Received: 8 April 2021 | Accepted: 12 April 2021

DOI: 10.1002/hyp.14226

INVITED COMMENTARY

HP TODAY WILEY

Hydraulic conductivity can no longer be considered a fixed property when quantifying flow between groundwater and surface water

Donald O. Rosenberry[1] | Peter Engesgaard[2] | Christine Hatch[3]

[1]U.S. Geological Survey, Lakewood, Colorado, USA

[2]Department of Geosciences and Natural Resource Management, University of Copenhagen, Copenhagen, Denmark

[3]Department of Geosciences, University of Massachusetts Amherst, Amherst, Massachusetts, USA

Correspondence
Donald O. Rosenberry, U.S. Geological Survey, MS 413, Denver Federal Center, Lakewood, CO 80225, USA.
Email: rosenber@usgs.gov

1 | INTRODUCTION

Hydrogeologists, biologists, geochemists, and ecologists who focus their interest at the sediment–water interface commonly use a Darcian approach to quantify exchange between groundwater and surface water:

$$q = Ki = kgi/\nu \tag{1}$$

where q is specific discharge per unit area (flow in either direction between groundwater and surface water) (L/T), K is hydraulic conductivity (L/T), i is hydraulic gradient (L/L), k is intrinsic permeability of the sediment (L^2), g is acceleration of gravity (L/T^2), and ν is kinematic viscosity of water (L^2/T). Although density and viscosity are part of the equation, K is nearly always related solely to sediment structure.

K is notoriously difficult to determine, both because it is among the most widely varying physical parameters (Freeze & Cherry, 1979) and because it is scale-dependent (e.g., Schulze-Makuch et al., 1999). Furthermore, spatial variability commonly is large in lacustrine and fluvial settings (e.g., Genereux et al., 2008; Glose et al., 2021; Käser et al., 2009; Sebok et al., 2015; Springer et al., 1999; Toran et al., 2015; Wang et al., 2016), for example, varying nearly 3 orders of magnitude over 160 m^2 (Kishel & Gerla, 2002). Order-of-magnitude estimates for K are often the best we can do.

Quantifying K at and near the sediment–water interface is more difficult yet because K also varies temporally (e.g., Dafny et al., 2015; Doppler et al., 2007; Genereux et al., 2008; Hatch et al., 2010; Wang et al., 2016). Sediment can be highly mobile in marine, lacustrine, and fluvial settings and the ability for water to flow through these shallow sediments can be altered by numerous physical, chemical, and biological processes.

We assert that considering K as a fixed parameter is inappropriate at and near the sediment–water interface where physical properties of the porous media are nearly constantly changing, sometimes altering K by several orders of magnitude. We attribute the bulk of this variability to surface-water dynamics, bias related to the direction of flow across the sediment–water interface, and biological activities that alter the physical and/or chemical structure of the sediment. Scientists need to view K in this setting as a variable, rather than a constant, and make repeated measurements to quantify and incorporate temporal variability in determinations of flow and associated chemical exchanges between groundwater and surface water.

The overly simplistic assumption that K is temporally invariant likely is invalid for virtually all settings where groundwater exchanges with surface water. All surface-water settings are highly dynamic compared to the broader groundwater-flow domain, where laminar flow prevails. Fluvial and high-energy marine settings are particularly obvious examples, but lacustrine and paludal settings also change in response to waves and currents and as shorelines move laterally with changing surface-water stage. Most groundwater-flow models facilitate separately assigning K for riverbeds and lakebeds, but nearly all studies apply a fixed value for K at those interfaces. Providing for temporal variation in this parameter that governs the connection between groundwater and surface water is usually difficult, is not automated, and as a result is rarely considered.

Given the global importance of the connections between groundwater and surface water, and the anthropogenic alterations of these exchanges, quantification of water and chemicals that move between these two water domains needs to improve, including the acknowledgment that *K* is temporally variable in addition to *q* and *i*. We argue for this paradigm shift because of improved knowledge and understanding of the influence of three processes mentioned above and detailed here:

1. Episodic or continual sediment erosion and deposition related to changing surface-water dynamics that alter the sediment grain-size distribution and associated hydraulic properties,
2. Flow-directional bias in *K* associated with buoyancy of fine-grained sediment during upward flow, compaction during downward flow, and changes in fluid viscosity related to difference in temperature of upward versus downward flowing water, and
3. Temporal variability in biological processes (bioturbation, bio-irrigation, biofilm growth, biogenic-gas production) that can increase or decrease the ability for flow across the sediment–water interface.

1.1 | Episodic deposition and erosion

It is well established that riverbed sediment erosion and deposition are directly related to water depth and surface-water velocity, high-flow events are particularly transformative, and streams and rivers are rarely if ever at steady state (Schumm & Lichty, 1965; Wolman & Miller, 1960). Therefore, with streams undergoing nearly constant degradation or aggradation, it is somewhat surprising that reasonable results for quantifying exchange between groundwater and surface water could stem from an assumption that *K* is fixed with respect to time. Credence is provided by the commonly large spatial heterogeneity, range, and associated uncertainty in determining *K* compared to the relatively small temporal changes. Even when limited to fluvial settings, reported values for *K* still range by nearly 8 orders of magnitude (Calver, 2001). Therefore, if estimates of *K* were within 1 or 2 orders of magnitude of true values, we often were satisfied they were close enough.

Numerous studies in fluvial settings have documented temporal changes in *K* associated with vertical flow (K_v), ranging from a factor of two to as large as 2 or 3 orders of magnitude, caused by changing grain-size distribution (e.g., Hatch et al., 2010; Korus et al., 2020; Mutiti & Levy, 2010; Rosenberry & Pitlick, 2009b; Sebok et al., 2015; Simpson & Meixner, 2012; Wang et al., 2016). Several studies have also reported temporal variability of K_v in lake sediments up to an order of magnitude (Kazmierczak et al., 2016; Wiese & Nützmann, 2009). Even minor sediment disturbance can create substantial change. At Mirror Lake in New Hampshire, USA, walking on the lakebed disturbed a surficial layer of fine-grained sediments only several mm thick and increased downward seepage by factors of 3 to nearly 8 (Rosenberry et al., 2010).

Variable *K* in fluvial settings is the result of multiple processes that can combine in complex ways. Flood flows can increase or decrease *K* depending on magnitude and duration of the flood, length of time since the previous flood, and whether the finer sediment fraction or the entire range of grain sizes is mobilized (e.g., Schalchli, 1995; Sebok et al., 2015; Simpson & Meixner, 2012; Wu et al., 2015). *K* can also increase with bedform amplitude and bankfull channel width and/or depth and then decrease with time since the last bed disturbance (Stewardson et al., 2016), particularly during extended low-flow or no-flow periods (Figure 1).

This is important not only to ecological and biogeochemical processes that drive much of the research at this important ecotone, but also for water-supply systems that make use of the sediment as a filter to minimize chemical treatment of public water supplies (e.g., Levy et al., 2011; Ray et al., 2003; Schubert, 2006; Zhang et al., 2011). Commonly termed "river-bank filtration" and growing in implementation world-wide, this practice has been in use for hundreds of years in Europe and provides about 16% of the potable water in Germany (Schmidt et al., 2003) and 7% in the Netherlands (Stuyfzand et al., 2006). This process usually induces fine-grained sediments to partially clog the riverbed, reducing K_v and associated well production. A riverbank-filtration installation on the Russian River, CA, USA, provides a good example. Reductions in *K* of 1 to 2 orders of magnitude decreased flow from the river to such an extent that sediments desaturated beneath the riverbed (Su et al., 2007). Biofilms were substantially responsible as they increased in areal extent and thickness during the summer low-flow period (Ulrich et al., 2015).

In some riverbank-filtration settings, fines and algal mats are flushed downstream during seasonal or other high-flow events, effectively resetting the bed sediment and largely restoring well production (Goldschneider et al., 2007). Reduced *K* in these settings could be permanent, however. If a downward hydraulic gradient exists during a flood, fine-grained sediments can clog the porous media beyond the depth altered by more typical ranges in streamflow. This deeper sediment clogging would only be removed by larger flood flows. Pumping of production wells during extreme events could exacerbate extreme-depth clogging that could be permanent because those fine-grained sediments likely would never be removed.

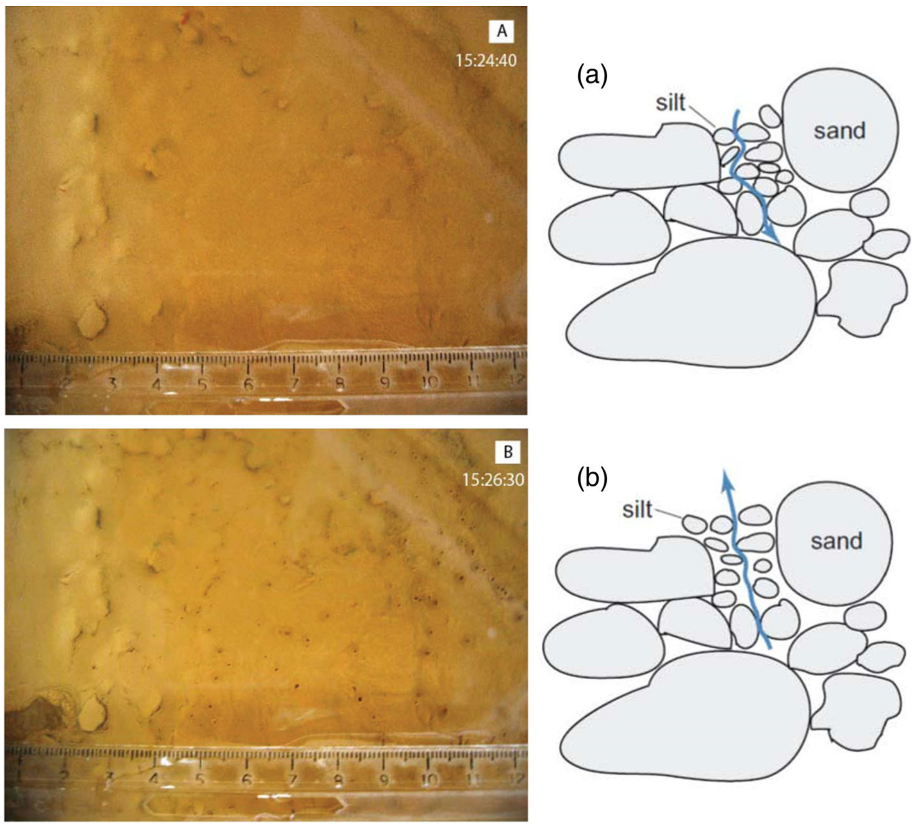
1.2 | *K* bias associated with upward versus downward flow

1.2.1 | Buoyancy effect

Several studies indicate that K_v at or near the sediment–water interface can vary depending on the direction of flow, both in lotic and lentic settings and particularly those where exchanges are fast and variable. During flow from surface water to the underlying sediments, a downward seepage force acts in concert with gravity to compress or transport fine-grained sediments deeper into interstices, creating additional flow restrictions and reducing K_v. This is common in fluvial settings where bedforms project into the flow field, even if there is no larger-scale downward hydraulic gradient (e.g., Jin et al., 2019; Packman & MacKay, 2003). During upward flow, the force associated with upward seepage opposes the gravity force exerted on each sediment grain. For silica-based inorganic sediments with a density of 2650 kg/

FIGURE 1 Pajaro River, California, USA, goes dry in the middle reaches during summer, revealing annual accumulation of fine sediment that has dried and cracked on the surface (foreground) (3.2-cm-diameter white monitoring wells in the middle distance provide scale). Inset shows organic floc on riverbed while still wet, with boot print providing scale. Photos by Christine Hatch and Donald Rosenberry (inset)

FIGURE 2 Downward flow (a) across the sediment–water interface promotes clogging and reduction in K_v, but upward flow (b) reduces clogging and develops visible preferential flowpaths. The combined surface area of the visible holes that formed with upward seepage was 0.5% of the total bed area (Modified from Rosenberry & Pitlick, 2009a.)



m^3, the sediment is made neutrally buoyant by upward seepage once the hydraulic gradient exceeds about 1.1 (Rosenberry & Pitlick, 2009a), increasing K_v. With no overlying sediment above neutrally buoyant grains at the sediment–water interface, grains are easily dislodged and preferential flow paths develop, further increasing K_v (Figure 2).

This bias in K_v related to flow direction was noticed decades ago during a study of the Little Plover River in Wisconsin, USA, and was locally coined the "flap-valve effect" (E.P. Weeks, personal communication, Weeks et al., 1965). Ratios of K_v for upward versus downward seepage were as large as 1.9 for sand and 2.7 for silt in a laboratory where gradients were measured in response to generated seepage (Rosenberry & Pitlick, 2009a). In the coarse-sand, fine-gravel bed of the South Platte River, Colorado, USA, ratios of K_v for upward versus downward flow ranged from 1.04 to 17.5 (Rosenberry & Pitlick, 2009b).

1.2.2 | Temperature/viscosity effect

Flow-directional bias can also occur where surface-water temperature is substantially different from temperature of the underlying sediments. Hydrogeologists have long known that viscosity can double, reducing *K* by half, when water temperature decreases from 25 to 0°C (Constantz et al., 1994). This factor-of-two variability in *K* is rarely considered because it is a small change relative to the uncertainty in determining *K* and also because groundwater temperatures typically vary little over time. However, in the shallow, near-shore margins of lakes, wetlands, streams, and oceans, where most of the exchange between groundwater and surface water normally occurs (McBride & Pfannkuch, 1975), temperatures commonly vary over this range. In higher-latitude settings, frozen surface water during winter can reduce the bed-sediment temperature near the shoreline to 0°C. During summer, the surface-water temperature in those same near-shore margins can exceed 30°C. The resulting temperature difference between surface water and water in the underlying porous media is often sufficiently large to have a substantial effect on viscosity and *K*. When surface-water temperature is warmer than the underlying porous media, *K* during upward flow is controlled by viscosity of the colder porous media, particularly for faster flows where upward advection is much faster than downward conduction of heat. For downward flow, surface water advects and conducts heat into the sediment, reducing viscosity and increasing *K*.

Temperature-generated bias in *K* related to flow direction either works in concert with, or in opposition to, bias related to sediment buoyancy or compression, depending on the season and flow direction. When working in opposition, directional bias associated with sediment buoyancy or compression likely is the dominant process in high-energy settings.

Simulations of exchange between groundwater and surface water can be improved if temporal variability in bed-sediment parameters are incorporated. Residuals were greatly reduced when hydraulic conductivity was changed seasonally by nearly a factor of 2 in a model simulating exchange between the Limmat River in Switzerland and underlying groundwater (Doppler et al., 2007). With the large increase and rapid adoption of heat as a tracer for quantifying exchange between groundwater and surface water (e.g., Briggs et al., 2014; Hatch et al., 2006; Irvine et al., 2017; Koch et al., 2016; Rau et al., 2014), it is surprising that temperature effects on *K* are not routinely addressed when quantifying that exchange with gradient-based approaches.

1.3 | Biological processes that alter *K*

Biological influences, such as bioturbation and bioirrigation of shallow sediment, biofilms and algal mats that grow on the sediment surface, and episodic accumulation and release of biogenic gas, can generate exchanges and/or alter sediment permeability at and near the sediment–water interface as much or more than purely physical processes. Assuming that exchanges are largely controlled by hydraulic gradients leads many to ignore these biologically driven processes. The following examples demonstrate the magnitude of this largely unaccounted-for influence.

Bioturbation is the process of altering the sediment matrix while bioirrigation is the passive and active flushing of water through the biologically enhanced porous matrix. Organisms, such as worms, clams, crabs, shrimp, and crayfish, create conduits beneath the sediment–water interface that collectively increase the areally averaged value of *K* (e.g., Baranov et al., 2016; Santos et al., 2012; Volkenborn et al., 2007). Organisms create these preferential flowpaths to bring oxygen into the sediment, for protection from predators and high-energy waves, and as a food supply. Bioirrigation has been measured at area-averaged rates on the order of 1 to 2 but up to 5 cm/d in Florida (Cable et al., 2006) and, depending on species and burrow density, at up to 200 to 1000 cm/d in shallow margins in Denmark (Santos et al., 2012). These exchanges can reach depths of 0.5 (Volkenborn et al., 2019) to several m (Santos et al., 2012) and can be orders of magnitude greater than gradient-driven exchange. In Lake Müggelsee near Berlin, Germany, densities of burrowing organisms as large as 2000 per m^2 can filter the entire lake volume in a single week (Baranov et al., 2016). The areal extent of these conduits of high permeability that serve to oxygenate the sediments also can be very large, effectively doubling the surface area when including the 3-dimensional distribution of burrows beneath the sediment–water interface (Volkenborn et al., 2019). The permeability of these networks of subterranean channels must be maintained by the organisms. Therefore, bulk *K* of the sediments, and associated fluid and chemical exchange, varies as the population dynamics of the various species evolve in response to seasons and episodic events, such as storms or disease or anthropogenic disturbances.

Biologically induced clogging from accumulation of organic sediments at and beneath the sediment–water interface can reduce *K*, particularly for downward flow, often by three orders of magnitude (e.g., Seifert & Engesgaard, 2007), as can algal and other areally extensive biological drapes at the sediment–water interface (Baveye et al., 1998). These reductions can occur over weeks to months and are variable depending on water temperature and river hydraulics (Newcomer et al., 2016; Ulrich et al., 2015). Biologically induced reductions in K_v decrease the depth of hyporheic exchange in fluvial settings and can also greatly affect biogeochemical processes and microbial transformations (Caruso et al., 2017). High-flow events often flush these drapes and restore streambed connectivity, but it depends on the extent of bed mobilization. A rainfall-induced streamflow pulse restored K_v at a setting where the bed remained flooded during a prolonged low-flow event (Chiu et al., 2020). Downstream, where the streambed dried during the low-flow period, accumulation of dead algal matter made the streambed more resistant to erosion and K_v decreased markedly once streamflow was reestablished.

Aquatic macrophytes can also create unexpected distributions of flow between groundwater and deeper surface water. At Lake Hampen in central Denmark, nutrient-enhanced macrophyte densities of up to 12 000 plants/m^2 created a thick rhizosphere that locally reduced groundwater discharge to the lake (Frandsen et al., 2012). The low-K rhizosphere altered the porous-media-based distribution of groundwater discharge to the lake and deflected groundwater and associated nutrients to discharge in deeper water farther from shore (Karan et al., 2014).

Decomposers of organic sediments serve a decolmation function that can increase K (Brunke & Gonser, 1997; Danielopol, 1984; Vanek, 1997), but they also generate gas that can either further increase or decrease K. If gas merely accumulates, it can occlude pores and reduce K, as has long been noted (e.g., Faybishenko, 1995). K can be reduced an order of magnitude or more by gas accumulation; reduction varies based on gas-production rate, temperature, and pressure. If accumulation is sufficient such that gas rises through the saturated sediment, preferential flowpaths are created when gas rises to the overlying surface water, creating high-K conduits for subsequent releases of gas and water (Santos et al., 2012).

2 | WHY DOES THIS MATTER?

The hyporheic zone associated with streams and rivers, and the equivalent hypolentic, hypopaludal, and intertidal zones associated with lakes, wetlands, and marine margins are complex and vitally important ecotones where biogeochemical processes wax and wane and are altered by constantly changing flows of water and chemicals across the sediment–water interface. Rates and reversals of flow across the interface govern many of these processes and are in turn largely governed by K. If we do not understand the degree of variability of K, we may erroneously attribute K-driven changes within this ecosystem to other processes.

Increasing exploitation of groundwater and surface water creates unintended anthropogenic alterations to natural fluxes, as well as unintended movements of contaminants across the interface that connects the groundwater and surface-water domains (Conant et al., 2019; Smith, 2005; U.S. EPA, 2000; Winter et al., 1998). Water-supply systems that induce surface water to flow to municipal pumping wells rely on the sediments to filter contaminants or pathogens of concern before the surface water reaches the well screen. An unknown breach of the lower-K clogging layer can lead to unexpected contaminant transport, such as during episodic high-flow events in rivers or when floodwater rises to extend over higher-K near-shore margins that can transmit water much more readily. Changing values of K can create unexpected increases in the flow rate, and concomitant decreases in travel time, that could result in unexpected levels of degradation of water quality for public water supply.

The relatively smaller influence of flow-directional bias in K could be viewed by pragmatists as an esoteric, primarily academic curiosity. However, K bias associated with flow direction may be globally significant in coastal settings where tidally driven reversals in flow direction occur two to four times a day. In these marine margins that extend along more than 600 000 km of shoreline, the collective influence could be very large. Estimates of global-scale discharge of groundwater to oceans vary widely but all are large and on the order of the global discharge of rivers to the ocean (Burnett et al., 2003; Moore, 2010). Tidally- and wave-driven exchanges are larger yet, estimated to be equivalent to the entire volume of the Earth's oceans once every 3000 years (Santos et al., 2012) at an average rate of about 1 cm/d. A flow-directional bias in K allows water to flow more readily to the ocean than from it, which in some settings may reduce the slope of the water table near the ocean margins, resulting in a lower water table and less freshwater storage. Bias in exchange can also affect water losses in managed rivers. Peaks in flow downstream of power-production dams cause reversed flow of river water to riparian margins (Rosenberry et al., 2013) where water can then be lost to evapotranspiration (Yellen & Boutt, 2015). These anthropogenically driven exchanges controlled largely by K can also affect river ecosystem health and water quality.

These examples indicate that our assumption of a time-invariant value for hydraulic conductivity at the sediment–water interface is almost always incorrect. We can improve our estimates of fluxes between groundwater and surface water by starting with the assumption that K fluctuates at multiple temporal scales and that we need to quantify or estimate those changes in K just as we consider and measure changes in hydraulic gradient. Effects of K-altering events, such as flood flows or stage changes that substantially move the shoreline, can be quantified with K measurements made pre- and post-event. Models that simulate exchange between groundwater and surface water should facilitate inclusion of temporal variability of K, making it an input variable, rather than a fixed parameter, with respect to time.

ORCID

Donald O. Rosenberry https://orcid.org/0000-0003-0681-5641
Peter Engesgaard https://orcid.org/0000-0002-5925-8757
Christine Hatch https://orcid.org/0000-0002-4996-1617

REFERENCES

Baranov, V., Lewandowski, J., & Krause, S. (2016). Bioturbation enhances the aerobic respiration of lake sediments in warming lakes. *Biology Letters*, *12*, 20160448.

Baveye, P., Vandevivere, P., Hoyle, B. L., DeLeo, P. C., & Sanchez de Lozada, D. (1998). Environmental impact and mechanisms of the biological clogging of saturated soils and aquifer materials. *Critical Reviews in Environmental Science and Technology*, *28*, 123–191.

Briggs, M. A., Lautz, L. K., Buckley, S. F., & Lane, J. W. (2014). Practical limitations on the use of diurnal temperature signals to quantify groundwater upwelling. *Journal of Hydrology*, *519*, 1739–1751.

Brunke, M., & Gonser, T. (1997). The ecological significance of exchange processes between rivers and groundwater. *Freshwater Biology*, *37*, 1–33.

Burnett, W. C., Bokuniewicz, H., Huettel, M., Moore, W. S., & Taniguchi, M. (2003). Groundwater and pore water inputs to the coastal zone. *Biogeochemistry*, *66*, 3–33.

Cable, J. E., Martin, J. B., & Jaeger, J. (2006). Exonerating Bernoulli? On evaluating the physical and biological processes affecting marine seepage meter measurements. *Limnology and Oceanography: Methods*, *4*, 172–183.

Calver, A. (2001). Riverbed permeabilities: Information from pooled data. *Ground Water*, *39*, 546–553.

Caruso, A., Boano, F., Ridolfi, L., Chopp, D. L., & Packman, A. (2017). Biofilm-induced bioclogging produces sharp interfaces in hyporheic flow,

redox conditions, and microbial community structure. *Geophysical Research Letters*, *44*, 4917–4925.
Chiu, Y.-C., Lee, T.-Y., Hsu, S.-Y., & Liao, L.-Y. (2020). The effect of hydrological conditions and bioactivities on the spatial and temporal variations of streambed hydraulic characteristics at the subtropical alpine catchment. *Journal of Hydrology*, *584*.
Conant, B., Robinson, C. E., Hinton, M. J., & Russell, H. A. J. (2019). A framework for conceptualizing groundwater-surface water interactions and identifying potential impacts on water quality, water quantity, and ecosystems. *Journal of Hydrology*, *574*, 609–627.
Constantz, J., Thomas, C. L., & Zellweger, G. (1994). Influence of diurnal variations in stream temperature on streamflow loss and groundwater recharge. *Water Resources Research*, *30*, 3253–3264.
Dafny, E., Tawfeeq, K. J., & Ghabraie, K. (2015). Evaluating temporal changes in hydraulic conductivities near karst-terrain dams: Dokan Dam (Kurdistan-Iraq). *Journal of Hydrology*, *529*, *Part 1*, 265–275.
Danielopol, D. L. (1984). Ecological investigations on the alluvial sediments of the Danube in the Vienna area - A phreatobiological project. *SIL Proceedings, 1922-2010*, *22*(3), 1755–1761.
Doppler, T., Hendricks Franssen, H.-J., Kaiser, H.-P., Kuhlman, U., & Stauffer, F. (2007). Field evidence of a dynamic leakage coefficient for modelling river–aquifer interactions. *Journal of Hydrology*, *347*, 177–187.
Faybishenko, B. A. (1995). Hydraulic behavior of quasi-saturated soils in the presence of entrapped air: Laboratory experiments. *Water Resources Research*, *31*, 2421–2435.
Frandsen, M., Nilsson, B., Engesgaard, P., & Pedersen, O. (2012). Groundwater seepage stimulates the growth of aquatic macrophytes. *Freshwater Biology*, *57*, 907–921.
Freeze, R. A., & Cherry, J. A. (1979). *Groundwater*. Prentice Hall.
Genereux, D. P., Leahy, S., Mitasova, H., Kennedy, C. D., & Corbett, D. R. (2008). Spatial and temporal variability of streambed hydraulic conductivity in West Bear Creek, North Carolina, USA. *Journal of Hydrology*, *358*, 332–353.
Glose, T. J., Lowry, C. S., & Hausner, M. B. (2021). Examining the utility of continuously quantified Darcy fluxes through the use of periodic temperature time series. *Journal of Hydrology*, *595*, 125675.
Goldschneider, A. A., Haralampides, K. A., & MacQuarrie, K. T. B. (2007). River sediment and flow characteristics near a bank filtration water supply: Implications for riverbed clogging. *Journal of Hydrology*, *344*, 55–69.
Hatch, C. E., Fisher, A. T., Ruehl, C. R., & Stemler, G. (2010). Spatial and temporal variations in streambed hydraulic conductivity quantified with time-series thermal methods. *Journal of Hydrology*, *389*, 276–288.
Hatch, C. E., Fisher, A. T., Revenaugh, J. S., Constantz, J., & Ruehl, C. (2006). Quantifying surface water–groundwater interactions using time series analysis of streambed thermal records: Method development. *Water Resources Research*, *42*, W10410. https://doi.org/10.1029/2005WR004787
Irvine, D. J., Briggs, M. A., Lautz, L. K., Gordon, R. P., McKenzie, J. M., & Cartwright, I. (2017). Using diurnal temperature signals to infer vertical groundwater-surface water exchange. *Groundwater*, *55*, 10–26.
Jin, G., Chen, Y., Tang, H., Zhang, P., Li, L., & Barry, D. A. (2019). Interplay of hyporheic exchange and fine particle deposition in a riverbed. *Advances in Water Resources*, *128*, 145–157.
Karan, S., Kidmose, J., Engesgaard, P., Nilsson, B., Frandsen, M., Ommen, D. A. O., Flindt, M. R., Andersen, F. Ø., & Pedersen, O. (2014). Role of a groundwater–lake interface in controlling seepage of water and nitrate. *Journal of Hydrology*, *517*, 791–802.
Käser, D. H., Binley, A., Heathwaite, A. L., & Krause, S. (2009). Spatio-temporal variations of hyporheic flow in a riffle-step-pool sequence. *Hydrological Processes*, *23*, 2138–2149.
Kazmierczak, J., Müller, S., Nilsson, B., Postma, D., Czekaj, J., Sebok, E., Jessen, S., Karan, S., Stenvig Jensen, C., Edelvang, K., & Engesgaard, P. (2016). Groundwater flow and heterogeneous discharge into a seepage lake: Combined use of physical methods and hydrochemical tracers. *Water Resources Research*, *52*, 9109–9130.
Kishel, H. F., & Gerla, P. J. (2002). Characteristics of preferential flow and groundwater discharge to Shingobee Lake, Minnesota, USA. *Hydrological Processes*, *16*, 1921–1934.
Koch, F. W., Voytek, E. B., Day-Lewis, F. D., Healy, R., Briggs, M. A., Lane, J. W., & Werkema, D. (2016). 1DTempPro V2: New features for inferring groundwater/surface-water exchange. *Groundwater*, *54*, 434–439.
Korus, J. T., Fraundorfer, W. P., Gilmore, T. E., & Karnik, K. (2020). Transient streambed hydraulic conductivity in channel and bar environments, Loup River, Nebraska. *Hydrological Processes*, *34*, 3061–3077.
Levy, J., Birck, M. D., Mutiti, S., Kilroy, K. C., Windeler, B., Idris, O., & Allen, L. N. (2011). The impact of storm events on a riverbed system and its hydraulic conductivity at a site of induced infiltration. *Journal of Environmental Management*, *92*, 1960–1971.
McBride, M. S., & Pfannkuch, H. O. (1975). The distribution of seepage within lakebeds. *U.S. Geological Survey Journal of Research*, *3*, 505–512.
Moore, W. S. (2010). The effect of submarine groundwater discharge on the ocean. *Annual Review of Marine Science*, *2*, 59–88.
Mutiti, S., & Levy, J. (2010). Using temperature modeling to investigate the temporal variability of riverbed hydraulic conductivity during storm events. *Journal of Hydrology*, *388*, 321–334.
Newcomer, M. E., Hubbard, S. S., Fleckenstein, J. H., Maier, U., Schmidt, C., Thullner, M., Ulrich, C., Flipo, N., & Rubin, Y. (2016). Simulating bioclogging effects on dynamic riverbed permeability and infiltration. *Water Resources Research*, *52*, 2883–2900.
Packman, A. I., & MacKay, J. S. (2003). Interplay of stream-subsurface exchange, clay particle deposition, and streambed evolution. *Water Resources Research*, *39*, 1097. https://doi.org/10.1029/2002WR001432
Rau, G. C., Andersen, M. S., McCallum, A. M., Roshan, H., & Acworth, R. I. (2014). Heat as a tracer to quantify water flow in near-surface sediments. *Earth-Science Reviews*, *129*, 40–58.
Ray, C., Melin, G., & Linsky, R. B. (Eds.). (2003). *Riverbank filtration: Improving source-water quality* Kluwer Academic Publishers. p. 364.
Rosenberry, D. O., & Pitlick, J. (2009a). Effects of sediment transport and seepage direction on hydraulic properties at the sediment-water interface of hyporheic settings. *Journal of Hydrology*, *373*, 377–391.
Rosenberry, D. O., & Pitlick, J. (2009b). Local-scale variability of seepage and hydraulic conductivity in a shallow gravel-bed river. *Hydrological Processes*, *23*, 3306–3318.
Rosenberry, D. O., Sheibley, R. W., Cox, S. E., Simonds, F. W., & Naftz, D. L. (2013). Temporal variability of exchange between groundwater and surface water based on high-frequency direct measurements of seepage at the sediment-water interface. *Water Resources Research*, *49*, 2975–2986.
Rosenberry, D. O., Toran, L., & Nyquist, J. E. (2010). Effect of surficial disturbance on exchange between groundwater and surface water in nearshore margins. *Water Resources Research*, *46*, W06518. https://doi.org/10.1029/2009WR008755
Santos, I. R., Eyre, B. D., & Huettel, M. (2012). The driving forces of porewater and groundwater flow in permeable coastal sediments: A review. *Estuarine, Coastal and Shelf Science*, *98*, 1–15.
Schalchli, U. (1995). Basic equations for siltation of riverbeds. *Journal of Hydraulic Engineering*, *121*, 274–287.
Schmidt, C. K., Lange, F. T., Brauch, H. J., & Kuhn, W. (2003). Experiences with riverbank filtration and infiltration in Germany. In *Proceedings of the International Symposium on Artifical Recharge of Groundwater* (pp. 117–131). Korea Water Resources Corporation.
Schubert, J. (2006). Experience with riverbed clogging along the Rhine River. In S. A. Hubbs (Ed.), *Riverbank filtration hydrology* (pp. 221–242). Springer.
Schulze-Makuch, D., Carlson, D. A., Cherkauer, D. S., & Malik, P. (1999). Scale dependency of hydraulic conductivity in heterogeneous media. *Ground Water*, *37*, 904–919.

Schumm, S. A., & Lichty, R. W. (1965). Time, space and causality in geomorphology. *American Journal of Science*, *263*, 110–119.

Sebok, E., Duque, C., Engesgaard, P., & Boegh, E. (2015). Spatial variability in streambed hydraulic conductivity of contrasting stream morphologies: Channel bend and straight channel. *Hydrological Processes*, *29*, 458–472.

Seifert, D., & Engesgaard, P. (2007). Use of tracer tests to investigate changes in flow and transport properties due to bioclogging of porous media. *Journal of Contaminant Hydrology*, *93*, 58–71.

Simpson, S. C., & Meixner, T. (2012). Modeling effects of floods on streambed hydraulic conductivity and groundwater-surface water interactions. *Water Resources Research*, *48*, W02515. https://doi.org/10.1029/2011WR011022

Smith, J. W. N. (2005). Groundwater–surface water interactions in the hyporheic zone. *Environment Agency Science Report SC030155/SR1*; pp. 65.

Springer, A. E., Petroutson, W. D., & Semmens, B. A. (1999). Spatial and temporal variability of hydraulic conductivity in active reattachment bars of the Colorado River, Grand Canyon. *Ground Water*, *37*, 338–344.

Stewardson, M. J., Datry, T., Lamouroux, N., Pella, H., Thommeret, N., Valette, L., & Grant, S. B. (2016). Variation in reach-scale hydraulic conductivity of streambeds. *Geomorphology*, *259*, 70–80.

Stuyfzand, P. J., Juhàsz-Holterman, M. H. A., & de Lange, W. J. (2006). Riverbank filtration in the Netherlands: Well fields, clogging and geochemical reactions. In S. A. Hubbs (Ed.), *Riverbank filtration hydrology* (pp. 119–153). Springer.

Su, G. W., Jasperse, J., Seymour, D., Constantz, J., & Zhou, Q. (2007). Analysis of pumping-induced unsaturated regions beneath a perennial river. *Water Resources Research*, *43*, W08421. https://doi.org/10.1029/2006WR005389

Toran, L., Nyquist, J., Rosenberry, D., Gagliano, M., Mitchell, N., & Mikochik, J. (2015). Geophysical and hydrologic studies of lake seepage variability. *Groundwater*, *53*, 841–850.

U.S. EPA. (2000). Proceedings of the ground-water/surface-water interactions workshop. *EPA/542/R-00/007*; p. 200.

Ulrich, C., Hubbard, S. S., Florsheim, J., Rosenberry, D., Borglin, S., Trotta, M., & Seymour, D. (2015). Riverbed clogging associated with a California riverbank filtration system: An assessment of mechanisms and monitoring approaches. *Journal of Hydrology*, *529*, 1740–1753.

Vanek, V. (1997). Heterogeneity of groundwater-surface water ecotones. In J. Gibert, J. Mathieu, & F. Fournier (Eds.), *Groundwater-surface water ecotones: Biological and hydrological interactions and management options* (pp. 151–161). Cambridge University Press.

Volkenborn, N., Polerecky, L., Hedtkamp, S. I. C., van Beusekom, J. E. E., & de Beer, D. (2007). Bioturbation and bioirrigation extend the open exchange regions in permeable sediments. *Limnology and Oceanography*, *52*, 1898–1909.

Volkenborn, N., Woodin, S. A., Wethey, D. S., & Polerecky, L. (2019). Bioirrigation. In J. K. Cochran, H. J. Bokuniewicz, & P. L. Yager (Eds.), *Encyclopedia of ocean sciences* (3rd ed., pp. 663–670). Academic Press.

Wang, L., Song, J., Zhang, B., Guo, H., Jiang, W., Wen, M., & Zhang, G. (2016). Spatial and temporal variations of streambed vertical hydraulic conductivity in the Weihe River, China. *Water*, *8*, 70. https://doi.org/10.3390/w8030070.

Weeks, E. P., Ericson, D. W., & Holt Jr. C. L. R. (1965). *Hydrology of the Little Plover River basin, Portage County, Wisconsin and the effects of water resource development*. U.S. Geological Survey Water-Supply Paper 1811; p. 78.

Wiese, B., & Nützmann, G. (2009). Transient leakance and infiltration characteristics during lake bank filtration. *Ground Water*, *47*, 57–68.

Winter, T. C., Harvey, J. W., Franke, O. L., & Alley, W. W. (1998). *Ground water and surface water: A single resource*. U.S. Geological Survey Circular 1139; p. 79.

Wolman, M. G., & Miller, J. P. (1960). Magnitude and frequency of forces in geomorphic processes. *Journal of Geology*, *68*, 54–74.

Wu, G., Shu, L., Lu, C., Chen, X., Zhang, X., Appiah-Adjei, E. K., & Zhu, J. (2015). Variations of streambed vertical hydraulic conductivity before and after a flood season. *Hydrogeology Journal*, *23*, 1603–1615.

Yellen, B., & Boutt, D. F. (2015). Hydropeaking induces losses from a river reach: Observations at multiple spatial scales. *Hydrological Processes*, *29*, 3261–3275.

Zhang, Y., Hubbard, S., & Finsterle, S. (2011). Factors governing sustainable groundwater pumping near a river. *Ground Water*, *49*, 432–444.

How to cite this article: Rosenberry, D. O., Engesgaard, P., & Hatch, C. (2021). Hydraulic conductivity can no longer be considered a fixed property when quantifying flow between groundwater and surface water. *Hydrological Processes*, *35*(6), e14226. https://doi.org/10.1002/hyp.14226