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Thomas Dépret, Clément Virmoux, Emmanuèle Gautier, Hervé Piégay, Mariya Doncheva, et al.. Lowland gravel-bed river recovery through former mining reaches, the key role of sand. Geomorphology, 2021, 373, pp.107493. 10.1016/j.geomorph.2020.107493 . hal-03793233

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Abstract

Whereas the geomorphic effects of in-channel mining have been widely documented, very few studies have focused on recovery trajectories after mining ended. This paper describes and quantifies the channel adjustment of extended and over-widened mining pits located in a low energy gravel-bed river, and identifies the factors that control their recovery rate. The study was based on aerial photographs and LiDAR analysis, grain-size sampling, bedload computation and geophysical measurements made during ground penetrating radar (GPR) and electrical resistivity tomography (ERT) surveys. We describe the spatio-temporal changes in the morphology of former pits as well as the current surficial longitudinal grain-size pattern in pits and in neighbouring upstream and downstream reaches. Our results show a wide range of geomorphic readjustments. Between 23 and 47 yr after the end of mining, the mining reaches have still not yet fully recovered. The planimetric recovery rate is closely correlated with the ratio of the volume deposited in the pits to their original volume. Finally, the infilled material is only partly composed of bedload and the mainly sandy fine sediments play a primary role in the recovery process. The abundance of sand, controlled by the lithology of the watershed, explains the unexpectedly high recovery rates of some pits despite the low energy of the river.

Keywords

Channel mining; Geomorphic resilience; Sand; Low-energy river

Highlights

- Characterisation of morphological and grain-size recovery of former in-channel pits
- Adjustment still underway 27-47 years after mining stopped
- Recovery rate controlled by pit volume, the volume of sediment supplied and pit trap efficiency
- The sand load plays an essential role in recovery processes
- Catchment lithology also crucial in recovery

1. Introduction

The alluvial resources provided by river corridors have been overused worldwide since the end of World War II because of the advent of concrete as a building material and to growing development. The detrimental social and environmental effects of this mining has led to control of, or to a complete ban on gravel mining in many countries in recent decades (Kondolf, 1997; Rinaldi et al., 2005). Gravel mining directly or indirectly modifies both the river bed geometry and the substrate, can lead to a profound and long-lasting alteration of aquatic and riparian habitats as well as to the destruction or weakening of fluvial engineering structures and loss of alluvial groundwater storage thereby reducing water resources (Kondolf, 1994; Bravard et al., 1997; Bravard et al., 1999a; Brown et al., 1998;

Koehnken et al., 2020). Although these effects and the factors that control them are widely documented (e.g., Bull and Scott, 1974; Collins and Dunne, 1989; Kondolf, 1994, 1997; Surian and Rinaldi, 2003; Simon and Rinaldi, 2006; Wyzga, 2007; Sreebha and Padmalal, 2011; Arróspide et al., 2018), very little is known about channel recovery after mining ends.

The present review focuses on in-stream dry-pit and wet-pit mining, which, as defined by Kondolf, (1994), both consist of dredging a pit in the channel of varying depth, width and length. We excluded bar skimming mining operations (Kondolf, 1994; Rempel and Church, 2009), defined as the scalping of the dry surficial layer of bars at low-flow. For these specific cases, both modelling and field approaches have already been explored. Physical, numerical or analytical modelling has focused on the spatio-temporal pattern of pit filling and/or how mining modifies hydrodynamics and bed morphology (e.g., Fredsoe, 1979; van Rijn, 1986; Kornis and Laczay, 1988; Lee et al., 1993; Gill, 1994; Barman et al., 2017, 2018, 2019; Neyshabouri et al., 2002; Cao and Pender, 2004; Chen et al., 2010). These authors particularly highlight the downstream migration of pits as they fill up as well as the degradation of the bed both upstream and downstream from the initial location of the pits. Most studies using these approaches used pit dimensions (pit width ≤ channel width, pit length ≤ 2 x channel width) that are considerably smaller than those of the pits reported in most rivers (e.g., Lee et al., 1993; Gill, 1994; Barman et al., 2017, 2018, 2019; Neyshabouri et al., 2002; Chen et al., 2010). Two main infill stages have been identified (Lee et al., 1993). During the first stage, termed the convective stage, the maximum scour depth remains relatively constant and the upstream boundary of the pit migrates up to the original downstream edge. The following diffusion stage corresponds to complete pit infill and is characterised by the downstream migration and the progressive reduction of the maximum scour depth. The infilling rate of pits is theoretically controlled by the amount of upstream sediment supplied and by pit trap efficiency (van Rijn, 1986), which in turn, is mainly controlled by pit geometry (Neyshabouri et al., 2002).

Field studies have also been conducted to understand pit infilling. Yuill et al. (2016) used time-series bathymetric surveys of a 1.6 km length borrow-pit whose width occupies around a quarter of the lower Mississippi width combined with a 3D morphodynamic model to describe the morphological changes in the pit and to quantify its infilling rate. These authors showed that the pit traps a very small fraction (5-11%) of the sandy bedload passing through it and that the infilling rate depends mainly on the supply of sediment and the dimensions of the pit. Lastly, the pit is hypothesised to be infilled after a period of 4.4 to 10 yr. Using a section of the gravel-bed Semois River (bankfull width = 50-55 m) dredged to a depth of two metres over a distance of one kilometre to prevent downstream flooding, and repeated topographical surveys, Gob et al. (2005) estimated the time needed to return to the initial state of the riverbed at ten years. Trap efficiency was assumed to be total. Many Italian rivers have undergone similar evolutionary trajectories characterised by incision and bed narrowing throughout the second half of the twentieth century because of the combination of human pressure, e.g., channelling and damming, and gravel extraction (e.g., Surian et al., 2009; Comiti et al., 2011; Scorpio et al., 2015; Scorpio and Rosskopf, 2016; Belletti et al., 2016). These authors investigated recovery processes at the reach scale, mainly highlighting bed aggradation and widening, 15-20 yr after mining ended. The magnitude of recovery was considerably less than the magnitude of the changes caused by mining and is mainly controlled by the degree of coarse sediment connectivity between hillslopes or floodplain and the riverbed. In the Middle Loire River (France), bed incision caused by gravel mining stopped within 15 to 20 yr following the end of mining, but aggradation was only observed in a few short reaches (Latapie et al., 2014). Morpho-sedimentary recovery has also been characterised in ephemeral Mediterranean systems 12 to 24 yr after mining ended (Rovira et al., 2005; Calle et al., 2017; Sanchis-Ibor et al., 2017). Using a sediment budget along an 11-km long reach of the fine gravel-bed Tordera River with a mean incision of 1.5 m, Rovira et al. (2005) determined that the pre-extraction bed level would not be recovered for around 420 yr, and that most aggradation only occurred during small floods in dry periods. Calle et al. (2017) reconstituted the evolution of morpho-sedimentary units of the Rambla de la Viuda from aerial photographs before, during and after gravel mining ended. Using LiDAR analysis and RTK-DGPS surveys, these authors identified a 3.5-m incision that continued to exist 12 yr after a major reduction in mining activity. They pinpointed the role of major floods in the recovery processes but considered the bed degradation to be irreversible at multi-decadal scale because of the lack of in-channel sediment connectivity. Sanchis-Ibor et al. (2017) quantified bed degradation in the Palancia River using LiDAR and aerial photographs and examined its recovery through a morphological recovery index computed by combining the width of the flow and active channels, the total sinuosity index and the channel count index from aerial photographs at each date.

In most rivers affected by mining, geomorphic recovery is usually reduced because of fairly low delivery of coarse sediments (e.g., Surian et al., 2009; Comiti et al., 2011; Belletti et al., 2016). Originating from the upstream part of the watershed or from the alluvial stock through lateral erosion, today most of these sediment sources are no longer available because of cumulative basin- and reach-scale effects of human activities (Downs and Piégay, 2019), particularly dams, which trap bedload sediments, and bank protection works (e.g., Nilsson, 2005; Lehner et al., 2011; Graf, 2006; Bravard et al., 1999b; Ollero, 2010; David et al., 2016; Dépret et al., 2017). In low-energy gravel-bed rivers, the potential for readjustment seems to be even lower because of the slowness of the processes. Broadly, even if these rivers are relatively common systems, knowledge of their morphodynamics remains scarce (Dépret at al., 2015, 2017; Lespez et al. 2015) and explains why management practices and restoration actions are often inappropriate (Lespez et al. 2015). In the case of the low-energy gravel-bed River Cher, the focus of this paper, sediments continued to be extracted from the riverbed until the early 1990s (Dépret et al., 2015, 2017) when this kind of mining was banned in France (Malavoi et al., 2011). At several locations along the river course, sediment mining occurred both in the riverbed and the alluvial plain. As a result, the river exhibits pit sections whose length and width are respectively more than ten and two times the width of the ‘natural’ active channel. These sections very likely act in the same way as floodplain pits captured by rivers, temporarily trapping all or most of the bedload (Mossa and Marks, 2011). Because of the analogy with the captured floodplain pits, we termed the over-widened sections of the Cher River pseudo-captures (PCs). PCs may also trap only part of the suspended load. Thanks to these features, PCs are interesting observatories to characterise and understand the mechanisms and potential for geomorphic recovery processes in low-energy gravel-bed rivers. More than 20 yr after the end of mining of the Cher River, recent aerial photographs show varying degrees of PC readjustment, some of which are surprisingly advanced. Following this observation, the aim of this paper is to explain the wide range of adjustment rates by analysing the pseudo-capture geomorphic recovery and by identifying the factors that control the recovery rate. In theory, the recovery rate is mainly controlled by the original volume of the PC, the volume of sediment originating from upstream and the trap efficiency of the PC. The smaller the original volume, the bigger the supply of sediment, especially the bedload volume, and finally, the more sediment trapped in the PC, and the faster the recovery rate. To test this hypothesis, we used an original interdisciplinary methodological approach combining time-series aerial photographs and LiDAR analysis, grain-size sampling, bedload computation and geophysical measurements made using ground penetrating radar (GPR) and electrical resistivity tomography (ERT) surveys. Underground investigations using the two geophysical methods allowed us to confirm some of the results of our analysis of surface data. Finally, we discuss (i) the importance of the suspended load and of the bedload in the recovery processes, (ii) the key role of sand in the recovery processes of low-energy rivers, (iii) the need to consider lithology when estimating recovery potential, and (iv) the relative importance of PCs in the diversity of aquatic and riparian habitats.

2. Study sites

Geological and hydrological context

After leaving its source at 713 m asl., the Cher River mainly flows through gorges across steep slopes or in deep valleys with a very narrow floodplain (upstream Cher in Figure 1). The upstream section of the river is located at the northwest end of the low altitude mountainous Massif Central, which mainly consists of crystalline and metamorphic rocks (Larue, 1981, 2011). After 63 km, the Cher alluvial plain begins (Alluvial Cher in Figure 1) and the river becomes single thread with numerous meandering reaches. After first occupying the Tertiary graben of Montluçon, which continues for 45 km to the confluence with the Aumance River (Larue, 1981, 2011; Simon-Coinçon et al., 2000), the river then crosses the sedimentary domain of the Parisian Basin, where it joins the Loire River at a 38 m asl. Here, the annual average discharge is 90 m^3 s^{-1} and the catchment area covers 13,615 km². The study section is the first 140 km of the Alluvial Cher between Montluçon and Vierzon. Here, the Cher River has developed a meandering pattern for the last millennia (Vayssière et al., 2020). In addition to intense gravel mining in the riverbed from 1950 to 1990, the river is densely equipped with ancient fluvial engineering works including bank protections and weirs that are major obstacles to local sediment supply through lateral erosion (Dépret et al., 2015, 2017; Vayssière et al., 2020). The hydrological regime is mainly influenced by a pluvio-evaporal oceanic climate and is characteristic of lowland rivers in the western part of mid-latitude regions. Maximum flow occurs in February and minimum flow in August. Because of the construction of the Rochebut dam in the early twentieth century, the hydrological regime has become partly artificial, particularly during low flow and especially upstream from the confluence with the Aumance River (Figure 1).

Gravel mining

The period of gravel mining was reconstituted by examining all aerial photographs taken between 1950 and 2016 (available on https://remonterletemps.ign.fr) and the archives of the former Directorate for Industry, Research and the Environment (*Direction de l'Industrie, de la Recherche et de l'Environnement*, French acronym DRIRE), local state agencies in the Cher and Allier administrative departments. On the aerial photographs, we looked at signs of mining activity indicated by the presence/absence of the equipment and infrastructure required for mining and sediment sorting. In the archives, we found some of the official authorisation documents granted to the mining companies (Rollet et al., 2008). Gravel mining began and expanded between the late 1940s and the early 1970s and ended between 1983 and 1993, except at one study site (PRE) where it ended in 1968 (Table 1).

Site selection

Twelve PCs were identified along the 140-km stretch of the river between Montluçon and Vierzon, in aerial photographs taken between 1950 and 2016 (available on https://remonterletemps.ign.fr). Among these, we selected six (Figure 1) by excluding those located immediately upstream or downstream from a weir, or complex planform configuration caused by a neighbouring floodplain pit capture after mining. The sites are named using the three first letters of the main municipality in which they are located, from upstream to downstream: AUD, BOU, BRU, SFL, STH and PRE (Figure 1). Their planimetric, hydraulic and grain-size features are summarised in Table 1. Details on the method used to identify these features are given in Section 3.

Figure 1 – Location of the study sites.

Table 1 – Hydraulic and geometric parameters, grain-size, and periods of gravel mining activity in the study reaches.

Site	**A (km^2)**	**L PC t_0 (m)**	**w PC t_0 (m)**	**w up t_0 (m)**	**w PC t_0 / w up t_0**	**Ws up (W m^{-2})**	**Ws PC t_0 (W m^{-2})**	**D_{50} up (mm)**	**D_{84} up (mm)**	**Start year**	**End year**
AUD	2112	807	131	49	2.7	7-16	4-6	18-44	39-67	1973	1983
BOU	3492	1107	145	48	3	8-15	2-4	8-21	18-41	1950/59-1967	1983
BRU	3924	1641	106	58	1.8	5-9	1-2	5-24	19-38	1950-1959	1989
SFL	4298	2079	77	45	1.7	32-52	2-3	11	24	pre-1950	1993
STH	4385	876	96	43	2.2	28-58	3-6	23	59	1950-1959	1986
PRE	4422	665	183	50	3.4	3-7	1-2	NA	NA	pre-1950	1968

A: Watershed area. L PC t_0: Initial length of PCs, with t_0 the ending of gravel mining activity. w PC t_0: Initial width of PCs. w up t_0: Initial width of active bed upstream of PCs. w PC t_0 / w up t_0: ratio between w PC t_0 and w up t_0. Ws up: Range of specific stream power upstream of PCs for discharge not exceeding 0.99 % of time (94.5 m^3 s^{-1} for AUD, 191 m^3 s^{-1} for the other sites) and along reaches whose length is between 12 and 60 times the active bed width (see Section 3 for explanations and the formula). This discharge was chosen to enable comparison of AUD and the other sites, whose hydrological regime differs because of the Aumance River confluence (Figure 1). It is higher than the critical discharge for bedload incipient motion (Dépret et al., 2015, 2017) and close to the bankfull discharge (85 m^3 s^{-1} for AUD, 178-245 m^3 s^{-1} for the other sites) (Dépret, 2014). The specific stream power is given by the following formula: $\rho_w gQS/w$, where ρ_w is the density of water (1000 kg m^{-3}), g is the gravitational acceleration (9.81 m s^{-2}), Q is the discharge (m^3 s^{-1}), S is the bed slope (m m^{-1}), and w is the bankfull bed width (m). Ws PC t_0: Minimum and mean specific stream power in PC discharge not exceeding 0.99% of time. D_{50} up and D_{84} up: Range of size of bed surface material for which respectively 50% and 84% of the number of sediment clasts are smaller (see Section 3 for explanations).

Between AUD and the five downstream PCs, the discharge is almost doubled because of inputs of water from the Aumance River, the main tributaries of the Cher River between Montluçon and Vierzon (Figure 1). Only minor tributaries join the Cher River between the reference gauge stations and the PCs (Montluçon for AUD and Saint-Amand-Montrond for the other PCs).

In AUD, the bedrock is composed of Eocene-Miocene sand and clays and of conglomeratic or non-conglomeratic Stephanian facies, (Turland et al., 1989a, 1989b). In BOU, the bedrock is Triassic sandstone, sands, argillite and dolomite (Lablanche, 1994; Lablanche et al., 1994). In BRU, the bedrock is Middle Jurassic oolitic and bioclastic limestone as well as of Early Jurassic bioclastic limestone, marl and marl-limestone and Eocene lacustrine limestone and clay (Lablanche, 1994; Lablanche et al., 1994). In SFL, the bedrock is Late Jurassic oolitic or stratified limestone (Debrand-Passard et al., 1997a, 1977b). In STH, the bedrock is stratified or unstratified Late Jurassic limestone and Eocene limestone and clay (Debrand-Passard et al., 1997a, 1977b). Finally, the bedrock in PRE is Eocene limestone and clay (Debrand-Passard et al., 1997a, 1977b).

3. Material and Methods

Our methodological approach combined time-series aerial photographs and LiDAR analysis, grain-size sampling, bedload computation and geophysical measurements using ground penetrating radar (GPR) and electrical resistivity tomography (ERT) surveys.

3.1 PC adjustment patterns and recovery rates

The magnitude of PC recovery was determined from planimetric, topographic and granulometric indexes, presented in that order.

Planimetric adjustments

Planimetric adjustment of the PCs was characterised from the end of mining to 2016 by digitising the surface colonised by perennial vegetation on all the available aerial photos with a GIS (ArcGIS 10.6 - ESRI) (Table 2). Vegetated surfaces were used to characterise planimetric recovery because they are the direct result of infilling of PCs and because they make it possible to precisely describe and quantify the spatial and temporal pattern of floodplain and channel reconstruction. Vegetated surfaces were digitised at a scale ranging from 1:500 to 1:800, allowing a good trade-off between their visual identification and the precise delimitation of their boundaries. Most of the photos from the French National Geographic Institute (*Institut Géographique National*, IGN) or the Centre-Val-de-Loire Regional Direction for the Environment, Planning and Housing (*Direction Régionale de l'Environnement, du de l'Aménagement et du Logement*, DREAL) used here were taken at low flow (Table 2). The most recent ones (i.e., most of those taken between 2002 and 2016) were already orthorectified (resolution range: 0.2-1 m). We rectified and georeferenced the others (maximum RMSE among PCs between 1.4 m and 3.2 m; resolution range: 0.25-1.7 m). In addition, we digitised the active channel (formed by flow channels and bare gravel bars) along a 4-km stretch upstream from the entrance of the PCs. From these data, two main metrics were used to describe the magnitude and the recovery rate. The first, named the width index, is the ratio between the mean width of the active channel in and upstream from the PCs. It was computed at t_0 and in 2016. The upstream width is considered here as the reference width. The closer the index is to 1, the more advanced the recovery. Widths were computed using the Fluvial Corridor Toolbox (Roux et al., 2015). The widths correspond to the mean length of cross sections roughly spaced at one active channel width (30 m for AUD, 40-50 m for the other PCs). The second metric describes the planimetric recovery rate. It corresponds to the percentage reduction in the initial surface area of the PC. It was obtained through the yearly ratio (expressed in percentage per year) between the surface area colonised by perennial vegetation in 2016 and the surface area of the PC at t_0.

In addition, the longitudinal recovery pattern was characterised by computing the width colonised by perennial vegetation (terrestrialised width) at 10-m intervals along the length of the channel and at each date. This procedure enabled us to reconstruct and quantify the spatio-temporal pattern of PC filling.

Table 2 – Temporal resolution at which the PCs' planimetric trajectories were reconstructed. Q = Discharge. Gauging stations: MTL = Montluçon; STA = Saint-Amand-Montrond.

Site	**t_0**	**First photo**	**Most recent photo**	**Study duration (yr)**	**N° of dates**	**Photo interval (yr)**	**Q range ($m^3 s^{-1}$)**	**Min monthly Q ($m^3 s^{-1}$)**	**Gauging station**
AUD	1983	1983	2016	33	9	3.7	1.3-6.9	3	MTL
BOU	1983	1983		33	12	2.8	1.9-54.6	7.1	STA
BRU	1989	1994		27	10	2.7	1.9-54.6	7.1	STA
SFL	1993	1994		23	11	2.1	1.9-25.3	7.1	STA
STH	1986	1986		30	12	2.5	1.9-15.1	7.1	STA
PRE	1969	1969		47	18	2.6	1.8-85.5	7.1	STA

Vertical adjustments

We computed the mean height of the emerged bare and vegetated surfaces above the low-water level in 2011 from a LiDAR DEM (provided by the DREAL Centre-Val-de-Loire and used 'as is') whose mean vertical precision of the original point cloud was +/- 0.089 m (minimum: +/- 0.15 m). We also extracted the low-water elevation from the LiDAR DEM (Biron et al., 2013). In the same way as for the planimetric adjustment, we examined the longitudinal pattern of the height of the emerged deposits in PCs by computing the mean height along cross sections spaced at 10-m intervals along the whole length of the channel. Here, we propose two hypotheses. First, a decrease in height in the downstream direction, expressing the progradation of the deposits through the PCs. Second, the higher the recovery, and hence the planimetric rate, the smaller the height downstream.

For the purpose of comparison, in the same way, we computed the mean height of the 'old' floodplain, namely, that of the deposits beyond the limits of the PC and prior to t_0. This height was determined on a narrow (30-m wide) strip along the PCs. A 10-m buffer zone was established between the edge of the PCs and the edge of the 30 m strip to exclude low bank heights from the computation.

Grain-size adjustments

Lastly, the recovery of PCs was characterised by grain size. We determined the grain size in the PC bed as well as upstream and downstream (at distances ranging between 2.5 km and 4 km). Sediment size was first determined from photos of coarse grain-size patches on each bar along the study reaches. The photos were taken vertically during the 2018 low-flow period and the target area was protected from direct solar radiation. The resolution of the photos was high enough to enable us to identify particles as small as coarse sand. On each photo, we used GIS (ArcGIS 10.6) to measure the apparent b-axis of 100 particles located at the intersection of a regular grid with spacing higher than the maximum longest particle axis (a-axis). Second, sediment size was determined through 400-particle Wolman samples taken on each riffle (the particles were measured with a calliper). Particles were sampled according to a regular grid covering the whole surface of the riffles.

We examined the longitudinal pattern of the D_{50} and the D_{95}. Here we hypothesise that grain-size sorting will continue until recovery is complete, the more advanced the recovery, the smaller the decrease in grain size expected downstream.

3.2 Size and origin of deposited sediments

We first estimated to what extent suspended load and bedload could have been trapped by PCs at t_0. For this purpose, we computed their trap efficiency at intervals of 10 m^3 s^{-1} over the whole discharge range throughout the study period and for each grain-size phi interval from fine silt (4-8 µm) to coarse gravel (32-64 mm). We used a simple theoretical reservoir trap efficiency formula developed from modified overflow rate models and for turbulent flows and varying discharge conditions. Originally proposed by the U.S. Environmental Protection Agency (1986, cited in Haan et al., 1994), it was subsequently modified by Verstraeten and Poesen (2000). We applied the formula to PCs because at least during the initial years after mining ended, they may be considered as weir reservoirs. The formula provides an insight into the trap efficiency of the PC at t_0, with a decrease in efficiency over time as the PC fills up.

$$TE = 100\left[\frac{AV_s}{1.2q_p}\right]$$

where TE is the trap efficiency (in %), A is the original surface area of the PC (m²), V_s is the 10°C clear water settling velocity of each grain size (m s^{-1}), extracted from Table 5.4 in Julien (2010).

We then investigated the role of the suspended load and the bedload in the recovery process by estimating the proportion of fine and coarse material (1 mm threshold) in the infilled volumes of BOU. For this purpose, we combined electrical resistivity tomography (ERT) surveys and grain-size sampling (Figure 2). We first conducted 20 ERT surveys and then took 15 samples at different resistivity measuring points (for resistivity values ranging from 82 Ω m to 3535 Ω m). ERT profiles were carried out using a Wenner-Schlumberger array (Loke, 2000) that has a strong signal to noise ratio, which means that with the electrode spaced 1 m apart, we can reach an investigation depth of 12 m, which is sufficient to map the entire thickness of the infilling of the PCs and the surface of the bedrock. The device we used was a multi-electrode Abem Terrameter LS with 64 electrodes. Field data were acquired in February 2019 and the water table level remained steady throughout the survey period. Data inversion was carried out by Res2dinv software (Loke and Barker, 1996) using a least square inversion technique. All ERT profiles had an RMS error <5% before five iterations. Eleven of the samples were taken manually at depths ranging from 0.25 to 0.8 m, with the mass of the largest particle being less than 1% of the total mass of the sample, as recommended by Church et al. (1987) to obtain a representative grain size. The four last samples were extracted from a 4-m core drilled in 2019 with a Cobra type TT power core drill. These different grain sizes enabled us to establish a relationship between resistivity value and the percentage of sediment volume of less than 1 mm. We then applied this relationship to each of the resistivity measuring points to estimate the percentage of sediment volume less than 1 mm in the vicinity of these points. Bedrock elevation was obtained through complementary GPR surveys and was validated thanks to the existence of logs from locations close to the PCs (see Section 3.3).

Figure 2 – Location of electrical resistivity surveys and grain-size samples in BOU PC.

To estimate the contribution of bedload to the recovery process, we compared the volumes deposited in the PCs (see Section 3.3) with a maximum estimation of the bedload volumes from Recking's formula (2010, 2013). Four calculation steps are needed to get these estimations:

(1) Calculation of the critical Shields number for the D_{84} surface:

$$\tau^*_{c84} = (5S + 0.06)\left(\frac{D_{84}}{D_{50}}\right)^{4.4\sqrt{S}-1,5}$$

where S is the bed slope (m m^{-1}), D_{50} is the median (m) and D_{84} is the 84th percentile of the bed surface grain-size distribution (m).

(2) Calculation of the Shields number applied to the D_{84} for a given discharge:

$$\tau^*_{84} = \frac{S}{(s-1)D_{84}\left[{}^{2}\!/_{w} + 74p^{2.6}(gS)^{p}q^{-2p}D_{84}^{3p-1}\right]}$$

where S is the bed slope (m m^{-1}), w is the bankfull width (m), g is gravitational acceleration (9.81 m s^{-2}), q is the specific discharge (m^3 s^{-1} m^{-1}) defined by the ratio of flow discharge to channel width, and $p = 0.3$ when, like in this case, $\frac{q}{\sqrt{gSD_{84}^3}} > 100$.

(3) Calculation of the Einstein dimensionless parameter:

$$\Phi = \frac{14{\tau^*_{84}}^{2.5}}{\left[1 + \left(\frac{\tau^*_{c84}}{\tau^*_{84}}\right)^4\right]}$$

(4) Calculation of the volumetric bedload transport rate (m^3 s^{-1}):

$$q_{sv} = w\sqrt{g(s-1)D_{84}^3}\,\Phi$$

where w is the bankfull width (m) and s is the relative sediment density, i.e., the ratio of sediment density to flow density (=2.65).

The data needed to calculate volumes are:

- The bed slope, computed from the LiDAR DEM. The reliability of the method was tested with a set of RTK-DGPS low-water levels. The bed slope was surveyed in 2010 and 2011 in 37 sections whose length was around 10 times the width of the active channel (length range: 320-900 m). The LiDAR slope was slightly underestimated but close to the 1:1 line and the correlation with the DGPS slope was strong and highly significant (LiDAR slope = 0.83 x DGPS slope + 0.0001; $R^2 = 0.83$, p-value < 0.0001). Furthermore, the relation is similar to that obtained by Biron et al. (2013) ((LiDAR slope = 0.91 x DGPS slope + 0.002; $R^2 = 0.75$), who concluded that the LiDAR slope can be confidently used to compute stream power.
- The bed width, obtained from GIS analysis and expressed by two parameters:
  (i) bankfull width, which allows computation of the specific water discharge (water discharge per unit width).
  (ii) active channel width, which allows calculation of the total bedload discharge.
- The grain-size distribution determined from photos as explained previously.
- Daily discharges at Montluçon gauging stations for AUD and Saint-Amand-Montrond gauging stations for the other sites (from http://www.hydro.eaufrance.fr/) (Figure 1).

In each site, several bedload volume estimations were obtained by applying the formula upstream to PCs to different reach lengths equal to 12, 18, and 24 times the width of the active channel. The slope, active channel width and bankfull width specific to each reach were used. The slope was obtained from 2011 LiDAR DEM. The bankfull width was obtained from the ratio of the active channel width to the bankfull width computed by Dépret et al. (2015). Different grain sizes were tested. We used the mean grain size available for each reach, the grain size of the last bar just upstream from the entrance to each PC, the grain size of the bar just downstream from the entrance to the PC, and the mean grain size of all bars in all the PCs. Among these combinations, we retained the highest bedload value. It was finally multiplied by two, as Recking et al. (2012) reported that at the decadal scale, the formula provides reliable estimations within a factor envelope of two. For comparison with the bedload volumes, the deposited volumes were corrected using porosity. Because field data were missing, we chose two values (0.1 and 0.5), which represent the wide porosity spectrum encountered in sand and gravel bed rivers (Frings et al., 2011; Liang et al., 2015; Tabesh et al., 2019).

Finally, we looked at the extent to which deposited material in PCs could have originated from the upper reach via vertical or lateral erosion. As no topo-bathymetric data were available, we used low-water elevation surveyed upstream from PCs from 1989-1995 to 2011-2016 as a proxy of vertical erosion (Gurnell et al., 2003; Wyzga, 2007). These data were provided by DREAL Centre-Val-de-Loire and different private engineering companies brought together for the purpose of 1D hydraulic modelling of flood hazards. On a section of the river roughly equal to 60 times the bankfull width (3 km for AUD, 3.6 km for the five other sites), we estimated the volume of sediment supplied to the riverbed by lateral erosion of the floodplain using the 2011 LiDAR as well as t_0 and 2011 aerial photos (acquired in 2013 for AUD and in 2010 for PRE). In the first step, eroded surfaces between t_0 and 2011 were determined using GIS following the method applied by Dépret et al. (2015, 2017). In the second step, their surface area was multiplied by the height of the floodplain above the riverbed. For this purpose, we estimated the height of the floodplain above the water level along each of the eroded surfaces, applying the method presented previously, to compare the height of deposits in the PCs and height of the floodplain along the PCs. We then added +0.5 m to account for water depth (considered as the mean water depth at low flow).

3.3 Controlling variables of the recovery rates

In theory, the planimetric recovery rate is mainly controlled by the original volume of the PC, the volume of sediment originating from upstream and the trap efficiency of the PC. The smaller the original volume, the bigger the supply of sediment, especially the bedload volume, and finally, the more will be trapped in the PC, and the faster the recovery rate. To test this hypothesis, we examined Spearman’s correlation between the 2011 recovery rate and the ratio of the volume deposited in the PCs to the original volume. This ratio expresses the three above-mentioned explanatory variables with the deposited volume in PCs combining the influence of the amount of sediments supplied to PCs as bedload or suspended load on one hand, and their trapping efficiency on the other hand. We used the 2011 recovery rate rather than the 2016 rate, because the explanatory variable we tested was obtained from the 2011 LiDAR DEM.

The original volume was defined as the volume between the low-water level and the bedrock level and was obtained by subtracting the low-water level DEM from the bedrock surface DEM. The low-water level DEM was built from the 2011 LiDAR DEM, assuming the water level at low flow remained constant from t_0 to 2011. This assumption seems to be confirmed by the comparison of low-flow level upstream and downstream from the PCs (no data were available for the PCs) from the end of the 1980s / beginning of 1990s which are clearly stable. In absence of topo-bathymetric data at t_0, we used the bedrock elevation as reference. Here we assume that because the mining companies wished to extract as much sediment as possible, they dug down to the bedrock. The bedrock elevation was estimated along several sections thanks to GPR transects of the PCs at low flow in 2018 (Figure 3). GPR images were acquired with a MALA Ramac CUII (Mala geosciences) coupled with a 100 MHz shielded antenna to reduce unwanted reflections from trees in areas of alluvial forest (Neal, 2004). Topographical surveys were made with an RTK DGPS. GPR observations were recorded at an in-line sampling interval of 0.2 m, a time window of 548 ns and a sampling frequency of 1079 MHz. GPR data were processed with Reflex® software (Jol, 2009; Sandmaier, 2016). The processing flow consisted of six steps. We started with static correction of the first arrival, which sometimes showed irregularities. Then we de-wowed data to remove very low frequency components. Next, we applied a gain correction and energy decay gain to preserve relative amplitude information. Data were then filtered to remove low and high frequency noise using a band-pass Butterworth filter. A topographic profile was then added to each profile to correct topographic effects. Finally, a time-depth conversion was applied based on the diffraction hyperbola fitting (Santaniello et al., 2013). No velocity analysis was performed during the field campaign. A specific velocity pattern was defined for each transect. Velocity ranged from 0.065 m ns^{-1} to 0.095 m ns^{-1}. The dominant bedrock geology is sandy-clay for AUD and BOU and limestone and clay for PRE, STH, SFL and BRU, so in all cases the dielectric constant of the bedrock is higher than the dielectric constant of PC filling, thus we were able to observe the bedrock upper boundary in the form of a more or less continuous reflector (Figure 4). In some cases, the GPR signal was too weak to detect the upper boundary of the bedrock. This can be explained by (1) a clayey deposit that weakened the GPR wave or (2) weathering of the upper part of the bedrock so there was no obvious boundary. In the later case, we chose a discontinuous picking method to map the bedrock upper boundary. The determination of bedrock elevation through GPR was enabled by the existence of logs located near or in the vicinity of the PCs. Logs were reported by the BSS-BRGM database belonging to the Mining and Geological Research Institute (*Bureau des Recherches Géologiques et Minières*) and by the Centre for Studies and Expertise on Risks, Environment, Mobility and Development (*Centre d'études et d'expertise sur les risques, l'environnement, la mobilité et l'aménagement*, French acronym CEREMA). With the GIS, the estimated elevations of the bedrock were then assigned to cross sections passing through each measurement point and perpendicular to the central line. A 1-m resolution bedrock surface DEM was then built from these cross sections.

Figure 3 – Location of ground penetrating radar surveys of the six mined river reaches.

Figure 4 – (A) 100 Mhz Ground Penetrating Radar (GPR) profile at BOU. The solid black line represents the surface of the bedrock. (B) The GPR trace (the vertical dashed green line in A) shows the negative-positive-negative reflection corresponding to the bedrock upper boundary (dashed red line).

The volume deposited in the PCs was calculated by subtracting the 2011 LiDAR DEM from the bedrock surface DEM. Such volumes are an underestimation of the actual volumes deposited because submerged deposits could not be taken into account (the 2011 active channel was excluded from the analysis). Furthermore, because of the reduction in the trap efficiency of PCs over time, the longitudinal sediment connectivity may have been restored before the LiDAR survey, and some deposited particles may have been remobilized and exported to downstream PCs.

4. Results

4.1 Channel adjustments

Planimetric adjustments

The changes in the width index indicate that the six sites have begun to recover as the index has decreased since the end of gravel mining, which happened between 23 and 47 yr before the most recent aerial photograph (2016) we used was taken (Figure 5A). Whereas it originally ranged between 1.71 and 3.66, in 2016, it was between 0.96 and 3.43. BOU has fully recovered with an index close to one in 2016. Among the five others, four have a relatively low index, ranging from 1.17 to 1.40. Only PRE, with an index that evolved from 3.66 in 1969 to 3.43 in 2016, has almost not recovered at all.

As shown by the annual percentage reduction in the original surface area, the recovery rate ranges from 0.12 to 1.99% yr^{-1} (AUD: 1.72%, BOU: 1.99%, BRU: 1.28%, SFL: 1.56%, STH: 1.52%, PRE: 0.12%). If we exclude PRE, the range is narrower, 1.28 to 1.99% yr^{-1}. As indicated by the very high coefficients of determination of the linear regressions between the percentage of reduction with respect to the original surface area and the time that has elapsed since the end of gravel mining, vegetation encroachment evolves mainly at a linear rate (Figure 5B).

Figure 5 – Changes in the width index (ratio of the width of the PC to the width of the control reach) (A) and percentage of the original surface area occupied by riparian vegetation (B).

Nevertheless, the width index alone is not sufficient to characterise the degree of planimetric adjustment. Although the Cher River is a single thread river, Fig. 6 shows that in 2016, all the sites are occupied by two or more channels separated by relatively big islands at least for part of the year. However, in the four sites (AUD, BOU, SFL, STH) that have recovered the most (based on changes in the width index), a main channel was clearly visible in 2016. The other branches, currently functioning as secondary or abandoned channels, are in the process of being filled and encroached by woody riparian vegetation.

Figure 6 – Extent of terrestrialization in 2016.

Overall, the longitudinal pattern of terrestrialization in 2016 mimics that of the longitudinal variability of the original width (Figure 7 and Figure 8). Most of the 2016 terrestrialized surfaces are located where the original width exceeds a threshold value roughly equal to twice the river bed width along the control reach. Table 3 shows that along these sections, the percentage of the total terrestrialized surface ranges from 69% to 86% depending on the PC. Lastly, except for AUD and BOU, which are the two PCs that have recovered the most, and PRE, which has almost not recovered at all, the 20% downstream part of the sites are the least readjusted (Figure 7).

Figure 7 – Longitudinal pattern of the terrestrialized width from the end of mining until 2013-2016. Only the years preceding major changes are shown.

Figure 8 – Relationship between terrestrialized width in 2016 and original width.

Table 3 – Percentage of terrestrialized area and of the length of PCs located where the original width of the PCs is double the width of the reference bed upstream.

	AUD	**BOU**	**BRU**	**SFL**	**STH**	**PRE**
% of terrestrialized surface	93	96	69	82	89	95
% of PC length	71	87	38	51	46	76

Vertical adjustments

In 2011, the mean elevation of sediment accumulations (bare and vegetated surfaces) above the low-water level ranged between 0.4 and 1.2 m, corresponding to rates of between 0.009 and 0.065 cm yr^{-1}. This is much lower than the mean elevation of the floodplain, ranging from 2.5 to 4.1 m above the low-water level (Table 4). Furthermore, the accumulation rate and the rate of planimetric change (computed from bare and vegetated surfaces) in 2011 are not significantly correlated (Spearman correlation test: $R^2 = 0.1$, $p = 0.54$), indicating different trajectories of change at the different mining sites.

Table 4 – Mean elevation (m) of sediment accumulation above the low-water level in PCs (bare and vegetated surfaces) and of the floodplains in 2011 (in m above the low-water level).

	AUD	**BOU**	**BRU**	**SFL**	**STH**	**PRE**
PC	0.8	0.9	0.6	1.2	1.0	0.4
Floodplain	2.5	4.0	4.1	4.0	3.7	2.8

Concerning the longitudinal pattern, we first expected that the relative depth of the sediment accumulation would decrease downstream, and second that the higher the planimetric evolution rate and the more advanced the recovery, the smaller the downstream decrease in sediment accumulation would be. However, as indicated by the black curve in Figure 9, such a decrease in downstream accumulation was only observed in AUD and BOU, which are the two PCs that have recovered the most from a planimetric point of view (Figure 5B). In the four others PCs, accumulation increased slightly downstream, but the trend was only statistically significant in PRE (Figure 9).

Figure 9 – Longitudinal pattern of sediment elevation above the low-water level in 2011.

Grain-size adjustments

Except in PRE, where only one bar was established, and except for D_{95} in SFL, there was a decreasing downstream trend in D_{50} or D_{95} that seems to indicate that grain-size adjustment is ongoing (Figure 10). The rate of decrease was even higher when the range of grain size in the PC was wide, pointing to a probable role for the range of available grain size upstream from the PCs. What is more, the rate was very low in BRU and SFL. All sites considered, the decrease was nevertheless only statistically significant for both D_{50} and D_{95} in AUD and BOU. It was also significant for D_{50} in SFL and in STH (but only for the Wolman samples in STH).

The D_{50} ranged from very fine or fine to medium gravels in BOU, BRU, SFL and PRE (respectively 8-17 mm, 4-11 mm, 3-11 mm, 3-12 mm) and from very fine or fine to very coarse gravels in AUD and STH (respectively, 2-42 mm and 7-47 mm).

Three distinct situations can be highlighted concerning the upstream and downstream context. In AUD, bedload transport appears to be relatively active and sediment availability quite high: the density of alluvial riffles and bars upstream/downstream from the PC was high (see spacing of riffle and bars in Table 5) and the grain-size along these two reaches resembled that of the PC (Figure 10). The minimum D_{50} in the PC is nevertheless much lower than the minimum D_{50} upstream/downstream. In the case of BOU and BRU, the density of riffles and bars in the upstream reach was much higher than in the downstream reach (Table 5 and Figure 10). Furthermore, the minimum D_{50} in the PC was close to the minimum D_{50} upstream/downstream. The upstream grain-size trend in BOU and BRU nevertheless clearly differed, as it was relatively stable in BOU but underwent a marked decrease in BRU. Then in SFL, STH and PRE, bed sediment availability, and hence likely bedload transport, appeared to be very low as demonstrated by the absence or almost complete absence of bars and alluvial riffles (Table 5 and Figure 10).

Figure 10 – Longitudinal grain-size pattern (D_{50} and D_{95}) in and around PCs.

Table 5 – Mean distance (in metres normalised by the active bed width in the control reaches) between riffles and bars in the PCs and along their upstream or downstream control reaches (computed on a length equal to that of the PCs) in 2018.

		AUD	**BOU**	**BRU**	**SFL**	**STH**	**PRE**
Riffle	**PC**	5	11	7	17	7	NA
	Control reach	10	9	NA	NA	NA	NA
Bar	**Control reach**	5	5	10	NA	33	NA

4.2 Size and origin of deposited sediments

Size of the infilling sediments

The trap efficiency formula indicates that at t_0, all PCs trapped all particles coarser than medium (0.25 mm) or coarse sand (0.5 mm), depending on the PC, and consequently the entire bedload (Fig. S1). The trap efficiency for particles finer than medium silt (16 µm) was almost zero whatever the discharge considered. The trap efficiency for coarse silts (32 µm) was very low at almost all discharges. All these observations imply that at t_0, most of the sandy fraction of the suspended load was deposited in the PCs and that conversely, most of the silty fraction was flushed through.

Figure S1 – Trap efficiency from very fine silt (4 µm) to medium sand (0.25 mm). Whatever the discharge considered, the trap efficiency of particles coarser than medium sand reached 100%.

Compared with the volumes deposited in the PCs, the estimated maximum bedload volume were low in the three upstream PCs (AUD, BOU and BRU), and much higher in the three downstream ones (SFL, STH and PRE) (Table 6). This indicates that, at least in the case of AUD, BOU and BRU, a large proportion of the deposited material was transported into the PCs by uniform or graded suspension and is thus composed of silt and/or sand.

Table 6 – Original volumes of PCs, deposited volumes in 2011 in PCs in 2011, maximum bedload volumes supplied to PCs from t_0 to 2011 (m^3), and maximum volumes delivered from the riverbed adjustment upstream to PCs by lateral erosion and by possible bed degradation from t_0 to 2011.

	AUD	**BOU**	**BRU**	**SFL**	**STH**	**PRE**
Original vol. of PCs	347,050	336,050	493,400	163,500	267,500	447,250
Deposited vol. in PCs in 2011	255,900	280,950	139,500	100,000	149,400	22,600
Max. bedload vol.	13,900	19,250	18,900	1,363,050	576,050	85,800
Upstream laterally eroded vol.	81,700	59,950	20,100	700	73,350	0
Upstream degraded vol.	29,750	34,350	41,550	32,500	31,200	35,900
Laterally eroded + Degraded vol.	111,450	94,300	61,650	33,200	104,550	35,900
Ratio of maximum bedload volumes / deposited vol. (porosity 0.5)	0.11	0.14	0.27	25	7.69	7.69
Ratio of maximum bedload volumes / deposited vol. (porosity 0.1)	0.06	0.08	0.15	14.29	4.35	4.17
Ratio of laterally eroded vol. / deposited vol.	0.32	0.21	0.14	0.01	0.49	0.00
Ratio of degraded vol. / deposited vol.	0.12	0.12	0.30	0.32	0.21	1.59
Ratio of total reinjected vol. / deposited vol.	0.44	0.34	0.44	0.33	0.70	1.59

A good example of the major role played by fine sediments in the infilling of the PCs, and their geomorphic recovery, is BOU, where the percentage of sediment size lower than 1 mm, considered as the approximate limit between bedload and suspension, was roughly estimated (Figure 11). We identified two main types of deposits located in distinct areas when we examined the five longitudinal profiles surveyed along the PC. The resistivity of the deposits located on the right-hand bank in the upstream half of the PC (P4 ERT profile) was very high (median 1435 Ω m) and the proportion of sediments < 1 mm was relatively low (35%). In the rest of the PC, resistivity decreased considerably (median between 101 Ω m and 182 Ω m for P10, P11, P12 and P13) and the proportion of sediments <

1 mm increased significantly (70%, 58%, 71% and 61% for P10, P11, P12 and P13, respectively) (Figure 12). Finally, particles < 1 mm were estimated to account for 56% of the total deposited volumes in 2011.

Figure 11 – Relation between resistivity and percentage of sediment volume < 1 mm in BOU.

Figure 12 – Electrical resistivity tomography in BOU.

Upstream delivery

Only a relatively small fraction of the deposited volumes can have originated from vertical or lateral erosion of the upstream reach. The hypothesis of a PC being infilled because of a massive sediment supply from headward erosion can very likely be rejected because the comparison of the low-water level upstream from the PCs from the end of the 1980s / beginning of 1990s until the early 2010s was clearly stable, except for SFL1 located 4 km upstream from the entrance of SFL (Fig. S2). In PRE, mining ended in 1969 so the vertical evolution of the bed for the first 20 yr following mining could not be characterised. A bias associated with the comparison of low-water levels is nevertheless possible, as a slight bed degradation (up to 0.1-0.2 m) might have not been detected using this method and/or the few available points of comparison might not be representative of the vertical changes to the riverbed upstream from the PCs. If we consider a 0.2 m bed degradation upstream from the PCs over a river reach equal to 60 times the bankfull width (3 km for AUD, 3.6 km for the other PCs) between t_0 and 2011 and if we consider that the whole degraded volume was trapped by PCs over the study period, we obtain a very low ratio between the degraded volume and the deposited volume (Table 6). Only PRE might have been filled by sediment originating from bed degradation, mainly because its deposited volume is very small. The values of the ratios are nonetheless overestimated because the deposited volumes in PCs are underestimated and the degraded volumes are probably also overestimated. A similar conclusion can be drawn concerning lateral erosion in the upstream reach (Table 6). The volumes involved are indeed much smaller than the infilled volumes. Furthermore, whereas we considered that all the laterally reinjected volumes reached the PCs, in fact, part of the volume is very likely stored in the riverbed upstream from the PCs (Table 6), as indicated by the many newly created areas colonized by vegetation upstream from the PCs between t_0 and 2011, especially in AUD, BOU, BRU and STH. Taking the degraded and the laterally eroded volumes together, the findings remain the same (Table 6).

Figure S2– Changes in the low-water level between 1989 and 2016 in the reach located upstream from the gravel pits (Minimum monthly discharge: 3.4 $m^3 s^{-1}$ for AUD and 7.1 $m^3 s^{-1}$ for the other sites).

4.2 Controlling factors

The overall planimetric recovery rate was highly and significantly correlated with the deposited volume / original volume ratio ($R^2 = 0.89$; p-value = 0.0048), thus confirming our initial assumption of a recovery rate controlled by the original volume of PCs, by the volume of sediment supplied from upstream and by the trap efficiency of the PCs.

Figure 13 – Relationship between 2011 planimetric recovery rate and the ratio between the deposited volume in 2011 and the original volume.

5. Discussion

5.1 The role of suspended load and bedload in the recovery rates

Our results indicate that at least in the three upstream PCs (AUD, BOU, BRU), the filling material is only partially composed of bedload sediments (Table 6). In contrast, in the three downstream PCs (SFL, STH and PRE), all the filling originated from bedload, because the deposited volumes are clearly lower than the bedload supply computed using Recking's formula (Table 6). This observation nevertheless seems quite paradoxical given the apparent limited bed sediment availability upstream from these PCs, as indicated by the total absence of bars and alluvial riffles (Figure 10 and Table 5). This absence does not seem to be the expression of the progression of a headward erosion front from PCs since the beginning of mining because bars are also rare in aerial photos acquired before sediment mining began. This is evidence for strongly limited transport conditions and overestimation of the maximum bedload volumes obtained from grain-size in the PCs. The paradox is even more striking when we compare the bedload volumes in the three upstream and the three downstream PCs. In SFL, STH and PRE, the maximum bedload estimations are respectively between 88 and 163, 27 and 50, and 2 and 5 times higher than in the three upstream PCs. Yet, in contrast to the downstream PCs, the reaches located immediately upstream from the three upstream PCs show a relative abundance of bars and alluvial riffles, suggesting active bedload transport (Figure 10 and Table 5). Expressed in terms of specific transport rates, the bedload was between 0.52 t km² yr^{-1} and 0.58 t km² yr^{-1} for the three upstream PCs whereas it was 46.7 t km² yr^{-1}, 13.9 t km² yr^{-1} and 1.3 t km² yr^{-1} for SFL, STH and PRE, respectively. The values for the three upstream PCs and PRE are in agreement with those provided by Houbrechts et al. (2008) for 16 rivers flowing in a relatively similar climatic and geological context, which range from 0.3 t km² yr^{-1} to 2.8 t km² yr^{-1} (periods: 2-31 yr, watershed areas: 0.26-2691 km², bankfull specific stream power: 17-140 W m^{-2}). In contrast, the bedload rates for SFL and STH are extremely unlikely, because they are close to the highest rates reported worldwide, and exceed several tens of t km² yr^{-1} (e.g., Powell et al., 1996; Rickenmann, 1997; Ziegler et al., 2014). If we apply a rate of 0.6 t km² yr^{-1} to SFL and STH, i.e., the maximum estimated rate amongst the three upstream PCs, we obtain bedload volumes between 2.9 and 5.4 times lower than the deposited volumes. All the above-mentioned considerations suggest that the filling material in SFL and STH is only partially composed of bedload sediments. Lastly, only PRE could have been completely filled with bedload because if we apply the same correction as the one we used for SFL and STH, the deposited volume is 2-3.6 times lower than that of the bedload. Nevertheless, considering that the specific stream power upstream to PRE is the lowest amongst the six PCs and that the maximum D_{50} in PRE is 3 mm, it can reasonably be hypothesised that the deposited volume is partly composed of suspended load. Finally, the importance of the suspended load in the recovery potential of PCs is emphasised by assuming 56% of the deposited volume in BOU in 2011 was made up of particles < 1 mm, as estimated from the combination of geophysical surveys and grain-size analysis (Figure 11 and Figure **12**). If we consider this grain size as a rough limit between bedload and suspended load processes, this implies that only 44% of the infilled volume in BOU can be attributed to bedload transport.

5.2 The influence of the sand load on the recovery processes

Whatever the PC considered, a large proportion of the deposited volume comprised of particles < 2 mm is hypothesised to be sand**Error! Reference source not found.**. At t_0, the trap efficiency for silt would be almost null for particles smaller than medium silt and very low for coarse silt**Error! Reference source not found.**. Furthermore, because of the progressive infilling of the PCs, trap efficiency in submerged areas decreases over time (Verstraeten and Poesen, 2000). As long as a PC remains submerged, silt is highly unlikely to be deposited. However, once the deposits emerge at low flow and have begun to be encroached by vegetation, the trap efficiency for suspended load is hypothesised to become significant (Corenblit et al., 2007; 2020). Trap efficiency on these elevated vegetated surfaces should thus result in a thick layer of silty material, transported as a uniform suspended load in the upper part of the water column. Yet, comparing the elevation in the 2011 LiDAR data and the DGPS data acquired in 2019 along the ERT survey sections in BOU on surfaces colonized by vegetation in 2011 shows almost complete absence of aggradation and thus probably a low silty load in the Cher River. For this comparison, we selected LiDAR points located less than 20 cm from DGPS points (giving a total of 526 points). Our comparison shows that for 75% of the paired points, the absolute difference in elevation is lower than the total error in elevation measurement (equal to 0.096 m and computed as proposed by Taylor (1997) when different components of a total error are considered as independent and are assumed to follow a Gaussian distribution: $\sqrt{E_{DGPS}^2 + E_{LiDAR}^2}$, with E_{DGPS}^2 the DGPS vertical error and E_{LiDAR}^2 the LiDAR vertical error equal to the mean absolute difference from control points). Moreover, the difference in 95% of the paired points is less than 0.2 m. Most of the suspended material and hence a large proportion of the total volume deposited in PCs is consequently hypothesised to be mainly composed of sand. Transported as a uniform suspension for its finest grain-size fraction, or as a graded suspension or as bedload (at low discharges) for its coarsest fraction, sand is therefore hypothesised to be the main determining factor in the morpho-sedimentary recovery potential of PCs in the Cher River. Furthermore, the role of sand in the recovery process probably goes even further because, at least up to a certain threshold (40% in Miwa and Parker, 2017, 50% in Dudill et al., 2018), the sand content increases gravel mobility, which, in turn, increases the bedload rate (e.g., Jackson and Beschta, 1984; Iseya and Ikeda, 1987; Wilcock et al., 2001; Curran and Wilcock, 2005; Miwa and Parker, 2017; Perret et al., 2018; Dudill et al., 2018). Such an effect is especially important in gravel-bed low-energy rivers as most of bedload very likely occurs under partial transport conditions, namely with shear stress barely higher than the critical shear. This implies that the role of gravels in the PCs recovery rates increases with an increase in the proportion of sand in the subsurface layer. For example, compared to a river deprived of, or with a very low sand content, the gravel mobility in the Cher River has very probably risen because the sand content is between 22% and 28% (Dépret, 2014). Broadly, the volume of sand load can be considered as a major control of the geomorphic recovery capacity of low energy gravel-bed systems.

5.3 The need to account for the lithology when assessing recovery potential

In addition, the role of the sand load in the Cher River PCs recovery highlights the importance of watershed lithology in recovery potential. As exemplified by Macaire et al. (1997, 2001, 2002, 2013), the amount of sand flux is strongly controlled by the lithological composition of the watershed. Macaire et al. (1997, 2001, 2002) studied the Holocene sediment yield and flux in two small lithologically contrasted basins in the Loire watershed, considered to be representative of the two main lithological units that make up the Loire watershed (like the Cher River), the Massif Central upstream and the Parisian Basin downstream. Macaire et al. (1997, 2001) showed that the total specific sediment yield and the total amount of sediment flowing out of the Chambon watershed, located in the Massif Central and mainly composed of granite, metamorphic and volcanic rocks, are largely dominated by a solid load at the expense of a dissolved load. The opposite was reported for the Négron watershed, located in the Parisian Basin and mainly composed of Cenomanian sand and marl, Oxfordian limestone and marl as well as of Turonian chalk (Macaire et al., 2001, 2002). Extrapolating their results to the whole Loire basin, Macaire et al. (2001) estimated that 91% of the total solid yield originated from the Massif Central. The remaining 9% would thus have been produced in the sedimentary context of the Parisian Basin. The minor role played by the Parisian Basin in the sediment delivery is supported by the very low proportion of sand originating from sedimentary rocks in the path of the Loire River through the Parisian Basin (Macaire et al., 2013). This means that if the lithology of the Cher watershed had been dominated by limestone, especially in its upper part, the sand load, and hence the recovery rate of the PCs, would have been much lower. It also implies that the recovery capacity of low-energy gravel-bed rivers increases with an increase in sand content and thus unquestionably depends on the geological context. Finally, it stresses the need for low-energy rivers to include watershed lithology when designing a recovery plan that incorporates morpho-sedimentary aspects. The same maintenance or restoration practices can indeed lead to significantly different results depending on the watershed lithology.

5.4 The role of PCs in habitat diversity

Shifting riverscape mosaic processes are known to trigger rejuvenation of riparian successions associated with pulse disturbance and their conservation or restoration is an issue in biodiversity conservation (e.g., Stanford et al., 2005; Tockner et al., 2010; Hauer et al., 2016; Diaz-Redondo et al., 2018;). The Cher River has been subject to longstanding anthropogenic inhibition of lateral mobility that has strongly limited such processes (Dépret et al., 2015, 2017; Vayssière et al., 2016, 2020). Moreover, its riparian forest, currently composed of a thin discontinuous strip, is in a remnant state because of the pressure exerted by farming and pasture in the floodplain. What is more, the riparian forest is also threatened by the drop in the alluvial water table caused by the bed incision (up to 2.5 m in some reaches) that followed gravel mining (Dépret, 2014). In this context, PCs play a paradoxical role. Drivers of the bed incision, and thus partly responsible for the deterioration of riparian forests, PCs are nevertheless currently transient hotspots for riparian diversity, in particular, by allowing pioneer vegetation to encroach. This is apparent when one compares the surface of newly deposited bars available for vegetation colonization inside the PCs and outside. The latter data are available between BRU and Vierzon for the period 1959-2016. Along this reach, the area of deposited surfaces colonized by vegetation in the PCs (BRU, SFL, STH, PRE and two others not included in the present study) represents 35% of the total area of deposited surfaces along this reach, whereas their course occupies only 10% of the total length of the river. Managers should thus pay particular attention to these woody pioneer habitats located inside the PCs. Nonetheless, their ecological benefits could be short lived because they are not able to self-maintain. With the progressive aggradation of their surfaces, which reduces stresses caused by flooding and increases the depth of the water table, the pioneer habitats will inevitably evolve towards the mature stage dominated by hardwood species similar to those composing the remnant riparian forest in the floodplain. The question remains of when this succession will occur. The almost complete absence of aggradation observed in BOU between 2011 and 2019 on surfaces occupied by pioneer vegetation (reported in Section 5.2) as well as the thinness of the layer of silty deposits on these surfaces suggest that the establishment of a hardwood forest could take at least a few decades starting now. A similar issue of sustainability applies to the aquatic habitats. PCs also increase their heterogeneity because of the creation of abandoned channels and to the diversification of geomorphic units (riffles, pools and bars), especially in the three downstream PCs surrounded by reaches that are almost completed deprived of such units (Fig. 10 and Table 4).

6. Conclusion

This study investigated the recovery of a low energy gravel-bed river more than 20 yr after gravel mining ended. Our specific objectives were to describe and quantify the morpho-sedimentary readjustment of extended and over-widened mining pits (we named pseudo-captures) and to identify the factors that control their recovery rate, which we defined using criteria of perennial vegetation encroachment. Thanks to an interdisciplinary methodological approach, we identified a wide spectrum of recovery magnitude closely correlated with the ratio of the volume of sediment deposited in pits to their volume when mining ended. In addition, we demonstrated that the infilled material is only partly composed of bedload and consequently that the suspended load, mainly sand, plays the main role in the recovery process. Whereas the watershed is separated into two main lithological units, with crystalline and metamorphic rocks upstream and sedimentary rocks downstream, most of the sand is assumed to have originated from the upper watershed, highlighting the importance of the surficial lithology in the recovery potential. The abundance of sand explains the unexpectedly high recovery rate of some pseudo-captures despite the low energy of the river. Our results highlight the need to further improve our knowledge in the morphogenic role of sand in gravel-bed rivers. Progress is required particularly in quantifying the proportion of sand in the total solid load of rivers as well as in identifying the conditions under which it is transported as suspended material or as bedload. Measuring fluxes remains a complex task and methods to achieve it are still rare and those that exist are associated with high uncertainty particularly because of the spatio-temporal variability at which the sand-size classes flows in suspension or as bedload (Camenen et al., 2019). Lastly, from a methodological point of view, our study illustrates how our understanding of geomorphic processes can be improved by the combined use of standard geomorphological methods, which mainly provide information concerning surface changes (grain-size, topography, planimetry, etc.), and the use of geophysical methods such as ground penetrating radar or electrical resistivity tomography, which make it possible to investigate the texture and structure of sediments beneath the surface.

Acknowledgements

This research was conducted in the framework of the *Plan Loire Grandeur Nature*. It was funded by the FEDER *Bassin de la Loire* and the *Agence de l'Eau Loire-Bretagne*. The authors thank Anaëlle Vayssière and Agnès Gauthier for their help with the core interpretation. Finally, the authors would like to express their gratitude to the editor, Scott Lecce, to Lorenzo Marchi and to a second anonymous reviewer for their constructive comments and suggestions that have significantly improved this paper.

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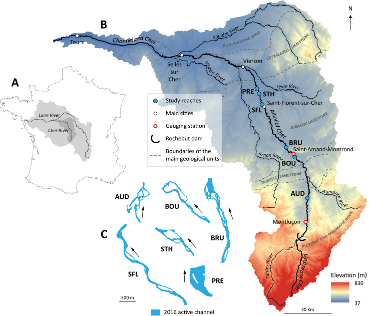
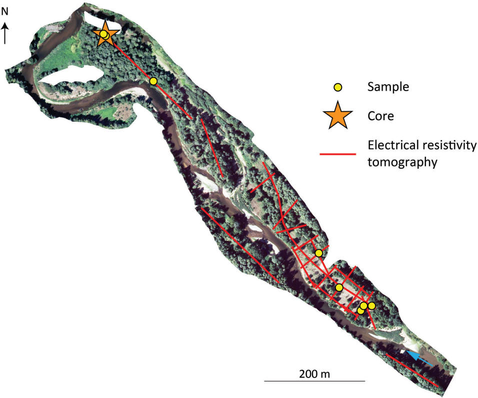
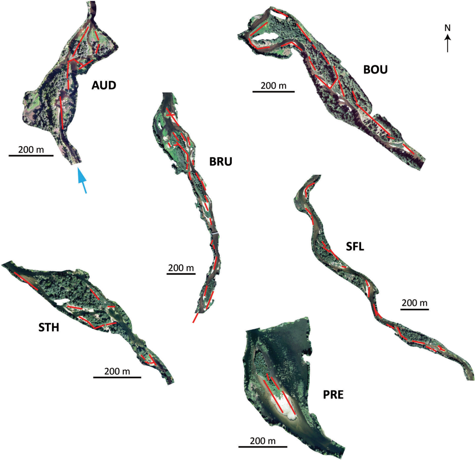
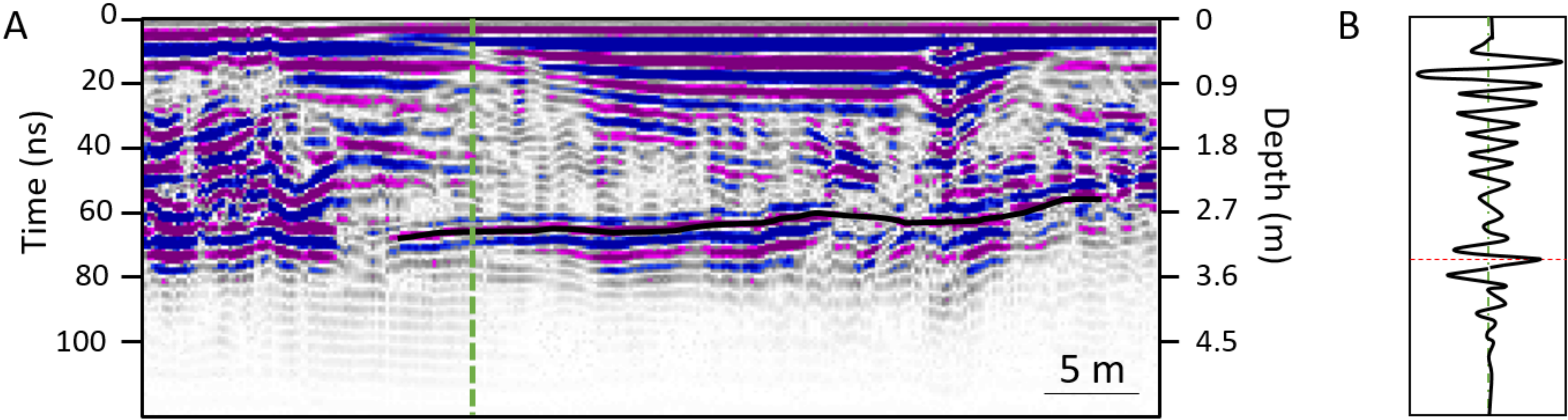
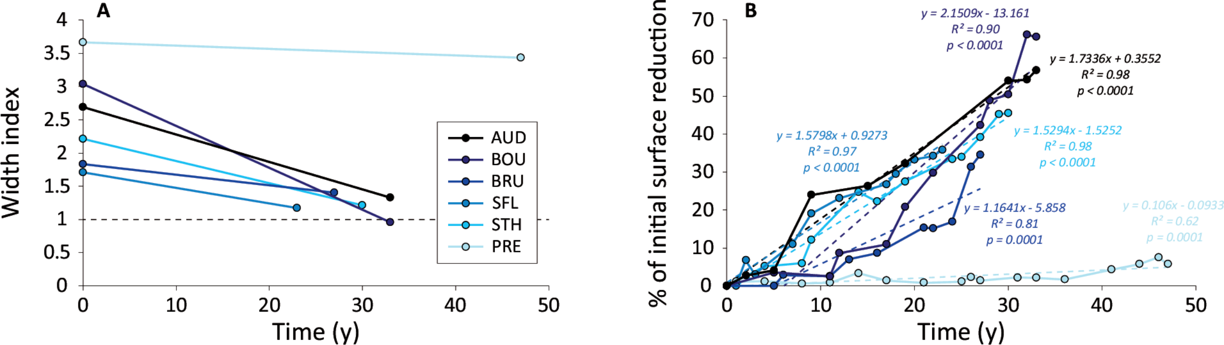
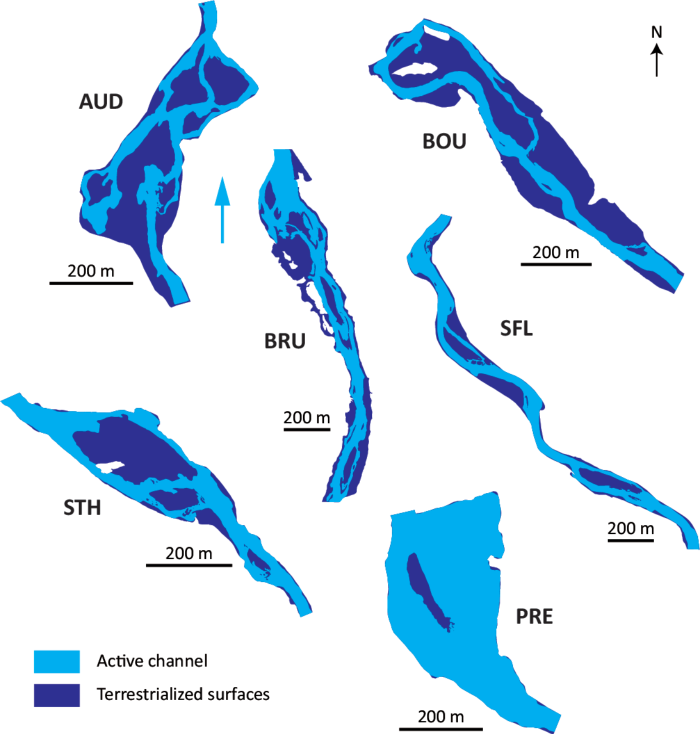

N
A
B
C
Loire River
Cher River
Tours
Channelized Cher
Sauldre River
PLIO-PLIOCENE SAND
CRETACEOUS LIMESTONE
Selles sur Cher
Vierzon
Fouzon River
Yèvre River
PRE
STH
SFL
Saint-Florent-sur-Cher
Alluvial Cher
JURASSIC LIMESTONE
BRU
Saint-Amand-Montrond
BOU
Arnon River
TRIASSIC LIMESTONE
AUD
Aumance River
Montluçon
PLUTONIC AND MAGMATIC ROCK
Tardes River
Upstream Cher
Study reaches
Main cities
Gauging station
Rochebut dam
Boundaries of the main geological units
Elevation (m)
830
37
30 Km
AUD
BOU
BRU
STH
SFL
PRE
300 m
2016 active channel

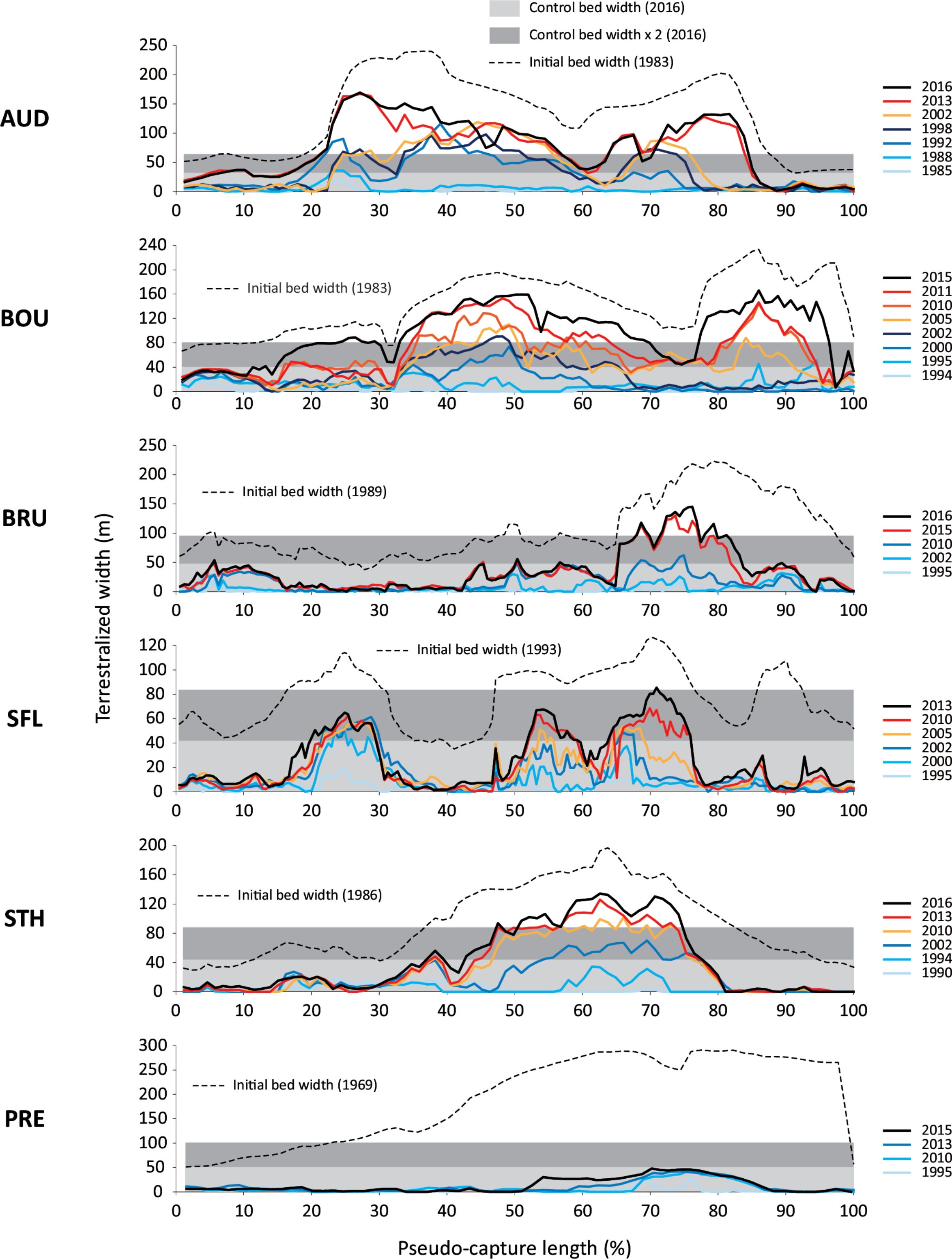

N
Sample
Core
Electrical resistivity tomography
200 m

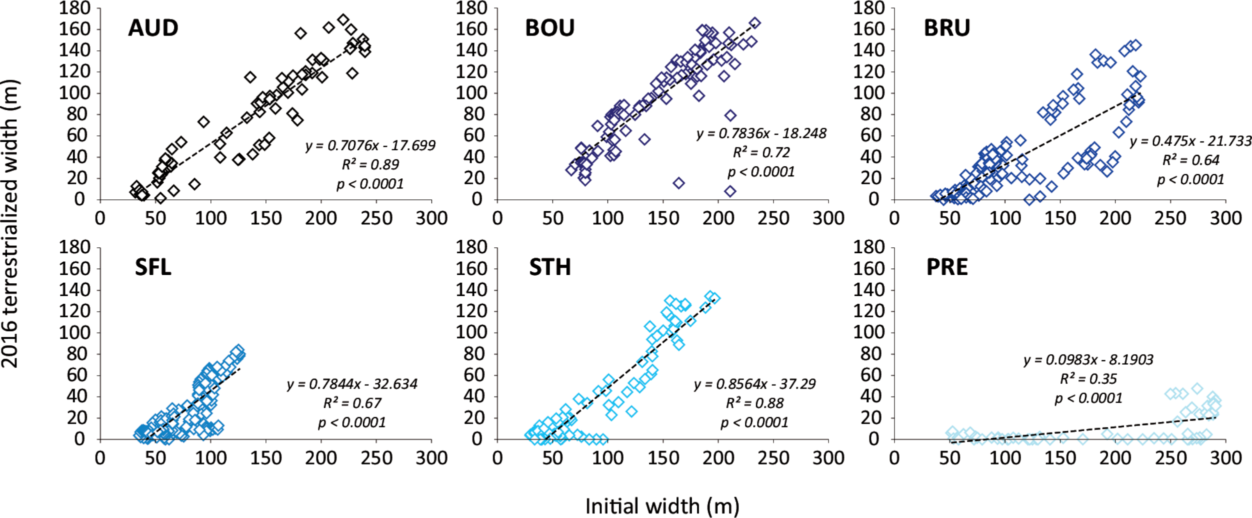

AUD
200 m
BOU
200 m
N
BRU
200 m
SFL
200 m
STH
200 m
PRE
200 m

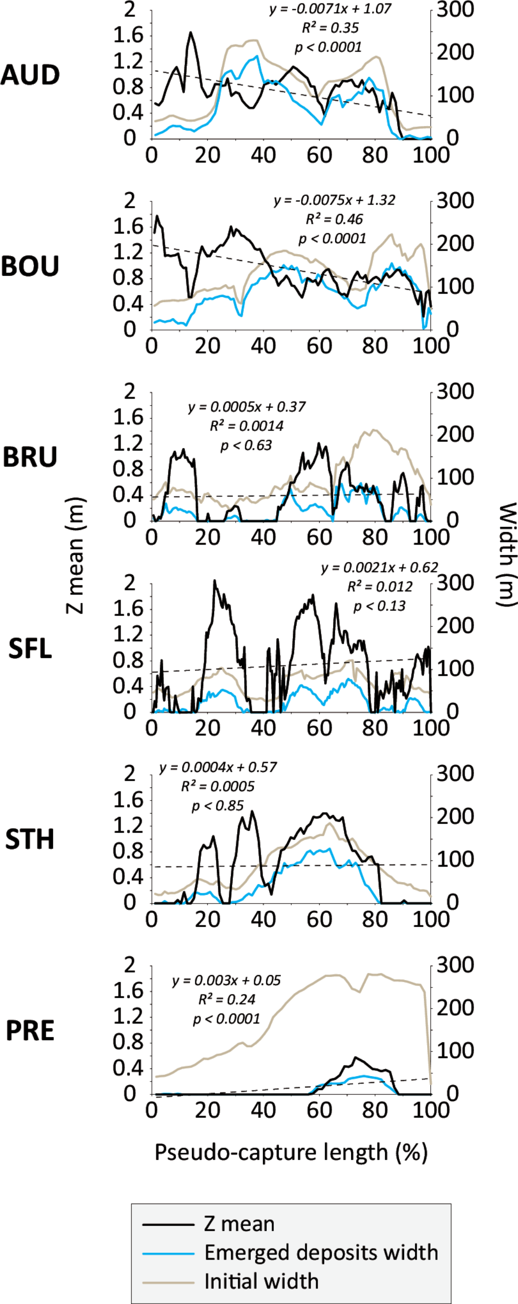

A
B
0
20
40
60
80
100
Time (ns)
0
0.9
1.8
2.7
3.6
4.5
Depth (m)
5 m

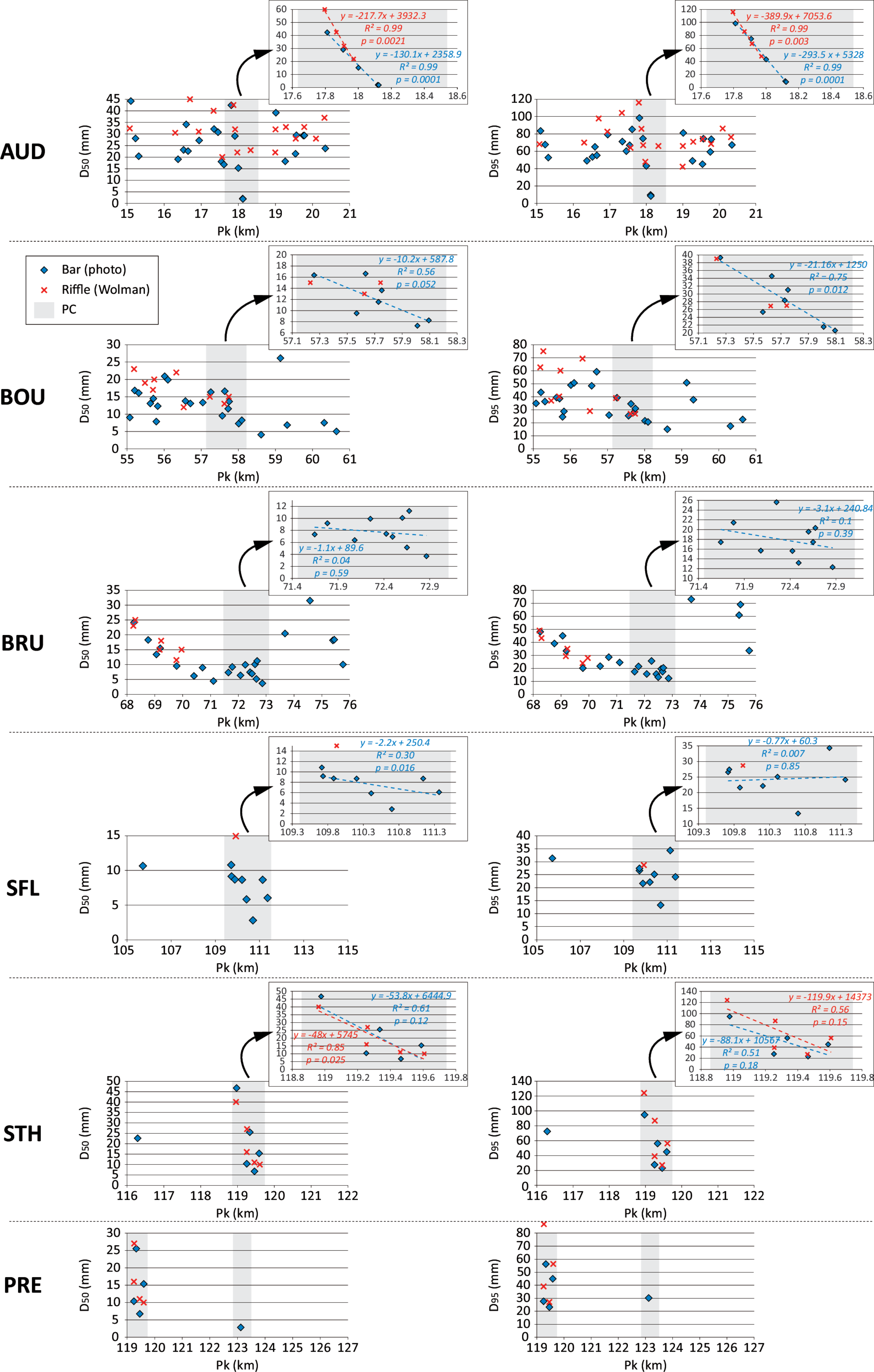

A
Width index
4
3.5
3
2.5
2
1.5
1
0.5
0
0
10
20
30
40
50
Time (y)
AUD
BOU
BRU
SFL
STH
PRE
B
% of initial surface reduction
70
60
50
40
30
20
10
0
0
10
20
30
40
50
Time (y)
y = 2.1509x - 13.161
R² = 0.90
p < 0.0001
y = 1.7336x + 0.3552
R² = 0.98
p < 0.0001
y = 1.5798x + 0.9273
R² = 0.97
p < 0.0001
y = 1.5294x - 1.5252
R² = 0.98
p < 0.0001
y = 1.1641x - 5.858
R² = 0.81
p = 0.0001
y = 0.106x - 0.0933
R² = 0.62
p = 0.0001

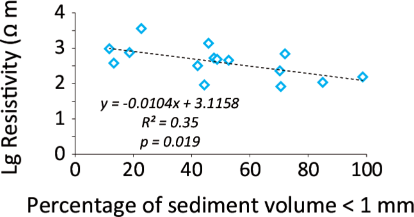

AUD
BOU
N
BRU
SFL
STH
PRE
200 m
200 m
200 m
200 m
200 m
200 m
Active channel
Terrestrialized surfaces

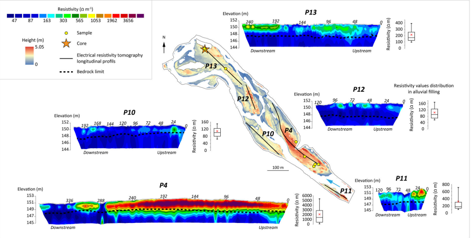

Control bed width (2016)
Control bed width x 2 (2016)
Initial bed width (1983)
AUD
2016
2013
2002
1998
1992
1988
1985
0
50
100
150
200
250
BOU
Initial bed width (1983)
2015
2011
2010
2005
2002
2000
1995
1994
0
40
80
120
160
200
240
BRU
Initial bed width (1989)
2016
2015
2010
2002
1995
SFL
Initial bed width (1993)
2013
2010
2005
2002
2000
1995
0
20
40
60
80
100
120
STH
Initial bed width (1986)
2016
2013
2010
2002
1994
1990
0
40
80
120
160
200
PRE
Initial bed width (1969)
2015
2013
2010
1995
0
50
100
150
200
250
300
0
10
20
30
40
50
60
70
80
90
100
Terrestralized width (m)
Pseudo-capture length (%)


2016 terrestrialized width (m)
Initial width (m)
AUD
y = 0.7076x - 17.699
R² = 0.89
p < 0.0001
BOU
y = 0.7836x - 18.248
R² = 0.72
p < 0.0001
BRU
y = 0.475x - 21.733
R² = 0.64
p < 0.0001
SFL
y = 0.7844x - 32.634
R² = 0.67
p < 0.0001
STH
y = 0.8564x - 37.29
R² = 0.88
p < 0.0001
PRE
y = 0.0983x - 8.1903
R² = 0.35
p < 0.0001
0
20
40
60
80
100
120
140
160
180
0
50
100
150
200
250
300


AUD
y = -0.0071x + 1.07
R² = 0.35
p < 0.0001
BOU
y = -0.0075x + 1.32
R² = 0.46
p < 0.0001
BRU
y = 0.0005x + 0.37
R² = 0.0014
p < 0.63
SFL
y = 0.0021x + 0.62
R² = 0.012
p < 0.13
STH
y = 0.0004x + 0.57
R² = 0.0005
p < 0.85
PRE
y = 0.003x + 0.05
R² = 0.24
p < 0.0001
0
0.4
0.8
1.2
1.6
2
0
100
200
300
0
20
40
60
80
100
Z mean (m)
Width (m)
Pseudo-capture length (%)
Z mean
Emerged deposits width
Initial width


AUD
BOU
BRU
SFL
STH
PRE
D50 (mm)
D95 (mm)
Pk (km)
Bar (photo)
Riffle (Wolman)
PC
y = -217.7x + 3932.3
R² = 0.99
p = 0.0021
y = -130.1x + 2358.9
R² = 0.99
p = 0.0001
y = -389.9x + 7053.6
R² = 0.99
p = 0.003
y = -293.5 x + 5328
R² = 0.99
p = 0.0001
y = -10.2x + 587.8
R² = 0.56
p = 0.052
y = -21.16x + 1250
R² = 0.75
p = 0.012
y = -1.1x + 89.6
R² = 0.04
p = 0.59
y = -3.1x + 240.84
R² = 0.1
p = 0.39
y = -2.2x + 250.4
R² = 0.30
p = 0.016
y = -0.77x + 60.3
R² = 0.007
p = 0.85
y = -53.8x + 6444.9
R² = 0.61
p = 0.12
y = -48x + 5745
R² = 0.85
p = 0.025
y = -119.9x + 14373
R² = 0.56
p = 0.15
y = -88.1x + 10567
R² = 0.51
p = 0.18


Lg Resistivity (Ω m)
4
3
2
1
0
0
20
40
60
80
100
y = -0.0104x + 3.1158
R² = 0.35
p = 0.019
Percentage of sediment volume < 1 mm


Resistivity (Ω m⁻¹)
47
87
163
303
565
1053
1962
3656
Height (m)
5.05
0
Sample
Core
Electrical resistivity tomography longitudinal profils
Bedrock limit
N
P13
P12
P10
P4
P11
100 m
Elevation (m)
Downstream
Upstream
Resistivity (Ω m)
Resistivity values distribution in alluvial filling
P13
240
192
144
96
48
0
P12
120
96
72
48
24
0
P10
192
168
144
120
96
72
48
24
0
P4
336
288
240
192
144
96
48
0
P11
120
96
72
48
24
0