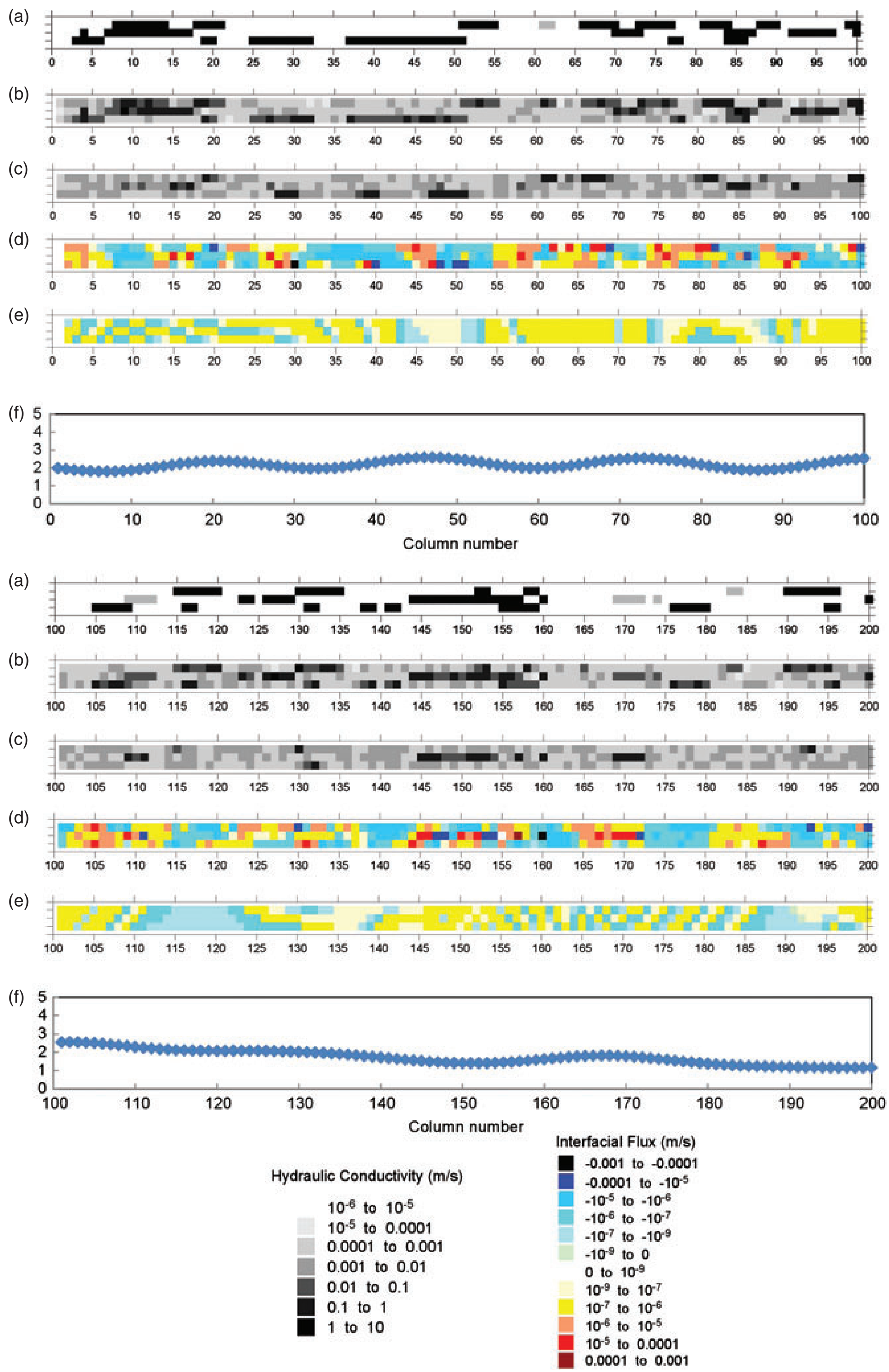

Figure 9. Combined information from the middle rows (rows 99 to 101) of the model grid. Results are shown in four sections of 100 columns each. Lateral mean flow is from left to right. In each section, the following information is presented. (a) Distribution of connected (black) and unconnected open-framework gravel (gray) in the uppermost active (i.e., head computed, not prescribed) grid layer. (b) Distribution of horizontal conductivity in the uppermost active layer. (c) Distribution of vertical conductance between cells in the uppermost active layer and the prescribed head cells above them. (d) Interfacial flux in the heterogeneous model. Positive flux is into the subsurface and negative flux is out of the subsurface. (e) Interfacial flux in the homogeneous model. (f) Channel topography.

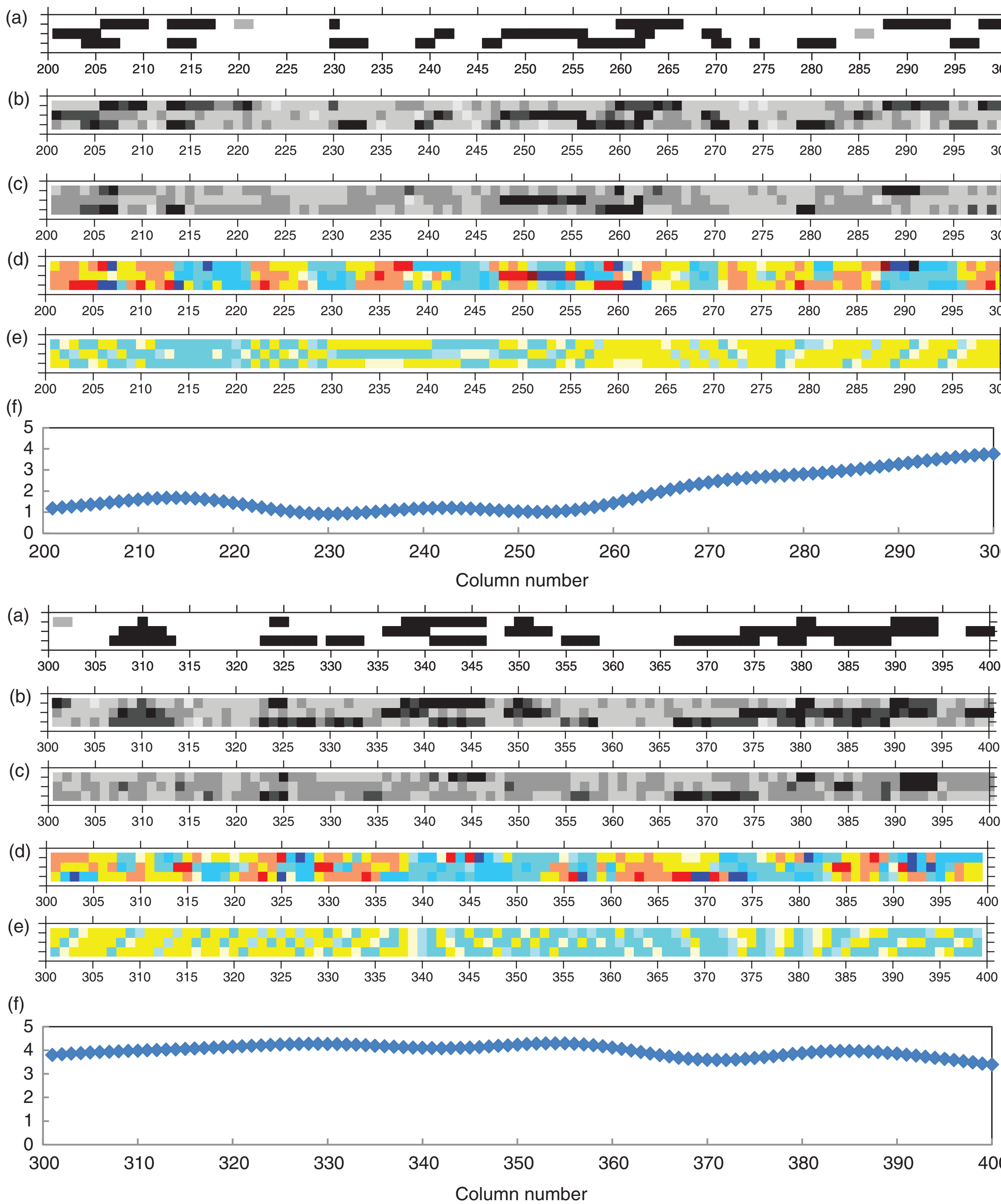

Figure 9. Continued.

in Equation 1 is the deepest point of a streamline emanating from the channel boundary in a homogeneous model. On the basis of the results of Stonedahl et al. (2010, their Figure 5), we expect this to be greater than 1 m. Using values that maximize N, we calculate a value of $N = 0.19$. The value is orders of magnitude greater than the values of 0.0017 and 0.0029 associated with the analyses by Sawyer and Cardenas (2009). Yet within the context of discussion of N by Cardenas et al. (2004), it is perhaps surprising to find a value of N less than 1 associated with a system in which the heterogeneity so strongly controls the hyporheic flow. Indeed, Cardenas et al. (2004) stated that "additional theoretical analysis or simulations ... are needed to confirm the definition, significance and critical values of N." The analyses here suggest that their metric might be more robust if modified to represent various aspects of percolation, including thresholds (e.g., Harter 2005; Guin and Ritzi 2008), and the length scales and tortuosity of connected clusters.

The analysis presented here is limited in addressing only the most pronounced aspects of permeability heterogeneity in sandy-gravel streambeds. Ramanathan et al. (2010) and Guin et al. (2010) created simulations representing more elements of the stratal architecture within these types of deposits, including the organization of stratasets at larger scales. Their simulations exceeded 150 million nodes. A goal for future work is to develop a model of hyporheic flow that includes their full heterogeneity model on a parallel-processing array. The Stonedahl et al. (2010) approach used to formulate the upper boundary condition in this study facilitated representing multiscale drivers of hyporheic exchange, in three-dimensions, and in a computationally tractable way. However, that approach has limitations articulated by Stonedahl et al. (2010) and those limitations translate to the analysis presented here. The size of their flume restricted the size of the meanders and made them more regular than natural channels. For example, the Sagavanirktok River has anastomosing segments that are both highly sinuous and individually divided (braided). Furthermore, groundwater flow patterns beyond the immediate vicinity of the channel were not represented in these studies. Another limitation is that these models have not represented the coupled system of surface and groundwater flow across the channel boundary. Cardenas and Wilson (2007) have made progress in simulating two-dimensional, turbulent, free-water flow over a dune coupled with porous flow through homogeneous sediment beneath the dune. However, it is not yet possible to simulate three-dimensional, fully coupled stream and groundwater flow with full feedback, with full representation of fluid physics at the channel boundary, and with streambed heterogeneity below it.

Conclusions

Stratasets of open-framework gravel in gravelly streambeds can exert a significant influence on hyporheic flow. The influence was quantified in a numerical model representing open-framework gravel stratasets with commonly observed attributes, including a volume fraction of about one-third of the streambed sediment, a hydraulic conductivity two orders of magnitude greater than other strata present, and a high degree of connectivity forming preferential-flow pathways. The influence of open-framework gravel pathways was much greater than the influence of geomorphic features such as channel bends and bedforms. About 70% of the total hyporheic exchange occurred across about 30% of the channel boundary at locations of open-framework gravel stratasets. The maximum local interfacial flux rates occurred at these locations, and were orders of magnitude higher than those at other locations. The patterns of interfacial flux caused by geomorphic features in an equivalent but homogeneous model were not observable in the model representing open-framework gravel stratasets. The composite flow rate through the model with open-framework gravel stratsets is an order of magnitude larger than the rate through an equivalent but homogeneous model. Thus, accounting for flow through open-framework gravels in the fluvial systems that contain them may be critical to correctly quantifying and understanding hyporheic flow.

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