1 Variation in reach-scale hydraulic conductivity of streambeds

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17 **Abstract**

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

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40 *Keywords:* hyporheic exchange; sediment dynamics; clogging; colmation; fine sediment

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42 **1. Introduction**

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

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

63 hydrology, channel morphology, and associated fluvial processes (Malard et al., 2002; Tonina
64 and Buffington, 2009).

65 Although measurements of streambed conductivity have been reported from a broad range of
66 stream types, few empirical studies link spatial (between sites) and temporal (with time)
67 variations in streambed hydraulic conductivity to flow, catchment characteristics, and other
68 geomorphic drivers. Point measurements of streambed hydraulic conductivity found in the
69 literature vary between 10^{-10} and 10^{-2} m/s (Calver, 2001), and reach-average values are
70 between 10^{-5} and 10^{-3} m/s (Genereux et al., 2008; Song et al., 2009; Chen, 2010; Cheng et al.,
71 2010; Min et al., 2012; Taylor et al., 2013). This upper limit on reach-average values is an
72 order of magnitude lower than might be expected for a uniform gravel [e.g., the Hazen
73 formula (Hazen, 1892) estimates hydraulic conductivity of 0.04 m/s for particle size
74 diameters of 2 mm]. This is because streambed sediments generally have a broad distribution
75 of particle sizes and because hydraulic conductivity is largely determined by the smaller size
76 fractions (Alyamani and Sen, 1993; Song et al., 2009; Descloux et al., 2010a). Consequently,
77 variation in hydraulic conductivity between reaches is likely the result of processes
78 controlling presence of fine sediments in the streambed rather than the coarse fraction.
79 Further, point-scale measurements vary considerably within a reach. In some rivers, sections
80 of streambed may be effectively impermeable but the streambed is rarely impermeable
81 throughout the river channel. The lowest reported value of 10^{-10} m/s, is five orders of
82 magnitude smaller than the lowest reported reach-average value.

83 In this study we model spatial and temporal variations in hydraulic conductivity to support
84 advances in our understanding of hyporheic processes and their ecological consequences at
85 the catchment scale. After describing a conceptual model of streambed hydraulic connectivity,
86 we use field data collected in 119 surveys of 83 stream reaches across continental France

87 (Datry et al., 2014) to fit and cross-validate an empirical model of reach-scale conductivity as
 88 a function of candidate geomorphic and hydraulic controls.

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90 **2. Conceptual model of streambed hydraulic conductivity**

91

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

93

94 Multiple processes likely influence the presence of fine sediments within the streambed and
 95 hence its hydraulic conductivity (Fig. 1). These processes drive fine sediment supply,
 96 retention on and within the streambed, and fine sediment removal. Fine sediment is supplied
 97 from scour of the upstream streambed or banks, and from erosion within the catchment

98 (Wood and Armitage, 1997). Worldwide, land clearance, logging, and mining have increased
99 catchment fine sediment supply whilst sediment control, sand mining, and trapping with
100 dams offsets some of these increases (Walling, 2006; Descloux et al., 2010b; Datry et al.,
101 2014).

102 Fine sediments are normally deposited on the streambed contemporaneously with coarser-
103 grained sediments (Lisle, 1989). In addition, suspended and bedload sediments may
104 encounter the streambed through various processes including slackwater deposition, biofilm
105 interception, and hyporheic exchange (Karwan and Saiers, 2012). Infiltrated fine sediment
106 can be trapped just beneath an armour layer on the streambed surface or transported farther
107 into the streambed by advection with downwelling pore water or through gravitational
108 settling and then trapped by straining, settling, or chemical adhesion within the coarse
109 sediment interstices (i.e., depth filtration) (Brunke, 1999; Blaschke et al., 2003; Cui et al.,
110 2008; Nowinski et al., 2011; Karwan and Saiers, 2012). Depth filtration has been observed to
111 extend into the streambed up to 0.5 m (Brunke, 1999; Blaschke et al., 2003; Olsen and
112 Townsend, 2005). Many of these processes contribute to clogging (Blaschke et al., 2003) or
113 colmation (Brunke, 1999), reducing the hydraulic conductivity and porosity of the streambed
114 sediments, thereby altering hyporheic zone functions (Packman and MacKay, 2003; Datry et
115 al., 2014).

116 For mobile streambeds, the effect of episodic scour-and-fill processes (or turnover) on
117 clogging and the implications for hydraulic conductivity are not well understood (Packman
118 and Brooks, 2001; Gartner et al., 2012). Bedload transport has been shown to inhibit clogging
119 in flume experiments (Packman and Brooks, 2001; Rehg et al., 2005) and in streams (Evans
120 and Wilcox, 2014). In contrast, streams with episodic bed mobilisation can exhibit a cyclical
121 clogging behaviour initiated by a high flow event flushing fine sediments from the streambed

122 (Genereux et al., 2008), followed by declining hydraulic conductivity with increased clogging
123 in upper streambed layers over time (Schalchli, 1992; Hatch et al., 2010), and finally reaching
124 a quasi-equilibrium state (Blaschke et al., 2003).

125 Although subject to little investigation, biological activity also influences streambed hydraulic
126 conductivity (Statzner and Sagnes, 2008; Nogaro et al., 2009; Statzner, 2012). Biofilm growth
127 is likely to enhance clogging (Mendoza-Lera and Mutz, 2013) and root growth and borrowing
128 of biota may create preferential flow paths and increase conductivity (Battin and Sengschmitt,
129 1999; Mermillod-Blondin and Rosenberg, 2006). For example, tubificid worms can dig
130 networks of galleries in fine sediment, creating preferential flow pathways and increasing
131 hydraulic conductivity (Nogaro et al., 2006). As with clogging by fine sediments, these
132 processes are likely to evolve over time but could be reduced or reset by scour of the
133 streambed.

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

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

153 completely saturated with fines (Blaschke et al., 2003) and hydraulic conductivity asymptotes
154 to some minimum value (K_s in Fig. 2). During the saturated phases, additional clogging is
155 balanced by processes that maintain conductivity of the bed sediment such as exfiltration,
156 resuspension, and bioturbation. In the following sections we undertake an initial evaluation of
157 this conceptual model based on a statistical analysis of extensive field measurements of
158 hydraulic conductivity across a number of rivers in France. This study makes opportunistic
159 use of an extensive hydraulic conductivity data set unique in terms of the large number of
160 sites included in the sample. Ideally, a test of this model would use time-series observations of
161 hydraulic conductivity at these sites. With such a data set, a parameterised model of the
162 relation indicated in Fig. 2 could be fitted to field data and fitted parameter values could be
163 related to relevant site characteristics, potentially using nondimensional forms of the relevant
164 variables. However, in this initial study only one (or in some cases two observations spaced
165 over several months) is available at each site. For this reason we have chosen a data-mining
166 approach for exploring influences on hydraulic conductivity choosing explanatory variables
167 based on the conceptualisation above.

168 **3. Methods**

169 *3.1. Study reaches*

170 Between February 2010 and October 2011, 153 field surveys of reach hydraulic conductivity
171 were made across 100 stream sites in France. This field program was part of a study to assess
172 use of sediment hydraulic conductivity as a measure of streambed clogging (Datry et al.,
173 2014). Of these sites, 18 were chosen according to their clogging conditions (9 clogged and 9
174 unclogged sites, as judged by local water managers). The other 82 sites were selected
175 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 km² with a median of 138 km². 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).

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

187 3.2. Measurements of reach hydraulic conductivity

188 For each reach, the following field protocol was used for estimating reach hydraulic
189 conductivity using point measurements (Datry et al., 2014). First, the mean wetted width was
190 estimated by measuring three randomly selected wetted widths. To include several sequences
191 of the available geomorphic units, the length of each reach (40 to 840 m) was then defined as
192 19 times the mean wetted width (Leopold et al., 1964). Along each reach, 10 transects were
193 sampled with a spacing of 2 times the mean wetted width. One to three measurements of
194 point hydraulic conductivity were then made at each transect. The position of each
195 measurement along each transect was randomly selected *a priori* in the laboratory. When
196 measurements could not be carried out at a given point because of bedrock or a water depth
197 >90 cm, the point was moved along the transect until the measurement could be carried out.
198 A total of 2482 measurements were recorded across the sites with between 14 and 30
199 measurements at each site (Datry et al., 2014).

200 Hydraulic conductivity was estimated using a falling head slug test (Lee and Cherry, 1978;
201 Butler, 1998; Baxter et al., 2003; Genereux et al., 2008) with the protocol described in detail
202 by Datry et al. (2014). The test involved inserting a mini piezometer (120 cm long, 1.7 cm
203 internal diameter, 4 cm screened area with 0.4 cm mesh screens) into the streambed to a
204 depth of 25 cm, so that the screened area was between 18 and 22 cm below the streambed
205 surface. When 25 cm could not be reached but the penetrating depth was >10 cm, the
206 measurement was carried out and the penetrating depth into the sediments was measured. If
207 the penetrating depth was <10 cm, the measurement was not made and the piezometer was
208 randomly displaced along the transect. The initial water level in the piezometer was recorded
209 and then water was added to the piezometer using a funnel. The time for the water level to fall
210 by a fixed height was recorded. If after two minutes, no change in water level was observed in
211 the funnel, zero infiltration was recorded and the hydraulic conductivity was recorded as zero

212 (or more correctly below the detection limit). Methods for calculating hydraulic conductivity
213 are provided by Datry et al. (2014). Although these tests only evaluate conductivity in the
214 vicinity of the mini piezometer, they are relatively low cost and require little time and
215 therefore represent an interesting tool for large-scale monitoring surveys. Point
216 measurements were averaged to provide an estimate of the reach hydraulic conductivity
217 (k_{reach})

218 *3.3. Identifying candidate predictors of hydraulic conductivity*

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

228 The sites were mapped to river segments in a digital representation of the river network of
229 France derived from a digital elevation model with 50-m resolution (Pella et al., 2012).
230 Estimates of mean flow (Q_{MAF}), valley slope (S_{valley}), and catchment area were available for
231 each segment. The variables describing the channel geometry come from the CarHyCE
232 database developed for characterizing the hydromorphology of rivers (Gob et al., 2014). Since
233 2009, the French National Agency for Water and Aquatic Environments (ONEMA) has been
234 building the database that collects hydromorphological data on more than 1000 French river

235 reaches. Data are collected following a standardized field survey allowing a detailed
236 description of the river channel morphology and of how it functions.

237 3.3.1. Reach-scale variables

238 At the reach-scale, grain size measurements (Wolman pebble count) are made to calculate the
239 84th quantile (D_{84}). This *coarse sediment size* is expected to predict the initial hydraulic
240 conductivity following a streambed disturbance and also the volume of interstices and hence
241 time required for clogging to develop. We expect larger coarse sediment sizes to produce
242 greater hydraulic conductivity as a result of greater initial hydraulic conductivity and longer
243 time scales for clogging.

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

252 On every cross section a minimum of seven measurements of depth are made and used to
253 characterize the streambed form through the calculation of the standard deviation in channel
254 depth, used here as a measure of *bedform amplitude* (H_b). Hyporheic flux is thought to
255 increase with bedform size measured either by their amplitude or wavelength and the
256 consequent size of periodic fluctuations in static or dynamic head with bedforms such as pool-
257 riffle sequences (Gooseff et al., 2006; Tonina and Buffington, 2011). We use H_b only because

258 bedform wavelength was not recorded for the sites, but it is likely to be correlated with
259 bedform amplitude. We expect increased downwelling flux to enhance rates of fine sediment
260 delivery to the streambed and promote clogging (i.e., reducing the timescale T_s in Fig. 2).
261 However, we recognise that features other than bedform, including channel sinuosity and
262 flow obstructions (e.g., logs), also produce hyporheic exchange and that these are neglected
263 here for simplicity.

264 The *mean stream power* (P) for the reach is included as a measure of sediment transport
265 capacity as suggested by Prosser and Rustomji (2000) and tends to be peak in the mid-
266 catchment (Knighton, 1999), corresponding to Schumm's (1977) transport zone. In this mid-
267 catchment transport zone, we might expect more frequent mobilisation of the streambed and
268 consequently less clogging and greater hydraulic conductivity. Mean stream power (P) was
269 calculated for each site using $\rho g Q_{MAF} S_{valley}$, with water density (ρ_w) of 1000 kg/m³ and
270 gravitational acceleration (g) of 9.81 m/s². Note that the valley mean slope is likely to be
271 slightly greater than the river gradient, particularly for a sinuous river, but reliable direct
272 measures of longitudinal river channel gradient were not available.

273 3.3.2. Time since last disturbance

274 Given the potential role of fine sediment flushing during high flow pulses and subsequent
275 clogging, we also include a predictor variable, which measures the ($\log_e$) *time since the last*
276 *streambed disturbance* ($\log_e T$) to indicate the potential extent of clogging (i.e., the x-axis in Fig.
277 2). We expect a decline in hydraulic conductivity with increasing time since the last bed
278 disturbance with an asymptote to some minimum value. To identify the most recent sediment
279 flushing event, we used Shields (1936) entrainment function to give the critical shear stress
280 for motion of bed sediments as $\tau_{*c} = \frac{1}{2} \frac{\rho_w}{\rho_s} (1 - \frac{\rho_w}{\rho_s})$. We used: dimensionless critical

281 shear stress (θ_c) of 0.06, which is within the range for hydraulically rough conditions and
282 mixed bed sediment (Gordon et al., 2004); 2650 kg/m³ as the density of sediment (ρ_s); and
283 D_{84} as the characteristic sediment diameter (d). We considered the threshold discharge for
284 bed disturbance to occur when the reach bed shear stress $\tau_{reach} > \tau_c$, and we used a power
285 function of discharge to estimate τ_{reach} . The exponent and coefficient of this power function
286 were calculated from estimates of τ_{reach} at the actual discharge during channel surveys and at
287 bankfull using $\tau_{reach} = \rho g R S_{valley}$. Discharge at the time of the survey (Q_{mes}) was measured
288 directly. The bankfull discharge was estimated using the Manning equation that gives
289 discharge (Q) as proportional to $AR^{2/3}$ assuming a constant Manning n coefficient and stream
290 gradient. The variables A and R are the cross-sectional area and hydraulic radius, respectively.
291 Therefore $Q_{bf} / Q_{mes} = (A_{bf} R_{bf}^{2/3}) / (A_{mes} R_{mes}^{2/3})$.

292 Streamflow records from the nearest available gauge (based on catchment area ratio) were
293 used to estimate the time since the last bed disturbance event when $\tau_{reach} \geq \tau_c$. Streamflow
294 gauge data is only used if the gauge was in operation for the period of survey and two
295 preceding years. Discharge was scaled by catchment area to adjust for the differences in
296 catchment area between the streamflow gauge and study site. The median catchment area
297 ratio (i.e., study site/gauge area) was 0.33 and varied between 0.001 and 6.7. The clogging
298 period was $\log_e$ transformed because the distribution of raw clogging times was highly
299 skewed toward small values. This method for estimating clogging period required a number
300 of approximations and is likely to have a larger relative error than the other predictor
301 variables.

302 3.3.3. Catchment-scale variables

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

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

324 Two metrics describe dam impacts on stream sediment loads: the *distance to the next*
325 *upstream dam* (L_{dam}) and the *proportion of catchment area that flows into an upstream dam*
326 ($A_{\text{dam}}/A_{\text{site}}$). 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 $A_{\text{dam}}/A_{\text{site}}$ 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 $A_{\text{dam}}/A_{\text{site}}$ ($r = 0.38$ and $r = 0.54$, respectively). Not unexpectedly, $A_{\text{dam}}/A_{\text{site}}$ 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

349 conventional generalised linear models (Hjort et al., 2014) and also successfully applied in
350 Ecology (Pittman et al., 2009).

351 We followed the procedure for fitting and evaluating BRT models recommended by Elith et al.
352 (2008) including steps to optimise and interpret the model. Models were fitted in R (R
353 Development Core Team, 2012) using the `gbm.step` function included with `brt.functions`
354 (version 2.9) provided as supplementary material with Elith et al. (2008). Three technical
355 parameters influence the selected trees (bag fraction, learning rate and tree complexity).
356 These parameters were optimised by searching the suggested range of values for each of
357 these and all possible combinations. The optimum parameters (i.e., those that produced the
358 maximum cross-validation predictive performance) were obtained with bag fraction = 0.5,
359 learning rate = 0.005, and tree complexity = 7.

360 Cross-validation was used for model fitting to estimate the optimal number of trees. This
361 cross validation is *internal*, i.e., the test data influences the model finally selected. Elith et al.
362 (2008) acknowledges that this internal cross-validation used for model fitting can still
363 overstate the predictive performance of the model. Hence, an external cross-validation is also
364 reported based on sequentially omitting one of the nine regions (Figure 3) when fitting the
365 model in turn and testing the model on this omitted data. This ensures test data are
366 independent of model fitting procedure.

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

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

377 4. Results

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

383

384

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

390
 391 Fig. 5: Reach mean hydraulic conductivity predicted by: (A) applying the final fitted BRT to all
 392 reaches; and (B) using a region-based external cross-validation procedure that ensures model
 393 calibration and test data are fully independent (dashed line is 1:1, 11 sites with average
 394 values $< 1 \times 10^{-6}$ m/s are considered below the detection limit and not included in the plot).

395
 396 Three comparisons were made between observed reach hydraulic conductivity and modelled
 397 values to evaluate the BRT model performance using the proportion of deviance predicted by
 398 the model. The final model, fitted using all data, explained 72% of the deviance in observed

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

409

410

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

414 The model results suggest that six out of the eight predictor variables are important in
 415 explaining variations in k_{reach} . Only the two dam metrics (L_{dam} or A_{dam}/A_{site}) make no

416 contribution and this is also the case for each of the nine models fitted for the external cross-
417 validation (Table 1). Different BRT models fitted to the two subgroups provide partial
418 dependence (Fig. 7) with similar trends for the six contributing predictor variables as those
419 using all data combined.

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

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

437

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

<i>Variable</i>	<i>Symbol</i>	<i>Units</i>	<i>Relative Contribution %</i>	<i>Range for external cross- validation</i>
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	$\text{Log}_e T$	$\text{log}_e(\text{days})$	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	$A_{\text{dam}}/A_{\text{site}}$	%	0	0 - 0

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].

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

452 $\times 10^{-4}$ 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
453 $\times 10^{-4}$ to 6.6×10^{-4} (Song et al., 2009) (all units in m/s). Despite the dominance of coarse-bed
454 rivers in our study, this upper limit is more than two orders of magnitude lower than
455 hydraulic conductivity expected for well-sorted gravel (estimated to be between 0.1 and 1
456 m/s by Bear, 1972). This result is consistent with our hypothesis that the fine sediment
457 fraction controls the hydraulic conductivity of streambed sediments.

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

476 we have used for estimating T despite the assumptions required given limited hydraulic and
477 streamflow data available in such a regional analysis.

478 Another two predictor variables contributing to our model (H_b and E_r in Fig. 6) produced
479 unexpected results, and these two variables were respectively the strongest and weakest
480 contributing variables to our model. The importance of bedform amplitude (H_b) as the
481 strongest contributing predictor to our BRT model is an interesting result with the possibility
482 of positive feedback between hyporheic exchange and hydraulic conductivity. The pressure
483 gradients producing hyporheic pumping through bedforms have a well-established positive
484 dependence on bedform amplitude (H_b), with hyporheic exchange velocities $u_m \sim H_b^a$ with a
485 $= 3/8$ to $3/2$ depending on water depth (Elliott and Brooks, 1997). So in streams with greater
486 bedform amplitude, we can expect stronger pressure gradients and consequently greater
487 hyporheic flux and flow velocities within bed sediments. These conditions may promote the
488 maintenance of flow pathways through the bed sediments by advection of fine sediment
489 deeper into the streambed and exfiltration fine sediments at upwelling zones. In addition,
490 greater hyporheic flux may support a greater biomass of organisms within the hyporheic zone
491 (Hendricks, 1993; Jones, 1995; Malard et al., 2002) and this may lead to the creation of
492 preferential flow paths as organisms move through and to bioturb sediments and hence
493 higher k_{reach} (Nogaro et al., 2006; 2009; Marmonier et al., 2012).

494 The model results (Fig. 2) suggest that increasing agricultural soil erosion risk (E_r) produces a
495 higher hydraulic conductivity of the streambed. However, we expected that these conditions
496 would lead to elevated fine sediment supply and increase the risk of clogging. This unexpected
497 result may be because few sites are situated in areas with high E_r levels and are statistically
498 not significant. Furthermore, the natural high fine sediment load (especially RhôneAlps) in
499 this region may explain why k_{reach} is insensitive to low levels of E_r (<65th percentile value

500 across the 119 sites). The lack of any influence by variables related to upstream dams (L_{dam} or
501 $A_{\text{dam}}/A_{\text{site}}$) may be because of the site-specific nature of these impacts or possibly because of
502 the reach-scale channel form variables are more effective surrogates for the effects of dams
503 on sediment regime.

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

513 On the basis of these observations, we propose that a dominant control on the hydraulic
514 conductivity of streambeds is the frequency of bed sediment mobilisation. We can consider
515 three scenarios where bed sediment disturbances are frequent, rare or intermediate. Where
516 bed disturbances are frequent (i.e., more than 6 per year), streambed hydraulic conductivity
517 is likely to be larger than at other sites and exhibit little variation associated with clogging
518 because of the limited time available for the effects of clogging processes to develop. In sites
519 where bed disturbances are rare (i.e., fewer than 1 per year), we expect that the hydraulic
520 conductivity is low as a result of a well-developed clogging layer at the surface of the
521 streambed, and variation in hydraulic conductivity may in fact be low where clogging has
522 been able to establish a quasi-equilibrium state. Finally, in streams with an intermediate
523 frequency of bed disturbances (i.e., 2-6 per year) we might expect variable hydraulic

524 conductivity depending on the stage in evolution of clogging effects since the previous bed
525 disturbance. If streambed disturbances are concentrated in a wet season, hydraulic
526 conductivity may vary seasonally. Additionally, in multichannel rivers, some channels may
527 experience more frequent disturbances that have higher streambed hydraulic conductivity.
528 Consistent with this model, Gartner (2012) observed that clogging of bed sediments in
529 regulated rivers (with less frequent bed disturbances) was greater than in unregulated rivers
530 and that sediment residence times were longer. These hypotheses and the central proposition
531 that the frequency of streambed disturbances moderates hydraulic conductivity warrant
532 further investigation.

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

543 Our analysis did not reveal any influence of dams on reach-scale conductivity, maybe because
544 the spatial distribution of our sites is not suited for identifying such effects. In particular, our
545 sites may be situated too far from dams to be strongly influenced by their functioning. In
546 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., 2010a).

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

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

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597

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