

1 **Title**

2 RIVER MORPHODYNAMIC EVOLUTION UNDER DAM-INDUCED BACKWATER: AN  
3 EXAMPLE FROM THE PO RIVER (ITALY).

4 **Running Title**

5 RIVER MORPHODYNAMIC UNDER DAM-INDUCED BACKWATER

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21 **Keywords**

22 Po River, backwater, drawdown, lateral river migration, meander, gravel-sand transition.

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24 **Highlights**

25 The Isola Serafini dam interrupts the Po River continuity creating a backwater zone that modifies  
26 river hydrodynamics for up to 30 km upstream.

27 Lateral migration rates of meanders reduce downstream, and coarse-grained channel-bars are  
28 progressively drowned and reworked over time.

29 The size of river bed sediment decreases downstream across the backwater zone.

30 The size of river bedforms increases downstream across the backwater zone.

31 Dam-induced backwater forced an up-flow shift of the gravel-sand transition.

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### 33 **ABSTRACT**

34 River systems evolve in response to the construction of dams and artificial reservoirs,  
35 offering the possibility to investigate the short-term effects of base level oscillations on fluvial  
36 architecture. A major effort has been dedicated to the understanding of river response downstream  
37 of large dams, where deep channel incisions occur in response to the removal of sediment that is  
38 sequestered within the upstream reservoir. Integrating field observations and numerical modelling  
39 results, this work quantifies the sedimentary and morphological changes of the Po River (Italy)  
40 upstream of the Isola Serafini dam to investigate the impact of dam-induced backwater on river  
41 morphodynamics. The construction of a reservoir generates a new base level that forces an  
42 upstream shift of alluvial lithofacies and a change in the planform geometry of the river. The lateral  
43 migration rate of the channel is up to 45 m/yr upstream of the influence of backwater flow and ca.  
44 10 m/yr at the transition from normal to backwater flow conditions (30 km from the dam). Within  
45 this reach, a reduction of the bed shear stress promotes deposition of coarse-grained sediment and  
46 the emergence of the gravel-sand transition of the river. The lateral migration of the channel  
47 continuously reduces over time, and rates  $<5$  m/yr can be observed within the reservoir backwater  
48 zone. This trend is accompanied by the drowning of channel-bars, the reduction of river

49 competence, and an increase in bedform length. Oscillatory backwater and drawdown surface water  
50 profiles can be observed closer to the dam, which are associated with varying low and high  
51 discharge events, respectively. While low-flow conditions, persisting for much of the year, allow  
52 the deposition of fine-grained sediment, high discharge events not only promote the resuspension  
53 and transport of fine material but also the progressive erosion of channel bars and the overall  
54 deepening of the thalweg. This study provides a clear picture of the river evolution in response to  
55 the construction of a hydropower dam that may be of help in predicting how other fluvial systems  
56 will respond to future human interventions. Moreover, the result of how oscillations in base level  
57 and backwater/drawdown profiles control river hydro-morphodynamics and sediment transport may  
58 provide new insights when reconstructing ancient fluvial and deltaic sequences.

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## INTRODUCTION

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During the last few decades, increasing attention has been dedicated to understanding how fluvial systems evolve in response to changing external forcing and how such signals could be identified in the rock record (Schumm, 1981; Marchetti, 2002; Miall, 2014). Due to a better knowledge of boundary conditions, field studies from late Quaternary systems offer the possibility of investigating the signature of tectonics, climate and/or eustasy in the evolution of fluvial systems (e.g. Blum and Törnqvist, 2000). The integration of direct observations with numerical simulations, and flume experiments, has allowed for a better understanding and ability to predict how allogenic processes influence river hydro-morphodynamics and, consequently, the evolution of subaerial and submarine landscapes (Paola, 2000; Van Heijst and Postma, 2001; Edmonds and Slingerland, 2009; Jerolmack, 2009; Kim et al., 2014; Bufe et al., 2016).

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An important step forward towards more accurately modelling connections between fluvial sedimentation and stratigraphy has been the recognition of backwater and drawdown zones upstream of the river mouth (Parker et al., 2008; Hoyal and Sheets, 2009; Lamb et al., 2012), where non-uniform flows occur in response to the river stage (decelerating and accelerating river flow

75 velocities associated with low and high flow conditions, respectively). Numerical simulations and  
76 field observations have recently demonstrated that non-uniform flow exerts an important control on  
77 the evolution of sediment routing across a range of depositional environments, from the continental  
78 realm to coastal and marine areas (Paola, 2000; Moran et al., 2017). For example, depending on the  
79 channel depth and river bed slope (Samuels, 1989; Paola, 2000), non-uniform flow may persist for  
80 tens to hundreds of kilometres upstream and influence the morphological evolution of the river by  
81 controlling the position of the gravel-sand transition (Venditti and Church, 2014), the rate of lateral  
82 migration of meander segments (Nittrouer et al., 2012), the river plume hydrodynamics (Lamb et al.,  
83 2012), the formation of distributary networks and the evolution of river deltas (Jerolmack and  
84 Swenson, 2007; Chatanantavet et al., 2012), and the behaviour of morphodynamic systems (Shaw  
85 and McElroy, 2016).

86 Much of this work has recognized that the backwater profile associated with a prograding  
87 river delta promotes sediment aggradation, and consequently reduces river bed slope and sediment  
88 transport capacity which, in turn, enhances downstream fining (Paola and Seal, 1995; Wright and  
89 Parker, 2005; Ferrer-Boix et al., 2016; Franzoia et al., 2017). This phenomenon is also evident  
90 when a delta progrades into a dam-induced reservoir (Petts, 1979; Church, 1995; Snyder et al.,  
91 2004). The study of reservoir sedimentation and of the impact of dams on the evolution of fluvial  
92 channels may reveal important insights regarding morphodynamic responses of rivers to external  
93 perturbations, such as changes in the elevation of the receiving basin free surface (i.e., air-water  
94 interface), or the sequestration of sediment. In particular, a big effort has been devoted to  
95 investigating river reaches downstream of large dams (Williams, 1984; Graf, 2006), where the  
96 response of fluvial systems has been quantified by coupling observations of planform, cross-  
97 sectional and bedform changes with measurements of water and sediment discharge, with the aim to  
98 design a set of scenarios for predicting future river evolution (Brandt, 2000; Nones et al., 2013).

99 This study focuses on a 70-km-long reach of the Po River (Italy) to investigate the upstream  
100 effect of a large hydropower dam (Isola Serafini, Fig. 1) on river evolution. The basic idea is to use

101 this trunk of the Po River as a natural laboratory to investigate the short-term effects of base level  
102 change and backwater hydrodynamics on fluvial architecture, usually studied in delta distributary  
103 channels (Jerolmack and Swenson, 2007). The planform geometry and river bed elevation of the  
104 reaches upstream of the dam have been monitored over time since the dam construction, allowing  
105 investigations of the river response to such perturbations. By integrating surface observations with  
106 sedimentological data (derived from river bed sediment samples and boreholes), hydrological  
107 measurements (water elevation and discharge) and numerical simulations, the present study, the  
108 first of this genre, aims to quantify how backwater and drawdown flows influence the system  
109 morphodynamics and affect sediment partitioning along the channel axis. The results obtained are  
110 discussed in the light of understanding the impact of human interventions on fluvial systems and the  
111 influence of backwater hydrodynamics on riverine sedimentology, sediment transport processes,  
112 channel kinematics and dune dynamics.

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## **STUDY AREA**

115 The Po River arises in the Western Alps and flows roughly West-East for 652 km towards  
116 the Northern Adriatic Sea, draining an area of ca. 74,500 km<sup>2</sup> (Fig. 1). The watershed can be  
117 divided into three parts on the basis of lithology and maximum elevation: an Alpine sector of  
118 crystalline and carbonate rocks (maximum relief 4,500 m above mean sea level, amsl), an Apennine  
119 sector mostly composed of sedimentary rocks with high clay content (maximum relief 2,000 m  
120 amsl) and a central alluvial area including the Po Plain and the delta (Fig. 1, Amorosi et al., 2016).

121 The annual hydrograph shows two peaks in discharge, normally in autumn and spring,  
122 generated by rainfall and snowmelt, respectively. The mean annual water discharge recorded at the  
123 Piacenza gauging station is 959 m<sup>3</sup>/s (period 1924-2009; Montanari, 2012), while the total annual  
124 sediment and freshwater discharges to the Northern Adriatic Sea are about 13x10<sup>9</sup> kg and 40-50  
125 km<sup>3</sup>, respectively (Syvitski and Kettner, 2007; Cozzi and Giani, 2011). On decadal timescales, the  
126 water discharge variability mostly reflects a change in the precipitation pattern that bears out a

127 sharpening of the extreme events, as observed in recent years (Zanchettin et al., 2008). For the time  
128 interval investigated in this study, an important impact on river dynamics are engineering structures,  
129 including embankments for flood control, rip-rap and longitudinal groins (to constrict the channel  
130 and facilitate navigation). A direct correlation with the changes observed in the planform geometry  
131 and these anthropogenic modifications is difficult to discern (Zanchettin et al., 2008). Moreover,  
132 looking over the last five decades, there no evidence of a statistically significant change in the flood  
133 hazards along the Po River (Domeneghetti et al., 2015), indicating stationarity of the hydrological  
134 series during the evaluated period. Zanchettin et al. (2008) show that extreme discharge events for  
135 the Po River ( $4,800 \text{ m}^3/\text{s}$ ) have a return period of around 50 years, while geomorphologically  
136 effective discharges (see Biedenharn et al., 1999), considered to have a return time of  
137 approximately one year range from  $1,000\text{-}2,100 \text{ m}^3/\text{s}$  for the Po River. Over the time period  
138 investigated for this study, there is a noted decrease of the dominant yearly discharge from ca.  $2,500$   
139  $\text{m}^3 \text{ s}^{-1}$  to  $1,500\text{-}1,000 \text{ m}^3 \text{ s}^{-1}$  (Guerrero et al., 2013).

140 The Isola Serafini dam, built for hydroelectricity production and completed in 1962,  
141 interrupts the continuity of the Po River at ca. 300 km upstream of its mouth (Figs. 1 and 2). The  
142 dam has eleven floodgates (width: 30 m each) designed to control the flow through the spillway and  
143 to maintain a normal retention level of about 41.5 m amsl (Fig. 3 A), thus forcing the river water  
144 surface to oscillate between backwater and drawdown profiles during low and high flows,  
145 respectively. Depending on flood intensity, the floodgates may open so as to prevent the river from  
146 overflowing, thereby allowing the downstream discharge of water and sediment (Fig. 3 B and C).  
147 The construction of the dam affected both upstream and downstream river hydrology, morphology  
148 and, consequently, the flux of sediment, nutrients, and dissolved material to the sea (Davide et al.,  
149 2003; Surian and Rinaldi, 2003; Bernardi et al., 2013).

150 In the 250 km reach upstream of the dam (watershed of about  $45 \times 10^3 \text{ km}^2$ ), the Po River  
151 course changes from a multi-channel braided to single thread meandering, with a bed slope  
152 reduction from 1.4‰ to 0.22‰ (Fig. 1). There is an associated fining of river bed sediments from

153 coarse gravel to medium/fine sand (Colombo and Filippi, 2010; Lanzoni, 2012). The present study  
154 focuses on the 70 km of river course upstream of the dam, where the modern channel morphology  
155 is characterized by a sequence of meander bends both upstream- and downstream-skewed (Fig. 2).  
156 This portion of the river is characterized by mean bed slope of 0.215 m/km, sinuosity index of 1.82,  
157 water surface elevation between 45 m and 41.5 m amsl, mean and maximum thalweg depths of ca.  
158 6 m and 22 m, respectively, and mean and maximum channel widths of 250 m and 550 m,  
159 respectively.

160 Today, river bathymetry (Fig. 2) is the result of interactions between natural processes and  
161 anthropic activities, as the Po River has been significantly affected by human interventions during  
162 the last century (Surian and Rinaldi, 2003; Lanzoni et al., 2012). In detail, gravel and sand mining  
163 from the river bed, particularly intense in the years between 1960 and 1970, and the construction of  
164 hydropower dams in the Alpine tributaries have promoted river bed degradation and a strong  
165 decrease in the sediment load, while the creation of longitudinal embankments to prevent flooding  
166 of the surrounding alluvial plain and structures used to maintain channel navigation downstream of  
167 the Isola Serafini dam have focused flow during dominant discharge events to a single channel  
168 (Lamberti and Schippa 1994; ADBPO, 2008). The modern gravel-sand transition has been  
169 recognized between the confluence of the Po River with the Tidone and Trebbia tributaries (see  
170 Figure 2 for the location of the rivers), where a subtle decrease in slope was created by the  
171 deformation of the pre-Quaternary substrate (ADBPO, 2005).

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## **MATERIAL AND METHODS**

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### ***Field data acquisition and analysis***

175 The planform geometry evolution of the Po River within the study area is quantified by  
176 using a combination of orthophotos (years 1954, 1962, 1991) and satellite images (years 2004, 2005,  
177 2014), with the year 1954 (before the construction of the Isola Serafini dam) as a reference for the  
178 quantification of river migration rates. Orthophotos were acquired by the IGM (Military

179 Cartographic Institute) and are available online (<http://www.igmi.or/voli>). Each photo, at 1:33,000  
180 scale, is rectified as WGS84 UTM33 N in Global Mapper by using 150 control points. Satellite  
181 images came from Landsat TM (years 2004 and 2005) and Landsat 8 (year 2014), both with 30 m  
182 of spatial resolution (data available at <http://earthexplorer.usgs.gov>). The river surface area for each  
183 year is obtained through image analysis in ArcGIS® by determining the pixel values associated  
184 with water and then measuring the total water surface area. This approach may neglect small  
185 planimetric variations associated with the river stage, allowing for larger channel widths during  
186 high-flow conditions. The position of the river centerline is located by tracking the midpoint  
187 perpendicular to river banks derived from the water surface area. Therefore, the centerline is  
188 positioned with horizontal errors in the order of few meters, depending on the resolution of the  
189 images available. The lateral migration rates are derived by comparing shifts in the river centerline.  
190 All the distances presented in this study are reported as kilometres upstream of the Isola Serafini  
191 dam.

192 The thalweg depths for the years 1954 and 2000 presented in figure 2 are derived from the  
193 regional topographic surveys of the river channel and the alluvial plain executed by the  
194 Interregional Agency for the Po River (AIPO, 2005). The data, referenced to the mean sea level,  
195 consists of 14 cross sections ca. 5 km spaced, comprised between the main levees (data available at  
196 <http://geoportale.agenziapo.it>). The modern thalweg depth and the 3D views of the river bed came  
197 from a multibeam bathymetric survey performed by AIPO in the years 2004-2005 by using a  
198 multibeam echosounder Kongsberg 3002 equipped with a DGPS Racal Landstar (Colombo and  
199 Filippi, 2010). Dune lengths are calculated manually from the multibeam bathymetry by measuring  
200 the linear distance of two consecutive dune crests; the values are then averaged along 300-m-long  
201 sections.

202 Synthetic borehole logs are derived from the core repository of the Emilia-Romagna Region  
203 and Lombardia Region (data available at <http://ambiente.regione.emilia-romagna.it> and  
204 <http://www.territorio.regione.lombardia.it>, respectively). Samples of river bed sediment, collected



205 by using a Van Veen grab, have been acquired during two campaigns in July and August 2014, both  
206 during low flow conditions (Fig. 4). The positioning of the samples has been obtained by using  
207 Trimble Marine SPS461 GPS receiver equipped by an internet-based (via GSM) VRS RTK for real-  
208 time corrections that provides vertical and horizontal resolutions of 10 cm. Grain-size analyses have  
209 been performed with a set of sieves for the  $>63 \mu\text{m}$  fraction (from mesh no. 7 to no. 230) and with  
210 SediGraph Micromeritics 5120 for finer fractions at ISMAR-CNR sedimentary laboratory. All the  
211 samples are treated with 30% diluted  $\text{H}_2\text{O}_2$  to remove organic matter and washed with distilled  
212 water to dissolve salts. Before SediGraph analyses the samples are dispersed in sodium  
213 hexametaphosphate and flocculation has been further avoided using mechanical agitation. Grain-  
214 size distributions were interpreted using the software Gradistat and are presented via the software  
215 graphics (Blott and Pye, 2001).

216 Water surface elevation data (Fig. 4), converted in water discharge by using the rating curve  
217 presented in Cesi (2004), derive from the Piacenza gauging station (source: <http://arpa.emr.it>, see  
218 location in figure 2). The comprehensive information on the water levels measured along the Po  
219 River is reported in the Annali Idrologici ([http://www.isprambiente.gov.it/it/progetti/acque-interne-  
220 e-marino-costiere-1/progetto-annali](http://www.isprambiente.gov.it/it/progetti/acque-interne-e-marino-costiere-1/progetto-annali)).

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### *Numerical model*

223 The 70-km-long reach of the Po River investigated in the present study has been simulated  
224 by implementing a one-dimensional model from the US Army Corps of Engineers' River Analysis  
225 System (US Army Corps of Engineers, 2015). HEC-RAS is a computer model developed by the  
226 Hydrological Engineering Center (HEC) that simulates the flow of water in rivers and has been  
227 applied in the study area to simulate the impact of backwater zones on the flow velocity and  
228 associated bed shear stresses. Fundamental hydraulics handled by the model is the water volume  
229 continuity equation and energy equation (De Saint-Venant, 1871), solved using an implicit finite  
230 difference scheme (Sturm, 2010). The transport capacity for each grain-size class has been

231 computed by using the Meyer Peter Müller formula and the Van Rijn particles fall velocity (White  
232 et al., 1975; Van Rijn, 1984; Gibson et al., 2010). A two-layer approach describing the river bed is  
233 applied to simulate armouring and sediment hiding/exposure coefficients, and the resulting particle  
234 mobility has been calculated using the active-layer approach (Hirano, 1971). Water flux entrains  
235 sediment particles from the active layer composed of coarse particles, which hide the substratum  
236 (i.e. inactive layer) of finer sediment, eventually resulting in grains having a reduced mobility.

237 The starting river profile is described using the 1954 regional topographic survey of the river  
238 channel and alluvial plain (AIPO, 2005), interpolating the field measurements to have a maximum  
239 fixed spatial resolution of 300 m, necessary to avoid model instability but also to adequately  
240 reproduce the hydro-morphodynamics (i.e., celerity of the water flow and changes of the bed  
241 elevation) of this reach. With geometry, water depths have been computed assuming changing  
242 Manning coefficients between the main channel ( $n=0.035$ , quite clean bed) and the floodplains ( $n$   
243  $=0.025$ , short grass), based on previous applications (Guerrero et al., 2013; Nones et al., 2018).  
244 HEC-RAS computes cross-section averaged shear stresses (Chow, 1959; US Army Corps of  
245 Engineers, 2015), while the critical values (Table 1) are computed following the approach proposed  
246 by Shields (1936).

247 A synthetic approach describing the water discharge input at the upstream boundary has  
248 been applied aiming to speed up the simulation. This consisted in reproducing the typical annual  
249 oscillation that yields two wet and dry periods, as observed in the Po River catchment (Montanari,  
250 2012), and in comparing the numerical outcomes with the observed values. Monthly discharges and  
251 their durations at the hydrological station of Piacenza are analysed for the period 1954-2005 and  
252 used to calibrate the bed roughness. A sensitivity analysis spanning the period 2000-2005 has been  
253 performed to compare the modelling results when applying the synthetic hydrograph (1-month  
254 resolution) with the values of water discharge measured at the Piacenza station. The reconstructed  
255 hydrograph, spanning the period 1954-2005, points out an overall decreasing in the monthly  
256 discharges during wet periods, which reduced from ca.  $6,000$  to  $2,000 \text{ m}^3 \text{ s}^{-1}$  when passing from the

257 first 30 years to the last decades, in accordance with other studies regarding the dominant  
258 discharges affecting the Po River morphodynamics (Guerrero et al., 2013). In the simulation, the  
259 water level at the downstream end is maintained fixed at 41.5 m amsl in order to reproduce the  
260 operation of the gates of the Isola Serafini dam (Fig. 3). The sediment input has been simulated  
261 proportional to the flow discharge and adjusted to obtain a 4-5 m lowering of bed level in the first  
262 30 years of the simulated period (i.e., 1954-1984), in agreement with historical data and with the  
263 results of previous modelling simulations (Lamberti and Schippa 1994; Guerrero and Lamberti,  
264 2011; Guerrero et al., 2013; Lanzoni et al., 2015).

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## RESULTS

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### *River bed elevation and channel dynamics*

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A comparison between thalweg elevations measured in 1954, 2000 and 2005 highlights that the river bed elevation decreases through time in the upstream portion of the study area, between 70 km and ca. 45 km (Fig. 2). Previous studies documented how the lowering of the river bed in this area has been produced by a combination of reduced sediment supply from tributaries and subsequent river bed mining (Marchetti, 2002; Colombo and Filippi, 2010). Between 45 km and 25 km upstream the dam, the bed elevation is generally stable, with slight aggradation observed in some locations (e.g., Section 7 in Fig. 2). Between 25 km and the dam, the thalweg incision progressively increases in the downstream direction: section 14, located just downstream of the dam, shows the highest magnitude of incision, with up to 10 m of river bed degradation (Fig. 2).

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A combination of orthophotos and satellite images provide the opportunity to quantify the planform geometry evolution of the Po River after the construction of the dam, highlighting the lateral mobility of channel bars, the stability of river banks and the progressive changes in the water surface area of the channel (Fig. 5). Lateral migration of few hundred meters is recorded in the upstream portion of the river (between 70 and 30 km), where the position of the river centerline changed continuously through time (Fig. 5). In this area, the lateral migration is not accompanied

283 by an increase in channel width, which shows a constant value of ca. 180 m. A closer look at the  
284 data highlights that the centerline migration rates are lower in the upstream portion of the river (i.e.,  
285 up to 25 m/yr between 70 and 55 km), and increase to values of up to 45 m/yr between ca. 55 and  
286 30 km in the years just after dam construction (1954-1962, Fig. 6). Lateral migration rates diminish  
287 between 30 and 25 km upstream of the dam, where the river flows near the city of Piacenza, and in  
288 its final part upstream of the dam, where the thalweg maintained a consistent position (Fig. 6). In  
289 this section the channel, water surface area progressively increases, starting from 30 km upstream  
290 of the dam and progressing downstream; the surface area also increased for this section of the river  
291 over time, from 1954 to 2014 (Fig. 5). Close to the dam, in meander loop 4, the channel width  
292 attains a maximum value of ca. 550 m. Migration rates decrease over time along this 25-km-long  
293 section, reaching extremely low values in the final reaches (Fig. 6). The condition of river bank  
294 stability is detectable along the 25 km of river course upstream of the dam, which persisted during  
295 the entire period investigated, namely between 1954 and 2014 (Figs. 5 and 6).

296 The planimetric evolution has been accompanied by morphological changes of channel bars,  
297 as detected in four snapshots between 1954 and 2004/2005 (Fig. 5). The upstream portion of the  
298 river, meander loop 1, shows the most dynamic conditions, with side bars that start to accumulate  
299 between 1954 and 1962 (yellow arrows in figure 5), soon after the construction of the dam, with  
300 continued growth since then. The point bar, furthermore, is progressively reworked and reshaped,  
301 testifying to the dynamism of this reach (orange arrows in figure 5). Moving downstream to  
302 meander loop 2, a gradual drowning of the side bars can be observed (between 1962 and 1991, red  
303 arrows in figure 5), accompanied by a low degree of point bar reshaping, with banks strengthened  
304 by riparian vegetation (blue arrows in figure 5). Farther downstream, side bars disappear soon after  
305 dam construction (between 1954 and 1962, in both meander loops 3 and 4, red arrows in figure 5),  
306 whereas sandy point bars are drowned faster close to the dam: between 1962 and 1991 in meander  
307 loop 3, and between 1951 and 1962 in meander loop 4 (red arrows in figure 5).

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### *River bed sediment and bedforms*

310 River bed sediment samples, collected along the 70 km of the study area, with more detailed  
311 sampling in meander loops 1, 3 and 4 (Fig. 7), show the presence of coarse-grained material ( $D_{50} =$   
312 100 mm) only in the upstream portion of the river (along about 40 km): a pebble to cobble river bed  
313 develops with patchy sediment distribution and alternate with a more sandy bed, where the average  
314  $D_{50}$  is 0.6 mm. Pebbles and cobbles noticeably accumulate where flow velocity increases (i.e., flow  
315 thalweg), and finer grained particles are inhibited from depositing (meander loop 1 in figure 7), or  
316 at the confluence with the main tributaries entering from the south. At this location, cobbles  
317 (sourced from the Apennines) remain in place without transport by the modern Po River (ADBPO,  
318 2005). Cobbles of different lithologies are also found in a few samples farther downstream and are  
319 probably materials accumulated during the operations for streambank stabilization. Approaching  
320 meander loop 3, sediment samples are almost entirely characterized by sand with a  $D_{50}$  between 0.5  
321 mm and 0.3 mm. Finer deposits, with high clay content, are found in low-velocity zones, for  
322 example on the inner side of the meander towards the drowned point bar (Fig. 7). Farther  
323 downstream, within the final reaches of the river next to the dam (meander loop 4, Fig. 7), bed  
324 sediment becomes finer and dominated by silt-size fractions, with an average  $D_{50}$  between 0.2 mm  
325 and 0.01 mm. This trend is well highlighted by the grain size distribution of the river bed sediment  
326 sampled in the thalweg along the final 15 km (Fig. 8). A comparison between aerial views of this  
327 final river reach, acquired before and 50 yr after the construction of the dam (1954 and 2004-2005,  
328 during low flow conditions), highlights how the downstream decrease in particles size is  
329 accompanied by the drowning of the system and the disappearance of coarse-grained channel bars.  
330 The stratigraphy of the study area, as highlighted by a series of boreholes (Fig. 9), is characterized  
331 by a layer of anthrosol above thick gravel deposits, often amalgamated or alternated with more  
332 sandy or clayey sediment. The borehole data support the hypothesis of a gravel-bed river along the  
333 entire study area, even where fine-grained sediments are accumulating nowadays, and therefore the  
334 presence of coarse-grained channel bars down to the location of the dam (Fig. 9).

335 The three-dimensional views of the river bed, derived from the multibeam bathymetry, show  
336 a variety of depositional and erosional features, both active and remnant. Figure 10 provides a  
337 summary of the most common features encountered progressing along the flow direction. Where  
338 the river is characterized by shallow depths and coarse-grained material (between 70 km and 45  
339 km), the river bed is almost flat and few detectable bedforms develop (“alpha” and “beta” in figure  
340 10). Farther downstream, fields of transverse, slightly sinuous, sand dunes develop, often showing  
341 bifurcation of the crests (“gamma” in figure 10). In the proximity of the dam (“delta” in figure 10),  
342 the largest dunes are detected, with heights and lengths on the order of 1 meter and 100 m,  
343 respectively. A summary of the mean grain sizes of river bed sediment and mean dune lengths  
344 plotted with distance upstream of the dam highlights a progressive change approaching the  
345 structure: a decrease in sediment grain size and an increase in dune size (Fig. 11).

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#### *Bed shear stresses*

348 The 1D numerical model implemented in HEC-RAS simulated water surface profiles and  
349 bed shear stress for different river stage conditions (Fig. 12). At low discharge, the river exhibits a  
350 typical backwater profile of the M1 type (for mild channel slopes). The water surface slope break  
351 during low flows ( $350 \text{ m}^3 \text{ s}^{-1}$ ) occurs at a distance of ca. 20 km upstream of the dam (Fig. 12, left).  
352 At higher discharges (i.e.,  $2,100\text{-}4,800 \text{ m}^3 \text{ s}^{-1}$ , return period of 10-50 years), an accelerated M2  
353 profile can be recognized, and a drawdown of the water profile happens close to the dam (Fig. 12,  
354 left). The M1 and M2 profiles observed at different river discharges are linked to different  
355 conditions of bottom shear stress: while backwater conditions force a strong reduction in shear  
356 stress, and consequently a spatial decrease in sediment transport capacity approaching the dam,  
357 drawdown promotes enhanced shear stress for increasing water discharge conditions (Fig. 12, right).  
358 Farther upstream, the model shows that the highest bed shear stress is located ca. 45 km for a  
359 discharge of  $4,800 \text{ m}^3 \text{ s}^{-1}$  (black curve in Fig. 12, right) because of a local increase of the bed slope,  
360 followed by a strong reduction moving downstream to 30 km that continues with a low value until

361 20 km upstream of the dam. In the upper part of the study area, between 70 and 40 km, the model  
362 gives values of the shear stress lower than the critical shear stress for gravel-size particles that  
363 reflects scarce mobility during low flow conditions. In this trunk of the river, field observations  
364 show the presence of cobbles and pebbles alternated with sandy beds (Fig. 7).

365 In the time interval investigated in this study, a decrease of the yearly dominant discharge  
366 from ca.  $2,500 \text{ m}^3 \text{ s}^{-1}$  to  $1,500\text{-}1,000 \text{ m}^3 \text{ s}^{-1}$  has been observed (Guerrero et al., 2013). Given that the  
367 dominant (or effective) discharge is the flow that cumulates most of the channel bed sediments  
368 (Biedenharn et al., 1999), the intermediate profiles of figure 12 (green and blue) are critical for the  
369 long-term watercourse morphodynamics, as reproduced by the HEC-RAS model.

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## DISCUSSION

372 The downstream hydrodynamic changes associated with a river entering a standing water  
373 body (a reservoir, a lake or the sea) have fundamental implications for sediment transport processes  
374 and channel dynamics. In recent years, many efforts have been made to understand how, how much,  
375 and how far upstream, the presence of a standing body of water affects the behaviour of the fluvial  
376 system itself (Fernandes et al., 2016). The study of anthropogenically modified rivers entering  
377 reservoirs and of natural systems flowing into lakes and seas helped in the understanding of the  
378 response of rivers to external perturbations (Collier et al., 1996; Grant et al., 2003), the  
379 morphodynamics of deltas and estuaries (Jerolmack and Swenson, 2007; Edmonds et al., 2009;  
380 Chatanantavet et al., 2012; Lamb et al., 2012; Bolla Pittaluga et al., 2015), and the evolution of  
381 continental and coastal landscapes (Edmonds and Slingerland, 2006; Geleynse et al., 2011). Here  
382 we use a trunk of the Po River where a backwater/drawdown zone is produced by a reservoir  
383 maintained after the construction of a dam; this can be considered a natural laboratory to investigate  
384 the morphodynamic evolution under gradually-varied flow conditions, in comparison with the  
385 observations made in river systems towards their outlets into oceans (Nittrouer et al., 2012).

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### *Morphological change within the backwater zone*

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A combination of morphological and sedimentological evidence, coupled with numerical simulations, shows that the backwater zone associated with the Isola Serafini dam extends upstream for ca. 30 km; this value may also be confirmed by applying the backwater length-scale equation  $L \approx H/S$  (S: channel bed slope, H: flow depth at the river mouth; Paola, 2000) to the study area, where  $H=6$  m and  $S=0.000215$ . Upstream of backwater influence (between 70 km and 30 km, approximately), the lateral migration rates of the river are high; modelling results show that the shear stresses calculated for this sub-reach are sufficient to transport coarse sand also during moderate flow conditions ( $1,000-2,000 \text{ m}^3 \text{ s}^{-1}$ , Fig. 12). The transition from normal to backwater flow conditions is marked by a sudden drop in the rates of lateral migration for the meanders (Fig. 6) and by a progressive reduction in the size of river bed sediment (Fig. 11). Modelling results demonstrate that the water surface slope break during low flow (at  $350 \text{ m}^3 \text{ s}^{-1}$ , Fig. 12) occurs at ca. 22 km upstream of the dam, in agreement with the field observations. Along the final reaches of the Po River and, in particular, close to the Isola Serafini dam, lateral migration rates decrease to extremely low values ( $< 5 \text{ m/yr}$ ), sandy channel bars are progressively drowned, and the river erodes its bed along the thalweg (Figs. 5, 6 and 7).

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To fully understand the morphodynamics of this reach, it is necessary to evaluate the hydrodynamic conditions during both low flow and high discharge events. During low flows, with a water discharge less than  $1,000 \text{ m}^3 \text{ s}^{-1}$  that persists for much of the year, a typical M1 water surface profile is observed (Fig. 12, left). This condition generates a drop in the bed shear stress to values below the critical threshold of bed-load transport for silt-sized sediments (Table 1), and is associated with the accumulation of fine-grained material in the channel axis and along channel bars (Figs. 7 and 8). During floods and high-discharge events, conversely, the river transport capacity changes drastically: the normal retention level is maintained at a constant elevation through the opening of the floodgates, producing an M2 water surface profile and the downstream increase in river flow velocity (i.e.,  $2,100 \text{ m}^3 \text{ s}^{-1}$  and  $4,800 \text{ m}^3 \text{ s}^{-1}$ , Fig. 12, right). As shear stress



413 increases, the river becomes erosional, removing fine-grained sediment accumulated during low  
414 flow river stages and partially reworking the antecedent coarse-grained channel bed sediment.  
415 Modelling simulations highlight that the drawdown condition is reached for water discharges  
416  $>2,000 \text{ m}^3 \text{ s}^{-1}$ , and generates high bed shear stress close to the dam, which promotes coarse-grained  
417 sediment transport as bedload (Fig. 12). The condition of low flow discharge and associated  
418 backwater profiles persists for much of the year, generating the overall downstream fining of river  
419 bed sediment observed in the model simulations, while pulses of coarse-grained material are  
420 associated with episodic high discharge events, characterized by a drawdown of the water surface  
421 profile.

422         The backwater/drawdown oscillations occurring during different river stages, first observed  
423 by Lane (1957), may help explain the reduction in lateral river migration, the deep incision of the  
424 channel thalweg, the formation of armoured beds, and the partial removal of sediment accumulated  
425 along the channel bars. Similar processes have been observed in the Teshio River (Ikeda, 1989) and  
426 in the Mississippi River (Nittrouer et al., 2012). In this latter example, lateral migration rates  
427 decrease from more than 120 m/yr to less than 20 m/yr along the last 600 km upstream of its outlet  
428 (Hudson and Kesel, 2000); this has been related to the backwater morphodynamics generated by the  
429 impact of the Holocene sea level rise on the river (Nittrouer et al., 2012). In the Lower Mississippi  
430 River, a 100-fold increase in sand transport has been observed during high-discharge events  
431 associated with an increase in shear stress by a factor of thirteen, compared to low flow conditions  
432 (Nittrouer et al., 2011). A decrease in lateral migration rates (Fig. 6) favoured the growth of riparian  
433 vegetation on streambanks and channel bars (Fig. 5), promoting stabilization of the river course. In  
434 detail, the effect of vegetation on bank stability is twofold, as it increases soil strength through the  
435 network of roots and reduces soil moisture content through canopy interception and  
436 evapotranspiration (Thorne, 1990; Simon and Collison, 2002; Wickert et al., 2013). Embankment  
437 construction for flood protection, introduced during the last few decades, has further increased  
438 bankline stability. The reduction of lateral migration coupled with the deepening of the channel

439 towards its outlet may promote a narrowing of the channel belt, which could be preserved in the  
440 stratigraphic record of the fluvial deposit. This concept can be useful when interpreting ancient  
441 fluvial-deltaic systems, as low values of the width-to-thickness ratio of the channel belt, associated  
442 with limited lateral migration rate (e.g., up to few channel widths), may be diagnostic of distributary  
443 channels (Gibling, 2006; Blum et al., 2013; Colombera et al., 2016).

444

#### 445 *Sedimentological changes within the backwater zone*

446 The hydrodynamic behaviour of the Po River throughout backwater/drawdown zones can be  
447 quantified not only by considering spatial information such as lateral river migration and channel  
448 deepening and widening, but also by examining the morphological evolution of channel bars and  
449 the grain size partitioning along the channel axis. Consecutive aerial surveys of the Po River during  
450 the last 60 years, since dam-induced backwater was created, show that the point bars and side bars,  
451 continuously evolving through time between 70 km and 30 km (Fig. 5), are often accompanied by a  
452 lateral migration of the channel through cutbank erosion. Farther downstream, in particular along  
453 the 30 km of river upstream of the dam, the establishment of a backwater zone coincides with  
454 channel deepening; this arises as repeated oscillations between M1 and M2, associated with the  
455 transition between discharge conditions, progressively erodes channel bars and the thalweg (Figs. 2  
456 and 5).

457 The morphological changes associated with the variation of the river transport capacity are  
458 reflected in river-bed grain size data: coarse sediments characterize the upstream portion of the  
459 system, with a progressive downstream fining trend observed along the 30 km river reach upstream  
460 of the dam. Interestingly, the upstream limit of the backwater zone nearly overlaps with the area  
461 where the gravel-sand transition of the river is observed. Pebbles and coarse sand that characterize  
462 the river bed in the upstream portion of the study area are gradually veneered by fine sand and silt  
463 moving downstream and, close to the dam, where the flow velocity is extremely low, by clay. River  
464 bed samples from the drowned point bars in meander loops 3 and 4 (Fig. 7) show, in detail, a

465 coarsening trend towards deeper water, with clay-rich sediment accumulating at the inner point bar  
466 (Fig. 7). This scenario is in marked contrast with the morphology of both the meanders observed in  
467 1954 (before the dam), where coarse-grained point bars are subaerially exposed (Fig. 8), and the  
468 presence of a gravel-bed river can be deduced by the stratigraphy of the boreholes (Fig. 9). Based  
469 on the observations made in meander loops 3 and 4, it is possible to conclude that a fining upward  
470 trend, with more frequent mud-drapes, should characterize the vertical sedimentary succession of a  
471 point bar where the river enters the backwater zone. This finding may be of consideration when  
472 interpreting ancient river deposits at outcrop or sediment-core scales (Shanley et al., 1992;  
473 Labrecque et al., 2011).

474         The change in river hydraulics and bed material composition observed throughout the  
475 backwater zone is also reflected by a change in bedform geometry (Fig. 10), specifically, showing a  
476 downstream increase in length (Fig. 11). This trend may be connected to a combination of the  
477 general fining of bed sediments, and the occurrence of high flow velocity during high-discharge  
478 events, as both factors contribute to increasing dune size (Southard and Boguchwal, 1990). It is  
479 possible that the large-scale dunes form only during high-discharge events and remain as relict  
480 features (i.e., not in flow equilibrium) during subsequent low flow conditions; this type of pattern  
481 has been observed along the Yangtze River (Chen et al., 2012). The general trend of downstream  
482 increase in dune height, highlighted in figure 10, could also be linked to increasing water depth  
483 progressing into the backwater segment. Moreover, the multibeam bathymetric data available in the  
484 study area show that the river bed is normally mantled by dunes along straight reaches and within  
485 the inner portion of meander bends, while pools and outer banks lack bedforms and are dominated  
486 by erosional forms, particularly in the deeper portions. This evidence agrees with the observations  
487 from the lowermost Mississippi River, which indicate a lack of bedforms in meander bends where  
488 faster flows generate sediment suspension and inhibit bedform development (Nittrouer et al., 2008).  
489 Future investigations supported by repeated multibeam bathymetric surveys could help provide a

490 detailed understanding of how backwater and drawdown hydrodynamics control the evolution of  
491 the river bed, including the formation of bedforms as is coupled to downstream sediment transport.

492

### 493 *Implications for the gravel-sand transition of the river*

494 It is widely recognized that the size of bed sediments in alluvial rivers decreases  
495 downstream (Church and Kellerhals, 1978; Frings, 2008; Franzoia et al., 2017), and among such  
496 variations the transition from gravel to sand is a fundamental boundary in the fluvial  
497 geomorphology, as it marks key changes in river planform geometry. Two processes have been  
498 proposed to explain sediment downstream fining and the emergence of a gravel-sand transition in a  
499 river (i.e., grain breakdown and selective transport; Schumm and Stevens, 1973; Hoey and  
500 Ferguson, 1994; Ferguson et al., 1996; Ferguson, 2003), and several studies pointed out the  
501 influence of backwater on river capacity as a fundamental mechanism (Sambrook Smith and  
502 Ferguson, 1995; Venditti and Church, 2014). The gravel-sand transition in the Po River has been  
503 recognized between the confluences with the Tidone and the Trebbia tributaries (Fig. 2), and has  
504 been associated to a slight decrease in river bed slope due to the deformation of the pre-Quaternary  
505 substrate (ADBPO, 2005). Interestingly, modelling results highlight a strong reduction of the bed  
506 shear stress between 40 and 20 km upstream of the dam (Fig. 12), where river hydrodynamics start  
507 conveying the backwater effect.

508 Orthophotos acquired in 1954 indicate that the river, within the study area and farther  
509 downstream of the Isola Serafini dam, was characterized by the presence of large point bars, side  
510 bars and both vegetated and un-vegetated mid-channel bars (Fig. 8). Conversely, none of these bars  
511 are present in the 2004-2005 satellite image along the 30 km upstream of the dam (Fig. 8). The  
512 stratigraphy derived from the boreholes shows armoured gravel beds below the modern anthrosol  
513 across the entire study area (see ADBPO, 2005 and Fig. 9), suggesting that the channel bars in 1954  
514 (Fig. 8) were probably coarse-grained. Combining such information, it is possible to argue that  
515 before the construction of the dam, when the base level of the river was at sea level, the gravel-sand

516 transition was located kilometres downstream with respect to the modern position. Therefore, the  
517 new base level created by the reservoir forced retrogradation of alluvial lithofacies and a change in  
518 the channel pattern (Schumm, 1993), as suggested by the erosion and drowning of the channel bars  
519 (Fig. 8). In addition, the emergence of the gravel-sand transition at the upstream limit of the  
520 backwater zone, where a strong decrease in bed shear stress promotes the deposition of coarse-  
521 grained sediment, suggests the dam-induced backwater as a primary control on its location. This  
522 result adds value to the importance of the backwater effect in establishing not only the downstream  
523 evolution of the river impacting the genesis of the distributary channel network of the delta  
524 (Jerolmack and Swenson, 2007; Edmonds and Slingerland, 2009; Chatanantavet et al., 2012; Lamb  
525 et al., 2012; Bolla Pittaluga et al., 2015), but also the river morphodynamics and associated  
526 lithofacies for tens of kilometres upstream. The data presented, moreover, may be further discussed  
527 in the light of better understanding the time-response of alluvial river systems to external  
528 perturbation and how changes of the river dynamics may be preserved in the sedimentary record.

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## CONCLUSIONS

531 The combination of field observations and numerical modelling simulations show that the  
532 effect of the dam-induced backwater on river hydro-morphodynamics propagates upstream for up to  
533 30 km. The dam interrupts the river continuity and generates a new base level, forcing a  
534 retrogradation of alluvial lithofacies and a change in the planform geometry. At the transition from  
535 normal to backwater flow, a strong decrease in water surface slope and associated bed shear stress  
536 promotes the deposition of coarse-grained material and the emergence of the gravel-sand transition  
537 of the river. Along the 30-km-long reach of the river affected by backwater, lateral migration rate of  
538 the meanders progressively reduces approaching the dam and is accompanied by a general fining of  
539 river bed sediment and an increase in dune length. Closer to the dam, M1 and M2 water surface  
540 profiles are modelled in response to low-flow and high-discharge events, respectively. During low-  
541 flows, which persist for much of the year, the bed shear stress is low enough to promote deposition

542 of fine-grained sediment (up to clay size), and during high-flow events, a significant drawdown can  
543 be observed, which enhances bed shear stress and promotes the removal of the fine-grained material,  
544 resulting in the progressive erosion of the channel bars and the overall deepening of the river  
545 thalweg.

546 In essence, this study shows the evolution of the upstream portion of a river after the  
547 construction of a dam and highlights the effects of base level rise, backwater and drawdown  
548 processes in controlling fluvial hydro-morphodynamics and sediment transport processes. These  
549 concepts, applied to source to sink studies of continental margin evolution, may provide better  
550 understanding of how far upstream a change in base level may affect fluvial architecture and facies,  
551 and of how M1 and M2 oscillations in the backwater zone govern the evolution of channel belts,  
552 control sediment yield to the receiving basin and, ultimately, disperse material to depositional  
553 basins.

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789

790

791 **Figure Captions**

792 *Figure 1.* Po River course and its catchment basin, with reported the main tributaries entering the  
793 river in the study area (black square). Point B marks the location of the Isola Serafini dam. DEM  
794 derived from SRTM 90m (available at [http://www.cgiar-csi.org/data/srtm-90m-digital-elevation-](http://www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1)  
795 [database-v4-1](http://www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1)).

796

797 *Figure 2.* Top: Aerial view of the study area derived from a combination of two images (date  
798 7/21/2004 and 5/8/2005, left and right sides of the white dashed line, respectively) acquired during  
799 low flow conditions, see figure 4 for details. Blue lines (numbered from 1 to 14) represent the  
800 location of river cross sections acquired in 1954. Note that section 14 is just downstream of the dam.  
801 Red dots (named from A to V) mark the location of the boreholes presented in figure 9. Orange  
802 squares (named from “alpha” to “delta”) highlight the location of the multibeam bathymetric  
803 sections presented in figure 10. The blue star marks the location of the gauging station near the city  
804 of Piacenza. Bottom: thalweg elevation derived from the 2005 multibeam bathymetric survey (MB,  
805 red line; source AIPO, 2005), from the 1954 river cross sections (blue squares) and from the 2000  
806 river cross sections (green square). The numbers in blue refer to the cross sections, as in the map.  
807 Yellow arrows mark the location of the four main tributaries (Tidone, Lambro, Trebbia and Nure  
808 rivers).

809

810 *Figure 3.* Schematic representation of the Isola Serafini dam (A, front view), with later views  
811 during low flow discharge events (backwater river profile and the gates are closed, B) and during  
812 high discharge events (drawdown river profile and the gates open depending of the discharge, C).

813

814 *Figure 4.* Top: water surface elevation (blue) and water discharge (red) data during June-September  
815 2014; cruise time is reported below. Bottom: water surface elevation of the Po River during the  
816 acquisition of the aerial views presented in Figure 2; both images are acquired during low flow  
817 conditions (black squares).

818

819 *Figure 5.* Top: extent of the Po River surface area obtained from five aerial surveys since 1954  
820 (before the construction of the Isola Serafini dam). Bottom: close view (age reported in years) of  
821 four-selected meander loops (1 to 4 moving downstream). In detail: orange and yellow arrows in  
822 meander loop 1 indicate channel bar and point bar accretion over time, respectively; blue arrows in



823 meander loop 2 mark river banks with increasing vegetation cover; red arrows in meander loop 2, 3  
824 and 4 highlight channel bars that are progressively eroded until disappear.

825

826 *Figure 6.* Lateral migration rate of the Po River centerline, calculated in four time intervals, plotted  
827 to distance upstream.

828

829 *Figure 7.* Detailed view of meander loops 1, 3 and 4 (see location in figure 4) showing the grain-  
830 size variability of river bed sediment. Blue (sediment  $< 0.063\text{mm}$ ), yellow (sediment between  
831  $0.063\text{-}2\text{ mm}$ ) and red ( $>2\text{ mm}$ ).

832

833 *Figure 8.* Morphology of the Po River along the final reaches located upstream of the Isola Serafini  
834 dam in 1954 and 2005. River bed samples acquired in this study are represented with coloured dots  
835 and are reported on the 2005 satellite image (each colour refers to a specific sediment sample). Red  
836 dots (named from K to V) mark the location of the boreholes presented in figure 9.

837

838 *Figure 9.* Schematic representation of the lithology of the boreholes available in the area (see  
839 location in figure 2). Note the presence of extensive coarse-grained deposit for the entire section  
840 investigated.

841

842 *Figure 10.* 3-D images of the river bed derived from the multibeam bathymetric survey of AIPO and  
843 2-D bathymetric sections (red lines). The multibeam data are located in figure 2. The 3-D images  
844 are selected to highlight the bedforms pattern observed in different river reaches.

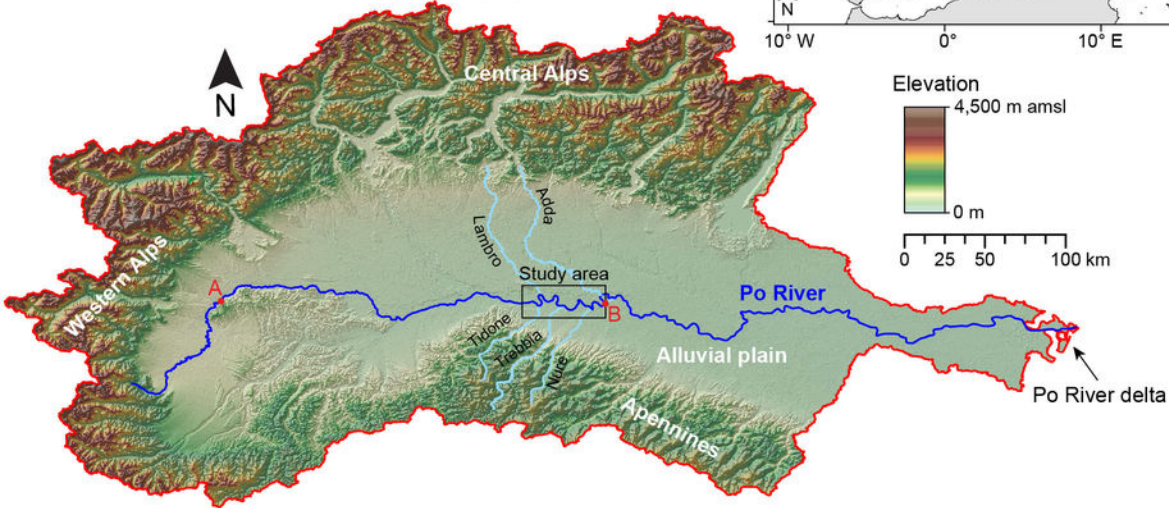
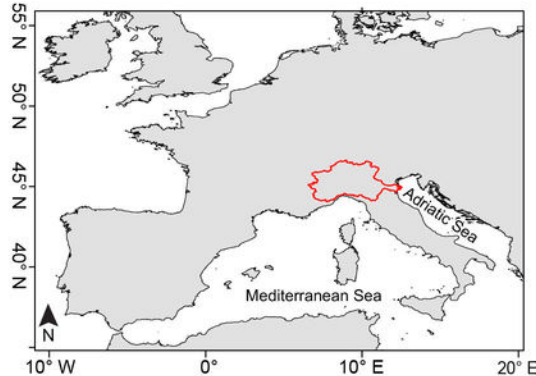
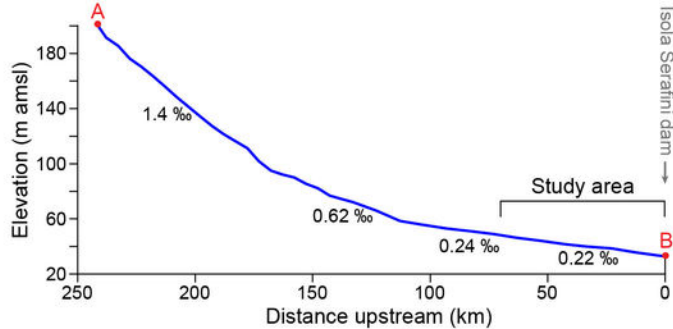
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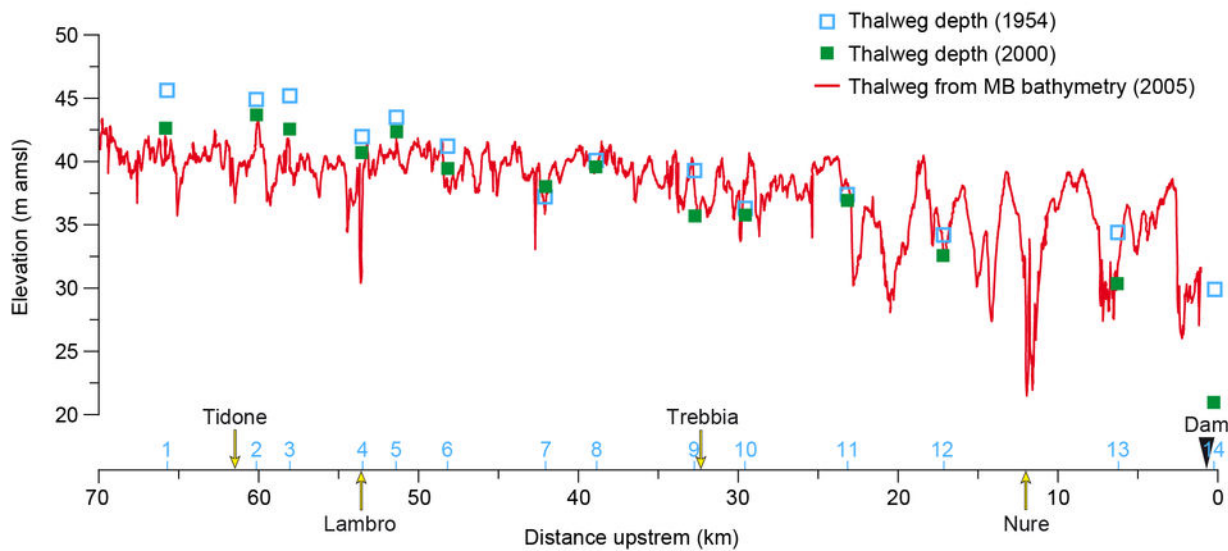
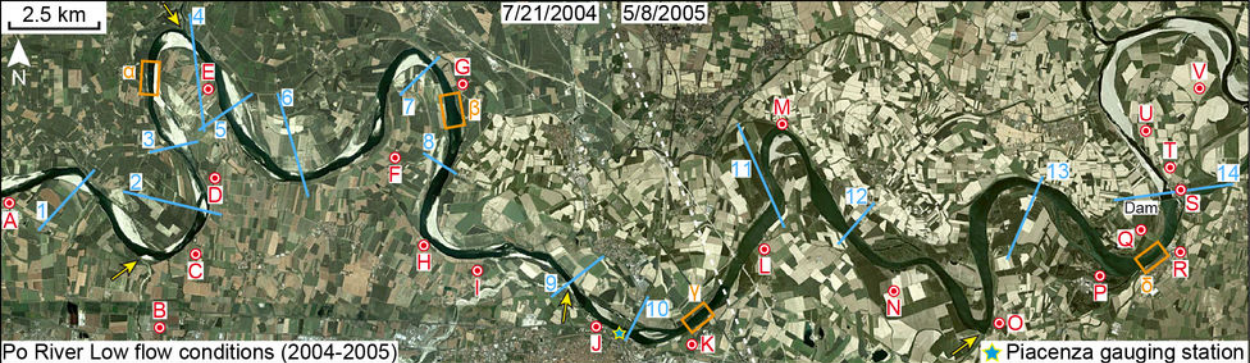
846 *Figure 11.* Variability of river bed sediment along the 70-km-long reach upstream of the Isola  
847 Serafini dam (blue dots, dimensions scale with  $D_{50}$  grain sizes). Vertical blue bars represent the  
848 range of grain sizes, while horizontal bars their spatial distribution along the river. Red diamonds

849 represent the average dune length derived from 300 m long 2D profiles extracted from the  
850 multibeam bathymetry. Vertical red bars represent the maximum and minimum dune length in each  
851 transect.

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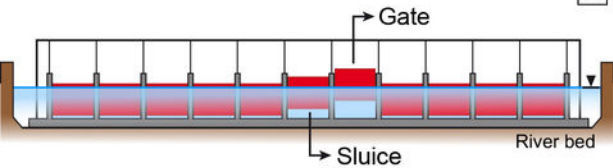
853 *Figure 12.* Simulated water surface profiles (left) and bottom shear stresses (right) calculated at  
854 different water discharges along the 70-km-long river reach of the study area.



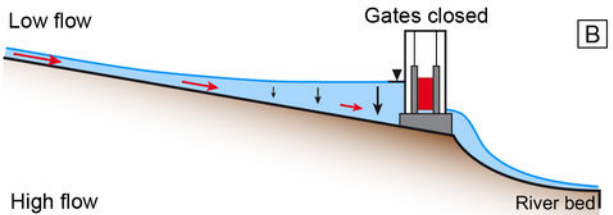


# Isola Serafini dam

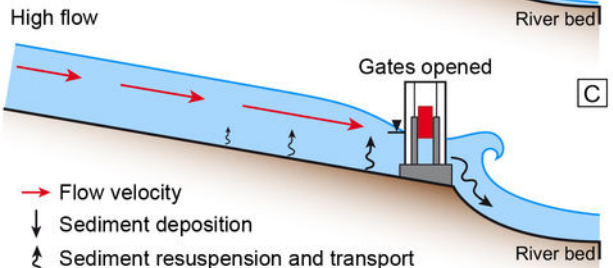
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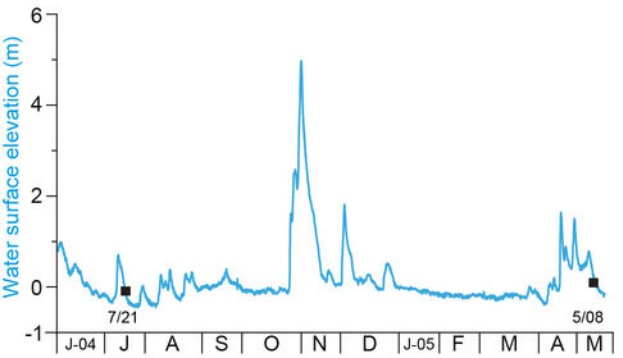
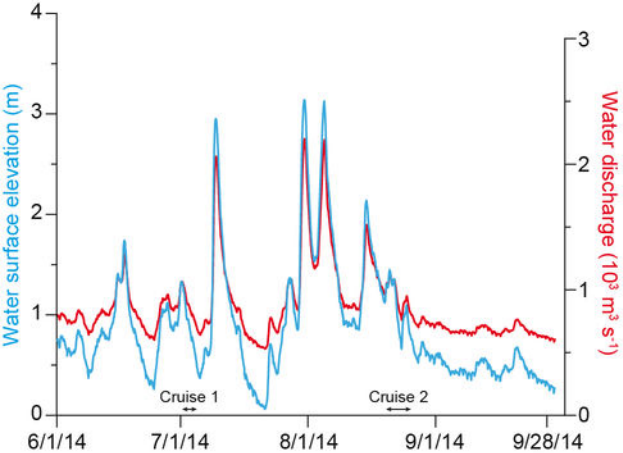
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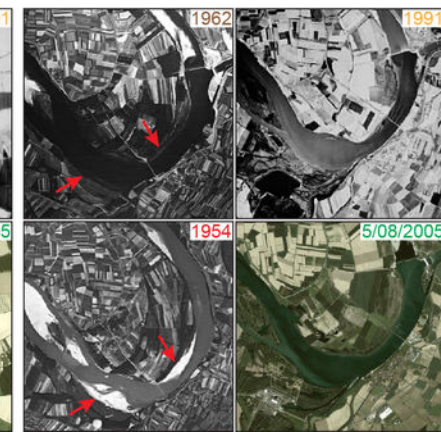
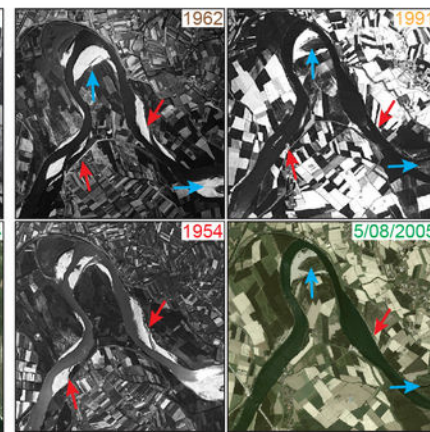
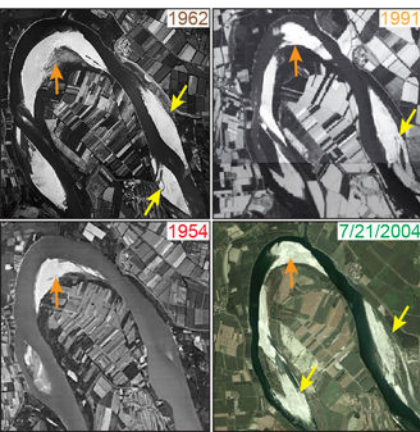
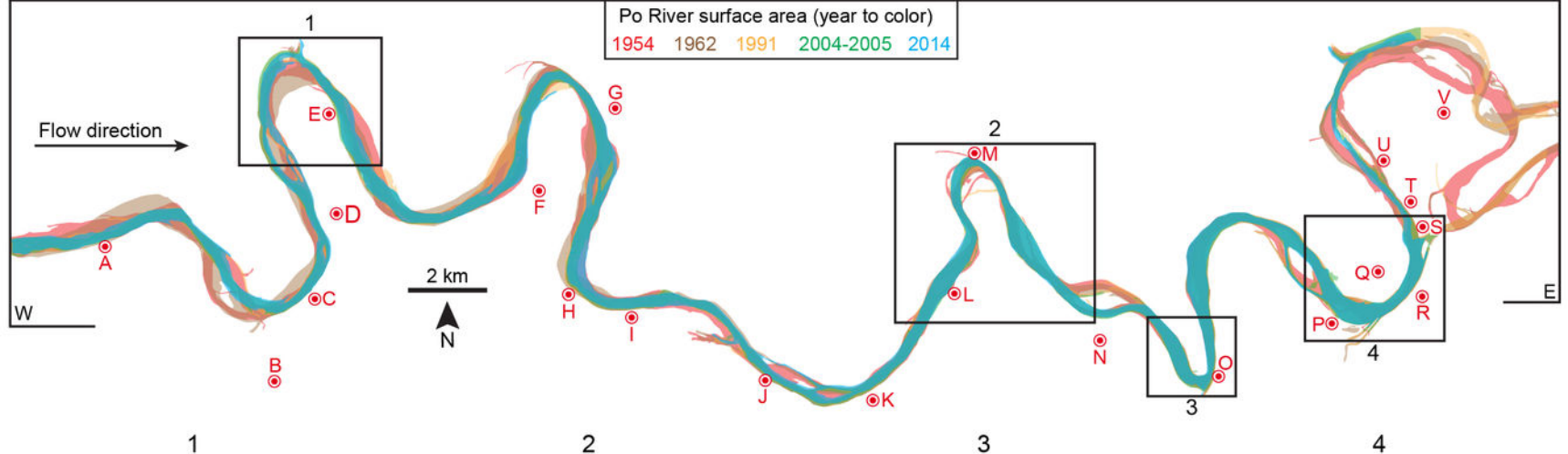


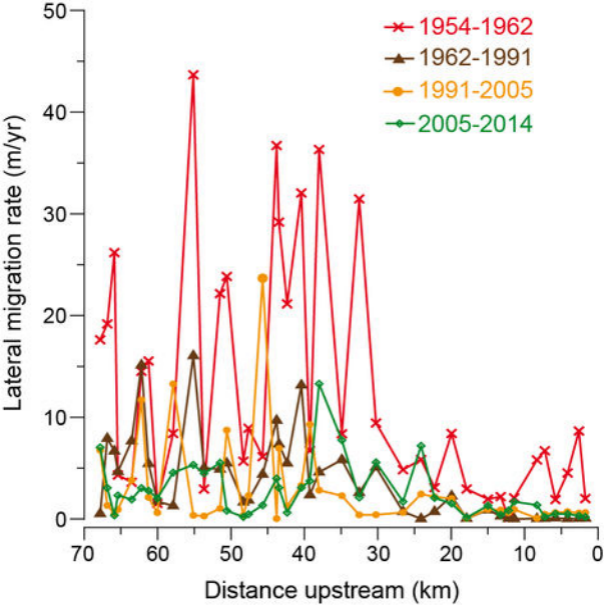
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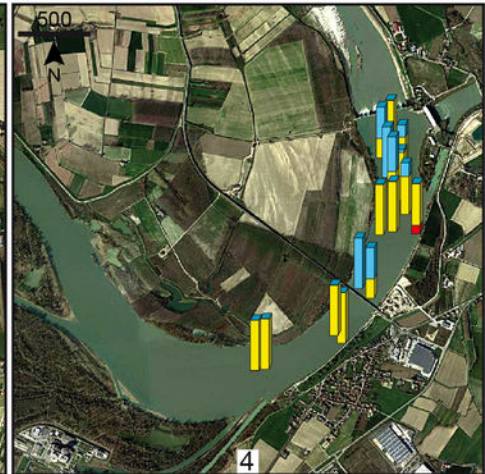
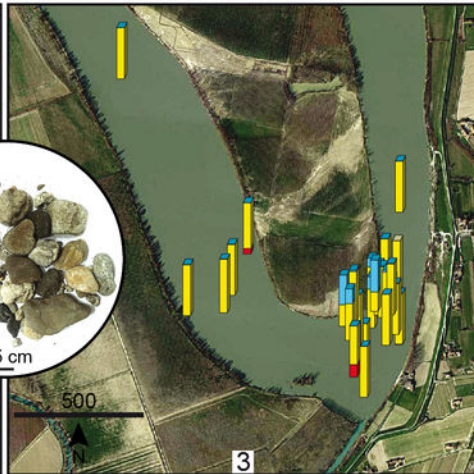
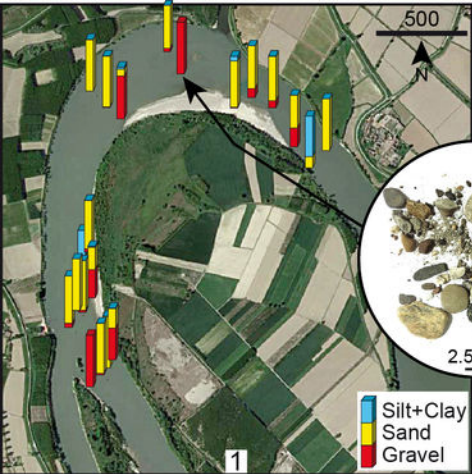
- Flow velocity
- ↓ Sediment deposition
- ⤴ Sediment resuspension and transport
- ▽ Water elevation at the dam, 41.5 m amsl

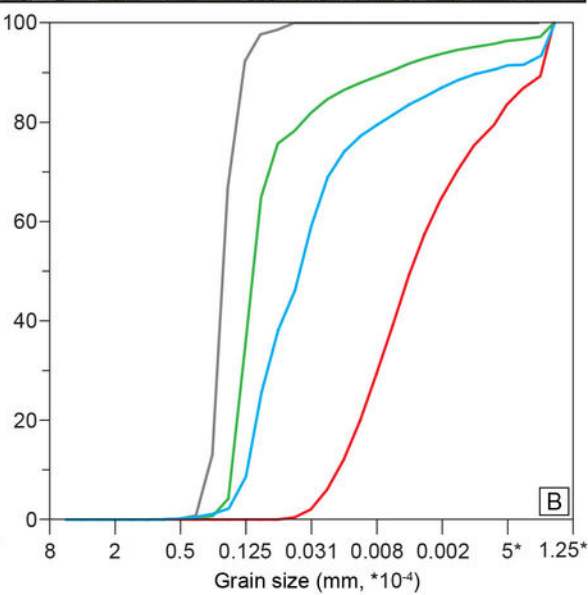
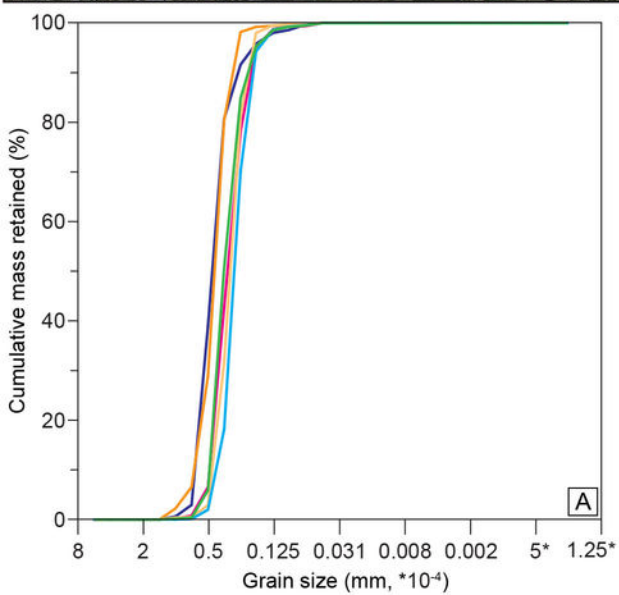
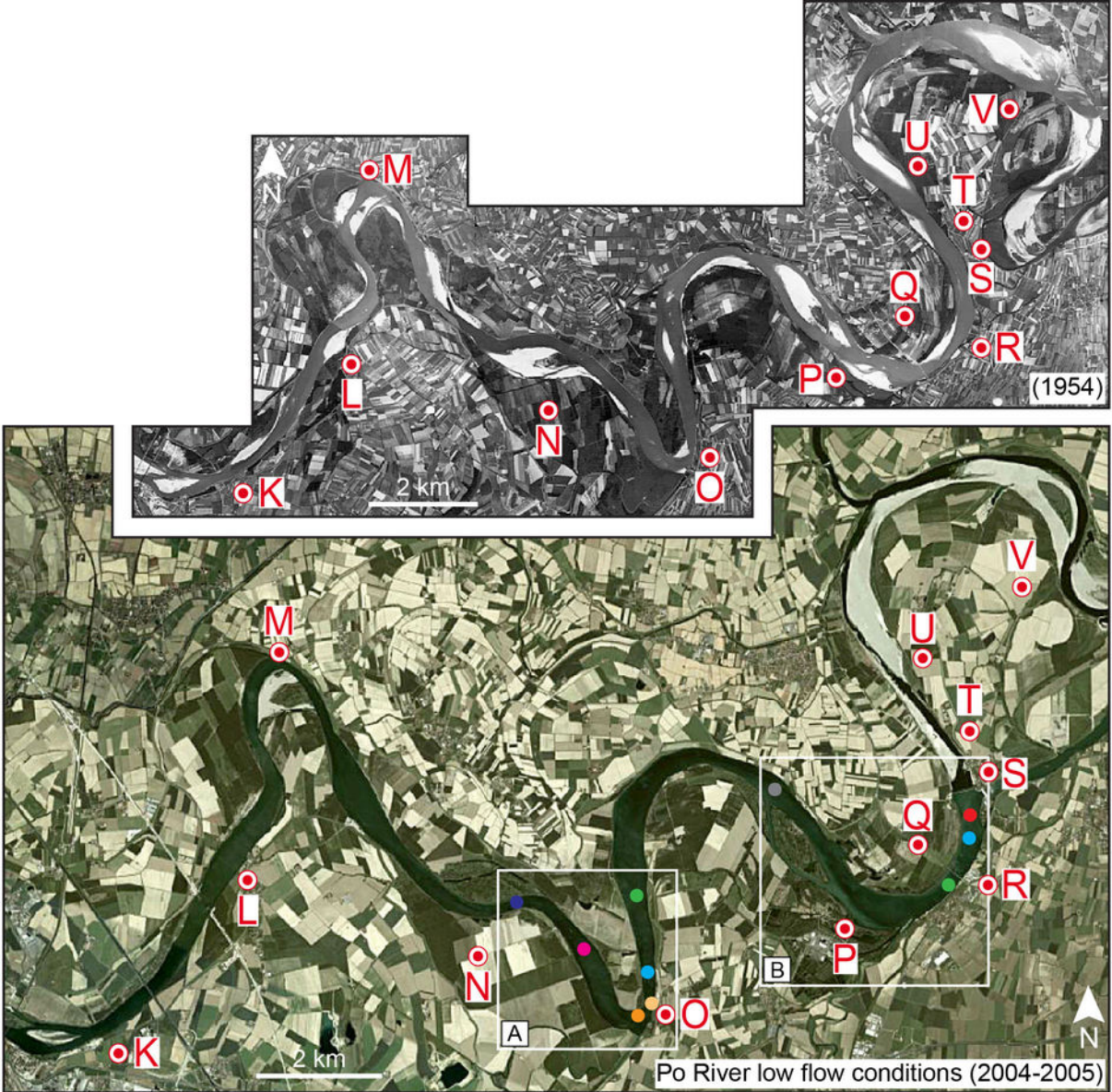


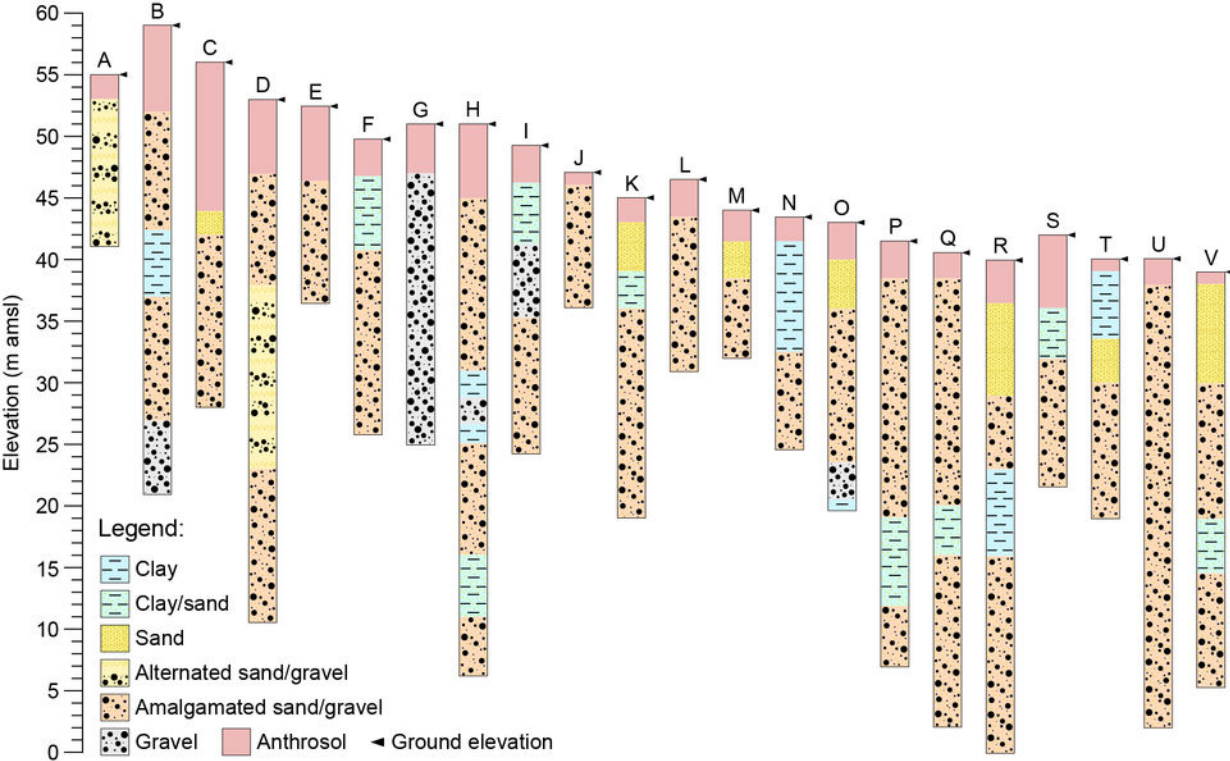


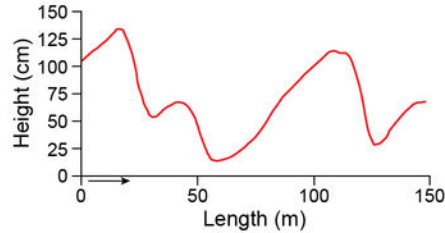
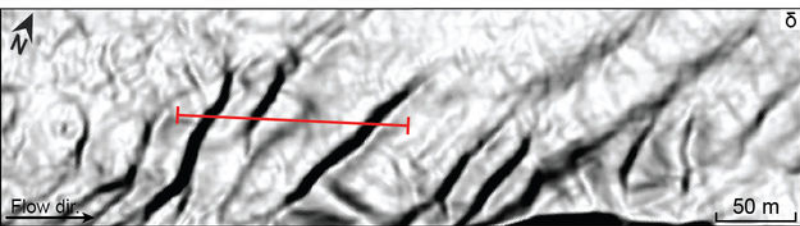
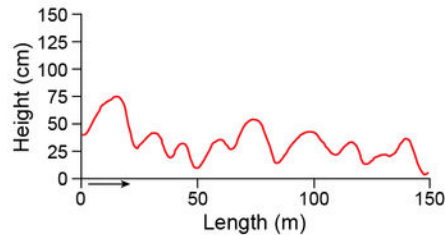
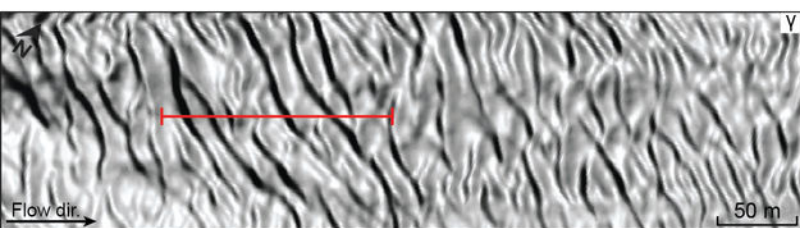
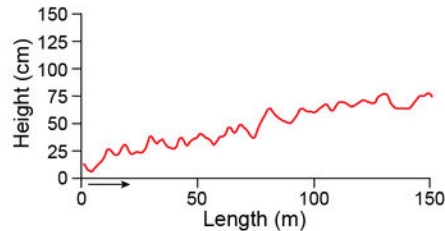
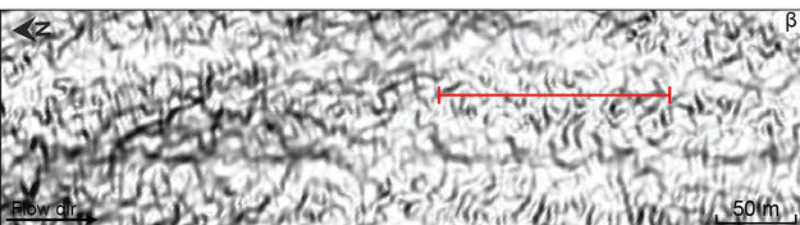
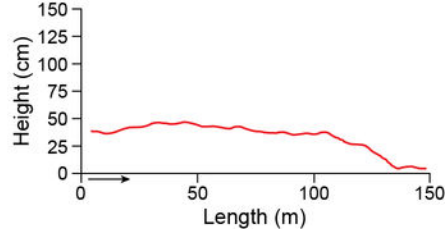
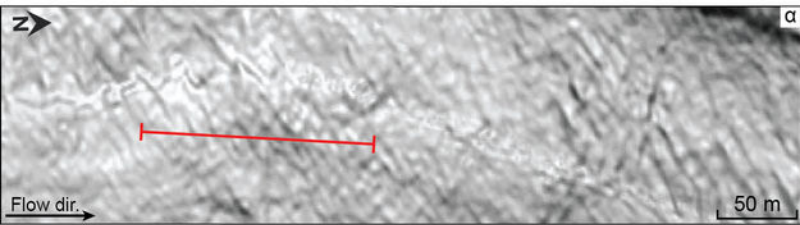


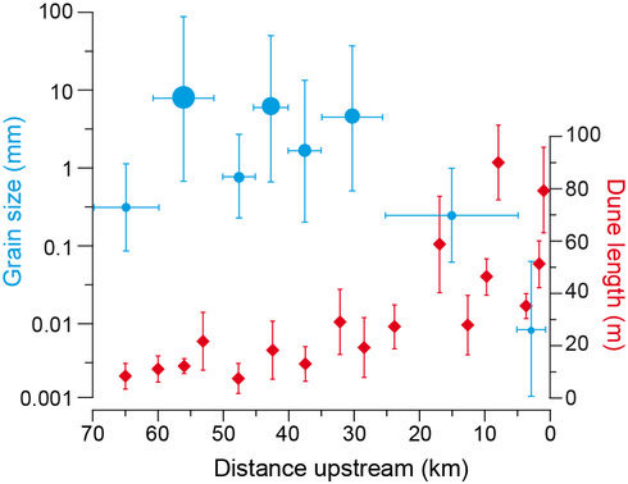


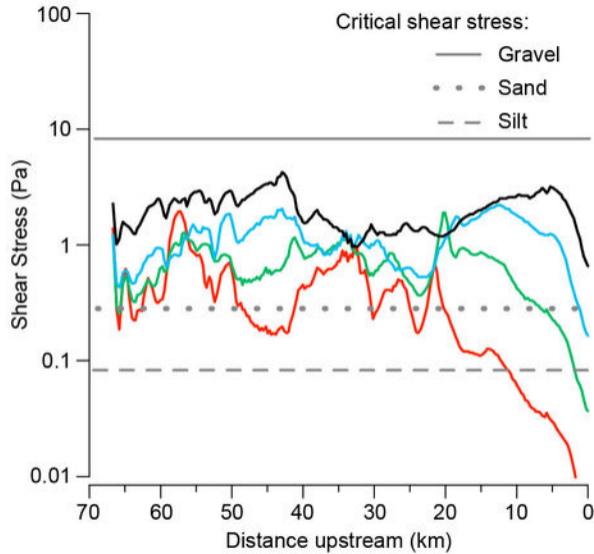
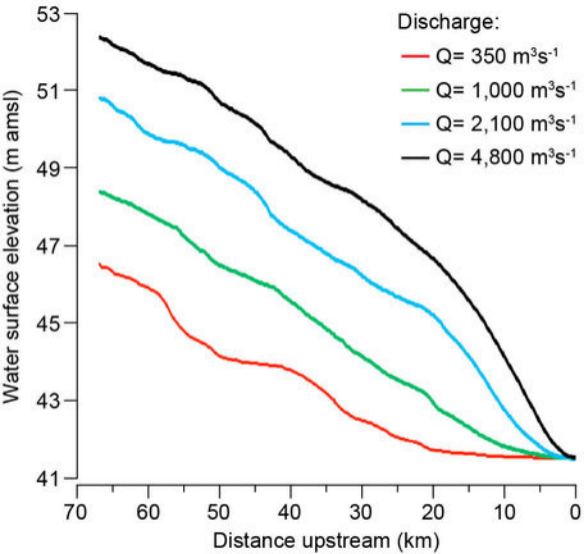












**Table 1.** Critical shear stress by particle-size classification (modified from Julien, 1995).

<i>particle classification</i>	<i>particle diameter</i> [mm]	<i>Shield parameter</i> [-]	<i>critical shear stress</i> [Pa]
coarse cobble	128 – 256	0.054 – 0.054	112 – 223
fine cobble	64 – 128	0.052 – 0.054	53.8 – 112
very coarse gravel	32 – 64	0.05 – 0.052	25.9 – 53.8
coarse gravel	16 – 32	0.047 – 0.05	12.2 – 25.9
medium gravel	8 – 16	0.044 – 0.047	5.7 – 12.2
fine gravel	4 – 8	0.042 – 0.044	2.7 – 5.7
very fine gravel	2 – 4	0.039 – 0.042	1.3 – 2.7
very coarse sand	1 – 2	0.029 – 0.039	0.47 – 1.3
coarse sand	0.5 – 1	0.033 – 0.029	0.27 – 0.47
medium sand	0.25 – 0.5	0.048 – 0.033	0.194 – 0.27
fine sand	0.125 – 0.25	0.072 – 0.048	0.145 – 0.194
very fine sand	0.0625 – 0.125	0.109 – 0.072	0.110 – 0.145
coarse silt	0.0310 – 0.0625	0.165 – 0.109	0.0826 – 0.110
medium silt	0.0156 – 0.0310	0.25 – 0.165	0.0630 – 0.0826
fine silt	0.0078 – 0.0156	0.3 – 0.25	0.0378 – 0.0630