

# Factors influencing streambed hydraulic conductivity and their implications on stream–aquifer interaction: a conceptual review

Sujay Raghavendra Naganna<sup>1</sup> · Paresh Chandra Deka<sup>1</sup> · Sudheer Ch<sup>2</sup> · William F. Hansen<sup>3</sup>

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**Abstract** The estimation and modeling of streambed hydraulic conductivity ( $K$ ) is an emerging interest due to its connection to water quality, aquatic habitat, and groundwater recharge. Existing research has found ways to sample and measure  $K$  at specific sites and with laboratory tests. The challenge undertaken was to review progress, relevance, complexity in understanding and modeling via statistical and geostatistical approaches, literature gaps, and suggestions toward future needs. This article provides an overview of factors and processes influencing streambed hydraulic conductivity ( $K$ ) and its role in the stream–aquifer interaction. During our synthesis, we discuss the influence of geological, hydrological, biological, and anthropogenic factors that lead to variability of streambed substrates. Literature examples document findings to specific sites that help to portray the role of streambed  $K$  and other interrelated factors in the modeling of hyporheic and groundwater flow systems. However, studies utilizing an integrated, comprehensive database are limited, restricting the ability of broader application and understanding. Examples of in situ and laboratory methods of estimating hydraulic conductivity suggest challenges in acquiring representative samples and comparing results, considering the anisotropy and heterogeneity of fluvial bed materials and geohydrological

conditions. Arriving at realistic statistical and spatial inference based on field and lab data collected is challenging, considering the possible sediment sources, processes, and complexity. Recognizing that the  $K$  for a given particle size group includes several to many orders of magnitude, modeling of streambed  $K$  and groundwater interaction remain conceptual and experimental. Advanced geostatistical techniques offer a wide range of univariate or multi-variate interpolation procedures such as kriging and variogram analysis that can be applied to these complex systems. Research available from various studies has been instrumental in developing sampling options, recognizing the significance of fluvial dynamics, the potential for filtration, transfer, and storage of high-quality groundwater, and importance to aquatic habitat and refuge during extreme conditions. Efforts in the characterization of natural and anthropogenic conditions, substrate materials, sediment loading, colmation, and other details highlight the great complexity and perhaps need for a database to compile relevant data. The effects on streambed hydraulic conductivity due to anthropogenic disturbances (in-stream gravel mining, contaminant release, benthic activity, etc.) are the areas that still need focus. An interdisciplinary (hydro-geo-biological) approach may be necessary to characterize the magnitude and variability of streambed  $K$  and fluxes at local, regional scales.

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✉ Sujay Raghavendra Naganna  
sujay.am14f03@nitk.edu.in

<sup>1</sup> Department of Applied Mechanics and Hydraulics, National Institute of Technology Karnataka, Surathkal, Mangalore 575025, India

<sup>2</sup> Ministry of Environment, Forest and Climate Change, New Delhi, India

<sup>3</sup> Hansen Hydrology, Lexington, SC, USA

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## Introduction

The purpose of this brief conceptual review is to discuss various geological, hydrological, anthropogenic, and biological factors that influence streambed hydraulic conductivity and

summarize the available laboratory and in situ methods of estimating streambed hydraulic conductivity ( $K$ ). The statistical and geostatistical methods employed for spatial interpolation of streambed  $K$  are qualitatively evaluated. The article addresses the current status and trend and suggests possible improvements to future work in the field of streambed hydraulic conductivity. However, due to the complexity of this subject, it is unlikely to present all the factors concerning streambed hydraulic conductivity and its connections between streams and groundwater. Due to the limited amount of research on this subject, and also the variability across landscapes and stream systems, it is inappropriate to try to describe all the conditions that may be encountered, but it is hoped that this review will help inform broadly on the subject.

The river basin acts as a system, distributing precipitation into the recognized surface network of channels and streams. The subsurface flow patterns and the stream–aquifer interactions via hyporheic zone are considerably more complex to characterize. It is not hard to imagine that most rivers dominated by gravel substrates have a permeable bed and an hyporheic zone with high porosity composed by gravels beneath it, and an intensive interaction exists between free-surface flow and subsurface flow (Gordon et al. 2004). But when dealing with interactions between surface water, hyporheic and groundwater, spatial and temporal variability of water transport within the saturated and unsaturated zone exists and is related to channel bed morphology, relic valley materials, hydraulic properties (such as hydraulic conductivity, porosity), the roughness of stream substrate, and geohydrological processes (Kalbus et al. 2006; Valerio et al. 2010; Fleckenstein et al. 2010). Factors such as the depth to impermeable surface or partitioning of subsurface flow among the interconnected domains of the soil macro-pore system and the surrounding geology add to the difficulty to understand, sample, and model these complex systems (Dillon and Liggett 1983; Cardenas 2009; Tang et al. 2015). Interdisciplinary information sharing and modeling approaches may be used to help discover anomalies, inconsistencies, and irregularities in the processes that affect these systems.

Wetlands, streams, rivers, and lakes are seldom isolated, but typically hydraulically coupled or connected to some extent with underlying groundwater reservoir, with water passing back and forth with the flow interaction pattern mostly dependent on rainfall inputs, water head changes, and substrate permeability (Castro and Hornberger 1991). Depending on the amplitude of topographic variations, geometric and hydraulic properties of the streambed; the travel time, length, and depth of the groundwater flow path vary significantly from the points of recharge or collection to the points of discharge or dispersal (Tóth 1962). Soil hydraulic properties are highly non-linear functions which may contain irregularities and pose variable circumstances. For instance, the hydraulic conductivity of a specific layer or substrate

may control the net water movement rates and exchange among surface and groundwater systems. Streambed hydraulic conductivity is a key physical parameter, controlling the water fluxes across the stream–aquifer interface. It is highly influenced by streambed substrate characteristics (e.g., structure, effective porosity, grain size, packing), streambed alteration processes (e.g., aggradation, degradation, sedimentation, colmation, and erosion), and the configuration of stream channel geometry, floodplain connection, and streambed morphology (e.g., dunes, anti-dunes, pool-riffle sequences) (Schumm et al. 1984; Schneidewind et al. 2015). A variety of natural and anthropogenic disturbances also contribute to conditions necessitating consideration of the history of geologic to anthropogenic changes. The streambed composition of various materials varying from bed load to fine-grained sediments does not exhibit constant properties or conditions of stable equilibrium, as they are subjected to alternate wet and dry conditions, swelling and shrinks, dispersion and flocculation, cracking, compaction, aggregation, experience bio-chemical changes, and structural rearrangements in channel adjustments (Reid and Frostick 1987; Rehg et al. 2005). The streambed hydraulic conductivity affects the groundwater residence time and the potential for processes such as filtration, mixing, chemical interchange, decomposition, and transport. Differences in material sources and types are also due to hydrologic position and scale, such as from higher gradient channels in the headwaters to low gradient valley rivers. Variations in groundwater residence and inflow and outflow rates can be associated with the hydraulic resistance of the clogging layer at the wetted perimeter of a streambed and river channel morphology (Katsuyama et al. 2010). Groundwater recharge through streambed depends on a variety of spatio-temporal factors such as the topography of the area, the velocity and depth of river stage, and the hydrogeologic setting of the underlying aquifer (Woessner 2000; Wang et al. 2016).

### Factors affecting streambed hydraulic conductivity

The accretion or erosion of streambed elements, transformations in channel morphology, and river network systems are affected by precipitation and flow events, underlying geology, soils, topography, erosion, sedimentation, organic inputs, and natural and anthropogenic disturbances. Furthermore, their specific influence and consequence may change with different topographic position and hydrologic scales. Hydraulic conductivity ( $K$ ) is the volume of water at the existing kinematic viscosity that will move in unit time through a unit area of the saturated substrates, such as geology, soil, or feature such as streambed or aquifer measured at right angles to the direction of flow under a unit hydraulic gradient. Changes in the hydraulic head above the substrate affect flow rate and are accounted for in Darcy's law. Exact boundaries and the

influence of most of the factors (mentioned in Fig. 1) is seldom sufficiently known due to the spatial and temporal complexity. Diverse geological histories and climatic conditions influence landscape and stream conditions resulting in the highly variable conditions. Most variables can be categorized as abiotic such as floods, stream morphology, sediment sources and composition, depth to impermeable sediment or bedrock, and chemistry. Other minor factors such as barometric pressure and moon position influence surface water levels and density which are often difficult to measure precisely (Toll and Rasmussen 2007). Biotic influences contribute to complexity such as large wood, leaves, root systems, shade, transpiration, colmation (clogging of the substrate with organic particles), macroinvertebrates, and beavers. Anthropogenic disturbances such as flow regime modification by dams, artificial ditching, river channelization, drainage of wetlands, and land use changes may influence streambed structure. Substrate modification may also be altered by runoff from impermeable surfaces, farming or clear-cutting of erodible or unstable landforms, or mining of channel substrates (Raghavendra and Deka 2015).

Climate change through geologic time has many implications to landform development in coastal areas due to sea level changes, deposit of marine terraces, etc. (Kemp et al. 2011). Glaciation expansion and retraction have modified some landscapes and stream channels exposing the extreme forces associated with snow accumulations and ice. Channel instability can be evidenced through changes in morphology via signs of degradation such as gully entrenchment, bank instability or aggradation due to excessive sediment loading, and channel infilling and braiding. The in-channel or near-channel degradation, therefore, can alter streams essentially into high water table swamps in a braided channel form to lower water table

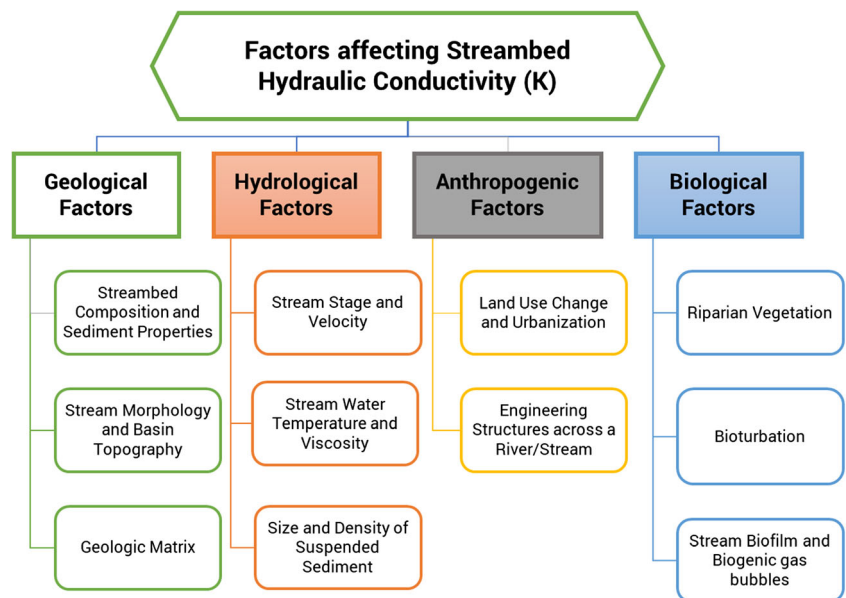
systems with increased drainage due to gullying (Trimble 2008). There are probably many such examples of local or area wide circumstances that have altered landforms that contribute to the variability of channel substrates and hydraulic conductivity. Some forces involved seem less extreme, such as wind-blown soils leaving loess deposits, tree blowdown or autumn leaf fall that can accumulate and clog channel substrates. Hurricanes, monsoons, floods, earthquakes, volcanism, plate tectonics, and severe wildfires are a few examples of extreme forces that can modify landscapes, stream morphology, and substrate materials (Schumm et al. 1984; Rosgen 1996). Within this context, one should also identify both upstream and downstream linkages which can be complex and difficult to generalize with respect to stream–aquifer connections and functions.

**Influence of geological and hydrological factors on streambed hydraulic conductivity**

*Effects of streambed composition and sediment properties*

The geology and hydrology based physical properties and factors are sometimes difficult to separate, but geology includes earth forming and modifying events over substantial time periods that have led to the base landforms and landscapes now present. The geologic factors that primarily influence streambed hydraulic conductivity are sediment particle size, substratum heterogeneity, longitudinal variations in impervious surfaces such as bedrock and sills, bed material depth, channel geometry, variations in hydraulic radius, and roughness due to natural and anthropogenic modifications (Jackson 1980; Stewardson et al. 2016). The other influencing factors that are uncommon include events such as consolidation

**Fig. 1** Factors affecting streambed hydraulic conductivity



or alteration of sediments through time, volcanic, tectonic, earthquakes, folding, and fracturing and shearing of bedrock materials (King and Wood 1994; Dale et al. 2005). Geology variations produce landform and material differences resulting in bedrock materials that influence streambed composition and differences in hydraulic conductivity. The existing stream channel may have undergone many changes through time as evidenced in its valley sediment record and variations in substrates. Some streams have a complex history of channel adjustment in elevation, position, and substrate change through time within the boundary of their valley reflecting periods of sediment accumulation along with channel adjustment through erosion, migration, degradation, or avulsion (Rosgen 2009).

Tables 1 and 2 provide representative values of hydraulic conductivity for various unconsolidated sedimentary materials and soils interfacing with permeability classes which are sometimes applied in the discussion. The range of hydraulic conductivity varies over several orders of magnitude for each sediment particle size class, reflecting variability that can be caused by settling, clogging of fine mineral or organic particles, compaction due to the weight of eroded materials, etc. Hence, the determination of  $K$  is not that simple or precise, due to uncertainty in stream–aquifer interactions resulting from the inherent complex-nested flow paths and particle size differences in substrate layers that pose difficulty to make generalizations.

The importance of hyporheic zone and its interchange potential with surface and subsurface flow includes not only the benefit of pollutant filtration from this process but also to what extent is this contributing to deep seepage so as to replenish groundwater or aquifer systems. The mechanisms/factors involved with deep seepage losses when conducting the water balance of these systems include the secondary porosity of the geologic units, including faults, bedding plane partings, dipping beds, joints, axial plane cleavage, crossbedding, facies, shear lineaments, and other fractures (Tóth 1962; Ingebritsen and Sanford 1999). Significant higher  $K$  at some sections of the streambed may be attributable to the underlying joint

**Table 1** Hydraulic conductivity of various unconsolidated sedimentary materials

Substrate particle type	Hydraulic conductivity (m/s)
Gravel	$3 \times 10^{-4}$ to $3 \times 10^{-2}$
Fine sand	$2 \times 10^{-7}$ to $2 \times 10^{-4}$
Medium sand	$9 \times 10^{-7}$ to $5 \times 10^{-4}$
Coarse sand	$9 \times 10^{-7}$ to $6 \times 10^{-3}$
Silt, loess	$1 \times 10^{-9}$ to $2 \times 10^{-5}$
Till	$1 \times 10^{-12}$ to $2 \times 10^{-6}$
Clay	$1 \times 10^{-11}$ to $4.7 \times 10^{-9}$
Un-weathered marine clay	$8 \times 10^{-13}$ to $2 \times 10^{-9}$

Source: Domenico and Schwartz (1990)

**Table 2** Substrate particle classification based on saturated hydraulic conductivity values

Permeability of soil (according the relative permeability)	Approximate range of saturated hydraulic conductivity (m/s)	Examples of soil types
Highly impermeable	$< 10^{-10}$	Clays with low, medium and high plasticity
Impermeable	From $10^{-8}$ to $10^{-10}$	Gravel loams, gravel clays and sandy clays, loams with low and medium plasticity
Lowly (poorly) permeable	From $10^{-6}$ to $10^{-8}$	Sandy loams, loamy sands and clayey sands, loamy gravels and clayey gravels
Permeable	From $10^{-4}$ to $10^{-6}$	Sands and gravels containing fine-grained fraction (5–15%)
Highly permeable	$> 10^{-4}$	Sands and gravels without or with very low fine-grained fraction ( $< 5\%$ )

Source: (Czech standard CSN 721020: replaced by ISO/TS 17892-11:2004) (ISO/TS.17892-11:2004 2004)

systems and bedding planes and discontinuities such as limestone inclusions (Miller et al. 2014). Dipping beds can change the direction of subsurface flow, while an interconnected system of fractures, even with some discontinuities could influence the streambed transfer of flow to groundwater or aquifer recharge. The extent and orientation of streambed fissures, sediment heterogeneity in both the horizontal and vertical direction, erodibility of materials that favor piping, dissolution such as in karst terrain contributing to underground caves, springs, etc., influence the groundwater flow path and extent (Harter and Rollins 2008).

Happ et al. (1940) categorized the sediment deposits in the sampled valleys in Mississippi, USA, into colluvial deposits (that which accumulates from adjacent hillslopes along valley sides), vertical and lateral accretion deposits over floodplains, channel fill deposits accumulated due to deficient stream power, alluvial fans associated with recent valley sediments originating from side channels or tributaries, splays (such as older silt deposits moved in streambank levee failure), and channel lag deposits which are relatively coarser materials sorted out in the process of stream action. Understanding on the channel morphology and sedimentation history renders substantial benefit in sampling and reporting of streambed  $K$  studies. The fundamental fluvial processes through time have implications applicable in measuring and modeling streambed  $K$ . The well-sorted, younger, fresh sediment deposits are likely to have a greater hydraulic conductivity than the formerly entrenched valley deposits due to more frequent and more

irregular vertical and lateral changes in texture. As time passes, the colluvial and alluvial deposits in the valleys with repeated wetting and drying cycles tend to settle, coalesce, and develop into soils with vegetation development and more recognizable horizons. Alluvial fan deposits at valley floors are stratigraphically complex, commonly displaying a high degree of heterogeneity in their sediment properties indicated by the variance of  $K$  (Hamill and Bell 2013).

Hyporheic zone composition and its degree of interconnection with the zones below may vary in both horizontal and vertical pathways due to the accumulation of abundant fines and humic compounds of hydrophobic nature (Kasahara and Wondzell 2003). Interbedding of sand and silt layers affects the frequency, openness, continuity, and amount of fine structure partings leading to anomalies in local and regional groundwater flow path (Xin et al. 2012; Midgley et al. 2013). The interaction of stream–aquifer interface via hyporheic zone is often influenced by colmation (a process of plugging of the streambed substrate pores by the settling of the suspended colloidal matter leading to consolidation of the hyporheic zone and minimization the bed hydraulic conductivity). Clogged streambed sediments are characterized by tight packing and a compact texture, with a low porosity. Colmation results in a low permeability sediment layer, known as a colmatage (Brunke 1998). If the diameter of clogging particles is larger than pore throats of the porous medium, the particulate matter gets entrapped at pore constrictions at various depths resulting in decreased hydraulic conductivity of the porous medium. Finer-grained riverbed is more vulnerable to external clogging than coarse-grained (Rosenberry et al. 2010). The colmation/decormation of porous media by accretion/erosion of organic and inorganic particulate matter reduces/increases the hydraulic conductivity of the porous media depending on the geometry of the porous medium (Baveye et al. 1998). The clogging layer does not seem to be stable over the long term, because of seasonal hydrostatic pressure variations. During the rising stage of storm or flood events, the relatively abrupt shift in stream power may entrain and move particles so that the initial hydraulic conductivity of the streambed substrates could be regained by reworking the subsurface sediment structure and decormation. Brunke (1999) provides an overview of the mechanism, factors, and processes involved in colmation and decormation, discussing the effects and differences between internal and external colmation. The factors that influence the temporal course of colmation, the clogging depth, and their effects on hydraulic conductivity are briefly summarized in Table. 3. Terms sometimes used in publications may discuss infiltration or porosity rates as a surrogate or relation to  $K$ .

Geogenic disturbances such as earthquakes may not only alter the flow pattern via drainage path modification but also may impact material types and delivery through landslides. The resultant soil hydraulic conductivity changes may occur

due to depth, compaction, channel filling, and its spatial variation as a consequence of redistribution of materials and dissipation of excess pore pressures both during and after seismic activity (Rai et al. 2005; Seid-Karbasi et al. 2011).

#### *Effects of stream morphology and basin topography*

Historically, channels adjust slowly to abruptly to reflect current conditions, but often leave remnant data that help define their past. Erosion, transport, and sedimentation are the key channel processes responsible for changes in river/stream morphology, with factors such as valley or channel gradient, substrate materials, and vegetation playing a role (Schumm et al. 1984; Boyer et al. 2006; Rosgen 2009). Rosgen (1994) classified/categorized stream patterns based on factors such as entrenchment, sinuosity, gradient, and cross-section and longitudinal profile of streams into nine primary forms, with about six substrate types for each form. Use of this or another acceptable classification system during data collection and field verification would help readers and technical experts understand conditions of research. Although researchers have expressed concerns or have issues with classifications made by Rosgen (1994) for channels or Happ et al. (1940) for valley sediments, there is a need for some standardization and communication tools to expand the ways to help transfer technology. One advantage to consider is the coupling of Rosgen sampling methods with the Rivermorph software which aids in storing, analyzing, and presenting data (Belcher and Athanasakes 2002). The channel types inferred via topographic maps, high-resolution aerial photos, typically require field verification, but sometimes recognizable from Light Detection and Ranging (LiDAR) digital analysis (James et al. 2007; Maceyka and Hansen 2016). The stream classification and associated methods provide background information useful for detailed characterization and assessment of streambed hydraulic properties.

Braided streams occur in rivers when the threshold level of sediment load significantly exceeds the single thread channel capacity resulting in the infilling or aggradation of the channel and to some extent valley surface with the development of multiple channels, often sinuous and low habitat. Braided streams for all their habitat limits may have enhanced stream, with highly hydrated floodplain and the potential for greater groundwater connection (Surian 2015). With ample and well-distributed rainfall, braided systems have been sometimes described as “swamped” for their lost efficiency in moving water resulting in the high water table (Trimble 2008). The braided pattern is anticipated to present the highest diversity and extent of surface and subsurface exchanges facilitating high hydrological connectivity and aquifer recharge (Tockner et al. 2009).

**Table 3** Studies related to effects of colmation on streambed hydraulic conductivity

Reference and test site	Type of colmation or decolmation	Influencing factors and their controls	Implications on hydraulic conductivity ( <i>K</i> )
(Carling 1984) Experimental flume	Siltation of experimental gravel bed	Shear stress, particle size, water depth, particle Reynolds number, Froude number, and suspended sediment concentration influence the degree of siltation. At Froude numbers > 0.70, no sediment transport was seen; the particles in suspension were entrapped only on the surface void space.	Due to siltation, there may be characteristic particle size distribution resulting in surface sealing, a decrease in hydraulic conductivity and macro-pore connectivity. However, turbulent resuspension of fine substrates influences the accretion on the surface layer of gravel.
(Cunningham et al. 1987) Experimental recirculating flume	Channel bed clogging in the sediment-laden flow. Impaction-type and depositional clogging	Average channel velocity, suspended sediment concentration, and its particle size distribution, hydraulic head, channel hydraulic radius, and kinematic viscosity are the hydraulic and sedimentation processes affecting channel bed clogging along with deposits of biological activity. Impaction-type clogging corresponds to a hydraulically rough bed condition where a fine layer of suspended sediment load is impacted into the pore spaces of the surface layer of the bed via turbulence. Depositional clogging can occur both in laminar and transitional flow conditions where initially the larger suspended particles settle out allowing for interstitial straining of fine sediments.	The suspended sediment concentration in the range of 200–1600 ppm solids content had no significant effect on the clogging process and infiltration rates. There exists a direct relationship between flow velocity and infiltration rate and an inverse relationship between suspended sediment content and infiltration rate. The infiltration rate may not be equal to the streambed hydraulic conductivity under unsaturated conditions due to the effect of greater matric suction in the drier soil beneath the wetting front.
(Schalchli 1992) Laboratory investigation of field samples collected from the Langeten River in Switzerland	Clogging of river bed Flushing of clogged layer	Clogging process is characterized by the dimensionless flow shear stress $\Theta$ (for $\Theta < 0.056$ , the clogging process is accelerated; for $\Theta > 0.060$ , the clogged layer starts to break up). If the groundwater is contributing to the river flow, the clogged particles are flushed out of the bed markedly leading to loose-packed bed texture.	The streambed tends to regain its initial hydraulic conductivity during a flood event, as the dimensionless shear stress of flow surpasses the critical value of the stability of the armor layer, thus disturbing the clogged layer/bed.
(Brunke 1999) In the transect across a calcareous prealpine gravel bed river, Toss River, Switzerland	Internal and external colmation Depth filtration	Internal colmation: The clogging of the gravel bed interstices directly below the armor layer may inhibit or disconnect the exchange processes between surface water from hyporheic water. External colmation: A thick layer of fine particulate matter (settling over the armor layer) may develop due to an extended period of low current velocity reducing the permeability of the streambed. There may be colmation and decolmation due to varying flow intensity and anthropogenic disturbances.	Due to internal and external colmation, there was a continuous reduction in the hydraulic conductivity of the bed up to 40–50 cm depth from the surface; however, it increased in the subsequent depths. External colmated bed sediments are characterized by a consolidated texture having reduced hydraulic conductivity and porosity. Lower levels of fines were observed near the bed surface. The internal colmation was severe and the proportion of fines < 1 mm in the bed increased with depth.
(Rehg et al. 2005) Experimental laboratory flume	Artificial clogging by injection of kaolinite clay	Clogging is a complex process which simply does not depend on the volumetric fraction of clay in the streambed. It varies temporally and is a function of both suspended load and transported bed material. The intense interaction between the suspended matter and bed sediments in the form of excess external clogging can potentially degrade benthic and hyporheic habitat. An immobile/stable bed is more likely to clog in the near-surface region and episodic mobilization of bed sediments result in clogged particles extending deeper into the bed.	Progressive accumulation of clay in the bed significantly decreases the effective streambed porosity and hydraulic conductivity. Hyporheic macro invertebrates have the potential to rework/alter the subsurface sediment structure and help maintain sediment permeability.
(Descloux et al. 2010) Catchment area of the Rhône River, Europe	Surface and subsurface colmation	The factors such as hydraulic head, heterogeneity of the particulate matter, and suspended sediment load influence on the subsurface	A curvilinear relationship with a good correlation was observed between the

**Table 3** (continued)

Reference and test site	Type of colmation or decolmation	Influencing factors and their controls	Implications on hydraulic conductivity ( <i>K</i> )
		colmation patterns that could result in the different hydrosedimentary functioning of the river. The vertical distribution of fine sediments in the surface and subsurface layers may vary significantly, with a higher proportion of fines in the deeper layers. The article provides a summary of different methods to assess embeddedness and colmation.	percentage of subsurface fine sediments and hydraulic conductivity. The hydraulic conductivity measurements could represent the valuable surrogate outcome of colmation levels. Higher fractions of fine sediments in the bed (> 20%) may result in a very low hydraulic conductivity.
(Datry et al. 2015) Several stream reaches in France	Internal and external colmation	There may be zones of anoxic and anaerobic conditions with biogeochemical processes such as fermentation during low flow periods leading to biological colmation (interstitial biofilm and surface algal mat).	The prior identified clogged reaches held much lower streambed hydraulic conductivity than the reference (unclogged) reaches.
(Pholkern et al. 2015) Ping River, Thailand	Clogging due to induced riverbank filtration Internal and external clogging	The lower velocity flow and higher turbidity levels lead to dense clogging. Due to the variation in the distribution of flow velocity gradient, the clogged areas are usually wider near the inside banks (point bars) and less extensive in the thalweg of the stream or outside turn. Induced infiltration through riverbanks leads to clogging in the first few centimeters of the riverbed causing obstruction to bank filtrate due to reduced hydraulic conductivity.	The clogging layer is found to decrease the riverbed permeability and abstraction rates in the case of induced infiltration through riverbank. The external clogging may have a lower impact on vertical <i>K</i> , while the internal clogging common to a greater depth decreases the <i>K</i> significantly.
(Stewardson et al. 2016) Several streams reaches in France	Internal and external clogging	The episodic scour-and-fill processes, biofilm growth and its straining, settling, or chemical adhesion on sediments may cause cyclical clogging behavior. The clogging of streambed evolves by initiation of bed mobilization, tailed by a transient phase and ends up with a saturated phase.	The streambeds with less frequent disturbances and higher suspended load flows have lower hydraulic conductivity.
(Grischek and Bartak 2016) Elbe River	Physical Mechanical Biological and Chemical clogging	Discussion of clogging types as Physical (deposition, filtration of suspended solids), Mechanical (gas entrapment), Biological (bacterial growth) and Chemical (precipitation reaction). The term leakage coefficient (reciprocal of infiltration resistance) is introduced which is the ratio of vertical hydraulic conductivity of the clogging layer to the thickness of the clogging layer	River water pollution by organics from anthropogenic activities can lead to unsaturated conditions beneath the riverbed thus affecting the hydraulic conductivity. During lower river stage in navigable rivers, the ship traffic can pose a stronger impact on the riverbed and its <i>K</i> value which promotes riverbed leakage.

Anastomosed channels are braided, but stabilized in place by well-vegetated settings such as rushes and sedges or dense bottomland with often fine grain substrates; however, the gravel beds are sometimes exposed by avulsions (i.e., flow diversions) that cause the formation of new channels on the floodplain (Makaske 2001). Anastomosed channels are fairly common in tropical and subtropical low gradient systems such as marine terraces and broad valley floodplains. Localized conduits of underground flow may develop in the alluvial systems due to soil piping, large wood burial from storms, the presence of remnant channels abandoned through time, and past or existing dredging or drainage operations that

may still have some function. Extensive cohesive deposits of fine sediments across portions of floodplains having a low hydraulic conductivity add to the variability of conditions found in some anastomosed sections (Heritage et al. 2009).

The quantum of sediment deposition is less in meandering-single thread river sections when compared to braided river sections. The stable single thread bankfull channel is able to move most of the sediment supply without aggrading or degrading and maintain regular floodplain access during extreme flow events. Progressive erosion and deposition of fine-grained sediments over the channel margins and on the adjacent floodplains are generally observed in the non-entrenched

single thread streams. The sediment eroded from the cut-bank side of the meander-river bend section includes fine-entrained particles that move extended distances as well as a coarser mix of particles that may get deposited down-current on the point bar side, ensuing the lateral movement of the river bend over time (Slatt 2006). In a meandering-river system, at zones where point bar accumulations of water washed sediments are accumulated, the stream water table interaction is favored with potential ramifications for groundwater-aquifer recharge (Ward et al. 2015). However, the water transfer may be greatest in thalweg, due to increased water velocity, less substrate depth, and resistance. The point bars will not be so effective, possibly due to clogging by fines or organic particles as compared to the lower velocity channel margins (Eekhout et al. 2013). Willis and Tang (2010) document the various patterns of fluvial meander migration which cause different deposit geometries and altered distribution of grain size, permeability, and porosity. Nowinski et al. (2011) studied the meander-scale changes in hydraulic conductivity of an alluvial aquifer adjacent to a stream and reports the temporal patterns of hydraulic conductivity. The lower initial hydraulic conductivity zones become further less permeable due to the hyporheic transport of fine materials that are flushed from upstream decolmated areas. Relatively rapid sampling approaches of the channel and point bar substrates have been used to help classify channel energy in the availability and movement of particles (e.g., pebble counts, riffle and base of point bar) (Rosgen 1994). Increased use and application of Rosgen or similar methods developed may increase our ability to apply surface metrics and observations in the process of correlating with and estimating subsurface conditions.

Straight channels are characterized by increased channel gradients, entrenchment (i.e., limited floodplain), high stream power, and erosion-resistant banks (Schumm et al. 1984; Makaske 2001). Without high gradient, natural straight channels are unusual without some geological control such as shear lineament. Alternate bars or side bars may still exist, but be poorly formed as a consequence of steeper gradients and higher velocities. The porosity and  $K$  of individual bars vary across the length of bar deposits, and those with gravel deposits are expected to have higher  $K$  in potential recharge zones during floods (Bridge 2003; Obana et al. 2014). Recharge from the straight river channels may be larger locally because of a fractured, sheared geology or due to coarse surficial deposits of higher  $K$  (Matsuda 2004). Stream degradation (channel entrenchment) into the gully form increases the rate of water removal from the land and adjacent water table along stream channels with associated potential implications and likely reductions in groundwater recharge and hyporheic flow. The base-level adjustments, gullies also favor channel expansion and drainage, favoring surface flow at the expense of water tables and base flows.

Higher gradient stream systems are a result of geology and topography, but sometimes affected by land uses in erodible situations that result in gullying into hillslopes. Waterfalls represent major elevation drops and interruptions in the stream or river current. High gradient sections without the major drops in elevation are cascades; a series of small falls and highly turbulent flow along a river. The hydraulic conductivity of the bed material (typically cobbles and boulders) near cascades which are generally disorganized (both laterally and longitudinally) is usually very high because of energy to remove fine materials (Ward et al. 2015). These higher gradient streams are more common to mountainous and hilly regions. Within mountainous or high gradient conditions where the channel is in close proximity to bedrock, with a dominance of coarse substrates present, it is difficult to assume that there is much groundwater connection as surface flow and rapid response are likely to dominate unless the geology is highly fractured allowing for flow to be removed and relocated to other areas (Tóth 1962). When there are abrupt changes in the slope of the streambed, some gaining segments can emerge within the watercourse where the channel base is lower than the level of the surrounding water table. The upwinnowing of groundwater via gaining segments has the potential to unclog the sediments and thus alter the bed composition.

Streambed topography often controls the distribution of water level and substrate changes that contribute to seepage forces and alteration of  $K$ . Regional stratigraphy such as sills, faults, and shear lineaments or presence of former debris dams or beaver ponds may lead to nested flow and detention/retention systems which may result in distinct groundwater storage or mixing zones. The fluvial and glacial systems produce fractal distributions of recharge, discharge, and associated subsurface flow patterns (Wörman et al. 2007). Streambed facets, substrates, and bedforms control reach-scale hyporheic pore water flows. The  $K$  tends to increase with the amplitude of bedforms within the reach (Stewardson et al. 2016). The spatial patterns of hyporheic flow are often controlled by bed topography and channel hydraulics (Harvey and Bencala 1993; Hansen 2001; Tonina and Buffington 2011). Due to topographic irregularities, such as ripples, dunes, and other typical bed forms in fine-grained systems, periodic pressure distributions are set up in the streambed, creating differences in the driving forces which lead to complex interfacial flux patterns (Zhou et al. 2014). Quantifying and reporting of these differences are a significant challenge, even at the experimental scale.

#### *Effects of stream stage and velocity*

Stream stage and velocity has implications on stream power and its ability to entrain, move, redistribute, and deposit channel particles. Water moving through the landscape from



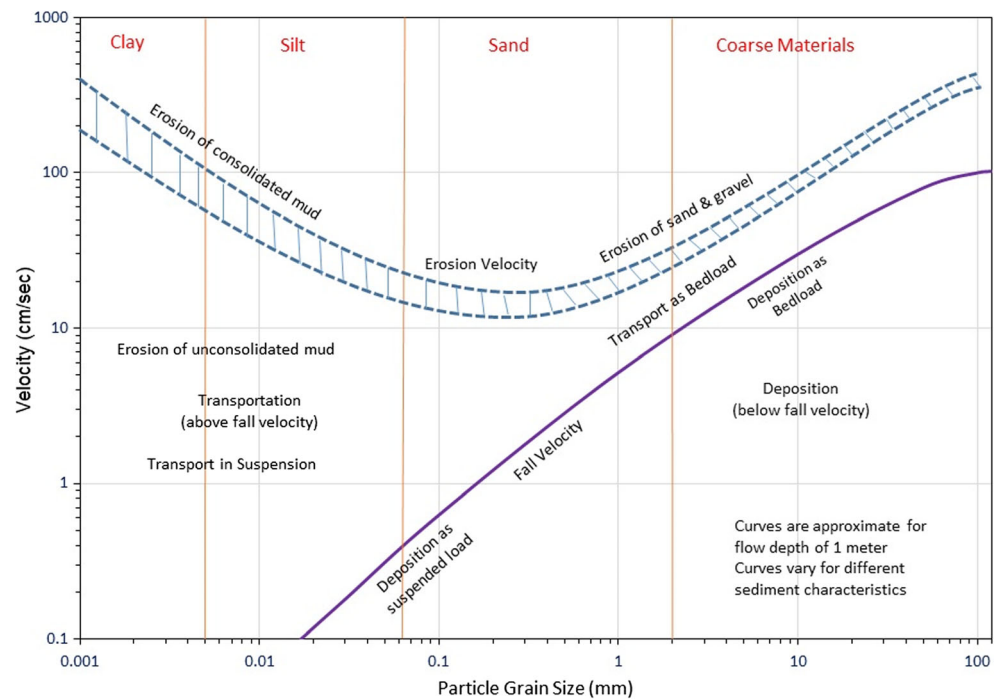
headwaters to outlet has many rates and scales, including seeps, springs, wetlands, falls, cascades, channels, streams, and rivers. Spatially, high gradient segments of stream channel such as riffles may be narrower with coarser substrates and a high velocity of flow, while lower gradient pool segments common to sinuous bends are wider and deeper with a tendency to accumulate finer materials with lower  $K$  values. Streams with bankfull channels having floodplain access tend to be relatively stable. Entrenched stream types such as gullies lack floodplain access, so velocity and stream power continue to increase above the normal bankfull stage without access to a floodplain. Stream velocity and water levels adjust not only to discharge rates but also to sharp bends, curves, falls or step-pool morphology, cascades (laminar to turbulent flow), and abrupt gradient drops into valleys (Allan and Castillo 2007). Stream features such as falls may be deceiving as their power is undeniable, which is expended abruptly in the elevation change, with velocities slowed or halted in a large pool feature at its base. Sharp stream bends act much as grade drops temporarily forcing velocities to slow with adjustments in water level. Relative to sediment transport, it is this combination of stream stage and velocity that induces helical rotation of the flow within the stream which facilitates the detachment, entrainment, and erosion in some areas accompanied by flushing, rolling, and depositing of various size particles and materials to other areas (Bisson et al. 2006). The hydraulic conductivity of streambeds is affected by a myriad of processes such as layering, sorting, and grain size, which varies by location and the extent of sediment supply. Stream velocity adds the ability to entrain, sort, transport, flush, embed, and compact materials while head controls siltation and particle sorting that may improve or limit the ability and timing of surface to groundwater connections. For hyporheic zones of coarse materials, conditions of flow with fine-suspended sediments may detain or retain mineral or organic sediments in the substrate materials (Bencala 2005). Whether these fines can be disturbed, flushed, or removed later would depend on disturbance mechanisms such as storm events that may move, roll, entrain, and/or suspend some of the large and fine particles for continuing transport downstream. Conditions of channel armoring with coarse materials, the growth of mosses, periphyton, or biolayers developed from impacted organics may add stability to limit substrate disturbance and flushing of fine particles. However, these generalities relative to stream channels describe conditions below bankfull flow. Above bankfull flows, for streams that have access to their floodplain, as stream stage rises, mean velocity and the ability to move sediment decline, resulting in deposition of materials onto the floodplain, sorting out with coarse materials settling first. Increase in water levels during storm events and conditions of flooding expands the extent and level of valley water tables thus affecting the flow to groundwater systems (Hewlett and Nutter 1970). Floods not only cause destructive consequences

but also have positive impacts like enrichment of flood plains by carrying sediments rich in nutrients and rejuvenate the river ecosystem (Jiang et al. 2008).

Natural disturbances can sometimes promote change and rejuvenation as the fine bed material is scoured away when disturbed during torrential flows or periods of groundwater influx that helps to backflush fines, resulting in an increase in the hydraulic conductivity (Hannula and Poeter 1995; Hatch et al. 2010). However, it is difficult to generalize these changes for all substrates, as some become embedded and compacted through time, may be stabilized by vegetation and do not easily break up during high flows. There is a substantial difference in the erosion capacity between clean and sediment-laden waters. Flood events can induce temporal changes in streambed elevation and particle size composition, influencing the bed's hydraulic properties and stream-aquifer fluxes during and after an event. Simpson and Meixner (2012) observed preferential entrainment of bed load, both coarse and fine sediments during the rising phase of flood hydrographs leading to overall bed coarsening and increase in vertical hydraulic conductivity. The introduction of large wood to stream systems adds complexity, diverts flow and energy, improves habitat, and alters the local substrate composition (Lassette and Harris 2001). During low flow/stage periods, streams will tend to carry only dissolved and very fine-suspended load. In the course of time, an equilibrium bed slope may be formed when aggradation and degradation of sediments are in balance, recognizing variances in slopes, and the inherent ability for particle retention or flushing remain within the riffles, runs, pools, and glides (de Llano 1993). The relationship between the size of sediment and the velocity required to erode it, transport it, and deposit it is explained by the Hjulström curve (Fig. 2). The critical erosion curve (derived based on 1-m flow depth) shows the minimum velocity required to lift a particle of a certain size. The critical deposition curve shows the maximum velocity at which a river can be flowing before a particle of a certain size is deposited. Silt and clay need higher velocities for entrainment than their size would suggest because they are cohesive (they stick together), but once entrained, they are difficult to settle and need very low velocities or static conditions (Keylock 2004; Arntzen et al. 2006).

Blasch et al. (2007) while modeling two sequential streamflow events, observed that the hydraulic conductivity of the streambed surface layer changed by about four orders of magnitude due to sediment redistribution from one event to the next. The rate of streamflow loss to the unsaturated zone beneath depends primarily on the stream stage, vertical hydraulic conductivity, and geometry of the streambed (Simonds and Sinclair 2002). In the case of stream disconnected from water table (i.e., losing stream), Wang et al. (2014) observed a linear relationship between the vertical seepage rate and streambed hydraulic conductivity. The vertical seepage rate had a linear relation with stream water depth during

**Fig. 2** The Hjulström curve for predicting sediment erosion and deposition. (Reproduced from source: Keylock, 2004)



laminar (low Reynolds number) flows and exponential relation during turbulent flows.

In gaining reaches of the stream, Chen et al. (2013) observed a decreasing trend of streambed hydraulic conductivity due to the upward winnowing of fine substrate from the deeper sediment cores; however, in losing reaches, hydraulic conductivity of deeper sediment cores increased due to silting of fine substrate in the porous top layers of the streambed. Rosenberry and Pitlick (2009) observed quite a dissimilar kind of situation where the magnitude of vertical hydraulic conductivity varied on the subject of seepage direction, the gaining stream reach holding a higher  $K$  than the losing stretch. This contrasting characteristic is site specific and may be attributed by the composition, weight of substrate materials and vibrations that compact and settle substrates, reducing or enhancing hydraulic conductivity. Of course, these circumstances would not be found everywhere and difficult to generalize, as sometimes fine materials from landslides bury channels, with coarse channel materials buried and possibly providing a conduit for high flow.

The influence of stream stage and velocity changes on hyporheic and groundwater flow and extent of connectivity suggests a similarity to hydrograph pulsing and topographic expanding and contracting area of contact as described in the variable source area concept (Hewlett and Hibbert 1967; Hewlett and Nutter 1970). Surface storm hydrographs are apt to exhibit rapid changes as the friction in open channel flow is much less than that of hyporheic and groundwater substrates. Flow permanence of gaining streams depends primarily on the gravity feed of stored water in water tables at

higher elevations flowing toward as they are intercepted by channels as low points in the landscape. When water tables are depleted to an elevation, channel depressions and water release do not provide a hydraulic gradient; underground flow has an increasingly longer path and contracting area contributing gravity flow to downstream perennial channels which continue to maintain a hydraulic gradient.

#### *Effects of stream water temperature and viscosity*

The density and viscosity of water are temperature sensitive variables. The hydraulic conductivity of a porous medium depends upon the density ( $\rho$ ) and dynamic viscosity of the fluid ( $\mu$ ) involved and also on the average size and shape of the pores in a porous medium. The relationship between water temperature, density, and viscosity are not linear, and as the water temperature approaches freezing, the density and viscosity increases (Prince 1984); however, the density decreases if the temperature falls below 4 °C as ice forms and floats on the surface. In fact, permeability is directly proportional to the unit weight of the fluid concerned and is inversely proportional to its viscosity (Hamill and Bell 2013). Changes in the atmospheric pressure and the temperature of the stream water influence the hydraulic conductivity of the porous media. Much like tidal action, changes in the position of the moon and weather patterns can influence the barometric pressure acting on both surfaces and to a lesser extent subsurface water levels, which contribute as hydrostatic pressures to move water through materials (Kemp et al. 2011; Liu and Higgins 2015). The effects of stream water temperature changes on

viscosity and hydraulic conductivity of streambeds are cited in literature extensively (Constantz et al. 1994, 2008; Su et al. 2004; Cardenas and Wilson 2007b; Cardenas and Wilson 2007a; Constantz 2008; Gerecht et al. 2011; Irvine et al. 2015). In a streambed, the vertical and horizontal distribution of temperature is a function of both the advective and conductive heat transport process (Soares et al. 2012). Constantz et al. (1994) suggest that high infiltration rates cause rapid convection of heat to the streambed. In a system where the stream recharges the aquifer, the streambed can experience daily fluctuations in temperature that are attenuated and delayed with distance and depth from the surface water body (Cardenas and Wilson 2007b). Cox et al. (2007) articulate that the parameters such as water temperature variation, scour, and clogging of the streambed influences the hydraulic conductivity of the streambed and observed a variation of 41% in  $K$  seasonally due to temperature variations alone. To evaluate the effect of water temperature on streambed vertical hydraulic conductivity, Dong et al. (2014) conducted in situ permeameter tests in Clear Creek, NE, USA, where the coarser sandy sediments had a greater increase in the extent of the vertical hydraulic conductivity value per 1 °C increase in water temperature than sediments composed of silt and clay layers.

The streambed temperature profiles have been used to quantify stream–groundwater interactions (Schmidt et al. 2006; Constantz et al. 2008; Essaid et al. 2008; Constantz 2008), delineate flow paths in the hyporheic zone (Conant 2004; Schmidt et al. 2006; Beach and Peterson 2013) and to estimate seepage and hydraulic conductivity of streambed (Constantz et al. 2004; Silver 2007). The streambed temperature will be highly variable in the case of a losing stream, but it will be relatively stable with little fluctuation in the case of a gaining stream (Lee et al. 2013). Stream thermal regimes are affected by outside influences such as air temperature, solar radiation, and heat loss as a result of evaporation. The in-stream geomorphic influences which facilitate heat transfer are processes such as hyporheic exchange of water, groundwater inputs from the adjacent water table, as well as the ability of the substrate to subdue heat exchange (Hester et al. 2009). The hydrogeomorphic processes that drive changes in sediment texture are fine-scale granulometric variables such as oxygen, interstitial space, and organic matter translocation (McKenzie-Smith et al. 2006) along with the sources of sediments, weathering, particle breakup from tumbling, chemicals, freezing, etc. Hatch et al. (2010) from their experiments noticed that a 30% decrease in fluid viscosity would yield a commensurate increase in hydraulic conductivity of the streambed. The temperature of water entering the soil changes as it moves through the unsaturated zone before entering the water table producing variations in soil temperature and water viscosity which can have a direct impact on hydraulic conductivity (Dong et al. 2014). Ronan et al. (1998) spotted increased infiltration rates beneath an ephemeral stream

and recognized the water viscosity effects on hydraulic conductivity from increased stream temperature. Lu et al. (2012) discusses the impact of temperature on the estimates of hydraulic conductivity and verified the inverse relationship of  $K$  value with the kinematic coefficient of viscosity. Beneficial to many streams and aquatic habitats, extreme temperature swings are limited and moderated by flow through hyporheic zone substrates relative to their capacity to absorb and dissipate heat over time. Groundwater influx to streams also contributes cool waters that help to moderate temperature extremes.

### **Influence of anthropogenic activities on streambed hydraulic conductivity**

#### *Effects of land use change and urbanization*

Urbanization, industrialization, and changes in the land use pattern have resulted in local changes in weather patterns that in some instances may approach irreversible disturbances to the hydrological processes. The increase in the impervious area associated with urban development affects the hydrologic cycle and consequently; there is an enhanced risk of channel instability and urban flooding (Niehoff et al. 2002). Urbanization often increases the discharging of pollutants to streams and water bodies from a variety of point and non-point sources. Physical changes as a result of soil compaction and vegetation losses within riparian zones may also alter water quality and hydrologic functions. The uncontrolled discharge of domestic, commercial, and industrial wastewater to nearby ponds or rivers has the potential to cause major ecological and sustainability problems for the river inhabitants (Yule et al. 2015). Deforestation increases the runoff two to six times over what would occur naturally, reduces the vegetation transpiration, raises the water table, and accelerates soil erosion, bank instability, and channel adjustments. Deforestation also affects the normal stream nutrient cycling, leaching, and transportation processes, some of which are desirable for ecological function (Biggs et al. 2002; Iwata et al. 2003). Due to deforestation, at landscape scales, the dense canopies, stabilizing root systems and increased filtering associated with stream buffers are destroyed, exposing streams to increased solar energy, erosion, and pollutants. Excessive snagging and removal of large wood and aquatic plants from the channel can initiate bed level lowering as well as impact habitat diversity (Erskine and Webb 2003). Thomas et al. (2004) found greater siltation and exposure in deforested streams contributing to the development of a fine-grained, organic-rich stream bed with increased presence to dominance or infilling by aquatic vegetation.

As demonstrated in the field studies of river contamination, riverbeds play a vital role in the transport, filtering, and fate of contaminants in river systems (Jang and Liu 2005; Barth et al.

2007; Gates et al. 2009; Venkatesha et al. 2012; Sehgal et al. 2012). Contaminants may enter and leave the bed of the river by a variety of mechanisms (Zhu et al. 2014). Diffuse (non-point) sources of contaminants such as activities or actions which expose soils to erosion and connect to lotic waters sometimes transport sediments along with pollutants. Point sources of pollution are directly discharged wastes or wastewater from the industry, domestic use, and community or urban releases from waste storage ponds or lagoons and by the use of agro-chemicals during irrigation and return flows (Nie et al. 2012). As long as the concentration of the contaminants is higher in the recipient water than in the pore water of the sediments, the net mass transport is directed into the sediments. Hence, the sediments act as a sink of contaminants (Forsman 2000), sometimes taking time for contaminant breakdown or continuing dilution of concentration as mixing occurs. However, contaminated sediment erosion and transport is a mechanism to consider if a breakdown does not occur. The hydraulic conductivity and porosity of the contaminated streambed sediments are the result of both erosion and depositional processes that take place in streams or rivers to produce a unique set of pore space geometries in the course of progressive adjustment and stabilization of the sediment layers through time (Messina and Biggs 2016). Weathering effects, which encourage for chemical breakdown of contaminated sediments, allow for settling and sediment embeddedness with trends to decrease particle size and the hydraulic conductivity of the streambed.

Intensive pumping of groundwater near streams has the potential to disconnect the aquifer from stream water inputs or changing the flow fields from the stream to the pumped aquifer system. The aquifer configuration and type of sediments in the stream channel controls the recharge rate of the underlying water table (Wang et al. 2011). Various types of other anthropogenic activities also influence streams, streambeds, and groundwater interactions, including hydrologic modifications associated with wetland draining, stream ditching, channel straightening, channel dredging and realignment for transport, dikes or dams of various types and intensities of flow alteration, culverts and bridges, dispersion of mine tailings into streams or lakes, and mining of in-stream materials such as sand and gravels. These anthropogenic activities can alter the chemistry, channel morphology and function, streambed substrate, particle size, composition, hydraulic conductivity, groundwater connectivity, flow permanence, and other factors that affect water, sediment, and pollutant movement (Jiang et al. 2010). Extensive best practices or professional guides have been prepared for many circumstances and anthropogenic activities to help prescribe reasonable approaches for protection of water quality, and it is likely that some of these measures such as forested stream buffers (Welsch 1991) may benefit in maintaining stream–aquifer connectivity and function.

### *Effects of engineering structures across or within a river/stream*

Erection of hydraulic structures are often designed specifically to alter the natural flow of water and, as a result, affect sediment and pollutant transport, leading to adjustments such as the localized growth or erosion of deposits and changes to the carrying capacity and functions of a river (Skalak et al. 2009). Fragmenting streams with dams, diversions, and less frequently road culverts disrupt the longitudinal connectivity and capacity of a stream. It is known that after the construction of a dam, dike, weir or a barrage, accretion starts upstream in response to the grade shift with lower stream velocity, whereas retrogression takes place downstream due to inefficient energy dissipation by the discharge of clean water from elevation. Another effect of aggradation is the increase in tortuosity, sinuosity, or even braiding of the channel upstream due to the loss in gradient and ability to transport sediment. The tendency for increasing tortuosity can be illustrated, by comparing the unaffected reference stream sinuosity and meander belts with sections affected by structures (Allan and Castillo 2007). In general, the larger and higher the dam, the greater the influence, while small or in-stream structures may produce only localized effects. Proper design and placement of small structures such as J-hooks and cross vanes can be used to reduce near bank stresses, effects of localized excessive sediment accumulations near bridges and habitat benefits (Rosgen 2001). Meandering alterations in low gradient systems may lead to adjustments or diversion of flow such as outflanking of hydraulic structures. Flow avulsion may not only be caused by diverting flow into a meandering channel that lacks capacity but also if the river shifts its course to join other low-lying rivers. These kinds of changes indeed alter the affected stream–groundwater connections.

The tail water releases of waters below the dam downstream are appreciably affected by the clean water effect, as these waters are hungry for sediment to the extent that stream degradation, bank and point bar erosion are typically significant. The channel gradient increases somewhat and this may extend over a considerable length downstream near large structures. The dam structure reduces the dominant discharge of a river and its fluctuations. The downstream channel tends to straighten, thus reducing the meander tendency (Mazumder 2004). Dams decrease the pollution effect considerably in the downstream part by lowering the pollution load coming from the source (Tahmiscioğlu et al. 2007). Downstream channels with lower than normal flows may be invaded by vegetation and loose channel capacity, leading to increased flooding during severe floods. Regular channel maintenance flows every year or two as capable are prescribed to help reduce vegetation encroachment and maintain channel capacity, even though degradation is a potential for some sections.

Measures to help maintain sediment transport through dams is sometimes employed to help reduce these effects.

The function of constructed embankments or levees along the flow of a river is to restrict access to the floodplain through a physical barrier or through channel adjustments in morphology that generally increases the stage and the velocity of the flood flow. With the increased velocity and potential for degradation, the total silt carried downstream to other low gradient areas or to the sea/ocean by an embanked river would be more than that of an unbanked river. Other conditions being the same, unless deposited before the basin outlet, the rate of extension of the delta with more sediment delivered would be greater. Greater extension of the delta with an increase to base level may cause portions (i.e., lower gradient) of the embanked river to rise and result in building up of the river bed. Hydraulic changes to rivers do not always produce the effects desired and are seldom permanent changes. In the case of stable rivers, embankments may raise the flood level but are less likely to induce a change in the river bed configuration (Garde and Raju 2010). Embankments may be supplemented by other flood control measures such as storage reservoirs, diversion of the river, and soil conservation practices. Failure of these structures sometimes causes much more damage than would have naturally occurred, and that is why flood preparation and managing of water controls previous and during eminent hazards are important elements to consider. The net effects of these structures to stream and groundwater interchange are seldom considered, analyzed, or known.

Bed load transport in mountainous streams varies with the geology, gradient, sources of material, flow intensity, stream power, and the riverbed structure. Corrugated metal culverts can be worn and damaged under high bed load transport conditions, suggesting bridges may be needed and cost-effective for sustainable passage (Hansen 1987). Dams or similar structures are likely to capture this load and may be a reason to avoid structures in high gradient systems. Zhang et al. (2010) articulates that the bed load transport rate varies in a range of 3–4 orders of magnitude in a mountain stream at the same place and under unaffected flow discharge, with the presence or absence of a step-pool system. Check dams are sometimes used to augment habitat and stabilize mountainous streams and landslide hazards. A point to remember is the higher the check dams, more hydrologically unstable they are. Located, designed, and installed inappropriately may cause more issues than leaving the stream alone. The construction of check dams can produce changes in channel cross-sectional geometry, grain size distribution, bed forms, bed slope, and riparian vegetation. Sediment and/or bed load retention behind check dams can facilitate localized channel widening or narrowing, armoring, and/or incision downstream (Wohl 2010). On watersheds with coarse soil textures, check dams may be particularly effective due to greater sediment retention and formation of a permeable bed, with a better mix of substrate

materials to add diversity (Polyakov et al. 2014). Low-rise check dams have some benefit for grade control such as in gully systems (Rosgen 1994). Spur dikes, deflectors, groins, rock vanes, cross vanes, and J-hooks are transverse structures that extend into or across the stream from the bank and reduce erosion by deflecting flows away from the bank and reducing bank stress (Rosgen 2001). Generally, two or three rock vanes (discontinuous, redirective structures) are constructed along the outer bank of a bend in order to redirect flows near the bank toward the center or thalweg of the channel. By designing structures to maintain a channel thalweg helps to reduce this potential for diversion. Some structures such as gabions can settle or move if not adequately anchored or keyed into channel, slope, or position. Continuing update and review of anthropogenic modifications and in-stream structures are needed to consider their effects to manage flows, flood damage, aquatic passage, habitat, stream morphology and stability, sediment and bed load transport, water quality, and stream–aquifer interactions.

#### *Influence of biological factors on streambed hydraulic conductivity.*

The removal of riparian (streamside) vegetation may lead to a reduction in bank stability and contribute to failure if soils are unstable. Activities that reforest or replace vegetation losses or allow for partial removal and regrowth are best to maintain stability as a general rule or best management practice. Vegetation removal may result in bank erosion or failure and stream siltation of fine sediments locally or downstream. The siltation generally reduces the number and capacity of pool sites and hydraulic conductivity due to finer materials and increases habitat homogeneity (leading away from pollution intolerant macroinvertebrates toward tolerant species). On slopes, over about 35%, special measures such as directional felling or adjusting logging systems may be needed to selectively fell, log, or thin stream sides.

Riparian ecosystems are biologically active zones of diversity and productivity and also often act as a sink for nutrients or pollutants from runoff waters (Sabater et al. 2000). Trees and shrubs physically constrain and retain the soil near the stream banks, thereby protecting the streambed and limiting the scouring effect of running water. In regions with high organic matter accumulation (e.g., leaf litter buried in alluvial deposits), an actively growing biofilm might clog sediment or soil pore spaces and thereby reduce the hydraulic conductivity of the streambed (Chestnut and McDowell 2000). Biofilms predominantly develop on small sediment particles because of their large surface areas. Biofilms have a low porosity and therefore result in localized areas of low *K*. Riparian vegetation helps to provide shade and cooler air temperatures that contribute to the cooler and consistent stream water temperatures. In cold weather circumstances, heavily vegetated stream

buffers may locally help to capture snow or retain warmth to buffer the heat exchange. The nutrient buffer capacity of riparian zones depends on the complex combination of landscape and hydrogeomorphologic variables (Gu et al. 2007; Garrett et al. 2012). The presence of roots in riparian and channel substrates can contribute to subsurface flow conduits or underground blockages. In addition, the presence and burial of wood debris within sediment deposition events can also add to substrate complexity and function (Lassetre and Harris 2001; Erskine 2002). Wood that remains saturated has limited ability to decompose, while wood frequently exposed to air and associated macro- and microorganism will decompose faster, increasing opportunity for voids, settling, and localized water passageways (Lassetre and Harris 2001). Song et al. (2010) provide evidence of bioturbation activities which can result in larger hydraulic conductivity streambed.

Tremendous complexity exists in fluvial systems and factors such as streamflow patterns, channel form, transport processes, and the distribution of riparian vegetation or woody debris that affect specific stream channel responses resulting in an irregular distribution of sediment supply (Abt et al. 1994). Organic sediment is derived from decaying accumulations of leaves, woody debris, algae, florae, and/or other organic material that develops falls or is delivered into streams (such as foliage). The microbial and macroinvertebrates attached to the debris or other inert matter contribute to the breakdown and processing of organic matter. Organic sediment transport through a fluvial system varies by location and season and is more difficult to quantify when compared to that of mineral sediment movement (Wynn and Mostaghimi 2006). Organic and timber debris are sometimes more notable in bedrock and boulder-bed streams; it crafts a part of the stream geomorphology and permits a stabilizing effect. Streambeds buried or dominated by very fine sediments and nutrient accumulation, exposed to the sun, and with little flow disturbance have increased risk of clogging due to algal blooms. In contrast, sediment-starved streams may be incapable of accumulating organic materials or providing diverse habitats for benthic organisms (Quinn et al. 2009). Aquatic plants or periphyton may shield the shoreline, or streambed substrates, consume nutrients and contaminants from the water, and may help stabilize the streambed sediments. They may also limit normal hyporheic water, particle mobilization, or transport and add roughness to reduce flow velocity which would promote finer particle settling.

The activities and response of various organisms that live in streambed hyporheic zone have the potential to breakdown clogged substrates and bioremediate contamination for the improvement in the groundwater quality relative to its future uses without added treatment (Whelan 2007). Sediment reworking by macroinvertebrates as well as others such as fish spawning may increase the  $K$  of river bed locally. Chemical processes within the hyporheic zone may result in mineral

dissolution or precipitation, which will increase and decrease water density and  $K$  seasonally. Hyporheic zones usually support a variety of substrate habitats that may accommodate, depending on water quality, a variety of invertebrate species including ostracods, copepods, amphipods, tardigrades, nematodes, oligochaete worms, rotifers, and early instars of aquatic insects. These fauna and flora within the hyporheic zone are significant contributors to organic and nutrient cycling in streambeds (Clarkin 2008).

### Laboratory and in situ methods of estimating streambed hydraulic conductivity

Streambed hydraulic conductivity ( $K$ ) measurements have been performed using a variety of techniques. In-stream methods of determining  $K$  include slug tests (Duwelius 1996; Cey et al. 1998; Springer et al. 1999; Ryan and Boufadel 2007), in situ permeameter tests (Lindgren and Landon 2000; Rosenberry 2000; Chen 2004, 2005; Chen et al. 2008; Genereux et al. 2008; Huang et al. 2014; Jiang et al. 2015), and seepage flux measurements using seepage meters coupled with measurement of hydraulic gradient through the streambed (Landon et al. 2001; Murdoch and Kelly 2003; Rosenberry 2008; Rosenberry et al. 2012). In addition, streambed samples can be collected for grain size analysis and  $K$  can be estimated from grain size distribution (Boadu 2000; Song et al. 2009). Although field permeameters and seepage meters coupled with hydraulic gradient measurements determine vertical hydraulic conductivity ( $K_v$ ), the slug tests measure horizontal hydraulic conductivity ( $K_h$ ), and the hydraulic conductivity ( $K$ ) values from grain size methods are non-directional (isotropic). Reynolds et al. (2000) employed positive-head tension infiltrometer (TI) and single-ring pressure infiltrometer (PI) and classical undisturbed soil core (SC) methods for measuring  $K$  of single-grain sand, structured loam, and cracking-clay loam soils. Cardenas and Zlotnik (2003) experimented with multilevel constant-head injection tests and multilevel slug tests to gather streambed hydraulic conductivity values. Piezo-Seep meter provided improved estimates of vertical saturated hydraulic conductivity, specifically in streambeds of medium to coarse textured sediments in the study conducted by Kelly and Murdoch (2003). Accurate evaluation of soil hydraulic conductivity, soil sorptivity, and matrix flux potential can be made in all types of soils using Guelph permeameter (Reynolds and Elrick 1985; Fares et al. 2000; Kodešová et al. 2010; MacDonald et al. 2012). For highly permeable gravel beds, Yamada et al. (2005) developed a packer test system based on Hvorslev's equation (Hvorslev 1951) for the in situ estimation of streambed hydraulic conductivity. Laboratory determined values may be more precisely measured, but rarely agree with field measurements. Field methods are generally more reliable than laboratory methods

due to the closer approximation to natural conditions and lower disturbance while sampling. (Scott 2000).

In addition, field sampling of streambed and valley bottom materials may exhibit variability of particle sizes resulting from various types of erosion and deposition processes of materials associated with vertical and lateral accretion, colluvial, splay, alluvial fan, and deltas (Happ et al. 1940; Trimble 2008). Piping in certain types of fine-grained soil and weathered substrate materials can lead to rapid movement of water and sediments underground. An understanding of the geology, climate, and hydrology forces of recent and past, and ample sampling of substrates should help to reveal and quantify the potential and/or presence of the variances caused by these variables. Decisions on how to conduct sampling for research and interpretations using in situ or laboratory readings may be an iterative process that applies to the specific circumstances and intent.

Other methods such as those described by Garrett et al. (2012) on end member analysis of water chemistry and others who use radionuclides or fluorescent dyes in tracing water flow rates or transfer are examples of alternative methods (Abbott et al. 2016). Using fluorescent dyes, visual responses are possible, but fluorimeters can detect and quantify low concentrations in the parts per billion (ppb) range, well below visualization concentrations (Baker and Lamont-Black 2001). A constant injection rate study of a known fluorescent dye concentration and monitoring concentrations downstream or in the hyporheic zone with fluorimeter at various levels can help define levels of dilution and rates of flow or exchange in layers (Holland et al. 2004). Applying hydrogeology and well-driller techniques of timed pumping or injecting water into wells with monitoring water levels in well or adjacent wells may also prove helpful in defining some of the local hydraulic conductivity processes. A sampling of coarse channel substrates with various methods of freezing materials (or freeze coring) has also been used to help remove relatively undisturbed samples for particle size analysis (Everest et al. 1980; Ulrich et al. 2015). Coating non-cored, relatively undisturbed frozen samples with paraffin may facilitate follow-up with laboratory testing for hydraulic conductivity.

### Statistical and geostatistical techniques employed for the assessment of spatial and temporal variations of streambed hydraulic conductivity

Previous studies have found a significant spatial and temporal variability in the valley and streambed sediments (Hannula and Poeter 1995; Reynolds et al. 2000; Landon et al. 2001; Hatch et al. 2010). Statistical measures are frequently used to design a sampling network and describe field data in terms of measures of central tendency and variability. Table 4 provides a detailed information regarding the statistical techniques

employed in the literature of streambed hydraulic conductivity assessment. Presently, there are a variety of procedures for generating interpolated hydraulic conductivity ( $K$ ) fields from a sparse set of  $K$  measurements. Geostatistical spatial characterization of a specified variable involves the generation of maps and predicting values of that particular variable at numerous unsampled locations. These methods are best used with care and consideration of geology, hydrology, soil, and topographic properties associated with the conditions as the distribution of particles, substrates, channel morphology, etc., are not necessarily random populations, but often affected by recognizable factors such as valley type, channel gradient, sinuosity, entrenchment, and stream type (Clément and Piégay 2005). Variogram models such as spherical, exponential, and Gaussian models are collectively used to quantify the spatial variability of random variables between two sites and describe monotonic features of spatial correlation as the lag distance increases. Cardenas and Zlotnik (2003) generated a 3D hydraulic conductivity field through the application of 3D kriging module of Mining Visualization System where a spherical variogram model is fitted to the field data. Jang and Liu (2004) used ordinary kriging and sequential Gaussian simulations (mean and individual) to estimate the spatial variability and distribution of the hydraulic conductivity. Chen (2005) observed a positive correlation between vertical hydraulic conductivity ( $K_v$ ) and the water depth from the variogram fitted using the least squares method. Semi-variograms and contour maps were developed by Genereux et al. (2008) to aid in visualization of the spatial variability in hydraulic conductivity. Cheng et al. (2011) fit an exponential model to the experimental semi-variogram along the flow direction at each test site to identify the horizontal and vertical correlation between independent samples.

Being able to incorporate more detailed surface elevation information such as obtained from LiDAR may help differentiate gradient, areas of floodplain flow concentration and movement, areas of floodplain depression or storage, areas of colluvial input and others with landslides, or severe erosion delivering and depositing sediments onto the valley or floodplain (James et al. 2007; Maceyka and Hansen 2016). The extent of remote sensing and geographic and hydrologic tools continues to expand and offer the potential for new approaches in design and analysis. Variogram models such as spherical, exponential, and Gaussian models are collectively used to quantify the spatial variability of random variables between two sites and describe monotonic features of spatial correlation as the lag distance increases.

At this stage, most of the experimental design and sampling is aimed at the understanding of the variability, processes, and functions that connect streams, hyporheic zones, and groundwater. Putting these components together to help identify and model stream network and landscape processes is still at an

**Table 4** Statistical and geostatistical techniques employed for assessment of variations of streambed hydraulic conductivity

Reference and test site	Statistical measures	Geostatistical measures	Applications	Implications
(Chen 2004) Republican, Platte and Little Blue rivers	<ul style="list-style-type: none"> <li>Measures of central tendency and variability</li> <li>Kruskal–Wallis test</li> </ul>	–	The <i>Kruskal–Wallis</i> test to determine the significance of the difference of $K_v$ values between two test sites of the same river and between the test sites of two rivers.	The <i>Kruskal–Wallis</i> test indicated a significant difference of the $K_v$ values between the Republican and Little Blue rivers and between the Platte and Little Blue rivers but found no such significant statistical difference between the Platte and Republican rivers. The Little Blue River is a smaller size hydrologic unit.
(Chen 2005) Platte River, USA	<ul style="list-style-type: none"> <li>Shapiro–Wilk test</li> <li>Kruskal–Wallis test</li> </ul>	<ul style="list-style-type: none"> <li>Variogram fitted using the least squares methods</li> </ul>	<p><i>Shapiro–Wilk</i> test to check whether a sample is from a normally distributed population.</p> <p><i>Kruskal–Wallis</i> test to determine the similarities of the <math>K_v</math> values from the test sites.</p>	$K_v$ values of the Platte River in south-central Nebraska showed a normal distribution. A positive correlation was observed between $K_v$ and the water depth which produced periodicity of $K_v$ values across the channel.
(Ryan and Boufadel 2007) Indian Creek, Philadelphia	<ul style="list-style-type: none"> <li>Kolmogorov–Smirnov test</li> </ul>	<ul style="list-style-type: none"> <li>Variogram analyses</li> </ul>	The $K_v$ measurements at each depth were checked for Gaussian distribution. The combined data set deviated from the Gaussian distribution. The normal distribution of $K_v$ (95% confidence level) at each depth was tested by Kolmogorov–Smirnov test.	Variogram analyses across the stream suggested symmetry—in the upper layers of thalweg) and fractality—in the lower layers). The variogram of $\ln K$ of the upper layer sediments suggests that fine sediments get trapped along the stream edge with low velocity and pressure head.
(Geneux et al. 2008) West Bear Creek, North Carolina	<ul style="list-style-type: none"> <li>Measures of central tendency and variability</li> </ul>	<ul style="list-style-type: none"> <li>Exponential-semivariogram models</li> <li>Contour maps</li> </ul>	Contour maps brought out anisotropic aspects of the $K$ distribution without an anisotropic interpolation routine. The streambed $K$ upstream and downstream of a small beaver dam were analyzed.	Slightly bimodal distribution of the $K$ was observed via histogram plotting. Higher $K$ in the center of the channel indicated gateway for water and chemical fluxes through the streambed. The average streambed $K$ was about 23% lower on the upstream of the dam.
(Leek et al. 2009) Touchet River, USA	<ul style="list-style-type: none"> <li>Kolmogorov–Smirnov (K-S) test</li> <li>Wilcoxon rank-sum test</li> <li>Median two-sample test</li> </ul>	–	$K$ -S, Wilcoxon rank-sum, and median two-sample tests determine the differences in empirical cumulative distribution function (CDF), mean, and median of $K$ between the two study sites and two depth intervals at each site.	All the three tests approved the similarities in empirical CDF, mean, or median among certain depth intervals. On the other hand, the spatial patterns of $K$ in channel sediments varied from one depth interval to another. Gaining and losing zones based on vertical hydraulic gradient varied among different depth intervals, suggesting the complexities of the water flow regime.
(Hatch et al. 2010) Fourth-order stream of the Pajaro River, USA	<ul style="list-style-type: none"> <li>Cross plots</li> <li>Ranges of <math>K_v</math></li> </ul>	–	Relationships between streambed hydraulic conductivity, discharge, and seepage rate were demonstrated through “cross plots.”	Local-scale variations in streambed seepage due to streambed topography were observed. The cross plot suggested a hysteresis relationship between streambed conductivity versus seepage rate. During higher discharge periods, increased $K$ and seepage rates were observed due to bed scouring, however, as the discharge reduces, the sediment deposition progresses decreasing the seepage and $K$ of bed.
(Chen 2011) Platte River, USA	<ul style="list-style-type: none"> <li>Kruskal–Wallis test</li> <li>Mann–Kendal test</li> </ul>	–	<i>Kruskal–Wallis</i> test to determine whether the $K_v$ values from the three depth ranges have the same mean. <i>Mann–Kendal</i> test to detect the trend of $K_v$ for channel sediments from the shallow to the deep part.	The results rejected the <i>Kruskal–Wallis</i> test hypothesis at 80% level of confidence indicating that two populations (bed depths) have different means. The <i>Mann–Kendal</i> test rejected the hypothesis at 95% level of confidence indicating a decreasing $K_v$ trend with depth. The hyporheic processes can affect streambed $K_v$ distribution patterns for about 9 m below the channel surface. The sites where the normal distribution of streambed $K_v$ was observed suggested sediments were well distributed and belong
(Cheng et al. 2011) Platte River, USA	<ul style="list-style-type: none"> <li>Jarque–Bera (J-B) test</li> </ul>	<ul style="list-style-type: none"> <li>Jarque–Bera (J-B) test</li> </ul>	( <i>J-B</i> ), ( <i>K-S</i> ), ( <i>S-W</i> ), and <i>Lilliefors</i> tests determine whether streambed $K_v$ at each test site is normally distributed.	



**Table 4** (continued)

Reference and test site	Statistical measures	Geostatistical measures	Applications	Implications
	<ul style="list-style-type: none"> <li>• Kolmogorov–Smirnov (K-S) test</li> <li>• Lilliefors test</li> <li>• Shapiro–Wilk (S-W) test</li> <li>• <i>t</i>-test</li> </ul>	<ul style="list-style-type: none"> <li>• Exponential-semi-variogram model</li> </ul>	<p>S-W test requires sample size <math>N (7 \leq N \leq 2000)</math>. The K-S and Lilliefors require sample size <math>N (N \geq 2000)</math>.</p> <p>A <i>t</i> test with unequal variance determines whether streambed <math>K_v</math> differs significantly between different test sites.</p>	<p>to a single population of <math>K_v</math>. For sites where normality tests failed, the sediments deposited belong to different hydrogeological processes, including geological conditions, geomorphic history, and physical transport processes.</p> <p>A part of the population of samples was reduced to independent samples by drawing out the spatially correlated samples by using an exponential-semi-variogram model.</p>
(Dong et al. 2012) Two point bars locations in Clear Creek, NE, USA	<ul style="list-style-type: none"> <li>• Kruskal–Wallis test</li> </ul>	–	<p>The Kruskal–Wallis test determines whether the <math>K_v</math> values from the point bars and from the streambed statistically belong to different populations.</p>	<p>The streambed and point bar <math>K_v</math> values were heterogeneous. The smaller <math>K_v</math> values in the point bar locations result due to a weaker hydrodynamic condition that deposits fine materials at the point bar edge.</p>
(Lu et al. 2012) Platte River, USA	<ul style="list-style-type: none"> <li>• Kolmogorov–Smirnov (K-S) test</li> <li>• Sign test</li> </ul>	–	<p>K-S test to determine for normal distribution, lognormal distribution, uniform distribution, and exponential distribution, respectively.</p> <p>Sign test–non-parametric test to determine if one group tends to produce different values from another group.</p>	<p>The <math>K</math> value and kinematic coefficient of viscosity of water had an inverse relationship.</p> <p>Since, <math>K</math> is a complex function of sediment packing, structure, heterogeneity, and several other factors, the <math>K</math> values determined from grain size analysis and empirical formulas do not represent any specific direction (anisotropic).</p>
(Min et al. 2013) Donghe River, China	<ul style="list-style-type: none"> <li>• Shapiro–Wilk Test</li> <li>• Spearman bivariate correlation analysis</li> </ul>	–	<p>Shapiro–Wilk normality test determines whether <math>K</math> or <math>\ln K</math> is normally or lognormally distributed.</p> <p>Spearman bivariate correlation analysis determines whether two non-normal distributed variables are significantly correlated at the 95% confidence level.</p>	<p><math>K</math> values determined from standpipe permeameter were normally distributed, at the 95% confidence level. However, the <math>K</math> determined using falling head permeameter was lognormally distributed although not normally distributed.</p> <p>There was neither a statistical correlation nor difference between the <math>K</math> values of the upper and lower layers.</p>
(Chen et al. 2013) Platte River and its tributaries, USA	<ul style="list-style-type: none"> <li>• Measures of central tendency and variability</li> <li>• Kruskal–Wallis test</li> </ul>	–	<p>Kruskal–Wallis test to determine if there are statistically significant differences between two or more groups of <math>K_v</math>.</p>	<p>A decreasing trend of <math>K_v</math> was realized in the gaining streams. Higher <math>K</math> values occurred in the parts of the channels with greater water depth Opposite vertical <math>K_v</math> distribution patterns was perceived in gaining and losing streams.</p> <p>The Kruskal–Wallis test indicated that <math>K_v</math> of the gaining streams differs from <math>K_v</math> of the losing streams. In the losing streams, <math>K_v</math> increased downward from the channel surface to depth of about 5 to 10 m. In the gaining streams, <math>K_v</math> values showed a decreasing trend due to the upward flux of suspended sediments, which enhanced the pore spacing and elevated the <math>K_v</math> values in the top layers of the streambed.</p>
(Dairy et al. 2015) 101 river reaches of French river monitoring network	<ul style="list-style-type: none"> <li>• Mann–Whitney–Wilcoxon test</li> <li>• Wilcoxon signed-rank test</li> </ul>	–	<p>Mann–Whitney–Wilcoxon test to compare the mean of <math>K</math> values between a priori clogged and reference reaches.</p> <p>Wilcoxon signed-rank tests for assessing seasonal differences in reach-scale <math>K</math> and to assess potential observer’s bias.</p>	<p>The distribution of streambed <math>K</math> was not normal across 101 stream reaches, and a large proportion of the values were null. The variability in <math>K</math> within-reach was high and influenced by penetrating depth and hyporheic exchanges.</p> <p>No significant influence of season on reach-averaged <math>K</math> values and bias in measures by two different observers</p>
(Wu et al. 2015) Dawen River, China	<ul style="list-style-type: none"> <li>• Kruskal–Wallis (K-W) test</li> <li>• Kolmogorov–Smirnov (K-S) test</li> </ul>	–	<p>K-W test determines whether two groups of the <math>K_v</math> values belong to the same population.</p>	<p>The sediment <math>K_v</math> values of the upper four layers before the flood season were distributed normally, while the <math>K_v</math> values of the fifth layer were lognormally distributed.</p> <p>The <math>K_v</math> values before and after the flood season statistically differed.</p>

**Table 4** (continued)

Reference and test site	Statistical measures	Geostatistical measures	Applications	Implications
(Wang et al. 2016) Weihe River, the largest tributary of the Yellow River, China	<ul style="list-style-type: none"> <li>• Kruskal–Wallis (K-W) test</li> <li>• Shapiro–Wilk test</li> <li>• Kolmogorov–Smirnov (K-S) test</li> <li>• Cox–Stuart test</li> <li>• Spearman correlation test</li> </ul>	–	<p>K-S test determines whether a set of measurements come from the hypothesized distribution such as the uniform, the normal or the lognormal distribution.</p> <p>The K-W test determines if streambed <math>K_v</math> values differ significantly between two test sites, between two sampling times or between two different layers.</p> <p>The Shapiro–Wilk test and K-S test were applied to verify the normal and log normal distributions of combined and individual datasets.</p> <p>Cox–Stuart test to detect the trend of median <math>K_v</math> values.</p> <p>Spearman correlation test to measure the strength of the relationship between <math>K_v</math> values and water depth.</p>	<p>The difference in <math>K_v</math> and its distribution patterns before and after the flood season were attributed to the infiltration of fine-grained particles.</p> <p>The median <math>K_v</math> values showed no evidence of significant trends with time.</p> <p>K-W tests with the Bonferroni correction suggested no significant differences in the <math>K_v</math> values at each site and weak evidence of differences of <math>K_v</math> between the upper and lower layers.</p> <p>Larger particles of streambed sediments were responsible for higher <math>K_v</math> values while the lower <math>K_v</math> values were observed at locations of fine-grained sediment.</p> <p>The clogging with fine sediments and embedding of substrates with fines along channel margins were observed rather than in the thalweg of the river, perhaps resulting in heterogeneous streambed hydraulic conductivity.</p>

early stage of consideration. Although publications have suggested that from headwaters to the final outlet of water in hydrologic basins may be processed several times through hyporheic and groundwater flow systems, documenting how this occurs spatially through time is a challenge for continuing research (U.S.EPA 2015). Expanding the use of installing, instrumenting, testing, and monitoring of water wells to sample these systems along the near upland-riparian-stream-river system may be especially helpful in identifying and quantifying the ongoing changes to be expected with natural environmental conditions and differentiate with systems modified by significant anthropogenic activity (Zhaohua Dai et al. 2010).

### Future scope of work

River ecology and the environment are extremely broad areas of research. The study of these or the practice of these requires an interdisciplinary approach. The hydraulic and sedimentological principles alone may not be enough to build sophisticated models or theories of river mechanics and groundwater connectivity. Variability of streambed hydraulic conductivity due to velocity fluctuations has been assessed mainly in uniform streams. Experimental investigations on the variability of streambed hydraulic properties under conditions of non-uniform flow, particularly in the downstream of protective structures in the tail water section, are very scant. For areas without stream and water quality protection guidance, consideration of stream and groundwater connections to riparian areas is important to recognize not only for habitat reasons but also for societal reasons intended for sustainable water quality and supplies that have direct to an indirect role in developmental activities. Whether landowners and managers can accept increased planning, conservation, and adjustment to water management practices may depend on not only increased awareness but also how reasonable or practical the proposals are.

The variability of geology, climate, hydrology, biology, and anthropogenic conditions affecting streambed hydraulic conductivity and connectivity to groundwater systems suggests a standard list of practices or cookbook approaches that may not be appropriate. In many developing countries, the issues such as land clearing, in-stream mining, development, urbanization, industrialization, and wastewater discharge into nearby streams without limits or after treatment adjustment invariably affect stream morphology, substrates, water quality, in-stream habitats, and the stream–aquifer functions and processes. Enough information about all the abovementioned activities has to be gathered to check whether the implemented Best Management Practices (BMPs) are sufficient to prevent, limit, or minimize impacts to stream–aquifer processes. The use of a combination of statistical, geostatistical, and remote sensing tools along with quantitative methods provides valuable

assistance for developing deterministic models by involving a number of influential variables, enabling presentation and analysis of spatial data, past and existing stream types, land use, and populations. Further, they may help to analyze for time change interactions to suggest effects of various practices and forecast trends, providing estimates of error and degree of reliability.

The phenomena of longitudinal and transverse silt transport invariably affect the streambed hydraulic properties and interaction between surface and groundwaters. Streambed alteration due to transverse circulation arising in channel bends in places where the channel splits up into branches calls for a more detailed study. Multi-scale and researcher's collaborative studies of stream-aquifer exchanges need to be carried out under varying conditions and consider the development of offering some guidance, suggestions, or recommendations on sampling such as those based on the concept of nested stream-aquifer interfaces and MIM (measurements-interpolation-modeling) methodological framework. Standardization of methods which can integrate varying levels of abiotic, biotic, and societal topics may help research studies not only increase awareness but also help to hoard a body of information with increased ability to access, share, compare, and stratify data to alternative categories such as geology, climate, topography, stream type, or substrates.

## Conclusions

Detailed characterization of streambed may be beneficial and significant in improving our understanding of streambed hydraulic conductivity and its role in the mechanism of pollutant filtering via hyporheic zone and groundwater recharge or discharge. This conceptual review provides an overview of the factors such as streambed composition, morphology, topography and other hydro-geo-ecological factors contributing to the variability of streambed hydraulic properties. Examples of research were provided that were intended to help inform others by compiling information on the hydraulic, chemical, and meteorologic gradients that help define and determine the fluxes of water, energy, and solutes between streams and aquifers. Streambeds are strongly heterogeneous in space and also in time. Flood events may change the structure of streambeds and cause erosion (degradation, aggradation, sediment transport, and/or material washing) resulting in high variability in streambed  $K$ . The stream sinuosity and facets (riffles, runs, pools, and glides) also process water, erosion, and sediment differently that add spatial complexity in association to the topographic position and scale, from headwaters, to stream or river valley, to outlet. It is very challenging to identify, stratify, and characterize the erosion/sedimentation events deterministically. The streambed hydraulic conductivity study should probably favor rivers or streams in equilibrium as

controls, not actively aggrading or degrading, before attempting to compare with modifications or activities that bring on unstable conditions where change may be ongoing. A persistent stream bed such as bedrock with stability controls is a stable system, but due to its energy and scouring, may not offer as much to this subject as one with a continuous supply of sediment materials. The hydraulic conductivity of the dynamic stream bed is difficult to predict and research. However, the in situ measurement of  $K$  values and other physical properties (such as streambed porosity, viscosity, water temperature, and particle size analysis) at regular interval for longer time periods can help in developing seasonal time series models which help to interpret the spatio-temporal variations of streambed  $K$ . Collaborative efforts may also be needed to help assimilate the modeling attempts to a wider variety of conditions.

The article mentions the effects of anthropogenic activities including engineering structures on river/stream functions and influences on the variability of streambed composition and hydraulic properties. Channel morphology changes due to aggradation, degradation, and sediment storage. With clean water releases from dams, downstream erosion, substrate flushing, and sorting are the processes contributing to  $K$  variability with some potential to change with time. Researchers are expected to continue to consider, add, and expand their ability to characterize the spatial and temporal variability of surface and groundwater interchange as a means to increase understanding of water interchange, channel filtering, etc. This may necessitate greater instrumentation costs, longer study durations, and expanded sampling networks within the stream corridor profile and quantify groundwater connections.

Land use changes, including hydrologic modifications, can alter erosion, sediment, water yield, stream type, stability, function, and habitat. Urban streams expand impermeable surfaces and storm water runoff sufficiently to modify channel morphology and the streambed. Land use changes to channel systems can be avoided or minimized with careful attention to BMPs or mitigation measures properly implemented. Hydrologic modifications such as dams, diversions, canals (dredging channel), and ditches are apt to produce substantial changes that may occur rapidly or with time. The body of knowledge is varied, but limited, as the study is both difficult and complex. Further studies and increased collaboration of interested researchers and professionals and combining of past and ongoing research effort would help target unknown areas. The coupling of various works into more situational or comprehensive models can inform relative to the importance, variety, quality, and quantities of flow involved in the stream-aquifer interaction. Improvements through time would be expected in modeling, sharing, and offering data for informed decisions on land use and activity changes to help identify any critical thresholds of environmental damage and ensure sustainable resource management.

## Appendix

### Definitions

**Base flow:** That part of the stream discharge that is not attributable to direct runoff from precipitation or melting snow; it is usually sustained by groundwater discharge (Isensee et al. 1989).

**Colmation:** It refers to the retention processes that can lead to the clogging of the top layer of channel sediments and decolmation refers to the resuspension of deposited fine particles (Brunke 1999).

**Hydraulic conductivity:** The volume of water at the existing kinematic viscosity that will move in a unit time under unit hydraulic gradient through a unit area measured at right angles to the direction of flow (in field aquifer tests) (ASTM D653-14 2014). The rate of discharge of water under laminar flow conditions through a unit cross-sectional area of a porous medium under a unit hydraulic gradient and standard temperature conditions [20 °C] (in laboratory testing).

**Hydraulic gradient:** The change in static head per unit distance in a given direction.

**Hydrograph:** A graph relating stage, flow, velocity, or other characteristics of water with respect to time.

**Hydrologic unit:** In hydrogeology, geologic strata can be distinguished on the basis of capacity to yield and transmit fluids. Aquifers and confining units are types of hydrologic units. Boundaries of a hydrologic unit may not necessarily correspond either laterally or vertically to lithostratigraphic formations.

**Infiltration:** The downward entry of water into the soil or rock.

**Permeability:** A measure of the relative ease with which a porous medium can transmit a fluid under a potential gradient and is a property of the medium alone.

**Piping:** The progressive removal of soil particles from a mass by percolating water, leading to the development of channels.

**Porosity:** The ratio, usually expressed as a percentage, of the total volume of voids of a given porous medium to the total volume of the porous medium.

**Riverbank filtration (RBF):** It is a low-cost water treatment/pre-treatment technology that is used in many countries around the world for water supply (Ray et al. 2008).

**Radionuclide:** A radioisotope (an unstable isotope of an element that decays or disintegrates spontaneously, emitting radiation).

**Seepage:** The infiltration or percolation of water through rock or soil to or from the surface. The term seepage is usually restricted to the very slow movement of groundwater. Deep seepage is the water lost to groundwater (or more appropriately drainage below the depth of interest). The drainage of soil water downward by gravity below the maximum effective depth of the root zone toward storage in subsurface strata.

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