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Variation in reach-scale hydraulic conductivity of streambeds

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ABSTRACT

Streambed hydraulic conductivity is an important control on flow within the hyporheic zone, affecting hydrological, ecological, and biogeochemical processes essential to river ecosystem function. Despite many published field measurements, few empirical studies examine the drivers of spatial and temporal variations in streambed hydraulic conductivity. Reach-averaged hydraulic conductivity estimated for 119 surveys in 83 stream reaches across continental France, even of coarse bed streams, are shown to be characteristic of sand and finer sediments. This supports a model where processes leading to the accumulation of finer sediments within streambeds largely control hydraulic conductivity rather than the size of the coarse bed sediment fraction. After describing a conceptual model of relevant processes, we fit an empirical model relating hydraulic conductivity to candidate geomorphic and hydraulic drivers. The fitted model explains 72% of the deviance in hydraulic conductivity (and 30% using an external cross-validation). Reach hydraulic conductivity increases with the amplitude of bedforms within the reach, the bankfull channel width-depth ratio, stream power and upstream catchment erodibility but reduces with time since the last streambed disturbance. The correlation between hydraulic conductivity and time since a streambed mobilisation event is likely a consequence of clogging processes. Streams with a predominantly suspended load and less frequent streambed disturbances are expected to have a lower streambed hydraulic conductivity and reduced hyporheic fluxes. This study suggests a close link between streambed sediment transport dynamics and connectivity between surface water and the hyporheic zone.

1. Introduction

Hyporheic zones (HZs) are the saturated sediments beneath and adjacent to river channels through which surface water exchanges and mixes with groundwater (White, 1993; Boulton et al., 2010). The HZ is a unique ecotone that supports a variety of hydrological, ecological and biogeochemical processes essential to river ecosystem function (Gibert et al., 1990; Boulton et al., 2010). By regulating the transfer of heat and mass across the sediment–water interface, the HZs play a critical role in temperature buffering (Arrigoni et al., 2008) and biogeochemical cycling (Mulholland and Webster, 2010). They are also permanent habitats for many microbes and invertebrates (Brunke and Gonser, 1999), provide refugia for surface invertebrates or fish (Dole-Olivier, 2011; Kawanishi et al., 2013), and are used by some fish for spawning (Geist et al., 2002). The occurrence and magnitude of processes occurring in HZs largely depend upon the hydrological flux between surface and ground waters (Findlay, 1995; Fischer et al., 2005).

Most laboratory-, field-, and model-based research of hyporheic zone processes has been at the scale of a short river reach (up to several meander wavelengths) or smaller, but efforts to scale up this research to an entire river catchment are very rare (Kiel and Cardenas, 2014). Such efforts will require an understanding of catchment-scale variations in the hyporheic flow regimes including hyporheic flux, residence time, and geometry of flow paths. These are largely determined by variations in pressure at the sediment–water interface and hyporheic zone/groundwater boundary, by bed mobility, and by the variable hydraulic conductivity of porous boundary material (Blaschke et al., 2003). In turn, all these factors vary with river hydrology, channel morphology, and associated fluvial processes (Malard et al., 2002; Tonina and Buffington, 2009).

Although measurements of streambed conductivity have been reported from a broad range of stream types, few empirical studies link spatial (between sites) and temporal (with time) variations in streambed hydraulic conductivity to flow, catchment characteristics, and other geomorphic drivers. Point measurements of streambed hydraulic conductivity found in the literature vary between 10^{-10} and 10^{-2} m/s (Calver, 2001), and reach-average values are between 10^{-5} and 10^{-3} m/s (Genereux et al., 2008; Song et al., 2009; Chen, 2010; Cheng et al., 2010; Min et al., 2012; Taylor et al., 2013). This upper

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limit on reach-average values is an order of magnitude lower than might be expected for a uniform gravel [e.g., the Hazen formula (Hazen, 1892) estimates hydraulic conductivity of 0.04 m/s for particle size diameters of 2 mm]. This is because streambed sediments generally have a broad distribution of particle sizes and because hydraulic conductivity is largely determined by the smaller size fractions (Alyamani and Sen, 1993; Song et al., 2009; Descloux et al., 2010). Consequently, variation in hydraulic conductivity between reaches is likely the result of processes controlling presence of fine sediments in the streambed rather than the coarse fraction. Further, point-scale measurements vary considerably within a reach. In some rivers, sections of streambed may be effectively impermeable but the streambed is rarely impermeable throughout the river channel. The lowest reported value of 10^{-10} m/s, is five orders of magnitude smaller than the lowest reported reach-average value.

In this study we model spatial and temporal variations in hydraulic conductivity to support advances in our understanding of hyporheic processes and their ecological consequences at the catchment scale. After describing a conceptual model of streambed hydraulic connectivity, we use field data collected in 119 surveys of 83 stream reaches across continental France (Datry et al., 2014) to fit and cross-validate an empirical model of reach-scale conductivity as a function of candidate geomorphic and hydraulic controls.

2. Conceptual model of streambed hydraulic conductivity

Multiple processes likely influence the presence of fine sediments within the streambed and hence its hydraulic conductivity (Fig. 1). These processes drive fine sediment supply, retention on and within the streambed, and fine sediment removal. Fine sediment is supplied from scour of the upstream streambed or banks, and from erosion within the catchment (Wood and Armitage, 1997). Worldwide, land clearance, logging, and mining have increased catchment fine sediment supply whilst sediment control, sand mining, and trapping with dams offsets some of these increases (Walling, 2006; Descloux et al., 2010; Datry et al., 2014).

Fine sediments are normally deposited on the streambed contemporaneously with coarser- grained sediments (Lisle, 1989). In addition, suspended sediments may encounter the streambed through various processes including slackwater deposition, biofilm interception, and hyporheic exchange (Karwan and Saiers, 2012). Infiltrated fine sediment can be trapped just beneath an armour layer on the streambed surface or transported farther into the streambed by advection with downwelling pore water or through gravitational settling and then trapped by straining, settling, or chemical adhesion within the coarse sediment interstices (i.e., depth filtration) (Brunke, 1999; Blaschke et al., 2003; Cui et al., 2008; Nowinski et al., 2011; Karwan and Saiers, 2012). Depth filtration has been observed to extend into the streambed up to 0.5 m (Brunke, 1999; Blaschke et al., 2003; Olsen and Townsend, 2005). Many of these processes contribute to clogging (Blaschke et al., 2003) or colmation (Brunke, 1999), reducing the hydraulic conductivity and porosity of the streambed sediments, thereby altering hyporheic zone functions (Packman and MacKay, 2003; Datry et al., 2014).

For mobile streambeds, the effect of episodic scour-and-fill processes (or turnover) on clogging and the implications for hydraulic conductivity are not well understood (Packman and Brooks, 2001; Gartner et al., 2012). Bedload transport has been shown to inhibit clogging in flume experiments (Packman and Brooks, 2001; Rehg et al., 2005) and in streams (Evans and Wilcox, 2014). In contrast, streams with episodic bed mobilisation can exhibit a cyclical clogging behaviour initiated by a high flow event flushing fine sediments from the streambed (Genereux et al., 2008), followed by declining hydraulic conductivity with increased clogging in upper streambed layers over time (Schalchli, 1992; Hatch et al., 2010), and finally reaching a quasi-equilibrium state (Blaschke et al., 2003).

Although subject to little investigation, biological activity also influences streambed hydraulic conductivity (Statzner and Sagnes, 2008; Nogaro et al., 2009; Statzner, 2012). Biofilm growth is likely to enhance clogging (Mendoza-Lera and Mutz, 2013) and root growth and borrowing of biota may create preferential flow paths and increase conductivity (Battin and Sengschmitt, 1999; Mermillod-Blondin and Rosenberg, 2006). For example, tubificid worms can dig networks of galleries in fine sediment, creating preferential flow pathways and increasing hydraulic conductivity (Nogaro et al., 2006). As with clogging by fine sediments, these processes are likely to evolve over time but could be reduced or reset by scour of the streambed.

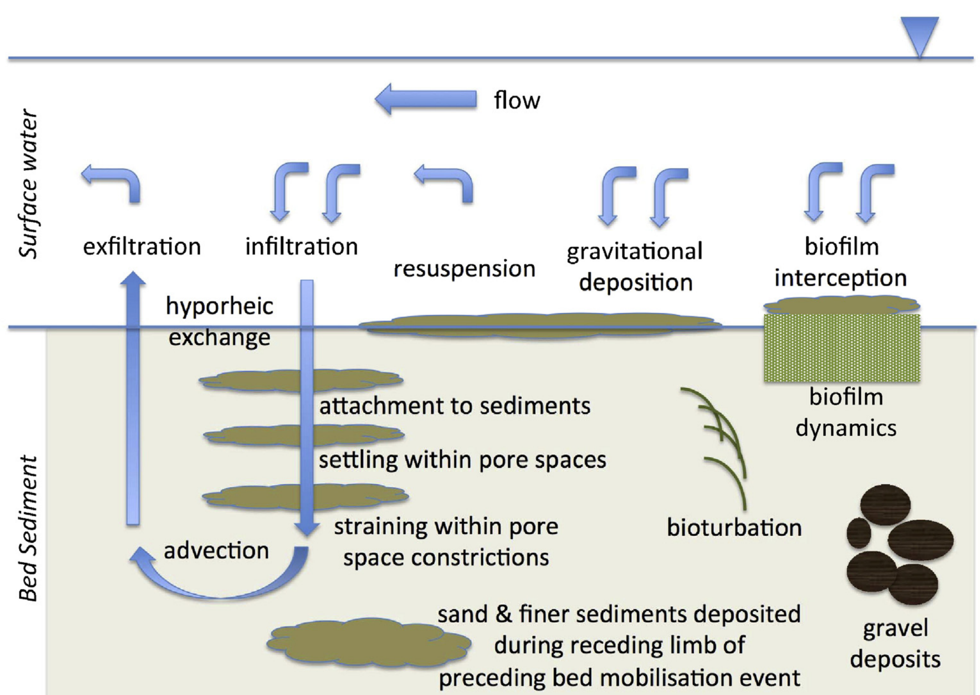

Fig. 1. Physical and biological processes affecting hydraulic conductivity of streambeds.

In this paper, we hypothesise that the evolution of clogging generally consists of three phases commencing with bed mobilisation, followed by a transient clogging phase and ending with a saturated phase, assuming sufficient time between bed mobilisation events (Fig. 2). The initial phase is produced by a flow pulse during which the bed is mobilised. These events flush the fines out of the coarse bed sediments (Genereux et al., 2008) and increase hydraulic conductivity to a maximum *initial* value (K_0 in Fig. 2). The value of K_0 will be controlled largely by the characteristics of the bed-material load deposited following the event. Clogging occurs progressively during the second phase (Schalchli, 1992; Hatch et al., 2010) over some timescale (T_s in Fig. 2). During the clogging phase, hydraulic conductivity is reduced by the continuous deposition of particles within the sediment matrix. The rate at which this happens will be determined by transported sediment concentration and size, hyporheic exchange flux (which is responsible for advective exchange of fines), and flow velocity (which regulates deposition and resuspension of fines). In the final or *saturated* phase, the streambed is completely saturated with fines (Blaschke et al., 2003) and hydraulic conductivity asymptotes to some minimum value (K_S in Fig. 2). During the saturated phases, additional clogging is balanced by processes that maintain conductivity of the bed sediment such as exfiltration, resuspension, and bioturbation. In the following sections we undertake an initial evaluation of this conceptual model based on a statistical analysis of extensive field measurements of hydraulic conductivity across a number of rivers in France. This study makes opportunistic use of an extensive hydraulic conductivity data set unique in terms of the large number of sites included in the sample. Ideally, a test of this model would use time-series observations of hydraulic conductivity at these sites. With such a data set, a parameterised model of the relation indicated in Fig. 2 could be fitted to field data and fitted parameter values could be related to relevant site characteristics, potentially using nondimensional forms of the relevant variables. However, in this initial study only one (or in some cases two observations spaced over several months) is available at each site. For this reason we have chosen a data-mining approach for exploring influences on hydraulic conductivity choosing explanatory variables based on the conceptualisation above.

3. Methods

3.1. Study reaches

Between February 2010 and October 2011, 153 field surveys of reach hydraulic conductivity were made across 100 stream sites in France. This field program was part of a study to assess use of sediment hydraulic conductivity as a measure of streambed clogging (Datry et al., 2014). Of these sites, 18 were chosen according to their clogging conditions (9 clogged and 9 unclogged sites, as judged by local water managers). The other 82 sites were selected randomly across nine regions in France (Fig. 3) and presumably covered a range of French stream types. Of the original 153 surveys, 119 surveys from 83 sites (Fig. 3) have been used here; 34 surveys were excluded because required complimentary data (channel and hydrology information, see below) were unavailable. The 119 surveys were in coarse-bed river reaches with riffle D_{84} bed sediment size corresponding to coarse sand (1–2 mm), gravel (2–64 mm) or cobble (64–256 mm) in 1, 31, and 51 reaches, respectively. Catchment areas varied between 5 and 1680 km^2 with a median of 138 km^2. Bankfull width varied between 2 and 120 m with a median of 11 m and 90% of rivers narrower than 26 m. Between 9 and 35 surveys were made in each region (Fig. 3).

3.2. Measurements of reach hydraulic conductivity

For each reach, the following field protocol was used for estimating reach hydraulic conductivity using point measurements (Datry et al., 2014). First, the mean wetted width was estimated by measuring three randomly selected wetted widths. To include several sequences of the available geomorphic units, the length of each reach (40 to 840 m) was then defined as 19 times the mean wetted width (Leopold et al., 1964). Along each reach, 10 transects were sampled with a spacing of 2 times the mean wetted width. One to three measurements of point hydraulic conductivity were then made at each transect. The position of each measurement along each transect was randomly selected a priori in the laboratory. When measurements could not be carried out at a given point because of bedrock or a water depth > 90 cm, the point was moved along the transect until the measurement could be carried out. A total of 2482 measurements were recorded across the sites with between 14 and 30 measurements at each site (Datry et al., 2014).

Hydraulic conductivity was estimated using a falling head slug test (Lee and Cherry, 1978; Butler, 1998; Baxter et al., 2003; Genereux et al., 2008) with the protocol described in detail by Datry et al. (2014). The test involved inserting a mini piezometer (120 cm long, 1.7 cm internal diameter, 4 cm screened area with 0.4 cm mesh screens) into the streambed to a depth of 25 cm, so that the screened area was between 18 and 22 cm below the streambed

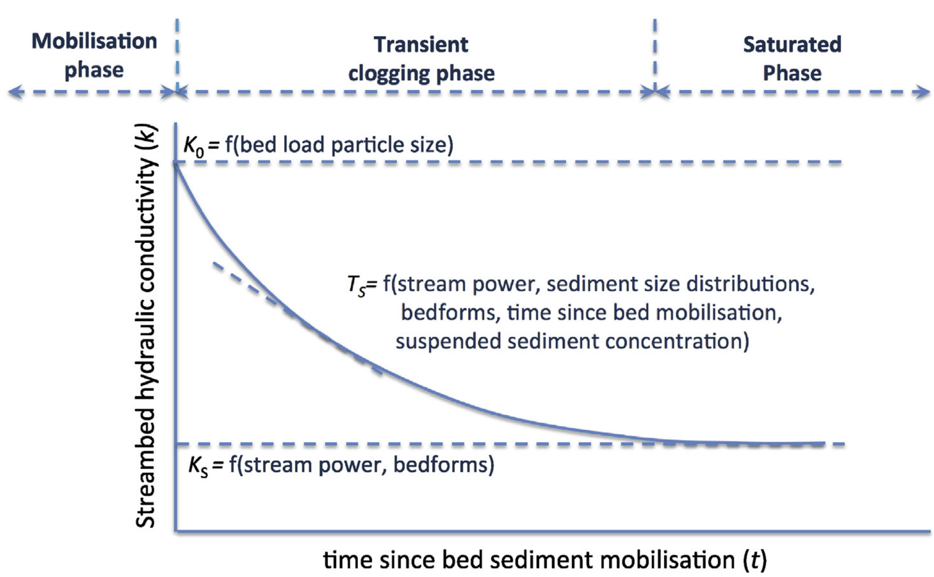

Fig. 2. Key variables implicated in process of streambed clogging.

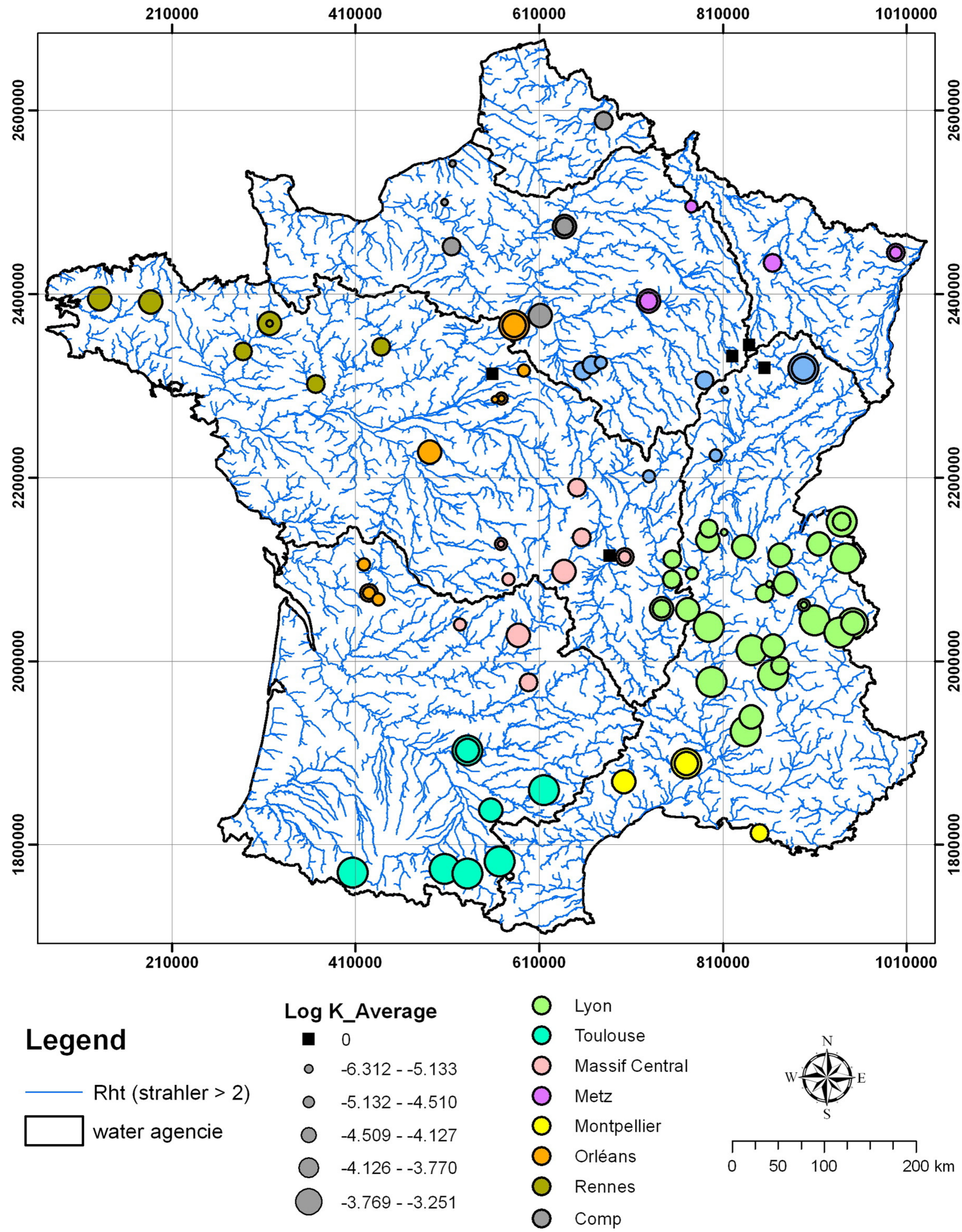

Fig. 3. Maps of hydraulic conductivity values (diameter of the circle, log-scale) in the studied reaches, regions used for the analysis (colours), and water agency boundaries (black lines).

surface. When 25 cm could not be reached but the penetrating depth was >10 cm, the measurement was carried out and the penetrating depth into the sediments was measured. If the penetrating depth was <10 cm, the measurement was not made and the piezometer was randomly displaced along the transect. The initial water level in the piezometer was recorded and then water was added to the piezometer using a funnel. The time for the water level to fall by a fixed height was recorded. If after 2 min, no change in water level was observed in the funnel, zero infiltration was recorded and the hydraulic conductivity was recorded as zero (or more correctly below the detection limit). Methods for calculating hydraulic conductivity are provided by Datry et al. (2014). Although these tests only evaluate conductivity in the vicinity of the mini piezometer, they are relatively low cost and require little time and therefore represent an interesting tool for large-scale monitoring surveys. Point measurements were averaged to provide an estimate of the reach hydraulic conductivity (k_{reach}).

3.3. Identifying candidate predictors of hydraulic conductivity

Our conceptual model (Fig. 2) suggests that streambed hydraulic conductivity depends on: an initial hydraulic conductivity immediately following a major streambed disturbance; the timescale for declining hydraulic conductivity with clogging; the time since the last streambed disturbance; and a final hydraulic conductivity when quasi-equilibrium is achieved. We identified a total of eight available predictor variables (Table 1) related to one or more of these characteristics. These predictors included four static reach-scale variables that can influence deposition and sorting processes within the reach (Figs. 1 and 2), a dynamic variable potentially reflecting the influence of the last disturbance (Fig. 2), as well as three catchment-scale variables related to sediment supply from the catchment and its alteration.

The sites were mapped to river segments in a digital representation of the river network of France derived from a digital elevation model with 50-m resolution (Pella et al., 2012). Estimates of mean flow (Q_{MAF}), valley slope (S_{valley}), and catchment area were available for each segment. The variables describing the channel geometry come from the CarHyCE database developed for characterizing the hydromorphology of rivers (Gob et al., 2014). Since 2009, the French National Agency for Water and Aquatic Environments (ONEMA) has been building the database that collects hydromorphological data on more than 1000 French river reaches. Data are collected following a standardized field survey allowing a detailed description of the river channel morphology and of how it functions.

3.3.1. Reach-scale variables

At the reach-scale, grain size measurements (Wolman pebble count) are made to calculate the 84th quantile (D_{84}). This *coarse sediment size* is expected to predict the initial hydraulic conductivity following a streambed disturbance and also the volume of interstices and hence time required for clogging to develop. We expect larger coarse sediment sizes to produce greater hydraulic conductivity as a result of greater initial hydraulic conductivity and longer time scales for clogging.

A field survey was undertaken at each site using an at-a-station hydraulic geometry approach (Navratil, 2005). The bankfull width and depth, averaged from measurements on 15 cross sections spaced at an interval of one bankfull width, are used to calculate the *bankfull channel mean width/depth ratio* (W/H). This is included as a candidate predictor variable characterizing the mode of sediment transport with ratios <10 indicating that suspended load dominates total load (Schumm, 1985). We expect that suspended-sediment-dominated streams will have reduced hydraulic conductivity as a result of higher levels of clogging with fine suspended sediments (i.e., reducing the time scale T_S for clogging in Fig. 2).

On every cross section a minimum of seven measurements of depth are made and used to characterize the streambed form through the calculation of the standard deviation in channel depth, used here as a measure of *bedform amplitude* (H_b). Hyporheic flux is thought to increase with bedform size measured either by their amplitude or wavelength and the consequent size of periodic fluctuations in static or dynamic head with bedforms such as pool-riffle sequences (Gooseff et al., 2006; Tonina and Buffington, 2011). We use H_b only because bedform wavelength was not recorded for the sites, but it is likely to be correlated with bedform amplitude. We expect increased downwelling flux to enhance rates of fine sediment delivery to the streambed and promote clogging (i.e., reducing the timescale T_s in Fig. 2). However, we recognise that features other than bedform, including channel sinuosity and flow obstructions (e.g., logs), also produce hyporheic exchange and that these are neglected here for simplicity.

The *mean stream power* (P) for the reach is included as a measure of sediment transport capacity as suggested by Prosser and Rustomji (2000) and tends to be peak in the mid-catchment (Knighton, 1999), corresponding to Schumm's (1977) transport zone. In this mid-catchment transport zone, we might expect more frequent mobilisation of the streambed and consequently less clogging and greater hydraulic conductivity. Mean stream power (P) was calculated for each site using $\rho g Q_{MAF}\, S_{valley}$, with water density ($\rho_w$) of 1000 kg/m^3 and gravitational acceleration (g) of 9.81 m/s^2. Note that the valley mean slope is likely to be slightly greater than the river gradient, particularly for a sinuous river, but reliable direct measures of longitudinal river channel gradient were not available.

3.3.2. Time since last disturbance

Given the potential role of fine sediment flushing during high flow pulses and subsequent clogging, we also include a predictor variable, which measures the ($\log_e$) *time since the last streambed disturbance* ($\log_e$T) to indicate the potential extent of clogging (i.e., the *x*-axis in Fig. 2). We expect a decline in hydraulic conductivity with increasing time since the last bed disturbance with an asymptote to some minimum value. To identify the most recent sediment flushing event, we used Shields (1936) entrainment function to give the critical shear stress for motion of bed sediments as $\tau_c = \theta_c g d(\rho_s - \rho_w)$. We used: dimensionless critical shear stress (θ_c) of 0.06, which is within the range for hydraulically rough conditions and mixed bed sediment (Gordon et al., 2004); 2650 kg/m^3 as the density of sediment (ρ_s); and D_{84} as the characteristic sediment diameter (d). We considered the threshold discharge for bed disturbance to occur when the reach bed shear stress $\tau_{reach} > \tau_c$, and we used a power function of discharge to estimate τ_{reach}. The exponent and coefficient of this power function were calculated from estimates of τ_{reach} at the actual discharge during channel surveys and at bankfull using $\tau_{reach} = \rho g R S_{valley}$. Discharge at the time of the survey (Q_{mes}) was measured directly. The bankfull discharge was estimated using the Manning equation that gives discharge (Q) as proportional to $AR^{2/3}$ assuming a constant Manning *n* coefficient and stream gradient. The variables *A* and *R* are the cross-sectional area and hydraulic radius, respectively. Therefore $Q_{bf} / Q_{mes} = (A_{bf} R_{bf}^{2/3}) / (A_{mes} R_{mes}^{2/3})$.

Streamflow records from the nearest available gauge (based on catchment area ratio) were used to estimate the time since the last bed disturbance event when $\tau_{reach} \geq \tau_c$. Streamflow gauge data is only used if the gauge was in operation for the period of survey and two

Table 1
Relative contribution of predictor variable to final BRT model fitted to all data.

Variable	Symbol	Units	Relative contribution %	Range for external cross-validation
Reach-scale predictors				
Riffle sediment size	D_{84}	mm	7.2	4–10
Bankfull channel mean width/depth ratio	W/H	–	23	18–40
Bedform amplitude	H_b	m	29	13–31
Mean stream power	P	W/m	21	16–31
Time since last streambed disturbance	Log_eT	$\log_e$(days)	11	6–18
Catchment-scale predictors				
Catchment agricultural soil erosion risk	E_r	–	8.4	6–9
Distance to the next upstream dam	L_{dam}	km	0	0–0
Proportion of catchment area that flows into an upstream dam	A_{dam}/A_{site}	%	0	0–0

preceding years. Discharge was scaled by catchment area to adjust for the differences in catchment area between the streamflow gauge and study site. The median catchment area ratio (i.e., study site/gauge area) was 0.33 and varied between 0.001 and 6.7. The clogging period was $\log_e$ transformed because the distribution of raw clogging times was highly skewed toward small values. This method for estimating clogging period required a number of approximations and is likely to have a larger relative error than the other predictor variables.

3.3.3. Catchment-scale variables

Anthropogenic disturbances to the rate and size distribution of sediment supplied to the survey sites is likely to be critically important with increased fine and coarse sediment loads expected to have opposing effects on hydraulic conductivity. Increased fine sediment supply is likely to increase the rate of clogging thereby reducing hydraulic conductivity. Increased coarse sediment supply may increase bed sediment mobility where bed sediment load is supply limited, limiting potential for clogging resulting in a greater hydraulic conductivity. Note that we do not include catchment-scale variables related to natural geomorphic controls such as catchment area, geology, and slope because the reach-scale channel form variables are more direct measures. In contrast, the utility of channel form variables as surrogates for catchment disturbances and their effect on sediment regimes is far from certain. The responses of river channels to such disturbances can be quite complex and may take decades to reach an equilibrium state associated with coarse sediment dynamics within the downstream river channel.

We include three catchment-scale predictor variables that relate to such disturbances in sediment supply. A measure of *catchment agricultural soil erosion risk* (E_r) is included as an indicator of the sediment load delivered to the stream network and is calculated based on the soil erosion risk developed by the Institut National de la Rescherche Agronomique (INRA) (Montier et al., 1998), aggregated at local basin scale. To focus on the anthropogenic fine sediment delivery, we weighted the INRA index by the agricultural practices that leave bare soil in winter (Chandesris et al., 2009). Then we calculated a catchment area-weighted average for each site.

Two metrics describe dam impacts on stream sediment loads: the *distance to the next upstream dam* (L_{dam}) and the *proportion of catchment area that flows into an upstream dam* (A_{dam}/A_{site}). These are calculated using the topological properties available from a national hydrographical network for France (Pella et al., 2012). The L_{dam} metric is the distance between survey site to the nearest upstream large dam (with a > 5 m), following the main river (i.e., not tributaries). Operations to flush sediment from dams may temporarily elevate clogging and hence reduce hydraulic conductivity for some distance downstream of the dam. Although the timing of such flushing operations is not available, we include L_{dam} as a measure of the potential for such an effect. The A_{dam}/A_{site} variable is calculated by dividing catchment area upstream from at least one large dam (i.e., with height > 5 m) by the catchment area of the survey site. This metric is included as a measure of the potential for reduced sediment loads at study sites as a result of sediment trapping in upstream dams.

The eight predictor variables showed weak correlations in only a small number of pair-wise comparisons (see Supplementary material) with only six pairs (out of a possible 28 pairwise correlations) showing correlation coefficients >0.3. H_b and P were positively correlated ($r = 0.67$) and both were negatively correlated with L_{dam} ($r = -0.42$ and $r = -0.58$ respectively) and positively correlated with A_{dam}/A_{site} ($r = 0.38$ and $r = 0.54$, respectively). Not unexpectedly, A_{dam}/A_{site} and L_{dam} are negatively correlated ($r = -0.84$).

3.4. Modelling method

We used a boosted regression tree (BRT) analysis to relate k_{reach} to the eight predictor variables. This is a machine learning technique and highly flexible for representing interacting and non-linear relations with few statistical assumptions required (Elith et al., 2008; Hjort et al., 2014). The BRT technique uses regression tree models, but these are combined with boosting which builds and combines a collection of models (Elith et al., 2008). It has produced models in physical geography that are more transferable to other regions than with conventional generalised linear models (Hjort et al., 2014) and also successfully applied in Ecology (Pittman et al., 2009).

We followed the procedure for fitting and evaluating BRT models recommended by Elith et al. (2008) including steps to optimise and interpret the model. Models were fitted in R (R Development Core Team, 2012) using the gbm.step function included with brt.functions (version 2.9) provided as supplementary material with Elith et al. (2008). Three technical parameters influence the selected trees (bag fraction, learning rate and tree complexity). These parameters were optimised by searching the suggested range of values for each of these and all possible combinations. The optimum parameters (i.e., those that produced the maximum cross-validation predictive performance) were obtained with bag fraction = 0.5, learning rate = 0.005, and tree complexity = 7.

Cross-validation was used for model fitting to estimate the optimal number of trees. This cross validation is *internal*, i.e., the test data influences the model finally selected. Elith et al. (2008) acknowledge that this internal cross-validation used for model fitting can still overstate the predictive performance of the model. Hence, an external cross-validation is also reported based on sequentially omitting one of the nine regions (Fig. 3) when fitting the model in turn and testing the model on this omitted data. This ensures test data are independent of model fitting procedure.

We used Friedman's (2001) method, implemented in gbm, to assess the importance of each predictor variable expressed as relative contribution percentage and to generate partial dependence plots. We also examined the sensitivity of these contributions to omission of each region in turn during the external cross-validation procedure.

As a further test of the generality of model results, we divided the data into two groups of regions, each of which included 68 samples. Group A includes the sites in the regions labelled COMP, DIJON, LYON, and MONTPEL (Fig. 3). Group B includes the sites in the regions labelled MC, METZ, ORLEANS, RENNES, and TOULOUSE (Fig. 3). We fitted BRT models independently to these two groups and assessed regional differences in the predictor influence and partial dependency plots.

4. Results

Hydraulic conductivity varied up to a maximum value of 5.6×10^{-4} m/s across the 119 reaches (Fig. 4). The lower detection limit using this equipment is uncertain, but for our purposes we consider 1.0×10^{-6} m/s to be the lower bound for this method of estimating reach-average values. The distribution of values was skewed toward lower values, and 9% of values were recorded at or below this lower detection limit.

Three comparisons were made between observed reach hydraulic conductivity and modelled values to evaluate the BRT model performance using the proportion of deviance predicted by the model. The final model, fitted using all data, explained 72% of the deviance in observed data (Fig. 5A). An internal cross-validation procedure used in fitting the model provides a better estimate of model predictive performance (Elith et al., 2008) and in this case the model explained 37% of the observed deviance. This cross-validation procedure can still overestimate predictive performance so we conducted a completely independent external cross-validation. In this case the model predicted 30% of the total deviance for the 119 reaches (Fig. 5B). In all cases, the model tended to underpredict high values of k_{reach} and over-predict low values. Most of the sites sit within one of four regions defined by river basin boundaries. Median residuals show no systematic bias between these river basin regions (Fig. 6). However residuals are lower for basins flowing to the northwest of France (i.e., Loire and Seine river basins).

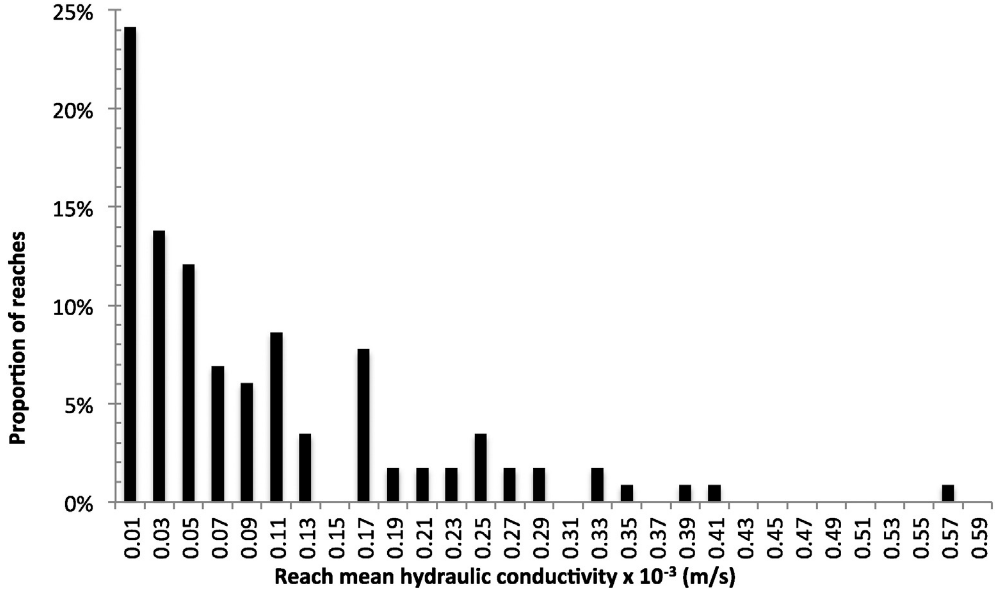

Fig. 4. Distribution of reach hydraulic conductivity for rivers in France (n = 119). Bars indicate the proportion of reaches in hydraulic conductivity classes (steps of 0.2×10^{-5} m/s). *X*-axis label are the central values of classes.

The model results suggest that six out of the eight predictor variables are important in explaining variations in k_{reach}. Only the two dam metrics (L_{dam} or $A_{\text{dam}}/A_{\text{site}}$) make no contribution and this is also the case for each of the nine models fitted for the external cross-validation (Table 1). Different BRT models fitted to the two subgroups provide partial dependence (Fig. 7) with similar trends for the six contributing predictor variables as those using all data combined.

The three most important predictor variables in the final BRT model are H_b, W/H, and P with a total relative contribution of 73% (Table 1). Partial dependence plots show a positive response to increases in these predictor variables (Fig. 6). The partial dependence plots indicate a threshold response to H_b and W/H. The variable k_{reach} transitions to higher values when H_b is between 0.8 and 1.4 corresponding to the 20th and 90th percentile values of H_b (Fig. 7). The transition with W/H is between 10 and 12 corresponding to the 65th and 80th percentile values of W/H (Fig. 6). Outside these ranges, k_{reach} is insensitive to variations in H_b and W/H, respectively. However, k_{reach} increase monotonically with mean stream power for the range of values up to the 80th percentile value (~400 W/m) with greatest sensitivity where stream power is <100 W/m.

The time since the last streambed disturbance provides an 11% relative contribution to the model (Table 1) with an approximate decline in k_{reach} with increasing time since the last streambed disturbance up to a year (i.e., $\log_e T = 5.9$) or more (Fig. 7). This is consistent with the expected declines in hydraulic conductivity with increased clogging by fine sediments over time (Fig. 2). Catchment erosion risk (E_r) contributes 8.4% to the model (Table 1) with increasing k_{reach} in the range 0.6 to 1.3 corresponding to the 65th and 90th percentile values of E_r.

5. Discussion

The upper limit for the range in k_{reach} values reported in this study (5.6×10^{-4} m/s) is consistent with published values including ranges of: 1.2×10^{-4} to 7.4×10^{-4} (Chen, 2010); 2.0×10^{-4} to 5.5×10^{-4} (Cheng et al., 2010); 0.2×10^{-4} to 1.3×10^{-4} (Genereux et al., 2008); and 1.3×10^{-4} to 6.6×10^{-4} (Song et al., 2009) (all units in m/s). Despite the dominance of coarse-bed rivers in our study, this upper limit is more than two orders of magnitude lower than hydraulic conductivity expected for well-sorted gravel (estimated to be between 0.1 and 1 m/s by Bear, 1972). This result is consistent with our hypothesis

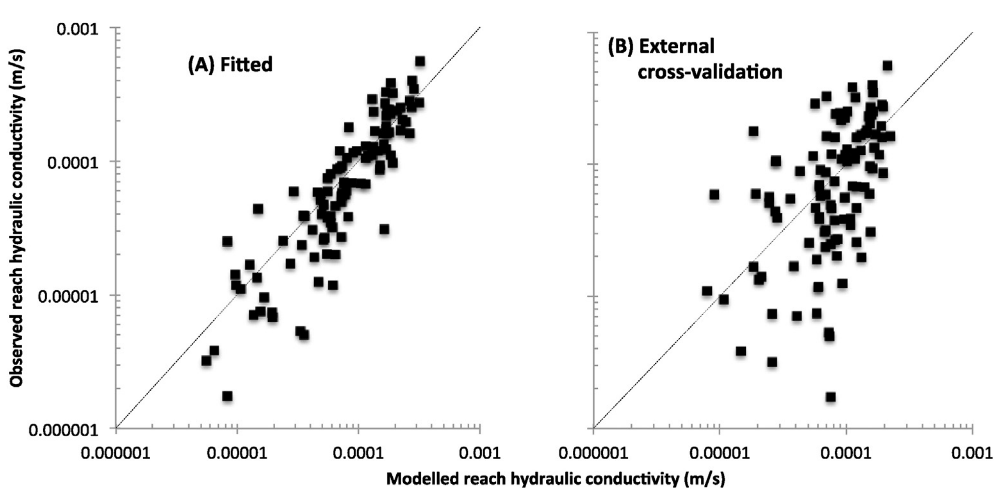

Fig. 5. Reach mean hydraulic conductivity predicted by: (A) applying the final fitted BRT to all reaches; and (B) using a region-based external cross-validation procedure that ensures model calibration and test data are fully independent (dashed line is 1:1, 11 sites with average values $<1 \times 10^{-6}$ m/s are considered below the detection limit and not included in the plot).

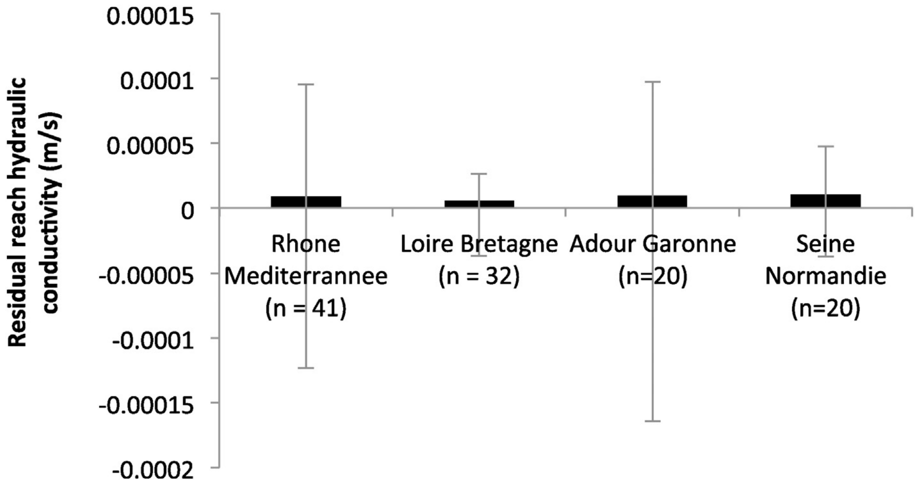

Fig. 6. Comparison of residuals in modelled reach-average hydraulic conductivity for four basins (bar indicates the median residual and whiskers indicates 5th and 95th percentile values).

that the fine sediment fraction controls the hydraulic conductivity of streambed sediments.

Of the six predictor variables contributing to our BRT model (Table 1), the partial dependencies for four of the reach-scale variables (i.e., D_{84}, W/H, P, and $\log_e T$ in Fig. 6) are consistent with those expected and support the conceptual model. Increasing hydraulic conductivity with coarser sediment size is consistent with higher initial conductivities following streambed disturbances and slow rates of clogging with larger interstitial pore volumes. The effect of increasing sediment transport capacity of the site (as measured by stream power P) appears to be increased hydraulic conductivity of the streambed. According to our conceptual model, this effect is expected to be the result of increased mobility of the streambed sediments and hence reduced opportunity for the development of clogging. The importance of bed sediment mobility is also supported by the results for W/H, with a step increase in hydraulic conductivity at W/H = 10, corresponding exactly with the threshold associated with the transition from suspended-sediment-dominated transport regime to a mixed bed-suspended load transport (Schumm, 1985). Finally, evidence supports a decline in hydraulic conductivity with time since the last disturbance of the streambed ($\log_e T$) suggesting the development of clogging effects over several months, and potentially continuing to develop for over one year. This is consistent with field estimates of the residence times of fine particles in river beds of between 4 and 300 days in unregulated rivers and longer in regulated rivers (Gartner et al., 2012). This result supports the method we have used for estimating *T* despite the assumptions required given limited hydraulic and streamflow data available in such a regional analysis.

Another two predictor variables contributing to our model (H_b and E_r in Fig. 6) produced unexpected results, and these two variables were respectively the strongest and weakest contributing variables to our model. The importance of bedform amplitude (H_b) as the strongest contributing predictor to our BRT model is an interesting result with the possibility of positive feedback between hyporheic exchange and hydraulic conductivity. The pressure gradients producing hyporheic pumping through bedforms have a well-established positive dependence on bedform amplitude (H_b), with hyporheic exchange velocities um ~ H_b ^ a with a = 3/8 to 3/2 depending on water depth (Elliott and Brooks, 1997). Therefore, in streams with greater bedform amplitude, we can expect stronger pressure gradients and consequently greater hyporheic flux and flow velocities within bed sediments. These conditions may promote the maintenance of flow pathways through the bed sediments by advection of fine sediment deeper into the streambed and exfiltration fine sediments at upwelling zones. In addition, greater hyporheic flux may support a greater biomass of organisms within the hyporheic zone (Hendricks, 1993; Jones, 1995; Malard et al., 2002) and this may lead to the creation of preferential flow paths as organisms move through and to bioturb sediments and hence higher k_{reach} (Nogaro et al., 2006, 2009; Marmonier et al., 2012).

The model results (Fig. 2) suggest that increasing agricultural soil erosion risk (E_r) produces a higher hydraulic conductivity of the streambed. However, we expected that these conditions would lead to elevated fine sediment supply and increase the risk of clogging. This unexpected result may be because few sites are situated in areas with high E_r levels and are statistically not significant. Furthermore, the natural high fine sediment load (especially Rhône Alps) in this region may explain why k_{reach} is insensitive to low levels of E_r (<65th percentile value across the 119 sites). The lack of any influence by variables related to upstream dams (L_{dam} or A_{dam}/A_{site}) may be because of the site-specific nature of these impacts or possibly because of the reach-scale channel form variables are more effective surrogates for the effects of dams on sediment regime.

As discussed in the paragraphs above, the results for D_{84}, stream_power, W/H, and $\log_e T$ all support the central importance of clogging and flushing with bed mobilisation as dominant processes controlling site-to-site variations in hydraulic conductivity. This is consistent with field observations of clogging following high flow events leading to a decline in hydraulic conductivity over time (Schalchli, 1992; Blaschke et al., 2003; Hatch et al., 2010) and also flume and field experiments where bedload transport has inhibited clogging (Packman and Brooks, 2001; Rehg et al., 2005; Evans and Wilcox, 2014). However, few field studies have taken time-series of hydraulic conductivity measurements and this could be an important area of future research.

On the basis of these observations, we propose that a dominant control on the hydraulic conductivity of streambeds is the frequency of bed sediment mobilisation. We can consider three scenarios where bed sediment disturbances are frequent, rare or intermediate. Where bed disturbances are frequent (i.e., more than 6 per year), streambed hydraulic conductivity is likely to be larger than at other sites and exhibit little variation associated with clogging because of the limited time available for the effects of clogging processes to develop. In sites where bed disturbances are rare (i.e., fewer than 1 per year), we expect that the hydraulic conductivity is low as a result of a well-developed clogging layer at the surface of the streambed, and variation in hydraulic conductivity may in fact be low where clogging has been able to establish a quasi-equilibrium state. Finally, in streams with an intermediate frequency of bed disturbances (i.e., 2–6 per year) we might expect variable hydraulic conductivity depending on the stage in evolution of clogging effects since the previous bed disturbance. If streambed disturbances are concentrated in a wet season, hydraulic conductivity may vary seasonally. Additionally, in multichannel rivers, some channels

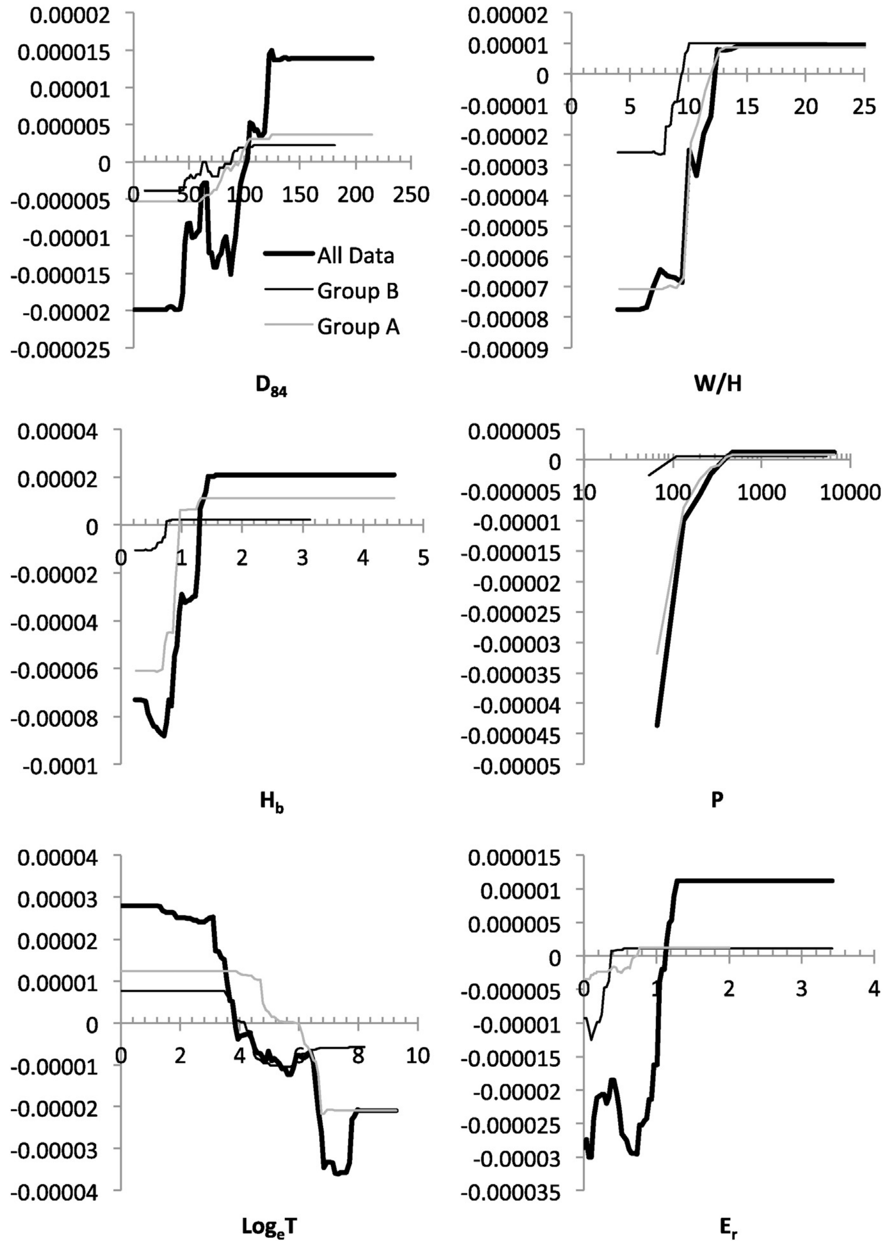

Fig. 7. Partial dependence plots for the eight predictor variables with BRT models fitted to all data and using half the data (i.e., groups A and B) [n.b. the *y*-axis indicates relative changes in reach hydraulic conductivity (m/s), and units for predictor variables (*x*-axis) is given in Table 1].

that experience more frequent disturbances may have higher streambed hydraulic conductivity. Consistent with this model, Gartner et al. (2012) observed that clogging of bed sediments in regulated rivers (with less frequent bed disturbances) was greater than in unregulated rivers and that sediment residence times were longer. These hypotheses and the central proposition that the frequency of streambed disturbances moderates hydraulic conductivity warrant further investigation.

Although our conceptual model emphasises the role of the fine sediment fraction and clogging as controls on temporal and spatial variations in streambed hydraulic conductivity, we anticipated that the coarse sediment size may influence the initial hydraulic conductivity following a streambed disturbance and also the rate of clogging. The results are consistent with this expectation, showing increased k_{reach} for sites with larger coarse sediments (D_{84}). The model did not reveal any interactions between the effects of D_{84} and $\mathrm{Log_eT}$, suggesting that D_{84} may not influence rate of clogging and hence that an effect on initial hydraulic conductivity is the more likely explanation for this result. However, time-series observations of clogging in streambeds of variable coarse sediment size would be necessary to confirm the influence of coarse sediment fraction on streambed hydraulic conductivity.

Our analysis did not reveal any influence of dams on reach-scale conductivity, maybe because the spatial distribution of our sites is not suited for identifying such effects. In particular, our sites may be situated too far from dams to be strongly influenced by their functioning. In addition, the variety of sediment management among dams (e.g., timing of flushing operations) may obscure any effect. Extending our study to reaches bypassed by dams or situated just downstream is likely to provide different results (Descloux et al., 2010).

Our results have sensible physical interpretation and our cross-validations indicated consistency across regions. However, the external cross-validations also indicated that the model only predicts a limited part of the observed deviance. A first explanation of this limited predictive success is the availability of physical predictors as well as their uncertainty when extrapolated spatially (Lamouroux et al., 2014). For example, our hypothesis of a constant Manning n (at a site) is a simplistic assumption that may influence our estimation of bed movement frequency. A second explanation is that some processes such as the interactions between groundwater and surface water and biotic influences on conductivity are not well taken into account by our eight predictors. More generally, comparable large-scale studies to predict grain-size distribution (Snelder et al., 2011) or the probability of intermittency (Snelder et al., 2013) across the French hydrographic network also revealed difficulties to obtain accurate reach-scale predictions. Overall, large-scale approaches such as ours are useful for identifying the drivers of observed hydraulic conductivity and their relative influence but cannot provide accurate predictions at the reach-scale.

Finally, stronger predictions may result from a more detailed investigation of variations within a river basin including consideration of spatial correlations along rivers associated with processes such as downstream fining of bed sediments. In this regard, it is reassuring that results do not show any systematic bias in the final BRT model between river basins. However, examination of trends within river networks would be worthwhile in a further study using closely spaced or continuous measurements along a river.

6. Conclusions

Streambed hydraulic conductivity can vary over several orders of magnitude potentially exerting a strong control on spatial and temporal variation in hyporheic flow regimes, including hyporheic flux and residence times. Hydraulic conductivity, even of coarse bed streams, is characteristic of sand and finer sediments indicating that processes of streambed clogging are critical. This empirical study found that hydraulic conductivity depends primarily on reach geometry (increases with bedform amplitude, bankfull channel width-depth ratio), stream power, and catchment erodibility but decreases with time since the last bed disturbance. These results suggest that streams with a predominantly suspended load and less frequent streambed disturbances have lower hydraulic conductivity. Based on this study we propose that the connectivity of surface water and hyporheic zones is dependent on river sediment dynamics including the frequency of streambed disturbance events and fine sediment loads during the intervening periods, which lead to clogging.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.geomorph.2016.02.001.

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